This collection of papers devoted to the Ice Age of northern Russia provides illustrated descriptions of landforms and sediments revealing former ice sheets of the arctic shelf that inundated northern Russia. It shows that a peculiar Siberian type of inland glaciation is inferred from preserved Ice Age features. This type of glacial environment implies arrested landscape evolution in continental climates with fossil glacial ice surviving within the conservative permafrost. The contributions here delve into the problem of the size and age of the last glaciation intensely discussed in the international literature. This is of broad interest because its solution is paramount for global climatic models and the reconstruction of Circum-Arctic paleoenvironments. It is also essential for understanding natural conditions of early human migration into the Arctic. Another point of interest is the book's discussion of the profound impact of reconstructed glaciers on the tectonic structure and distribution of petroleum reserves.

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978-1-5275-4928-9
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Cover image *Fossil glacier*

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Cambridge Scholars Publishing

The Northern Pleistocene of Russia

Valery I. Astakhov

THE NORTHERN PLEISTOCENE OF RUSSIA

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This book first published 2020

Cambridge Scholars Publishing

Lady Stephenson Library, Newcastle upon Tyne, NE6 2PA, UK

British Library Cataloguing in Publication Data
A catalogue record for this book is available from the British Library

ISBN (10): 1-5275-4928-3
ISBN (13): 978-1-5275-4928-9

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INTRODUCTION

This book consists of 26 research papers by the same first author and editor. They are selected by their relevance to the large geological object, namely the Northern Pleistocene of Russia. This object is a grey-coloured polymineral siliclastic formation up to 300-400 m thick mantling the great plains and neighbouring uplands of arctic and subarctic Russia. The papers describe geological results and paleoenvironmental inferences obtained by study of several decades.

The study was originally initiated from the practical needs of geological cartography and prospecting. They were in a great measure spurred by the incessant discussion of the age and mode of former glaciation which for decades haunted the Russian Quaternary literature and since the 1980-s was also the theme of many research papers worldwide.

Fundamental knowledge of the recent geological history of continental terrains is based on studies of paleoclimatic cyclicity and its material manifestations. In this respect a special attention has always been paid to the formerly glaciated regions of temperate and arctic lands where the cyclicity is most evidently expressed in the form of paleolandscape and sedimentological zonality.

The classical studies of the Pleistocene stratigraphy and paleogeography have been performed in North America and Western Europe relating the former Pleistocene Laurentide and Scandinavian ice sheets and mountain glaciers of the Alps. The general textbooks information is largely based on many decades of meticulous studies of these regions by hundreds of researchers (Elias & Mock, 2013). These works were mostly confined to the same geologic environment profoundly influenced by the permanent factor of the Atlantic Ocean.

However, the largest formerly glaciated area of Eurasia extending over 60 degrees of longitude south of the Barents and Kara seas and north of 59-60° parallels is much less known. Without understanding the peculiar surficial geology of this area, where the Atlantic influence is considerably weakened, our knowledge of the recent history of the dry lands would be incomplete and even deficient. The recent geological history of eastern subaerial and subaquatic lands with continental climates domineered by peculiar geological processes exhibits a considerable idiosyncrasy not readily understood within the frame of the classical models. Another feature of these terrains – the poor accessibility and scarce population – is an excuse for the gaps in the Quaternary science of the northern terrains.

The natural history interest in the Northern Pleistocene has been driven by its record of climatic fluctuations of the last million years. Without this record it is hardly possible to build the consistent global models of climate change and understand the development of the arcto-alpine biota, origin of great river systems, prehistoric human migrations and many other phenomena. This is the reason of the growing attention of the international community to the surficial sediments of the Russian North which resulted in several collaborative research projects lately performed on the Northern Pleistocene and partly reflected in my publications.

These papers were published during half-century in various journals and collections. Although almost all of them (except for the first one) were in English, their accessibility for an interested student is problematic: part of them was written in pre-digital times and can only be found in fairly obscure editions. These considerations inspired me to collect the disjointed publications into one easily accessible volume.

My first forays into the Barents and Kara seas catchment areas in the 1960-s were met with the cardinal controversy in popular ideas on the origin of the thick, predominantly diamictic sedimentary cover of the West Siberian Plain and the Pechora Basin. One group, basically geological, tried to apply the classical model of European glacials and interglacials to the thick, mostly diamictic Quaternary of the northern plains, whereas the other school insisted on the profound influence on the sedimentary cover of frequent transgressions of the Arctic Ocean with minor addition of alpine and Norwegian type glaciers from the highlands of the Urals and Central Siberia.

The last, so-called hypothesis of glacio-marinism, was bred by drift ideas of XIX century which were adapted to the northern lowlands. It was based on the fine-grained composition of the very thick diamicts which in places contained marine shells and microfauna. Accordingly, the northern diamicts were often interpreted as glaciomarine sediments even by those who maintained the European paradigm of Pleistocene climatic fluctuations and accepted former ice sheets (e.g. Arkhipov, 1971; Lazukov, 1971; Zubakov, 1972). The ideas of glacio-marinism, and sometimes even antiglacialism, were eagerly supported by permafrost workers without sedimentological background and by bedrock geologists who could not see anything glacial in the pre-Quaternary record.

A glimpse of the way out appeared in the 1970-s when the comprehensive glacial sedimentology was first applied to the key sections of the Quaternary on the Pechora and great Siberian rivers. It was found that most of so-called glaciomarine diamicts were lowland facies of glacial deposits left by huge continental ice sheets (Guslitser, 1973; Kaplyanskaya & Tarnogradsky, 1975; Astakhov, 1981). Simultaneously the progress in geological mapping with wide use of remote sensing methods firmly established the glacial nature of the surficial cover of Quaternary sediments and its origin not from mountain-based but from shelf-based ice sheets (Astakhov, 1974a, 1976, 1079; Astakhov, Fainer, 1975).

There was also the old problem of the extent and age of the last glaciation which was hovering in the background of the incessant discussion on the origin of the diamict formations. This problem was getting more acute with the growing number of radiocarbon dates which after the first enthusiasm of the 1960-70-s added a considerable uncertainty in the important correlation of the northern glacial and interglacial events with the Quaternary chronology of better studied Central Russia and Western Europe.

The classical concept, which domineered in the Russian literature in 1950-60-s, accepted the Saalian glaciation as the Pleistocene maximum ascribing a modest size to the early stage of the post-Eemian glaciation (Yakovlev, 1956; Ganeshin, 1976). With the advent of radiocarbon dating the revised classical chronology of the Late Pleistocene in the 1970-90-s was replaced by the popular concept of the Late Weichselian age of the last northern glaciation. At the same time the northern paleogeography was considerably influenced by the so-called maximalist idea of the huge Pan-Arctic ice sheet which during the last ice age presumably covered all arctic shelves and adjacent subarctic plains, i.e. was commensurate with Middle Pleistocene ice sheets (Grosswald, 1983, 1998).

The maximalist model, an obvious replica of the North American paleogeography (e.g. Denton & Hughes, 1981), was fashionable in the West but in Russia was not that popular, especially among geologists dealing with ground truth. The analysis of the situation based on the knowledge of key sections, glacial topography, periglacial environments and statistics of conventional radiocarbon dates made me firmly reject the maximalist paleogeography and return to the classical idea of an early ice advance in the Late Pleistocene reinforced by the modern concept of shelf ice dispersal centres (Astakhov, 1992, 1998).

The problem of the age and volume of the last ice sheet of Eurasia is crucial for the solution of the global paleohydrological equation (Peltier, 1994). The last ice sheet of northern Eurasia is supposed to be responsible for 14 m of global sea level. As a result, a growing number of Pleistocene investigators from Western Europe and America have been over the last 25 years attracted to studies of the Russian Northern Pleistocene.

The modern research history started in 1993 with concerted field work by Norwegian and Russian scientists within the collaborative research project PECHORA (Palaeo Environments and Climate History of the Russian Arctic) supported by the Norwegian agencies and with the German-Russian project ‘Taimyr’. These efforts were followed by many other joint expeditions into the Russian Arctic with participation of Swedish, Danish and American geologists. The various international projects were coordinated by the programs QUEEN (Quaternary Environments of the Eurasian North) and APEX (Arctic Paleoclimate and its Extremes) supported by the European Science Foundation. The field works were supplemented by state-of-the-art means of remote sensing and analytical research in best laboratories of Europe.

The decisive contribution to the chronology of the latest Quaternary events was made by comprehensive dating programs which included radiocarbon, luminescence and cosmogenic exposure methods supported by paleontological and geocryological data (Svendsen et al., 2004). This author participated in all these studies from the beginning as an organizer, photointerpreter and field geologist.

I tried to organise the ingredient papers of this collection both thematically (in five Chapters) and chronologically (within each Chapter), but could not escape some repetitions of statements, and in rare cases even pictures. Replications are hardly avoidable in a book on the same subject embracing research efforts of several decades in different social and technological environments and published in disparate editions. However, the repetitions might not be too conspicuous because most readers would probably be interested in particular articles rather than in the entire collection. On the other hand, the repeated attempts to solve the same questions might be educating for those striving to understand the trend of reasoning and change of arguments over the decades.

It is my hope that the complicated story told in this book will be judged not only by its academic value but also by the persistent, if not always successful, efforts to unravel the glacial history of the huge area fascinating for students of paleoenvironments.

The papers appear in their original wording except for corrections of obvious grammar and terminological lapses, common for a not native English writer. Also Introduction and Summary are added. All references are assembled in the end of the book.

I am sincerely thankful to my coauthors named in the titles of papers 5, 8, 11-13, 15-17, 18, 19 and 25 for their assistance in the research and

permissions to use these reports of the collaborative work for the present publication. They cannot be blamed for possible linguistic or terminological errors since I have edited all the texts single-handedly.

My gratitude is also due to the publishers of the below mentioned journals and collections who originally dealt with the articles included into this book and only slightly modified for this edition.

CHAPTER I

MODE OF PLEISTOCENE GLACIATION

This chapter consists of four papers written in the 1970s (papers 1 and 2) and in 2003-2004 (papers 3 and 4). The old papers present my initial attempts to understand the fundamentals of northern ice sheets which for the first time were considered as basically originated in the low terrains and on the dry shelf and not as products of mountainous ice caps. The later papers 3 and 4 already took this new paradigm for granted and tried organize the data collected over the decades by many Russian geologists in a form accessible for external users. Paper 3 specially concerns indications of ice flow pattern and discrepancies in available proxies. Paper 4 presents the overview of geological mapping data which were used for estimates of shape and size of former ice sheets of different ages. This collection of Russian geological information was ordered by the INQUA project `Quaternary Glaciations: Extent and Chronology`– the first world-wide compendium published in 2004. The data described in this chapter, largely obtained by the Soviet Geological Survey, served as the cartographic and stratigraphic basis for several international research projects started in the 1990s which provided the bulk of results discussed in the following chapters.

1. Geological Evidence of Kara Sea Centre of Inland Glaciation

Valery I. Astakhov

Doklady Akademii Nauk SSSR 1976, Vol. 231, № 5,
(originally in Russian, translated by author)

The idea of centres of ancient glaciation in the West Siberian North was first put forward by V.A. Obruchev (1930). Later P.S. Voronov (1951) discovered Pleistocene ice advances from the low Yamal Peninsula onto the Pai-Hoi Ridge. Presently a Late Pleistocene ice sheet on the Barents Sea shelf is advocated by M.G. Grosswald. Lately structural analysis of the Middle Pleistocene Sanchugovka Formation has revealed its glacial origin. This led to a suggestion that the remains of arctic malacofauna in this formation were delivered from the shelf by a Kara ice sheet (Kaplyanskaya & Tarnogradsky, 1975). Nevertheless, the overwhelming majority of investigators of West Siberia have until now considered only influence of mountainous ice dispersal centres. The research of recent years has obtained decisive evidence of a Kara ice dome.

The crucial fact is the lately discovered traces of ice motion from lowlands onto the margin of the Central Siberian Plateau (Astakhov & Fainer, 1975). A large number of diamictic and sandy ridges is observed north of river Podkamennaya Tunguska where the convex distal slopes of these crescentic ridges face eastwards and south-eastwards upslope of the Plateau. Radial eskers and striae in the bedrock traced up to 450 m isohypse are orientated from north-west to south-east or west-east. The upslope ice motion is indicated also by mineral composition of the tills with predominant products of redeposition of Mesozoic sediments of West Siberia, mainly of quartz.

A number of large blocks of Cretaceous kaolin sands are found far to the east of their *in situ* occurrence. The flat summits of trapp mountains some 600 m a.s.l. are covered by alien tills full of Mesozoic pollen (Fainer, Mitachkina, 1974). Only east of the 92° E the Central Siberian

Plateau shows glacial features orientated downslope NE-SW with the tills predominantly consisting of trapp boulders, heavy minerals and feldspars in the matrix (Astakhov & Fainer, 1975).

Rose-diagrams of long axes of pebbles in tills with the West Siberian mineral association in the Middle Yenissei catchment area everywhere demonstrate prominent NW and WNW peaks. All these facts unanimously indicate a frontal advance of Middle Pleistocene glaciers from the north-west into the Yenissei valley and farther eastwards up to altitudes of at least 600 m.

Quite a similar pattern is observable at the same latitudes along the western margin of the lowland. The transportation of pebbles into the Urals from the lowland has been known for a long time (Sirin, 1947; Lider, 1964). Numerous sand and sand-gravel hillocks and west-east striking eskers cover the eastern slope of the Northern Urals up to 600 m (sources of rivers Lopsiya, Khuntynya, Mazapatya, Manya, Bolshaya and Malaya Sosva). Such landforms making in plan a huge arc with the convex side facing west are traced southwards up to Burmantovo settlement. In this area B.V. Ryzhov (1974) has described three formations of matrix-supported tills with numerous fragments of Mesozoic and Paleogene rocks of West Siberia at altitudes up to 500 m. Pebble long axes at the upstream river Severnaya Sosva indicated ice motion from NNE. Farther to the south in the area of Ivdel-Polunochnoye clast-supported till is practically devoid of boulders from the Central Uralian rocks (Rabinovich, 1961).

Scattered data from central West Siberia confirm the diffluent flow of Middle Pleistocene ice. The diagram of pebble orientation from the Samarovo Till near the city of Khanty-Mansiysk shows two distinct peaks: west-east (common for a marginal zone) and more pronounced north-eastern (Chernov, 1974b).

The sub-latitudinal orientation of the ice limits and of the accretion ridge of the Siberian Hills is most naturally accounted for by ice motion from the north and is hardly explainable by the assumed model of centripetal movement of mountain glaciers.

The pattern of linear parallel ridges which occur only in glaciated terrains is very significant (Fig. 1). In the Meso-Cenozoic succession they are represented by tight folds rapidly flattening downwards. Some investigators tried to explain these ridges by non-glacial factors such as permafrost, gravitation tectonism, etc. The only weak point of the glaciodynamic version of their origin was the orientation of the parallel ridges conflicting with the traditional paleogeography: the distal slopes of these arcuate ridges often face the presumed mountainous ice dispersal centres. Evidently this configuration of the glacigenic disturbances excellently agrees with the concept of a lowland ice dispersal centre in the West Siberian North (Fig. 1). These ridges should be considered as ice-pushed features. Their glacial origin is underlined by the reverse, i.e. downslope, orientation of parallel ridges in the area of piedmont glaciation of the Late Pleistocene. The pattern of the ice-pushed ridges is also conformable with the distribution of accretion hummocks such as kame plateaus (Fig. 1).

The diffluent motion of the lowland glaciers is evidenced also by the material composition of the Middle Pleistocene formations of the central part of the lowland. The north-south strip of the Taz Peninsula – Nadym-Pur interfluve is practically devoid of Uralian and trapp boulders. If the most resistant fragments of the Central Siberian trapps sometimes are noticed almost in the centre of the lowland, the amount of Uralian pebbles is negligibly small already in the southern Yamal and around Khanty-Mansiysk. Even in the Muzhinsky Urals the pebbles are mostly composed of Cretaceous clays and Paleogene opokas but not crystalline rocks of the Urals (Shumilova, 1974).

The base of the Pleistocene of the Taz Peninsula does not contain gravels at all. All investigators of northern West Siberia point out that the scarce fragments of crystalline rocks are mostly flint and quartz, obviously redeposited many times. The divergent transportation of terrigenous materials from the centre of the lowland is more evident from the matrix composition: the typical West-Siberia assemblage with dominant quartz and abundance of epidote and zoisite in the heavy crop is traceable at the eastern slope of the Urals as well as on the table summits of trapp monadnocks of Central Siberia (Ryzhov, 1974; Fainer, Mitachkina, 1974).

The thickness distribution agrees with the idea of the upslope ice dispersal: the relatively thin Quaternary of the central lowland thickens to the margins of the Plain together with the elevation growth. In parallel the Middle Pleistocene diamictons enriched with marine faunas are concentrated in the lower reaches of rivers Yenissei and Ob, whereas the Salemal (Sanchugovka) Formation of the central interfluve Nadym-Pur is practically devoid of marine shells.

Judging by the sub-concentric pattern of marginal features the Middle Pleistocene ice dispersal centre was located on the Yamal, Taz and Gydan peninsulas. However, the literary data and aerial images indicate a very young glacial topography there. Fresh kames, eskers, marginal canals, sandurs and initial valleys are especially characteristic for southern Yamal and Gydan. The lake-and-hummock relief around the lake Yarroto is built of sand with fragments of local rocks and practically devoid of Uralian boulders (M.N. Boitsov and S.G. Maksimenko, mapping report of 1953).

It is underlain by brecciated varved clay (probably of the Middle Pleistocene). From here fresh radial features (striae and eskers) were traced by aerial photos to the central Pai-Hoi Ridge, where transport of boulders and organic remains from north-east to south-west was documented (Voronov, 1951). In the south of the Taz Peninsula the margin of the last ice sheet is possibly outlined by festoons of parallel ridges, shallow upthrusts in the Paleogene rocks (data by Yu. F. Andreyev) and local gravitational maxima, presumably connected with minor flow-folding along the ice margin (Fig. 1).

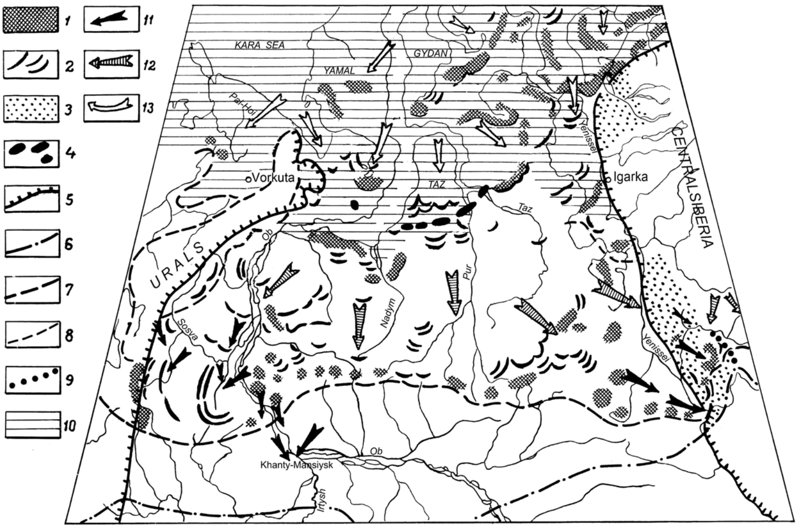

Fig. 1. Pleistocene glacial features in northern West Siberia.
Symbols: 1 – accretion hummocky terrains; 2 – parallel ridges of Meso-Cenozoic rocks (push moraines); 3 – area of Mesozoic erratics in Paleozoic Central Siberian Plateau; 4 – local gravitational maxima (after M.N. Boitsov and S.G. Maksimenko, 1957); 5 – 9 – limits of 5 – Mesozoic formations of West Siberia; 6 – maximum glaciation; 7 – Taz glaciation; 8 – Late Pleistocene piedmont ice sheet; 9 – Mesozoic and trapp mineralogical provinces in Middle Pleistocene tills of Central Siberia (after Yu. Fainer); 10 – provisional area of Late Pleistocene ice on the Plain. Ice flow directions of: 11– Middle Pleistocene glaciers (by orientation and composition of pebbles); 12 – Taz glaciers (suggested by marginal landforms), 13 – Late Pleistocene Kara Ice Sheet.

The paleoglaciologically estimated maximum ice thickness of the Late Pleistocene Barents Ice Sheet is up to 3.2-3.5 km (Grosswald & Chernova, 1972). The Middle Pleistocene ice of northern West Siberia must have been thicker. Direct measurements indicate its great thickness even in the marginal zone. Judging by the altitudes of foreign tills in the Urals and at the Yenissei Siberia marginal ice at 63° N was at least 0.8 km during the maximum glaciation and ~ 0.5 km thick in the time of the subsequent Taz glaciation. The great thickness of ice masses in the north of the Plain is indicated by boulders of igneous and metamorphic rocks of the Taimyr Peninsula scattered over the Putorana Plateau at altitudes over 1000 m (Urvantsev, 1957).

The occurrence of trapp and Uralian boulders in the West Siberian Plain is explained by the initial growth of ice sheets in the mountainous borderlands. Subsequently the ice dispersal centres shifted onto the shelf. As a result, glaciers moving upslope mixed originally deposited Uralian and Central Siberian material with products of assimilation of the Meso-Cenozoic rocks of West Siberia. This is the explanation of the meagre content of coarse fragments in the Middle Pleistocene sediments and its high concentration in the non-redeposited Late Pleistocene moraines of the Transuralia and Lower Yenissei area.

Apart from the landforms orientation and regularities of transport of terrigenous material the concept of the Kara ice dispersal centre is a logical explanation of the occurrence of arctic malacofauna in the diamicts of the Sanchugovka Formation and their evidently glacigenic structures (Kaplyanskaya & Tarnogradsky, 1974), transportation of bauxite fragments eastwards to the western slope of the Central Siberian Plateau (Fainer & Mitachkina, 1974), the southward displacements of crests of the local anticlines in the Meso-Cenozoic cover (Rudkevich, 1961), etc. More precise paleogeographic reconstructions in the coming years would demand detailed studies of the variations of the Quaternary sediments composition over the area and in the succession and careful analysis of accretion landforms in northern West Siberia.

2. New Data on the Latest Activity of Kara-shelf Glaciers in West Siberia

Valery I. Astakhov

International Geological Correlation Programme, Project 73/1/24 `Quaternary Glaciations in the Northern Hemisphere`, Rep. N 5, Šibrava V., Shotton F., Eds, Prague: Czechoslovak Geological Survey, 1979, p. 22–31.

The hypothesis of centres of former glaciation on low coastlands of West Siberia was advanced by I.A. Molchanov and V.A. Obruchev around 1920-s but until recently it did not greatly influence research of the northern Pleistocene. Paleoglaciological speculations about a Middle Pleistocene ice sheet on the shelf (Voronov, 1968) were not substantiated by geologists because of lack of field data. The authors of the modern view on the former Siberian glaciation did not doubt the mountain origin of the Würm ice sheets of West Siberia and discussed only a hypothetical confluence of Uralian and Taymyr-Putorana glaciers on the Kara sea shelf (Arkhipov et al., 1976). Such conclusions were based on geological investigations between latitudes 50° and 60° which were rather detailed in the valleys and too scarce in the wide interfluvial terrains.

The situation changed in the seventies as the All-Union Corporation `Aerogeologia` began areal investigations and mapping of glacial topography of North Siberia and the Urals using small-scale aerial photos and also satellite images. The starting point for revival of the shelf-centres concept was the discovery of extra-local moraines in the area of the right bank of the Middle Yenissei. These data proved the frontal advance of Middle Pleistocene ice masses south-eastward onto the Central Siberian Plateau (Astakhov, Fainer, 1975).

New data on the glacial topography of Peri-Yenissei Siberia and the Northern Urals combined with a revised interpretation of published literature reestablished the concept of the Kara-shelf as a centre of former glaciation (paper 1). The main proofs are as follow:

1) `Anti-orographic` configuration of end moraines of the Middle Yenissei and Northern Urals: the convexities of the morainic ridges face mountain ranges, not the lowland.

2) Numerous West Siberian erratics found in the Urals and in the Central Siberian uplands.

3) Glacial landforms in the West Siberian Plain including accretion hummocks and parallel-scalloped ridges in the Arctic area of the latest glaciation which are parallel to the Kara Sea coast.

4) Orientation of the long axes of pebbles, commonly normal to the terminal moraines.

5) Lowland, i.e. West Siberian, composition of the tills which occurs not only in the central part but also at the margins of the Plain.

6) Definite areal coincidence of West Siberian erratics and `anti-orographic` morainic amphitheatres.

7) Drift thickness which increases from the central part of the lowland towards the mountain margins.

8) Absence of topographic evidence of downslope ice movement except for the youngest moraines in the Norilsk region and on the Trans-Uralian piedmont of the Polar Urals.

9) Absence of crystalline pebbles in the breccia-like till of the Kara Sea floor covered by Holocene marine sediments (Astakhov, 1977).

Further photogeological mapping in West Siberia has provided the opportunity of a practical examination of the Kara glacial centre concept, particularly relating to the terminal formations in the Arctic area of the latest glaciation. New satellite and high-altitude aerial images have revealed more push moraines concentrically orientated around the Kara Sea coast. The interpretation of high-altitude aerial photos of the Gydan Peninsula found amphitheatres of parallel ridges opening northward and bordered by outwash plains from the south and by lacustrine depressions from the north.

Such a ridge of a mixed push and accretion origin, 30-60 m high, was recognized in the Taz-Messo interfluve. Landsat imagery showed also horseshoe-shaped parallel ridges about 80 km long on the Yenissei west-bank area stretching out from the Muram Lake across the upper Turukhan River up to the Osetrovaya-Pokoinitskaya interfluve. The axes of these amphitheatres indicate late Pleistocene ice movement from the Ob Estuary south-eastward (the first case) and from the Yenissei Estuary southward (the second case).

The most distinct glacial end formations were investigated in 1977-1978 in southern Yamal and Polar Trans-Uralian piedmont. Two chains of parallel ridges joining at an acute angle were recognized from high-altitude photos of the area. The first one stretches out northward in the Yuribey left bank area. It consists of crescentic festoons of interfluve hills 60-100 m above sea level which form the western borders of the biggest lake basins of the Yamal Peninsula – Yarroto 1st, Yarroto 2nd, Mengakoto and Tetanto (Fig. 1). These lakes, with water levels about 25-35 m a.s.l. are placed in a sandy plain of 40-50 m a. s. l. This plain was considered to be a Kazantsevo (Eemian) marine terrace (German et al., 1963; Trofimov et al., 1975). The `Kazantsevo terrace` comprises light-grey, inclined quartz sands rhythmically interbedded with silt, lenses of redeposited peat, fragments of driftwood and thin beds of fine gravel. According to the author's observations (together with K.E. Simonov) these sands form also festoons of parallel ridges rising10-50 m above the `Kazantsevo terrace` level.

The most detailed investigations were made at the Handy-Hoi Ridge which is 20 km long and 3 km wide. The ridge bounds the Tetanto Lake basin from the west (H and 6 in Fig. 1). The internal structure of the ridge was studied at numerous natural exposures along the consequent valleys. Angles of dip of the `Kazantsevo` sands increase from 0-5° on a flat foot to vertical ones at the crest of the ridge in pace with the increasing elevation (Fig. 2). Along the axis of the ridge the sands are always crumpled into steep, sometimes isoclinal folds. Dark grey fissile silt layers 3-8 m thick containing very scarce pebbles of crystalline rocks and balls of underlying sand and varved clay appear there together with lenses of coarse gravel.

Extreme contortion is registered along the westward shifted crest of the asymmetric ridge where the exogenous-tectonic structure becomes actually of alpine type: uneven ferruginated thrust planes dipping eastward divide blocks of tilted sand beds at almost every outcrop. Layers of incompetent till-like silt sometimes form shear dykes and tongues which are concordant with the general direction of lateral pressure (Fig. 3). Where distinct blocks are absent, in these sections one can see disharmonic folding of at least two structural stages divided by sharp angular unconformity. The upper sands are crumpled with dips from 20-40° of the lower sands up to 50-90°. The beds at the foot of the western steep slope of the ridge abruptly flatten to the angles of 3-10° and become conformable to the wavy surface of the `Kazantsevo terrace`. The flat crest of the ridge is topped with deflation armors consisting of pebbles and boulders sometimes 1-1.5 m in diameter. Along the crestline small gravel cones 1-5 m high are scattered. Numerous (more than 150) measurements of bedding showed strike coincidence of the steep-dipping beds with arcuate stripes on the aerial pictures bounding Tetanto Lake from the west.

Such ridges were formerly regarded by investigators either as end moraines of the Uralian glaciers (Strelkov, 1959) or as coastal bars of a retreating sea (Trofimov et al., 1975). The inadequacy of these explanations can be seen from Fig. 2 which shows an asymmetric diagram of the dip of the internal strata conformable to an asymmetric profile of the ridge itself. This means that the arcuate ridge (H in Fig. 1) was formed simultaneously with contortion of the beds and by the same process. The dominant eastward dip of the bedding and the crescentic form of the parallel ridge-and-groove features clearly indicate severe lateral pressure from Tetanto Lake, i.e. from the east.

The structure and topography of the ridge are identical to ice-pushed ridges well known on the plains of North America and Northern Europe (Flint, 1971) and described in Bielorussia as `glaciotectonic garlands` (Levkov, 1978). The largest lakes of the Yamal Peninsula, located in the rear of the glaciotectonic garlands, are most likely remnants of dead tongues of ice that advanced from the Ob Estuary and Gydan Peninsula. These sandy ice-pushed ridges formed by frontal contortion of the proglacial outwash are the reason why the whole terminal formation of the lake district of Yuribey River has been termed the `Sandy Belt`.

West of the Sandy Belt the glacial ridges acquire another orientation. At the upper Yorkuta River festoons of ice-pushed ridges with their front facing eastward stretch out south-westward along the right bank of the Tanlova River. This system is traced farther along the left bank of river Heyaha towards the west-east orientated length of river Shchuchya where it transforms into the thick till ridges of the Sopkay amphitheatre (Fig. 1). The latter was always considered to be a typical end·moraine of the Uralian glaciation (Trofimov et al., 1975). ·

Yet photointerpretation data do not support this idea. Zones of hummock-and-lake landscape do not fringe the Polar Urals as is shown on all the maps of Quaternary deposits but they are perpendicular to the mountain front (Fig. 1). To clear up the origin of the hummocky ridges the author in 1975 traced them from the eastern front of the mountains to the middle course of river Heyaha, i.e. about 100 km from west to east. It was thus established that the uplands about 100-300 m a. s. l., barring the former lake of the Shchuchya River basin from the north, are not till ridges but ice-moulded Paleozoic bedrock forming roches moutonnées. The flat late-glacial lacustrine depression of the middle course of river Shchuchya, about 50 m a. s. l., occupies the lowest central part of a Paleozoic ring structure filled with Jurassic deposits.

Typical terminal moraines are situated along the south and east perimeter of the ring where they cap narrow ledges of Paleozoic basement and form hummocky till ridges 30-60 m high. As the altitude of the bedrock surface declines eastward, so do the levels of the accretion ridges from 200-300 m at the mountain foot to 100 m farthest east at the Sopkay Ridge. The whole system of end moraines on the right bank of the Shchuchya River strikes in a W-E direction except for the longitudinal part of the Sopkay Ridge·about 15 km long.

The hummock-and-lake topography of the Sopkay amphitheatre is analogous to the landscape of the Valday Ridge but the composition makes an important difference. There is a striking lack of crystalline boulders in the drift of the Sopkay Ridge despite its being much nearer to an assumed mountain source of ice than the Valday Ridge. Surficial boulders are scattered over the ridges only in the piedmont area where the moraines rest immediately on the bedrock basement. A small amount of gravel is concentrated in little kames that occupy inter-moraine hollows.

The terminal moraines as seen in natural outcrops consist of dark-grey silt and clay with lenses of laminated lacustrine sediments. Immediately east of the last outcrops of Paleozoic rocks at about 68° E coarse fragments in till are so scarce that measurements of pebble orientation usually run into insurmountable obstacles. Crystalline fragments are practically lacking at the base of the Sopkay till. The basal till 1-3 m thick is a viscous breccia consisting of small pieces of varved clay and Mesozoic-Paleogene siltstone. Upwards in the succession the clayey fragments crush into crumbs and then grade into surrounding till which incorporates very scarce well rounded pebbles of Uralian rocks and balls of underlying quartz sand.

The basal clayey till is observed almost everywhere at the exposures along the W-E stretch of the Shchuchya River. The terminal ridges prolongating the Sopkay Ridge north-eastwards along the right bank of the Heyaha River also consist of clayey till, with practically no pebbles. The whole chain of high accretion ridges stretching out from the Polar Urals to the upper river Yorkuta (Fig. 1) may be called the `Clayey Belt`.

Moraines of the Clayey Belt extend to the Baydarata Estuary basin from the south and probably join the glacial disturbances on the right bank of river Yuribey. Pebble orientation diagrams for 5 sites in the Shchuchya river valley show a regular strike of long axes normal to the strike of the morainic ridges. A north-west strike of the long axes of pebbles extending to Laborova settlement and still farther along the eastern slope of the Polar Urals, dominate at the Sopkay Ridge which faces south-eastwards. Orientation diagrams with northerly peaks varying to north-east prevail at the transverse stretch of river Shchuchya. This indicates ice movement from the south-east coast of the Baydarata Estuary.

Orientation of abundant stoss-and-lee features of Paleozoic limestone to the north from the transverse stretch of river Shchuchya agrees with the orientation diagrams (Fig. 1). The stoss slopes dip northwards between 350 and 20°. The same orientation is common for polished rocks and longitudinal grooves that can be seen in aerial photographs of the eastern slope of the Urals; for instance, in the Big Hadata-Little Hadata interfluve, and around the Laborova settlement. Farther north the traces of moving ice extend north-westwards round the Polar Urals to Mt. Konstantinov Kamen (Fig. 1).

Signs of longitudinal ice movement such as eroded summits, grooves, and blocks of northern dunite are observed along the eastern slope of the Urals up to 500 m above sea level. Typical alpine topography made by valley glaciers can only be seen at a higher altitude. Mountain gravel moraines are sometimes located at the foot of the Polar Urals, but they are obviously connected with the late glacial transverse troughs dissecting the longitudinally eroded surfaces. Large morainic amphitheatres of local valley glaciers joining the extra-local moraines of the Clayey Belt are recognised on aerial photos only in the Longot-Yegan valley much farther south (Fig. 1).

The characteristic properties of glacial topography studied in the field can be easily extrapolated to the western slope of the Urals where high altitude aerial photographs also show signatures of exclusively longitudinal movement of the last ice sheet. Such features are represented by the sub-longitudinal lobe basin of the Kara River. The basin is bounded in the south by a zone of fresh-looking hummock-and-lake landscape stretching against Mt. Khoidype (Fig. 1).

Thus, modern data on the glacial topography of the Arctic Trans-Uralian region definitely testify to southward movement of the latest ice masses from the Baydarata Estuary round the adjacent high massifs of the Polar Urals. Marks of south-westward ice-movement discovered in 1951 across the NW orientated Pai-Hoi Ridge should obviously be referred to the Baydarata Ice Stream, not to a hypothetical upland ice-cap at the Yamal Peninsula as proposed by P.S. Voronov. Moraines of the Clayey Belt are most likely formed by ice-masses of the West Kara Sheet moving south-eastward and by excavation of clayey Cenozoic deposits of the Baydarata glacial depression.

As can be deduced from the altitudes of ice eroded surfaces on the slopes of the Urals, the thickness of the ice of the Sopkay Stage, even taking into account a postglacial uplift of the Urals of about 100 m, may amount to 300 m in the Kara and Shchuchya river valleys and a minimum of 500 m in the Baydarata Estuary. The ice-eroded surfaces at 500 m

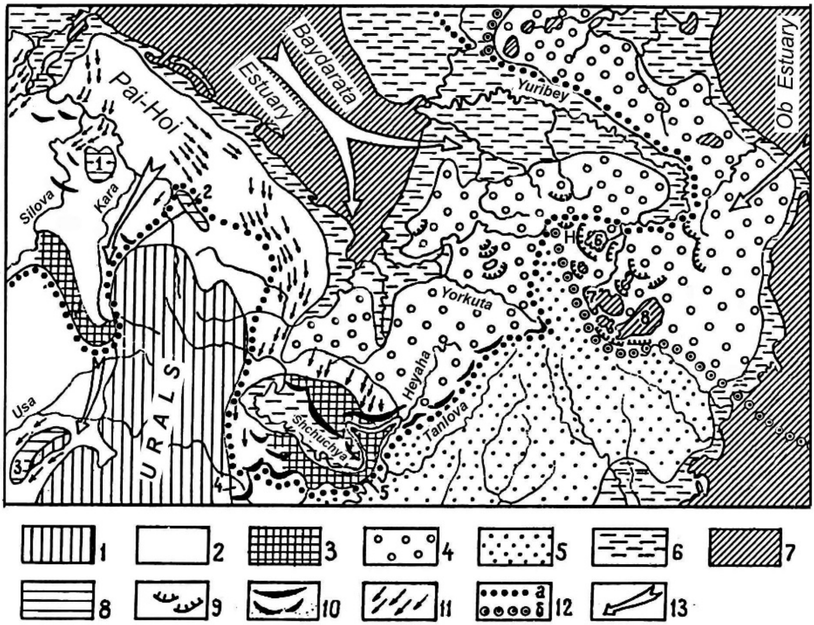

Fig. 1. Glacial features of the Polar Urals and south of the Yamal Peninsula. Symbols: 1 – alpine mountain massifs; 2 – piedmont and low mountains graded by continental ice; 3 – hummock-and-lake landscape of terminal moraines; 4 – accretion plains with Upper Pleistocene till and outwash; 5 – outwash plains; 6 – flat plains of late glacial lacustrine basins; 7 – present reservoirs; 8 – large glacial monadnocks; 9 – push moraines= glaciotectonic garlands; 10 – end moraines=. push-and-accretion till ridges; 11 – orientation of stoss-and-lee features and grooves; 12 – assumed boundaries of: a – West Kara and b – Central Kara ice streams; 13 – ice flow. The figures indicate: 1 – Mt. Yeduney and 2 – Mt. Konstantinov Kamen in the Pai-Hoi Range; 3 – Mt. Yenganape; 4 – glacial amphitheatre of river Longot-Yegan; 5 – Sopkay Ridge; 6 – Tetanto Lake east of Handy-Hoi Ridge (H); 7 – Yarroto 2nd Lake; 8 – Yarroto 1st Lake.

above sea level, and up to 600-700 m on the northern end of the Urals, may refer to a preceding larger glacial stage.

The terminal formation of the Sandy Belt should be considered as a result of ice movement from the Ob Estuary and Gydan Peninsula. It may be traced south-eastward via the push garlands of river Hadutte on the Taz Peninsula that were discovered during geological mapping, and by the above mentioned ridges in the Taz-Messo interfluve. All of them mark the southern margin of the Central Kara Ice Sheet. The distinctive push style of the Sandy Belt moraines and the relatively thin basal till are probably connected with a short duration glacial advance and maybe with a second advance of a thinner ice sheet of the latest glaciation on the Taz and Gydan peninsulas correlated with the West Kara Ice Sheet.

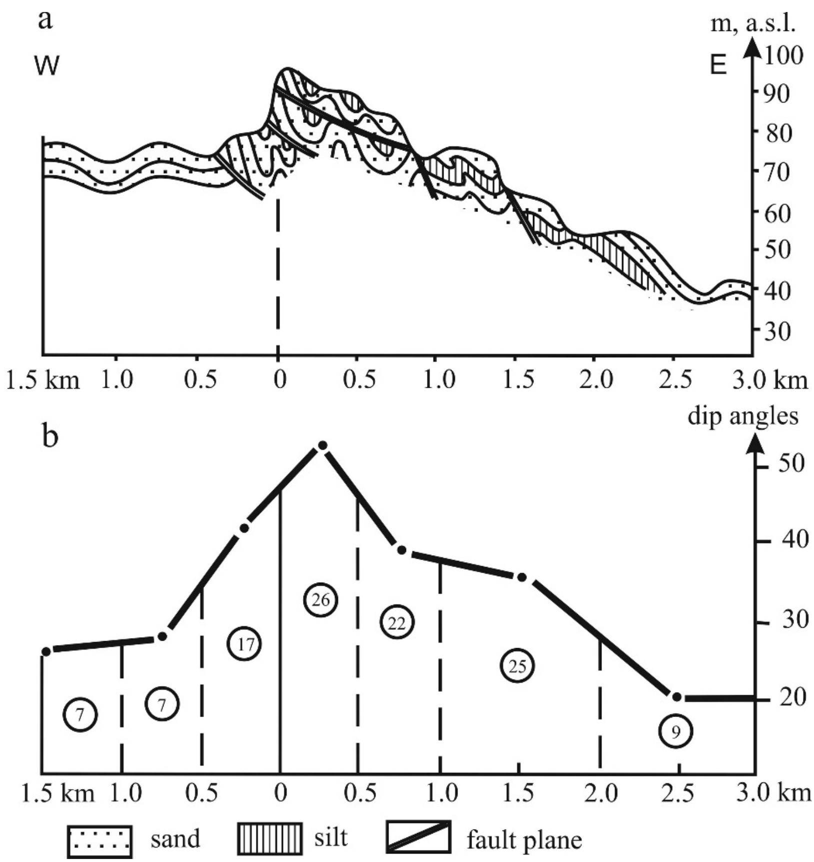

Fig. 2. Glaciotectonic structure of Handy-Hoi Ridge west of Tetanto Lake. Symbols: a – simplified cross-section; b – diagram of average angles of dip. The figures in the circles indicate number of measurements for calculation of average dip within the intervals of 0-0.5, 0.5-1.0, 1.0-2.0, and 2.0-3.0 km from the crest of the ridge.

Little is known on the absolute chronology of the described glacial stage. Moraines of a preceding advance in the Salehard region, i.e. about 50 km south, lie on deposits with radiocarbon dates of 25-28 000 years BP (Arkhipov et al., 1977). These investigators are inclined to consider the Sopkay Stage to be about 13-15 000 years BP. A date of 15 000 years was recently obtained from the second terrace of the upper river Yuribey which is enclosed in the described morainic belt (A.I. Spirkin, personal communication). This date would not permit the Sopkay stade to be younger than 17-18,000 years BP but the question must remain open until additional dates from the Sopkay deposits are received.

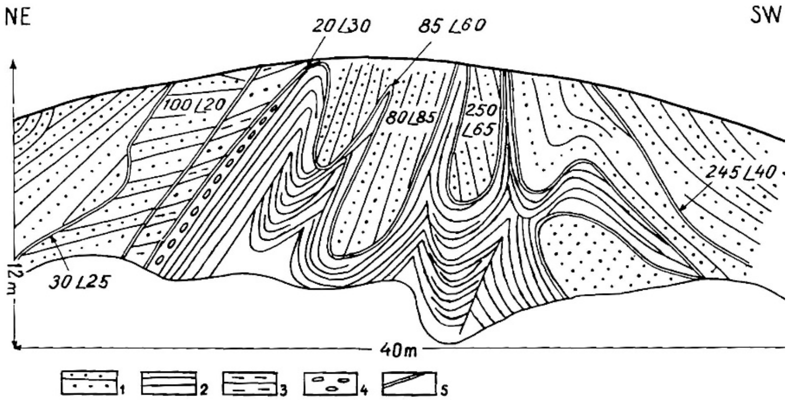

Fig. 3. Deformed Quaternary deposits near the crest of Handy-Hoi Ridge in exposure at 7 km WSW from the mouth of Tetan-Tanyo creek.
Symbols: 1 – sand; 2 – silty till; 3 – fine silty sand; 4 – gravel; 5 – thrust line.

In any case the modern photogeological research demonstrates quite clearly that the northern lowlands of West Siberia in the Late Würm-Wisconsin were attacked by continental glaciers approaching from the north.

3. MIDDLE PLEISTOCENE GLACIATIONS OF THE RUSSIAN NORTH

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QUATERNARY SCIENCE REVIEWS, 2004, VOL. 23 (11-13), P. 1285–1311.

Introduction

The volume and extent of Middle Pleistocene ice sheets exceeded those of the Late Pleistocene by far, especially on the eastern flank of glaciated Eurasia (e.g. Ganeshin, 1976). Accordingly, their impact on the geological structure and environments of the arctic and subarctic regions was more profound. However, geological research within the QUEEN framework has largely been focused on the Late Pleistocene history of the Russian North. Only a few sections of Middle Pleistocene drift have been studied by QUEEN members on the Russian mainland. Hence the bulk of data discussed below is derived from Russian literature.

Over the last half-century various attempts have been made to synthesize data on Middle Pleistocene glaciations collected by hundreds of Russian researchers. The only work in which all Quaternary of the Russian Arctic and Subarctic is discussed – the monumental volume by Sachs (1953). Another outstanding contribution is a stratigraphic monograph with a Quaternary map of European Russia by Yakovlev (1956). In the 1960-70s the huge influx of data, especially from geological surveys, led to graphical generalisations in the form of synthetic Quaternary maps for the entire Soviet Union (Ganeshin, 1976) and separately for each of the superregions of northern Eurasia such as the Russian Plain, the Urals, West and Central Siberia. These maps are principal sources of hard data about the size of former ice sheets obtained by generations of mapping geologists. Lately they were used to compile the digital map of Pleistocene ice limits as part of the INQUA project (Astakhov, 2004b).

After Sachs and Yakovlev numerous articles and regional monographs were dealing with the Middle Pleistocene glacial deposits of the North in terms of stratigraphy (Arkhipov & Matveyeva, 1964; Lazukov, 1970;

Zubakov, 1972; Yakhimovich et al., 1973; Kaplyanskaya & Tarnogradsky, 1974; Arkhipov et al., 1986, 1994; Velichko & Shik, 2001), lithology (Kuznetsova, 1971; Zemtsov, 1973a, 1973b; Sukhorukova et al., 1987; Andreicheva, 1992; Andreicheva et al., 1997), paleogeography and geomorphology (Isayeva, 1963; Troitsky, 1975; Arkhipov et al., 1976). Structural geological and sedimentological works are noticeably scarcer (e.g. Zakharov, 1968; Kaplyanskaya & Tarnogradsky, 1975; Astakhov et al., 1996). Rare attempts to reconstruct dimensions and flow patterns of the ice sheets were mostly based on till lithologies (Zubakov, 1972; Sukhorukova et al., 1987) and occasionally on glaciological considerations (Voronov, 1964).

The volume of existing data makes a comprehensive overview impossible for any journal paper. Hundreds of papers discussing various aspects of the pre-Weichselian glacial history from different standpoints cannot be reviewed here. My task is limited to briefly describing the most reliable geological results in pre-Weichselian glacial geology of northern Russia beyond the realm of Fennoscandian glaciations (Fig. 1) in order to present data for comparison and to highlight weak points that might be of interest for further research. A selection of key evidence in a diverse area ca 4.5 million km^2, although it might seem arbitrary, is actually scale-dependent. It is dictated by the author`s mapping experience, according to which superregional conclusions are derived mostly from examination of large geological features, whereas small details of geological structure often yield only results of local significance. The questions of immediate concern will be i) size and flow pattern of former ice sheets, ii) possible correlations of ice advances within the QUEEN area.

General geological setting

Northern Russia as an area of inland glaciation is very different from the classical glaciated regions of northwestern Europe and North America. It is largely soft-rock flatland extending offshore as sedimentary basins of the Barents and Kara seas (Fig. 1). The weak substrate is responsible for the great thickness (up to 300-400 m) and predominantly fine-grained composition of the Quaternary cover. Only in the east looms the Putorana Plateau, up to 1600 a.s.l., a major source of hard-rock clasts, built of horizontally lying dolerites and lavas. Other, smaller sources of erratics are narrow Paleozoic ranges of the Ural and Byrranga Mountains and low ridges of Timan, Novaya Zemlya and Pai-Hoi (Fig. 1). The rest of the area is underlain by Mezosoic and Cenozoic sand, clay, silt, opoka, diatomite

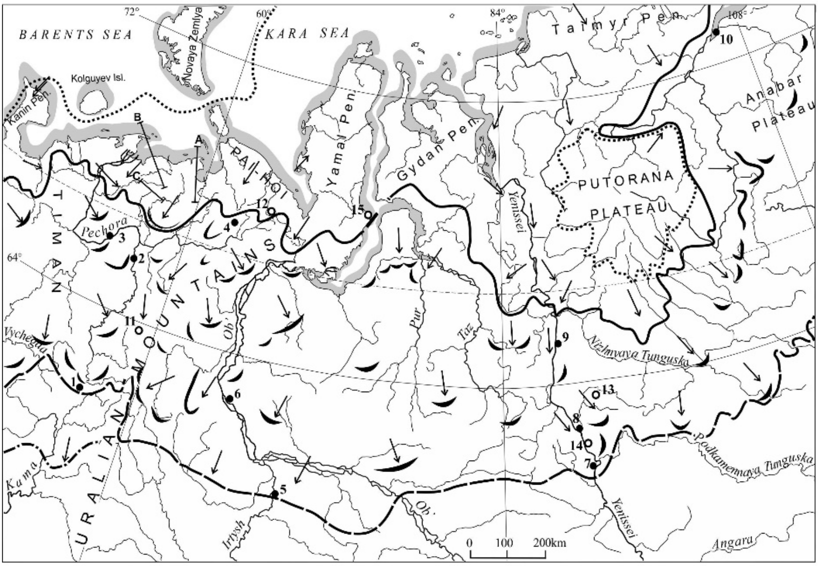

Fig. 1. Location map and glacial features of northern Russia.

Ice limits**:** dotted line – Late Weichselian, solid line – Early Weichselian; broken line is glacial drift limit in Siberia and limit of Moscow glaciation west of the Urals; dash-and-dot line is drift limit of European Russia (Don glaciation).

Black arcs are largest ice-pushed features. Lines A and B are geological profiles in Fig. 2, C in Fig. 7. Black circles are key sections of Middle Pleistocene interglacial formations sandwiched between tills: 1 – Lake Chusovskoye (Stepanov, 1974); 2 – Rodionovo (Loseva & Duryagina, 1973); 3 – Kipiyevo (Guslitser & Isaychev, 1983); 4 – Seyda (Russian-Norwegian project PECHORA, 1998); 5 – Semeika (Kaplyanskaya & Tarnogradsky, 1974); 6 – Belogorye Upland (Arkhipov et al., 1978); 7 – Khakhalevsky Yar (Levina, 1964; Zubakov, 1972); 8 – Bakhtinsky Yar (Arkhipov & Matveyeva, 1964; Zubakov, 1972); 9 – Pupkovo (Zubakov, 1972); 10 – Novorybnoye (Kind & Leonov, 1982).

Open circles indicate lowland tills in the mountains: 11 – clayey tills with clasts of piedmont Carboniferous limestones at 500-600 a.s.l. (Astakhov, 1974a; see also Fig. 4, A); 12 – Weichselian end moraine of northern origin at 560 m (paper 18, Ch. IV); 13 – three tills with lowland erratics on flat summit 600 m a.s.l. (Fainer et al., 1976).

Deepest buried valleys filled with tills: 14 – borehole at Lebed, 342 m b.s.l. (Arkhipov & Matveyeva, 1964; Zubakov, 1972); 15 – borehole at Kosa Kamennaya, 367 b.s.l. (Arkhipov et al., 1994).

Arrows are ice flow directions inferred from ice-pushed ridges and clast indicators.

and therefore is apt to produce few visually recognisable clasts. Consequently, the traditional method of reconstructing former glaciers by mapping boulder trains would inevitably point to the Central Siberian uplands and the smaller salients of folded Paleozoic rocks as the only sources of moving ice. The seaward sloping sedimentary basins of the Pechora, West Siberian and North Siberian lowlands have generally been viewed as an arena of marine incursions from the Arctic Ocean.

The stratigraphic methods employed and conclusions obtained are different in two natural zones: a) the Arctic with marine and glacial formations alternating, and b) the Subarctic, where only terrestrial stratified sediments occur sandwiched between thick diamictons. These zones, being crossed by the Uralian Range, give 4 stratoregions to be discussed separately: two in European Russia and two in Siberia. In all regions 3 to 5 diamict sheets up to 60 m thick each are found in borehole profiles (Arkhipov, 1971; Zubakov, 1972; Lavrushin et al., 1989). Tills of the penultimate glaciation can be locally seen in exposures of the Arctic beneath Upper Pleistocene strata, whereas diamictic formations of preceding ice advances normally lie below sea level. The most continuous till sheet, observed in the subarctic zone as the second from the surface, is believed to form the drift limit along 60° N in West Siberia, deviating northwards in Central Siberia and southwards in European Russia (Fig. 1). Older tills of more restricted occurrence are known just from boreholes.

The Middle Pleistocene till sheets gradually decline northwards to get progressively more eroded in the High Arctic, especially on the sea floor (Fig. 2). The occurrence of pre-Weichselian tills is mantle-like: they can be found both in buried valleys at 300-350 m b.s.l. (locations 14 and 15 in Fig. 1) and on plateaus 500-600 m a.s.l. (locations 11 and 13 in Fig. 1). The thickest accretions of Middle Pleistocene sediments, up to 200-300 m, are known from the areas of rougher sub-Quaternary relief along the Urals and Central Siberian Plateau, whereas the flat Timan Ridge and central West Siberian lowlands often bear a discontinuous cover of glacial deposits less than 50 m thick. Above the Arctic Circle the Middle Pleistocene sediments are either totally eroded or fill in deep depressions. The latter are normally obscured by the Weichselian glacial complex which is often glaciotectonically stacked to attain 100 m in thickness (Fig. 2). The described regional structure reflects a lowland position of principal ice dispersal centers.

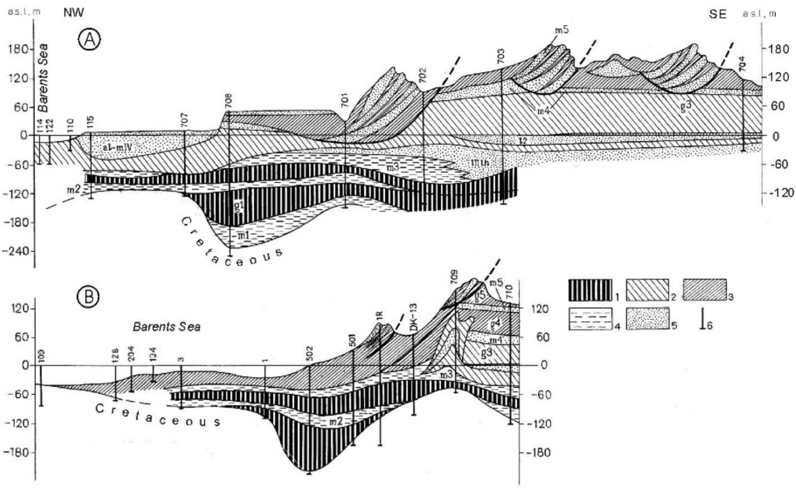

Fig. 2. Profiles of Quaternary formations on the Barents Sea coast (from Lavrushin et al., 1989, simplified). See Fig. 1 for location. Diamictons: 1 – deep buried (pre-Holsteinian); 2 – close to sea level (Saalian); 3 – surficial (Weichselian); 4 – stratified silt and clay; 5 – sand; 6 – borehole. Till sheets are g1 to g5 and marine formations are m1 to m5 (m5 is probably a glaciotectonic repetition of m4 –V.A.); lIIlh is a lacustrine formation with Likhvin-type pollen spectra according to Lavrushin et al. (1989).

Since the 1950s many authors, based on the low content of pebbles, local occurrence of marine fossils and the old idea of montane ice dispersal centres, interpreted thick diamict sheets, partly or entirely, as glaciomarine formations (Sachs, 1953; Lazukov, 1970; Zubakov, 1972). However, sedimentological analysis of fine-grained Middle Pleistocene diamictons in their stratotype sections proved beyond any reasonable doubt that they were deposited by huge ice sheets that advanced southwards (Guslitser, 1973; Kaplyanskaya and Tarnogradsky, 1974; 1975). The pattern of ice pushed ridges (Fig. 1) and dispersal of lowland clasts also indicate ice streams diverging from lowlands to the adjacent highlands (Astakhov, 1974a, 1977), contrary to ice flows from the mountains predicted by the glaciomarine hypothesis. The distribution and thickness of glacial deposits, as well as the pattern of their imbricate accretions (Fig. 1, 2), is in accordance with N-S directed ice advances from the shelf but is hardly explainable by ice dispersal from the highlands.

Size and flow pattern of ice sheets

European Russia

Ice sheets inferred from erratics

The ice limits in central European Russia since the XIX century have been attributed to activity of the Fennoscandian ice dispersal center, as suggested by the distinct boulder trains and configuration of the marginal formations. For the last half-century three major Middle Pleistocene ice advances called Oka, Dnieper (the maximum glaciation) and Moscow have been correlated with the Elster, Saale and Warthe glaciations of Central Europe (Yakovlev, 1956, Goretsky et al., 1982). In northern European Russia, however, influx of glacial ice from the northeast was acknowledged quite early (Ramsay, 1904). The contributions of individual ice domes were thoroughly discussed by Yakovlev (1956). Based on provenance of erratics in the European North, Yakovlev distinguished three major areas of ice sheet growth: 1) Fennoscandia, 2) Novaya Zemlya with the adjacent Barents Sea shelf and 3) the Urals.

The dark-grey Novaya Zemlya till is a thick, fine-grained diamicton with strong NE-SW or N-S fabrics. It is devoid of Fennoscandian erratics but contains some fragments of Novaya Zemlya pink and black limestones (Andreicheva, 1992). Large limestone blocks transported from the Kara Sea coast towards SW across the Pai-Hoi Range have been known for a long time (Voronov, 1951). Many authors also noted numerous boulders of foreign rocks scattered over flat-topped mountains of the Polar Urals up to 1000 m a.s.l. (Yakovlev, 1956). Fragments of western Uralian rocks occur east of the main watershed, limestone clasts from the low western piedmont being often found atop glacially smoothed hills 500-600 m high (Savelyev, 1966).

According to Yakovlev (1956) ice streams directed from Novaya Zemlya to the SSW reached the Volga catchment area, where they coalesced with SE flowing Fennoscandian ice of the Dnieper glaciation to form the largest ice sheet of European Russia. This ice sheet was thought to have penetrated into the southern steppes by two huge ice streams forming the Dnieper Lobe in the Ukraine and the Don Lobe in southeastern Russia. In this model ice from northeastern sources covered practically all sedimentary basins of the Russian Plain north of 58°N.

The subsequent penultimate glaciation left the reddish brown till widely observable on the surface west of the Pechora. This till is unanimously correlated with the Moscow glacial complex of Central

Russia. It contains numerous western erratics, including the characteristic nepheline syenite from the Kola Peninsula, and therefore must have been deposited by a Fennoscandian ice sheet. East of the Pechora the latter coalesced with ice streams originating from Novaya Zemlya and the Urals (Yakovlev, 1956; Potapenko, 1974; Lavrov et al., 1986; Andreicheva, 1992).

Individual ice domes are not easily identified by the erratic dispersal alone on which many authors rely heavily. A straightforward interpretation of statistics on pebble composition and orientation may be controversial. For example, tills of western provenance have long been known on the Uralian piedmont south of 64° N. Heaps of boulders of Paleozoic sedimentary rocks from the west occur even at the very foot of the highest range of the Northern Urals. However, farther to the west clasts of sedimentary rocks are mixed with central Uralian crystalline rocks, i.e. the percentage of Uralian clasts in the clayey basal till apparently increases westwards.

Varsanofieva (1933) explained this paradox by ice of northern origin streaming along the Urals and fanning off westwards into the lowlands and eastwards into the mountains. North of 64° N, just west of the highest massif of the Subpolar Urals, clayey tills are even more enriched in Uralian erratics, which led to the idea of thick ice which presumably flowed westwards from ice domes positioned over the Urals (Yakovlev, 1956; Chernov, 1974a). Pebble orientation and mineralogical composition of the till matrix were also used to suggest a separate Uralian ice dome (Kuznetsova, 1971; Andreicheva, 1992).

However, the question is not that simple. The most comprehensive work was done by Lavrov et al. (1986) who mapped the glacial topography and measured pebble composition in hundreds of till samples a quarter of a cubic meter each. The generalised results unambiguously show the persistent N-S and NE-SW dispersal of glacial clasts over the northeastern Russian Plain during the maximum Pleistocene glaciation (Fig. 3, A), west-east direction for the till of the penultimate glaciation west of the Pechora and again N-S ice flow east of this river (Fig. 3, B). The maps by Lavrov et al. (1986) offer no signs of a Uralian influence except at several locations with transverse fabric (Fig. 3, B) reported by Kuznetsova (1971). The pebble content pattern and the main vector of till fabric point directly towards the NE across Pai-Hoi and Novaya Zemlya into the Kara Sea, without disturbance by ice flows from the Urals or from the Barents Sea.

Other evidence of ice dispersal

The signals from pebble composition and mineralogy of tills are often mixed. On the western carbonaceous piedmont of the Northern Urals (63 to 61° N) Varsanofieva (1933) found that the clayey till of western provenance was covered in places by a clast-supported diamicton full of central Uralian erratics at 20 km west of the mountain front. Based on these data she suggested two glaciations: one in the form of a regional ice sheet which advanced along the Urals and a subsequent glaciation in the form of restricted piedmont ice flows from old alpine troughs. The situation is mirrored on the Siberian slope of the Subpolar and Northern Urals between 65 and 62°N (Sirin, 1947; Ber, 1948), where a till of eastern (Siberian) provenance mantling the hills up to 500 m a.s.l. is superposed

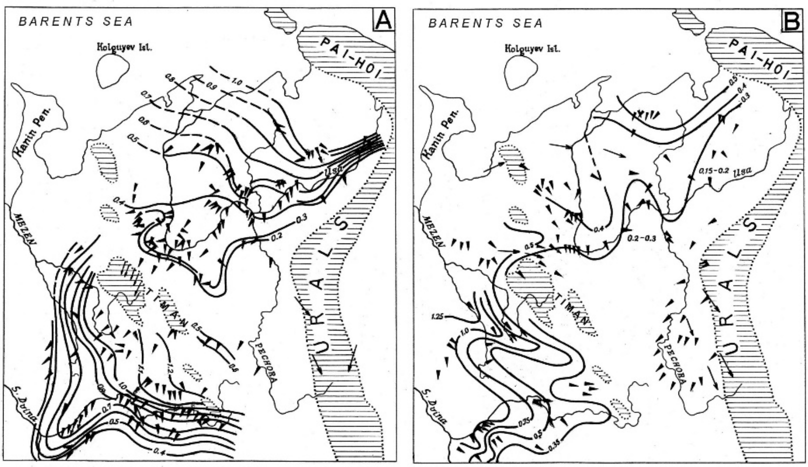

Fig. 3. Pebble content and ice flow indicators in Middle Pleistocene tills of northeastern European Russia: A – maximum glaciation; B – penultimate glaciation. By Lavrov et al. (1986) with minor additions from Matveyeva (1967) and Gornostay (1990) for the Timan Ridge and from Astakhov (1974b), Ryzhov (1974), Astakhov et al. (1999) for the Urals. Isolines show volumetric content of pebble fraction 1 to 5 cm in till samples about a quarter of cubic meter each. Black wedges are long axes of pebbles. Arrows are other ice flow indicators: striae, boulder pavements, eskers, transport paths of erratics. Hatched are Paleozoic salients of the Timan Ridge and Urals. Note the pebble content decreasing downglacier and increasing locally in the lee of the Timan Ridge at southwestern corner of Fig. 3, A.

by a cobbly diamicton of local (central Uralian) provenance. Sirin (1947) concurred Varsanofieva to suggest two glaciations on the eastern slope: a maximum ice sheet which advanced from the West Siberia onto the Urals and a local glaciation centred in the Uralian axial zone of metamorphic rocks.

Thick sequences of lowland drift later found on both slopes of the mountain range amply confirm ice advances onto the Northern Urals from both NW and NE but reject local ice caps. The thickest clayey till at 500-600 m a.s.l. in the axial zone (Fig. 4, A) contains fragments of black Carboniferous limestones picked up in the low piedmont some 20 km to the west (Astakhov, 1974a). The clayey till is capped by the 50 m high kame-like hummocks built of well-washed sand with small and roundish cobbles of central Uralian origin but obviously deposited by the same inland ice sheet, most probably of MIS 6. In contrast, younger local moraines filling alpine troughs higher than 600 m (Fig. 4, A) consist of only clast-supported tills derived from metamorphic rocks. The Siberian slope of the Northern Urals, where no alpine moraines have been mapped, is in places covered by thick matrix-supported tills of eastern provenance (Fig. 4, B) containing fragments of soft Mesozoic rocks of the West Siberian basin and the West Siberian mineralogical assemblage (Ryzhov, 1974).

In the lowlands no landforms testifying to ice flow from the Urals have ever been mapped (Fig. 1). Moreover, even on the western piedmont of the Northern Urals, 200-300 m a.s.l., eskers and striae are oriented either N-S or even NW-SE (Astakhov, 1974b). This is quite compatible with ice flow directed towards the mountains and not with a Uralian ice dispersal centre.

The 90-m thick sequence of fine-grained Middle Pleistocene glacial sediments of a kame field found in the central metamorphic zone (Fig. 4, A) has evidently never been affected by any glaciers of montane origin. Similar kames built of fine sand are also known in the east of the central range (Ber, 1948). According to this author`s observations, in several places of the Northern Urals well-preserved kames are mantled by thin clast-supported diamicton full of fragments of Uralian quartzites and schists. This sediment, which is better be interpreted as ablation till, may well belong to the same ice advance.

Thus, no second ice advance is needed to account for pebbles of Uralian provenance in the surficial diamicts of the piedmonts. The crystalline pebbles mostly originate from mountains higher than 500 m and therefore are rare in the basal till but abundant in the ablation till, which was enriched in clasts from nunataks during the lowering of the ice surface. This agrees with the fact that the percentage of Uralian crystalline rocks in diamictons of the Pechora Basin is minimal in the oldest tills of thicker ice sheets, but clearly increases upwards in the till succession (Andreicheva, 1992). This can be explained by the growing influence of montane sources with progressive thinning of each subsequent ice sheet, but should not be taken as a proof of a separate ice dome over the Urals.

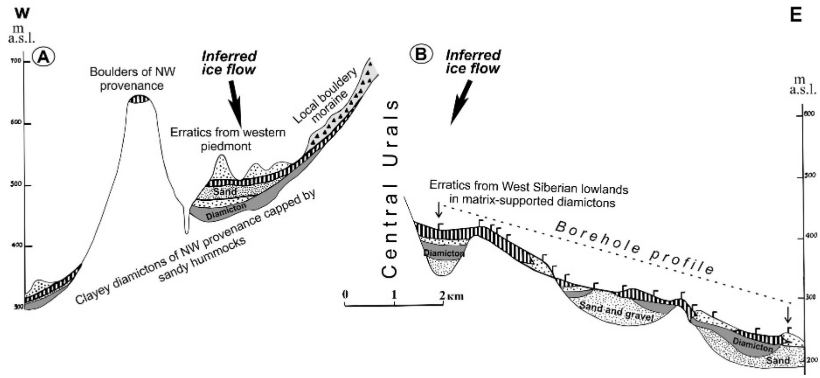

Fig. 4. Lowland tills on western (A) and eastern (B) slopes of the Northern Urals. A – river Telpos catchment, 63°50′ N/ 59° E (from Astakhov, 1974a); B – between rivers Manya and Mazapatya, N 62°15′/E 59º 50′ (from Ryzhov, 1974, simplified). Note kames of lowland glaciation undisturbed by Uralian glaciers in Fig. 4, A. Lowland erratics in basal tills (Fig. 4, A) and esker orientation in Fig. 3, B are at odds with fabrics of surficial diamicton measured by Kuznetsova (1971) in the upper Pechora area.

The above facts indicate that diamictons with west-east fabric and predominant central Uralian clasts may not represent separate glaciations but belong to the ablation subcomplex of the same ice age. Stagnating ice sheets, while thinning, might acquire reverse surface gradients forcing supraglacial boulders from Uralian nunataks to slide westwards. This could happen in the end of every ice age, thus adding to the volume of crystalline clasts already delivered to the lowlands by small alpine glaciers which probably acted prior to the main ice advances from the shelf. If the thin diamictons of Uralian provenance had been formed subglacially by westward ice flow, as Varsanofieva (1933), Kuznetsova (1971), Andreicheva (1992) and others suggested, there should have been abundant features of glacial erosion and deposition transverse to the longitudinal topography and structures of the Urals. Nothing of the kind has ever been detected by remote sensing or surface mapping. On the contrary, all large structural and geomorphic features suggest ice flow towards or parallel to the Urals.

The NNE-SSW ice flow reflected in the clast distribution is also supported by large-scale striations across the Paleozoic structures of the Timan Ridge (Fig. 3, A).

The most reliable indicators of thick upslope flowing inland ice are large glaciotectonic disturbances which in places form arcuate ridges up to 250 km long (paper 18, Ch. IV). In the Pechora Basin such features are oriented in accordance with ice flow directions from NE and NW (Fig. 1). Huge rafts of soft Mesozoic rocks, including rock salt, have been incorporated into the till of northeastern provenance. The till of the penultimate glaciation contains large basalt blocks from Timan (Guslitser, 1973). In the peri-Timan area ice flow from the west is also clearly indicated by surficial ice-pushed features. For example, along the arctic slope of the Timan Ridge a large allochthonous stack of westward dipping imbricated slices of Mesozoic and Quaternary rocks testifies to eastward glaciotectonic transport over a distance of some 40 km (Gornostay, 1990). Orientation of small glaciotectonic structures is more scattered.

In the Arctic zone ice flow features directed from lowlands into the mountains are abundant as well. A sand pit on river Hanmei near town of Labytnanghi on the eastern piedmont of the Polar Urals, examined during the PECHORA project expedition in 2000, displays Middle Pleistocene glaciofluvial gravels lying atop a very clayey till (Fig. 9). Surprisingly, these sediments contain mostly fragments of soft Mesozoic rocks from West Siberia but no clasts of resistant ultramafic rocks from the nearby Rai-Iz massif. Gravelly cross-beds all dip towards the Urals.

Farther northwards the Late Pleistocene matrix-supported Sopkay moraines strike W-E, i.e. transverse to the Urals, reflecting an ice flow from the shelf (Astakhov, 1979). Even the youngest boulder moraines on the northwestern tip of the Urals at 560 m a.s.l. show only ice push from the north, i.e. from the Kara Sea (paper 18 in Ch. IV).

The SE-facing arcuate ridges on the Siberian slope (Fig. 1), outlining a piedmont morainic apron south of the Arctic Circle (Astakhov, 1997), can hardly be of a Uralian origin, because they i) occur only downglacier of old wide troughs crossing the narrow mountain range and ii) do not have any symmetrical counterparts on the western, more humid slope. Therefore, the morainic apron of the eastern piedmont most likely originated from outlet glaciers that drained a Middle Pleistocene ice sheet of European Russia via Uralian through valleys to the SE. This suggestion agrees with the boulder trains of northwestern provenance traced across the Urals (Yakovlev, 1956).

For the interpretation of the ice flow patterns it is crucial that the sedimentological evidence of lowland glacial advance into the Urals are in accord with the pattern of ice-pushed ridges and terminal moraines in the lowlands. Large ice-pushed ridges generally strike parallel to the arctic shoreline and nowhere fringe the Uralian range except narrow morainic aprons along the foot of the mountains south of 67° N (Astakhov, 1977, 1986b, 1997). Where discrepancies exist between the lithological composition of the glacial deposits and the pattern of ice-pushed ridges, the latter should be taken as decisive evidence, because fresh large-scale geomorphic features can only be produced by the latest ice flows.

The above mentioned horseshoe-shaped end moraines of the Kara ice sheet, shoved up-valley in the Polar Urals (Astakhov et al., 1999), consist of heaps of local boulders re-orientated by ice push from the north. These and many other similar observations are good evidence that the Urals were generally overriden or bypassed by ice streams that originated north of the mountains, most likely on the western Kara shelf. The low and narrow mountain range of the Urals was not a major ice dispersal center either in the Late Pleistocene (Astakhov et al., 1999), or in pre-Eemian times of greater continental ice sheets.

Dominant ice dispersal centre

Thus, the available data indicate that during the Middle Pleistocene northern European Russia was the realm of a powerful inland ice that flowed from the NE upslope and eroded mostly soft Meso-Cenozoic sediments to deposit unusually thick and clayey tills. Lateral erosion of the Ural Mountains by transit ice streams contributed a small amount of central Uralian clasts to the generally fine-grained drift. Although the main ice streams were basically directed to the south and southwest, inland ice also flowed laterally across the Urals, overriding their flat-topped mountains when the ice was over 1 km thick. When the European ice sheet was thinner, it could reach the Siberian slope of the Urals only in the form of outlet valley glaciers.

Karpukhin and Lavrov (1974) considered the Yakovlev`s Novaya Zemlya ice sheet too large and preferred to explain the NE-SW fabric in the Upper Volga area by a diverted flow of Fennoscandian ice. According to them traces of a Novaya Zemlya glaciation and erratics of NE provenance occur mainly east of river Severnaya Dvina (Fig. 1). Immediately west of 50° E the drift limit turns almost straight south in accordance with the margin of Fennoscandian ice. For comparison, Velichko et al. (1977), based on the peculiar composition of the fine-grained Don Till, suggest that even the huge Don Lobe on 52° N was produced by the northeastern ice dispersal centre.

The strong NNE-SSW orientation of various glacial features, persistent over thousands of kilometers from the Pai-Hoi Range on 68° N to the Upper Volga on 58° N, is truly amazing, especially compared to the radially diverging pattern of Fennoscandian ice flow. It seems impossible for a small elongated ice dome over the narrow Novaya Zemlya archipelago to produce the extensive ice streams which deflected the Fennoscandian ice. The lowland ice sheets of northeastern origin, which emanated such long strings of erratics and flow features, obviously were too thick to be obstructed by most ice advances from Fennoscandia.

Yakovlev (1956), who saw the difficulty, suggested an additional accumulation area on the adjacent Barents Sea shelf. However, no traces of ice flow from the NW onto the Russian mainland have been mapped for the maximum glaciation: anywhere east of 43° E all features are directed NE-SW or N-S. Features of W-E and NW-SE orientation and clasts of western provenance first appear in surficial tills of the penultimate (Vychegda=Moscow) glaciation, when the influence of the northeastern ice dispersal centre decreased (Matveyeva, 1967; Andreicheva, 1992; Andreicheva et al., 1997). This underlines the dominant position of the Kara Sea shelf as a major source of inland ice in northern European Russia throughout the Pleistocene. Only during the penultimate glaciation the shelf ice domes were overpowered by a Fennoscandian ice sheet which advanced across the Timan Ridge at right angles to the ice flow from the Kara Sea.

Siberia

Signatures of ice motion

East of the Urals the thickness of inland ice and its flow pattern are more controversial issues, especially for the West Siberian Plain. Historically there have always been competing hypotheses of large Middle Pleistocene glaciers versus concepts of thinner local ice sheets that were assumed to reflect the drier Siberian climates. The first paradigm is commonly maintained by practicing geologists, who observe ubiquitous glacial features in various areas and at various altitudes, whereas the second trend of thinking is largely motivated by general paleogeographic considerations.

One of the main glaciation centres was inferred long ago from the pyroxene abundance in till matrix and the mafic rock fragments scattered over the Central Siberian Plateau built of Palaeozoic sedimentary formations, Triassic basalts and dolerites (Obruchev, 1931; Urvantsev, 1931; Sachs, 1953). Three morainic belts rich in dolerite boulders concentrically surround the Putorana Plateau (Isayeva, 1963). In the east the Putorana ice coalesced with a small ice sheet of the Anabar Plateau. It is interesting that during the penultimate glaciation the easternmost Anabar ice sheet, similarly to the European situation, was overpowered by eastward ice flow from Putorana (Andreyeva & Isayeva, 1974).

Ice flow from the High Arctic was inferred early basing on characteristic granite boulders transported onto the dolerite plateaus of Putorana from the Nordenskjold Archipelago and northern shores of the Taimyr Peninsula across the c. 400 m high Byrranga Mountains. The phenomenon was first thought to result from a reverse topographic gradient of the Ice Age (Urvantsev, 1931). Today it is seen as evidence of an ice dome thicker than 2 km on the Kara Sea shelf. This arctic ice sheet prevented the Putorana ice from flowing northwards (Andreyeva, 1978; Kind & Leonov, 1982).

The difference between glacial features of Central Siberia and West Siberia is striking. In Central Siberia U-shaped valleys, deeply cut into bedrock and in places barred by boulder-rich morainic arcs, radiate from the Putorana ice dispersal centre. They are accompanied by smoothed, striated, sometimes fluted and drumlinised surfaces, by occasional eskers and glaciofluvial fans (Isayeva, 1963; Arkhipov et al., 1976). There are such typical signatures of wet-based sliding as boulder pavements on the Yenissei (Troitsky, 1975).

On the contrary, eskers, flutes and marginal ridges are totally absent west of the Yenissei. They are replaced by the englacial glaciotectonic imbrications, kame fields and late glacial sandurs. Especially characteristic are thick fine-grained West Siberian tills with ubiquitous local and far-travelled blocks of soft sediments, including Quaternary and Paleogene loose sands with original lamination preserved. In many cases sedimentary rafts hundreds of meters long were transported over hundreds of kilometers (Shatsky, 1965, Zakharov, 1968; Kaplyanskaya & Tarnogradsky, 1974). The glaciotectonised structure of the glacial deposits, that is common for most sections of northern West Siberia, often precludes visual tracing of stratigraphic contacts. Such tills cannot have been deposited by lodgment, but most likely reflect the structure of dirty basal parts of huge stagnant ice sheets which slowly melted out in a degrading permafrost environment (Astakhov & Isayeva, 1988; Astakhov et al., 1996).

The wet-based versus dry-based features at the same latitudes in Siberia clearly point out to the difference between the downslope sliding over rigid bedrock in Central Siberia and slow glaciotectonic motion of the entire ice-permafrost couplet upslope in West Siberia. Especially suggestive are the huge composite imbrications and the lack of any ice retreat features. In West Siberia vast ice fields seem to have simultaneously lost mobility to decay for a very long time, being protected from beneath by thick permafrost (paper 8 in Ch. II). Masses of buried glacial ice are ubiquitous in the Upper Pleistocene tills of Arctic Siberia. Fossil ice may have even survived from Middle Pleistocene glaciations conserved in the West Siberian permafrost, as suggested by very thick massive ice sometimes found in deep boreholes.

The described combination of glacial features indicates large ice sheets that were somewhat sluggish due to the cold stored in the lithosphere over the entire Pleistocene. This accummulated cold is evident from numerous boreholes in West Siberia, which find a thick perennially frozen layer at 150-200 m from the surface even on 59-60° N, i.e. far south from the present-day continuous permafrost of the Arctic. Detached blocks and drag structures in glacial drift of the region suggest that the cold glaciers proceeded by involving into motion a 300-400 m layer of very unstable although frozen clayey substrate (paper 8 in Ch. II). This type of inland glaciation is dynamically different from other glaciated regions, which is also evident from the poor preservation of interglacial formations. Inter-till paleosols, common in North America, have never been observed in West Siberia.

The ice flow pattern is hard to recognise in the West Siberian Plain, because the predominantly soft bedrock, consisting of Cretaceous and Paleogene sand, silt, clay, opoka and diatomite provide neither streamlined features, nor enough clasts for statistical analysis. Before the onset of modern glacial sedimentology and glaciotectonic research in the 1970-s the only ice flow indicators used for reconstructions were clasts of Uralian and Central Siberian crystalline rocks. Although already Obruchev (1931), basing on the configuration of the few known morainic chains, suggested a lowland ice dome on the Taz and Gydan peninsulas, subsequent mapping failed to find its material signatures, and ensuing overviews acknowledged only ice dispersal paths from the Urals and Central Siberian uplands (Sachs, 1953; Zarrina et al., 1961; Lazukov, 1970; Zubakov, 1972; Arkhipov et al., 1977).

Nevertheless, Zemtsov (1973a, 1973b), who interpreted the composition of several thousand boulders and mineralogical samples of till matrix, concluded that there was a central zone in the north of West Siberia with no fragments of Uralian rocks or central Siberian dolerites. Tills of the central zone contained mostly material of local Mesozoic and Cenozoic formations with admixture of Taimyr rocks. Zemtsov thought this zone to have been influenced by an additional ice dispersal centre on the Taimyr

Peninsula, and he did not rule out a possibility of ice flow from a hypothetical lowland ice dome by Obruchev (1931). Also Kaplyanskaya and Tarnogradsky (1975) interpreted marine fossils found in diamictons of the Lower Yenissei area as signatures of ice flow from the Kara Sea.

Puzzling erratics

Again, as in European Russia, in West Siberia a discrepancy exists between the strong signatures of upslope ice flow from lowland ice domes (Astakhov, 1976, 1977) and the dominant petrographic and mineralogical composition of the glacial drift, which seemingly is evidence to the contrary. E.g. the most comprehensive study of till composition in West Siberia (Sukhorukova et al., 1987) reveals three major clastic provinces roughly coinciding with the earlier conclusions by Zemtsov (1973a, 1973b): Central Siberian, Uralian and West Siberian (Fig. 6). Sukhorukova et al. (1987) maintain that these provinces reflect the relative significance of three major ice dispersal paths: from the Putorana Plateau, from the Urals and from the Kara Sea shelf.

First, this interpretation contradicts the pattern of ice-pushed features which everywhere only reflects ice flow directed upslope (Astakhov, 1977). Second, it is glaciologically impossible: there are no obstacles in the very low central West Siberian Plain that would prevent Uralian and Central Siberian ice streams from moving farther south beyond the mapped drift limit (Fig. 1 and 6). Especially strange is the sudden increase of Uralian clasts in the Ob tills south of 64° N, which made Sukhorukova et al. (1987) suggest an ice dome in the Northern Urals instead of the Polar Urals, as would follow from the general N-S ice direction of local boulder trains. In addition, the till fabric measurements along the Yenissei show only ice movement towards the south and not westwards across the valley. The suggestion by Kaplyanskaya and Tarnogradsky (1975) that upland ice divides, formed in the beginning of each ice age, afterwards might have shifted into the lowland, only partially helps.

It must be taken into account that the West Siberian tills are typical lowland diamictons consisting largely of fine-grained material derived from the underlying Mesozoic and Paleogene sediments. The admixture of hard rock pebbles is very low (less than 1%) and decreases northwards. Another interesting feature is flat-iron and wedge-shaped forms among the pebbles, noticeably more frequent than in other glaciated regions. The ideal shape of glacial abrasion is often found in the same pebble sample on very dissimilar materials, such as hard dolerites and soft opokas, unlike the situation in Urals and Central Siberia, where wedge-shaped clasts mostly consist of schists and limestones.

Also, up to 40% of the West Siberian glacial pebbles show palimpsest aqueous roundness. Clasts in till are normally found to be well sorted by their size, which should be characteristic of fluvial rather than glacial transport (Sukhorukova et al., 1987). All these are unmistakable signs of very long paths of glacial and fluvial transport that involved multiple redepositions which certainly distorted the initial boulder trains. Therefore, the mapped ice limits, in agreement with the flow pattern reflected in ice-pushed features, show only the direction of the final, most powerful ice advances from the north but not any earlier paths of clast dispersal.

However, the progressive mixing and integration of pebble and mineralogical composition over several ice ages cannot explain the occurrence of large angular boulders from the Urals and Central Siberia, which are occasionally found in the central West Siberian lowlands. These can probably be accounted for by the same mechanism as suggested for the Pechora Basin, i.e. by superglacial transport of stones from mountainous borderlands towards central lowlands caused by the reversed ice surface gradient after the late glacial collapse of the lowland ice domes.

Ice thickness and major ice dome

The mapped configuration of the glacial drift limit (Fig. 1) and large directional features are clear indications of thick ice upslope moving during glacial culminations, irrespective of the predominant mechanisms of crystalline clasts transport to the lowlands. In the adjacent mountains of the Urals and Central Siberia there is ample evidence of upslope ice flow, such as lowland tills found high in the mountains (Fig. 4, loc.11, 12, 13 in Fig. 1).

The crucial indication of southbound flow of thick Middle Pleistocene ice in West Siberia is the system of arcuate ice-shoved imbrications of intricately disturbed soft rocks, often topographically expressed as hill-hole pairs. The system of the largest ice pushed ridges runs roughly parallel to the coast of the Arctic Ocean (Fig. 1). The largest zones of thrusted and tightly folded sediments are up to 200 km long and 20-25 km wide (Zakharov, 1968; Troitsky, 1975; Arkhipov et al., 1976; Astakhov, 1979; 1986b). The disturbances may penetrate down to 400 m below the surface, which is probably the world record. The tectonic style of these structures indicates their deformation in a frozen state and is evidence of deep crumpling of soft substrate during ice movement (paper 8 in Ch. II).

These structures, normally consisting of para-autochthonous slices or detached blocks of local provenance, as a rule occur far upglacier from the ice margin. Along the ice margin they are often replaced by huge rafts of far-travelled Mesozoic, Paleogene and Quaternary sediments (the famous Yugan, Samarovo and Semeyka erratics of soft Jurassic and Paleogene rocks hundreds of meters long). They sometimes have been transported as far as 600 km downglacier (Shatsky, 1965; Kaplyanskaya & Tarnogradsky, 1974; Arkhipov et al., 1976). The above facts imply a thickness of the West Siberian ice sheets in the order of kilometers, not hundreds of meters as was thought in the 1950-60s (e.g. Lazukov, 1970; Zubakov, 1972).

Independent evidence of very thick Middle Pleistocene ice in West Siberia is provided by numerous glacial valleys with irregular bottom profiles that are buried by drift, in places 300-400 m thick (Arkhipov & Matveyeva, 1964; Zubakov, 1972; Arkhipov et al., 1976, 1994). These overdeepened, predominantly N-S striking valleys, never occur in the proglacial zone. Their glacial origin can be readily seen in the longitudinal profile of the Quaternary thickness across the drift limit based on numerous boreholes along the Ob and her left tributaries. In the periglacial area Pre-Quaternary bottom profiles, gently sloping parallel to the present-day fluvial thalwegs, do not show any overdeepening (1 to 4 in Fig. 5). Immediately north of the maximum glaciation limit the buried bottom of the Ob valley becomes very irregular, plunging deep below sea level (5, 6 in Fig. 5), which cannot be explained by normal fluvial processes.

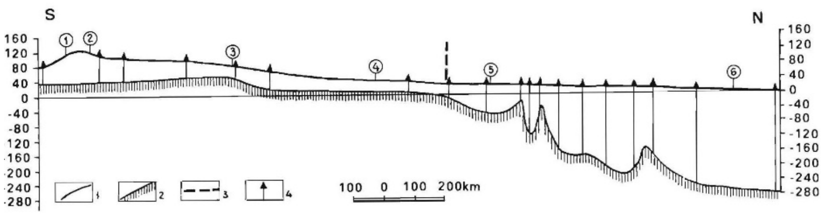

Fig. 5. Longitudinal profile of buried valleys of the Ob river system (from Astakhov, 1991). Symbols: 1 – present thalwegs; 2 – sub-Quaternary surface; 3 – glacial drift limit; 4 – profile of boreholes across the valleys. Circled numbers: (1) source of river Turgai flowing to the Aral Sea; (2) source of river Ubagan; (3) mouth of river Ubagan; (4) mouth of river Tobol; (5) River Irtysh mouth; (6) River Ob mouth. Compare normal fluvial profiles of extraglacial valleys in the left with bumpy overdeepened trough of glaciated area in the right of the figure.

The principal ice dome over the Kara Sea shelf and adjacent lowlands only got generally (although not unanimously) accepted after remote sensing had revealed the pattern of ice pushed ridges and after foreign erratics found in the mountains were considered (Astakhov, 1976, 1977, 1979). Lowland tills of Mesozoic provenance of the West Siberian Basin (Fig. 4, B) and erratic boulders found high in the Urals indicate that above the Arctic Circle inland ice covered summits more than 1 km high, whereas

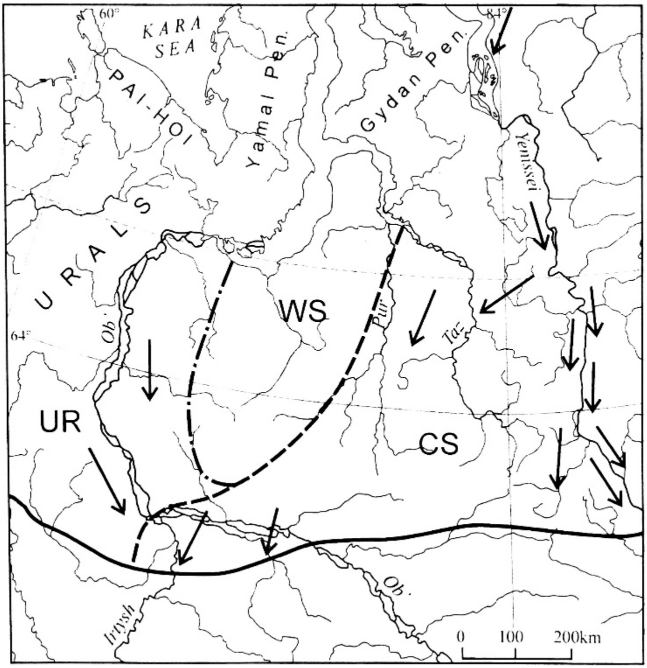

Fig. 6. Distribution of erratics in tills of the West Siberian maximum glaciation (from Sukhorukova et al., 1987, simplified). UR – Uralian province of crystalline and Paleozoic clasts; CS – province of Central Siberian clasts: dolerites, carbonate rocks and sandstones; WS – province of no Uralian and Central Siberian stones, with predominating quartz, opokas, diatomites, siltstones, sandstones of West Siberian Mesozoic and Cenozoic formations. Taimyr boulders sporadically occur in both WS and CS provinces.

Arrows are ice flow directions inferred from fabrics and erratics. Dispersal of Central Siberian and Uralian erratics is not conformable to configuration of the ice limit (bold line), orientation of ice-pushed ridges (Fig. 1) and with lowland tills deposited in the mountains (Fig. 4).

at 62-63° N the trimline occurs at ~ 500 m a.s.l. In the western part of the Central Siberian Upland three till sheets with lowland erratics, 33 m thick altogether, were described on the table-like summit of a dolerite monadnock 618 a.s.l. at 150 km from the drift limit (location 13 in Fig. 1, Fainer et al., 1976). The surface of the inland ice, which brought Mesozoic material from the northwest and deposited it atop the inselberg at loc. 13, Fig. 1, must have been higher than 700 m a.s.l. (Fainer et al., 1976) and at least 900 m upglacier in the Yenissei valley, south of loc. 9 (Fig. 1), where till of the maximum glaciation was found below sea level (Zubakov, 1972). These observations make the West Siberian ice sheet at least 500 m thick at 64°N at 150 km from the drift limit and ~ 1000 m thick at 65° N, i.e. at 350 km upglacier from the margin of the maximum ice sheet.

Voronov (1964), basing on empirical profiles of Antarctic and Greenland ice sheets and the known configuration of the drift limit in Siberia, calculated the maximum thickness of the West Siberian ice sheet as 3.5 km on the Yamal Peninsula. Judging by the above geological facts, this estimate seems to be realistic. To overpower and divert southwards ice streams from the Putorana Plateau, as evidenced by the foreign tills east of the Yenissei, the lowland ice dome must have been really thick and probably occupied the entire Kara Sea shelf.

Chronology of ice advances

European Russia

General stratigraphic framework

The modern chronological concepts of Middle Pleistocene glaciations in northern European Russia are heavily dependent on the official stratigraphic scale of Central Russia, either directly, as in the Arkhangelsk Region of prevailing Fennoscandian glaciers, or in disguise of local startigraphic labels, as in the Timan-Pechora-Vychegda Region of dominant ice advances from the northeast (Guslitser et al., 1986).

The stratigraphic cornerstones of central European Russia are classical interglacial formations of biogenic and limnic sediments at the Likhvin and Mikulino stratotypes, indicating climates warmer than the present. They were traditionally correlated to the Holsteinian and Eemian. The most continuous till sheet is poorly topographically expressed and for a long time has been associated with the Dnieper maximum of the Fennoscandian glaciation in the Ukraine, close to 48° N, and with the German Saale glaciation.

Table 1. Correlations of pre-Weichselian glacial events of northeastern European Russia with other regions

This paper	Guslitser et al., 1986 + Velichko and Shick, 2001	Central Russia	Northern Germany	MIS
Sula interglacial	*Sula interglacial*	*Mikulino interglacial*	*Eem*	***5***
Vychegda glacial	**Vychegda glacial**	**Moscow glacial** *Kostroma interstadial* **Dnieper glacial**	**Warthe** *Interstadial* **Drenthe**	**6**
	Rodionovo interglacial	*Interstadial* Cold stadial *Chekalin interglacial*	*Wacken-Dömnitz*	*7* **8** *9*
	Pechora glacial	Kaluga cold stage	**Fuhne**	**10**
Chirva= Rodionovo	*Chirva interglacial*	*Likhvin interglacial*	*Holstein*	***11***
Pechora glacial	**Pomusovka glacial**	**Oka glacial**	**Elster**	**12-14**
Visherka interglacial	*Visherka interglacial*	*Roslavl= Muchkap interglacial*	*Cromer*	***15***
Maximum glacial	**Beryozovka glacial**	**Don glacial**		**16**
Kolva transgression	*Tumskaya interglacial*	*Ilyinka interglacial*		***17***
Pre-Kolva glacial	**Kama glacial**	**Likovo glacial**		**18**

Another, less extensive pre-Eemian glacial complex, mapped mainly north of 54° N by its distinct glacial topography, is attributed to the Moscow glaciation, presumably equivalent to the Warthe glaciation (Yakovlev, 1956; Goretsky et al., 1982). The till lying under the Likhvin marker beds for many years was thought as representing the oldest Oka (Elsterian) ice sheet of more limited extent.

The most problematic was the third interglacial formation (Odintsovo, or Roslavl) found between two uppermost Middle Pleistocene tills. This sequence, showing a characteristic pollen profile with two deciduous peaks, is notably different from both Likhvin and Mikulino pollen successions. In formal stratigraphic schemes and in general maps of the Quaternary it was for many years placed between the Moscow and maximum Dnieper glaciations (Krasnov, 1971; Ganeshin, 1976), although there always were dissidents insisting on a much older age of the Roslavl strata.

This scheme with two separate Saalian glaciations was also applied to northern European Russia, where the Novaya Zemlya till was associated with the maximum Dnieper ice advance and the younger till containing Fennoscandian erratics – with the Moscow glaciation. The official regional stratigraphic scheme offers correlation units (climatostratigraphic horizons) replicating central Russian climatoliths (Guslitser et al., 1986). Several glacials and interglacials were distinguished from litho- and pollen stratigraphy in parallel with the old Central Russian stratigraphic scale, in which Saalian tills were separated by the very warm interglacial called Odintsovo (Roslavl) around Moscow and Rodionovo on the Pechora.

However, in the 1970s it was discovered that the till of the Don Lobe was deposited earlier than the Dnieper Till of the Ukraine (Velichko et al., 1977, Velichko & Faustova, 1986). The Don Till is overlain by the Muchkap interglacial strata with two characteristic optima of broad-leaved trees and with remains of typically Tiraspol (Cromer) rodents. A similar fauna was described from the stratotype sections of the Roslavl interglacial. This implied that the maximum glaciation on the Don was separated from the Dnieper Till by at least two interglacials: Roslavl (Muchkap) and Likhvin. Thereby an intra-Saalian interglacial had to be abandoned (Shik, 1989).

The new stratigraphy was proven beyond any reasonable doubt when sediments with the Likhvin (Russian Holsteinian) floras were found atop the Roslavl sequence (Biryukov et al., 1992). In the stratigraphic scheme of Central Russia, presently used by the Russian Geological Survey, no early Saalian till or overlying intra-Saalian interglacial formation (MIS 8 and 7) are mentioned. The only till sandwiched between the Likhvin and Mikulino formations is the Moscow glacial complex correlated with MIS 6 which presumably laterally merges with the Dnieper Till of the Ukraine (Shik, 1989).

Still, there are researchers who accept the new pre-Holsteinian Don-Muchkap cycle but also insist on two separate glaciations (the Dnieper and Moscow) between the Likhvin and Mikulino warm stages (e.g. Sudakova & Faustova, 1995). Recently an attempt to fill in the gap between the Moscow till and Likhvin strata has been undertaken by Velichko and Shik (2001). They suggest as climatochronological units 2 paleosols and 2 loess units found directly on top of the Likhvin lacustrine formation with *Brasenia* flora (Table 1). In this new scheme they place the Dnieper glaciation at the beginning of MIS 6, implying that the maximum drift limit in the Ukraine is somewhat older than the Moscow Till within the same ice age. The maximum glaciation of the eastern Russian Plain is firmly outlined by the Don Lobe (probably MIS 16) reaching south to the 50-th parallel.

Stratigraphy in the Arctic

The modern stratigraphic scheme first did not affect the remote northern areas, where two Saalian tills (Pechora and Vychegda) were conventionally correlated with the Dnieper and Moscow glaciations. These tills are separated by a warm interglacial with two climatic optima described at Rodionovo as a counterpart of the Odintsovo/Roslavl (2 in Fig. 1) (Guslitser et al., 1986; Loseva et al., 1992; Duryagina & Konovalenko, 1993; Andreicheva et al., 1997). The situation is more complicated in the Arctic, where inter-till marine formations are not readily correlated with the central Russian stratotypes, or with terrestrial interglacials of the Middle and Upper Pechora catchment.

Drilling projects by the Geological Survey provided a wealth of information on the structure of drift in the Arctic. Two such borehole profiles supported by field observations (A and B in Fig.1) reveal at least 4 independent diamict sheets separated by 4 marine formations (Fig. 2). The topmost glaciotectonised complex with blocks of marine sand is obviously a deposit of the shelf-based Weichselian glaciation (Astakhov et al., 1999; Svendsen et al., 2004). It is separated from the thickest till body g3 by sands with an Eemian fauna (m4 by Lavrushin et al., 1989). The g3 till lies above sea level and cannot be anything but a Saalian till. Just below sea level it is underlain by lacustrine sediments with Likhvin type pollen spectra merging seawards with marine formation m3 (Lavrushin et al., 1989). The two lowermost marine formations m1 and m2 and intervening till sheets g1 and g2 cannot be correlated directly with any of the southern stratotypes.

The old view of one boreal transgression separating two last glacial events (Ramsay, 1904) was challenged by Yakovlev (1956) and by geologists who reported the first drilling results in the White Sea basin (Biske & Devyatova, 1965). They maintained that two boreal transgressions with similar mollusk fauna occurred: the Boreal transgression *s. stricto*, an arm of the Eemian sea, and the so-called Northern transgression, which was conventionally correlated with an intra-Saalian event.

The second shallow-water Atlantic transgression was later supported by drillings in the Pechora catchment area, where two sandy formations with boreal fauna were found alternating with diamictons. The upper boreal formation with *Arctica islandica* had been known long ago from many sections above sea level as a counterpart of the Eemian, whereas the lower formation with characteristic extinct mollusk *Cyrtodaria jenisseae (angusta)* was reported from beneath a thick diamicton just below sea level (Zarkhidze, 1972b).

A third marine sequence of deep-water silts, the so-called Kolva Formation, found well below sea level at the base of the Quaternary cover, contains mostly a subarctic fauna with characteristic *Propeamussium groenlandicum* (Yakhimovich et al., 1973) and probably corresponds to unit m1 in the Lavrushin`s profiles (Fig. 2). In some boreholes it is underlain by a diamicton. From the Kolva Formation Gudina (1976) described arctoboreal foraminifers with characteristic *Miliolinella pyriformis,* which she correlated with the Ob and Turukhan strata in West Siberia and with the marine Holsteinian. The latter correlation is at odds with palynological investigations which find Likhvin-type spectra stratigraphically higher than the Kolva Formation (Fig. 2). In one borehole a diamicton was discovered at the base of the Kolva Formation (Yakhimovich et al., 1973).

Stratigraphy in the Subarctic

Terrestrial interglacial sediments sandwiched between tills are known mainly from three natural exposures in the Pechora catchment area. The best studied is the Rodionovo section (2 in Fig. 1) with very compact, slated and slightly distorted peat up to 3.5 m thick contained in clay and silt within a sandy inter-till fluvial sequence. In the arboreal pollen spectra spruce, pine and birch are dominant with admixture of *Abies* and *Alnus* (Loseva & Duryagina, 1973). Two climatic optima were inferred based on a small admixture of broad-leaved trees pollen (Duryagina & Konovalenko, 1993), which led the referred authors correlate this sequence with the Odintsovo/Roslavl interglacial of Central Russia.

In the Kipiyevo section (3 in Fig. 1) lacustrine strata with a similar pollen assemblage also contain large *Unio* shells, thick coniferous logs with `formidable growth of tree-rings` and nuts of *Ajuga reptans,* which now lives only in oak forests some 400 km to the south (Guslitser & Isaychev, 1983). The Kipiyevo interglacial strata are dissected by ice wedge-casts and overlain by sediments containing a late Middle Pleistocene assemblage of teeth of pied and grey lemming, similar to those found between the Likhvin lacustrine strata and the overlying till south of Moscow. The evolutionary level of the Kipiyevo lemmings is somewhat higher than in the Likhvin stratotype, which made Guslitser and Isaychev (1983) suggest a Moscow-Warthe age for this rather archaic fauna.

They also describe another find of similar rodent remains in the Akis section 6 km downstream of loc. 2, Fig. 1. At this site lemming teeth were collected from cross-bedded sand overlying the same Novaya Zemlya till as is at the base of the Kipiyevo sequence. The evolutionary level of the Akis lemmings is only slightly higher than in the Lower Saalian beds of the Likhvin section, again with no progressive Late Pleistocene morphotypes present. Guslitser and Isaychev (1983) relate the Kipiyevo interglacial to the Saale-Warthe interglacial (MIS 7) and the underlying till to the `early Saale`.

Recent dating attempts by the Russian-Norwegian PECHORA project confirm an OIS 6 age for the surficial till of the subarctic zone. The dated section is located in present-day tundra on river Seyda, formally in the Arctic (4 in Fig. 1) but featuring only terrestrial sediments. A compact peat layer 1 m thick and 300 m long, with forest pollen spectra even richer than in Rodionovo, is contained in a thin sand sheet at the base of the 40 m thick stacked till sequence. The peat first yielded a finite radiocarbon date and was thought to represent the Middle Weichselian (Lodmashchelye section by Arslanov et al., 1987).

However, more detailed sampling of this sequence by J.I. Svendsen and M. Henriksen provided much older ages. An uninterrupted series of samples of the inter-till peat analysed for U/Th ratio in the laboratory of St. Petersburg University yielded ages of ca 200 $\pm$ 30 ka BP in the middle of the peat layer. Younger values obtained from the top and bottom of the peat are accounted for by post-depositional influx of younger uranium (analyst Yu. Kuznetsov). OSL dating on quartz particles in the peat produced ages of 180 $\pm$ 13, 185 $\pm$ 12 and 191 $\pm$ 37 ka, whereas the surrounding sand was dated to c. 144 ka (3 datings) and once to 173 $\pm$ 75 ka. Glaciofluvial and glaciotectonised sands in the upper part of the overlying glacial complex sequences yielded OSL values of 148 ± 10, 149 ± 13, 152 ± 11, 156 ± 16, 160 ± 9 and 170 ± 13 (Table 2).

Thus, there are two different sets of OSL dates: one is close to 150 ka and another is close to 200 ka. The consistency of the dates and the large measured dose rates, according to A. Murray, make these dates apparently reliable. The 50 ka difference in OSL ages of the two sets of samples probably means that a Late Saalian glacier picked up a peat layer ca 200 ka old and deposited it as a stratiform raft together with younger glaciofluvial sand. A Saalian age of this sequence is supported by OSL dates of 109 ± 8 from the top and 143 ± 12 ka from the base of the overlying aeolian sand. A Late Saalian OSL date of 152 ± 9 ka was also obtained from the sand between the peat and the upper till at Rodionovo (Table 2, sample by O. Maslenikova).

The age of the lower, `Novaya Zemlya` till of the Pechora basin is less certain. Its correlation depends on the age of the Rodionovo interglacial. MIS 7 suggested by local geologists from comparison with the central Russian Odintsovo-Roslavl sequence (Guslitser et al., 1986) looks like a miscorrelation, because the Roslavl strata around Moscow are certainly pre-Likhvin, or pre-Holsteinian (see above).

How many ice advances?

This question heavily depends on the correlation of the rare interglacial sequences. It is interesting to compare the paleontological characteristics of the Rodionovo Formation with the Chirva interglacial strata known from boreholes in the south of the subarctic zone (Vychegda catchment area) and correlated by pollen with the Likhvin interglacial (Duryagina & Konovalenko, 1993). These and other authors (Guslitser et al., 1986; Loseva et al., 1992; Andreicheva et al., 1997) presume that the Chirva predates the Rodionovo.

However, both formations contain almost identical botanical taxa, including indicative Likhvin species *Osmunda claytoniana* and *O. cinnamomea* (Table 7 in Duryagina & Konovalenko, 1993). Both pollen diagrams feature 5 zones and two climatic optima with an admixture of broad-leaved trees, both, unlike the older Visherka horizon, contain no Tertiary relicts and only rare exotic (Balkan-Caucasus) elements such as *Betula sect. Costatae, Picea sect. Omorica, P. sect. Strobus.* The slightly richer floristic composition of the Chirva strata is easily explained by the more southerly location of the studied site. As mentioned above, the rodent assemblages above the Kipiyevo (Rodionovo) strata are very similar to those found on top of the Likhvin type sequence. The only OSL date available from the base of the Rodionovo peat gave a fairly old value 334 ± 29 ka, more appropriate for MIS 9 or 11 than for MIS 7 (sample by O. Maslenikova, Table 2).

Table 2. Optically stimulated luminescence dates by the PECHORA project according to A. Murray, the Nordic Laboratory for Luminescence Dating, Risø, Denmark.

Lab. no	Sample no	Location	Age, ka	Dose rate, Gy/ka	Paleo dose, Gy
002527	99-1238	Rodionovo, above peat	152 ± 9	1.20 ± 0.04	183 ± 6
002526	99-1237	Rodionovo, below peat	334 ± 29	0.97 ± 0.04	323 ± 22
992529	98-3078	Seyda 1, aeolian	109 ± 8	1.73 ± 0.09	189 ± 8
992528	98-3077	Seyda 1, aeolian	143 ± 12	0.95 ± 0.05	136 ± 8
002540	99-4226	Seyda 1 sand in till	148 ± 10	1.49 ± 0.06	220 ± 10
002539	99-4225	Same	149 ± 13	1.38 ± 0.05	206 ± 15
002541	99-4227	Same	152 ± 11	1.50	230 ± 13
002542	99-4228	Same	160 ± 9	1.74	279 ± 9
992512	98-3064	Same	156 ± 16	1.87 ± 0.09	291 ± 24
992513	98-3065	Same	170 ± 13	1.79 ± 0.09	305 ± 13
992515	98-3029	Seyda 1, sand below till	144 ± 14	1.76 ± 0.09	254 ± 18
992511	98-3063	Same	144 ± 13	1.44 ± 0.07	207 ± 14
962523	95-0070	Same	145 ± 42	2.15	289
992502	98-3035	Seyda 1, sand above peat	185 ± 12	1.82 ± 0.09	336 ± 4
992501	98-3033	Seyda 1, intergl. peat	180 ± 13	1.39 ± 0.07	251 ± 10
962524	95-0071	Seyda 1, intergl. peat	191 ± 37	1.96	347

962525	95-0072	Seyda 1, sand below peat	173 ± 15	1.68	271
012584	01-0197	Sangompan, below rhythmite	82 ± 5	1.61 ± 0.07	131 ±4
012583	01-0196	Same	93 ± 7	0.78 ± 0.05	72 ±2
022518	01-0190	Same	77 ±6	1.54 ± 0.06	118 ± 8
022519	01-0191	Same	72±5	1.29 ± 0.05	93 ± 4
012586	01-0134	Pyak-Yaha, above rhythmite	133 ± 11	1.59 ± 0.08	211 ±14
012585	01-0133	Same	138 ± 8	1.51 ± 0.07	208 ±5
012581	01-0115	Same	125 ± 10	1.77 ± 0.08	221 ±14
012582	01-0116	Same	137 ± 9	1.66 ± 0.08	227 ±9
012579	01-0119	Pichuguy-Yaha, below rhythmite	197 ± 15	0.91 ± 0.06	179 ±8
012580	01-0120	Same	192 ± 16	0.78 ± 0.05	150 ±7
012548	00-0505	Aksarka 2, coversand	17.3 ±0.8	1.70 ± 0.07	81 ±3
012547	00-0504	Aksarka 2, loess-like silt	19.8 ±0.9	2.13 ± 0.08	81 ±3
002550	00-0503	Aksarka 2, sand	97 ±8	1.90	184 ±9
002549	00-0500	Aksarka 2, sand with wood	84±10	1.94	168 ±16
002545	00-0450	Sopkay, sandur	167 ± 22	1.17	192 ±23
002546	00-0451	Sopkay, sandur	96±13	1.54	148 ±17
012544	00-0418	Yerkata, aeolian sand above till	58 ±4	1.84 ± 0.07	106 ±5
012543	00-0416	Same	62 ± 5	1.89 ± 0.08	118±7
012542	00-0414	Yerkata, lacustrine sand above till	68±5	1.69 ± 0.07	115±6
012541	00-0413	Same	59 ± 4	1.94 ± 0.07	114 ± 5

The most important is the stratigraphic position of the interglacial formations in borehole profiles of the Arctic zone (A, B and C in Fig. 1). Loseva et al. (1992) relate the pollen spectra of the terrestrial inter-till sequence in boreholes 8-y, 5-y, 3-y and 71 (Fig. 7) to the Chirva interglacial, but the marine formation at the same level in the SE boreholes 754 and 755 is correlated with the younger Rodionovo interglacial. Consequently, the overlying till in the NW boreholes is thought to be the Pechora till (MIS 8 in their correlation), but its lateral counterpart in the SE is referred to the younger Vychegda Till (MIS 6) (Fig. 7). Neither the lithological composition of the tills, nor the landscape features support such a differentiation. In the profile described by Lavrushin et al. (1989) slightly farther to the east (A in Fig. 1) the lateral change from the Likhvin lacustrine formation to the marine formation m3 (Fig. 2), suggested by the authors, seems more logical.

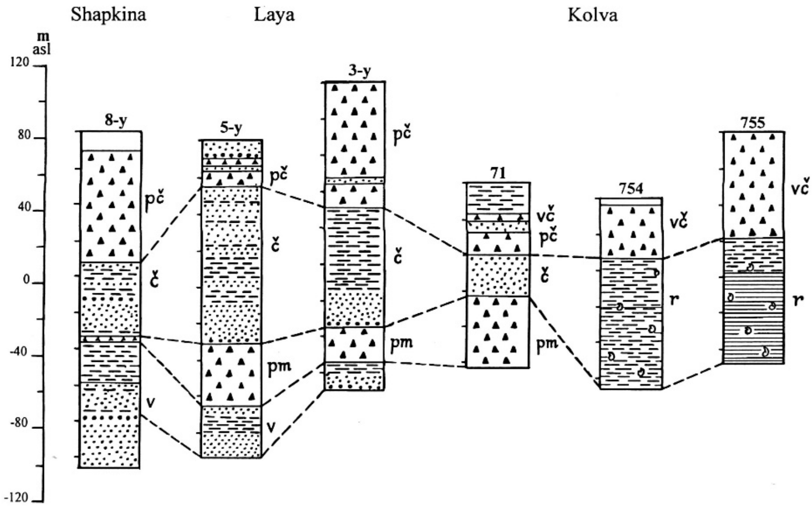

Fig. 7. Middle Pleistocene formations in boreholes of profile C (Fig. 1) (from Loseva et al., 1992, simplified). Letters indicate correlation with regional climatostratigraphic horizons assumed by Loseva et al. (1992): v – Visherka interglacial, pm – Pomusovka glacial, č – Chirva interglacial; pč – Pechora glacial; r – Rodionovo interglacial, vč – Vychegda glacial. Broken lines are a possible alternative correlation.

The Chirva and Rodionovo interglacials have never been found in superposition (Fig. 2, 8). Therefore, it appears that the uppermost Middle Pleistocene interglacial, the Rodionovo on the Pechora, is identical with the Chirva interglacial of the Vychegda catchment area. Many nterglacial sequences covered by the upper till in the subarctic zone, and lying just below sea level in the Arctic, probably belong to the same interglacial interval, most likely represented by the marine formation with *Cyrtodaria angusta.* The *Cyrtodaria* strata are in places found up to 70 m a.s.l. (Zarkhidze, 1972b), which gives a rough idea of the large isostatic depression caused by the thick preceding ice sheet.

In the compromise scheme by Velichko and Shik (2001), accepting the different Chirva and Rodionovo interglacials, the Pechora glaciation is not correlated anymore with the Dnieper or other maximum ice advance of Central Russia. Instead it is shifted downwards to MIS 10, probably in order to accommodate at MIS 12 another glaciation called the Pomusovka by Guslitser et al. (1986) (Table 1). This results in the correlation of the mighty Pechora glaciation of NE provenance with the thin Kaluga loess on top of the Likhvin stratotype sequence. However, the Pechora Till, widely observable in many subarctic sections and covered by the distinctly interglacial Rodionovo Formation, is still the most salient feature of the northern drift. Therefore, it better correlates with a central Russian till covered by the Likhvin strata, the Oka Till, which has been always thought as Elsterian in age. There is no lithostratigraphic evidence of a laterally consistent Pomusovka glacial complex.

Whether the Chirva/Rodionovo interglacial corresponds to MIS 9 or to MIS 11 is uncertain, but its stratigraphic position makes it the best candidate for correlation with the Russian Holsteinian, i.e. the Likhvin *s. stricto*. The preceding Visherka interglacial sequence, containing Tertiary relics such as *Liquidambar* and *Pterocarya* plus exotic *Tsuga* and *Ilex,* not known from younger interglacials, consequently would better correspond to the Roslavl/Muchkap strata overlying the Don Till of the maximum glaciation, as suggested by Velichko and Shik (2001) (Table 1).

In this respect sections close to the glacial drift limit are of particular interest. The best picture of the geological structure of the Pechora-Volga interfluve area is given by the geotechnical drilling data described by Stepanov (1974, 1976). Two of his profiles intersecting at a right angle are presented in Fig. 8. The surficial till g3 correlated by Stepanov with the Dnieper glaciation is covered by only one interglacial alluvial formation of the modern valleys with fresh-water mollusks, rich diatom flora and characteristic Mikulino pollen spectra. A modern area of the same concentration of the fossil flora can be found in southern Germany and Poland. The same is true of the Mikulino type flora of central Russia.

The interglacial formation between g3 and the underlying g2 till (Fig. 8) contains a flora similar to the Likhvin assemblage with indicator plants

Osmunda claytoniana, Pinus sect. Strobus, Picea sect. Omorica, Tilia tomentosa, present inhabitants of the western foothills of the Alps, according to L. Tyurina. The plant macrofossils identified by P.I. Dorofeyev are typical for the Singil floras of the Russian `Mindel-Riss` (Stepanov, 1976). A single find of forest elephant *Palaeoloxodon sp.* is known from a similar sequence on river Kolva, tributary to river Kama (Yakhimovich et al., 1973). The overlying g3 till should be the Uralian counterpart of the Vychegda till. The underlying dark-grey till g2, low on Uralian boulders, contains mostly clasts of sedimentary rocks and in this respect is not different from the Pechora till.

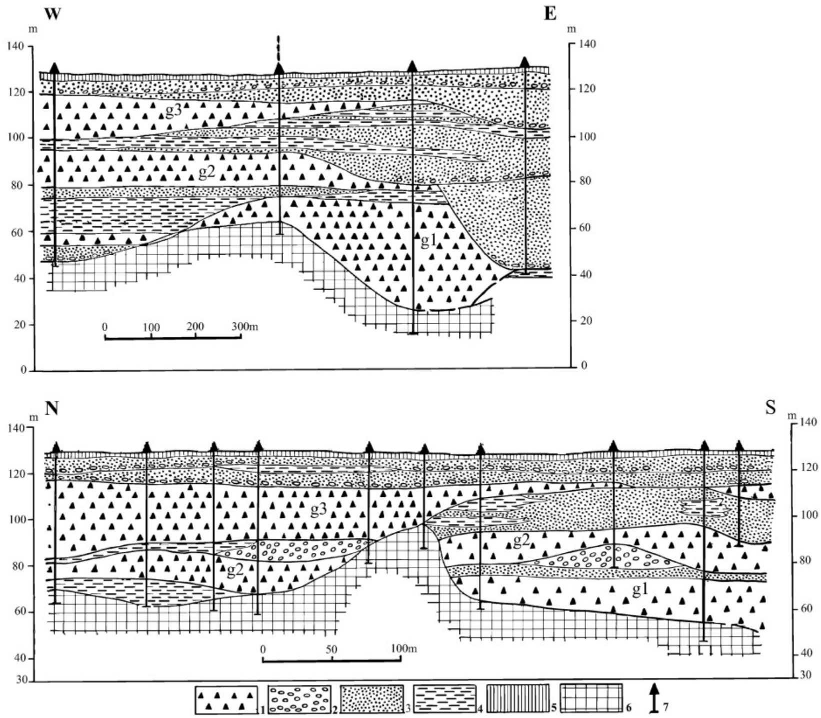

Fig. 8. Borehole profiles of Middle Pleistocene tills at Lake-Chusovskoye, Pechora-Kama interfluve, 61° N (from Stepanov, 1974, simplified). 1 – diamicton; 2 – gravel; 3 – sand; 4 laminated silt and clay; 5 – loess-like silt; 6 – bedrock; 7 – borehole. Broken line shows intersection of two perpendicular profiles. Till units are numbered upwards in the succession: g1, g2, g3 irrespectively of their possible age.

The lower inter-till formation (between g2 and g1) shows pollen spectra of mixed forests with very few pollen grains of *Carpinus, Corylus* and exotic trees. The high percentage of pyrite grains is typical for the oldest interglacial of the eastern Russian Plain (Stepanov, 1974). If the lower interglacial sequence, locally labeled the Solikamsk Formation, can be correlated with the Roslavl/Muchkap strata, then the underlying overconsolidated g1 till (the Kama Till of Stepanov) should be a counterpart of the Don till of southern Russia.

The alternative interpretation puts the Solikamsk interglacial into MIS 17-19 and the Kama Till into MIS 20 (Zubakov, 1992), which makes the Kama Till contemporaneous with the Mansi Till of West Siberia presently attributed to the Matuyama chron (Volkova & Babushkin, 2000). If the latter interpretation is correct, then the maximum glaciation of northeastern Russia is represented by g2 till of Stepanov`s profile (Fig. 8). The limit of this maximum glaciation shown in the left lower corner of Fig. 1 was for many years attributed to the Dnieper (early Saalian) glaciation (Krasnov, 1971; Ganeshin, 1973) which later proved to be wrong (Velichko et al., 1977; Shick, 1989).

The correlation between the Arctic region of marine transgressions and the Upper Pechora-Vychegda catchments remains problematic. A formal comparison says that the 3 main glacial complexes underlying the Upper Pleistocene of the Arctic (Fig. 2) probably correspond to the 3 tills of the Pechora-Kama interfluve (Fig. 8). However, g1, g2 and g3 in both profiles are just diamict units numbered by this author and cannot be viewed as synchronous with similarly designated tills in the Arctic profiles. Actually, it is not easy to find in the Arctic Pleistocene counterparts to all warm intervals of the southern record. Two boreal transgressions are more or less satisfactorily reflected in the warm Mikulino and Likhvin floras of the south, whereas there is no evidence to synchronise the cool Kolva marine strata with the Visherka or Solikamsk interglacial formations with their exotic plant remains. It is possible that the Kolva interglacial is missing in the terrestrial record of the Subarctic zone. This question will stay open until more reliable means of long-distance correlation are found.

The above brief analysis of the available stratigraphic data indicates that there are at least three readily recognisable glacial complexes and major pre-Weichselian ice advances in northeastern European Russia, which certainly does not preclude their subdivision into minor glacial stages. The fourth ice advance preceding the Kolva transgression in the Arctic might tentatively be correlated with the Kama Till of the Pechora-Volga intefluve.

Siberia

General stratigraphic situation

The Siberian glacial chronology is more arbitrary due to the formidable size of the country and poor applicability of pollen analysis. Unfortunately the dominant boreal forests have a monotonous composition with practically no broad-leaved trees presently growing, except sparse lime in southwestern West Siberia. Therefore, in distinguishing interglacials one has to rely on N-S shifts of very broad biogeographical zones, which would demand hundreds of meticulously studied sections. In literature one can usually see generalisations on a continental scale based on a handful of sites and, which is even worse, with European labels attached to totally unrelated objects, i.e. Atlantic terms applied to a non-Atlantic environment. This certainly hinders development of an independent Siberian Pleistocene stratigraphy.

Many geologists, exasperated by the difficulties encountered in Siberia, took to elaborating local stratigraphic scales by exclusively chronometric methods which normally should only be used for long-distance correlation. This poor practice, exemplified by the popular misuse of radiocarbon dating, led to corruption of the traditional stratigraphic nomenclature, which is not reliable anymore, being permanently reshuffled after each laboratory `discovery` (Astakhov, 2001). To lesser degree the same happens with the Middle Pleistocene, although the limited number of available chronometric methods helps to keep the old stratigraphic terminology in a better shape.

Two stratigraphic markers, identified in key sections of the West Siberian basin, have over several decades been the cornerstones of glacial history and Quaternary mapping. These are the interglacial formations of the Kazantsevo transgression in the Arctic and the Tobol alluvium in central and southern West Siberia. The Kazantsevo Formation consists mostly of shallow-water facies containing rich arctoboreal mollusk fauna with characteristic boreal species such as *Arctica islandica* indicating water temperatures 4 to 8°C above the present-day Kara Sea temperature (Sachs, 1953; Troitsky, 1975). This formation is conventionally correlated with the Eemian strata of the Boreal transgression in European Russia. This correlation seems to be confirmed in Yenissei Siberia by several ESR dates on marine shells in the range 108–134 ka (Sukhorukova, 1999). The main drawback of this marker formation is its being limited to the Arctic, where it is overlain by till and often badly glaciotectonised.

The marker significance of the Kazantsevo Formation is somewhat diluted by the fact, that, as in European Russia, there are distinct traces of another boreal transgression. First, boreal mollusk shells and foraminifers have been locally found in Middle Pleistocene diamictons (Arkhipov & Matveyeva, 1964; Kaplyanskaya & Tarnogradsky, 1975). Second, there are inter-till strata with rare boreal mollusks (Zubakov, 1972). Third, some Kazantsevo marine sequences contain shells of extinct bivalve species *Cyrtodaria jenisseae (angusta)* (Sachs, 1953; Kind & Leonov, 1982; Svendsen et al*.,* 2004) and probably belong to earlier interglacials.

The Tobol strata are much better preserved in the area of their classic occurrence along the transverse Ob and lower Irtysh, mostly between 62 and 56° N, i.e. just within the drift limit and in the periglacial area. They are washed, diagonally bedded quartz sands with thick lenses of green-bluish clayey silts, partly covered by till of the maximum glaciation. Literature on the Tobol Formation is extensive, especially regarding paleontological questions. A comprehensive overview of geological data is given in a collection of papers edited by Arkhipov (1975), and also in numerous works by palaeobotanists.

Pollen spectra, as always in Siberia, are not very characteristic. Sequences in the present taiga zone mainly show a predominance of arboreal *Betula* and rich herb assemblages with a minor coniferous component, which is common for the Siberian forest-steppe. Some successions reveal two coniferous peaks below and above the main forest-steppe phase. The interglacial nature of the formation is clear from abundant fresh-water mollusks, especially *Corbicula fluminalis (tibetensis)* presently living only in Central Asia. However, a rather cool and relatively humid climate is indicated by the rich macrofossil flora with exotic aquatic ferns *Azolla interglacialica, Selaginella selaginoides, Salvinia natans,* etc. This assemblage, known as the `Flora of the Diagonal Sands`, is similar to the Singil flora of the Russian Plain and has for several decades been the main indicator of the so-called `Siberian Mindel-Riss` and an argument for its correlation with the Likhvin and terrestrial Holsteinian. Later *Azolla* remains were also found in Upper Pleistocene sediments of the Lower Ob (Arkhipov et al., 1977).

The mammal fauna is the most controversial issue, probably due to the all-pervading redeposition of osteological material by the huge laterally migrating rivers. More frequent are finds of bones of the Tiraspol (Cromer) mammals. There is only one known site with forest fauna of the Singil type represented by *Palaeoloxodon, Megaloceros, Bison, Eolagurus, Arvicola*, etc. In many places the `Diagonal Sands` contain mammal species ranging from pre-Tiraspol to typical Late Pleistocene lemming faunas. The explanation is that the `Diagonal Sands` may be deposited at the same topographic level by slow meandering rivers during different interglacials. The Tobol Formation proper, according to Arkhipov (1975), chronologically ranges from the second half of Mindel to the Mindel-Riss interval, i.e. belongs to two interglacials.

This interpretation was supported by a wedge of glaciolacustrine varves (the Semeika Formation) found in the middle of the Tobol Formation on the Irtysh, where it is also overlain by till of the maximum Samarovo glaciation. The lower alluvial sequence, called the Talagaika Formation, is thought to represent a pre-Holsteinian interglacial (Kaplyanskaya & Tarnogradsky, 1974). In the present stratigraphic scheme only the *Corbicula* sands between the Samarovo Till and the Semeika varves are related to the Tobol climatostratigraphic `Horizon` *s.stricto* (Volkova & Babushkin, 2000).

Correlation of the Tobol `Horizon` (stage) with the Holsteinian is apparently supported by TL dates of 300 ± 75 and 313 ± 80 ka and ESR date 306 ± 20 ka on a *Corbicula* shell. However, in the periglacial zone *Corbicula* shells yielded ESR ages 285, 219 and 174 ka (Arkhipov, 1989). Later the palaeomagnetic excursion Biwa-II-Semeika along with the ESR date of 396 ka was reported from the basal part of the Tobol alluvium with *Corbicula* shells (Volkova & Babushkin, 2000).

Anyway, there are two independent interglacial markers, one of marine and another of terrestrial origin that can be seen in West Siberian exposures. The Arctic marker, if identified correctly, should provide the upper boundary of the Middle Pleistocene tills. The Subarctic marker separates surficial Middle Pleistocene tills from pre-Holsteinian glacial events known only from glacial deposits of buried valleys.

Stratigraphy in the Arctic

The till overlying the Kazantsevo strata in most cases are correlated with the Weichselian, whereas the underlying till is commonly related to MIS 6 (Svendsen et al., 2004). These ice advances are distinguished by their geographic distribution, because the post-Kazantsevo glaciation is limited to the Arctic, whereas the pre-Kazantsevo ice sheet advanced far beyond the Arctic Circle. The main problem is that in many cases either marine strata with a boreal fauna are older than the Kazantsevo *s. stricto*, or the Eemian interglacial is represented by non-marine facies. The first case is known from the Pupkovo section in the Yenissei valley (locality 9 in Fig. 1), where marine strata with interglacial pollen spectra and rare finds of a boreal fauna, including *Arctica islandica,* are sandwiched between two tills well beyond the limit of Late Pleistocene glaciation (Zubakov, 1972).

Troitsky (1975), who insisted on the Eemian age of this marine formation, had therefore to extend the Weichselian ice limit along the Yenissei as far south as 64° N, which is refuted by periglacial evidence (Astakhov et al., 1986). A similar situation is in the central part of the West Siberian basin close to the Arctic Circle, but well beyond the currently accepted Weichselian ice limit (Svendsen et al., 2004). In the Samburg borehole a thick marine formation with an arctoboreal fauna was found beneath a Middle Pleistocene till (Zubakov, 1972).

Non-marine Eemian is found on the Lower Ob, where two main stratigraphic concepts have been competing. The classical concept relates most of surficial diamictons and varved sequences to glacial or glaciomarine formations of the Middle Pleistocene (Lazukov, 1970, Zubakov, 1972). According to Zubakov, this interpretation is based on pollen spectra characteristic of southern taiga found in the sand incised into the thick varved rhythmite at the Pyak-Yaha section of the southern bank of the Ob in the present forest-tundra (Fig. 9). Only thin lenses of diamict and gravel materials which sometimes occur on the surface suggest a Late Pleistocene glaciation (Lazukov, 1970).

In contrast, the concept of the `young stratigraphy`, based mostly on sparse radiocarbon dates obtained in different sections, ascribes a Late Pleistocene age to all units of stratified and non-stratified drift observed in natural sections. In the latter scheme the Kazantsevo marker is represented by sand with boreal foraminifera identified in boreholes just below sea level (Arkhipov et al., 1977). This concept is currently accepted in the official stratigraphic scheme (Volkova & Babushkin, 2000).

Arkhipov et al. (1994), applying their radiocarbon-based `young stratigraphy` to the entire arctic Pleistocene, suggested a Holsteinian age for another interglacial formation at 100-200 m b.s.l., the so-called `Ob marine Strata`. This older interglacial formation, underlain by only one till, contains an assemblage of arctoboreal foraminifera of `*Miliolinella pyriformis* zone`. The thick diamictic sequence positioned between the two interglacial marine formations in this scheme belongs to MIS 6-8. Arkhipov et al. (1994) offered TL-dates on the sampled cores: 153 ± 15 ka for their `Kazantsevo Formation` and 246 ± 23, 306 ± 26, 366 ± 31, 370 ± 31 ka for the `Ob Strata`.

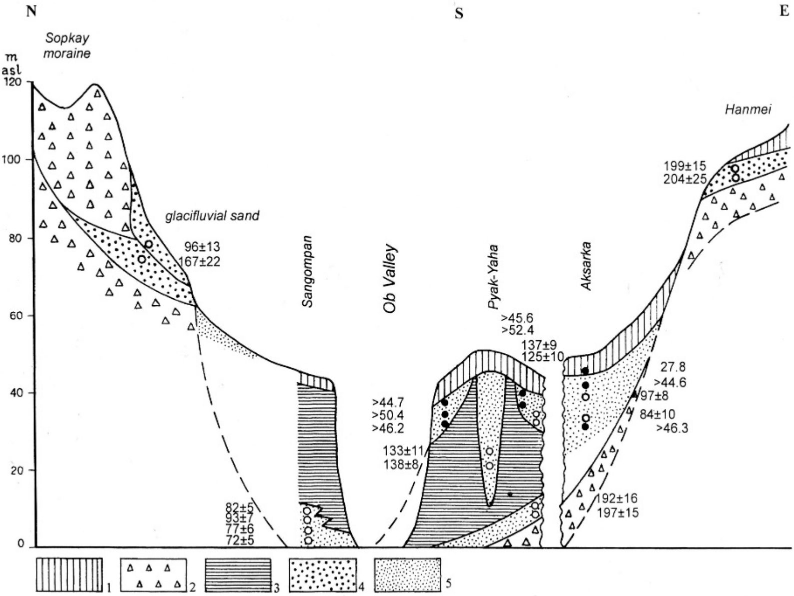

Fig. 9. Schematic profile across the River Ob valley on the Arctic Circle south of Early Weichselian Sopkay moraines (not to horizontal scale) according to Russian-Norwegian PECHORA project.
Symbols: 1 – loess-like silt; 2 – diamicton; 3 –silt/clay rhythmite changing into varves; 4 – glaciofluvial sand and fine gravel; 5 – laminated sand with peat lenses. Open dots are OSL, blackened dots are radiocarbon dates; age values are indicated in kiloyears BP. Note on southern bank non-finite radiocarbon dates and 4 OSL ages 125 to 138 ka from interglacial sand overlying rhythmite of mid-Weichselian age by Arkhipov et al. (1977).

The latest results from the Russian-Norwegian PECHORA project clearly refute the `young stratigraphy`. Weichselian tills and related glaciolacustrine sediments are found only on the northern bank of the river Ob (Fig. 9). The fluvial or deltaic sands with forest pollen spectra incised into the thick varved sequence of the southern bank yielded 4 OSL dates of 125 to 138 ka (samples 115, 116, 133 and 134, Pyak-Yaha, in Table 3), together with non-finite radiocarbon dates supporting an Eemian age of these sands. Glaciofluvial sands beneath the varved sequence, as well as on high interfluves of the Uralian piedmont, are OSL dated to c. 200 ka (samples 119 and 120 Pichuguy-Yaha in Table 3), which is consistent with a Middle Pleistocene age of the last ice advance on the Arctic Circle (Fig. 9).

The underlying Middle Pleistocene till and terrestrial sand with peat are stratigraphically above the Arkhipov`s `Kazantsevo` with TL-date of 153 ka, which means that the TL method underestimates the age, as compared to OSL dating. The better reliability of OSL dating has been independently confirmed by U/Th dating of the Seyda peat and by the above pollen data. Therefore, the sand with boreal foraminifera below sea level, Eemian by Arkhipov et al. (1977), is most likely Middle Pleistocene, probably Holsteinian.

Consequently, the deep lying Ob interglacial formation must be pre-Holsteinian, similar to the Kolva Formation of the Pechora Basin with the same foraminifer assemblage (Gudina, 1976). This implies that the thickest (more than 100 m) till sequence, sandwiched between two interglacial marine formations, is not Saalian but older. Another important implication is that, accepting the traditional correlation of the Ob Strata with the Tobol alluvium, the latter must be pre-Holsteinian. However, it is possible that the Tobol alluvium corresponds in the Arctic to the upper marine formation with boreal foraminifers lying just b.s.l., i.e. to the `Kazantsevo strata` by Arkhipov et al. (1977, 1994). In this case the lower marine formation, the Ob Strata, may correlate with the Talagaika or even older interglacial deposits of the south (see below).

Middle Pleistocene tills of the eastern glaciated Arctic are poorly studied. The thick fine-grained diamictons directly underlying the Kazantsevo Formation with boreal fauna were related to the marine or glaciomarine Sanchugovka Formation with a sparse, predominantly arctic fauna (Sachs, 1953; Lazukov, 1970; Zubakov, 1972) until Kaplyanskaya and Tarnogradsky (1975) thoroughly investigated the type section and found that the formation consisted of basal tills with rafts of marine sediments. The diamictons contain not only an arctic fauna, but also shells of boreal and Cretaceous mollusks. The Sanchugovka Till is believed to represent MIS 6, although Arkhipov (1989) thinks that below sea level there are real marine strata with arctic foraminifera of MIS 7 underlain by tills of MIS 8.

One of the rare sections in which two interglacial marine formation are separated by the Murukta Till can be seen is Novorybnoye at the mouth of Khatanga river (10 in Fig. 1). The upper marine unit, which is not covered by till, is commonly interpreted as a deposit of the Eemian transgression, whereas the lower marine formation, containing foraminifera of the *Miliolinella pyriformis* zone, and in another section farther to the west also *Cyrtodaria angusta,* is thought to represent the Siberian equivalent of the

Holsteinian. The alternative stratigraphic model suggests an Eemian age for the lower marine unit and a mid-Weichselian age for the upper marine formation (Kind & Leonov, 1982). The latter interpretation being paleontologically questionable, is also at odds with the latest results of the QUEEN program which refute a Late Pleistocene glaciation that far in the east (Svendsen et al., 2004). The traditional interpretation of this sequence in which the earlier transgression is thought to be Holsteinian (e.g. Gudina, 1976) suggests a Saalian age for the Murukta Till, which agrees well with its widespread occurrence in Central Siberia.

Stratigraphy in the Subarctic

Best identifiable in this zone is the Samarovo Till of the maximum glaciation named after the settlement close to the Irtysh mouth, where the famous thick sequence of stacked tills and Paleogene opoka rafts has long been known (Shatsky, 1965). The additional diamicton unit on top of the Samarovo Till north of the Irtysh mouth is thought to represent the penultimate Taz glaciation. The Samarovo Till, resting on the Tobol alluvium, is getting thinner upstream and disappears south of the Semeika village (5 in Fig.1). North of this point a couple of additional tills beneath the Tobol Formation are known from boreholes. The lowermost interglacial formation (the Talagaika alluvium) north of 63° N is overlain by a double diamict formation up to 70 m thick called the Shaitan Till which is thought to correspond to the Semeika glaciolacustrine clay on the Irtysh (Volkova & Babushkin, 2000). The lowermost till (the Mansi Till) is found beneath the Talagaika interglacial at the bottom of a drill well north of the drift limit on the Irtysh (Arkhipov, 1989).

There are two main problems with the till count in this zone. The first is connected with the Taz glaciation which is thought to predate the Eemian and deposit a till of limited distribution on top of the Samarovo glacial complex. Originally the Taz Till was mapped in the upper reaches of river Taz, where it is separated from the underlying Samarovo strata by sands with ambiguous paleontological characteristics. Later such a till was distinguished all over West Siberia. Although no interglacial formation with abundant organics has ever been found between the Samarovo and Taz tills, the alleged Shirta interglacial, called `interstadial` by more cautious geologists, persists in the regional stratigraphic schemes (Volkova & Babushkin, 2000). Interglacial organics are in general rare in Siberian inter-till formations. This led Lazukov (1970) to suggest that all till sheets along the Ob river are deposited by the maximum Samarovo glaciation, whereas Arkhipov (1989, Arkhipov et al., 1978) tried to subdivide this sequence by means of thermoluminescence dating (A in Fig. 10).

According to Arkhipov the upper sand and silt in Kormuzhikhanka section on the Belogorye Upland (locality 6 in Fig. 1) is Eemian, and the overlying soft diamicton is therefore an Early Weichselian till (Fig. 10, A). The till under the Eemian TL values belongs to the Taz glaciation of MIS 6, and the till next downwards – to the Samarovo glaciation of MIS 8. The lowermost till in this sequence is related to the Late Shaitan glaciation of MIS 10-12 (Arkhipov, 1989). This correlation does not contradict the latest PECHORA results in the Arctic Ob area.

However, taking into account the too young TL dates in the Arctic zone, this lowermost till might be even older. The uppermost Kormuzhikhanka diamicton is soft, mantle-like, low on pebbles and without structures of ice flow or shear planes at the base. It is neither traceable regionally, nor associated with other glacigenic sediments. Therefore, it should be better viewed as a solifluction bed, or a flowtill derived from residual Middle Pleistocene ice, but not as a signature of a Late Pleistocene ice advance (Zolnikov, 1990). The latter interpretation fits the PECHORA data indicating no Late Pleistocene glaciation on the Ob along the Arctic Circle (Fig. 9).

Another problem is the correlation of the Ob-Irtysh ice advances with the Central Siberian events. As was pointed out, south of the Late Pleistocene ice limit on the Yenissei there are two tills separated by a marine formation with rare boreal shells (locality 9 in Fig. 1). Zubakov (1972) relates the upper till to the last Middle Pleistocene glacial event called the Yenissei glaciation, MIS 6. This event is supposed to be identical with the Taz glaciation of West Siberia. Isayeva (1963) acknowledged this till as a counterpart of her second (Nizhnyaya Tunguska) belt of end moraines running around the Putorana Plateau between the drift limit and the Late Pleistocene Onyoka moraines. The Nizhnyaya Tunguska moraines in the northeast merge with the Murukta moraines, which in the stratigraphic scheme of Central Siberia are positioned above the Eemian level (Isayeva et al., 1986). This seems to be a miscorrelation because, according to the QUEEN results, the Early Weichselian maximum is represented by the youngest belt of the Onyoka moraines *sensu* Isayeva (1963) (see the Taimyr discussion in Svendsen et al., 2004). Therefore, both Yenissei and Murukta tills should belong to MIS 6.

The till underlying the Pupkovo marine strata (locality 9 in Fig. 1) is traditionally correlated with the Samarovo glaciation of MIS 8, which is thought to have reached the drift limit also on the Yenissei (Zubakov, 1972). However, it is more likely that the earlier boreal transgression, as on the Pechora and Ob, relates to the Holsteinian or MIS 11, thus making the lower till in the Yenissei bluffs Elsterian in age. The official stratigraphic scheme keeps all these strata in the Bakhta `Superhorizon` (Volkova & Babushkin, 2000) exemplified by the Bakhtinsky Yar section (B in Fig. 10, loc. 8 in Fig. 1).

There are no indications of warm paleoclimates in the inter-till sediments traditionally correlated with the `Shirta interglacial` (or interstadial) (Arkhipov & Matveyeva, 1964). This sequence is a rare case of the lower till overlying interglacial alluvium with a bone of *Alces latifrons.* The latter is characteristic of the Tiraspol (Cromer) mammalian complex (Arkhipov & Matveyeva, 1964; Arkhipov, 1975), which implies that the lower till in the Yenissei sections might well be Elsterian. Upstream of Bakhtinsky Yar a thicker till formation was found below the Bakhta Strata in a borehole penetrating down to -342 m. This so-called Lebed Till is the lowermost member of the official stratigraphic scheme of Central Siberia (Isayeva et al., 1986).

Another key section is Khakhalevsky Yar close to the drift limit (C in Fig. 10, loc. 7 in Fig. 1). It represents the Turukhan alluvium (Arkhipov & Matveyeva, 1964), or Panteleyeva Formation (Zubakov, 1972), sandwiched between two tills. The coarse channel alluvium overlying the lower till grades upwards into floodplain silts with gyttja and peat lenses, changing farther upwards into proglacial varves distorted by the maximum ice advance. The succession is capped by till and glaciofluvial sand. Pollen spectra of the channel alluvium show an upward increase of pollen of coniferous forests with predominance of *Picea* and *Pinus sibirica.* In the floodplain silts they are gradually replaced by *Betula* dominated parklands with abundant herbs, *Ericaceae* and *Selaginella selaginoides* (Levina, 1964).

By these and other similar spectra the Turukhan alluvium is correlated with the Tobol alluvium of the Ob and Irtysh (Arkhipov, 1975). However, pollen spectra of the preceding Talagaika interglacial are not much different. Therefore, it cannot be excluded that the maximum glaciation on the Yenissei correlates not with MIS 8, as suggested by Arkhipov for the Samarovo glaciation, but with the preceding cold stages MIS 10 or 12.

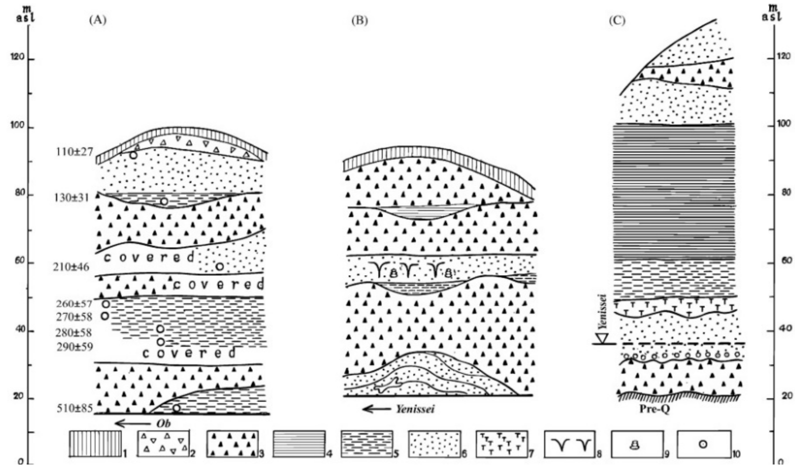

Fig. 10. Key sections of Middle Pleistocene tills in West Siberia.
A – Kormuzhikhanka section at locality 6 in Fig. 1 (from Arkhipov, 1989; Arkhipov et al., 1978, simplified), B – Bakhtinsky Yar section at locality 8 in Fig. 1 (Arkhipov & Matveyeva, 1964), C – Khakhalevsky Yar section supplemented by a borehole at locality 7 in Fig. 1 (Levina, 1964).

The latter option is even more likely for the Central Siberian Upland, where the poorly preserved tills of the maximum glaciation are in a stark contrast with the inner belt of topographically expressive Nizhnyaya Tunguska moraines (Isayeva, 1963). There are authors who correlate the maximum glaciation limit of the eastern Central Siberia with the thick sequence of diamictons and stratified drift found by boreholes at -240 m and lower on the floor of the overdeepened Yenissei valley at Lebed (locality 14 in Fig. 1). By its position under the sedimentary complex of the maximum Yenissei glaciation the Lebed double till sequence is similar to the Shaitan tills on the Ob (Volkova & Babushkin, 2000). However, the Lebed tills might even be pre-Elsterian in age.

How many ice advances?

The most pressing stratigraphic problem concerns the age of the two surficial tills – the Taz and Samarovo – overlying the Tobol horizon. They are traditionally related to MIS 6 and 8 correspondingly (Table 3). The alternative that both belong to MIS 6 is unlikely, because glacial landforms are not known atop of the Samarovo Till covered by loess-like silts up to 15 thick. If the Tobol Horizon proves to be pre-Holsteinian, and real interglacial organics are found between the Taz and Samarovo tills, then the Taz Till may relate to the Saalian (MIS 6) and the Samarovo Till to the Elsterian.

The latter option was considered by Arkhipov based on TL and ESR datings. He concluded that the Tobol Formation *s. stricto* might be pre-Holsteinian and therefore the Samarovo glaciation might be Elsterian. However, this alternative would upset the traditional correlation pattern and therefore not desirable for the time being (Arkhipov, 1989). Anyway, the great thickness of the tills underlying the penultimate interglacial in all major buried valleys suggests that Elsterian or earlier ice sheets of West Siberia were very large. It is quite possible that several ice advances are still hidden in the 70-100 m thick diamict sequences of the buried valleys, where no good interglacial formations have been distinguished so far. The latest results of deep coring in arctic West Siberia show that assemblages of arctoboreal foraminifera (previously perceived as the Holsteinian `zone of *Miliolinella pyriformis`)* occur at three different stratigraphic levels (Volkova & Babushkin, 2000). This makes correlation of the arctic tills with the terrestrial record in the south even more difficult.

The above data allow to infer four major pre-Eemian ice advances in the Siberian record (Table 2). The oldest Mansi Till reflects a more restricted ice sheet. In the present official stratigraphic scheme of West Siberia (Volkova & Babushkin, 2000) this ice advance is related to the Matuyama chron of reverse polarity, i.e. is thought to be older than 780 ka, although Arkhipov (1989) preferred to place it at MIS 18. The thickest Shaitan tills separated by stratified drift with arctic foraminifera are certainly pre-Holsteinian but within the Bruhnes chron, their more precise correlation with the European record being premature. The correlation with MIS 12-16 suggested by Arkhipov (Table 3) is only based on very old and therefore hardly reliable TL dates. The maximum glaciation (MIS 8, or 10, or 12) probably produced the thickest (up to 3.5 km) ice sheet that grew over the Kara Sea shelf and eventually overrode nearby mountain ranges up to an altitude of 1 km. The last Middle Pleistocene ice sheet called the Taz (also the Yenissei or the Murukta), judging by the fresh glacial landscapes, was formed in MIS 6. It was thick enough to cover almost the same area as the maximum glaciation and flow southwards over the low mountains.

Table 3. Pre-Weichselian glacial chronology of Siberia (after Arkhipov, 1989).

Glaciations/ *Interglaciations*	***Arctic zone*** Ages of marine formations, ka	***Subarctic zone*** Ages of terrestrial formations, ka	MIS
Kazantsevo	ESR 105 ± 10.5; 120 ± 13; 121.9; 134.8	TL 130 ± 25	***5e***
Taz			**6**
Shirta	ESR 170 ± 10	ESR 196.8 ± 20.6 TL 180 ± 40; 190 ± 36	*7*
Samarovo			**8**
Tobol	ESR 306 ± 21	ESR 326.9; 396 TL 260 ± 56 to 390 ± 80	**9-11**
Late Shaitan Early Shaitan		TL 510 ± 65; 550(561) ± 110(140)	**12-16**
Talagaika		TL 660 ± 180; 740 ± 170	***17***
Mansi			**18**

Conclusions

1) The huge ice sheets, much larger than the Weichselian ones, at least 4 times covered the Russian mainland and adjacent shelves prior to the Eemian.
2) The main centre of ice accumulation was the Kara Sea shelf, additional sources of inland ice being Fennoscandia and the Barents Sea shelf in European Russia, and the Putorana Plateau in Siberia. The Ural Mountains played mostly the passive role of an orographic barrier in the ice flow pattern.
3) The extent and geological work of the coalesced ice sheets imply an Antarctic type of glaciation with ice thickness up to 3.5 km. Unlike Antarctic and most of North Atlantic ice sheets, North Russian continental glaciers acted on predominantly soft and perennially frozen substrate, which was deeply affected by pervading glaciotectonism.

4) The thickest tills are found beneath interglacial formations similar to the Holsteinian, suggesting that the most extensive ice sheets are pre-Saalian not only in European Russia, but probably also in Siberia.

5) During the maximum (Cromerian) glaciation and subsequent ice ages (MIS 16 to 10) the influence of the Fennoscandian ice dome was limited in northern European Russia. The Fennoscandian ice sheet, however, culminated during the penultimate glaciation (MIS 6), when shelf sources of inland ice were less powerful than in pre-Holsteinian times.

6) The drift limit in Northern Russia is time-transgressive, being certainly pre-Holsteinian in northeastern Europe, probably MIS 8 in West Siberia and getting older east of the Yenissei.

7) There are several unsolved questions concerning correlation of the marine transgressions with terrestrial interglacial events. The reliability of the correlation of Russian pre-Eemian interglacials with their western European counterparts noticeably decreases northwards and eastwards partly because of the insufficient data available, and also due to the fading biotic signal of the climatic fluctuations.

Acknowledgements

This work is part of the current Russian-Norwegian PECHORA project supported mostly by the Norwegian Research Council. It is also a contribution to the European Science Foundation program QUEEN. OSL dating was performed by A. Murray at the Nordic Laboratory for Luminescence Dating, Risø, Denmark. Radiocarbon and U/Th dates have been obtained in the Geochronological Laboratory of St. Petersburg University, Russia, under the guidance of Kh. Arslanov. Editing efforts by J. Ehlers, A. Raukas and an anonymous referee have considerably improved the text. The author offers his sincere thanks to the mentioned persons and institutions.

4. PLEISTOCENE ICE LIMITS IN RUSSIAN NORTHERN LOWLANDS

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QUATERNARY GLACIATIONS – EXTENT AND CHRONOLOGY. PART 1: EUROPE. EHLERS J. AND GIBBARD P.L., EDS. ELSEVIER: 2004, P. 309–319.

Introduction

This paper is an explanatory note to the digitized map of ice limits compiled by the author for the Russian North, beyond the limits of the Fennoscandian glaciation. The glacial limits presented here are cartographically based on data from the Russian Geological Survey obtained during general geological mapping at standard scales of 1:200,000 and 1:1,000,000. It also incorporates results of special studies in critical areas performed by various researchers, the author included. The outermost limit of Pleistocene glaciation is mostly derived from standard geological maps and drilling projects, whereas the Late Pleistocene ice limits are largely products of stratigraphic and photogeological studies during the last two decades. The result is presented in simplified form in Fig. 1.

Although the principal ice limits in Central Russia were already established at the beginning of geological mapping by the Emperor`s Geological Committee in the late 19th century, the size of Pleistocene glaciers in the north and east, excluding the Fennoscandian ice dome, was poorly known until after extensive surveys of these remote areas in Soviet times. Indeed many sheets of the National Geological Map of the USSR appeared in the 1940-50s. This information was summarised in synthetic maps of Quaternary deposits at scales 1:2,500,000 and 1:1,500,000 (Yakovlev, 1950; Zarrina et al., 1961) derived from the first generation geological maps. The second generation of geological maps, that included complementary Quaternary maps, began to appear in the 1960s. The new survey and drilling data were incorporated into more accurate small-scale maps of the Quaternary deposits of European Russia and the entire Soviet

Union (Krasnov, 1971; Ganeshin, 1976). Many maps of the second generation were also published in the 1980-90s, albeit at a slower pace. The most inaccessible areas of Central Siberia did not have published Quaternary maps by the end of the 20th century, when the Russian Geological Survey had already commenced compiling the third generation maps in digital form.

The maximum Pleistocene ice limit, related to pre-Eemian glaciations, has only undergone minor changes during the last decades. However, younger ice limits have been under incessant stratigraphic discussion. The number of known ice limits in each given region depend very much both on the logistical accessibility and geographical peculiarities of the area. Several ice limits mapped in European Russia are hardly traceable beyond the Urals, because photogeology, the main tool in the north, does not produce good results in swampy lowlands. Also, topographically expressive ice marginal zones of the Central Siberian ice sheets cannot be directly connected with limits of the Barents-Kara ice sheets in the flat West Siberian Plain. Several morphological boundaries discussed in academic works (e.g., Arkhipov et al., 1980, 1986; Grosswald, 1980, 1993) are not used in the present map, if they are not supported by stratigraphic evidence. Three main sets of data are presented in the digital map:

1). Ice limits, subdivision of which into certain and uncertain ones is not exactly the same for different glaciations. In the case of Middle Pleistocene glaciations, a certain ice limit in the lowlands is a line distal to the southernmost sections that show glaciotectonised and till-covered interglacial formations. In the uplands of Central Siberia this line is drawn between the last mapped ridge-like diamicton accumulations and heavily dissected fluvial landscape to the south. This line is considered uncertain where no distinct morainic features have been mapped, and only diamicton facies in rare boreholes or erratics on the surface suggest the glacial drift limit. In the case of the Early Weichselian glaciation, the ice limit is shown as certain where it can be drawn between sections of interglacial sediments of Eemian type (mostly marine with warm-water mollusk fauna), overlain by till, on the one hand, and not covered by till, on the other.

An uncertain Early Weichselian ice limit is a line interpolated between known stratigraphic sites using such signatures of former glaciation as occurrence of buried bodies of stratiform massive ice in the lowlands of West Siberia and/or assemblages of ice-pushed ridges in higher terrains. In many cases marginal features are very spectacular, such as the double Laya-Adzva ridge in the Pechora Basin, which can be seen even from the

Moon. Nevertheless the lack of Eemian sites makes this author designate the ice limit there as uncertain. Similarly, the limit of Late Weichselian ice sheet in Central Siberia is drawn as certain where sequences of Weichselian sediments with finite radiocarbon dates have been reported as overlain by till. Where no finite dates are known from sediments beneath till, it is shown as uncertain, even if the limit is clearly expressed by very fresh end moraines. All these distinctions are omitted in the simplified overview map (Fig. 1).

2). Topographically expressive features of glaciotectonic accretion are shown in the digital map as red arcs the radii of which can be taken as representing former flow lines.

There are two kinds of these features dependent on the quality of the substrate. Where substrate is solid bedrock that allowed an easy sliding, these horseshoe-shaped ridges may be positioned close to the outer till limit and thereby considered as end moraines (Fig. 2, a). The internal structure of terminal ridges is poorly known and is normally described as chaotic agglomeration of stony or clayey diamictons distorted together with underlying sand and sometimes massive ice (e.g., Kind & Leonov, 1982; Astakhov et al., 1999). The ridges of sedimentary basins are generally larger and predominantly occur far upglacier irrespective of any marginal formations. In all the sections they invariably demonstrate a regular imbricate structure, with folded slices of various soft rocks divided by listric thrusts and sometimes by clay or ice diapirs (paper 2, Ch. I and paper 8, Ch. II). They are often expressed in the landscape as typical `hill-hole pairs` (Fig. 2, b).

In the Middle Pleistocene glaciation area, the imbricate assemblages are mostly built of Paleogene and Cretaceous sediments, such as the largest arc between the Urals and the Ob river, the `Malososvinsky Amphitheatre`. The depth of these glacial disturbances may reach 400 m and their width up to 20-25 km. The morainic arcs on the map only show crests of such large structures. These forms, derived from glacial crumpling of perennially frozen clayey formations, must have originated under very thick ice (paper 8, Ch. II) and therefore cannot be called end moraines. They provide evidence of glacial overriding but not of a quasi-stationary ice margin. In the northern plains where the ice front was fringed by a large proglacial lake and, where glacier motion was obstructed neither orographically nor thermally, extension ice flow failed to produce any terminal elevations.

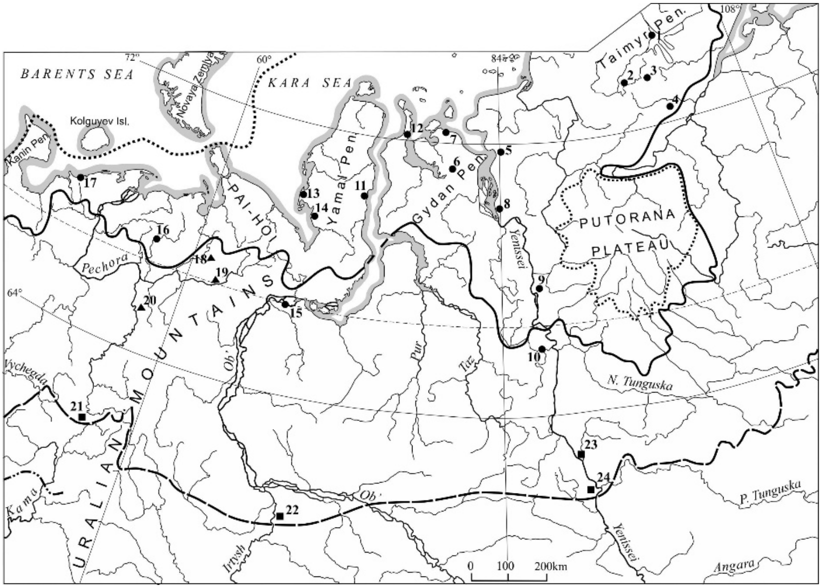


Fig.1. Major Pleistocene ice limits of northern Russia. Dotted line – Late Weichselian, solid line – Early Weichselian, dashed line – Saalian, dash-and-dot line – Don, MIS 16. Detailed maps are at http://booksite.elsevier.com/9780444534477/ *Black circles* are major sites with dated Middle and Late Weichselian sediments not covered by till, with their ages: 1 – Cape Sabler, 39 to 17 ka (Kind & Leonov, 1982; Möller et al., 1999); 2 – River Logata, 45 to 28 ka, 3 – Lake Logata, 45 to 25 ka (Fisher et al., 1990); 4 – Lake Labaz, > 48 to 20 ka (Siegert et al., 1999); 5 – Mokhovaya mammoth, 35.8±2.7 ka , 6 – Gyda mammoth, 33.5±1 ka (Heintz & Garutt, 1964); 7 – Leskino mammoth and plants, 30.1±0.3 and 29.7±0.3 (paper 10, Ch. III); 8 – Cape Karginsky, 15.3±0.2 (Kind, 1974); 9 – Igarka permafrost pit, >50 to 35 ka (Kind, 1974); 10 – Farkovo, 42 to 34 ka (Kind, 1974); 11 – Syo-Yaha, 40 to 17 ka (Vasilchuk et al., 2000); 12 – Mongotalyang, 31 to 21 ka (Vasilchuk et al., 1984); 13 – Marresale, 28 to 26 ka (Forman et al., 2002); 14 – Mutny Mys, 30.5 to 27.5 ka (Gataullin et al., 1998); 15 – Salehard moraines covered by interglacial sediments (paper 9, Ch. II); 16 – bones in river Kolva terraces, 37 to 26 ka , 17 – Timan Beach, 52 to 13 ka (Mangerud et al., 1999). *Black triangles* are surficial Paleolithic sites from Mangerud et al., 1999: 18 – Pymva-Shor, 26 to 10 ka, 19 – Mamontovaya Kurya, 37 to 24 ka, 20 – Byzovaya, 33 to 25 ka. *Black quadrangles* are interglacial lacustrine and fluvial sediments of Holsteinian type covered by till: 21 – Lake Chusovskoye (Stepanov, 1974); 22 – Semeika (Kaplyanskaya & Tarnogradsky, 1974); 23 – Bakhtinsky Yar (Zubakov, 1972); 24 – Khakhalevsky Yar (Zubakov, 1972). See also the detailed List of key sites in the end.

3). Named (or numbered) sedimentary sequences (Table 1), constraining temporal and spatial brackets for former ice sheets, are shown mostly for the Late Pleistocene glaciation area, where geochronometric methods can be employed. They are of four kinds: i) interglacial formations of Eemian type which were apparently not overridden by glaciers, ii) glacially disturbed or till-covered Eemian sediments, indicating a Late Pleistocene ice advance, iii) Weichselian sediments with successions of `old` (40 to 15 ka BP) radiocarbon dates, lacking any sign of overriding by a Late Weichselian ice advance, and iv) Weichselian sediments, with finite radiocarbon dates, covered by a till.

The most important and well-dated sections are shown in Fig. 1, others can be found in the digital map. For the Middle Pleistocene glaciation area with till-covered interglacial formations of Likhvin/Tobol (i.e. Holsteinian) type only a few famous sites have been selected. These formations are known for their characteristically forest pollen spectra that include exotic taxa, indicating warmer and more humid environments than at present.

The West European terms Early, Middle and Late Weichselian are rarely applied in the Russian North for specific geological objects, the local names being more popular. However, because of the size of the area considered, with many semi-formalised stratigraphic labels, the author has used these terms below in their geochronological sense common in Russian literature, i.e. as equivalents of the subdivisions of the Wisconsinan.

Pleistocene Glacial Maximum

The spatial resolution of pre-Eemian ice limits, presented here, is mainly in accord with the general maps of Quaternary deposits (Krasnov, 1971; Ganeshin, 1976; Zarrina et al., 1961). However, in many places the configuration of ice margins is corrected using information from later geological maps (Bobkova, 1985; Chumakov et al., 1999; Potapenko, 1998; Rudenko et al., 1981, 1984) and several specialised studies, including Stepanov (1974) for the Pechora-Volga interfluve (21 in Fig. 1), Kaplyanskaya and Tarnogradsky (1974) for river Irtysh (22 in Fig. 1), Astakhov and Fainer (1979) and Zubakov (1972) for Yenissei Siberia (23, 24 in Fig, 1), Arkhipov et al. (1976) and Isayeva (1984) for Central Siberia.

The maximum glaciation was traditionally mapped in the USSR as a counterpart of an early Saalian ice advance in Western Europe, which was stratigraphically traced eastwards across Poland and the Ukraine (Yakovlev, 1956). In the 1980s it was discovered that the till of the southernmost lobe of the Middle Pleistocene glaciation in Central Russia, unlike in the Ukraine, was overlain by interglacial sediments with fauna of the Tiraspol (Cromer) Complex and by 4-5 paleosols of interglacial character. Accordingly, this so-called Don glaciation, presumably c. 0.5-0.6 ma old, could no longer be correlated with either the Saalian in Western Europe, the Dnieper glaciation in the Ukraine, or with the Samarovo glaciation of Siberia (Shik, 1995; Velichko, 1991). In new 1:1,000,000 maps the Pleistocene drift limit is related to the Don glaciation (Marine Isotope Stage 16 or 14 by different estimates), even in the very east of the Russian Plain (Chumakov et al., 1999) (Fig. 1).

This age of the maximum glaciation is supported by the coring results around lake Chusovskoye on the Pechora-Volga interfluve (21 in Fig. 1), where two or three more tills occur beneath the superficial pre-Eemian till. Interglacial lacustrine sediments, with pollen of Likhvin (Holsteinian) type, have been found between the two upper tills of this area. The unique find of forest elephant *Palaeoloxodon sp*. is also related to this interglacial sequence (Stepanov, 1974). The uppermost till was previously related to the maximum glaciation of the East European Plain, presumably Saalian (Krasnov, 1971), which is currently split into two different stages – the Dnieper (Saalian) and the Don (Cromerian) (Shik, 1995). It is likely that the new Pleistocene maximum (the Don ice advance) is represented by one of the pre-Holsteinian tills in the lake Chusovskoye area (paper 3, Ch.1).

The Samarovo glacial maximum of West Siberia is still considered to be of Saalian age. This age is inferred from the underlying Tobol interglacial alluvium (Arkhipov, 1975), which contains forest fauna, shells of Central Asia fresh-water mollusk *Corbicula tibetensis (fluminalis),* shows normal magnetic polarity and yields thermoluminescence dates in the range 260 to 390 ka BP. A bone of the elk *Alces latifrons* has been found in sediments beneath the till at Bakhtinsky Yar on the Yenissei (23 in Fig. 1). Overlying the Samarovo Till, only one horizon of interglacial soils and peats has been described. Beneath this till two pre-Holsteinian tills can be distinguished in the sedimentary sequences filling buried valleys, overdeepened to 200-300 b.s.l. (Zubakov, 1972; Arkhipov, 1989). These ancient tills have been found proximally to the Samarovo glaciation limit, rather close to it. A pre-Holsteinian glaciation may be the most extensive in the east of Central Siberia, as suggested by Bobkova (1985).

The lingering stratigraphic uncertainty is connected with the relationship between the Moscow Till of Central Russia and Dnieper Till in the Ukraine. Both are currently related to the Marine Isotope Stage 6 by Russian geologists (Shik, 1995), but in the Ukraine the Dnieper Till is commonly thought as belonging to an earlier Saalian substage, as suggested by several thermoluminescence dates c. 280 ka (Gozhik, 1995).

A similar problem is discussed in Siberia where the less extensive Taz Till is thought to represent MIS 6, whereas the maximum Samarovo glaciation is related to MIS 8, based on a few thermoluminescence dates (Arkhipov, 1989). Traditionally the Samarovo stage was correlated with the Dnieper glaciation. Although, unlike in European Russia, both pre-Eemian tills are known in superposition, the independence of the Taz glaciation is questionable, because no unequivocal interglacial formations have ever been described from between this and underlying Samarovo glacial complex. In the present digital map the boundary of the Taz glaciation, commonly drawn along hummocky plateaus and morainic ridges of unknown age (e.g., Arkhipov et al., 1976, 1986), is omitted as unreliable.

Today the only clue for the spatial differentiation between Saalian and pre-Saalian ice sheets is glacial topography which is partly preserved in the Taz area and totally absent in older glacial landscapes. However, this feeble tool can hardly be employed in the primordially flat lowlands of central West Siberia, where only rare drilling profiles, undertaken for mapping and geotechnical projects, help to interpolate the drift limit between key sections in river valleys. Immediately west of the Urals, the Saalian ice limits are controversial in different survey maps. Thus the author has assigned the Saalian limit in the Volga-Pechora interfluve using available literature and personal experience in the Pechora Basin. In the extreme northeast of Central Siberia the pre-Eemian ice limits cannot so far be suggested for want of any reliable data.

Late Pleistocene ice limits

The first cartographic concepts of the last glaciation of northern Russia without the area of the Scandinavian glaciers were put forward by Sachs (1953) and Yakovlev (1956), who authored fundamental syntheses of previous investigations. They established the pattern after which many subsequent models have been developed. It was proven beyond any doubt that Late Pleistocene glaciers were much smaller than pre-Eemian ice sheets. On the basis of mapped boulder trains and occurrence of fresh-looking hummock-and-lake landscapes the latest ice sheets of the West Siberian and Pechora Lowlands were thought to have been confined to the Arctic and fed from upland ice domes of the Urals, Novaya Zemlya, Putorana Plateau of Central Siberia and Byrranga Mountains in the Taimyr Peninsula. Yakovlev (1956) also suggested an additional ice dome on the dry shelf of the Barents Sea.

The Late Pleistocene age of the uppermost till was inferred from the underlying interglacial marine formation with warm-water mollusk fauna considered to be a product of the Eemian transgression. These marine sediments, called the Boreal Strata in Europe and Kazantsevo Formation in Siberia, were described in many places as glacially disturbed and overlain by the uppermost till. The latter was correlated with the Early Weichselian (Sachs, 1953; Yakovlev, 1956). The limits of this glaciation were drawn around main Paleozoic uplands, but only in the west was the marginal zone found to parallel the Barents Sea coast. Later, Lavrov (1974) extended the limit of the last Barents Sea ice sheet farther south and east, maintaining its coalescence with the Uralian ice sheet.

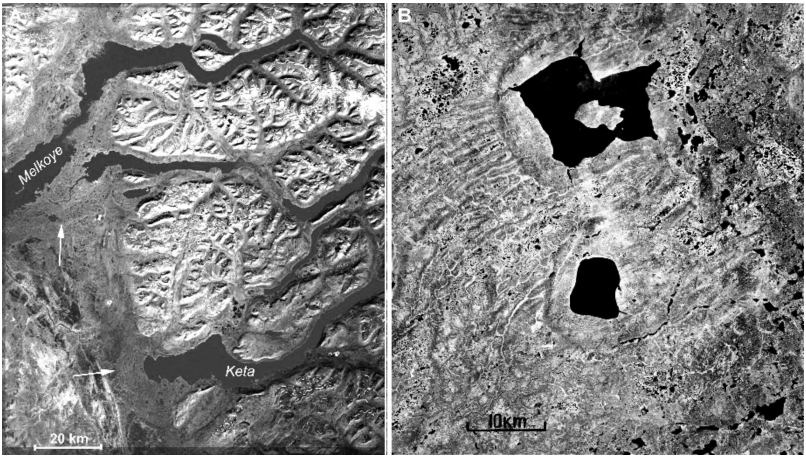


Fig. 2. Ice-pushed ridges in the Yenissei catchment (satellite images).
A. Horseshoe-shaped Norilsk end moraines (marked by arrows) bounding deep fjord-like lakes (Melkoye and Keta) in the western foothills of the Putorana Plateau. B. Two hill-hole pairs in the sedimentary basin west of the Yenissei, inside the Late Pleistocene glaciated area, 60 km upglacier from the western margin the Early Weichselian ice sheet; the northern glaciotectonic ridge built of stony diamicton and cobbly sand is 130 m high above lake Makovskoye which is 65 m deep in the western part (Arkhipov et al. 1976).

The wide use of remote sensing data in the 1970-s revealed many push moraines, the configuration of which totally contradicted the mountain glaciation hypothesis and testified to upslope ice flow from coastal lowlands towards Urals and Taimyr highlands, a direction supported by till composition (papers 1 and 2, Ch, I; Andreyeva, 1978). The mapped pattern of ice-pushed ridges could only form if the thickest inland ice resided on the low coastlands and totally blocked the northbound drainage. Beyond the Kara ice sheet only traces of small alpine glaciers were mapped in the Urals (Astakhov, 1979; Arkhipov et al., 1980). At that time, very large morainic loops overlying interglacial marine sediments were also mapped in the southern part of the Taimyr Peninsula. These morainic assemblages together with boulder trains pointed to a former ice divide north of the Byrranga Mountains, i.e. in the northeastern Kara Sea (Andreyeva, 1978; Grosswald, 1980; Kind & Leonov, 1982; Isayeva, 1984). When ice domes on the Kara Sea shelf were demonstrated to have existed, the paleogeographic paradigm changed, and totally different ideas of Weichselian ice limits in the Arctic began to be discussed.

At the same time the advent of many finite radiocarbon dates tipped the scale of the discussion towards concepts of a very extensive last glaciation. The last ice sheet was viewed as a counterpart of the well-studied Late Wisconsinan Laurentide and Late Weichselian Fennoscandian glacial systems. Grosswald, the most ardent proponent of the maximalist model, maintained that Weichselian ice sheets culminated in northern Russia ~ 20 ka ago and extended southwards very far from the Arctic Circle, close to the Middle Pleistocene ice limit (Grosswald, 1980, 1993). This model was applied for Quaternary mapping of the Pechora Basin based on photogeology (Arslanov et al., 1987).

However, subsequent testing of the maximalist concept in West Siberia has led to the rejection of a Late Weichselian age for the most of the Kara Sea catchment area. It appeared that finite radiocarbon dates from sediments overlying the uppermost till greatly outnumbered those from beneath the till. There are several successions of `old` radiocarbon dates in the right stratigraphic order derived from soft superficial silts, in places with syngenetic ice wedges, plainly indicating a lack of any glacial overriding (Fig. 3). These are also sites 1, 4, 11, 12, 13, 14 (Fig. 1). Well-preserved frozen mammoth carcasses on the surface (sites 5, 6 in Fig. 1) underline the absence of glacial activity during the last 35 ka. Thus only moderately-sized Late Weichselian ice sheets, mainly north of the Polar Urals and in the Pechora Basin, were deemed possible on the basis of the available radiocarbon evidence (papers 6 in Ch. II and 10 in Ch. III).

A new investigation of the Weichselian morainic system was initiated in 1993 by a joint Russian-Norwegian team in the area between the Uralian Mountains and Timan Ridge (the PECHORA project). During this project marginal formations of the Weichselian maximum have been mapped using photogeological interpretation and dated by the radiocarbon and luminescence methods (Astakhov et al., 1999; Mangerud et al., 1999).

This work has established the Weichselian maximum along the Markhida Line, north of the Arctic Circle, and demonstrated the age of this ice advance to be ~ 70-90 ka BP. The main evidence for an older Weichselian age of the last ice dam across the Pechora Lowland is beach formations of the proglacial Lake Komi postdated by alluvial terraces containing Paleolithic artefacts and abundant remains of mammoth fauna. The latter have been radiocarbon dated to 25-37 ka BP. Traces of Late Weichselian glaciation east of the Fennoscandian moraines have been found neither on the plains of European Russia, nor in the Urals below the 600 m isohypse.

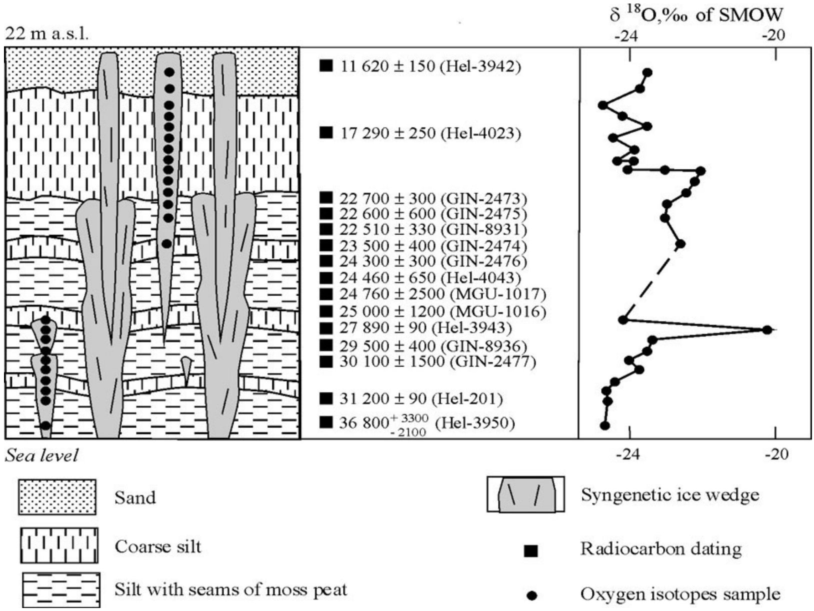


Fig. 3. Periglacial silts in Syo-Yaha section on the eastern coast of Yamal (11 in Fig. 1), adapted from Vasilchuk et al. (2000). Note the two generations of telescoped ice wedges that were growing simultaneously with accumulation of organic-rich silts, the latter of probable aeolian origin. No sign of any glacial activity has been found overlying this periglacial sequence. In other sections this formation is underlain by the latest till with blocks of fossil glacial ice.

A late Weichselian glaciation of this area suggested by Arslanov et al. (1987) and Grosswald (1993) is impossible because of many well-dated sedimentary sequences, including three open Paleolithic sites (18, 19, 20 in Fig. 1), overlain only by aeolian and alluvial sediments (Mangerud et al., 1999). International research teams have lately confirmed the absence of Late Weichselian ice in the most of the Yamal and Taimyr Peninsulas (Forman et al., 2002; Möller et al., 1999; Siegert et al., 1999), yet they have not attempted to map former ice limits.

The Weichselian ice limits suggested herein (Fig. 1) are based on results of the PECHORA project in European Russia and previous works by various Russian investigators in Siberia adapted by the author. The main post-Eemian ice sheet of Siberia evidently predates Middle Weichselian interstadials, which can be judged from the mapped pattern of radiocarbon dated sequences not covered by tills (List in the end). Especially important are undisturbed surficial formations of loess-like silt and moss peat with syngenetic ice wedges and long series of radiocarbon dates from 37 to 11 ka BP in the right stratigraphic order: Cape Sabler (Kind & Leonov, 1982; Möller et al., 1999), Syo-Yaha and Mongotalyang sections (Vasilchuk et al., 1984; Vasilchuk et al., 2000), and Marresale section (Forman et al., 2002) (sites 1, 11, 12, 13 in Fig. 1).

Earlier the limit of Weichselian glaciations east of the Urals was suggested south of the Arctic Circle (papers 6 and 7 in Ch. 2), based on the work by Arkhipov et al. (1977) who reported finite radiocarbon dates from presumably Middle Weichselian sediments overlain by till at the Salehard moraines. The latest study of various sections in this area by the PECHORA project found neither tills nor glacial disturbances within the range of radiocarbon method. Finite radiocarbon dates have been obtained only from fossil plants and mammal bones associated with a well-pronounced periglacial formation up to 9 m thick consisting of aeolian and slope deposits.

The Late Weichselian till by Arkhipov et al. (1977) proved to be small lenses of soliflucted diamictic material at the base of the periglacial mantle, which also displays all kinds of permafrost disturbances. Underlying fluvial sands have been dated by optically stimulated luminescence (OSL) to 80-90 ka (site 15 in Fig. 1) (paper 14, Ch. III). The only sign of the last glaciation in the Salehard area is thick varved rhythmites, which probably correspond to the Sopkay morainic belt mapped along 67.5°N (paper 2, Ch. I). Now it is clear that the Sopkay moraines mark the maximum extent of post-Eemian glaciers, i.e. Early Weichselian glaciers did not reach the Arctic Circle (Fig. 1).

In the Yenissei area the principal uncertainty is connected with the stratigraphic problem of distinguishing between interglacial marine formations of different ages. Only one marine formation (the Kazantsevo strata) with boreal mollusks, indicating an influx of atlantic water, has for decades been identified in natural sections and correlated with the Eemian.

This traditional correlation is geochronometrically confirmed in four sections of the interglacial marine sediments, between 68 and 73.5°N, by ESR dates in the range 109–134 ka (Sukhorukova, 1998). A subtill ESR date of 122 ka is known from the type site at Cape Karginsky (8 in Fig. 1) (Arkhipov, 1989). Superficial tills containing boreal shells are therefore commonly attributed to a Weichselian ice sheet. However, in some reconstructions this approach has led too far. E.g., according to Troitsky (1975), the last ice sheet ended in a very long Yenissei ice tongue reaching as far south as site 23 in Fig. 1.

The problem is that typical boreal fauna, such as *Arctica islandica*, sometimes occurs also in Middle Pleistocene tills, indicating that there was at least one transgression of warm-water sea older than the Eemian. Zubakov (1972) placed this marine event between the last two Middle Pleistocene ice advances, presumably c. 170 ka ago. A similar interglacial transgression is also known in northeastern European Russia (Yakovlev, 1956). Therefore, some interglacial sites with marine fauna, shown in the digital map, might be not Eemian but older, thereby calling to a conservative approach in drawing the Late Pleistocene glacial limit.

In the present map the Weichselian ice limit on the Yenissei is shown just south of the Arctic Circle (Fig. 1) as it was originally mapped by geological surveys (Zarrina et al., 1961), with minor modifications. The main signature of the last ice sheet is the impressive glaciokarst hummocks and lakes described by Zemtsov (1976) and glaciotectonic `hill- and-hole pairs` (Fig. 2). South of the Arctic Circle the fresh-looking glacial landscape is truncated by a flat intra-valley plain at 45-55 a.s.l. composed of thick glaciolacustrine rhythmites. The rhythmites are overlain by alluvium radiocarbon dated to 34-42 ka at Farkovo (10 in Fig. 1) and by sinkhole silts containing frozen logs with dates from 35 ka to infinite at Igarka (9 in Fig. 1) (Kind, 1974).

Many `old` finite and non-finite radiocarbon dates are known from sediments and mammoth remains overlying the uppermost till (papers 6, Ch. II and 10, Ch. III). Besides the subtill marine sediments with boreal fauna and fresh glaciokarst topography there are other indications of a Late Pleistocene age of the last ice advance. These are thick (5 to 60 m) stratiform bodies of massive foliated ice with erratics which often occur within the hummocky landscape above the 66th parallel. The massive ground ice is believed to be mostly remnants of glacier sole preserved in the thick Pleistocene permafrost (Kaplanskaya & Tarnogradsky, 1986; papers 5 and 8, Ch. II). This direct signature of former glaciation is instrumental in delineating the Weichselian ice margin in the central lowland between the Ob and Yenissei where no ice limits have been mapped by the Geological Survey. In the Gydan and Yamal peninsulas the massive buried ice is sometimes overlain by cold-water marine silts with *Portlandia arctica.* However, it is very unlikely that buried glacial ice, normally found at low altitudes, could survive the warm Eemian transgression. Thus, the area with known sites of massive buried ice (papers 6 and 8 in Ch. II) was probably occupied by the last ice sheet.

The Last Glacial Maximum (LGM) ice limit is now identified well offshore in the Barents Sea based on marine drilling and seismic data (Gataullin et al., 2001). The only refugium for Late Weichselian ice on the Russian mainland is the Putorana Plateau, where several finite radiocarbon dates were obtained from beneath well-pronounced end morainic arcs (Isayeva et al., 1976; Isayeva, 1984; Bardeyeva et al., 1980; Bardeyeva, 1986). These piedmont morainic arcs of the Norilsk Stage (Fig. 2, a) reflect snouts of valley glaciers which probably were outlets of a flat ice cap of Norwegian type.

Several problems remain unsolved. In European Russia there are huge morainic loops protruding south of the Markhida Line, namely, the Laya-Adzva Ridge and adjacent ridges, unequivocally indicating a former ice flow from the north-east. No reliable Eemian sequences have been discovered in this area. Therefore, this ice stream is probably from the Early Weichselian (Mangerud et al., 1999), as shown in the digital map, but the Middle Pleistocene age maintained in the regional stratigraphic scheme cannot be ruled out yet. The statistics on OSL dates in the Pechora Basin suggest that there were two Weichselian ice advances: c. 80-100 and c. 60 ka BP (Mangerud et al., 2001a), which agrees with OSL dates from the northern Taimyr Peninsula (Alexanderson et al., 2001) and west of the Timan Ridge (Houmark-Nielssen et al., 2001). However, an unambiguous lithostratigraphic proof of two Weichselian glacial complexes is still lacking. Also, dimensions of Late Weichselian glaciers in the Urals, on the Putorana Plateau and possibly on the northern shore of the Taimyr Peninsula are still disputable.

List of key stratigraphic sites (Fig. 1) for determination of Pleistocene glacial limits

Eemian interglacial sequences not overlain by till

Sula 22, Sula 21, Urdyuzhskaya Viska (Mangerud et al., 1999); Aksarka (Lazukov, 1970; Astakhov, 2001); Hutty-Yaha; Varka-Sylky (Troitsky, 1975); Bol. He-Yaha, Lysomarra, Pancha, Russkaya-1 (Shatsky, S.B. et al., 1956, manuscript, Tomsk); A-434, B-59, IL-60 N-114 (Kind & Leonov, 1982).

Eemian interglacial sequences overlain by till

Sopka, Vastiansky Kon, Kuya (Mangerud et al., 1999); More-Yu-1, More-Yu-2, More-Yu-3, (Astakhov & Svendsen,, 2002); Golodnaya Guba, Vashutkiny lakes, Silova-Yaha (Loseva & Duryagina, 1983); Boreholes 703, 704, 710 (Lavrushin et al., 1989); Nurma-Yaha, Voivareto, (Dolotov M.S.et al., 1981, manuscript, Moscow); Yuribei, Tanama (Lavrov A.S. et al., 1983, manuscript, Moscow); Lukova Protoka, Krestianka (Sachs, 1953); Rogozinka , Igarsky Yar, Cape Karginsky (Troitsky, 1966); Russkaya-2, Russkaya-3, Bol. Heta-1 (Shatsky, S.B. et al., 1956, manuscript, Tomsk); Bol. Heta 2, Bol. Heta-3, Osetrovaya, Lodochnaya-1, Lodochnaya-2, Hikigli-1, Hikigli-2, Uhelengde (Strelkov & Troitsky, 1953); Karaul, Dudinka (Astakhov et al., 1986); Boreholes 28-BH; 31-BH (Arkhipov et al., 1973); Turukhan-1, Turukhan-2, Turukhan-3 (Komarov, V.V., 1980, manuscript, Krasnoyarsk); Potapovo-1, Potapovo-2, A-79, A-81, A-267, A-329, A-410, A-430, B-117, IL-254, borehole bh-5 (Kind & Leonov, 1982).

Radiocarbon-dated, non-glacial Weichselian deposits not overlain by till

Podkova, Yarei-Shor, Pymva-Shor, Mamontovaya Kurya, Byzovaya, Timan Beach (Mangerud et al., 1999); Marresale, Mutny Mys (Forman et al., 2002; Gataullin & Forman, 1997); Syo-Yaha, Mongotalyang (Vasilchuk et al., 1984; Vasilchuk, 1992); Syadei, Lysukanye, Parisento (Bolikhovsky, 1987); Yuribei, Gyda, Mongoche-Yaha (Avdalovich & Bidzhiyev, 1984); Mokhovaya, Gyda (Heintz & Garutt, 1964); Leskino, Kureika (Astakhov, 1998a); Igarka Shaft, Karasino, Farkovo (Kind, 1974); Cape Sabler (Kind & Leonov, 1982; Möller et al., 1999); A-50 (Kind & Leonov, 1982); F-8, F-9, F-17, R-10, R-59 (Fisher et al., 1990); Labaz (Siegert et al., 1999); Kotuy-1 (Bardeyeva, 1986).

Radiocarbon-dated, non-glacial Weichselian deposits overlain by till

Mal. Romanikha-1, Mal. Romanikha-2 (Isayeva et al., 1976); Maimecha (Bardeyeva et al., 1980); Amnundakta, Kotuy-2 (Bardeyeva, 1986).

Interglacial sequences of Likhvin/Tobol (Holsteinian)-type overlain by Saalian tills

Lake Chusovskoye (Stepanov, 1974); Semeika (Kaplyanskaya & Tarnogradsky, 1974); Bakhtinsky Yar, Khakhalevsky Yar (Zubakov, 1972).

Acknowledgements

This work was supported by the Research Council of Norway via the project Paleo Environment and Climate History of the Russian Arctic (PECHORA) and by the EU project Ice Sheets and Climate in the Eurasian Arctic at the Last Glacial Maximum (Eurasian Ice Sheets) (grant ENV4CT97-0563). This paper is also a contribution to the European Science Foundation Program Quaternary Environment of the Eurasian North (QUEEN) coordinating both the projects mentioned. Partial support was provided by Institute of Remote Sensing Methods for Geology (NIIKAM), the Russian Geological Survey. Olga Maslenikova from the latter Insitute assisted with the figures.

CHAPTER II

PECULIARITIES OF SEDIMENTARY RECORD

The following 5 papers reflect mine and my coauthors` attempts to reveal the nature of geological processes governing the terrestrial evolution of northern Russia in the Quaternary. In this respect one of the crucial factors is the old stable permafrost which is intimately involved with all landscape features, especially glacial processes and related geomorphic and sedimentological phenomena (papers 5 and 7). Russian science has a long and fruitful history of geocryological research but only rare works directly attended the glacier-permafrost interaction because traditionally surficial and subterraneous glaciations were studied separately.

The lessons inferred from the discovery of shelf centres of inland glaciation (Ch. I) told us that many previous misinterpretations of the Quaternary sedimentary record in Siberia stem from disregard for the permafrost which deeply influences all glacial processes. The perennially frozen substrate proved to be responsible for the scarcity of streamlined landforms and hummock-and-lake topography in the Arctic (papers 6 and 8), for especially deep glacial disturbances in northern West Siberia and for the confusing sedimentary record in areas of retarded deglaciation (papers 8 and 9). The information obtained on the permafrost influence will be eventually used in chronological and tectonic interpretations.

5. THE 'ICE HILL': AN EXAMPLE OF RETARDED DEGLACIATION IN SIBERIA

VALERY I. ASTAKHOV AND LIA L. ISAYEVA

QUATERNARY SCIENCE REVIEWS, 1988, VOL. 7, P. 29–40

Introduction

The Pleistocene ice sheets of Eurasia are usually presumed to have completely disappeared some thousands of years ago. But in the Arctic lowlands this is not quite accurate because numerous remnants of the former glaciers have recently been described from several localities in West Siberia. This paper deals with a site of ground ice interpreted as a fossil glacier in the Lower Yenissei valley.

Northern Siberia abounds in thick layers of ground ice with controversial genetic labels. The matter has been under discussion since Toll (1897) suggested a glacier origin for large ice beds. Some thirty years ago this concept was abandoned by the overwhelming majority of investigators, partly because of considerable progress in the field of geocryology. As many sophisticated freeze-thaw mechanisms were inferred for modern permafrost conditions, the simple idea of buried ice fields appeared unlikely. Besides, many believed that severe permafrost and thick ice sheets were mutually exclusive. It is not surprising that the once fashionable hypothesis, which postulated restricted glaciation (e.g. Lazukov, 1972; Zubakov, 1972) and predominance of aqueous sedimentation in the Siberian Pleistocene, was readily adopted. In the 1970s engineering works on ground ice seriously considered only ice segregation and ice injection hypotheses (Vtyurin, 1975).

One of the consequences of such an approach was the geocryologists` belief of being able to cope with the problem without special geological investigation. As a result, the Quaternary geologists were deterred from a close examination of ground ice.

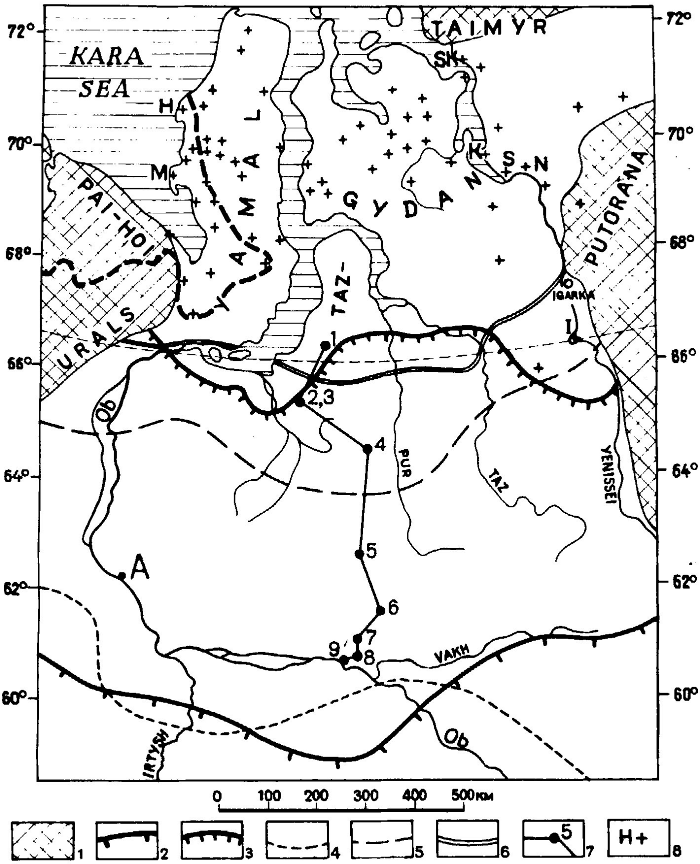


Fig. 1. Location map with principal geocryologic phenomena of West Siberian Plain. Limits of: 1 – maximum inland glaciation (Gm); 2 – last glaciation (Gl); 3 – maximum extent of lower permafrozen layer (Pm); 4 – of discontinuous permafrost (Pi); 5 – joined layers of continuous permafrost (Pc). Other symbols: arrows – direction of ice flow for the last glaciation; quadrangle – study area in Fig. 2a, crosses – sites of thick stratiform ground ice.

Lately a shift in opinion has taken place because it has become evident that the West Siberian lowlands were invaded by south-moving ice masses several kilometers thick (Astakhov, 1976, 1979; Kaplyanskaya & Tarnogradsky, 1975).

The discovery of contorted layers of banded ground ice syngenetic to surrounding mineral matter (Kaplyanskaya & Tarnogradsky, 1976) is the most interesting geological result relevant to the debate. The glacial origin of such ice has given rise to the notion of `primordially frozen tills` for the entire ground-ice body. By now, the known sites of relict glacial ice are so numerous, that it is clear that vast areas of the West Siberian extreme north consist of still frozen till.

The geological consequences of this discovery are as yet unclear because the buried ice beds in the Arctic tundras decay only slowly and thus do not produce extensive diagnostic surficial features. Thus the southernmost fields of buried ice that have survived under comparatively milder climates may be important for drawing uniformitarian inferences.

In 1972 a field of such buried ice was found in the Yenissei valley near the Arctic Circle in thickly forested terrain with discontinuous permafrost. The staff of the Igarka Permafrost Station, who first described the site, initially used an injection mechanism for the so-called `Ice Hill` (Karpov & Grigoryev, 1978). Subsequently, being unable to account for its stratiform structure, they preferred a glacial hypothesis (Grigoryev & Karpov, 1982). From a geological point of view this appeared somewhat puzzling because the ice layer was reported to lie on a low valley plain which had conventionally been regarded as an alluvial terrace.

During the last decade the locality has been visited by a number of investigators collecting data on permafrost, ice structure and composition, stratigraphy and till fabric, etc. All the available data could constitute a formidable volume. But our purpose is to demonstrate a sequence of events via the major structural and geomorphic features to show how the feature originated.

Geomorphic situation

The Yenissei valley between the towns of Turukhansk and Igarka contains constructional landforms, which have conventionally been described as terminal features of the west-moving Putorana ice sheet of the Late Pleistocene. Recent photogeological mapping, however, has revealed that the small hummocks and ridges form several transverse festoon-like belts outlining a former ice stream that advanced from the Kara shelf dispersal centre southwards (Figs 1 and 2). The festooned pattern implies eastward, i.e. uphill, motion of small lobes of the Yenisei ice stream (Fig. 2a). The `Ice Hill` belongs to one of these east-facing lobate features within a large west-east ice-marginal belt, which was previously described as the

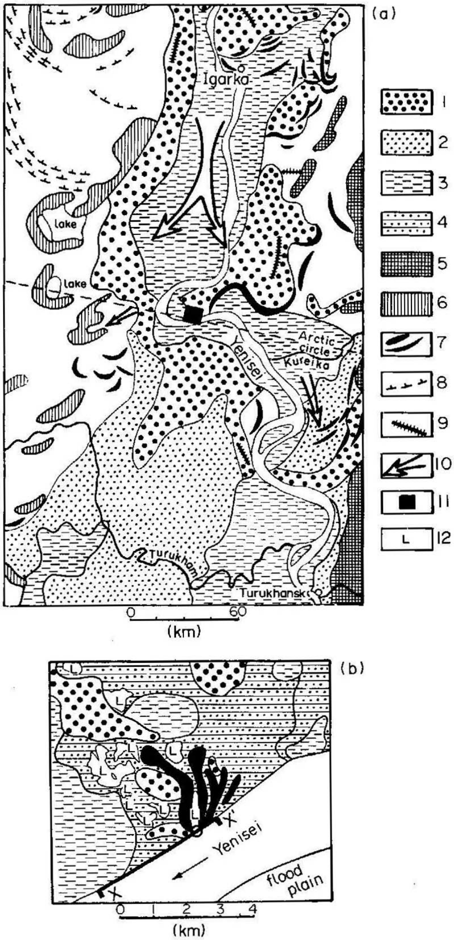


Fig. 2. Glacial morphology of study area according to interpretation of aerial photos. (a) General map of Igarka region. (b) Detailed map of the `Ice Hill` environs. Symbols: 1 – hummock-and-lake landscape of ablation diamicts and sands; 2 – sandy outwash plain; 3 – flat terrace-like surfaces (mainly rhythmites of Igarka lacustrine plain); 4 – kettled `alas` terrace built of sediments of coalesced thermokarst ponds; 5 – outliers of Paleozoic solid bedrock; 6 – ridges of large push moraines (accumulations of contorted soft rocks); 7 – small accretion ridges (mainly diamictons and sands); 8 – glaciotectonic linear features (presumably fractures and thrust planes); 9 – eskers; 10 – direction of the Yenissei Ice Stream; 11 – location of Fig. 2b; 12 – lake.

The small circle in Fig. 2b indicates the thermocirque (Fig. 3), $x - x^1$ is the line of cross-section in Fig. 6.

`Denezhkino stage` (Arkhipov et al., 1980). In front of the Denezhkino moraines the generally north-flowing Yenisei makes a sharp turn to the south-west, cutting the eastern flank of a large marginal festoon (Fig. 2a). Eight kilometres farther downstream (34 km down from the Kureika river mouth) the Yenisei runs through a subparallel set of east-facing arcuate ridges (Fig. 2b). The ridges 0.5-2 km long, 150-300 m wide and 45-75 m a.s.l. are subdivided by swampy hollows and elongated lakes. These ridges consist of a sediment complex 50-60 m thick with ice atop of the sandy `basement` some 10-15 m thick. The diagonally bedded sand, with thin peat interlayers, is the Kasantsevo Formation which is assumed to be Eemian in age (Zubakov, 1972).

The `thermocirque` of the `Ice Hill` was initiated where the ridges joined, as the result of localised thermal erosion by the Yenissei River. A hollow, with an ice floor of about 250 m across, was evident by 1980 as a result of rapid retreat of the walls (Fig. 3). Since 1981 the maximum point of thermo-erosion activity has shifted eastwards, and a new deeper hollow has developed. Contemporaneously the melt-water channel, cutting into the ice floor, broke through the south wall, and a new area of discharge, the Eastern gully, appeared (Fig. 3).

By then the older ice floor in the western part of the cirque became coated with stony mud 2-3 m thick containing peat blocks and tree trunks. Presently it is protected by dried out and refrozen diamicton, packed, cracked, and overgrown with weeds. In 1983 the older floor was reminiscent of a terrace, and was undercut by a new thermo-erosion scarp 1.5-3 m high. The height of the scarp depends on the topography of the surrounding woodland and the surface inclination of the ice, and increases up to a height of 20 m in the north-east where a constructional ridge has built up on the ice surface (Fig. 5a).

The accelerated retreat of the northern walls due to differential insolation that induced asymmetric thawing could drain the Pike Lake shortly (Fig. 3). The natural frozen dam between the lake and the older cirque was only 3-4 m wide in 1983. The discharge of this lake 15 m deep and with a water-table 37 m above the Yenissei could result in a sudden re-exposure of the older ice floor causing rapid melting of the entire ice body, which actually happened several years later.

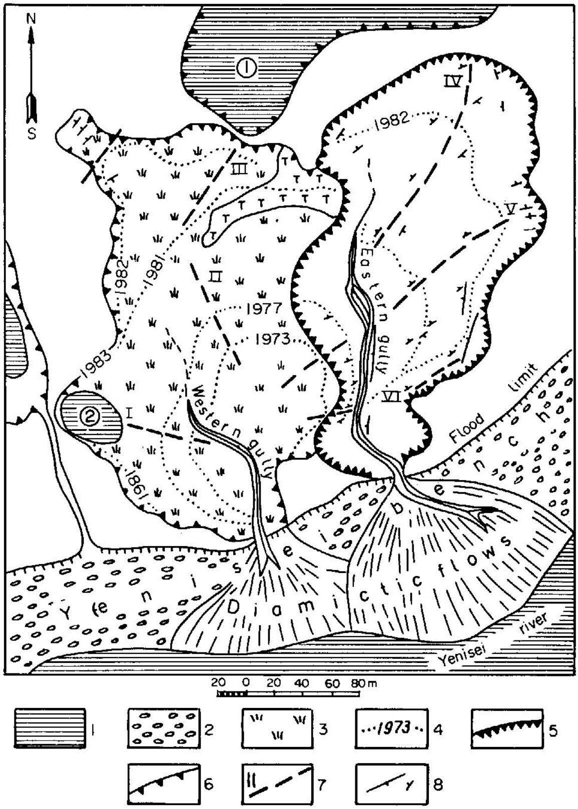


Fig. 3. Thermocirque of the `Ice Hill` (compiled from Ye. Karpov and I. Zolnikov data). Symbols: 1 – present water surfaces; 2 – channel alluvium; 3 – older ice floor; 4 – position of the thermoerosion scarp as reported in years marked; 5 – steep thermoerosion scarp; 6 – gentle thermoerosion scarp; 7 – axes of ice swells; 8 – strike and dip of ice folia. The peat outlier is indicated by Ts. Numbers: 1 – Pike Lake; 2 – Round Lake.

Ice structure

The ice body exposed at the thermocirque floor is believed to be stratiform (or sheet-like) from borehole data and annual observations made by Karpov (Igarka Permafrost Station). The ice bed is more than 30 m thick and, according to electrical survey data, may reach a thickness of 60 to 70 m. The volume of the ground ice far exceeds a million cubic metres

(Grigoryev & Karpov, 1982). The ice surface declines gently to the north-east.

The gentle-sloping ice surface is diversified by dome-like bulges up to 50 m wide. These were formerly interpreted as pingos (Karpov & Grigoryev, 1978). Further observations, however, showed that distances between the 'domes' increase annually. Thus, the bulges can be shown to be linear 'swells` (or undulations), some of which diverge radially from the node at the south-western part of the thermocirque, where the highest point of the ice surface appears to have existed (Fig. 3). The pattern of radial 'swells' also shows in the rate of wall retreat. The initial almost circular thermo-erosion patch has grown irregularly because of the accelerated retreat of the walls along axes of the swells where the ice was more exposed (Fig. 3).

The ice swells are subparallel to the constructional ridges, and the axes of the swells can be extrapolated along the inter-ridge troughs. This is confirmed by a section of the north-eastern wall, where the maximum thickness of the overlying deposits is estimated between swells IV and V. This means that the constructional ridges and inter-swell depressions appear co-axial (Fig. 5a). The inverse correlation between the ground surface and ice surface reliefs coupled with the similarity of their patterns seems to make some genetic sense to be discussed below.

Chemical analyses show that the ice consists of ultra-freshwater with a small $Ca(HCO_3)_2$ residual uncommon for local ground water (Karpov & Grigoryev, 1978). The oxygen isotope ratios (-21.2‰) are the same as in deeper layers of a modern glacier in Severnaya Zemlya, in contrast to -14.1‰ in the ground segregated ice from lacustrine sediments of the Igarka observation shaft (R. Vajkmäe, *oral commun.,* Tallinn, 1984).

The basic elements of the internal structure are bands of clear bluish ice 5-10 cm thick, of milk-white ice 2-5 cm thick, and of dark dirty ice up to 15 cm thick. The alternating bands are generally parallel but converge locally or truncate each other. Apart from the thin dirty bands of silt and clay, a large number of coarser mineral inclusions of various shape, size and composition (from tiny clayey clots to roughly polished angular boulders as in Fig. 4b) are scattered randomly throughout the entire ice body. Faceted and striated wedge-form pebbles are common. Fine gravel clasts tend to concentrate along the inter-band joints where they acquire a planar orientation. On these characteristics the banded structure is believed to be glacial foliation.

Banded ice constitutes several linear anticlines, which roughly coincide with the linear swells of the ice surface. The anticlines are divided by wider box-like synclines. The bands are truncated by the ice surface at a sharp angular unconformity (Figs 4a and 5a). The anticlines are usually asymmetric: the converging gentler limbs of the two anticlines in Fig. 5a dip at 20 to 40° whereas the steeper outward limbs dip at 60-90°. These two anticlines make the cores of swells IV and V (Fig. 3). The ice surface meets the steep bands at this point with sharp unconformity.

The main anticlines are often complicated with crest-line undulations, small conformable faults, and accessory recumbent folds of about 2-3 m wide. Considerable fault displacements can be inferred from the sharp trend discordance between the axes of swells II and III (Fig. 3).

It should be added, that Solomatin (1981), who studied the ice in thin sections, found the crystals to be elongated in accord with the foliation, or radially divergent from the cores of the anticlines. He reached positive conclusions on the metamorphic nature of the ice which shows clear signs of the friction-regelation process.

Sedimentary record within the thermocirque

The most complete sequence was recorded in 1983 at the north-eastern wall (Fig. 5a).

Unit A (the topmost)

(A_1) Yellow-brown and whitish silty sands, laminated, changing upwards into banded silt with fossil weeds, 2 m. These sediments of shallow ponds occur only along the crest of the ridge, at the slopes being replaced by loess-like silt overlain by Holocene peat.

(A_2) Boulders in brown-yellow sandy silt, 1 m thick. This short lens disappeared in 1983.

(A_3) Dark-grey and brown silt with gravel seams in several cup-shaped lenses 1-4 m thick.

Unit B

(B_1) Dark grey clayey diamicton with few boulders and lenses of poorly sorted and washed sand, and fine gravel, 3-5 m thick. The sand lenses are about 0.5 m thick and sometimes up to 10 m long.

(B_2) Dark grey, laminated silt with 2-3 cm bands of black clay, with floaters and fragments of clay, up to 3 m thick. This lacustrine bed thins in a north-easterly direction.

(B_3) Brown-grey poorly sorted sand 0.5-0.8 m thick visible all around the 'cirque'. A radiocarbon date of 43.1 ± 1 ka BP (GIN-1894) was obtained from wood fragments by Yu. Fainer (personal communication).

Unit C

(C_1) Dark-grey, diamictic, stony silt with small lenses of sand. The stones are frequently faceted, wedge-shaped and striated. A slight angular unconformity is observable at the base, but pressure and drag structures were not found. The ^{14}C age of wood fragments is more than 50 ka BP (GIN-1892) (personal communication by Yu. Fainer). This bed persists around the thermocirque.

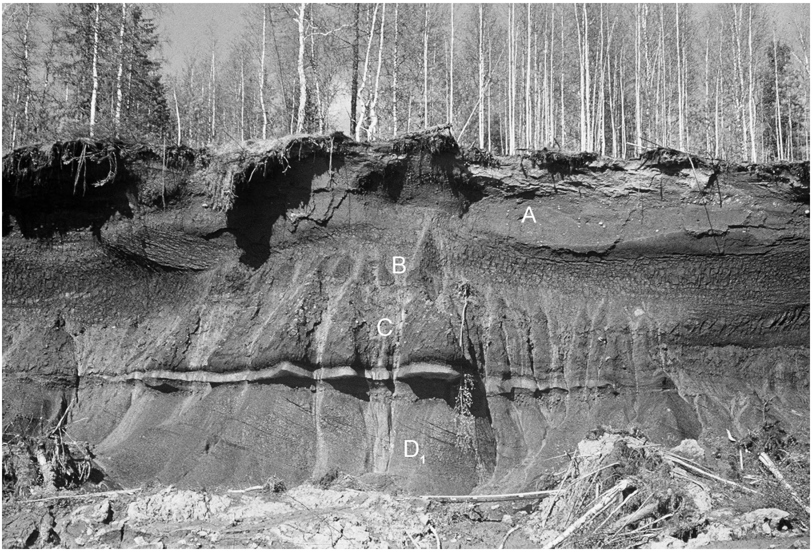
(C_2) Dark-grey, laminated silt, similar to B_2, up to 5 m. The bed is discontinuous and has been observed near the crest of the ridge only since 1983 (Figs 4a and 5a).

(C_3) Poorly sorted, polymict sand with plant detritus changing into fine gravel. The thickness, as with B_3, varies from 0.1-0.2 m in the southern cirque (1970s), to 0.8-1.5 in the north (1983). Locally this bed contains an uneven layer of frosted, firn-like ice with air bubbles, 0.3-0.5 m thick (Fig. 4a).

The sequence is underlain by the already described ice which has an apparent thickness of about 7-10 m (bed D_1). Drill-holes show that the ice is underlain by a compact diamicton 1-3 m thick (bed D_2). Thawed brownish diamicton D_2 can be seen at the Yenissei bench exposures, and was traced downstream where it attains a thickness of 10-15 m (Fig. 6). It consists mostly of sandy silt with pebbles, clasts of underlying alluvial sands with undisturbed lamination, drag folds, shear planes, thin folia, small grooves and thrusts dipping west by north. The long axes and basal planes of the pebbles in D_2 have the same dip, i.e. perpendicular to the strike of the ridges whereas pebble orientation in the younger and softer diamictons B_1 and C_1 varies considerably. According to measurements by I. Zolnikov (manuscript, 1985), the pebbles prefer the down-slope directions of the ice surface. Pebble roundness shows a clear trend from the ice of bed D_1 (minimum total roundness) to beds A_3, B_1 and C_1 (maximum roundness).

All these deposits, except for the top 1-1.5 m, are perennially frozen. The cryogenic structures range from pore-filling ice in the sandy beds to thin veins of various patterns including reticular and ice-gneiss systems. Thin veins in C_1 and C_2 make a brick-like pattern with cells of 20 x 10 x 5-10 cm which lie parallel to the surface of the massive ice but never join it. The same patterns have been laboratory simulated in water-soaked sediments by simultaneous freezing from the top and below (Solomatin, 1981).

a



b

Fig. 4. Photographs of the main wall of the thermocirque (see also Fig. 5). (a) General view 1981. For indices see text; white strip at the sole of C is a firn-like ice C_3; stratified clay of C_2 was not exposed in 1981 (cf. Fig. 5). (b) A boulder suspended in foliated glacial ice D_1. Photo by Ye. Karpov.

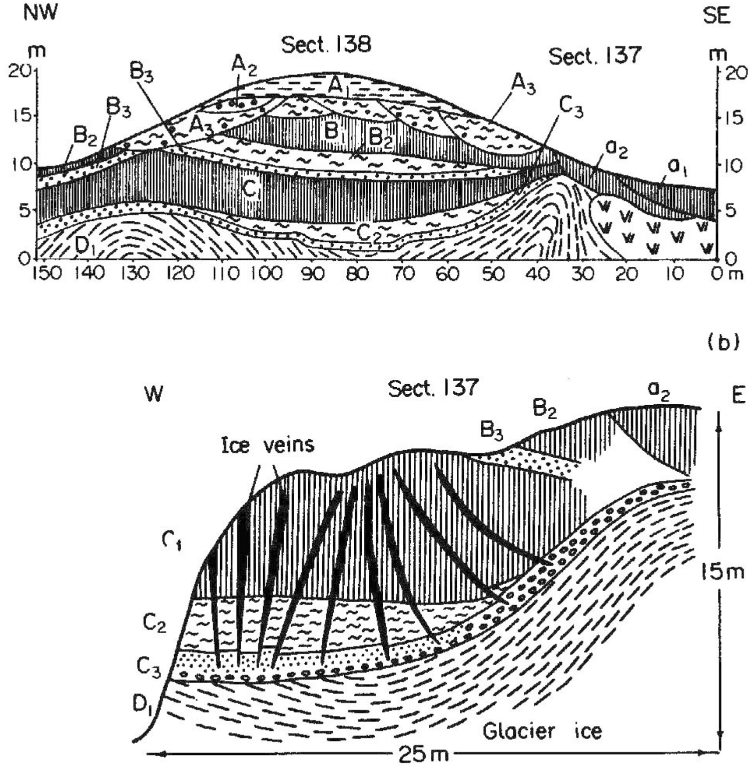


Fig. 5. Exposures of the main wall of the thermocirque as observed in 1983.
(a) Section of accretion ridge between swells IV and V (see Fig. 3). (b) Buttress of the main wall at swell V. See Fig. 6 for symbols and text for indices.

Another spectacular feature is a system of long veins of clear ice up to 40 cm thick which infills sub-vertical cracks, cutting units B and C, but dying out in C_3 sand. Viewed in steep walls they appear as palmate fans opening upwards, if located in ground buttresses, and diverging down, if occur above steeper slopes of the ice surface (Fig. 5b). In the latter case they are slightly bent according to the slope, though they are generally normal to the ice surface. At the contact between beds C_2 and C_3, the ice veins produce a stepped series of microfaults with displacements of a few centimeters. This suggests that these epigenetic ice-filled veins are tensional fractures which were initiated by the slow mass movement of frozen ground over the ice surface.

Other evidence of frozen ground slumping is provided by flat box-like grooves 0.3-0.5 m deep and some tens of metres wide on the ice surface (Fig. 5a). These grooves, of sub-longitudinal strike, are filled with fine gravel and the firn-like ice of bed C_3. Gravel clay and angular fragments of the bubbly ice are fused together to make a dense mixture coating the bottom of the grooves and closely resembling friction breccia. While there are other cryogenic structures not displaying signs of frozen mass slumping, all structures have no direct connection with the massive ice bed (D_1) and their formation is subsequent to the deposition of most of the ice covering sediments.

Apart from the beds already described, there are some younger deposits on the slopes of the ridges lying above the main sequence. First, woody peat with icy gneiss-like structure, up to 5 m thick, abutting against the upper parts of the ridges and filling intervening hollows. A residual block of such peat is preserved in the northern part of the cirque (Fig. 3). The ^{14}C dates from the peat range from 1.4 to 5.6 ka BP (Grigoryev & Karpov, 1982).

Second, the forested riverside slope of the `Ice Hill` is covered with a series of younger diamicton sheets 1-3m thick (beds a_1 and a_2 in Fig. 5a). The diamictons are similar to bed B_1 but contain a number of logs and distinct boulder pavements inclined to the Yenissei bench. A piece of wood from the topmost sheet gave a ^{14}C age of 1,270 ± 100 BP (GIN-1893) (personal communication by Yu. Fainer).

Finally, there is the diamicton cover protecting the higher cirque floor which is not older than 10 years. The diamicton contains thin (1-5 cm) lenses of fine sand and laminated silt deposited in small pools among the chaotic agglomeration of mud, peat blocks, tree trunks and stones, which some years ago were observed on the higher ice floor. The formation of the diamicton continues in the summer months, when stony mud carrying downslope oriented wedge-shaped boulders creeps and slides down along the younger ice floor to the Eastern gully. The gully collects the semi-liquid debris and sometimes spills it onto the Yenissei bench, where the mud flows build up a fan consisting of alternating dark diamictons and alluvial channel sands (Fig. 3). The debris is derived from thaw slumping of the ice-covering deposits and partly from the (D_1) ice.

The structural difference between diamicton D_2 and the younger diamictons suggests different origins. D_2 is compact, sandy, rich in glaciodynamic structures and is basal till. The other diamicton sheets, as it is evident from current processes, were deposited mainly by debris flow, and closely resemble what some call flow till (Hartshorn, 1958; Boulton, 1971), or sediment flow (Lawson, 1979). The alternations of diamictons and waterlain sediments containing glacial clasts, the slope dependent fabric, and surface morphology support this postulated origin.

All three varieties of flow till described by Boulton (1971) in Svalbard are found in the `Ice Hill` diamictons, and show a succession of units decreasing in compactness and increasing stratification from the basal diamicton D_2 to modern mudflows. In the overlying sequence diamicton C_1 is the most compact and least stratified one, and it resembles the compact `stable` flow till described from Svalbard. B_1 is similar to the semi-plastic flow till described from Svalbard (Boulton, 1971). The modern mudstone sheets in the thermocirque are formed by liquid flowage. The more continuous units consist of less plastic diamictic matter and are more extensive than the liquid flows which are confined within the walls of the modern thermocirque.

Background sedimentary record

Quaternary exposures along the right bank of the Yenissei are extremely variable, and the `Ice Hill` succession is not repeated in adjacent outcrops (Fig. 6). Upstream the nearest ridge (section 140) consists of different sediments. Crudely stratified dark diamicton of B_1 type, 3-5 m thick, only covers the slopes and the crest of the ridge, while the core consists of washed and well sorted yellow sands more than 20 m thick. The sands occur in blocks 1-5 m wide tilted in various directions. About 10 m above the Yenisei flood limit the sand beds with seams of plant detritus form a block-like unit with many small tension cracks at both flanks. On the top of this unit the gently dipping sands contain small (1-1.5 m wide) recumbent slump folds. Higher in the succession the dip of the sand beds decreases from 60-80° to 20-30°, and the uppermost sand unit becomes conformable with the diamicton cover on the slopes. A wood fragment from the sands 14 m above the flood limit yielded a date of 39,600 ± 600 BP (GIN-3751) (F. Kaplyanskaya, personal communication).

Imbricated or other glaciodynamic structures have not been found in this exposure, which supports the idea that the sands were accumulating and then collapsing within a deepening ice-floored hollow. The sandy ridge should, therefore, be regarded as an inversion glaciokarst feature. Whether or not a till unit occurs between the collapsed sands and underlying glacially contorted alluvial sands cannot be ascertained because of a thick talus (Fig. 6). Both sandy formations are similar although the underlying one is more regularly bedded.

The possibility that the bottom alluvium has been squeezed diapir-like into englacial cavity seems unlikely on structural grounds, and also because the ridge-forming sands are widespread around the `Ice Hill`: e.g. the wider northern extensions of the ridges (Fig. 2b) consist of thick washed sands of the same texture which are not covered by till. Wide, fan-shaped flat-topped ridges, with steep, sometimes stepped slopes, rise to 75 m above sea level. The form of the ridges resembles the association eskers with delta moraines (Embleton & King, 1968), though not all the ridges can be shown to be feeder channels.

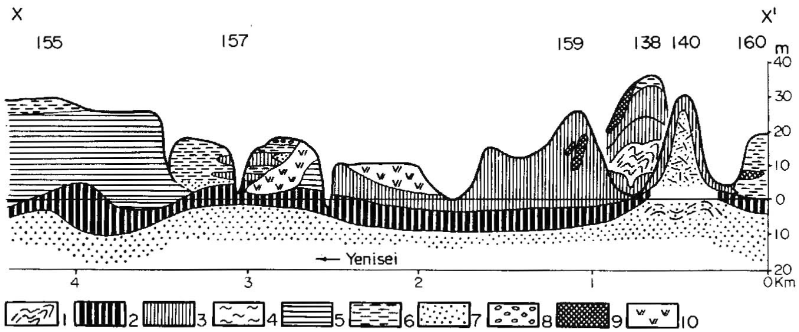


Fig. 6. General cross-section along the right bank of the Yenissei 3.2-7.4 km downstream of river Denezhkina mouth (location see in Fig. 2b). Zero point of the height scale is at the Yenissei flood limit which equals 18 m a.s.l. Symbols: 1 – contorted glacial ice (D_1); 2 – dense diamicton (D_2); 3 – loose diamictons; 4 – crudely laminated clays; 5 – varved clay and silty rhythmites; 6 – laminated silts 7 – washed sands with fine gravel; 8 – pebbles and boulders; 9 – wood peat; 10 – talus.

Most of the Yenissei valley is occupied by a constructional plateau of much the same height as the `Ice Hill` ridges, and often even higher (above 60-80 m).

Its surface is rather flat, with many small mounds and kettles but without lakes. The plateau is composed of lacustrine rhythmites 30-40 m thick which continue for hundreds of kilometres along the Yenissei (Fig. 2a). The topmost silt-sand, alternating with laminae of allochthonous peat, grades downward into black varved clays with numerous marlekors. The rhythmites are underlain by diamicton D_2 of variable thickness (Fig. 6, left). The lacustrine formation appears to replace the `Ice Hill` sequence laterally, although the age relations are uncertain. We believe that cyclic units covering the ice are somewhat younger because no tongues of flow tills have been observed in the rhythmites (Fig. 6). Some 8 km upstream, at the distal flank of the Denezhkino morainic belt, flow-tills do occur, but in an area where no ground ice was found.

Some idea of the age of the glacio-aqueous beds can be inferred from radiocarbon dates on younger deposits. There are numerous sink-holes in the glaciolacustrine plateau, and tree trunks and stumps from silts filling a thermokarst hole in the Igarka observation shaft were dated to 53,400 ± 300 (GIN-140), 39,000 ± 460 years (GIN-328) and >50 ka BP (GIN-327) (Kind, 1974).

The Yenissei alluvial sequence begins with the `Second Terrace` (30-40 m a.s.l.) which consists of coarse channel sands changing upwards into fine sandy rhythmite with seams of plant detritus. This narrow and topographically poorly defined strip of alluvium bears strong former cryoturbations in the topmost layer. Just below the cryoturbations, at a depth of 4.5 m, a date of 32,500 ± 400 BP (GIN-99) was obtained from plant detritus (Kind, 1974). We have obtained a similar date of 31,100 ± 800 (GIN-3674) from bark-covered stems of tundra shrubs at the same stratigraphic level near the mouth of river Kureika (Astakhov & Isayeva, 1985). This suggests that the channelling process ceased and was replaced by floodplain sedimentation in a cold environment about 30-35 ka BP. A very cold climate is also inferred from ice wedge casts in the alluvium of the `First Terrace` which is believed to have been formed about 10-11 ka BP (Zubakov, 1972).

Subsequent events are recorded in the sediments of the thermokarst lakes that compose a terrace-like pediment with kettles and bogs around the arcuate ridges of the `Ice Hill` (Figs 2b and 6). This `alas` plain is at the same elevation (30-40 m a.s.l.) as the `Second Terrace`, but channel sands are rare, and laminated silts and silty fine sands of lacustrine appearance predominate. The topmost units are thick peats, and in some places gravelly sand. Lamination of the silts is often disturbed, showing underwater slumping in thermokarst ponds. Not far from the `Ice Hill` the `alas` formation contains wisps of soft dirty diamicton 1-4 m thick (section 157 in Fig. 6). Many pieces of birch with intact bark are included in these diamictons, and two of them have provided dates of 9,490 ± 100 (GIN-3669) and 7,080 ± 40 BP (GIN-3670) at depths of 4 and 1.5 m below the surface (Astakhov & Isayeva, 1985).

The exposures of the thermokarst pediment nearest to the `Ice Hill` show only thick soft diamicton. Upstream of the `Ice Hill` an apparently slumped, thick lens of wood peat with sharp contacts has been observed in the lower part of a silty unit (section 160 in Fig. 6). The ^{14}C dates are: 7,850 ± 40 (GIN-3673) from the peat, 8,080 ± 50 and 8,150 ± 40 BP (GIN-3672) from the wood (Astakhov & Isayeva, 1985). It is noteworthy that the surficial peats of the `Ice Hill` are not older than 5600 years (Grigoryev & Karpov, 1982).

Origin of the `Ice Hill`

The data described hitherto do not support an injection and segregation hypotheses for the origin of the ice bed. The erosional nature of ice-ground interface, the strike of thin sand-gravel beds all around the thermocirque, and the mantling shape of the sediments covering the ice body, are consistent with a buried ice idea. A possibility of a repeated glacial advance that could lay down B_1 and C_1 diamictons (Grigoryev & Karpov, 1982) is rejected because the adjacent lacustrine plains have neither superposed basal tills, nor glacial disturbances, and because of the flow nature of the ice-covering series. The view of Solomatin (1981) who believed that the ice body was buried by ablation moraine is more realistic.

It could, of course, be argued that the ice bed was primarily segregated, glacially contorted, then finally buried, as Mackay (1971) suggested for the ice beds of Arctic Canada. But this is unlikely, because of the suspended boulders (Fig. 4b). The debris in the ice D_1 in the superposed sequence, and in modern mud flows is practically identical. The volume of the ice-covering sediments is considerably larger than the volume of the underlying diamicton D_2 (Fig. 6). Therefore large masses of dirty ice must have melted away to produce the volume of glacially polished stones, silt and sand that comprise the ice covering sequence. If the melted ice was glacial it is unlikely that the remnant ice is any different in origin. And this ice, with large erratics (Fig. 4b), lies most appropriately between the basal and the ablation tills.

Finally, the spatial relationships of the diamicton show that it covers sandy ridges and lacustrine rhythmites, and is the result of a long-term process of melting that commenced in the Late Pleistocene to form the modern hummock-and-lake topography. The latter landform assemblage dominates the Lower Yenissei basin; and cannot be attributed to local thermokarst sinking, but only to multi-stage melting of extensive ice fields.

The sum of the site`s characteristics show that an application of Boulton's (1972) Spitsbergen model of inversional development of ice-controlled topography is appropriate in this instance. In particular (1) the arcuate pattern of ridge-and-trough topography; (2) direct correlation between the areal patterns and inverse correlation between the heights of the land and ice surfaces; and (3) the thick sequence of ablation sediments, lend further support to this view.

The ablation sediments show three cycles of melting: units A, B and C of the thermocirque series. B and C are similar; a thin mantle of washed sand at the bottom and silts deposited in shallow ponds with stone-mud masses on the top.

The structure of the sedimentary sequence makes us suspect a cold and rhythmically changing climate. The most striking feature is the thin laterally persistent bed C_3 coating the ice relief. It is repeated, although without the gravel lenses, in B_3. This and the generally cyclic structure of the sequence do not permit melting out of the sand. Thus we suppose that no appreciable seasonal thawing occurred and, moreover, the ice mass was increased by the firn sheet C_3. The sand particles may have been wind-transported to the ice hollow; and the gravel pockets may be a residual cover from a preceding episode of ablation. Any possibility that the cyclic sequence reflects a rhythmic alternation of clear and dirty ice is unlikely because there are no thick debris-rich bands in the surviving ice.

Some climatic amelioration during the deposition of rhythm C presumably led to the formation of pools with silty sediments. This could be the origin of the fairly compact unit C_1, which may have slumped because of an increased thickness of the active layer. At this time the ice surface probably became forested and soil-covered because the overlying sand B_3 contains driftwood dated to ~ 43 ka. The thick logs with similar dates above the lacustrine rhythmites at Igarka suggest that the Igarka Lake had been drained by this time. Following the cooling marked by B_3, the processes were repeated but with the difference that meltwaters were more abundant, as is inferred from the continuous waterlain bed B_2, and the stratified diamicton B_1.

The numerous ponds of cycle A imply a climatic amelioration when meltwater was more abundant. The ice slopes, however, must have eroded so as to cause the melting of the basal ice. Lacustrine bed A_1 could have been deposited in shallow ponds on undulating and mud surface of the ice plain. It is also likely that the distinctly stratified diamicton coating of the sandy ridge (sect. 140 in Fig. 6) was formed at this time. The climate, although sufficiently ameliorated to allow the ice relief to be degraded was, nevertheless, still sufficiently cool so as to prevent any continuous drainage through the dead ice mass.

Because the ice-covering sediments were frozen immediately after deposition, and because they lack any taliks beneath the lacustrine lenses, this shows that the climate was not warmer than the modern one. The present climate is very cool (-8 or -9°C mean annual temperature), but rather humid (mean annual precipitation about 500 mm), with the result that there is a thick snow cover which does not allow severe permafrost (Shevelyova & Khomichevskaya, 1967). Annual precipitation of the inter-Weichselian phase of deglaciation should have been somewhat lower to produce conditions for permafrost preservation, as well as to allow progressive downwastage of the dead ice. Nowadays such conditions, with very thick permafrost and deep seasonal thawing, are widespread in East Siberia.

The history of the `Ice Hill` and its environs may be modelled as follows:

(1) Stagnation of an east-moving lobe of the Yenissei ice stream that had deposited the compact basal till-diamicton D_2. Widening of arcuate crevasses parallel to the lobe margin led to their gradual development into ice-floored troughs. The crevasses only penetrated the rigid upper ice and became pinched out in the basal layers of the contorted plastic ice. The `Ice Hill` block was isolated from the main ice field by Igarka Lake that originated as a proglacial lake and developed into a propagation bay (Fig. 2a). At this phase supraglacial debris was being transported to this and other such lakes. The lake level was higher than the bottoms of the arcuate troughs (Fig. 6), and a through drainage did not exist.

(2) Formation of the cyclic ice-covering sequences of the `Ice Hill` by intermittent creeping of thawed englacial materials down the degrading ice slopes of the troughs. During the course of this gradual climatic amelioration Igarka Lake was drained and the resultant lacustrine plain (Fig. 2a) was refrozen, forested and dotted with numerous small thermokarst holes. The radiocarbon dates show that this occurred about 40 to 50 ka BP.

Afterwards the formation of glaciokarst hollows with channel sands occurred (sect. 140 in Fig. 6). The lacustrine deposits of unit A (Fig. 5a) could be tentatively associated with the beginning of the Middle Weichselian interstadial. The climate remained continental, and the massive ice of the `Ice Hill` was sealed by frozen sediments. At this time the surrounding ice relief was probably graded and transformed into an ice plain. At the same time the alluvium of the Second Terrace of the Yenissei was accumulating.

(3) A further cooling is shown by the Second Terrace rhythmites and aeolian deposits; this commenced about 30-35 ka BP. Thermokarst processes ceased, and the soil-covered thin ice fields survived until next climate warming.

(4) The end of the last cold stage of the Pleistocene caused a resumption of the glaciokarst process. The ice plain was forested and swamps developed. As the former ice troughs were choked by drift, the local drainage shifted laterally towards strips of more exposed ice, a process which marked the beginning of the main topographic inversion at the `Ice Hill` locality. Numerous sinking ponds, gradually etching the forested ice fields, coalesced to form a thermokarst plain, above which the

choked former troughs became progressively more conspicuous as till and sandy ridges. At the end of this process the sink-holes were filled with lacustrine sediments with diamicton wisps (sect. 150, 160), and were drained to build up the `alas` terrace body. The process continued during the Early Holocene (it may have commenced during the latest Weichselian interstadials) and ended when the large fields of dead ice were destroyed.

(5) During the Middle Holocene a new thermokarst cycle began to perforate less protected remnants of the buried ice to produce the present lakes. Their bottoms were lowering down through the buried ice beds. The sediment-filled lakes of the preceding stage appeared as the kettled `alas` terrace which means a second, partial topographic inversion occurred. Many of the younger lakes were drained off and more swamps developed, especially where ice remnants were thin. Numerous peat bogs, which originated some 5-6 ka BP, are now scattered all around the `alas` plain and on the slopes of the till ridges.

The general subsidence of the hummocky terrain is continuing although this process is relatively slow. The most effective means of deglaciation is displayed by the persistent northern shift of the west-east sector of the Yenissei river channel: from the wide left-bank floodplain to the rough thermokarst country at the right bank. Other Yenissei channel sectors show both right and left-bank shifts dependent upon location of the nearest hummock-and-lake terrain (Fig. 2a). The formation of the `Ice Hill` thermocirque was the last event in this thermo-erosion process, during the course of which a headward eroding gully intersected the thinly covered part of the large buried ice layer.

So far as can be evaluated from the abundance of modern steep-shored lakes and their intricate shape, buried glacier ice, (though modified considerably) is not uncommon at these latitudes. There are at least three known localities in the Igarka region. An exposure of contorted dirty ice with boulders has been recorded 60 km south and 80 km west of the `Ice Hill` by Lavrushin (1959) who, however, did not discuss its origin. Such bodies which survive in the south of the permafrost zone should be mapped as hazards for the construction industry. No water-soaked sediments could safely be deposited on the steep ice slope of swell V (Fig. 5).

It is well known from model experiments, theoretical calculations and salt tectonics studies, that two-layered systems with a lower layer of lesser density and viscosity are very unstable. Such instability leads to·diapiric protrusions of the lower layer in the form of waves and domes (Ramberg, 1968). In our case the emerging of the ice along the swells caused by a difference in density at the rock-ice interface is favoured by the two additional circumstances. Firstly, as it is evident from the section in Fig. 5a, the overburden is only 1.5-2 m thick at swell V compared with 15-20 m in the adjacent depression, which provides a ten-fold difference in lithostatic pressure. Secondly, there are ready-made anticlines which are in accord with both stress fields induced by the former glacier and by the modern overburden. This phenomenon may well be called inherited glaciotectonics.

Thus it is possible that the drift-buried ice swells were less conspicuous initially, and diapirism which tilted the drift-ice interface then occurred during the course of Holocene heating. This kind of adjustment of the ice-rock system to changing thermal conditions may account for the tension fractures which dissect the deposits covering the ice (Fig. 5b).

General discussion

The foregoing leads to several conclusions which are not in full accord with some current ideas. Many maintain that the recent (about 20 ka BP) ice advance was from the mountains down to the West Siberian Plain. The extreme north of the Plain, where conspicuous landforms were not reported, was regarded as a zone of marine transgressions and of older glaciations. This has been contradicted by investigations showing west-east ice-marginal features demonstrating an ice movement from the Kara Sea coastal lowlands (Astakhov, 1976, 1979; Arkhipov et al., 1980; Kind & Leonov, 1982). The fresh appearance of the glacial topography between 66° and 69°N was taken to show a Late Weichselian age. The example of the `Ice Hill`, however, shows that the Late Pleistocene glacial maximum occurred much earlier, as inferred from numerous `old` radiocarbon dates overlying the basal till D_2.

The scarcity of accretion hummocks in the extreme north cannot be ascribed entirely to glaciolacustrine activity because thick limnic formations have only been mapped there at altitudes below 70 m. For the higher till and outwash plains (mostly in the Gydan Peninsula) the idea of Kaplyanskaya & Tarnogradsky (1977) about older frozen tills seems more suitable. They suggest that north of 69° N expressive glacial landscapes have not yet formed because debris-rich basal layers of the last ice sheet have never been subject to surficial melting and glaciokarst activity. This means that in permafrost areas the appearance of glacial topography cannot be used reliably for morphostratigraphic correlations.

Another important question concerns the origin of the permafrost. It is widely known that the base of the West Siberian permafrost gradually rises from 400-500 m below the ground in the extreme north to 200-300 m near the Arctic Circle (limit Pc in Fig. 1). Farther south the permafrost is subdivided by thawed rocks into two layers, of which the upper continues south as a number of thin frozen `islands` and ends near the limit marked as Pi in Fig. 1. The lower layer plunges southwards. Near limit Pm (Fig. 1) its surface lies at a depth of 150-250 m and its base is at depths of 400-500 m (Zemtsov, 1976; Trofimov, 1977; Shpolyanskaya, 1981). That is, the lower layer in the south of the frozen zone is thicker (though less continuous) than the entire permafrost along the Arctic Circle. Zemtsov was the first to describe the lower layer and he believes that it is a relict from the most extensive glaciation of the Middle Pleistocene (limit Gm in Fig. 1); and that the upper layer bas been formed as a result of Late Holocene cooling. Some geocryologists regard the lower layer as a Late Weichselian relict (e.g. Balobayev et al., 1983), a view based on computerized simulations of the permafrost history which show that a frozen layer thaws completely in 10-20 ka of climatic amelioration (Shpolyanskaya, 1981). The isolation of the lower layer is ascribed to extensive downwasting of the permafrost during the Holocene climatic optimum.

The preservation of the fossil glacier of the `Ice Hill` near the surface shows quite clearly that along the Arctic Circle deep thawing of interfluvial terrains did not occur either in the Weichselian interstadials nor in the Holocene. This is also supported by the wood (with `old` radiocarbon dates) from the Igarka observation shaft which bears no traces of decay, thus suggesting that it was frozen immediately after burial and not thawed since (Shevelyova & Khomichevskaya, 1967). In spite of the Late Pleistocene age of the permafrost in the Igarka lacustrine rhythmites the frozen layer is only 20-40 m thick, which indicates that the Late Weichselian freezing was not excessively deep. Thus, the thick permafrost at 100 km to the south (about 360 m in a borehole on the left bank of the Yenissei river west of the `Ice Hill`) should be regarded as a relict of Early Weichselian cooling along the edge of a former ice-sheet. The relict nature of the permafrost in the hummocky terrain around the `Ice Hill` is evident from the temperature profile in the mentioned borehole which shows a discontinuity in heat flow from a near-zero gradient in the frozen layer, to a normal one immediately below (Shpolyanskaya, 1981).

In this way it has been concluded that the frozen layer near the Arctic Circle was formed mainly in the Pleistocene. This is reasonable if the closeness of the limits of the last glaciation and continuous permafrost layer is taken into account (Gl and Pc in Fig. 1). The divergence of these boundaries in the Yenissei basin may be attributed to the persisting marginal and propagation lakes of the Late Pleistocene (Fig. 2a) and subsequent channel activity of the Yenissei. Those water bodies evidently heated the thick marginal permafrost locally and dissected it into a number of islands.

The same is suggested for the central part of the West Siberian Plain where recent boreholes have revealed that the solid layer of continuous permafrost in the south ends abruptly (Balobayev et al., 1983). It is believed that this sharp southern edge of the permafrost coincides with a glacial margin and a shore of a large proglacial lake; after it had drained only a thin surficial frozen layer had time to form (limit Pi in Fig. 1) (Balobayev et al., 1983). This assumption of the Late Weichselian age of the Arctic permafrost is contrary to the views of other geocryologists who maintain its modern origin. The `Ice Hill` shows that the thick upper permafrost is even older than the Late Weichselian.

The distribution of the lower relict layer is further instructive because its southern limit (Pm in Fig. 1) is close to the glacial drift limit (Gm) in the lowest central-western part of the plain, and is easily related to the well- known field of proglacial varved clays of the Middle Pleistocene. Pm deviates to the south of Gm only in the Ob-Yenissei interfluve where altitudes exceed 100 m and where vast lacustrine plains do not occur. It has nothing to do with modern climate.

All this implies that Pleistocene glaciations and permafrost, both surficial and underground, were closely associated. Moreover, the modern distribution of permafrost is indicative of glacial history. Perennially frozen rocks, including fossil glaciers, were formed mainly in the Pleistocene, and underwent only minor changes in the Holocene. The preservation of thick `warm` permafrost with stagnant glacial ice shows that computerized models underestimate the temperature inertia of sedimentary rocks and exaggerate the activity of geothermal heat flow. Some authors suggest that endothermic consolidation of Mesozoic and Paleogene montmorillonite clays could provide an effective screen for the geothermal heat flow (Ostryi & Sakhibgareyev, 1971).

On sedimentological grounds it is clear that the ablation series of the `Ice Hill` is similar to descriptions from Spitsbergen, Denmark and elsewhere from marginal environments. But the unthawed tills, with clasts of underlying sediments, such as diamicton D_2 of the `Ice Hill`, can hardly be identified as lodgement or melt-out tills of the sedimentological models of Boulton (1971), Lawson (1979) and Shaw (1982). Those permanently frozen tills are most appropriately interpreted in terms of the Shantser-Lavrushin model (Lavrushin, 1976) as layers of basal ice-debris mixture exfoliated during active ice phases. Their ubiquitous occurrence in Arctic Siberia (Kaplyanskaya & Tarnogradsky, 1977) implies that `melting out` is a diagenetic, not a depositional process and, therefore, is not necessary for genetic definition of basal tills. The same tills in the south of glaciated West Siberia are thawed, but this fact adds little to the comprehension of their genesis. The way by which the till was released from an active glacier sole (exfoliation, plastering-on, etc.) is the most essential characteristic for a glacial geologist striving to obtain information about behaviour of ancient glaciers. The way by which the till was modified afterwards may provide some evidence of postglacial climates. Thus we consider fossil ice to be a basal glacial deposit *(sensu stricto)* which, under certain conditions, may persist infinitely.

Kaplyanskaya & Tarnogradsky (1977), referring mainly to the extreme north of today, have compared an incomplete deglaciation in Siberia against a complete one in Europe. It is reasonable to suppose that there would have been differences in the process of deglaciation as well. Bearing in mind possible existence of such difference in the past, it seems appropriate to distinguish *rapid* (Atlantic) and *retarded* (Siberian) climatic types of deglaciation. In both cases glaciers may have acted on frozen bedrock but subsequent events differed.

The rapid deglaciation in Atlantic environments was supported not only by direct solar action but also by intensified import of heat and moisture due to the resumed cyclonic activity and ocean proximity. Rapid melting provided plentiful drainage which fashioned numerous supra and englacial channels. This led to rapid and effective surface dissection of the dead ice by thermoerosion and to bare glaciokarst activity. The increased roughness of the ice surface decreased its albedo and considerably amplified absorption of solar energy. This process, with positive feedback, caused dead ice disintegration into a great number of small blocks. The latter were apt to accelerated destruction accompanied by debris flowage. The initially not very thick permafrost was destroyed by direct heating. The abundance of free water may account for the much referred to wet-based conditions that induced diapiric protrusions, ice-pressed forms, etc.

In Siberia ice sheet disintegration was rapid only at the start due to the `dissecting deglaciation` in the extreme north. Such deglaciation is thought to have been caused by sea-level raising and calving (Grosswald, 1983). Much of the ice was destroyed by lateral thermal erosion both in marine and lacustrine environments. The remains were preserved in the form of huge ice-fields protected by an ablation debris mantle on top and thick permafrost below. Atlantic cyclonic influences did not penetrate to Siberia during the Late Pleistocene, so that continental conditions were stable. Thus the ice fields decayed mainly by means of covered glaciokarst (thermokarst) accompanied with sporadic creeps and slumps of the ablation mantle. Surviving Early Weichselian dead ice in Siberia was, therefore, predetermined not only by its spatial distribution, but also by persisting glacial climates in Europe.

Retarded deglaciation should be regarded as a specific type of landscape evolution. Because areal downwasting is too slow, deglaciation proceeds mostly by local fluvial and lacustrine erosion, dissecting the entire ice-permafrost body into separate blocks; these decay from the sides rather than from the top. Such processes produce a stepped thermokarst terrain.

The `Ice Hill` case is a reminder of Boulton's warning (1972) that flow sheets alternating with stratified drift may be erroneously interpreted as results of repeated glacial advances. Such sheets are even more misleading in areas of retarded deglaciation where they may occur in real interglacial sequences, marking episodes of climatic amelioration, i.e. having the opposite to basal tills environmental inferences. The situation may be worsened by slip planes, boulder pavements, and buckle folds (which can readily be confused with glaciodynamic features, if viewed in separate sections).

The case of the `Ice Hill` is especially relevant for the northern stratigraphy, where persistent controversy on the inter-Weichselian transgression of the Arctic Ocean, preservation of mollusk shells in boulder clays, very recent glacial advances and so on is endemic. Many key sections, therefore, need revision, especially those where, except for a diamictic sheet, no other members of glacigenic sequence have been observed.

Such an instance is the Markhida section near the mouth of the Pechora river, which Grosswald (1983) interpreted as evidence of an Early Holocene surge of the Barents Sea Ice Sheet. So far as can be judged from the published account, the situation is similar to that of the `Ice Hill`. A thin coat-like Markhida diamicton is underlain by lacustrine sediments with numerous ^{14}C dates from 8 ka to 10 ka BP, and is overlain by a peat dated to 4,000 to 9,300 BP. A further investigation in search of flow, slump or drag structures would be instructive.

PS. Such investigation was performed later (Tveranger et al., 1995).

Acknowledgements

We are greatly indebted to those colleagues mentioned in the text who kindly allowed us to make use of their unpublished data. Special thanks are due to L.D. Sulerzhitsky, Geological Institute (Academy of Sciences of the U.S.S.R.), who supplied all the radiocarbon dates. Very helpful suggestions were made by F.A. Kaplyanskaya and V.D. Tarnogradsky, National Geological Research Institute. This work would have been impossible without the encouragement and support in field study by the staff of Igarka Permafrost Station (Institute of Permafrost, Siberian Branch of Academy of Sciences of the U.S.S.R.) and its chief R.M. Kamensky. We are also grateful to Professor D.Q. Bowen from London University who assisted us with the manuscript.

6. The Last Glaciation in West Siberia

Valery I. Astakhov

Sveriges Geologiska Undersökning, 1992, Ser. Ca 81, p. 21–30.

Introduction

The world greatest sedimentary basin, topographically expressed as the West Siberian Plain, has for several decades been a battlefield for exceptionally controversial ideas regarding the extent and age of the last glaciation. The solution of this problem is of no small importance, as it would affect global paleoenvironmental models for the last 100 000 years. The maximum and minimum paleoglaciological models (e.g. Hughes, 1981; Grosswald, 1980; Biryukov et al., 1988) depend on certain assumptions that can only be verified by direct evidence from the eastern flank of glaciated Eurasia (Fig. 1). Such evidence is not easy to organize because the available field data have been collected by numerous investigators armed with different techniques, and do not fit into any simple pattern. Hence, no universally accepted model of the last glaciation has ever existed for West Siberia and the adjacent Kara Sea shelf. In simplified terms, the discussion boils down to the question of whether or not there was 2-3 million km^3 of ice spread over the Kara shelf and adjacent dry lowlands in the Post-Eemian, and if so, when exactly?

There are two opposing answers:

i) The montane glaciers of the Urals and Middle Siberia expanded downslope but the resultant piedmont ice sheets hardly reached the Ob and Yenissei valleys even during the Early Weichselian maximum. In the Late Weichselian the glaciation was reduced to an orthogonal system of alpine glaciers. The present northbound drainage of the Plain has never been interrupted (Lazukov, 1972; Zubakov, 1972).

ii) The glaciation developed in much the same way in West Siberia as it did elsewhere in Northern Europe and America and reached its maximum in the Late Weichselian, when thick inland ice filled up the northern part of the sedimentary basin. The Arctic ice sheet dammed the northb-flowing rivers to produce huge proglacial reservoirs with a water table of about 1.5

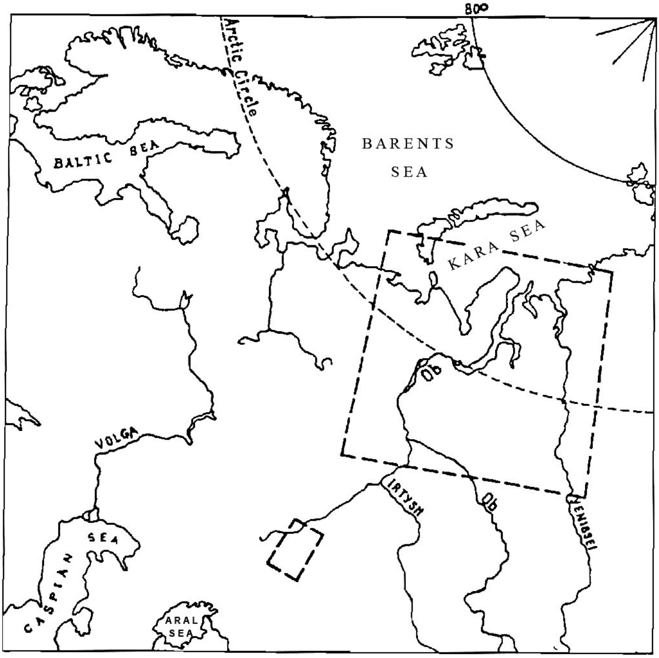


Fig. 1. Location map. Heavily dashed line shows location of Fig. 8 and 9.

million km^2. The drainage was diverted southwards via the Turgai Valley to the Aral Sea (Volkov et al., 1978; Arkhipov et al., 1980; Grosswald, 1980, 1983).

This author, who helped to disprove the minimum model by showing that it did not fit the spatial pattern of glacial features (Astakhov, 1976, 1979; Arkhipov et al., 1980), originally tried the maximum model. A closer examination of the periglacial sediments nevertheless has led to a rejection of the idea of great proglacial lakes in the Late Weichselian (Astakhov, 1989). Thus a reassessment of the available data along with newly discovered facts to create a comprehensible model became a must.

Glacial geology

The arena of the last glaciation in West Siberia is mostly low perennially frozen terrain beyond the Arctic Circle. Expressive hummock-and-lake topography occurs only along the Yenissei valley and around the Polar Urals. The predominant landscapes are tundras and sparse woodlands with numerous muskegs. Underneath there are silts and clays usually covered by washed sand 5 to 20 m thick. The Quaternary sediments 50-100 m, in places more than 300 m thick, overlie soft Mesozoic-Paleogene clays, sands and diatomites of a sedimentary basin some 6-8 km deep. Paleozoic hard rocks are only exposed in the Urals and Central Siberian uplands (Fig. 2).

The topmost diamict formation contains a sizeable amount of clasts mostly near the mountainous borderlands. In the central part of the Arctic lowlands, not higher than 100 m a.s.l., hard rock pebbles are scarce and usually derived from the local Cretaceous and Paleogene gravels. Instead, the Arctic diamictons contain mollusk shells, foraminifera and numerous blocks of underlying soft deposits, both Quaternary and pre-Quaternary. Many geologists have perceived this as evidence of montane ice sheets which discharged into a marine reservoir (e.g. Lazukov, 1972; Zubakov, 1972), but this concept is incapable of providing any realistic pattern of ice dispersal paths.

The glaciomarine hypothesis lost its foundation when Kaplyanskaya and Tarnogradsky (1975) made extensive clearings in the Yenissei bluffs to amply demonstrate that the would-be glaciomarine diamicton is actually basal till with drag folds, thrust planes, streamlined sand balls and other typical glaciodynamic structures. Simultaneously the features showing the ice dispersal from the West Siberian Plain upslope onto the Middle Siberian Uplands were found (Astakhov & Fainer, 1975; Astakhov, 1976).

There are still researchers who think that major sources of the Weichselian inland ice were located in the mountains (Biryukov et al., 1988). This idea can only be partially substantiated for the Putorana Plateau, around which there are small terminal moraines left by short outlet glaciers, and no signs of a local ice cap have been found in the Polar Urals (Arkhipov et al., 1980). The long-distance erratics that occur on the flat summits of the Urals should be connected with thicker Middle Pleistocene ice masses which advanced southwards across the Urals.

The Late Pleistocene features in the mountains only allow us to reconstruct a network of alpine glaciers, which were coalescing in places to form small piedmont ice sheets of the Alaskan type. The north-facing slopes of the Polar Urals are smoothed and striated by unmistakably allochthonous ice masses which were bypassing this narrow range in two southbound streams (Fig. 9). It is significant that terminal moraines produced by valley glaciers are lacking in the striated piedmonts but they appear immediately to the south of the limit of the last inland glaciation (Astakhov, 1979; Arkhipov et al., 1980). This is evidence of a simultaneous culmination of Arctic inland and Uralian alpine glaciers. It seems natural to suggest that the Urals, being a narrow mountainous strip, could never have accomodated

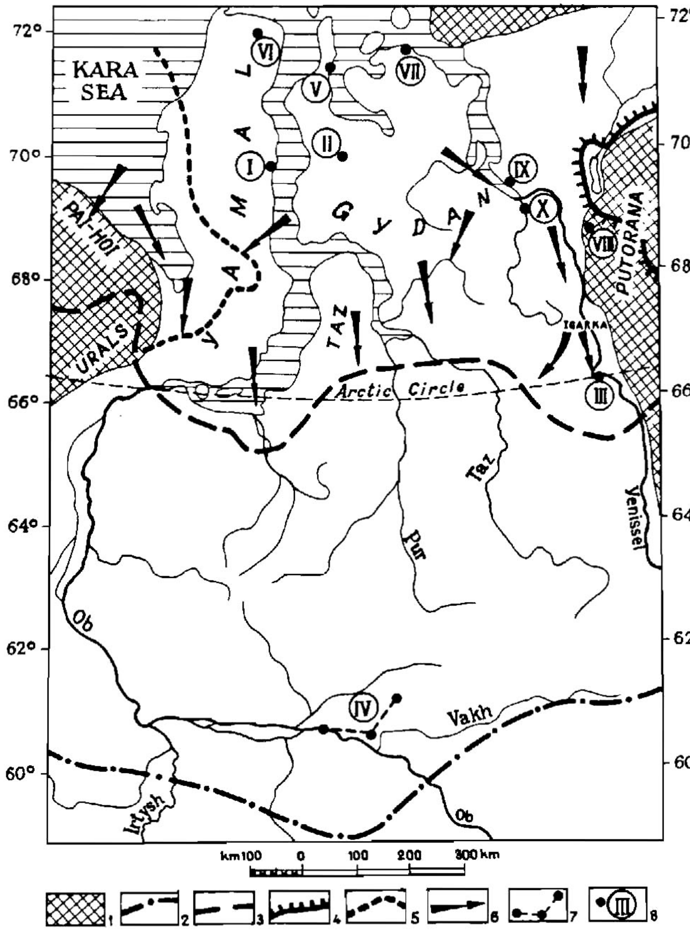


Fig. 2. Location of sections and main glacial limits. Symbols: 1 – hard rocks uplands; limits of: 2 – maximum glaciation (Middle Pleistocene), 3 – last glaciation, 4 – Late Weichselian glaciation of Central Siberia, 5 – West Kara Ice Stream (probably Late Weichselian); 6 – ice flow direction for the last glaciation; 7 – section in Fig. 7 – profile of dated sections in Fig. 7; 8 – radiocarbon dated sections mentioned in the text.

any substantial ice cap of local origin.

The decisive evidence for a thick ice sheet that was expanding from the Kara Sea is the pattern of arcuate glaciotectonic ridges which consist of imbricate slices of underlying soft rocks. The convex distal slopes of the ridges commonly face south or the bordering mountains in the west and east, outlining former ice streams which advanced from the Arctic lowlands upslope to 500 m a.s.l. (Astakhov, 1976; Arkhipov et al., 1980). The paleoglaciological calculations based on this evidence give approximate thicknesses ranging from 2 to 3 km near the ice divide of the last Kara ice sheet.

The Weichselian age of the last Kara glacier is determined by the position of the topmost diamicton over the last interglacial marine formation of West Siberia (Fig. 3). This formation, locally termed Kazantsevo (Sachs, 1953), contains shells of boreal mollusks such as *Arctica islandica* and has conventionally been correlated with the Eemian. The boreal mollusks, which do not currently occur east of the Kola Peninsula, indicate that bottom temperatures of the interglacial sea were 2-4°C higher than the present temperatures of the Kara Sea, which are below zero (Troitsky, 1966).

As the Kazantsevo sediments have only been found within the limits of the last glaciation they are glacially disturbed almost everywhere. Their altitudes are less than +10 m along the Arctic Circle and reach 60-80 m farther northwards. The upper part of the sequence does not show any transitional cool facies, from which it has been inferred that the Kazantsevo transgression occurred before complete recovery of the Middle Pleistocene glacioisostatic depression (Troitsky & Kulakov, 1976). Furthermore, it is quite possible that the Kara Sea shallows were already dry at the onset of the Weichselian inland glaciation because the marine formation in places is capped with peat deposits (Sachs, 1953).

The occurrence of boreal fauna in the upper till may be used as an indication of the last glacial limit (Fig. 2), which is not expressed topographically everywhere. Only in the Ob and Yenissei Valleys can conspicuous hummocky landscapes, entrenched meanders and deflected courses of large rivers be observed, and the relief in the central, lowest part of the Plain is mostly graded by outwash. The available glacial landscapes nevertheless are so fresh that many investigators could not help believing in the Late Weichselian age of the last glaciation (Arkhipov et al., 1980; Grosswald, 1980).

The radiocarbon dates, although numerous, do not help to elucidate the problem of age, if taken indiscriminately. The finite dates occur both below and above the uppermost till, as in Fig. 3. The main reasons for this are widespread redeposition by the upslope moving cold glaciers and contamination by younger carbon during the permafrost thawing. Therefore, only those sites should be chosen for further discussion where the two sources of error can be more or less reliably eliminated. Such sites of prime importance are numbered from I to X in Fig. 2. The radiocarbon dates are selected by this author if they (i) show a consistent temporal succession in a given section, and/or (ii) were obtained from apparently local, preferably perennially frozen material (e.g. birchwood with bark).

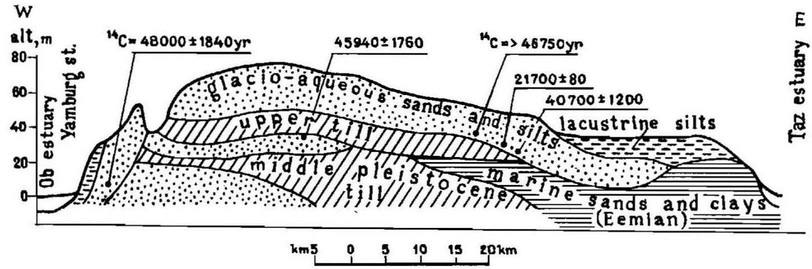


Fig. 3. Cross-section of the Taz Peninsula along the 68th parallel (after A. Lavrov, simplified).

The selected dates provide quite different statistics from those cited by Grosswald (1980) to prove the Late Weichselian age of the last glaciation. Most of the `old` finite dates occur in the sediments overlying the uppermost till, which is obviously at variance with the fresh appearance of the glacial topography. Another discovery by Kaplyanskaya and Tarnogradsky (1976) showed the way out. These geologists found that the frozen tills merge imperceptibly into large stratiform bodies of ground ice with dirt bands, ice flow and other glaciodynamic structures persisting both in the massive ice and in the surrounding diamicton. This led them to the suggestion that the vast flatlands of the Siberian Arctic are composed of buried glacial ice which has survived as a constituent of `primordially frozen tills` within the thick, stable permafrost. Due to the incomplete deglaciation in Siberia the debris-laden basal parts of the stagnant ice sheet only in the south of the permafrozen area have had enough time to melt out and produce the rough glaciokarst topography contrasting with the gently undulating flatlands of the High Arctic (Kaplyanskaya & Tarnogradsky, 1977).

Valuable information elucidating the process of deglaciation in Siberia and the conditions for the burying and preservation of fossil ice has been obtained at the famous site of buried glacial ice in the Yenissei valley called the Ice Hill (site III in Fig. 2). Here thick basal ice has survived in the cores of conspicuous depositional ridges, being buried by flow tills and other ablation sediments (paper 5 in Ch. II). A series of radiocarbon dates from in situ material show that the live ice phase occurred here long before 35 ka (Fig. 4). After the initial disintegration of the ice sheet, probably due to a sea level rise ~ 50 ka ago, the remaining fields of stagnant ice slowly decayed in a very cold, dry Weichselian climate.

The active glaciokarst inversion that has formed the present ice-controlled topography commenced only in the Holocene. Hence the `retarded` deglaciation of Arctic regions should be set off against the rapid deglaciation in temperate environments. Retarded deglaciation may proceed infinitely long if climatic conditions facilitating the preservation and/or growth of permafrost are granted (paper 5 in Ch. II). Sedimentological and stratigraphic inferences of retarded deglaciation are illustrated by the block-diagram modelling the Ice Hill results (Fig. 5). The principal point is that the glacial complex is clearly subdivided into a basal and an ablation series (or subcomplexes).

Only the basal subcomplex, consisting of sheared compact till, detached blocks of interglacial sand and chunks of fossil glacial ice, can be regarded as a stratigraphic signature of the last glaciation, while the ablation subcomplex, comprising flow tills, outwash sands, lacustrine silts and varved clays, reflects the multistaged process of retarded deglaciation. The ablation subcomplex at the Ice Hill contains plant detritus and logs with dates from more than 46 ka to 1200 years ago. The sink-hole sediments (Fig. 4) show that the glaciokarst here was especially active 7-9 ka ago and decelerated about 5 ka ago, when the remnant ice became protected by thick flow tills. This process, which has turned the solid Pleistocene permafrost into the present discontinuous one, is still under way in places where the protective sedimentary mantle has been destroyed by headward erosion or human activity.

The late glacial record in the Arctic

The ablation subcomplex is thick and well developed mainly along the limit of the last glaciation, where the solid permafrost is already replaced by the discontinuous permafrost. In the extreme north, where massive ground ice is especially plentiful (Fig. 9), the basal subcomplex is usually overlain by sandy outwash and in places by late glacial marine sediments. The latter are massive or crudely varved silts and clays with occasional thin shells of *Portlandia arctica* (Troitsky, Kulakov, 1976), which very often contain Arctic foraminifera and no pebbles. Where the clays overlie thick glacial ice a distinct thaw contact can be observed (Péwé & Brown, 1989). This shows that the initial disintegration of the last ice sheet was followed by inundation of the glacioisostatic depression by cold saline water. The bottom temperatures of the late glacial sea must have been zero or less to prevent the basal parts of the stagnant glacier from melting out.

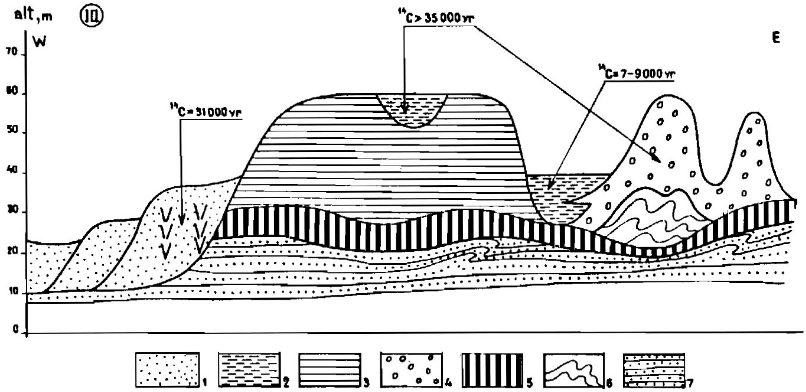


Fig. 4. Idealized cross-section of the Ice Hill (III in Fig. 2) and Igarka environs (paper 5, Ch. II). 1 – postglacial alluvium; 2 – sink-hole silts; 3 – glaciolacustrine rhythmites of the Igarka Lake; 4 – ablation sediments of the Ice Hill; 5 – basal till; 6 – fossil glacial ice; 7 – interglacial alluvium.

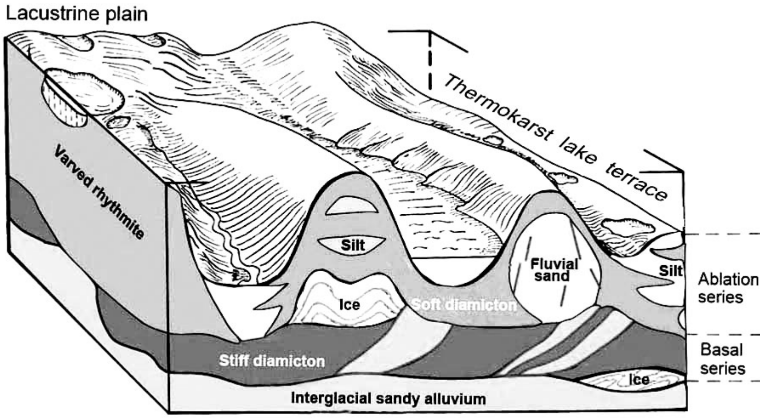


Fig. 5. Block model of the West Siberian glacial complex as derived from the Ice Hill (paper 5, Ch. II).

The age of the late glacial transgression is uncertain, though Troitsky and Kulakov (1976) suggested 14 to 11 ka BP as estimated from the number of varves. This formation is too thick for this, however (up to 15-20 m), and it does not mantle coastal terraces but makes up the cores of low interfluves at 20 to 60 m a.s.l. Mapping geologists later obtained several non-finite radiocarbon dates from the *Portlandia* formation on the Yamal Peninsula. But a stratigraphically more important fact is that the late glacial clays are overlain by thick washed sands with peat lenses, the latter yielding radiocarbon dates ranging from 25 to 40 ka BP (M. Dolotov et al., unpubl. report, 1981). All this makes an Early-Middle Weichselian age for the late glacial marine formation more likely. Another point against a younger age for the *Portlandia* strata is the occurrence of terrestrial sediments dated 30 ka BP and less atop the washed sands, to be discussed below.

As for the low terraces which occur in places along the Kara Sea coast at altitudes of 6-11 m, they do not as a rule bear any signs of marine origin (Troitsky & Kulakov, 1976).

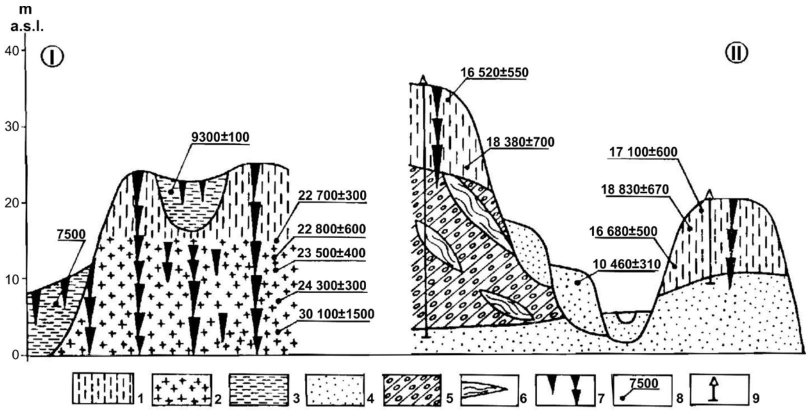


Fig. 6. Arctic formation of icy loess at localities I and II (Fig. 2). I – northern Yamal (after Vasilchuk et al., 1984). II – northern Gydan (after Bolikhovsky, 1987). Symbols: 1 – loess-like silt with horizontal veins of segregated ice; 2 – same with thin peat seams; 3 – sink-hole laminated silts; 4 – alluvial and marine (?) sand; 5 – frozen till; 6 – fossil glacial ice (perceived by Bolikhovsky as marine ice); 7 – syngenetic ice wedges; 8 – radiocarbon dates, yrs BP; 9 – borehole.

The absence of Holocene raised shorelines is another indication of a very early isostatic recovery of the Arctic lowlands. A postglacial marine terrace has been described only on the western coast of the Yamal Peninsula, where it is 8-10 m high, On the whole, the late glacial sediments fall into the following groups: (i) marine silts and clays of the *Portlandia* transgression; (ii) fresh-water varved clays and silts deposited in lakes blocked locally by dead ice; (iii) washed sands with peat and plant detritus, both fluvial and windblown; (iv) laminated silts and sands with wisps of diamict material deposited in thermokarst sink-holes; (v) loess-like silts with ice wedges and peat lenses. The waterlain sediments as a rule yield non-finite or `old` finite radiocarbon dates, while the wind-blown sediments are usually 30 ka old or younger.

The most important stratigraphic feature is the pale brown silt with long syngenetic ice wedges, capping locally the sandy plains of the Yamal and Gydan Peninsulas. This was described by Vasilchuk et al. (1984) and Bolikhovsky (1987), who presented a series of finite radiocarbon dates (localities I, II and V in Fig. 2, Fig. 6). The oxygen isotope ratio in a thick ice wedge from section I ranges from -21.4 to -24.8 per mil, in contrast to values of -13 to -15 in modem surficial water, which allows Vasilchuk et al. (1984) to infer that annual air temperatures 30-22 ka ago were 4-9°C lower than at present, and to conclude that no inland glaciation could have taken place since 30 ka BP. A similar succession of radiocarbon dates, 30,200 ± 800, 25,100 ± 200, 21,900 ± 900 years BP, has been obtained from apparently the same formation near the sea level at locality V, Fig. 2 (*ibid.).* A thinner loess-like silt mantle with mammoth bones at locality II (Fig. 6) gave radiocarbon dates of about 16-18 ka BP (Bolikhovsky, 1987).

The above authors without much argument refer to the loess-like silts as waterlain sediments, but Bolikhovsky still notes that this formation is quite analogous to the famous ice-loess Yedoma Formation of the eastern Siberian Arctic, which contains huge ice wedges and numerous remains of large mammalia. The formation of icy loess is highly characteristic of all non-glaciated Arctic lowlands, especially those of Beringia, where it is an obvious counterpart of the European loess and bears all the signs of aeolian origin (Tomirdiaro, 1980).

The aeolian nature of the West Siberian Yedoma is evident for anyone familiar with such sediments from its mantle-like occurrence on various terraces (II, Fig. 6) and its uniform grain-size composition. It is highly improbable that any aqueous agent could deposit more than 20 m of coarse silt without any sand or clay seams, whereas the thin peat seams present in section I (Fig. 6) are quite consistent with frozen tundra-steppe environments where wind-blown silt could accumulate in local wet depressions as well. Slow, unperturbed sedimentation is evident from the numerous horizontal veins of segregated ice which formed annually at the base of the active layer. These veins, upturned by growing vertical ice wedges, are clearly seen in the picture of section I taken by Péwé, a known authority on the Alaska and Yakutia loesses, who believes that the West Siberian Yedoma is typical aeolian loess (Péwé & Brown, 1989). It is significant that the icy loess sequence continues below sea level (I in Fig. 6), which means that icy loess probably accumulated also on the dry sea shelf, as is known in Beringia.

The deep Holocene sink-holes with minor ice wedges (I in Fig. 6) signify a sharp climatic amelioration ~a 10 ka ago. A frozen mammoth carcass dated to 10,000 ± 70 years BP has been found in a river terrace 7 m high near locality II (Sulerzhitsky, 1995), and a close date of 9,540 ± 50 years has been obtained for a beaver dam at locality IX, Fig. 2. The latter is too far north for the present beavers. This dam rests upon a fluvial sequence of a gully with the date of 15,300 ± 200 years BP near the bottom (Kind, 1974), which suggests a resumption of fluvial activity about 15 ka ago, correlating with a similar event in the southern part of West Siberia.

The above radiocarbon dates from the ice-loess sequence of the Arctic tundra correlate very well with the dates of 31-32 ka BP obtained from twigs with bark at the base of the aeolian sand overlying the Second alluvial terrace of the Yenissei near the Ice Hill (paper 5, Ch. II). Farther northwards, but still within the boreal forests with discontinuous permafrost, the sink-hole silts of the Igarka Permafrost Station contain stumps and logs dated from 35 ka BP to non-finite (Kind 1974, see also Fig. 4). All these apparently reliable dates are positioned above the latest till and unaccountable for if the Late Weichselian age of the last glaciation is assumed.

The following radiocarbon dates are worth mentioning because they have been obtained from the northernmost sites. A frozen log taken by this writer from washed sand on river Hole-Yaha, site VI in Fig. 2, is dated to 35,320 ± 570 years BP (LU-1137), while sink-hole silts at locality VII, containing a solifluction layer with *in situ* mammoth bones at depth 5 m, yielded the dates 30,100 ± 300 (GIN-3742) from a bone and 29,700 ± 270 years BP (GIN-3743) from the enclosing silt (Kaplyanskaya et al. 1985, unpubl. report). All three dates are indisputably positioned above the uppermost till.

The most controversial site in the Arctic is X, Fig. 2, on river Malaya Heta, where radiocarbon dates of 35,000 ± 900, 39,100 ± 1000, 40,300 ± 800 and 43,500 ± 700 years BP have been obtained from peat seams beneath diamicton interfingering with varve-like clays (Kind, 1974). These dates made Troitsky (1967) believe in a Late Weichselian ice sheet which dispersed from the Putorana Plateau westwards onto the Gydan Peninsula. However, this idea is refuted by the glacial landforms pattern (Arkhipov et al., 1980) and by till fabric measurements (Sukhorukova & Gaigalas, 1986) which unanimously show an SSE direction for the last ice stream along the Yenissei (Fig. 2). Moreover, the Malaya Heta limnic clays with a thin diamict layer overlie a flat alluvial terrace less than 30 m a.s.l. All this leads one to suggest the same story as at the Ice Hill (paper 5, Ch. II), i.e. the limnic clays and diamictons could have formed as parts of the ablation series in the course of retarded melting of a stagnant ice body in the Middle Weichselian or in the Holocene.

Such a possibility is consistent with the geology of site VIII, Fig. 2, where a date of 34,300 ± 500 years BP has been obtained for a thick varved formation filling a montane valley at altitudes 300-340 m. This sequence, capped not by till but outwash gravels, accumulated over 6-7 thousand years in an ice-dammed lake (Golbert et al., 1971).

Thus, the available fragments of the sedimentary record give us a notion of the main events in the Arctic after the disintegration of the last ice sheet. First there was a marine ingression, probably contemporaneous with the large englacial lakes, such as the Igarka Lake along the Yenissei (paper 5, Ch. II), both events being beyond the range of radiocarbon dating. Then thermokarst sink-holes and small peat bogs with dates of 44 to 35 ka BP were formed simultaneously with outwash deposition caused by a growing rate of the dead ice melting. After this, about 31-32 ka BP, the climate became much drier and colder, with the growth of ice wedges and pronounced aeolian activity. The aeolian processes predominated at least up to 16 ka BP (Fig. 6), and the most active thermokarst process started 10 ka ago, when boreal forests appeared beyond the 70th parallel, i.e. much farther than their present limit.

Sediments of the periglacial zone

Due to the general northward slope of the Plain the issue of the proglacial drainage has always been a crucial point in paleoglaciological reconstructions. After the existence of a former ice dam across West Siberia had been established (papers 1 and 2, Ch. I), the hypothesis of a huge proglacial lake became a permanent topic of the discussion. Fragmentary sequences of limnic sands and silts were connected to postulate a system of fresh-water reservoirs reaching up 120-130 m a.s.l. (the Mansi and Pur Lakes by Volkov). Such reservoirs are thought to have been parts of a great transcontinental system of proglacial drainage which connected Lake Baikal and the Mediterranean (Volkov et al., 1978; Grosswald, 1980). This writer has been trying for a decade to find geological evidence in support of the idea, but has failed.

First, no topographic signs of fresh shorelines have been found along the prescribed isohypses on the eastern and western borders of the Plain, there being only gentle slopes with thick loess and sandy dunes. Moreover, this obviously aeolian mantle can be easily traced down onto the central lowlands with altitudes 40-60 m and up to the Kazakh loess plateaus with altitudes more than 200 m. Volkov (1971), who amply demonstrated the aeolian nature of the silts and sands mantling the south of the Plain, still maintains that the Upper Weichselian subaerial association thins out to the north being replaced by aqueous sands and silts deposited in large proglacial lakes (Arkhipov et al., 1980).

To solve the problem of spatial interrelations between subaerial and aqueous environments a special investigation was carried out in the lowest central part of the Plain, where the postulated lake must have been the deepest – up to 80-100 m (IV in Fig. 2). The results have been published (Astakhov, 1989) and boil down to the following.

The bottom of the would-be proglacial lake is a stepped swampy flatland with numerous oval, circular and doughnut-shaped knolls 5 to 10 m high and up to 1 km across. The surficial sediments are mostly pale brown, non-stratified loess-like silts, but the knolls are commonly built of sandy rhythmites. Natural exposures along the transverse length of the Ob show a very complicated facies composition of the Upper Pleistocene succession with an apparent thickness of 20 m (see figures in Astakhov, 1989).

The most conspicuous member is a pale sand-silt rhythmite which forms topographic highs, attaining a thickness of 15 m in places but pinching out abruptly in a lateral direction where it is replaced by a brown, at the base greenish, loess-like dusty silt which tends to be the thickest in the inter-knoll lows. The silts contain only very rare pollen grains of *Artemisia* and *Ephedra* but pollen of arboreal *Betula* occurs in the limnic rhythmites (Arkhipov et al., 1980). The loess-like silts contain several horizons of ice-wedge casts and local paleosols with hones of horses, mammoths and reindeer. Such horizons can be traced from the silts into the rhythmites, tying these two different types of sediment up into a loess-limnic association which covers not only the low alluvial plain but the adjacent uplands as well (Fig. 7). Along the river this formation is underlain by cross-bedded sand and peaty laminated clay not appreciably different from the present floodplain sediments.

Radiocarbon dates show that the bulk of the loess-limnic association belongs to the interval between 10 and 36 ka BP. The thickest paleosol with spruce cones and cryoturbations in the lower part of the succession (Kiryas in Fig. 7) is beyond the range of the radiocarbon method, but there is a thermoluminescence date of 120 ± 16 ka for it (Arkhipov et al., 1980). It is very important that the syngenetic ice-wedge casts propagate from the floodplain alluvium dated 33 ka and older, upwards into the loess-limnic sequence, dated 25 ka and younger (Fig. 7). This means a gradual replacement of fluvial processes by aeolian and, locally, limnic sedimentation in permafrost environments.

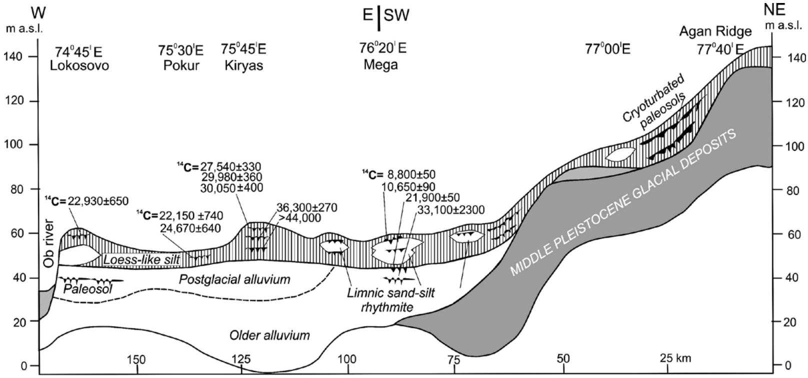


Fig. 7. Subaerial association of central West Siberia along transverse River Ob (profile IV in Fig. 2). Late Pleistocene sandy alluvium penetrated by boreholes in the Ob valley is covered by unfrozen subaerial formation of loess-like silts with paleosols and lenses of sink-hole rhythmites. Note local elevations above the limnic lenses testifying to relief inversion in the course of permafrost degradation. The subaerial formation is traceable on all topographic elements from fluvial terraces to morainic uplands.

The same transition from frozen floodplains to loess-like sediments was described in detail in the Lipovka section on river Tobol, 58°N, by Kaplyanskaya and Tarnogradsky (1974). This section of the Second Terrace contains in situ larch stumps dated 30,700 ± 300, 30,560 ± 240, 30,200 ± 250 BP, and 4 m lower 31,300 ± 800 years BP. The paleosol is disturbed by ice-wedge casts forming polygons 25-50 m across. Such polygons can nowadays develop only north of the 68-th parallel, and perennially frozen floodplains are no longer found south of the Arctic Circle.

There is a number of radiocarbon dates for the alluvium of the Second Terrace in the Ob and Irtysh valleys, ranging from non-finite ones to 24.5 ka BP (Arkhipov et al., 1980; Krivonogov, 1988), while the alluvium of the First Terrace yields dates ranging from 14.5 to 11 ka BP (Zubakov, 1972; Krivonogov, 1988). No alluvial facies is known in the south of West Siberia within the time span 24–15 ka BP, and the first sign of general humidification is the thick paleosol on the Middle Ob dated to 14,880 ± 440 years BP (Arkhipov et al., 1980).

It has been inferred from all this evidence that the central-southern part of West Siberia (between 62° and 56° N) had a Weichselian climate continental enough to provide solid permafrost of the Yakutia type. Since

24-22 ka BP the continentality increased even more which induced the extinction of boreal forests and predominance of aeolian sedimentation even in large river valleys. The wind-blown material was a plentiful source for the sand-silt rhythmites piling up in the sinking ponds which were very shallow, judging by the numerous ice-wedge casts. The frozen loess steppe with thermokarst ponds had apparently developed since the end of the Eemian (Fig. 7) but became dominant only in the Late Weichselian.

The icy loess plateaus sagged to set off inverted sinkhole deposits as the low knolls when the permafrost table dropped in the course of the Holocene climatic amelioration. The stepwise topography of the Ob flatlands (Fig. 7) is due to this thermokarst inversion of the landscape and has nothing to do with normal fill-in-fill alluvial terraces (Astakhov, 1989).

Although varved sediments of a large proglacial lake with a water level up to 70 m a.s.l. have long been known in the Yenissei valley, they are definitely older than 40 ka (Kind, 1974), and merge into the late glacial succession of the Igarka Lake, which in turn was formed before the Second Terrace (Fig. 4). No Lower Weichselian lacustrine strata have so far been dated in the Ob catchment area, but loess-like silts with peat seams are known on the right bank of the Ob near the mouth of the Irtysh, 72 m a.s.l. These apparently subaerial sediments have yielded several non-finite radiocarbon dates and thermoluminescence dates of 64 ± 8, 75 ± 15, 80 ± 13 and 80 ± 11 ka BP (Arkhipov et al., 1987). At any rate, the uninterrupted periglacial successions leave no place for a widespread lacustrine formation of the Upper Weichselian anywhere in West Siberia.

The principally aeolian nature of the Middle-Upper Weichselian sediments is clearly visible when they are traced southwards from the Kara Sea. In the Arctic there are mostly icy Yedoma silts, but farther south the same silts, already thawed, are interspersed spatially with sand dunes and bear numerous signs of recent permafrost disturbances. It is significant that thick sheets of loess-like silts mainly occupy the higher southern and eastern banks of large rivers south of the Arctic Circle (Fig. 8), so that it looks as if creeping sands propelled by northerly and westerly winds had been obstructed by the large rivers, and only suspended dust could cross the valleys to be deposited on the lee bank terrains. This pattern also, by the way, invalidates the reconstruction of the predominant easterlies for the West Siberian Weichselian (Velichko, 1980). South of the area pictured in Fig. 8 this formation changes imperceptibly into typical arid loess with carbonate shots and immobile sand dunes of the Kazakh steppes, the latter indicating wind direction from SSW (Volkov, 1971).

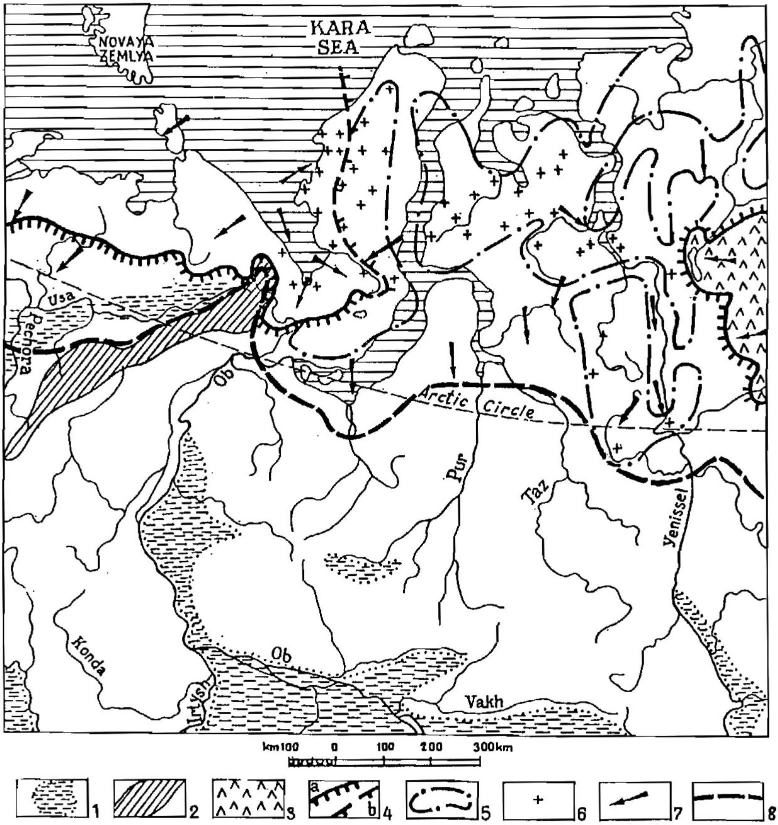


Fig. 8. Weichselian environments of West Siberia. 1– areas of predominant loess sedimentation; 2 – alpine glaciation of the Urals; 3 – the Late Weichselian plateau ice cap with outlet glaciers; 4 – limits of the Late Weichselian inland ice: a – topographically expressed, b – inferred from geological evidence; 5 – fields of Early Weichselian stagnant ice covered by a mantle of frozen sediments, inferred from location of fossil glacial ice and hummock-and-lake topography; 6 – known sites of massive ground ice; 7 – direction of ice flow as inferred from the topography, till fabric and glacial disturbances; 8 – limit of the Early Weichselian active inland ice.

The characteristic thickness of the subaerial formation 15-20 m is persistent everywhere except on the higher bedrock plateaus, where it is usually 2-3 m thick. The maximum thickness of the Middle-Upper Weichselian periglacial formation is recorded in the dry Turgai Valley, where it attains 70-80 m. This valley crossing the high Tertiary plateaus connects the West

Siberian Plain with the Aral Sea basin (Fig. 9), and it is only by this route that the proglacial waters could have drained southwards when the solid ice dam completely blocked the Arctic passages (Fig. 8). Since the present surface of the Turgai Valley reaches up to 126 m a.s.l., Volkov et al. (1978) suggest that this was the level of the last proglacial lake. The available radiocarbon dates 28,800 ± 800 years BP at a depth of 75 m, 19,140 ± 500 at 34 m and 10,800 ± 210 and 11,600 ± 160 years at 4 m (Fig. 9) seem to confirm the Late Weichselian southbound drainage (Astakhov & Grosswald, 1978; Grosswald, 1983). But the southward-transported gravel is only a very thin veneer over the bedrock floor of the valley, which lies at 40 m a.s.l. at most. The bulk of the succession consists of limnic clays, aeolian silts and colluvial diamicts.

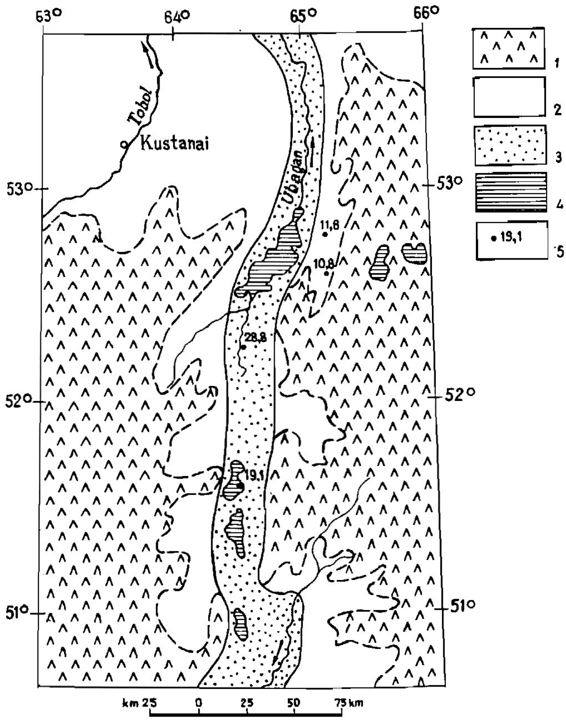


Fig. 9. The Turgai Valley: passage from West Siberia to the Aral Sea (for location see Fig. 1). 1 – higher Tertiary plateau, 180 – 300 a.s.l.; 2 – lower Tertiary plateau, 130 – 180 a.s.l.; 3 – Upper Pleistocene sediments of the spillway; 4 – present salt lakes; 5 – location of radiocarbon samples, age in thousands years BP (see text).

This means that the 126 m level of the present valley bottom can hardly pertain to glacial drainage but is the proper result of a subsequent burial of the narrow spillway by wind-blown and slope-creep sediments. The north-south strike of the valley made it a convenient trap for eastward-moving aeolian material. The Weichselian pluvial episodes may have produced abundant solifluction flows from the thawed clayey slopes. Furthermore, before it became filled in, the Turgai Valley offered a good opportunity for the Early Weichselian proglacial reservoirs to spill over southwards when their levels reached 40-50 m a.s.l.

Discussion

The above data imply that neither the minimum model (Lazukov, 1972; Zubakov, 1972) nor the maximum model (Volkov et al., 1978; Arkhipov et al., 1980, Grosswald, 1980) for the last glaciation in West Siberia amply conforms with the geological facts. The hypothesis of montane ice dispersal centres is mistaken in overlooking the former solid ice dam across the Plain. The pattern of glaciotectonic features concentric to the Kara Sea coast, the till fabric and the till overlying the Eemian marine formation even in the centre of the Arctic lowlands (Fig. 3) leave no alternative. The hypothesis of a thick Late Weichselian ice sheet should be rejected either. Since the latter is the most extensively elaborated one (Grosswald, 1980, 1983) here is the list of the main arguments against it:

– The most reliable non-finite and `old` finite radiocarbon dates are clearly positioned above the uppermost basal till and fossil glacial ice.

– No signs of a large proglacial lake within the range of the radiocarbon method have been found south of the last glacial limit.

– Instead, a thick subaerial formation with loess sheets, paleosols, ice-wedge casts and remnants of terrestrial mammalia is widespread over the West Siberian Plain, including the glaciated area.

– There are no raised Holocene shorelines on the low Kara Sea coastlands.

– Instead, much older marine sediments with late glacial fauna are present.

– Two pre-Holocene alluvial terraces with mammoth remnants occur in the river valleys of the glaciated area.

The new facts confirm the classical view that the last glaciation in West Siberia is older than that in Europe. Sachs (1953) maintained this opinion chiefly based on the numerous finds of mammoths above the latest till, which are very rare in Europe. The radiocarbon dates, if selected carefully, seem to be in full agreement with this. I would like to stress that the radiocarbon dates need not be the crucial evidence. My main argument is: *it is highly improbable that such a long succession of postglacial events could have taken place during the last 10-15 ka.*

The model of the last glaciation suggested herein (Fig. 8) is empirical, i.e. its sole purpose is to accommodate the available geological facts. Still, these data allow one to speculate about the last glaciation in West Siberia as follows. The Early Weichselian glaciation probably started with the appearance of ice sheets on the dry Kara shelf, more likely in the north-eastern part of the present sea. Surficial ice was growing on a soft substrate which had been already perennially frozen. The growing ice sheet was deflecting the West Siberian drainage routes to the west. After the ice sheet had reached the front escarpment of the Polar Urals, several valley proglacial lakes with water table up to 70 m a.s.l. appeared. These drained via the Irtysh and Tobol valleys, and the Turgai spillway down south into the Aral Sea (Fig. 9).

The end of the Early Weichselian ice sheet was probably caused by a decline of precipitation and the global rise in sea level. The initial disintegration was very rapid due to dissecting bays, which propagated from the rising Arctic Ocean (Grosswald, 1983). This seems to be the only option, as recession moraines and other features of frontal deglaciation are absent in West Siberia. Before the isostatic rebound a thick sequence of marine clays with arctic fauna was deposited in the bays, among fields of stagnant ice. The collapse of the ice cap occurred beyond the range of the radiocarbon method, possibly ~ 50 ka BP, or earlier. The remnant ice fields fused together with the thick permafrost soon became covered with a protective veneer of ablation and aeolian sediments, which signified the beginning of retarded deglaciation (paper 5, Ch. II). During the slow isostatic recovery some of the remnant ice fields could have reached the snow line again and partially reactivated, of which there are some indications not to be discussed here.

The proglacial valley lakes seem to have broken through the stagnant ice to resume the present drainage pattern very early, possibly during the regression of the *Portlandia* sea. The alluvium of the Second Terrace had been forming up to 25-30 ka ago, when it was supplanted by aeolian sediments due to progressing aridification. For most of the Late Weichselian the West Siberian Plain, and probably the Kara shelf, was a perennially frozen tundra-steppe with ice wedges growing everywhere and some bush available only along the major rivers. The low flatlands south of the Arctic Circle were dotted with innumerable thermokarst lakes. The whole sedimentary basin belonged to the `periglacial hyperzone` with a uniform continental climate and a mixed tundra/steppe flora and fauna (Velichko, 1973).

The aridity of the environment reached its peak between 22 and 16ka ago, when not even peat seams could form in the High Arctic and fluvial sedimentation seems to have stopped in the river valleys. No substantial warmings are recorded in the periglacial succession except for several more humid episodes with the correspondent deepening of sink ponds and appearance of alluvial facies, mainly before 22 and after 16 ka BP. The permafrost ceased to grow and its table dropped only around 10 ka ago, when the boreal forests advanced northwards and very quickly appeared even in the present arctic deserts of the Taimyr Peninsula (Sachs, 1953). A probable explanation of such a drastic change in vegetation (e.g. the above mentioned beaver dam 9.5 ka BP on the 70th parallel) is that residual boreal forests were available in some nearby river valleys for the entire Late Weichselian.

The most enigmatic question concerns the exact dimensions of the Late Weichselian glaciers. There is not much doubt about the Late Weichselian age of the valley glaciers which drained the remnant ice cap of the Putorana Plateau (Fig. 8), Their end moraines interfingering with varved clays are dated to 19,900 ± 500 and 10,700 ± 200 years BP (Kind, 1974). There is a series of radiocarbon dates from 43 to 25 ka underlying the uppermost till on the western Yamal Peninsula (Gataullin, 1988) but their reliability is questionable. A late Weichselian age of the West Kara Ice Stream (Fig. 8) seems to agree with the submerged field of fresh glacial landforms east of Novaya Zemlya (Dibner, 1970) and with the marine terrace in the south-western corner of the Kara Sea (Sachs, 1953; Gataullin, 1988).

On the other hand, the western Yamal till is overlain by 5 m of loess-like silts with mammoth bones and ice wedges, dated to 13,970 ± 140 and 13,830 ± 260 years BP (Gataullin, 1988). The marine terrace is not Holocene but Weichselian, judging by the available dates of about 15 ka (*ibid.).* It is not quite clear why this ice stream should have advanced upslope along the Polar Urals when the northern Yamal lowlands must have been already free of moving inland ice. The only conceivable explanation is that the West Kara Ice Stream of the Late Weichselian was obstructed in the north and east by very thick stagnant ice (Fig. 8). All this makes the age of the ice stream conjectural, though its glacial features are quite expressive (paper 2, Ch. I).

At any rate it is clear that there was very little Late Weichselian moving ice east of the Urals. The reason for this can be found in the peculiarities of the atmospheric and oceanic circulation, as discussed in detail by Velichko (1980). According to his analysis, the progressive Weichselian cooling increased the volume of the Siberian anticyclonic system, which was gradually supplanting the Atlantic cyclonic system. The diminishing precipitation in the Late Weichselian was being intercepted by progressively more southwesterly terrains. The Scandinavian Ice Sheet was naturally the first to seize all that there was left of the reduced cyclonic activity. Such a veering of Weichselian ice dispersal centres from the high Arctic down to the Atlantic is viable for North America as well.

7. The Mode of Degradation of Pleistocene Permafrost in West Siberia

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Quaternary International, 1995, Vol. 28, p. 119–121.

Continuous permafrost 300 to 600 m thick is mapped in West Siberia beyond the Arctic Circle. To the south, it splits into two uneven layers to form a swallow-tail structure in cross-section. The surficial layer, which is about 50 m thick, occurs discontinuously between 66° and 64° N (in isolated patches - down to the 60th parallel) and is thought to be a product of the Late Holocene cooling. The second, much thicker layer, separated from the surficial permafrost by a 100-200m thick talik, extends to the 60th parallel and must be a Pleistocene relict. Its fossil nature appears in near-zero temperatures and non-gradient curves of the vertical temperature profile (Yershov, 1989).

The relict layer is commonly thought to be of the Late Weichselian age when solid permafrost reached the 51st parallel, i.e. shifted by 1500-1600 km south from its present limit (Baulin et al., 1984). Only Zemtsov (1976) related it to the Middle Pleistocene maximum glaciation. The study of fossil glacial ice contained in the thick discontinuous permafrost near the Arctic Circle has shown that relict permafrost must have survived since Early Weichselian time (paper 5, Ch. II).

There are different ideas about the process which changed the ubiquitous solid permafrost of the Pleistocene into the present discontinuous swallow-tail structure south of the Arctic Circle. The usual reasoning based on temperature profiles, freeze-and-thaw structures in sediments and computer simulations does not allow unambiguous conclusions, particularly if geological data are disregarded. Balobayev et al. (1983) have suggested that the abrupt southern boundary of the modern continuous permafrost reflects a shoreline of a Late Weichselian proglacial lake that fringed the ice sheet along the Arctic Circle. Sedimentological research was discovered no traces of large proglacial lakes (Astakhov, 1989, 1992).

Most investigators believe that the deep talik over the relict permafrost developed during the Holocene climatic optimum. The popular view, ascribing great mobility to the West Siberian permafrost, suggests that south of the Arctic Circle a permafrost layer approximately 500-600 m thick could form and vanish during the Late Pleistocene-Holocene (Baulin et al., 1981). This apparently contradicts geological data about the conservative nature of Siberian permafrost. The above mentioned lower Weichselian fossil ice and fresh-looking (not decayed) frozen wood of infinite radiocarbon age has survived within the permafrost near the surface (paper 5, Ch. II).

These facts show that: (i) no deep climatic fluctuations affected the permafrost since the Early Weichselian except the Holocene warming; and (ii) the Holocene thawing of the Arctic permafrost tens of metres deep (Baulin et al., 1981) must have been local. The apparent discrepancy between the data on great inertia of the Pleistocene permafrost in the Arctic and deep taliks registered south of the Arctic Circle must be attributed to heterogeneity of the degradation process in the different geographical zones (Astakhov, 1990a).

The divergence of geological processes in the Arctic from central West Siberia can be judged from the comparison of the sedimentary records. Facies related to the stable Weichselian permafrost show an arid and very continental climate from the Kara Sea to the Kazakh Steppe. They are thick loess sheets (changing laterally into sand dunes), which are frozen with syngenetic ice veins in the Arctic, and thawed with ice wedge casts in the centre and south of West Siberia.

The only zonal difference appears in the volume of waterlain facies within this principally aeolian formation. In the Arctic, aqueous sediments are mostly comprised of glaciofluvial sands and glaciolacustrine varves underlying the loess and dunes. Within the uppermost aeolian, mostly sandy formation, waterlain sediments are scarce, especially during 22 to 15 ka BP (Astakhov, 1991). Late Weichselian sediments in thermokarst sink-holes have not been described, even along the major river valleys.

The situation is somewhat different south of the 64th parallel where numerous lenses of sandy, limnic rhythmites with dates from infinite to 22 ka BP are encompassed by loess-like silts with paleosols and ice-wedge casts. They are especially abundant along the transverse Ob valley and the crest of the Siberian Hills. Such rhythmites, up to 15 m thick (the `Kolpashevo sand` by Arkhipov et al., 1980), have nothing to do with large ice-dammed lakes, because they:(i) occur at all hypsometric levels from 40 to 120 m; (ii) contain beds of loess, paleosols, cryoturbated horizons and ice-wedge casts; (iii) make oval and doughnut-shaped hillocks of thermokarst inversion topography; and (iv) are never traced laterally for more than a few hundred metres (Astakhov, 1989, 1992). These features indicate that in central West Siberia, thermokarst was active in the Middle Weichselian, probably along large drainage ways. No sink-hole limnic sequences with dates 21-15 ka BP are known south of the Arctic Circle.

The difference between the Arctic zone of solid permafrost and southern zone of discontinuous permafrost became much more conspicuous in the course of the Holocene warming. In the Arctic where thermokarst landscapes have developed south of the 68th parallel, sinkhole sediments are scarce and normally only 2-4 m thick. Thicker limnic sequences can be found mainly over locally exposed bodies of fossil glacial ice where thermokarst accelerates and turns into glaciokarst (paper 5, Ch. II). The oldest peats in the Arctic are dated between 10,000 BP and 11,400±200 years BP (Trofimov, 1986).

These values cannot be taken as the actual age of the permafrost degradation because prior to 9.5 ka BP the climate stayed continental enough to allow growth of ice wedges. Thicker thermokarst sediments with slumped peat and without syngenetic ice wedges were formed later, mainly 9-7 ka BP. The Arctic sink-hole sequences are normally non-continuous because the thermokarst ponds were permanently shifting laterally, leaving behind flat terrace-like surfaces. The latter ones are often erroneously described as marine terraces, although real Holocene marine formations have never been noticed along the Kara Sea coast of West Siberia (Troitsky & Kulakov, 1976).

South of the Arctic Circle, sediments from thermokarst lakes are widespread. The most salient feature of the deep surficial talik between the 60 and 64th parallels is hummocky landscape, sometimes misinterpreted as glacial in origin. The knolls and some ridges, 5-10 m high, built of rhythmically bedded sands and silts with ice-wedge casts, are well-rounded and interspersed with bogs and oval residual lakes of the same dimensions, i.e. several hundred metres across. The all-pervading cryoturbations and lack of pebbles in these sediments along with the loess environment of the inter-knoll lows rule out glaciokarst genesis for this landscape. It must have originated in the course of thawing of surficial permafrost which led to general sagging of the surface and inversion of former thermokarst ponds. The thermokarst inversion was dependent on the initial ice content in the soft clayey sediments underlying the loess cum sink-hole terrain. This resulted in sagging, which produced stepped plains often falsely perceived as alluvial terraces (Astakhov, 1989).

The main problem is the timing of the degradation process. The deep regional talik of central West Siberia means that in the process of the Holocene climatic amelioration there was a moment when the permafrost table dropped and became detached permanently from the layer of seasonal freezing. This moment of transition of the main perennially frozen body into the relict state should be associated with the start of the regional topographic inversion which resulted in the present hummocky loess-thermokarst landscape. This could not have happened before the Pleistocene ice wedges thawed and were covered by organic sediments.

The initial thawing in the south (on the 52nd parallel) has been dated to 11,600 ± 600 years BP (Astakhov & Grosswald, 1978). In central West Siberia, on the transverse Ob, a peat layer which slumped into an ice-wedge crack has yielded the age of 10,650 ± 90 years BP (Arkhipov et al., 1980). These slumped peats are covered by a thin mantle (1-2 m thick) of `warm` loess without permafrost disturbances but with rodent burrows (Astakhov, 1989, 1991). The `warm` loess sheet, between 60° and 64° N, blankets the inverted thermokarst knolls without any signs of pinching out. This indicates that the regional thermokarst inversion occurred after its accumulation. The `warm` loess on the eastern slope of the Southern Urals is associated with sand dunes with Mesolithic artifacts, and on the transverse Ob it contains organics dated to 8800 ± 50 years BP (Arkhipov et al., 1980; Astakhov, 1989).

Therefore, the general thermokarst inversion and transition of the Pleistocene permafrost into its present relict state should probably not be dated earlier than 8000 years BP. The most active sinking of thermokarst ponds on the Arctic Circle is dated by slumped peats on the Yenissei with radiocarbon ages of 9,490 ± 100, 8,150 ± 40, 8,080 ± 50 and 7,850 ± 40 years BP (paper 5, Ch. II). The development of the deep sink ponds with limnic rhythmites obviously predates the general subsidence of the permafrost table south of the Arctic Circle. Such subsidence has not occurred north of the Arctic Circle.

Thus, the available geological evidence leads to the conclusion that the process of degradation of the Pleistocene permafrost in West Siberia consisted of two stages. The first, reversible stage started when surficial water bodies warmed enough to perforate permafrost by numerous sinking ponds and produce local taliks of various shape. In central West Siberia this happened in the Mid-Weichselian when thick lenses of limnic rhythmites (the Kolpashevo sands) accumulated along the transverse Ob. In the High Arctic only thin seams of peat could form in topographic lows of the perennially frozen semi-desert. The very cold and arid climate of the Late Weichselian led to restoration of solid continuous permafrost all over the Plain.

In the beginning of the Holocene (possibly starting with the Allerød), local thermokarsting spread over West Siberia excluding the High Arctic, where ice wedges still could have developed.

After the Boreal arid phase the second, irreversible stage of permafrost degradation took place south of the Arctic Circle. The climate became so warm that direct solar heating of the rocks was sufficient to produce the regional talik separating the permafrost from the layer of seasonal freezing. The irreversibility of this process is underlined by the surficial permafrost which developed south of the Arctic Circle during the Late Holocene cooling (about 3 ka BP) to form the present swallow-tail structure. This stage never happened in the Arctic where the Middle Holocene heat wave led only to more sink-holes wandering over the frozen flatlands. Although the local thermokarst considerably lowered the permafrost table by lateral shifting of heated water bodies, the Arctic climate stayed cool enough to maintain, throughout the entire Holocene, the connection of seasonal freezing with the Pleistocene cold stored in the lithosphere.

8. PLEISTOCENE PERMAFROST OF WEST SIBERIA AS A DEFORMABLE GLACIER BED

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PERMAFROST AND PERIGLACIAL PROCESSES, 1996, VOL. 7, P. 165–191.

Introduction

Subglacial permafrost is usually referred to as a factor controlled by thermal zonality of ice sheets and, in its turn, greatly affecting glacial erosion and deposition (Sugden, 1977, 1978; Moran et al., 1980, Hughes, 1981). It is often thought that glacial deformation of frozen rocks is hindered by their higher shear strength (e.g. van der Wateren, 1985). If the glacial stress is high enough to overcome the frozen rocks' resistance, the released friction heat would preclude the existence of subglacial permafrost (Hart et al., 1990). As a result, glaciotectonism in frozen rocks is considered as possible, but unlikely (Hart & Boulton, 1991). Formation of thick till sheets is also thought not to be characteristic for frozen glacial beds. However, many years of our field experience in northern Russia with glacial drift up to 300 m thick and pervading deformations of soft bedrock suggest a different approach.

Most of the current ideas in the international literature have been derived from geological data on former ice sheets developing either on hard bedrock or in sedimentary basins of temperate climates. A different case is the world's largest sedimentary basin of West Siberia, half of which is still perennially frozen (Fig. 1, 2). It provides an opportunity to better understand geological consequences of the glacial ice/permafrost interaction during the Ice Age.

This opportunity stems from the well-established fact that the present permafrost of West Siberia is a principally Pleistocene phenomenon at various stages of its adjustment to the Holocene climate. South of 59° N it is completely extinct, but in the northern boreal forests it exists beneath a

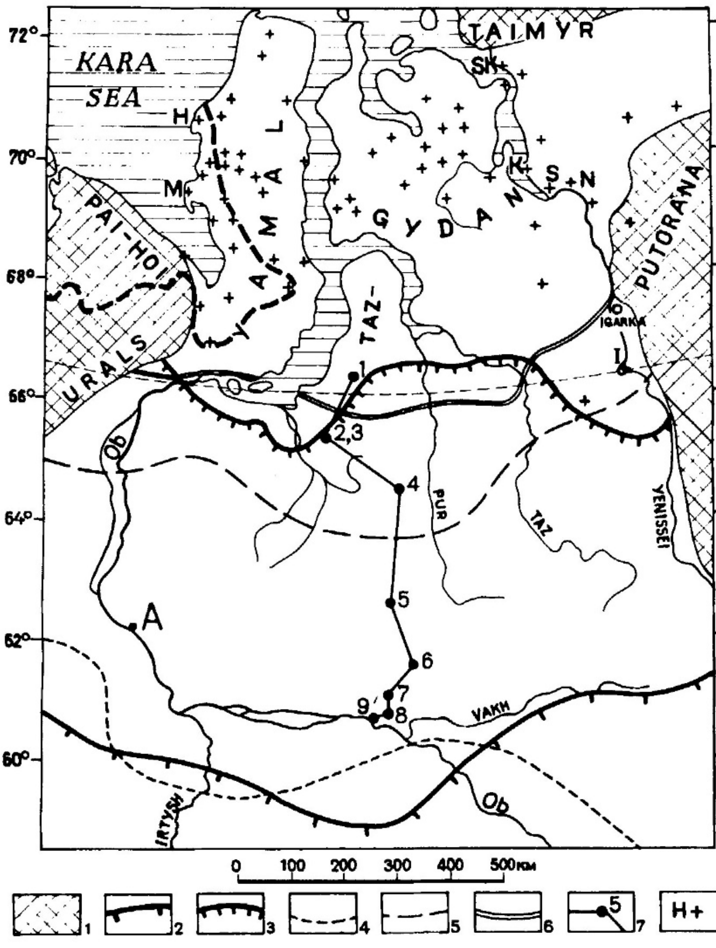

Fig. 1. Location map with principal geocryological phenomena. 1 – Paleozoic uplands. Limits of former glaciations (Arkhipov et al., 1986; Astakhov, 1992): 2 – Middle Pleistocene, 3 – Weichselian; thick broken line outlines the Baydarata Ice Stream. Limits of present permafrost zones (Yershov, 1989): 4 – deep-seated relict permafrost, 5 – discontinuous surficial permafrost, 6 – continuous surficial permafrost. Other symbols: 7 – borehole profile in Fig. 2. 8 – stratiform bodies of massive ground ice according to Badu et al., 1982; Grosswald et al., 1985, and observations by the authors. Locations mentioned in the text: A – Atlym, H – Harasavey, I – Ice Hill, K – Karaul, M – Marresale, N – Nikitinsky Yar, S – Selyakin Mys, SK – Sopkarga.

thick layer of thawed sediments as a non-gradient zone of sub-zero temperatures up to 150-250 m thick. Close to the Arctic Circle the relict layer, retaining non-gradient temperature profiles in boreholes, surfaces to coalesce with the modern surficial permafrost ~50 m thick formed in the course of the Late Holocene cooling. Only well beyond the Arctic Circle, in tundra with annual air temperatures below -8°C, small (1-2°C/100 m) gradients appear in the continuous permafrost 300-550 m thick (Yershov, 1989).

The Pleistocene age of the main volume of the present permafrost is amply evident from old ice wedges, buried or truncated by Holocene deposits (Yershov, 1989), and from fossil glacial ice abundant in the surficial till (Kaplyanskaya & Tarnogradsky, 1976, 1986). Whereas the upper member of the two-layered permafrost of the Subarctic (Fig.2) may be ascribed to the cooling of the last 3 ka, the lower layer must be much older. It is a Pleistocene relict, modified by Holocene climates, even if it occurs on the surface directly beneath the present active layer, as at the Ice Hill on the Yenissei (I in Fig.1). In the latter case non-finite and old radiocarbon dates from alluvial and ablation sediments, overlying the surficial till with large stratiform bodies of fossil glacial ice, show that the ermafrost has survived at least since the Early Weichselian (Astakhov & Isayeva, 1988). The relict layer, which occurs in the area of the maximum glaciation (Fig. 1), may even have formed in the Middle Pleistocene (Zemtsov, 1976).

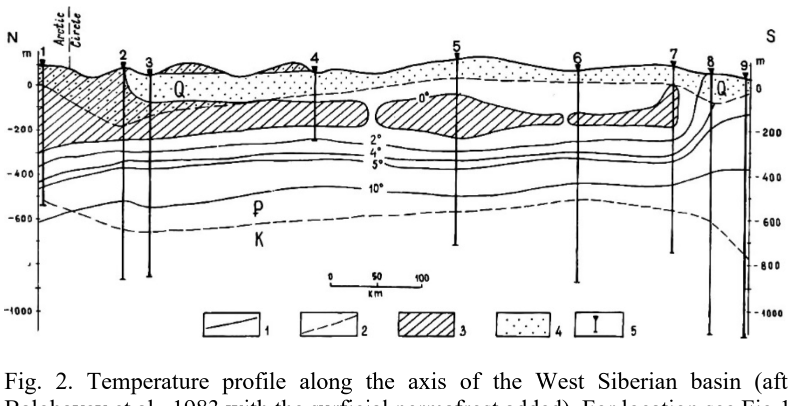

Fig. 2. Temperature profile along the axis of the West Siberian basin (after Balobayev et al., 1983 with the surficial permafrost added). For location see Fig.1. Symbols: 1 – isotherms, 2 – stratigraphic boundaries, 3 – perennially frozen sediments, 4 – Quaternary, 5 – boreholes.

These data, revealing the conservative nature of the West Siberian permafrost, are better understood if the high thermal inertia of the Meso-Cenozoic sedimentary rocks 3-4 km thick, which are 35-40% marine clays, is taken into account. The geophysical evidence is provided by deep boreholes in which observed geothermal gradients are distorted by 40-60% as compared to the normal ones (Fig. 3). The latter occur only deeper than 3-4 km. Upward in the succession the gradients are increased in the northern glaciated part of the basin and decreased in the south. All soft Meso-Cenozoic rocks down to the basement are cooler by 25-30°C than expected for stationary conditions (Kurchikov & Stavitsky, 1987).

The distorted gradients, reflecting the climate history along the north-south profile (Fig. 3), can be clearly subdivided into 3 sets according to the paleogeographical zonality. The reduced gradients in the southern periglacial zone (curves a, b and c) are explained by interglacial and Holocene heating from above which led to the extinction of the Pleistocene permafrost.

Curves d, e and f, belonging to the zone of the Middle Pleistocene glaciation, show reduced gradients only in the uppermost part of the sedimentary column, where the fossil Pleistocene permafrost still persists (cf. Fig. 2). The increased gradients at depths more than 0.4-0.9 km are evidently the trces of the Pleistocene cold wave. The gradients in the Arctic with continuous solid permafrost increase northwards (curves *g, h* and *i*) in

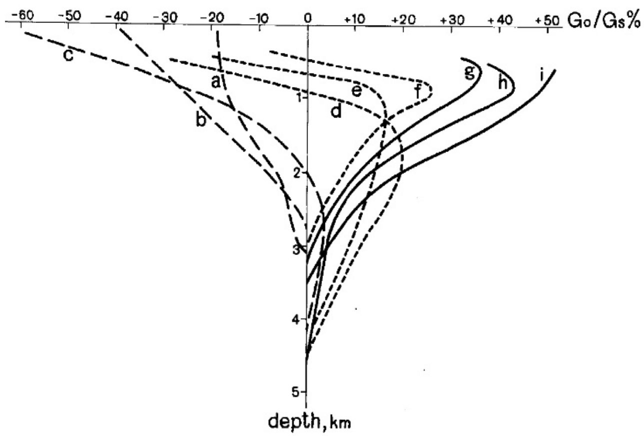

Fig. 3. Deviations of observed geothermal gradients (Go) from stationary gradients (Gs) in boreholes of West Siberia at northern latitudes: a – 56°, b – 58°, c – 60°, d – 62°, e – 64°, f – 66°, g – 68°, h – 70°, i – 72° (after Kurchikov & Stavitsky, 1987).

pace with the progressive cooling of the upper lithosphere in this direction. It is noteworthy that curves *g* and *h* show a slight decrease in the uppermost layer ~0.5 km thick. This agrees with the geological evidence on the present solid permafrost in the Arctic, which, being essentially a Pleistocene phenomenon, used to be much colder prior to the Holocene heat wave. Curve *i* relates to the High Arctic, where the Holocene heating was not sufficient.

The above data imply a very low efficiency of the local geothermal heat flow blocked from above by the cold stored in the lithosphere during the Pleistocene ice ages. Judging by the depth of distortion of the temperature field of West Siberia (Fig. 3), accumulation of the cold started long before the maximum glaciation, interglacial warmings being too short for restoration of the normal temperature gradient. Another implication is that the Pleistocene ice sheets, operating in this sedimentary basin, were predominantly cold-based, as Hughes (1985) has suggested, based on the present permafrost thicknesses.

Apart from the peculiarities of the thermal conditions, the northern West Siberian Basin is characterised by displacements of subsurface Meso-Cenozoic strata, which have been described practically from everywhere within the drift limit (Rostovtsev, 1982), i.e. from the area of presently frozen rocks (Fig. 1). These are innumerable folds, imbricated fold structures, thrust sheets and diapiric protrusions, forming arcuate tracts up to 10-25 km wide and up to 200 km long, accompanied by erratic blocks of soft sediments of various sizes. Soft surficial rocks are almost entirely reworked by epidermal tectonism over great expanses of the West Siberian North.

Thus, in West Siberia large-scale glaciotectonic disturbances coexist with old conservative permafrost. One may argue that the Pleistocene subglacial permafrost, deduced from the present temperature gradients and the ubiquitous deformations of the glacier bed, are incompatible and, therefore, must be heterochronous phenomena. However, geological observations on structures formed beneath the former ice sheets definitely point out that glaciotectonic disturbances in West Siberia developed mainly in perennially frozen sediments. Direct evidence has been found in the Arctic. Indirect evidence is obtainable from the area south of the Arctic Circle, where subglacial permafrost is already extinct but can be inferred from sedimentary and tectonic structures analogous to those in the Arctic. These observations also show that in West Siberia ice sheets were i) eroded mostly by glaciotectonism and ii) deposited mostly by accretion of stagnated debris-laden sheets of basal ice, building up the subglacial permafrost.

Glacial disturbances within present permafrost

The exposed upper part of the present continuous permafrost in Arctic West Siberia contains a variety of structures, originating from the base of the Late Pleistocene ice sheet and still containing remnant glacial ice. The basal ice/sediment complex, studied since the 70s (Kaplyanskaya and Tarnogradsky, 1976), includes such ice flow features as glaciotectonites with streamlined clasts of frozen sediments, changing laterally into normal matrix-supported diamictons of basal tills.

Western Yamal Peninsula

A case of complex subglacial deformations has been studied on the western coast of the Yamal Peninsula near Marresale Polar Station, 70° N (M in Fig.1). Permafrost is only 120 to 300 m thick there (Yershov, 1989) versus 300–500 m thickness in the northern Yamal and farther east. Its surface temperature is -5 to -8°C. The permafrost is underlain by a layer 80 to 150 m thick with negative temperatures but unfrozen due to high content of soluble salts.

The oldest exposed sediments are laminated clayey silts and silty clays with seams of fine sand and plant detritus, labelled the Marresale Formation. Their salinity is 0.4-0.6 weight percent (Gataullin, 1991). They do not contain visible ice inclusions, though the moisture content may attain 33 weight percent, according to a personal communication by Gataullin (1992), who maintains deltaic origin and Eemian age for this formation (Gataullin, 1991). The formation of the Labsuyaha sands up to 50 m thick fills in wide channels incised into the Marresale silts. The sands contain about 38% of interstitial ice (Gataullin, 1988). There are two radiocarbon dates on driftwood from the sands: 31,000 ± 400 and 42,000 ± 1000 years BP (Bolikhovsky et al., 1989).

These interstadial deposits are overlain by the Kara Till up to 20 m thick (Kaplyanskaya & Tarnogradsky, 1982; Gataullin, 1988). The till occurs in two facies. The first facies consists only of apparently local material: clasts of the Labsuyaha sands suspended in the matrix derived from the Marresale silts, or vice versa. The second facies is diamicton built of clasts of local fine-grained material with an admixture of erratic pebbles. When thawed, both facies, as is observable in the seasonally thawed layer on the surface of the coastal cliff, are not appreciably different from analogous sediments occurring in the non-frozen zone of West Siberia. In deeply incised thermocirques, or in special clearings made to remove the present day active layer, the till facies appear primordially frozen (Kaplyanskaya & Tarnogradsky, 1976, 1986) without any traces of subsequent thawing.

The ice content varies widely and may reach 80%, with layers of almost clear ice up to 1 m thick being visible. The icy till has striped appearance due to alternating bands of clear and dirty ice. The bands participate in deformation structures typical for debris-rich ice of modern glaciers (e.g. Lavrushin, 1976; Lawson, 1979). The same deformed banded structures are clearly seen in the thawed till on cliff buttresses, although such steep slopes are unfavourable for preservation of original structures in sediments. Formerly thawed but subsequently refrozen till 5-6 m thick also displays well preserved structures of glacier flow when found beneath former thermokarst ponds. This refrozen till now contains much less ice as compared to the primordially frozen till. The two varieties of the Kara till are separated by sharp thaw boundaries. On the contact of the primordially frozen till with overlying subaerial sandy silts there is another, 30-50 cm thick, band of low ice content, which must be a former active layer.

Thin (in one case up to 4 m thick) lenses of ice-poor diamicton with traces of gradational bedding sometimes overlie the basal till and may be interpreted as flowtills. At the northern end of the exposed cliff small occasional depressions on top of the basal till are filled with varved clay.

The described sediments are mantled by the Baydarata sandy silt 1.5 to 5-6 m thick, which, due to its banded appearance, is often perceived as a fluvial-limnic deposit. Syngenetic ice wedges only 0.5 to 1 cm wide and large epigenetic ice wedges up to 5-6 m high are numerous in this sediment. The wedges, partly penetrating into the underlying till, differ from the till ice by their vertical banding and yellowish colour.

The sandy silt also abounds in upright weeds and roots, which are radiocarbon dated to 13-14 ka (Gataullin, 1988; Bolikhovsky et al., 1989). The weeds *in situ*, the lack of clay seams and the mantling occurrence of the Baydarata silt on diverse topographic elements point to its being aeolian loess, analogous to the Yedoma Formation of East Siberia and Alaska. The Marrresale succession is topped with various Holocene deposits, limnic fills of thermokarst sink-holes and fluvial sediments in particular.

Many investigators, including the present writers (Kaplyanskaya & Tarnogradsky, 1982), have studied glaciotectonic disturbances persisting in the coastal cliff for 30 km north and 3 km south of Marresale. The cliff cuts into an elongated, slightly sinuous bulge 50-60 m high, striking NW-SE.

The most complete geological profile along the entire cliff was obtained by Gataullin (1988, 1991). North of Marresale it reveals multiple open and tight folds accompanied by faults, embracing only the Marresale silts and Labsuyaha sands. The strike of the fold axes and faults is 100-120° SE. The younger formations, truncating the folds at a sharp unconformity, are very thin, or absent at the crest of the bulge and increase their thickness towards adjacent low terrains. The wavelength of the folding is 100-200 to 500 m, the apparent amplitude being 10-20 to 50-80 m. The limbs of the major folds are complicated by disharmonic subsidiary folding and buckling of sand seams. The faults, mostly thrusts, and axial planes of the folds, all dipping south by south-west, indicate that a horizontal stress was applied by an ice lobe of the Baydarata Ice Stream (paper 2, Ch. I, paper 6, Ch. II). The entire folded structure may be perceived as an anticlinorium, most of which has been removed by erosion (Gataullin, 1988).

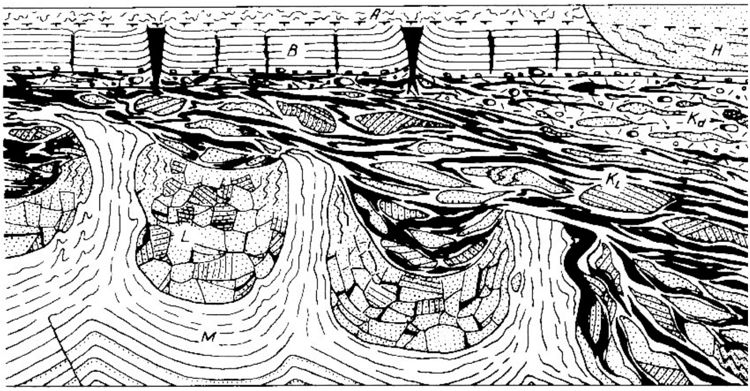

Fig. 4. Reconstructed structure of sedimentary formations exposed in coastal bluff 30 m high at Marresale, western coast of Yamal (M in Fig. 1). M – clayey silt of Marresale Formation, folded and occasionally faulted, in the upper part involved in diapirism; L – the brecciated Labsuyaha sands; K_l – Kara till, local facies; Kd – Kara till, diamict facies with foreign material; B – Baydarata sandy silt with large ice wedges and minor syngenetic ice veins; H – Holocene limnic sediments with collapse structures; A – active layer. Thin lines are sedimentary and deformation structures; thick lines and blackened areas are ice; sands are shown by dots.

The Kara till attains 12 m of thickness, and even more if folded, on the southern flank of the anticlinorium, where it exposed in a cliff 3 km long at the Polar Station (Kaplyanskaya & Tarnogradsky, 1982, Fig. in p. 78).

The further description relates to this section which displays overprinting of structures of different morphological types and generations, as well as various ground ices participating in them (Fig. 4). The disturbances of the first type, represented by the described large open folds with minor thrusts, embrace here only clayey sediments of the Marresale Formation. They are preserved in the lower part of the cliff beneath the superimposed disturbances of the second type.

The second type is the result of squeezing-up of vertically protruding clayey sediments of the Marresale Formation and subsidence of the overlying Labsuyaha sands jointly with lower layers of the Kara till. Some of the protrusions are true mushroom-shaped diapirs, which is evident from their overhanging lateral contacts closer to the top. The apparent height of the diapirs is up to 20 m, often much less. In plan they are elongated with the axes striking 165° SE. The inferred width is 10-20, possibly up to 50 m. The planar bedding of the original clays become progressively more contorted upwards, merging into tectonites with subvertical flow structures, in which the original bedding is almost indiscernible. The boundary between the gently folded clay of the cliff base and the diapirs is not a sharp unconformity, as was erroneously shown in the figure by Kaplyanskaya & Tarnogradsky (1982), but is a transitional zone of pulling out of the strata from the hinge parts of the major anticlines (Fig. 5).

The subsidence basins between the diapirs are filled up with the Labsuyaha sands, the upper part of which is often transformed into sandy local till. The sands of the basins are brecciated and consist of randomly oriented angular blocks up to several metres long. At the margins of the basins the sands are sometimes broken by normal faults into sub-parallel slabs tens of meters long, inclined towards the basins lows. The blocks mostly retain the original bedding of the sands, although sub-vertically oriented in places. On the whole, the observable deformations of the sands are almost exclusively brittle. Only in one case the sand was conformable to an overhanging diapiric contact, making a fold-like conchiform structure.

A removal of surficial thawed material in an intra-diapir basin has revealed thin, 0.5-4 cm, ice veins coating edges of the sand blocks. Within the layer of seasonal thawing such veins are replaced by fine-grained mineral material. The veins are probably a result of partial friction melting of the interstitial ice during the deformation with subsequent regelation.

In a gully at 0.3 km south of Marresale a diapiric overhang has been observed both in thawed and perennially frozen state (Fig. 6). Both cases demonstrate contrasting deformational behaviour of the clay versus sand sediments. Veins of clear transparent ice, probably again of friction-regelation nature, are contained in extension fractures of the sand massif close to a diapir wall. Apart of this, ice seams up to 40 cm thick, sub-parallel to the sand bedding, have been found in some sand blocks. These may be segregation ice veins developed in the Labsuyaha sands prior to their deformation.

Fig. 5. Cross-section of folded clayey silts in coastal cliff 17 m high, at 150 m north of Marresale, western Yamal (M in Fig.1). The hinge parts are pulled out by diapirism to make small tight folds with axes oblique to axial planes of the major folds. Sand bands look white in photo.

The above mentioned local sandy till of the basins contains much more ice, sometimes in the form of thick layers and lenses, than the parental fluvial sand, and because of that shows only ductile (fold) deformations resultant from the subsidence. One such ice layer more than 8 m long and 1 m thick, folded and steeply dipping conformably to a diapir wall (Fig.7), has been observed at 1.5 km south of Marrresale (Tarnogradsky, 1982). The ice has been identified as buried glacial ice on account of its glaciodynamic foliation and angular debris of foreign pure clay. The difference between the local till and the parental Labsuyaha sands is well pronounced only in perennially frozen state. When thawed a boundary between the sandy formations is imperceptible.

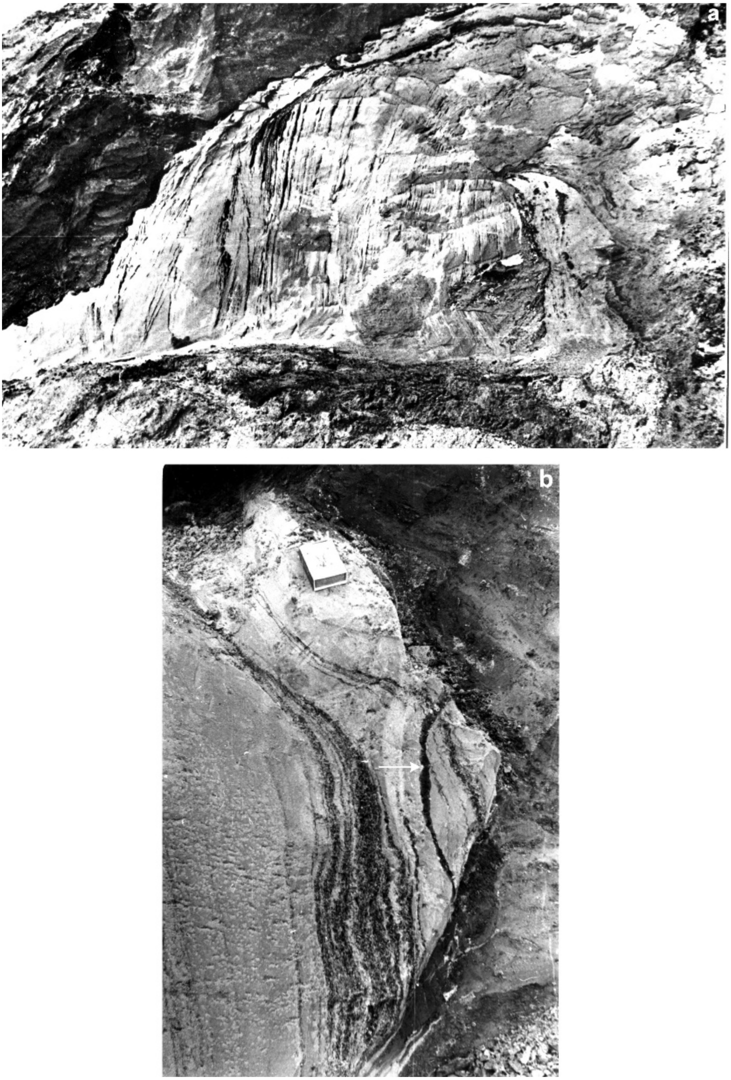

Fig. 6. Diapir contact at 0.3 km south of Marresale, 15 m a.s.l., western Yamal. a – diapir mushroom (dark clay) hanging over brecciated sands with variously dipping remnants of sedimentary bedding; clearing 15 m high viewed from south. b – detail of the same viewed from north (matchbox for scale); sub-vertical stripes are sedimentary bedding in a sand block; arrow shows extension crack filled with dark transparent ice which separates a slightly displaced fragment of frozen sand.

The ice has been identified as buried glacial ice on account of its glaciodynamic foliation and angular debris of foreign pure clay. The difference between the local till and the parental Labsuyaha sand is well Glaciodynamic structures, typical for basal layers of glaciers (Lavrushin, 1976, 1980), constitute the third type of deformations in this section. They may be seen in the local till of the subsidence basins, but best of all are represented in the upper part of the sequence, where both local and diamict facies of the Kara Till lie flat, truncating caps of the diapirs. These small-scale structures of the permanently frozen icy till are alternating bands of debris and ice (Fig. 7, b), gently dipping shear planes, boudinaged sand clasts, recumbent flow folds, crenulations, and so forth. The crenulated dirty ice looks the same as observed by French and Harry (1990, Fig. 3A) in the buried glacial ice of Banks Island. The form of the recumbent folds and their noses indicate an ice flow to NE (Gataullin, 1990, see also Fig. 1).

The distortions of the original glaciodynamic structures by ice wedges penetrating into the till may be considered as the fourth type of disturbances. Besides, collapse structures can be observed in the Holocene limnic sequences of lakes formed atop icy sediments.

The stages and depths of deformation identified in this section reflect varying responses of frozen sediments to stress and temperature fields caused by changing thicknesses and flow patterns of a lobe of the Baydarata Ice Stream. The leading factor was probably advection of warmer basal ice flowing from the deepest part of the western Kara shelf to produce reverse temperature gradients in subglacial permafrost of the Yamal Peninsula.

Thus, the recent glacial history of the site must have commenced with low temperature permafrost formed immediately after deposition of the Labsuyaha sands. The clayey sediments of the Marresale Formation, heavily loaded by a subsequent ice advance, flowed laterally to form the folded compression bulge along the ice margin. Its proglacial position makes a frozen state of the folded sediments very likely. This is also indicated by the generally brittle deformations of the Marresale sands which must have been stone frozen. Occasional ductile deformations in the sands are probably due to a low strain rate and/or locally high ice content.

Incompetent behaviour of the clays is explained by their persistent plasticity in a wide range of negative temperatures (Tsytovich, 1973), which was in this case enhanced by the high salinity of the clays. Later the encroaching ice stream overrode the proglacial `forebulge` and extensively eroded it. The warmer basal ice deformed more readily than the frozen bedrock, which may be the reason for the shallow penetration of disturbances of this stage. The debris-laden basal ice is partly preserved in the intra-diapir basins as the frozen local till with large clasts of the Labsuyaha sands. These must be a product of plucking which never occurs under sliding wet-based glaciers. The frozen Marresale clays beyond the basins might have been involved in basal motion, but only in a thin subglacier layer where their plasticity was the greatest.

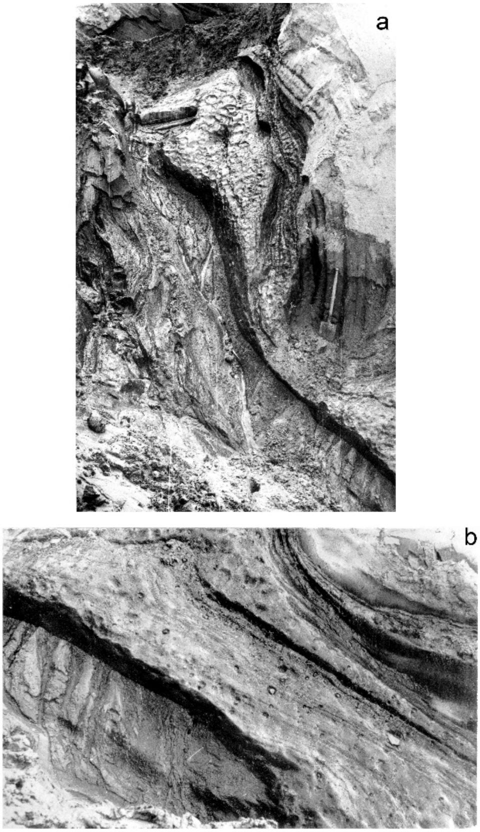

Fig. 7. Glacier ice (pitted surface) plunging along with local sandy till into an intra-diapir basin, at 1.5 km south of Marresale, western Yamal. A – general view; shovel 1 m long is at a vertically oriented block of stratified sand. B – close-up of the downward continuation of the ice bed; ice bands with different debris content and melting out fragments of pure clay are visible; pictured wall is 1.5 m high.

At the end of the first ice advance the insulation by glacier ice and release of deformation heat must have increased temperatures of the frozen bed, in a layer minimum 20 m thick. The diapiric protrusions are the result of this warming and reduction of shear strength of the clays, whereas the sands stayed solid frozen. Their frozen deformation is evident from the ice layers, seams and veins in the surrounding sands, which sank into intradiapiric depressions along with dirty basal ice. The straightforward diapirism without any drag features makes us to assume a temporary stagnation of the lobe, maybe related to the brief retreat of glaciers in the Late Weichselian (Broecker & Denton, 1990). The diapirs may have originated at intersections of large cracks of the stagnant ice and crests of the anticlines. Crevasses are the normal cause of subglacial diapirism which is usually associated with water-soaked sediments (e.g. Sharp, 1985). We have no indications of wet-base conditions at Marresale.

A new influx of basal ice at near-zero temperatures, restored erosive activity of the ice stream, resulting in the sharp unconformity between the diapirs and overlying primordially frozen till with glaciotectonic banding. The diapirs have survived mostly south of Marresale. In other places along the cliff only roots of the diapirs are sometimes visible. Extension strain of moving ice was concentrated in its basal layer, which left the previous structures below the caps of the diapirs almost undeformed.

The final stagnation of the ice stream was followed by melting of the bulk of the clean ice. The debris-rich basal ice has survived, protected by the cold stored in the permafrost, from beneath and by a thin active layer of melted out debris from above. Subsequent accumulation of the loess with ice wedges enhanced the protective quality of the ablation mantle. Holocene thermokarst by short-lived sink ponds produced only localized melting out of the ice/debris mixture. Many melted out patches of the originally icy till refroze again in the end of the Holocene.

The above reconstruction of deformation events is not necessarily accurate. However, one may be certain about two immediate inferences. First, the large disturbances were imposed by glaciers on perennially frozen sediments. Among them all major types of glacial deformations – proglacial, subglacial and dead ice related (Hart & Boulton, 1991) – are discerned. Second, it is the preservation of ice flow structures in melted out but originally frozen tills. This phenomenon was long ago inferred by Lavrushin (1969, 1976) from comparison of ancient tills with live basal ice, but in the western Yamal Peninsula it may be observed *in statu nascendi*. It is clear that in the described situation basal tills cannot be deposited via particle by particle lodging but only via stagnation and exfoliation of entire layers of dirty basal ice with their incorporation into pre-existing permafrost.

Fig. 8. Mylonite-like ice/sand complex steeply dipping in the core of a diapir, at Harasavey, western coast of Yamal (H in fig.1). Formerly ice-cemented contorted sand is seen in seasonally thawed layer in left upper corner. Stick 1.2 m long.

The Late Pleistocene till is also exposed farther north in the coastal cliff mapped in detail by Gataullin (1984). At Harasavey settlement (H in Fig. 1) and 20 km farther north it makes a discontinuous layer 1-2 to 15-20 m thick (Kaplyanskaya, 1982; Tarnogradsky, 1982). In its lower part the till is primordially frozen with varying ice content. The ice occurs in bands and laminae participating in small and generally planar glaciodynamic structures. The upper part of the till with reticulate and massive cryogenic structures is obviously refrozen after melting out beneath small water bodies. The till is again underlain by the glacially eroded Middle Weichselian Labsuyaha sands, which lie mostly below sea level. The visible upper part of the sands is very rich in ice constituting the cryogenic matrix and partly forming concordant ice layers. In places ice is prevailing with the sand particles being suspended in it.

The icy sand and sandy ice make multiple diapiric protrusions, only caps of which are observable in the cliff. The rest of the piercing ice/sand mass has been studied by boreholes (Krapivner, 1986). The sand-ice strata dipping divergently from the diapirs axes are buckled into small isoclinal folds showing an upward flow direction. Cores of the diapirs are in places built of almost pure ice (Fig. 8). The caps of the diapirs are glacially reoriented in a subhorizontal plane with many complicated refold and drag structures (Fig. 9). The ice of the Harasavey diapirs is poor in water soluble salts (0.1 to 0.19 g per litre) and thought to be segregation ice formed from glacial meltwater (Tarnogradsky, 1982). Probably here (and at Marresale) the lower boundary of the permafrost due to insulation by glacial ice at a time before its stagnation went up into the Labsuyaha sands. Below it the sands were water-soaked being an aqueduct for pressurized subglacial water discharging from under the central wet-based part of the Baydarata Ice Stream. With thinning of the stagnated glacier the

Fig. 9. Upper part of the diapir in Fig. 8 subhorizontally reorientated by glacial stress. The protruding ledges is almost pure ice contained in icy sand; the wall is about 3 m high.

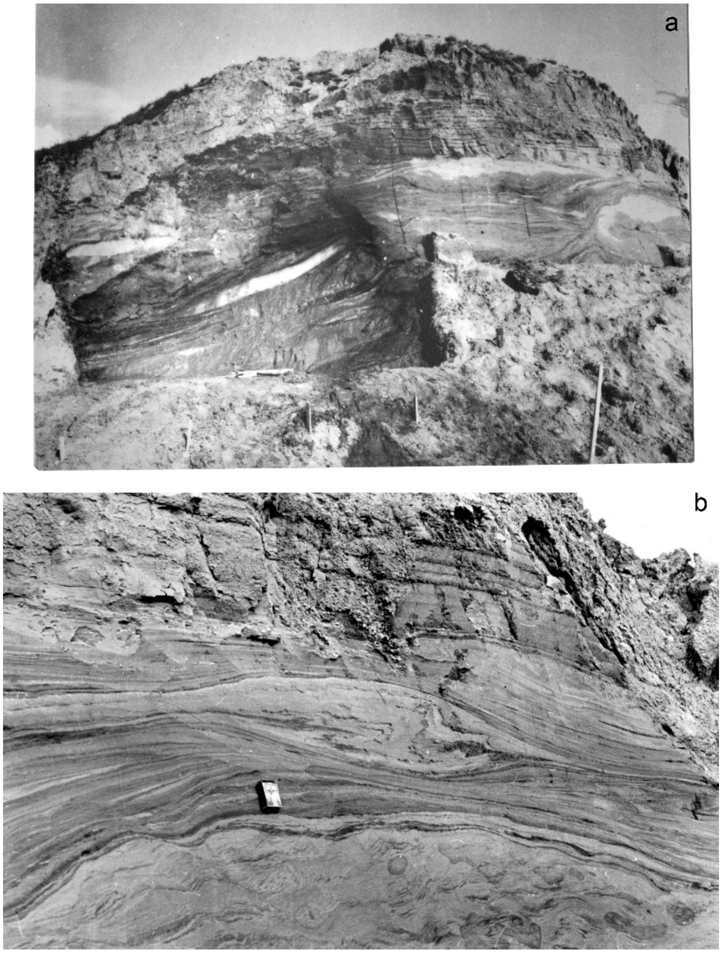

Fig. 10. Glaciotectonically displaced and reorientated material of ice/sand diapir at Harasavey, western Yamal. (a) general view of exposure, vertical lines are 1 m apart; left – frozen sediments: lower part – primordially frozen glaciotectonite with ice bands; milk-white ice band terminates at the base (partly dotted) of a refrozen talik beneath a former lake, in which sand-silt rhythmite seen at the top was deposited; right – same on the seasonally thawed surface of the cliff; (b) close-up of right (thawed) part of the exposure featuring fluidal structure of the glaciotectonite preserved after two thaw events induced by i) the former lake (above dotted line in Fig.10(a) and ii) seasonal thaw on the surface of exposure; bottom of 10(b) shows the upper part of a large sand ball with palimpsest lamination.

frost front moved downwards, which coupled with the influx of pressurized water facilitated the formation of abnormally large volumes of segregated ice. The resultant ice/sand sandwiches, loaded by fractured stagnant ice, produced the present diapirs. Then the readvancing glacier reorientated the diapir caps and deposited the overlying till.

At Harasavey, as well as at Marresale, special clearings show that the melted out sand, even very icy, retains its original intricate structure (Fig. 10).

Lower Yenissei region

Traces of subglacial permafrost are also reliably established in the Lower Yenissei reaches, where the present permafrost is up to 560 m thick and have surface temperatures -7○ to -9○C (Yershov, 1989). The characteristic site is Selyakin Mys (S in Fig. 1), where a network of large, syngenetic in the upper part ice wedges (Fig. 11, 12) and tabular ice bodies of small entirely frozen ponds (Solomatin, 1986) have survived under an epigenetically frozen Zyryanka (Weichselian) till up to 10 m thick. However, in most cases tills are primordially frozen and contain well preserved fossil glacial ice (Fig. 13) up to 60 m thick (Kaplyanskaya & Tarnogradsky, 1976, 1993; Solomatin, 1986; Karpov, 1986; Astakhov & Isayeva, 1988). The exposed tills belong to Late Pleistocene glaciations, although preservation of older primordially frozen tills within the glacial drift filling deep (up to 300 m) troughs is not excluded.

The till sheets with occasional inclusions of buried glacial ice are normally represented by icy diamict beds intercalating with and laterally substituting apparently chaotic assemblages of balls, angular blocks and giant rafts of stratified sands and clays with marine Quaternary and Cretaceous fossils. The matrix of these glaciomelanges (Komarov, 1987) is partly diamict with ice and erratics, partly consists of tectonized clayey sediments. The sand clasts are mostly built of marine sediments with interglacial (boreal) fauna or of stratified drift. The clayey clasts, commonly consisting of marine sediments transformed into allochthonous glaciotectonite, often contain arctic fauna.

The fauna finds is the reason why the entire succession is perceived by some geologists (e.g., Bryzgalova & Bidzhiyev, 1986; Krapivner, 1986) as a marine formation *in situ*. However, most of these sedimentary inclusions are foreign for the area and must have been delivered by glaciers from the Kara Sea shelf and adjacent lowlands (Kaplyanskaya & Tarnogradsky, 1975). The predominant north-south direction of Middle and Late Pleistocene ice streams from the Kara Sea ice dispersal centres is amply evidenced by the pattern of ice-pushed ridges, till fabrics, striations, mineralogy of tills, etc. (papers 1 and 2 in Ch. I). Deep glaciotectonic disturbances of both Quaternary and Pre-Quaternary sediments have long been known in the region (Troitsky, 1975).

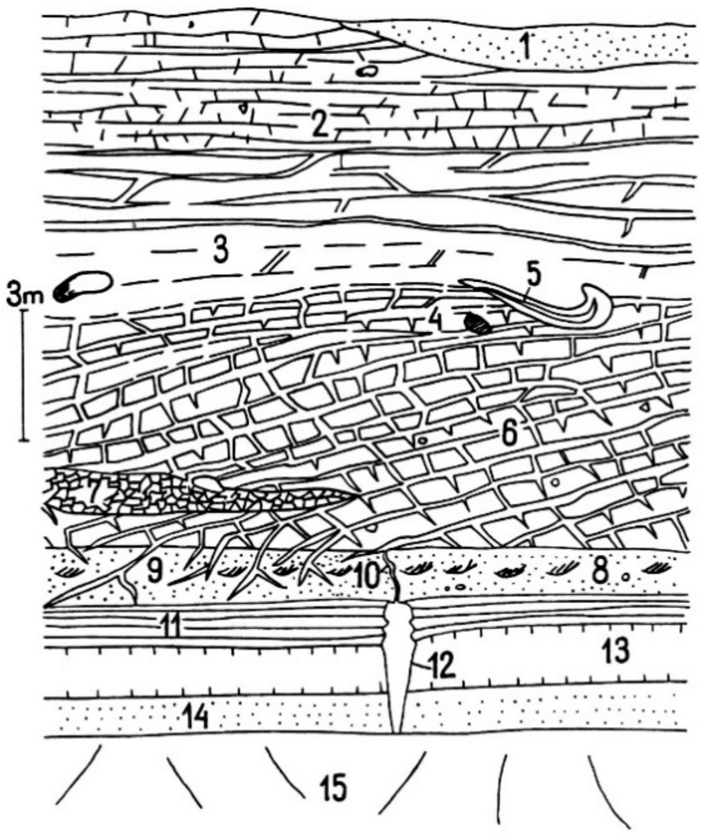

Fig. 11. Field sketch of cryogenic structures exposed in a thermocirque in the upper part of a Yenissei bluff at Selyakin Mys (S in Fig.1). 1 – iceless sandy drift, 2 – diamicton with vague tightly spaced reticular structure, 3 – bedded diamicton, 4 – inclusions of dark iceless clay, 5 – inclusions of laminated sand, 6 – diamicton with widely spaced reticular structure, 7 – inclusions of sandy silt with tightly spaced reticular structures, 8 – massive sand with plant detritus, 9 – same with widely spaced reticular structure, 10 – crack above ice wedge, 11 – laminated sandy silt, 12 – ice wedge, syngenetic in the upper part, 13 – massive transparent ice, 14 – iceless sand, 15 – talus. 2–7 – epigenetically frozen till.

Fig. 12. Ice wedge (centre) penetrating into massive ice (bottom) in sediments underlying Late Pleistocene till at Selyakin Mys Fig.11). Wall is about 3 m high.

The structure of drift terrains built of tills and melange is not completely understood yet. However, most known glaciomelange assemblages have been described as constituting folded and imbricated structures. Such position of allochthonous formations implies at least two stages of subglacial deformation. The first one is longitudinal extension characteristic of non-marginal glacial transport. The second stage is longitudinal compression, particularly typical for ice margins. Both types of deformation must have occurred in the frozen state of the rocks involved because: i) sand blocks coud be detached whole only when the parental sand is solid frozen and ii) ice beds also participate in compression structures.

Folded glacial ice in marginal position has been described at the `Ice Hill` (paper 5, Ch. II, location I in Fig. 1). Upthrusted and contorted debris-rich ice (Fig. 14) is also part of the glaciomelange layer within an imbricated stacking of various Quaternary rocks about 50 m thick at Karaul (K in Fig. 1). Besides, 2-5 m thick slices of diamictons and Kazantsevo (Eemian) sands with shells of boreal mollusks, dipping west at angles 25-35° participate in the imbricate structure. The topmost diamict sheet contains numerous large blocks of the Eemian sands and bodies of glacial ice.

The upthrusted beds are expressed on the adjacent plateau as an east-facing bow of small arcuate ridges indicating a glacial stress from the west as well. Such stress could be applied by the eastern flank of an ice lobe which occupied the Yenissei valley.

The glaciomelange on the Lower Yenissei, as in other regions of West Siberia, reveals the rheological difference between frozen sand and clay. Frozen sands and even sandy silts (if they are poor in clay particles) show almost exclusively brittle deformation, low ice content provided. Ductile deformations predominate in clay and ice-rich sediments.

For example, sand transformed into large-block breccia (Fig. 15, a) takes part in a layer of allochthonous glaciomelange, exposed between two diamict beds at 20 to 31 m from the top of the 54 m high bluff of Nikitinsky Yar (Fig. 1). The detailed description of this key section is available in Kaplyanskaya and Tarnogradsky (1975, Fig. 3). Downwards in the succession (34-41 m from the top) a layer of allochthonous glaciotectonite, composed of mylonised silt and clay with marine fauna (Fig. 15, b), has been encountered. Judging by the meltwater seeping through the talus, the diamict till and melange of this sequence also contain buried ice. The topmost diamict bed exposed in the thermocirques has been observed in primordially frozen state with spectacular recumbent folds in fossil glacial ice (Kaplyanskaya and Tarnogradsky, 1978, Figs. 1, 2, Solomatin, 1986, Figs. 34, 36). When exposed in the layer of seasonal thawing the till reveals numerous clasts of loose sand.

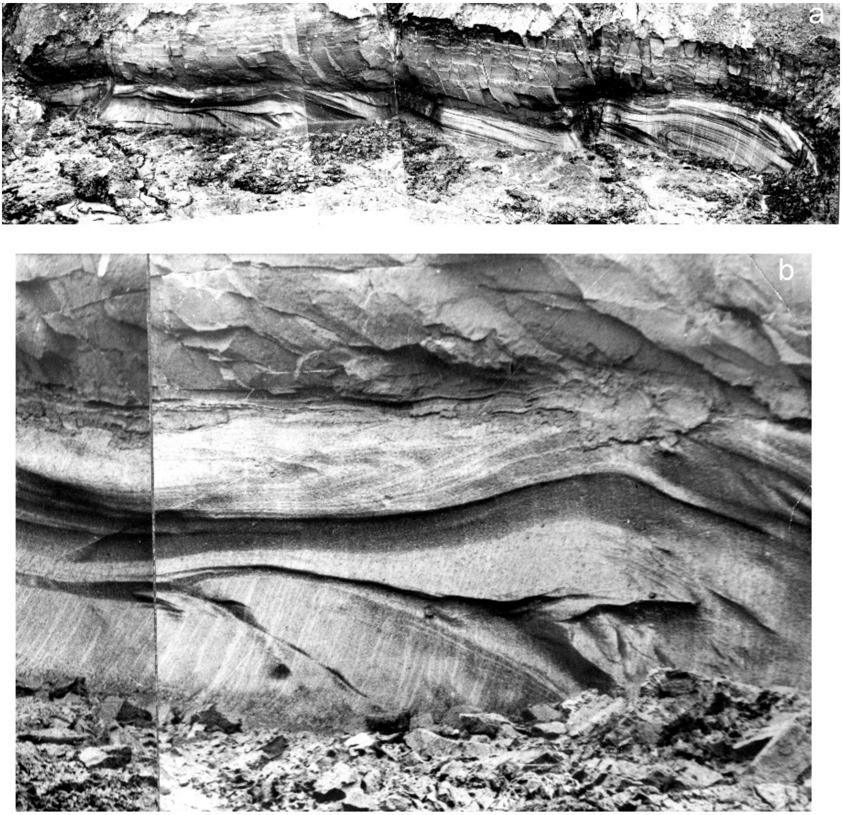

Fig. 13. Relict dirty glacier ice with flow structures buried under epigenetically frozen flowtill at Sopkarga, eastern bank of the Yenissei Estuary (SK in Fig. 1). A – exposure about 20 m long. B – close-up of the left end; rare small erratics are visible.

Fig. 14. Contorted and cleavaged debris-rich glacier ice with small erratics within glaciomelange at Karaul, eastern bank of the Yenissei (K in Fig.1).

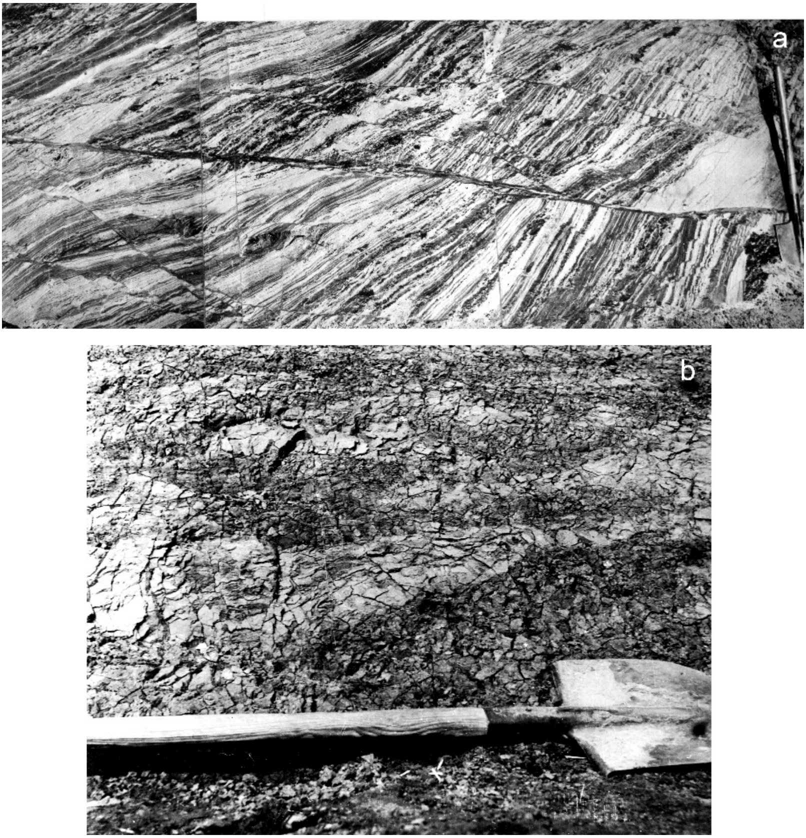

Fig. 15. Contrast deformation behaviour of different rocks participating in glaciomelange at Nikitinsky Yar, eastern bank of the Yenissei (N in Fig.1).
a – large-block breccia consisting of loose sand (light) with silt laminae (dark); shovel 1 m long for scale. b – mylonised marine sediments: dark – more plastic silty clay, light streaks – clayey silt

Glacial disturbances in thawed rocks

South of 64-65° N, where the present surficial permafrost is sporadic, or non-existent, structures very similar to the described above are widely observable in thawed sediments, but it is not always easy to connect them with former subglacial permafrost. The most spectacular evidence of former subglacial permafrost is large inclusions of non-glacial stratified sand within the Pleistocene glacial drift. They range from sand balls several centimetres across up to large stratiform rafts hundreds meters long and have been described throughout glaciated West Siberia from the drift limit (Kaplyanskaya & Tarnogradsky, 1974) up to the northern tip of the Yamal Peninsula (Astakhov, 1981).

The sands, if thawed, are so loose that they obviously must have been detached, entrained and deposited in frozen state. This is especially evident in many cases of long-travelled rafts composed of pre-Quaternary sand with their primary sedimentary structures preserved. One of the largest is the sand slab from 80 to 120 m long resting upon basal till in the right bank of Irtysh river close to the village of Semeyka near the limit of the maximum glaciation (Kaplyanskaya & Tarnogradsky, 1974). The quartz sand with lignite seams, according to palynological and carpological data, originated from a Paleocene formation (Nikitin, 1988), which occurs *in situ* not less than 650 km NE from this place.

However, such erratics, although ubiquitous, is evidence only of frozen sand substrate which occurs not everywhere within the basin. In many places the substrate is composed of various clays and silts, or sand underlain by clayey rocks. In areas of the lithologically diverse substrate with underlying clay the most conspicuous and often large-scale glaciotectonic disturbances are observable.

A very instructive case is presented by the well-studied Atlym disturbances on the right bank of the river Ob (A in Fig.1). At this site the cross-section of glaciotectonic structures has directly been measured in the river bluffs up to 90 m high, providing a detail profile 15 km long. Most of the observed lithological boundaries dip westwards at various angles. The same directions are recorded for numerous overturned folds, thrust planes, imbricated slices and clay injections. The measured contacts of the local sedimentary formations have been extrapolated below the water level using the borehole data (Fig. 16). The resultant generalized profile shows a very complicated zone of deep crumpling of the Palaeogene strata up to 25 km across (Astakhov, 1990b).

This zone is of interest not only because of its size, but also because it occurs at the base of the drift and, being truncated by a basal till with a pronounced unconformity, is not expressed in topography. The basal till, containing slabs of loose Oligocene sand, proceeds over undisturbed Paleogene formations southwards to the limit of the Middle Pleistocene glaciations. Predecessors perceived the structure as basically folded, either ascribing it to unspecified glaciotectonism (Li & Kravchenko, 1959), or remaining undecided about its genesis (Nalivkin, 1960).

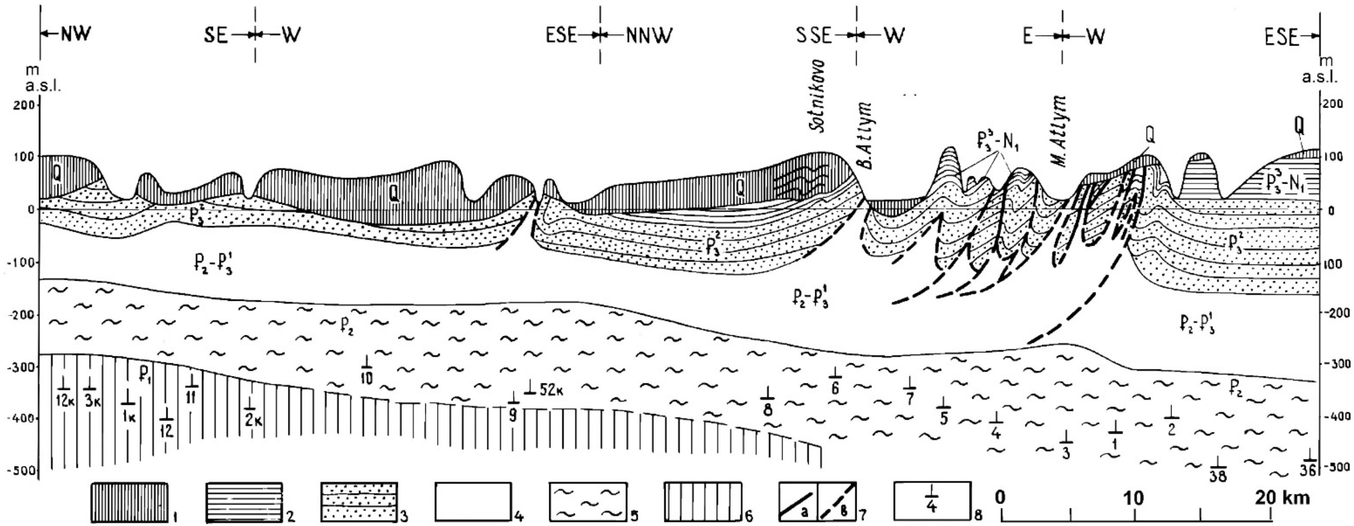

Fig. 16. Geological profile along the north-eastern bank of the river Ob near Maly Atlym settlement (from Astakhov, 1990b). Location is A in Fig.1. 1 – glacial tills and outwash, 2 – terrestrial silty rhythmites, Upper Oligocene-Lower Miocene, 3 – terrestrial sands, Middle Oligocene, 4 – marine clay with siderite lenses, Eocene-Lower Oligocene, 5 – Eocene diatomites and opokas, 6 – Paleocene marine clay, 7 – faults: a – observed, b – inferred, 8 – bottom of borehole with its number.

Our detailed survey has revealed that the most characteristic features here are listric faults growing steeper and more tightly spaced upstream, i.e. downglacier. In this direction simple folds change into tight recumbent ones accompanied by numerous zones of mylonisation and thin injections of the underlying Eocene clay. At the distal end sub-vertically dipping and heavily foliated Paleogene rocks are abruptly replaced by subhorizontal strata lying in the normal stratigraphic order with the Eocene clay positioned 250 m below the surface. The general impression is that the shear strain increasing downglacier was suddenly terminated by an unsurmountable obstacle.

Conversely, in the upglacier direction the structure gradually becomes simpler, with gentle slightly asymmetric folds prevailing and only occasional clayey injections breaking through the overlying Oligocene sand. It is noteworthy that while upglacier the Eocene clay is abnormally thin (about 100 m), at the distal end its thickness may reach 200-250 m (boreholes 1 to 6), the normal thickness being ~ 170 m (borehole 36). This is an indication of extension upglacier and compression downglacier. In borehole 3 steep angles of dip and slickensides have been observed up to the depth of 310 m below the river level (Li & Kravchenko, 1959). Thus, the entire zone of the disturbed sediments is ~ 400 m thick. The underlying Meso-Cenozoic beds down to the Paleozoic basement, which is 2.8 km deep, are only slightly undulated.

For the present discussion the most important feature of the Atlym disturbances is the difference in the deformation style between the various Paleogene formations. This difference, dependent on mechanical properties of the rocks, is amply evident in minor structural complications of the extremely deformed distal part of the zone.

The most competent behaviour is observable in the Oligocene sand which shows no crenulation but only numerous sharp-edged blocks, often displaced and stacked into a pile of `chips` (Fig. 17, top). Sometimes sand blocks are thinly sliced along minor listric upthrusts (Fig. 18). In extreme cases the sand is fracture cleavaged or turned into a typical friction breccia consisting of small angular splinters. Such deformations normally develop only in cemented sandstones, which is not the case here. Therefore, the sand must have been cemented by ice during the deformation. This is confirmed by a 120 m long raft of the same sand within the basal till truncating the distorted Paleogene. The sand raft shows no strong internal deformation, having probably been incorporated into glacial ice before the maximum compression was achieved.

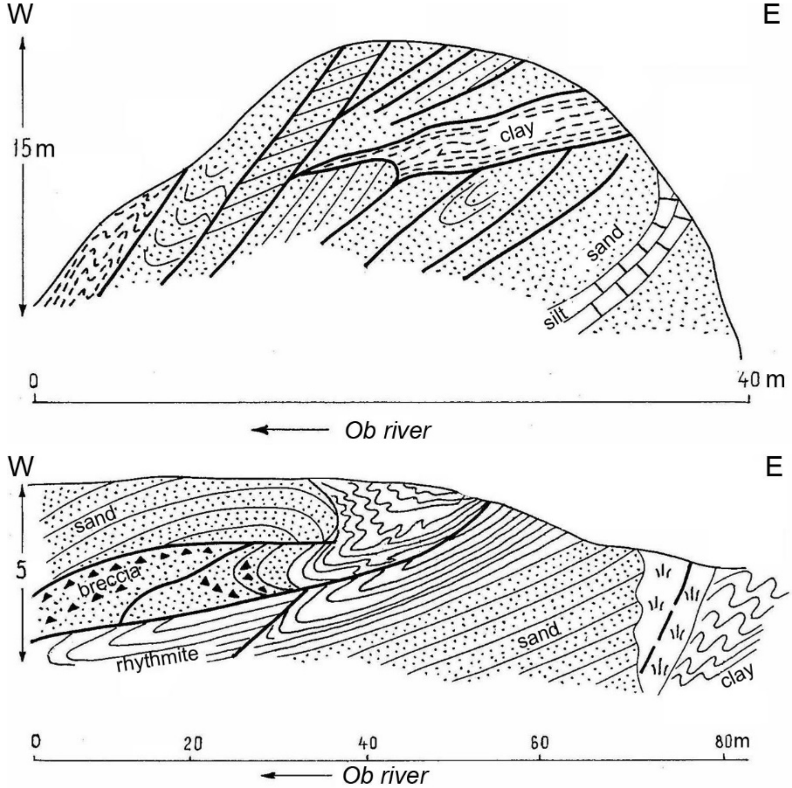

Fig. 17. Tectonised Paleogene sediments observable in river bluffs at 3.5 km (top) and 13 km (bottom) from the eastern end of the Atlym disturbances (Fig. 16). Top section exposes blocks of the Middle Oligocene loose sand chipped and stacked in frozen state with small injections of Eocene marine clay. Bottom section displays a ski-like ductile thrust in folded and crenulated Upper Oligocene-Lower Miocene silty rhythmites; Middle Oligocene sand is not crenulated but brecciated along the thrust plane. The disturbed sediments are positioned at 40-50 m below the base of the glacial drift.

The Upper Oligocene-Lower Miocene silty rhythmites feature typical ductile deformations, namely small harmonic and recumbent folds with crenulated bedding (Fig. 17, bottom). They also must have been frozen as follows from their normal position atop the Middle Oligocene sand. Extreme ductile deformations have been observed in the Eocene montmorillonite marine clay which participates in the above structures only as tabular injections up to 200 m across, or minor clastic dikes 0.3–1 m thick (Fig. 17). Similar features could theoretically develop in thawed clays, but in this case it is very unlikely. First, the internal structure of the injected clay is not just flow foliation but also crenulation cleavage. Second, the contacts with the surrounding sand are mostly flat slickensides with practically no sand xenoliths incorporated into the protruding clay. Third, in several places pulled apart siderite lenses within the clay are bent into small tight folds, which demands a relatively competent medium to translate the huge stress. These and other similar features, revealing ductile non-fluid rheology, tell us that the overlying sediments were pierced by frozen clay.

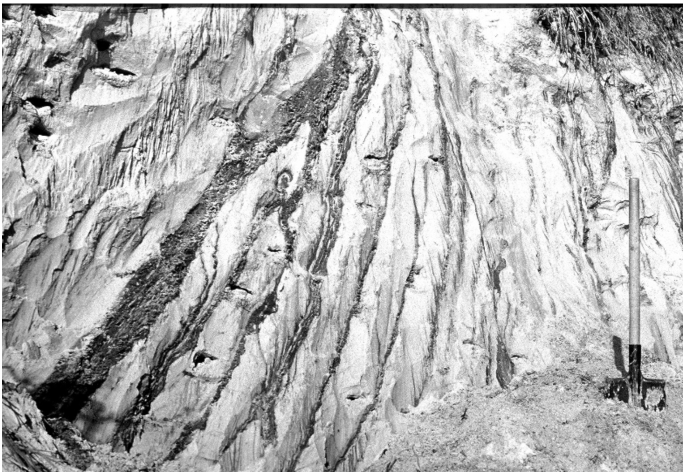

Fig. 18. Middle Oligocene sand with laminae of clayey silt exposed at 8 km upglacier from the eastern end of the Atlym disturbances (Fig. 16). The photo shows thin slices of medium-grained quartz sand upthrusted along silt laminae truncating sand lamination.

The ductile deformations of the frozen clay of the Atlym disturbances is normal because clay always retains combined water, and according to numerous experiments, up to -7°C shows dynamic viscosity by an order lower than pure ice. Rheological behaviour of frozen clay is more time dependent (elastoviscous) than that of crystalline ice, shear strength of the former being by ten times reduced if a prolonged stress is applied. After the initial resistance is overcome, stressed frozen clay would creep faster than pure ice (Tsytovich, 1973). An important consequence is that the maximum deformation within the glacier/frozen clay couplet would occur beneath the ice/rock interface with resultant `floating` of the glacier upon a clay pillow.

Thus, the mode of deformation of the Paleogene sediments, quite different from their behaviour in the present thawed state, is the evidence of former permafrost at least 300 m thick under a Middle Pleistocene glacier at about 300 km from the drift limit. An interpretation of this case in terms of glacial geology is presented in Fig. 19. Stage 1 shows a glacier advancing over perennially frozen Paleogene sediments with shear strength sufficient to support not very thick ice. With the ice thickness increasing (stage 2) the shear resistance limit is first overcome in clay formations closest to the sole of the glacier. This means that the dynamic sole of the glacier (a plane along which the maximum deformation occurs) starts to split with a part of the sliding component being transferred into the frozen clay, which is capable of faster deformation than ice itself. The frozen overlying sand provides a competent dam downglacier, where it is thicker, and gets crumpled and broken into slabs upglacier, where it is thinner.

Stage 3 shows further growth of ice thickness with the dynamic sole being fully established within the frozen clay upglacier. The result is downglacier flow of sheared frozen clay supporting the overburden of the glacier coupled with the frozen sand. Breaking of the competent sand roof reaches a limit somewhere downglacier where the sand is thick enough to resist any glacial stress. This means a drastic reduction of the effective velocity of the clay flow and its thickening against the sand obstacle with steep listric thrusts developing in the ice/sand couple. This would lead to a spring of the dynamic sole upwards into pure ice downglacier with another minimum of shear strength. Then the glacier advances as a whole block with two dynamic soles: within the frozen clay upglacier and within the ice downglacier. A very likely result of the latter position is stagnation of the basal dirty ice downglacier, leading to deposition of frozen tills, as suggested by Lavrushin (1976, 1980).

The model in Fig. 19 accounts also for major glaciotectonic features, which in a sedimentary basin do not need thawed bedrock or an ice margin to develop. In the case of the Atlym disturbances a marginal environment is very unlikely because of the great width and depth of the distorted zone, which would demand large ice thicknesses.

The main prerequisite is rheologically contrasting sedimentary formations which are available everywhere. Moreover, an originally diverse subglacial topography is also not necessary for large-scale deformations of the substrate and entrainment of large erratic blocks. The springing behaviour of the dynamic sole of the glacier, reacting to the different rheologies of the substrate, is an ample reason for detachment and distant transportation of large slabs of sediments. For better understanding erosional and tectonic activities of a cold glacier it seems useful to consider a possible split of its dynamic sole which may (sequentially or simultaneously) occur in i) englacial, ii) basal, and iii) subglacial positions. The different positions of the zone of maximum deformation within the ice/permafrost couple accounts for large-scale shaping of landscape by glaciers, i.e. deepening of upglacier depressions by excavation and building up of downglacier elevations by accretion of till sheets.

Discussion

Paleoglaciological reconstructions of former ice sheets usually take into account 3 main forms of ice motion: i) basal sliding of warm-based glaciers over a rigid bed, ii) internal deformation of basal ice of cold-based glaciers upon a rigid bed (Drewry, 1986), and, especially lately, iii) deformation of thawed glacial drift with high pore-water pressure beneath glacier ice (e.g. Alley, 1991). In West Siberia, as the above data suggest, the fourth form of motion by deformation of frozen sediments under cold-based glaciers was widespread in the Pleistocene. This process has already been described by Echelmeyer and Wang (1987) on a much smaller scale in a thin layer beneath a cold-based alpine glacier.

Our inspection of frozen relicts of the Pleistocene glaciation in Siberia shows ubiquitous traces of this process over vast expanses, sedimentary sheets tens to hundreds meters thick being involved into subglacial deformation. Such a thick deformable bed means its generally lower shear strength as compared to glacier ice. This could happen if thick clayey strata and/or other sediments with high ice content are available. Water saturated clayey sediments can retain their plasticity in a wide range of negative temperatures due to preservation of combined water in liquid state (Tsytovich, 1973; Williams & Smith, 1989). The common for West Siberia high salinity of pore water would facilitate the plastic-frozen state of clays and retard their freezing at sub-zero temperatures.

Various ground ices, if voluminous enough, would also produce subglacial zones of low shear strength. In such cases even coarse-grained sediments could react to prolonged stress as incompetent rocks. The zones of weakness may consist of stratiform ice as well as of ice-supported sands. The latter case (represented by the Harasavey diapirs) suggests dilation decreasing shear strength of a sand massif. A weak element, namely dead glacial ice, is also present in tills deposited in frozen state.

Favourable conditions for the downward shift of the zone of maximum deformation are provided by any increase of a temperature gradient in frozen substrate and overlying ice, because it implies a colder and therefore more rigid glacial ice. Conversely, if warmer ice flowed over colder permafrost, glacial deformation of the sedimentary substrate was insignificant (the Marresale and Selyakin Mys cases).

At last, deformations could concentrate at the base of permafrost, if it was not too deep for shear strains. Such shallow permafrost might develop as a result of refreezing hydrographic taliks just before an ice advance. Especially favourable conditions for the zone of weakness appeared when the base of permafrost moved upward under a thickening ice sheet (Romanovsky, 1993), particularly if discharge of the meltwater was impeded.

As a result, the dynamic sole of the cold-based glacier shifted down into incompetent strata of the substrate to a depth allowed by increasing normal stress. The deformation of simple shear was developing in clayey, icy, or wet sediments until a plane of décollement emerged in the form of a shear zone or drag fault (more adequately labelled ductile fault in Russian literature, e.g. see Fig. 17). Due to friction heat and pulsating ice movement a multigelation regime of recurrent freezing and thawing must exist along such a fault or shear zone. This regime with positive feedback, however, would hardly produce enough heat for all-round thawing of subglacial permafrost.

More competent, usually sandy, members of the frozen sediment package moving above the subglacial dynamic sole in the substrate, were boudinaged and partly transformed into parautochthonous cataclastic breccia. Subsequently they could be glacially entrained and turned into long-travelling sediment rafts and clasts. Unimpeded subglacial displacement in this basically extension zone supported by glacial shear stress could provide forward motion of the entire glacier/permafrost complex with large glaciotectonic nappes developing. In other cases subglacial extension of incompetent frozen sediments ceases very soon, changing into a conjugate zone of compression and stacking of frozen rocks downglacier.

The transition to a compression zone may happen, although not necessarily, at the ice margin. This zone may develop either upglacier from the margin or in the proglacial position. In the former case a normal reason for the change of deformation style is cooling and hardening of subglacial sediments, thus impeding their horizontal flow. The result would be tight folding and stacking of slabs of frozen sediments (and often ice) separated by listric thrusts. This process probably accounts for many glaciomelanges. In the proglacial position compressional deformation may be topographically expressed as a squeezed-up end moraine with the same imbricate structure.

The change of subglacial extension to compression deformations often occurred in non-marginal environments, induced by lithological, orographic or thermal barriers. The purely lithological barrier formed by the downglacier thickening of hard frozen Paleogene sands is exemplified by the Atlym disturbances. A similar effect could be achieved by advection of colder ice producing a rigid dam downglacier.

Large compression disturbances, marginal as at Marresale, and non-marginal, as at Atlym, are ubiquitous in glaciated West Siberia. When studied in boreholes and natural exposures (e.g. Zakharov, 1968; Astakhov, 1979; Sergiyenko & Bidzhiyev, 1983; Kaplyanskaya & Tarnogradsky, 1986; Generalov, 1987), they show deeply crumpled sedimentary successions hundreds of metres thick, which in many places change to horizontally lying strata only below 300-400 m from the surface. The size and amplitude of these imbricate structures tempt some geologists to connect them with deep-seated tectonism (e.g. Sergiyenko & Bidzhiyev, 1983; Generalov, 1987). Such attempts are refuted not only by the epidermal character of the disturbances, but also by the bow-shaped form of their topographic expression and fundamental fact of their occurrence exclusively within the drift limit (Astakhov, 1986b).

The argument is usually focused on parautochthonous imbrications of the Atlym type, where the parental undisturbed strata are close by, and Pleistocene tills are not directly involved into the compression structures. However, the same tectonic style is also observed in Quaternary successions and in those cases, where compressed and folded slices could be identified as definitely foreign, i.e. transported from afar along a glacial thrust. One of the most spectacular examples of the compressed allochthon is known in Arctic European Russia, where an imbricate assemblage of Mesozoic and glacial sediments has been found stacked against a Paleozoic ridge at 40-50 km from their *in situ* occurrence (Gornostay, 1990). A bow-shaped zone of erratic Cretaceous sandstones 0.5-4 km wide and 60 km long, described on river Lyamin in West Siberia (Generalov, 1987), is probably another such allochthon.

In West Siberia large imbricate structures are very often accompanied by numerous erratic rafts of sedimentary rocks, although many far-travelled rafts are known lying atop of undisturbed younger sediments. Such are the largest studied sedimentary erratics, the stratiform slabs of Paleogene (Samarovo) and Jurassic (Yugan) rocks near the limit of the maximum glaciation. The Samarovo erratic at least 400 m long and 30 m thick and Yugan erratic is more than 160 m long and up to 14 m thick (Shatsky, 1965). Transportation of these and the above mentioned Semeyka erratic is conceivable if a good part of basal ice was, at least temporarily, relatively rigid and immobile, thus allowing the upper ice layers with entrained rafts of frozen sediments to slide over distances of hundreds of kilometres.

The subglacial compression, expressed in assemblages of tightly spaced folded slices and protrusions of substrate sediments mixed with glacial ice, in the non-marginal position is compensated by accelerated ice flow with entrained material in higher layers of glacier (Weertman, 1976). This may take the form of a large plane of décollement inside the glacier, along which blocks of frozen sediments, emerging from their subglacial source (Lavrushin, 1976), can be transported very far downglacier. Such a change of subglacial deformation to englacial sliding above obstacles is described here as springing-up of the dynamic sole of the glacier (Fig. 19). In this context, glaciers on wet beds represent only a special case of glacier sliding, geologically recorded in lodgement tills. Sliding may also occur within a frozen sedimentary substrate with resultant ductile thrusts, or englacially, which can be recorded in long-travelled rafts of soft sediments.

Both processes – of longitudinal extension over flowable rocks and compression at basal or subglacial obstacles – make a conjugate couple responsible for the saltant behaviour of the dynamic sole of the glacier, which probably provides the most effective mechanism of erosion and distant transportation by cold-based glaciers instead of abrasion and regelation adfreezing of debris characteristic for warm-based glaciers.

Apart of the extension/compression structures of active ice motion, subglacial diapirism is widespread in glaciated West Siberia as well. It includes not only the described mesoscale protrusions but also very large topographically expressed diapiric massifs built of clays and diatomites (Generalov, 1987). The diapirism is generally thought to be induced by differential loading by stagnant glacial ice (Astakhov, 1986). The above described small-scale diapirism in the Yamal Peninsula obviously occurred in perennially frozen rocks due to low shear strength of clays (Marresale) and/or low density of buried ice (Harasavey).

Vaccillations of the base of thin subglacial permafrost in the area adjacent to thawed bed could (at a stage of aggradation) intensify segregation of massive ground ice, which only indirectly, via parental meltwater, was connected with inland glaciation. Probably a similar process has been described for subglacial bodies of intrasedimental ice in the Canadian Arctic by Rampton (1991).

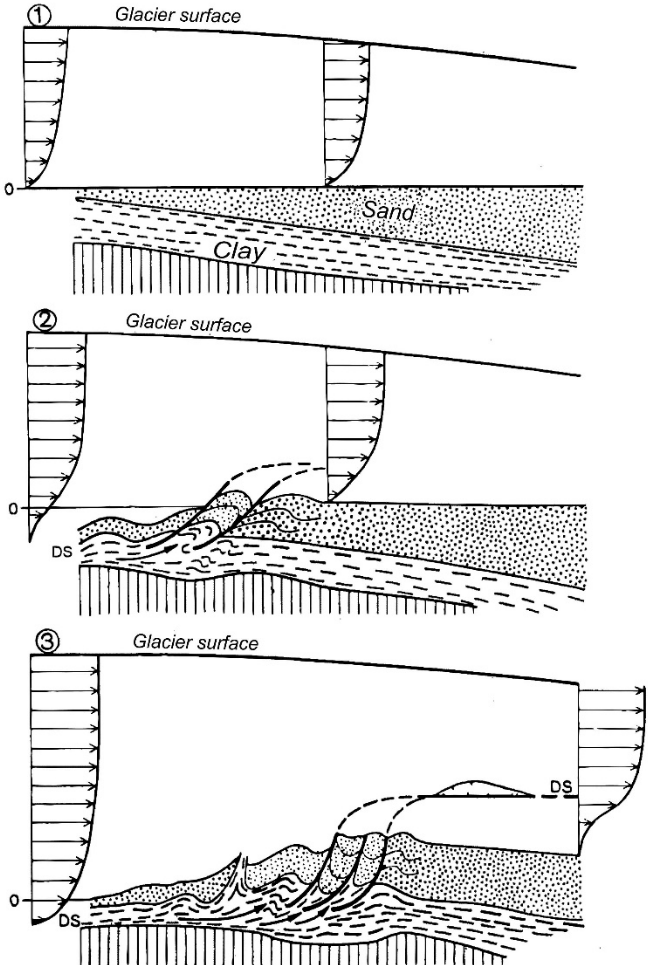

Fig. 19. Origin of deformations in frozen glacier bed consisting of clay and sand as inferred from the Atlym disturbances (Fig. 16). Stages 1 to 3 refer to thickening of advancing glacier. Parallel arrows show vertical distribution of summary velocities within ice/permafrost couple. DS is dynamic sole of the glacier.

Thermal zonality of Pleistocene ice sheets (Sugden, 1977; Hughes, 1981) is still an open question for West Siberia. A thawed substrate so far may be suggested for deeper parts of the Kara Sea shelf, where ice divides were situated. Farther south it must have changed into the freezing zone with intense erosion and glaciotectonism (Hughes, 1981). Most of the glaciated area was covered by cold-based ice with exception of large ice streams along major river valleys. Especially numerous traces of wet-based conditions are observed along the Yenissei. E.g., boulder pavements, typical for lodgement by warm-based glaciers (Eyles, 1983), have been described in the Middle Pleistocene tills at several localities between 64 and 62° N (e.g. Troitsky, 1975). Drumlinised surfaces are noted in aerial photos immediately east of the Yenissei, where relatively rigid Permian sandstones are close to the surface.

The predominantly frozen glacier bed is closely connected with the manner of deposition of thick successions of Middle Pleistocene tills in central parts of glaciated West Siberia, which almost invariably contain clasts and rafts of very loose sands. Such tills cannot be deposited by classical lodgement process implying friction `smearing` of glacial bed as a result of particle by particle release of debris from the melting sole of a glacier (Drewry, 1986). The model of frozen basal exfoliation suggested by Shantser and elaborated by Lavrushin (1976, 1980) is more applicable. This model envisages accretion of basal tills by progressive stagnation of thick layers of debris-laden basal ice. A possible cause for the wide spreading of this process is a relatively fast influx of cold ice from upper layers of ice sheets towards the base (Kaplyanskaya & Tarnogradsky, 1993). Another reason is the stagnation of basal ice over lithological barriers, as suggested in Fig. 19.

Exfoliated icy tills incorporated into subglacial permafrost are the last elements of disintegrated ice sheet to thaw because of their basal position and protection by the cold stored in the underlying old permafrost. They lose their ice much later than the main glacier body disappears, i.e. already under interglacial climates and in a closed subterraneous system. Over great expanses of West Siberia (between 66 and 60° N) the melted out tills are still underlain by deep-seated relict permafrost blocking the terrestrial heat flow. The slow melting out of thick sequences of stagnant ice protected from beneath by the persisting permafrost has evidently provided the rare hydraulic and gravitational conditions (Paul & Eyles, 1990) necessary for preservation of glaciodynamic structures.

In the area of present permafrost structurally same tills have not melted out and are described as primordially frozen tills with relict glacial ice. Therefore, we think that a basal till is deposited not at the moment of aggregation of particles by melting out, but when the debris-laden ice becomes stagnant, i.e. at the end of glacial transportation. Melting out may be considered as a kind of diagenetic process (Astakhov & Isayeva 1988) dependent on the post-depositional climate history.

Conclusions

The discussed geologic and thermometric data suggest that Pleistocene ice sheets of West Siberia were mostly cold-based. The subglacial frozen sediments were not a passive rigid bed but were involved into the glacier strain system. In many cases this resulted in the all-round mobility of large sedimentary masses, which were capable of lateral movement over considerable distances. The dynamic sole of the glacier could dive deeply into a weaker zone of the sedimentary substrate to produce a plane of décollement, thus providing a mighty mechanism of subglacial erosion. The mobilized masses of frozen sediments stopped by lithological, orographic or thermal barriers were stacked to form imbricate allochthons. Glacier ice participating in such imbricate structures is preserved within the present permafrost zone. The subglacial barriers could make the dynamic sole spring up into an englacial position, thereby providing entrainment and distant transportation of giant sedimentary rafts.

In West Siberia glacially entrained debris was mostly deposited by basal exfoliation of stagnating sheets of dirty ice, thus building up the subglacial permafrost. Such icy tills were melting out much later, under interglacial conditions, and not everywhere. In the zone of present permafrost the basal tills often retain their primordially frozen state. They are the most conservative element of inland glaciation during interglacial shrinking of the cryosphere. In sedimentary basins frozen basal tills survived longer being protected from beneath by the cold stored in the sedimentary substrate.

Acknowledgements

Our thanks are due to Vladimir Solomatin for the useful field discussion at Selyakin Mys and Valery Gataullin for his helpful comments on the Yamal geology. This work has been partly supported by the Russian Science Foundation for Basic Research (RFBR) grant 95-05-14512. The chemical analyses have been carried in the Central Laboratory of Krasnoyarskgeologia Corporation.

9. The Stratigraphic Framework for the Upper Pleistocene of the Glaciated Russian Arctic: Changing Paradigms

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Global and Planetary Change, Vol. 31/1-4, P. 281–293

Preface

This paper is inspired by the growing number of Quaternary studies performed in the Russian Arctic by West European scientists, mostly under the aegis of the QUEEN programme. The objective of the paper is twofold: i) to briefly outline the development of stratigraphic ideas for the Late Pleistocene of the Russian Arctic, responsible for origin of the existing terminology, and ii) to summarize recent results leading to a revision of the conventional stratigraphic wisdom. Such a review seems timely, because foreign researchers are not necessarily familiar with the origin or spelling of common names in Siberian stratigraphy, and even less with the ideas behind them. Moreover, many Russian writers, especially those not professionally involved in the northern glacial geology, often misuse traditional labels, thus contributing to the notorious confusion in Quaternary stratigraphy of the Russian Arctic.

Siberia

Classical stratigraphy

The principal marker strata of the Arctic Quaternary, first recognised by V.N. Sachs as \`horizons\` on the Lower Yenissei around Ust-Port (Sachs & Antonov, 1945), were as follows (upwards in the succession):

1) maximum glaciation stony till (only in boreholes);
2) Messo fluvial sands, named after the river Messo-Yaha;

3) cold marine sediments of the Sanchugovka Formation: diamictic and stratified clays with poor mollusk fauna with predominance of *Portlandia,* named after the Sanchugovka creek;

4) Kazantsevo marine sands with rich boreal fauna, especially *Arctica islandica*, named after the Kazantsevo village;

5) Zyryanka till and outwash, deposited by mountain glaciers which spread onto the adjacent lowlands, named after the Zyryanka creek;

6) Karginsky postglacial terrace 20–30 m a.s.l. named after the Cape Karginsky and consisting of marine sediments with an arcto-boreal mollusk fauna, upstream changing into deltaic and alluvial formations.

Later the uppermost Sartan \`horizon\` was added basing on descriptions of the final generation of alpine moraines overlying high alluvial terraces on the river Sartan, eastern Verkhoyansk Range, more than 2000 km east of Ust-Port area.

The above scheme was extrapolated by Sachs over the entire Russian Arctic in his grand volume (Sachs, 1953) which contained observations made by early travelers, results of small-scale geological mapping and of stratigraphic studies by the author himself. Sachs thought that the Messo, Sanchugovka and Kazantsevo formations were all deposited during an interglacial analogous to the European Eemian. The last inland (Zyryanka) glaciation occurred in the form of thin ice sheets fed by mountain glaciers. The Karginsky marine and fluvial terraces were never covered by lowland tills and represented a climate warmer than the present one, although colder than the climate of the Kazantsevo transgression. Only in the mountains of northeastern Siberia Late Pleistocene glaciers survived the Karginsky warming and reached their maximum extent during the final, Sartan phase of the last glaciation. Sachs concluded that the Zyryanka lowland glaciation culminated before the last Fennoscandian glacial maximum, probably earlier than 30-35 ka ago, which view was supported by numerous finds of mammoth remains on top of the latest till. He estimated the age of the Sartan alpine phase as between 20 and 10 ka ago.

However, without any radiocarbon dates available he could not decide whether it corresponded to the maximum of the last Fennoscandian glaciation or to its final stages (Sachs, 1953, pp. 510–517). Later the above events were used for the names of Siberian \`climatostratigraphic horizons\` (or climatoliths) which, according to the Russian Stratigraphic Code (Bekker et al., 1992), are material representatives of thermochrons and cryochrons in a given region.

What puzzled Sachs (and many other geologists) was the rare occurrence of diamicts within his Zyryanka \`horizon\`, which over the vast expanses of Siberian glaciated flatlands was represented mostly by sands atop of marine Sanchugovka and Kazantsevo deposits. Sachs (1953) tried to explain this phenomenon by a very thin lowland ice which soon got stagnant producing mostly kames. Strelkov (1965) advocated the idea of passive ice fields over the Siberian plains. Such explanations did not appear very convincing, especially in the view of the wide extent of the reconstructed Zyryanka ice sheets, presumably produced by small mountain glaciers. Therefore, in many publications of the 1960-70s the Zyryanka glaciation shrunk to modest-sized piedmont aprons and alpine glaciers.

Later revisions

Geologists, who in the 1950s mapped West Siberia south of the Arctic Circle, found no marine formations but only very thick tills separated in places by terrestrial sands. The lowermost till traced south to the drift limit at 60° N was labeled the Samarovo Till. The upper till of more limited occurrence was named the Taz Till, and intervening sand without distinct paleontological characteristics was attributed to the so-called Shirta interstadial (or interglacial). This glacigenic sequence, covered by thick loess-like silts, was unanimously related to the Middle Pleistocene. A big surprise came when the Taz Till was proven to laterally merge into the marine Sanchugovka Formation. This led to the idea of the huge Sanchugovka transgression in the lowlands coinciding with the Middle Pleistocene sheet glaciation of uplands (Arkhipov, 1971; Lazukov, 1972; Zubakov, 1972). Since then in all regional stratigraphic schemes the Middle/Upper Pleistocene boundary was placed between the Sanchugovka and Kazantsevo Formations.

The mortal blow to the scheme of Sachs was dealt by his disciple and malacofauna expert S.L. Troitsky, who studied the Lower Yenissei stratotypes and found that the Karginsky terrace was built of two unrelated formations: the basal marine deposits with arctoboreal fauna, not much different from the Kazantsevo fauna, and overlying terrestrial sediments, till included (Troitsky, 1966). Sachs had to agree with this conclusion. Since the sediments with arctoboreal fauna were related to the Eemian, all record of the Karginsky transgression at its most typical low-altitude sites evaporated.

Then Troitsky found that terrestrial sediments of the `Karginsky horizon` in the parastratotypic section of the same low terrace on river Malaya Heta were again covered by till. Because the sub-till alluvium yielded a couple of finite radiocarbon dates he suggested that i) the last inland glaciation of West Siberia was synchronous with the last ice sheets in Europe and America, ii) the Zyryanka name should apply to all Late Pleistocene tills, iii) the Sartan label be abandoned, iv) real `Karginsky terraces` should be looked for outside the limits of the last glaciation and its proglacial reservoirs (Troitsky, 1967). These conclusions actually turned the Sachs` scheme upside down: a Zyryanka Till appeared on top of the Karginsky `horizon`, and the basic idea of an early glacial culmination in Siberia was replaced by a paleogeography not much different from the European and North American models.

Further application of the radiocarbon method in its conventional form deepened the alienation from the spirit of the classical scheme, though preserving its empty shell in the form of traditional stratigraphic labels. The landmark in this process was the monograph by Kind (1974), who applied radiocarbon geochronometry to the Sachs` scheme to compose a geochronological scale for Siberia, apparently similar to the North American one. She obtained several old finite dates on shells from the marine strata at the Cape Karginsky stratotype of Sachs (= the Kazantsevo marine Formation, according to Troitsky) and at the same time acknowledged their sub-till position.

Whereas Troitsky (1967) attributed this till to a late advance of Zyryanka glaciers, Kind (1974), based on finite dates from sub-till beds with palaeontological indicators of a climate by 3-4°C warmer than the present one, suggested a Karginsky Interglacial in the interval of 50 – 22 ka ago, correlative to the Middle Valdai, Middle Wisconsin and isotope stage 3. According to her, it was followed by a large ice sheet synchronous with the Late Valdai and classical Wisconsin. She used the Sartan label for her Late Pleistocene glacial maximum. Kind also maintained that the Younger Dryas event was reflected in the large end moraines of the Norilsk stage at the foot of the Putorana Plateau, treated by Sachs (1953) as a counterpart of the Sartan stage.

It is obvious that the names of the Sachs scheme were applied by Kind (1974) to geochronological stages of quite different paleogeographical meaning. Unfortunately, this change happened not because the sedimentary events described by Sachs were refuted, but solely on the base of presumed validity of all radiocarbon dates. The main differences of the radiocarbon stratigraphy from the classical scheme were that the second Late Pleistocene (Karginsky) interglacial appeared and that the last ice sheet was rechristened. Because of the wide use of Kind`s scheme by mapping geologists, her geochronological terms stuck and got unwittingly applied to some geological bodies, which Sachs would have never thought as suitable for such labeling. Some geologists even call thick marginal accumulations of tills and glaciotectonised sediments in lowlands `Sartan moraines`. This violation of the Stratigraphic Code (Bekker et al., 1992) cannot be excused by the implied geochronological sense, because the alpine Sartan moraines *s. stricto* have never been properly dated and may theoretically belong to quite a different stage, or to any part of that 22 – 10 ka cold `Sartan` interval by Kind.

Thus, the victorious radiocarbon method turned the classical Upper Quaternary stratigraphy into a double entendre. Since that time some researchers, especially those who never professionally dealt with glacial geology of the Siberian Arctic, habitually employed words Zyryankian, Karginskian, Sartanian for abstract geochronological intervals roughly equivalent to the triple subdivision of the Weichselian/Wisconsinan time, whereas others, mostly northern glacial geologists, used the same terms in their original stratigraphic sense. Unfortunately, the same terms are often loosely applied to much better studied periglacial formations which correlation with the above glacigenic and marine sediments of the Lower Yenissei is highly questionable. The confusion is so deep-rooted that today there is no easy way out. Though Sachs himself lived to see his original terminology being corrupted and attached to unrelated deposits, he was by then already deeply involved with the Mesozoic research and preferred not to intervene, leaving us to sort out the confusion created by the indiscriminate use of the radiocarbon method coupled with disregard for the Stratigraphic Code.

There was an attempt to `escape` the lowland Sartan glaciation leaving it at the mountains where it rightly belongs. Arkhipov et al. (1977) and Arkhipov (1984) suggested the Zyriankian climatostratigraphic superhorizon consisting of three horizons: the Lower and Upper Zyryankian glacial horizons separated by the Middle Zyryankian interstadial horizon within the radiocarbon brackets 50 to 22 ka. But it did not help much, because each time the question arose one had to explain what was actually meant. Later Arkhipov (1990) returned to the Sartan label for the last inland glaciation.

The present official stratigraphic scheme of the West Siberian Plain holds the names of the climatostratigraphic horizons (regional representatives of global climatic events): Upper Zyryankian=Sartanian and Lower Zyryankian=Yermakovian separated by the Karginskian interstadial, all subdivisions of the Zyryankian superhorizon. The Sartanian and Karginskian are accepted without stratotypes (sic!), because the Sartan moraines *s. stricto* have no relation to the lowland glaciation, and the mentioned marine strata at the Cape Karginsky are dated by ESR to 121.9 ka (Arkhipov, 1990). This terminological mess is further complicated by the geochronological value attached to the `Zyriankian superhorizon`, which is roughly equivalent to the Wisconsinan (isotope stages 4 – 2) but not to the West European Weichselian (isotope stages 5d – 2) by Mangerud (1989).

Modern data

Apart of the discussion on nomenclature, material foundations of the classical scheme were questioned by reinvestigations of the Sachs sections. The most serious attempt was undertaken by Kaplyanskaya and Tarnogradsky (1975) in the Ust-Port stratotypic area. In laterally extensive sections they found that the Sanchugovka Formation was actually basal till in the form of glaciotectonites with rafts of marine sediments transported from the Kara Sea. In accordance with the Stratigraphic Code they preserved the geographical name of the Sachs` sedimentary formation by coining the term Sanchugovka Till. They also found boreal fauna of the Kazantsevo type in that till, which further complicated the question, suggesting either another boreal transgression in the Middle Pleistocene or a Late Pleistocene age for the Sanchugovka Till.

Later they discovered that the Zyryanka diamicton in many places contained thick bodies of buried glacial ice imperceptibly merging into surrounding `primordially frozen till` (Kaplyanskaya & Tarnogradsky, 1976). The works of the above authors made it clear that in the northern plains with the thick continuous mantle of glacigenic formations no good results can be achieved by using only narrow logs or boreholes. Lateral tracing of sedimentary bodies in order to assess their structural position and the governing sedimentary processses is essential for understanding the stratigraphy!

Meanwhile the very core of the classical paleogeography of the Siberian Ice Age was shaken, when it was found that the pattern of glaciotectonic ridges and marginal formations together with till petrography conforms sources of inland ice not in the mountains but on the low coastlands and the shelf of the Kara Sea (Astakhov, 1976, 1977). The would-be Uralian moraines of the last stages of the Zyryanka glaciation (Arkhipov et al., 1977) proved to be marginal features of thick Arctic inland ice that flowed upslope in and around the Polar Urals up to 500–600 m a.s.l. (paper 2, Ch. I). These results led to reconstruction of a Late Pleistocene ice sheet more than 2 km thick and to rejection of the classical idea of free northward discharge of Siberian rivers during the maximum of the Late Pleistocene glaciation in favour of a large ice-dammed lakes (Volkov et al., 1978; Arkhipov et al., 1980).

The reconstruction of a thick Kara Sea ice sheet helped to get rid of the old puzzle of the Zyryanka glacial `horizon`. It has become clear that what

Sachs and his followers described as the Zyryanka Formation were partly piedmont varieties of basal till, and partly outwash and other ablation sediments deposited in the lowlands, whereas the lowland basal tills and varves due to the widespread redeposition of mollusk shells by Kara shelf glaciers were perceived as the Sanchugovka marine formation. This can be easily demonstrated on the Gydan and Yamal peninsulas, where thick clayey diamicts underlain by marine sands contain thick bodies of fossil glacial ice (papers 6 in Ch. II and 10 in Ch. III). Yet some researchers still look in vain for Late Pleistocene tills in the overlying sandy outwash of the arctic plains.

In 1980-81 two special boat trips along the Yenissei river were undertaken by a number of Russian geologists from various institutions to check stratotypes described by Sachs, Troitsky, Arkhipov and Zubakov (Astakhov et al., 1986). They confirmed the conclusions by Kaplyanskaya and Tarnogradsky (1976) and found that most of other Yenissei sequences were also heavily glaciotectonised, the normal superposition in many cases being questionable. The result of this collective inspection was even more disastrous than expected, since it proved almost impossible to directly correlate between basic units of individual sections within the stratotypic area around Ust-Port (Fig. 1).

The Kazantsevo marker sand with abundant shells was found only in isolated outcrops, probably as large rafts of the interglacial strata, and could hardly be laterally traced even within the stratotypic area some 40 km wide. The validity of the conventional radiocarbon method in the local

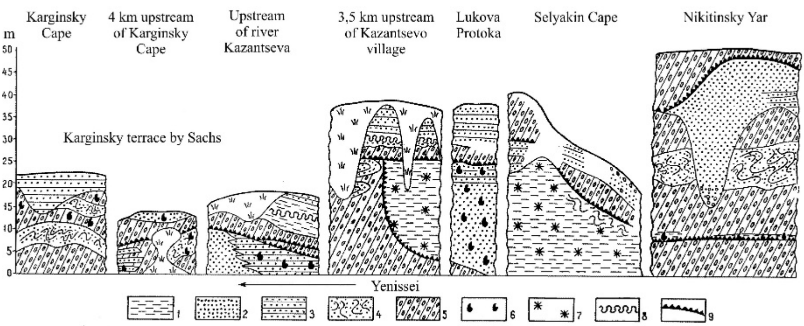

Fig. 1. General structure of Quaternary deposits exposed on the right bank of the Yenissei in the stratotypic area of Ust-Port (after Astakhov, 1984). Symbols: 1 – silty rhythmite, 2 – crudely bedded sand, 3 – laminated sand, 4 – disharmonic folding, 5 – clayey diamicton, 6 – mollusk shells, 7 – marlekors, 8 – cryoturbations, 9 – shear planes and thrusts.

conditions of cyclic thawing of the permafrost was also questioned, because several finite dates were obtained from the very base of the Sanchugovka Formation (Astakhov et al., 1986).

Another source of stratigraphic mistakes was revealed in the study of deposits of fossil glacial ice which were first described by Kaplyanskaya and Tarnogradsky (1976). Buried glacial ice preserved within the long-existing Pleistocene permafrost is apt to accelerated melting during climatic ameliorations, especially in the Holocene, producing thick diamictic sequences with all kinds of radiocarbon datable materials. Such flowtills, characteristic of retarded deglaciation in continental climates, are readily confused with basal tills, leading to erroneous ideas of very young ice advances (paper 5 in Ch. II). The bona fide basal till of the Lower Yenissei, containing bodies of thick glacial ice, was found to be overlain by alluvial and limnic sediments dated to more than 30,000 radiocarbon years (Astakhov & Isayeva, 1985, 1988).

By that time sections on the Yamal, Gydan and Taimyr peninsulas, where the thickest part of the last ice sheet resided, had already yielded consistent series of old radiocarbon dates derived from postglacial successions at Cape Sabler, Mongotolyang and Syo-Yaha (Kind & Leonov, 1982; Vasilchuk et al., 1984). A special study of the alleged sediments of a huge ice-dammed lake on the Ob river (Volkov et al., 1978) found only a loessic sequence with lenses of sink-hole deposits yielding finite radiocarbon dates (Astakhov, 1989).

Likewise, many silty formations of the Arctic, often related to lacustrine activity, proved to be parts of the ice-rich loess-like mantle not much different from the Yedoma Formation of eastern Siberia. E.g., the intermittent (if not continuous) growth of long ice wedges through the monotonous silt formation at least 30 m thick at Cape Sabler, Taimyr (Derevyagin et al., 1999), indicates predominantly subaerial and shallow limnic sedimentation for the last 30 ka but hardly a stable lacustrine environment, as was originally suggested (Kind & Leonov, 1982).

From the above and other evidence produced by various investigators it was concluded that the last invasion of inland glaciers from the Kara Sea shelf onto the Siberian mainland could only have occurred beyond the radiocarbon age limit (papers 6 in Ch. II and 10 in Ch. III), and therefore all finite radiocarbon values from beneath the arctic tills must be due to contamination by younger carbon. The indiscriminate use of doubtful radiocarbon dates in previous years coupled with sedimentological misinterpretation is put forward as the main cause of the Siberian geochronological confusion (*ibid.).*

European Russia

Classical stratigraphy

In the Russian European North two tills separated by marine strata with warm-water fauna (containing boreal to lusitanian mollusk species) have been known since the turn of the century. Later S.A. Yakovlev (1956) developed a scale with 3 Neopleistocene glaciations and two interglacial transgressions of warm Atlantic water. He related his oldest Neopleistocene glaciation to the Warthe stage and the preceding `Northern Transgression` – to an inter-Saalian interglacial. The Boreal Transgression proper, which occurred later, was followed by the Second Neopleistocene glaciation centred on Novaya Zemlya. This ice sheet, correlated to the Early Weichselian, in his opinion occupied the Lower Pechora flatlands but neither reached the Pai-Hoi Range, nor coalesced with the Scandinavian ice. The Second glaciation was followed by the Onega Transgression which left boreal fauna at present day altitudes up to 220 m in the Kanin Peninsula. Yakovlev and many other geologists saw traces of such a transgression in mollusk shells scattered over the step-like coastal plains. The Third Neopleistocene glaciation took place only in Fennoscandia (Yakovlev, 1956).

Later revisions

The deep-rooted `Siberian` inability to distinguish between shelly tills and real marine sediments in the 1960-s spread across the Urals and caused the resurrection of drift hypotheses in the form of so-called `marinism`. As a result, thick diamictic sequences along the southeastern low coast of the Barents Sea were mapped by Vorkuta geologists as glaciomarine or cold marine strata named the Rogovaya Formation, and only Uralian piedmont diamicts were described as glacial deposits (Popov & Afanasyev, 1963). This Rogovaya Formation, presumably of Saalian age, was supposed to represent all clayey surficial sediments, except the staircase of younger marine terraces.

The only difference between this scheme and the Siberian one by Sachs (1953) was as regards the Eemian marine strata which in the scheme of Vorkuta geologists (Popov & Afanasyev, 1963; Zarkhidze, 1972a, 1972b) existed only as a terrace about 100 a.s.l., called the More-Yu Formation, whereas similar sands with mollusk shells at higher altitudes were related to the Vashutkiny Formation of the final Middle Pleistocene. The Karginsky term was borrowed from Siberia for presumed marine terraces at 40-60 m a.s.l. along the Barents Sea coast. Sands with boreal mollusk fauna underlying the topmost diamictic formation were regarded as the Padymei Formation of Pliocene-Early Pleistocene (Zarkhidze, 1972b).

The above stratigraphy, with varying terminology, is still in use by local exploration and mapping companies, who deal predominantly with drilling logs, but professional glacial geologists do not accept marinist classifications. A team of Moscow geologists led by A.S. Lavrov, who mapped most of the Pechora and Severnaya Dvina catchment areas and were the first to massively employ radiocarbon dating, related all surficial pre-Holocene strata to the Late Weichselian glaciation. The hummocky sand fields with mollusk shells were mapped as kames and other ablation features. Already in the 70-s they found that the `Rogovaya Formation` was an ill-assorted collection of geological objects of various ages, including radiocarbon dated Middle Weichselian and Holocene sediments (Arslanov et al., 1981a).

The official stratigraphic scheme of the region presently is based on climatostratigraphic subdivisions. The Vychegda (a Middle Pleistocene) till, according to this scheme, is overlain by the Sula marine Strata correlated with the Eemian. The Weichselian series north of the Arctic Circle is subdivided into the Laya (Lower Valdai) Horizon represented by periglacial sediments, the Byzovaya (Middle Valdai) interstadial Horizon with Paleolithic artefacts and the Polar (Upper Valdai) Horizon consisting of the surficial tills and stratified glacial drift (Guslitser et al., 1986).

Modern data

Recent studies by the Russian-Norwegian team (the PECHORA project) discovered that the surficial Markhida till deposited by the last ice advance from the shelf is out of reach for radiocarbon dating. Luminescence dating of beach sediments of the last ice-dammed lake, Lake Komi, yielded values of ~ 80-100 ka (Mangerud et al., 1999, 2001a). Predecessors erroneously ascribed an Early Holocene age to the Markhida ice advance (Grosswald et al., 1974; Arslanov et al., 1987). However, application of the Siberian principle of retarded deglaciation in the Pleistocene permafrost environment (paper 5, Ch. II) to the Markhida sequence (Tveranger et al., 1995) helped to differentiate between the genuine basal till of an older ice advance and diamicts of flowtills and solifluction sheets deposited during the Early Holocene warming and melting of Pleistocene permafrost with buried glacial ice, thus getting rid of the strange Holocene glacier surge suggested by Grosswald et al.

(1974). In spite of persistent search during six field seasons no sediments related to interglacial or warm interstadial climate have been found above the Markhida till (paper 18 in Ch. IV and Mangerud et al., 1999). Sub-till sediments with forest pollen spectra described by Arslanov et al. (1981a, 1987) and Golbert et al. (1973) as Middle Weichselian most likely belong to the Eemian or even older interglacials. Some sites with old finite dates (35–47 ka), such as Urdyuzhskaya Viska and Lodmashchelye (Seyda) (Arslanov et al., 1981a, 1987), when resampled and AMS-dated yielded non-finite values. Thus, the situation with the Late Pleistocene of the European Arctic proved to be very similar to that in Siberia, where conventional radiocarbon dating often gives too young values, especially on shells and driftwood (papers 6 in Ch. II and 10 in Ch. III).

Similarities with West Siberia are not confined to the spurious radiocarbon dates. A special study by the PECHORA project of the Rogovaya, Padymei, Vashutkiny and More-Yu formations suggested by Vorkuta geologists (Popov and Afanasyev, 1963; Zarkhidze, 1972a, 1972b) at their stratotypic localities, revealed thick heavily glaciotectonised rhytmites and diamictons with shell fragments resembling the Sanchugovka Formation of the Yenissei (Fig. 2). Sands with mollusk shells occur any place in the 50-60 m high bluffs. Their normal position is beneath the diamicton, as in sect. 9. In logs made by local geologists the sands in sub-till position are called the \`Padymei Formation\`. When overlying the diamict unit, as in section 11, they are described as the \`More-Yu Formation\` of a 100-m high Eemian marine terrace (Zarkhidze, 1972a, 1972b). A little digging laterally reveals that the sand with boreal fauna in this section occurs as large blocks within the \`Rogovaya Formation\` diamicton (Fig. 2).

When traced beyond the bluff exposures, the steep-dipping marine sands, as in sect. 13, were found protruding as narrow ridges on nearby plateaus at 100 to 200 m a.s.l. At these altitudes the marine sands were described as the Middle Pleistocene \`Vashutkiny Formation\` by local geologists. Facies structure, lithology and fossils of the sands are similar in any position and most probably reflect one marine event, which is underlined by very close values of U/Th datings on *Tridonta* shells (Fig. 2). Various ridge-like heights in arctic tundras east of the Lower Pechora normally consist of up to 100 m thick imbricate stackings of the Upper Pleistocene diamicts and sands, as is evident from geological profiles compiled by Lavrushin et al. (1989). The all-pervading glaciotectonism imposed by the Early Weichselian ice sheet on the Eemian marine sediments is the obvious reason for several boreal transgressions to have been described from redeposited shells found on the surface up to 220 m

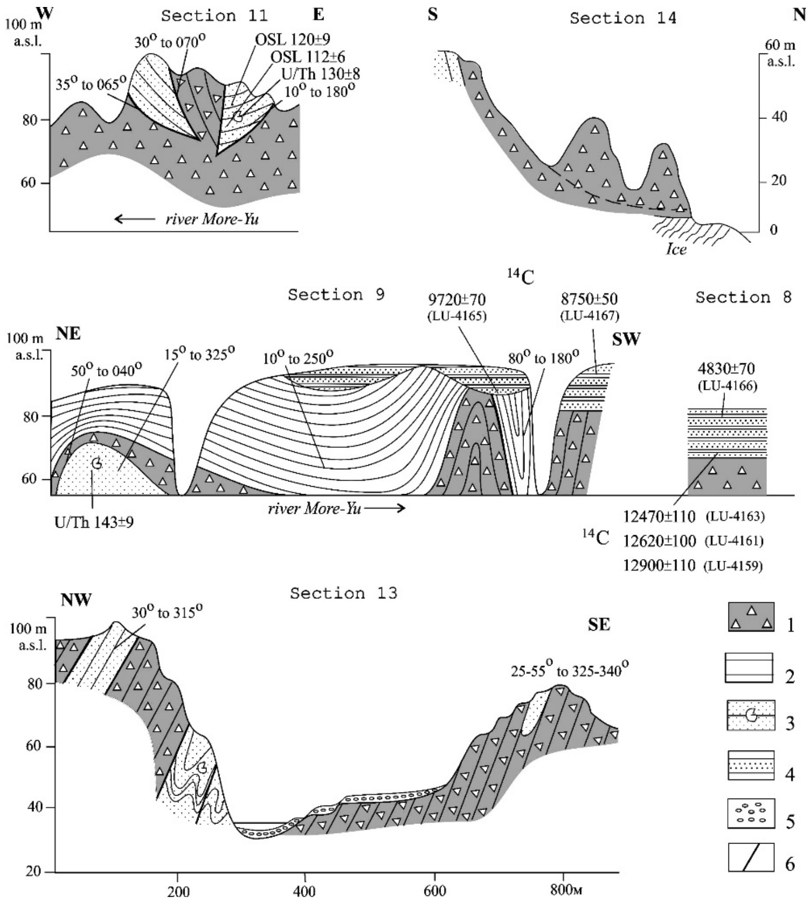

Fig. 2. Sections of Upper Pleistocene sediments on river More-Yu, 67°50′ N, 60–60°30′ E (numbered downstream) – results of 1998 by the Russian-Norwegian PECHORA project.

on the Kanin Peninsula and even higher in the Pai-Hoi Range (Yakovlev, 1956).

Thus, modern data from the European Arctic give roughly the same picture as in West Siberia. Again, as in Siberia, log descriptions without lateral tracing in disregard for the heavy glaciotectonism proved to be the main source of stratigraphic errors. The actually observable structure of the Upper Pleistocene on both sides of the Urals reflects deep distortions of the subsurface strata by a thick, upslope moving Early Weichselian ice sheet. In the Polar Urals imprints of this thick ice is retained in the form of fresh-looking morainic arcs at altitudes up to 560 m, inserted into the alpine valleys by ice streams that flowed southwards from the Kara Sea shelf (paper 18 in Ch. IV).

Large-scale glaciotectonic reworking of the Eemian basement, characteristic for Siberia and the Peri-Uralian Arctic, tends to be less pronounced farther westwards. At least around the Pechora mouth till-covered Eemian formations, unlike Siberia, can be traced in many sections (Astakhov et al., 1999; Mangerud et al., 1999). West of the Timan Ridge no large-scale disturbances of the so-called `Boreal Strata` have been reported (Yakovlev, 1956). This change may be explained by moderate erosive activity of wet-based ice sheets in the west, as opposed to the cold-based ice sheets of Siberian type that heavily excavated the weak substrate in the east (paper 8 in Ch. II).

To complete the comparison with West Siberia: during the field season of 1998 massive glacial ice with pebbles was found at the base of a left-bank bluff built of diamicts on river More-Yu 60° E (paper 18 in Ch. IV). Although in nearby sections the overlying sediments of postglacial lakes yield radiocarbon dates not older than 13 ka (section 9 in Fig. 2), the diamicton containing the fossil glacier ice cannot be Late Weichselian, because there are many sections in this area where the topmost member is thick aeolian sand of this age. E.g., the cover sand on a right tributary of the More-Yu (Ute-Yaha creek) yielded a consistent series of 4 optical luminescence dates in the range 26 to 21 ka. This is quite compatible with older luminescence dates obtained on sediments deposited during the last glaciation of this region and with numerous radiocarbon values in the range of 37–27 ka from postglacial sediments with mammoth fauna and Paleolithic artifacts (Mangerud et al., 1999).

Ubiquitous postglacial aeolian and permafrost phenomena, not detected by previous investigators, account for many features of periglacial sedimentation earlier perceived as signatures of a Late Weichselian glaciation (paper 18 in Ch. IV). Discontinuous aeolian sheets on top of the uppermost till of the Pechora Basin, indicating very cold and arid Late Weichselian climate, have obvious counterparts in Siberia, where they are intimately associated with synsedimentary ice-wedges and finds of mammoth fauna (paper 6, Ch. II).

The Karginsky problem

The Karginsky Interglacial suggested by Kind (1974) was readily picked up by micropaleontologists, who attributed to this event assemblages of foraminifers, sometimes even warmer than the Kazantsevo microfauna. The idea of a warm intra-Weichselian transgression quickly spread over the Russian Arctic and even reached the shores of the Kola Peninsula (Gudina, 1976). However, the Kola geologists falsified it rather promptly by showing that the finite radiocarbon dates were derived from sediments U/Th dated to the Eemian (Arslanov et al., 1981b).

In Siberia the problem was first dealt with by S. Troitsky (1966), who showed that mollusk assemblages were basically the same in the Kazantsevo and Karginsky formations. Then he considered the coastlands of West Siberia and found no postglacial marine terraces except the low Holocene beaches (Troitsky & Kulakov, 1976). All terrace-like surfaces there happened to be either covered by till or built of sediments older than the radiocarbon age limit. The stepped lowland of the Yamal Peninsula in the range of altitudes from 20 to 100 m a.s.l. proved to be composed of glacigenic materials (Astakhov, 1981; Gataullin, 1988; papers 1, Ch. I and 8, Ch. II)). The only sedimentary body in the Yamal that could be related to a Karginsky marine terrace (Vasilchuk et al., 1984) is loess-like silt with long ice wedges, most probably terrestrial in origin (paper 6, Ch. II). The only postglacial marine formation of Arctic West Siberia, consisting of cold-water clay, sometimes varved, with *Portlandia arctica* (Troitsky & Kulakov, 1976), directly overlies fossil glacier ice (paper 6, Ch. II) and has no relation to a transgression of warm Atlantic water.

A similar situation can be seen everywhere along the low coastlands of the Barents Sea. The low coastal platform, that was traditionally mapped as the Karginsky marine terrace, proved to be covered by till and limnic sediments (Arslanov et al., 1987, Lavrushin et al., 1989). Participants of the PECHORA project, that proved an Early Weichselian age of the last ice sheet of the region, have also found no postglacial strandlines. Where the uppermost till is removed by postglacial erosion, the exposed base of the coastal lowland shows a thick marine formation with interglacial mollusk fauna, sometimes glacially disturbed and not different from the Eemian marine sediments in the sub-till position (paper 18 in Ch. IV; Mangerud et al., 1999). Like in Siberia, finite radiocarbon dates are sometimes derived from this basal marine formation.

As mentioned above, in places the terrace-like coastal surface is covered by thick aeolian sands of Late Weichselian age (Mangerud et al., 1999). Most serious attempts to pinpoint the Mid-Weichselian interstadials (Golbert at al., 1973; Arslanov et al., 1981a, 1987) are connected not with this terrace but with sub-till fluvial sediments overlying the Eemian marine strata.

The only area, where the warm-water Weichselian transgression has not yet been disproven by direct geological observations, or falsified by redating, is the Taimyr Peninsula and the arctic islands. However, the suggested Karginsky sedimentary formation there has never been reliably described in terms of sedimentology and structural geology. On the other hand, one cannot exclude the possibility of a deep Mid-Weichselian glacioisostatic trough accomodating Atlantic water (Sachs, 1953). The available evidence is, however, too scant for any positive conclusion. It boils down to a number of finite radiocarbon dates obtained on shells, driftwood and walrus bones collected on various terrace-like surfaces of Taimyr and Severnaya Zemlya (Kind & Leonov, 1982; Bolshiyanov & Makeyev, 1995).

Other authors, who analysed analogous dating results, reject the Mid-Weichselian transgression and conclude that the dated materials originate from Eemian sediments (Fisher et al., 1990). This conclusion seems to be the most reliable for the time being, especially in view of the fact that the `Eemian` U/Th and ESR dates have been already obtained from the allegedly Mid-Weichselian marine formations (Arkhipov, 1990; Arslanov et al., 1981b).

Conclusions

The above review purports to demonstrate that the regional stratigraphic schemes of the Russian Arctic cannot accomodate the sedimentological and geochronometric results obtained during the last two decades, the conventional nomenclature being neither adequate, nor unambiguous. The obsolete method of log descriptions in isolated sections used for developing the conventional stratigraphy is a poor tool for deciphering the laterally variable structure of glaciated sedimentary basins. Whereas in marine geology the log approach, justified by apparently valid presumptions of undisturbed superposition and mostly supported by seismic stratigraphy, may not lead too far away, in terrestrial geology of glaciated plains the log approach has no such excuses. For terrestrial studies the best way forward seems to be either by obtaining as wide a view of the glacial complex as possible by mapping the country (e.g., paper 18 in Ch. IV) or by measuring lengthy cross-sections with important structures carefully plotted (e.g., Gataullin, 1988).

Fig. 3 illustrates the basic difference between the conventional stratigraphic thinking and results obtained by applying the principle of lateral tracing. The bottom diagram in Fig. 3 shows the basic features of arctic sedimentary sequences that cannot be explained in terms of the

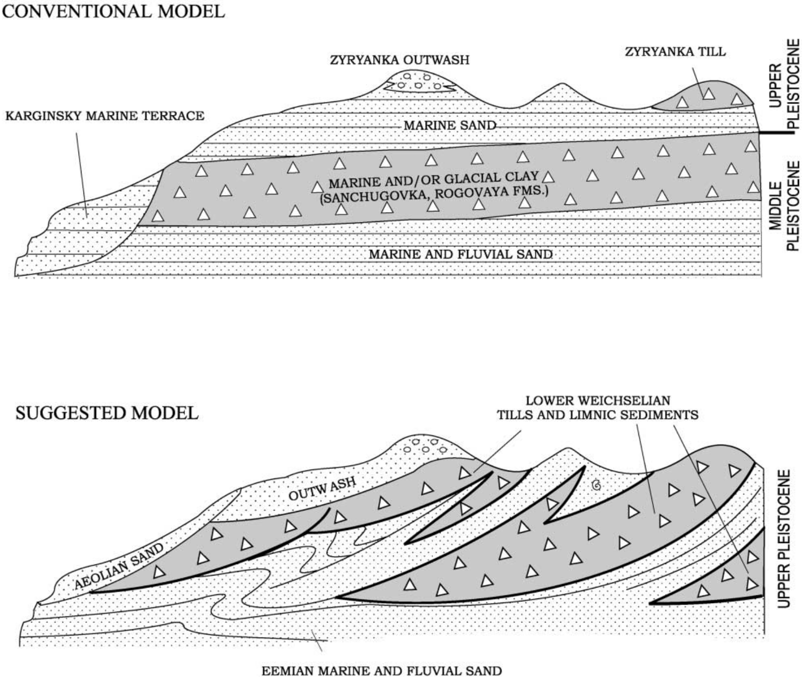

Fig. 3. The conventional and the author's models of relations between Pleistocene sedimentary formations exposed in glaciated plains of Arctic Russia.

traditional stratigraphic nomenclature. The sedimentary complex of the last ice age should not be termed the Zyryankian any more, as follows from comparison of top and bottom sketches in Fig. 3. In the lowlands the term originally implied only the ablation sediments, i.e. an upper part of what is the last glacial complex in modern understanding. The lowland Upper Pleistocene basal tills, containing marine fossils, were customarily mapped as Sanchugovka, or Rogovaya formations and the same diamict horizons near the uplands were commonly described as a Zyryanka Till. The Sartan label, derived from the latest stage of an alpine glaciation, is even less suitable for the lowland glacigenic formations rich in rafts of marine sediments.

Another major source of stratigraphic errors is a disregard for permafrost and aeolian features, often confused with glacial and reservoir sediments, which can lead to the application of radiocarbon time brackets to wrong or non-existent sedimentary events: to glacial stages instead of warming, as is the case with the Markhida and other flowtills, or to inferring lacustrine events instead of arid phases, as has been done for many ice-rich loess-like sequences.

It would thus be wise in the future to refrain from attaching the worn-out Siberian stratigraphic terminology, in its original or later acquired meaning, to newly studied sequences or paleogeographical stages. The conventional stratigraphic scheme, derived from the very complicated and poorly studied glacigenic sequences of West Siberia, is especially counter-productive if applied to the periglacial record which, in principle, could yield more reliable information on the succession of paleoclimatic events.

The old geological method of applying local geographical names to sedimentary formations is still very much advisable, even if it is not always clear how far such local events can be extrapolated. To avoid the difficulty when alluding to regional objects this author prefers using such descriptive terms as `the last glaciation`, `the uppermost glacial complex`, etc. For long-distance correlation and geochronological interpretations the Siberian scheme is necessary neither. It seems better to compare regional (e.g., North Siberian) sedimentary events with subdivisions of the more reliable European scheme – Late Weichselian, Middle Weichselian, etc.

Acknowledgements

This paper is a contribution to studies of the arctic Pleistocene performed in the course of the Russian-Norwegian PECHORA project funded by the Research Council of Norway and the EC sponsored project Eurasian Ice Sheets (Contract no. ENV4-CT97-0563) coordinated by the European Science Foundation program QUEEN. The author is also indebted to Christian Hjort and Christine Siegert whose thorough reviewing have considerably helped to clarify the message of the above text.

CHAPTER III

CHRONOLOGY OF LAST GLACIAL CYCLE: SIBERIAN DATA

These 8 papers describe attempts to solve the old puzzle of the age of the last glaciation in Siberia by geochronological methods including the regional stratigraphic and sedimentological descriptions. The papers of this chapter are particularly concerned with the geochronometric database in order to get reliable correlation signals for comparison of the Siberian record with much better investigated glacial history of Western Europe. The high time for such studies arrived in the mid-1990-s when the mode of the Pleistocene glaciation and basic regional peculiarities of glacial sedimentation in Siberia, which might hinder application of geochronometric methods, had been roughly clarified (Chapters I and II). After that, more technologically sophisticated dating efforts became imperative.

They arrived on the agenda owing to the synergetic effect by integrating the Russian geological expertise with modern laboratory techniques brought in the 1990-s to the Russian North by researchers from Scandinavian countries, Germany and United States. The collaboration of scientists of different backgrounds armed with diverse research tools proved especially fruitful for solution of a typically interdisciplinary problem – the age and extent of the last glaciation (e.g. the project QUEEN of the European Science Foundation, Svendsen et al., 2004). This author was in the midst of this collaboration, both as a performer and consultant.

The papers of this Chapter tell a part of this international collaboration story from my side. Paper 10 summarises results by Russian investigators principally solving the problem of the age of the last glaciation in West Siberia but insufficient for the European North-East and the Central Siberian Arctic. It is clear from the map of paper 10 in which the area west of the Urals was still covered by Late Weichselian glacial ice – an assumption refuted by further research (Chapter IV).

10. The Last Ice Sheet of the Kara Sea: Terrestrial Constraints on its Age

Valery I. Astakhov

Quaternary International, 1998, Vol. 45/46, P. 19–29

Introduction

A previous paper of 1992 (6 in Chapter II) discussed geological evidence from the West Siberian Plain that can be used for reconstruction of the last glaciation of the Kara Sea. More details on ice dispersal features of this glaciated area can be found in Grosswald (1994), with whom I fully agree in this respect. The major disagreement between Grosswald and the interpretation presented in paper 6, Ch. II, relates to the age of the last shelf-centered ice sheets of Arctic Russia. In this paper I would like to emphasize the temporal aspect of the problem by presenting more geochronological data collected by Russian geologists but not easily accessible by the English speaking community.

Ice dispersal pattern

The last ice sheet of the Kara Sea commonly evokes two questions: i) whether it existed at all, and ii) if it did exist, when? In my opinion, the evidence of the Post-Eemian ice dispersal from the shelf onto adjacent dry lands is overwhelming (Astakhov, 1976, 1979; Andreyeva, 1978; Arkhipov et al., 1980; Kind & Leonov, 1982; Grosswald, 1980, 1994). The ice flow features directed upslope, i.e. southwards, from the Kara Sea are in full accord with the very small size of morainic loops of alpine glaciers mapped along the borders of the Urals and eastern Taimyr mountains (e.g. Gesse et al., 1963; Makeyev & Berdovskaya, 1973). The well pronounced striae, flutes, grooves and eskers, trending across the Paleozoic folds, are accompanied by numerous erratics transported southwards from the Kara Sea coasts (Voronov, 1951; Tarakanov, 1973; Andreyeva, 1978; Astakhov, 1979; Grosswald, 1994, etc.). Such features, striking normally to the arcs of ice-pushed ridges (Fig.1), certainly rule out any subsequent ice flow from highlands.

Fig. 1. Location map and some features of the last sheet glaciation. 1 – ice-pushed ridges by terrestrial geological surveys, photogeological interpretation and marine seismics (Astakhov, 1976, 1979; Arkhipov et al., 1980; Kind & Leonov, 1982; Epstein & Gataullin, 1993); 2 – same, inferred from bathymetric maps; 3 – limit of last ice advance over the uppermost interglacial formation; 4 – suggested limit of a possible ice advance in 30–15 ka BP span; 5 – striae and flutes across Paleozoic structures; 6 – radiocarbon dated sequence overlying the uppermost till (see Table 1); 7 – same within the possible West Kara ice sheet; 8 – profiles in Fig. 2.

Nevertheless, many authors (e.g. Biryukov et al., 1988) disregard this clear evidence and advocate the idea of Late Weichselian ice caps positioned upon the low mountains, which presumably advanced east, west and north onto the Siberian lowlands and shelves. Such highland ice sheets should have left large terminal moraines concentric to the corresponding mountainous massifs, which they did not. Moreover, the northernmost tip of the Urals is devoid of any, even small morainic loops, which could be ascribed to montane glaciers. The available morainic ridges are transverse to the Uralian range and too large for any conceivable valley glaciers (Astakhov, 1979; Arkhipov et al., 1980).

Since the ice directional features show ice dispersal from topographic lows upslope, not vice versa, the hypothesis of highland ice domes can only be supported by such ambiguous evidence as: i) the occurrence of hummock-and-lake topography mostly along the highlands, ii) the hard-rock clasts being more plentiful toward the mountains, and iii) the presumed lack of surficial tills in the Arctic lowlands.

The idea of a driftless area in the Arctic lowlands stems from observations of the 1960s when diamictons with marine fossils and scarce pebbles were mistaken for glaciomarine sediments. This error was first revealed by Kaplyanskaya & Tarnogradsky (1975) to be later confirmed by other investigators (Chapter I and Komarov, 1986). As a result, a surficial till, lying on top of interglacial marine sediments and containing fossil glacial ice, is now mapped everywhere in the Arctic (Fig. 2.).

The distribution of highland clasts is hardly a weighty argument because northern highlands are always surrounded by fields of hard-rock pebbles regardless of the actual glacial history. Nobody has ever been able to demonstrate that the Uralian and Mid-Siberian clasts were transported to the lowlands specifically during the LGM. Such pebbles often originate from non-glacial gravels and are always accompanied by more informative balls of soft rocks derived from topographic lows.

The question of the fresh hummock-and-lake landscapes is more complicated, but sufficient explanations of its peri-montane location are available after research of the last two decades. The scarcity of hummock-and-lake landscapes in the high Arctic can partly be explained by retarded deglaciation, owing to which large slabs of basal glacial ice still persist within the thick permafrost as constituents of `primordially frozen tills` that need further warming to produce expressive glaciokarst features (Kaplyanskaya & Tarnogradsky, 1977; Astakhov & Isayeva, 1988).

On the other hand, the lowlands were mostly occupied by central parts of the ice sheet relatively poor in debris. Resultant thinner tills could readily be destroyed by the late glacial sea penetrating into the isostatic trough amongst fields of stagnant ice, as well as by lakes and streams, as can be seen in Fig. 2. These late glacial aqueous agencies led to a flatter topography.

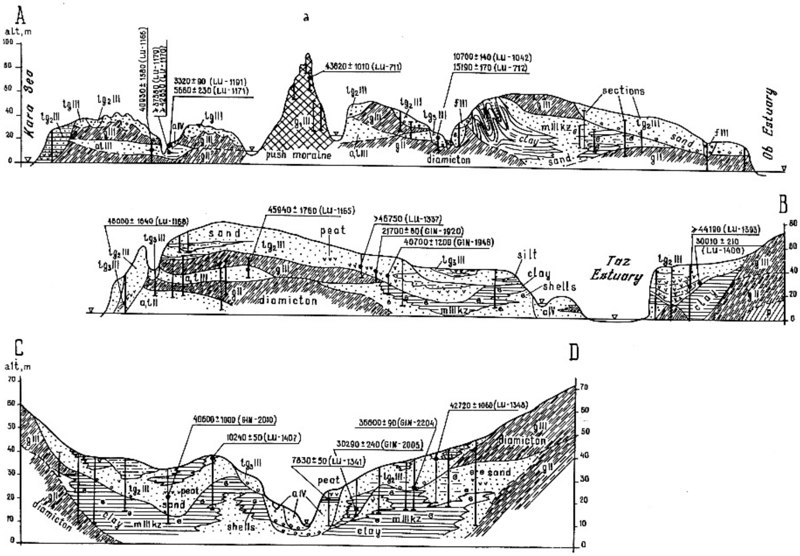

Fig. 2. Geological profiles of Arctic West Siberia compiled by A. Lavrov and L. Potapenko (mapping report, 1983), based on photogeological survey data. For location see Fig. 1. Interpretation symbols by Lavrov and co-authors: gII – Middle Pleistocene till; gIII – Upper Weichselian till; mIIIkz – Kazantsevo Formation with shells of boreal marine mollusks; a,1III – alluvial and limnic sediments; lg, lg_2, lg_3III – glaciolacustrine sand, clay and loess-like silts; aIV – Holocene alluvium; Pg – Paleogene sediments. Black dots indicate radiocarbon samples, a – Late Weichselian ice limit suggested in paper 2 of Chapter I.

Conversely, the debris-laden marginal parts of the Kara ice sheet, affected by upslope compressive ice flow, were stagnating much farther south at higher altitudes, where they survived late glacial inundations to produce thick tills and rough topography due to perforating activity of late glacial sinking lakes. Close to the highlands the till of the last ice age may be 40-50 m thick (Kind & Leonov, 1982; Komarov 1986; Astakhov & Isayeva, 1988).

It is also significant that in the marginal parts of the former ice sheet pebbles in the till are oriented longitudinally, i.e. parallel to the mountain front (Astakhov, 1979; Sukhorukova & Gaigalas, 1986), which is another phenomenon not explained by the highland ice theory.

Thus, the most expressive surficial features such as push moraines, glacial striae, orientated pebbles unanimously indicate to the last ice advance from the low Kara Sea coastlands upslope. Only the most prominent of them are pictured in Fig.1, but they are sufficient to demonstrate where the last glacier came from.

Judging by the sub-parallel (not radial) striae and flutes across Novaya Zemlya, Vaigach and Pai-Hoi, these narrow Paleozoic ranges could not be major ice dispersal centers (Grosswald, 1994). The Novaya Zemlya archipelago is fringed along the northwest by a huge submerged double ridge which is, according to Russian seismic surveys and geotechnical drilling (Epstein & Gataullin, 1993), built of diamicton more than 60 m thick. As traced northeastwards by bathymetry this ridge is 700 km long (Fig. 1). Its spatial pattern does not support the idea of narrow ice caps positioned along the crest of Novaya Zemlya (Biryukov et al., 1988), and fits with the submerged hummocky terrain in the western Kara Sea by Dibner (1970), thus suggesting a rotund ice sheet spreading westward over Novaya Zemlya.

Geochronological data

Now we attend to the most acute problem of the age of the last Kara glacier. The available evidence is probably not sufficient to correlate its advances with Scandinavian and Laurentide glacial stades. Still the data accumulated and presented below seem to be ample to at least eliminate certain geochronological speculations about the Post-Eemian glacial maximum.

The popular idea of the Late Weichselian age of the LGM in the Russian Arctic is shared by both groups of scientists: those who profess the hypothesis of restricted ice domes upon highlands (Biryukov et al., 1988) and those who accept the Kara shelf ice domes (Volkov et al., 1978; Arkhipov et al., 1980; Grosswald, 1980). There are others who notice discrepancies in the evidence presented and either leave ample room for doubts (Kind & Leonov, 1982), or even entirely reject Weichselian inland ice (Fisher et al., 1990). The Late Weichselian age of the LGM is most ardently advocated by Grosswald and Siberian investigators Arkhipov and Volkov. Three sets of arguments are used: i) the fresh appearance of some hummock-and-lake topography in the Arctic; ii) limnic rhythmites in the central West Siberian Lowland which are perceived as sediments of a huge proglacial lake with a level up to 120-130 m; iii) finite radiocarbon dates from beneath the topmost till sheet.

Now that relict glacial ice in the Siberian Arctic has been identified (Kaplyanskaya & Tarnogradsky, 1977; Astakhov & Isayeva, 1988) it is clear that basal parts of former ice sheets, fused together with thick permafrost, can survive through minor climatic fluctuations until the permafrost is completely destroyed by a major warming. In the Kara Sea catchment area such a warming occurred only in the span of 9 – 3 ka BP, when south of the Arctic Circle the permafrost table dropped well below the level of seasonal freezing. Yet north of the 68th parallel the ancient continuous permafrost survived even after the Mid-Holocene warming (Baulin et al., 1984).

The present fossil glacial ice could have been left by any Weichselian glaciation. These days deep sink lakes are developing on buried glaciers regardless of the age of the corresponding ice advance. The resultant glacial karst landscapes are most conspicuous in the southern Arctic, between 68 and 66° N, where permafrost became discontinuous in the course of the Holocene warming (Kaplyanskaya & Tarnogradsky, 1977; paper 5, Ch. II).

The question of the existence of a Late Weichselian proglacial lake has been discussed elsewhere (Astakhov, 1989, papers 6, Ch. II and 14, Ch. III). Special research has revealed no real signatures of an extensive fresh-water sea but only sporadic lenses of sink-hole limnic sediments incorporated into a discontinuous mantle of loess-like silts with ice-wedge casts, paleosols and mammal bones. The lenses of limnic rhythmites make numerous oval and doughnut-shaped knolls at all altitudes from 40 to 150 m, irrespective of underlying surface. Local horizons of ice-wedge casts and sometimes buried tundra soils are suspended within the shallow-pond fine sand-silty sediments up to 10-15 m thick (paper 14, Ch. III). Hummock-and-bog landscapes built of the loess-like and stratified silts widespread in central-southern West Siberia far south from the drift limit are often perceived as glacial karst forms by investigators unfamiliar with permafrost processes.

Actually these knolls are not glaciokarst, but typical thermokarst features, the difference being in significantly lower content of the parental ice formed subterraneously. Glacial karst develops on thick relatively clean subaerial ice where sinking and lateral shifting of deep thaw lakes are not impeded by surrounding mineral matter. Thermokarst ponds, evolving normally along polygonal ice wedges, are more shallow and restricted in their development by ground massifs with low ice content and by thickness of the permafrost. That is why inverted thermokarst knolls, unlike kames, are generally flatter, often clustered in polygonal patterns and commonly have, due to isotropic thermoerosion of frozen soil, a fairly regular oval or circular form in plan. Fig. 3 illustrates the origin of the West Siberian landscapes south of the LGM limit as understood by this author, the principal idea being borrowed from special works by Boitsov (1961) and Kaplyanskaya (in Mikhankov, 1973).

Thus, south of the suggested limit of the last glaciation (Fig. 1) there is no geological phenomena that would demand a high-level impounding by an extensive lake to explain the Late Weichselian paleogeography. On the contrary, the hypothesis of the Late Weichselian ice advance has to explain why in the flat lowland the great north-flowing rivers failed to flood the whole country and arrest the development of permafrost features, soils and large mammals. The thick sediment fill of the Turgai Valley with the dates ca 19 and 29 ka BP (in Grosswald, 1983) cannot be taken as reliable evidence of a Late Weichselian spillway directed to the Aral Sea, because they are mostly lacustrine clay, loess, aeolian sand and mass-movement diamicts. Only the thin gravel over the bedrock bottom relates to a south-directed overflow of proglacial water. It occurs at altitudes ca 40 m, suggesting not the 120 m but much lower level of the overflow and not necessarily during the Late Weichselian (paper 6, Ch. II).

At last, we should consider the glaciated area itself. The thick and diverse sedimentary succession overlying the topmost till includes glaciomarine and varved clays, outwash sand, multi-staged sink-hole rhythmites and frozen loess-like silts with long syngenetic ice wedges (paper 6, Ch. II).). The classical concept viewed the numerous mammoth remains scattered throughout the Arctic as evidence of a relatively old age of the last glaciation of West Siberia as compared to the European one (Sachs, 1953). Modern authors consider radiocarbon dates as far more important evidence. Grosswald in particular has quoted 23 finite dates ranging from 26 to 52 ka BP as recovered from sediments underlying the topmost till of the Kara Sea catchment area (Grosswald, 1980, 1983; Grosswald & Goncharov, 1991). He disregards much more numerous `old` radiocarbon dates obtained from sediments postdating the latest till. His idea is that such dates, being derived from redeposited organics, indicate only maximum ages (Grosswald & Goncharov, 1991).

This idea, although not groundless, has led its authors too far. In the work mentioned, they, basing on several finite dates, had to push the Late Weichselian ice limit 500 km down south as compared to Fig. 1 of this paper. If such dates are only taken into account, one could draw the limit even farther south in the periglacial zone, because there are finite radiocarbon dates from beneath a till even on 61° N (Arkhipov et al., 1980). The error stems from underestimating the contamination by younger carbon which happens very frequently, especially in areas where ground waters were very active during the degradation of the Pleistocene permafrost. Sulerzhitsky, a radiocarbon expert, quotes a number of finite dates from Eemian sediments. In his opinion most dates on shells and even logs are too young. According to many years of his dating experience in the Arctic, more stable results are received on bones. The experimental dating of driftwood from the same formation has shown a rapid decrease of radiocarbon content in logs recently released from permafrost, which is explained by bacterial activity (in Kind & Leonov, 1982).

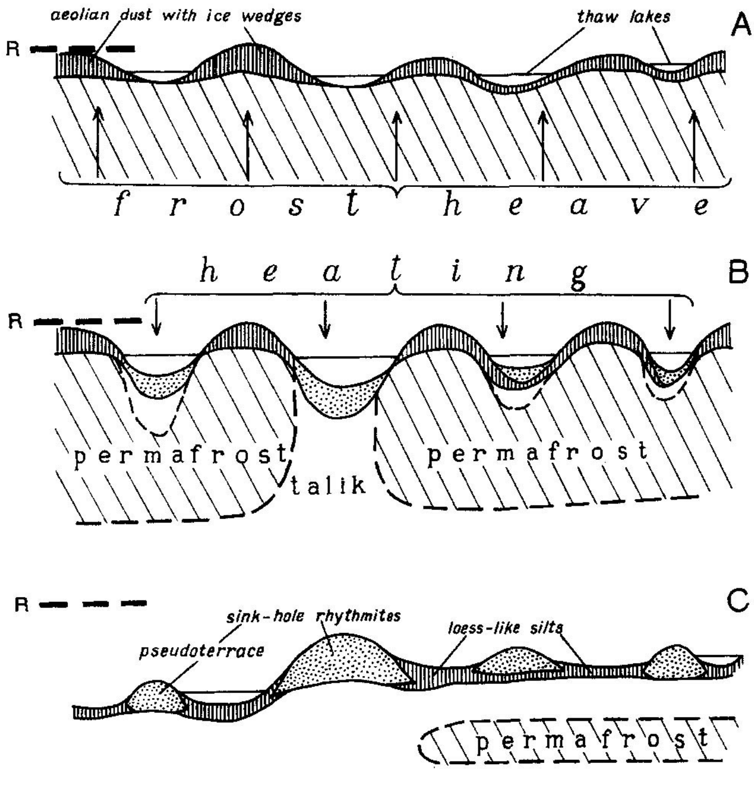

Fig. 3. Origin of the loess-limnic association of West Siberia resultant from thermokarst inversion of Weichselian periglacial landscapes. A – solid Weichselian permafrost; B – initial climatic amelioration with silt and sand accumulating in sinking lakes; C – present state after the sagging of perennially frozen terrains; R – reference plane.

Taking into account all these complications, no single radiocarbon date can be referred to as completely reliable. We do not have any other option than to consider the whole collection of available dates in search for age indications. A typical spectrum of radiocarbon dates in the West Siberian Arctic is shown in Fig. 2. One can see a variety of radiocarbon ages in sediments overlying the uppermost till with occasional finite dates in the underlying strata.

The general distribution of the finite radiocarbon dates older than 15 ka for the Kara Sea catchment within the limit of the last glaciation is presented in Fig. 4. The bulk of them (160) is derived from sediments not covered by till, and only 48 dates (23%), including all shell samples, originate from beneath the topmost till. Most of the sub-till dates has been obtained on materials notoriously prone to redeposition and contamination such as mollusk shells, driftwood and plant detritus. Almost all such dates are isolated for a studied sequence: no successive series of radiocarbon ages being known in the sub-till position except Malaya Heta section with 5 finite dates (Kind, 1974). But in the latter case there is no sedimentological proof that the overlying diamicton is not flowtill which is very likely, as suggested by the well-studied sequence in the same Yenissei river valley (paper 5 in Ch. II).

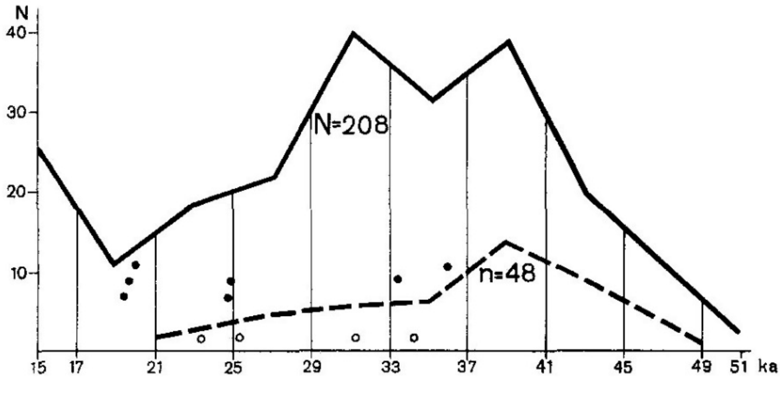

Fig. 4. Distribution of finite radiocarbon dates in the Kara Sea catchment area covered by Weichselian tills (within limit 3 in Fig. 1). Solid line – total number of dates in 4 ka intervals (N=208); broken line – number of dates from sediments underlying the uppermost till (n=48). Black dots represent dates obtained on mammoth remains from sediments overlying the topmost till, circles – dates obtained on shells from sediments underlying the till.

The dates from above the surficial till, on the contrary, frequently occur in series of 3 to 12 successive ages (Table) and mostly originate from mammal remains, local herbs, shrubs, peat and partly plant detritus. It is noteworthy that all dated mammoth remains, including obviously *in situ* frozen carcasses, have been found in sediments overlying the uppermost till.

The entire population of the radiocarbon dates may be sorted out to get a more clear indication of the age of the surficial till. Table 1 presents 92 dates, of which 76 is 17 ka and older, all from sediments indisputably postdating the last glaciation. The location of the dated sequences is shown in Fig. 1. To make the selection of radiocarbon dates in Table 1 less arbitrary the following negative criteria have been applied: i) no dates on shells; ii) no isolated dates (except remains of mammals and large logs from perennially frozen sediments); iii) no sections with significant age inversions. Some comments on the list of dates selected for Table 1.

Most of the dates are not likely to yield maximum but rather minimum age estimates. It is especially obvious in the case of frozen mammoth carcasses which cannot be transported too far. The comparison of dates on outer and inner layers of mammoth tusks is needed to assess how thoroughly young humic acids were removed in laboratory (Makeyev et al., 1979). An additional proof of reliability of the mammoth dates is provided by the very similar ages obtained on fossil plants from the stomach of the beast (site 5) or from the enclosing sediments (site 8).

The dated tooth of woolly rhinoceros (site 11) was extracted from the lower jaw with a full set of teeth preserved. The jaw was found in 1974 in a sandy Yenissei terrace at the mouth of P. Tunguska river (Fig. 1). According to Grosswald & Goncharov (1991) this terrace at 80 m a.s.l. must be very young, as they ascribe the Late Weichselian maximum (ca 20 ka BP) to the higher terraces at 120 and 160 m a.s.l., presumably left by the proglacial lake. This date of ~ 21.5 ka BP obtained on the intact jaw of the terrestrial beast strongly supports the alternative view, according to which the Middle-Late Weichselian of the proglacial area is recorded in this low alluvial terrace descending to the north along the Yenissei valley (Laukhin, 1981) and not in a limnic sequence.

A remarkable succession of radiocarbon dates ranging from 12 to 35 ka BP at site 13 (Cape Sabler in Lake Taimyr) was obtained from undisturbed surficial limnic silts and sands with seams of local plant remains (Kind & Leonov, 1982). The non-finite date on a frozen mammoth (site 6) comes from the area glaciated during the Late Weichselian according to Grosswald and Goncharov (1991), but the site is likely to remain beyond the limit of Weichselian glaciations (Fig. 1). The

Table 1. Radiocarbon dates from sediments overlying the topmost till of the of the Kara Sea catchment area

Site	Date, years BP	Dated material	Reference
		A. Mammals and related plants	
1	19,640±330 (LU-654A)	Mammoth tusk, outer part	Makeyev et al., 1979
	19,270±130 (LU-654B)	Mammoth tusk, inner part	*ibid.*
2	19,970±110 (LU-688)	Mammoth tooth	*ibid.*
3	24,910±200 (LU-749A)	Mammoth bone	*ibid.*
	24,960±210 (LU-749C)	Same	*ibid.*
4	11,500±60 (LU-610)	Mammoth tusk	*ibid.*
5	11,450±250 (T-297)	Mammoth flesh	Heintz & Garutt, 1964
	11,700±300 (MO-3)	Accompanying plants	*ibid.*
6	>53,170 (LU-1057)	Mammoth flesh	Arslanov et al., 1980
7	36,000±4300 (T-169)	Same	Heintz & Garutt, 1964
	33,500±1000 (T-298)	Same	*ibid.*
8	30,100±300 (GIN-3742)	Mammoth bone	Find by Kaplyanskaya in Astakhov, 1992
	29,700±270 (GIN-3743)	Accompanying plants	*ibid.*
9	25,400±300 (GIN-2210)	Mammoth bone	Avdalovich & Bidzhiyev, 1984
	32,600±1300 (GIN-2026)	Underlying peat	*ibid.*
	34,500±1000 (GIN-2027)	Same	*ibid.*
10	25,400±300 (GIN-2210)	Mammoth bone	*ibid.*

11	21,445±145 (TUa-675)	Tooth of woolly rhinoceros	Find by Astakhov
		B. Plant remains	
12	35,800±500 (Gin-1498)	Stump	Kind et al., 1981
	35,800±1000 (GIN-1497)	Plant detritus	*ibid.*
	>46,000 (GIN-1500)	Plant detritus	*ibid.*
	>48,600 (GIN-1499)	Stump	*ibid.*
13	12,100±100 (GIN-1528)	Detritus, depth 1 m	Kind & Leonov, 1982
	11,600±200 (GIN-1527)	Same, depth 4 m	*ibid.*
	12,000±150 (GIN-1289)	Same, depth 5 m	*ibid.*
	17,750±300 (GIN-1290)	Same, depth 7 m	*ibid.*
	18,400±1000 (GIN-1526)	Same, depth 8 m	*ibid.*
	21,400±1100 (GIN-1525)	Same, depth 11 m	*ibid.*
	24,900±700 (GIN-1291)	Peat, depth 15.5 m	*ibid.*
	24,200±800 (GIN-1524)	Detritus, depth 17 m	*ibid.*
	30,300±400 (GIN-1521)	Detritus, base of section	*ibid.*
	29,600±1000 (GIN-1522)	Same	*ibid.*
	30,400±600 (GIN-1523)	Same	*ibid.*
	34,500±2000 (GIN-1292)	Same	*ibid.*
14	29,600±1100 (GIN-3818)	Peat	Fisher et al., 1990
	29,900±1200 (GIN-3819)	Same	*ibid.*
	40,600±900 (GIN-3803)	Plant detritus	*ibid.*
	43,800±1500 (GIN-3825)	Same	*ibid.*

15	25,700±400 (GIN-3480)	Plant detritus	*ibid.*
	32,070±300 (GIN-3494)	Peat	*ibid.*
	32,500±400 (GIN-3479)	Same	*ibid.*
	38,000±700 (GIN-3838)	Peat	*ibid.*
	44,400±800 (GIN-3839)	Plant detritus	*ibid.*
16	16,780±80 (GIN-3939)	Same	*ibid.*
	25,700±400 (GIN-3934)	Same	*ibid.*
	32,710±400 (GIN-3940)	Same	*ibid.*
17	28,490±450 (MGU-687)	Peat	Danilov & Parunin, 1982
	35,050±600 (MGU-686)	Peat	*ibid.*
	32,000±1600 (GIN-2876)	Plant detritus	Sulerzhitsky et al., 1984
18	35,400±300 (GIN-140)	Log	Kind, 1974
	35,800±600 (GIN-76)	Plant detritus	*ibid.*
	39,000±460 (GIN-328)	Log	*ibid.*
	>50,000 (GIN-327)	Log	*ibid.*
19	31,100±800 (GIN-3674)	Twigs with bark	Astakhov & Isayeva, 1988
	32,500±400 (GIN-99)	Plant detritus	Kind, 1974
20	42,780±1135 (SOAN-2519)	Peat	Sukhorukova et al., 1991
	39,340±750 (SOAN-2518)	Same	*ibid.*
	45,520±1270 (SOAN-2517)	Same	*ibid.*
21	34,200±1000 (GEN-2872a)	Plant detritus	Astakhov *et al.,* 1986
	35,200±1500 (GIN-2872b)	Wood	*ibid.*
22	32,700±1500 (GIN-2189)	Peat	Avdalovich & Bidzhiyev, 1984
	38,000±500 (GIN-2190)	Same	*ibid.*

23	21,900±900 (GIN-2469)	Peat	Vasilchuk et al., 1984
	25,100±220 (GIN-2471)	Same	*ibid.*
	28,600±800 (GIN-2638b)	Same	*ibid.*
	30,200±800 (GIN-2470)	Same	*ibid.*
24	17,100±600	Plant detritus	Bolikhovsky, 1987
	16,830±670	Same	*ibid.*
	16,680±500 (MGU-1047)	Same	*ibid.*
	16,520±550	Same	*ibid.*
	18,380±700	Same	*ibid.*
25	38,600±100 (GIN-1926)	Peat	Avdalovich & Bidzhiyev, 1984
	37,100±400 (GIN-1928)	Same	*ibid.*
	39,000±1500 (GIN-1927)	Same	*ibid.*
	38,900±1200 (GIN-1929)	Same	*ibid.*
	38,800±600 (GIN-1930)	Same	*ibid.*
26	22,700±300 (GIN-2473)	Peat seam	Vasilchuk et al., 1984
	22,600±600 (GIN-2475)	Same	*ibid.*
	23,500±400 (GIN-2474)	Same	*ibid.*
	24,300±300 (GIN-2476)	Same	*ibid.*
	39,100±1500 (GIN-2477)	Same	*ibid.*
27	35,320±1570 (LU-1137)	Frozen log	Astakhov, 1992

	C. Dates from the western sector		
28	14,590±300 (Ri-285)	Peat seam	Gataullin, 1988
29	13,830±260 (Tln-1059)	Fossil weeds	*ibid.*
	13,970±140 (Tin-1026)	Same	*ibid.*
	13,340±200 (GIN-5196)	Fossil moss	Bolikhovskaya & Bolikhovsky, 1992
	13,280±150 (GIN-5197)	Same	*ibid.*
30	15,310±650 (LU-1188)	Peat seam	Malyasova, 1989
31	8210±90 (LU-1462)	Peat seam	*ibid.*
	10,550±160 (LU-1466)	Basal peat	*ibid.*
	9230±100 (LU-1464)	Peat seam	Krasnozhon et al., 1982
	15,120±120 (LU-1446)	Basal peat	*ibid.*

successions of dates at locations 23, 24 and 26 quoted by Astakhov (1992) have been obtained from thick loess-like silts with very long syngenetic ice wedges that were developing through the depositional history of this postglacial formation (Vasilchuk et al., 1984; Bolikhovsky, 1987).

A date of site 19 from alluvium at the mouth of river Kureika has been obtained on small shrubs with bark intact (paper 5 in Ch. II) to support another date from the same Second Terrace on plant detritus (Kind, 1974). A nearby section 18 in the permafrost observation pit of Igarka town contains perennially frozen logs and stumps in sink-hole silts that have never melted out since deposition. The pre-Late Weichselian dates from sites 19 and 20 are of interest because they originate from loess-like sediments overlying the `Late Weichselian till` by Grosswald and Goncharov (1991).

Dates younger than 17 ka BP are included in Table 1 only if they either continue a successive series of older ages, or occur in the western sector (Fig. 1) where a Late Weichselian age for the uppermost till is plausible. Any randomly chosen date from this list may prove to be wrong, but the whole constellation unambiguously points in one direction: the uppermost till east of 70° E must have been deposited prior to 40-50 000 radiocarbon years BP, most likely beyond the range of the radiocarbon method. The assemblage of old radiocarbon dates overlying the uppermost till in Table 1 even numerically leaves Grosswald's statistics far behind, let aside the reliability of the ages obtained. It is not accidental that among the dates cited by Grosswald and other advocates of the Late Weichselian ice sheet no measurements are mentioned on undisputably *in situ* materials such as remains of large mammals. If the aforementioned negative criteria of selecting were applied to sub-till finite dates, only the 5 values in the span of 35–44 ka BP from Malaya Heta section (Kind, 1974) would be left to offset more than 70 similar dates obtained from above the till.

These 5 dates, being in discrepancy with the rest of the radiocarbon evidence, obviously call for a reinvestigation of the site, especially of the overlying diamicton, which might be a flowtill deposited well after the ice sheet stagnation.

Discussion

The above data can serve as constraints on speculations about the last ice sheet centered on the Kara Sea shelf. The size of ice-pushed ridges and position of the maximum ice limit (3 in Fig.1) permit a suggestion that thicker inland ice was occupying not only present shallows between Severnaya Zemlya and northern Yamal, but the western Kara Sea along with the Eastern Novaya Zemlya Trough up to 400 m deep as well. In this case the southern ice margin must have been distanced from the ice divide suggested in Grosswald (1980) by 500 to 1000 km (as in Scandinavia), but not that far south as in later models by Grosswald and Goncharov (1991) and Grosswald (1994).

The maximum ice thickness, which may have varied from 2 to 3 km, is difficult to estimate more precisely until we know more about marine geology of the Kara Sea. Judging by the thick Pleistocene permafrost containing fossil glacial ice, the glacier must have been very cold, which implies thicker ice. On the other hand, the very soft clayey substrate comprised of Cretaceous, Paleogene and Quaternary deposits, may have flattened the ice dome. The striated and polished surfaces on the northern tip of the Urals and on the right bank of the Yenissei show that the ice sheet reached 400-500 m in thickness even very close to its margin (Arkhipov et al., 1980), which is more consistent with the thicker ice model. In any case, the observed signatures of cold and thick upslope moving ice are incompatible with the idea of mountainous ice dispersal centers. It is even difficult to imagine how such an ice cap could reside upon a rugged range of the Polar Urals 1 km high and only 50-80 km wide.

The available geological evidence from both glaciated and proglacial areas (paper 6 in Ch. II) along with the radiocarbon data presented above do not permit any ice advances from the Kara shelf onto the Siberian landmass within the range of the radiocarbon method.

However, the age of the last ice sheet in the deepest western part of the Kara Sea between Yamal and Novaya Zemlya is less certain. Submerged hummocky relief in this part of the sea was mapped long ago (Dibner, 1970). The available ice direction features and erratics on the adjacent uplands indicate that there was a recent ice advance from the western Kara Sea to the south and west (Arkhipov et al., 1980; Grosswald, 1994) and also to the east onto Yamal Peninsula (Astakhov, 1979; Gataullin, 1988). The huge morainic ridges on the adjacent Barents Sea shelf (Fig. 1) tell the same story.

Several finite radiocarbon dates are known from sediments underlying the upper till of this area, but their validity may be no better than elsewhere in the Kara Sea basin. The lack of old radiocarbon dates from sediments overlying the till seems to be more important (Figs. 1 and 2 and Table). The available dates at least limit the active ice phase for the western sector to times earlier than 15 ka BP. Especially intriguing are 15 ka BP dates from peats of Novaya Zemlya. Peat seams of this archipelago have yielded also dates 10.5, 8.2 and 6.4 ka BP. Presently peat accumulation occurs only on the mainland not farther north than 68°N (Krasnozhon et al., 1986; Malyasova, 1989). This implies that several episodes with climate warmer than the present one have occurred since the last ice advance.

The western Kara till cannot be very young, which follows from the overlying loess-like silts 5 m thick on the western coast of the Yamal (site 29), containing syngenetic ice wedges and 13-14 ka BP dates (Gataullin, 1988; Bolikhovskaya & Bolikhovsky, 1992). The Younger Dryas age for this glaciation can be safely eliminated, but a Late or Middle Weichselian ice advance from the western Kara Sea is still possible, although the configuration of this younger ice sheet (4 in Fig. 1), inferred from the available evidence (Fig. 2) looks rather artificial at its southeastern margin. More evidence is needed to decide whether the West Kara glacier was just a remnant part of the large Early Weichselian ice sheet or represented a second Weichselian activation of shelf inland ice. Such evidence is obtainable from Arctic European Russia.

The most significant inference from the data on Novaya Zemlya and adjacent areas is the constraint on the age of the last glacial maximum of the Barents Sea. Even if a Late Weichselian age of the last ice sheet is eventually proven, it would apply only to shelf areas immediately east and west of Novaya Zemlya, the huge marginal ridges (1 in Fig. 1) included. A more extensive ice cover of the Barents Sea shelf would inevitably demand a more or less symmetrical ice flow in the opposite direction, of which no signs exist in West Siberia east of 70° E.

The major part of the Barents shelf must have been glaciated earlier: otherwise the fresh ice flow features on the adjacent dry land, especially across Novaya Zemlya (Grosswald, 1994), would have been directed eastwards. The narrow strip of Novaya Zemlya highlands, likewise the Urals, could never accomodate an extensive ice dome responsible for such large terminal features as pictured in Fig. 1.

The geological and geochronological evidence for an early culmination of the last Kara ice sheet raises the issue of the maximum time limit for this glaciation. On the whole it is predetermined by the occurrence of the uppermost till on top of the Kazantsevo interglacial sediments (Fig. 2). This marine formation is conventionally correlated with the Eemian because it contains numerous shells of boreal mollusks such as *Arctica islandica* (Sachs, 1953) which presently can survive only west of the White Sea.

There are also works describing another (Karginsky) marine formation with boreal fauna. Many authors, taking into account finite radiocarbon dates from marine sediments, suggest a second transgression in the span of

50 to 26 ka BP, i.e. correlative to isotope stage 3 (Kind & Leonov, 1982). Sulerzhitsky, a radiocarbon expert, however, is sure that all such dates are obtained from the Eemian organics (Fisher et al., 1990).

On the other hand, in European Russia the last boreal transgression is reliably correlated with the Eemian, and the previous one is often referred to as intra-Saalian (Yakovlev, 1956). Zubakov (1986), who thinks that the finite radiocarbon dates are not the proof of a Weichselian warm-water sea, has suggested a similar framework for West Siberia, placing the Kazantsevo Formation at 230–170 ka BP (isotope stage 7) and Karginsky Formation with finite dates on shells – at 130–110 ka BP (stage 5). If the latter correlation is correct, then the last glaciation of the Kara Sea could even belong to stage 6. This option is difficult geologically. The uppermost till contains large fields of fossil glacial ice (Astakhov & Isayeva, 1988; Astakhov, 1992) which in the latter case should have survived through the very warm Eemian transgression.

In the view of this important fact and data described elsewhere in this paper the only reasonable solution is to place the last ice sheet of the Kara shelf and adjacent Siberia in the span of 110 to 50 ka BP, probably within isotope stage 4. This, of course, refers only to the live ice phase because large isolated masses of dead ice are still buried in the surficial drift.

Acknowledgements

My sincere thanks are due to Dr. Alexander Lavrov, Aerogeologia, Moscow, for the permission to use his unpublished geological profiles and to Dr. John Inge Svendsen, University of Bergen, for helping out to date the rhinoceros tooth.

11. THE AGE OF MAMMOTH FAUNA ON THE LOWER OB

V. I. ASTAKHOV, KH. A. ARSLANOV AND D. V. NAZAROV

DOKLADY EARTH SCIENCES, 2004, VOL. 396 (4), P. 538–542.

Translated from Doklady Akademii Nauk, 2004, vol. 396 (4), P. 253–257.

This paper discusses the dating results on fossil organics on the Ob River along the Arctic Circle (Fig. 1) in order to shed more light on the problem of the last arctic glaciation. The dated samples were collected by Astakhov and his associates during field seasons 2000 and 2001. Radiocarbon measurements have been performed by Arslanov in the Geochronological Laboratory of St. Petersburg University by the liquid-scintillation method (Arslanov, 1987). The bones have yielded collagen that has been purified of contaminants (humic acids, saprophytes, and others) using an improved technique (Arslanov & Svezhentsev, 1993). Radiocarbon content has been measured with a higher statistical reliability (Table 2). OSL ages in Fig. 2b have been obtained by optically stimulated luminescence on quartz grains of sand samples (Murray & Wintle, 2000) by A. Murray in the Nordic Laboratory for Luminescence Dating, University of Aarhus (Denmark). A comprehensive description of OSL results in the Russian Arctic is presented in the special issue of Quaternary Science Reviews 2004 containing the final report of the Quaternary Environment of the Eurasian North program.

Dating mammoth fauna is directly related to the problem of the last glaciation. In the north of Europe, remains of large mammals are either older than 25 ka or younger than 15 ka. This gap is attributed to the low productivity of glacial deserts during the Last Glacial Maximum (LGM) 25–15 ka ago. However, far beyond the Urals and east of the 75th meridian, bones of large animals dated within this interval are not rare in the glaciated terrains of Gydan, Taimyr, and even Severnaya Zemlya (Makeyev et al., 1979; Sulerzhitsky, Romanenko, 1997; Vasilchuk et al., 1997). On the other hand, glacial features of that age are totally absent on the flat mainland of the Russian Arctic, which has been amply proven by recent efforts by Russian, West European, and American geologists in the Pechora Basin, Yamal, and Taimyr (Astakhov et al., 2000; Forman et al., 1999; Mangerud et al., 2002; Svendsen et al., 1999). One of the most convincing arguments for an older age of the last glaciation is provided by numerous remains of the mammoth fauna not covered by diamicts.

The well-exposed Quaternary sections of the Lower Ob area have always been considered crucial for the Pleistocene stratigraphy and paleogeography of West Siberia. Many researchers have concluded that the last glaciation either did not reach this area at all (Zubakov &. Levkovskaya, 1969) or covered it partly in the very beginning of the last glacial cycle, i.e. earlier than 50 ka (Lazukov, 1970). However, according to the official viewpoint, adopted in the Regional Stratigraphic Scheme, the last glacier invaded this area ~ 18-20 ka ago (Volkova & Babushkin, 2000). This idea was based on several finite radiocarbon dates obtained by scientists of the Siberian Division of Russian Academy of Sciences in the Salehard area from the lenses of diamicton interpreted as tills of an ice sheet (Arkhipov et al., 1977).

We have not found *bona fide* surficial tills anywhere in this area (Fig. 1). Sparse lenses of thin diamicton are spatially interspersed not with glacial formations but with sheets of loess-like silt. At the base, they show only permafrost disturbances. Therefore, these diamicts should be attributed to solifluction processes.

Plant remains *in situ* have been collected in a deep sand pit of Aksarka village (Fig. 1, section 18) and in fresh headwalls of triangular ravines cutting the southern bank of the Ob River along the former ice wedges close to the Pyak-Yaha and Pichuguy-Yaha creeks. The bones were mainly picked at the foot of the north-facing bluff of the Nadym Ob at 3-4 km downstream of the Pichuguy-Yaha creek and near the mouth of the Pyak-Yaha creek (Figs. 1, 2; sections 25, 26). The subvertical bluff, rising to 40 m above the shallow wave-cut platform, is persistently undercut by storm floods and spring ice floes. Local fishermen commonly find fresh, unrounded bone fragments of mammoth fauna along the aforesaid stretch of the river bank. It was not too difficult for us to collect many large bone fragments of mammoth, muskox, horse, woolly rhinoceros and reindeer in 3-4 hours (Table 1). The frequency of bone finds grows towards creek valleys concurrently with descent of the bluff and loess cover.

This mammoth fauna site is not unique. Well-preserved bone remnants in a similar position are reported by local hunters also in the upstream of the left-bank area. Right-bank finds of mammoth remains, unfortunately undated, have been described in the upper part of the right-bank terrace 2 (Zaionts & Ziling, 1972). We also found splinters of bones and teeth on the eastern shore of Shuryshkary Sor, just north of the village (Fig. 1, section 31), where an outcrop of thick autochthonous peat is known.

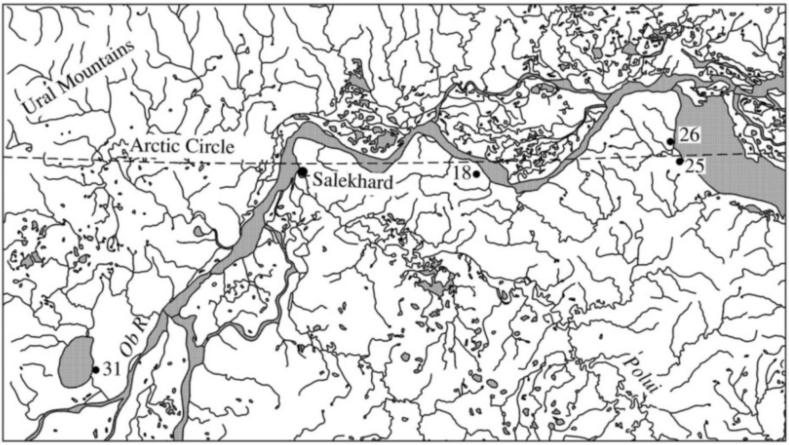

Fig. 1. Location of dated Quaternary sequences on the Lower Ob.

The base of the Nadym Ob bluffs is built of compact matrix-supported diamicton with characteristic structures of basal till (Fig. 2a). OSL dates of ~200 ka have been obtained from beneath the till. The ancient till is overlain by inorganic silt rhythmite grading into varved clay. This glaciolacustrine formation of the high southern bank of the Ob River is older than 140 ka, which follows from the luminescence dates obtained in the sand-filled channel incised into the rhythmite (Fig. 2b).

The rhythmite is in many places truncated by cross-bedded fluvial sand with fine gravels and peat interlayers. The sand, most probably, relates to the last interglacial, which is indicated by southern taiga pollen spectra (Zubakov &. Levkovskaya, 1969) and non-finite radiocarbon dates (Table 2, samples 117, 118, 129, 131, 132, 485 and LG-13). The top of the interglacial formation is blown out, cryoturbated, and everywhere covered by loess-like (flowable if wet) silt that mantles the bluff highs and descends over the terraces and slopes of small tributary valleys (Fig. 2).

The cover silt is the most natural source of bones, because their ages are mostly less than 40 ka, whereas the fluvial sand is obviously much older (Table 2). One bone fragment has yielded an age of more than 43 900 years,

Table 1. Pleistocene mammal remains found on the Lower Ob (identified by I.Ye. Kuzmina)

Sample no.	Locality	Identification
112	Pichuguy-Yaha, downstream part of sect. 25	1. Two tubular bones and rib of mammoth 2. Lower part of right humerus of adult horse 3. Lower part of radius bone of muskox 4. Antler of large reindeer and a bone of short-legged reindeer
122	Pichuguy-Yaha, upstream part of sect. 25	1. Upper part of muskox cranium with horn roots 2. Shinbone, humerus, ribs of mammoth 3. Three bones of reindeer - caxal bones and cannon-bone 4. Five muskox bones- vertebrae, fragment of scapula, a cannon-bone and a cannon-bone of another individual
123	Pyak-Yaha, downstream part of sect. 26	1. Rib and fragment of mammoth tusk 2. Caxal bone of horse
126	Pyak-Yaha, downstream part of sect. 26	1. Ribs of adult mammoth 2. Calcaneum of woolly rhinoceros 3. Right humerus of mammoth 4. Right radius bone of horse 5. Caxal bone and fragment of humerus of reindeer
135	Pyak-Yaha, upstream part of sect. 26	1. Fragment of mammoth thigh-bone
176	Shuryshkarsky Sor, sect. 31	1. Fragment of mammoth bone 2. Horse diaphysis
183	Shuryshkarsky Sor, sect. 31	1. Fragment of mammoth tooth

Table 2. Radiocarbon dates from Quaternary sediments of the Lower Ob area by geochronological laboratories of St. Petersburg University (index LU) and the A.P. Karpinsky Geological Research Institute (index LG)

Sample no.	Locality	Stratigraphic position	Material	^{14}C age, yr BP (lab. no.)
117	Pichuguy-Yaha, sect. 25, downstream part	Rippled sand between loess-like silt and glaciolacustrine rhythmite, 35.5 m a.s.l.	Plant detritus	>52400 (LU-4769)
118	Same	Same, 37.5 m a.s.l.	Same	>45900 (LU-4767)
112	Same	Picked up under the bluff	Mammoth bone	25050±220 (LU-4787)
122a	Pichuguy-Yaha, sect. 25, upstream part	Same	Same	15480±70 (LU-4783)
122b	Same	Same	Muskox cranium with horns	16380±50 (LU-4992)
132	Pyak-Yaha, sect. 26	Cross-bedded sand above glaciolacustrine rhythmite, 28.5 m a.s.l.	Plant detritus	>46200 (LU-4764)
129	Same	Sand with current ripples, 33 m a.s.l.	Wood	>50400 (LU-4759)
131	Same	Peat between sand and loess-like silt, 33 m a.s.l.	Peat	>44700 (LU-4765)
135	Pyak-Yaha, sect. 26, upstream part	Picked up under the bluff	Mammoth bone	>43900 (LU-4784)
126	Pyak-Yaha, sect. 26, downstream	Same	Same	35200±210 (LU-4785)
123	Same	Same	Bones of mammoth and horse	9530 ±120 (LU-4788)

170	Shuryshkarsky Sor, sect. 31	Autochthonous peat, 10 m a.s.l.	Twigs	>39700 (LU-4761)
185	same	Same	Birch in bark	>47200 (LU-4768)
178	Same	Upper part of loess-like silt 20 m a.s.l.	Twigs	12800±150 (LU-4763)
183	Same	Picked up under the bluff	Mammoth tooth	>30700 (LU-4757)
176	Same	Same	Bones of mammoth	17340±230 (LU-4786)
485	Pit at Aksarka, sect. 18	Ice wedge cast in sand, 17 m a.s.l.	Peat	>45000 (LU-4553)
502	same	Base of loess-like silt, 40 m a.s.l.	Peaty silt with twigs	27840±310 (LU-4555)
LG-13	Pyak-Yaha, sect. 26	Bed of silty clay in gravelly sand	Wood	>57000 (LG-13)

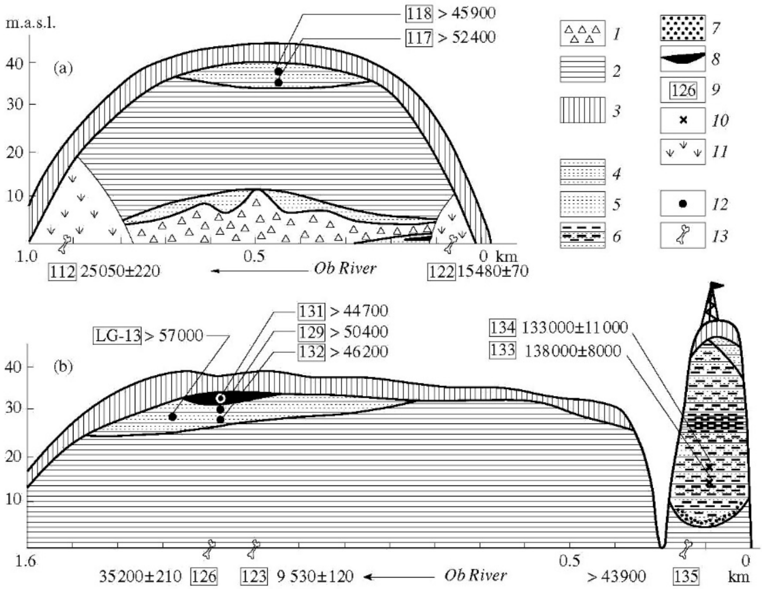

Fig. 2. Location of dated samples in two sections of the southern bank of the Nadym Ob. Zero distance is in 3 km downstream of the Pichuguy-Yaha creek for Fig. 2a (sect. 25) and 0.3 km downstream of the Pyak-Yaha creek for Fig. 2b (sect. 26). (1) Basal till (diamicton); (2) glaciolacustrine rhythmite (clay and silt); (3) subaerial mantle (loess-like silt); (3a) Fluvial sediments; (4) cross-bedded sand with gravels; (5) laminated sand; (6) sand with silt interlayers; (7) gravel; (8) peat; (9) sample no.; (10) OSL date; (11) scree. Radiocarbon dates from (12) plant remains and (13) mammal bones.

and therefore, in principle, might have originated from the fluvial sand (Fig. 2b). However, the majority of the bones have no traces of aqueous transportation. On the contrary, cavities of the well-preserved muskox skull contained remains of loess-like silt. Fallen blocks of similar silt are often observable where the bluff is lower.

The described chronological relations persist in other Ob sections. Thus, fluvial sand has yielded several radiocarbon values greater than 45 ka in the Aksarka sand pit, whereas a lens of mire silt, found at the base of a cover silt 6-7 m thick, is ~ 28 ka old (Table 2, samples 485, 502). A very young age of the bulk of the subaerial cover is also supported by the date of 12 800 years at 1 m below the top of the loessic mantle of Shuryshkary Sor (Table 2, sample 178) and by a still younger (~ 9.5 ka) age of bone sample 123.

Occurrence of much older mammoth remains cannot be ruled out on the River Ob. However, for the present discussion, the radiocarbon values within the 25–15.5 ka interval (Table 2, samples 112, 122a, 122b, 176) are of prime interest. They, for the first time, demonstrate that, during the so-called Last Glacial Maximum, the arctic Transuralia was not a glacial desert but a pasture for a characteristic Upper Paleolithic community of large mammals, including not only mammoths, woolly rhinos, muskoxen and reindeer, but also horses. This fauna, together with the loessic appearance of the synchronous sediments, is evidence of dry, snow-poor and windy – but habitable – frozen steppe in the second half of the Late Pleistocene. Results obtained for the Lower Ob along with the data by earlier researchers (Makeyev et al., 1979; Sulerzhitsky & Romanenko, 1997; Vasilchuk et al., 1997) give reason to believe that such landscapes dominated over the entire Siberian Arctic.

Acknowledgments

This study was carried out under the Russian-Norwegian PECHORA project `Paleo Environment and Climate History of the Russian Arctic` sponsored by Norwegian Research Council within the QUEEN program (Quaternary Environment of the Eurasian North), European Science Foundation. We are very grateful to I. Kuzmina (Zoological Institute, Russian Academy of Sciences) for identification of the bone remains and to A. Matyushkov (A.P. Karpinsky Geological Research Institute), who helped in field work. Our thanks are also due to S. Chernov, T. Tertychnaya and F. Maksimov (St. Petersburg State University) for radiocarbon dates and to A. Murray (University of Aarhus), who provided luminescence ages.

12. The Age of Interglacial Peat on the Lower Ob

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DOKLADY EARTH SCIENCES 2005, VOL. 401(2), P. 298–302.
Translated from Doklady Akademii Nauk, 2005, vol. 400(1), P. 95–99.

Results of dating and pollen analysis of a thick fossil peat on the Lower Ob are presented in this paper in order to improve the Late Pleistocene stratigraphy of West Siberia. Field investigation and sampling was carried out in 2001 by Astakhov, Matyushkov and Nazarov. Radiometric dating was performed in Laboratory of Geochronology, St. Petersburg University by Arslanov, Maksimov and Kuznetsov. Pollen analysis is by Razina.

Peat deposits, ubiquitous in the Holocene of West Siberia, are very rare in the Pleistocene sequences. This is one of the reasons for many uncertainties in understanding paleoenvironments and climatic rhythmicity of the last interglacial-glacial cycle of the region. After a long discussion fossil peats in the modern subzones of southern taiga (Gornaya Subbota on the Irtysh, 60º N) and middle taiga (Karymkary on the Ob, 62º N) have been acknowledged as Eemian, i.e. synchronous to marine isotope substage 5e (Arkhipov & Volkova, 1994). However, the most voluminous peat layer in the Ob valley, found in the southeastern corner of the Shuryshkarsky Sor (floodplain lake) in sparse northern taiga, 66º N, was by Novosibirsk geologists attributed to the time span of 44–50 ka BP which is a first part of their `Karginsky interglacial`. This conclusion stems from Arkhipov`s et al. (1977) idea that the Eemian on the Lower Ob occurs as a marine formation below river level. However, the modern dating by optically stimulated luminescence (130-140 ka BP) indicates to the alluvium of the 40-m terrace of the Nadym Ob with peat and forest spectra as a representative of the Eemian interval (paper 11, Ch. III)

Therefore, deeper marine strata there, as well as upstream in Shuryshkary, cannot be younger than the Middle Pleistocene (paper 3, Ch. I)

In the early 1970s Arkhipov et al. (1977) described a 0.7 m thick peat layer overlain by coarse sand at the low water level, 10 m a.s.l., and another peat 0.25 m thick at 15.9 m a.s.l., all parts of their `Shuryshkary strata`. According to them the section was found at 2.5 km from the mouth (?) of the Shuryshkary Sor. In 2001 we could only find the lower but thicker (up to 1.1 m) peat traceable between 1 and 1.5 km north of Shuryshkary settlement, section Shur-1 (for location see Fig. 3 in paper 20, Ch. IV). Where the peat layer plunges below the lake level, in two clearings 70 m apart, we observed the following sediments downwards in the succession:

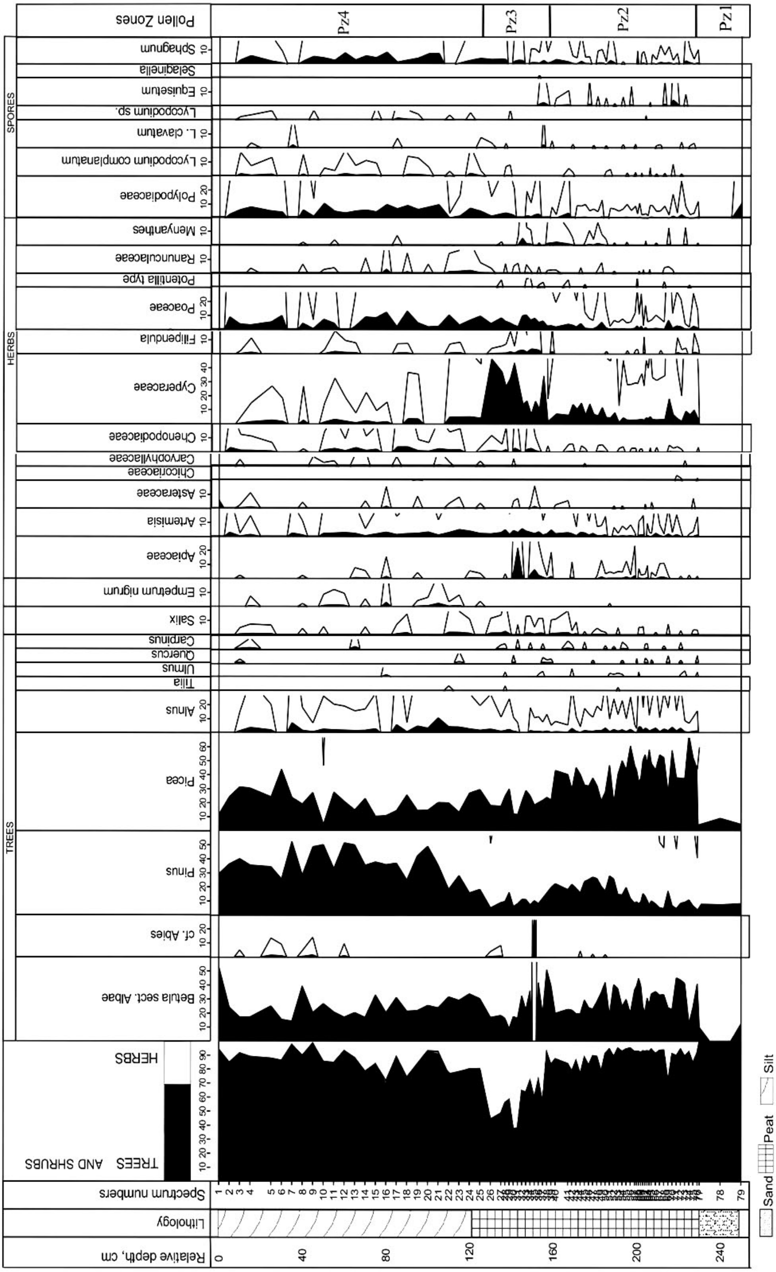
1. 0–5 m Pale-brown loess-like silt, cryoturbated at the sole.
2. 5–9.2 m Pale-grey medium sand, poorly sorted, crudely stratified, with silt laminae. Lenses of fine gravel appear closer to the base. Basal beds, dipping southwards at 15º, are disturbed by fossil frost cracks. The strata get gentler and more planar upwards. Both contacts are sharp, uneven, with angular unconformities.
3. 9.2–10 m Pale or whitish fine sand, well washed, with north directed current ripples draped by plant detritus.
4. 10–11.4 m Tan clayey silt with occasional pin-stripe laminae of white fine sand and grey clay. An interbed of yellow-grey fine sand with ideally symmetric, microlaminated wave ripples 1 cm high and 5 cm apart, occurs at 15 cm from the top. Below there are seams of washed-on plant detritus.
5. 11.4–12.5 m Black and brown compact peat, at the top consisting of sedges and numerous willow twigs, below rich in flat-pressed stems of spruce and birch, with occasional larch cones. The upper decimeter is mostly peaty clay. Fine seams of sand occur at 20-30 cm below the top. Further downwards the peat is clean, slaty, without visible mineral inclusions.
6. 12.5–13 m Pale-grey, non-stratified, water-soaked silty sand, with a net of ferruginated fine cracks at the top.
7. 13–13.5 m Bluish sticky varved clay with marlekors, outcropping everywhere below the lake level. Under the clay Arkhipov locally noticed a compact diamicton of the so-called `Hashgort Till`. This singular till of the area occurs in many other places along the Ob river level.

The peat deposit looks like a result of overgrowing of a shallow lake. The overlying beds 3 and 4 without traces of strong wave or tidal activity should be treated as sediments of a shallow but not bottom-frozen lake. Bed 2 is probably deposited by unstable low energy streams. Bed 1 is aeolian-permafrost feature. Such a facies composition is quite normal for the upper part of the Ob Second Terrace where no real alluvial strata have been described. The unique peculiarities of the peat layer are: i) its lateral consistency for at least 500 m; ii) the absence of permafrost disturbances which normally deform other peat deposits. The lack of cryogenic features is probably explained by fast submergence of the bog in the end of the peat accumulation. Cryogenesis started only after draining off the lake (base of bed 2). No admixture of mineral detritus is perceived by naked eye below l1.7 m level, but the ash content 13 to 30% indicates some influx of dust into the bog. The peat layer is sampled in a continuous vertical trench cut by an ax.

Pollen analysis was performed for beds 4–6. The percentage diagram (Fig.) is composed on 79 samples collected each 2 cm in the peat (bed 5) and at 5 cm intervals in the silt (bed 4). The samples were treated by acetolysis using HF. Pollen zone (PZ) 1 is represented by the spectra of the pollen poor sand (bed 6). PZ 2 marks the ecologically optimal conditions of the peat accumulation at the lower half of bed 5. Arboreal pollen dominates (up to 90%), spruce and birch shares being equal (40-50% each) with pine subdominant (20-30%). Admixture of deciduous (broad-leaved) pollen without traces of corrosion and redeposition – elm, oak, hornbeam – occurs regularly (up to 1-2%). The presence of *Osmunda* is characteristic. Arboreal pollen decreases notably in PZ 3 (the upper part of the peat deposit with mineral laminae) versus the increasing content of sedges, palustrine and aquatic plants.

The deteriorating ecology was initially attributed to a cooling (Arkhipov et al., 1977). However, it is more likely to be a result of higher ground water and periodic submergence of the floodplain bog. The minimal floristic change supports this conclusion. Besides, arboreal pollen, accompanied by typical for boreal forest spores of mosses, ferns and club mosses, dominates again in PZ 4.

The curves in Fig. resemble those in the diagram by Votakh (Arkhipov et al., 1977). However, she noticed no broad-leaved arboreal pollen**,** possibly due to wider gaps between samples. One would hardly insist on broad-leaved trees growing around that bog, but their area might have been not too far away, which is a sign that the ancient forest belonged to the southern taiga subzone. It is important that the part of spruce among the arboreal pollen is only 5% in sub-recent spectra from the Ob floodplain (Volkova et al., 1988), whereas spruce evidently dominates in the fossil spectra (20 to 50%).



Percentage pollen and spore diagram of beds 4–6 from the Shurushkarsky Sor section (compiled by V. Razina).

Arkhipov et al. (1977, 1994) concluded that climate during peat accumulation was only slightly warmer than the present one. Our data would rather suggest a shift of ancient vegetation zones 300-400 km northwards. This suggestion agrees with findings of pollen of middle and southern taiga in synchronous sediments on the Arctic Circle (Zubakov, Levkovskaya, 1969). We subscribe to the previous conclusion that the difference between the modern and interglacial climates is more pronounced in northern parts of West Siberia (Volkova et al., 1988).

The first attempt of geochronometric correlation was undertaken by Arkhipov et al. (1977) who reported radiocarbon dates more than 40 ka (SOAN-646 and 647) from the main peat deposit and from the overlying peat interlayer as well. We have also obtained non-finite values from the main peat body: more than 39.7 ka on willow twigs (LU-4761) and below it – more than 47.2 ka on birch in bark (LU-4768). Finite ages have been provided by optically stimulated luminescence (Murray & Wintle, 2000), although the sampled sand interlayers of beds 4 and 6, being too thin, are not quite suitable for this method. Their quartz grains are subject to higher radiation levels from the enveloping clays.

Our samples from from bed 6 in Nordic Laboratory for Luminescence Dating, Denmark, have been dated to 82 ± 4 (Risø lab № 012587) and 100 ± 6 ka BP (№ 012588). Due to the above complications these ages should be treated as minimum estimates. This suggestion is supported by the abnormally low value 57 ± 4 ka (№ 012589) from the thinnest sand interlayer of bed 4. However, the age 100 ± 5 ka BP (№ 012590) from bed 3 might be closer to reality because of the better sampling situation.

Lately the uranium-thorium isochronous method (Hejnis, 1992; Geyh, 2001) has been widely used for correlation of thick buried peats. It was successfully applied for dating the peat of the Mikulino stratotype (Kuznetsov et al., 2003) and interglacial peat at Bedoba village north of the Angara river (Arslanov et al., 2004). With that experience the Shur-1 section looked promising because the peat was isolated between clay beds. To estimate the applicability of the method for this peat deposit we first studied changes of uranium content through the vertical profile. The U content in leachates from the upper and lower layers (some $0.2\text{-}0.25\text{x}10^{-6}$ gr per gr of peat) proved to be very close to that from the inner layers of the peat ($0.2\text{-}0.3\text{x}10^{-6}$gr), which testifies to a lack of noticeable import of uranium by ground water. Evidently the inner part of the peat bed, protected by impermeable clays, has not been contaminated by migrating uranium.

Table 1. Results of radiochemical and alpha spectrometric analyses of U and Th isotopes in peat samples from section Shur-1 (leaching)

Depth, cm	Ash content,%	^{238}U dpm/g	^{234}U dpm/g	230Th dpm/g	^{232}Th dpm/g	$^{230}Th/^{234}U$	$^{234}U/^{238}U$
35-38.5	29.99	0.222± 0.010	0.284± 0.011	0.232± 0.005	0.292± 0.005	0.817± 0,037	1.282± 0.069
38.5-42	18.31	0.165± 0.010	0.220± 0.012	0.190± 0.007	0.240± 0.008	0.861± 0,056	1.333± 0.096
42-46	15.17	0.147± 0.006	0.192± 0.007	0.165± 0.005	0.193± 0.005	0.860± 0,042	1.304± 0.067
46-50	26.15	0.229± 0.009	0.289± 0.011	0.255± 0.007	0.356± 0.009	0.881± 0,041	1.263± 0.062
50-52.5	13.26	0.143± 0.004	0.175± 0.005	0.138± 0.004	0.164± 0.005	0791± 0,033	1.219± 0.041
52.5-55.5	19.84	0.246± 0.004	0.300± 0.005	0.193± 0.006	0.220± 0.006	0.644± 0,023	1.218± 0021

Table 2. Results of radiochemical and alpha spectrometric analyses of U and Th isotopes in peat samples from section Shur-1 (total dissolution)

Depth, cm	Ash content,%	^{238}U dpm/g	^{234}U dpm/g	^{230}Th dpm/g	^{232}Th dpm/g	$^{230}Th/^{234}U$	$^{234}U/^{238}U$
35-38.5	29.99	0.481± 0.013	0.519± 0.014	0.488± 0.011	0.584± 0.012	0.942± 0.033	1.109± 0.034
38.5-42	18.31	0.336± 0.011	0.360± 0.012	0.367± 0.009	0.436± 0.010	1.021± 0.042	1.070± 0.043
42-46	15.17	0.274± 0.006	0.310± 0.007	0.294± 0.004	0.366± 0.005	0.946± 0.026	1.133± 0.028
46-50	26.15	0.481± 0.009	0.516± 0.010	0.492± 0.015	0.677± 0.017	0.953± 0.034	1.073± 0.029
50-52.5	13.26	0.237± 0.007	0.262± 0.007	0.265± 0.006	0.340± 0.007	1.011± 0.035	1.105± 0.044
52.5-55.5	19.84	0.426± 0.013	0.471± 0.014	0.451± 0.010	0.549± 0.012	0.957± 0.035	1.106± 0.037

For dating purposes concentration of isotopes of U, Th and their activity ratio were determined in a successive series of samples through the interval of 35 – 55.5 cm from the top of the peat. The peat samples were dried out and burnt in a muffle oven at 800ºC. Isotopes of U and Th were extracted by leaching and by total dissolution. Leaching was carried out by treating samples with 7M HNO_3 solution during 6 hours of permanent stirring. Total dissolution was achieved by treating samples with concentrated solutions of HNO_3, HF, HCl and $HClO_4$ acids. We used the known radiochemical methods of co-precipitation of U and Th isotopes on iron hydroxide and their separation by anion-exchange chromatography. Then U and Th isotopes were applied to platinum discs by electrolytic precipitation. Their alpha activity was measured using surface-barrier silicon detector and pulse analyzer AI-1024. The results are presented in Tables 1 and 2.

For determination of the U/Th age we used 4 samples with the minimum ash content (13–26%) from the inner layers of the peat (35–50 cm). Because the values of the 4 samples are too close to each other on the graph of $^{230}Th/^{232}Th$-$^{234}U/^{232}Th$, making composition of an isochronous line difficult, we applied an alternative method of determination of $^{230}Th/^{232}Th$ ratio (thorium index) in detrital fraction (Geyh, 2001). The dependence of relative scattering of age on thorium index was plotted for both methods of isotope extraction. Thorium index is 0.19 for total dissolution method and 0.10 for leaching. Inserting the corrected values of $^{230}Th/^{234}U$ into the formula for age calculation give 133 ± 14 ka BP for method of leaching and 141.1 ± 11.7 ka for total dissolution.

The obtained values, within the error bar of measurements, correspond to the U/Th ages of the Mikulino peat of Central Russia (Kuznetsov et al., 2003), interglacial sediments of Central Siberia in the Bedoba section (Arslanov et al., 2004) and marine isotope substage 5e. The latter commenced between 128 and 140 ka ago by estimates of different authors (Lowe & Walker, 1997). Therefore, the Shur-1 peat is reliably attributed to the warmest Eemian (Mikulino) interglacial, although its pollen characteristic is markedly different from the classic diagrams of the last European interglacial. Correspondingly, the Shuryshkary warming occurred not 45-50 ka BP (Arkhipov et al., 1977; Arkhipov &Volkova, 1994) but in the very beginning of the Late Pleistocene. Since the studied peat layer rests atop of the uppermost `Hashgort Till`, the latter cannot be Late Pleistocene. The west bank of the Lower Ob, as well as the east bank (paper 11, Ch. III) was evidently not covered by Late Pleistocene glacial ice.

This study was carried out for the PECHORA (Palaeo Environment and Climate History of the Russian Arctic) project funded by the Norwegian Research Council. It was also a part of the international program QUEEN (Quaternary Environments of the Eurasian North) of European Science Foundation. Our thanks are due to A. Matyushkov, National Geological Institute for valuable field assistance, to A. Murray, University of Aarhus, for OSL-dating and to S.B. Chernov for alpha-spectromeric measurements of peat samples.

13. THE AGE OF THE KARGINSKY INTERGLACIAL STRATA ON THE LOWER YENISEI

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DOKLADY EARTH SCIENCES, 2005, VOL. 403(5), P. 673–676.
Translated from Doklady Akademii Nauk, vol. 403, No. 1, 2005, P. 63–66.

New geochronometric data on a key section of the Karginsky terrestrial sediments on the Malaya Heta River, a western tributary of the River Yenissei, demonstrate that the corresponding warm Pleistocene interval is significantly older than has previously been assumed.

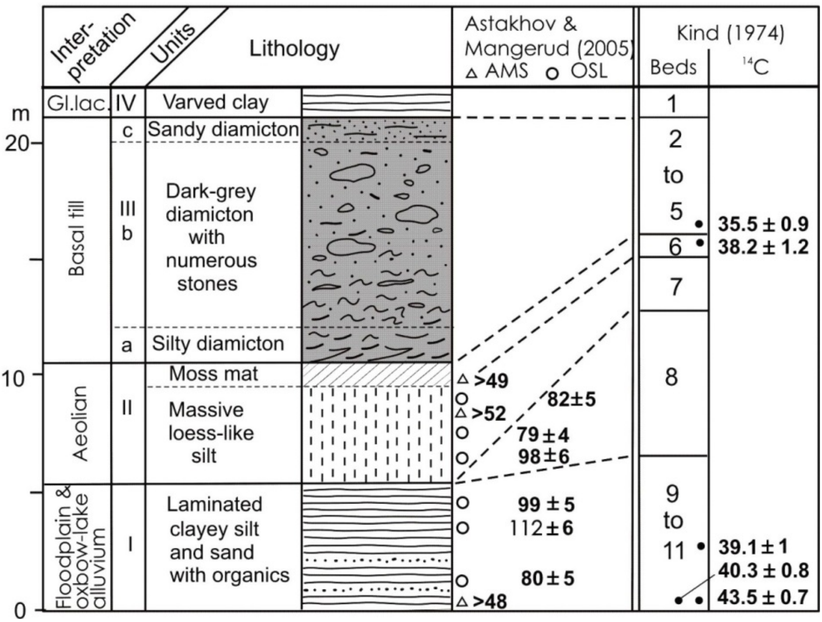
The Karginsky warm interval is considered to be a representative of marine isotope stage 3 (MIS 3) in Siberia. Accordingly, the subsequent glaciation has been related to the last glacial maximum, MIS 2 (the Late Weichselian by the European scale). This geochronological convention is based on two key sections in the Ust-Yenissei Port area of Siberian Quaternary stratotypes (Sachs, 1953): the marine formation at Karginsky Cape and alluvial sediments of the \`Karginsky terrace\` on the Malaya Heta River. Finite radiocarbon dates obtained for both sections (Troitsky, 1967; Kind, 1974) led N. Kind to propose the \`Karginsky interglacial\` (nowadays often called the \`Great Karginsky Interstadial\`), which occurred in the interval 50 to 22 ka B.P. and featured a climate that was not worse than the present one. An MIS 3 age for the marine formation was later invalidated by an ESR date of 121.9 ka, and the Karginsky Cape was therefore rejected as a stratotype (Arkhipov, 1990).

A young (MIS 2) age for the upper till deposited by the Barents-Kara Ice Sheet in Siberia and in the Pechora Basin is presently refuted by a number of \`old\` radiocarbon dates from the overlying sediments (Astakhov, 1998; Mangerud et al., 2002). Hundreds of radiocarbon and luminescence dates obtained within the European project Quaternary Environments of the Eurasian North (QUEEN) since 1995 indicate that the last glaciation in these lowlands occurred before 50 ka and that the Late

Weichselian ice sheet did not reach the shores of the Kara Sea (Mangerud et al., 2002; Svendsen et al., 2004).

The Malaya Heta section is therefore crucial both for the age and correlation of the Karginsky thermomer and as the most important remaining support for the Late Weichselian age of the last Kara Sea till in West Siberia. However, the section has not been studied since 1966. These are the reasons for our visit there in August 2003 with a view to applying modern geochronometric methods.

The section is located at 69°32' N and 84°40' E, 17 km upstream of the Malaya Heta River mouth in its eastern river bank, cutting into a shrub tundra surface 20-30 m a.s.l. Radiocarbon dates by Kind (1974) within the range of 45 to 35 ka BP were recovered from the floodplain alluvium with lenses of oxbow-lake clay overlain by glacial deposits. This sequence was described in the upstream part of the bluff in 1966 at a very low river level.



Geochronological results of the PECHORA II project versus Kind`s results for the Malaya Heta section. Zero line is the river level in August 2003. Stippled lines show the relation of sedimentary units identified by the authors of the present communication to beds in (Kind, 1974).

In 2003, this part of the bluff was mostly covered by scree, and we could observe only the upper part of Kind's section (beds 1-7). Nevertheless, the main elements of Kind's log (figure) were readily recognizable in our three clearings along a bluff stretch 200 m long. Having only two days at our disposal, we sampled mostly the better cleaned central section, approximately 100 m downstream of Kind's section. The glacial part of the sequence was thicker and the subtill sequence was thinner in our central section (figure). We have distinguished the following lithostratigraphic units (from the river level to the top):

Unit I is a dark gray and brownish, finely laminated, compacted clayey silt with thin laminae of sand in the lower part and of plant remains higher up. Numerous freshwater mollusks (gastropods and bivalves) are scattered throughout the unit, and peaty seams sometimes contain branches and thin logs. In our central clearing, the top of the unit was 5.5 m above the river (figure). Unit I roughly corresponds to beds 9–11 of Kind, who considered it to be floodplain alluvium of a large river, the underlying channel sand having been described only from boreholes (Kind, 1974). Kind's bed 8, oxbow-lake clay with gyttja, probably pinches out downstream, as it was not observable in our main section.

Unit II is more variable, both laterally and vertically. It is mostly brown-pale, massive, soft, loess-like coarse silt of an earthy appearance with seams and patches of mossy peat, 5-6 m thick, in the central section (figure). Stratified fine sand and coarse silt appears in the upper part of the unit exposed in the upstream section, where the top of the unit is positioned 4 m higher than in the central section. The latter facies displays lens-like and trough-like wavy lamination, in places with tiny current ripples and small ice wedge casts, and probably correlates with bed 7 in Kind (1974). Peaty inclusions increase in abundance upwards. At the top of the unit, they make a marker bed of silty moss mat 0.5-0.9 m thick, corresponding to Kind's bed 6. She described this part of the sequence as floodplain alluvium as well. We interpret the unit as mainly aeolian sediment, partly modified by rivulets and mire accumulation in a drier and colder environment.

Unit III is 9 m thick and consists of the following three diamictic beds:

The lower bed, 1.5 m thick, is a yellow-brown, massive, silty diamicton with rare pebbles, peat balls, and attenuated tails of fine sand. The granular texture is due to small clasts of clay and silt. Shear planes dip toward 335° NW. The base is sharp, but contains drag structures from the underlying peat.

The middle bed is a dark gray homogeneous diamicton 5.5 m thick. It is silty at the base and more sandy upward the section. It differs from the lower and upper beds by a much higher content of pebbles and larger, often striated stones. Laminae at the hinge of a large fold dip toward 345° NW and shear planes toward 295° NW and 25° NE.

The upper bed is a brown-yellow sandy diamicton, 1 m thick, with many tiny fragments of silt and clay and discontinuous bands of light fine sand. Shear planes dip 345° NW.

Unit IV is a gray varved clay (1 m thick), probably deposited in an ice-dammed lake in the course of deglaciation. The top of the bluff is covered by a solifluction mantle or peat.

The dark gray diamicton IIIb corresponds to Kind's basal till up to 3 m thick. Our diamictic beds IIIa and IIIc are described in Kind (1974) as alluvial silts (beds 5 and 2 + 3, respectively). This was rather common in the 1940s – 1970s, when tills were identified as such only if crystalline cobbles, indicating the Putorana provenance, were present. However, the fine-grained composition of diamictons IIIa and IIIc is quite typical for lowland tills of West Siberia, where no source of stones is available. We interpret the entire unit III as basal tills deposited by an ice sheet flowing toward south-southeast, i.e., from the Kara Sea up along the Yenisei, as indicated by directional structures. This dominant southward flow was earlier established from the pattern of push moraines (Arkhipov et al., 1980) and till fabrics (Sukhorukova & Gaigalas, 1986).

Kind concluded that beds 8-11 (unit I) were deposited in a climate warmer than the present. This conclusion is based on diverse paleontological evidence, including pollen of dense boreal forest dominated by *Picea* and *Pinus sibirica* with some *Abies,* a rich diatom flora (170 species), freshwater mollusks (14 species), and plant macrofossils. N. A. Halfina reported several Baikal diatom species, indicating that unit I was probably deposited on the Yenissei floodplain, partly in oxbow lakes (Kind, 1974). Pollen spectra from beds 6 and 7 (our unit II) are dominated by birch and arctic herbs, suggesting a colder climate, which is consistent with our interpretation of unit II as a periglacial subaerial deposit.

Based on finite radiocarbon dates (figure), Kind attributed all subtill sediments (our units I and II) to the `Malaya Heta warming` with a postulated age of 43–33 ka. She correlated it with the Hengelo interstadial in Europe (Kind, 1974) and the maximum of the last warm-water transgression on Taimyr (Kind & Leonov, 1982). The dates were obtained by L. D. Sulerzhitsky on peat and plant detritus using conventional ^{14}C dating. However, Sulerzhitsky stressed that perennially frozen sediments often yield too young ages due to contamination by young carbon,

Geochronometric data from the Malaya Heta section obtained by the PECHORA II project in 2003. Radiocarbon AMS dates by Poznan Lab., Poland (T. Goslar), optical luminescence dates by Nordic Lab., Risø, University of Aarhus, Denmark (A. Murray).

Sample no.	Lab. no.	Age, ka B.P.	Material	Height above river, m	Sample no.	Lab. no.	Age, ka BP	Material	Height above river, m	Number of aliquots	Water content, wt %
217	Poz-5079	>49.0	Moss	9.7	212	042524	82 ± 5	Coarse aeolian silt	9.2	23	43
210	Poz-5078	>52.0	25 cm branch of spruce (?)	9.1	211	042523	79 ± 4	Same	8.5	27	35
214	Poz-5018	>48.0	Small birch log	0.5	209	042522	98 ± 6		6.0	24	39
					206	042521	99 ± 5	Sand laminae in clayey silt	4.8	27	39
					205	042520	112 ± 6	Intercalation of sand and clayey silt	3.2-3.6	27	37
					202	042519	80 ± 5	Clayey silt	0.6	26	39

Note: Sample 217 is tissue of *Drepanocladus cf. revolvens* and *Scorpidium turgescens*.

especially for ages beyond 30 ka, and he usually treated them as minimum ages (Kind & Leonov, 1982). This conclusion is supported by statistics on radiocarbon dates from the Kara Sea catchment area (paper 10, Ch. III).

We used the AMS method for radiocarbon dating. The advantage here is the possibility of dating very small samples (a few milligrams) and thus specially selecting the least contaminated constituents. The three samples all yielded non-finite ages (based on the two standard deviation criterion) of more than 48-52 ka (table), and thus indicate that the sediments are older than suggested by Kind (figure).

Our finite ages have been obtained by a modern version of the optically stimulated luminescence (OSL) method (Murray & Wintle, 2000). The dates (table) again clearly demonstrate that unit I is much older than postulated by Kind. We conclude that unit I and thus the Karginsky warm period belong to the last interglacial, i.e., to the Mikulino/Eemian or MIS 5. This age is also consistent with the interglacial-type organic remains described from unit I.

However, our OSL values are somewhat lower than the generally accepted age of MIS 5e (130-117 ka) and U/Th ages of the last Siberian interglacial (Arslanov et al., 2004; Astakhov et al., 2005). The reliably Eemian sediments on the Sula River, a western tributary to the Pechora, also yielded OSL dates about 10% younger (Mangerud et al., 1999). The reason for this is not yet fully understood. For age calculations, the present-day porosity was used assuming that the Malaya Heta sediments have been water-saturated since the time of deposition. However, unit I is now compacted, i.e., water content might have been higher in the past. The underestimation of mean water content for the time elapsed since the deposition of unit I may lead to an overestimation of the annual dose rate, which to some extent can explain the excessively young dates in this case.

The OSL dating of subaerial unit II yielded even lower ages, and we therefore suggest that the unit reflects periglacial conditions in the beginning of the last glacial cycle (MIS 5d or MIS 5b). This means that the overlying till can be much older than LGM, for example, of MIS 4 or even MIS 5b age, as established for the Taimyr and Pechora tills (Svendsen et al., 2004; Mangerud et al., 1999).

We fully agree that the Malaya Heta warming was contemporaneous with the last interglacial transgression of the Yenisei and Taimyr, which is identified as the Karginsky transgression in (Kind, 1974; Kind & Leonov, 1982). However, the OSL datings (and the ESR date from Karginsky Cape, Arkhipov, 1990) indicate that this transgression occurred during MIS 5 rather than MIS 3 (based on the finite radiocarbon dates). Therefore, the Siberian chronostratigraphic interval corresponding to MIS

3 cannot be called the Karginsky interval, because its sedimentological representatives have been found neither at Karginsky Cape nor in the `Karginsky terrace` of Sachs (1953). Moreover, the low altitude of the Malaya Heta alluvium makes unlikely a distant southward propagation of the last interglacial sea along the Yenissei.

Acknowledgments

The above data have been obtained in the Russian- Norwegian project Paleoenvironment and Climate History of the Russian Arctic (PECHORA II) funded by the Norwegian Research Council. The authors are also grateful to A. Murray (Nordic Laboratory for Luminescence Dating, Denmark) and T. Goslar (Poznan Radiocarbon Laboratory, Poland) for dating the samples, A. Matyushkov (VSEGEI) for assistance in studying and sampling the section, E. Schoelberg and H. Birks (University of Bergen) for identification of wood and mosses, respectively, and J.-I. Svendsen (University of Bergen) for critically reading the manuscript.

14. EVIDENCE OF LATE PLEISTOCENE ICE-DAMMED LAKES IN WEST SIBERIA

VALERY I. ASTAKHOV

BOREAS, 2006, VOL. 35, P. 607–621

Introduction

Understanding the Quaternary history of northern Eurasia depends very much on the knowledge of the mode of glaciation and proglacial drainage in West Siberia, which, due to its size (about 2 million sq. kilometers) and peculiar natural setting, is the key region and the best test ground for any paleogeographic concept. First, the zonality of recent geological processes and biotic changes clearly manifest themselves in this sedimentary basin filled with soft, lithologically monotonous, Meso-Cenozoic formations, which include the Quaternary up to 300-400 m thick. Second, the very low, north-sloping West Siberian Plain, drained by two great northbound Eurasian rivers, the Yenissei and the Ob-cum-Irtysh (Fig. 1), is extremely sensitive to any hydrographic and sea level changes. Third, various glaciations left their indelible imprints in the landscape of the huge lowland.

However, the great extent and poor accessibility of the region have resulted in controversial paleogeographic reconstructions which have been debated in the Russian literature since the 19th century. E.g., the Paleozoic ranges bordering the West Siberian Plain were for a long time considered the principal sources of inland ice (Sachs, 1953; Lazukov, 1970; Zubakov, 1972). Only in the 1970s was the major role of ice accumulation on the Kara Sea shelf, as well as a modest contribution from mountain ice dispersal centres, recognized (papers 1 and 2, Ch. I). Still, estimates of the dimensions of the last glaciation in the literature are vastly different – from small piedmont glaciers (e.g. Arkhipov et al., 1977) to a huge coherent ice sheet (Grosswald, 1980; Grosswald & Goncharov, 1991).

One of the hypotheses, especially popular in the international literature, was used as a base for several geophysical and paleogeographic models (e.g. Grosswald 1980, 1983; Peltier 1994). This so-called maximalist hypothesis, copying the North American paleogeography, suggested a very young (c. 20 ka ago) and enormous (about 2500 km across) ice sheet which grew over the Kara Sea shelf and intruded into the present zone of boreal forests far below the Arctic Circle. The huge ice sheet was associated with a world-greatest ice-dammed lake reaching up to 130 m a.s.l., which presumably flooded the southern half of the West Siberian Plain and spilled into the Caspian and Black seas via the Turgai Pass. This major change of the West Siberian hydrology was suggested as the crucial link in the totally reorganised continental drainage resulting in huge volume of fresh water reaching the Mediterranean (Volkov et al., 1978; Grosswald, 1980, 1983).

Many Russian authors have pointed out flaws in the maximalist reconstructions and argued that discharge of the great Siberian rivers into the Arctic Ocean did not cease during the Late Pleistocene (e.g. Laukhin, 1981; Vasilchuk et al., 1984). This controversy resulted in two incompatible models for the West Siberian Last Glacial Maximum (Velichko, 1993). Field checks in the 1980s found very little in support of the maximalist model. Radiocarbon and sedimentological evidence was either ambiguous (e.g. Kind & Leonov, 1982), or contradicting the maximalist scheme (Vasilchuk et al., 1984; Astakhov & Isayeva, 1988; Astakhov, 1989). Data collected in West Siberia led to rejection of a large Late Weichselian glaciation, as well as proglacial lakes higher than 70 m a.s.l. Sediments of the LGM time indicated mostly frozen loess environments, and only a small ice sheet in the southwestern corner of the Kara Sea close to the Urals was deemed possible (papers 6, Ch. II and 10, Ch. III).

A new history of paleogeographic research started in 1993 when the Russian-Norwegian PECHORA project, the Russian-German `Taimyr` project and other collaborative projects were launched. Most of the international activity was coordinated by the European Science Foundation program QUEEN (Quaternary Environments of the Eurasian North). Comprehensive sedimentological descriptions, together with hundreds of radiocarbon AMS and luminescence dates, have been obtained at many key sections over the Russian Arctic. These more detailed studies, coupled with the Russian geological results of the 1980s, removed the last props from the maximalist hypothesis. All glacial features on the arctic plains proved to belong to ice sheets older than 50 ka (Mangerud et al., 2002). Only small patches of Late Weichselian inland ice on the Siberian mainland were tentatively proposed for the northern tip of the Taimyr Peninsula and the Putorana Plateau (Svendsen et al., 2004). With the removal of a Late Weichselian ice dam across West Siberia a large ice-dammed lake of this age became superfluous.

In arctic European Russia (Fig. 1), where topography is not that flat as in West Siberia, shorelines of proglacial Lake Komi (LK in Fig. 1) were mapped in front of Early Weichselian moraines at 100-110 m a.s.l. (paper 18, Ch. IV). Its beach sediments were dated to 80-100 ka BP by optically stimulated luminescence (OSL) (Mangerud et al., 2001a). Slightly younger (60 to 80 ka) OSL ages have been obtained for proglacial lakes at 80 to 140 m a.s.l., whose sediments are located in the hilly northern Taimyr Peninsula (Alexanderson et al., 2001). For the West Siberian Plain a rough outline of the Early Weichselian drainage system was suggested using mostly topographic evidence (Astakhov, 1992; Mangerud et al., 2001b). Recently a hypothesis of large early Weichselian proglacial lakes (although much smaller than the Grosswald lakes of the Late Weichselian) has been elaborated based on sparse geological evidence and some theoretical considerations (Mangerud et al., 2004).

The validity of all these models depend on geological data which are surprisingly few, considering the size and crucial paleogeographic role of the area. The aim of this paper is to review and discuss available geological evidence in order i) to evaluate current hypotheses of the Late Pleistocene proglacial drainage system in West Siberia, and ii) to reveal weak points in the existing knowledge which should be addressed in future research. Unfortunately, the bulk of the factual material has over decades been collected only as a byproduct of various geological projects which were not specifically aimed at paleolimnology. No special research project has ever been undertaken in West Siberia. This fact and the size of the area make presentation of geological data in a systematic manner impossible for a journal paper. The following selection of relevant data picked up along the major waterways of this super-region may seem arbitrary, but those willing to form their own conclusions can hardly obtain a more formal summary for the time being.

The referred data are organized geographically, along large stretches of the major rivers. These bear informal names, common in Russian literature: the Lower Yenissei for the stretch downstream of river Podkamennaya Tunguska, the Lower Ob for the Ob valley downstream of the Irtysh mouth, the Middle (or transverse) Ob for west-east striking piece of the Ob valley. The cited radiocarbon ages are uncalibrated.

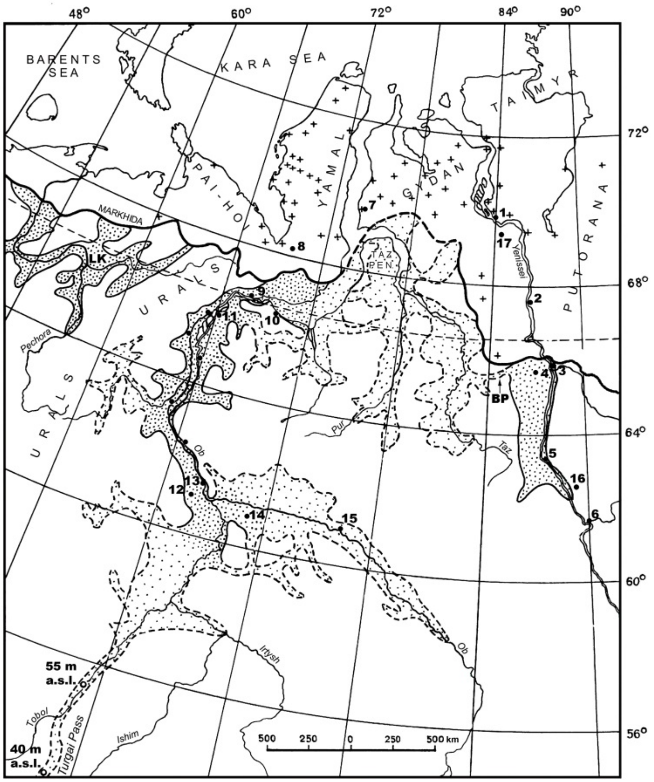


Fig. 1. Proglacial drainage features in West Siberia. Thick line is the limit of the last ice sheet of an early Weichselian age. Densely dotted area is the latest proglacial reservoir inferred from varved sequences within lowland below 60 m a.s.l. on the Ob and below 70 m a.s.l. along the Yenissei (Arkhipov & Lavrushin, 1957; Zarrina et al., 1961). West of the Urals, the proglacial Lake Komi (LK) level is c. 100 m (paper 18, Ch. IV). Sparsely dotted area are hypsometrically possible but sedimentologically not proven proglacial reservoirs. Crosses – sites of massive buried ice, remnants of the last ice sheet (papers 8, Ch. II and 10, Ch. III). Black numbered circles are sites mentioned in the text: 1 – Cape Karginsky (Fig. 2); 2 – Igarka; 3 – Bolshoi Shar; 4 – Farkovo; 5 – Alinskoye; 6 – Podkamennaya Tunguska mouth; 7 – Tadibei-Yaha; 8 – Yerkata (Fig. 3); 9 – Sangompan (Figs 4, 6); 10 – Pyak-Yaha (Fig. 6); 11 – Pitlyar (Fig. 5); 12 – Seul (Zaionts & Ziling, 1972); 13 – Yelizarovo (Lazukov, 1970); 14 – boreholes in thick clayey rhythmites (Zaionts & Ziling, 1972); 15 – Mega (Fig. 8). Not numbered black circles are other sections of glaciolacustrine rhythmites. Open circles indicate bedrock altitudes (coring data by Boboyedova, 1971). BP is a through valley Verkhnyaya Baikha-Pokolky, a possible spillway of the Yenissei proglacial lake.

Lower Yenissei

Glaciated terrains

The classical area of Late Pleistocene glaciation on the arctic stretch of the Yenissei River (Fig. 1), mapped by the Geological Survey in the 1950s (Zarrina et al., 1961), is conspicuous due to its expressive hummock-and-lake landscape and spectacular hill-hole pairs (Zemtsov, 1976). Originally these moraines were taken as evidence of a downslope westward ice flow from the Putorana Plateau. Later it was found that on the eastern bank of the Yenissei accretion ridges indicate eastward upslope movement onto the Paleozoic bedrock of the Central Siberian Upland (Astakhov, 1977; Arkhipov et al., 1980). Taken as a whole, this pattern is evidence of a divergent flow of the N-S directed Yenissei ice stream fed by Taimyr and Kara Sea ice domes. The principal ice flow along the Yenissei valley from NNW or due north is also supported by till fabrics (papers 5 and 6, Ch. II).

Impressed by the young appearance of the hummock-and-lake topography, several authors assigned a Late Weichselian age to this glaciation (Troitsky, 1967; Kind, 1974), although the bulk of stratigraphic and geochronometric evidence indicated an older age (Zubakov, 1972). Closer studies of the morainic landscape revealed that in places it is underlain by thick slabs of fossil glacier ice (Figs. 1, 2) that has been melting in the Holocene giving the very fresh appearance to the glaciokarst topography (Kaplanskaya & Tarnogradsky, 1986; papers 5 and 6, Ch. II).

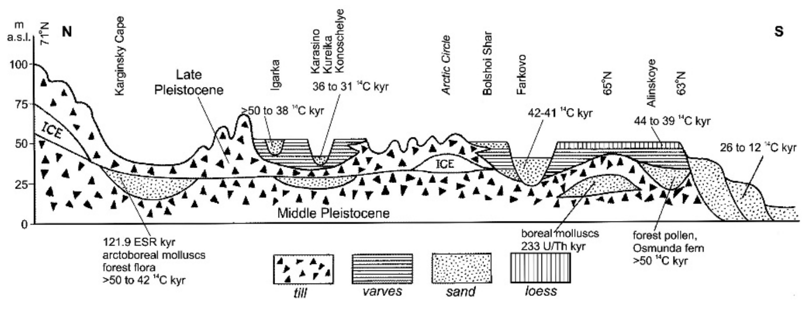


Fig. 2. Idealized profile of the Quaternary stratigraphy along the Yenissei with key sections indicated. Horizontally is not to scale.

A section of interglacial marine sediments at Cape Karginsky (site 1 in Fig. 1) with finite radiocarbon dates was previously used as evidence for a

Late Weichselian age of the overlying Yenissei till (Kind, 1974). Later, this marine formation was ESR-dated to 122 ka (Sukhorukova, 1999). Another support of a Late Weichselian glaciation is alluvium on nearby river Malaya Heta (17 in Fig. 1), where sub-till sediments yielded 5 radiocarbon dates ranging from 44 to 35 ka BP (Kind, 1974). However, this fluvial sequence, presently located in tundra, contains arboreal pollen, abundance of other organics of interglacial type, and therefore should correlate with the Eemian, which has lately been proven by OSL dates exceeding 80 ka (Astakhov & Mangerud, 2005). Therefore, no stratigraphic proof of a Late Weichselian age of the last glaciation is now available in the Lower Yenissei stratotypes.

Finite radiocarbon dates are known from various parts of the Siberian Quaternary and sometimes occur even in Middle Pleistocene sequences (Astakhov et al., 1986). Old radiocarbon dates (~ 30-40 ka) from beneath the surface till of West Siberia, cited by Grosswald (1980), are statistically overwhelmed by much more numerous and consistent successions of dates of the same order, or non-finite, obtained on postglacial organics, including frozen mammoth carcasses (paper 10, Ch. III). Finite dates recovered from beneath the Yenissei tills are probably due to contamination by young carbon which is common in permafrost areas, especially for wood fragments.

Many examples of unlikely correct young radiocarbon ages from northern Siberia are reported by Sulerzhitsky (1998, see also Kind & Leonov, 1982). He explains this by rapid consumption of athmospheric carbon dioxide by autotrophic bacteria dormant in permafrost which become activated after melting out of perennially frozen organics. Whatever the explanation, contamination by young carbon is normal for perennially frozen terrains, whereas cases of contamination by ancient carbon are rare.

As was previously stated, `concerning old radiocarbon ages, it is useful to keep in mind that such samples are insensitive to contamination by old carbon but very sensitive to contamination by young carbon. For example, any sample contaminated with 10% infinitely old carbon appears only about 850 years too old, whereas a contamination of an infinitely old sample by only 1% of modern carbon would yield a date of about 37 ka` (Mangerud et al., 2002: p. 117). In any case, it is clear that no individual date out of the paleogeographic context can be taken as a correlation signal.

The topmost glacial complex is chronologically better constrained by dates from overlying non-glacial sediments. Between the Arctic Circle and the city of Igarka (site 2 in Fig. 1) a 30-40 m thick formation of varved clay, grading upward into a silt and sand rhythmite with moss-mat seams, is inserted into the fresh hummock-and-lake glacial landscape. The lacustrine plateau, up to 60 m a.s.l., in places rises above the surrounding hummock-and-lake landscape underlain by fossil glacial ice, and therefore must have originated as an ice-contact feature (paper 5, Ch. II). The lacustrine rhythmite contains a monotonous fauna of fresh-water ostracods, *Lymnocythere baltica* and *Cytherissa lacustris* (Zaionts & Ziling, 1972).

In the observation pit of the Igarka Permafrost Station silt with cobbles overlies the varved clay. The silt, containing *in situ* frozen logs without any traces of decay, have yielded radiocarbon dates in the range of 39 to more than 50 ka BP. Immediately upstream, at Karasino, Konoshchelye and Kureika (Fig. 2), the upper part of alluvium of a high postglacial terrace (locally called the `Second Terrace`) contains *in situ* fragile twigs dated to 36–31 ka BP (Kind, 1974; Astakhov & Isayeva, 1988). The mentioned data (Fig. 2) plus the long series of radiocarbon dates from overlying subaerial silts elsewhere (Astakhov 1998; Mangerud et al., 2002; Svendsen et al., 2004) are enough to conclude that the expressive morainic ensemble on the Yenissei and the corresponding glaciolacustrine formation were most probably produced by an early Weichselian ice advance, beyond the range of radiocarbon method.

Periglacial area

The above glaciolacustrine formation is traceable upstream below the Arctic Circle where it merges with the flat swampy Farkovo plain at 45-55 m a.s.l. This plain is built of silty and clayey rhythmites overlain by 5-15 m of laminated fine sand with periglacial flora in peat and moss seams. The entire Yenissei depression, 100 km wide, at 66° N is filled with this lacustrine formation, which pinches out southward by 62.5° N (Arkhipov & Lavrushin, 1957). At Bolshoi Shar (site 3 in Fig. 1) a transition from the Early Weichselian alluvium to the varved sequence, covered by glaciofluvial sand, has been described (Fig. 3). At Farkovo (site 4 in Fig. 1) alluvium overlying the glaciolacustrine rhythmites yielded radiocarbon dates of ~ 41-42 ka (Kind, 1974). The stratigraphic position of the proglacial varved formation atop of the last interglacial was established at Alinskoye (site 5 in Fig. 1), where it contains a tundra flora and is underlain by alluvium with southern taiga pollen spectra and interglacial fern *Osmunda cinnamomea* (Zubakov, 1972).

On the stretch of the Yenissei between 66 and 63° N, loess-like silt with buried soils blankets the Farkovo lacustrine plain and higher terrains indiscriminately. The loessic cover on the lacustrine plain about 50 m a.s.l.

at site 5 (Fig. 1) and in the nearby Kangotovo section contains buried soils and basal peat which yielded 7 radiocarbon dates in the range 39 to 44 ka that are considered minimal (Sukhorukova et al., 1991). At Bakhta (16 in at Fig. 1), 120 km to the south, a plateau at 100 m a.s.l. built of pre-Weichselian tills is capped by a loess-like silt with a paleosol dated to 35 and 34 ka BP (Astakhov et al., 1986).

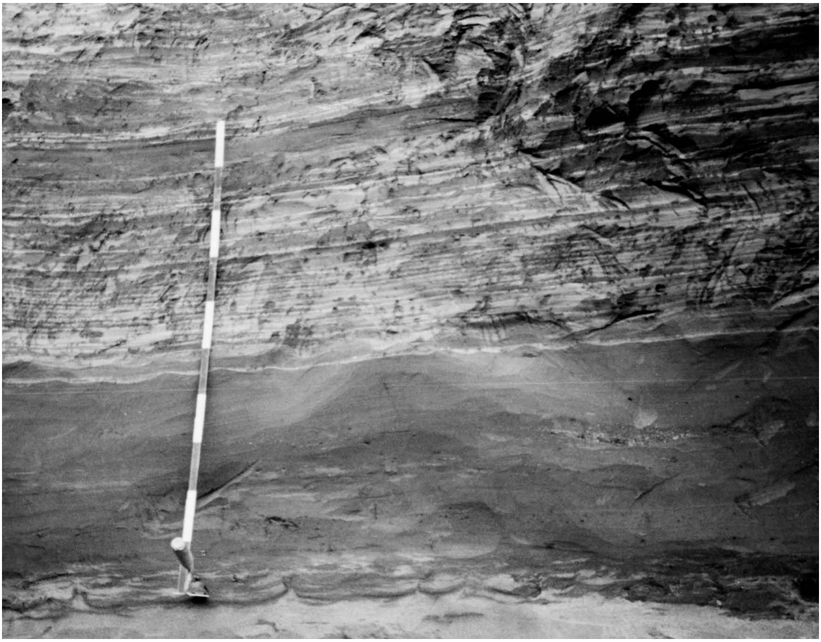


Fig. 3. Early Weichselian alluvial sand and silt grading up into glaciolacustrine varved clay at Bolshoi Shar, 66° N/88° E (site 3 in Fig. 1). Stick with 10 cm intervals.

Exactly this loessic area was originally chosen by Volkov et al. (1978) and Grosswald (1980) as a bottom of their Late Weichselian `Yenissei Lake` which presumably flooded all terrains up to 130 m a.s.l. Later, based on several questionable finite radiocarbon dates from beneath the till, the same area was pictured by Grosswald & Goncharov (1991) as a former long tongue of a Late Weichselian ice sheet. In this version of the maximalist reconstructions the ice protuberance extends to 61-62° N, i.e. by 500 km south of the mapped moraines. This glacier snout was presumably fringed by a narrow high-stand lake. The above authors saw signatures of such a lake in stratified silts and sands at altitudes ~ 120 m atop of peats with radiocarbon dates of 42 to 24 ka. Peats on the so-called `80 m terrace`, dated to 9.5-11 ka, were presumably covered by an ice-dammed lake of the terminal Late Pleistocene (Grosswald & Goncharov, 1991).

Lacustrine genesis and Late Weichselian age of the high-altitude sands is in disagreement with the established stratigraphic interpretation of the lower fluvial terrace at 30-40 m a.s.l. Its alluvium contains periglacial pollen, bones of large mammals and radiocarbon dates in the range of 26 to 12 ka BP and therefore represents the Late Weichselian interval (Laukhin, 1981) (Fig. 2). This is confirmed by LGM date 21,445 ± 145 years BP (paper 10, Ch. III) obtained by AMS method on a mandible of a woolly rhinoceros with teeth intact that was washed out from this terrace at the mouth of Podkamennaya Tunguska (site 6 in Fig. 1). This terrace and the above mentioned higher terrace are well expressed in topography and nowhere show signs of glacial overriding or deposition in deep water.

Therefore, there must be another explanation for stratified sediments at high elevations. Laminated aeolian coversand, sinkhole limnic rhythmites or low-energy fluvial sands commonly occur as lenses within many subaerial formations. In periglacial conditions aeolian coversand is abundant at all altitudes. The structure of the loess-thermokarst formation on the Middle Ob is especially instructive (Astakhov, 1989, see also Fig. 8). The Yenissei valley is hardly an exception. The only evidence for a high-level Late Weichselian lake in Yenissei Siberia is young peaty deposits overlain by stratified sand of unknown origin. The rest of geological facts lead to the same conclusion as in the glaciated terrain: the last glacial advance and corresponding ice-dammed lake occurred beyond the limit of radiocarbon method, i.e. prior to 50 ka BP. During the Late Weichselian large mammals were grazing along the northbound Yenissei.

Former drainage features

The geological data accumulated along the Yenissei can be explained by an early Weichselian proglacial lake with a level not higher than 70 m that probably spilled westward onto lower swampy plains of the Taz and Pur catchment areas (Fig. 1). Arkhipov & Lavrushin (1957) argued that the proglacial lake had no outlet, while other authors thought this idea improbable and suggested westbound spillways in the form of narrow valleys downcutting into old morainic plateaus (e.g. Zarrina et al., 1961). The most promising, although not investigated, spillway is located immediately south of the Turukhan end moraines close to Farkovo. This narrow through valley, now employed by diverging rivers Verkhnyaya Baikha and Pokolky (BP in Fig. 1), connects the Farkovo palaeolake with the Taz catchment area in the west.

At the beginning of ice sheet disintegration the Farkovo Lake propagated northward into the glaciated area in the form of the Igarka ice-contact lake, which eventually broke through stagnant ice to initiate the modern drainage pattern (paper 5, Ch. II). Postglacial sediments, yielding a wide range of radiocarbon dates, including infinite ones, are represented by two Pleistocene terraces with mammoth fauna and by the discontinuous mantle of loess-like silts with paleosols and peats. These terraces, descending downstream along the Yenissei, reflect the normal northbound drainage which was restored prior to 50 ka ago and has never been obstructed afterwards.

Lower Ob

Sediments of the Ob valley

A Weichselian proglacial lake in the lower course of the Ob River was mapped by Zarrina et al. (1961) as a low-level reservoir draining into the Kara Sea along the southeastern margin of a large Uralian ice sheet. In those times another option was unthinkable, since glaciers were thought to flow exclusively downslope, from mountains onto the West Siberian Plain. Later, when the former Kara Sea ice domes were reconstructed, their flow lines were immediately discovered around the northern tip of the Polar Urals, whereas traces of contemporaneous alpine glaciers were found only south of the ice sheet margin at 67° N (paper 2, Ch. I).

The proglacial reservoir according to Zarrina et al. (1961) was confined to an intra-valley flatland. It was described by several authors as a boggy terrace-like plain 40-55 m a.s.l., in places up to 150-180 km wide, continuously traceable from 60° to 67° N along the Lower Irtysh and Ob rivers (Fig. 1). Its surface is undulating due to the gentle mounds and hollows 4-8 m high, probably of thermokarst origin. South of the Irtysh River mouth the intra-valley plain is decorated with a swarm of old aeolian ridges up to 12-17 m high uniformly elongated in NE direction (Zaionts & Ziling, 1972). These authors stress that, unlike the lower riverine surfaces, this terrace-like plain keeps the same altitudes for more than a thousand km irrespective of the descent of the present-day riverbed.

The base of the intra-valley lowland consists of glacial diamicts overlain by fluvial sand. The topmost part is represented by terrestrial sediments, including loess-like silts, aeolian sands, low energy fluvial sands, sinkhole silts and peats. Sandwiched between these formations, typical glaciolacustrine varves and silty rhythmites up to 30-40 m thick constitute the middle part of the terrace sequence. The conventional key section is Sangompan (site 9 in Fig. 1) on the northern bank of the Ob River just above the Arctic Circle, presently in forest-tundra environment. Well-sorted, laminated and cross-bedded sand, up to 12 m, containing macrofossils of southern boreal forest, such as *Menyanthes trifoliata* and *Carex rostrata,* and arboreal pollen spectra, is observable at the base (Lazukov, 1970; Arkhipov et al., 1977). Also fresh-water ostracods and brackish-water diatoms are known from this sand which is radiocarbon dated to more than 50 ka (Zubakov, 1972). The sand gradually changes upward into silty rhythmite and varved clay (Fig. 4) up to 27 m thick with sparse tundra pollen.

Fig. 4. Transition from fluvial or estuarine sand with forest flora to cold-water clayey rhythmite at Sangompan, 66°48′ N/67°45′ E (site 9 in Fig. 1). The sand has mean OSL age 81 ka BP (Fig. 6)

Lazukov (1970) interpreted the sequence as fluvial and lacustrine sediments of the last interglacial. Zaionts & Ziling (1972) suggested an estuarine/deltaic origin and Mid-Weichselian age for this formation. Most authors think that only the sand unit is interglacial, whereas the overlying clayey rhythmite with varves was deposited in an Early Weichselian proglacial lake (Zubakov, 1972; Arkhipov et al., 1977). The last option seems more reasonable, since the upward change from forest to tundra vegetation concurs with the facies succession.

The intra-valley varved sequence becomes sandy upstream and in places lies with non-erosive contact atop of well-pronounced peaty soils, according to Zolnikov (1991), who traced it along the river up to 64.5° N. The present author observed similar rhythmites with symmetric sand ripples overlying terrestrial peat or frost disturbed aeolian sand and loess-like silt at the eastern bank of the Ob at 65° N (Fig. 5) and farther upstream at 62° N.

Along the longitudinal stretch of the Ob valley the lacustrine rhythmite in many places contains fresh-water ostracod assemblages with dominant species *Cytherissa lacustris* and *Limnocythere baltica.* Insect fauna is represented by cold-water species, including *Tahinus sp.* Dry-land insects appear above the rhythmite in the upper part of the terrace sequence, from where also mammoth fauna and many lenses of peat were described. The fauna assemblages are characteristic of Late Pleistocene formations (Zaionts & Ziling, 1972). This age is also evident from its position within the Ob valley, where the described sequence always occurs incised into the thick Middle Pleistocene diamictic sediments. However, the two rhythmite formations near the Ob mouth (locations 9 and 10 in Fig. 1) are not synchronous.

Although an early Weichselian age of the Sangompan varves (9 in Fig. 1) seems to be unanimously accepted, a very similar rhythmite with fresh-water diatoms on the opposite bank of the Ob has met different interpretations. Arkhipov et al. (1977) correlated it with their presumably Middle Weichselian glaciomarine inter-till clay unit at Harsoim. Zubakov & Levkovskaya (1969) thought that the southern bank rhythmite was a glaciomarine formation of a Middle Pleistocene age. This geochronological conclusion is based on the arboreal pollen spectra and a radiocarbon date more than 57 ka BP obtained from a fluvial sand deeply incised into the rhythmite near Pyak-Yaha creek (Fig. 6, site 10 in Fig. 1). These authors refer the Pyak-Yaha sand to the last interglaciation. Therefore, according to Zubakov (1972), the last ice advance across the Ob valley occurred in the end of the Middle Pleistocene. Lazukov (1970) thought that both banks were glaciated in the Early Weichselian and Arkhipov et al. (1977) – even in the Late Weichselian.

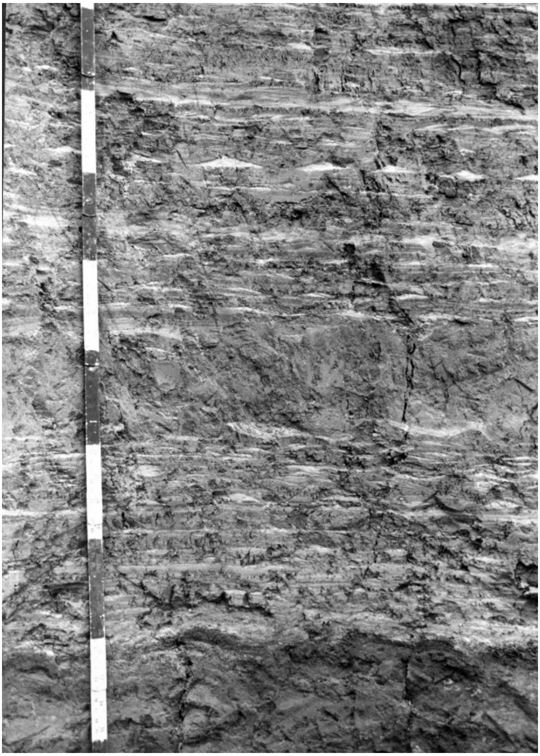


Fig. 5. Clayey silt rhythmite with symmetric ripples of fine sand overlying a thin peat on the eastern bank of the Ob at Pitlyar, 65°52′ N/65°53′ E. Scale in 10 cm intervals. Beyond the photograph the rhythmite gradually changes upwards to fine sand with clay laminae and plant detritus OSL dated to 72 ± 4 and 75 ± 4 ka BP, and downward from the peat to cross-bedded sand OSL dated to 66 ± 4, 62 ± 3, 108 ± 6 and 76 ± 4 ka BP (mean value 78 ka BP).

Recent geochronological results

New data by the Russian-Norwegian PECHORA project help to constrain the age of the last ice-dammed lake (paper 3, Ch. I). In the southern bank sections a varved rhythmite occurs between two sands with peat interlayers (Fig. 6). The lowermost sand overlies a till, and the uppermost sand is covered by loess-like silts with peat and solifluced diamict lenses at the base. In a sand pit at Aksarka village, 29 km south from site 9 in Fig. 1, the base of the loessic cover was radiocarbon dated to 27,840±310 years BP (LU-4555). Peat in an ice-wedge incised into the underlying laminated sand is ≥44,600 years old (LU-4553). A piece of wood from the sand yielded another non-finite date ≥46,300 (LU-4580).

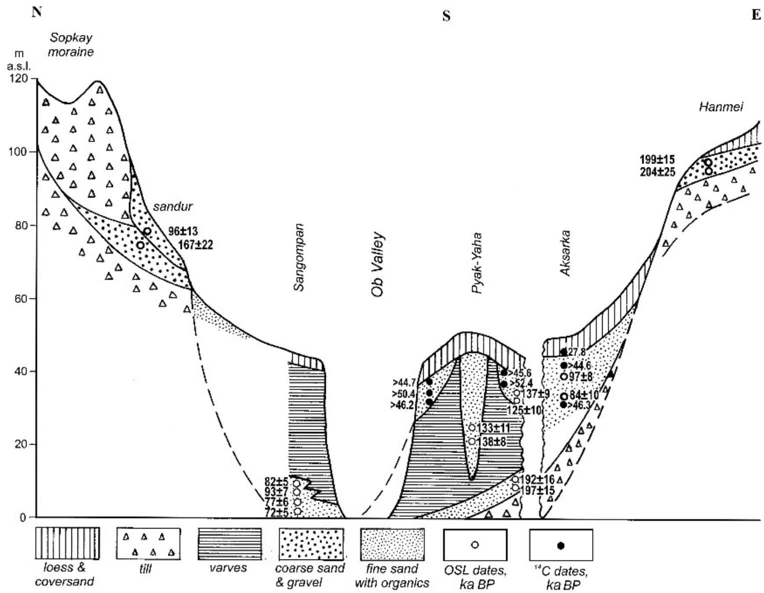


Fig. 6. Principal profile of dated Quaternary formations along the River Ob valley on the Arctic Circle, sites 9 and 10 in Fig. 1. Horizontally is not to scale (from Astakhov, 2004). See Fig. 4 for the base of the Sangompan varved formation.

Finite age values have been obtained by OSL (optically stimulated luminescence) (Murray & Wintle, 2000). At Aksarka OSL method has produced ages of 84 ± 10 ka BP at 31.5 m and 97 ± 8 ka BP at 39 m above the river. The infinite radiocarbon age of the sand at Pyak-Yaha (Zubakov & Levkovskaya, 1969) is confirmed by our dates ≥46 200 (LU-4764) and ≥50 400 (LU-4759). Even a meter thick peat at the base of the loess-like silt is old: ≥44 700 years BP (LU-4765). In a section 6 km downstream from Pyak-Yaha again infinite radiocarbon dates ≥52 400 (LU-4769) and ≥45 600 (LU-4767) have been obtained from apparently locally deposited lenses of plant detritus found in a sand between the surficial loess-like silt and the lacustrine rhythmite (Fig. 6).

Crucial are the OSL dates in the range of 138 to 125 ka which suggest an Eemian age for the upper sand with peat and forest pollen and a Middle Pleistocene age for the underlying rhythmites of the southern bank.

However, the rhythmite sequence of the northern bank is apparently younger, judging by a series of OSL dates in the range 93 to 72 ka (mean value 81 ka) at Sangompan obtained from the underlying sand which displays a gradual transition to the rhythmite (Figs. 4, 6). A similar series of OSL dates has been obtained in the upstream section at Pitlyar where rhythmite is sandwiched between mean OSL values 78 and 73 ka (Fig. 5). Thus, in the Lower Ob natural exposures two glaciolacustrine varved formations (Middle Pleistocene and Early Weichselian) are intercalated with three sandy formations (Middle Pleistocene, Eemian and Early Weichselian) (paper 3, Ch. I).

The Lower Ob sequence is mantled by loess-like silts and crudely stratified coversand up to 9 m thick (Fig. 6). Numerous remains of mammoth fauna originate from this subaerial formation. Radiocarbon dates on the bone material have given ages 25, 17, 16 and 15 ka BP, suggesting that during the Late Weichselian interval this unglaciated area was a habitat of grazing mammals (paper 11, Ch. III). This situation is similar to the Yenissei valley.

To test this chronological concept glaciofluvial sands on a higher plateau have also been examined. A sand atop of a clayey till without Uralian erratics in a road pit 13 km to NW from the city of Labytnanghi, the western bank of river Hanmei, have yielded OSL ages 204 ± 25 and 199 ± 15 ka BP (paper 3, Ch. I). Although OSL ages older than 150 ka cannot be treated as reliable, it is clear that the glaciofluvial sand mantling the eastern Uralian piedmont was deposited prior to the Late Pleistocene. Even in the Uralian montane valleys fine-grained sediments have yielded similarly old OSL ages suggesting a limited alpine glaciation of the Late Pleistocene. Isolated fresh morainic landscapes of the Late Pleistocene age mapped in the Uralian valleys are more common on the western, more humid slope of this mountain range (paper 18, Ch. IV).

Ice limit updated

No fresh glacial features or boulder trains have been found by members of the PECHORA on the Arctic Circle within the eastern Uralian piedmont whose flat solifluction surface is in places decorated by tors. The piedmont is cut off from the above described lacustrine plain by a knickline or a low escarpment persisting along the 60 m isohypse. Only at 60-70 km north of the Arctic Circle both the piedmont and the adjacent lowland are traversed by a west-east striking chain of kames and fresh-looking morainic hillocks. These are the Sopkay moraines, unmistakable signs of a frontal ice flow from the north into the Urals. Only south of 67°

N small alpine moraines appear at the foot of the Urals (paper 2 in Ch. I and paper 18 in Ch. IV).

These data mean that in works of the 1970s (Arkhipov et al., 1977; Arkhipov et al., 1980; Grosswald, 1980) the extent of the last ice sheet was vastly exaggerated. Even in QUEEN related publications (e.g. paper 10, Ch. III) the Late Pleistocene ice limit was drawn 150-250 km too far south, along the Salehard moraines which later proved to be Middle Pleistocene (paper 3, Ch, I). The stratigraphic data for the new age model of the Lower Ob sequences are presented in Fig. 6. In view of the new results the last ice limit should be located along the northernmost Sopkay moraines which are continuations of the Markhida moraines of the Pechora Lowland, Harbei and Halmer moraines of the western Urals and (Fig. 1). The latter morainic systems collectively outline the southern margin of an early Weichselian ice sheet centred on the Kara Sea shelf (paper 18, Ch. IV).

The Sopkay moraines were originally thought to belong to a stade of the Late Weichselian glaciation (Arkhipov et al., 1977; Astakhov, 1979). However, the recent studies have unambiguously shown that the last Yamal ice sheet disintegrated prior to 45 ka BP (Forman et al., 2002). The Sopkay glacial maximum is sandwiched between the above mentioned OSL ages of about 80 ka at the base of the Sangompan varved sequence (Fig. 6) and 70-60 ka dates from the sand above massive foliated ice on Yerkata River in the southern Yamal Peninsula (Fig. 7).

The limit of the last ice sheet is marked by topographically expressive end moraines mainly in the south of the Yamal Peninsula and in the Yenissei catchment area. Isolated glaciotectonic ridges in between these areas do not outline a continuous ice limit, which is drawn across the central lowland tentatively, using the occurrence of thick deposits of massive buried ice (Fig. 1) that are largely interpreted as relicts of a Late Pleistocene ice sheet (Kaplanskaya & Tarnogradsky, 1986; Astakhov & Isayeva, 1988; Astakhov et al., 1996).

Especially important are stratiform ice bodies in the eastern Yamal and on the Tadibei-Yaha River, the Gydan Peninsula (site 7 in Fig. 1). The massive ice at Tadibei-Yaha is foliated, abounds in thin dirt bands and contains a large recumbent fold. The glacial origin for massive ice bodies of this and other key sites was confirmed by a group of international permafrost experts based on the ice deformations, dirt bands and thaw contacts (Péwé & Brown, 1989). The chain of relict glacier ice sites across West Siberia (Fig. 1) excludes a contemporaneous drainage to the north. Besides, an ice dam in arctic West Siberia is necessary to explain the configuration of ice-pushed ridges on Yamal and Gydan peninsulas which were evidently produced by ice flows diverging from the central lowlands (Astakhov, 1979; Arkhipov et al., 1980; Grosswald, 1980; Astakhov & Mangerud, 2005).

An early Weichselian age of massive stratiform ice in natural exposures is inferred from OSL dates in overlying sediments 72, 65, 63, 59, mean 65 ka BP, in southern Yamal (Astakhov, 2004; Mangerud et al., 2004) and thermoluminescence dates 78 and 79 ka on the Yenissei (Kostyayev et al., 1992). Such ice beds, which are in places up to 60 m thick and 2 km across (Solomatin et al., 1993), are underlain by a till and interglacial sands of unknown age. It is very unlikely that the fossil glaciers found at some 20-30 m a.s.l. (Fig. 1) could survive the warm-water Eemian transgression.

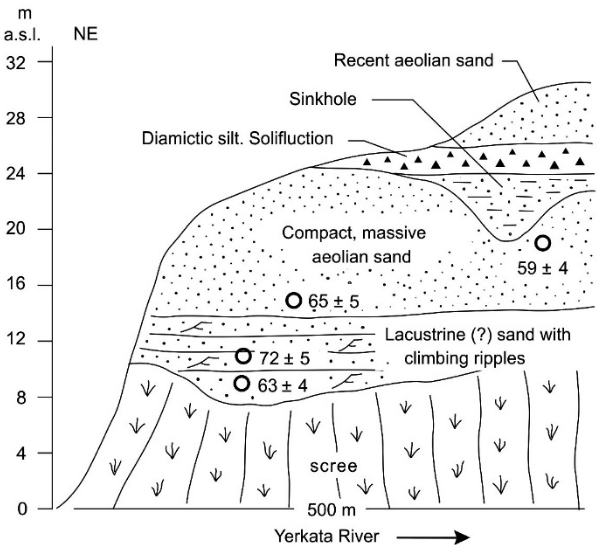


Fig. 7. OSL dates, ka BP, from sediments overlying the upper glacial formation on river Yerkata, southernYamal, site 8 in Fig. 1 (Mangerud et al., 2004). Lower part of the section exposed 1 km downstream displays massive, foliated, distorted ice covered by varved clay some metres thick.

The new, less extended, ice limit still implies an ice-dammmed lake that had no outflow toward the Arctic Ocean, i.e. the only possible drainage was in a southern direction. However, first attempts by the PECHORA project to find Late Pleistocene varved sequences in the central West Siberian Arctic south of the ice limit in Fig. 1 have not brought positive results (Nazarov, 2005). For the range 100 to 70 ka (by OSL) only a sandy terrace at 30 m a.s.l. has been found so far along the Taz Estuary (D. Nazarov, pers. comm. 2005). A shallow ice-dammed lake can be tentatively suggested for flatlands around the Taz Peninsula lower than 60 a.s.l. (the sparsely dotted area in Fig. 1). Alternatively, this area may belong to a system of wide shallow straits connecting the Yenissei and Ob proglacial lakes.

The peculiar shape of the southern margin of the last ice sheet of West Siberia with an embayment in the central lowlands (Fig. 1) demands an explanation. This shape might be caused by thermal diversity of the substrate, which consists of soft sediments in the centre of the Plain but hard Paleozoic rocks along the Urals and Yenissei. The Paleozoic bedrock, showing drumlinoids, eskers and other wet-base features, could provide a fast southward sliding of the westernmost and easternmost ice streams. On the contrary, in the Gydan Peninsula the ice sheet was frozen to its soft bed, as follows from the deposits of relict glacier ice, Fig. 1, and the total absence of wet-base features. Therefore, its advance was considerably hindered and could proceed mostly by glaciotectonic squeezing and stacking of the perennially frozen substrate (paper 8, Ch. II).

Age of ice-dammed lake

The thick varved sequence incised into the Middle Pleistocene drift at site 9 and farther upstream (Fig. 1) overlies all tills in this area. Middle Pleistocene ice limits have been mapped at hundreds of kilometers to the south, close to the 60th parallel (Zarrina et al., 1961). A Late Weichselian age of the Lower Ob rhythmites is rejected by numerous old radiocarbon dates from atop of the rhythmite, including the 28 ka date from the base of the aeolian mantle (Fig. 6) and several Late Weichselian dates on mammal bones (paper 1, Ch. III). The OSL dates 72 to 93 (mean value 81 ka BP) from the sand gradually changing to the Sangompan varves (Fig. 6) are consistent with the mean OSL age of 78 ka from the base of the rhythmite at Pitlyar (Fig. 5). These OSL values coupled with the mean OSL age of 68 ka from overlying sediments in Yamal (Fig. 7) and Pitlyar (Fig. 5) is definite evidence of an early Weichselian age of the last ice-dammed lake on the Ob which probably was a counterpart of the Yenissei proglacial lake. Slightly higher elevations inferred for the Yenissei ice-dammed lake agree with its suggested overflow to the west (Fig. 1).

Middle Ob

The Middle Ob River bluffs become lower upstream of the Irtysh River mouth, obscuring sediments of the proglacial lake. The southernmost exposures of an unmistakably lacustrine silt/clay rhythmite or finely laminated clay are known at 61° N, on river Seul (Zaionts & Ziling, 1972), and close to Yelizarovo village (Lazukov, 1970), where such sediments about 10 m of observable thickness lie beneath subaerial silt and sand at 40 m a.s.l. (sites 12 and 13 in Fig. 1 respectively). The lacustrine sediments contain a rich ostracod fauna, with *Cytherissa lacustris* predominating. In the opinion of Zaionts & Ziling (1972) they are estuarine in origin. This interpretation seems unlikely because the varved clay formation persists at the same altitudes from the Ob mouth at least for 1600 kilometers upstream (Fig. 1).

A level of the former intra-valley lake in the south could not be very high, because on the eastern bank near the Irtysh mouth at 72 m a.s.l. it is replaced by a thick sequence of loess-like silts not covered by lacustrine sediments. This loessic mantle with infinite radiocarbon dates has yielded thermoluminescence ages of 64 ± 8, 75 ± 15, 80 ± 13 and 80 ± 11 ka BP (Arkhipov et al., 1987). The easternmost site of lacustrine sediments, up to 15 m thick, with ostracods and fresh-water mollusks, was described from boreholes on Salym River, 45 m a.s.l. (Zaionts & Ziling, 1972, 14 in Fig. 1). Lacustrine facies are nowhere exposed farther upstream, being replaced by younger fluvial and subaerial sediments. The varved clay, known on the Middle Ob south of 60° N at altitudes 60 to 90 m, is overlain by interglacial sand and therefore relates to the Middle Pleistocene (Zubakov, 1972).

The Middle Ob valley is a key area, where the hypothesis of the huge Late Weichselian proglacial lake (up to 130 m a.s.l.) was first tested (Astakhov, 1989). This stretch of the Ob valley has a step-like topography with small flat knolls and bogs at 40-50 m a.s.l. and, according to Volkov et al. (1978) and Grosswald (1980), should be covered by deep-water facies of a Late Weichselian high-stand proglacial reservoir called `Mansi Lake`. However, the only sequence Volkov was able to refer to as a candidate for this lacustrine formation is crudely laminated sand lying atop of fluvial silt and sand at Mega (Arkhipov et al., 1980, site 15 in Fig. 1). Arkhipov also put forward loess-like silts of the surficial Urtam formation as deep-water sediments overlying fluvial strata of the `Second` (Kolpashevo) terrace which is radiocarbon dated at ~ 30 ka and older (Arkhipov et al., 1980).

A closer inspection reveals that the sandy rhythmites of limnic appearance at Mega section always pinch out laterally within the massive loess-like silt which is the main constituent of the sedimentary cover (Fig. 8A, site 15 in Fig. 1). Most significant is the occurrence of numerous ice-wedge casts on various levels and in lithologically different sediments, testifying to the dominant subaerial or shallow-water environments. The casts are intimately associated with discontinuous remnants of buried soils (Fig. 8B) from which mammal bones and spruce remains were reported (Arkhipov et al., 1980). Finally, the loess-like mantle with rhythmite lenses occurs not only over riverine terraces but also on the highest interfluvial terrains up to 150 m a.s.l., where it is even thicker and interspersed with old dunes (Fig. 2 in Astakhov, 1989).

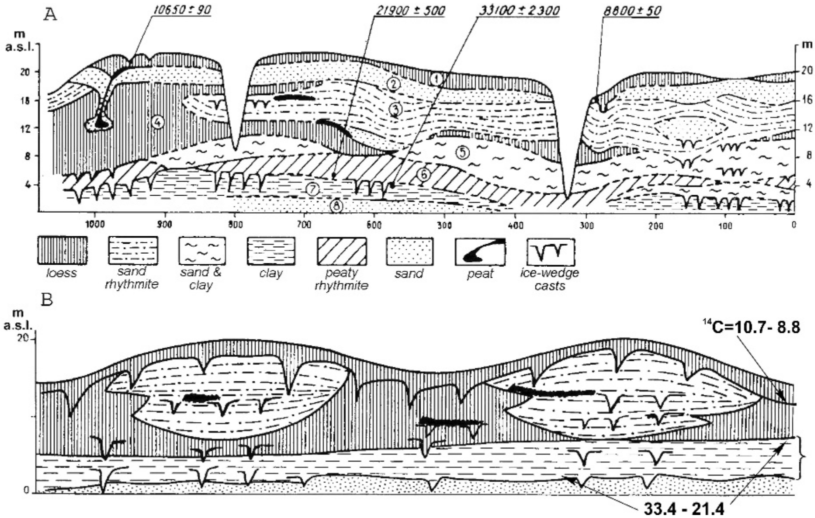


Fig. 8. Periglacial sediments on the northern bank of the Middle Ob, site 15 (Fig. 1). A. Section of Middle Weichselian fluvial sediments overlain by Late Weichselian aeolian/thermokarst formation on Mega Channel, 13 km upstream of Megion town, 61°03′ N/76°20′ E. Circled numbers are lithostratigraphic units described by Astakhov (1989). Conventional radiocarbon dates noted above the section are from Arkhipov et al. (1980). B. Principal cross-section of the terrace-like plain 40-50 m a.s.l. along Mega Channel (approximately 1 km long). Note the topographic highs above rhythmite lenses versus lows above non-stratified loess-like silt. Regularly undulated plain is a result of terrain sagging and inversion of sink ponds in the course of degradation of Pleistocene permafrost. Radiocarbon brackets in the right of 8B are by Arkhipov et al. (1980).

The irregularly stepped topography of the Ob valley, dotted by oval and doughnut-shaped knolls, by no means looks like the bottom of a huge lake. Although monolithic permafrost was ubiquitous here not long ago, which is evident from its relicts observed in boreholes some 150-200 m below the ground surface (Zemtsov, 1976; Astakhov et al., 1996). The sand-rhythmite mounds alternating with inter-knoll loessic depressions (Fig. 8B) are better explained by uneven sagging of the perennially frozen steppe accompanied by inversion of sink ponds in the course of the Holocene permafrost degradation (Astakhov, 1989, 1998).

The bulk of surficial Pleistocene sediments along the Middle Ob can be determined as a Late Weichselian aeolian formation with lenses of thermokarst lake sediments on top of the Middle Weichselian alluvium. The alluvium is easily traceable downstream, where it in many places is distorted by casts of syngenetic ice-wedges propagating upwards into the loessic sequence (Astakhov, 1989). Similar observations were reported by Krivonogov et al. (1993) at Lokosovo, a hundred km downstream from Mega. They dated peat slumped into an ice wedge cast in the alluvium to more than 50 ka BP. In the overlying sediments of a local lake the age of 46 325 ± 1800 years BP has been obtained. The loessic cover dated at the base to 22 930 ± 650 years BP suggests basically arid environments in the Late Weichselian which is confirmed by *Artemisia* and other herbaceous pollen. Only in fluvial and rhythmite sand facies pollen of cold-climate birch parklands is recognised (Arkhipov et al., 1980; Krivonogov et al., 1993).

Thus, no sediments of a large proglacial lake crop out upstream of site 13 in Fig. 1 on the Middle Ob, where only diverse facies of younger Weichselian formations 15-20 m thick are observable. Counterparts of the Lower Ob glaciolacustrine rhythmites were either eroded away during a Mid-Weichselian fluvial cycle or buried under young fluvial and subaerial sediments. The area of the last ice-dammed lake of an early Weichselian age in the north is drawn based on the above stratigraphic data and altitudes of the topographic knicklines ~ 70 m a.s.l. along the Yenissei and 60 m a.s.l. in the Lower Ob valley. However, east and south of the Irtysh mouth, where the surface is considerably modified by former permafrost sinkholes and aeolian dunes, a large lake is delineated very tentatively along the 60 m isohypse (Fig. 1).

The sum of stratigraphic evidence available in the Ob valley testifies to changes of Weichselian environments from a cold proglacial lake older than 50 ka towards normal fluvial regime after the draining of the valley lake, and finally to cold and arid landscapes with a predominance of aeolian sedimentation. Judging by the radiocarbon dates, that main arid phase occurred after 22 ka BP, i.e. exactly at the time of the greatest West Siberian lake postulated by Volkov et al. (1978) and Grosswald (1980).

Southwestern region

Weichselian formations in Irtysh and Tobol valleys

In search of an outlet of ice-dammed lakes the distal part of the proglacial drainage system should be considered. Along the Tobol and Irtysh rivers Late Pleistocene deposits can only be seen in two old riverine terraces, whose heights gradually increase upstream concurrently with altitudes of the present riverbed. They are built of fluvial sediments changing upwards into loess-like silts. Zaionts & Ziling (1972) think that this is the sign of a lateral transition from estuarine and deltaic facies of the Lower Ob to normal alluvium of the \`Second Terrace\`. This interpretation is stratigraphically untenable for the Middle Ob, where fluvial sediments are notably younger than the lacustrine rhythmites of the Lower Ob (cf. Figs. 6, 8).

A similar situation occurs on rivers Irtysh and Tobol, where a number of radiocarbon dates in the range 32 ka BP to infinite are known from the basal strata of the \`Second Terrace\` alluvium with remains of mammoth fauna. Fossil floras reflect a change from middle taiga associations (slightly colder than the present-day southern taiga environments) upwards to northern taiga and forest-tundra with severe permafrost (Kaplyanskaya & Tarnogradsky, 1974; Krivonogov, 1988). At Lipovka, a key section of the \`Second Terrace\` on Tobol River, 58° N, 15 m thick alluvium is overlain by 10-15 m of coversand and loess-like silt. A gradual transition upwards from the alluvium via cryomorphic palaeosols into loess-like silt and coversand with cryoturbations was reliably dated at Lipovka by radiocarbon analysis of *in situ* larch stumps to 30,700 ± 300, 30,560 ± 240 and 30,200 ± 250 years BP. Wood at 4 m below yielded 31,300 ± 800 years BP (Kaplyanskaya & Tarnogradsky, 1974).

These authors and Krivonogov (1988) considered the coversand as lacustrine sediments. However, its weak and wavy stratification, lack of ripples, abundance of permafrost disturbances, indistinct paleosols, mammalian bones, and the mantle-like occurrence on various topographic elements definitely indicate that it is subaerial, mostly aeolian in origin and is basically the same surface formation as described along the Ob and Yenissei rivers. Attempts to find lacustrine beaches or knicklines along the Urals at altitudes 120-130 m prescribed by the maximalist hypothesis (Volkov et al., 1978) revealed only thick loess sheets.

Lacustrine rhythmites of the Lower Ob type are definitely older than the `Second Terrace` formations (Fig. 9). They are nowhere exposed along the Irtysh River and probably have been eroded away. Alluvium of the lower `First Terrace`, which contains mostly northern taiga plant macrofossils with arctic species, yielded radiocarbon dates 16 770 ± 160 and 14 310 ± 70 years BP (Krivonogov, 1988) and in the upper part, just below the last generation of ice-wedge casts there are three radiocarbon dates ~ 11-12 ka BP (Zubakov, 1972).

The `Second` and `First` terraces are readily traced and mapped downstream to the Ob mouth (Lazukov, 1970). In the Lower Ob valley they are built of sandy alluvium and definitely incised into the above described intra-valley lacustrine plain. The lower part of the `Second Terrace` alluvium contains arboreal pollen spectra, whereas the upper part of alluvium is characterised by tundra pollen and ice-wedge casts. This alluvium is conventionally interpreted as a Mid-Weichselian formation. Tundra pollen recovered from the `First Terrace` alluvium of the Lower Ob reflects a colder climate, presumably of the terminal Weichselian (Lazukov, 1970; Zubakov, 1972).

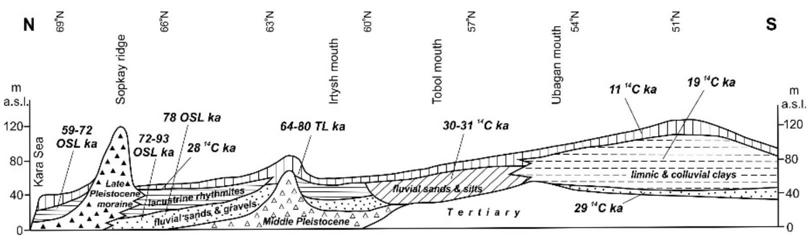


Fig. 9. Correlation of Upper Pleistocene formations within valleys of rivers Ob, Irtysh, Tobol, Ubagan (approximately along 67° E). Topmost layer is aeolian and low-energy fluvial sands and loess-like silts with occasional paleosols. Note the reverse gradient of the bedrock directed southward to the Caspian basin.

Sedimentary infill of the Turgai Valley

Southwards the gradually rising floor of the Tobol River valley and her tributary river Ubagan merges with the dry Turgai Valley containing small salty lakes (Fig. 1). This through valley extends across the Kazakhstan steppes and semi-deserts, being the only passage connecting the West Siberian lowlands with the Aral and Caspian seas. The highest point of the present surface of the Turgai Pass, 126 m a.s.l. at 51° N, was suggested by Volkov et al. (1978) and Grosswald (1980) as the threshold of their Late

Weichselian ice-dammed lake which presumably spilled into the Aral and Caspian basins.

However, several drilling profiles across the Turgai Valley uncovered everywhere a thick fine-grained succession consisting mainly of limnic and paludinal clays with silt lenses, rich in organics, carbonates and gypsum. These clays interfinger with colluvial clayey breccia and are topped by aeolian sand and silt up 5 to 15 m thick. The thickest sediment fill (almost 90 m) is recorded at the highest point of the present day passage at 51° N (Fig. 9). Fluvial facies occur only as thin (0.2 to 1.5 m thick) basal gravel or sand mantling the bedrock floor at 30-40 m a.s.l. South of 51° N basal lenses of fluvial sand up to 15-20 m thick are more frequent. Farther south sand may reach 50 m in thickness, probably forming a fan inclined to the Aral Sea (Boboyedova, 1971).

This predominantly clayey sequence was related by mapping geologists to the late Middle-Pleistocene and Late Pleistocene, based mostly on pollen spectra which showed two arboreal and two steppe phases of vegetation alternating. Boboyedova (1971) explained the arboreal phases by humidification of the southern landscapes during Middle Pleistocene and Late Pleistocene glaciations of West Siberia, the steppe phases reflecting the last interglacial and the Mid-Weichselian interstadial. According to her, a rising level of a Middle Pleistocene proglacial lake in Siberia caused an overflow to the south with a major channel incised into the soft Tertiary bedrock down to 40-50 m a.s.l. After the final drainage of the proglacial reservoir, the Turgai Valley was gradually filled with sediments of residual lakes and also by colluvial and aeolian sheets.

The general scenario for sediment filling the Turgai Pass seems likely, although the timing can be corrected using the later obtained radiocarbon dates. At 52° N, 113 m a.s.l., a piece of driftwood from the base of the clayey sequence at 75 m from the surface yielded 28 800 ± 800 years BP, a peaty soil at the depth of 34 m was dated to 19 140 ± 500 years, and twigs from the base of loess on the `First` river terrace are 11,600 ± 160 and 10 800 ± 800 years old (Astakhov & Grosswald, 1978; Grosswald, 1983, see also paper 6, Ch. II for the map of the Turgai Pass). These dates suggest that the main sequence of the Turgai Valley is a stratigraphic counterpart of the alluvial and subaerial formations of the terraces in the Irtysh, Tobol and Middle Ob valleys (Fig. 8, 9), evidently postdating the varved rhythmites of the Lower Ob valley (Fig. 6). Therefore, the latter cannot be younger than the fluvial gravel mantling the bedrock floor of the Turgai Pass. The through valley itself might have been originally cut during the Middle Pleistocene, afterwards used for a spillway in early Weichselian times and filled with sediments much later.

Proglacial drainage

According to the borehole data (Boboyedova, 1971), the bedrock bottom of the Turgai Pass descends from its highest point 55 m a.s.l. at 55° N southward and at a point 500 km south, where the present dry valley floor is highest (126 m a.s.l.), it is already at 40 m a.s.l. (Fig. 1, 9). This implies that the bulk of the valley sequence originated much later than the bedrock threshold of the spillway. The silt/clay infill mostly reflects an evolution of the Middle-Late Weichselian environments from stagnant fresh-water towards salty lakes and then to dry loess steppe. A similar succession of events in the course of the progressive aridification is also recorded in intra-valley sedimentary formations of West Siberia (see above).

Thus, the nature of sediments of the only possible southern outlet of the West Siberian Basin does not support the idea of a Late Weichselian overflow of a proglacial lake at 125-130 m a.s.l. suggested by Volkov et al. (1978) and Grosswald (1980). No fluvial facies is recorded on this level and 70 m downwards. The limnic and subaerial origin of the bulk of the Turgai valley sediments allows the southbound drainage only at altitudes not higher than 60 m and prior to 29 ka (Fig. 9). The available stratigraphic data are not enough to exactly pinpoint a time span of the last overflow from West Siberia into the Aral Sea, but they do not contradict a much smaller early Weichselian ice-dammed lake reconstructed in the north.

The geological record of the Late Weichselian provides no evidence of the huge amount of fresh water coming into the Caspian Sea from Siberia that was postulated by Grosswald (1980) for the Late Weichselian Khvalyn Transgression of the Caspian Sea with an overflow channel to the Black Sea at 26 m a.s.l. If the above discussed West Siberian proglacial lake (Fig. 1) is contemporaneous with Lake Komi in European Russia, i.e. 100 to 80 ka old, as calculated by Mangerud et al. (2001a), its overflow via the Turgai Pass probably happened well before the Khvalyn Transgression. The end of the interglacial Late Khazar Transgression, when the Caspian-Black Sea threshold was just below present sea level, Caspian water became less saline and colder by 5°C, and Caspian mollusks migrated to the Black Sea, seems more appropriate for the Early Weichselian influx of proglacial water from Siberia. This episode on the

Lower Volga has been dated by ESR, U/Th and luminescence within the range of 117–85 ka BP (Shkatova & Arslanov, 2004).

Judging by the available luminescence dates, the West Siberian proglacial lakes and southbound drainage should have developed between 80 and 70 ka BP (Fig. 5, 6 and 7), i.e. somewhat later than Lake Komi in European Russia and Late Khazar Transgression of the Caspian Sea. A similar time span is recorded by OSL dating on proglacial sediments farther east, on the Taimyr Peninsula (Alexanderson et al., 2001). In theory this might mean a delayed build-up of the ice dam in Siberia. However, this temporal disparity might also be accounted for by imprecision of OSL dating method which error margin can be as wide as 20%.

Conclusions

● Traces of proglacial lakes are common along the great Siberian rivers and very scarce outside their valleys. The available geological data point out to a low level of the last ice-dammed lake in West Siberia (not higher than 70 m a.s.l.) and its early Weichselian age.

● No signs of a Late Weichselian high-stand proglacial lake that carried most of the Siberian runoff into the Mediterranean (Volkov et al,. 1978; Grosswald, 1980, 1983) have been found in the geological record of West Siberia, even within river valleys. The Late Weichselian was the driest period with predominant subaerial sedimentation.

● Last major reorganisation of the drainage pattern of West Siberia apparently occurred in early Weichselian times, roughly simultaneous to the similar event in adjacent northeastern Europe (Mangerud et al., 2001a, 2001b). Judging by the sedimentary record of the Turgai Valley its impact on the water balance of the southern seas seas was not of the magnitude postulated by Grosswald (1980) for the Late Weichselian.

● Normal northward drainage, which formed two regional fluvial terraces, was restored in West Siberia before 50 ka ago and has never ceased since.

● Lacustrine rhythmites, traceable at least for 1600 km along the Ob valley and dated to 70-80 ka BP, together with the configuration of glacial features demand a solid ice dam at the lowest central part of arctic West Siberia. However, environmental reconstructions for this ice-dammed lake, being still enigmatic in many aspects, call for additional field studies, first of all in the central lowland along the provisional ice limit.

Acknowledgements

This work was supported by the Russian-Norwegian project PECHORA, Norwegian Research Council, and by the Institute of Remote Sensing Methods for Geology (NIIKAM), Russian Geological Survey. Dating by optically stimulated luminescence was performed by A. Murray in Nordic Laboratory for Luminescence Dating, Risø, Denmark. Radiocarbon ages with prefix LU have been obtained in the Geochronological Laboratory of St. Petersburg University, Russia, under the guidance of Kh. Arslanov. A. Matyushkov and D. Nazarov assisted in obtaining field data. My sincere thanks are due to all these persons and institutions.

15. The Geochronometric Age of Late Pleistocene Terraces on the Lower Yenissei

V. Astakhov and J. Mangerud

Doklady Earth Sciences, 2007, Vol. 416(7), P. 1022–1026.
Translated from Doklady Akademii Nauk. 2007, vol. 416(4), P. 509–513.

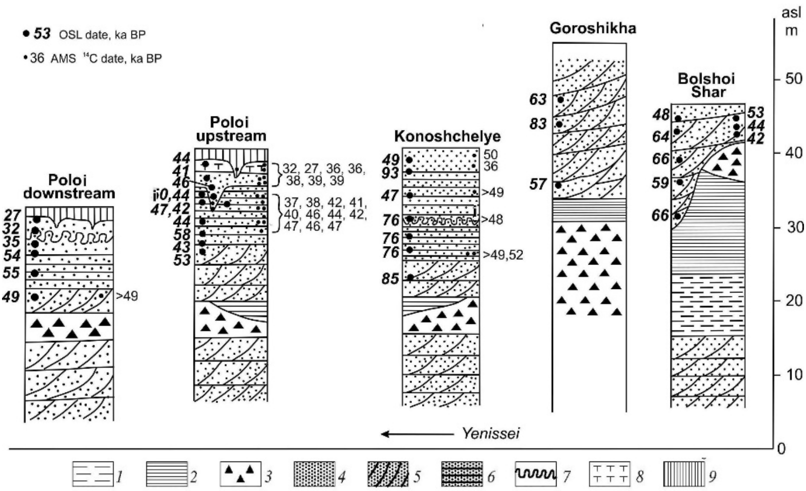
New radiocarbon AMS and optically stimulated luminescence (OSL) dates from key sections of postglacial terraces of the Yenisei River near the Arctic Circle demonstrate that the bulk of the Second Terrace alluvium was deposited during the early marine isotope stage 3 (MIS 3) some 45-60 ka BP. The glaciofluvial cover of the Third Terrace was formed by melting Putorana glaciers approximately 60 to 75 ka BP.

The terrace sequences of the Yenisei River between the cities of Turukhansk and Igarka (Turukhansk District, 66°-67° N) have been used since the 1970s as stratigraphic markers for reconstructing climatic fluctuations in Siberia and their correlation with Late Pleistocene events of Western Europe. Following Sachs (1953), many authors correlated all waterlain sediments, which cover the Zyryanka glacial complex at 30-60 m a.s.l. with the Karginsky marine strata representing the last boreal transgression of Siberia north of 69° N. Based on conventional radiocarbon dating, the Karginsky thermochron with interglacial climate was attributed to the interval of 50 to 22 ka BP (Kind, 1974; Arkhipov, 1990; Zubakov, 1972), i.e., to MIS 3. This correlation was often restricted to the Second Terrace at 30-40 m a.s.l. The Third Terrace at 50-60 m a.s.l. was considered as the Zyryanka glaciofluvial plain of MIS 4, and the First Terrace at 20-25 m a.s.l. was thought to have formed during MIS 2 (Zubakov, 1972; Lavrushin, 1963).

Recently obtained AMS, OSL, and U/Th dates have demonstrated that the Karginsky warming in key sections is older than 80 ka BP. Consequently, they correlate not with MIS 3, but with MIS 5 (probably, 5e) (Arslanov et al., 2004; Astakhov et al., 2005; Astakhov & Mangerud, 2005). Therefore, it was necessary to obtain updated chronometry of the fluvial strata in the Turukhansk Yenissei key sections, where they clearly overlie tills and glaciolacustrine rhythmites of the last (Zyryanka) glaciation.

Methods

In 2003, we resampled the key exposures and obtained 25 AMS-radiocarbon dates from T. Goslar at the Poznan Laboratory, Poland (table). Because 10 mg of organic material is sufficient for an AMS dating, we preferred to date only delicate plant remains, such as leaves, seeds, and moss tissue, which are least likely to survive redeposition. These tiny fossils were selected under a binocular microscope. We consider the resultant AMS dates to be more accurate than earlier published conventional dates measured on bulk samples (Arkhipov, 1990; Zubakov, 1972). The calibration of radiocarbon dates is still controversial for the period discussed in this paper, although one should probably add 3-5 ka to the ^{14}C age in order to record `calendar years` (van der Plicht, 2004) which could be compared with OSL values.



Sections of terraces on the Peri-Polar stretch of the Yenissei River with radiocarbon and luminescence dates, ka. (1) floodplain silt; (2) glaciolacustrine clayey rhythmites; (3) till; (4) subaerial sand, in places with peat seams; (5) cross-bedded channel sand; (6) laminated sand with silt seams or moss mats; (7) cryoturbation; (8) peaty bed; (9) loess-like silt.

We also present 35 dates obtained by measuring optically stimulated luminescence (OSL) of quartz grains (Murray & Wintle, 2000) at the Nordic Laboratory of Luminescence Dating in Risø, Denmark, under the guidance of A. Murray (table). The OSL dates are here cited with an error given as one standard deviation of the laboratory counting procedure in the same manner as for ^{14}C dates. However, other uncertainties of OSL dating, such as incomplete bleaching, fluctuating water content, migration of nuclides and instability of the signal lead to additional errors, which often are larger than for the radiocarbon analysis. For example, our OSL ages have been calculated as though the sediments were continuously water-saturated since deposition. In our opinion, this is a reasonable hypothesis, although it may lead to an overestimation of the age. If we extrapolate the low water content of the sampling moment to the entire postdepositional period, then the calculated ages would have been up to 15% lower. Due to all such uncertainties we consider only series of OSL dates as meaningful. The reliability of the ages is further verified by the internal consistency of each series and by comparison with ^{14}C series.

Distribution of dates

Intercalated fluvial and lacustrine formations have been repeatedly described from the Bolshoi Shar bluff (66° N) exposing the Third Terrace at 50-60 m a.s.l. (Lavrushin, 1963; Zubakov, 1972; Kind, 1974). At 2 km from the upstream end of the 6 km long bluff, we discovered a thus-far unknown till with cobbles of Putorana dolerites and basalts, sandwiched between the underlying glaciolacustrine clayey rhythmite and the overlying gravelly sand. This cross-bedded sand is dark brownish due to an abundance of mafic clasts. This unit is interpreted as a sandur deposited by the same Putorana ice cap that produced the mentioned till (figure).

There are two sets of dates from the Bolshoi Shar section. The right-hand series in the figure is taken at 2 km downstream from the southern end of the bluff above the till bed, whereas the left-hand series is taken 2.8 km farther downstream. The sand, up to 30 m thick, fills a 600-m-wide channel there and penetrates the entire glaciolacustrine unit. The date of 48 ka is obtained on a sample from the aeolian (or lacustrine) mantle of the terrace. The unweighted mean of the seven remaining OSL dates from the sandur is 56 ± 5 ka BP. Eliminating the two youngest dates, we get the nucleus of the five most consistent OSL dates with a mean value of 62 ka.

We have also obtained OSL dates from two sites of the Second Terrace: (i) at Konoshchelye (66°17.4′ N) 29 km upstream from the settlement of Kureika and (ii) at the Poloi upstream section (66°42.2′ N).

In both cases, one can see a till overlain by a glaciolacustrine rhythmite at the base of the terrace (figure). Higher up, a cross-bedded sand is overlain by very striking sediments: light fine parallel- or ripple-laminated sand with numerous moss mats. At Konoshchelye, the sand locally fills 1- to 3-m-deep troughs with dense moss mats along the banks and purer sand in deeper parts. This sand with forest pollen was interpreted as a signature of a warm climate of the Karginsky Stage (Kind, 1974; Zubakov, 1972). A ^{14}C date of >36 ka BP was reported by Zubakov who related it to the Igarka interstadial (Zubakov, 1972). A finite ^{14}C date of 32 ka obtained by Kind from the cryoturbated layer in the Konoshchelye section (figure) was used for dating the `Konoshchelye cooling` (Kind, 1974). However, we obtained a non-finite ^{14}C date (>48 ka) and an OSL date of 47 ka from this layer and a ^{14}C date of >49 ka just above this layer (figure).

We sampled only one exposure of the First Terrace, the Poloi downstream section (66°43.6′ N). The upper part of the red-yellow sequence appeared to be aeolian sand and the underlying light alluvial sand yielded dates not different from the Second Terrace (figure).

Discussion

The OSL ages from the Bolshoi Shar sandur (table and figure) suggest that this site was last glaciated from the Putorana Plateau during MIS 4. This conclusion contradicts our former suggestion that the last Putorana glacier reached the Yenissei as early as 90 ka BP (Svendsen et al., 2004). The latter age should better be attributed to the underlying glaciolacustrine rhythmite formed during the damming of the Yenisei River by a glacier that advanced from the north. The Boshoi Shar sandur was formed by an already free northbound drainage during the advance of the second Late Pleistocene ice from the uplands. It is clear that the Third Terrace here is of glaciofluvial origin, as was suggested by Lavrushin (1963) and Zubakov (1972). Only Kind (1974) considered the sand as Karginsky alluvium. Analogous sand above the till and varved rhythmites in a bluff 1.5 km upstream from the settlement of Goroshikha (66°23.2′ N) has yielded similar OSL ages (mean value 68 ka) (Figure).

On the whole, our AMS and OSL dates from the Second Terrace at Konoshchelye and Poloi show older ages than the earlier published conventional radiocarbon dates. However, the conventional dates of 35, 40, 41 and 40 ka BP at Farkovo on the Turukhan River obtained from alluvium of the Second Terrace with washed plant detritus and forest pollen spectra (Kind, 1974) are very similar to our dates from the Poloi upstream section.

Dates: AMS radiocarbon (Roman print) and OSL (italics)

Field sample	Height above river, m	Age ^{14}C, ka BP	Lab no.	Field sample	Height above river, m	Age OSL, ka BP	Lab no.
Poloi downstream (First Terrace, 66°42.2′ N), *aeolian sand*							
				406	21.8	***27 ±2***	042576
				405	21.3	***32 ±2***	H52529
				404	20.4	***35 ±2***	H52528
				403	19.6	***54 ±4***	042575
Alluvial sand							
				401	17.6	***55 ±3***	042574
407	15	>49	Poz-4929	409	15	***49 ±3***	042577
Poloi upstream (Second Terrace, 65°43.6′ N), *aeolian sand with peat*							
391	30	10.7 ± 0.06	Poz-8307				
389	31.3	31.8 ± 0.5	Poz-8306	390	31	***44 ±3***	042567
388	30.8	26.7 ± 0.2	Poz-5086				
387a	30.2	36.2 ± 0.7	Poz-8304	386	30	***41 ±2***	H52527
387b	30.2	35.5 ± 0.7	Poz-8305	393	29.5	***46 ±3***	042568
383	29	38.4 ± 0.8	Poz-5085	385	28.8	***44 ±3***	042566
380a	27.4	39.2 ± 1	Poz-4924				
380b	27.4	39 ± 1	Poz-4925				
Underlying laminated alluivial sand with moss mats							
379	29.9	37.1 ± 0.8	Poz-4922				
378	29.4	37.7 ± 0.9	Poz-8301				
				377	29.2	***50 ±3***	042565
382a	27.9	42.5 ± 1.4	Poz-4926	381	27.8	***42 ±2***	H52526
382b	27.9	40.5 ± 1.1	Poz-4928	376	27.8	***47 ±2***	H52525
375	27.5	40.3 ± 1.2	Poz-8300				
384a	27	46 ± 2.5	Poz-8303				
384b	27	43.6 ± 1.8	Poz-8922				
373	25.5	42.2 ± 1.4	Poz-4921	374	25.8	***44 ±2***	H52524
371	23.6	47.4 ± 2.8	Poz-8299	372	24.5	***58 ±4***	042564
369	23.3	46 ± 2	Poz-4920				

368	23	46.5 ± 2.6	Poz-8297	367	22.6	***43 ±3***	H52523
				366	22	***53 ±3***	042563
Konoshchelye (Second Terrace, 66°17.4′ N), *alluvial sand*							
308	32.6	50 ± 7	Poz-5083				
307	32	35.6 ± 0.8	Poz-8294	304	32.5	***49 ±3***	H52520
302	29.5	>49	Poz-8293	306	31.5	***93 ±6***	042548
297	27	>48	Poz-8291	298	27	***47 ±4***	H52519
				295	24.2	***76 ±5***	042547
294	23.2	>52	Poz-5082	292	21.9	***76 ±4***	042546
291	21	>49	Poz-5081	289	21	***76 ±4***	042545
				288	15	***85 ±9***	042544
Goroshikha (Third Terrace, 66°23.2′ N), *sandur*							
				309	42	***63 ±4***	042551
				311	38	***83 ±6***	042550
				312	31	***57 ±4***	042549
Bolshoi Shar (ThirdTerrace, 66° N), *sandur*							
				275	37	***53 ±4***	H52513
				274	36.2	***44 ±7***	042540
				273	35.5	***42 ±5***	H52512
				251	32.5	***48 ±4***	H52509
				250	31	***64 ±5***	042531
				249	30	***66 ±4***	042530
				248	28.5	***59 ±4***	H52508
				247	25	***66 ±5***	H52507

The basal dark brownish cross-bedded sand at Konoshchelye may be contemporaneous to the sandur at Bolshoi Shar and Goroshikha. It might be even older, judging by the OSL series of 85 to 76 ka at Konoshchelye (figure). There is a persistent geochronometric difference between similar sand units with moss mats at Konoshchelye and Poloi. At Poloi, the light sand with moss mats yielded an exceptionally consistent series of old finite ^{14}C dates ranging from 47 ka in the lower part to 40 ka in the upper. This series of decreasing ages continues higher above the alluvial sand. Certainly, a single radiocarbon age close to the limit of the method should be regarded with skepticism. However, the stable series of 18 dates (table) in our case strongly suggests their reality. The ^{14}C dates are also supported by a number of parallel OSL ages in the same range.

At the Poloi upstream section, the alluvial sand is overlain by aeolian sand with two levels of large ice wedge casts (figure). The dates around the oldest ice wedge are crucial for determining the moment of the final alluvial deposition. The wedge cast filled with aeolian sand (^{14}C dates 38, 39, 39 ka; OSL dates 41, 44, 46 ka) cuts into alluvial mossy sand with ^{14}C dates of 44, 42, 41 ka and an OSL date of 42 ka. Thus, alluvial deposition ceased very close to 40 radiocarbon ka BP or to 44 luminescence ka BP. The aeolian sand is overlain by a peat bed with sand interlayers. The lowermost peat layer yielded ^{14}C ages of 35 and 36 ka, whereas the younger peat yielded 27 and 32 ka BP, supporting the interpretation presented above. The infill of the uppermost ice wedge cast, which pierces all three peats, has yielded a ^{14}C date of 10.7 ka BP indicating the early Holocene degradation of the permafrost.

At Konoshchelye, non-finite ^{14}C ages are accompanied by OSL ages higher than at Poloi (table), indicating that this characteristic sand is really older at Konoshchelye. The absence of driftwood tree trunks and the predominance of moss and delicate shrub twigs in this alluvium of the river flowing from the south suggest that the climate, which was milder than immediately before and after the interstadial of early MIS, was more severe during this interstadial than today in the subzone of northern taiga. The large ice wedge casts at Poloi are located approximately 350 km south of the present-day limit for active ice wedges in mineral soils, suggesting drier and possibly colder winters than today.

Surprisingly, the OSL dates from the First Terrace of the Poloi downstream site just above the till indicate synchronicity of this alluvium with that of the Second Terrace. In other words, the First Terrace is an erosional feature at this site. The uppermost radiocarbon dates of the Poloi upstream section (39 to 27 ka), all collected from aeolian sand, suggest that the river level fell rapidly from 35 to approximately 20 m a.s.l. after formation of the Second Terrace. Taking also into consideration the old age of the main part of the Konoshchelye sequence, we infer that the Yenisei terraces mostly have high basements and their sediments are not necessarily genetically and chronologically related to the formation of the terrace staircase.

Geochronological conclusions

The sand of the Third Yenisei Terrace at Bolshoi Shar and Goroshikha was evidently formed by meltwater streams from a Putorana ice cap during MIS 4, some 75 to 60 calendar ka BP. This formation can also be traced as lithologically similar dark cross-bedded sand at the base of the alluvial mantle of the Second Terrace. The overlying light-colored alluvial sands of the Second Terrace with moss mats and forest pollen at the bottom and cryoturbations at the top are approximately 56 to 45 luminescence or 47 to 40 radiocarbon ka old at Poloi, but still older at Konoshchelye. Previously, Kind (1974) assigned them to the Middle Karginsky warming 42 to 35 ka BP. Zubakov (1972) assigned them to the Igarka thermochron, which presumably lasted from the beginning MIS 3 to 34 ka BP. Although the new dates show older ages than those estimated by Kind and Zubakov, the containing strata are still younger than the stratotypic Karginsky alluvium (paper 13, Ch. III). The composite sequence is topped by aeolian sand and silt with cryoturbations and peat layers formed slightly later than 45 luminescence or 40 radiocarbon ka BP but before MIS 2.

Acknowledgments

The discussed data have been obtained in the course of the Russian-Norwegian research project PECHORA II (Paleo Environment and Climate History of the Russian Arctic) supported by the Research Council of Norway. The authors are also very grateful to A. Matyushkov (VSEGEI) and O. Kazansky (Igarka Geocryological Laboratory) for valuable assistance in the field research and to H. H. Birks (University of Bergen) who selected microscopic fossils for AMS dating.

16. Correlation of Upper Pleistocene Sediments in Northern West Siberia

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Quaternary Science Reviews, 2010, Vol. 29, p. 3615–3629.

Introduction

The northern West Siberian Plain covering approximately 2 million km^2 is part of a huge sedimentary basin with economic and scientific importance. This basin contains a thick sequence of Quaternary formations and is a key area for stratigraphic correlations across northern Eurasia. Only the upper part of the Pleistocene sequence relating to the chronological interval of marine isotope stages (MIS) 6 to 2 is exposed in large coastal and riverine bluffs. The rest of the Quaternary cover down to 200-300 m below sea level is understood from rare boreholes.

Due to poor accessibility of this flat, perennially frozen area, the late Quaternary stratigraphy and paleogeography has been disputed for a long time (e.g. Sachs, 1953; Troitsky, 1966, 1967; Lazukov, 1970; Zubakov, 1972, 1974; Kind, 1974; Arkhipov, 1990, 1998; Astakhov, 1998). Nevertheless, the Upper and Middle Pleistocene formations studied along the lower course of the River Yenissei since the 1940s have served as stratotypes of regional stratigraphic subdivisions for Siberia (Sachs, 1953; Kind, 1974).

Some problems concerning the last glaciation were resolved in the 1990s using previously collected Russian data (paper 10, Ch. III). However, controversies remain regarding the correlation between disparate sections in West Siberia and established western European Quaternary stratigraphy because of the lack of suitable chronologic control. Collaborative research of Russian, Scandinavian, German and American scientists in the past 15 years has yielded new understanding Quaternary environments in Siberia. The QUEEN umbrella program (Quaternary Environments of the Eurasian North) supported by European Science Foundation summarized important results obtained from various arctic areas (Svendsen et al., 2004), which, however, only marginally concerned West Siberia. Crucial dating results on the West Siberian Pleistocene have been obtained by American-Swedish expeditions (Forman et al., 2002) and then by the Russian-Norwegian PECHORA and ICEHUS projects. The data by the two latter projects are partially published in Russian periodicals but have not yet been made available to the wider international community.

This paper summarizes chronologic control for disparate sections across northern West Siberia and provides criteria for regional correlation and relation to Quaternary events of West Europe. The foremost task is to establish firmly the position of the main climatic extremes of the time interval MIS 6 to 2 within the local sedimentary record. This paper will demonstrate that sedimentary sequences from West Siberia that reflect temperate conditions and a high sea level are correlative to MIS 5, whereas the decidedly coldest interval with a low sea level is correlative to MIS 2. This analysis demonstrates that the regional stratigraphic record can now be reliably compared to other late Quaternary records in the Northern Hemisphere.

Conventional Siberian stratigraphy

The stratigraphy of the tundra along the southwestern shore of the Kara Sea and in lower reaches of Ob, Pur, Taz and Yenissei rivers has received limited scientific attention and has been studied only in disparate sites. The key stratigraphic areas that have been discussed persistently in the literature are the lower courses of Ob and Yenissei rivers (Sachs, 1953; Troitsky, 1966, 1967; Lazukov, 1970; Zubakov, 1972; Kind, 1974). Rare sections have been studied on the shores of the Yamal Peninsula (Vasilchuk, 1992; Forman et al., 2002). The rest of the flatlands are known mainly from photogeology and occasional boreholes. Regional correlation has been difficult because the lithostratigraphy is largely restricted to repeated successions of diamictons with scarce pebbles, silty rhythmites and sorted sands. The principal correlation tools for several decades have been regional stratigraphic schemes adopted by special Quaternary congresses and approved by the National Stratigraphic Committee (e.g. Arkhipov, 1990; Volkova & Babushkin, 2000).

The history of prior research starts in the 1940s when Sachs (1953) put forward his famous system of regional `horizons` derived from the arctic Yenissei River area and which was later extrapolated across all Siberia. Sachs's scheme described from Ust-Port sections between Cape Karginsky and river Malaya Heta (Fig. 1) was as follows:

The Late Quaternary Interglacial Series

1) Messo Horizon of fluvial sands, named after Messo-Yaha river;
2) Sanchugovka Horizon based on a very thick formation of diamictic and stratified clays with a poor arctic mollusk fauna with predominance of *Portlandia arctica* (named after Sanchugovka creek);
3) Kazantsevo Horizon derived from marine sands with rich boreal fauna, especially *Arctica islandica* (named after Kazantseva river and Kazantsevo village).

The Late Quaternary Glacial Series

4) Zyryanka Horizon of tills and outwash, deposited by mountain glaciers which spread onto the adjacent lowlands (named after Zyryanka creek);
5) Karginsky Horizon based on a late glacial terrace 20-30 m a.s.l. consisting of marine sediments with arctoboreal mollusk fauna, changing upstream into deltaic and alluvial formations (named after Cape Karginsky).
6) Sartan Horizon based on the final ice advance which left small morainic loops in Central Siberian mountains.

The two units reflecting temperate environments, the Kazantsevo and Karginsky `horizons`, were associated with two marine incursions of the Arctic Ocean that punctuated the Late Pleistocene history of the region. According to Sachs's descriptions (1953) the Kazantsevo strata contained the warm-water mollusk fauna with *Cyprina (Arctica) islandica* and the extinct species *Cyrtodaria jenisseae Sachs* (=*Cyrtodaria angusta Nyst et Westendorp* according to Merklin et al., 1979). The last, less extensive Karginsky sea, was associated with arctoboreal fauna, again indicating warmer water in contrast to the modern Kara Sea.

Subsequently, Sachs's subdivisions were used in regional stratigraphic schemes as climatostratigraphic horizons. These according to the Stratigraphic Code of Russia are kryomers or thermomers associated with either cold or temperate conditions, respectively (e.g. Zubakov & Borzenkova, 1990; Astakhov, 2006b). They are regional correlation units whose stratigraphic volume is similar to the chronostratigraphic stages and substages in the western European literature. On the other hand, the boundaries of the climatostratigraphic horizons are isochronous only within the given region. Such regional correlation horizons are known in the American Stratigraphic Code as geologic-climate units (ACSN, 1961).

The formal West Siberian stratigraphic schemes have repeatedly been modified since 1960 by special congresses. The main difference from

Sachs's scheme is that the lower boundary of the Upper Pleistocene was later placed at the base of the Kazantsevo strata.

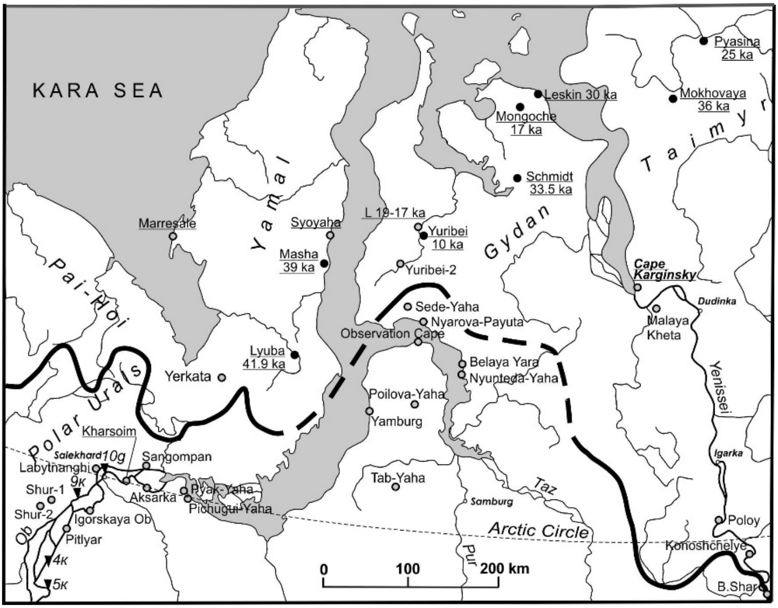


Fig. 1. Locations of principal Upper Pleistocene sites in northern West Siberia (see Fig. 5). Thick line is the early Weichselian ice limit (Astakhov, 2006a). Shaded dots are geochronometrically important sites with: i) names underlined – from literary sources (Kind, 1974; Vasilchuk, 1992; Forman et al., 2002), ii) names not underlined – authors' sections. Black triangles are boreholes in Fig. 5. L are dated mammoth bones from loess-like silts (Bolikhovsky, 1987). Black dots are well-preserved mammoth remains not covered by diamicton with their radiocarbon dates ka BP: mammoth carcass by F. Schmidt – 33.5 ± 1 ka BP, mammoth carcass on river Mokhovaya – 35.8 ± 1.2 ka BP, mammoth on river Pyasina – 25.1 ± 0.5 ka BP, Yuribei mammoth – 10 ka BP (Sulerzhitsky, 1995); mammoth leg by F. Kaplyanskaya at Cape Leskin – 30.1 ± 0.3 ka BP (paper 10 in Ch. III); baby mammoth Masha – 39.1 ka BP (Tomirdiaro & Tikhonov, 1999), baby mammoth Lyuba – 41.9 ka BP (Kosintsev, 2008), Mongoche mammoth – 17 ka BP (Gilbert et al., 2007).

Troitsky (1966), who analysed the mollusk fauna, came to the conclusion that both marine formations belonged to one boreal transgression, a view not shared by other paleontologists. Most importantly, he identified glacial diamictons and varved clays atop of the Karginsky terrace which was the basis for splitting the Zyryanka glaciation into two lowland ice advances: the early pre-Karginskian and the late Gydan stage correlative with the Late Weichselian (Troitsky, 1967). The latter idea was adopted by Kind (1974) who named the Late Zyryanka ice advance the `Sartan glaciation` borrowing the name from the Sachs's alpine late glacial stage. She described the underlying Karginsky strata as fully interglacial on the basis of their mollusk fauna, pollen spectra and diatoms. Based on finite conventional radiocarbon dates from beneath the Late Zyryanka glacial formation, she correlated the Karginsky interglacial with the early European Pleniglacial and MIS 3.

After this work, the Karginsky and Sartan horizons in all Siberian stratigraphic schemes were correlated to MIS 3 and 2, respectively (Arkhipov, 1990; Volkova & Babushkin, 2000). Accordingly, the preceding Kazantsevo thermomer was customarily placed at the MIS 5 level (Table 1).

Table 1. Conventional correlation units of the Siberian Upper Pleistocene (Arkhipov, 1989)

Siberian regional scale	**Chronometric ages, ka**
Upper Zyryanka=Sartan glacial (MIS 2)	^{14}C=10 – 22(23)
Middle Zyryanka=Karginsky interglacial (MIS 3)	^{14}C=31–>45; ESR= 51.6 ± 12.8
Lower Zyryanka=Yermakovo glacial (MIS 4)	TL=75 ± 11; 80 ± 11(13); 100 ± 17; 110 ± 27
Kazantsevo interglacial (MIS 5e)	ESR=121.9; 134.8 optimum TL=130 ± 25
Taz glacial (MIS 6)	

Another cornerstone of the Siberian stratigraphic framework is distinct foraminifer assemblages. Micropaleontologists have established two assemblages with boreal species called the Kazantsevo and Karginsky assemblages. An unexpected conclusion from this work was that the Karginsky assemblage contained more warm-water species, including even some lusitanic ones (Levchuk, 1984; Gusskov & Levchuk, 1999).

A peculiar feature of the Siberian stratigraphic model is that the two last horizons of the Upper Pleistocene have been accepted without stratotypes (Arkhipov, 1990). The reason for that is because the type section of the marine formation at the Cape Karginsky yielded an ESR date of 121.9 ka BP which obviously contradicted the radiocarbon correlation with MIS 3 suggested by Kind (1974). The Sartan name was derived from alpine moraines of the Verkhoyansk mountain range which have never been geochronometrically dated and their correlation with the uppermost glacial complex in the West Siberian Plain is dubious.

Such usage, unrelated to the original stratotypes, made the Siberian stratigraphic subdivisions equivalents of the radiocarbon intervals that do not need any regional names. The lack of stratotypes for interglacials and of associated chronologic control confounded correlations and climatic interpretations beyond the immediate study area.

The correlation difficulties outlined above led to the arbitrary application of the Siberian stratigraphic nomenclature to disparate sedimentary bodies located at significant distance from the area of its origin. Even the main stratigraphic marker – the `warmest` horizon of the Upper Pleistocene presumably correlative to MIS 5 – has never been reliably traced across Siberia. Thus, in the Gydan Peninsula some authors identify interglacial marine strata as the surficial sands that cover interfluves 50-60 m a.s.l., whereas others find interglacial sand beneath thick diamictons. In the Yamal Peninsula some scientists relate the thick deltaic sequence of the Marresale Formation to the lowermost Upper Pleistocene, whereas others argue that the first thermomer of the Upper Pleistocene is represented by marine sand with the `Kazantsevo foraminifers` (Levchuk, 1984; Gusskov & Levchuk, 1999).

The name `Kazantsevo` can also be found in descriptions of various parts of the Siberian non-glacial Quaternary. These could be surficial sands without any organics, or peat layers with forest flora, or even marine strata of the younger Karginsky transgression, provided non-finite radiocarbon dates were obtained. Such arbitrary usage was encouraged by the formal stratigraphic schemes which firmly associated the Kazantsevo name with the main thermomer of the Upper Pleistocene.

This problem with the Last Interglacial led to uncertainty in assessing the extent of the uppermost glacial sedimentary sequence. The last glacier in the form of piedmont ice sheet crossed the Ob River valley near the Arctic Circle in the beginning of the Late Pleistocene (Lazukov, 1970) to leave sandy moraines near Salehard (Fig. 1). For the last glaciation this author borrowed the name Zyryanka used by Sachs (1953) on the Lower Yenissei River. In his correlation, Lazukov (1970) relied on lithostratigraphy and the geomorphic position of different sedimentary formations. Zubakov (1972) used mostly pollen diagrams and very rare radiocarbon dates. According to him, Late Pleistocene glaciers were confined to the Uralian piedmont and never reached the Ob River valley, whereas all sediments to the east of this valley belong to the Middle

Pleistocene. On the contrary, Arkhipov et al. (1977), who stressed the correlation value of foraminifers and ^{14}C dates, believed that during MIS 2 ice sheets were centered on the Polar Urals and expanded from there onto the West Siberian Plain beyond the Arctic Circle. The regional stratigraphic schemes of West Siberia adopted the Arkhipov's view (Arkhipov, 1990).

The most recent lowland glaciation was labeled Sartan and correlated with MIS 2 following Kind (1974). According to Arkhipov (1998) this glaciation, although less extensive than ice sheets of MIS 4, still covered all arctic West Siberia including the Salekhard Hills, Gydan, Taz and Yamal peninsulas. The key point of this maximalist model is that it was associated with the huge ice-dammed Mansi Lake which had a level up to +130 m and which spilled southwards via the Turgai Pass. Arkhipov (1998) saw a trace of this freshwater inner sea in the Urtam mantle of loess-like silts covering all terraces above the floodplain.

However, radiocarbon statistics showed that the most reliable conventional dates, including those in the 30-40 ka BP range derived from frozen mammoth carcasses, were clearly positioned above the uppermost morainic complex with blocks of fossil glacial ice. Surficial ice-bound loess-like silts yielded a long series of radiocarbon dates starting from 37-40 ka BP. Therefore, the last glaciation of northern West Siberia, whatever its name, must predate this series (paper 10, Ch. III). The central lowland of periglacial West Siberia, instead of deep-water Mansi Lake sediments, displays loess-like silts with ice-wedge casts and paleosols around the age of the LGM (paper 14, Ch. III).

Siberian Quaternary literature often produces an impression that the regional stratigraphic scale of the Upper Pleistocene is broadly similar to that of Western Europe and that minor differences between the schemes can be rectified by slight changes in local boundaries and terminology. In reality the difference between the two schemes is fundamental. It can be illustrated by comparison of the climatic chronology of northern Western Europe (Fig. 2A) with the curve of climatic fluctuations in West Siberia (Fig. 2B). In Western Europe paleoclimate records demonstrate a steady cooling up to the final Pleistocene. Minor interstadial reversals of this pre-Holocene trend did not result in restoration of the modern natural zonality. From 74 to 11.5 ka BP, tundra-steppe or at best forest-tundra types dominated. In Siberia, on the contrary, a return to warmer climate is suggested during the last glacial cycle and associated with sea level rise, but has no analogue in other parts of northern Eurasia.

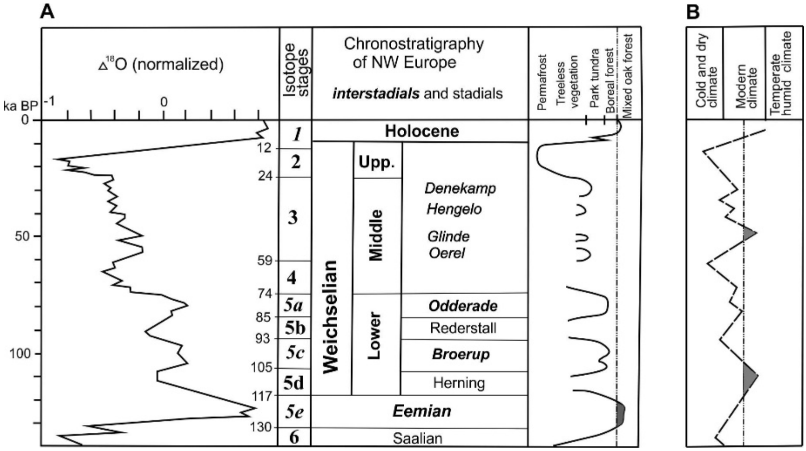


Fig. 2. Comparison of Late Pleistocene climatic chronologies of northern Western Europe (A – by Mangerud, 1989) and West Siberia (B – by Arkhipov & Volkova, 1994).

Sometimes the Karginsky warming in Siberian literature is called `megainterstadial` (Arkhipov & Volkova, 1994). This does not change the essence of the Siberian model, namely of a predominantly forested landscape during this interval. Indeed, according to microfauna analyses the Karginsky transgression was the warmest within the Pleistocene (Gusskov & Levchuk, 1999). The Siberian curve (Fig. 2B) features two interglacial/glacial cycles and not one as in Europe (Fig. 2A). Zubakov and Borzenkova (1990) think that this must be simply a chronometric blunder and that a revision of this discrepancy is long overdue.

Methods

In the course of the Russian-Norwegian projects of 2000–2005 we studied natural sections along the subarctic Ob River, in the southern Yamal, Taz and Gydan peninsulas and on the Lower Yenissei River (Fig. 1). Unlike the studies of the 1950–1970s we used lithostratigraphic criteria only for short-distance correlations and for assessment of reliability of paleontological and geochronometric data. The latter two methods proved to be most instrumental in determination of main correlation levels.

The main thermomer of the Upper Pleistocene is defined as embracing sediments with maximum organic content supported by geochronometric values close to the astronomically dated marine isotope substage 5e, ~130 ka. Pollen spectra, plant macrofossils, mammal remains, marine mollusks and other paleontological indicators are used for correlation if plentiful and diagnostic of clearly defined paleoenvironments. We used microfossil data only from strata with rich organics, clearly positioned in the local stratigraphic succession, avoiding reworked sediments with numerous clasts.

Our attempts to find and date previously described poor foraminifer assemblages from sediments with megaclasts (e.g. Arkhipov et al., 1977) have been unsuccessful. Repeated sampling in such cases did not yield foraminifers, and dating by modern techniques produced much older ages (Forman et al., 2002; Astakhov et al., 2007). Often microfossils are not primary, but reflect reworking from older deposits. We believe that randomly obtained and non-reproducible analyses of foraminifers and pollen from diamictic sequences abundant in the Siberian literature (e.g. Lazukov, 1970; Arkhipov et al., 1977; Levchuk, 1984) should be questioned because of ambiguity of their stratigraphic position.

The correlation of disparate sections has been facilitated by radiocarbon, luminescence and U/Th ages. It is well known that in the Siberian north the conventional method of measuring ^{14}C in large volumes of organics often yields unreliable results in the range beyond 30 ka BP. Many authors noted obviously too young ages obtained not only on mollusk shells but also on driftwood and mixed assemblages of plant detritus (e.g., Zubakov, 1974; Astakhov, 1998; Mangerud et al., 2002; Astakhov & Mangerud, 2005, 2007). Sulerzhitsky (1998) relates this contamination by young carbon to the reproductive outbreak of dormant autotrophic bacteria immediately after fossil organics melt-out of permafrost.

Whatever the reason, the radiocarbon dating method is more sensitive to contamination of samples by young than by old carbon (Zubakov, 1974; Mangerud et al., 2002). The AMS dating technique using carefully selected microscopic volumes of organic materials is more reliable. AMS method often yields much older ages than conventional radiocarbon dating which are independently confirmed by parallel luminescence dates. Especially important, and even crucial for correlation, is that AMS ages of small twigs, moss stems, vascular plants are often non-finite whereas conventional dates on driftwood or randomly assembled large volumes of plant detritus gave finite values from the same formations (e.g. papers 13 and 15 in Ch. III). Unfortunately AMS techniques cannot provide a correlation tool for sediments older than 50 ka.

Lately the uranium-thorium method has demonstrated a possibility for dating sediments 100–300 ka old. However, this technique demands thick peat layers (which are extremely rare in the Siberian Pleistocene) and their isolation by clayey beds from migrating younger uranium (Arslanov et al., 2004). Because of this limitation we have only two peat deposits in Siberia dated by the U/Th method. Constraints on occurrence and quality of mollusk shells, which are normally too thin and scarce in the Siberian Arctic, limit use of the ESR method as well.

Therefore, the only practical option for far-distance correlation of sediments beyond the range of radiocarbon method is luminescence dating. In its optical stimulation form (OSL) it is the most universal method applicable to various sand deposits, although a summary error in age estimates may sometimes reach 20%. The accuracy of OSL dating depends mainly on solar resetting of the original sand grains, estimates of environmental radiation rate, on water content estimates, etc. In general, the dating range of OSL technique is about 150–200 ka. After many experiments with TL, OSL on feldspar grains and other techniques optically stimulated luminescence of quartz grains measured using the single aliquot regenerative dose (SAR) protocol was chosen as the best (Murray & Olley, 2002).

The applicability of this method was statistically tested in northern European Russia on Paleolithic sites with numerous radiocarbon dates (Mangerud et al., 2002) and on a type section of Boreal marine strata, a local analogue of the Eemian formation (Murray et al., 2007). In the latter case 16 samples gave mean age value 112 ka, i.e. 10% less than the middle of astronomically estimated Eemian age interval. The offset is smaller for Weichselian sediments. Outliers (exceeding 2 standard deviations) which sometimes occur in series of 5-10 OSL ages are best eliminated before calculating the mean OSL age of the formation.

An exemplary test for the OSL method in Siberia is provided by the stratigraphy of postglacial fluvial terraces of the Yenissei River stretch crossing the Arctic Circle (Fig. 3) (paper 15, Ch. III). The Poloi upstream section of the second alluvial terrace provided a series of 18 AMS dates within the span of 47–26 ka BP. Small shrub twigs in bark, buried almost *in situ,* were dated. The dates may be grouped in two clusters with mean values 47 ka in basal beds and 36.8 ka BP in the upper beds. The parallel series of OSL dates also gave two clusters with mean values of 48.1 and 43.7 ka BP correspondingly. The difference between ^{14}C and OSL values is not unexpected because the luminescence signal, being closer to astronomic age, on this level should theoretically exceed radiocarbon ages by 4-5 ka.

Our chronometric database contains 121 OSL and 59 ^{14}C uncalibrated dates (Tables 2 and 3) including those published by us earlier (mostly in the Russian literature). Previously published OSL dates: 1–4, 86–93 (Astakhov, 2006a; Astakhov et al., 2007), 5–7, 19–24, 43–58, 63–68, 103–111 (Nazarov, 2007), 8–18, 25–31, 69–85 (paper 15, Ch. III), 37–42 (paper 13, Ch. III), 97–102 (paper 12, Ch. III), 112–115 (Mangerud et al., 2004). Previously published ^{14}C dates: 1–4, 22–27, 50–54 (Astakhov, 2006a; Astakhov et al., 2007), 5–10, 30–49 (paper 15, Ch. III), 11–13 (paper 13, Ch. III), 14–21, 28, 29 (Nazarov, 2007), 55–59 (paper 12, Ch. III).

Italicised are OSL ages recently corrected by A. Murray against the published dates. The corrections are made mostly for possible influence of additional cosmic radiation connected with varying depths of burial. For this discussion we also use screened dates by other authors (mostly AMS) from Yamal (Vasilchuk et al., 2000; Forman et al., 2002) and radiocarbon dates on undisputably *in situ* remains of mammoths, mostly frozen carcasses or well-preserved bones (black dots in Fig. 1).

Eastern sections

The original stratotypes by Sachs (1953) on the River Yenissei were partly dated in the 1960s by the conventional radiocarbon method (Kind, 1974) except sections of marine sands with boreal mollusks close to Kazantsevo village. The Cape Karginsky marine formation with arctoboreal fauna and plant macrofossils from a forested environment yielded several old finite dates (Fig. 4A) and was correlated with MIS 3 (Kind, 1974). More recently, the ESR date of 121.9 ka BP was obtained on a shell from this site, and as a result the Cape Karginsky section was deleted from stratotypes of the regional stratigraphic scheme (Arkhipov, 1990). Moreover, the marine formation was renamed the `Kazantsevo strata`, contrary to the stratigraphic codes of USSR and Russia, and also strongly disapproved by the International Stratigraphic Guide (Salvador, 1994). Nazarov and Henriksen (2010) have since obtained 6 OSL dates with mean value of 111 ka BP from this marine formation. It is clear that its paleontological and chronometric characteristics are best suited for environments of the Last Interglacial at MIS 5 level.

Table 2. Optically stimulated luminescence (OSL) dates for northern West Siberia by Russian-Norwegian projects. Measurements are by Nordic Laboratory for Luminescence Dating, Denmark.

No.	Locality	Sediment	Field sample no.	Lab. no. Risø	w.c.%	Paleodose Gy	Dose rate Gy/ka	Age, ka BP
1	Aksarka	Alluvial sand	00-0500	002549	26	168±16	1.94±0.11	84±10
2	Aksarka	Alluvial sand	00-0503	002550	27	184±9	1.90±0.11	97±8
3	Aksarka	Loess-like silt	00-0504	012547	29	42.4±0.6	2.13±0.08	19.8±0.9
4	Aksarka	Loess-like silt	00-0505	012548	29	29.4±0.4	1.70±0.07	17.3±0.8
5	Belaya Yara	Marine sand, moss mat	04-2525	052549	28	285±13	2.07±0.09	137±9
6	Belaya Yara	Same	04-2526	052550	27	214±15	1.78±0.07	120±10
7	Belaya Yara	Same	04-2530	052551	27	298±9	1.79±0.07	166±9
8	Bolshoi Shar	Glaciofluvial sand	03-247	H52507	25	47±2	0.71±0.04	66±5
9	Same	Same	03-248	H52508	34	62±3	1.05±0.04	59±4
10	Same	Same	03-249	042530	28	58±3	0.88±0.04	66±4
11	Same	Same	03-250	042531	31	59±3	0.92±0.04	64±5
12	Same	Same	03-251	H52509	34	55±4	1.15±0.05	48±4
13	Same	Same	03-273	H52512	32	34±4	0.81±0.04	42±5
14	Bolshoi Shar	Same	03-274	042540	29	31±4	0.71±0.04	44±7
15	Bolshoi Shar	Same	03-275	H52513	25	43±2	0.80±0.04	53±4
16	Goroshikha	Glaciofluvial sand	03-309	042551	29	51±2	0.90±0.04	57±4
17	Goroshikha e	Glaciofluvial sand	03-311	042550	35	72±4	0.87±0.04	83±6

18	Same	Glaciofluvial sand	03-312	042549	41	75±3	1.19±0.06	63±4
19	Igorskaya Ob	Aeolian silt and sand	04-2507	052545	26	90±4	1.23±0.06	73±5
20	Same	Same	04-2508	052546	24	86±3	1.19±0.05	72±4
21	Same	Same	04-2509	052547	45	166±7	1.87±0.07	89±5
22	Same	Same	04-2510	052548	37	159±8	1.94±0.07	82±6
23	Same	Fluvial sand	04-2503	052543	32	178±7	1.53±0.06	116±7
24	Same	Fluvial sand	04-2504	052544	29	175±10	1.69±0.07	104±7
25	Konoshchelye	Alluvial sand	03-288	042544	26	40±2	0.47±0.04	85±9
26	Same	Same	03-289	042545	30	112±4	1.47±0.06	76±4
27	Same	Same	03-292	042546	47	122±3	1.61±0.07	76±4
28	Same	Same	03-295	042547	30	109±5	1.44±0.06	76±5
29	Same	Same	03-298	H52519	27	94±6	2.02±0.08	47±4
30	Same	Same	03-304	H52520	33	100±4	2.04±0.08	49±3
31	Same	Same	03-306	042548	29	102±3	1.1±0.05	93±6
32	Labytnanghi	Glacilacustrine. sand	01-0204	022522	30	255±14	1.78±0.07	143±10
33	Same	Glacilacustrine rhythmite	01-0205	012591	32	227±10	1.73±0.08	131±9
34	Same	Glacilacustrine rhythmite	01-0206	012592	35	174±5	1.13±0.05	154±9
35	Same	Glaciolacustrine sand	01-0202	012593	31	219±13	1.52±0.06	145±11
36	Same	Glaciolacustrine sand	01-0203	012594	42	188±4	1.37±0.06	137±8
37	Malaya Heta	Aeolian silt	03-212	042524	43	142±6	1.73±0.07	82±5
38	Same	Same	03-211	042523	35	148±4	1.88±0.08	79±4

39	Same	Same	03-209	042522	39	161±5	1.63±0.08	98±6
40	Same	Alluvial sand	03-206	042521	39	160±3	1.62±0.06	99±5
41	Same	Alluvial sand	03-205	042520	37	163±4	1.46±0.06	112±6
42	Same	Alluvial silt and clay	03-202	042519	39	131±7	1.64±0.06	80±5
43	Nyarova-Payuta	Alluvial sand	04-2574	052555	28	122±6	1.13±0.06	108±8
44	Same	Same	04-2575	052556	28	152±7	1.76±0.07	86±5
45	Same	Same	04-2580	052557	27	96±4	0.9±0.04	106±7
46	Same	Same	04-2583	052558	28	138±5	2±0.08	69±4
47	Nyunteda-Yaha	Marine sand	04-2559	052537	31	253±11	1.75±0.07	145±9
48	Nyunteda-Yaha	Same	04-2568	052538	29	251±11	1.82±0.08	138±9
49	Nyunteda-Yaha	Same	04-2569	052539	29	309±14	1.99±0.08	155±10
50	Nyunteda-Yaha	Same	04-2558	052554	33	262±11	1.82±0.07	144±9
51	Observat. Cape	Same	04-2597	052540	29	257±11	2.08±0.08	124±8
52	Observat. Cape	Same	04-2598	052541	30	257±9	2.03±0.08	127±7
53	Observat. Cape	Marine sand	04-2614	052542	31	278±10	2.11±0.25	132±17
54	Observat. Cape	Same	04-2591	052559	31	274±10	2.06±0.08	133±8
55	Observat. Cape	Same	04-2592	052560	34	278±11	1.98±0.07	141±8
56	Observat. Cape	Same	04-2613	052561	31	309±14	2.04±0.08	152±10
57	Observat. Cape	Fluvial sand	04-2625	052562	27	136±6	1.76±0.07	77±5
58	Observat. Cape	Fluvial sand	04-2626	052563	32	145±6	1.96±0.09	74±5
59	Pichugui-Yaha	Fluvial sand beneath rhythmite	01-0119	012579	27	179±8	0.91±0.06	197±15

60	Pichugui-Yaha	Fluvial sand beneath rhythmite	01-0120	012580	27	150±7	0.78±0.05	192±16
61	Pichugui-Yaha	Alluvial sand	01-0115	012581	32	221±14	1.77±0.08	125±10
62	Pichugui-Yaha	Same	01-0116	012582	32	227±9	1.66±0.08	137±9
63	Pitlyar	Same	01-0153	022512	22	102±3	1.55±0.08	66±4
64	Pitlyar	Same	01-0154	022513	21	104±4	1.68±0.07	62±4
65	Pitlyar	Same	01-0159	022514	20	196±5	1.82±0.08	108±6
66	Pitlyar	Same	01-0160	022515	26	151±3	1.98±0.09	76±4
67	*Pitlyar*	*Glaciolacust. sand*	*01-0161*	*022516*	*36*	*150±3*	*2.08±0.08*	*73±4*
68	Pitlyar	Glaciolacust. sand	01-0162	022517	35	149±6	1.99±0.07	75±4
69	Poloi downstream	Alluvial sand	03-401	042574	27	74±2	1.34±0.06	55±3
70	Same	Aeolian sand	03-403	042575	26	80±3	1.48±0.07	54±4
71	Same	Same	03-404	H52528	27	48.6±1.9	1.37±0.06	35±2
72	Same	Same	03-405	H52529	30	48.6±1.7	1.53±0.06	31.7±1.8
73	Same	Same	03-406	042576	24	44±2	1.66±0.07	27±2
74	Same	Alluvial sand	03-409	042577	28	77±3	1.57±0.06	49±3
75	Poloi Upstream	Alluvial sand with moss mats	03-366	042563	32	93±2	1.75±0.07	53±3
76	Same	Same	03-367	H52523	32	68±3	1.60±0.07	43±3
77	Same	Same	03-372	042564	30	81±4	1.40±0.06	58±4
78	Same	Same	03-374	H52524	31	80±3	1.81±0.07	44±2
79	Same	Same	03-376	H52525	34	68.2±1.2	1.47±0.06	47±2

80	Same	Same	03-377	042565	30	63±3	1.26±0.05	50±3
81	Same	Same	03-381	H52526	38	55.7±1.7	1.33±0.05	42±2
82	Same	Aeolian sand with peat lenses	03-385	042566	27	49±2	1.12±0.06	44±3
83	Same	Same	03-386	H52527	26	55±2	1.34±0.06	41±2
84	Same	Same	03-390	042567	34	56±2	1.28±0.05	44±3
85	Poloi Upstream	Same	03-393	042568	28	54±1	1.17±0.05	46±3
86	Pyak-Yaha	Alluvial sand	01-0133	012585	33	208±5	1.51±0.07	138±8
87	Pyak-Yaha	Same	01-0134	012586	30	211±14	1.59±0.08	133±11
88	Sangompan	Same	01-0196	012583	27	72±2	1.41±0.06	93±7
89	Sangompan	Same	01-0197	012584	31	131±4	1.61±0.07	82±5
90	*Sangompan*	*Alluvial sand*	*01-0190*	*022518*	*30*	*93±4*	*1.16±0.05*	*80±5*
91	*Sangompan*	*Alluvial sand*	*01-0191*	*022519*	*30*	*118±8*	*1.41±0.06*	*84±7*
92	Sangompan	Subaerial sand	01-0193	022520	22	53.8±1.1	2.60±0.12	20.7±1.1
93	Sangompan	Subaerial sand	01-0194	022521	28	48±2	1.65±0.06	29.2±1.8
94	Sede-Yaha	Marine sand	05-3037	062523	29	228±6	2.37±0.09	96±5
95	Same	Same	05-3039	062524	25	247±7	2.20±0.09	112±6
96	Same	Same	05-3048	062525	27	222±11	1.56±0.07	142±10
97	Shur-1	Sand under peat	01-0167	012587	26	161±4	1.97±0.08	82±4
98	Same	Sand under peat	01-0168	012588	27	187±5	1.88±0.09	100±6
99	Same	Sand lamina in clay	01-0172	012589	47	93±6	1.64±0.06	57±4
100	Same	Sand above peat	01-0173	012590	34	159±3	1.59±0.06	100±5
101	*Shur-2*	*Subaerial sand*	*01-0186*	*022510*	35	*99±2*	*1.31±0.06*	*75±4*

102	*Shur-2*	*Subaerial sand*	*01-0187*	*022511*	28	*103±2*	*1.64±0.08*	*63±4*
103	Tab-Yaha	Alluvial sand	03-1346	042591	23	42.3±1.1	1.75±0.17	24±2
104	Same	Same	03-1336	042592	25	20.5±0.8	0.66±0.04	31±2
105	Same	Same	03-1337	042593	23	30.6±0.6	0.82±0.05	37±3
106	Same	Glaciofluvial sand	03-1334	042594	25	115±4	0.44±0.04	260±23
107	Same	Glaciofluvial sand	03-1342	042595	24	212±10	0.7±0.04	305±24
108	Yamburg	Marine sand	03-1358	042587	23	170±6	1.02±0.05	167±10
109	Same	Same	03-1359	042588	25	195±6	1.2±0.05	162±9
110	Same	Same	03-1360	042589	26	247±11	1.67±0.08	148±10
111	Same	Same	03-1361	042590	26	213±6	1.72±0.07	124±7
112	Yerkata	Lacustrine sand	00-0414	012542	33	115±6	1.59±0.07	72±5
113	Same	Lacustrine sand	00-0413	012541	34	114±5	1.83±0.07	63±4
114	Same	Aeolian sand above lacust.sand	00-0416	012543	29	118±7	1.81±0.07	65±5
115	Same	Aeolian sand above lacust.sand	00-0418	012544	29	106±5	1.78±0.07	59±4
116	Yuribei-2	Glaciolacustrine sand above till	05-3077	062526	32	84±3	1.41±0.06	59±3
117	Same	Same	05-3078	062527	26	71±2	1.29±0.06	
118	Same	Same	05-3079	062528	31	89±3	1.52±0.06	
119	Same	Same	05-3083	062529	31	89±2	1.28±0.05	
120	Same	Same	05-3084	062530	29	89±2	1.29±0.05	
121	Yuribei-2	Same	05-3085	062531	28	120±4	1.83±0.07	

Table 3. Radiocarbon ages obtained by Russian-Norwegian projects in northern West Siberia. Prefixes for conventional date are LU – Geochronological Laboratory of St. Petersburg University, T – Trondheim Laboratory, Norway. AMS dates: Poz – Poznan Laboratory, Poland

No.	Locality	Sediment	Sample no.	Lab. no.	Dated material	Age, ka BP
1	Aksarka	Loess-like silt	00-502	LU-4555	Moss & branches	27.84±0.31
2	Same	Ice wedge cast	00-485	LU-4553	Peat	≥44.6
3	Same	Alluvial sand	00-486	LU-4580	Wood	≥46.3
4	Same	Soliflucted diamicton	00-494	T-15160	Wood	45±2.34
5	Konoshchelye	Alluvial sand	03-308	Poz-5083	Shrub twigs	50±7
6	Same	Same	03-307	Poz-8294	Shrub twigs	35,6±0.8
7	Same	Same	03-302	Poz-8293	Shrub twigs	≥49
8	Same	Same	03-297	Poz-8291	Peat	≥48
9	Same	Same	03-294	Poz-5082	Shrub twigs	≥52
10	Same	Same	03-291	Poz-5081	Shrub twigs	≥49
11	Malaya Heta	Same	03-214	Poz-5018	Branch	≥48
12	Same	Loess-like silt	03-210	Poz-5078	Branch	≥52
13	Same	Loess-like silt	03-217	Poz-5079	Moss mat	≥49
14	Nyunteda-Yaha	Pick up under bluff	04-2543	LU-5344	Mammoth tooth	≥44.8
15	Same	Same	04-2538	LU-5345	Mammoth tusk	44±1.8
16	Same	Same	04-2547	LU-5346	Mammoth bone	18.71±0.11
17	Nyunteda-Yaha	Pick up under bluff	04-2546	LU-5347	Rhinoceros bone	41.4±1.3

18	Same	Marine sand	04-2562	LU-5368	Moss mat	≥50.7
19	Same	Marine sand	04-2564	LU-5367	Moss mat	≥50.5
20	Observat. Cape	Fluvial sand	04-2609	LU-5363	Peat	≥47.5
21	Observat. Cape	Fluvial sand	04-2611	LU-5364	Peat	40.2±1.26
22	Pichugui-Yaha	Alluvial sand	01-118	LU-4767	Plant detritus	≥45.9
23	Same	Alluvial sand	01-117	LU-4769	Plant detritus	≥52.4
24	Same	Pick up under bluff	01-112	LU-4787	Mammoth bone	25.05±0.22
25	Same	Same	01-122a	LU-4783	Mammoth bone	15.48±0.07
26	Same	Same	01-122б	LU-4992	Muskox skull	16.38±0.05
27	Pitlyar	Glaciolacustrine rhythmite	01-165	LU-4766	Moss mat	≥45.6
28	Poilova-Yaha	Alluvial sand	03-1354	LU-5105	Moss mat	27.5±0.28
29	Poilova-Yaha	Alluvial sand	03-1355	LU-5104	Moss mat	28±0.26
30	Poloi downstream	Aeolian sand	03-407	Poz-4929	Plant detritus	≥49
31	Poloi upstream	Aeolian sand	03-391	Poz-8307	Peat	10.7±0.06
32	Same	Same	03-389	Poz-8306	Peat	31.8±0.5
33	Same	Same	03-388	Poz-5086	Shrub twigs	26.7±0.2
34	Same	Same	03-387a	Poz-8304	Peat	36.2±0.7
35	Same	Same	03-387b	Poz-8305	Peat	35.5±0.7
36	Same	Same	03-383	Poz-5085	Shrub twigs	38.4±0.8
37	Same	Same	03-380a	Poz-4924	Shrub twigs	39.2±0.1
38	Same	Same	03-380b	Poz-4925	Moss mat	39±0.1
39	Poloi upstream	Alluvial sand	03-379	Poz-4922	Moss mat	37.1±0.8
40	Same	Same	03-378	Poz-8301	Moss mat	37.7±0.9

41	Same	Same	03-382a	Poz-4926	Shrub twigs	42.5±1.4
42	Same	Same	03-382b	Poz-4928	Shrub twigs	40.5±1.1
43	Same	Same	03-375	Poz-8300	Moss mat	40.3±1.2
44	Same	Same	03-384a	Poz-8303	Same	46±2.5
45	Same	Same	03-384b	Poz-8922	Same	43.6±1.8
46	Same	Same	03-373	Poz-4921	Same	42.2±1.4
47	Same	Same	03-371	Poz-8299	Same	47.4±2.8
48	Same	Same	03-369	Poz-4920	Same	46±2
49	Same	Same	03-368	Poz-8297	Same	46.5±2.6
50	Pyak-Yaha downstream	Alluvial sand	01-129	LU-4759	Wood	≥50.4
51	Same	Same	01-131	LU-4765	Peat	≥44.7
52	Same	Pick up under bluff	01-126	LU-4785	Mammoth bone	35.2±0.21
53	Same	Same	01-123	LU-4788	Mammoth and horse	9.53±0.12
54	Pyak-Yaha upstream	Same	01-135	LU-4784	Mammoth bone	≥43.9
55	Shur-1	Autochthonous peat	01-170	LU-4761	Shrub twig	≥39.7
56	Same	Autochthonous peat	01-185	LU-4768	Birch with rind	≥47.2
57	Same	Soil in loess-like silt	01-178	LU-4763	Shrub twig	12.8±0.15
58	Same	Pick up under bluff	01-183	LU-4757	Mammoth tooth	≥30.7
59	Same	Pick up under bluff	01-176	LU-4786	Mammoth bone	17.34±0.23

Radiocarbon statistics demonstrate that the majority of finite dates from the Kara Sea basin originate from fine sediments that overly the last glacial complex (paper 10, Ch. III). The only remaining support for the idea of LGM ice sheet over the West Siberian Plain was the section of the Karginsky Horizon recorded in terrestrial facies on river Malaya Heta in the present southern tundra. Here an alluvium below glacial diamicton contains evidence of a warmer climate such as forest pollen, freshwater mollusks and Lake Baikal diatoms. The diatoms from Lake Baikal could reach the Arctic only by a free waterway, not ice-dammed. These alluvial sediments in the 1960s received 5 conventional ^{14}C dates in the range of 35–44 ka (Fig. 4B). This led Kind (1974) to suggest the `Malaya Heta warming` as a part of the Karginsky Interglacial. Redating of the subtill strata by modern methods (paper 13, Ch. III) yielded only non-finite AMS dates and OSL ages ranging from 79 in the aeolian sand atop of the alluvial formation to 112 ka BP in the floodplain alluvium proper (Fig. 4B).

Thus, both classical sections show that the glacial complex overlying the formations from the Last Interglacial could be deposited in any part of the Weichselian time and not necessarily during LGM. A pre-LGM age of the glacial complex of the arctic River Yenissei, called Late Zyryanka by Troitsky (1967) and Sartan by Kind (1974), is clearly indicated by overlying mammoth carcasses with radiocarbon dates of 36–17 ka BP (Fig. 1).

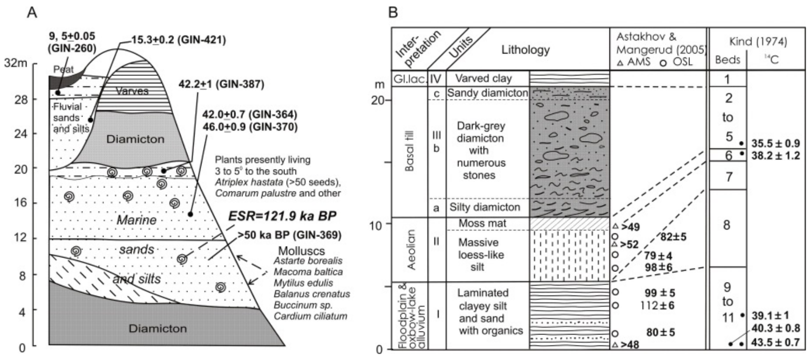


Fig. 4. Classical stratotypes of the Karginsky marine and alluvial formations: A – Cape Karginsky, 69°57´ N/ 83°33´ E, B – Malaya Kheta river, 69°32´ N/84°41´ E (Fig. 1). Conventional radiocarbon dates by Kind (1974) are black dots. ESR date (Arkhipov, 1990) and modern AMS and OSL dates (Astakhov & Mangerud, 2005) refute old correlation of these interglacial formations with MIS 3 (Kind, 1974).

The area some 300 km to the south is stratotypic for the Yermakovo glacial complex correlated with MIS 4. An early Weichselian age of the last ice advances has recently been confirmed by many AMS and OSL dates from postglacial terraces in the range of 66–30 ka BP (Fig. 3).

Besides, even older ages have been obtained from the Konoshchelye alluvium (paper 14, Ch. III), which according to Kind (1974) should represent a `Konoshchelye cooling` 30-33 ka BP within the Karginsky Interglacial. The cooling is expressed in parallel bedded sands with moss laminae but without any driftwood which is rather peculiar for a north-bound river crossing all forest zones of Siberia: the present-day alluvium of the Yenissei River is full of logs. According to the available dates (Table 3) the cold environment alluvium at Konoshchelye was deposited ~ 70-80 ka BP, probably immediately after the break-up of the ice front which dammed the River Yenissei near the Arctic Circle (paper 14, Ch. III). The second ice advance from the Central Siberian uplands ca 60 ka BP did not dam the Yenissei but produced a thick free sloping sandur (paper 15, Ch. III). Anyway, the series of OSL dates indicates that the surficial fluvial formations on the peri-Polar River Yenissei are younger than the Karginsky interglacial strata at 300-350 km downstream.

The most probable representative of the Last Interglacial complex was found in the River Yenissei eastern bank farther upstream. This is alluvium of a high terrace reported from Alinskoye (63°20′ N) and Mirnoye (62° 20′ N). The alluvium contains peat with spectra of southern boreal forest with abundant *Abies* pollen and spores of the East Asiatic ferns *Osmunda cinnamomea* and *Osmunda sp.* presently alien for Russia. The radiocarbon dates are non-finite (Zubakov, 1972, 1974).

Dated sediments of the central arctic lowlands

The basement of the exposed Upper Pleistocene in the West Siberian Arctic found within plateaus above 80 m a.s.l. is composed of thick diamicton and rhythmite formations. This glacial complex was originally interpreted as the Sanchugovka marine strata because of sparse arctic mollusks (Sachs, 1953). However, the sedimentological properties and contacts of the diamicton with overlying varves are typical for sediments of a large ice sheet, the same as were discovered at the Sanchugovka stratotypes of the Middle Pleistocene glacial complex on the Lower Yenissei (Kaplyanskaya & Tarnogradsky, 1975).

Low accretion plains at levels 25-30 m and 40-50 m a.s.l. are built of sediments incised into the Middle Pleistocene glacial complex and were previously interpreted as Karginsky or Kazantsevo marine formations. A sandy sequence up to 40 m thick (Belaya Yara, Nyunteda-Yaha and Sede-Yaha sections) (Figs. 1 and 5) has yielded 10 OSL dates, mean value 135.5 ka BP (Table 2). The character of the basal contact, sedimentary structures and abundant plant detritus is evidence of a rapidly increasing erosion rate during deposition in a shallow basin. This event could be connected with the final stages of the Middle Pleistocene deglaciation and beginning of a Late Pleistocene marine transgression. Similar marine sediments overlie glacial diamicton on the west coast of Taz Peninsula at Yamburg (67°55.5′ N/74°49′ E) (Fig. 1) where 4 OSL dates in the range of 167–124 ka have been obtained (Table 2).

Laminated marine sands, silts and clays of a similar age, as suggested by 6 OSL dates with mean value of 135 ka BP (Table 2), are exposed at Observation Cape on the northern tip of the Taz Peninsula (Figs. 1 and 5). The littoral facies contain numerous *in situ* shells of bivalves and gastropods accompanied by their abundant burrows. Typical boreal and arctoboreal species *Buccinum undatum, Macoma balthica, Modiolus sp.* and *Mytilus edulis* have been identified. Together with the geochronometric data they imply an interglacial transgression, probably contemporaneous with the Eemian. This marine formation can serve as the stratigraphic marker for correlation between the Ob and Yenissei regions. Its closest paleoclimatic and geochronometric counterparts are the Shuryshkary peat and alluvial sands on the Lower Ob River (paper 12, Ch. III) the Malaya Heta alluvium (paper 13, Ch. III) and the Karginsky marine strata on the River Yenissei (Arkhipov, 1990).

South of the Gydan Ridge there is no evidence of a Late Weichselian ice advance. Only fluvial sands, silts and clays with rare peat lenses and ice-wedge casts are incised into the described marine formations at Nyarova-Payuta and Observation Cape (Tables 2 and 3). According to OSL dates and non-finite radiocarbon ages they were formed in an earlier part of the Weichselian, approximately from 108 to 69 ka BP. This chronological interval probably corresponds to cold substages of MIS 5d to MIS 4. These sediments by OSL dates can be correlated with the fluvial sands and subaerial deposits overlying and underlying the Upper Pleistocene glaciolacustrine rhythmite on the Lower Ob River (paper 14, Ch. III)).

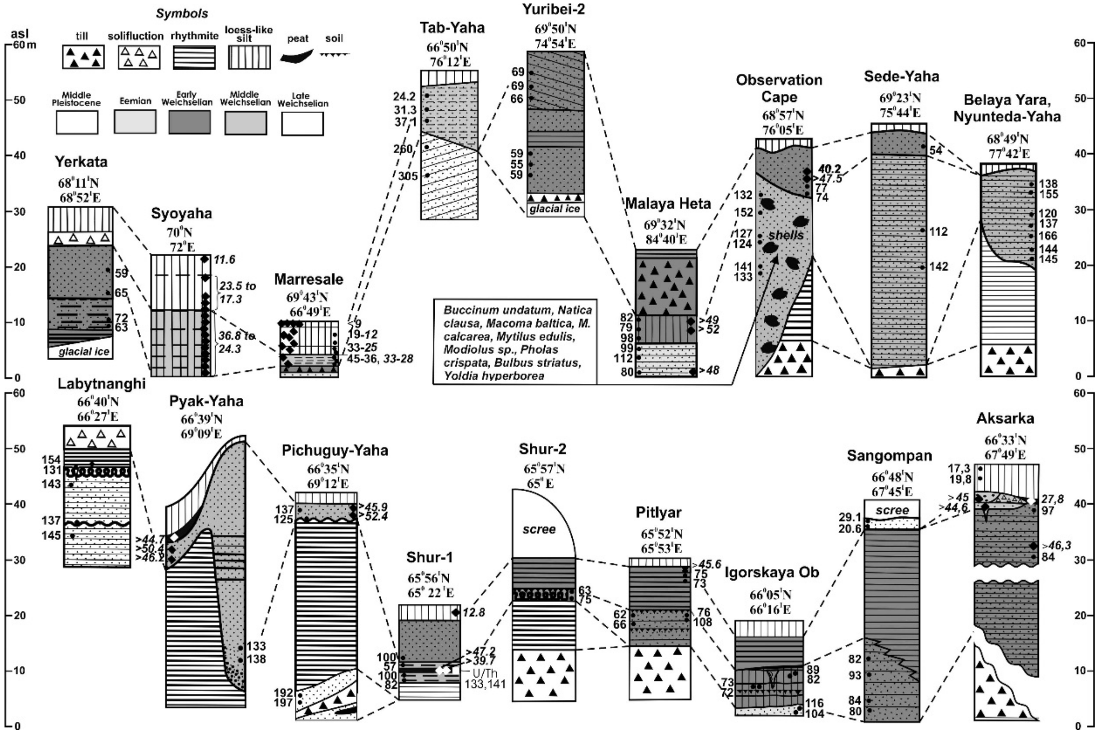


Fig. 5. Dated Upper Pleistocene sections correlated across northern West Siberia (see locations in Fig. 1). Numbers are ages ka BP: OSL – Roman, radiocarbon – italicized.

Thick diamicts and deformed massive ice are exposed only north of the Gydan Ridge crest. Fossil glacial ice in diamicton with typical for basal glacial debris sedimentary structures makes the base of the succession described at Yuribei-2 section (Figs. 1 and 5). They are overlain by outwash sand and varves (Nazarov, 2007). The two latter members make a transgressive/regressive series capped by a progradation delta. This sequence was probably formed during the retreat of the last glaciation. The OSL samples from beneath and above the varves have yielded ages 59 ± 3, 55 ± 3, 59 ± 3, 69 ± 4, 69 ± 4 and 66 ± 4 ka BP suggesting a last ice advance from the Kara Sea shelf onto the West Siberian Plain not later than the Middle Weichselian. Another and stronger indications that the uppermost glacial complex is older than MIS 2 are frozen mammoth carcasses from the surface sediments with ^{14}C ages within the span of 36–17 ka BP (black dots in Fig. 1).

Possible stratigraphic counterparts of this glacial complex are the ice-rich Kara diamicton in western Yamal (Forman et al., 2002) and basal diamicton and outwash of the last Putorana ice sheet at Goroshikha and Bolshoi Shar sections on the River Yenissei (paper 15, Ch. III).

The youngest dated fluvial sediments are represented by alluvium at Poilova-Yaha 68°07.30′ N/76°15′ E (Fig. 1, Table 3) and Tab-Yaha section (Figs. 1 and 5, Table 2) on the Taz Peninsula. According to three OSL dates and two ^{14}C dates in the range of 37–24 ka BP alluvial terraces were formed during MIS 3. No traces of a Middle Weichselian high sea level suggested by some authors (Volkova & Babushkin, 2000) have been encountered in studied sections from this area. Mammoth and woolly rhinoceros bones with ^{14}C ages >44.8, 44 ± 1.8 and 41.4 ± 1.3 from the beach of the Taz Estuary (Table 3) are also not consistent with a marine incursion of this age.

The late dry and frosty Pleistocene is marked by the discontinuous mantle of loess-like silts with long syngenetic ice wedges. The $\delta^{18}O$ values of ice wedges dated to 13-16 ka BP in the north of the Gydan Peninsula suggest winter temperatures lower than modern ones but higher than for the interval 22–11 ka BP in the Yamal Peninsula (Vasilchuk, 1992). The cover silt often contains scattered bones of mammoths.

The richest accumulation of mammoth bones is known on river Yuribei close to the mouth of river Lysukansye (L in Fig. 1). Conventional radiocarbon dates on bones from the thick loess-like silt on both banks of the river are 18,380 ± 700 (MGU-1049), 17,100 ± 600 (MGU-1020), 16,830 ± 670 (VSEGINGEO-16-9-85), 16,680 ± 500 (MGU-1047) and 16,520 ± 550 years BP (MGU-1019) (Bolikhovsky, 1987). The same substage is characterized by a mammoth bone at lake Parisento 17,500 ±

300 years old (GIN-7576), mammoth carcasses found on river Uribe (10,000 ± 30 years old (LU-1153) (Sulerzhitsky, 1995) and on river Mongoche-Yaha (16,990 ± 70 years BP, GrA-35614) (Gilbert et al., 2007) (Fig. 1). The mammoth bone ^{14}C dated to 18,710 ± 110 years BP in our Table 3 was picked up at the base of Nyunteda-Yaha bluff and probably also originates from the cover silt capping the sequence.

Western sections

Sections in the valley of the Lower Ob River close to the Arctic Circle along the western border of the sedimentary basin are crucial for identifying the main Upper Pleistocene marker beds of the Last Interglacial. During initial geological mapping the Sangompan Formation of cold-water lacustrine rhythmites atop of fluvial sand with plant remains was attributed to the `Kazantsevo Horizon`. This sequence occurs within a wide terrace incised into the thick Middle Pleistocene diamicts of the Salehard Formation (Lazukov, 1970). However, Zubakov found interglacial alluvium on a higher terrace at Pyak-Yaha, the southern bank of the Ob River, in the zone of sparse northernmost taiga. This alluvial sand contains peat lenses with southern taiga pollen and non-finite radiocarbon date (Zubakov & Levkovskaya, 1969). Later Arkhipov et al. (1977) using rare finite ^{14}C dates suggested a position of the Last Interglacial in a quite different part of the 200-300 m thick succession. They identified it in marine sands and silts with the Kazantsevo boreal foraminifer assemblage which was recovered by boreholes below sea level, in places down to 60 m b.s.l. This formation occurs under thick diamictons, varves and sands of the Lazukov's Salehard Formation (Fig. 6). The overlying peat deposits with non-finite radiocarbon dates were attributed by Arkhipov et al. (1977) to a `Lower Karginsky warming` presumably 40-50 ka old but with a climate close to the modern one.

We reinvestigated the peat-containing sediments (paper 12, Ch. III)). The thickest (up to 1.1 m) and most extensive (minimum 0.5 km wide) peat layer is located in northern taiga 1 km north of Shuryshkary settlement (our Shur-1 section) (Figs. 1 and 5). This very compact autochthonous sedge peat with tree trunks and cones is a rare feature for Siberia. It rests on diamictons and varves and is covered by shallow-lake silt, coarse fluvial sand and 5 m thick loess-like silt. The pollen spectra are full of spruce with admixture of broad-leaved trees pollen, whereas subrecent samples contain only 20% of spruce pollen. Our radiocarbon samples produced non-finite dates. OSL samples from peat-containing silty sand yielded two 100 ± 5(6) ka ages and one 82 ± 4 ka BP. A thin sand seam in the silty clay above the peat gave an anomalously young age of 57 ± 4 ka BP possibly reflecting higher environmental dose rate from the surrounding clay beds. We consider these OSL dates minimum estimates of the interglacial peat age.

Uranium-thorium analysis was applied because the peat layer appeared well-sealed by clayey layers from migrating younger uranium. The $^{230}Th/^{234}U$ ratio was measured along the entire column of peat thickness. The plateau of this curve is considered a signature of the original U/Th ratio not contaminated by young uranium and therefore was used for calculation of the age. Two methods of isotope extraction gave slightly different age estimates: 133 ± 14 ka BP by total dissolution and 141 ± 11.7 ka BP by leaching. These values are somewhat higher than U/Th ages obtained on Mikulino (Eemian) peats of Central Russia and a peat layer in Central Siberia (Arslanov et al., 2004). The geochronometric data at Shur-1 section are incompatible with the idea of MIS 3 age but can be reconciled with the MIS 5 time span, possibly MIS 5e. The idea of the Eemian time is further strengthened by the detailed pollen diagram indicating a climate warmer than the present one (paper 12, Ch. III). The combined paleoecological data and geochronometric constraints indicate that Shuryshkary warming commenced prior to ca 50 ka, most probably in MIS 5 and possibly in the Eemian.

We also revisited the Pyak-Yaha section where Zubakov and Levkovskaya (1969) reported pollen spectra of southern taiga and a ^{14}C date exceeding 57 ka BP (Fig. 1). Our 3 radiocarbon samples from peats also yielded non-finite ages (Fig. 5), and underlying alluvial sands here and at the next Pichugui-Yaha section of the same terrace produced OSL ages 138 ± 11, 137 ± 9, 133 ± 11 and 125 ± 10 ka BP (paper 11, Ch. III).

The above data mean that position of the main Upper Pleistocene thermomer correlative with the Eemian in the west of the Plain is established in terrestrial formations above sea level. Another implication is that the uppermost glacial complex of the valley of the Lower Ob is older than the Upper Pleistocene. The last glacial cycle in the Ob River valley is represented only by proglacial varves of the ice-dammed lake which are the upper part of the Sangompan Formation. We OSL dated this formation in several sections, the Sangompan site proper (Fig. 1) included.

The lower part of the formation is composed of well-washed fluvial or estuarine sand with forest macrofossils carried by the River Ob from the south. The upper inorganic member of clayey rhythmite ca 30 m thick upstream of Sangompan is sandier and thins to 3-5 m. The shallowing is evident at Shur and Pitlyar where the rhythmite contains regular symmetrical microdunes 1-2 cm high (paper 14, Ch. III).

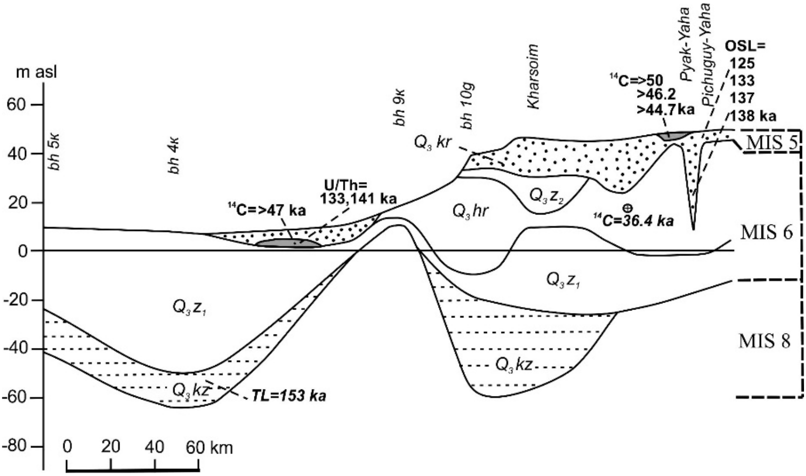


Fig. 6. Recent geochronological results in the Lower Ob valley by ICEHUS (regular print) (Astakhov, 2006a; Astakhov et al., 2007) as compared to results by Arkhipov et al. (1977, 1992) (italicized). Indices by Arkhipov et al.: Q_3kz – marine silt and clay with Kazantsevo foraminifer assemblage, Q_3z_1, Q_3z_2 – Lower Zyryankian and Middle Zyryankian glacial and lacustrine sediments. Correlation with MIS is by Astakhov et al. (2007). Q_3hr – Harsoim interstadial marine clay, Q_3kr – Karginskian alluvial sand. Strata below sea level are according to boreholes by Glavtyumengeologia Corporation.

The base of the rhythmite is crucial for stratigraphic reasoning. In most sections it has a sharp but non-erosive contact and lies conformably upon peat and aeolian deposits. These traces of abrupt flooding are replaced by a transition facies in the downstream Sangompan section. Pale horizontally bedded sand in this section grades upwards into laminated sand with clay bands. The clay bands occur with greater frequency up sequence transitioning to a clayey rhythmite (Fig. 4 in paper 14, Ch. III). This facies has yielded 4 OSL ages between 93 and 80 ka, with a mean age of 85 ka BP which determines the maximum age of the ice-dammed lake (Fig. 5). Similar sands at Aksarka (Fig. 1) gave several non-finite radiocarbon dates and OSL dates of 84 ± 10 and 97 ± 8 ka BP (Tables 2 and 3). Upstream of Sangompan the sands beneath the rhythmite show slightly younger ages: mean 78 ka at Pitlyar and 79 ka at Igorskaya Ob. Similar ages 75 ± 4 and 73 ± 4 ka BP have been obtained from the uppermost shallowed sandy part of the rhythmite member at Pitlyar (Table 2).

The minimum age of the uppermost glacial formation can be assessed on the Yamal Peninsula, north of the marginal chain of Sopkay moraines where, in addition to the data in Tables 2 and 3 of the present paper, two series of 38 reliable ^{14}C and 7 IRSL (infrared stimulated luminescence) dates have been published (Vasilchuk et al., 2000; Forman et al., 2002). The Siberian schemes presume that the Sopkay moraines belong to the last kryomer called Sartan and correlated with MIS 2 (Arkhipov, 1990; Volkova & Babushkin, 2000).

However, the thick and old succession of postglacial sediments atop of the very icy ground moraine, termed the Kara Diamicton in the key section on the western coast of Yamal Peninsula at Marresale (Fig. 1), does not support this idea. This section is now dated by 23 AMS and 7 luminescence age determinations. The base of the succession is represented by 4 m thick lens of Varyaha shallow limnic silt with peat laminae that formed during a brief climatic amelioration within the generally cold and dry environment, judging by the rare herbal pollen with *Artemisia* (Andreev et al., 2006). The Varyaha interstadial has yielded 6 radiocarbon dates of 33–28 ka BP and 3 IRSL dates of 45–36 calendar ka BP. It is topped by a series of aeolian silt and sand (Oleny and Baidarata formations) with syngenetic ice wedges and lenses of fluvial sand with 17 successive radiocarbon dates from 28 to 12 ka BP and 4 IRSL dates of 33–13 ka BP (Forman et al., 2002). This part of the succession with thick ice wedges bears only rare grains of herbal pollen (Andreev et al., 2006). The bluff is capped by Holocene peat.

The major deficiency of the Marresale sections is the hiatus between 25 and 15 ka BP. It is compensated by another section on eastern Yamal at the Syoyaha river mouth where loess-like silt with syngenetic ice wedges yielded a successive series of 15 ^{14}C dates from 37 to 11 ka BP (including 8 AMS dates). The lower peaty beds of the Syoyaha sequence with ^{14}C dates of 37–28 ka probably correspond to the Varyaha interstadial. The very light isotopic composition of the ice wedges indicate that January temperatures during 22–11 radiocarbon ka BP were at least 6-10 °C and annual temperatures were by 3-6 °C lower than the present ones. This means that summer temperatures were not lower than the present ones. The interval of 25–17 ^{14}C ka BP, missing at Marresale, was studied at Syoyaha in a continuous succession with 9 dates (Vasilchuk et al., 2000). A similar sequence was described close to the northern tip of the Gydan Peninsula (Vasilchuk, 1992).

The old (an early Weichselian) age of the last Yamal glaciation is confirmed by the Yerkata section of aeolian and lacustrine sands overlying varves and fossil glacial ice and by finds of frozen mammoths (Fig. 1).

The sands at Yerkata have yielded 4 OSL dates ranging from 72 to 59 ka BP (Mangerud et al., 2004). Well- preserved mammoth babies drowned ca 39 ka (Masha) and 41.9 ka ago (Lyuba) are clearly positioned atop of the uppermost glacial complex (Fig. 1).

Discussion

The above account of recent dating efforts in the northern plains of West Siberia suggests a revision of the conventional stratigraphic scheme. There are clear indications that current formal stratigraphic schemes based on sparse and questionable geochronometric data are of limited value.

The Last Interglacial

The position of the warmest Late Pleistocene stage has always been ambiguous in regional stratigraphic schemes. Incorrect correlation is widespread along the Ob River where the Kazantsevo marine interglacial formation, presumably synchronous with the Eemian, was previously placed well below sea level (Arkhipov et al., 1977, 1992). However, this formation is overlain by the thick Salehard glacial and glacio-aqueous strata into which the Sangompan terrace is incised. The mean OSL age of the Sangompan varved sequence is 85 ka BP, practically the same as the mean 82.4 ka age of Lake Komi sediments west of the Urals (Astakhov et al., 2007). The fluvial sands, loess-like silts and blown-out soils under the varves probably reflect an interstadial episode some 80-90 ka, just before the main ice advance from the north. Therefore, the conventional radiocarbon dates 43 and 34 ka BP obtained earlier from beneath the rhythmite at Igorskaya Ob (Arkhipov, 1997) and 28.6 ka BP from the Aksarka sand (Arkhipov et al., 1977) are obviously too young. We obtained a 27.8 ka radiocarbon date much higher up in the succession from the base of the coversand at Aksarka (Astakhov et al., 2007).

The Shuryshkary interglacial peat dated to 140–100 ka BP is positioned atop of Arkhipov's Hashgort Till (part of the Salehard Formation by Lazukov, 1970) which in turn overlies the Kazantsevo marine sediments below sea level. The Pyak-Yaha interglacial peat occupies the same high position (Fig. 6). Therefore, the underlying diamicts, varves and marine sediments, which are placed in the Upper Pleistocene of the Siberian schemes, are most likely Middle Pleistocene.

The implication is that the Kazantsevo foraminifer assemblage (Levchuk, 1984; Arkhipov et al., 1992, Arkhipov, 1997; Gusskov and Levchuk, 1999) is not an indication of the Late Pleistocene age. A TL date of 153 ± 15 ka BP from this formation (Arkhipov et al., 1992) can hardly be taken as a signature of the Late Pleistocene, especially in view of more numerous OSL and U/Th dates in the bracket of 140–100 ka BP from the strata lying higher up in the succession (Fig. 6). The only finite date of 34.8 ka BP from the thick Salehard Formation underlying alluvial peat deposits is too young because of its stratigraphic position and also because it originates from a piece of peat extracted from the deep-water marine clay at Harsoim (Arkhipov et al., 1977).

The stratigraphic position of the Kazantsevo marine strata *s. stricto* on the Lower Yenissei is far from being clear. They cannot be traced laterally even in the stratotypic area of Ust-Port which leads to a well-grounded suspicion of their being just detached blocks. The Malaya Heta alluvium constrains the southern limit of the last interglacial sea on 69° N. Therefore, interglacial marine strata on the River Yenissei south of 69°N should belong to older, pre-Karginskian (pre-Eemian) thermomers. The U/Th date of 233 ka on boreal shells at Pupkovo, 65° N, i.e. far south from the Malaya Heta section, led Zubakov (1974) to correlate the Kazantsevo transgression with MIS 7.`

There are other indications that the Kazantsevo thermomer is older than the Upper Pleistocene. According to Sachs (1953) the main criteria for identifying the Kazantsevo littoral sequence are i) its subtill position, ii) the occurrence of warm-water, manifestly boreal mollusks, such as *Cyprina (Arctica) islandica*, iii) sporadic appearance of the extinct species *Cyrtodaria jenisseae Sachs.* The latter Siberian mollusk is not different from the well-known *Cyrtodaria angusta Nyst et Westendorp* (Merklin et al., 1979) which occurs in the Pechora Basin only beneath thick Middle Pleistocene diamictons of the Padimei Formation (Zarkhidze, 1972). The same pattern appears valid in West Siberia: marine strata with *Cyrtodaria jenisseae* were described in the subtill position well south of the limit of the Late Pleistocene ice sheet in the Samburg borehole (67° N) (Fig. 1), also on Limbya-Yaha river, an eastern tributary to the Taz River (67°20′ N) (Zubakov, 1972).

Thus, the remaining candidate for the warm marine event in the beginning of the Late Pleistocene is the Karginsky Formation proper with the ESR date 121.9 ka BP and OSL ages ~ 111 ka BP. This formation, evidently a counterpart of the Eemian, is also well expressed at Observation Cape (Fig. 1) where we have 6 OSL dates of mean value 135 ka BP from a marine formation with typical boreal fauna (Fig. 5). On the Lower Ob River peats with U/Th values 135–141 ka BP and fluvial sands with 125–138 ka OSL ages probably reflect the same warm event.

Chronology of the last glaciation

The stratigraphic position of recent glacial events is less confusing. The sedimentary complex of the last inland glaciation traditionally bore the label `Zyryanka` from the creek between the Cape Karginsky and Malaya Heta sections (Sachs, 1953). This name can be retained for surficial morainic deposits with fossil glacial ice and associated glacio-aqueous sediments. However, Sachs himself miscorrelated this formation with glacial deposits underlying the Karginsky thermomer which according to our work are part of the Middle Pleistocene. N. Kind (1974) acknowledged the overlying glacial complex, unfortunately attaching to it the Sartan label derived from alpine moraines. The following misuse of the Sartan name as the synonym of the last inland glaciation of Siberia prompts us to abandon it altogether and restore Sachs's usage of the Zyryanka name for the last glaciation, although Sachs (1953) never tried to lay the Zyryanka moraines atop of his Karginsky Horizon. This revision is advisable also on the grounds that in Sachs's understanding the Zyryanka glaciation was not the final Late Pleistocene but older which is confirmed by modern OSL and ^{14}C dates.

A geochronometric bracket for the last inland glaciation is constrained by mean OSL age of 85 ka BP for the Lower Ob proglacial lake and ages of the Gydan sandur at Yuribei-2 (mean 63 ka BP from 6 dates, Fig. 5) and Yenissei sandur (mean 59 ka BP from 11 dates, Fig. 2). This age bracket, which is similar to the OSL range of the Pechora Basin (Astakhov et al., 2007), has lately been supported by OSL dating of the last glacial invasion into the Polar Urals from the northwest. The sandur emanating from an end moraine inserted into the Urals has a mean age of 73 ka BP on 13 OSL dates (Nazarov et al., 2009). However, the mean age of the Konoshchelye postglacial alluvium of 75 ka BP on the Yenissei River (Fig. 2) suggests that the glaciation proceeded in two stages: the first is ~ 80-90 ka and the second is ~ 60 ka BP which is similar to double Weichselian ice advances in northern European Russia and on the Taimyr Peninsula (Svendsen et al., 2004).

Another interstadial event can be constrained using radiocarbon chronology which is more reliable now owing to many finds of *in situ* mammoth remains and some remnants of plant organics in postglacial sediments. The principal sections in the Yamal Peninsula give us an approximate time span of the weak climate amelioration in the middle of the last glacial cycle which can be readily compared with MIS 3 and the Middle Pleniglacial of Western Europe.

In Marresale section these are AMS radiocarbon dates 37–28 ka BP from the Varyaha interstadial lenses of shallow lake silts with peaty laminae. The key sections of Marresale and Syoyaha can be easily correlated with even better sections from farther east: Cape Sabler with 60 ^{14}C dates from icy silts with peaty laminae and plant remains and the Yedoma Formation in the Lena delta where already 90 ^{14}C dates are available, the oldest being 48 ka BP.

The isotopic composition of the ice wedges, plant macrofossils and macrofauna remains in these sections unequivocally indicate an extremely continental, i.e. frosty and arid climate with warmer summers, of the Middle and Late Weichselian time in Siberia which developed into the present relatively humid environment only at the beginning of the Holocene (Hubberten et al., 2004; Sher et al., 2005). The icy subaerial silts continuing below sea level with the long series of finite radiocarbon dates exclude any marine incursions for the last 50 ka. A very similar situation is recorded in the Yamal key sections.

In the Ob River valley, where permafrost is already extinct, the final Pleistocene is recorded in the mantle of coversand and loess-like silts with occasional mammoth bones and OSL dates from 30 ka BP (Sangompan) to 18 ka BP (Aksarka) (Fig. 5). Many radiocarbon dates give similar age estimates for the subaerial cover – from 29 ka BP at the base of Aksarka coversand to 12 ka BP in a thin paleosol at Shuryshkary. The fine-grained composition of the subaerial mantle, as well as the radiocarbon ages of megafauna bones 25, 17.5, 16.4, 15.5 ka BP underline the absence of active ice during the LGM time (Astakhov, 2006a; Astakhov et al., 2007). The $\delta^{18}O$ values in ice wedges of the cover silt formation of eastern Yamal and northernmost Gydan suggest the extremely cold winters from 22 to 11 radiocarbon ka BP (Vasilchuk, 1992).

New stratigraphic scheme

Objective stratigraphic and associated chronometric data that establish the geochronological position of two extreme paleoclimatic events on the West Siberian Plain – the mildest climate of the Last Interglacial and the harshest climate of the last glacial – can now be used to revise the regional stratigraphic scheme.

The main features of the new scheme (Table 4) are: i) the traditional four-fold subdivision; ii) new stratotypes for the climatostratigraphic horizons (warm or cold intervals) of the second half of the Upper Pleistocene instead of the non-existent Sartan and Karginsky stratotypes of the former schemes; iii) the Varyaha interstadial suggested instead of the Karginsky thermomer; and iii) the new Malaya Heta (Karginsky) horizon as the main thermomer of the Upper Pleistocene based on the old stratotypes.

Table 4. Regional stratigraphic scale of the Upper Pleistocene suggested for northern West Siberia (after Astakhov, 2006b)

Climato-stratigraphic horizons: kryomers and *thermomers*	Key sections	Characteristic sedimentary formations	Geochronometric marks, ka BP	Approximate correlation: Western European chronostratigraphy	Approximate correlation: MIS
Syoyaha	Syoyaha, Marresale parastratotype	Icy loess, aeolian sands, ice wedges	^{14}C=28–12; OSL, IRSL= 33–13	Upper Weichsel	**2**
Varyaha	Marresale, Syoyaha parastratotype	Sinkhole limnic and palustrine silts with mammoths, alluvium, interlayers of loess-like silt, ice wedges	^{14}C=37–28; OSL, IRSL= 45–35	Middle Weichsel	***3***
Zyryanka	Rivers Zyryanka, Malaya Heta, Gydan Yuribei, Sangompan	Complex of last glaciation: diamictons, fossil ice, varves, sandurs	OSL=108–54, mean 75 of 34 dates	Early Weichsel	**4**
Malaya Heta= Karginsky	Malaya Heta river; Cape Karginsky, Observation Cape, Shuryshkary	Forest alluvial beds, peat deposits, marine strata with boreal fauna	ESR=122; U/Th =141, 133 OSL=160–100, mean 127 of 26 dates	Eem	***5***

Note: the Zyryanka kryomer probably comprises sediments of two ice advances.

The suggested subdivision avoids the contradictions of the official scheme (Volkova & Babushkin, 2000) and still retains the principal reference sections by Sachs (1953) and Kind (1974), changing only their distant correlation with the European and oceanic scales.

The new system needs of course approval by a regional convention. Therefore we derive the name of the main thermomer from the Malaya Heta terrestrial stratotype in order to avoid the Karginsky name for this substage because of the ambiguous stratigraphic and chronologic association. However, we retain the Karginsky label in parenthesis in case a nomenclature commission keen on the priority article might prefer to use it for this temperate stage.

Conclusions

(1) New chronological data presented here suggest that the Karginsky warming is correlative with the Eemian and not with the cold Middle Pleniglacial at a low sea level stand. The last (Karginsky) invasion of warm Atlantic waters into the Siberian Arctic occurred within MIS 5. The last inland glaciation of West Siberia with all its stadials developed in the time span of 90–60 ka BP, i.e. approximately at the level of late MIS 5 to MIS 4. It corresponds to the Early Weichselian – early Middle Weichselian time of Western Europe. Only the loess-like mantle with ice wedges is correlative to the Late Weichselian/Late Valdai time.

(2) The main thermomer of MIS 5 with the mildest climate is reliably identified as a lateral sequence of sedimentary formations: peat deposits and alluvium of the Lower Ob River, Karginsky alluvial and marine sediments with arctoboreal fauna and their analogues in central arctic lowlands overlain by the topmost diamicton with fossil glacial ice. On this basis, a Middle Pleistocene age of the underlying glacial strata and Kazantsevo marine sediments with the extinct mollusk *Cyrtodaria angusta* is inferred.

(3) The postglacial sedimentary mantle is mostly represented by loess-like silts with long ice wedges. Local limnic and fluvial lenses within this subaerial cover contain the famous finds of frozen mammoth carcasses radiocarbon dated to 42-17 ka BP. The paleoclimatic extreme – the most continental, cold and arid environment of MIS 2 time is connected with the topmost and thickest part of the icy loess.

(4) This study indicates that OSL correlation signal in arctic sediments is reasonably compatible with the geological record and

radiocarbon chronometry. The OSL ages, though variable in accuracy beyond the 100 ka limit, are consistent with the sedimentary succession.

(5) The reported stratigraphic and geochronometric criteria for correlations across arctic West Siberia give a firmer basis for other Northern Hemisphere correlations. They indicate that despite the asynchronicity of glacial maxima in Europe and Siberia, the general trend of paleoclimate change, including sea level fluctuations, is comparable. This knowledge allows us to at last correlate the main events of the Late Pleistocene in both regions and rectify the stratigraphic scale of West Siberia.

Acknowledgements

This paper is a result of ongoing Russian-Norwegian collaboration and a contribution to the European coordination program APEX (Arctic Paleoclimate and its Extremes). The field study and dating has been carried out, since year 2000, within research projects PECHORA and ICEHUS supported by the Research Council of Norway (co-ordinator John Inge Svendsen). We greatly appreciate the logistical assistance by Ecological Station of Yekaterinburg Institute of Ecology in Labytnanghi (director V. Shtro) and Geocryological Laboratory of Institute of Permafrost in Igarka (director O. Kazansky), both subsidiaries of Russian Academy of Sciences. We are also thankful to two anonymous reviewers who meticulously combed the manuscript in order to improve the English and make the presentation more correct without damaging the message.

17. To the Chronology of the Last Ice Age on the Lower Yenissei

V. I. Astakhov and J. Mangerud

Doklady Earth Sciences, Vol. 455(1), P. 219–222.
Translated from Doklady Akademii Nauk, 2014, vol. 455, No. 1, P. 48–51.

Introduction

The first Late Pleistocene ice advance is inferred from the Yermakovo basal till buried beneath the chain of terminal moraines south of the Arctic Circle near settlement Angutikha (Arkhipov et al., 1980). The Yermakovo diamicton in the formal stratigraphic scheme of West Siberia is related to the interval of 100 to 50 ka BP, whereas the end moraines barring the Yenissei are thought to belong to the last Pleistocene kryochron ~ 23–10 ka BP correlated with MIS 2 (marine isotope stage 2) (Kind, 1974; Volkova & Babushkin, 2000). The young age of the terminal moraines is in discrepancy with the \`old\` radiocarbon dates from post-glacial sediments (Astakhov & Isayeva, 1985; Astakhov & Mangerud, 2007).

Study object

The events of the last Ice Age are best recorded in Bolshoi Shar section of the right-bank Yenissei terrace. This is a longitudinal bluff up to 10 km long and 45 m high located on the eastern bank across of the mouth of the River Yenissei channel called Bolshoi Shar. Its upstream end is at 24 km downstream from the Nizhnyaya Tunguska river mouth, 66º N/88º E. The downstream end of the bluff is only 20 km south from the Angutikha Line of the Late Pleistocene glacial maximum. In the upstream end of the bluff periglacial alluvium overlain by sediments of an ice-dammed lake of the beginning of the last (Zyryanka) Ice Age has been repeatedly described since 1950-s (Lavrushin, 1963; Kind, 1974). The uppermost coarse sands have been related to the Karginsky interglacial alluvium by Kind (1974) who obtained infinite radiocarbon dates from this section.

In 2003 a new study of the Bolshoi Shar sequence was performed in the course of the Russian-Norwegian research collaboration. A previously unknown till in the upper part of the sequence was found by tracing the basic members of the sequence for 6 km from the upstream end (Figure). Quartz sand grains analysed in Nordic Laboratory for Luminescence Dating, University of Aarhus, Denmark, according to the techniques in Murray & Wintle (2000), have yielded 22 optically stimulated luminescence (OSL) dates (Table). Eight dates from this collection were already published (paper 14, Ch. III). The dates are treated as a statistical sample (Svendsen et al., 2004; Astakhov, 2006).

Lithostratigraphy

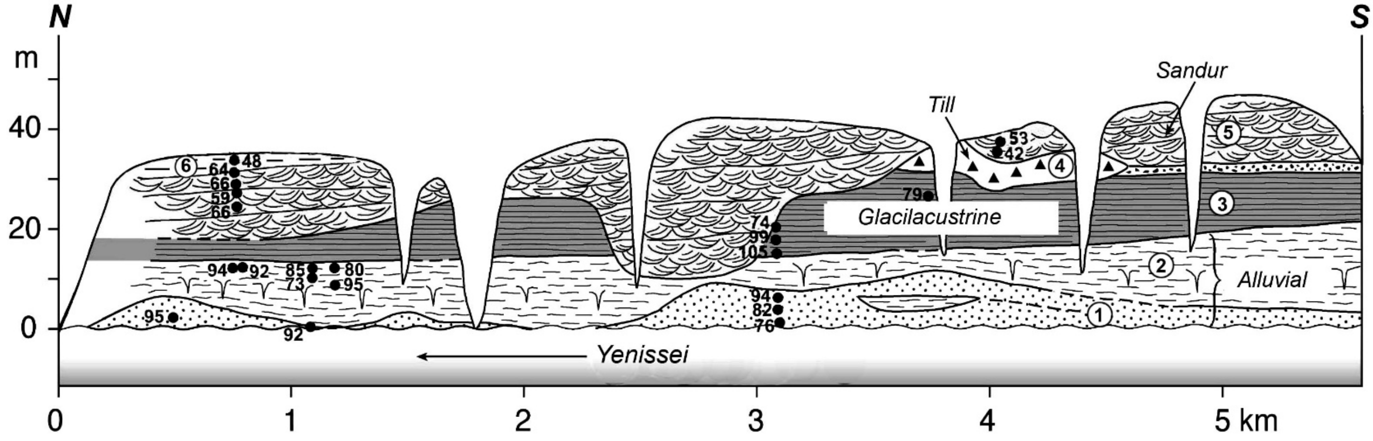
The following units are described in several clearings from the water level upwards: (Fig. 1):

1). Light-yellow sand, fine and medium-grained, micaceous, well washed, with north-dipping cross-beds. Lenses of fine plant detritus and brown silt are present. Apparent thickness is up to 10 m. Channel alluvium.

2) Gray-bluish silt, compacted, conformably lying, with seams of fine sand every 2-3 cm. At 7-10 m level there is an interlayer of reworked peat which every 20 m sinks into large ice-wedge casts 3-4 m deep. Lenses of fine sand with ripple marks occur in the upper part. Up to 15 m thick. Floodplain alluvium.

3) Bluish-gray, stiff, clayey rhythmite with seams of fine sand and plant detritus, changing into silty rhythmite upwards and into varved clay downwards. Summer layers are 0.5-1 cm and winter layers are several mm thick. The base of the unit is a 1 m thick transition member consisting of laminated fine sand and silt with symmetric ripple marks. The basal contact is flat, non-erosive, with load casts in fine sand and with calcareous nodules. In the upstream end of the bluff this member is capped by a 2-3 m thick pale-pink silty clay looking as a profile of deep weathering. Up to 13 m thick. Sediments of a cold lake with a distinct seasonality.

4) Newly discovered dark-gray massive diamicton, sandy at the base and bouldery in the upper part. Contains flattened sand rafts 1 to 5 m long. Cobbles and boulders, consisting of the Mid-Siberian trapps, are poorly rounded, their polished facets dip north-eastwards. Up to 7 m.



Geological profile of Yenissei eastern bank, upstream part, against the mouth of the Bolshoi Shar channel. Figures at the black dots are OSL ages, ka. Numbers of lithostratigraphic units are encircled.

5) Brownish poorly sorted sand, coarse and medium-grained, with west and north-west dipping cross-beds and trough cross series, with gravel lenses and boulder pavements at the base. Sand grains are predominantly trapp fragments with a meagre admixture of quartz. Between 2.3 and 3.2 km of the measured section the sand up to 23 m thick fills in a sub-latitudinal steep-walled trough (Fig. 1).

6) Lenses of laminated fine sand with silt interlayers, up to 2 m. Probably sediments of shallow ponds.

7) A discontinuous mantle of pale loess-like silt with ice-wedge casts, ~ 2 m.

Facies interpretation

There are two basically different formations clearly observable in the Bolshoi Shar section. The lower formation of light sand and bluish silty and clayey rhythmites (members 1 to 3) reflects fluvial sedimentation changing into lacustrine sedimentation in a glacial climate. The Ice Age environment is evident from the lack of driftwood in the alluvium and non-arboreal pollen spectra with abundance of *Artemisia* and *Betula nana,* which are unusual for the present boreal forest. The tree limit of that time was located far away in the south (Lavrushin, 1963; Kind, 1974; Astakhov & Mangerud, 2007). The same is indicated by the large pseudomorphs after former ice-wedges which do not presently grow in floodplains on this latitude. The north-bound fluviatile discharge was terminated by the valley lake which appeared due to an ice dam along the Arctic Circle (Lavrushin, 1963; Astakhov, 2006).

The upper brown-tinted formation with predominance of trapp material (members 4 to 6) resides with a sharp unconformity and basal erosional troughs. All direction structures indicate an eastern provenance. The diamicton of unit 4 was deposited subglacially by ice advancing from the Putorana Plateau. The overlying coarse sand (unit 5) belongs to a sandur climbing eastwards towards the fresh-looking hummocky moraines of the Yenissei–Malaya Shorikha interfluvial plateau. Unit 6 was probably deposited in small residual ponds on the sandur.

Geochronology

The predecessors distinguished only one, early Late Pleistocene proglacial complex, labelled the Yermakovo in the official stratigraphic scheme (Volkova & Babushkin, 2000). According to Kind (1974), the glaciolacustrine rhythmites of this age in the Bolshoi Shar section are overlain by coarse sand of the Karginsky interglacial alluvium.

The clear subdivision of the Bolshoi Shar sequence into two formations of different ages is supported by the results of OSL dating. The basal alluvium changes gradually into the rhythmites which is also evident from the OSL ages indiscriminate in both units (Table, Fig.).

The mean age of the lower formation is 86.7 ± 5.6 ka from 15 measurements. This is practically the same as 85 ka, the age of initial accumulation of the varved clay on the Lower Ob (paper 14, Ch. III) and very close to the age of the sediments of the major ice-dammed lake on the Pechora which is ca 90 ka (the weighted mean is 82 ± 1.2 ka on 27 dates) (Svendsen et al., 2004). All these ages provide a consistent chronometric signal for the maximum Late Pleistocene ice advance from the Kara shelf.

The sands of the upper glacial complex yield very different OSL values. They, together with 3 dates 63, 83 and 57 ka from the same sandur in a nearby section of Goroshikha (paper 15, Ch. III), constitute a sample of 10 dates with the mean value of 60 ± 4 ka. Thus, according to OSL dating the upper glacigenic formation proves to be by 20 ka younger than the lower glacial formation of Bolshoi Shar and by 40-60 ka younger than the alluvial and marine strata in the Karginsky key sections (papers 13 and 15, Ch. III).

The postglacial Pleistocene is represented by the sand alluvium of the II Terrace (the Konoshchelye Terrace by Kind, 1974), which has been thoroughly dated at Poloi. The upper part of this terrace yielded a consistent series of 18 AMS radiocarbon dates in the range of 47–27 ka BP supported by 11 OSL dates (paper 15, Ch. III). The Karginsky and Konoshchelye strata by the traditional radiocarbon chronology, accepted by the regional stratigraphic scheme, are placed between two glaciations of the Late Pleistocene called the Yermakovo and Sartan according to (Kind, 1974; Volkova & Babushkin, 2000). The new data reliably put both glacial events within the time span between the Karginsky interglacial and Konoshchelye interstadial *s. stricto.*

Conclusions

Clear indications of two major glacial events of the Late Pleoistocene are recorded in the Bolshoi Shar section. These are: i) the maximum ice advance from the Kara ice dispersal centre southwards up to the 66°10′ parallel with ice-damming the Yenissei discharge and ii) an advance of Putorana glaciers with a free discharge of meltwater into the Yenissei valley, which resumed after the draining of the valley lake. The OSL

Table. Optically stimulated luminescence dates from Bolhoi Shar section measured by A. Murray, Luminescence Laboratory, University of Aarhus, Denmark

Sediment	Sam-ple №	Lab. №	Wt.cn. %	Paleo-dose, Gy	Dose rate, Gy/ka	Age, ka BP
Medium sand, alluvium	243	042529	32	86±4	0.93±0.05	**92±7**
"	259	042535	*28*	*85±3*	1.11±0.05	**76±5**
"	255	042532	41	99±2	1.05±0.06	**95±6**
"	260	042536	35	88±3	1.08±0.06	**82±6**
"	261	H52510	42	104±3	1.11±0.05	**94±5**
Fine sand, alluvium	257	042533	36	80±2	0.95±0.05	**85±5**
"	258	042534	28	77±3	1.05±0.05	**73±5**
"	237	042525	33	64±3	0,68±0,04	**95±7**
"	238	042526	33	85±6	1,07±0,05	**80±7**
"	239	042527	36	89±4	0,98±0,04	**92±6**
"	239a	042528	38	89±6	0,95±0,05	**94±8**
Fine glaciolacustrine sand	268	042537	47	120±4	1,14±0,06	**105±7**
"	271	042538	40	95±4	0,96±0,05	**99±7**
"	272	042539	38	82±4	1,10±0,05	**74±5**
"	262	H52511	36	86±2	1.08±0.05	**79±5**
Coarse glaciofluvial sand	247	H52507	25	47±2	0.71±0.04	**66±5**
"	248	H52508	34	62±3	1.05±0.04	**59±4**
Medium glaciofluvial sand	249	042530	28	58±3	0,88±0,04	**66±4**
"	250	042531	31	59±3	0,92±0,04	**64±5**
"	273	H52512	32	34±4	0.81±0.04	**42±5**
Fine sand	275	H52513	25	43±2	0.80±0.04	**53±4**
"	251	H52509	34	55±4	1.15±0.05	**48±4**

dating of the Bolshoi Shar sands suggests that the maximum south-bound ice advance occurred ~ 90-80 ka BP, whereas the second glacial stage in the form of a Putorana ice sheet took place ~ 60 ka BP. The new data confirm the chronology of the last Ice Age derived from the Pechora and Taimyr sections by the European program QUEEN (Svendsen et al., 2004), but do not support the traditional correlation of the Upper Pleistocene still employed in West and Central Siberia (Kind, 1974; Volkova & Babushkin, 2000).

Acknowledgements

The above results are obtained in the course of the collaborative research of the Russian-Norwegian PECHORA II project (Paleo Environment and Climate History of the Russian Arctic) supported by Research Council of Norway. This paper, composed with the help of the grant № 11-05-00624 from Russian Foundation for Basic Research and of a grant from St. Petersburg University, is also a part of the European coordination program APEX (Arctic Paleeoclimate and its Extremes). Authors are thankful to A. Matyushkov, VSEGEI, for the cooperation on the section, to O. Kazansky, Igarka Geocryological Laboratory, for the logistic assistance and to A. Murray, University of Aarhus, Denmark, for processing and analysing the OSL samples.

CHAPTER IV

LATE QUATERNARY GLACIATIONS OF NORTHEASTERN EUROPEAN RUSSIA

Three papers of this Chapter are devoted to the northeastern corner of European Russia studied by numerous mapping and scientific expeditions since the 1930-s .The region with the Pechora waterway, railroad from Moscow and local roads around cities of Vorkuta, Pechora, Naryan-Mar, Usinsk is better accessible for investigators than Siberia but was the last to receive sufficient geochronological information for solving the old problem of age and extent of the last glaciation. This paradox is largely due to the over study: too many efforts with abundance of conventional (and controversial) radiocarbon and pollen data giving a wrong impression of too young glacial events. As a result a Late Weichselian ice sheet was almost unanimously accepted for this area and was taken for granted when the Russian-Norwegian collaboration project PECHORA started field and laboratory explorations in 1993.

This dominant paleogeographic idea is felt even in the publication presenting data against a young glaciation in Siberia (paper 10 in Chapter III) but still accepting, although with some reservations, a young ice sheet in the west. Only in the 1998, under the weight of the growing geochronological data, the members of the Russian-Norwegian team had to completely abandon the idea of a Late Weichselian ice sheet for northern European Russia beyond the limit of the Scandinavian glaciation (Boreas, 1999, vol. 28). Further on, over the years of work on different international projects, northern Russia, including the Urals, was totally freed from Late Weichselian inland ice. This Chapter offers a part of the voluminous evidence that allows to revert to the classical idea of early Late Pleistocene glaciation for the Russian mainland and adjacent arctic seas.

18. Marginal Formations of the Last Kara and Barents Ice Sheets in Northern European Russia

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Boreas, 1999, Vol. 28, p. 23–45

Introduction

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 by three coordinated papers in this issue of Boreas, presenting results of six years 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 that 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 to the sediments and chronostratigraphy discussed in Mangerud et al. (1999), where it is concluded that the Barents and Kara ice sheets most probably did not reach the Russian maninland 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).

Quaternary mapping of Northern Russia by local geological surveys has for decades been largely influenced by non-glacial theories of the drift origin, making 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 have been 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 at the scale of 1:200 000 for vast areas west of 60° E (Lavrov, 1977; Lavrov et al., 1986, 1991). These data have been used for paleoglaciological 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, basing 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-centered 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 within 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 NE 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 south-western 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 an 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).

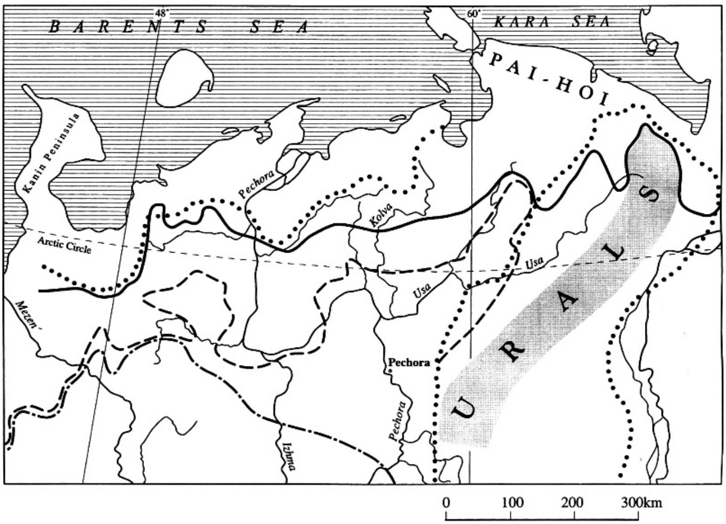

Fig. 1. Map of northern Russia with Weichselian ice-sheet limits according to different authors: dotted line – Early Weichselian by Yakovlev (1956), dashed line – Late Weichselian by Lavrov (1977) and Arslanov et al. (1987); solid line – Early-Mliddle Weichselian (Markhida Line in Fig. 2) by present authors.

Methods and principles of mapping

Aerial photographs and satellite images

Our data on surficial glacial features have been mostly derived from stereoscopically studied aerial photographs at basic scales of 1:50 000 and 1:35 000 (Figs. 6-12). We used altogether about 4 500 aerial photographs and 60 photomosaics, mostly obtained by aircraft surveys between 1988 and 1991. These are of better quality than the images of the 1940-60s 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

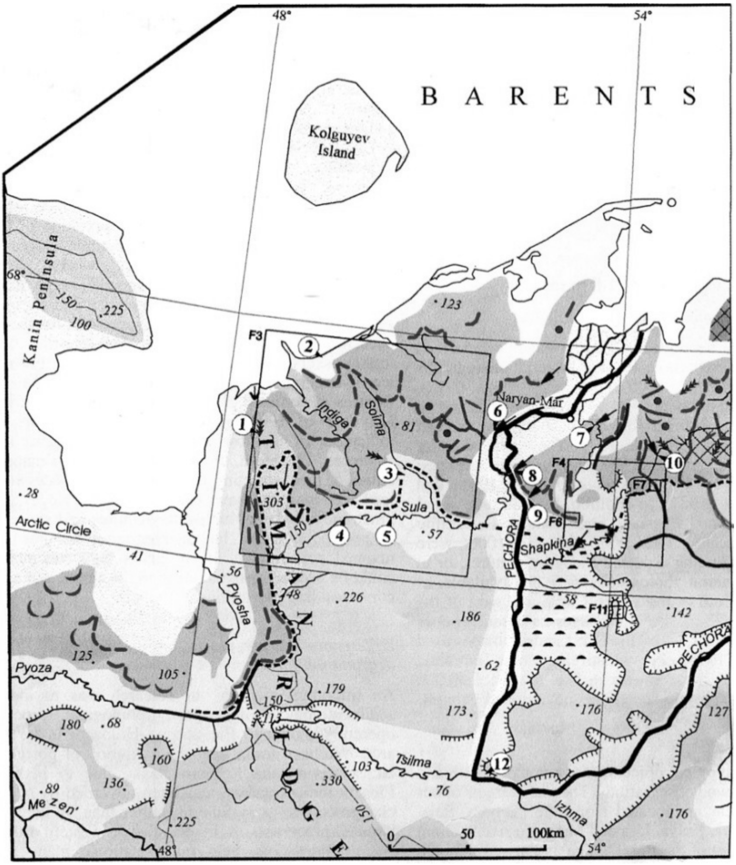

Fig. 2. Map of glacial and periglacial features. Non-studied areas are shown in grey shades, depending on altitudes. Rectangles with numerals relate to figures in the text. The broken line – inferred ice-sheet limit named the Markhida Line, i.e. the southern boundary of hummock-and-lake landscapes of Markhida, Harbei and Halmer types. Black arrows with circled numerals – key sections in Mangerud et al. (1999): 1 – Harius Lakes; 2 – Timan Beach; 3 – Urdyuzhskaya Viska; 4 – Sula sect. 7; 5 – Sula sect. 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: 21 – Byzovaya; 22 – Pymva-Shor; 23 – Mamontovaya Kurya. Sites on the Yamal Peninsula according to Gataullin and Forman (1997) and Gataullin et al. (1998): 24 – Marresale; 25 – Mutny Mys. See centrefold for this image in colour.

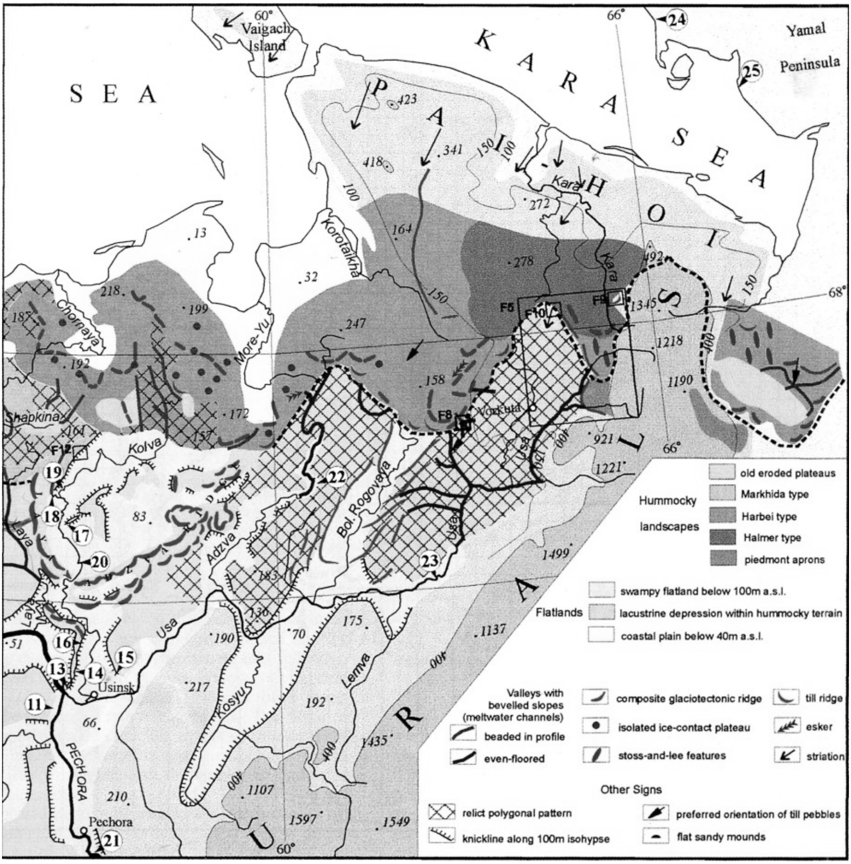

Fig. 2. Continued

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 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 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 of 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 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, features are shown in more detail in the maps of special important areas (Figs. 3-5) 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 generalised version of our original photointerpretation maps compiled at the scale of 1:200 000. Some features are shown in more detail in the maps of special 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 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 consists of 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 (paper 2, Ch. I). 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 (Figs. 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: Markhida type in the west comprising the Varsh, Indiga and Markhida moraines proper, Harbei type in the middle, including the Sopkay moraines east of the Urals, and 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 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.

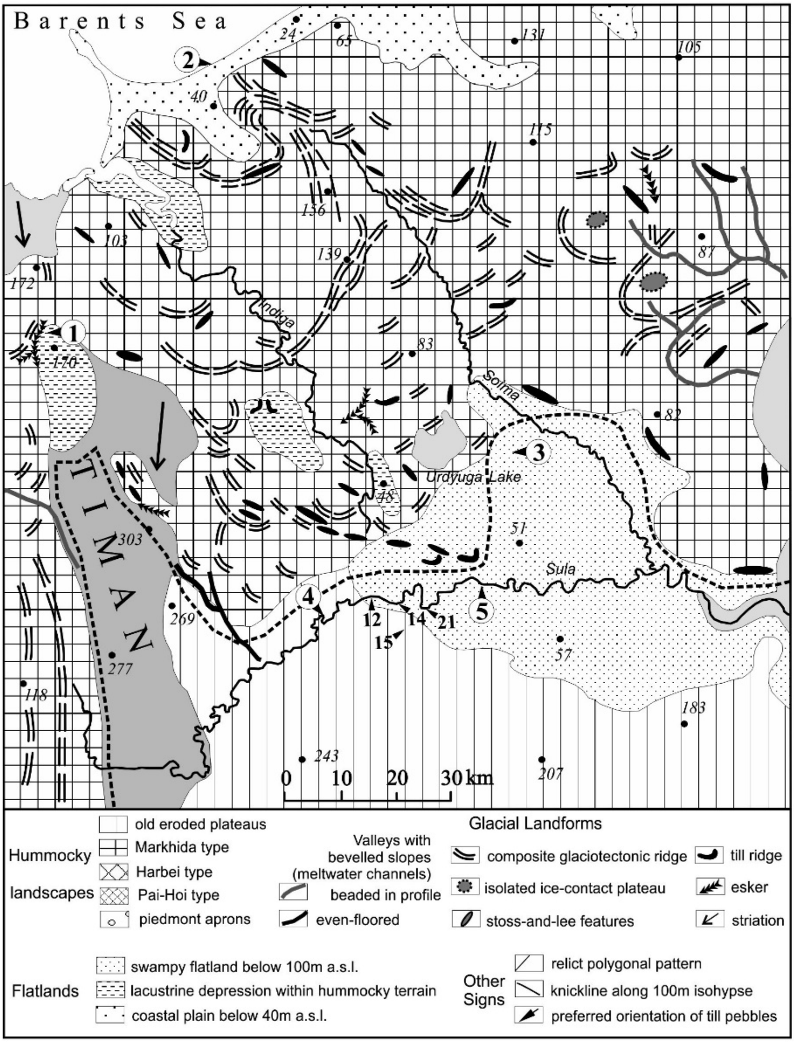

Fig. 3. Photointerpretation map of Timan-Sula area (location in Fig. 2). Circled numerals correspond to the site numbers in Fig. 2. Numbered arrows without circles indicate studied exposures. The broken line is the inferred ice-sheet limit corresponding to the Markhida Line.

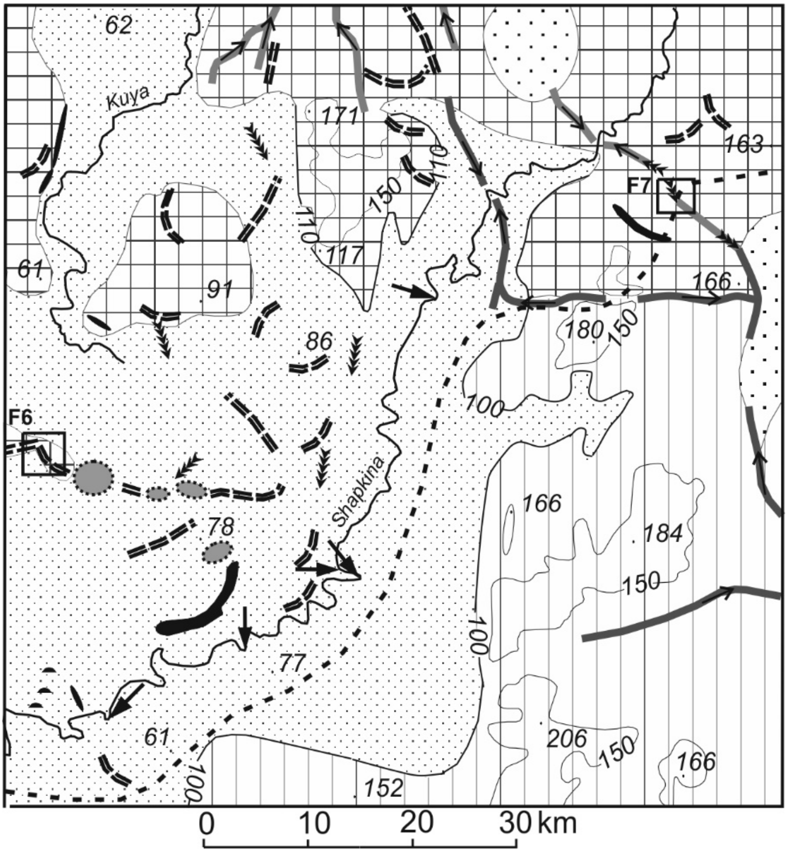

Fig. 4. Photointerpretation map of Shapkina river valley (see location in Fig. 2). Explanation is in Fig. 3. Rectangles are locations of Figs. 6 and 7. The broken line is the inferred ice-sheet limit corresponding to the Markhida Line. Note the mapped shorelines about 100 m a.s.l. that seem to be incised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

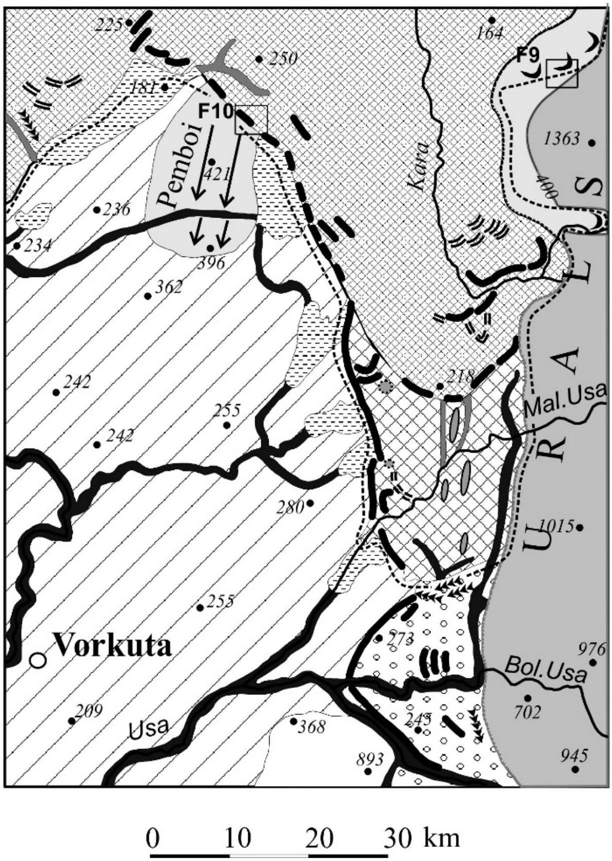

Fig. 5. Photointerpretation map of western slope of Polar Urals (see location in Fig. 2). Explanation is in Fig. 3. Rectangles are locations of Figs. 9 and 10. Note that the area inside the reconstructed ice-sheet lobe along the Urals includes two types of landscape, the Halmer type in the north and 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 from the north.

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) and occur either in confined depressions or atop the higher plateaus.

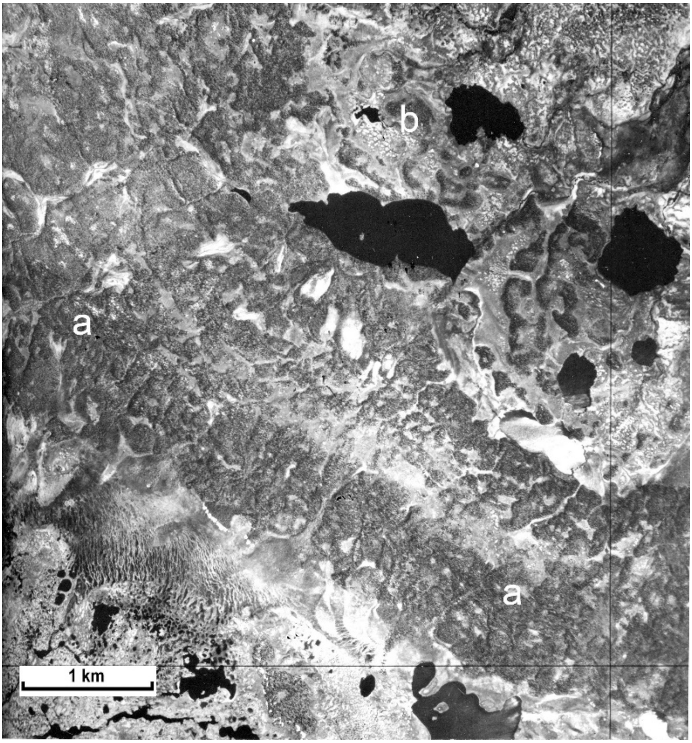

Fig. 6. Aerial photograph of the Markhida marginal ridge (see location in Figs. 2 and 4). (a) – faintly expressed parallel ridges along the forested distal slope. The flat crest of the ridge at 70-80 m a.s.1. is possibly eroded by a proglacial lake; (b) – swampy flatland at 30-40 m a.s.l., the deepest part of the proglacial Lake Komi.

The most expressive glacial features 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,

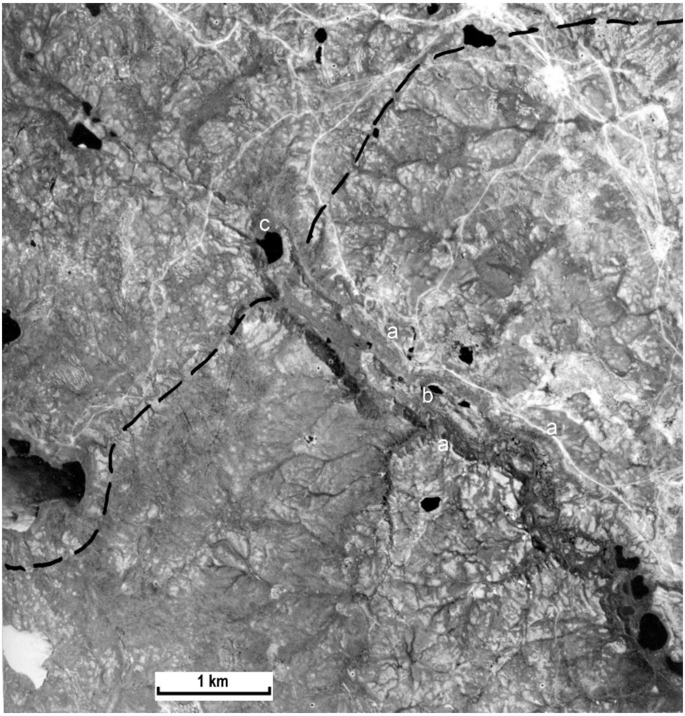

Fig. 7. Aerial photograph of the Markhida Line (dashed) at the left bank of river Shapkina (see location in Figs. 2 and 4). A plateau built of pre-Weichselian tills at 140-160 m a.s.l. east of the Markhida Line is cut by a dry valley, 30-40 m deep, with bevelled slopes. The valley is a continuation of 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 northwest.

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 (paper 2, Ch. I) and also by the lack of mantling diamicton on the flattened surfaces of large composite ridges. Where the mantling diamicton 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 being obscured by a solifluction mantle (Fig. 6).

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 aeolian sand (Mangerud et al., 1999).

Flat swampy depressions or lakes often occur under the proximal slopes of composite ridges constituting hill-hole pairs typical for glaciotectonism (Levkov, 1980). The glaciotectonic origin of the 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 the NE (Tveranger et al., 1998). 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 Siberia (papers 2 in Ch. I and 8 in Ch. II). In West Siberia their occurrence at hundreds of kilometres 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 diamicton, 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 SW-NE orientated valley of the lower Pechora River follows such a streamlined relief, reflecting ice movement from the NE. Here the normal thickness of the basal till overlying Eemian sand is 2 to 5, 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 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 in composite ridges up to 100 m a.s.l. 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, 1998b). 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 of 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 upon all topographic elements above floodplains. 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 River Pechora north of 67.5° N, are evidence of former 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, sometimes marine mollusk 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 ventifacts are common decorations on the sandy surfaces along the large glaciotectonic ridges (paper 2, Ch. I).

Hummocky morainic landscape of Harbei type

East of the Markhida moraines, beyond the catchment area of the River Pechora, 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 an appearance younger than the Markhida type. Deep glaciokarst lakes of intricate configuration and of various sizes are characteristic for this landscape, although shallow thermokarst ponds, sometimes surrounded by boggy hasyreis, are also common. The density of small (1-6 hectares) lakes is estimated to 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 Harbei-To Lake (Fig. 8). All Harbei type terrains west of the Urals are underlain by 300 to 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 surrounding flats). Chains of such ridges are mostly parallel to the southern boundary of the Harbei landscape, outlining N-S oriented lobate basins (Fig. 2). One such chain, probably a retreat moraine, can be traced for 80 km in a SE-NW direction between rivers More-Yu and Chornaya. 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 Paleozoic structures, is visible in satellite images of the south-western Pai-Hoi (Fig. 2). The ridge, which is 80 km long and up to 200 m wide, is striking N-S, 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 to 250 m a.s.l.) near the source of river Adzva (Vashutkiny Lakes), where they mark the southern boundary of the Harbei landscape.

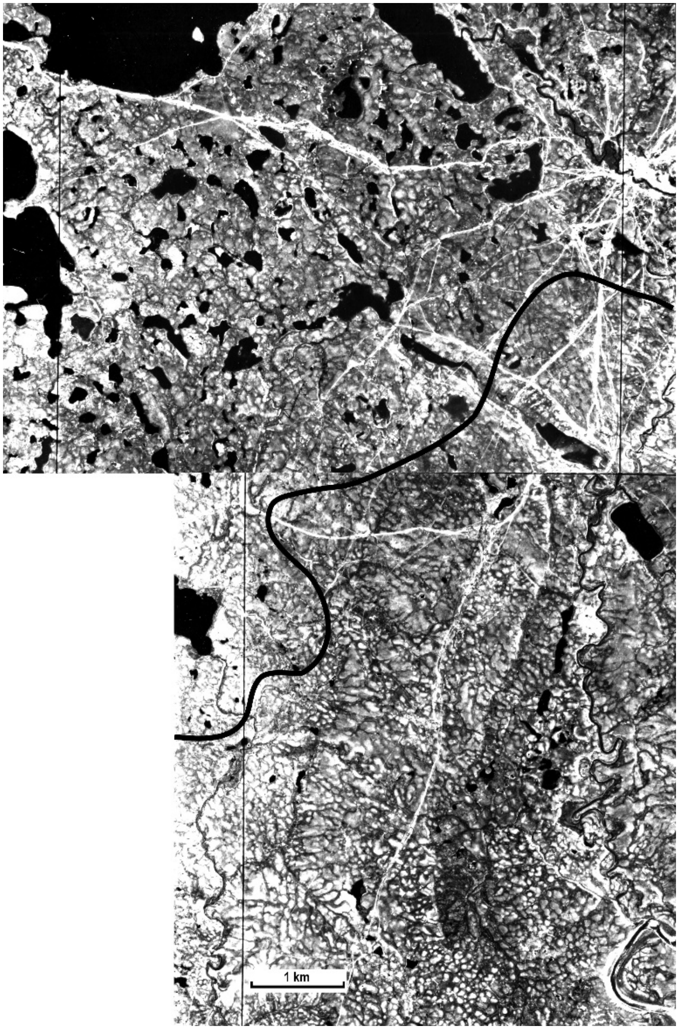

Fig. 8. Aerial photomosaic of the Markhida Line (solid) south of Harbei-To Lake (location in Fig. 2). Glaciokarst landscape in the upper left part of the picture consists of small morainic hummocks and kames interspersed with deep lakes (black) of intricate configuration; the periglacial landscape to the east and at the bottom of picture is shaped by eroded permafrost polygons. Proglacial drainage system left the beaded system of long SE-orientated lakes; they change across the ice limit to NW into a subglacial channel with an uneven floor marked by minor elongated lakes.

More characteristic for the Harbei landscape are isolated oval or round table-like plateaus 30 to 40 m high and 30 to 35 km^2 in area, 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 as glaciolacustrine ice-contact features that originated amongst fields of stagnant glacier ice at latest stages of ice sheet disintegration.

The thickness of glacial sediments overlying the Eemian marine formation may reach 80 to 120 m on higher plateaus (180 to 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 to 2 m) layers of ice/diamicton mixture with the strongly preferred northeast orientation of numerous angular and wedge-shaped pebbles (paper 19, Ch. III).

The ice is covered by a diamicton of similar composition, with a distinct thaw contact, which is interpreted as melt-out till. In appearance and structure this banded ice is similar to the fossil glacier ice described in Siberia (papers 5 and 8 in Ch. II), 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 Ridge, are actually remnants of stagnant glacial ice surviving within the thick Pleistocene permafrost (paper 19, Ch. III). 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 (paper 8 in Ch. II).

East of the Urals a hummocky landscape of the Harbei type is represented by the Sopkay moraines (paper 2, Ch. I) including the NW-SE 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 Paleozoic limestones (Fig. 2). Solifluction flows occur on long clayey slopes, but they do not obscure the glacial hummocky relief.

The lithologically diverse substrate is very uneven, which favoured formation of a large intra-morainic 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 to 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 south-western Kara Sea (paper 2, Ch. I).

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 hard basement of Permian and Triassic conglomerates. These moraines are named after river Halmer-Yu. 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 (paper 5 in Ch. II).

The Halmer moraines, outlining a distinct lobate basin with a N-S axis along the upper river Kara, have sharp boundaries with the surrounding bedrock uplands (Figs. 2 and 5). The bedrock frame of the lobe in the east is represented by the frontal escarpment of the Urals with N-S orientated 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 marginal ridge 200 to 250 m wide and 15 to 20 m high 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 Permian-Triassic sedimentary rocks are most probably the principal source of stones in the Halmer till of the western foothills but 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 north-western 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 to 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. (Fig. 9, a) has an asymmetric profile with the steepest (35°) slope along its north-orientated 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 occurs *in situ* only at lower altitudes to the north of the ridge. Local down-valley orientated end moraines are only noted higher than 600 m a.s.l. and always very close to the present day glaciers.

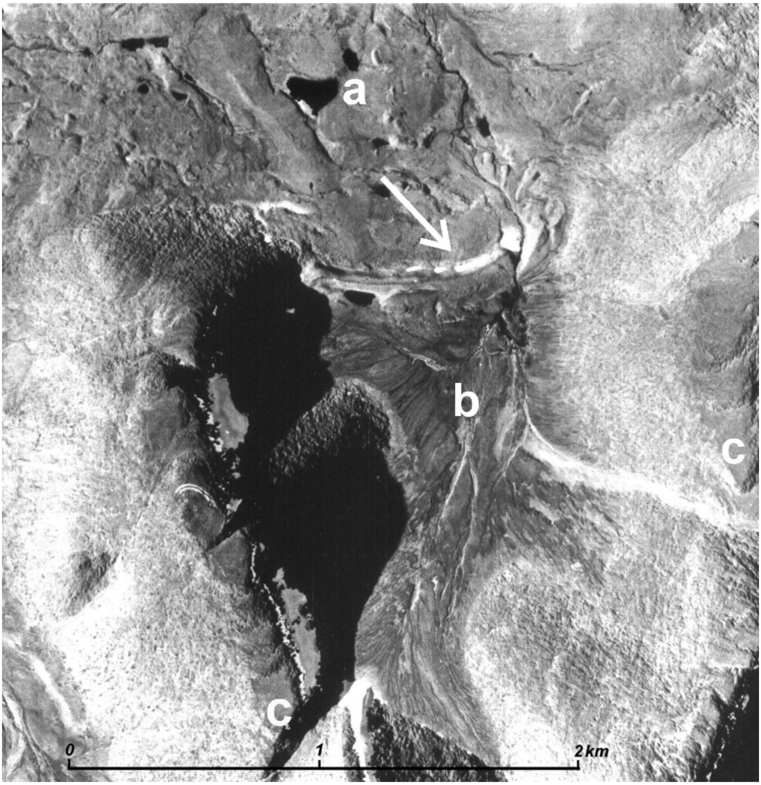

Fig. 9. Aerial photograph of north-western margin of Polar Urals (location in Figs. 2 and 5). (a) – morainic ensemble inserted into a horseshoe-shaped marginal ridge (arrow) is located at 560 m a.s.1.; (b) – flat valley bottom devoid of moraines. (c) – cryoplanation terraces on unglaciated summits at 700-800 m a.s.1.

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 south-western 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 south-westwards from the Baydarata Estuary coast (Voronov, 1951).

Fig. 10. Aerial photograph of Halmer landscape in western foothills of Polar Urals (location in Figs. 2 and 5). (a) – Pemboi Plateau built of gently dipping Permian conglomerates at 200-260 m a.s.1. is heavily striated by an older ice flow from the NNE. Extremely fresh glaciokarst landscape (b) is separated from river Halmer-Yu by a narrow ridge (black arrows) at 230 m a.s.1. and 30 m above the marginal chain of elongated lakes

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 W-E, pointing to the upper Usa valley, whereas the axis of the Halmer lobe is N-S and parallel to the Urals (Figs. 2 and 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 to 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 down river Usa 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 to 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, shist and other rocks of the central Ural Mountains. These materials are much coarser and of different provenance than that 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 (paper 2 in Ch. I). 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 and 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 orientated 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 as 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 W-E striking eskers. The end of the N-S striking morainic lobe grows oblate, and when widening, issues westwards three small (0.5 to 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. 7–9). 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 N-S striking flat-topped ridge, 10 to15 m high and 100 to 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 upper river Shapkina close to the ice margin (Figs. 2 and 4). Some of them are situated in straight dry valleys with uneven floors and beaded profiles, characteristic for 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, independent of the present day drainage, is the most reliable morphological combination for delineating the ice-sheet margin on uplands (Figs. 7 and 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 NE (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 to 5 km wide inner ridge with the proximal slope 20-30° steep rising up 40 to 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 river Kolva 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 to 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 river Rogovaya (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 such ridges undercut by the Seyda River we observed a clayey till with strong NE fabric, some 40 m of apparent thickness. The long Rogovaya morainic loop, parallel to the Urals, is important evidence of 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 to 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 floodplain. The tightly spaced valley network occurs on a variety of substrata, including folded Paleozoic rocks of the Timan Ridge. Modern, 10 to 100 m thick surficial permafrost occurs in the tundra of the river Usa 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 as correlating with the Moscow (Saalian) glaciation. However, Arslanov et al. (1987) and Grosswald (1993) draw their Weichselian ice limit within this area (Fig. 1). Our observations show that the upper till of the southern lakeless plateau along 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 the River Pechora 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 aeolian 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 aeolian 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, dune sand 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 for 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 to 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 the East Siberian type (Popov, 1962a). In the present day climate only small ice wedges grow in the north near the coast (Yershov, 1988).

The former ice-wedge cracks are mostly filled with loess-like silts or soliflucted diamicts. 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 till (Kaletskaya, 1962). We have observed that this diamicton in some places 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 the relict permafrost polygons (Popov, 1962a).

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 getting obliterated by solifluction and aeolian sediments and are, therefore, not easily recognisable in aerial photographs, especially in boreal forests.

Radiocarbon dates from organic infills in large ice-wedge casts along rivers Shapkina and Kuya 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.

Another set of features interpreted by us as the result of former permafrost are regularly spaced sandy mounds on the unfrozen low flatlands (40 to 60 m a.s.l.) and on the River Pechora terraces to the south of river Shapkina (Fig. 2). These oval or pancake-shaped mounds are normally several hundred metres across, less than 10 m high and weakly elongated in WNW 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 as reflecting 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, 1998a, 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, 1962a) 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 recognised in aerial photographs by numerous light-coloured spots of present-day blowouts stripped of vegetation. The sand occurs both in the form of dunes and as mantle-like cover sand. It normally consists of 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 River Pechora: 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 along the major rivers there are tens of kilometres of wide swampy flatlands. The flatlands 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 (from 90 to 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 and 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 river Kolva (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 to 1.5 m) has been observed in a northernmost pit on river Haryaha, 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 originated as 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 river Kolva. Commonly the floor of Lake Komi, when exposed at altitudes of 70 to 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 the upstream 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 Paleozoic 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. Knicklines can be followed across this watershed 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 river Pyoza, a knickline, seen on high-resolution satellite images, persists along the 110 m isohypse, in places highlighted by sand bars. We therefore conclude that Lake Komi continued across the described watershed to coalesce with a similar reservoir in the river Mezen catchment area.

East of the River Pechora a very pronounced knickline along the 100 m isohypse can be traced continuously from the village of Ust-Tsilma (12 in Fig. 2) to river Shapkina. In places it outlines some narrow headlands of older till and outwash massifs protruding into the proglacial lake (Fig. 11). North of river Shapkina a knickline at 110 m a.s.l. intrudes into the Markhida landscape by embayments with outliers of morainic plateaus

(Figs. 2 and 4). Farther northwards the former shorelines disappear being replaced by hummocky terrains with lower altitudes.

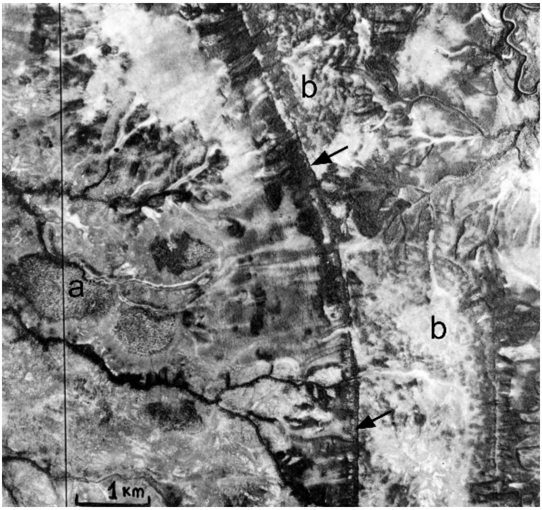

Fig. 11. Aerial photograph taken south of river Sozva in central part of Pechora Basin (location in Fig. 2). A well-expressed shoreline, traced along the 100 m isohypse by a narrow sand bar (arrows), separates swampy plain at 85 to 95 m a.s.l. from (a) pre-Weichselian glaciofluvial plateau at 105 to 120 m a.s.l. (b) 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.

Strandlines are well developed inside the Laya-Adzva moraines (Fig. 2). They penetrate far northwards along river Kolva, 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 river Rogovaya (Fig. 2). Here, the Harbei moraines at 130 to 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. A gravelly infill of a meltwater channel seems to cut into the lake bottom at 80-90 m a.s.l. along the upper Adzva river, incised into a higher plateau. This means that meltwater discharge from

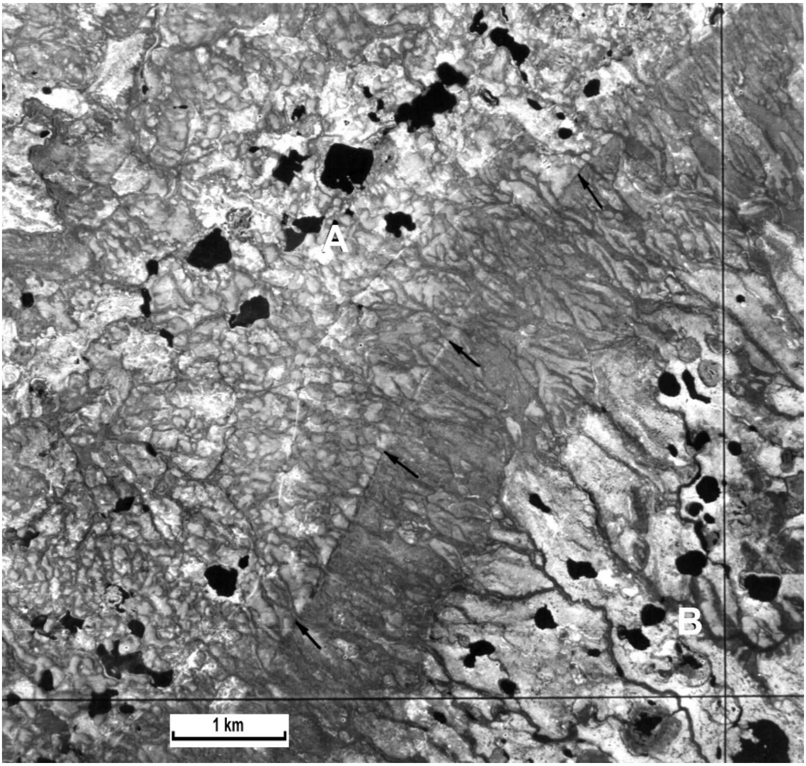

Fig. 12. Aerial photographs of Lake Komi shoreline at 110 m a.s.l. indicated by arrows north of Haryaha River inside the Laya-Adzva morainic ridge (location in Fig. 2). (A) – hummocks-and-lakes landscape of Markhida type on a plateau at 140-150 m a.s.l.; lakes of intricate shape are often encircled by dry boggy platforms; note solifluction streams across the former shoreline. (B) – former floor of Lake Komi at 90 m a.s.1. with shallow thermokarst lakes and mature hasyreis.

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 floor 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 (paper 2, Ch. I). 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 of 13 to 12 ka BP from silty rhythmites of a drained lake on the west-east stretch of the More-Yu River (paper 19, Ch. IV).

Coastal lowland

Along the Barents Sea coastline there is a flat seawards inclined lowland at altitudes 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 the Eemian marine sediments. Moreover, the flatland gradually climbs inlands and intrudes into the morainic landscape by long sandy tongues which in places rise up to 60 m a.s.l. On the western shore of the Pechora Estuary there are some narrow, 10 to 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 thin diamicton overlying an interglacial marine formation and covered by sand (Lavrushin et al., 1989). Along the lower stretch of the river More-Yu thick aeolian 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 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 and 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 and 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 rivers Shapkina and More-Yu (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 3 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 most fresh 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 and 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 River Yenissei in Siberia, where fresh morainic landscapes are developed over the Early Weichselian basal till with large stratified bodies of fossil glacier ice (paper 5, Ch. II). 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 river Kolva 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 localised 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 NE (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 aeolian silts (12.2 to 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 and 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 of the hummocky moraines where buried glacier ice melted over a considerable span of time. Two alluvial terraces, which along with the floodplain 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 Paleolithic 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 (*ibid.).*

So far we have not been able to identify spillways from Lake Komi. One possibility, which 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 north-western 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 south-westward 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. 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 were 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 shallow Kara and Barents Seas shelves. 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 Rogovaya 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 of 1993 to 1998, were funded by the Research Council of Norway. From 1998 on 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 has been 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 field work. We appreciate the criticism by Steve Forman, which helped to improve the text. To all these persons and institutions we offer our sincere thanks.

19. Age of Remnants of a Pleistocene Glacier in Bolshezemelskaya Tundra

V. I. Astakhov and J. I. Svendsen

Doklady Earth Sciences, 2002, Vol. 384(4), P. 468–472.
Translated from Doklady Akademii Nauk, 2002, vol. 384(4), P. 534–538.

New data on geological structure and geochronometry of Quaternary sediments, containing the first find of relict glacier ice in European Russia, are discussed below to constrain the geological age of the last inland glaciation of the arctic plains.

Relict glacier ice in perennially frozen Siberian lowlands, known since the 70-s (Kaplyanskaya & Tarnogradsky, 1976; Solomatin, 1977)], has been comprehensively studied (Astakhov, 1986; Karpov, 1986; Solomatin, 1986; Astakhov & Isayeva, 1988). However, in European Russia buried glacier ice was reported only from arctic islands with present day glaciers (Novaya Zemlya). Known cases of distorted massive ground ice found in boreholes of Bolshezemelskaya Tundra were attributed to recent injections of ground water (Oberman, 1988).

During the field work of August 1998 massive ground ice, not appreciably different from repeatedly described relict glaciers of West Siberia, was found in the middle course of river More-Yu (Fig. 1) within the sub-latudinal zone of hummock-and-lake landscapes called Harbei moraines (Astakhov et al., 1999). The massive ice occurs at the base of river bluffs 20-30 m high beneath a thick diamicton called the `Rogovaya Formation` by Vorkuta geologists (Zarkhidze, 1972), or the Late Valdai basal till by Moscow photogeologists (Arslanov et al., 1987). The ground ice in the end summer is covered by thick diamict slides, and its existence can only be suspected from fresh thermoerosion scars and from tales by travelers who noticed the ice immediately after spring floods. In August 1998 only massive ice in section 14 (Fig. 1, 2) was readily observable.

The exposed ice was located in the centre of a large overgrown semi-circular hollow cut into a diamicton plateau with salients of dislocated sands. Two lower diamictic hillocks, segment-like in plan and descending stepwise in profile (sect. 14 in Fig. 3), are concentrically inserted into the main cirque-like hollow, their arcuate bases running parallel to the brow of the plateau. The chord of the inner diamictic segment is the linear river bluff 30 m high with massive ice at the base. This topographic ensemble makes one suspect a sliding origin for the bulk of the diamict mass atop of the ice deposit. Large-scale slides of frozen rocks usually develop over a particularly icy substrate, which would suggest rising of the ice surface towards the plateau.

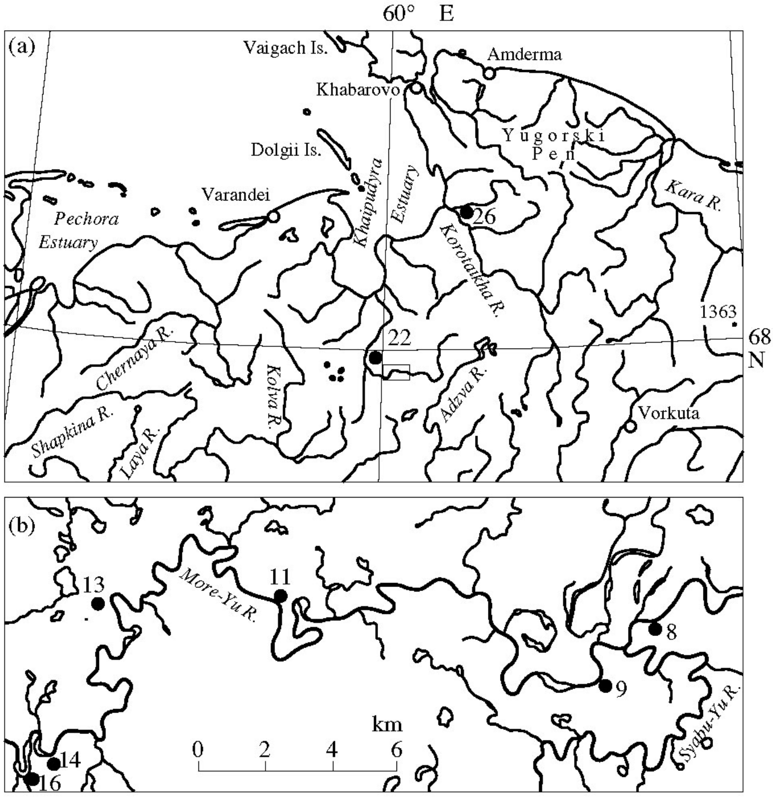

Fig. 1. Location maps. Numbered dots are mentioned sections. (a) Overview map. The rectangle in Fig. 1a shows location of Fig. 1b, which is the map of the middle course of river More-Yu r.

Only a part of the ice body cleaned by meltwater from 6 m above the river was accessible for study. The cleaner part appeared as a concave slope 12 m long and 4 to 7 m wide. The exposure is actually much larger, but in the end of the melting season most of it was covered by fast creeping bluish-grey mud. The icy surface has a striped appearance due to alternating bands of transparent, glassy, slightly bubbly ice (1 to 10 cm thick) and thinner (0.5 to 5 cm) clayey dirt bands. The transparent ice contains silt and clay particles suspended. Slightly bent ice bands in places wedge out laterally being truncated by overlying bands, which is a good signature of glacial foliation. Downslope the dirt bands thicken and merge into a layer of massive gritty ice/debris mixture with concordant lenticular boudins of transparent ice 0.5-1 cm thick.

Fig. 2. Massive ice at the base of section 14 (Figs. 1, 3). (a) Ice/debris mixture, (b) alternating bands of transparent and dirty ice with suspended cobbles; (c) gray diamicton; (d) dark gray clayey diamicton.

The ice/debris mixture is rich in angular, often wedge-shaped pebbles up to 2 cm long. They are less numerous in the overlying banded ice, where angular cobbles 0.1-0.2 m long occur as well (b in Fig. 2). The surface of the dirty ice looks pitted due to small melt-out dimples containing splinters of pure clay. Apart of the exceptionally high ice content, the composition of the ice/debris mixture is not much different from the overlying grey diamicton. No traces of injection contacts have been noticed.

Measured ice bands dip at 30, 40, 45 and 55° to 185, 170, 160 and 145° respectively, i.e. they are steeply inclined to S and SSE. Pebbles consist of grey slates, sandstones and limestones, common for the Pai-Hoi Paleozoic, and of Quaternary clays. 3 fragments of shells were also noticed. Directions of elongated pebbles are 65, 35, 65, 70, 75, 55, 30, 55, 355, 70, 70, 25, 350, 275, 50, 360, 15, 70 and 30°, i.e. 12 of 20 measurements showed axes of the NE-SW strike, coinciding with the general direction of ice flow inferred from other evidence (paper 18, Ch. IV).

Isotopic composition of this ultra-freshwater ice is characterised by measurements of $\delta^{18}O$ from 3 varieties of the ice equalling -20, -20.4 and -20.4‰, whereas water from a nearby creek is isotopically heavier: $\delta^{18}O$=-13‰. The abnormally light isotopic composition of the massive ice suggests its atmospheric origin in a very continental climate (of present-day Taimyr or Pleistocene type), and the creek water is isotopically closer to modern ground water of the Pechora basin, whose $\delta^{18}O$ is -12 to -15‰ (Vasilchuk & Kotlyakov, 2000). These results repeat the data from the type site of relict glacier ice on the Yenissei called the `Ice Hill` (Vaikmae et al., 1993).

Glacial origin for the dirty massive ice is further supported by its position within the perennially frozen Quaternary formations. Grey diamicton, 1.5-2 m of apparent thickness in a vertical wall (c in Fig. 2), is composed of the same materials as mineral inclusions in the ice and conformably overlies the thaw surface of the ice body. This diamicton in places displays internal cabbage-like structures formed by folded seams of coarse sand. Such structures were previously recorded in freshly melted out diamicton at Harasavey, western Yamal, where they copy hinges of recumbent folds in nearby relict glacier ice. The ice and grey diamicton in sect. 14 are unconformably overlain by a thick, more dark and clayey diamicton traceable far upslope (d in Fig. 2).

Similar diamictons with orientated pebbles and large detached blocks of waterlain sand, obviously varieties of basal till, are exposed all along the river. The parental sand can be seen in sect. 9 (Fig. 3), where a sheared zone, composed of diamict and sand bands alternating, at the base of till rests atop of gently folded well-washed quartz sand with lenses of fine gravel and abundant shells of marine mollusks.

In sect. 11 the same sand with boreal mollusk fauna and beds of gravel overlies the thick grey diamicton. Zarkhidze (1972), who thought that all

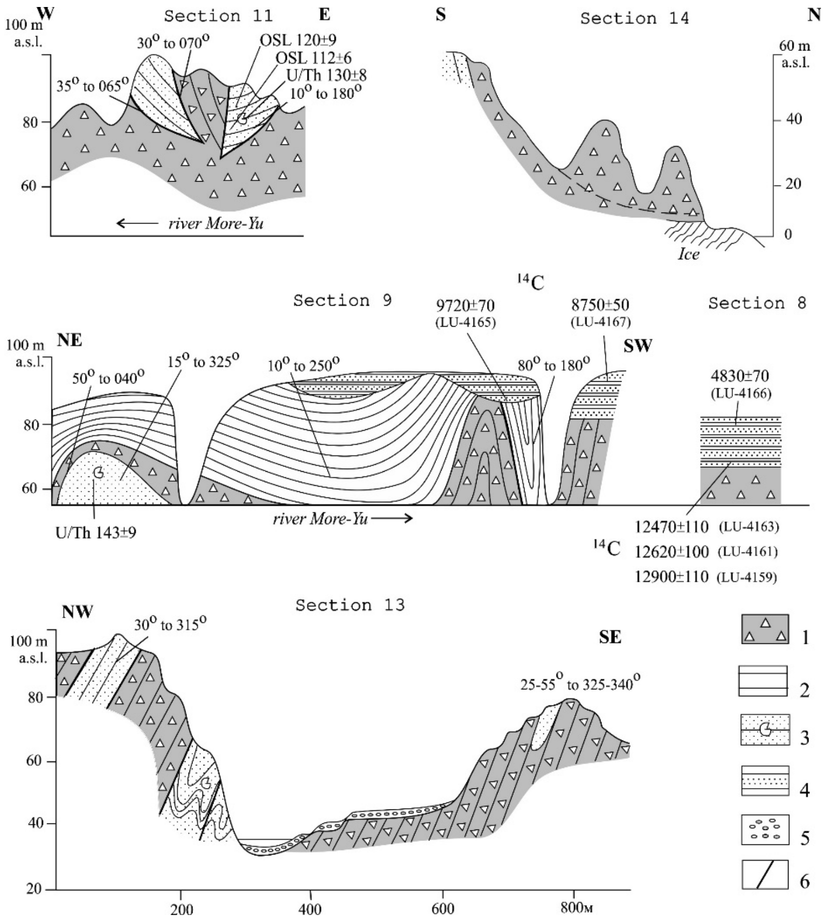

Fig. 3. Sections of Quaternary sediments in the middle course of river More-Yu (complemented from paper 9, Ch. II; for location see Fig. 1b). OSL and U/Th ages are indicated in kiloyears (ka); radiocarbon ages – in radiocarbon years B.P. (1) glacial diamicton; (2) glaciolacustrine silty rhythmite; (3) marine sand with shells; (4) limnic sand with driftwood; (5) fluvial gravel; (6) fault.

sediments of this area were marine, assigned this sand to `More-Yu Formation` of Eemian age, while relating the sand beneath the diamicton to `Padimei Formation` of the late Middle Pleistocene. However, tracing and measuring contacts in sect. 11 revealed no horizontally lying beds, all sand/diamicton contacts being faults. This means the large blocks of marine sand were squeezed up along thrust planes into distorted diamicton (Fig. 3). Along More-Yu only postglacial sediments lie horizontally truncating the upper glacial complex at an angular unconformity (sect. 8 and 9 in Fig. 3).

Interglacial sands are readily traceable from river bluffs onto the interfluvial plateau, where outcrops of steeply dipping marine sands stick out as linear ridges at 95–140 m a.s.l. (sect. 13 in Fig. 3). In plan the ridges appear slightly convex towards SSW upstream of sect. 13 and towards SE farther downstream, reflecting directions of glacial stress. River More-Yu, following the strike of the sub-latitudinal zone of glaciotectonic compression with buried basal ice, could originate as a marginal channel during the ice sheet disintegration. Salients of dislocated interglacial sand have also been uncovered by shallow pits dug on the eastern plateau 170-180 m a.s.l., close to Vashutkiny Lakes, where they were first described by Zarkhidze as an individual `Vashutkiny Formation` of the final Middle Pleistocene (1972).

Age of the marine sands can be assessed from U/Th datings on mollusk shells. Taking into account the scattering typical for this method, the values from beneath the till (sect. 9) and atop of it (sect. 11) probably refer to the same age. Also the right sand block in sect. 11 has yielded optically stimulated luminescence (OSL) dates: 112 ± 6 and 120 ± 9 ka at 30 and 35 m above the river respectively (Fig. 3). Thus, both geochronometric methods concurrently indicate an Eemian age of distorted marine sands, probably synchronous to substage 5e of the oceanic isotope time scale.

From these data we infer that an individuality of the mentioned sandy `formations` (Zarkhidze, 1972), is an artefact stemming from the invalid presumption of undisturbed stratification. These `formations` were probably carved out of sediments of the single Late Pleistocene interglacial transgression deformed by a subsequent advance of inland ice.

The Late Valdai age was previously suggested for this ice advance (Arslanov et al., 1987). Lately the main glacial event of the Pechora Basin was driven back to the very beginning of the last ice age on the strength of OSL dates in the range 76–93 ka obtained from sediments of ice-dammed Lake Komi (Mangerud et al., 1999). A minimum age of this glaciation can be estimated from the following data.

The upper till is overlain by horizontally lying sediments of postglacial lakes previously related to the Holocene (Arslanov et al., 1987). A series of radiocarbon dates on wood from our collection has shown that active thermokarst and formation of large lakes in this area started not later than 13 radiocarbon kiloyears BP (sect. 8 in Fig. 3). Older postglacial sediments are represented by aeolian sand, which is the best material for luminescence dating. Such sand, up to 15 m thick, in sect. 22 on river Ngutayaha, 10 km to the NW from sect. 16 (Fig. 1), has yielded 4 OSL dates 26 ± 3, 26 ± 3, 26 ± 3 and 21.8 ± 1.4 ka upwards in the succession. A mammoth bone lying nearby was dated by radiocarbon to 39,320 ± 960 years BP (LU-4178). Six OSL dates in the range 16–13 ka have been obtained from an aeolian formation in sect. 26 on river Korotaikha (Fig. 1, A). Two mammoth teeth picked up at the base of this section gave AMS dates 34,600 ± 1300 (Ua-14891) and 35,150 ± 2030 radiocarbon years BP (Ua-14890). Mammoth teeth are notably fragile material that cannot stand multiple redepositions, even less likely in pairs; in this case they certainly belong to the sediments capping the till. The oldest postglacial ages are provided by two OSL dates, both 46 ± 3 ka, from limnic sand in sect. 16 (Fig. 3) at 26 and 27 m above river More-Yu.

Thus, geochronometric data from the eastern Bolshezemelskaya Tundra indicate that the last glacier, which left fragments of its dirty sole on river More-Yu, disintegrated not later than 50 ka BP, which coincides with the results in Yenissei Siberia (Astakhov, 1986; Astakhov & Isayeva, 1988) and with the modern estimates of the age of the last glaciation obtained from other evidence in the Pechora Basin (Astakhov et al., 1999; Mangerud et al., 1999). The maximum expansion of the arctic ice sheet in European Russia is probably determined by OSL dates in the range 80–100 ka on sediments of the highest stand of the proglacial lake (Mangerud et al., 2001). The thick monolithic permafrost of Bolshezemelskaya Tundra must have already formed by this time, most likely in the course of the Boreal Sea regression.

The material discussed above is a result of the Russian-Norwegian project PECHORA funded by Norwegian Research Council and coordinated by the program QUEEN of European Science Foundation. OSL dates were obtained by A. Murray, Nordic Laboratory of Luminescence Dating, Denmark, using optically stimulated luminescence (Murray & Wintle, 2000). Conventional radiocarbon (index LU-) and uranium/thorium datings were performed in Laboratory of Geochronology, St. Petersburg University, supervised by Kh. Arslanov. AMS dates (index Ua-) were obtained in Laboratory of Uppsala University, Sweden. R. Vaikmäe, Institute of Geology, Estonian Academy of Sciences, determined oxygen isotope ratios. Authors sincerely thank all mentioned persons and institutions.

20. Late Quaternary Glaciation of the Northern Urals: A Review and New Observations

Valery I. Astakhov

Boreas, 2018, Vol. 47, P. 379–389

Introduction

The Ural Mountains separating Europe and Asia play a crucial role in many Quaternary issues relevant to northern Eurasia. Acting as a major orographic barrier for the westerly transfer of moisture, this mountain system has posed problems for correlating East European and Siberian Quaternary records as a consequence of (i) the wide gaps in the sedimentary succession in the mountains, and (ii) the radical changes in floristic composition, with the absence of standard indicators of European interglacial climates (broad-leaved trees) beyond the Urals precluding the reliable correlation of pollen diagrams. The lack of interglacial sequences coupled with the scarcity of radiometric data has led to significant ambiguities in attempts to reconstruct the Quaternary history of the Urals. The resultant disparate assessments of both the extent and age of Pleistocene glaciations has in turn significantly contributed to persistent miscorrelations of the stratigraphic time scales between the Russian and West Siberian plains.

However, the northern part of the mountain system provides distinctive landscape evidence of interaction between glacial and periglacial phenomena as well as clear indicators for former glaciation in the form of trimlines and erratic materials at high altitudes that are absent in the adjacent lowlands. This evidence is partly published in Russian (e.g. Boch & Krasnov, 1946; Sirin, 1947; Gesse et al., 1963; Troitsky et al., 1966; Astakhov, 1974; Chernov, 1974a; Surova et al., 1974) and can be found within National Geological Survey reports on various mapping projects. Observations for the Uralian Quaternary glaciation collected since the 19th century have been previously summarized within the stratigraphical compendium of Yakovlev (1956), although this standard reference work has subsequently been challenged by various researchers. The main idea maintained by the majority of investigators is that during Quaternary glaciations the Ural Mountains was a major source of inland ice, which presumably dispersed westwards and eastwards (e.g. Ganeshin. 1973; Chernov, 1974a; Sukhorukova et al., 1987). This conventional concept is still employed in some modern works including new geological maps (e.g. Andreicheva, 2002; Shishkin, 2007).

The last two decades have seen an influx of geochronometric and remote sensing data provided by the application of modern techniques in a series of international research projects undertaken in the Russian Arctic (Mangerud et al., 1999, 2004, 2008; Astakhov, 2004, 2013; Svendsen et al., 2004). These data focus on the late Quaternary events as expressed in the modern topography, namely the penultimate and most recent glaciations in Marine Isotope Stage 6 (MIS 6) and MIS 2, respectively. This increase in the availability of data combined with the persistence of ongoing controversies in the assessment of the region's glacial history have provided the stimulus for this attempt to integrate the available morphological and geochronological data to outline a new framework able to facilitate stratigraphical correlations across the continental watershed.

Modern data on the glaciated Polar Urals have recently been summarized by collective efforts. However, this work (Svendsen et al., 2014) is limited to the Late Pleistocene of the area within the Arctic Circle and does not concern adjacent parts of the Urals and older glacial features. The present paper is an attempt to view the modern stratigraphical data in a wider geomorphic context in order to arrive at more comprehensive paleoglaciological conclusions.

Data and methods

This study is performed on the basis of two data sets: cartographic and geochronological. The cartographic analysis aims at paleoglaciological interpretation of the map pattern of glacial features. Their spatial layout is studied using middle- and small-scale geological maps of the Quaternary deposits compiled by the National Geological Survey, and aerial and satellite images interpreted by this author (e.g. Gesse et al., 1963; Astakhov et al., 1999; Shishkin, 2007; Astakhov, 2011). The timing of glacial features is provided by numerous dates obtained from measuring optically stimulated luminescence of quartz grains (OSL method) in the Nordic Laboratory for Luminescence Dating, Risø, Denmark, according to the technique described by Murray & Wintle (2000). The collection of

OSL dates is supported by cosmogenic beryllium-10 dating of large boulders (Mangerud et al., 2008). and by U/Th dating of interglacial peat (paper 12, Ch. III). The dates employed in this study have over the last two decades been reported by members of the Russian-Norwegian research project PECHORA (Mangerud et al., 1999, 2004, 2008; Astakhov, 2004, 2006; Svendsen et al., 2004, 2014; Astakhov et al., 2005; Nazarov et al., 2009).

Geographical setting

The northern Urals *sensu lato* is a longitudinal mountain chain positioned between 60 and 68.5° N. It is traditionally subdivided into the Polar Urals (north of 65°40′ N), Subpolar Urals (64–65°40′ N) and Northern Urals *sensu stricto* (south of 64° N) (Fig. 1). The adjacent lowland located to the west is referred to in Russian as Preduralye, i.e. Cisuralia. The lowlands east of the mountains are referred to as Zauralye (meaning Transuralia) and form part of the West Siberian Plain. The Urals proper, i.e. the exposed Paleozoic folded rocks, are topographically expressed as the Central Uralian Mountains and are bordered by the western and eastern piedmonts or foothills. The dominant ridges of the Central Urals at 1000–1500 m a.s.l. are composed of metamorphic schists, quartzites and igneous rocks whereas the foothills, which are 200–500 m high, comprise Paleozoic folded sedimentary formations. The Urals are dissected by an orthogonal system of river valleys and glacial troughs. The trough-like valleys crossing the narrow portion of the Polar Urals between 66 and 67° N are especially deep with their floors occurring at 200-300 m a.s.l.

Annual precipitation diminishes from 900 to 1000 mm on the western slopes of the range to 500 mm in the eastern piedmont. The difference is especially marked in winter: in the windward western Urals snow cover is 1.2-1.5 m thick whereas on the leeward eastern slope it never exceeds 0.6 m. The boreal forest tree line descends from 700 m a.s.l. at 60° N to 200 m a.s.l. at 66° N (67° N on the eastern slope) to gradually disappear beyond the Arctic Circle. Above the tree line there is a zone of frost weathering resulting in the formation of block fields with distinct stone streams that are located 100-150 m lower on the western slope than on the Siberian slope. The climatic snowline in the western Polar Urals north of 67° N is ~1600 m a.s.l. reaching 2000 m on the drier Siberian slope. More than a hundred small glaciers with areal extents of ~ 0.3 km^2 are scattered along the watershed. They are contained mostly within glacial cirques with east-facing aspects. With none of the glaciers reaching the modern-day snowline, this orientation reflects their reliance upon the accumulation provided by avalanches in these lee-side locations (Troitsky et al., 1966).

Traditional views on former glaciation

Numerous fragments of Uralian Paleozoic rocks found over the adjacent lowlands led to the early idea of ice dispersal centres formed upon the mountains. This paradigm, dominant since the 19th century, has been maintained within the majority of the literary sources. At the same time, field geologists pointed to the existence of scattered erratics of lowland rocks on montane plateau over 1 km high as evidence for the prior occurrence of thick glacier ice. This is the reason why Yakovlev (1956) in his trans-regional overview had to limit the Uralian ice dispersal centre to late Quaternary times although other authors believed in the presence of Uralian ice domes even during the Pleistocene maximum glaciation (e.g. Boch & Krasnov, 1946; Sukhorukova et al., 1987).

To explain the erratic boulders of lowland origin in the mountains, Yakovlev (1956) suggested a remote source of older and thicker ice that presumably flowed southwards from Novaya Zemlya and the Barents Sea shelf. He attributed many Uralian erratics found in the West Siberian Plain to the expanding Novaya Zemlya ice sheet that overrode the mountains up to 1300 m high in mid-Quaternary times. The erratics (Fig. 1) therefore provide the main line of evidence for the presence of thick Middle Pleistocene ice sheets within the European Arctic. Additional evidence for the southward flow of ice uphill from the Arctic shelf was also found in the form of large striae accompanied by Paleozoic blocks transported from the Kara Sea shore to the southwest across the Pai-Hoi Ridge (Voronov, 1951).

In the Northern Urals the clasts from adjacent lowlands and both piedmonts were found at altitudes of up to 500 m a.s.l. To explain their occurrence amongst a background of widely dispersed and abundant Central Uralian clasts, Yakovlev (1956) proposed the development of a separate ice sheet upon the highest and widest mountainous massif of the Subpolar Urals at 64–65° N. Ice streams from this hypothetical Uralian ice dome could then disperse Central Uralian clasts southwards. The southwards diversion of the ice streams from the foothills towards the watershed of the mountain chain was related to the obstructing influence of lowland ice sheets in the north.

The concept of a Middle Pleistocene ice dome in the highest and widest Subpolar Urals based solely on fragments of Uralian resistant rocks such as quartzite and granite scattered over the adjacent plains has persisted in the literature (e.g. Ganeshin, 1976; Chernov, 1974a; Sukhorukova et al., 1987). This conventional idea still influences the development of more recent format

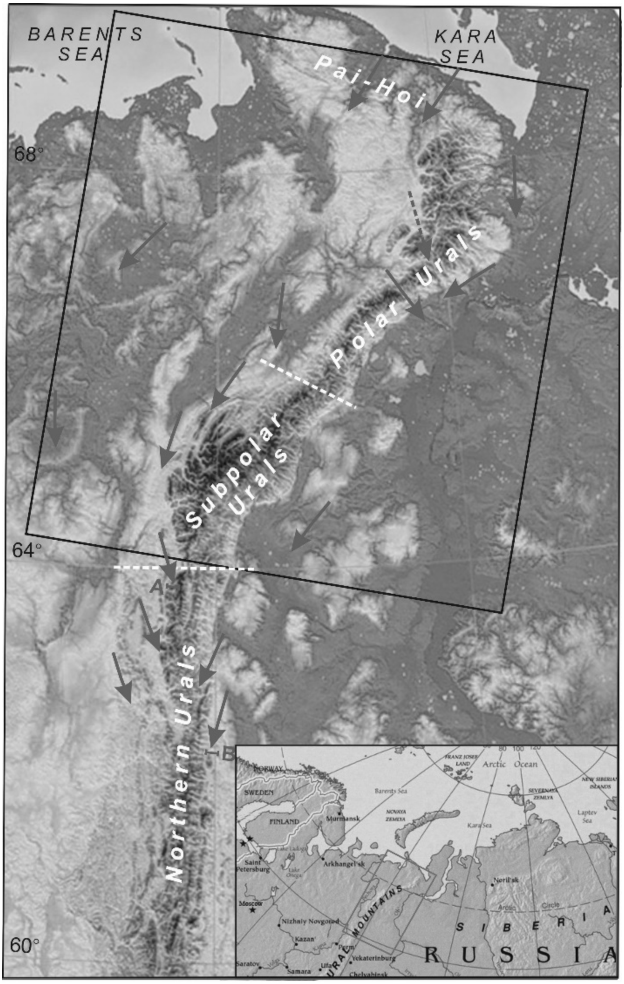

Fig. 1. Orographic subdivisions of the northern Urals with main directions of Middle Pleistocene ice motion by various authors. Broken arrows are ice flows inferred from geomorphic directional features; solid arrows – boulder trains. Short bars A and B indicate profiles in Fig. 2. See centrefold for this image in colour.

geological models including new state geological maps produced in GIS format (e.g. Shishkin, 2007). A Late Pleistocene ice dispersal centre in the Polar Urals has also been invoked to explain boulder trains in recent research (e.g. Andreicheva, 2002).

However, the pattern of the Uralian erratics provides ample opportunity for alternative hypotheses of the glacial history of the region as the boulder trains normally contain mixed assemblages of resistant clasts deposited and reworked by various ice advances (paper 3, Ch. I). A voluminous discussion has been focused on the extent and age of Late Pleistocene glaciation. A Uralian origin was ascribed to Late Pleistocene glaciers within both early literary sources and overview Quaternary maps (Yakovlev, 1956; Gesse et al., 1963; Troitsky et al., 1966; Krasnov et al., 1971; Ganeshin, 1976; Velichko et al., 1989). According to these sources, the Late Pleistocene glaciation of the Urals evolved in two independent stages: (i) the development of a montane ice sheet that expanded onto the nearby lowlands, and (ii) the subsequent development of a network of valley glaciers that in places coalesced to produce thin piedmont ice fields.

The survey geologists, who mapped morainic assemblages in the 1960s, found that the alpine glaciers of the second Late Pleistocene stage in the Polar Urals produced isolated terminal lobes no further than 3 to 8 km from the western mountain front (Gesse et al., 1963). Therefore the preceding glacial maximum of the Urals was attributed to the Early Valdai cold stage (Yakovlev, 1956; Krasnov et al., 1971) correlated with MIS 4.

Following subsequent reports of finite radiocarbon dates beneath till in Cisuralia, the paradigm shifted to a Late Valdai (MIS 2) age of the main ice advance (Surova et al., 1974; Velichko, 1993; Velichko et al., 1997). Grosswald's idea (Grosswald, 1993) of a huge Late Weichselian ice sheet reaching 64° N, however, was shared only by Lavrov & Potapenko (2005) who reported subtill radiocarbon dates in the order of 45–32 ka BP in the Pechora Lowland. Based on these relatively old radiocarbon dates, Faustova & Velichko (1992) suggested that a smaller Novaya Zemlya-Uralian ice sheet might have culminated by 15–20 ka earlier than the Late Weichselian glacier of Fennoscandia.

The basis of the alternative scheme

Traces of penultimate (pre-Eemian) glaciation

The glacial topography of the peri-Uralian plains is clearly subdivided into two zones: north and south of the Arctic Circle. The geomorphic evidence of different ages for these two zones is unequivocal. The subdued topography of the forested sub-arctic zone with a well-developed arborescent drainage system appears to be significantly older than the hummock-and-lake dominated and perennially frozen tundra occurring north of 67-68° N. The ice-pushed ridges scattered in the southern zone of

Cisuralia are accompanied by boggy lakeless lows and are blanketed by loess-like silts and solifluction sheets (paper 18 in Ch. IV).

The modern influx of data was initiated in the late 1970s when the first results contradicting the above classical scheme appeared from the interpretation of high-altitude aerial and satellite images. The first issue relates to the sub-arctic landscape of the penultimate glaciation commonly attributed to the end of the Middle Pleistocene. The conventional idea of its Uralian origin is difficult to reconcile with the features that have been mapped using satellite imagery and middle-scale geological surveys. No morainic arcs emanating from the Urals have been mapped in the lowlands where only ice-pushed ridges transverse to the strike of the Uralian range can be discerned. Eskers and striae on the western piedmont of the Northern Urals, 200-300 m a.s.l., are orientated either N–S or even NW–SE but not normally to the axis of the range as should be expected for a mountain ice sheet.

The idea of Middle Pleistocene ice sheets that encroached from both lowlands onto the Northern Urals (Yakovlev, 1956) was strongly supported by the discovery of fine-grained lowland diamictons and varved clays in the mountains up to at least 600 m a.s.l. (Astakhov, 1974, 2004a). The pattern of lowland ice streams converging along the mountain axis (Fig. 2) leaves no place for montane ice domes south of 64° N. The 90-m-thick sequence of fine-grained Middle Pleistocene glacial sediments containing lowland erratics found in the central metamorphic zone of the Northern Urals (Fig. 2A) has never been affected by Late Pleistocene glaciers (Astakhov, 1974a). No striae or other indications of Middle Pleistocene glacier motion downslope from the central Urals have been found on either of the two piedmonts south of 64° N (Astakhov, 2004a). Even north of 64° N, morainic assemblages and polished valleys in the foothills clearly indicate the dominant ice flow from the north along the mountain range (Fig. 3).

The transverse system of alpine troughs with steep slopes is evidently younger as observable in the Subpolar Urals where they truncate the main longitudinal system of Middle Pleistocene glacial troughs. The latter are divided by residual ridges with sharp crests, such as Mt. Saledy (Fig. 4), polished by the southward flow of thick allochthonous ice. These long hogbacks are consistent with the NW–SE orientation of many transverse morainic ridges (Fig. 3). The most spectacular longitudinal monadnock is Mt. Sablya (Sabre) 1497 m a.s.l. This tourist attraction is named after its blade-like crest decorated by younger glacial cirques (Fig. 4).

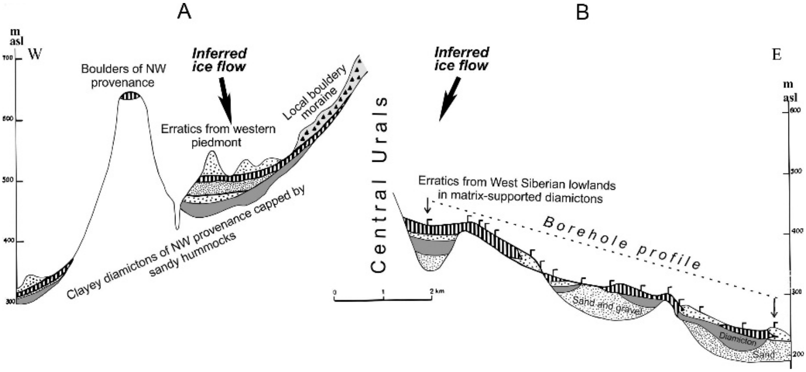

Fig. 2. Glacial sediments with lowland erratics in the Northern Urals (modified from Fig. 4 in paper 3, Ch. I). Adapted after Astakhov (1974a) (A), Ryzhov (1974) (B). Location is in Fig. 1.

Although the surficial features show only longitudinal or even upslope ice flow in the subarctic zone during the penultimate glaciation, they do not exclude preceding montane ice caps that could have formed over the mountains but were obliterated by subsequent ice advances from the Arctic. This follows from the evidence of central Uralian clasts found as erratics on the adjacent plains. However, it is not the decisive evidence because the Uralian clasts could also have been transported to the plains by very thick foreign ice overriding the mountains (Yakovlev, 1956).

Last continental glaciation

In the 1970s evidence for ice flow from the Arctic shelf into the Urals was also discovered for the Late Pleistocene. Fresh-looking moraines and striae orientated longitudinally were mapped around the northern tip of the Urals in the Arctic zone of the last glaciation. This appeared consistent with the N–S till fabrics and striations occurring on the eastern piedmont. The fresh hummock-and-lake landscape fringing the western slope of the Polar Urals and comprising two longitudinal lobes designated as Harbei and Halmer moraines was also identified (Fig. 5, for description see paper 18 in Ch. IV).

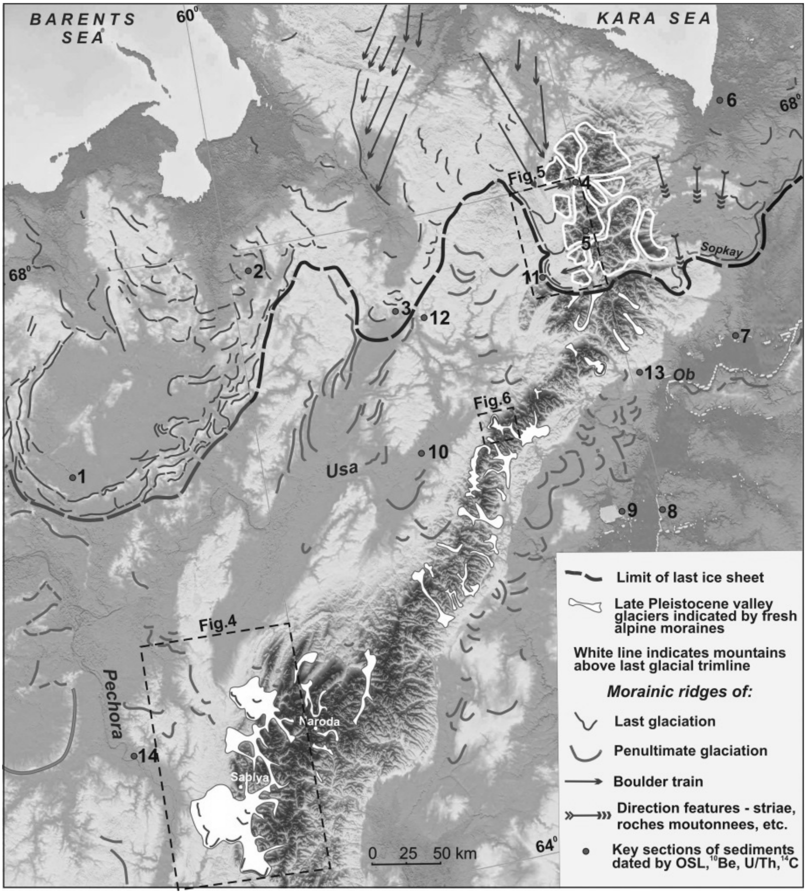

Fig. 3. Principal Pleistocene glacial features of the Polar and Subpolar Urals based on interpretation of aerial and satellite images and ground truthing by the Russian-Norwegian PECHORA project (modified after Astakhov, 2011 and maps at http://booksite.elsevier.com/9780444534477). Numbered stratigraphic sites: 1 – Podkova (Mangerud et al., 1999); 2 – river More-Yu (paper 19, Ch. IV); 3 – river Ileymusyur (Mangerud et al., 2004); 4 – river Bol. Kara (Nazarov et al., 2009); 5 – Chernov glacier (Mangerud et al., 2008); 6 – river Yerkata (Mangerud et al., 2004; Astakhov, 2006a); 7 – Sangompan and 8 – Pitlyar (paper 14, Ch. III); 9 – Shur peat (paper 12, Ch. III); 10 – Meskashor and 11 – river Bol. Usa (Svendsen et al., 2014); 12 – river Seyda (paper 3, Ch. I); 13 – river Hanmei (paper 14, Ch. III), 14 – Byzovaya Paleolithic site (Heggen et al., 2010). See site 11 also in Figs 5 and 7. See centrefold for this image in colour.

It is significant that big boulders from the central Urals are extremely rare in the Halmer hummocks, which are mostly built of well-drained and rounded fine gravels derived from the Permian conglomerates exposed west of the lobe. These features could only be explained by Late Pleistocene streams of Arctic ice that flowed southwards and bifurcated against the northern extremity of the mountain range to flow southwards along both Uralian flanks (paper 2 in Ch. I). Only immediately south of the margin of the ice sheet did a large lobe of fresh-looking moraines with Central Uralian boulders appear on the western piedmont close to the River Bolshaya Usa (Bol. Usa) (Fig. 5).

The former intrusion of non-local ice into the Urals from the northwest is supported by the presence of foreign push moraines located at altitudes 250–560 m a.s.l. (paper 18 in Ch. IV). Ground observations confirm that the distal convex slopes of these crescentic moraines facing upstream in montane valleys (e.g. Astakhov, 2011: fig. 25.8). The ice flow from the Kara Sea coast is independently indicated by southerly vectors of boulder trains in the surficial till that have been traced by geological mapping across the Pai-Hoi Ridge with eastward diversions into the Uralian valleys (Shishkin, 2007). The pattern of these and other directional features such as striae indicates that Late Pleistocene ice streams from the Kara shelf intruded into the Polar Urals almost up to 67° N, blocking and assimilating small local glaciers (Fig. 3).

The thickness of the external ice has recently been estimated through the interpretation of aerial photographs and the determination of the altitudes of cryoplanation terraces that produce long staircases located above the glacial trimlines. The trimline level descends from 400 to 500 m a.s.l. at the northern tip of the range (26 measurements) to 200 m a.s.l. on river Bol. Usa, indicating a decrease of ice thickness along 100 km of the mountain front from 300-400 m a.s.l. in the northern tip of the Urals to 100 m a.s.l. at the ice-sheet margin along the River Bol. Usa.

Another important conclusion from this geomorphic pattern: the cryoplanation terraces, which are periglacial features incompatible with glacial erosion (Boch & Krasnov, 1946), indicate that the Late Pleistocene ice sheet did not cover mountain slopes and summits higher than 500-600 m a.s.l. This is the main argument against the reconstruction of the Ural ice cap suggested in Svendsen et al. (2014: p. 422). Additionally, the Kara ice-sheet moraines inserted into the mountains often do not show any traces of collision with local glacigenic features because such features are generally absent in the same valleys.

Fig. 4. Subpolar Urals in Landsat image. Forested slopes are dark green. G – older generation of hummocky moraines; g – fresh morainic landscape of smaller morainic hummocks and ridges; F – heavily dissected landscape of fluvial erosion without visible glacial features. See centrefold for this image in colour.

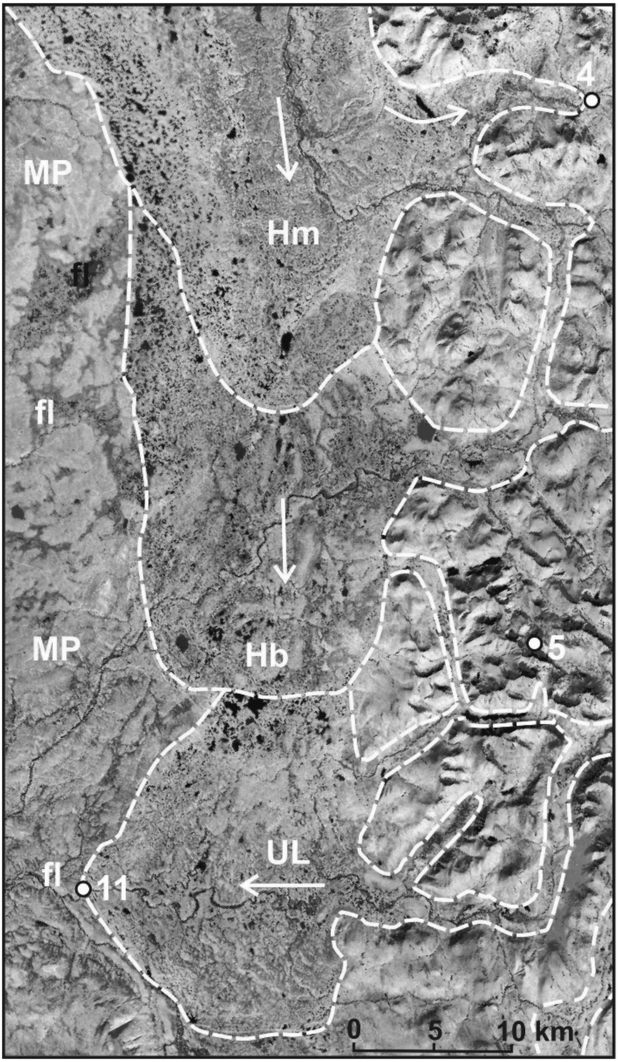

Fig. 5. Moraines of main ice lobes of the last glaciation (outlined by broken lines) in the western piedmont of the Polar Urals north of 67° N as seen in Landsat satellite images. Hm – fine-gravel hummocks of Halmer ice lobe; Hb – gravel moraines of Harbei lobe; UL – boulder moraines of Usa lobe; MP – Middle Pleistocene moraines; fl – outwash (modified after paper 18, Ch. IV). Arrows indicate inferred ice flow directions. 4, 5 and 11 are dated sections in Fig. 3. See centrefold for this image in colour.

The crescentic morainic ridge at 560 m a.s.l. inserted from the NW into a montane valley on the northern tip of the mountain range was reported previously (Fig. 9 in paper 18, Ch. IV); Astakhov, 2011: Fig. 25.8). This ridge has no counterparts at similar altitudes. Its unusual elevation might be a result of local tectonic uplift. Irrespectively of the local uplift on the order of 100–200 m, the advancing ice should have been not less than half a kilometre thick on the shore of the Kara Sea some 70 km to the north (Fig. 3). This ice thickness was sufficient to produce the transverse Sopkay morainic belt located on the eastern piedmont and smoothed mountain slopes of the eastern Polar Urals. Traces of the youngest ice that flowed from the Kara Sea are clearly seen on the eastern piedmont where the clay-rich Sopkay moraines with N–S fabrics are interspersed with polished and striated limestone bedrock surfaces. Stoss-and-lee features on Paleozoic limestones occurring north of the Sopkay moraines display orientations varying from 350° to 10° N. These directional features together with the clay-rich composition of the thick surficial till that contains few Central Uralian clasts unequivocally indicate ice flow from the north and not from the nearby mountain valleys (paper 2, Ch. I)

Alpine glaciation

Traces of local glaciation are mapped south of 67° N where they consist of (i) fresh-looking hummock-and-ridge morainic assemblages in the proximal foothills, (ii) U-shaped troughs, and (iii) numerous cirques located above the current tree line, mostly in the upper parts of the western slopes of the glacial troughs. The termini of individual valley glaciers of the Polar Urals are recognizable south of 67° N as morainic loops not more than 5-8 km wide (Figs 3, 6). North of 67° N, the distal ends of the glacial troughs radiating from the mountains are not observable, were most probably assimilated by the advance of the Kara Sea ice sheet (Fig. 3).

The largest moraine derived from westbound valley glaciers is a stony piedmont lobe (Fig. 5) that occurs in the glacial trough of the river Bol. Usa. With a width of ~ 20 km, the Usa morainic lobe is considered too wide to represent an end moraine associated with ice flow from local glaciers draining the area mostly below 1000 m a.s.l. Other alpine moraines produced by glaciers that discharged from higher mountains are noticeably smaller in this regard (Fig. 6). The Usa morainic lobe probably originated from the merging of local alpine glaciers with outlet glaciers of the Kara ice sheet. The valley glaciers interconnected in a reticulate network that separated the higher mountain massifs as nunataks above the ice-sheet surface (Fig. 3).

The transverse U-shaped valleys of the alpine glaciers are longest and most frequent in the western part of the Subpolar Urals where the mountain chain is at its widest. In comparison, the eastern slope is devoid of deep and long glacial troughs (Fig. 4). Similarly, the distribution of hummocky moraine comprising Central Uralian glacigenic material is abundant on the western flank of the Subpolar Urals but totally absent on the eastern slope. Judging by the fresh hummocky moraines, such as those along the river Bolshoi Patok, alpine glaciers coalesced to form an Alaskan-type piedmont glacier at the western foot of the Subpolar Urals between 64 and 65° N. However, even isolated terminal horseshoe-shaped moraines are not found in the eastern foothills where an arborescent pattern of fluvial erosion dominates (Fig. 4). The dimensions of the western morainic apron are similar to the modern-day Malaspina Glacier. However, the comparatively modest elevation and relatively small feeding area of the Subpolar Urals suggest that the piedmont glacier was limited to not more than 200 m in thickness in comparison to the thickness of ~ 600 m for the Malaspina Glacier.

This stark contrast in the nature of the glacial features on the eastern and western flanks of the Subpolar Urals illustrates the strong influence of topography on ice accumulation during the last glacial cycle: the western slope was heavily glaciated whereas the leeward eastern slope was too dry to produce valley glaciers. This implies a significant reduction in the amount of snow accumulation on the Siberian flank of the Urals in the Late Pleistocene imitating the modern precipitation pattern. A similar orographic effect is felt in the wide northern part of the Polar Urals above 67° N where no traces of young alpine glaciation can be found on the eastern leeward slope.

An unusual pattern of glacial features is observed along the narrowest chain of the Polar Urals between 67 and 65.5° N where the morphological impact of the precipitation difference is indistinct. In contrast to the 150-km wide Subpolar Urals, fresh-looking alpine troughs with small terminal morainic arcs occur on both slopes of the Polar Urals where the mountain range is only 40 km wide (Fig. 3). The U-shaped rocky valleys with glacial cirques often cross-cut the narrow mountain chain (Fig. 6). Such through valleys are clearly related to former transfluent glaciers. The apparent lack of orographic asymmetry in the distribution of the young glacial features in this part of the range probably relates to the influence of the through valleys that allowed westerlies and associated snow accumulation to influence the eastern flank of the Urals.

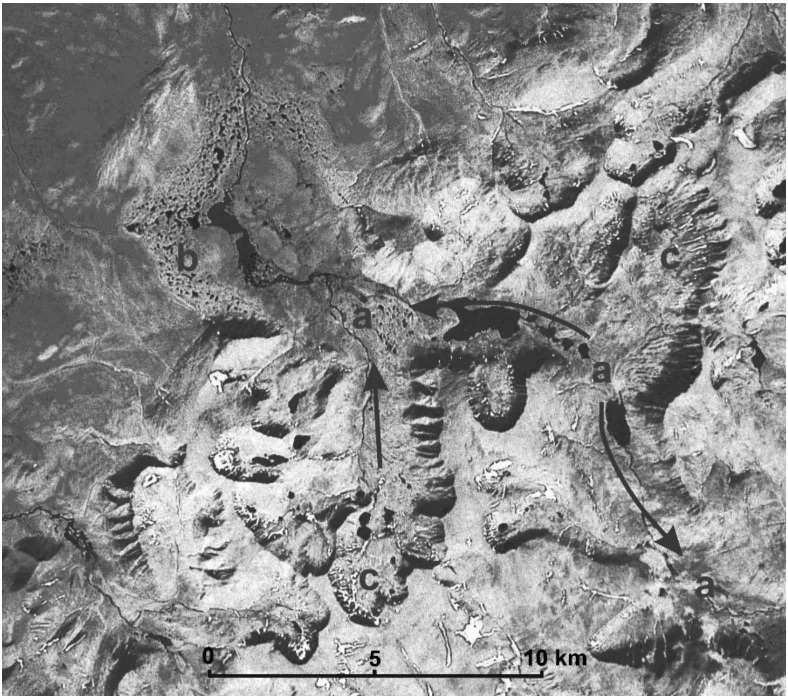

Fig. 6. Satellite view of Hoila transfluent glacial trough (a–a) terminating at the morainic loop (b) on the western piedmont, Arctic Circle. Ice-flow directions are marked by arrows, feeding cirques by (c). See Fig. 3 for location. See centrefold for this image in colour.

The age of the glacial features

Penultimate glaciation

The largest erosional features of the western mountain front, especially the floors of the U-shaped valleys and residual hogbacks like Mt. Sablya, indicate ice motion from the NE along the Urals. This longitudinal ice flow evidently occurred earlier than the alpine glaciation. The alpine features are either superimposed on or truncate traces of the longitudinal ice flow in the western foothills (Figs 3, 4).

The crucial evidence for the age of the sub-arctic glacial features can be found in the lowlands occurring west of the river Pechora at 67° N where a lowland built of marine sands and not covered by till intrudes into the well eroded morainic upland. This Sula Formation containing abundant Atlantic malacofauna is a signature of a paleoclimate warmer than the current climate. Based on a continuous series of 16 OSL dates, mean value 112 ka, it is reliably correlated with the Eemian (Murray et al., 2007).

A pre-Weichselian age for the last ice advance in the sub-arctic Cisuralia is also confirmed by the following OSL dates that are obtained from sediments overlying the topmost till at key sites indicated in Fig. 3:

• Seyda, site 12. OSL dates from overlying outwash 145, 146, 151, 157 ka, mean value 152 ka (Fig. 9 in paper 22, Ch. V).

• Byzovaya, site 14. OSL dates from outwash 249, 145 ka (Heggen et al., 2010).

There are also numerous radiocarbon dates with ages of up to 40-50 ka BP recovered from surficial Paleolithic sites 10 and 14 in Fig. 3 (Mangerud et al., 1999; Svendsen et al., 2004; Heggen et al., 2010). These facts in combination with the 10 subtill OSL dates, mean value 174 ka, and U/Th dates c. 200 ka at Seyda (12 in Fig. 3, fig. 9 in paper 22 , Ch. V) clearly indicate a late Middle Pleistocene age for the glacial landscapes of subarctic Cisuralia (paper 22, Ch. V).

• Shur, site 9. U/Th dates 133 and 141 ka on peat accompanied by OSL dates 100 and 82 ka from sand. Undisturbed by ice, 1 m thick continuous peat layer with interglacial pollen and old U/Th dates seems to be a good indication of pre-Eemian age of the underlying till in Transuralia (paper 12 in Ch. III).

• Scattered OSL dates at several Transuralian sites also indicate a pre-Weichselian age for the surficial outwash. For example, surficial glaciofluvial sand at Hanmei, site 13, is dated to 204 and 199 ka. Farther north close to the Sopkay moraine ages 167 and 96 ka have also been obtained from glaciofluvial sands (Fig. 6 in paper 14, Ch. III)).

Last glaciation

The arctic zone of Cisuralian plains with numerous ice-pushed ridges, kame plateaus and small hummocks and ponds is related to the last glaciation by practically all the investigators of Quaternary history in spite of their varied views on the age and dynamics of recent glaciers. The dominant concept of the 1970-80s ascribed a Late Weichselian age to the most pronounced crescentic moraines located at the foot of the mountain range at 300-400 m a.s.l., which are a product of younger valley glaciers (Fig. 6). However, according to modern data the firmly established Late Weichselian age relates only to the much smaller moraines around the small glaciers located at altitudes above 500 m a.s.l. (Mangerud et al., 2008).

Thus, large boulders dated by cosmogenic ^{10}Be yielded six ages in the range of 28 to 14 ka (mean 21 ka) only within 1 km of the tiny Chernov Glacier (site 5 in Figs. 3 and 5). Downstream along the glacial trough, four large boulders displayed ^{10}Be ages of 60 to 50 ka with a mean value of 58 ± 3 ka (Fig. 7). These data point to the limited extent of Late Weichselian glaciers even on the western slope of the Polar Urals (Mangerud et al., 2008). In the troughs of the eastern Polar Urals the Late Weichselian is represented only by alluvial fans with no traces of young glaciers. This agrees with the very dry, polar desert climate of the Late Weichselian in Cisuralia conditioned by its lee position east of the Kara Sea Ice Sheet (Astakhov & Svendsen, 2011). Thus, the latest data on chronometry of the uppermost glacial complex totally disagree with the idea of a Late Weichselian age for lowland glaciation and large alpine glaciers.

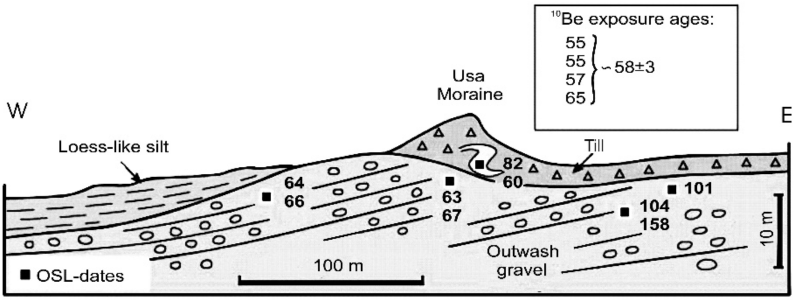

Fig. 7. Dated glacial sediments on river Bol. Usa, sect. 11 in Figs 3 and 5. OSL dates are obtained from sand pockets in glaciofluvial gravels and boulder till at the terminus of the Late Pleistocene glacier, ^{10}Be exposure dates are from quartzite boulders upstream within the same river valley in the mountains (Svendsen et al. 2014). All dates are in kiloyears ago.

The relative antiquity of the last major ice advance is indicated by the OSL dates obtained from the terminal moraine of a foreign glacier tongue inserted from the western foothills into the mountain valley of river Bol. Kara (site 4 in Fig. 3 and 5) (Nazarov et al., 2009 and D. Nazarov, oral communication, 2009). These authors reported 13 OSL dates on the sandur sand and gravel located on the distal flanks of this moraine: 82, 80, 80, 80, 79, 76, 76, 75, 74, 73, 66, 60 and 47 ka. If two uppermost dates from the cover sand sheet are excluded, the remaining 11 dates are tightly clustered around the mean value of 76.5 ka.

Other geochronometric evidence of the age of the last glaciation includes OSL dates recorded from glaciofluvial gravels at the margin of the piedmont morainic apron on the river Bol. Usa (site 11 in Figs. 3 and 5). The values display considerable scatter. However, if the obvious outlier of 158 ka is eliminated, the remaining 8 dates give the same mean age of 76 ka (Fig. 7). A pre-Late Weichselian age for the Usa glacial complex is independently indicated by ^{10}Be exposure dates that display a mean value 58 ka for big boulders sampled upstream on this river (Mangerud et al., 2008). This age for the last major glacial advance agrees with the ^{14}C age ~ 40 ka BP (11 dates) obtained from alluvium incised into the latest varved formation at the Paleolithic site Mamontovaya Kurya (10 in Fig. 3) (Svendsen et al., 2014).

The above OSL values are within a reasonable range from the statistically sound mean OSL age of 82 ka measured on 27 samples from sediments of the major proglacial lake in Cisuralia (Mangerud et al., 2004).

The Sopkay moraines on the eastern piedmont (Fig. 3) indicate the position of the limit of the last glaciation. Originally they were considered a Uralian product arbitrarily related to the Late Weichselian or even the Younger Dryas stadial. However, the mapped morainic ridges proved to be clay rich and contained very few Uralian clasts. The pebble and striae orientations occurring parallel to the mountain chain suggest ice flow from the north (see paper 2 in Ch. I). Immediately to the south, sediments of a proglacial lake in the Ob valley at 50-60 m a.s.l. (Sangompan, 7 in Fig. 3) yielded four OSL dates with the mean age of 81 ka. This result is supported by six OSL dates with a mean value of 78 ka recorded from the proglacial rhythmites upstream at Pitlyar (8 in Fig. 3) (paper 14 in Ch. III). This implies that the Sopkay ice advance occurred simultaneously with the Harbei ice advance of the Kara ice sheet west of the Urals.

Sediments overlying the uppermost glacial complex of the Arctic provide a minimum age for the last ice advance from the shelf. On river More-Yu (site 2 in Fig. 3), there are two OSL dates of 46 ka obtained from lacustrine sand and three dates on mammoth bones from 35 to 40 ^{14}C ka BP (paper 19, Ch. IV). Fluvial sands capping the Harbei moraines at Ileymusyur, site 3, yielded OSL values of 77 and 75 ka BP (Mangerud et al., 2004). East of the Urals, within the limits of the last ice sheet, on river Yerkata (6 in Fig. 3) lacustrine and aeolian sands above the basal diamicton with large bodies of relict glacier ice yielded OSL dates 72, 65, 63, 59 with a mean age 65 ka (Fig. 7 in paper 14, Ch. III). In total, there are 18 OSL dates that range from 46 to 77 BP and a number of old radiocarbon dates from overlying sediments (sites 1, 10, 14 in Fig. 3) that support a pre-Late Weichselian age for the uppermost glacial complex.

Svendsen et al. (2014) suggest an MIS 4 age for the major advance of Late Pleistocene ice into the Urals. However, taking into account the wide scatter of the OSL dates, the age of this ice advance cannot be specified more precisely than the Middle/Early Weichselian period (Mangerud et al., 1999). Building up a large ice sheet on the shelf would probably take a considerable period during the Early Weichselian. Judging by the rarity of nested Late Pleistocene moraines, the ice retreat was comparatively fast and irreversible. This is consistent with the traces of the Barents Sea high-stand that occurred during the Middle Weichselian (Larsen et al., 2006; Mangerud et al., 2008).

In any case the available chronometric record is sufficient to relate the last major ice sheet of the Polar Urals to an early part of the last glacial cycle and not to the Late Weichselian. These dates also imply that the alpine glaciers that survived south of the Kara Ice Sheet margin cannot be younger than MIS 4.

Unfortunately, there are no geochronometric data for the Late Pleistocene alpine glaciers occurring south of the Arctic Circle. Surova et al. (1974) suggested a Late Weichselian age for a marginal moraine on river Lagorta located on the eastern piedmont of the Polar Urals. However, this dating is highly speculative based solely on a hypothetical rate of proglacial lacustrine sedimentation.

It is theoretically possible that the small Late Weichselian glaciers of the Arctic could have expanded to form more substantial valley glaciers, especially in the higher Subpolar Urals. This may be inferred from the series of nested crescentic moraines that occur along the long glacial troughs of the western Subpolar Urals. However, a Late Weichselian age for these troughs seems unlikely in the view of the small extant glaciers which are not larger in the Subpolar Urals than in the Polar Urals. In waiting for the crucial geochronometric data from the mountains south of the Arctic Circle, it would be prudent to presume for the time being that only smaller recessional moraines in the heads of troughs of the western Subpolar Urals might have formed during the Late Weichselian.

The age of Transuralian moraines

The data described in the previous section also provide an insight into the origin of the large and degraded moraines located on the Transuralian plain of West Siberia, that occur distally from the small morainic arcs of the alpine maximum on the eastern piedmont (Fig. 3). They comprise arcuate accumulations of coarse diamicton that are covered by vast solifluction sheets with a dense network of parallel periglacial rills. These large subdued morainic ridges have previously been considered as traces of the Late Pleistocene Uralian glaciation (Krasnov et al., 1971; Ganeshin, 1976). However, the largest moraines located in the lee of the narrowest mountain range surprisingly have no symmetrical counterparts on the wetter western slope where only moraines running transverse to the mountain front are mapped. This paradox is emphasized by the lack of younger moraines on the eastern slope of the Subpolar Urals that contrasts with the abundant moraines on its western slope (Figs 3, 4).

The described antiorographic pattern excludes an alpine origin for the large Transuralian moraines. They should instead be attributed to outlet glaciers of non- Uralian origin – ice streams derived from a thick Cisuralian ice sheet retreating to the northwest. Such an ice sheet south of the Polar Circle could conceivably exist only during the Middle Pleistocene. It was probably not thick enough to override the Uralian summits but may have occasionally discharged eastwards in the form of outlet glaciers that occupied the through valleys of the Polar Urals. If this is true, then the penultimate ice sheet, locally related to the Vychegda MIS 6 glaciation, was much thicker in Cisuralia than in Siberia. According to almost all authors the Vychegda ice sheet coalesced with the main Saalian ice sheet of Fennoscandia whose southeastern limit is mapped close to Moscow (Krasnov et al., 1971; Ganeshin, 1976; Astakhov, 2011).

The evidence of the end moraines of transfluent outlet glaciers has important paleogeographic and stratigraphic implications. The ice flow from the west through the Uralian valleys forming the large terminal moraines on the plain close to the river Ob means that at the end of the Middle Pleistocene, when Arctic ice flowed along the western slope of the Urals, the West Siberian Plain must have been relatively ice free. This poses a big question mark on the conventional correlation of the Vychegda glaciation of Cisuralia with the Taz glaciation of West Siberia (Astakhov, 2011, 2013).

Conclusions

• An assessment of both recent and previously obtained data contradicts the traditional concept of the Urals acting as a major centre of inland glaciation. Instead, during the late Quaternary, the mountain range produced only valley glaciers. The Urals should therefore be viewed as a long and narrow rocky barrier that acted as an obstacle to the southward flow of ice streams from the Arctic shelf.

• The primary element of the revised glacial history, as described in this paper, relates to the advance of inland ice from the Arctic shelf. Middle Pleistocene ice streams are responsible for surficial glacial features south of 67° N whereas Late Pleistocene continental ice formed the landscape of the Arctic.

• The last maximum glaciation occurred in early Weichselian times 90–70ka BP. The invading Arctic ice incorporated alpine glaciers from the Urals north of 67° N whereas independent valley glaciers survived south of the Kara ice-sheet margin. South of 65° N on the windward western slope, these alpine glaciers coalesced to produce large piedmont glaciers similar in size to the present-day Malaspina Glacier.

• No reliable traces for montane ice domes during the two last glacial cycles are discernible in the modern land surface. This conclusion is at variance with Svendsen et al. (2014), who suggested a Late Pleistocene ice cap over the Polar Urals. The position of the mapped trimlines in the valleys below the cryoplanated summits does not support the idea of a montane ice sheet.

• Taking into account the erratics of central Uralian rocks scattered over the adjacent plains, older montane ice caps cannot be excluded. However, the bulk of the Uralian erratics could have been delivered onto the plains by passing streams of external ice rather than by local glaciers. Such a case is presented by the large moraines of Transuralia that were probably left by outlet glaciers of a Cisuralian ice sheet of the late Middle Pleistocene.

• The local Uralian glaciation fully depended on the orographically asymmetrical and usually insufficient supply of snow, especially in the second half of the last glacial cycle when Arctic desert existed west of the Polar Urals in the lee of the Late Weichselian ice sheet. This resulted in limited extent of the Late Weichselian glaciers, even in the windward western Urals.

The revised pattern of late Quaternary glaciations presented above may be used as a model for the events that occurred during the earlier stages of the Ural glacial history.

Acknowledgements

This study was supported by the St. Petersburg State University. Most of the chronometric dates were obtained in the course of the Russian-Norwegian projects PECHORA (1993–2003) and ICEHUS (2005–2010) funded by the Research Council of Norway. The satellite images and base map are kindly provided by the Centre for Remote Sensing of Russian Geological Research Institute, St. Petersburg. My sincere thanks are due to Dr. R. Waller, University of Keele, who checked the English language. I am also grateful to the referees who helped to clarify the message.

CHAPTER V

TRANSREGIONAL SYNTHESIS

The last chapter is an assemblage of works not strictly related to a certain terrain but rather applying the knowledge collected in the previous chapters to solution of several problems common for entire northern Eurasia. Paper 21 dwells on the impact on the Quaternary history of the super-continental size and climatic diversity of the Russian mainland with implications for chronological and paleoclimatic reasoning. Paper 22 is an overview of the available data on the chronology of ice advances and interglacial episodes of northern Russia and their relations to the glacial history of Western Europe. Paper 23 is an analysis of paleoenvironmental proxies in order to recognize the trend of the Late Pleistocene non-glacial geological events which happened on the Russian mainland during the last 50 thousand years, looking for parallels with Western Europe. Paper 24 suggests a transregional correlation of the Late Pleistocene events within northern Russia. Paper 25 is a cartographic exercise using glacial landforms to correlate ice limits across northern Russia in order to obtain material support for chronological conclusions. And paper 26 is an attempt to apply the knowledge of glacial geology to the problem of recent tectonism of a great petroleum basin and adjacent uplands.

21. Geographical Extremes in Glacial History of Northern Eurasia

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Polar Research, 2008, vol. 27, p. 280–288

Introduction

The huge Eurasian landmass presently displays a wide variety of environments along its northern margin. The geographical extremes of the modern situation are evident by a comparison of the west against the east along the same latitude: the mild and rainy climate of western Norway contrasts sharply with the frosty and dry climate of Yakutia, close to the northern Cold Pole. However, the response of this climatic asymmetry to Pleistocene global coolings and warmings, especially to the inception and development of glaciation, has been debated. The problem has been discussed in the Russian literature since the 19th century (Voyeikov, 1881; Gerasimov & Markov, 1939; Velichko, 1980). Western European textbooks focused, as a rule, on Atlantic paleoenvironments and hardly ever mentioned the notable west-east climatic gradient distinctive of Eurasia. This is understandable since the western European peninsula of Eurasia is too small for studies of long-distance changes.

The climatic gradient has persistently been held responsible for various geographical and paleogeographic phenomena. For example, Gerasimov and Markov (1939) believed in a decisive importance of the west-east gradient and elaborated Voyeikov's ideas into a theory of the asynchronous development of Quaternary glaciers across Eurasia. They suggested that i) warmer intervals of the Quaternary history in eastern Siberia were more beneficial for glaciation than the cold ones, ii) former western Siberian ice sheets did not exceed 500-700 m in thickness and iii) the main sources of glacial ice were mountainous areas.

In the 1970s it became clear that big continental ice masses were basically contemporaneous throughout the Northern Hemisphere. Hence, the initial idea of asynchronous ice advances at two extreme flanks of the Eurasian continent was transformed into the concept of glacial asymmetry with an eastward reduction of Pleistocene ice volume (Velichko, 1980). The latter version of paleogeographic extremes in glacial history, did not, however, question the previous conclusions about mountains as the main source of glacial ice east of the Fennoscandian ice sheet. There have been persistent attempts to theoretically reconstruct a large Weichselian ice cap over the Urals (e.g., Velichko et al., 1987), despite clear geological indications to the contrary (papers 2 in Ch. I and 18 in Ch. IV).

The discovery of the Kara shelf ice dispersal centre and much thicker Pleistocene ice in Siberia than had previously been assumed (papers 1 and 2 in Ch. I) stimulated another extreme in paleogeographic thinking. The `continental theory` was challenged by glaciologists who disregarded the fundamental difference of oceanic versus inland environments and simply transferred the glacial model of the smaller North America to the huge Eurasia. The resulting reconstructions suggested improbably large and contiguous circum-arctic Late Weichselian ice sheets (Grosswald, 1998). The maximalist model enjoyed support from geophysicists whose models needed plenty of continental ice to balance the sea level change, whereas only few geologists were charmed by the idea.

The problem re-emerged with the multiyear efforts in the study of the Quaternary Eurasian northern margin by international research teams assembled under the European Science Foundation umbrella programme Quaternary Environment of the Eurasian North (QUEEN) which terminated in 2004. These basically European-Russian collaborative projects, supplemented by several ventures with American participation, have since 1993 attended to the Late Pleistocene geological history of the northern Eurasian landmass from Scandinavia to the Lena delta and the adjacent Arctic islands and deep seas. The geochronological, sedimentological and geomorphic data that were obtained shed new light on the old paleogeographic problem of reconstructing former ice sheets (Svendsen et al., 2004). New information on the interglacial intervals during the Pleistocene have been less spectacular.

This paper discusses the principal events of Eurasian glacial history, as demonstrated by QUEEN research, in relation to popular theoretical models and possible palaeogeographic inferences. The purpose of the paper is to draw the attention of western European geologists working in Russia to the peculiar mode in which the extremely continental environment may respond to global climate forcing, and possibly to provoke a discussion of further research themes along the lines embodied in the Arctic Paleoclimate and Its Extremes (APEX) programme.

Modern and Holocene climatic asymmetry of Eurasia

Continents are known to have climate gradients normal to their coastlines. The huge landmass of Eurasia reveals the most pronounced west-east gradient, reflecting a transition from maritime to continental climates along a given latitude. This natural feature is often described (and taken into account by general circulation models) as an eastward increase in continentality caused by the fading transfer of heat and moisture from the Atlantic. Continentality drops again only very close to the Pacific. It is normally measured by two indirectly linked parameters: annual precipitation per square unit (P) and annual amplitude of mean monthly air temperatures, i.e., the difference between July and January means (J–J index). Maritime climates, typical for western Europe, are characterized by J–J <25°C and P>700 mm, whereas in the continental climates of eastern Eurasia J–J may reach 50-60°C (Fig. 1) and P<400 mm.

Although the size of ice sheets depends first of all on the availability of moisture, the present-day precipitation pattern does not show a straightforward affinity to the distribution of former glaciers. Annual precipitation, being dependent on seasonal air flows, is a mobile and spatially highly variable parameter, whereas glaciation is a steady feature averaging regional conditions over centuries. The J–J index, governed mostly by radiation balance, also indirectly reflects the level of precipitation. It is a more stable characteristic and gives a pattern easily comparable to former glacial phenomena (Fig. 1). The map shows two climatic extremes of northern Eurasia: humid Atlantic coasts, with J–J = 15°C, versus the very continental climate of perennially frozen eastern Siberia near the Cold Pole, with J–J = 50-60°C.

The influence of climatic asymmetry on the geological history can be more readily perceived from the circum-arctic change of Holocene environments. Results of integrated analyses of pollen profiles around the Arctic have provided evidence of temporal shifts of Holocene climatic optima (Bennike et al., 2001). Figure 2 shows that the Holocene optimum gradually shifted westwards from the so-called insolation maximum in East Siberia and Alaska, 9-10 ^{14}C ka, to West Siberia 6.5 ^{14}C ka, then shifted further westwards to Europe 5.5 ^{14}C ka and finally to eastern Canada around 4 ^{14}C ka (Bolshiyanov, 2000). The time-transgressive Holocene altithermal suggests a similar temporal pattern for other paleoclimatic events.

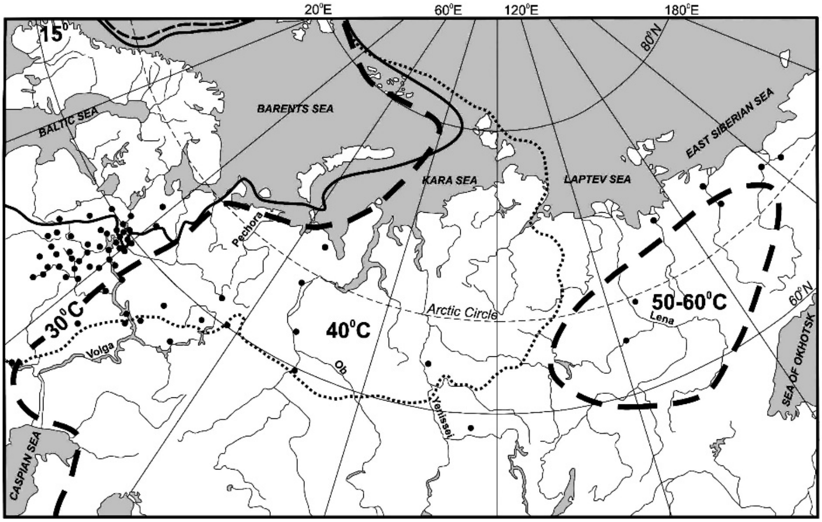

Fig. 1 The spatially changing continentality of northern Eurasia against limits of ice sheets (see Svendsen et al., 2004). The solid line indicates the extent of the last ice sheet about 20 ka BP. The dotted line shows the Quaternary glacial maximum. Broken lines are isolines of differences between July and January mean air temperatures (Berg, 1938). Black dots are thick peat deposits from the last interglacial, often referred to as `Karginsky` in Siberian literature (from numerous Russian publications).

Glacial asymmetry of Eurasia

The QUEEN results have firmly established the asymmetric configuration of Late Quaternary ice sheets of northern Eurasia. The first Weichselian ice sheet that culminated at about 80-90 ka BP was very small in Scandinavia but occupied the entire Kara Sea shelf and a good part of the present dry land above the Arctic Circle (Svendsen et al., 2004). The second ice advance at about 60 ka BP covered most of Scandinavia and the Baltic Sea and reached south of the Arctic Circle in western European Russia (Larsen et al., 2006). The early Weichselian expansion of Arctic ice sheets was accompanied by small alpine and piedmont glaciers, mostly along the western slopes of mountain ranges (e.g., papers 18 and 20, Ch. IV). The last Late Pleistocene ice sheet, roughly corresponding to marine isotope stage (MIS) 2 (also called the Last Glacial Maximum [LGM] by marine geologists), covered only north-western Europe and most of the

Barents shelf, leaving the central and eastern Eurasian Arctic landmass ice-free.

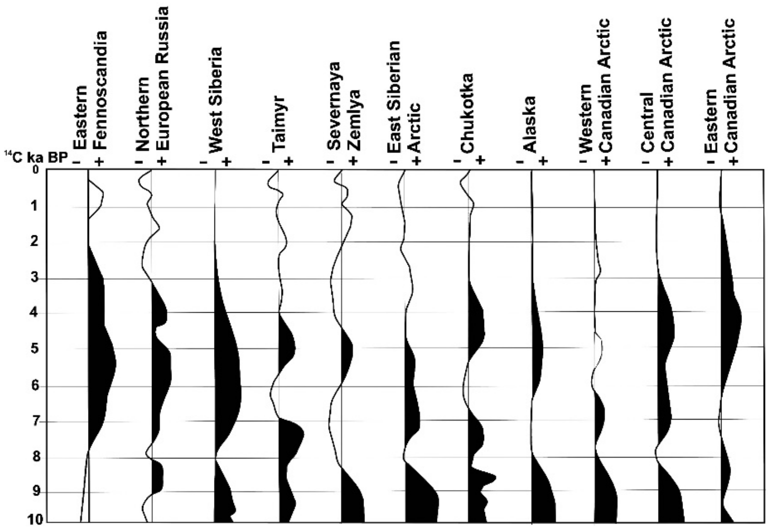

Fig. 2. Circum-Arctic change of Holocene climatic optima inferred from pollen records contained in radicarbon-dated terrestrial sediments (modified from Bolshiyanov, 2000).

Figure 3 illustrates the reverse succession of the latest glacial events in Siberia versus Europe in terms of ice volume (Svendsen et al., 2004). The same pattern was recorded in the mountains of southern Siberia on the basis of luminescence dating (Sheinkman, 2004). The former glaciers seem to have been progressively west-biased in the course of the Late Pleistocene glacial cycle, i.e., the volume of each successive ice sheet more distinctly gravitated towards the North Atlantic margin.

In this respect the Eurasian continent markedly differs from North America, where larger ice sheets were fairly symmetrical in relation to the longitudinal continental axis (Flint, 1971). This difference is doubtless a result of the pronounced growth of continentality in eastern Eurasia, whereas in North America, due to the modest size of the landmass, continentality never developed to the same extent and did not prevent oceanic influences from the Atlantic and Pacific to join forces and produce the largest ice sheets of the Northern Hemisphere.

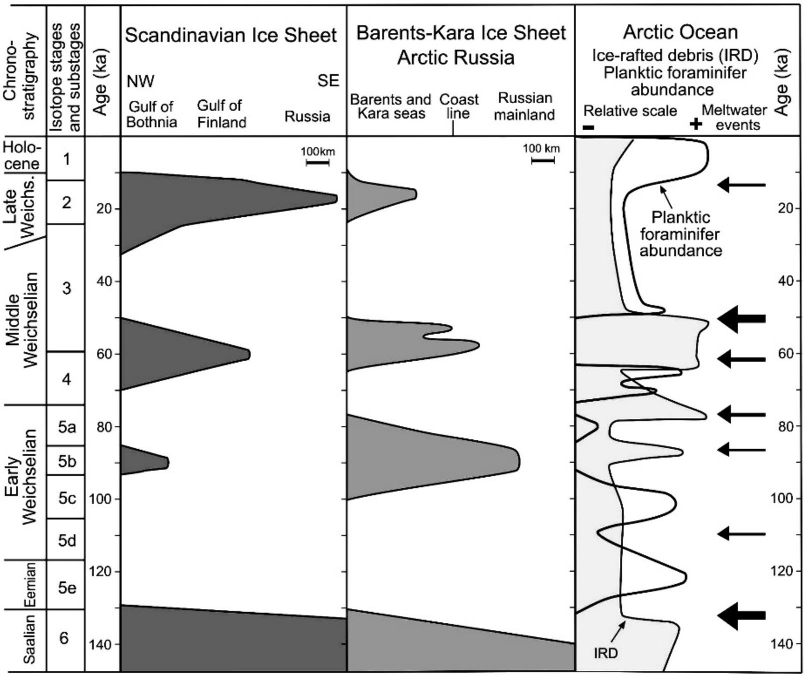

Fig. 3 Time–distance diagram of late Quaternary ice sheets of Eurasia (modified from Svendsen et al., 2004.) The right-hand column shows results of the analysis of bottom sediments from cores of the Arctic Ocean; arrows indicate freshwater fluxes inferred from the isotope composition (Spielhagen et al., 2004).

The relation of ice sheet size and the climatic gradient can be seen in Fig. 1, in which the former basically wet-based Scandinavian glaciations are confined within the present day J–J isoline of 30ºC. The `grey zone` of cold-based glaciations of the West Siberian type is located within J–J values 30 to 45ºC. Farther eastward, with J–J values growing up to 50-60ºC, former ice sheets are replaced by thick conservative permafrost, which is virtually underground glaciation without interglacials. The active surficial glaciation of the Atlantic seaboard may be viewed as the extreme versus the passive subterraneous glaciation of eastern Siberia. Even within the glaciated area there are two extreme cases: predominantly wet-based ice sheets in western Europe versus cold-based Siberian glaciers in the `grey zone` with J-J of 40-45°C. This does not mean that an ice sheet can form in West Siberia at the present level of precipitation, but the map shows the tendency felt through the Pleistocene. The extreme modes of the

Eurasian glaciation are evident also from a comparison of the fluted and drumlinized subglacial surface of Europe with the West Siberian glaciated plains that lack eskers and drumlins but are full of glaciotectonic and stagnant ice features (paper 8 in Ch. II).

The climatic gradient is detected in the deglaciation pattern as well. The continental climate and thermal inertia of the deeply cooled Siberian lithosphere, with reduced geothermal gradients, led to very slow, retarded deglaciation and preservation of fossil glacial ice in the Arctic for tens of thousands of years (paper 5 in Ch. II). It is significant that the Siberian permafrost persistently grew, irrespective of intermittent glacials and interglacials. Even under the present Laptev Sea it is at least 400 ka old (Romanovsky et al., 2004). Such short retreats from the general Quaternary cooling trend as the Holocene warming were only able to temporarily remove the upper skin of the thick permafrost by oceanic transgression and thermokarst. In contrast, the thin-skinned permafrost in the Atlantic realm was an ephemeral feature which only sporadically affected glacier behaviour. The short-lived marginal permafrost could not impede the fast deglaciation during Atlantic terminations.

Post-QUEEN results

Post-QUEEN results have emphasized the asymmetric distribution of Late Pleistocene ice volume. The change of paradigms is especially evident from the revision of the age of the youngest alpine morainic ensembles which were traditionally attributed to MIS 2 (the radiocarbon time span of ca. 25-10 ka BP), and everywhere correlated with the European Weichselian and the classical Wisconsinan of North America (Kind, 1974). Already during the QUEEN activity it became obvious that the largest alpine glaciers east of Fennoscandia developed and vanished well before MIS 2. The MIS 2 age of the alpine maximum correlation was refuted by optically stimulated luminescence dating in the Urals (Svendsen et al., 2004) and by bottom sediments cores from deep lakes in the Taimyr Peninsula distally barred by the alpine morainic loops. The pollen data from postglacial sediments in southern Taimyr proved to be incompatible with the idea of MIS 2 age of the underlying till (Hahne & Melles, 1997).

Later on, Mangerud et al. (2008) dated boulders in the glacial troughs of the Polar Urals using the cosmogenic isotope ^{10}Be and found that values around 20 ka are confined to moraines not farther than 1 km distal from a present-day glacier. Further downstream, boulders were intact for 50-60 ka. Finally, horseshoe-shaped alpine moraines in the Verkhoyansk Mountains, the stratotypic area for the youngest `Sartan` Siberian glaciation, yielded optically stimulated luminescence ages exceeding 50 ka (Stauch et al., 2007). It seems that after MIS 2 the climate of the northern margin of the continent became so dry that glaciers could not survive, even in the mountains.

Thus, the latest results stress the insignificant role of mountainous areas as ice sources during the Pleistocene. This conclusion is in harmony with the well-known fact of Middle Pleistocene glaciers overriding the Ural Mountains and the mountains of central Siberia (paper 3, Ch. I) and periglacial evidence of an extremely dry Late Pleistocene climate in eastern Siberia (Hubberten et al., 2004; Sher et al., 2005).

Asymmetry of non-glacial environments

The west-east gradient is supposed to be less pronounced in periglacial environments governed by uniform continental climate during sea lowstands. The mammoth tundra-steppe (called `hyperzone` by Velichko) extended across all of Eurasia and the dry Arctic shelves. Along the margin of the Late Weichselian ice sheet there is a wide belt of aeolian dune sands and niveo-aeolian coversands traced from the Netherlands to central Russia (Zeeberg, 1998) and farther to the south-eastern shores of the Barents Sea (Astakhov et al., 2007; Astakhov & Svendsen, 2011). A much wider periglacial zone is marked by sheets of cold loess with ice wedge casts. These features are very similar all along the last Scandinavian ice margin. However, Weichselian continental climates with permafrost in Western Europe were certainly more humid and less frosty than in eastern Siberia.

Much more arid conditions of the Late Pleistocene periglacial landscapes in eastern Siberia are recorded by the Yedoma-type subaerial formation of thick silts with long syngenetic ice wedges, sometimes called the `Ice Complex`. The traditional interpretation of this formation by Russian permafrost scientists as waterlain is hardly applicable as it conflicts with the monotonous granulometric and mineralogic composition, the mantle-like occurrence and the predominance of xerophilous flora and tundra-steppe fauna. Investigators who specially studied the eastern Siberian loess-like silts arrived at a more plausible origin of these sediments as mostly aeolian, deposited in harsh continental climate (Tomirdiaro, 1980; Péwé & Journaux, 1983).

Most interesting are palaeoclimatic indicators recently studied in the Upper Pleistocene icy silts of the Lena delta (the Bykovsky Yedoma). The oxygen isotope composition of long syngenetic ice wedges reveals extremely cold winters – mean January temperatures 7-8°C lower than today for the interval of 60-10 ka (Hubberten et al., 2004). On the other hand, insect assemblages testify to summers 6-8°C warmer than today in an arid environment with a precipitation level close to Antarctica. The aridity, rather than the summer temperatures, prevented the landscape from being forested during the LGM (Sher et al., 2005). These temperature deviations – winters 7-8°C colder than today and summers 6-8°C warmer – mean that the Late Pleistocene J–J index for the east Arctic coastlands was about 55-60°C, i.e. by 15°C higher than the present continentality. The implication is that the extremely continental climate of mountainous eastern Siberia (Fig. 1) during the last glaciation spread over the dry Arctic shelf.

The continentality in the Pleniglacial of northern Western Europe, measured from former frost cracks and insect assemblages, gave J–J values from 28°C to 33°C decreasing to 20°C during the final arid Pleniglacial after 20 ka (Huijzer & Vandenberghe, 1998). This means that the difference in continentality of about 25-28°C between the Arctic and Atlantic coasts during the coldest phase of the Weichselian Pleniglacial was practically the same as at present. There are not enough data to estimate the LGM J–J value in the eastern Siberian mountains, where now it reaches 60°C. Continentality of the time around 20 ka BP probably exceeded this value.

The structure of interglacials is more mysterious, but the strong west–east gradient is evident in warm intervals as well. The density of geologically studied peat deposits dating from the last interglacial west of the J–J 30ºC isoline is spectacular not only in western Europe but also in central Russia (Fig. 1). But their number decreases eastwards. Between the Urals and Yenissei only three thick (~ 1 m) peat lenses originating in the last interglacial are known, whereas present-day peat bodies (in places up to 10 m thick) occupy about 400 000 km^2 in West Siberia. According to Smith et al. (2004) the West Siberian peatlands contain 7-26% of all terrestrial carbon stored during the Holocene – a tremendous contrast between the last and present interglacials.

The meagre amount of interglacial peats in Siberia cannot be explained solely by the scarcity of investigators since there is no lack of marine interglacial sites. The distribution of peat deposits in Fig. 1 can be related to the much drier interglacial climate in Siberia. Despite the well-known fact that Eemian Atlantic water penetrated deep into the Taimyr Peninsula, the interglacial assemblage of small mammals in Trans-Uralia is patently arid (Maleyeva, 1982). Also, the lack of forest fauna in eastern Siberian warm intervals has been intriguing for a long time (Sher, 1991). Several interglacial peat deposits are recorded in the maritime lowlands of eastern

Siberia (Fig. 1). Still, the latest paleobotanic data from the New Siberian Islands indicate an Eemian climate significantly drier if compared to the present day (Kienast et al., 2006). This looks like another extreme versus the humid climate of the classical Eemian interglacial of Western Europe (e.g. Zagwijn, 1996).

In general, the west-east climate gradient appears to have been more pronounced during the last interglacial than in glacial and present-day environments. This conclusion does not fit with the estimates, based on pollen analyses, of Eemian climate in Siberia as having been more humid (Velichko et al., 1991). Local palynologists are, however, more cautious about estimating precipitation based on pollen spectra. They note that computerized general circulation models agree better with a diminished precipitation in the last interglaciation of eastern Siberia (Lozhkin & Anderson, 1995). East of the Urals, the diversity of plant taxa decreases, further limiting the applicability of paleobotanic data for paleoclimatic reconstructions. In addition, chronological control of warmer intervals of the Late Pleistocene in Siberia based on spurious `finite` radiocarbon dates is very poor (Sher, 1991; Sher et al., 2005), as evident from other geochronometric data (paper 14, Ch. III).

Extending the Late Pleistocene phenomenon of increasing aridity in Siberia back through geological history calls for caution. A wet climate of the Likhvin interglacial (the Holsteinian of central Russia) with Singil flora and forest elephants is well known and readily traceable to West Siberia, where huge alluvial plains with remnants of Pacific-type vegetation reflect a climate more humid than the present one. No reliable climate indicators are known from interglacial terrestrial sequences in eastern Siberia.

Paleogeographic and stratigraphic inferences

After the QUEEN studies of the last glacial cycle in northern Eurasia (Svendsen et al., 2004) it became evident that the northeastern shelves and coastlands of the Arctic Ocean were first to experience the cooling impact of the Ice Age to form extensive ice sheets. With the continentality increasing through a glacial cycle, i.e. precipitation and temperature progressively diminishing, Arctic and Siberian ice sheets were waning but subterraneous glaciation was growing. In Western Europe this trend is perceived as a result of progressive cooling. In terms of the entire continent, the leading trend of the glacial history looks like a progressive aridification of the northern margin (which was previously suggested by Velichko for the Late Pleistocene). The Late Weichselian ice sheet may be considered as a byproduct of such development that eventually resulted in the lopsided glaciation restricted to the westernmost Eurasian margin. The ice barrier growing in Scandinavia and on the western Barents Sea shelf probably amplified aridification of the rest of the continent.

Discussing the problem, it is necessary to take into account the quality of paleoclimatic signals for distinguishing cryochrons and thermochrons which in the Siberian Arctic differs from the simple relation \`cold-warm\`, as accepted in Europe. The climate of Arctic Siberia has always been cold, even during the thermochrons when the treeline could reach the coast. However, during global interglacials the Arctic Ocean encroached far onto the low Eurasian margin, thereby changing the extremely continental climate of Siberia into a less continental one.

The higher annual temperatures were mostly due to milder winters, whereas the increased precipitation and cloudiness could lead to even colder summers during interglacials, decreasing the J–J value. The growth of bogs and arboreal communities was accompanied by more active thermokarst. In contrast, cryochrons were manifested by greater amplitudes of annual temperatures, a replacement of limnic-palustrine and fluvial sedimentation by aeolian processes, a decreasing number of hygrophilous species and a growing diversity of cryoarid flora and fauna. In such an environment the principal signature of a thermochron is not temperature but decreased aridity and greater organic content in the mineral mass. Summer temperatures, being predetermined by cloudiness rather than by annual temperature background, can hardly indicate thermochrons in Siberia. Paleontological evidence for Siberian cryochrons suggests drier and better heated soils than in present-day cloudy summers (Sher et al., 2005).

In simplified terms, we conclude that whereas in the extreme west of Eurasia a change from interglacial to full glacial meant changing a warm and wet climate to a cold and dry one, in the extreme east it was instead a cold and dry climate that became doubly cold and dry.

The west-east change of climatic parameters influences other environmental elements, such as vegetation, as exemplified by the pollen diagrams in Fig. 2. The westward temporal shift of Holocene pollen optima shows unequivocally that pollen-based environmental chrono-taxons of north-western Europe, such as Boreal, Atlantic etc., not to mention regional Pleistocene subdivisions, are not applicable to the east of the continent. Their use (or rather misuse) by pollen analysts for Siberia introduces unnecessary noise into the discussion and may confuse geologists.

The asymmetric pattern of late Pleistocene glacial history means that the upper glacial sedimentary complex of different regions fits into different chronostratigraphic brackets. It is late MIS 5 in the north-east dominated by the Kara Sea ice dispersal centre, late MIS 4 in the Russian European north-west where the Barents Sea ice had its influence, and MIS 2 in the realm of the Fennoscandian glaciation. This pattern, taken as a model for earlier glacial cycles, may help to resolve some stratigraphic problems. The available geological data suggest that the ultimate drift limit (the margin of the Pleistocene maximum glaciation) is spatially diachronous. There is some evidence that this limit is pre-Holsteinian in central Siberia and is younger (i.e., it corresponds to MIS 8) in western Siberia (paper 4 in Ch. I). The Saalian ice maximum of Kara Sea origin, which is MIS 8 in western Siberia, is naturally replaced by the Fennoscandian Moscow glaciation in European Russia, which happened during MIS 6. The pattern repeats throughout glaciated Europe: the Dnieper ice advance of MIS 8 in the Ukraine is replaced by the classical Saalian MIS 6 of western Germany and the Netherlands (Ehlers & Gibbard 2004).

The drift limit in European Russia coincides with the margin of the oldest Don ice sheet (MIS 16) close to the 50th parallel, i.e., 1500 km south from the edge of the Baltic Shield (Velichko et al., 2004). However, in Western Europe pre-Elsterian tills appear only in Denmark at 56.5°N, just 200 km south of the Scandinavian Mountains (Houmark-Nielsen, 2004). The very asymmetric position of the Don lobe margin, more than 2000 km from the centre of the Fennoscandian ice dome is explained by an influx of additional glacial ice from the north-east (Velichko et al., 1987), i.e. from the Arctic ice dome located on the Kara shelf. In post-Holsteinian times a similar mighty ice invasion from the north-east over half of the Russian Plain occurred probably during MIS 8, whereas the next ice advance of MIS 6 was plainly dominated by the Scandinavian source (Astakhov, 2004a; Velichko et al., 2004).

Perspectives of future research

The results obtained so far still have a way to go for a final solution. Even the best studied Pleistocene in the extreme west (Great Britain) offers quite a number of riddles, such as Middle Pleistocene ice limits. We certainly need much more data on the extreme eastern flank of glaciated Eurasia. For example, the mechanism of inception of glaciation in the High Arctic during Pleistocene ice ages is not known due to the lack of geological data. It is not clear whether the Kara Sea shelf was first covered by inland ice due to the eastward shift of precipitation or whether the shelf's ice-cover originated from saline water due to extreme cold. The MIS 2 and MIS 4 ice limits in the QUEEN reconstruction (Svendsen et al., 2004) seem too speculative for the Kara Sea shelf and need further investigation. Any fresh data from the floor of the Kara Sea would be welcome.

There is hardly any doubt that in eastern Siberia the amplitude of Pleistocene climatic swings was considerably smaller than on the Atlantic seaboard. However, if the mode of ice sheets shrinking to the east is more or less clear, the fading biotic signal of interglacials still awaits proper study and measurement, especially as these relate to the transition from thermochrons to cryochrons. Although new drilling results from deep eastern Siberian lakes are spectacular, they are not enough. To better understand peculiar environments of the interglacials in the extreme east we need more detailed sedimentological and paleontological descriptions from both lake cores and natural sections throughout the Russian Arctic.

The diminished amplitude of Pleistocene environmental changes in the east poses a serious problem in terms of employing appropriate climatic indicators. For example, the discrepancy between the pollen record and magnetic and luminescence age models in the best studied sequence of Lake Elgygytgyn in Chukotka is puzzling. The boundaries between palynological `warm` and `cold` stages seem to deviate far from the prescribed chronostratigraphic brackets (Brigham-Grette et al., 2007). It is possible that conventional palynological approaches to paleoclimate reconstructions are not adequate for eastern Siberia. Since climatic reconstructions from scarce remains of monotonous Siberian vegetation are often ambiguous, looking at other proxies, such as Arctic paleosols and insect assemblages, is advisable. Considerable work has already been done on the oxygen isotope composition of fossil ice in Siberian permafrost (Vasilchuk & Kotlyakov, 2000), but this method has not yet exhausted its potential for climatic reconstructions.

Acknowledgements

This work has been carried out in the course of the project Ice Age Development and Human Settlement in Northern Eurasia (ICEHUS), supported by the Research Council of Norway.

22. PLEISTOCENE GLACIATIONS OF NORTHERN RUSSIA – A MODERN VIEW

VALERY I. ASTAKHOV

BOREAS, 2013, VOL. 42, p. 1–24

Introduction

Views on the volume and chronology of former glaciers of northern Russia have varied over time between (i) recurrent ice sheets of a similar size – the `Atlantic model`, based on the better studied cases of Western Europe and North America; and (ii) various, volumetrically diverse, glacial systems dwindling eastwards and progressively diminishing since the Middle Pleistocene – the `continental concept`, taking into account the huge extent of the Eurasian landmass with its permanent continental climate and old conservative permafrost.

Ideas concerning the geographical extent of former glaciations, as derived from Russian geological surveys and being less affected by theory, naturally drift towards concepts of ice thinning out in the east (Astakhov, 2011). However, the numerous radiocarbon dates that became available in the 1960-80s favoured the Atlantic model, in its most extreme form incorporating enormous Pan-Arctic ice sheets for the Last Glacial Maximum (LGM) (e.g. Grosswald, 1980). This radical idea which disregarded the harsh continental climate of Siberia with the predominance of underground glaciation (ice in permafrost) evidently emanated from the North American example. The maximalist models were eagerly adopted by geophysicists who were anxious to balance global sea-level fluctuations with huge volumes of Siberian inland ice.

However, the maximalist scenario was immediately challenged by Russian paleogeographers, who objected to it as climatically implausible and offered instead a minimalist alternative for the last glaciation (e.g. Velichko et al., 1997). Aside from this theoretical discussion, a different concept of the last glacial cycle has been elaborated in the course of the international research program QUEEN that started in 1993 (Svendsen et al., 2004).

Size and *modus operandi* of former glaciers

The conventional model

Ice volume, crucial for the global hydrological equation, is normally inferred from the calculated horizontal and vertical extent of former glaciers. Siberian ice sheets have traditionally been regarded as only a few hundred metres thick. Ice-flow patterns, inherently connected with the above parameters, have for many decades been deduced from the distribution of crystalline clasts. The crystalline erratics were essential for the classical idea of mountain ice dispersal centres elaborated after the Alpine example (Sachs, 1953; Yakovlev, 1956).

However, this approach is hardly a great help in vast sedimentary basins with no solid rocks on the surface. Thick lowland diamictons would inevitably consist of redeposited fine-grained materials with an admixture of stones from the Paleozoic uplands of the Urals, Timan and Central Siberia, irrespective of changes in flow patterns (paper 3, Ch. I). The arctic lowland diamictons, in places containing noticeable amounts of foraminifers and mollusk shells, have often tempted geologists to imagine former presence of a cold sea with floating ice shelves fed by montane glaciers of the Urals and Central Siberia (e.g. Sachs 1953; Zubakov, 1972). Somewhat thicker ice has often been suggested for the Subarctic zone, where marine fossils are scarce but crystalline erratics occur over a broad range of elevations.

Yakovlev (1956) in his *opus magnus* advocated a different viewpoint based on exotic indicator erratics. He postulated Middle Pleistocene ice more than a kilometre thick, flowing from Novaya Zemlya to the central regions of the East European Plain and across the Urals into West Siberia. Also Voronov (1964), using simple geometric relations of ice sheets and their limits, calculated a former ice dome on the Kara Sea coastal plains, reaching 3.5 km in thickness, only 0.8 km less than the present thickness of the Antarctic ice sheet.

Nevertheless, the majority of authors have for decades argued for Late Pleistocene glaciers in the form of thin piedmont ice sheets fed by highland valley glaciers, specifically the Ural Mountains (Yakovlev, 1956), the Putorana Plateau, the Byrranga Mountains of Central Siberia and the Verkhoyansk Range in East Siberia (Sachs, 1953; Arkhipov et al., 1976)

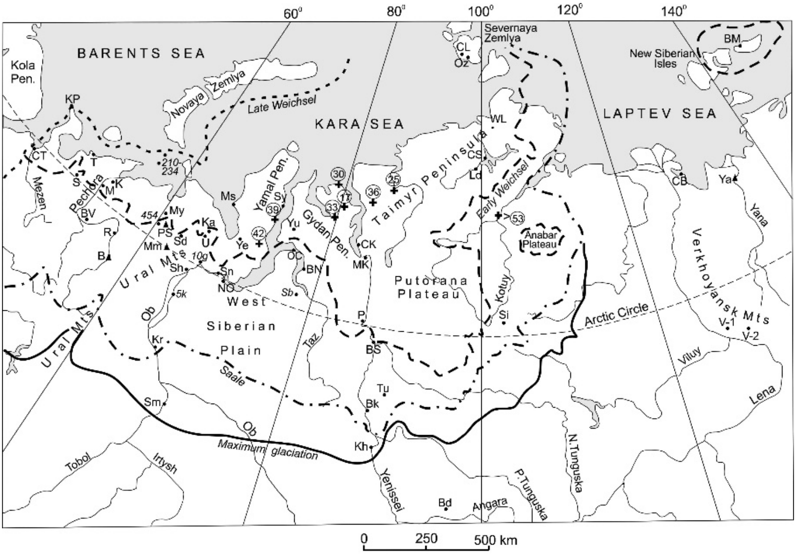

Fig. 1. Key sections and ice limits of the Middle and Late Pleistocene in northern Russia. *European part:* BV – Bolshaya Volma (Zarkhidze, 1972); CT – Cape Tolstik (Larsen et al., 2006); K – Kuya (Astakhov & Svendsen, 2008); Ka – river Kara (Nazarov et al., 2009); KP – Kanin Peninsula (Demidov et al., 2006); M – Markhida, T – Timan Coast (Mangerud et al., 1999; Henriksen et al., 2008); My – More-Yu (paper 19, Ch. IV); R – Rodionovo (Andreicheva, 2002; Arslanov et al., 2005); S – Sula, Sd – Seyda (Figs 10, 11; paper 3, Ch. I); U – Polar Urals (Mangerud et al., 2008). *Siberia:* Bd – Bedoba (Laukhin, 2011); Bk – Bakhta, Kr – Karymkary, Kh – Khakhalevka, Sm – Semeyka (Fig. 7); Sh – Shur (paper 12, Ch. III); BN – Belaya Yara and Nyunteda, BS – Bolshoi Shar, OC – Observations Cape, P – Poloy, NO – Nadym Ob, Sn – Sangompan, Ye – Yerkata, Yu – Yuribei (paper 16, Ch. III); CB – Cape Bykovsky (Sher et al., 2005); CK – Cape Karginsky, MK – Malaya Heta (Fig. 8); CL – Changeable Lake (Raab et al., 2003); CS – Cape Sabler, Ld – Ledyanaya (Möller et al., 1999); BM – Bukhta Mira (Basilyan et al., 2008); Ms – Marresale (Forman et al., 2002); Oz – Ozernaya (Möller et al. 2006); Si – Sida (Bardeyeva, 1986); Sy – Syo-Yaha (Vasilchuk et al., 2000); Tu – Mt. Bolshaya Tundrovaya (Fig. 3); V-1, V-2 – Verkhoyansk Range (Stauch & Gualtieri, 2008); WL – White Lake (Alexanderson et al., 2001). *Paleolithic sites (triangles):* B – Byzovaya (Mangerud et al., 2002; Svendsen et al., 2010); Mm – Mamontovaya Kurya (Svendsen et al., 2010); PS – Pymva-Shor (Mangerud et al., 1999); Ya – Yana (Pitulko & Pavlova, 2010). Mammoth carcasses are indicated by crosses; encircled are their radiocarbon ages, ka BP (Astakhov & Nazarov, 2010; Astakhov, 2011). *Boreholes (italics): 5k, 10g* (Arkhipov et al., 1992), *210–234* – boreholes on sea floor (Polyak et al., 2000), *454* (Zarkhidze, 1972), *Sb* – Samburg (Zubakov, 1972).

(Fig. 1). The conventional view of Pleistocene glaciations in West Siberia is illustrated in Fig. 2A.

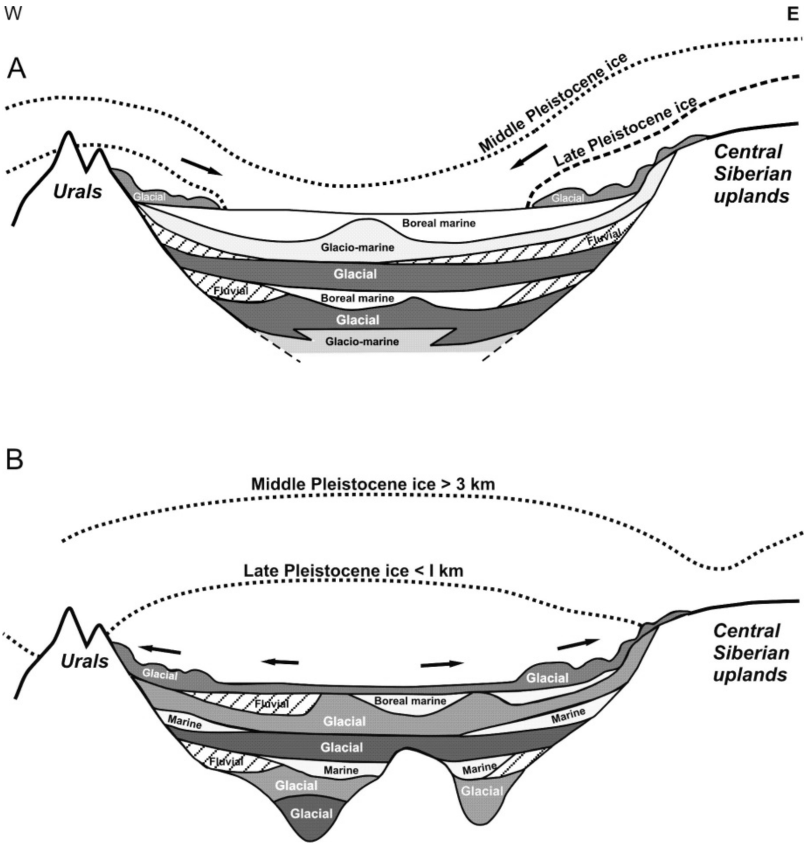

Fig. 2. Models of Pleistocene sedimentation in West Siberia as idealized cross-sections of sedimentary formations of different origin. A. Conventional (after Sachs, 1953; Yakovlev, 1956; Zubakov, 1972; Arkhipov et al., 1976). B. Modern (from Kaplyanskaya & Tarnogradsky, 1975; Astakhov, 1977; 2004). Note the opposite ice-flow directions (arrows) and different ice thicknesses inferred from a different genetic interpretation of thick diamicton formations.

In the 1960s and `70s the conventional glaciation model was under criticism, mostly from dissident anti-glacialists. They advocated much smaller alpine glaciers, refuting glacial geology in general, but providing no sound sedimentological or chronological evidence for such small former glaciers. The relevant discussion on this issue can be found in various Russian overviews (e.g. Arkhipov et al., 1976).

The modern model

Remote sensing data, widely employed since the 1970s, have helped in the discovery and mapping of numerous geomorphic features of ice flow directed from coastal lowlands towards Paleozoic uplands, thus contradicting the classical idea of predominantly down-slope ice flow from montane sources. Diamictons with foreign material and erratics of lowland provenance, located in the mountains, agree with the pattern of push moraines directed upslope (Astakhov, 1977, 2004a). Thorough sedimentological studies of key sections in the Arctic lowlands have demonstrated that fine-grained diamictons with marine fossils contain large detached blocks of marine sediments (Fig. 3) that must have been subglacially deposited by thick ice sheets advancing from the Kara Sea shelf (Kaplyanskaya & Tarnogradsky, 1975).

These findings imply that downslope ice flow from mountainous areas was restricted to the beginning of a glacial cycle. The following glacial maxima are attributed to thick ice domes on coastal lowlands and to dry arctic shelves that discharged ice streams upslope (Kaplyanskaya & Tarnogradsky, 1975; Astakhov, 1977, 2004). The new results were immediately adopted for maximalist glaciological reconstructions (Grosswald, 1980). The predominance of shelf-based accumulation centres was also later supported by international studies in the Russian Arctic (e.g. Forman et al., 2002; Svendsen et al., 2004; Möller et al., 2006, 2011).

The major ice domes, reconstructed on the Barents and Kara Sea shelves and coastal lowlands, imply thick ice also farther inland. Pervasive glaciotectonic deformations, penetrating into soft clay and sand of up to 300-400 m beneath the glacier sole, is evidence of the great thickness of former glacial ice in subarctic West Siberia (paper 8, Ch. II). Diamictons of lowland provenance, found on high flat mountains up to 600 m a.s.l., as well as in overdeepened valleys down to 200-300 m b.s.l., strongly support this interpretation (Fig. 4, also Fig. 4 in paper 3, Ch. I).

In Central Siberia the range of altitudes of erratic materials indicates Middle Pleistocene ice at least 800 m thick just 100 km upglacier from the ultimate ice limit. The Eemian shorelines on Severnaya Zemlya and the Taimyr Peninsula, located at ~140 m a.s.l., suggest Saalian ice more than 3 km thick in the eastern Kara Sea (Möller et al., 2006). Farther west the Saalian ice was even thicker, judging by the shorelines up to 420 a.s.l. m in Novaya Zemlya (Sachs, 1953). The thickness of the Saalian ice of the Kara

Sea is estimated as 4.5 km by a modern model (Lambeck et al., 2006). Even in the Late Pleistocene ice-sheet thickness exceeds any previous estimates, as indicated by horseshoe-shaped moraines on the northern tip of the Urals, pushed from the northern piedmont uphill to 560 m a.s.l. (paper 18, Ch. IV).

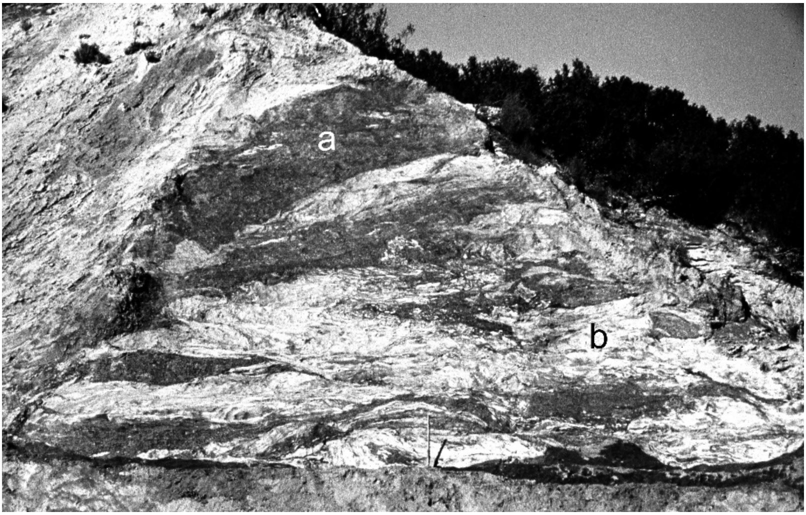

Fig. 3. Key section of Sanchugovka Formation in Nikitinsky Yar (MK in Fig. 1), northern bank of the Yenissei, 69°45′ N/84°40′ E, 1-m stick for scale. Note the upthrusted pale blocks of frozen waterlain sand (b) within dark diamicton (a) with dispersed Cretaceous and Quaternary shells. Interpreted as marine formation by Sachs (1953) but basal glaciotectonic melange by Kaplyanskaya & Tarnogradsky (1975). Photograph by the last authors.

The new glaciation model (Fig. 2B) also takes into account the map pattern of ice limits (Fig. 1) and ubiquitous glaciotectonic features. The numerous arcuate imbrications, typical for glaciated West Siberia, display alpine-type folded and overthrust structures (Arkhipov et al., 1976). They commonly occur far upglacier from the former marginal zones and are better explained by locally obstructed motion of very thick ice. A former ice front in West Siberia is normally marked only by proglacial varves.

The widespread occurrence of proglacial ice-dammed lakes is probably the reason for the absence of push moraines along the margins of ice sheets in West Siberia (papers 3 in Ch. I, 14 in Ch. III). Large push moraines, produced by active ice margins, are more readily encountered west of the Urals (paper 18, Ch. IV), where the Laya-Adzva double morainic ridge, more than 200 km long, is the most prominent (Fig. 5). However, many

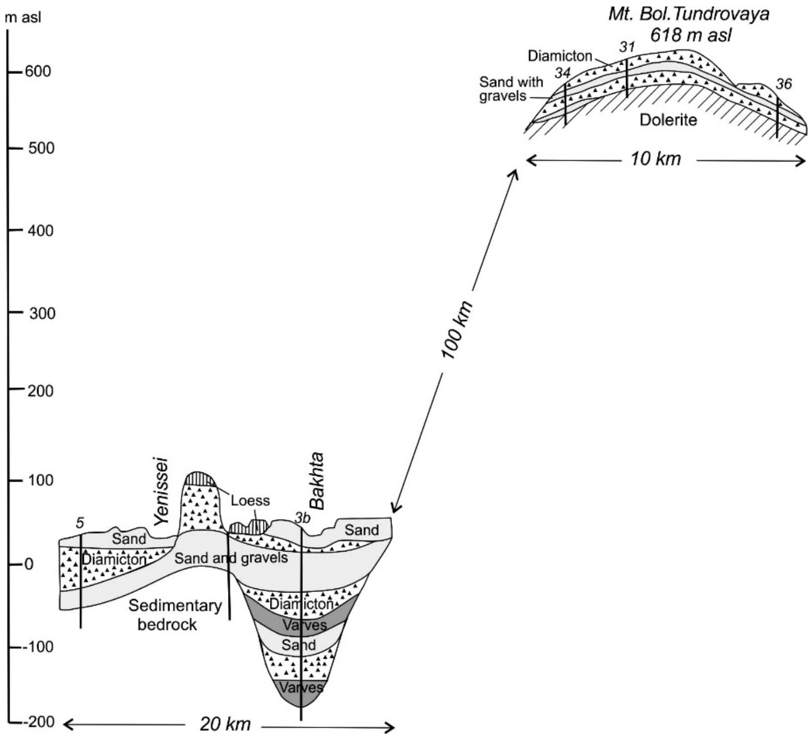

Fig. 4. Middle Pleistocene glacial deposits with lowland erratics in Yenissei Siberia From Yu. Fainer: in Arkhipov et al. (1976). Note the difference between altitudes of foreign diamictons penetrated by boreholes in the valley (Bk in Fig. 1) and on the dolerite inselberg (Tu in Fig. 1).

morainic arcs in the Pechora Basin are believed to have been produced subglacially by the squeezing of clayey materials from beneath thick stagnant ice and not by marginal ice push (Lavrov, 1981). On the whole, the modern estimates of Pleistocene ice thicknesses, based on former ice limits, the glaciotectonism and altitudes of lowland erratics, agree better with the maximum estimates of Voronov's (1964) model than other models.

Another important discovery is the numerous thick slabs of buried ice (Kaplanskaya & Tarnogradsky, 1986) with the characteristic foliated and folded structure of dirty glacial ice (Fig. 6A). Such ice bodies, ubiquitous in perennially frozen Siberia, can be up to 60 m thick and kilometres across (papers 5 and 8 in Ch. II). Their size and association with thick frozen diamictons together with the petrographic properties of metamorphic ice described in thin sections (Solomatin, 1986) exclude the hypotheses of segregation or injection of modern ground ice favoured by many geocryologists. Thick fossil glacier ice in extensive outcrops has been encountered, at odds with expectations, even in the far east of the Russian Arctic (Fig. 6B, C). This stratiform ice is truncated by a marine formation, probably Eemian, and therefore must be a remnant of a Middle Pleistocene ice sheet (Basilyan et al., 2008). Fossil glaciers beneath interglacial marine deposits are also known in Novaya Zemlya (Sachs, 1953). In West Siberia, however, fossil glacier ice occurs only in the area of the Late Pleistocene glaciation (paper 14 in Ch. III).

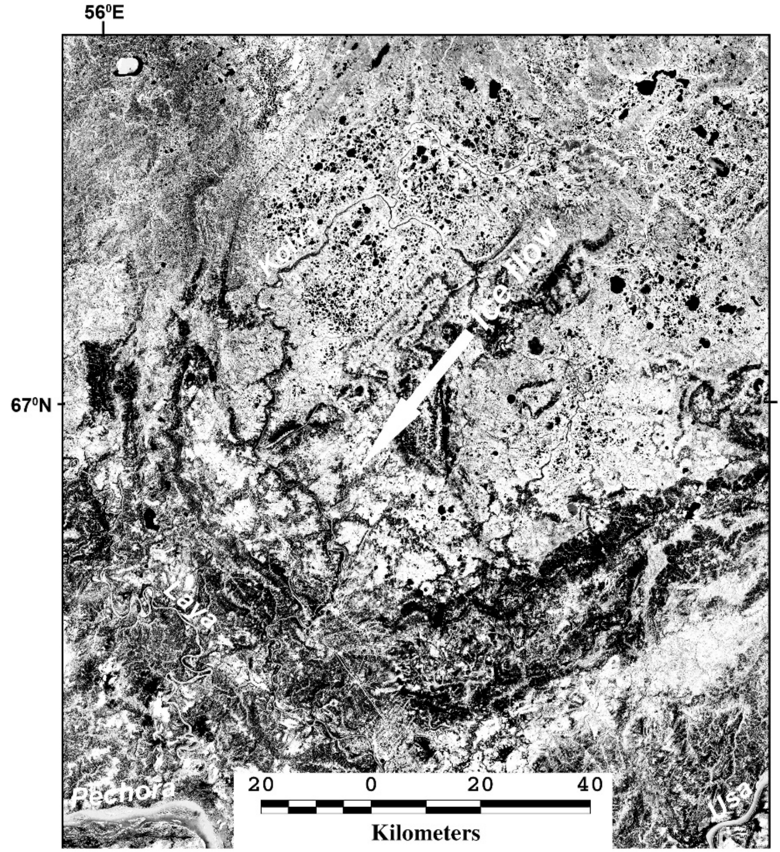

Fig. 5. Laya-Adzva morainic ridges indicating a marginal lobe of the Last Kara Sea ice sheet in the Pechora Basin between 56° and 60° E (satellite view).

The large bodies of glacier ice, tens of metres thick and laterally merging into frozen diamictons, are thought to be constituents of `primordially frozen tills` (Kaplanskaya & Tarnogradsky, 1986). These result from subglacial deposition typical for perennially frozen lowlands. Such an ice/debris mixture is simply a buried dirty sole of the former glacier, that is, glacial deposit *sensu stricto* (paper 8 in Ch. II). According to the INQUA definition (Dreimanis, 1989), it *cannot be termed `till` because it has never melted out*. South of the Arctic Circle the basal diamictons are melted out but the underground melting, prolonged for thousands of years, is a diagenetic rather than a depositional process. On the whole, Siberian subglacial deposits cannot be termed tills because they consist not only of diamictons but also of large amounts of glacial ice, detached blocks of stratified drift and pre-Quaternary sediments. In this respect the old term `ground moraine` seems more appropriate (paper 5 in Ch. II). I will refrain from using both `ground moraine` and `till` labels except for stratigraphic names produced by other authors.

Although very thick diamictons with well-preserved structures of moving ice, as in Fig. 3, are ubiquitous in Siberian sedimentary basins, a lodgment process is hardly conceivable for the base of cold glaciers acting on soft perennially frozen substrata. The arctic subglacial deposits are in many cases formed by the progressive stagnation and exfoliation of basal ice. This way it was so debris-laden and rigid that it became detached from the active ice and joined with the subglacial permafrost while the rest of the glacier slid farther over this new icy substrate (Shantser, 1966). Several stages of this process can be directly observed live at the base of present glaciers of the Canadian Arctic (e.g. Evans, 2009: Fig. 4).

Signatures of wet-based ice flow such as flutes, drumlins, eskers, boulder pavements, lodgment tills, large striae, etc. are virtually absent in the perennially frozen sedimentary basins. In Siberia, however, they occur in the Paleozoic uplands or along valleys of the great rivers (e.g. Möller et al., 2011). In such environments basal sliding is facilitated either by wet soils of subglacial taliks or by friction heat released at the interface of ice and hard bedrock.

Astakhov et al. (paper 8, Ch. II), who described the subglacial features of West Siberia, presume that cold glaciers in perennially frozen sedimentary basins were largely able to advance by sliding along englacial thrust planes or by cataclastic flow of deforming frozen clay of the substrate. The large hill-hole pairs together with far-travelled blocks of undisturbed sand (Kaplyanskaya & Tarnogradsky, 1974, 1975) imply that Siberian cold glaciers constituted one entity with the permafrost hundreds

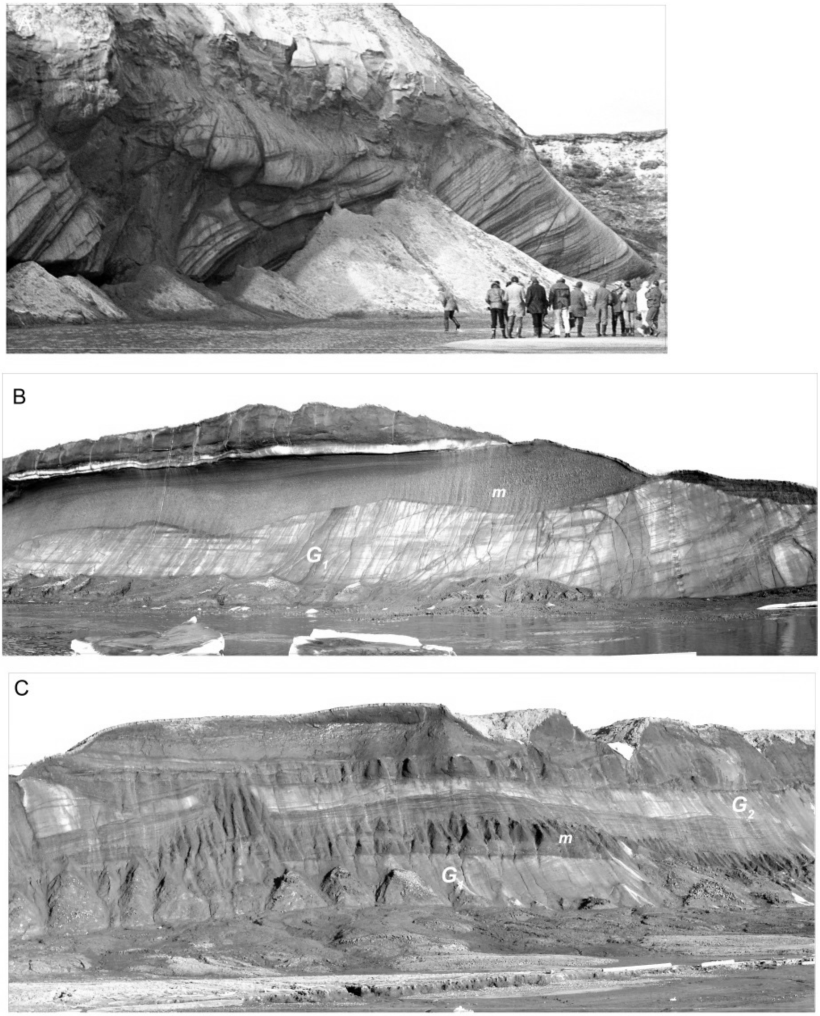

Fig. 6. Subglacial deposits in arctic Siberia. A. Deformed fossil ice with dirt bands, small faults and thrusts covered by melt-out silt in the Gydan Peninsula, 70° N/75° E (Yu in Fig. 1). B and C. Easternmost and oldest fossil glaciers exposed in coastal cliffs of the New Siberia Island, MB in Fig. 1. B. Glacier ice of Middle Pleistocene age (G_1) overlain by marine clay (m) with distinct thaw contact. C. Same members as in (B), overlain by another sheet of fossil glacial ice (G_2). Bluffs are 35 m high. Photos are by V. Tumskoy. See centrefold for this image in colour.

of metres thick. These glaciers vigorously eroded the soft substrate by glaciotectonic excavation when their motion was topographically, lithologically or thermally obstructed. The cold ice slid far southwards along englacial overthrusts, leaving dead slabs of dirty basal ice behind. The numerous clay diapirs and erratic blocks of loose sand transported over hundreds of kilometres are indications of the irregularly obstructed ice motion (paper 8 in Ch. II).

The mode of deglaciation in the northern lowlands is also peculiar and differs from that in the Paleozoic uplands. The Putorana and Anabar plateaus display discontinuous belts of morainic ridges, forming concentric patterns (Fig. 1) (Arkhipov et al., 1976). Ice-retreat morainic loops are also observed on the eastern piedmont of the Polar Urals and in the Taimyr Peninsula, especially on the southern slope of the Byrranga Mountains (Kind & Leonov, 1982; Möller et al., 2011). In these cases, glacier ice, descending over the hard slippery bedrock and releasing friction heat, accelerated basal sliding and formed a lobate pattern of retreating ice margins.

In the sedimentary basins, however, the pattern of parallel, upglacially recurrent marginal features is rarely seen. In West Siberia, where the largest push moraines were formed subglacially far away from the ice margin, the lack of ice-retreat landforms is very conspicuous (papers 3 in Ch. I, 8 in Ch. II). Although glaciofluvial deltas, end moraines and valley trains occur in places in the European Arctic (paper 18 in Ch. IV) they are unknown in the West Siberian lowlands, where the ice-retreat features are predominantly kames and proglacial lacustrine formations.

It seems that after rapid initial disintegration, caused by sea-level rise, the ice-sheet decay proceeded mostly in the form of huge fields of stagnant ice, eventually buried by meltout debris. The slow degradation of stagnant ice fields with recurrent thermokarst inversions of topography was governed by the postglacial climate, which in the Late Pleistocene Arctic was hardly favourable for steady melting. The expressive glaciokarst landscapes of disintegrating ice sheets, such as hummock-and-lake topography, appeared largely in the Holocene, irrespective of the age of active ice advance. This prolonged `retarded deglaciation` is conducive to stratigraphic blunders, as multiple ablation formations post-dating the principal ice advance would readily provide a wide range of dates and diverse climatic signals (papers 5 and 8 in Ch. II).

The features discussed in the above publications imply that former ice sheets over the Russian North, owing to the perennially frozen soft substrate, behaved in many cases in a distinctly different way from ice sheets in the Atlantic environments of Western Europe and North America.

Taking into account that the glaciated Siberian lowlands covered millions of square kilometres, the above glacial features indicate not just a local peculiarity of the Ice Age but rather an alternative climatic type of inland glaciation. Thick sluggish glaciers merged with the old conservative permafrost must have grown and decayed very slowly, the decay sometimes being urged by the meager influx of solar energy (paper 5 in Ch. II). The areas west of the Urals display an intermediate situation, with the Siberian features of glacier/permafrost interaction mixed with glacial formations of Atlantic type (paper 18, Ch. IV).

Conventional chronology of former glaciations

Middle Pleistocene

Quaternary formations below the Bruhnes-Matuyama boundary in Russia are formally referred to as the Eopleistocene, with the overlying strata belonging to the Lower Pleistocene (now labelled the Lower Neopleistocene). The long and humid interglacial of Likhvin (=Holsteinian) is conventionally placed at the beginning of the Middle Neopleistocene. Sediments of the last interglacial-glacial cycle constitute the Upper Pleistocene (=Upper Neopleistocene) (Resolutions, 2008). These subdivisions of the General Stratigraphic Scale are filled with regional correlation units termed `climatoliths`, which are either `kryomers` or `thermomers` (similar to the `geologic-climate units` of the American Stratigraphic Code). The climatoliths corresponding to glacial periods or major glacial stages serve as chronostratigraphical landmarks of Pleistocene glacial history instead of European substages.

The climatoliths of the regional stratigraphic scheme of glaciated Siberia, based on works by Arkhipov and colleagues in West Siberia, were endorsed by the National Stratigraphic Committee in 2000 as a standard (Table 1). Many researchers use these subdivisions for other Siberian regions, without trying to identify them with the real sedimentary bodies of West Siberia. The understanding of continental glacial history is seriously hindered by such indiscriminate usage of problematic chronostratigraphic labels. This is briefly overviewed below.

The subdivisions of the early Middle Pleistocene are derived from Ob and river Irtysh sections between 59° and 64° N, whereas the upper part is inferred mostly from arctic Yenissei exposures. All glacial sediments show only normal magnetic polarity (Arkhipov, 1989). The lowermost `Mansi Moraine`, namely the diamicton lying on top of the oldest fluvial sand, sporadically shows in boreholes of the overdeepened Irtysh valley. It is covered

Table 1. Conventional schemes of glacials (bold) and interglacials (*italics)* in northern Russia

Unified Stratigraphic Scheme of West Siberia (Volkova & Babushkin, 2000)			Timan-Pechora-Vychegda Region (Guslitser et al., 1986)	Correlation units for European Russia (Resolutions, 2008)	Northern Europe	MIS
Climato-chronological units	Ages, ka by Arkhipov (1989)					
	^{14}C and TL	ESR on shells				
Sartan	^{14}C=10-22(23)		**Polar glacial**	**Ostashkov** (Upper Valdai)	**Late Weichsel**	**2**
Karginsky	^{14}C=22(23)-50	51.6±12.8	*Byzovaya interstadial*	*Leningrad interstadial (Middle Valdai)*	**Middle Weichsel**	*3*
Yermakovo	TL=100±17; 110±27		**Laya cold stage**	**Kalinin** (Lower Valdai)	**Early Weichsel**	**4**
Kazantsevo	TL=130±25	121.9; 134.8	Sula interglacial	*Mikulino interglacial*	*Early Weichsel* *Eem*	*5*
Taz			**Vychegda glacial**	**Moscow glacial**	**Saale**	**6**
Shirta	TL=180±40; 190±36	196.8±20.6; 146.9 ; 161.8; 193.0	*Rodionovo interglacial*	*Gorka interglacial*	*Dömnitz/ Wacken*	*7*

Samarovo	TL=200±38;230(240)±51 (54)		**Pechora glacial**	**Vologda glacial**	**Fuhne**	**8**
Tobol	TL=260(270)±56 300(308)±75; 313±75; 380±65; 390±80	306.2±20.8; 298±28; 336±30; 356±34	*Chirva interglacial*	*Likhvin interglacial*	*Holstein*	*9-11*
Upper Shaitan	TL= 420(445)±110; 510±65		**Pomusovka glacial**	**Oka glacial**	**Elster**	**12**
Tiltim	TL=550±110 ; 560±140 600±70		*Visherka interglacial*	*Muchkap interglacial*	*Cromer*	*13*
Lower Shaitan			**Beryozovka Glacial**	**Don glacial**		**16**
Talagaika	TL=660±160; 740±170		*Tumskaya interglacial*	*Ilyinka interglacial*		*17*
Mansi			**Kama glacial**	**Likovo glacial**		**18**

by pre-Holsteinian interglacial alluvium of the Talagaika Formation (Fig. 7) and higher up by the thickest double diamicton formation, locally labelled the `Shaitan Moraines`. The Shaitan diamictons can occasionally be seen at the base of higher river bluffs some 200-300 km upglacier of the drift limit (e.g. Kormuzhinanka in Fig. 7).

Their low stratigraphic position suggests that they are counterparts to the double Elsterian tills and `Oka Moraine` of Central Russia. Southwards, the Shaitan diamictons are replaced by glaciolacustrine clays and silts of the Semeika Formation sandwiched between TL ages 600 and 380 ka (Fig. 7).

The full skeleton of a *Mammuthus trogontherii Polig,* recovered from the base of the Semeika Formation (Kosintsev et al., 2004), is seemingly in accordance with these otherwise very suspicious dates. A few older and even more doubtful TL dates were reported from the Matuyama Chronozone below the Mansi diamicton but are not verified by different methods or in other sections. The luminescence (and any other) dates cannot be taken for an `absolute age`, although they may be useful as correlation signals.

The Elsterian chronostratigraphic position for the Shaitan diamictons and Semeika clays is also inferred from the overlying Tobol Formation, alluvium of a great river with shells of the freshwater Central Asia mollusk *Corbicula tibetensis* and remains of southern taiga plants. The Holsteinian age of this alluvium is supported by an ESR (Electron Spin Resonance) age of 326.9 ka on *Corbicula* shells and by TL sediment ages of the same order (Fig. 7). The overlying Samarovo glacial complex, extending to the drift limit, has for decades been associated with an Early Saalian (MIS 8) chronostratigraphic level (Arkhipov, 1989; Arkhipov & Volkova, 1994).

The younger Taz glacial complex, underlying the strata of the last interglacial, is traditionally correlated with the late Saalian `Moscow Moraine` of Central Russia. The `Taz Moraines` of the final Middle Pleistocene are widespread on the surface of West Siberia between 63° and 69° N, although they never extend to the ultimate ice limit (Fig. 1). The independence of the Taz ice advance from the Samarovo maximum is dubious because the intervening Shirta sands do not contain typical interglacial organics.

This fact provoked disputes about the age of the Samarovo glaciation, which, in the absence of proven Samarovo-Taz interglacial strata, might also relate to MIS 6. However, the lack of an erosive contact between the Tobol alluvial strata and Samarovo glaciolacustrine silts (Kaplyanskaya & Tarnogradsky, 1974) does not support the latter idea. A gradual transition of the interglacial formation upwards into the glacial sequence permits the Samarovo glacial to be even older than MIS 8, provided that the Tobol strata really belong to MIS 11.

Anyway, the great thickness of the diamictons, underlying the Tobol interglacial strata in all major buried valleys, suggests that the Elsterian or earlier ice sheets of West Siberia were very thick. It is quite possible that several ice advances remain indiscernible within the 70-100 m thick diamict sequences of the buried valleys, where no reliable interglacial formations have been reported so far.

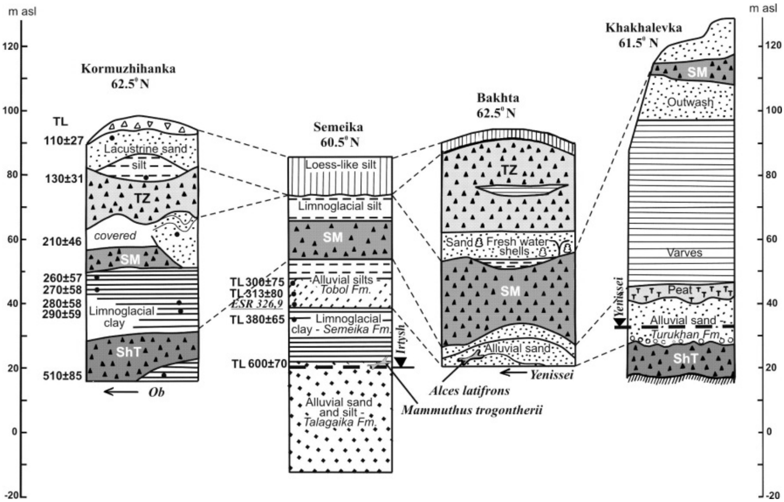

Fig. 7. Middle Pleistocene glacial formations of West Siberia as measured in natural sections along the Ob (Kr, Sm) and Yenissei rivers (Bk, Kh in Fig. 1). Glacial diamictons (shaded) according to the Stratigraphic Scheme of West Siberia (Volkova & Babushkin, 2000) belong to: Sht – Shaitan glaciation of early (pre-Holsteinian) Middle Pleistocene; SM – Samarovo (maximum) glaciation (presumably MIS 8); TZ – Taz glaciation of late Middle Pleistocene (MIS 6). Tobol and Talagaika alluvial formations contain pollen of boreal forest (by Levina 1964; Zubakov 1972; Kaplyanskaya & Tarnogradsky 1974; Arkhipov 1989).

The subdivision of the Middle Pleistocene glacial history based on the Yenissei river sections (Fig. 7) is also adopted for Central Siberia (Isayeva et al., 1986). The Turukhan alluvium with taiga pollen from Yenissei boreholes is conventionally accepted as a counterpart of the Tobol alluvium, although this correlation has never been tested by geochronometry. Bakhtinsky Yar at the Bakhta river mouth (Bk in Figs 1, 7) is a rare section where two Middle Pleistocene diamictons lie directly on interglacial alluvium with a bone of *Alces latifrons.* This elk is typical for the Tiraspol

(Cromer) mammalian complex. Therefore an Elsterian age of the lower glacial complex on the Yenissei cannot be excluded. Upstream, the base of the succession is represented by a thick diamict formation found in a borehole at 342 m b.s.l. This so-called `Lebed Moraine` (diamicton) constitutes the lowermost unit of the standard stratigraphic scheme for Central Siberia (Isayeva et al., 1986).

In the Arctic, the situation is more complicated because hypothetical marine analogues of the subarctic interglacial strata have few and often dubious TL ages. These have been obtained on sand cores from terrestrial wells drilled for geological surveys (Arkhipov et al., 1992). Traditionally, the Ob marine strata with a subarctic foraminifer assemblage, lying deep below the present sea level, were taken for a northern counterpart of the Tobol and Turukhan alluvial formations. However, the Ob assemblage of arcto-boreal foraminifers, previously identified by Gudina (1976) as the Holsteinian `zone of *Miliolinella pyriformis`,* was later found in cores at three different stratigraphic levels and therefore cannot be used as a biostratigraphic marker (Volkova & Babushkin, 2000). Several glacial events in the Arctic are deduced mostly from the count of thick diamictons and intervening stratified formations with rare foraminifers.

A similar scheme with a four-fold climatostratigraphic subdivision for each series (`zveno`=link in the Russian terminology) of the Neopleistocene was long ago formally adopted for the Timan-Pechora-Vychegda region west of the Urals based on sections in the Subarctic zone (Table 1). This scheme assumes a Late Weichselian age of the Late Pleistocene ice sheet and two post-Likhvin inland glaciations in the Middle Pleistocene (Guslitser et al., 1986; Andreicheva, 2002). The `Vychegda Moraine` with Fennoscandian erratics of the final Middle Pleistocene is accepted there as a counterpart of the `Moscow Moraine` of Central Russia, which is correlated with Saale and MIS 6. The underlying Rodionovo peat was thought to represent the MIS 7 time span, whereas the thickest exposed diamicton of the Pechora glaciation was traditionally correlated with MIS 8 (Table 1).

The diamict formations of the Lower Neopleistocene (i.e. the early Middle Pleistocene) are identified in boreholes beneath strata with pollen spectra of Likhvin (Holstein) type. The oldest glaciations, labelled Kama, Beryozovka and Pomusovka, left fine-grained diamictons of lowland provenance even in the Urals, and therefore should be associated with shelf-based ice domes. Only the Vychegda diamictons contain erratics readily recognizable as having been transported from the north-west, partly from Fennoscandia (Guslitser et al., 1986; Astakhov, 2004a).

A different and thicker sedimentary succession is found in the arctic part of the Pechora Basin. Profiles of deep wells drilled from the Barents Sea shore southwards through the 200-250 m thick Quaternary succession uncovered five thick diamictons intercalated with five marine formations (Lavrushin et al., 1989; Astakhov, 2004a: Fig. 2). The lowermost marine silts of the Kolva Formation contain a fairly cool subarctic fauna. The next marine formation, labelled Padymei, contains boreal elements, including the extinct mollusk *Cyrtodaria angusta* (Zarkhidze, 1972). In the local stratigraphic schemes two lower formations are related to the Lower and Middle Pleistocene, although no geochronometric dates are available. The two Upper Pleistocene marine formations, observable in natural sections, have recently been dated by optically stimulated luminescence (OSL) and U/Th dating techniques (see below).

The crucial issue for dating glaciations and their correlation with the European glacial events is the stratigraphic position and astronomical age of the interglacial strata. Within the Middle Pleistocene these stratigraphic markers are the terrestrial Rodionovo and Tobol formations in the subarctic Pechora and Ob river catchments.

Late Pleistocene

The weakness of the traditional schemes of glacial history (Table 1) is evident in the best studied upper part of the composite stratigraphic succession. Formations crucial for the Late Pleistocene history are represented by organic-rich beds at two stratigraphic levels. In Siberia these markers are two marine formations with boreal fauna, indicating successive inflows of Atlantic water into the Arctic: the Kazantsevo and Karginsky, as originally outlined from sections on the Lower Yenissei (MK to CK in Fig. 1). They were distinguished by mollusk faunas (Sachs, 1953) and later by radiocarbon dates (Kind, 1974).

The older formation of near-shore marine sediments, the Kazantsevo, has for decades been correlated with the Eemian of Europe and used as a representative of the main thermomer at the base of the Siberian Upper Pleistocene (Table 1). The Kazantsevo formation contains common boreal species such as *Arctica islandica* and the indicator extinct mollusk *Cyrtodaria jenisseae (=C. angusta* in modern systematics) (Sachs, 1953). The younger Karginsky formation is associated with an even warmer sea, inferred from a boreal foraminifer assemblage with lusitanic elements (Gudina, 1976; Kind & Leonov, 1982; Levchuk, 1984).

This fact, together with the persistent Siberian correlation of this last Atlantic ingression with the cool MIS 3 stage, has for many years been puzzling for non-Siberian geologists. The only argument in favour of such a correlation has been provided by a number of finite radiocarbon ages, obtained by conventional dating of shells and by bulk dating of plant material. Similar ages have also been reported from alluvial sediments with interglacial flora from the Central Siberian uplands (Kind, 1974; Kind & Leonov, 1982; Bardeyeva, 1986).

According to Sachs (1953), the most extensive Kazantsevo marine basin appeared prior to the main glaciation of the Late Quaternary. The latest Karginsky ingression with a similar arctoboreal fauna was suggested to postdate the main stage of the Late Pleistocene Zyryanka glaciation. In the 1970s, the Karginsky marine formation and its fluvial counterparts were assumed to separate two independent glaciations of the last glacial cycle: the major Early Zyryanka (Yermakovo) and minor Late Zyryanka (Sartan) (Table 1). Because of the correlation of the Karginsky strata with MIS 3, the Late Zyryanka glaciation was synchronized with the Late Valdai-Late Weichselian glaciation of Europe (Table 1) (Kind, 1974; Isayeva et al., 1986; Arkhipov, 1989; Volkova & Babushkin, 2000). Maximalists treated the Zyryanka and even Taz glacial formations as Late Weichselian. A more detailed history of the Upper Pleistocene stratigraphy can be found in paper 16, Ch. III, where a comparison of Siberian events with the European Late Pleistocene is presented.

In the European Arctic, the last interglaciation has long been associated with a marine formation labelled the `Boreal Strata`. The boreal mollusk fauna of this formation, containing *Cerastoderma edule, Arctica islandica, Spisula elliptica* and other species of ice-free seas, contrasts with the present-day subarctic fauna of the White and eastern Barents seas, and especially with the arctic fauna of the icy Kara Sea (Funder et al., 2002). The Boreal Strata have always been correlated with the last terrestrial interglacial formation and the corresponding thermomer termed Mikulino, a counterpart of the West European Eemian (Table 1). Another warm event of the European North, inferred from rich forest pollen in terrestrial facies, is often related to a prolonged intra-Weichselian interstadial (the Byzovaya in Table 1) on the basis of several finite radiocarbon dates (e.g. Arslanov et al., 1987; Andreicheva, 2002).

Updated glacial chronology

Dating interglacial markers

Marine formations. The most important recent results concern re-dating by different chronometric methods of the main stratigraphic markers, namely sedimentary units with paleontological signatures of climates similar to or warmer than the present one. The postulated ages for two Weichselian ice advances in the Siberian schemes are MIS 4 for the Early Zyryanka and MIS 2 for the Late Zyryanka. This depends on the assumed ages – Eemian for the Kazantsevo Formation and late Middle Weichselian for the Karginsky Formation (Table 1). Therefore, the *experimentum crucis* would demand redating the Karginsky interglacial strata, presumably deposited between these two glacial events.

Whereas traditional correlation was performed mostly using conventional radiocarbon dates, derived from random bulk samples, and occasional thermoluminescence (TL) dates, the modern chronometric estimates are obtained by accelerator mass spectrometry (AMS) ^{14}C analysis, OSL techniques, ESR, uranium series and cosmogenic exposure dating. Most important is that the modern dates, reported since 1993, are far more numerous and thus allow for testing the obtained ages by statistics and also by comparison of parallel dating by different methods. For example, the ^{14}C ages were statistically constrained at key sections of Byzovaya (22 dates), at Cape Sabler (60 dates), at Cape Bykovsky (90 dates) and at Yana (35 dates) (sites B, CS, CB, Ya in Fig. 1) (Möller et al., 1999; Svendsen et al., 2004, 2010; Sher et al., 2005; Pitulko & Pavlova, 2010).

The new techniques usually give older ages for a given sedimentary unit (e.g. AMS ages of small twigs, moss stems, vascular plants are often infinite), whereas conventional radiocarbon dates on driftwood or randomly assembled large volumes of plant detritus yield finite values from the same formations (Mangerud et al., 1999; Forman et al., 2002; Astakhov & Mangerud, 2005; Astakhov & Nazarov, 2010). The validity of correlation signals provided by modern geochronometry is proved by parallel OSL and ^{14}C dating of the Paleolithic sites (Mangerud et al., 2002; Heggen et al., 2010; Svendsen et al., 2010) and also by the tight clustering of similar radiocarbon ages at a given stratigraphic level (Pitulko & Pavlova, 2010). In spite of the age scatter, increasing downwards and becoming large in the Middle Pleistocene sediments, the mean values of Late Pleistocene OSL series are verified and supported by similar chronometric values provided by other methods such as ESR and U/Th dating (Mangerud et al., 2002; Astakhov & Mangerud, 2005, 2007).

The modern dates amply demonstrate that the last interglacial warming of the Siberian Pleistocene occurred not during MIS 3, as suggested by conventional wisdom (Table 1), but much earlier, most probably within the MIS 5 time-span, as elsewhere (paper 16 in Ch. III)). The new chronometric values from the Karginsky marine and fluvial strata with interglacial organics in the stratotype sections of the Lower Yenissei (MK and CK, Fig. 1) are much older than those suggested by conventional radiocarbon dates (see Fig. 2 in paper 16, Ch. III)). The first dating revision was the 121.9 ka

ESR age from Karginsky marine shells (Arkhipov, 1990), followed by six OSL dates with a mean age of 111 ka from the same sequence (Nazarov et al., 2009). The nearby terrestrial counterpart of the Karginsky Formation, when redated, yielded only infinite ^{14}C ages and OSL ages between 112 and 80 ka BP (Astakhov & Mangerud, 2005).

Four marine formations with foraminifer assemblages, intercalated with four units of subglacial deposits (`tills`), have recently been dated on the Severnaya Zemlya Islands using OSL and ESR techniques (Möller et al., 2006). The two lowermost formations, containing high arctic fauna, have been dated to 300-400 and 220 ka BP. The lowermost marine formation M-I is tentatively correlated with the Tobol alluvium of West Siberia, whereas M-II probably relates to MIS 7. Hence a MIS 10 or MIS 12 age is assigned to the lowermost till. The two uppermost tills probably correlate with the MIS 8 and MIS 6 stages, respectively. These ages are supported by subarctic (arctoboreal) elements in the mollusk and foraminifer assemblages in the overlying M-III formation, which connects it with the last influx of warm Atlantic water into the High Arctic.

The position of the M-III shoreline at 140 m a.s.l. suggests a very deep glacioisostatic depression from the previous ice sheet. ESR ages between 90 and 80 ka BP from this marine unit are seemingly in conflict with OSL ages from the same marine formation, ranging from 176 to 142 ka BP. However, based on the warmer fauna, Möller et al. (2006) correlate M-III with the Kazantzevo and Eemian marine deposits. The correlation with the Eemian is very likely, whereas that with the Kazantsevo is hardly valid. In a new stratigraphic scheme the correlation should be with the Karginsky Formation, which is reliably associated with the youngest and warmest marine invasion in the Siberian Arctic (Gudina, 1976; Kind & Leonov, 1982; Levchuk, 1984).

Terrestrial sequences. The quality of modern geochronometric signals can also be judged from the interlacial key sections in terrestrial facies, both in the Pechora Basin (Sula) and in West Siberia. The Nadym Ob and Cape Karginsky sections in the standard regional schemes are positioned within the Middle Weichselian, but all recent AMS radiocarbon ages in sediments with interglacial organics have shown infinite values. OSL ages are scattered between 155 and 100 ka BP, with a mean value of 124 ka for 48 measurements in all sampled interglacial sequences (Fig. 8). This mean OSL age is consistent with the astronomical age of the MIS 5e substage,

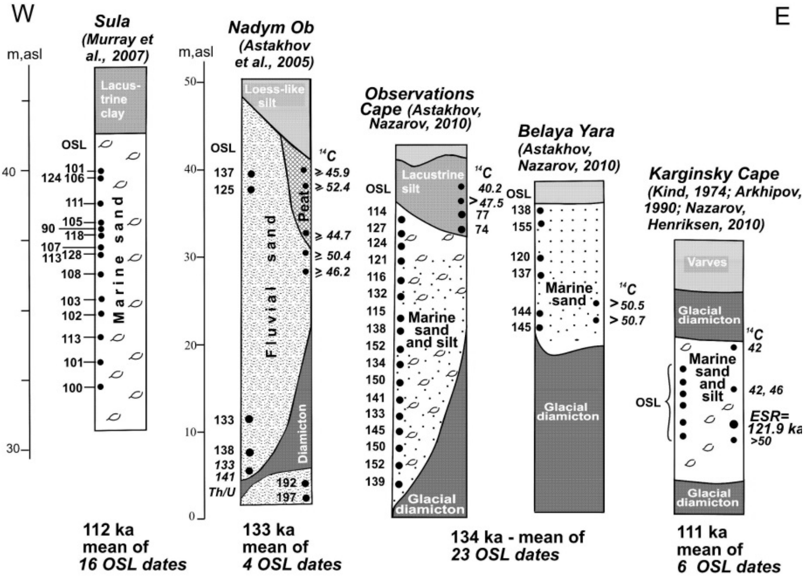

Fig. 8. Dated sediments of the last interglaciation exposed in glaciated arctic Russia. Locations are in Fig. 1: S – Sula, NO – Pyak-Yaha and Pichuguy-Yaha on the Nadym Ob; OC – Observations Cape; BN – Belaya Yara; CK – Cape Karginsky. All dates are in ka.

which is further supported by U/Th dating. Ages of 133 and 141 ka BP have been obtained on shells from glacially disturbed marine sands west of the Urals (My in Fig. 1) (paper 19 in Ch. IV).

Thick Karginsky peats yield similar U/Th dates: 133 and 137 ka at Shur-1 on the Ob river (Sh in Fig. 1) (paper 12 in Ch. III), and 127 and 148 ka at Bedoba in Central Siberia (Bd in Fig. 1) (Laukhin, 2011). The OSL ages between 112 and 80 ka BP from the Karginsky key section of alluvial facies seem somewhat too young, but do not contradict the older part of the OSL population. They are, however, at odds with the conventional radiocarbon ages in the range of 44 to 35 ka (paper 13, Ch. III). Thus, the modern geochronometry in many cases places the Eemian chronostratigraphic level much higher up in the regional sedimentary succession than the Kazantsevo interglacial stage of the Siberian schemes. It seems that all sediments below the Karginsky interglacial deposits should relate to the Middle Pleistocene.

Middle Pleistocene sequences. The chronometric base for correlation of Middle Pleistocene glacial events has also been obtained by re-dating inter-till peat beds. West of the Urals, the Seyda 1 section (Sd in Fig. 1) shows a 1-m-thick peat layer that can be continuously traced for 300 m, separating two thick diamict beds (paper 12, Ch. III). This peat was previously interpreted as belonging to a Late Pleistocene interstadial, based on a radiocarbon date of 43.8 ka BP (Arslanov et al., 1987). However, the pollen spectra suggest a climate much warmer than present, as they indicate dense boreal forest, incompatible with the present-day frozen tundra (Fig. 9). Ten OSL dates on quartz grains from the same interglacial bed give a mean age of 174 ka BP, which agrees with the mean age of 152 ka of five samples from the overlying sediments. U/Th dating of the same peat yielded ages closer to 200 ka (paper 3 in Ch. I). An evidently Middle Pleistocene age of the entire sequence is in full accord with its position outside the Late Pleistocene glaciation (Fig. 10).

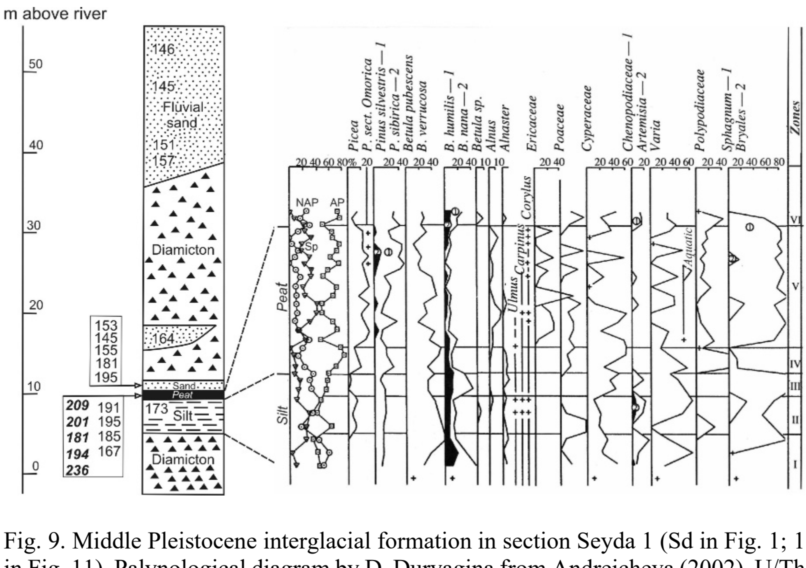

Fig. 9. Middle Pleistocene interglacial formation in section Seyda 1 (Sd in Fig. 1; 1 in Fig. 11). Palynological diagram by D. Duryagina from Andreicheva (2002). U/Th dates *(italicized)* and OSL dates (Roman print) are from Astakhov (2004a). OSL dates later were corrected by A. Murray for cosmogenic radiation influx dependent on burial depth. All dates are given in ka.

Similarly, a 3.5-m-thick peat sandwiched between two glaciotectonized diamictons in the present middle taiga subzone at Rodionovo (R in Fig. 1) has yielded six U/Th ages in the range 240 to 186 ka BP. The pollen diagram from the peat indicates southern boreal forest with admixture of broad-leaved trees (Arslanov et al., 2004).

The above geochronometric data imply that in the subarctic Pechora Basin two glaciations of the final Middle Pleistocene, locally called the Pechora and Vychegda (Table 1), are divided by a pronounced interglaciation correlated with MIS 7.

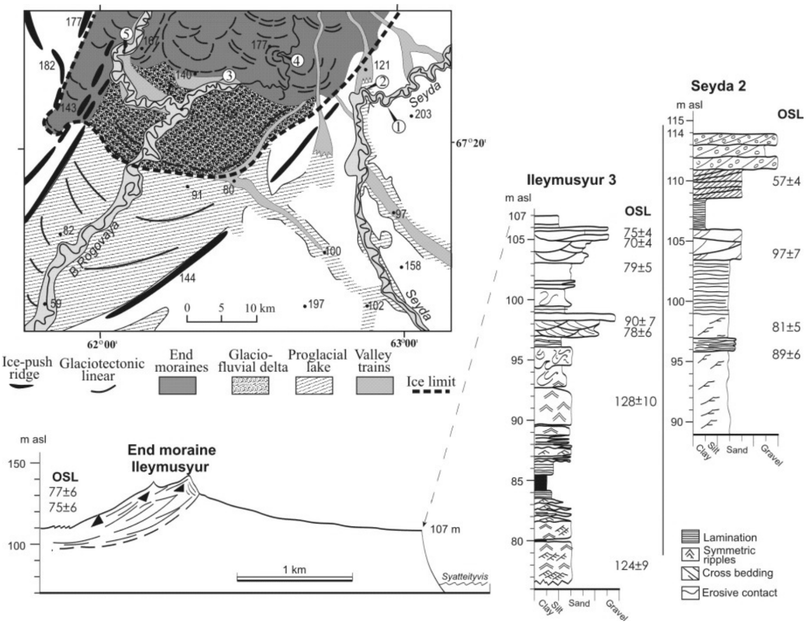

Fig. 11. Mapped and OSL-dated marginal formations of the last Kara ice sheet on rivers Bolshaya Rogovaya and Seyda; Sd in Fig. 1 (Mangerud et al., 2004; Astakhov et al., 2007). Key sections (encircled numbers on the map): 1 – Seyda 1, thick Middle Pleistocene glacial formations (see Fig. 10); 2 – Seyda 2, proglacial lacustrine silts overlain by deltaic gravels; 3 – Ileymusyur 3, proglacial lacustrine silts overlain by outwash; 4 – Site 4 on Syatteityvis creek, sand of III fluvial terrace.

A similar, `intra-Saalian`, interglaciation was suggested long ago by Yakovlev (1956), based on the boreal fauna of the so-called `Northern Transgression`. Recently, several authors working west of the Pechora, in the Mezen catchment area, rejected an older boreal transgression, suggesting only the Eemian incursion of Atlantic water (Larsen et al., 2006). Directly to the east in the Pechora Basin and in Siberia, however, two boreal formations with slightly different faunas have been described many times in natural sections and from deep boreholes (Sachs, 1953; Zubakov, 1972; Gudina, 1976; Levchuk, 1984).

Marine sediments of the lower stratigraphic level, termed the Padimei Formation on the Pechora, contain shells of the extinct mollusk *Cyrtodaria angusta.* Such strata are known from sections along rivers Laya, Nyamda-Yu and Yangarei, and also under thick double diamictons below sea level, as revealed by boreholes such as no. 454 (Fig. 1) (Zarkhidze, 1972). Another important fact is that the subtill *Cyrtodaria* strata occur south of the Weichselian ice limit, for example on river Bol. Volma, 65°30' (BV in Fig. 1).

Similarly, in West Siberia the *Cyrtodaria* strata are identified from the Samburg borehole (Sb in Fig. 1) under a thick diamicton and well outside the southern limit of the last glaciation (Zubakov, 1972). Therefore these strata should belong to the Middle Pleistocene.

Boreholes along the Lower Ob provide evidence that sediments with the Kazantsevo foraminifer assemblage are overlain by thick diamictons and varves. The surficial peat on top of this sequence yields interglacial pollen and U/Th and OSL dates of Eemian age. Therefore the underlying glacigenic sequence is Middle Pleistocene (see Fig. 6 in paper 16, Ch. 3). Thus, the subtill position of the Kazantsevo marine sediments cannot be used as evidence of a Late Pleistocene age of the overlying glacial strata. It appears that the Kazantsevo Formation, with extinct mollusks and with foraminifer assemblages indicating slightly cooler water than the Karginsky microfauna, is more likely to be related to the MIS 7 time span.

A similar situation occurs in the central Siberian Arctic, where diamictons of the `Murukta Moraine`, mantling lowlands of southern Taimyr and between the Putorana and Anabar plateaus, have been attributed to an early Weichselian ice advance. The inner belt of the Putorana Plateau, the Onyoka moraines, was thought to represent the Late Zyryanka (Late Weichselian) maximum (Arkhipov et al., 1976; Isayeva et al., 1986).

However, the Murukta diamictons underlie distinctly interglacial sediments: the Karginsky marine strata with lusitanic forams in southern Taimyr (Kind & Leonov, 1982), and alluvium with forest pollen and southern boreal diatoms on river Sida (Si in Fig. 1) (Bardeyeva, 1986). According to the modern correlation of the Karginsky interglacial with the Eemian, the distal Murukta moraines east of the Putorana Plateau were probably left by the extensive ice sheet of MIS 6 (Fig. 1). Therefore, the Onyoka belt should relate to the Early Weichselian glacial maximum (Svendsen et al., 2004).

Maximum ice advance. In the 1990s it was found that most of the `old` finite conventional radiocarbon dates originated from sediments overlying the uppermost glacial complex of West Siberia, thus indicating a pre-Late Weichselian age of the last glaciation of the Kara Sea catchment area (paper 10 in Ch. III). Later, based on AMS and OSL dates reported by various international teams working in the Russian North, it was concluded that all finite pre-Holocene radiocarbon ages were derived from sediments overlying the youngest glacial formation. This leaves extremist models without chronological support for LGM ice also in the Pechora Lowland, Yamal and Taimyr peninsulas (Mangerud et al., 1999, 2002; Möller et al., 1999; Forman et al., 2002; Svendsen et al., 2004).

Even in the High Arctic, on the Severnaya Zemlya islands, close to a present-day ice cap, the last ice advance proved to be beyond the radiocarbon dating range (Raab et al*.,* 2003). The absence of a Late Weichselian ice sheet on the Severnaya Zemlya was known decades ago from finite radiocarbon dates on mammoth bones scattered close to the present ice caps (Makeyev et al., 1979). The uppermost formation of glacial deposits, overlying the marine strata with subarctic (i.e. warmer than present) fauna, is truncated by marine terraces (unit M-IV) dated by OSL and radiocarbon to 50-40 ka BP. Therefore the age of the last ice sheet of the Severnaya Zemlya is Early to Mid-Weichselian (MIS 5d-MIS 4) (Möller et al*.,* 2006).

All recent dating results from mainland Russia positively indicate that the maximum Late Pleistocene ice advance from the arctic shelf took place ~ 90-70 ka BP. This largest ice sheet of the Late Pleistocene was nevertheless confined within the Arctic Circle and was dwarfed by the Saalian glaciers that advanced 700-800 km farther southwards (Svendsen et al., 2004). The cornerstones for delimiting the maximum Late Pleistocene ice advance are numerous sections of beach gravels and proglacial varves of river valleys dammed by arctic ice. An example of dating marginal formations west of the Urals by OSL is given in Fig. 10. The entire population of 29 OSL ages from beach sediments of proglacial Lake Komi in the Pechora Basin has a weighted mean value of 82 ± 1.3 ka BP (Mangerud et al., 2004).

Similarly, east of the Urals the mean age of four OSL samples from the base of the surficial varved sequence at Sangompan on the Ob river (Sn in Fig. 1) is 81 ka BP (paper 14 in Ch. III). Four OSL ages from postglacial sands of southern Yamal (Ye in Fig. 1) have a mean value of 65 ka (Mangerud et al*.,* 2004), which agrees with the postglacial record of western Yamal with a minimum IRSL age of 45 ± 4 ka BP (Forman et al., 2002).

The main ice advance of the last glaciation was identified by various teams within an OSL span of 100-70 ka BP everywhere along the arctic seaboard from the White Sea shores in the west (Larsen et al., 2006) to the

Taimyr Peninsula in the east (Svendsen et al*.,* 2004; Möller et al*.,* 2008, 2011). This wide range of OSL ages does not necessarily mean an equally long time period for the ice advance and retreat of the Late Pleistocene glacial maximum. The difference in timing of the ice advance over vast areas of arctic Russia may be attributable to an insufficient accuracy of OSL dating, which at this chronological level would readily disperse ages by up to 20% (cf. OSL ages on the Eemian sediments in Fig. 8).

On the other hand, a long life of the first Late Pleistocene ice sheet is suggested by dates on terminal moraines inserted into the Uralian valleys by lowland ice streams and never overridden by younger alpine glaciers. Such a horseshoe-shaped ridge facing upstream was described on the river Kara at 68° N/65°48′ E (Ka in Fig. 1). This moraine is fringed by meltwater sands and gravels, which yield 13 OSL ages with a mean value of 73 ka BP (75 ka if an outlier of 43 ka is excluded) (Nazarov et al*.,* 2009). A pre-LGM age of the major inland glaciation of arctic Russia is independently confirmed by the dating of other montane moraines (see below).

Second Weichselian ice advance. There are different chronometric and geometrical considerations on this part of the glacial history. All authors agree that such an ice sheet must have been much smaller than the previous one. However, there is a discrepancy between the OSL ages of Middle Weichselian ice advances in the Arkhangelsk region (river Mezen catchment area) and in the Pechora Basin. Unlike the rest of northern Russia, in the Arkhangelsk region two ice advances are postulated, ~ 65-70 ka and 45-50 ka BP, with the Mezen Transgression in between (CT in Fig. 1) (Larsen et al., 2006).

In contrast, in the adjacent Pechora Basin only one Middle Weichselian ice advance, at ~ 50-60 ka BP, pre-dated by a marine incursion at ~ 60-70 ka BP, has been identified (Svendsen et al., 2004). Theoretically, a small difference between close OSL ages could be a signature of asynchronous ice advances. However, given the notorious scatter of OSL ages, such fine distinctions of dated events, as suggested in Larsen et al. (2006), seem premature. The most significant result in the Arkhangelsk area is the description of the Mezen marine formation with subarctic fauna, which was deposited between two ice advances. This implies that the Barents Ice Sheet probably disappeared in the Mid-Weichselian.

A way to resolve the chronological discrepancy may be found in better statistics. The latest dating efforts in the Pechora Basin have provided more OSL dates from beneath the so-called Markhida Till. The subtill marine sands have yielded 15 dates with a mean age of 62 ka at Markhida section (M in Fig. 1) and 7 dates with a mean age of 72 ka BP at Kuya (K in Fig. 1) (Astakhov & Svendsen, 2008; Henriksen et al., 2008). Six OSL dates with a mean age of 72 ka BP have recently also been obtained from alluvial sand beneath the varved sequence at Mamontovaya Kurya (MK in Fig. 1) (Astakhov et al., 2010).

On the boundary between West and Central Siberia (P in Fig. 1), a sandur associated with a glacial diamicton of eastern provenance has yielded 11 OSL dates with a mean age of 59 ka BP (paper 15, Ch. III) whereas the underlying periglacial alluvium and varves of the previous ice advance from the north have 15 OSL dates in the range between 95 and 73 ka, with a mean age of 87 ka BP (Astakhov et al., 2010). Taking into account a proposed second ice advance on the Taimyr Peninsula that produced an ice-dammed lake dated by eight OSL measurements to a mean age of 60 ka BP (Alexanderson et al., 2001, 2002; Möller et al., 2011), the age of the second Weichselian ice advance ~ 50-60 ka BP suggested by Svendsen et al. (2004) seems ascertained.

Further support for the Mid-Weichselian ice advance has recently been obtained also in arctic West Siberia, *terra incognita* in the QUEEN reconstruction of the 60-ka glaciation (Svendsen et al., 2004: fig. 15). A sequence of meltwater sands with an interbed of varved clay above the icy subglacial diamicton in the western Gydan Peninsula (Yu in Fig. 1) has yielded six OSL dates with a mean age of 62 ka BP (paper 16, Ch. III). In principle, these relatively young OSL ages do not necessarily belong to an independent ice sheet: theoretically they may result from the retarded melting of stagnant ice fields of the previous Weichselian ice advance.

Late Weichselian cryochron. The timing of the last ice advance in the Russian Arctic is dependent on the assessment of radiocarbon ages of non-glacial sediments. The concept of a Late Weichselian age of the last glaciation in Siberia was originally based on finite ^{14}C dates from subtill sediments of temperate climate (Kind, 1974). Similarly, a long and warm Middle Weichselian interstadial in the Pechora Basin was based on finite radiocarbon dates (Arslanov et al., 1987; Andreicheva, 2002). In both cases these radiocarbon ages are disproved by modern dating results.

For example, the basal layer of the type section of the Middle Weichselian interstadial in the Pechora Basin (the famous Paleolithic site Byzovaya; B in Fig. 1) yields 22 radiocarbon ages at around 28-30 ^{14}C ka BP. This bed contains artefacts and mammoth fauna in soliflucted diamicton but no indications of a temperate climate (Svendsen et al., 2010). The oldest Paleolithic site (37-40 ka ^{14}C BP) on the Arctic Circle in the present northern taiga (MK in Fig. 1) (Svendsen & Pavlov, 2003) is found in the II alluvial terrace with dominant non-arboreal pollen (Halvorsen, 2000). Even in the present-day environment of dense taiga the content of arboreal pollen in Weichselian interstadial spectra does not exceed 50%

(Henriksen et al., 2008). Thus, the `warm` pollen assemblages, presumably characteristic of Weichselian interstadials on the Pechora (Arslanov et al., 1987), in all probability originate, as in Siberia, from interglacial sediments with too young radiocarbon dates.

The modern dating efforts throughout arctic Russia, based mostly on AMS techniques, put `old` finite ^{14}C dates into quite a different geological context, which is clearly seen from many recently studied sections of postglacial sands and silts (Fig. 11). These are subaerial or fluvial sediments found on the surface, much higher up in the sedimentary succession than the last (Karginsky) interglacial. No significant warming or sea-level rise is recorded in the mantling aeolian sediments or periglacial alluvium, located near present sea level and containing cryoarid plant remains, abundant mammoth fauna and ice wedges (Kienast et al., 2001; Sher et al*.,* 2005; Astakhov & Svendsen, 2008; Astakhov & Nazarov, 2010). Weak and short temperate events within the generally arid treeless environment allowed early humans to migrate into the Arctic, as indicated by the Paleolithic sites well dated between 40 and 20 ^{14}C ka BP in European Russia (PS, Mm and B in Fig. 1) (Svendsen & Pavlov, 2003; Svendsen et al., 2010) and at 28.5 to 27 ka BP in East Siberia (Ya, Fig. 1) (Pitulko & Pavlova 2010).

The old permafrost, recorded in sections of periglacial sediments, has never been affected by serious pre-Holocene warmings, as is evident from the numerous well-preserved carcasses of frozen mammoths in arctic Siberia (crosses in Fig. 1) (paper 16, Ch. III**).** The absence of significant climate and environmental changes in the former mammoth steppe of the Siberian Arctic is strongly supported in a recent study that integrates sedimentary DNA, pollen and macrofossil data from a number of sections with silty to organic-rich sediments, spread around Lake Taimyr on the Taimyr Peninsula, at 75° N, for example CS in Fig. 1. The results `imply that the vegetation cover in the interior of the Taimyr Peninsula has remained fairly stable during the Late Pleistocene from *c.* 46 to 12.5 cal. kyr BP` (Jørgensen et al., 2012: p. 2000).

Only minor climatic changes took place in this generally arid permafrost environment, as is shown also by the study of the *Coleoptera* fauna, the best indicator of summer air and soil temperatures, by Sher et al. (2005). These authors analysed the Yedoma formation of periglacial silts with long ice wedges and mammoth bones in the Lena delta (CB in Fig. 1). They found steppe insect assemblages persisting throughout the succession from 47 to12 ^{14}C ka BP, which indicates summer temperatures higher than present ones even within the coldest interval of 23–15 ^{14}C ka BP. The warmest and driest stage started at ~ 15 ^{14}C ka BP. Thus, well-studied sections of periglacial sediments in Siberia bear no indications of nearby ice sheets younger than the Mid-Weichselian.

The only site on the arctic seaboard related to the Late Weichselian ice advance east of Fennoscandia is the tip of the Kanin Peninsula (KP in Fig. 1), where erratics of Fennoscandian provenance lie on the rocky surface dated by the cosmogenic nuclide exposure (^{10}Be) method to 18.6 ka BP. East of this point all OSL ages from the terrestrial outwash are older than 50 ka BP (Demidov et al., 2006). Farther eastwards, the Late Weichselian ice limit is located on the Barents Sea floor according to seismics and geotechnical boreholes (Svendsen et al., 2004). Boreholes nos 210–234 (Fig. 1) penetrated fine-grained marine sediments up to 50 m thick on top of the uppermost till. They yield nine AMS ^{14}C dates on foraminifers, which give ages in the range of 39.4 to 23.6 ka BP. Northwards, across the inferred Late Weichselian ice limit, the oldest radiocarbon ages from postglacial marine sediments do not exceed 15 ka BP (Polyak et al., 2000). In the Kara Sea this ice limit is inferred from bathymetric and seismic data (Svendsen et al.. 2004; Polyak et al., 2008).

There is one point on the Siberian mainland where an indication of a possible Late Weichselian ice sheet has been reported, namely White Lake in the northern Taimyr (WL in Fig. 1), where two AMS ages of c. 20 ka BP were obtained on shells found in an ablation till on top of buried glacier ice at 120 m a.s.l. (Alexanderson et al., 2001). Based on this observation a small ice lobe, barely touching the mainland, was suggested in the QUEEN reconstruction of Late Weichselian ice sheet (Svendsen et al*.,* 2004).

This part of the reconstruction seems suspicious because the nearby Severnaya Zemlya with present ice caps was populated by mammoths during the LGM (Makeyev et al., 1979). A lack of Late Weichselian ice is also indicated by several sections at Changeable Lake (Raab et al., 2003) and along the Ozernaya river (Möller et al., 2006) (Cl and Oz in Fig. 1). As the mollusks dated to 20 ka BP from the North Taimyr marginal belt must have originally lived a hundred metres below the present shoreline, it is hard to imagine the configuration of a glacier that could bring the shells up to their present position. As previously discussed, radiocarbon ages from Pleistocene mollusk shells in other Siberian areas too often yielded abnormally young ages (paper 10 in Ch. III). Thus, without further evidence from other parts of the North Taimyr ice marginal zone, its age remains uncertain.

A small ice sheet on top of the Putorana Plateau is pure speculation so far (Astakhov, 2011). Sediments with finite radiocarbon ages under the Putorana piedmont moraines contain, amongst other organics, even extinct *Cyrtodaria angusta* shells. This means that they are not evidence of a Late Weichselian age of the following glaciation here (Svendsen et al*.*, 2004).

A chronological pattern of the sedimentary events of the last glacial cycle of the major arctic plains is outlined in Fig. 12, based on OSL and ^{14}C dates available in the literature.

The alpine chronology presents a special issue which may shed light on the size of Late Weichselian inland glaciers beyond the Barents and Kara ice sheets. Especially relevant in this respect are spectacular exposure ages, calculated from ^{10}Be content in free-lying boulders in the Polar Urals. Moraines, located within 1 km of a small present-day glacier at 400-600 m a.s.l., have yielded six beryllium ages ranging from 28 to 14 ka BP. In contrast, boulders scattered down-glacier along the valley produced ages mostly in the range 60-50 ka BP (Mangerud et al*.*, 2008). This spatial pattern implies that LGM ice at 67°N (U in Fig. 1) was not much larger than the present-day tiny glaciers, whereas the main advance of valley glaciers occurred before 60-50 ka BP. Similar results are reported from the 2.5 km high Verkhoyansk Range of eastern Siberia, where the youngest terminal moraines have yielded OSL ages over 50 ka (V-1 and V-2 in Fig. 1) (Stauch & Gualtieri, 2008).

The above chronometric data, however surprising they may seem, are in good accord with the absence of LGM glaciers in the vicinity of present ice caps in the high Arctic, Severnaya Zemlya, at 79°N (Makeyev et al., 1979; Raab et al., 2003; Möller et al., 2006). These results testify to a very low precipitation level during the LGM in Siberia. This is in harmony with the dry and frosty periglacial environments of the mammoth steppe, inferred from surficial sediments scattered from the Pechora to Taimyr and the Lena delta. Anyway, the modern data imply that the minimalist model for MIS 2 glaciation of northern Russia (Velichko et al., 1997) is not minimal enough.

The evolution of ice volumes in northern Russia during the last glacial cycle is illustrated by a time-distance diagram derived from QUEEN projects (see Fig. 2 in paper 21). Although the timing of the main glacial stages is similar in all northern Eurasia, the trend of ice volume changes over Siberia is reversed as compared with the Atlantic margin (Svendsen et al*.*, 2004): while Weichselian ice sheets in Fennoscandia became larger during each successive glacial stage, in central northern Russia the concurrent ice sheets gradually diminished, probably owing to the progressing lack of humidity in the entire Eurasian Arctic. Unlike the Laurentide glaciations, the ice sheets of northern Russia have become successively smaller during the last 200 ka.

However, the available geochronometric data suggest that time brackets for the Siberian glacial and interglacial events roughly correspond to the Fennoscandian chronostratigraphic units. This result can be summarized in a preliminary correlation chart (Fig. 13).

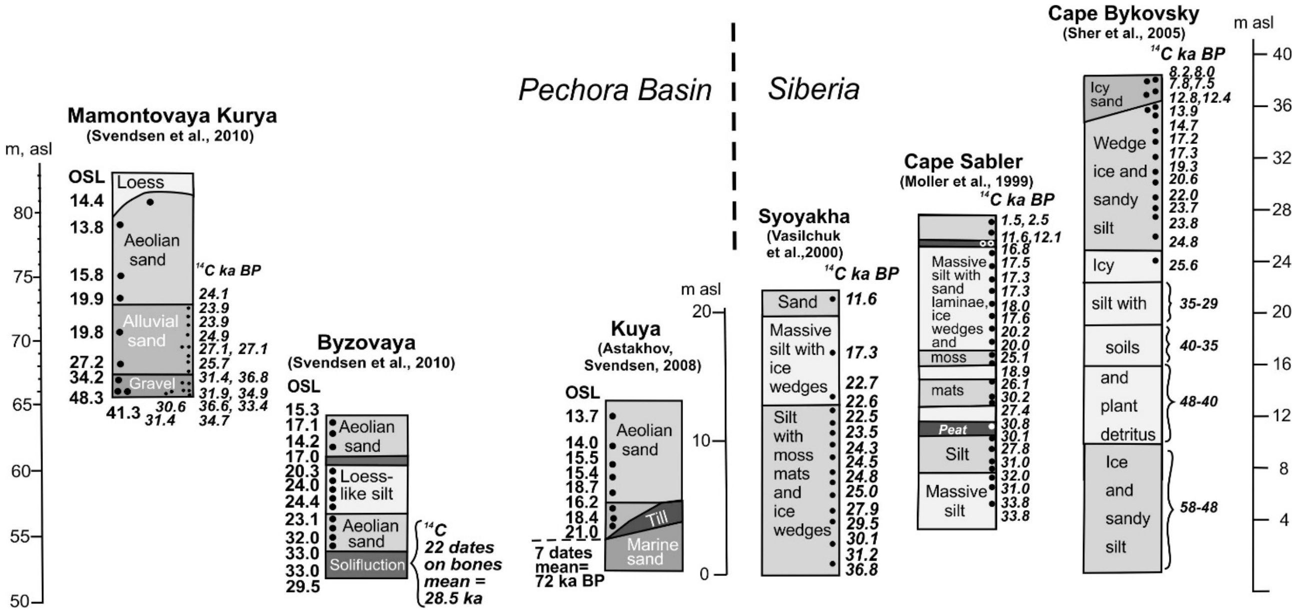

Fig. 11. Major postglacial sequences chronometrically correlated from Pechora Basin to Lena delta.
Key sections in Fig. 1: MK – Mamontovaya Kurya, 66°34′ N/ 62°25′ E; B – Byzovaya, 65°01′ N/57°23′ E; K – Kuya, 67°35′ N/53°30′ E; S – Syoyaha, 70°10′ N/ 72°34′ E; CS – Cape Sabler, 74°33′ N, 100°32′ E; CB – Cape Bykovsky, 71°47′ N/129°25′ E. OSL and ESR ages are shown in Roman print, radiocarbon and U/Th dates are italicised. All dates are given in ka.

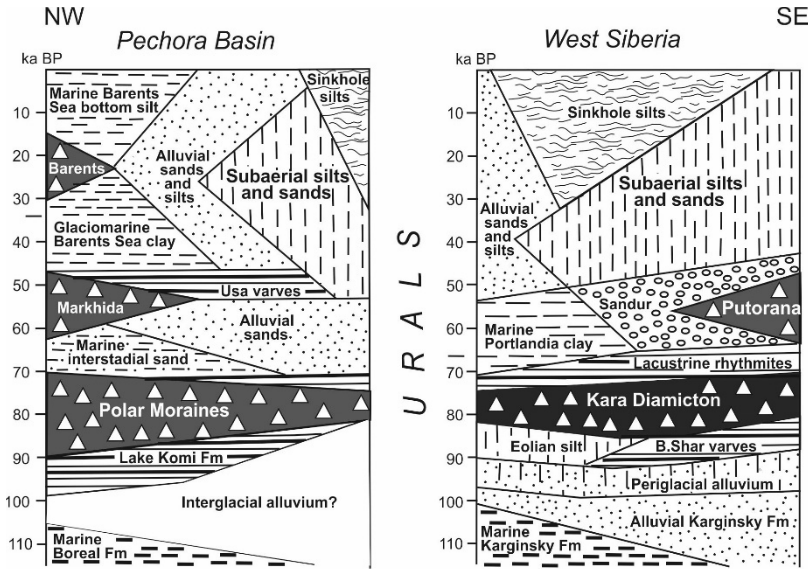

Fig. 12. Chronological brackets for Late Pleistocene sedimentary events in the Russian Arctic west and east of the Urals. Glacial events are shaded. Dates are from Vasilchuk et al. (2000), Forman et al. (2002), Mangerud et al. (1999, 2002, 2004), Astakhov & Mangerud (2005), Astakhov et al., 2005, 2007, 2010), Astakhov & Svendsen (2008), Henriksen et al. (2008), Astakhov & Nazarov (2010), Svendsen et al. (2010), Nazarov & Henriksen (2010), Svendsen et al. (2010).

Problematic issues

Numerous problems are far from being solved, which is not surprising given the size of northern Russia and the scarcity of Quaternary scientists.

First, the extent of older ice sheets is still to be ascertained, especially in the east of central Siberia, where even the limit of glacial erratics is drawn tentatively (Arkhipov et al., 1976; Astakhov, 2011).

Second, the age of the maximum glaciation needs to be specified in each separate region, because the ultimate ice limit is probably time-transgressive from west to east. The research history of Central Russia indicates that the latitudinal age difference may be great (papers 3, 4 in Ch. I). The problem would demand painstaking stratigraphic work on Middle Pleistocene interglacial sediments in attempts to correlate, for example, the Tobol alluvium of the Ob-Irtysh sections with the Yenissei and Lena catchment areas.

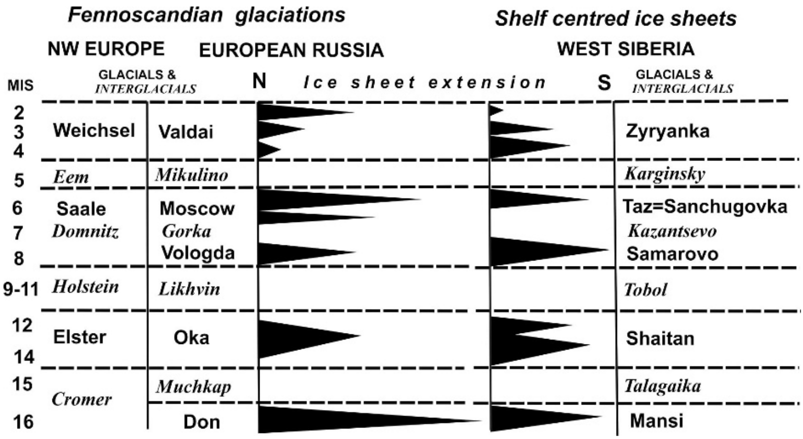

Fig. 13. Suggested correlation of main Siberian and European glacial events.

Third, the configuration of certain ice sheets of the Late Pleistocene is far from reliable. Especially enigmatic is the extent of the second ice advance ~ 60 ka BP in West Siberia and of the Late Weichselian glaciation on the Putorana Plateau. The same can be said about the ice limits in the eastern Taimyr Peninsula. Crucial for calculating former equilibrium-line altitudes should be the mapping of LGM alpine glaciation: it might grow in size southwards along the Uralian Mountains (Mangerud et al., 2008).

Fourth, the age brackets for glaciations, especially the pre-Weichselian ones, lack precision because of the existing uncertainties and sometimes discrepancies between OSL, ESR and U/Th dating techniques.

Fifth, the question of changing ice-dispersal paths is open so far. When the evidence of the southbound ice motion first emerged in the Siberian literature, Flint's idea of shifts of ice accumulation centres through a glacial cycle from mountains to lowlands was employed for the explanation of the crystalline clasts in the Siberian diamictons and the geography of the glaciated area (Kaplyanskaya & Tarnogradsky, 1975; Arkhipov et al., 1976; Astakhov, 1977). This model is also used for the Kara Sea ice sheet in modern works (Möller et al., 2006).

However, the same authors, based on directions of glaciotectonic deformations, suggested an additional upland ice dome on the Pai-Hoi Range following the Early Weichselian shelf-based ice sheet (Lokrantz et al., 2003). A reactivation of upland ice sources ~ 60 ka BP after the major ice advance from the shelf ~ 80-90 ka BP was also documented in the Yenissei valley (papers 15, 17 in Ch. III). The paucity of evidence leaves a wide berth for speculations on the inception of ice sheets and their evolution through glacial cycles.

Conclusions

Quaternary studies of the last two decades in northern Russia mark a departure from the conventional model of glacial history elaborated on the alpine example towards a better documented model of shelf-based ice sheets.

- The former cold and thick glaciers combined with stable icy permafrost constitute the peculiar Siberian type of Pleistocene glaciation. Glaciers of the Russian North probably grew within the same time spans as in Europe but their dynamics and degradation patterns were different. After initial fast disintegration, ice decay in Siberia was arrested (or retarded) owing to the large stock of cold accumulated in the lithosphere, protecting debris-covered stagnant ice from further melting. `Primordially frozen` subglacial deposits (Kaplanskaya & Tarnogradsky, 1986) are widespread in the Arctic, while wet-based features, such as those related to lodgment processes, are confined to specific areas.
- The most interesting chronological results of the last decade concern the age of the sedimentary formations of temperate climate underlying the upper glacial complex. In various regions they proved to be Eemian *sensu lato.* Together with `old` finite radiocarbon dates from the overlying silty mantle this proves the Early and Mid-Weichselian age of the last ice sheets. Late Weichselian glaciers were evidently minuscule in the Russian Arctic. Chronological correspondence between the last glacial events of the Russian Arctic and Scandinavian ones now seems to be firmly established.
- Modern geochronometric results have also changed the chronological position of two last interglacials. OSL, ESR and U/Th dating proved an Eemian age of the Karginsky temperate event, previously attributed to MIS 3, and pressed the penultimate Kazantsevo interglacial down to MIS 7.
- New dating efforts also established a MIS 7 age for the northern penultimate interglaciation based on the sedimentary formations that were earlier considered to be either enigmatic intra-Saalian or related to MIS 5. Now glacial correlations across northern Eurasia are much better constrained, although not without uncertainties in intraregional correlations of Middle Pleistocene events. Combining the old data and the modern results of the last two decades, an approximate correlation of the principal Siberian landmarks of glacial history with the European ones may be

proposed (Fig. 13).

- New chronometric techniques have amplified the continental trend in paleoglaciological considerations, suggesting older and smaller glaciers for the Late Pleistocene. The reconstructed Late Weichselian ice sheets have drastically shrunk on the eastern lee-side of the Scandinavian and Barents Sea ice sheets (Svendsen et al., 2004). In the modern view, really large ice sheets developed only during the more humid Middle Pleistocene. Since then, the volumes and areal extents of glacial ice decreased stepwise, reflecting the progressive aridification of northern Eurasia.

Acknowledgements

This overview was supported by research grant no. 11-05-00624-a from the Russian Foundation for Basic Research and by a grant from St Petersburg State University. This is a contribution to the European coordination program APEX (Arctic Palaeoclimate and its Extremes). I am indebted to my Norwegian colleagues Professor Jan Mangerud and Professor John Inge Svendsen, who over the last two decades have contributed greatly to the Upper Pleistocene stratigraphy of the Russian Arctic and to my understanding it. My thanks are also due to Per Möller and an anonymous referee who meticulously read the manuscript and helped to clarify the message. Dr. Monty Charles of Birmingham University has kindly checked the language.

23. The Postglacial Pleistocene of the Northern Russian Mainland

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Quaternary Science Reviews, 2014, Vol. 92, 388–408

Introduction

The postglacial history of northern Eurasia is comprehensively studied only in the area of Fennoscandian glaciations where it is rather short, 15 ka at most. The remaining larger and older part of the Eurasian North, ice-free since at least 50-60 ka BP (Svendsen et al., 2004), is far less known. The main stages of the Holocene history of northern Russia have been extensively discussed (e.g. Velichko et al., 1997; Andreev & Tarasov, 2007), whereas the Pleistocene postglacial events were only briefly reviewed in the summary of the Late Weichselian periglacial environments with reference to Middle Weichselian climates by various QUEEN (Quaternary Environment of the Eurasian North) research teams (Hubberten et al., 2004).

In the wake of the QUEEN projects the chronological control of the postglacial sedimentary record has improved both quantitatively, owing to numerous chronometric measurements, and qualitatively due to new proxies obtained by several international research teams. New results have been largely published in the Russian literature (e.g. Astakhov & Mangerud, 2007; Nazarov, 2007; Astakhov & Svendsen, 2011; Astakhov & Nazarov, 2010; Derevyagin et al., 2010).

Apart of the QUEEN overview (Hubberten et al., 2004) most recent studies were restricted to limited areas or focused on especially important or interesting cases studied by investigators with different agendas. Therefore, to outsiders many new results would seem contradictory or derived from unrelated topics. It is my contention that despite the different methods applied and sometimes different conclusions suggested for various parts of the Russian Arctic the general succession of geological events of the postglacial Pleistocene history is basically the same for different terrains of the glaciated territory.

The following overview is aimed at collating published evidence of the nature, succession and timing of geological events postdating the last regional glaciation of northern Russia mostly between 48° and 110° E (Fig. 1). To fulfill the task some geological results from the periglacial zone are included, drawing on my own research findings. For the more problematic area of East Siberia I refer to literary sources but this information is necessary because a couple of sections in this region contain uniquely continuous records of Late Pleistocene history. For chronological control I rely on previously published dates from the international literature with lists of ages obtained by ^{14}C, OSL and other chronometric techniques, supplemented by dates available only from Russian sources.

The geographic names used in this text for the major parts of the continental landmass are northeastern European Russia including the Pechora Lowland, Cisuralia and the Barents Sea coast, West Siberia for the area between the Urals and Yenissei river, Central Siberia for the area between the Yenissei and Lena rivers, and East Siberia for the lands farther eastwards.

Existing knowledge

There are contradictory conclusions regarding the second half of the northern Late Pleistocene. The most evident are different ideas regarding the chronology of postglacial events that are connected with different genetic labels, i.e. glacial and marine origins ascribed to certain surficial features support the short version of the postglacial history. In contrast, the focus on subaerial and permafrost processes has led to the adoption of a much longer succession of postglacial events.

The Late Pleistocene of northern Russia was originally viewed as fundamentally different from the Fennoscandian history. The classical viewpoint acknowledged the thick sedimentary cover of postglacial formations and the finds of frozen mammoth carcasses as evidence of a much longer time interval that had elapsed since the disintegration of the last regional glaciation. The latter was considered as an early Weichselian event preceding the maximum ice advance in north-western Russia correlative with the Late Weichselian (Yakovlev, 1956; Zubakov, 1972).

However, the advent of conventional radiocarbon dating led to a new paradigm of a younger glaciation which, it was thought, covered all northern plains and coalesced with the Late Weichselian ice sheet of Fennoscandia. The main argument for a short postglacial history was provided by several finite radiocarbon dates that were recovered from beneath the uppermost diamicton in various sections (e.g. Kind, 1974; Arslanov et al., 1987; Arkhipov, 1998).

The concept of a short postglacial history was criticized as contradicting the statistical distribution of the entire population of radiocarbon ages and the remains of megafauna with old radiocarbon dates found near the surface (papers 6 in Ch. I and 10 in Ch. III). The restoration of the classical concept of a long postglacial history, however, was not eagerly accepted in the Russian literature, in which the idea of a young continental glaciation dominated since the 1970s. Thus, the MIS 2 age of the last regional glaciation called Polar on the Pechora and Late Zyryanka (Sartan) in Siberia has now been accepted by various official stratigraphic schemes (Guslitser et al., 1986; Isayeva et al., 1986; Volkova & Babushkin, 2000).

The recent efforts of the international community in stratigraphic studies of the Late Pleistocene of the Russian Arctic (Fig. 1) within the QUEEN program, summarized by Svendsen et al. (2004), demonstrated conclusively the relative antiquity of the last regional glaciation. The new geochronometric techniques proved that finite conventional radiocarbon dates from beneath the upper glacial complex were mostly too young, thus firmly establishing the classical concept of a long postglacial history. However, some dissenting voices insisting on a MIS 2 age for the last glaciation based on the rejection of all chronometric ages beyond the conventional ^{14}C still exist in the Russian literature (e.g. Lavrov & Potapenko, 2005).

The published paleoenvironmental record consists of:

– Descriptions of geological structures and sedimentological properties of postglacial rocks.

– Geomorphic evidence including remote sensing data.

– Geocryological descriptions with oxygen isotope analyses of ground ice.

– Paleontological and archaeological descriptions.

The proxies available for the area of the last regional glaciation can be compared with and are supported by similar information obtained from several best studied sections in adjacent periglacial terrains (marked by letters Ma, BP and Ya in Fig. 1).

The discussion in the Russian literature has been traditionally based on paleonthological and geocryological material aided by conventional radiocarbon dates and occasional sedimentological descriptions (e.g. Kind, 1974; Kind & Leonov, 1982; Vasilchuk, 1992). The situation has changed since 1993 when new lines of evidence have been added to the discussion due to field and laboratory studies within the framework of the Russian-

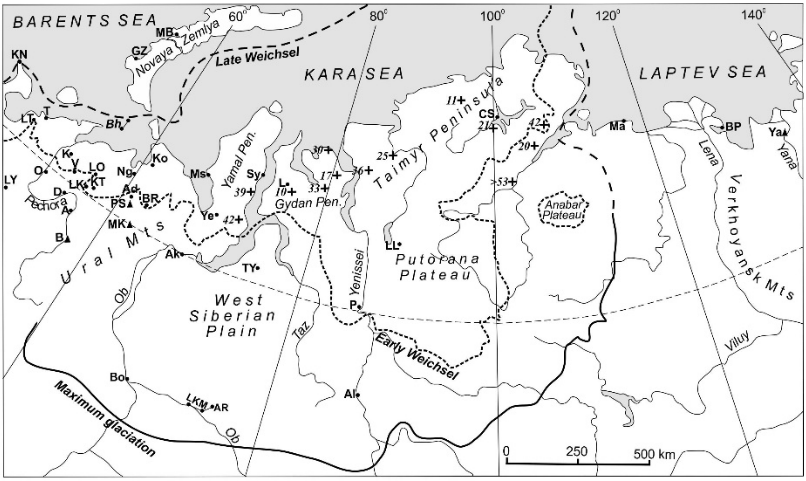

Fig. 1. Dated sections of the postglacial Pleistocene of northern Russia.

European part. Bh – boreholes 210–234 on sea floor (Polyak et al., 2000). *River sections*: A – Akis, Pechora, BR –Bol. Rogovaya, KT – Kolva, Ng – Ngutayaha (Astakhov et al., 2007), Ad –Adzva, K – Kuya, Ko –Korotaikha, V - Vonda (Astakhov & Svendsen, 2011), D – Denisovka, O – Okunyovo (Lavrov & Potapenko, 2005). *Coastal sections:* GZ – Gusinaya Zemlya, MB – Mityushikha Bay (Serebryanny & Malyasova, 1998), KN – Kanin Nos (Demidov et al., 2006), T – Timan Coast (Mangerud et al., 1999). Lakes: LK – Kormovoye, LO – Oshkoty (Henriksen et al., 2003), LT – Timan (Paus et al., 2003), LY – Yamozero (Henriksen et al., 2008). *Paleolithic sites:* B – Byzovaya (Heggen et al., 2012), PS – Pymva-Shor (Mangerud et al., 1999), MK – Mamontovaya Kurya (Svendsen et al., 2010). **Siberia.** Ak – Aksarka, River Ob, TY – river Tab-Yaha (Astakhov & Nazarov, 2010); Al – Alinskoye, Yenissei (Sukhorukova et al., 1991); BP – Bykovsky Peninsula (Sher et al., 2005); CS – Cape Sabler (Kind & Leonov, 1982; Möller et al.,1999); I – Igarka (Kind, 1974); L – Lysukanse (Bolikhovsky, 1987; Astakhov & Nazarov, 2010); LKM –AR – profile Lokosovo-Kiryas-Mega-Agan Ridge (in Fig. 7, paper 6, Ch. II); LL – Lake Lama (Hahne & Melles, 1999); Ma – Mamontovy Klyk (Schirrmeister et al., 2008); P – Poloi (paper 15, Ch. III); Ms – Marresale (Forman et al., 2002); Sy – Syoyaha (Vasilchuk et al., 2000); Ya – river Yana (Pitulko et al., 2004; Pitulko & Pavlova, 2010); Ye – river Yerkata (paper 14, Ch. III). *Mammoth carcasses* with their radiocarbon ages, ka BPe indicated by crosses: river Khatanga – >53.2, Lyuba – 41.9, Masha – 39.1, river Mokhovaya – 35.8, Schmidt – 33.5, Cape Leskin – 30.1, rivers: Pyasina – 25.1, Mongoche – 17, Mamonta – 11.5, Yuribei – 10 (Astakhov & Nazarov, 2010 and references therein); lake Arilakh – 42.4, Fish Hook – 20.6, Jarkov – 20.2 (MacPhee et al., 2002).

Black triangles are Paleolithic sites. Thick lines are major ice limits.

Norwegian, Russian-German, Swedish-American and other international projects. The projects under the aegis of European programs QUEEN and APEX have brought better sedimentological expertise, diverse paleontological and geochemical data and, which is especially valuable, numerous geochronometric measurements carried out by modern, more reliable techniques such as AMS radiocarbon and optically stimulated luminescence (OSL) (Mangerud et al., 1999; Forman et al., 2002; Schirrmeister et al., 2002; Svendsen et al., 2004).

Remote sensing data helped to understand the recent evolution of permafrost and glacial landscapes (Astakhov, 1995, 1998b; Astakhov et al., 1999; Grosse et al., 2007). Substantial information on the final Pleistocene has been obtained by coring the bottom sediments of extant lakes (Hahne & Melles, 1999; Henriksen et al., 2003; Andreev et al., 2003, 2004). An additional influx of paleoenvironmental data is connected with studies of oxygen isotope composition of ground ice, megafauna and fossil insects assemblages in arctic Central Siberia (Hubberten et al., 2004).

The post-QUEEN studies added a multitude of AMS radiocarbon and OSL dates supported by paleontological finds that further clarify the succession of postglacial events (Andreev & Tarasov, 2007; Andreev et al., 2011; Astakhov & Mangerud, 2007; Nazarov, 2007; Henriksen et al., 2008; Astakhov & Nazarov, 2010; Astakhov & Svendsen, 2011). For understanding the post-glacial Pleistocene we owe much to archaeological research at northernmost Paleolithic sites which, beside the artifacts, brought detailed geochronometric and sedimentological evidence from thoroughly excavated sections (Pitulko et al., 2004; Pitulko & Pavlova, 2010; Svendsen et al., 2010; Heggen et al., 2012).

Very important are large numbers of dates from the Taimyr Peninsula in Central Siberia which provide numerically robust estimates of the age of the principal sedimentary formations. More recently, analysis of fossil DNA has also been used (Jørgensen et al., 2012). New proxies illuminating past climates have come from the adjacent region of East Siberia, or western Beringia, obtained by paleopedological research of periglacial formations (Gubin et al., 2008).

Below I will overview and discuss the geological, geomorphic and paleontological data necessary for geochronological and paleonvironmental inferences. The locations of the best dated sedimentary sequences are indicated in the map of Fig. 1. Most of them overlie the uppermost glacigenic complex. To establish main landmarks of the postglacial history in the sedimentary record it is necessary to identify and date features produced by major surficial agencies.

Half of the important sections are concentrated west of 65° E due to a more intensive analysis of northeastern European Russia relative to the rest of the Russian mainland. Several sections beyond the Late Pleistocene ice limit are considered because the postglacial Pleistocene environments were uniform over great distances independently of ice limits (sites B, LKM–AR, LY, Ma, MK, BP, Ya etc. in Fig. 1). Frozen mammoth carcasses, if properly dated, are valuable chronological landmarks due to their unambiguous *in situ* position above the uppermost glacigenic formation.

Waterlain sediments

The postglacial sedimentary formations are typically more fine-grained and richer in organics than older Weichselian deposits. The spatial extent of waterlain sedimentary bodies is more restrictive than the contemporaneous subaerial formations but their stratigraphic importance cannot be overestimated because they are normally richer in fossils and provide a multitude of geochronometric data.

Marine sediments

Marine muds and sands with arctic fauna and finite radiocarbon dates occur mostly in submerged positions (Polyak et al., 2000) and sometimes on low terraces belonging to arctic islands (Möller et al., 2006). On the coastal lowlands of Yamal and Gydan peninsulas they sometimes contain shells of *Portlandia arctica* and are known as the `Portlandia Clay` overlying the uppermost diamicton or even fossil glacial ice at altitudes normally <30-40 m above sea level (Troitsky & Kulakov, 1976; Astakhov, 1992). The postglacial marine sediments according to AMS radiocarbon dates are related to MIS 3 time and therefore constrain the age of the underlying till. Their position in boreholes (Bh in Fig. 1) is crucial for tracing the Late Weichselian ice limit on the sea floor.

The easternmost Late Weichselian ice margin was located on the tip of the Kanin Nos (KN in Fig. 1). At this site erratics of Fennoscandian origin lie on a rocky surface that is dated by cosmogenic nuclide exposure (^{10}Be) method to 18.6 calendar ka BP. East of this point all OSL ages from the terrestrial outwash on the surface are older than 50 ka BP (Demidov et al., 2006). The most important landmark for delimiting the Late Weichselian ice sheet on the sea floor is the profile of geotechnical boreholes no. 210–234 (Bh in Fig. 1) which penetrated marine muds up to 50 m thick that are underlain by till and overlain by Holocene silts (left column in Fig. 2).

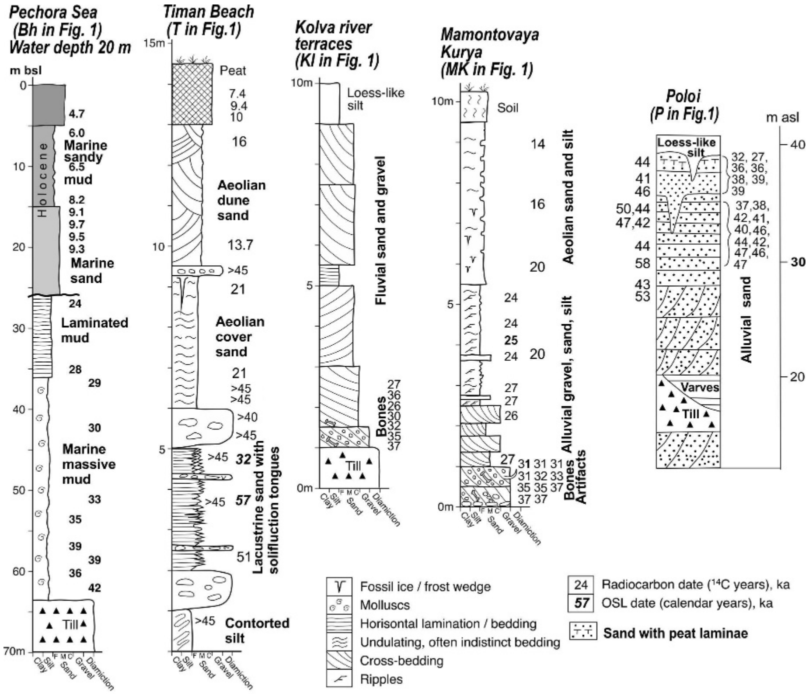

Fig. 2. Key dated sections of the postglacial sediments in the Pechora Basin (Bh, T, MK and B in Fig. 1) compared with alluvium of the Yenissei (P in Fig. 1). Adapted from Mangerud et al. (2002) and Astakhov & Mangerud (2007). Note that sedimentary units within the chronometric range of 50 ka are nowhere covered by glacial deposits.

Arctic foraminifera from the muds have yielded ten AMS ^{14}C ages in the range of 39.4 to 23.6 ka BP (Polyak et al*.,* 2000). The area of the glaciation traced by shallow seismics is located on the Barents Sea floor north of this profile where oldest postglacial radiocarbon ages are younger than 15 ka BP (Svendsen et al., 2004). Still farther eastwards the Late Weichselian ice limit is tentatively extrapolated over the Kara Sea floor using bathymetric and seismic data (Polyak et al., 2008).

The absence of recent isostatic uplift on the low coastlands of arctic Russia is emphasized by their present-day subsidence and by the position of Early Holocene peats on the shallow shelf that was submerged by the Holocene Flandrian transgression (Veinbergs, 1982). The idea of the staircase of Pleistocene marine terraces on coastal plains used in old

Russian maps is now indefensible because sedimentary bodies continuously traced along the shores of the Yugorski and western Yamal peninsulas are distributed independently of altitudes of the retreating coastal bluffs (Astakhov et al., 1999; Manley et al., 2001; Forman et al., 2002). The lack of raised marine levels in periglacial Siberia was recently confirmed by studies of the Laptev Sea coasts where terrestrial formations are traceable to at least to -10 m (Schirrmeister et al., 2011). The absence of high Holocene marine terraces along the Russian Arctic shores east of the White Sea is one of the strong arguments against MIS 2 glaciation of the mainland.

Fluvial sediments

The alluvial formations of the great rivers compose regional terraces which are numbered according to their relative position above the Holocene floodplain as First, Second and Third Terrace, with the present-day floodplain having the zero number. The alluvial sediments are lithologically more diverse and coarser than other postglacial formations. The Third Terrace on the Pechora, Ob and Yenissei rivers is covered by sands and gravels of stratified drift of the last ice age ~ 60-70 ka old and its base in places consists of interglacial strata (Zubakov, 1972). Thus, strictly speaking, only the Second and First Terraces belong to the postglacial Pleistocene. That their elevations are very similar in lowland river valleys complicates their identification.

The alluvium of the Second Terrace is only ~ 10 m thick on the Pechora and reaches 20 m on the greatest river Yenissei (Fig. 2), i.e. it is generally thinner than the alluvium of Holocene floodplains. The Second Terrace largely contains channel facies of basal gravel that changes upwards into cross-bedded and laminated sands. The reduced sediment thickness commonly makes it a strath terrace, its base under gravelly alluvium being either bedrock or glacial drift. Peculiar features of this alluvium are its distinctly braided or anastomosed character, reduced floodplain facies and lack of oxbow-lake lenses (Mangerud et al., 1999). This facies composition, typical for periglacial alluvium, is consistent with diverse paleontological evidence of treeless landscapes. For example, the channel alluvium on the Yenissei, north-flowing in the present-day zone of dense boreal forest, contains only shrub twigs but no logs, unlike the Holocene alluvium (paper 15 in Ch. III).

The essentially treeless environment of the Second Terrace alluvium of the Pechora Basin is indicated by pollen spectra of the well-studied Paleolithic site Mamontovaya Kurya situated on the Arctic Circle in present northern subzone of the taiga (or boreal forest) belt (MK in Fig. 1) (Halvorsen, 2000). Numerous radiocarbon ages from this alluvium range from 37 to 24 ^{14}C ka BP and luminescence ages are between 48 and 24 calendar ka (Svendsen et al., 2010). In places this sandy alluvium yield younger ages, e.g. on river Adzva (Ad in Fig. 1) where four OSL dates range from 32 to 18 ka BP (Astakhov & Svendsen, 2011).

In West Siberia the alluvial terraces are in age similar to the terraces of the Pechora Basin. For example, the sand with mammoth bones in the Second Terrace of the river Tab-Yaha (TY in Fig. 1) have yielded three OSL dates of 37, 31 and 24 calendar ka BP upwards in the succession, whereas moss mats from the same formation on river Poilovo-Yaha ~ 100 km to the north have been radiocarbon dated to 28 and 27.5 ka BP (Nazarov, 2007; Astakhov & Nazarov, 2010). The alluvium of the Yenissei can be up to 47 ^{14}C ka old (Fig. 2, right column).

The First Terrace alluvium is usually thicker, more like the Holocene floodplain formation which is ~ 20 m thick on the Pechora and ~ 30 m thick on the Yenissei. There is a full set of normal alluvial facies: channel sands, floodplain laminated silts and sands and clayey lenses of oxbow-lake facies. The noticeable difference of this fill terrace as compared to the Holocene floodplain is the cover sand or dune sand invariably present on the wavy surface of the First Terrace. Deposits of aeolian sand in places totally obscure the alluvium (Astakhov & Svendsen, 2011). The range of ^{14}C ages measured at Denisovka and Okunyovo terraces on the Pechora (D and O in Fig. 1) is 12.7 to 10.4 ka (Lavrov & Potapenko, 2005). The First Terrace of river Sula has yielded 6 OSL dates in the range of 16.8 to 11.7 calendar ka BP (Astakhov & Svendsen, 2011). Similarly, radiocarbon ages of 16.8 to 11 ka are known from the First Terrace of the Ob drainage system in West Siberia (paper 14 in Ch. III).

Lake sediments

The limnic silts and sands of the final Pleistocene have been studied in detail by international drilling projects of present large lakes (Hahne & Melles, 1999; Henriksen et al., 2003, 2008; Paus et al., 2003) with the thickest and oldest limnic sequences found in deeper lakes of glacially scoured depressions. Within the hilly terrains of arctic Central Siberia on the Taimyr Peninsula (above 74° N) the bottom sediments may attain 60 m in thickness (lakes Levinson-Lessing and Taimyr close to CS in Fig. 1). These silty sequences are partly varves indicating stable sedimentation rates ~ 0.7 mm per year. Assuming an approximate constancy of this rate an Early Weichselian age of the lake inception is inferred (Ebel et al., 1999; Niessen et al., 1999). Although only the upper part of deep lakes sediments 22 m thick and ~ 30 ^{14}C ka old was cored and properly analysed, their pollen spectra provide important records of paleoenvironmental changes (Hahne & Melles, 1999; Andreev et al., 2003, 2004). These archives are more reliable than those obtained from alluvial facies in which fossils are readily redeposited.

The long lacustrine records of Taimyr are apparently continuous whereas in northern European Russia the entire Weichselian sequence from the large and shallow lake Yamozero on the Timan Ridge ~ 20 m thick (LY in Fig. 1) shows two hiatuses which are probably subaerial breaks caused by arid episodes (Henriksen et al., 2008). Present-day lake bed sediments contain long pollen records but are poorly dated due to the lack of organics in these oligotrophic systems. Even at the latitudes ~ 60° N in West Siberia sandy rhythmites of small thermokarst lakes are almost devoid of organics except for rare tundra soils (Astakhov, 1989).

The problem is aggravated by unreliable luminescence dating of deep-water silts or fine sand recovered in cores. This is most evident in the cores of Lake Yamozero where nine OSL dates from under the warmest interval with pollen of broad-leaved trees have yielded a mean value of ~ 72 calendar ka, totally incompatible with their presumably pre-Eemian age (Henriksen et al., 2008).

Attempts to date sediments of the final Pleistocene by radiocarbon are more successful, especially for the interval of 13 to 10 ^{14}C ka BP when incipient lakes became relatively rich in organics due to denser shrubs (Astakhov & Svendsen, 2002, 2011). This stage of lacustrine development is chronometrically close to the alluvium of the First Terrace.

Stratigraphically most important fine-grained lake deposits with organic remains left by thermokarst ponds which developed over the periglacial tundra-steppe during more humid episodes. The best dated are the Varyaha silts interbedded with thin peats, 4-5 m thick, which are sandwiched between the icy glacial diamicton and aeolian sand at Marresale, western Yamal (Ms in Fig. 1). They have yielded six ^{14}C AMS ages in the range of 32.7 to 28 ka BP and three IRSL ages 45 to 36 ka BP (Forman et al., 2002). No significant warming is reflected in the herbaceous pollen spectra of these sediments of periglacial lakes (Andreev et al., 2006).

Such shallow lakes scattered over the perennially frozen tundra after the permafrost degradation occur as lenses of fine sand and silt 5 to 15 m thick building flat inverted hummocks (e.g. Fig. 8 in Astakhov, 2006). They yield rare dates of MIS 3 interval like the 33-35 ka ^{14}C BP (see chapter 4). Similar sediments fill late glacial thermokarst depressions often hosting minor Holocene lakes. They have been partly described and dated in natural sections as sand-silt rhythmites with occasional varved members which started to accumulate ~ 13 ^{14}C ka BP (paper19 in Ch. IV).

Permafrost features

Sedimentary features conditioned by permafrost environments are pervasive because all Russian North was perennially frozen for most of the Pleistocene. Relict permafrost structures are abundant in all the above described formations. Moreover, the very existence of such sedimentary bodies as the yedomas is due to the old stable permafrost. However, underestimation of permafrost induced landscape changes in geological studies of the formerly glaciated terrains of European Russia and Siberia used to be a major source of some puzzling interpretations of surficial topography, fine-grained sediments and ice limits. Spatial intermingling of permafrost features with glacial landforms and sediments often hinder their discrimination. An instructive case is described below mostly based on the Pechora Lowland data.

Ice wedges, the most common and easily identified features of permafrost areas, form expressive polygonal patterns concordant either with the present-day or past climate (Fig. 3). Ice wedges growing in frost cracks of solid permafrost are found everywhere in the arctic mainland but only east of the Urals is their volume comparable to or even exceeding the volume of the ice-containing mineral mass (Romanovsky, 1993). The regularity of ice wedge patterns is crucial for drawing the line between permafrost and glacial landforms in remote sensing imagery. Also, oxygen isotope ratios measured in ice wedges is the main tool in estimating winter temperatures of Late Pleistocene-Holocene environments (Vasilchuk, 1992; Derevyagin et al., 1999, 2010; Hubberten et al., 2004).

In the European Arctic, ice wedges, mostly inactive nowadays, have never grown to such a formidable size as in Siberia. In glaciated areas south of the Arctic Circle ancient ice wedges are generally replaced by ice-wedge casts filled with taberal (thawed) sediments with convoluted structures. They are unalienable parts of the postglacial sedimentary record.

Presently active ice wedges often project above wet concave polygons (Fig. 3A). They are typical for East Siberia but in Europe and West Siberia normally occur in mineral materials only above 68° N. Especially wide ice wedges forming large (50 to 150 m across) polygons which are found on thin degrading permafrost with non-gradient temperature profile in the vicinity of Vorkuta city (~ 67° N/63° E) are Pleistocene relicts (Popov, 1962a). The relict nature of the large polygons is reflected in their convex profile and gentle, sub-rounded form in plan (Fig. 3B). The bounding water-logged furrows with shrubbery mark partly thawed ice wedges 0.6 to 1.5 m thick terminating at 0.3-0.6 m below the table of the present permafrost (Astakhov & Svendsen, 2011).

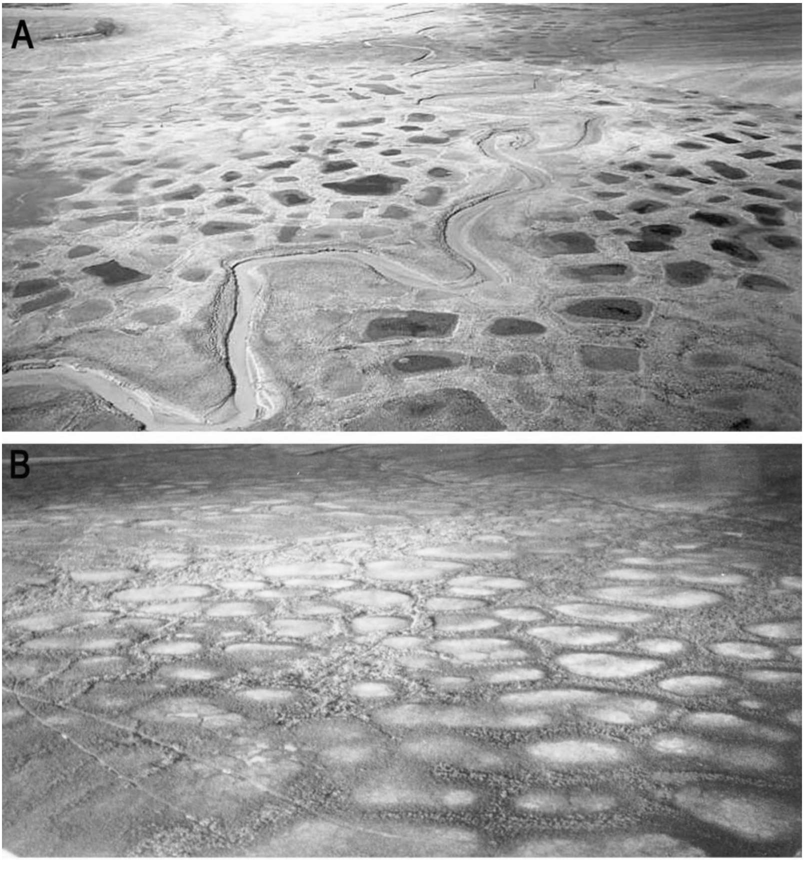

Fig. 3. Helicopter view of tundra polygons bounded by ice wedges. A. Concave active polygons in the Yamal Peninsula (photo by Ó. Ingólfsson https://notendur.hi.is/oi/ siberia_photos.htm). B. Convex relict polygons south-west of the Vorkuta city, Cisuralia. Green stripes are shrubs along the bounding relict ice wedges (Astakhov & Svendsen, 2011). See centrefold for this image in colour.

The importance of the relict frost polygons for delimiting and dating former ice sheets is evident from two opposing models of the last glaciation in the Pechora Lowland. The maximalist model put forward in the 1970s and maintained until recently draws the limit of the Late Weichselian ice far southwards, close to 63th parallel (Arslanov et al., 1987; Grosswald, 1998). This concept is heavily dependent on the glacial origin ascribed to the flat oval sandy hummocks underlain by organics with ^{14}C ages of 33-35 ka BP (Lavrov & Potapenko, 2005).

However, this interpretation is in conflict with several well-known Paleolithic sites on the surface between 65 and 67° N which were reliably dated to 40–20 ka BP. These and other stratigraphic facts made the participants of the PECHORA project to assign ages of 90 to 60 ka BP to Weichselian ice sheets and limit their advances to areas above the Arctic Circle (Mangerud et al., 1999, 2002). Therefore a closer look at the hummocks south of the area of the last glaciation was essential, based on a comparative analysis of similar landforms in glacial and non-glacial landscapes including West Siberia. The resultant explanation puts glacial processes aside and ascribes the flatbread-shaped accretion hummocks of the Pechora Lowland as results of recent degradation of postglacial permafrost (Astakhov, 1998a, b; paper 18, Ch. IV).

In the Pechora Lowland south of the Arctic Circle, where surficial permafrost is already extinct, the pattern of the large rounded frost polygons is imitated by a system of gentle forested hummocks 5 to 10 m high built of laminated fine sands and silts. These flatbread-shaped elevations look as dark-grey ovoids in an image of the lowland east of the Pechora River (Fig. 4). They are labeled \`kames\` by Lavrov and Potapenko (2005) who use the sandy mounds as evidence of their Late Weichselian ice sheet. This age is inferred from the radiocarbon dates 33,520 ± 470 and 34,540 ± 1570 years BP recovered from under the oval piles of laminated sand.

However, the consistent shape and size of the flat hummocks copying the oval form of adjacent lakes, their regular pattern, the preferred mass orientation and occurrence on low terraces are patently different from usually chaotic patterns of ice disintegration. A comparison of this thawed landscape with presently frozen coastal flats shows that the sandy ovoids mirror the configuration of shallow thermokarst lakes wandering over modern solid permafrost (Fig. 5A, B). An analogous pattern of flat sandy hummocks and lakes is widespread in formerly frozen West Siberia 500 km southwest of the Weichselian ice limit (Fig. 5C, see also Fig. 8 in paper 14, Ch. III).

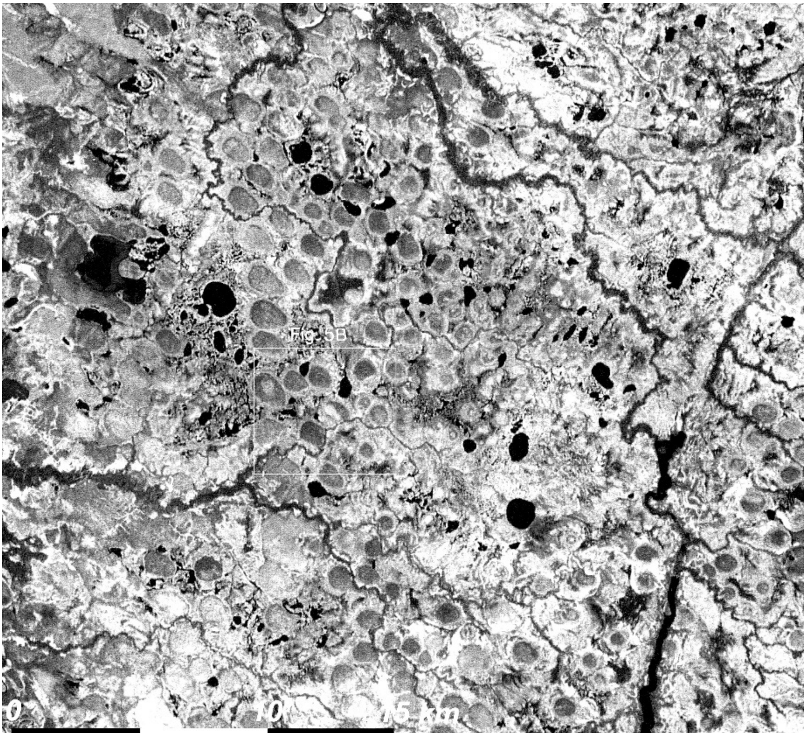

Fig. 4. Satellite view of flat forested hummocks (dark-grey ovals) on a thawed low terrace of the Pechora River east of O in Fig. 1, south of river Sozva, 66°20′ N. Their shapes replicating contours of adjacent lakes (black) are similar to rounded polygons in Fig. 3B. The white quadrangle indicates aerial photo in Fig. 5B.

The WNW-ESE elongation of oval lakes, common also in present-day European tundras, is probably caused by anisotropic wave thermoerosion due to dominant winds or insolation asymmetry (Boitsov, 1961). The fine-grained composition of the flat mounds and lack of cobbles or diamict sheets usual for kame plateaus supports a non-glacial origin for these features. Generic kames are steeper and higher, often mantled by ablation diamicts or gravels and do not occur on riverine terraces. Besides, they make chaotic agglomerations of hummocks, sometimes 50 m high, with intervening deep lakes of intricate configuration but not a regular pattern of separated features as in Fig. 4 and 5 (Astakhov, 1998b).

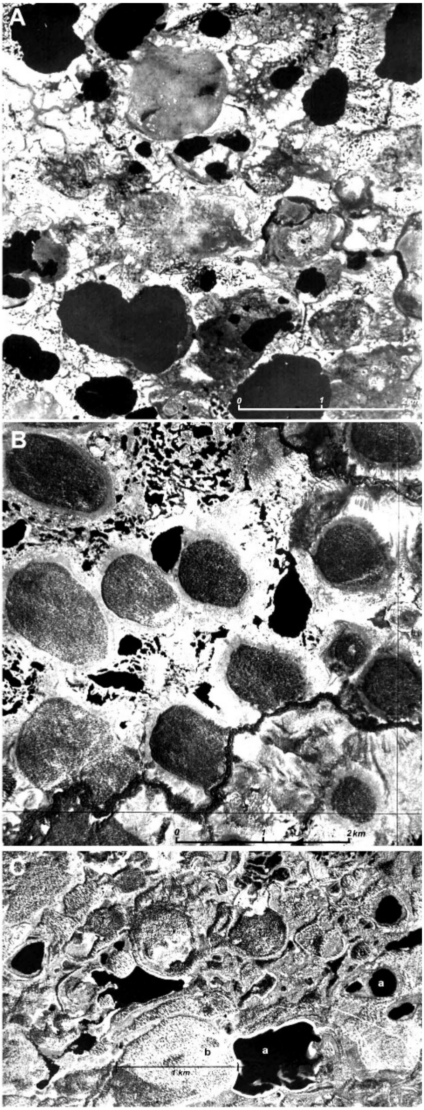

Fig. 5. Similar patterns of live and inverted thermokarst lakes (aerial photos). A. Present thermokarst lakes (black) on the perennially frozen Barents Sea coast 240 km north of Fig. 4, at 68°30′ N/55°20′ E. B. Dark-grey flatbread-shaped hummocks on extinct permafrost (quadrangle in Fig. 4) mirror shallow thermokarst lakes. C. Analogous system of lakes (a) and inverted sandy hummocks (b) in thawed central West Siberia, 63° N/82° E (Astakhov, 1998b).

Thus, flat hummocks scattered over riverine terraces is evidence of not a former ice sheet but thermokarst inversion of topography caused by permafrost degradation (Boitsov, 1961). The above mentioned plant detritus 33-35 ^{14}C ka old obtained from under the hummocks is probably left by Mid-Weichselian thermokarst lakes which spotted the Pechora terraces along thawing thick ice wedges of the preceding frosty interval. This is also suggested by their rhythmic pattern similar to polygonal tundra (cf. Fig. 3B and 4, 5). Sand-silt sediments of the local thermokarst lakes are eventually inverted into flat hummocks as a result of the final permafrost degradation in the Holocene which leads to regional sagging of the entire terrain (Fig. 6A).

The morphological difference between glaciokarst features of ice disintegration and thermokarst features of permafrost degradation is schematically shown in Fig. 6B. Glacial karst develops on thick relatively clean ice where sinking and lateral shifting of thaw lakes are not confined by surrounding mineral matter. This results in numerous deep lakes of intricate configuration abundant in kame landscapes. Deep lakes and englacial cavities on different levels of disintegrated glacial ice when filled by melt-out sand and gravel and finally projected on each other produce mounds of meltwater deposits in the form of steep hummocks (Fig. 6B, right).

Conversely, shallow thermokarst ponds intitiated by thawing ice wedge patterns are restricted in their development by ground massifs with low ice content and by the thickness of the permafrost. Their thin bottom sediments after sagging of degraded permafrost produce low hummocks positioned slightly below the reference plane (Fig. 6B, left). The resultant knolls, which are imprints of inverted thermokarst ponds, are generally flatter than kames, often cluster in polygonal patterns and commonly have, due to isotropic thermoerosion of frozen soil, a fairly regular oval or circular form in plan (Astakhov, 1998a, b).

The important role of permafrost in structure of the postglacial formations is also evident from the diamictic facies that blankets gentle slopes in various environments and at different elevations. In the Uralian foothills this is usually soft clayey diamicton 0.5 to 4 m thick containing up to 30-35% of rock fragments or pebbles. In the flatlands such muds with shattered pebbles occur either as lenses or tongues within aeolian and lacustrine formations (Fig. 7A) or fill ice-wedge casts. In the wells drilled across the Pechora valley south of the Arctic Circle diamict sheets attain 10 m in thickness (Yudkevich & Simonov, 1976).

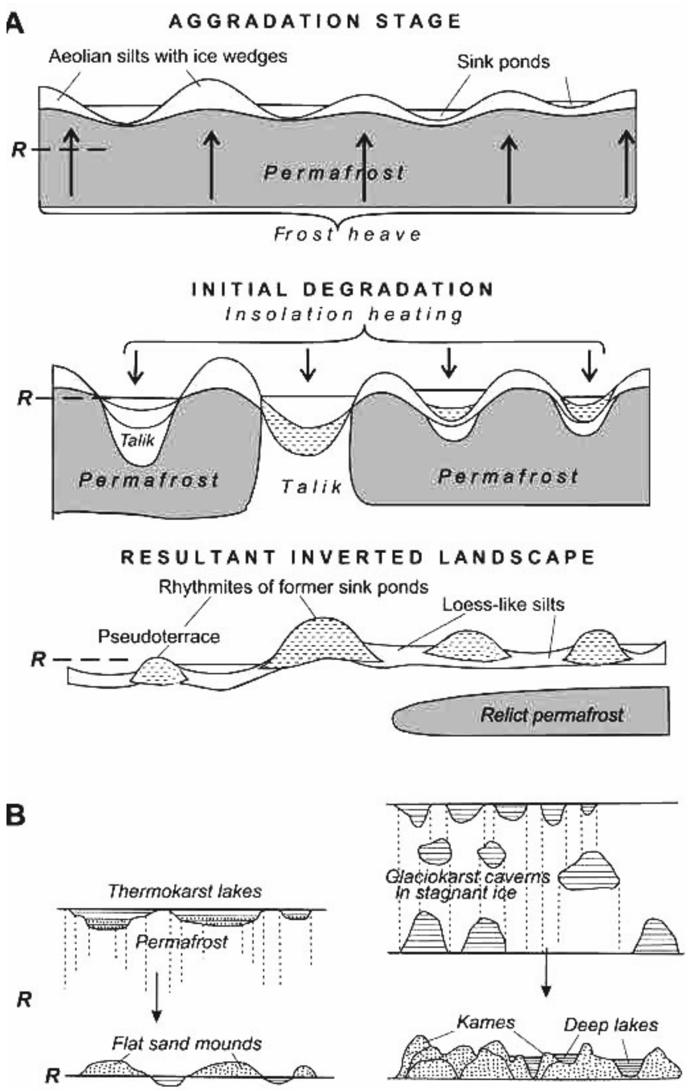

Fig. 6. Schematic profiles of geomorphic features resulting from cryosphere degradation. A. Stages of permafrost development leading to inverted thermokarst topography. B. Comparison of thermokarst (left) and glaciokarst (right) topographic inversions (adapted from Astakhov, 1998a, b). Ref – reference plane.

Around the city of Vorkuta the thin veneer of the pebbly diamicton in olden times was mapped as `the Upper Till`. However, the diamict sheets are readily traced from higher plateaus onto riverine terraces thus revealing their solifluction origin. Their internal structure is often convoluted by creep and plastic flow. They are accompanied not by glacial disturbances and varved clays but by wind-blown sands and loess-like silts (Astakhov & Svendsen, 2011).

Solifluction sheets of diamictic material have been studied and dated in excavations of Byzovaya Paleolithic site where they contain numerous mammal bones and artifacts marking episodes of relative humidification within generally arid periglacial environment (Heggen et al., 2012). Solifluction sheets, judging by aerial photographs are widespread on slopes of the Central Siberian uplands. Their general properties were described by geocryologists (e.g. Gravis, 1969) but slope deposits of glaciated Central Siberia have not been studied stratigraphically.

Subaerial association

Lithologically diverse waterlain sedimentary bodies are limited in their extent whereas the thin mantle of terrestrial materials, usually less than 25 m thick, often poor in organics, is ubiquitous on morainic plateaus and riverine terraces indiscriminately. This complex consists mostly of soft fine-grained sands and silts with intervening paleosols, tongues of soliflucted diamicts, thin peat lenses, sinkhole silts, and narrow sand strips of low-energy temporal streams. The complex is designated as the `subaerial association` (Volkov, 1971) because of its predominantly silt and fine sand lithologies, its universal occurrence on all northern plains irrespectively of large water bodies and the frequent signs of high aridity during the deposition.

The complex contains numerous ice wedges which grow in size and number eastwards to reach maximum in the `Ice Complex` of East Siberia where mineral matter is often volumetrically subordinate (Romanovsky, 1993). In European Russia and West Siberia below the Arctic Circle they are replaced by pseudomorphs after ice wedges. The best dated and paleontologically studied sequences of this formation with different genetic labels have been reported from several archeological excavations (Mangerud et al., 1999; Pitulko et al., 2004; Svendsen et al., 2010; Heggen et al., 2012) and the icy bluffs of arctic Siberia (Forman et al., 2002; Möller et al., 1999, Sher et al., 2005, Schirrmeister et al., 2002; Vasilchuk, 1992).

Aeolian activity, pervasive in the present treeless Arctic, should have been much more powerful in Pleistocene continental climates, as it is acknowledged worldwide. However, Russian arctic research for decades was focused on aqueous and glacial processes at the expense of subaerial formations. Studies of aeolian sedimentation have been limited to the periglacial area of Central and East Siberia (e.g. Kolpakov, 1986; Tomirdiaro, 1980), this despite their importance for our understanding of the postglacial Pleistocene history. The features of northeastern European Russia are discussed below in more detail because there the surficial mantle due to the Holocene climatic amelioration is generally iceless and more accessible for field studies.

Sand deposits

The surficial mantle of predominantly fine sands of the northeastern Russian Plain is depicted in old geological maps as marine and glaciolacustrine sediments. Such attributions are still maintained by Lavrov and Potapenko (2005). However, the Russian-Norwegian investigations of 1993–2002 challenged this point of view in favour of the subaerial origin accepted by all participants of the PECHORA project (Astakhov et al., 1999; Mangerud et al., 1999).

There are two types of surficial sand accumulations. The most massive sand up to 20 m thick sporadically occurs in all depressions but it is thickest in close vicinity of the Late Weichselian ice sheet along the south-eastern Barents Sea coast (Fig. 1). The dry sandy tundra with numerous blowouts is composed of fine-grained, sometimes medium, predominantly quartz sand with long diagonal bedding (Mangerud et al., 1999). The thick sand forms a flatland strip several km wide at 20 to 60 m a.s.l. (Fig. 2 in paper 18, Ch. IV) which was previously interpreted as a marine terrace of the final Pleistocene (Arslanov et al., 1987).

However, along the 150 km of its coastal length the narrow flatland retains its peculiar composition of monotonous, pale-yellowish loose sand without beds of clay or gravel, with rare narrow sand wedges, weak soils, involutions and peat lenses. No aquatic organics has been found in this formation but a large mammoth tusk ~ 48 ^{14}C ka old was picked up at the base of the bluff. OSL dates from 17 to 13.7 ka have been obtained from dune sands at Timan Beach (T in Fig. 1) (Mangerud et al., 1999), on rivers Kuya (K in Fig. 1) and Vonda (V in Fig. 1) (Astakhov & Svendsen, 2011). The formation of thick sands is distinctly alien to the modern wet mossy tundras and boreal forests.

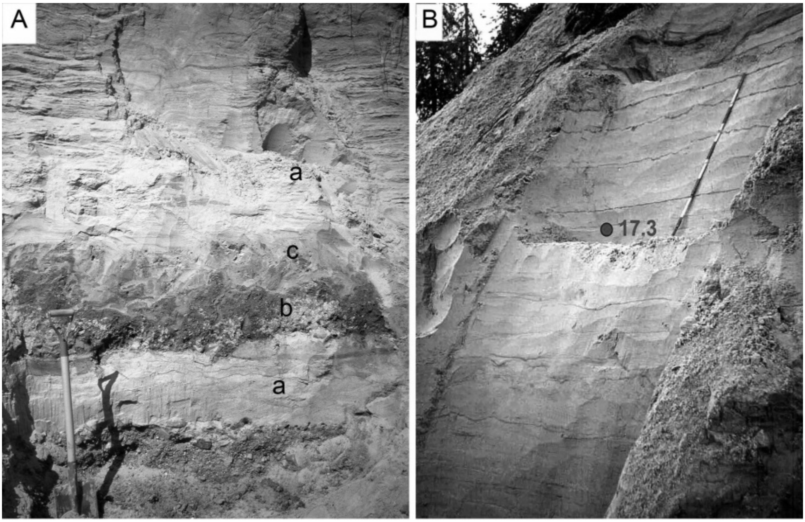

Fig. 7. Coarser subaerial sediments of the Pechora Lowland. A. Subaerial sequence at Akis (A in Fig. 1): aeolian sand (a) containing a wisp of soliflucted diamict (b) with arid soil (c) on top (photo by author). B. Aeolian medium to fine sand on river Vonda (V in Fig. 1). Parallel rusty bands are casts of former veins of segregated ice formed at the base of each subsequent active layer syngenetically with accretion of sand. Red number is OSL age, ka BP. 1 m ruler for scale (photo by O. Nikolskaya). See centrefold for this image in colour.

Regular sets of ripple marks or other clear signs of aqueous sedimentation have not been found in this sandy formation. Long diagonal bedding and the yellowish colours of the loose sands indicate an aeolian genesis. Their deposition in very continental climates is also clear within river valleys where, besides narrow ice-wedge casts, other traces of former permafrost such as wisps of solifluction diamicts are seen (Fig. 7A). Fig. 8B depicts a section of soft sand, at least 15.5 m of apparent thickness, dated to 17 ka BP (V in Fig. 1) (Astakhov & Svendsen, 2011). The peculiar feature of the sand is a regular sequence of subhorizontal rusty seams cutting at low angles thick laminae of pale-yellow medium sand. The rusty seams are identified as casts of former ice veins of lenticular cryogenic structure. The veins normally form at the base of the active layer by segregation during its seasonal refreezing (Romanovsky, 1993). Their presence indicates accumulation of aeolian sand simultaneously with permafrost aggradation. Similar striped sands have also been described on river Kolva and other tributaries to the Pechora (Astakhov et al., 1999; Astakhov & Svendsen, 2011).

Ancient sand dunes also occur in places beyond the Urals, far from the Late Weichselian ice sheet. Wind abraded quartz grains with matte surfaces occur on the surface of West Siberian sandy plains beneath Holocene bogs (Velichko & Timireva, 2005). In the southern Yamal Peninsula aeolian sand has yielded OSL ages ~ 50-60 ka BP.

Another distinct type of sand deposits in the Pechora Basin is a discontinuous cover of tan or rust tinted unevenly laminated fine sand with coarse silt laminae that are draped over low terraces and gentle slopes (Fig. 8A). At the valley bottoms these cover sands may be interstratified with dune sand and massive loess-like silt (Fig. 8B, see also photos in Heggen et al., 2012). Fine-grained cover sands are often more compact than dune sand and prone to solifluction if wet (Fig. 9A). The cover sand blanket is usually 2-3 m thick, in places up to 7 m. It was originally mapped as late-glacial lacustrine sediment (Lavrov & Potapenko, 2005).

However, the regional grouping shows a spatial trend from dune sands along the Barents Sea coast to cover sands of the Pechora valleys and to loess-like silts farther southeastwards. This distribution is the reverse of what would be expected for waterlain sedimentation because the coarser facies occurs at the lowest altitudes close to sea level and the finest facies occupies the Uralian piedmont. The modern interpretation of the laminated cover sand views it as a result of snowdrift traction of sand and coarse silt by strong winds over a flat frozen surface. Spring melting of sand-snow drift is probably the reason for the wavy or dimpled surface of individual laminae (Mangerud et al., 1999; Astakhov & Svendsen, 2011). This type of cover sand is well-known as niveo-aeolian sediment in the periglacial zone of western Europe (Koster, 1988). Patches of cover sand are common in northeastern European Russia. They occur sporadically in West Siberia (Fig. 8B, 10), but they are very rare farther eastwards where the governing anticyclonic climates were probably not especially favourable for sand drifts.

Whereas the dune sands normally keep the uppermost position in the aeolian sequence their bottom beds in places interfinger with cover sands (Fig. 9A). OSL dates 26 to 15 ka from the cover sand facies on rivers Adzva, Bol Rogovaya, Kuya, Ngutayaha (Ad, BR, K, Ng in Fig. 1) suggest that it is often slightly older than most of the dune sands in European Russia: younger cover sands are also known, for example, on the river Korotaikha (Ko in Fig. 1) where they have yielded 6 OSL dates in the range of 14.7 to 13.1 ka (Astakhov & Svendsen, 2002, 2011).

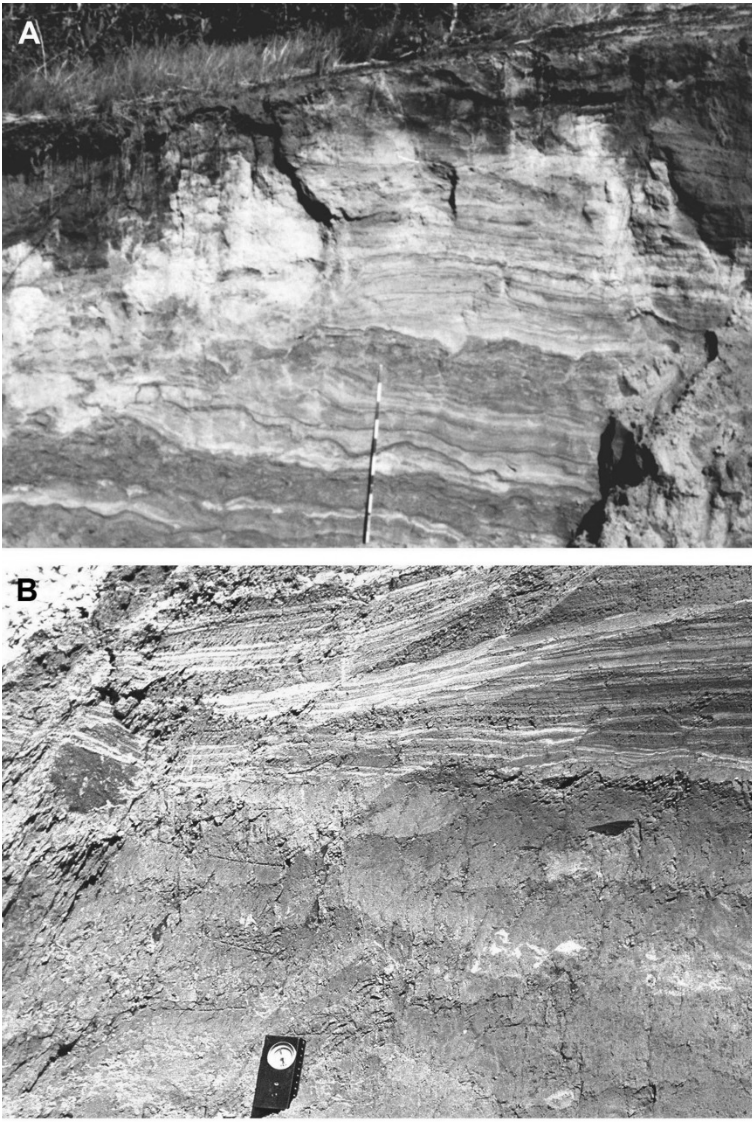

Fig. 8. Fine-grained subaerial sediments in northern flatlands. A. Niveo-aeolian coversand with wavy lamination on a high terrace of river Kolva, Pechora Lowland, 1 m stick for scale (Astakhov & Svendsen, 2011). B. Loess-like silt (bottom) with coversand (top) at Kiryas on the River Ob (Fig. 13, LKM in Fig. 1). See centrefold for this image in colour.

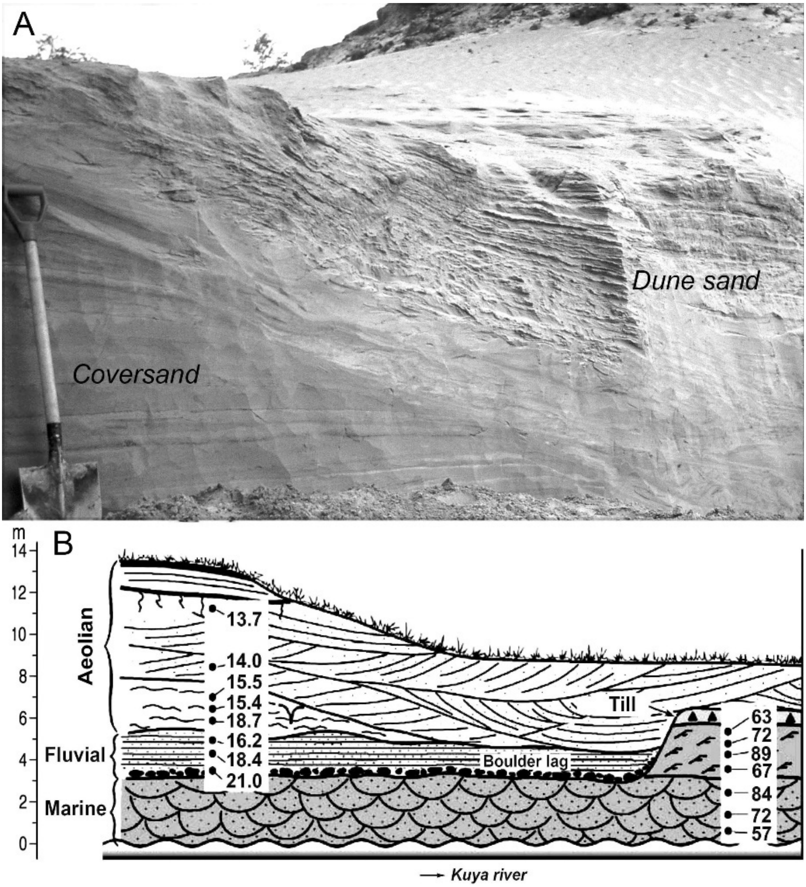

Fig. 9. Postglacial aeolian sands on the Lower Pechora, river Kuya section (K in Fig. 1). A. Compact fine-grained cover sand with silt laminae and traces of sediment flow overlain by loose dune sand. B. OSL dates ka BP in the Kuya sequence (Astakhov & Svendsen, 2011).

Similar facies probably occur also farther east but because of their relative scarcity they have not attracted much attention yet. In West Siberia cover sands are dated to MIS 2 time at Aksarka on the Lower Ob (Ak in Fig. 1) where they are up to 7 m thick (Fig. 10A). In the western Yamal at Marresale (Ms in Fig.1), the Oleny cover sand has ^{14}C dates from 28.3 to 25.1 ka BP and IRSL dates 33-30 ka BP (Forman et al., 2002). In the Pechora Basin and central West Siberia the mantle of cover sands with loess-like silts is commonly mistaken for lacustrine formations.

Loess-like silts

Loess-like silt is the most common sediment of any periglacial zone. Thawed loess-like silts are known as `cover loams` in Central Russia and West Siberia. In the Russian North they are observable either as a thin veneer on valley slopes or as a thick ice-silt mixture constituting the yedoma loess-ice formation of the Siberian Arctic. Better opportunities to trace and understand key sedimentary features exist in the west of the periglacial zone, where loess-like silts are thinner but permafrost is less icy or already degraded. In the Subarctic Pechora Basin loess-like silts are as a rule located distally from Weichselian ice limits, for example, along the western Uralian piedmont where they are up to 7 m thick (Astakhov & Svendsen, 2011). The topmost member of the silty cover found in river valleys yield OSL ages in the range of 24 to 11 ka BP on both sides of the Urals (Astakhov et al., 2004, 2007). In places it contains thin buried soils with shrubs and mammal bones ^{14}C dated to ca 12-13 ka BP, as at Akis (A in Fig. 1) (Astakhov & Svendsen, 2011).

Fig. 10. Unfrozen subaerial formation of West Siberia in Aksarka sand pit (Ak in Fig. 1). A. Thick laminated cover sand dated by OSL with a lense of silt with plant remains at the base dated by radiocarbon. B. Close-up of the cover sand with small ice wedge casts, 1 m stick for scale.

The distribution of loess-like silts in West Siberia is patchy, leaving great expanses of northern lowlands covered by dune sand or occasional blowouts. In the eastern Yamal Peninsula, yedoma-type frozen silts more than 20 m thick are exposed down to present sea level at famous section of Syoyaha (Sy in Fig. 1). They contain thin moss mats, bands of segregated ice and telescoped syngenetic ice wedges.

This sequence has a long series of ^{14}C dates from 37 to 11 ka BP. The interval 25 to 17 ^{14}C ka was studied there in a continuous succession with nine dates (see Fig. 3 in paper 4, Ch. I). Plant remains from the wedge ice have yielded AMS ages in the range of 21–14 ka BP, indicating that the large ice wedges formed during the MIS 2 time span. The δ^{18} O values are ~ -23% as compared to ~ -17% for Holocene wedge ice in this area (Vasilchuk et al., 2000). The Syoyaha icy silts and similar sediments on the Gydan Peninsula were described in the 1980s by geocryologists as either an estuarine formation (Vasilchuk, 1992) or counterparts to the East Siberian yedomas (Bolikhovsky, 1987).

The latter author also described numerous mammoth bones from the West Siberian yedoma of the Gydan Peninsula (L in Fig. 1). The bones have yielded 6 ^{14}C ages in the range from 18,380 ± 700 to 16,520 ± 550 years BP (paper 16 in Ch. III). The striking similarity of the Syoyaha silts to the Alaska loess was acknowledged by Péwé and Brown (1989). Afterwards the Syoyaha section was used as evidence of continuous postglacial sedimentation of subaerial nature and a stratotype for the uppermost Pleistocene (paper 16 in Ch. III).

The thawed loess-like silts occur in many places in the central lowland of West Siberia at 40-50 a.s.l. beyond the limits of Late Pleistocene ice sheets. They are interstratified and laterally replaced with cover sands (Fig. 8B) and rhythmites of small thermokarst ponds (paper 14 in Ch. III). The silty formation atop of the Middle Pleistocene moraines is definitely older than similar silts in the Arctic. For example, a thick sequence at Bogdashkino on the Ob river, 61°10′ N, has yielded a series of thermoluminescence dates in the upward succession: 110 ± 15, 80 ± 11, 75 ± 10, 65 ± 8 and 65 ± 8 ka BP (Arkhipov et al., 1987). U/Th dates ~ 105 ka BP are reported from peat at the base of the periglacial silty sequence at Kiryas (LKM in Fig. 1) (Laukhin, 2011). In the Yenissei valley the loess-like silts with paleosols and peat lenses are generally thinner, with a maximum 5 m thick at Alinskoye (Al in Fig. 1) where they yielded a succession of ^{14}C ages of 44.2, 36.9, 31.9, 22.8 and 15.7 ka BP (Sukhorukova et al., 1991).

Thick loess-like silts with interlayers of fine sands and moss mats have been thoroughly described and dated in the Taimyr Peninsula at Cape Sabler section (Kind & Leonov, 1982; Möller et al., 1999). A similar surficial formation proceeds eastwards in the periglacial zone of the East Siberia where massive loess-like silts containing voluminous ice, buried soils and peat lenses may be more than 40 m thick, their age exceeding 47 ^{14}C ka (sites Ma and CB in Fig. 1). The increased thickness of the East Siberian yedoma is largely due to the abnormal volume of ground ice. The yedoma silts often appear pseudo-laminated because of many thin horizontal bands of segregated ice marking former active layers in the growing pile of silt (Romanovsky, 1993). The silts are not lithologically or structurally different from analogous sediments of West Siberia (Bolikhovsky, 1987) but east of the Yenissei river they are progressively richer in ice wedges and organic remains. The yedoma silts, as a rule, contain grass roots, vascular plants, organic detritus, lenses of moss mats, fossil insects and remains of mammoth megafauna.

Features of wind erosion

Local sources of wind-blown sand are readily found on the arctic plains built of predominantly glaciofluvial, marine and alluvial sand/gravel formations of the Upper Pleistocene which were actively deflated in periglacial climates. High interfluvial plateaus bear unmistakable signs of recent strong wind erosion that contrasts with the moderate aeolian activity in the present mossy tundra. They are especially numerous along the Barents Sea coast and in the Yamal Peninsula.

There are two kinds of wind-eroded landforms: horseshoe-shaped parallel ridges of push moraines and isolated conical residuals. A number of such features are shown in the map of the Pechora Basin (Fig. 1 in Astakhov & Svendsen, 2011). Multiple conical hillocks with pebble armours are abundant in the Siberian Arctic, for example, in the central Yamal Peninsula. These features for a long time puzzled arctic researchers who advanced various hypotheses of their origin related to glacial or snow action (e.g. Danilov, 1965). The aeolian explanation appeared in the Russian literature only after the works of 1990s (Astakhov et al., 1999; Mangerud et al., 1999).

In the Pechora Lowland the conical hillocks were previously mapped as `kames`, i.e. features of ice disintegration (Lavrov & Potapenko, 2005) which at a closer look cannot be accepted for the following reasons. The isolated hillocks have distinctly concave profiles of eroded slopes and pointed summits (Fig. 11 top), unlike the gentle convex profiles typical for classical kames. The tops of the conical hillocks are devoid of vegetation and are usually armoured by cobbles with traces of wind polishing. The parental sands observed in shallow pits often contain no visible pebbles which implies a large mass of sand has been blown away to provide enough stones for the cobbly veneer (Astakhov et al., 1999).

Similarly shaped cones of smaller size occur in numbers along the crests of arcuate push moraines built of deformed marine or fluvial sands. The small pebbly knolls follow the strike of steeply dipping beds (Fig. 11, bottom) and must be deflation residuals after the removal of sand and silt particles from the outcropping sand formations. When tundra plateaus are composed of gravelly sands, their deflated outcrops stand out as the highest and sharpest hillocks at 150-200 m a.s.l., as observed between the Pechora and Urals along the 68th parallel (Astakhov & Svendsen 2011).

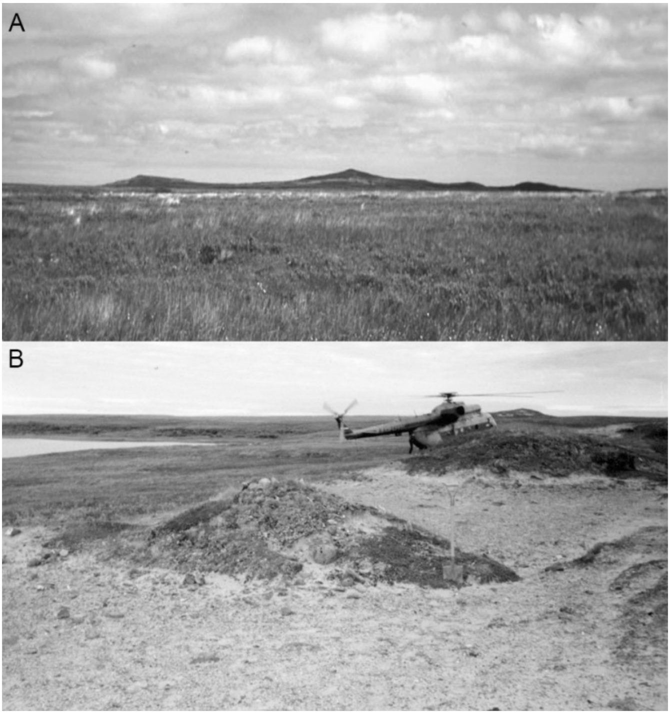

Fig. 11. Deflation residuals in arctic European Russia (Astakhov and Svendsen, 2011). A. Isolated conical hillocks west of the Pechora mouth (kames by Lavrov & Potapenko, 2005). B. Small conical knolls armoured by cobbles and gravels along the crest of the glaciotectonic ridge at Vashutkiny Lakes, 67°57′ N/61°40′ E. See centrefold for this image in colour.

The entire structure of the postglacial Pleistocene mantle of the European Arctic west of the Urals is schematically shown in the profile of Fig. 12. It includes the main types of postglacial deposits: sand and gravel alluvium, solifluction diamicts, dune sands, fine cover sands and loess-like silts. East of the Urals, in the West Siberian lowlands the postglacial Pleistocene has a somewhat different composition: solifluction sheets are rare, alluvial formations are more voluminous although they are invariably blanketed by thick loess-like silts and coversands with lenses of limnic rhythmites (Fig. 7 in paper 6, Ch. II, for details see Fig. 8 in paper 14, Ch. III). Origin of this formation is further discussed below.

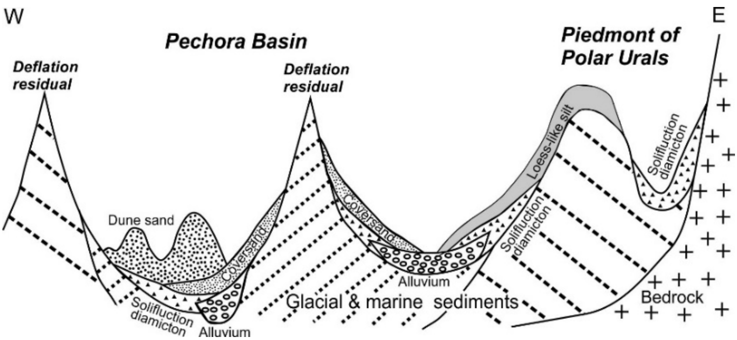

Fig. 12. Idealised profile of the Pleistocene postglacial subaerial association west of the Polar Urals (Astakhov & Svendsen, 2011).

Paleontological proxies

The available paleontological data in general are in good agreement with the sedimentological evidence of basically arid and cold climates. They add valuable information about non-analogous Pleistocene environments which in many cases were more beneficial for terrestrial fauna than the Holocene conditions.

Paleobotanic evidence

Pollen spectra from alluvial sediments, commonly full of redeposited grains, might be totally misleading along the great Siberian rivers carrying warmer water with diverse flotsam from the dense southern taiga to the Arctic. It is a common case when forest pollen spectra are in apparent discrepancy with the lack of driftwood in alluvium of great northbound rivers (e.g. paper 15 in Ch. III). Lithologically uniform sediments of deep modern lakes provide clayey traps for pollen grains and are more valuable for reconstructions of past environments although oligotrophic arctic lakes are very poor in organic remains.

Pollen analysis gives a consistent although generalized picture of the postglacial vegetation. In all studied sections of the Eurasian North a sudden change of Late Pleistocene non-arboreal pollen assemblages with xerophytes to Holocene spectra of boreal forest or shrub tundra has been recorded at ~ 10-11 ^{14}C ka BP (Andreev et al., 2003, 2004, 2006, 2011). Within the last 30 ^{14}C ka of lake sedimentation in the Taimyr Peninsula no intervals of vegetation similar to the Holocene mesic communities of shrub tundra and boreal forest have been described: all Pleistocene pollen assemblages are patently periglacial, i.e. contain cryoxeric mixture of steppe and tundra plants. Minor changes to more mesic vegetation are recorded for short intervals richer in shrub pollen and poorer in pollen of *Poaceae, Cyperaceae* and *Artemisia* (Andreev et al., 2003; Sher et al., 2005).

Only in the present dense forest of subarctic European Russia (~ 65° N) Middle Weichselian lacustrine sequences show intervals with arboreal pollen up to 50% (Henriksen et al., 2008). The inferred shrub tundra with some trees soon changed to herb-dominated cold steppe of the Late Weichselian time. Close to the Barents Sea, just above the Arctic Circle, the Late Weichselian pollen assemblages of a lacustrine sequence are totally dominated by cryoxeric herbaceous communities (Paus et al., 2003). Even the organic-rich Varyaha sediments of shallow interstadial ponds in western Yamal (Ms in Fig. 1) contain pollen of grass-sedge tundra with an admixture of xerophytes and shrubs, thus indicating cold and dry climate, slightly milder than LGM environments (Andreev et al., 2006).

Several Russian authors, however, reported rich forest spectra from presumably Middle Weichselian sediments in the northern Pechora Lowland (Arslanov et al., 1987; Lavrov & Potapenko, 2005). But these forest beds were found under the upper glacigenic complex which is now reliably proven to be older than 50 ka. Therefore the forest beds with finite radiocarbon dates were most likely deposited in MIS 5 time, a problem that is endemic for conventional radiocarbon dates in the Russian North (Mangerud et al., 2002; Sher et al., 2005; Astakhov & Nazarov, 2010).

Pollen spectra from continuous silty sequences of yedoma type everywhere in the Siberian Arctic also indicate treeless, herb/grass dominated vegetation for the last 40 ka of the pre-Holocene history with an admixture of shrub pollen only in certain Mid-Weichselian intervals and in the Late Glacial (Vasilchuk, 1992; Andreev et al., 2003; Sher et al., 2005; Andreev & Tarasov, 2007).

Plant macrofossils which are locally transported give more reliable account of postglacial environments than far transported pollen. Thus, the well-known Cape Sabler sequence on the Taimyr Peninsula (Fig. 13, CS in

Fig. 1) was first erroneously ascribed to lacustrine activity. However, the dominance of xerophytes over hydrophytes in macrofossil herb assemblages recovered from thick loess-like silts with peat interlayers and mammoth bones (Kienast et al., 2001) led to the adoption of basically subaerial origins (Möller et al., 2011).

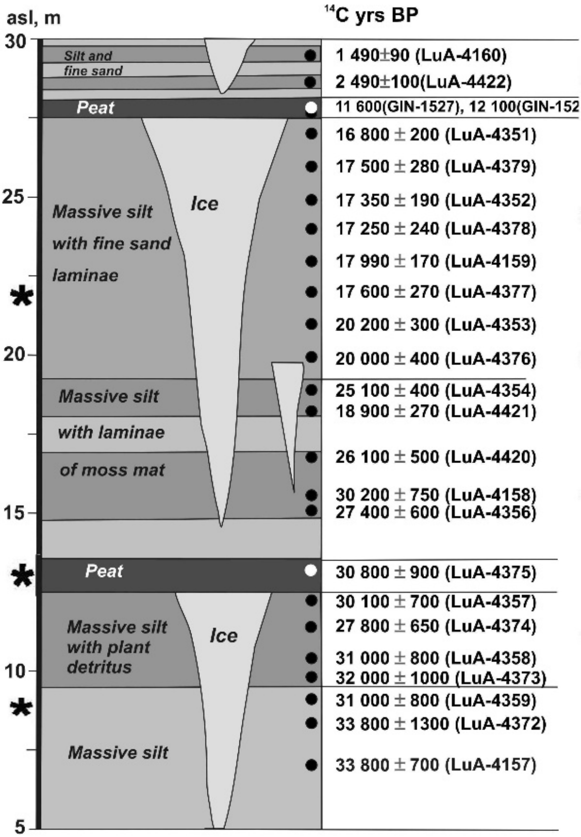

Fig. 13. Key section of postglacial sediments at Cape Sabler, Lake Taimyr (CS in Fig. 1) with syngenetic ice wedges. Uncalibrated radiocarbon ages on plant remains are in the right. Stars indicate levels sampled for plant macrofossils. Adapted from Möller et al., 1999; Derevyagin et al., 1999; Kind & Leonov, 1982; Kienast et al., 2001.

Mammoth fauna

Pollen indications of persistent periglacial character of arctic vegetation are in accord with megafauna remains found in subaerial and fluvial formations. The frozen mammoth carcasses scattered close to the surface of arctic Siberia have yielded a wide range of radiocarbon dates, from infinite to 10 ka BP (Fig. 1). It is significant that mammoth fauna, although less abundant, existed along the Laptev Sea shores even during the coldest interval of 24–15 ^{14}C ka BP (Sher et al., 2005).

This fact is in harmony with mammoth bones of this age found around the present glaciers of Severnaya Zemlya (Makeyev et al., 1979) and on the Arctic Circle in the lee of the Urals (paper 11, Ch. III). Eight ^{14}C dates in the range of 19 to 16 ka BP have been obtained at Lysukanse, the Gydan Peninsula (L in Fig. 1) on various mammoth bones recovered from Yedoma silts (Bolikhovsky, 1987). In this respect Arctic West Siberia is similar to the Taimyr Peninsula which is rich in mammoth remains (Mac Phee et al., 2002). The mammoths which wandered in the MIS 2 time span do not suppport the idea of Late Weichselian cold desert all over the West Siberian north based on aeolian sand grains recovered from under the Holocene peats (Velichko & Timireva, 2005). Sher et al. (2005) more correctly stated that the LGM environment in Siberia was probably less favourable for grazers, but not intolerable.

The typical representatives of postglacial fauna – mammoth, horse and saiga, sometimes muskox and woolly rhinoceros – suggest open tundra-steppe landscapes with abundant grasses and herbs providing ample forage for large grazers. Plants remains recovered from the yedoma of the Laptev Sea region indicate low snow cover which explains the survival of megafauna in the extremely cold winters of LGM time. `Because of lacking snow cover, the nutrient rich grasses and forbs would have been available for herbivores` (Kienast et al., 2005, p. 298). The steppe megafauna cannot tolerate snow cover thicker than 30 cm or uneven surfaces associated with mires, but both circumstances appeared only in the Holocene (Tomirdiaro, 1980; Sher et al., 2005).

These observations contrast with the absence of mammals remains of the LGM time west of the Urals in the vicinity of the Barents Ice sheet where, unlike in the Siberia tundra-steppe, abiotic polar desert dominated. Mammal bones, plentiful in Weichselian sediments older than 24 ^{14}C ka BP, reappeared in flatlands of northern European Russia about 14 ^{14}C ka BP. Scarce remains of LGM time fauna (mostly rodents) are found only in rare refugia such as Uralian caves or Pymva-Shor Paleolithic site with the unique hot springs (Mangerud et al., 1999).

Other organics

Important paleoclimate results are obtained by analyses of fossil beetles (*Coleoptera*) which have been studied in the periglacial zone of East Siberia. Such fauna in general are the best proxy of summer temperatures of air and soil because these factors are decisive for insects breeding (Sher & Kuzmina, 2007).

The most continuous record has been obtained from the longest and best studied section in the cliffs of Bykovsky Peninsula (BP in Fig. 1, 15). Here the *Coleoptera* assemblages show a mixture of steppe and tundra species, even more spectacular than the mixed megafauna (Sher et al., 2005). The authors found steppe insects persisting throughout the succession from 47 to 12 ^{14}C ka BP which implies summer temperatures higher than present ones even within the coldest interval of 24-15 ^{14}C ka BP. The beetle analysis allows the postglacial Pleistocene to be divided into four stages. The contribution of thermophilous and xerophilous species was more noticeable in the first half of the middle Weichselian interval, especially in the assemblages dated to about 45 to 48 ^{14}C ka where they reach 30 to 50%.

These steppe species now `occupy isolated areas in Central Yakutia, and the northernmost occurrences of some of these species are known on the relict steppe patches in the upper basins of the Yana and Indigirka rivers, where the mean July air temperature is 12-14°C (i.e., at least 5-7° higher than on the Bykovsky Peninsula)` (Sher et al., 2005, p. 541). Counted together with dry tundra species the percentage of relatively thermophilous forms gradually decreased from 48 to 34 ^{14}C ka. The percentage of arctic species simultaneously increased almost reaching the LGM levels (more than 50%) (Sher & Kuzmina, 2007).

The second half of the middle Weichselian (34–24 ^{14}C ka) is marked by a low share of xerophilous species (usually below 7%) and abundance of mesic tundra insects with an admixture of high arctic species, implying more humid and cooler summers (Sher & Kuzmina, 2007). Most of the Late Weichselian time, from ~ 24 to 15 ^{14}C ka, was characterized by low percentage of xerophiles, disappearance of thermophiles, and dominance of arctic tundra species in summers that were still warmer than today. A sharp climatic turnover from the Late Weichselian peak of cold to the warmest and driest stage started at ~ 15 ^{14}C ka BP when thermophilous insects reappeared together with the increased mammoth population (Sher et al., 2005).

The above reconstruction of the Weichselian environment is supported by independent data on fossil soils studied in East Siberia (Gubin et al., 2008) and is probably applicable to all northern Siberia where the mammoth tundra-steppe coexisted with formation of the icy Yedoma sediments. Yedoma silts are mostly syngenetic cryopedoliths with low organic content formed in the active layer under regular influx of aeolian silt. Especially valuable for paleogeographic reasoning are fossil burrows containing rodent, plant and beetles remains, obviously *in situ.* The burrows with heaps of seeds stored by rodents allow one to calculate the thickness of active layer in dry habitats as 60-80 cm. Three relatively fertile epigenetic paleosols have been identified by Gubin in the late MIS 3 interval, the uppermost paleosol being 28-26 ^{14}C ka old. These soils with remains of sedges, grasses and mosses formed under the same cryoxeric conditions but with reduced influx of aeolian material. In contrast, the MIS-2 sediments with grass remains and very high rate of dust accumulation contain no visible soils (Gubin et al., 2008).

The richest concentrations of various organics, including mammal bones, are found together with artifacts at Paleolithic sites B, MK, PS and Ya (Fig. 1, 14 and 15).

Other indications of paleoclimate

Extremely arid and cold climates with an antarctic level of annual precipitation ~ 50 mm had to be assumed for the eastern periglacial zone to balance the glaciological models of the mapped configuration of the Late Weichselian Barents-Kara ice sheet (Svendsen et al., 2004).

Winter temperatures

In the radiocarbon 40–20 ka interval the treeless landscapes in all Arctic Siberia from the Yamal Peninsula (Vasilchuk, 1992) to the shores of the Laptev Sea (Derevyagin et al., 1999; Hubberten et al., 2004) experienced mean winter temperatures by 6-9°C lower than present, as indicated by the oxygen isotope ratio of relict ice wedges. Slightly warmer (but colder than present by 3-5°C) winters are isotopically recorded for the late glacial time ~14–11 ^{14}C ka BP in the northern Gydan Peninsula (Vasilchuk, 1992). A similar relative `warming` is inferred from isotope oxygen values of ice wedges of the south-western shores of the Laptev Sea for 13-14 ^{14}C ka BP with a change to colder winters ~ 11-12 ka ^{14}C BP (Dereviagin et al., 2010). The latter author also notes slightly warmer winters for the 39–35 ^{14}C ka BP interval.

Winter temperature estimated by oxygen isotope ratios derived from ice wedges are not very helpful in subdividing the succession of past environments because winters have always been cold in Siberia and still are. The numerous well-preserved carcasses of frozen mammoths of various ages (from >50 to 10 ^{14}C ka BP, crosses in Fig. 1) indicate that the old permafrost, which was growing during the Pleistocene, has not experienced serious warmings during the last 50 ka. The absence of significant climate and environmental changes in the former mammoth steppe of the Siberian Arctic is strongly supported by a recent integrated study of sedimentary DNA, pollen and macrofossil data obtained in a number of silty and organic-rich sequences around Lake Taimyr on 75° N (CS in Fig. 1). The reported results `imply that the vegetation cover in the interior of the Taimyr Peninsula has remained fairly stable during the Late Pleistocene from c. 46 to 12.5 cal kyr BP` (Jørgensen et al., 2012, p. 2000).

Summer temperatures

Variations within the range of very low winter temperatures do not bear significant ecological consequences. What matters for biota is the temperature level and precipitation of summertime, i.e. of the vegetation period. However, the palynological estimates, often biased by redeposited pollen, are too general because vegetation signals in pollen spectra are moderated over great expanses, especially given the windy Pleistocene environments. Besides, the pollen record shows little variation through the periglacial Weichselian interval (Sher et al., 2005) as follows from monotonous spectra recorded in sections Ma and BP (Fig. 1), on Bol. Lyakhovsky island (Andreev et al., 2011) and on the Yamal Peninsula (Andreev et al., 2006).

The temperature and precipitation of Weichselian summers are more accurately estimated from quite different and better localized proxies: i) plant macrofossils, ii) insect assemblages whose breeding is strongly constrained by summer air and soil temperatures, and iii) paleosols. Plant macrofossils analysed in Cape Sabler section (CS in Fig. 1) for the Middle and Late Weichselian time spans suggest `distinctly increased summer temperatures and decreased annual mean precipitation. These environments seem to have been more favourable for the terrestrial life of the Arctic than today, as indicated by higher species diversity` (Kienast, 2001, p. 280). The LGM thermal depression is only indicated by the disappearance of hydrophytes and boreal plants, i.e. by increasing aridity. The remains of southern plants and steppe insects demonstrate that the north Siberian summers were warmer than at present not only in Mid-Weichselian but also during the most part of the Late Weichselian cold stage. Late Weichselian vegetation of the Laptev Sea coast appears similar to present cryoxeric communities some 300 km to the south where the mean January temperature is lower by 11°C but the summer mean temperature is by 7°C higher than at Bykovsky Peninsula (Kienast et al., 2005).

Discussion

Origin of the subaerial formation

Origin of the postglacial fine-grained cover formation is crucial for environmental interpretations. Most problematic are universally distributed loess-like silts whose role varies regionally. In the vicinity of the Late Weichselian ice sheet coarser facies such as dune sands and coversand predominate while the volume of the silty component increases eastwards.

There are two opposing concepts explaining deposition of the surficial fine-grained formation of the periglacial zone. The local concept in its extreme form visualizes loess-like silts as cryogenic weathering crust (Konishchev, 1981) which is hard to reconcile with the common occurrence of loess-like mantle on sandy plains. More popular is the idea of local water bodies: rivers, lakes or temporary streams responsible for silt deposition, also professed mostly by geocryologists. The alluvial version of the local concept, especially popular in former works by Russian geocryologists (e.g. Popov, 1962b) was supported by some geologists (e.g. Lavrushin, 1963). A modern version of the local concept ascribes origin of the yedoma sediments to meltwater streams transporting cryogenic silt downslope from local perennial snowfields (Schirrmeister et al., 2008, 2011).

However, the majority of geologists, including this author, prefer a wider concept which attributes the mountains of Pleistocene silt accumulated worldwide to cryogenic weathering as a source and wind as a means of transportation (Kriger, 1965). This concept explains deposition of massive silt anywhere on dry land devoid of dense vegetation independently of local topography. Climatically conditioned pedogenesis would predetermine the final lithological appearance (facies) of wind-blown dust: it could become carbonate loess in southern steppes or loess-like silt with abundant ground ice and cryoarid paleosols in the arctic treeless plains.

The draping occurrence of the pre-Holocene subaerial formation of northern Russia on all topographic elements including low coastlands is a signature of the airborn mode of dust transportation. A typical example of the Siberian subaerial association can be observed in the lowest central part of the West Siberian Plain where some authors visualized the bottom of a huge proglacial fresh-water sea dammed by the Late Weichselian ice sheet (Arkhipov, 1998; Grosswald, 1998). No deep-water sediments have been encountered in the sections along the River Ob (paper 14 in Ch. III). Instead a thick silty mantle with local accumulations of sand is observable on all topographic elements from fluvial terraces to morainic uplands (Fig. 7 in paper 6, Ch. II).

The preferable position of this mantle on the southern and eastern lee sides of large river valleys is a positive sign of airborn transportation (paper 6, Ch. II). The sediments are unfrozen loess-like silts with paleosols, mammoth bones, multi-floor ice-wedge casts and thick lenses of sinkhole rhythmites overlying the Late Pleistocene sandy alluvium of the Ob valley (paper 14 in Ch. III).). Local elevations of silt surface above the limnic lenses testify to relief inversion in the course of permafrost degradation as modelled in Fig. 6A.

Within the area of the Weichselian glaciation the best studied section of the subaerial formation, which links West and East Siberian yedomas, is located at Cape Sabler on the Taimyr Peninsula (Fig. 13, CS in Fig. 1).

This well-dated silty sequence was initially interpreted as a lacustrine formation (Kind & Leonov, 1982; Möller et al., 1999). However, its patently subaerial nature soon became evident from the narrow range of the grain size, multi-floor ice wedges, terrestrial peat beds, mammoth bones and remains of xeric insects. Important evidence supporting a subaerial origin is given by assemblages of plant macrofossils with predominance of xeric species over hydrophilous plants (Kienast et al., 2001). As a result, Möller et al. (2011, p. 378) changed the conclusion stating that the `fine sand and silt, rich in organic detritus and also thick units of silt-soaked peat, was deposited in a terrestrial setting with peat bog and aeolian deposition, interrupted by occasional lacustrine floods`. Peat layers associated with thick syngenetic ice wedges probably reflect more humid episodes in the same permafrost environment (Derevyagin et al., 1999).

The subaerial, largely aeolian origin of the silty Yedoma Formation in the periglacial area of East Siberia was originally demonstrated by Tomirdiaro (1980) from the bluffs of the Laptev Sea coast and strongly supported by Péwé and Journaux (1983) in their study of cold loess of central Yakutia. From a sedimentological point of view the most clear signatures of predominantly aeolian transportation of loess-like silts in all northern environments are :i) the monotonous grain-size, mostly coarse silt and fine sand particles without clay and gravel products of aqueous sorting, ii) the admixture of mineral grains alien to the local rocks, iii) the position on various topographic elements, iv) leeward distribution along major river valleys (paper 6, Ch. II).

The draping occurrence is not easily recognizable in the lowlands but evident enough in nearby uplands. Thus, Late Pleistocene icy loess in association with aeolian sand and ventifact horizons were mapped in the Lena valley along the Arctic Circle as climbing the slopes of the Central Siberian Upland up to 314 m a.s.l. (Kolpakov, 1983).

The detailed study of the mineralogical composition of coarse silt grains by Tomirdiaro and Chornyenky (1987) reveals the dominance of rock fragments with subordinate amount of quartz and small portion of heavy minerals. The most characteristic fact in the northernmost yedomas is the abundance of volcanic rock fragments, including ash particles, whereas no such rocks or volcanoes are known in the adjacent uplands. The only feasible explanation is a long-distance transportation of the dust in the upper atmospheric layers (*ibid.).* The permanent aeolian influx of mineral grains is evidenced by microscopic analysis of cryoarid soils which are unalienable part of the Yedoma Formation (Gubin et al., 2008). The originally aeolian dust is transformed by permafrost, slope creep, ephemeral water bodies and turned into loess-like silt by arctic pedogenesis.

Admittedly, sedimentological analysis is not easy in Arctic East Siberia where ice bands and wedges prevail in many sections and obscure the structure of the mineral matter. That is the reason why some authors prefer to refrain from genetic labels and allude to the yedoma silts as `polygenetic` (Sher et al., 2005). This vague term is not very explanatory and would be better replaced by the more definite `subaerial association` (Volkov, 1971) or `subaerial syncryogenic association` coined by Romanovsky (1993). The latter notion implies accumulation of mineral materials on growing permafrost by various agencies without the need for permanent water bodies. This means that shallow ponds, creeks, slope processes, etc. could contribute to pedogenetic transformation of the aerially transported dust. The sedimentological indications of subaerial genesis of the periglacial mantle are fully supported by the overwhelming paleontological evidence of high aridity which is not compatible with the aqueous hypotheses (Tomirdiaro, 1980; Kienast et al., 2001, 2005; Sher et al., 2005).

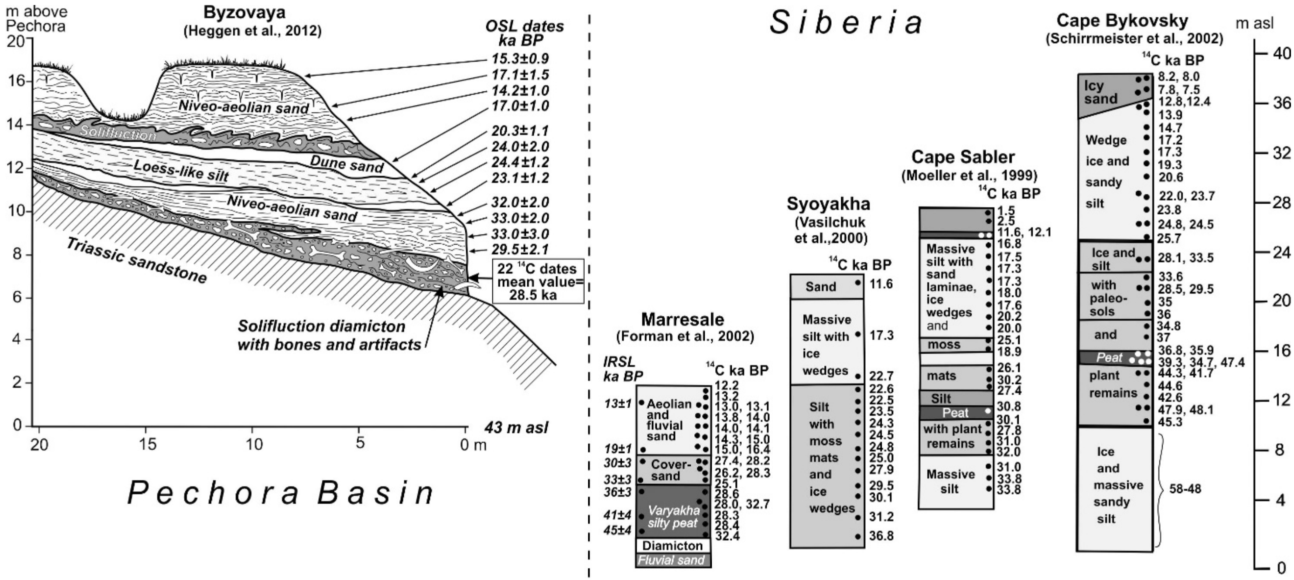

Fig. 14. Comparison of best dated postglacial sequences of unfrozen European Russia (B in Fig. 1) with perennially frozen sediments in glaciated (Ms, Sy and CS) and periglacial Siberia (BP). Key sections in Fig. 1: B – Byzovaya Paleolithic site, 65°01´N /57°23´ E; Ms – Marresale, 69°43´N /66°49´ E, Sy – Syoyaha, 70°10´N /72°34´ E; CS – Cape Sabler, 74°33´N /100°32´ E; BP – Bykovsky Peninsula, 71°47´ N/129°25´ E. Radiocarbon ages are shown by Roman print, OSL dates, ka BP, are italicised.

Geochronological implications

The above described chronometric data are sufficient for correcting the chronology of recent geological events and their correlation with the European record. Principally, the long well-dated sedimentary succession posterior to the last regional glaciation supports the QUEEN conclusion about the relative antiquity of the last ice advances onto the arctic lowlands from the shelf ice domes. The well-studied sections of fine-grained postglacial sediments (Fig. 2, 13, 14) and frozen mammoth carcasses (Fig. 1) reliably testify that the last glaciation of the arctic plains occurred prior to 40 ^{14}C ka BP.

Karginsky problem

The lack of reliable evidence for temperate environments and high sea levels in the pre-Holocene time is an important fact (see 3.1). E. g., the shore staircase along the Barents and Kara seas is presented either by Holocene terraces (Troitsky & Kulakov, 1976) or by eroded Eemian marine formations (Astakhov et al., 1999) whereas cold-water marine muds 42 to 24 ^{14}C ka old are found at 50-80 below sea level (Fig. 2). However, in conventional stratigraphic schemes the Karginsky warm event is referred to as a marker of MIS 3 interval (Isayeva et al., 1986; Volkova & Babushkin, 2000). It was originally defined and correlated with the Middle Pleniglacial in two sections of subtill interglacial marine and alluvial sediments with finite radiocarbon dates on the Yenissei (Kind, 1974).

Afterwards various workers have attached the `Karginsky` label to very different climatic events with the only justification in radiocarbon dates. Much effort was spent seeking to reconcile warm with cold episodes from sequences with similar finite radiocarbon dates. The dates from `warm` intervals eventually proved to be too young. Therefore, the Karginsky interglacial sediments with luminescence dates older than 100 ka were correlated to the Eemian (papers 13 and 16, Ch. III). Judging by the paleoclimatic indications of a climate warmer than the present one, such as forest pollen and arctoboreal mollusks, the Karginsky interval should be compared with MIS 5, although the chronometric scatter from 160 to 100 ka is too wide for more exact correlation.

The overlying glacial complex everywhere in Arctic Russia was dated to the interval of 90–60 calendar ka BP (Svendsen et al., 2004). Arguments for a Late Weichselian age of the last ice advance (Arkhipov, 1998; Lavrov & Potapenko, 2005) are refuted by the postglacial sites of the old Paleolithic in European Russia and frozen mammoth carcasses in Siberia scattered above the uppermost glacial complex (Fig. 1, 14).

An attempt to revive a Karginsky transgression of MIS 3 level was made on the Taimyr Peninsula based on two finite AMS dates obtained on foraminifers in two far distanced sections: 31 ka at 57 m a.s.l. and 39 ka BP at 12 m a.s.l. (Gusskov et al., 2008). However, the continuous and paleontologically studied terrestrial silts exposed close to sea level with far more numerous ^{14}C dates (CS, Ma and BP in Fig. 1) do not show any warm or marine events within the entire radiocarbon range (Möller et al., 1999; Schirrmeister et al., 2002, 2008; Sher et al., 2005; Astakhov & Nazarov, 2010). Therefore, the reported two dates are probably too young which is no wonder because they are derived from foraminifers collected in the 1960-s and subaerially exposed since. Marine organics found far above sea level normally originate from the redeposited Arctic Eemian.

Disregard for these observations can lead to erroneous correlations as evident from the confusion in the stratigraphic terminology. For example, the Kiryas sequence with forest-tundra pollen in the present subzone of middle taiga (LKM–AR in Fig. 1) and similar sequences within the radiocarbon range 40–27 ka are related by some authors to a `Karginsky regiostage` (Laukhin et al., 2011). In the view of the data discussed above, the Kiryas peat with indications of tundra vegetation on 61° N cannot have a relation to the Karginsky interglacial strata with forest pollen found on 69.5° N. Some permafrost investigators call Karginsky even the cold strata of the Ice Complex lying well above the Eemian level (Schirrmeister et al., 2002, 2008).

The Karginsky correlation could be correct if the cited authors did not attach the MIS 3 label to this warm interval. According to the mass of modern chronometric evidence this epoch of relatively more humid and temperate climate with trees growing close to the modern shoreline can only be correlated with the Eemian in spite of finite radiocarbon dates reported from this stratigraphic level (papers 13 and 16 in Ch. III). References to finite radiocarbon dates as the signature of the `Karginsky interval` is indefensible because the same range of radiocarbon ages occurs anyplace globally and cannot bear any regional geographic name. The best dated sediments of MIS 3 interval both in European Russia and in Siberia nowhere show a climate resembling the Holocene. On the contrary, traces of former permafrost in presently thawed sediments with `old` radiocarbon dates (Astakhov, 2006; Astakhov & Svendsen, 2011) and periglacial character of alluvium of the Second Terrace (see 3.2) together with paleontological data firmly reject any significant warmings within the last 40 ka of the Pleistocene history.

Main stratigraphic marker

Nowadays a better correlation is provided by the excavated and well dated Paleolithic strata which can be used as a stratigraphic marker for the postglacial Pleistocene of the northern Russian mainland. It allows to chronologically pinpoint the episode of a milder climate indicating penetration of early humans into the Arctic and Subartctic within the generally harsh postglacial environments. The obvious chronometric counterparts are Byzovaya (B in Fig. 1, 14) in Europe and Yana (Ya in Fig 1, 15) in East Siberia. The Byzovaya cultural layer is found in solifluction diamicton with artefacts and hundreds of bones over the thalweg of an ancient ravine in present-day unfrozen boreal forest. The mean age of mammoth bones from the cultural layer is 28.5 ka ^{14}C BP (~ 32-34 calendar ka) calculated from 22 dates and supported by OSL ages from 33 to 14 cal ka BP from the overlying subaerial cover (Fig. 15) (Mangerud et al., 1999; Heggen et al., 2012).

World's northernmost Paleolithic site Yana has 39 radiocarbon dates obtained both from the cultural layer and organics of the surrounding sediments. The cultural layer within the loess-like sequence locally called `II Terrace` and excavated at four different spots has yielded radiocarbon age within the interval of 28.5 to 27 ^{14}C ka BP (Pitulko & Pavlova, 2010). The mean value of seven ^{14}C ages directly from the cultural layer is 27.9 ^{14}C ka (Fig. 15). Thus, these two sites, although within very narrow sedimentological and chronometrical interval, provide an east-west correlation level across the 2,600 km stretch of the Eurasian mainland. Most probably both episodes belong to an interstadial event of a transcontinental significance.

This correlation level is a part of a wider interval of relatively mild periglacial environments recorded in loessic, limnic and alluvial formations. Alluvium of the Second Terrace of the Pechora Lowland (MK and KT in Fig. 1, Fig. 2) containing abundant megafauna bones and oldest artifacts is radiocarbon dated within the range of 37 to 24 ka BP but the base of this succession is not exposed. The Yenissei alluvium broadens the radiocarbon interval up to 47 ka BP (Fig. 2).

Unfortunately, in the glaciated area a full section reflecting the complete series of sedimentary events from the end of the last ice advance to the start of LGM cooling has not been found yet. The Byzovaya cultural layer has for a long time been used in the formally adopted stratigraphic scheme of the northeastern Europe as a representative of the main Late Pleistocene interstadial of the same name (Guslitser et al., 1986).

A longer series of chronometric ages and especially the pollen diagram reflecting typical periglacial vegetation is obtained from the section of

Mamontovaya Kurya (MK in Fig. 1, 2, 14) (Halvorsen, 2000) which can serve as a parastratotype of the Byzovaya interstadial. Coarse channel alluvium with artifacts at the base of the terrace of Mamontovaya Kurya deposited between 50 and 38 OSL ka BP contain mammoth bones with^{14}C dates from to 37 to 32 ka B. The overlying fluvial sands have yielded a successive series of 8 ^{14}C dates in the range 31 to 24 ka BP. The coversand and silt ~ 10 m thick have provided OSL ages from 20 to 14 ka BP (Svendsen et al., 2010). Similar dates (32 to 17 ka) are recovered from the alluvial sand of river Adzva (Ad in Fig. 1) overlain by aeolian sand dated to 18–15 ka BP (Astakhov & Svendsen, 2011).

However, the conventional chronostratigraphy has for a long time erroneously placed the Byzovaya interstadial before the regional Polar Glaciation correlated with the Late Weichselian. In the 1990-s it became clear that the last glaciation with its two ice advances along the lines Harbei+Laya-Adzva and Markhida predated the Byzovaya interstadial, and the overlying sediments of the Polar Stage have no traces of ice advances on the mainland (Mangerud et al., 1999). Signatures of the Late Weichselian ice sheet were found only on the Barents Sea floor and Arctic islands (Svendsen et al., 2004).

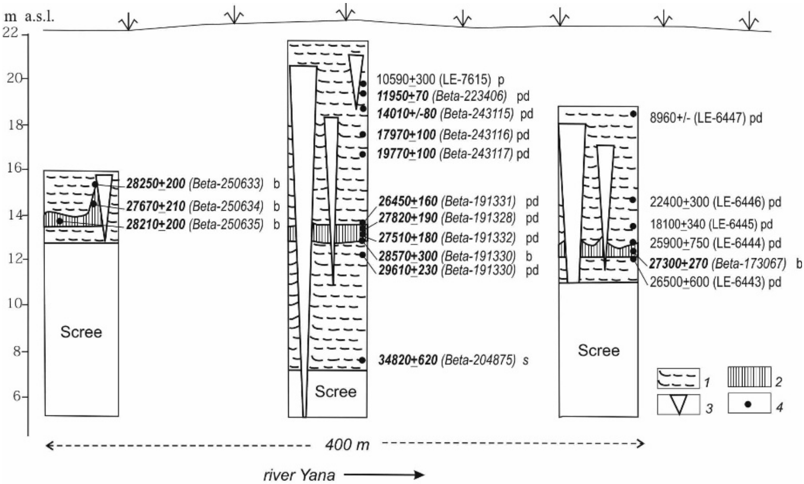

Fig. 15. Radiocarbon dated northernmost Paleolithic site on river Yana (Ya in Fig. 1) (adapted from Pitulko & Pavlova, 2010). Symbols: 1 – pale-brown silt with ice bands and fine sand laminae; 2 – silt with organic remains and artefacts; 3 – ice wedge; 4 – radiocarbon date. Bold italics – age by AMS, years BP, regular print – age by conventional radiocarbon method, years BP. Dated materials: b – bone collagen, p – peat, pd – plant remains, s – soil.

The best dated reference section of the periglacial zone is the yedoma at Bykovsky Peninsula (BP in Fig. 1, 14) where, besides numerous organic remains, more than 90 ^{14}C dates ranging from 58 ka to Holocene values have been obtained. This rich sequence ~ 40 m thick directly above sea level has enough radiometric and paleontological data to trace main environmental changes through the second half of the Late Pleistocene. The ^{14}C AMS ages show a linear distribution against the height of the section and therefore represent the best chronological control for arctic terrestrial sequences (Schirrmeister et al., 2002; Kienast et al., 2005).

The main substages of this record suggested by Sher et al. (2005) are Middle Weichselian I:`warm` tundra-steppe – 47 to 34 ^{14}C ka BP and Middle Weichselian II: `cool` tundra-steppe – 34 to 24 ^{14}C ka. In their diagram of insect assemblages an additional peak of arctic species approximately between 34 and 31 ^{14}C ka seems to separate two substages of relatively warmer summers. The early substage is also indicated by a peat layer with ^{14}C dates of 39 to 35 ka. In the Cape Sabler yedoma a thick peat 30-31 ^{14}C ka old (Fig. 13) covered by silts with moss mats probably belongs to the later substage which in this section is marked by a set of radiocarbon dates from 31 to 25 ka.

Both Paleolithic layers of Byzovaya and Yana coincide chronometrically with the level of minimum arctic insects in the diagram by Sher et al. (2005). It is noteworthy that two minor peaks of arboreal pollen in the Mamontovaya Kurya diagram radiocarbon dated to 37 and 27 ^{14}C ka roughly coincide with the substages of insect fauna MW I and MW II correspondingly on the Bykovsky Peninsula (see paragraph 6.3). These chronological levels were probably peaks of biological activity in the Russian North including vegetation, animals and humans.

There is a pronounced maximum of arctic beetles ~ 34-30 ^{14}C ka which appears to be a colder interval between two optima of the major interstadial in the Bykovsky sequence (Fig. 8 in Sher et al., 2005). This subdivison superficially resembles the tripartite Karginsky scheme by Kind (1974) with Konoshchelye cooling at 33-30 ^{14}C ka BP, if one forgets that the Kind`s scheme was derived from patently interglacial formations containing forest flora and not from ice-rich postglacial formations parental to the modern radiocarbon dates.

Thus, the postglacial sedimentary record of northern Siberia cannot be accommodated into the conventional chronostratigraphic subdivisions named Karginsky and Sartan on the Yenissei (Kind, 1974; Volkova & Babushkin, 2000) because the sedimentary formations of these names are obviously older than the subaerial formation discussed above. New type sections representing MIS 3 and MIS 2 intervals are available on the Yamal Peninsula.

These are the Varyaha interstadial silty peats of small sinkholes sandwiched between the Kara diamicton with fossil glacial ice and the Oleny coversand with ice wedges at Marresale (Ms in Fig. 1). This unit 4-5 m thick, close to sea level, covers an interval of 32 to 28 ^{14}C ka (6 dates) with 3 IRSL ages of 45 to 36 calendar ka BP (Forman et al., 2002). The key position of the Varyaha unit is supported by pollen analysis which indicates dry treeless tundra with more shrubs upwards in the succession. The overlying sands contain mostly redeposited pollen (Andreev et al., 2006). In Arctic West Siberia most of famous mammoth carcasses chronometrically belong to the same interval from 42 to 25 ^{14}C ka BP (Fig. 1).

The Marrresale sequence does not embrace the initial events of the interstadial but 6 AMS dates in the range 32 to 28 ka BP firmly correlate it with the archeological marker Byzovaya-Yana and the peat layer in the Sabler section (Fig. 13), the lower peaty beds of the Syoyaha sequence with ^{14}C dates of 37 to 28 ka (Fig. 14) and cold-sea muds on the Barents Sea shallows (Fig. 2) This stratigraphic marker is probably correlative to the Briansk Soil of Central Russia and Denekamp interstadial of NW Europe.

Another transregional marker at the start of the Holocene humidification is paleobotanically well expressed and was discussed elsewhere (Velichko et al., 1997; Andreev & Tarasov, 2007).

Paleoenvironmental trend

Using multi-proxy approach, Sher et al. (2005) suggest a division of the periglacial Middle Weichselian into two substages and infer uni-directional trend from warmer to cooler summers from 47 to 24 ^{14}C ka BP without major fluctuations in vegetation and climate. The stability of the grass/herb-dominated environment with aridity persisting through all MIS 3 and MIS 2 stages of northern Siberia based on recent integrated research led them to the conclusion that `Within this new paradigm, the LGM environment was just an impoverished variant of the MIS 3 tundra-steppe` (Sher et al., 2005, p. 564). They date this stage of extreme continentality to 24–15 ^{14}C ka BP (appr. 30 to 17 calendar ka) which roughly corresponds to the interval between two relatively humid events in the glaciated regions (Fig. 16).

The periglacial conditions of the second half of the Late Pleistocene dominated not only the Arctic but all Siberian plains as well. This can be inferred from the well-studied Lipovka alluvial section on river Tobol, 58° N/67° E, where larch stumps *in situ* yielded several dates ~ 30-31 ^{14}C ka BP. At that time the present-day southern taiga was replaced by periglacial forest-tundra with large syngenetic ice wedges which grew 1000 km south of their present occurrence (Kaplyanskaya & Tarnogradsky, 1974).

The harsh Siberian climate of the last glacial cycle after 50 ka BP was increasing (with minor humid episodes) to the extreme continentality caused by the then emerged Arctic shelf. The cold winters vs warm summers can be explained by anticyclonic air circulation with cloudless skies all through the year. The Paleoarctic anticyclone with its northern and northeastern winds, inferred from volcanic rock fragments of the northernmost Yedoma silts, is also reflected in sand dune orientation (Tomirdiaro & Chornyenky, 1987).

The abundant fauna remains in Siberia indicate that high aridity rather than low temperatures prevented the postglacial landscapes from being forested during the LGM. The temperature deviations – winters by 7-8°C colder than today and summers by 6-8°C warmer – mean that the Late Pleistocene continentality index (January-July average temperature difference) for the Laptev Sea coasts was about 55-60°C, i.e. by 17-23°C higher than the present continentality index which is 38°C for the Lena delta. For comparison, the Pleniglacial continentality in NW Europe was 28–33°C, decreasing to 20°C after 20 ka BP during the arid final Pleniglacial (Huijzer & Vandenberghe, 1998).

This environmental continuity is less spectacular in European Russia in the proximity of the Fennoscandian glaciation. The difference between interstadial and more arid intervals seems to be the most pronounced in European Russia (Fig. 16). The treeless MIS 3 environment on the Pechora was relatively humid, judging by the periglacial alluvium of the Second Terrace and rich remains of megafauna. In contrast, ubiquitous aeolian sediments and absence of mammal bones indicate extremely arid environments of polar desert during MIS 2 (Astakhov & Svendsen, 2011).

The Late Pleistocene continentality, which was twice higher in Siberia than in periglacial Western Europe, is amply recorded by the icy yedoma-type formations, a product of old and very stable permafrost (Romanovsky, 1993). However, the cloudless summers, maybe not that hot as in Central Siberia, were felt even in the western Russian Arctic. The most peculiar feature of the yedoma on the Bykovsky Peninsula (BP in Fig. 1) is the maximum of steppe insects accompanied by increased number of mammal remains in the final Pleistocene starting with 15 ^{14}C ka BP (Sher et al., 2005).

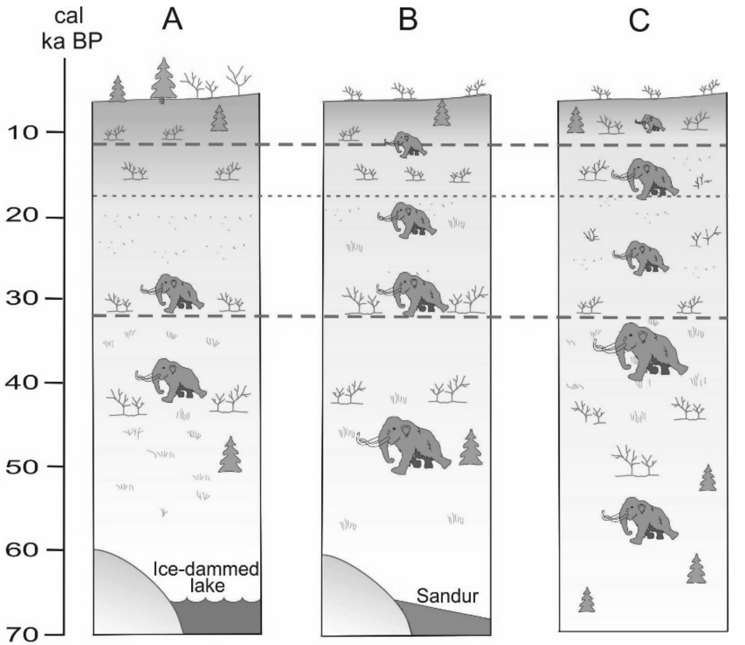

Fig. 16. Sketch of evolution of postglacial environments in northeastern European Russia (A), northern West Siberia (B) and northern East Siberia (C). Density of yellow colours depicts increasing aridity of periglacial landscapes; greenish shades designate relatively mesic environments. Blue lines are best dated correlation levels. See centrefold for this image in colour.

The warm peak ~ 15 ^{14}C ka BP of the Bykovsky Peninsula succession has interesting counterparts in the west. There are two conventional radiocarbon dates 15,310 ± 650 years (LU-1188) and 15,120 ± 120 years (LU-1446) from thin peat seams reported from both islands of Novaya Zemlya at the latitudes of 72° (GZ in Fig. 1) and 73°40′ N (MB in Fig. 1) correspondingly (Serebryanny & Malyasova, 1998). Today the cold summers of the western High Arctic do not permit peat accumulation sufficient for bulk radiocarbon sampling.

In contrast, the Novaya Zemlya late glacial spectra contain pollen not only of periglacial but also thermophilous plants such as *Rubus chamaemorus* indicating dwarf willow-herbaceous communities *(ibid.)*. Thus, the dates, if correct, suggest that late glacial summers must have been warmer than the present ones even as far west as the Barents Sea islands. In addition, the Gusinaya Zemlya peat (GZ) lies ~2 m above sea level which disagrees with the idea of Holocene high marine limits and therefore with Late Weichselian ice on 72°N.

Scattered indications of the late glacial climate amelioration are known also on the glaciated mainland. For example, Kind (1974) reported 15.3 ka radiocarbon date from the narrow fluvial incision in the uppermost glacial complex of the Lower Yenissei (see Fig. 4 in paper 16, Ch. III). A muskox skull dated to 14 radicarbon ka is the first bone found on the Lower Pechora after the sterile 24–14 interval (Mangerud et al., 1999). Similar scattered dates are known from thick soils in southern West Siberia (Hubberten et al., 2004).

Ice wedges 14–11 ^{14}C ka BP old on the Gydan Peninsula yielded $\delta^{18}O$ values from 22.6% to 19.9%, suggesting milder winters, but still 3–5°C colder than the present ones (Vasilchuk, 1992). A climate amelioration is indicated by the First Terrace alluvium and sediments of resumed thermokarst sink ponds dated to 13–10 ^{14}C ka and to 16.8–11.7 OSL ka in the Pechora Lowland (Astakhov & Svendsen, 2011) and by similar features in Central Siberia and Beringia (Sher et al., 2005). It seems premature to correlate these scantily dated events with the European environmental stages of Bölling and Allerød as some palynologists are too eager to do. However, a start of climate amelioration close to 15 ^{14}C ka BP is suggested between two best dated markers of the postglacial Pleistocene: Byzovaya-Varyaha-Yana interstadial and the base of the Holocene humid event (Fig. 16).

Conclusions

- The Pleistocene terrestrial history following the last regional glaciation of the Russian mainland is at least 40 ka long which is an independent evidence of relative antiquity of the last ice sheet of the northern plains (Fig. 16).
- The postglacial Pleistocene record indicates that the perennially frozen Eurasian landmass grew colder and drier reaching the extreme continentality ~ 30 to 17 calendar ka BP. Minor climatic changes in Siberia did not disturb the historically stable environments beneficial for the rich periglacial biota.
- In Siberia mammoth steppe survived even during the coldest and driest Late Zyryanka time. Precipitation was very low but due to the warm summers the active layer was wet enough to support steppe-tundra grasslands and forage for large grazers.

- The amplitude of Late Pleistocene climate change was larger in the Russian European northeast where abiotic polar desert formed in the lee of the Barents Ice Sheet.
- Interglacial or similar periods of milder climates and high sea levels have not been proven for the last 40 ka of the pre-Holocene history of the Russian North. The increasing continentality of the postglacial climate parallels the Pleniglacial history of western Europe.
- Indications of interglacial rise of sea level and arboreal vegetation found in northern sedimentary formations, such as the Karginsky warm stage, suggest their correlation with the Eemian and not with the Middle Pleniglacial or MIS 3.

Acknowledgements

This work was stipulated by the Research Plan of St. Petersburg University and supported by the research grant № 11-05-00624-a from the Russian Foundation for Basic Research. It is the final contribution to the European Coordination Program APEX.

24. Stratigraphic Correlation of Late Pleistocene across Glaciated Northern Russia

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Proceedings of International Conference INQUA-SEQS 2014 `The Quaternary of the Urals: Global Trends and Pan-European Quaternary Records. Ekaterinburg, URFU, Russia, 9–12.

Synchronization of Pleistocene events across northern Russia has long been problematic due to the size of the dry land extending across 70° of longitude. Paleontological correlation fails to achieve reliable results because of the environmental gradient aggravated by the orographic obstacle of the Urals. The last factor is crucial for the paleobotanical comparisons which are hindered by the absence of broad-leaved trees beyond the Urals. The latest efforts are largely based on cartography and geochronometry. The cartographic correlation is performed by tracing major ice-marginal formations from the Baltic Sea to Central Siberia (paper 25, Ch. V).

However, paleoclimatic landmarks of distant terrains can only be correlated by geochronometry. The international studies of the last two decades have provided a multitude of dates by diverse techniques permitting to obtain correlation signals by a statistical approach irrespective of the validity of each single date. For this the population of dates is organized in clusters spatially linked with major sedimentary associations with distinct paleoclimatic signatures, thus composing mutually supportive climatostratigraphic and geochronometric levels. Since the number and quality of chronometric data decrease downwards in the succession the traditional approach of starting from the oldest formations is hardly fruitful. I.e. the lowermost exposed interglacial is conventionally correlated by its paleoclimatic characteristics with the Holsteinian. This correlation is tentatively supported in West Siberia by several dubious TL and ESR dates (paper 22, Ch. V) but in adjacent regions where no dates are available it is no more than a conjecture.

1). Therefore, as a first step, the most reliable chronostratigraphic marker should be considered. In the Russian North this is the sedimentary complex of the uppermost postglacial Pleistocene best studied by sedimentology and supplied with numerous radiocarbon dates. This complex is largely a subaerial cover consisting of loess-like silts and aeolian sands with intervening lenses of fluvial and lacustrine sediments. The paleoenvironmental signature of the complex is unambiguous: the sedimentary structures and organic remains reflect dominance of perennially frozen tundra-steppe with intervening cool interstadials in climates drier and frostier than today at lower sea levels (e.g. Hubberten et al., 2004).

Hundreds of available dates allow me to attach this typically periglacial association to the time span of 60 to 11 ka BP everywhere from the Taimyr Peninsula in Central Siberia to the Timan Ridge in European Russia. This chronological interval is firmly based on robust estimates by parallel series of AMS radiocarbon and OSL dates supported by finds of frozen mammoth carcasses and artifacts of the early Upper Paleolithic (paper 23, Ch. V). This correlation is validated by less reliable but numerous OSL dates within the range of 100 to 60 ka derived from the underlying glaciofluvial and lacustrine deposits of the last regional glaciation (Svendsen et al., 2004).

2). The next downward correlation level is represented by the interglacial marker strata: marine, fluvial and palustrine deposits which are found in the subtill position above the Arctic Circle. Their interglacial nature is evident from warm-water aquatic fauna and arboreal pollen spectra clearly contrasting with treeless continental environments of the overlying periglacial complex. The chronometric correlation has long been ambiguous because the strata with rich organics yield only non-finite or patently incorrect finite radiocarbon dates.

The latest synchronisation efforts are largely based on OSL dates (~75 measured samples) which on this chronological level are naturally more scattered than on the first one. The 60 dates (a couple of outliers being excluded) on exposed interglacial marine sequences are distributed within the range of 155 to 100 calendar years BP. The mean values of the OSL chronometric series are centred on 111-112 ka (river Sula, European Russia and Cape Karginsky, Yenissei Siberia), and on 133-134 ka levels (the tip of the Taz Peninsula, West Siberia) (papers 16, Ch. III and 22, Ch. V). These estimates, although diverging by twenty ka, look sufficient for correlating the last interglacial of northern Russia with the Eemian *s. lato* within the MIS 5 time span. This correlation is supported by ESR dates 100 to 134 ka from marine formations of Yenissei Siberia (Katzenberger & Grün, 1985) and by U/Th dates ~ 130-140 ka on shells in European Russia and thick peat

Table. Correlation of late Quaternary sediments of the Northern Pleistocene of Russia

MIS	Age estim-ate, ka	Measured ages, ka BP	Lower Pechora and European Northeast	Lower Ob and Yamal Peninsula	Lower Yenissei and Taimyr
2 *3*	60-11	^{14}C=50-10; OSL=85-13 IRSL=45-13 TL=110-65	Two riverine terraces, aeolian sands and silts, solifluction sheets and fluvial sands with mammoth bones, Byzovaya artifacts	Syoyaha icy loess-like silts, Varyaha peaty silts, frozen mammoths	Two river terraces, Cape Sabler icy loess-like silts with peat, frozen mammoths
4	100–60	OSL=105-50; TL=80-78	Upper glacial complex (Polar moraines, Lake Komi sediments)	Upper glacial complex: Kara diamictons, Sangompan	Yermakovo, Zyryanka, Sartan moraines, Igarka varves
5	140-100	ESR=135-122 OSL=155-100 U/Th=141-130	***Sula*** *marine strata with boreal mollusks*	***Shur*** *peat, exposed marine strata with boreal fauna*	***Karginsky*** *marine and alluvial strata*
6	150-140	OSL=164-145	Vychegda glacial complex	Salehard moraines and varves	Sanchugovka and Murukta moraines
7	200-150	OSL=195-153 U/Th=240-186 TL=153; 178	***Rodionovo*** *and* ***Seyda*** *peats, marine strata with Cyrtodaria angusta*	*Buried marine strata with Kazantsevo forams*	***Kazantsevo*** *marine strata with Cyrtodaria angusta*

on the River Ob (paper 16, Ch. III). The stratigraphic position of this correlation level is clear from the underlying glacigenic complex of MIS 6 age widespread in the Subarctic where it is known as the northern counterpart of the Saalian (the Vychegda Till in northeastern European Russia, Taz Moraines in West Siberia). The Murukta Till of Central Siberia, previously ascribed to the MIS 4 time span, should be correlated with the Saalian because it is clearly overlain by marine sediments with the warmest (boreal-lusitanic) microfauna (Svendsen et al., 2004; Astakhov, 2013).

3). The chronometrically poorly studied interglacial formations positioned below the second glacigenic complex of the Arctic can be ascribed to a MIS 7 age in the view of the above correlation results. These are exposed marine strata with extinct mollusk *Cyrtodaria angusta* called the Kazantsevo in Siberia and the *Cyrtodaria* Strata in European Russia. Similar formations with the Kazantsevo assemblage of boreal foraminifera were found by boreholes below sea level on the Lower Ob (Arkhipov et al., 1994) and separated by a thick till from surficial peat deposits with dates ~130-140 ka. West of the Urals this correlation level is supported by the thick peats with interglacial floras with U/Th and OSL dates ~ 200 ka BP at Rodionovo and Seyda (paper 22, Ch. V).

The Table above lists major Pleistocene formations of northern Russia correlated from west to east. Interglacial formations are italicized. For the data see also papers 22 and 23, Ch. V and references therein (Astakhov, 2014b).

25. Glaciomorphological Map of the Russian Federation

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QUATERNARY INTERNATIONAL 2016, VOL. 420, P. 4–14

Introduction

VSEGEI, a research subsidiary of the Russian Geological Survey, in 2013 completed a new map of Quaternary formations of the Russian Federation, scale 1:2 500 000. This overview map covers a huge territory of northern Eurasia from Scandinavia to the Bering Strait plus adjacent sea floors. Designed for large-scale decision making and education purposes, similar to the old map of the USSR (Ganeshin, 1976), the new map is for the first time compiled in GIS-format. The compilation was burdened by the long-standing problem of stratigraphic correlation of thick glacigenic formations across the entire Eurasian north. They are basically inorganic whereas sparse radiometric dates illuminate mostly the uppermost sequences. As a primary approach to a solution of the problem we compiled an auxilliary map of the same scale showing spatial distribution of crucial glacigenic features potentially instrumental for inter-regional stratigraphic correlations of surficial glacigenic formations.

This product, called Glaciomorphological Map, depicts topographically expressed glacigenic features, as well as stratigraphically inferred ice limits, minor ice retreat zones, ice streams, large erratics, etc. Here we present a generalised version of this map in which only larger glacial features are retained (Fig. 1). Minor and out-of-scale landforms such as drumlins, ice-walled-lake plains, etc. are omitted for the sake of clarity of the compressed map suitable for a journal publication.

Fig. 1. Glaciomorphological Map of the Russian Federation (continued in next page). Localities referred to in the text: 1 – glaciofluvial plateau upon the Don Till, 2 – Surena glaciofluvial ridge, 3– southernmost exposures of lowland tills in the western Uralian foothills, 4 –Rogovaya ridges, 5 – the highest point of the Valdai Upland, 346 m a.s.l., 6 – Laya-Adzva double ridge, 7 – southernmost sections of lowland till in the Urals, 8 – Yugan megablock of Jurassic sediments in Samarovo Till, 9 – Agan morainic ridge with detached blocks of Paleogene sand, 10 – glaciotectonic parallel ridges in Fig. 4, 11 – morainic ridges up to 130 m high, 12 – Sopkay moraines, 13 – Bol. Tundrovaya inselberg 618 m a.s.l. covered by tills with lowland erratics, 14 – Mokoritto morainic loops, 15 – Norilsk end moraines. See centrefold for this image in colour.

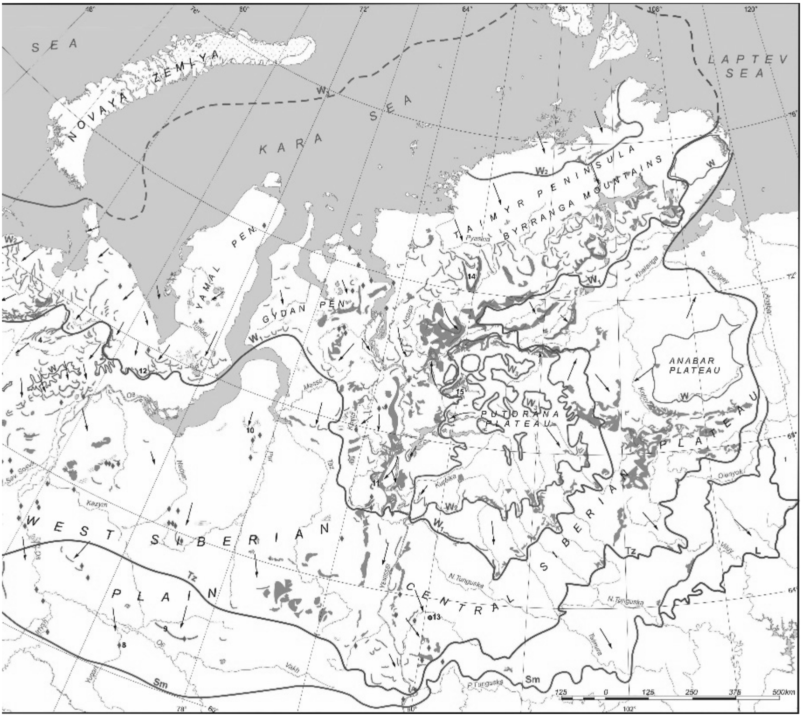

Fig. 1. Continued

For timing ice limits we rely on numerous Russian and foreign publications, particularly on international works of the last two decades with comprehensive geochronometric database, if they are not outright controversial. We do not question the published stratigraphic results but try to assimilate them and bring together conclusions from various terrains wherever it is possible without violating the geological common sense. Our principal aim is to present in concise form the main cartographic evidence of the Eurasian glacial history which can be further used for stratigraphic and paleogeographic considerations on the time span 600 to 11 ka BP. Former mountain glaciers of East and South Siberia are not discussed here because they are far distanced from the main glaciated area and their dubious timing requires special consideration.

State-of-the-art

Previous attempts of small-scale overviews of glacial morphology were of two kinds: i) restricted by a region and age of the mapped glacial features, and ii) comprehensive, i.e. embracing all ice advances and all regions of glaciated Russia. The restricted efforts showed mostly ice margins within the area of the last Scandinavian glaciation of European Russia and adjacent lands (e.g. Zarrina & Krasnov, 1965; Chebotareva, 1977; Kalm, 2012). A special glaciomorphological map was compiled for the area of the Moscow glaciation (Goretsky et al., 1982).

Similar maps derived from results of geological surveys were also created for other regions of the Russian Plain. Some publications considered Siberian glacial features (e.g., Arkhipov et al., 1976, 1986). The most comprehensive overview of ice limits and accompanying glacigenic formations is contained in the old Quaternary Map of the USSR, scale 1:2 500 000 (Ganeshin, 1976), which was based on various regional research and partly on the sheets of SGM-1000/1 – the State Geological Map, scale 1:1000 000, the first generation.

The recent attempt of mapping this huge area in ArcGIS format is represented by the compendium on world glaciations by Ehlers *et al.* (2011) which provides digital maps based on topographic tiles of scale 1:1000 000. This spatially all-embracing product is available in separate layers at http://booksite.elsevier.com/9780444534477/. Unfortunately the composition of these maps is rather erratic. Different tiles vary considerably in degree of generalization: some correspond to scale 1:1000000 (parts of western Europe), others are very schematic. The variety of regional contents is largely dependent on idiosyncrasies of individual authors. E.g. in central European Russia ice limits, key sites and some proglacial lakes are shown, but maps of the last glaciation area also contain subtler features. In arctic Russia and Siberia only ice limits, key sites and major ice-pushed ridges are depicted (Astakhov, 2011), though not in detail demanded by the scale of the map. The principal point is that various maps of this project are not quite compatible and therefore hard to use for interregional correlations.

Content of the map

Our idea of the present work was to produce a map of all regions of inland glaciation in Russia in a uniform manner irrespective of local peculiarities and diverse study levels. The essential tool and a principal database for this approach can be found in Quaternary maps of the Russian

Federation with standardized legends. Such maps are prescribed as mandatory in the thematic packets of State Geological Map, scale 1:1000000, issued by Geological Survey for each quadrangle 6 by 4°. We used 35 such Quaternary sheets of the second generation (SGM-1000/2) published between 1974 and 2004 and 18 digital maps of the 3d generation (SGM-1000/3) compiled and partly published in 2011-2013. These maps, depicting polygons of genetically and chronologically different Quaternary formations, marginal and other glacigenic features, are accompanied by regional profiles. The interregional correlations and timing the ice limits are based mostly on recent stratigraphic overviews (Astakhov, 2004a, 2006, 2013; Demidov et al., 2004, 2006; Svendsen et al., 2004; Larsen et al., 2006; Astakhov et al., 2007; Möller et al., 2011, 2015; Velichko et al., 2011). In cases of discrepancy between the published stratigraphic conclusions, which is normal in poorly studied terrains, we have to rely on remote sensing imagery (e.g. Astakhov et al., 1999; Astakhov, 2004b, 2011, 2013; Kalm, 2012).

The major features of the map (Fig. 1) are the landscape zones and limits of topographically expressed and stratigraphically independent major glacial events: in European Russia D – Don glaciation, commonly associated with MIS (Marine Isotope Stage) 16, alternatively MIS 14 (Fig. 2), Ms – Moscow glaciation, MIS 6 (papers 3, 4 Ch. 1, 22, Ch. V, Astakhov, 2011); in Siberia L – Lebed, probably Elsterian, Sm – Samarovo, MIS 8 and Tz – Taz, MIS 6. The Late Pleistocene glaciation is subdivided in W_1 – early Weichselian, ca 80-90 ka BP, W_2 – middle Weichselian, 50-60 ka BP and W_3 – the Late Weichselian glaciation, the classical Valdai, MIS 2. Note that the Russian Middle Valdai and Siberian Middle Zyryanka climatoliths are traditionally related to the time span of MIS 3 only, i.e. chronologically shorter than the European Mid-Weichselian substage. At this scale the restricted area of the Late Pleistocene glaciation in uplands does not permit a spatial differentiation of several glacial stades of the Weichselian age. The maximum ice limit designated as W is actually drawn along pre-Late Weichselian moraines.

To lighten understanding the glacial zonation of the mapped area we offer a tentative correlation of the Russian glacial events with the chronological units of NW Europe following the general pattern by Cohen and Gibbard (2011) (Fig. 2). The geographical names in the Russian columns designate major paleoclimatic events – glaciations and interglaciations, which are, however, not equivalents of European chronostratigraphic stages. Here we use the European names in purely chronological sense. Actually, stages in terms of the International Stratigraphic Guide are employed in Russia only for the pre-Quaternary and not for the Quaternary record where the chronological volume of the operational units is inferior to typical stages 2 to 10 ma long. In the Russian geological tradition units of lower rank –`links`, similar to pre-Quaternary substages, and `steps`, corresponding to marine isotope stages – are mostly employed for mapping and correlating finely detailed Quaternary subdivisions. On the regional level the material basis of a step, according to the Stratigraphic Code of Russia (Zhamoida, 2006), is a climatolith. Each climatolith comprises rocks deposited during a cold (cryochron) or warm (thermochron) interval. Large paleoclimatic events, such as the Valdai or Moscow glaciation, may chronologically correspond to three climatoliths each.

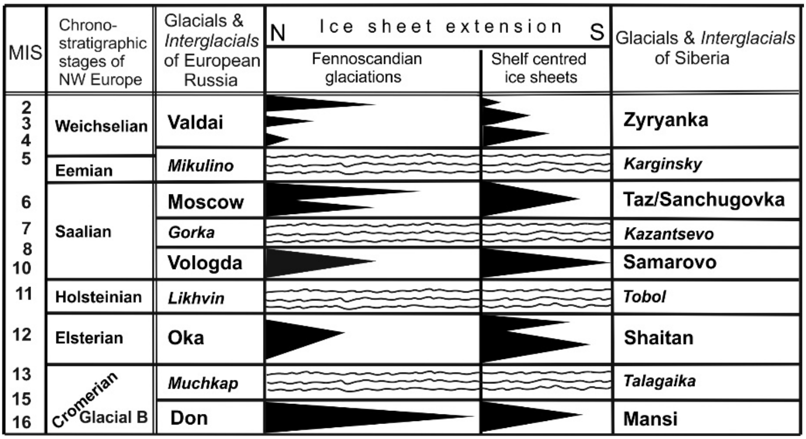

Fig. 2. Correlation of Russian glacial events with the Pleistocene chronology of NWt Europe. Black triangles show relative positions of ice margins, wavy lines are symbols of interglacial environments, also indicated by italics.

Another point in Fig. 2 is the chronological position of the Siberian interglaciations. In formal stratigraphic scales of Siberia the last (Karginsky) interglaciation is traditionally correlated with MIS 3, whereas the penultimate (Kazantsevo) interglaciation is attributed to the time interval of MIS 5. However, the international dating efforts of the last decade proved beyond any reasonable doubt that the Karginsky thermochron (temperate event) at high sea level with atlantic fauna occurred within 160 to 100 luminescence or ESR ka BP (papers 12, 13, 16, Ch. III, 22, Ch. V). Consequently, we have to push both interglaciations by one cycle down in the climate-chronological scale.

Minor ice limits within the area of the last Scandinavian glaciation, irrespective of their stratigraphic significance, are drawn by geomorphic reasoning along sub-parallel marginal belts to mark positions of the retreating ice margin within the time span of 20 to 11 ka BP. Kalm (2012) already tried to connect the Baltic ice-marginal zones with their Russian counterparts. We followed suit with some revisions based on recent geological surveys.

The map also shows selected features relevant to the structure of ice marginal zones. These are hummocky morainic assemblages and kame plateaus, individual large morainic ridges, far-transported sedimentary megablocks, outstanding glacial disturbances, inferred ice flow directions. The original map of 1:2 500 000 scale contains also ice-contact lines, ice stream axes, structural escarpments, eskers, drumlins, large erratics, ice-walled-lakes plateaus, kame plateaus, and separately hummocky moraines in more detail. Such minor features are omitted in the journal version for the sake of clarity demanded by the smaller scale.

Ice marginal zones of European Russia

Don glaciation

The most salient feature is the southernmost limit of the huge Don glaciation reaching 50° N. The Don glacial complex occupies the low Oka-Don Plain which is a southern part of the Russian Plain between the Central Russian and Volga Uplands. This maximum inland glaciation was for several decades considered a counterpart of the Saalian (Drenthe) glaciation. However, in the 1970-s late Tiraspol (Cromerian) assemblages of small rodents were found both above and beneath the Don Till which placed this glaciation before the Elsterian. This age is amply confirmed by the stratigraphic position of the Don Complex under 7-8 loess-paleosol couplets and atop of the lowermost paleosol of the Bruhnes magnetochron (Velichko et al., 1977, 2011).

No wonder that after several cycles of periglacial activity the original glacial topography of the Don Lobe is hardly visible, although some subdued sandy ridges of ice-contact nature are still preserved. Such a terrace-like flat glaciofluvial ridge proceeds along the eastern bank of the Don for 165 km (1 in Fig.1). Farther east there is the Surena glaciofluvial ridge striking north-south along the axis of the Don Lobe (2 in Fig. 1). The Surena ridge is thought to have served as an ice divide during the retreat stage of the Don glaciation (Kholmovoy, 1981).

The eastern margin of the lobe is poorly expressed in topography and in many places is conjectural. Especially problematic is its northeastern margin in the river Kama catchment area shown by the broken line (Fig.1). In the old map before the advent of the concept of Don glaciation the drift limit attributed to the Saalian maximum was placed approximately along 59.5° N (Ganeshin, 1973). However, such position disregards two southernmost sections of clayey till on river Chusovaya close to its mouth (3 in Fig.1) (Vereshchagina, 1965). The tills are probably left by an ice stream of the Don glaciation which reached 58° N along the western slope of the Urals.

The clasts of the Don diamictons consist mostly of local sedimentary rocks with a small admixture of crystalline pebbles in the western part of the lobe. According to Velichko et al. (2011) the Don tills, unlike other tills of the western Russian Plain, lack the major Fennoscandian indicator erratics which together with south-western orientation of many pebbles suggests a more north-easterly ice dispersal centre, probably on the arctic shelf.

Moscow glaciation

The Moscow glaciation, commonly related to the Saalian, left the largest zone of glacial landscapes in central and northeastern European Russia. In the westernmost part of Russia the Moscow ice limit is not topographically expressive but the retreat stages are indicated by three wide marginal ridges with many kames (Fig. 1). The hummocky uplands west of Moscow that evidently obstructed ice streams of the last Valdai glaciation are massive plateaus with Quaternary more than 100 m thick with abundant glacial disturbances and detached blocks (Fig. 1). Very prominent is the hilly system of the Klin-Dmitrov Ridge more than 200 km long and up to 285 m high perched on a salient of the Paleozoic rocks north of Moscow.

Farther to the north-east wide marginal belts are scarcer being replaced by numerous arcuate ridges of push moraines and wide troughs of meltwater discharge. Lakes are almost extinct, the postglacial drainage network is rather dense but the original landscape of ice disintegration with numerous till hummocks and kames is well preserved. The conspicuous feature of the Moscow landscapes is numerous boggy inter-hillock hollows replacing former glacial lakes and meltwater channels. Such depressions are often filled with interglacial peats and gyttjas of Mikulino climatolith roughly corresponding to the Eemian stage. The slopes of morainic uplands and floors of former lakes are slightly graded by the blanket of loess-like silts 2-3, in places up to 6 m thick.

The orientation of ice marginal belts and pebbles together with petrography of clasts indicate frontal ice motion from Fennoscandia to the southeast (Goretsky et al., 1982). The NW-SE ice directional structures are traceable northwards across the Upper Volga region to the Severnaya Dvina catchment area (Fig. 3) where this ice advance is called the Vychegda glaciation. The Vychegda glacial complex is traced by crystalline indicator erratics of Fennoscandia up to the Arctic Circle and as far to the east as the Pechora river valley (Lavrov & Potapenko, 2005). The ice directional structures amply demonstrate that the Moscow ice sheet was the largest Scandinavian glacier which supports its chronological correlation with the Saalian.

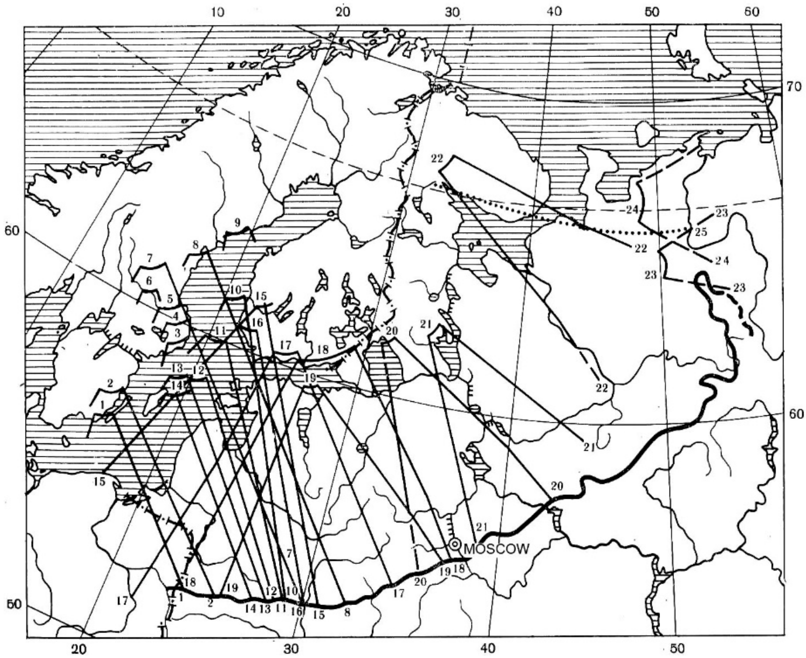

Fig. 3. Locations of sources of Fennoscandian erratics found in the Moscow Till (numbered are resultant lines of transportation) (Goretsky et al, 1982).

In the Pechora basin the contemporaneous glacial system bears unmistakable signs of ice flow from the north-east to south-west. These are massive hummocky plateaus of arcuate configuration accompanied by pushed morainic ridges and kame plateaus. They are divided by wide sandy flats of former proglacial lakes and sandurs. The directional structures invariably show dominant ice flow along the Urals with ice streams deviating to the southeast into the mountains. This NE-SW ice flow is amply reflected in the parallel Rogovaya ridges (4 in Fig. 1). The directional structures are discussed in more detail in paper 3, Ch. I.

Last glacial cycle

The area of the last glaciation is clearly subdivided into two quite different parts affected by i) Fennoscandian ice sheet and ii) shelf ice domes. These glacial systems differ by their age, geographic distribution and landform assemblages.

Late Valdai glaciation

The classical Valdai glaciation that covered all northwestern European Russia is an undisputable part of the Weichselian ice sheet of Europe (Chebotareva, 1977), and therefore in the map is designated as W_3 meaning that the Late Valdai glaciation occurred during the Late Weichselian time span (25 to 11 ka BP). The glacial features of this age are best studied and mapped everywhere in scale 1:200 000. Glacigenic landforms, such as various hillocks and numerous lakes, including huge excavated hollows of lakes Ladoga and Onega, occur all over the country up to midways between St. Petersburg and Moscow. Bare diamictons, heaps of sandy stratified drift and lacustrine clays lie on the surface without any loess cover.

The most conspicuous ice retreat features in the map (Fig. 1) are sub-latitudinal elevations 30 to 70 km wide and hundreds of kilometers long decorated with innumerable morainic hillocks, kame plateaus, ice-walled-lake plateaus, valley trains and accessory morainic ridges of various orientation. They make a concentric system of ice retreat zones. The chains of morainic hummocks interspersed with kames are in places broken by wide lacustrine and outwash hollows. The marginal systems are commonly of mixed composition: glaciotectonic push structures with far-travelled detached blocks of sedimentary rocks and sometimes huge crystalline boulders are interspersed with diamictic wavy plains and sand-gravel accumulations of ablation sediments.

Ice disintegration features such as kames are not obligatory parts of the marginal chains, they are often just accessories of the large push moraine assemblages. Th highest parts of the marginal zones (250-300 m a.s.l.) are usually located upon protrusions of Paleozoic rocks which are exposed only in deep postglacial river valleys. In general the height of marginal ridges and plateaus increases southwards together with growing thickness of the glacial drift (Malakhovsky & Markov, 1969). Along the western border of Russia, where the Quaternary cover is thin, especially on the Precambrian basement of Karelia, eskers up to 10 km long and drumlin fields appear. Such features of wet-based sliding are normally absent in the rest of the area of thick Quaternary underlain by permeable sedimentary rocks.

The marginal belts of the Valdai deglaciation were identified during geological mapping of the 1960-s as `glacial stages` with own geographical names. There are seven such belts: Yedrovo (Bologoye, Valdai maximum), Vepsa, Krestsy, Luga (Karelia, Keiva I), Neva (Siamozero, Keiva II), Rugozero, Kalevala (Malakhovsky & Markov, 1969; Chebotareva, 1977). Similar morphostratigraphic units are designated as `marginal zones` in the Baltics (Kalm, 2012) and `glacial phases` in Poland (Marks, 2012).

There is no good stratigraphic evidence of repeated ice advances in the course of ice retreat towards Finland. Because of the lack of interstadial formations reliably traced northwards the Valdai recessional belts cannot be treated as climatostratigraphic units. At the present knowledge they probably mark just brief standstills or ice margin oscillations in the range of first kilometers. Besides, there is simply not enough time for readvances since all area 700 km wide from the Valdai ice limit in western Russia to the Salpausselkä moraines in Finland (12.5 ka) was deglaciated during 5.5 ka (Lasberg & Kalm, 2013).

The map in Fig. 1 together with Table 1 suggests spatial and temporal correlations of Russian marginal belts with similar deglaciation features in the Baltics, Poland, Germany and Finland. The maximum of the Late Valdai glaciation is associated with Bologoye-Yedrovo moraines ~ 18 to 20 ka old which are probable counterparts of the Poznan phase in Poland, and the Frankfurt in Germany. In Western Europe the Weichselian maximum was reached somewhat earlier at the Brandenburg-Leszno phase ~ 21-24 ka BP (Böse et al., 2012; Marks, 2012). Glacigenic landforms of the Bologoye-Yedrovo zone are not very expressive which is explained by thinning of distal marginal ice (Chebotareva, 1977). The thickest drift of predominantly diamictic composition with numerous Fennoscandian erratics is found in the insular uplands which are built of glaciotectonically stacked tills of various ages. Such uplands served as local ice divides during active ice flow (Malakhovsky & Markov, 1969).

Most spectacular is the highest and widest Vepsa morainic belt making the massive continental watershed of the Valdai Upland 250-300 m a.s.l. and about 600 km long. The Quaternary is up to 200 m thick there with topmost point at 346 m a.s.l. (5 in Fig. 1). The next Krestsy line is drawn along the ice-contact escarpment separating the morainic upland from the proximal glacial depression of the Baltic Lowland at 50–100 m a.s.l. The Lowland was probably deglaciated during the Bølling (Chebotareva, 1977). Thus, the Krestsy standstill appears closer in age to the northern Luga than to the Vepsa moraines.

Farther northwards the lower moranic ridges and isometric massifs are often isolated and generally conform to a distinctly lobate pattern of thinner ice. The Luga morainic belt is topographically less spectacular, and being broken by wide lacustrine basins and meltwater channels, is harder to trace laterally. However, this belt is rather important for St. Petersburg region stratigraphy because the Luga Till is commonly discernible as a special kinetostratigraphic and lithological unit. The festoons of the Luga moraines merging with the Karelian phase in the north delineate late glacial ice streams along the largest lakes Ladoga and Onega which were still active when the adjacent plateaus were already ice free. The Luga belt, morphologically traced northwards, is probably synchronous with Keiva II moraines of the Kola Peninsula (Fig. 1).

The configuration of Keivas is peculiar: they circumvent the practically bare bedrock of the eastern Kola Peninsula without any orographic obstacles. This has since 1960-s been explained by former existence of a stagnant and probably dry-based local glacier called the Ponoi Ice Cap (Ekman & Iljin, 1991). The Kuloi Plateau probably hosted a similar passive glacier (Fig. 1).

In our map we omit the locally recognized morainic chains if they are hard to trace and connect laterally. E.g. the eastward continuation of the Pandivere marginal zone of Estonia was traditionally associated with the so-called `Neva stage` (Kalm, 2012). However, no marginal ridges along river Neva have ever been found. The Neva till is just a mantle of ablation diamicton ca 2 m thick. The Pandivere morainic chain apparently descends to the floor of Gulf of Finland. It emerges on the Karelian Isthmus (far from the Neva channel) as the kame plateau Väremenselkä without any traces of glacial stress (Malakhovsky & Markov, 1969). This zone is presumably traced across Lake Ladoga to appear on its north-eastern shores as the Siamozero morainic belt (Chebotareva, 1977). Thus, the enigmatic `Neva stage` is most likely just a local melting event immediately preceding the Baltic Ice Lake which can hardly be taken for a special morphostratigraphic unit.

Marginal formations in the Kola Peninsula, Russian Karelia and on the bottom of the shallow White Sea are represented by assemblages of low till hummocks, small morainic ridges, drumlinoids, kame plateaus, eskers and glaciofluvial deltas. They designate a set of minor oscillatory advances

Table 1. Correlation of ice retreat zones of last Scandinavian glaciation (from Böse et al., 2012; Marks, 2012; Lasberg & Kalm, 2013; Chebotareva, 1977, Malakhovsky & Markov, 1969; Larsen et al., 2006; Ekman & Iljin, 1991, Saarnisto & Saarinen, 2001.

Ka BP	Germany	Poland	Lithuania	Latvia & Estonia	St. Petersburg, Novgorod, Tver regions	Karelia, Kola, Arkhangelsk regions	Finland
11.3						Kalevala II+III	Salpausselka III
118							Salpausselka II
12.5						Rugozero	Salpausselka I
				Palivere			
13-14				Pandivere		Siamozero (Keiva I)	
14-15		South Middle Bank	North Lithuanian	Linkuva, Haanja	Luga	Karelia (Keiva II)	
		Slupsk Bank	Middle Lithuanian	Gulbene	Krestsy	Kenozero	
16-17	Mecklenburg/ Gerswalder	Gardno	South Lithuanian	Dagda			
17-18	Pomeranian	Pomeranian	Baltija	Major glacier expansion	Vepsa	Valdai maximum	
18-21	Frankfurt	Poznan	Gruda- (Nemunas maximum)		Yedrovo (Valdai maximum)		
21-24	Brandenburg (Weichselian maximum)	Leszno – (Weichselian maximum)					

reflected in the thin and discontinuous drift mantle. Quaternary sediments thicker than 50 m are found only locally in the buried valleys.

The fast sliding advances and retreats were facilitated by the proximity of the slippery, downglacier sloping crystalline basement. The most expressive are Rugozero moraines and younger phases of Kalevala II and III, which are counterparts of the Finnish moraines Salpausselkä I, Salpausselkä II and III accordingly (Table 1). The continuation of Kalevala II and III moraines in Norway is the Tromsø-Lungen ridge.

Correlation of marginal formations of western Russia with dated Weichselian phases of the Baltics and Poland is not always straightforward because of the uncertainty of morphological criteria and problems with compatibility of different dating methods (Kalm, 2012). More or less reliably numerical age is assigned to the Valdai maximum as 17-18 ka BP inferred from OSL (optical luminescence), radiocarbon and cosmogenic exposure dating in the Arkhangelsk and Vologda regions (Demidov et al., 2006; Larsen et al., 2006; Lunkka et al., 2001). This means that the maximum extension of the eastern margin of the Late Weichselian ice sheet in Russia was delayed by ~ 10 millenia as compared to its western margin in Britain (Clark et al., 2012).

The Vepsa moraines which compose the major morainic belt of the Valdai Upland is a probable counterpart of the Baltija Ridge and Pomeranian phase of the North German Lowland dated to 16-17 ka BP by cosmogenic beryllium and radiocarbon (Kalm, 2012; Marks, 2012). Varvochronology combined with radiocarbon and paleomagnetic analyses indicate that the Siamozero moraines are 13.3 calendar ka old (Saarnisto & Saarinen, 2001). The cited authors conclude that it took 1000-1200 years for the ice margin to retreat from the Luga to Siamozero moraines and another 800 years for retreat from the Siamozero to Rugozero (Salpausselkä I, 12.5 ka old) moraines. Therefore, the Luga moraines must be ~ 14-15 ka old (Table 1).

Older Weichselian glaciations

The northeastern Russian Plain is within a totally different glacial realm governed by ice advances from the Kara and Barents Sea shelf. In this area the predominantly fine-grained tills contain considerable amounts of redeposited organics, including marine biota. The shelf ice domes expanded over the entire arctic coast of Russia to the east of the White Sea but not farther south than the Arctic Circle. The idea of the 1960-70-s about the arctic glaciation being contemporaneous with the classical Late Valdai ice sheet after the dating efforts of the last two decades proved to be erroneous as based on too young conventional radiocarbon dates. The Weichselian maximum ice advance in the Arctic could not occur later than 50 ka BP which is evident from the early Paleolithic sites on the surface (Svendsen et al., 2004).

Two independent glacial stages are identified in the arctic Pechora Basin by morphology and OSL dating: the Harbei, 80-90 ka BP, and Markhida, ~ 60 ka BP (Svendsen et al., 2004). The maximum Harbei stage is expressed in topography as a latitudinal chain of push moraines along the Arctic Circle, including the huge arcuate system of the Laya-Adzva Ridge 220 km long (6 in Fig. 1, for satellite image see Fig. 5 in paper 22, Ch. V). This boundary is traced by photogeology into the foothills of the Polar Urals where the Harbei moraines intrude into the mountain valleys. To the south the Harbei moraines are replaced by sands and silts of the proglacial Lake Komi which impounded all river valleys of the Pechora catchment up to 100 m isohypse (paper 18, Ch. IV). 29 OSL dates yielded the mean weighted value of 82 ± 3 ka BP for the lacustrine shoreline sands and thereby for the Harbei ice dam (Mangerud et al., 2004).

The marginal ridges of the Markhida stade are less topographically expressive, probably because of the thinner and not very active marginal ice. Commonly they are arcuate glaciotectonic imbrications built of underlying diamictons and marine sands and bounded upglacier by lakes or boggy hollows. Lavrov (in Chebotareva, 1977) thinks that they are evidence not of ice push but of squeezing up of the clayey substrate from beneath a stationary ice margin. Accretion end moraines consisting mainly of ablation sand deposited by stagnant ice are better known in the area of the Vychegda glaciation.

A similar chronological scheme but with 4 ice advances was developed also in the river Mezen catchment area, Arkhangelsk region (Larsen et al., 2006). The scale of our map (Fig. 1) allows to retain only two major stages which are readily correlated with two Early-Middle Weichselian glaciations elsewhere along the Arctic coast. These two stages, according to Larsen et al. (2006), are separated by the Mezen transgression with subarctic marine fauna which is dated by OSL to 65-75 ka BP. Similar marine sands were later described on the Lower Pechora below the Markhida Till.

In old Weichselian landscapes of the northern plains the retreat zones have a low profile or are non-existent as compared to the Fennoscandian area. Shelf-based arctic ice sheets probably disintegrated very fast under the influence of raising sea level followed by very slow deglaciation in the cold climate of the Weichselian pleniglacial.

Glaciated Urals

Traditionally the Urals have been viewed as a major ice dispersal centre based on the Uralian composition of erratics scattered in the adjacent plains. Accordingly, ice limits were drawn in the plains concentrically to the mountains. The Uralian ice dispersal centers were pictured not only in older maps such as Ganeshin (1976), but in later research publications as well. However, we cannot accept this model because the concentric system does not fit with the clearly observable pattern of morainic ridges mapped by numerous projects of geological survey and confirmed by remote sensing imagery (e.g. Astakhov et al., 1999).

Additional sources of inland ice on the arctic shelf were first suggested for the Middle Pleistocene from exotic boulders perched on flat mountain summits 1000-1100 m high. Uralian stones found in West Siberia also testify to ice motion across the Urals to SE (Yakovlev, 1956). Middle Pleistocene ice that overrode the narrow mountainous range is also inferred from clayey tills with lowland pebbles found in the Northern Urals, 62.5-63.5° N, at altitudes up to 600 m (papers 3,4, Ch. I, and 22, Ch. V). Later foreign clayey till was found even on the Siberian slope of the Middle Urals in Kachkanar quarry, 58°40′ N (7 in Fig. 1) (Sigov, 1971).

The fresh non-Uralian moraines discovered in the 1970-s around the northern tip of the Polar Urals demonstrated that also in the Late Pleistocene the range was circumvented by southbound ice streams (paper 2 in Ch. I). Later a former presence of foreign ice in the Urals was confirmed by finds of horseshoe-shaped moraines at altitudes 250 to 560 m inserted into the mountains from the north-west (Astakhov et al., 1999; Astakhov, 2011).

The latest ice flow along the Urals is also indicated by southern vectors of boulder trains in the surficial till. The bedrock sources of these boulders are located on the Kara Sea shores. The large erratics are transported to the southwest across the Paleozoic Pai-Hoi Ridge (Shishkin, 2007). Now it is evident that Late Pleistocene ice streams from the Kara shelf intruded into the northernmost Polar Urals almost up to 67° N blocking and assimilating small local glaciers. Judging by the fresh hummocky moraines alpine glaciers coalesced to form an Alaskan type ice sheet only on the windward western piedmont between 64 and 65° N at the foot of the highest and widest massif of the Peri-Polar Urals (Fig. 1).

The eastern slope in precipitation shadow is practically devoid of local glaciation traces which appear only farther north along the narrow and low range of the Polar Urals. Although even there geological mapping between 67 and 65.5° N revealed a very limited extent of local tadpole-shaped moraines which do not extend farther than several kilometers from the mountain front on both slopes of the Polar Urals.

In the 1950-60-s the last glacial maximum of the Urals was attributed to the Early Valdai, but later, under the impact of the would-be finite `old` radiocarbon dates, the paradigm shifted to the MIS 2 age of the main ice advance. The recent data on chronometry of the uppermost glacial complex totally disagree with that fashionable idea of the 1970-80-s. Thus, the large boulders dated by cosmogenic ^{10}Be in the range of 28 to 14 ka BP (mean 21 ka) occur within 1 km from the extant miniglacier Chernov. Downstream along the glacial trough boulders already have ages 60 to 50 ^{10}Be ka. This is obvious evidence of insignificant size of Late Weichselian glaciers on the western slope of the Polar Urals (Mangerud et al., 2008). No such young moraines have been found on the leeward eastern slope.

A relative antiquity of the last major ice advance is indicated by the MIS 4 age of frontal sandurs provided by OSL dating. Thirteen dates, the mean value of 73 ka, are obtained from the sandur east of the horse-shaped moraine of a foreign glacier inserted into a mountain valley (Nazarov et al., 2009). The glaciofluvial sands from the margin of the piedmont morainic apron on river Bol. Usa yielded similar ages of 67 to 63 ka BP. The MIS 4 age of the moraines agrees with ^{14}C values ca 40 ka BP (11 dates) from alluvium incised into the youngest varved formation at Paleolithic site Mamontovaya Kurya (Svendsen et al., 2014).

The above dates imply that alpine glaciers that survived south of the Kara ice margin were not younger than MIS 4. This fact sheds light also on the origin of large piedmont moraines east of the small morainic loops of the maximum alpine glaciation (magenta arcs in Fig. 1). Earlier these large subdued and soliflucted moraines have been taken for traces of the Late Pleistocene Uralian glaciation. But they have no symmetric counterparts on the wetter western slope. This excludes a Uralian origin for the large eastern moraines. They must have been left by trans-Uralian ice streams derived from a thick European ice sheet retreating to the northwest in the late Middle Pleistocene. Thus, the last glacial maximum of the Urals was reached in the MIS 4 time. The main invasion of inland ice from the Kara Sea incorporated disjointed alpine glaciers which coalesced to produce a thin local ice sheet only on the western piedmont south of 65° N.

West Siberian ice limits

Beyond the Urals ice limits are hard to trace across the flat and swampy West Siberian Plain with low, barely noticeable topography. The drift limit is identified largely from rare river exposures and borehole profiles. It is sometimes designated by large erratic blocks of Paleogene sands, opokas and Mesozoic clays, such as the famous Yugan megablock of Jurassic clays brought from a source ~ 600 km to the northeast (8 in Fig. 1). As is the rule of thumb in this lowland, ice-pushed ridges and glaciotectonics assemblages occur only uplacier and never along ice limits. Due to the north-sloping topography glaciers contacted with deep proglacial lakes which precluded any considerable compression along the wet-bed ice margin. Marginal features are better expressed in relatively high and dry borderlands along the Urals and Central Siberian uplands. Diamictic ice-pushed features, such as the Agan ridge (9 in Fig. 1), are scarcer than in European Russia.

West Siberian diamictons are, as a rule, poor in clasts and their fine-grained composition does not benefit topographic expression. In many places they are replaced by glaciotectonic parallel ridges which are modest landscape manifestations of imbricate stacks of soft rocks, partly Quaternary but mostly Mesozoic and Paleogene clays, sands, diatomites and opokas (Fig. 3 and 10 in Fig.1). However, the arcuate configuration of parallel ridges and their orientation in plan reveal ice flow directions as well. Narrow sandy ridges, owing to selective erosion, normally stick out for several meters above clayey intervening troughs. But in places arcuate composite ridges backed by deep proximal lakes produce expressive hill-hole pairs more than a hundred meters high, such as two horseshoe-shaped ridges just west of the Yenissei (11 in Fig. 1). For satellite image see Fig. 2b in paper 4, Ch. I. In such cases they are designated in Fig. 1 as morainic ridges (small horseshoe-shaped arches painted magenta).

The drift limit has traditionally been assigned to the Samarovo glaciation of MIS 8 age based on the underlying Tobol alluvium correlative with the Holsteinian. The next boundary to the north, the Taz glaciation, is related to MIS 6 because its tills are overlain by interglacial formations indicating by paleobotanic proxies a climate warmer than today. The Taz tills are also covered by thick loess-like silt with a paleosol.

The limit of the Weichselian glaciation has for a long time been very problematic due to the flat landscape and poor accessibility of arctic West Siberia. Greatly controversial positions of Late Pleistocene ice limits (e.g. Arkhipov et al., 1976, 1986) were prodded by misinterpretations of glacial topography. Many investigators used the so-called freshness of glacial landscapes as a correlation tool. However, special research revealed the retarded or delayed deglaciation in Siberia conditioned by the old, thick and very persistent permafrost. Therefore, fresh hummock-and-lake landscapes with many glaciokarst hollows formed only during the Holocene warming everywhere irrespective of the age of the ice advance. The implication is that they, unlike in European Russia, cannot serve as a correlation tool. Buried fields of glacial ice are still persistent in the Arctic and in many places kames and other ice-contact forms have not appeared yet (Astakhov & Isayeva, 1988).

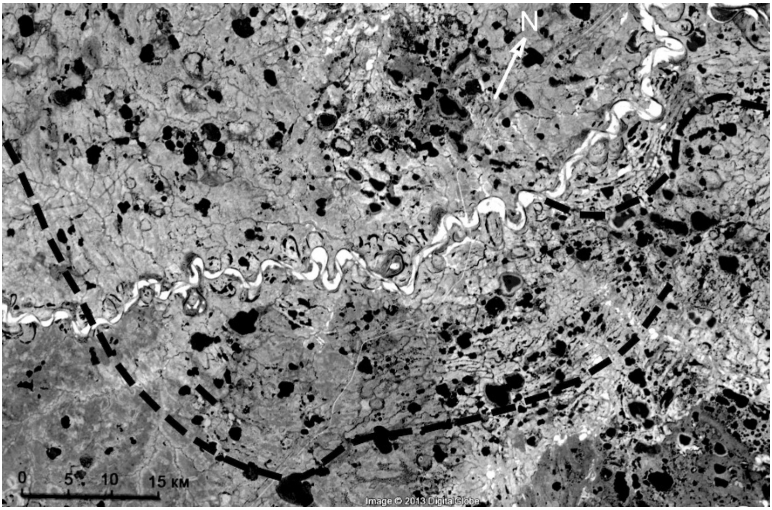

Fig. 4. Satellite image of glaciotectonic parallel ridges built of Paleogene rocks at the northern flank of the Urengoi gas field. Broken lines indicate the distal edges of the low arcuate ridges which are surficial manifestations of disharmonic imbricate structures.

The latest dating efforts by the optical luminescence method allowed to locate the Late Pleistocene maximum limit beyond the 67°N, much more northerly than it was previously suggested e.g. by Arkhipov et al. (1976, 1986). East of the Urals the Sopkay moraines (12 in Fig. 1) are crucial for the position of the limit of the last glaciation. Originally they were considered a product of the Uralian glaciation arbitrarily dated to the Late Weichselian or even the Younger Dryas stadial. However, the mapped morainic ridges proved to be very clayey with directional structures parallel to the Urals (papers 2, Ch. I and 18, Ch. IV). Immediately to the south sediments of a proglacial lake in the Ob valley at 50-60 m a.s.l. yielded a series of OSL dates with the mean age of 81 ka BP. This implies that the Sopkay ice advance occurred simultaneously with the Harbei ice advance west of the Urals (paper 14 in Ch. 3) and therefore should be indicated as early Weichselian (W_1) as well. The moraines and glaciotectonic ridges are photogeologically traced across the Yamal Peninsula to Gydan. The uppermost till dated by radiocarbon and OSL proved to be not younger than the overlying frozen mammoths ^{14}C 39-42 ka old and older than 69 luminescence kiloyears (paper 16, Ch. III).

In general, there are virtually no ice retreat features in the central West Siberian Plain where ice sheets seem to have disintegrated immediately with the onset of interglacial climate and sea level rise. On the eastern border of the plain several chains of kame landscapes, hummocky moraines, glaciotectonic parallel ridges and sometimes N-S striking eskers together with pebble orientations and other directional structures testify to a mighty N-S ice stream along the Yenissei valley. But latitudinal morainic chains can hardly be taken as indications of successive ice margins, they rather reflect irregular melting of stagnant ice in valley lakes after the initial disintegration of the last ice sheet (paper 14 in Ch. III). The southernmost belt of push moraines and ablation hummocks along the Arctic Circle is fringed by a deep valley lake with varved formations covered by sandurs at 60-70 m a.s.l. The glaciolacustrine sediments yielded a series of OSL dates in the range of 95 to 73 ka BP, mean value ca 85 ka which is similar to the dating results from the Ob proglacial sediments (Astakhov et al., 2010).

Thus, the last ice advance from the Kara shelf onto the West Siberian Plain must have occurred very early in the last glacial cycle, indicated as W_1 in the map. No traces of Late Weichselian ice advances have been found in arctic West Siberia. They have no stratigraphic place amongst loess-like silts over 37 ^{14}C ka old and numerous finds of frozen mammoth bodies atop of the uppermost glacial complex (paper 16 in Ch. III).

Ice limits in Central Siberia

Central Siberia is very topographically and geologically diverse which predisposed great variety of ice marginal formations. The drift limit is traced by mapping geologists across the high Central Siberian Plateau built of gently lying Paleozoic sedimentary rocks and Triassic volcanogenic formations. The Putorana and Anabar plateaus display concentric belts of isolated pebbly morainic ridges delineating former ice caps which formed largely over the dolerite Putorana Plateau up to 1600 m high (Fig. 1) (Arkhipov et al., 1976). They are often associated with meltwater canyons and gravel outwash terraces. There are two parallel ice limits drawn mostly by photo interpretation and thought to correspond to the West Siberian Middle Pleistocene ice margins – the Samarovo and Taz, although no chronometric evidence of this correlation is available. The drift limit is probably time transgressive. In the east it seems older judging by the absence of thick moraines and by the only signatures of former ice motion – crystalline boulders from Triassic volcanogenic formations of the Putorana Plateau and Precambrian rocks of the Anabar Plateau.

The oldest and most poorly expressed ice limit is tentatively correlated with the Lebed Till deeply buried in the Yenissei valley (paper 22 in Ch. V). The Lebed glaciation is thought to be of an Elsterian age. Along the Samarovo ice limit patches of till hummocks are preserved better. The Middle Pleistocene maximum glaciation limits are lost in the east due to the lack of ground data. Extensive fields of diamict hillocks and sand-gravel kames are mapped only within the Taz ice limit, especially in the east where Anabar moraines with Precambrian clasts and Putorana moraines with mafic boulders coalesce.

A peculiar feature of the Middle Pleistocene glaciation is the occurrence of tills with lowland pebbles and mineralogical assemblages in the western part of the Central Siberian Plateau. The best example is the Bol. Tundrovaya table mountain which is a dolerite inselberg 618 m a.s.l. (13 in Fig. 1). This summit is covered by double diamictons separated by gravelly sand which is evidence of very thick ice just at 100 km upglacier from the drift limit (see Fig. 4 in paper 22, Ch. V). There are other facts, such as bauxite blocks brought to the plateau from their sources in West Siberia, which indicate the thickest Middle Pleistocene ice was in the West Siberian Plain and not on the Central Siberian Plateau (paper 22 in Ch. V). The corresponding ice flow is indicated by arrows in Fig. 1.

The largest marginal features, however, are displayed within the zone of Late Pleistocene glaciations which occurred after the Karginsky interglacial transgression with boreal (atlantic) fauna. The most spectacular are marginal formations of the Taimyr Peninsula which are wide morainic belts and ridges built of diamictons, outwash sand and partly of stacked Quaternary marine and Cretaceous deposits. Even though eroded by the dense postglacial gully network the marginal features are enormous: the largest Jangoda-Syntabul composite ridge almost continuously proceeds for up to 800 km which is probably the record (Kind & Leonov 1982).

Very expressive morainic loops are parts of the next marginal chain to the north. The lobate ridges of the Mokoritto system (14 in Fig. 1, see also Fig. 28.3 in Möller et al., 2011 for satellite image) are parallel concentric indicating a local frontal ice retreat which is unusual for arctic Siberia. This is probably due to the position of these marginal formations over the Paleozoic bedrock on the southern slope of the Byrranga Mountains. The sloping bedrock facilitated basal sliding by friction heat released at the wet base. This was beneficial for frontal retreat and for developing arcuate and lobate patterns. In contrast, the perennially frozen substrate of West Siberia was prone to deep excavation of soft sedimentary bedrock by dry-based glaciers which afterwards disintegrated areally (paper 22 in Ch. V).

The recent dating efforts suggest that the major marginal belts are much older than the MIS 2 age that was previously maintained by Kind & Leonov (1982) and Arkhipov et al. (1986). The age of two main stages are inferred from OSL dating as Early and Middle Weichselian according to the summaries by Svendsen et al. (2004) and Astakhov (2011, 2013). The main marginal belt of the Taimyr Peninsula is OSL dated to 80-90 ka BP (Svendsen et al*.,* 2004; Möller et al., 2011). Accordingly, the previous more extensive Murukta glaciation cannot belong to the Late Pleistocene because the thick Murukta tills are covered by interglacial marine and terrestrial (Karginsky) sediments.

No reliable radiometric indications of MIS 2 glaciers are available beyond sparse conventional ^{14}C dates ~ 20-30 ka which are probably too young. The expressive morainic loops of the Norilsk moraines barring deep fjord-like lakes of the north-western Putorana Plateau (15 in Fig. 1), have long been taken for signatures of the Sartan glaciation ~ 20 ka old or the Younger Dryas stadial. But these lakes contain very thick postglacial silts older than 20 ka (Svendsen et al., 2004) and therefore were probably excavated by older ice advances.

We cannot adopt the Late Weichselian age of the morainic lobe on the northeastern coast of the Taimyr Peninsula (Svendsen et al., 2004; Möller et al., 2011, 2015). It is based exclusively on the feeble support of two dates ~ 23 ^{14}C ka BP on shells from till. To accept this idea one should believe that on 77° N during the LGM lowstand there was an open sea inhabited by *Hiatella* and *Astarte* which were then brought by a narrow shelf glacier from -100 m to +120 m without affecting nearby Severnaya Zemlya populated by mammoths. We find easier to believe in excessively young ^{14}C ages which plague the Siberian shell dating.

The idea of the age of the second stage of the Late Pleistocene glaciation of Central Siberia is given by OSL dating of ice-dammed lake sediments in the North Taimyr ice-marginal zone (Möller et al., 2011) and of sandurs of the last ice advance from Putorana into the Yenissei valley close to the mouth of river N. Tunguska (Astakhov & Mangerud, 2007; Astakhov et al., 2010). In both cases OSL dating yielded sets of values ca 60 ka BP. This is the reason for the Middle Weichselian age of the second stade of the last glaciation indicated as W_2 in the map.

Ice limits of presumably MIS 2 stage in Fig. 1 are purely conjectural because no dates are available for the thin discontinuous and bouldery moraines topping the Putorana Plateau.

Conclusions

The presented map is destined to help in sorting out the history of former ice sheets in the southeastern sector of the Scandinavian glaciation and their relations with northeastern ice sheets fed by shelf ice domes and uplands of Central Siberia. Although the mapped pattern of the Scandinavian glaciations is hardly sensational, ice limits beyond the Scandinavian realm suggest that Pleistocene ice sheets, both in Siberia and in European Russia, were significantly influenced by mighty ice dispersal centers on the arctic shelves. Arctic sources of inland ice had outstanding influence during earlier stages of glacial cycles (MIS 8 and MIS 5-4). The Scandinavian ice sheet was most active during the Late Saalian (MIS 6) and Late Weichselian (MIS 2) stages.

The importance of inter-regional comparison of mapped glacial features is quite evident in the Urals which is the main obstacle for trans-Eurasian correlations of Quaternary formations. E.g., the pattern of large moraines on the Siberian flank of the Polar Urals has no counterpart on the windward European slope. This asymmetry suggests the origin of the large east-Uralian morainic loops from trans-Uralian overflow of inland ice in the late Middle Pleistocene and not from local glaciers of Weichselian times according to the conventional wisdom. The mapped ice limits do not support the traditional idea of the Urals as a substantial source of Pleistocene inland ice. This low and narrow mountain range has always served mainly as an obstacle for ice streaming from the arctic shelves.

Acknowledgments

The Glaciomorphological Map of the Russian Federation was ordered by the Russian Geological Survey and compiled in VSEGEI under the guidance of Dr. Petrov, Director General. This paper was prepared according to the research plan of St. Petersburg University and supported by the Russian Foundation for Basic Research, grant 14-05-00579.

26. Glaciotectonism in a Great Sedimentary Basin

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RUSSIAN GEOLOGY AND GEOPHYSICS, 2019, vol. 60, No. 12, p. 1353–1367.

Introduction

Recent tectonic history of economically important northern sedimentary basins is deeply influenced by Quaternary glaciations. Comparing the structure of a large sedimentary basin against the confining highlands could be instructive for understanding the scale and imprint of glacial processes on the regional tectonic evolution and distribution of mineral resources. The major petroleum province of Russia, the West Siberian Basin (Fig. 1), located between the Uralian mountains and Central Siberian Plateau, presents such an exemplary case. In order to assess the nature of recent tectonism and its significance this paper offers a short overview of the macroscopic neotectonic features of the region accompanied by a detailed cross-section of the largest mesoscopic structure. The data available on features of different scales are presented and discussed separately. Their basically glaciotectonic nature is considered as indications of deep influence of Quaternary glaciation on the recent history of the Russian North and as material for comparison with other northern sedimentary basins (Astakhov, 2019).

There are popular definitions of glaciotectonism as related `to the processes of glaciotectonic deformations` (Aber & Ber, 2007, p. 6) and `Glaciotectonics is glacially induced structural deformation of bedrock or sediment masses as a direct result of glacier-ice movement or loading` (*ibid.* p. 7). I prefer a broader option relating to glaciotectonism any changes in geological structure best explained by ice load applied vertically and tangentially. Such changes, as will be shown, are not limited to glacioisostatic movements or thin-skinned deformations of sediments.

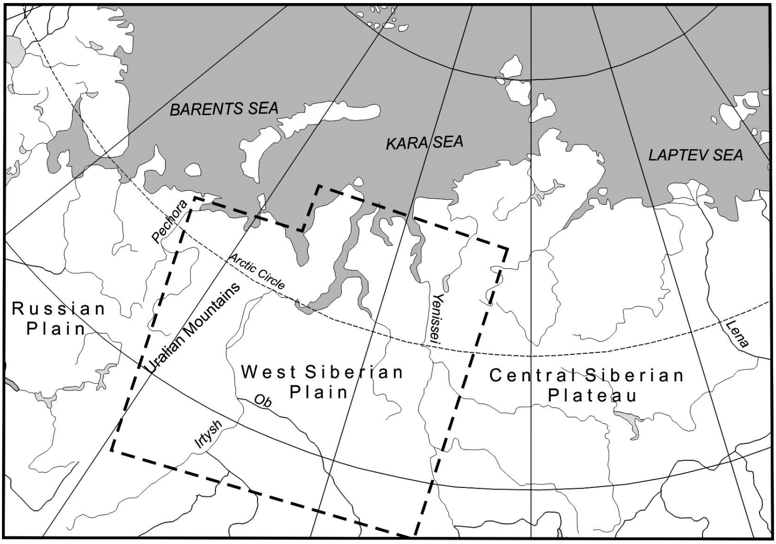

Fig. 1. Location map; study area (Fig. 4) is marked by broken line.

Geological setting

The West Siberian Plate stretching between 55 and 73° N is the world greatest sedimentary basin of more than 2.2 mln km^2 in area (Fig. 1, 2). Its basement is a cratonised system of Paleozoic folded zones with blocks of the Precambrian. The last paroxysm of deep-seated tectonism produced a system of south-north striking Triassic rifts filled with terrigenous rocks and basalt lavas. The overlying unconsolidated Mesozoic-Cenozoic strata, commonly 3-4 km thick but up to 9 km thick in the Arctic, is the body of the West Siberian hydrocarbon province containing more than 50% of Russian petroleum and natural gas reserves (Kontorovich et al., 1975; Ulmishek, 2003). Its topographic manifestation, the West Siberian Plain, is a swampy flatland descending northwards from 300 to 100 m a.s.l. to eventually merge with the Kara Sea shelf. The northern half of the basin is covered by glacigenic and marine Quaternary formations up to 400 m thick. The flat relief and poor accessibility of the swampy boreal forest and perennially frozen tundra led to many controversial interpretations of Quaternary history (e.g. Zubakov, 1972).

The petroleum is discovered in permeable Mesozoic strata composing gentle elongated undulations striking parallel to Paleozoic borderlands of the Urals and Central Siberia. The common petroleum traps are very gentle local anticlines formed by sedimentary draping over basement salients less than 2-3° steep and sealed by Upper Cretaceous clayey formations. A good deal of folds' amplitude is due to consolidation of thick marine clays (Kontorovich et al., 1975). The local synsedimentary anticlines show practically no post-Oligocene growth. They normally flatten upwards from the Middle Jurassic strata to horizontally lying Paleogene clays, excluding the arctic terrains where structural traps in places show a Late Cenozoic amplitude growth ~ 20-50 m (Kuzin, 1983). According to geochemical studies the explored petroleum reserves were formed mostly in Neogene-Quaternary times (Rylkov et al., 1976) which stimulates the interest in recent tectonics.

Neotectonic macrostructure

Distribution of macroscopic features

The neotectonic pattern of the West Siberian Plain shows a latitudinal zonality which is transverse to the principal sub-longitudinal strike of major structures of the sedimentary basin. Such zonality is especially pronounced in the regional trend and in the pattern of the largest structures. Only in the south of the Basin all Cenozoic formations are evenly spread as thin blankets. In the north the Pleistocene cover is ~ 100 m, in places even 300-400 m thick, whereas Neogene and Upper Paleogene strata are mostly absent (Kontorovich et al., 1975). The age of sub-Quaternary sand, clay and diatomite formations appears progressively older towards the Kara Sea concurrently with the descent of the surface. I.e. the pre-Quaternary stratigraphic gap widens northwards (Fig. 2).

Along the borders of the Plain there are some south-north striking depressions outlined by fault escarpments. The most pronounced is the Yenissei Depression which is seen in satellite imagery as a young graben expanding northwards (Fig. 3 and 8 in Fig. 4). The Quaternary is up to 342 m thick there as measured in a borehole at the southern end of this rift-like trough (Arkhipov et al., 1976; Zubakov, 1972).

Another characteristic feature is young clay diapirs involving Tertiary marine clays, opokas and diatomites. Beyond the drift limit their amplitudes are barely few metres, maximum first tens of metres which is accountable by the gravitational adjustment to fluvial escarpments up to 50-60 m high. However, north of the drift limit the diapirs can be hundreds of metres high (Generalov, 1987). These large diapirs without deep roots may be explained only by much higher gradients of lithostatic and fluid pressure, probably caused by disjointed slabs of stagnant glacier ice (see next section).

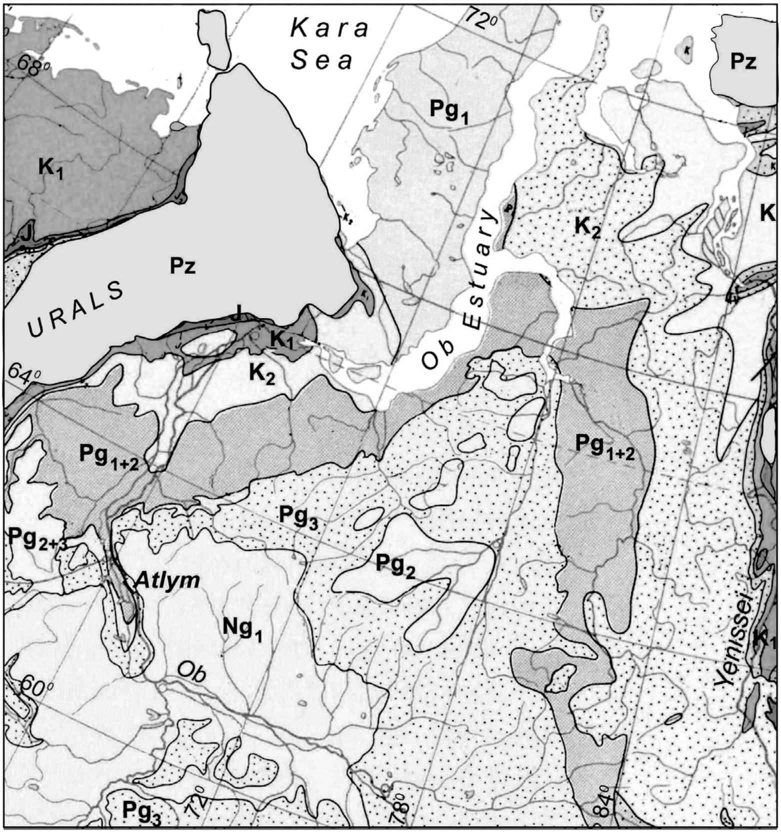

Fig. 2. Schematic geological map of northern sedimentary basins.
Background of Atlym disturbances in Fig. 10 are formations of: Pz – predominantly consolidated Paleozoic rocks of the Uralian and Central Siberian borderlands; principal Mesozoic and Cenozoic formations: J – Jurassic sands overlain by clay, K_1 – Lower Cretaceous shales with sandstone interlayers, K_2 – Upper Cretaceous sandstones and shales, Pg_1 – Paleocene clays, Pg_{1+2} – Paleocene and Lower Eocene clays, Pg_2 – Eocene opokas, diatomites and clays, Pg_{2+3} – Upper Eocene–Lower Oligocene clays with siderite and sand interlayers, Pg_3 – Oligocene sands and silts, Ng_1 – Miocene silty rhythmites.

It is noteworthy that the south-north change of the sedimentary basin is concurrent with changes in the neotectonic macrostructure of the adjacent highlands. A prominent example is presented by the Central Siberian Plateau along the eastern border of the Plain. Its height steadily increases from 250-350 m a.s.l. close to the drift limit up to 1300-1700 m in the Arctic Putorana Plateau (Fig. 3). The latter is deeply dissected by rectilinear network of fjord-like valleys which reflects a system of extension fractures and faults caused by a young dome-like uplift (Fig. 4). The Putorana Plateau is a clearly structural antiform being located in the deepest Paleozoic-Triassic depression of the north-western corner of the Siberian Craton called the Tunguska Basin (Staroseltsev, 1985). The fresh tectonic escarpments cutting young glacial features are a probable cause of the Holocene seismicity there (Maksimov, 1970).

Another spectacular example of the south-north neotectonic change is presented by the Ural Mountains making the western border of the Plain. Independently of the persistent south-north strike of all Paleozoic structural and formational zones the mountain chain steadily grows northwards from the drift limit (Fig. 4). South of this line the Urals are mostly gentle hills with many monadnocks and very wide pre-Quaternary valleys. The Middle Urals close to the city of Ekaterinburg merge imperceptibly with peneplanated piedmonts and display only rare summits exceeding 450 m a.s.l. The most salient feature of the Middle Urals is the lack of topographic contrast between the Urals and adjacent Cis-Uralian and Trans-Uralian plains of similar altitudes (Rozhdestvensky, 1971)

Actually, the Middle Urals are hardly noticeable from the air when jet-flying across the range. The Southern Urals are considerably higher (1000-1500 a.s.l.) but consist mostly of gentle ridges with concave slopes built of most resistant rocks such as quartzites. The subdued ridges are separated by much wider longitudinal valleys which are too large for the extant rivers. The well graded slopes with patches of weathering mantles and Paleogene sedimentary formations indicate that the major elevations of southern mountains belong to pre-Quaternary erosional cycles. Pleistocene fluvial erosion left only narrow valleys less than 100 m deep (Bashenina, 1948).

The general subdued appearance of the old relief of the Southern and Middle Urals changes to quite a different aspect of alpine topography in the much higher Sub-Polar and Polar Urals reaching 1400-1800 m a.s.l. The principal difference in the age of landscapes of southern and northern Urals is evident from comparison of volumes of the negative and positive landforms: whereas old gentle-sloped valleys of Tertiary erosion cycles make

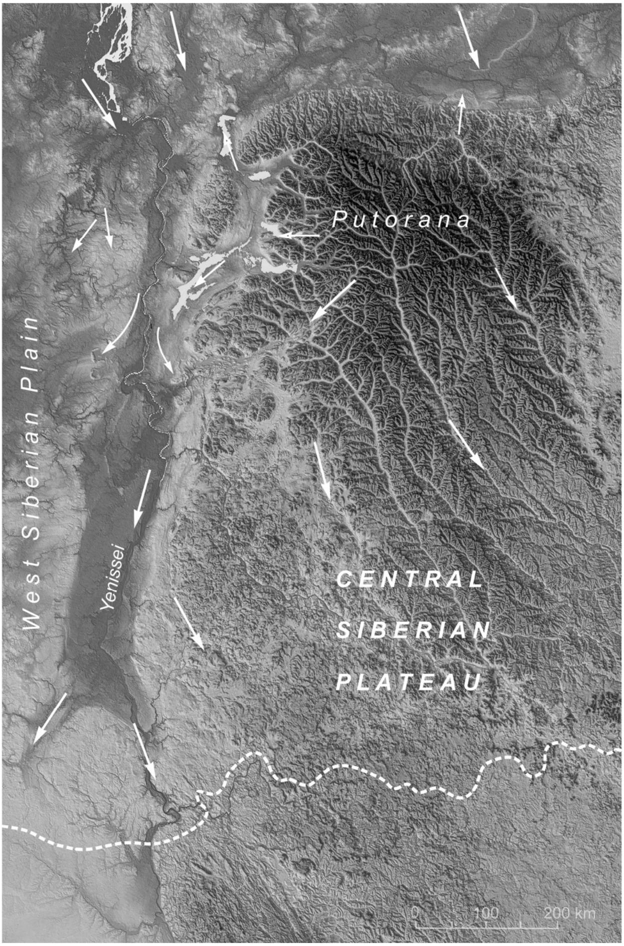

Fig. 3. Digital model of eastern West Siberia and Central Siberian margin. Broken line is the glacial drift limit, arrows indicate former ice flows. Change of light shades of boreal forests in the south to dark colours of montane tundra in the Arctic reflect gradual increase of elevations from 300 to 1300-1700 m contrary to the deepening of the Paleozoic-Triassic basin.

the general picture in the south, positive features, such as uplifted alpine blocks, dominate the northern landscapes. In short, the southern part of the Urals is a residual range and the northern part is rejuvenated low mountains (Astakhov, 1986).

Most revealing are the borders of the West Siberian Plain. South of the drift limit they are poorly expressed, both in the west and east. Only in the glaciated area they appear as tectonic escarpments growing northwards from tens to hundreds of meters concurrently with the gradual descent of the Plain (Fig. 4). The result is the scissor faults with offsets up to 1000 m separating the northern highlands from the adjacent lowlands whereas the unglaciated Urals and Central Siberian Plateau merge gradually with the Russian and West Siberian plains.

In the West Siberian Plain large hummocky uplands 200-300 m a.s.l. built of thick Quaternary sands and diamicts occur only north of the 60-61° N. In plan they compose a huge horseshoe of the Siberian Hills roughly parallel to the glacial drift limit recognised early in the XX century (Fig. 4). These uplands are in places bordered by linear escarpments several tens of metres high conciding with deep-seated geophysical linears. The mapping geologists interpreted the linears as basement faults reactivated by Pleistocene glacioisostasy (Zaitsev, Meshalkin, 1987).

Discussion

The popular ideas for the origin of the large transverse landforms and for the Neogene stratigraphic gap in the north have long been invoking a Late Cenozoic tectonic inversion which presumably affected the northern part of the Basin. Many geologists believed that the northern, most depressed part of the Basin was suddenly uplifted in the Neogene to subside again in the Quaternary (e.g. Rudkevich, 1974). This hypothesis of the regional oscillations of deep-seated tectonism seems hardly compatible with the stable subsidence of the Basin inherited since the Jurassic and with the numerous erosional remnants of Oligocene-Miocene sands and clays north of 64° N (Fig. 2). Besides, in the south of the Plain a thick cover of correlative terrigenous sediments commensurate with the hypothetical Neogene uplift of the north is absent.

The distribution of the Pleistocene formations against the background of the large landforms suggests an alternative explanation for the neotectonic macrostructure of the West Siberian Plain. The decades of geological mapping and studies of glacial erratics and thick diamictic formations have reliably established that the West Siberian north was at least 5 times covered by Middle and Late Pleistocene ice sheets (Arkhipov et al., 1976, 1986; Sukhorukova et al., 1987; Astakhov, 1977, 2004, 2011, 2013). Paleoglaciological estimates based on mapped ice limits give maximum Middle Pleistocene ice thickness 3.8 km in the central West Siberian Arctic (Voronov, 1968) and up to 4.5 km for the ice dispersal

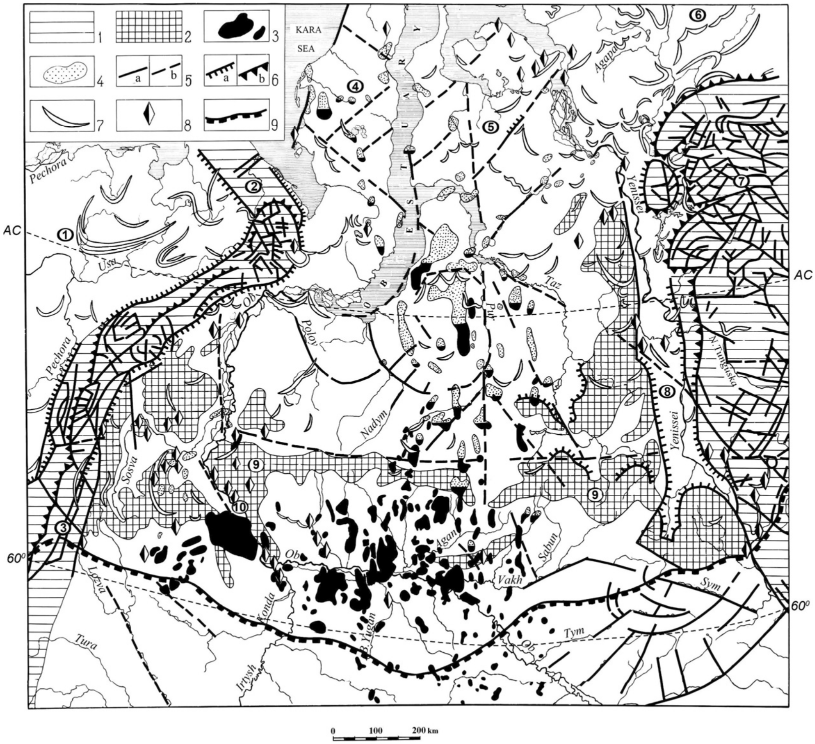

Fig. 4. Glaciotectonic map of West Siberia and adjacents uplands (modified after Astakhov, 1986). Geological and geomorphological features: 1 – Paleozoic uplands; 2 – Quaternary uplands within West Siberian Plain; 3 – oil fields; 4 – natural gas deposits; 5 – large linears observable in satellite imagery: a – orohydrographic, b – discontinuously traced in landscape; 6 – tectonic escarpments: a – tens of metres high, b – hundreds of metres high; 7 – ice-pushed ridges – topographically expressed imbricate assemblages of soft rocks; 8 – large detached blocks of Meso-Cenozoic rocks; 9 – glacial drift limit; physiographic elements indicated by circled numbers: 1 – Pechora Lowland, 2 – Pai-Hoi Range, 3 – Urals, 4 –Yamal Peninsula, 5 – Gydan Peninsula, 6 – Taimyr Peninsula, 7 – Putorana Plateau, 8 – Yenissei Depression, 9 – Siberian Hills.

centre on the Kara Sea shelf (Lambeck et al., 2006). For the northern part of the sedimentary basin this means isostatic subsidence ~ 1 km with the commensurate compensation uplift of adjacent borderlands. This value is close to the amplitude of the entire Paleogene subsidence. The regional glacioisostatic subsidence should have been followed by a rebound of a comparable magnitude. Also, the ice flow and static pressure would have caused large-scale lateral redistribution of the non-consolidated substrate rocks in the form of voluminous glaciotectonic disturbances.

Glacial origin of the large-scale geomorphic aspect of the northern part of the West Siberian Plain is also clearly demonstrated by the south-north change of the Quaternary thickness as observed in several drilling profiles across the valleys of the Ob catchment. The data were obtained by Soviet geotechnical projects of Hydroproject Corporation exploring possible routes for diversion of Ob water to the southern deserts (Fig. 5). The Quaternary sediments, normally 30-40 m thick in the south, drape the smooth north-bound profiles of the former and present waterways.

However, north of the drift limit Quaternary sediments filling the buried valleys suddenly thicken up to 100-300 m. The even and smoothly graded concave bottom of the Tobol-Irtysh stem valley of the periglacial area plunges abruptly below sea level in the glaciated area. Downstream, along the river Ob, the pre-Quaternary bedrock profile is obviously complicated with big humps (Fig. 5). All this suggests that the buried valleys of the West Siberian north are excavated not by riverine but by glacial activity. The form of the Ob buried valley in Fig. 5 is very similar to bumpy subglacial tunnels of northern Europe and America attributed to discharge of over-pressurized subglacial meltwater (Kehew et al., 2012).

The available evidence of recent tectonics indicates that only rare features inherit ancient structural directions whereas the general pattern is produced by zonality transverse to the ancient longitudinal trends. This \`geographical\` or rather paleogeographical zonality of the topography and recent tectonism mirrors the distribution of Pleistocene ice thickness. The process of the slow adjustment of basement blocks to the increasing sedimentary load during the Mesozoic-Cenozoic was drastically accelerated in the Pleistocene when the rapid ice accumulation disturbed the system of isostatically balanced blocks. The rapidly applied load is the most probable explanation for the Pleistocene faults of high amplitudes incongruous with stable plains and old residual mountains (Astakhov, 1986).

Glacioisostatic effects in the Ural-Siberian north demonstrate a cardinal difference with the Fennoscandian pattern. Forebulges, i.e. peripheral compensation uplifts, that used to form around the former ice sheet, now are almost extinct in Fennoscandia. They are replaced by a gentle circular trough commensurate in volume with the central glacioisostatic dome (Mörner, 1980). These epeirogenic oscillations imply elastic deformation of the lithosphere. In the Ural-Siberian North, on the contrary, most conspicuous are marginal compensation uplifts indicated by isostatic horsts of the Urals and Putorana and probably also the Siberian Hills in the centre of the West Siberian Basin. These uplifts matched by the persisting subsidence of ice-loaded Arctic West Siberia are evidence of residual deformation of the lithosphere (Astakhov, 1986).

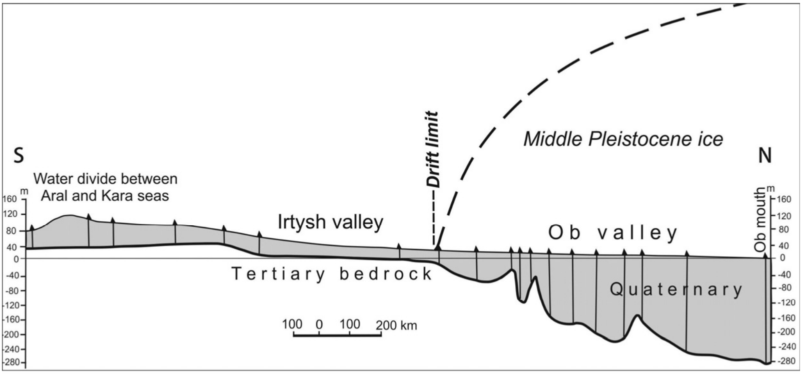

Fig. 5. Distribution of Quaternary thickness along the south-north striking river valleys of the Kara Sea and Aral Sea catchments according to transverse drilling profiles (black triangles) of Hydroproject Corporation (after Astakhov, 1991). Note the sudden increase of Quaternary thickness north of drift limit.

The residual deformation versus elastic rebound is probably due to the antithetic structure of the deep-seated lithospheric layers of the Ural-Siberian region in contrast to Fennoscandia. It can be judged from the depth of the sharp change of seismic velocities called the Moho boundary which geophysicists commonly take for the bottom of the `crust`. In Fennoscandia the Moho boundary plunges down to the centre of the isostatic dome from 30 to 47 km (Mörner, 1980). Conversely, in West Siberia the seismic M boundary is the deepest (47 km) under the marginal highlands whereas at the inferred ice dispersal centre in the Arctic lowlands M level is only 35-40 km deep (Karus et al., 1984).

Radial glaciotectonism and hydrocarbon reserves

The concept of the structural reorganisation of the sedimentary basin by glaciotectonism may be useful for solving some problems of petroleum geology in West Siberia such as origin of anomalous fluid pressures, explanations of geothermic peculiarities, saturation of traps, etc. E.g. the glacial interpretation of the neotectonic macrostructure can be applied to a solution of the old puzzle of the geographical separation of natural gas and oil deposits. The major oil fields discovered in Upper Jurassic–Lower Cretaceous strata at depths of 1.5 to 3 km are located within a west-east striking belt in the centre of the Plain (60–63° N) which is transverse to the general south-north strike of the main Mesozoic structures. The Subarctic belt to the north of the Siberian Hills contains a mixture of oil, gas-condensate and gas deposits. In the Arctic huge gas fields sealed by Paleogene clay formations decidedly predominate (Fig. 4).

There were attempts to explain the geographic separation of gas and oil reserves by changes of hydraulic pressure in petroleum-bearing strata caused by the fluctuations of sea level (Kuzin, 1983). However, the fluid pressure drop ~ 20-30 atmospheres that could be provided by a hypothetical Pliocene regression of the Arctic Ocean is too small for degassing of ground water of Mesozoic aquifers. The rate of fluid pressure decrease in such a case would be insufficient for generation of the giant gas fields of West Siberia such as the Urengoi pool ~ 120 km long (Korzenstein, 1970), speaking none of the dubious geological evidence of such a regression.

The alternative hypothesis sees the reason for the geographical separation of different types of hydrocarbons in lateral migration of fluids within aquifers under glacial load and in enhanced gas generation in the perennially frozen zone of hydrate formation (Trofimuk et al., 1979). Unfortunately this model employed an outdated reconstruction of ice sheets which were presumably thinning towards the sub-longitudinal axis of the basin which is hardly compatible with the actual distribution of petroleum reserves (Fig. 4). Also the deepest Pleistocene permafrost is found not in the Arctic zone of the main gas accumulation but in the oil belt some 600-700 km to the south. This hypothesis is welcomed by petroleum geologists neither because a lateral migration of viscous oil through compact sandstones over distances of hundreds of kilometers is hardly feasible. Glacially induced lateral migration is more probable for ground water with gaseous hydrocarbons.

The huge ice load that produced some 200-300 atmospheres of additional pressure should have inevitably affected the underground hydraulic system. Some immediate effects of unevenly applied recent ice load can be directly measured in drilled petroleum fields. Measurements in boreholes down to the depth of 2.7 km revealed that aquifer porosity in the Arctic is by 5-7% lower than that of the same sandstones in central regions along the transverse Ob river (Fig. 6). This fact is explained by thicker and more persistent glacial ice of the Arctic whereas close to the drift limit on 60° N only a minor ice load was sporadically applied (Gorelov, 1975).

A glacial effect is felt in northward inclination of water/hydrocarbon contacts in petroleum traps. The inclination of petroleum/water contact is also known in petroleum fields of the Norwegian shelves. In that case glacioisostasy was considered responsible for depth changes, tilting of the traps and expansion of formation gas being used for estimates of Pleistocene spillage and loss of petroleum from the traps during the Pleistocene (Zieba et al., 2016).

In an ideal environment the free-water level of a trap is horizontal. In a real situation this level is tilted because the buoyant force is interfered with by the hydrodynamic force. E.g. water/oil contacts may have been tilted up to 1:10 with gas in reservoirs compressed (Forsberg, 1996). Whatever the mechanism, the northward inclination of water/oil interface and piezometric surfaces of water-bearing strata were directly measured in boreholes of all major petroleum fields of central West Siberia in the area of Middle Pleistocene ice sheets (Tsaryov, 1976).

Taking into account the modern data on the ice limits and geographical distribution of the Pleistocene ice thickness (paper 22 in Ch. V) a simpler explanation of the Quaternary separation of the oil and gas reserves in West Siberia is possible. The sharp vertical oscillations of the Arctic part of the Basin on the order of hundreds of metres pressurized by periodical ice load should have caused fast changes of underground thermobaric situation. This would result in vertical shifts of zones of generation of different phases of hydrocarbons with irreversible replacement of the liquid phase by the gaseous phase. Gas volume is influenced by pressure both through compression and dissolution when ice load is applied. Conversely, when the ice load is removed gas volume in aquifers is instantly increased by dilatation and degasification.

Thus, an abrupt drop of formation pressure during ice sheet disintegration would result in mass degassification of ground water similar to the effect of the champagne cork. Recurrent changes of formation pressure due to intermittent ice load of Pleistocene glacials and interglacials should have been by an order higher than the pressure drop connected with the hypothetical Pliocene sea regression. This mechanism is called the `ice pump` (Riis, 1992).

In the view of this simple mechanism, far-distance lateral migration of petroleum is not necessary for spatial separation of oil and gas reserves. Vertical migration of fluids would concentrate gaseous phase closer to the surface with liquid phase lagging in deeper strata. The southern petroleum belt contains mostly oil in shallow anticlines of Lower Cretaceous and Upper Jurassic rocks within the depths of 3 to 1.5 km (Kontorovich et al., 1975). The shallow gas deposits must have already dissipated there through not very reliable seals. In the Arctic predominating natural gas is found in more expressive structural traps sealed by thicker Paleogene clays and monolithic permafrost at depths of 0.7 to 1.2 km whereas oil deposits occur sporadically and much deeper.

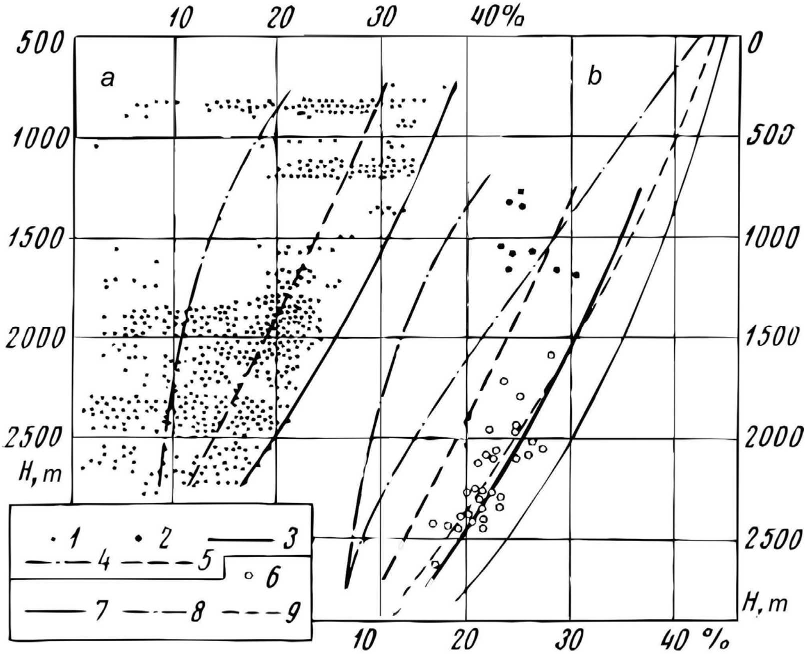

Fig. 6. Open porosity of Cretaceous rocks (percentage) against depth of samples in Arctic West Siberia (a) and in the transverse Ob region (b). Symbols: 1–5 for a, 6–9 for b; 1 – single measurements, 2 and 6 – mean values used for calculating oil reserves, 3,7 – weakly-cemented sandstones, 4 and 8 – clayey sandstones, 5 and 9 – mean porosity of oil-reservoir sandstones (Gorelov, 1975).

It is significant that the Arctic belt of gas fields is located mostly in the area of the last glaciation and of maximum thickness of preceding ice sheets. Therefore the amplitude of oscillations of reservoir pressure was probably highest in the extreme north where it more significantly facilitated periodic degassing of ground water with formation of the huge gas bubbles close to the surface (Fig. 4). This process was not that powerful in the southern belt with lower ice thickness and less frequent ice advances. Besides, the last ice sheet along the drift limit occurred by a 100-150 ka years earlier than in the Arctic. Therefore shallow gas pools of the southern belt had ample time to disappear by gradual dissipation through unreliable seals of the patchy permafrost and discontinuous clay formations.

Thus, the peculiar distribution of oil and gas reserves of West Siberia is probably due to pumping the ground fluid system by recurrent Pleistocene ice sheets which greatly enhanced release of hydrocarbons and their vertical migration to the actual structural traps.

Imbricate mesostructures

General pattern

One of the disputable phenomena of Arctic and Subarctic lowlands is thick imbricate assemblages of distorted and steep-dipping surficial formations of Meso-Cenozoic soft rocks appearing in plan as large bow-like curvatures. These epidermal disharmonic displacements of Meso-Cenozoic rocks known as \`exotectonic disturbances\` (Zakharov, 1968) have been mapped practically everywhere north of the drift limit (Rostovtsev, 1982). In places they are not visible on the surface being covered by thick Quaternary. But more often they are expressed in topography as systems of parallel ridges (e.g. Fig. 4 in paper 25, Ch. V) sculpted by selectively eroded steep-dipping beds of alternating sands, clays, sometimes opokas and diatomites of any age abutting on a deep depression from the concave side (Zakharov, 1968).

In the international literature they are commonly described as composite ridges or hill-hole pairs if such assemblages are steep enough (Aber & Ber, 2007). In West Siberia and in the Pechora Basin such features range from hill-hole pairs first kilometres long (papers 2 and 3 in Ch. I) to composite ridges of record length up to 220 km and width up to 25 km (Fig. 4). The arcs of parallel ridges and hill-hole pairs are generally open to the north, sometimes to the west or east but never to the south (papers 1 in Ch. I, 18 and 20 in Ch. IV). Their map pattern is clearly concentric in relation to the Arctic lowlands and Kara Sea shelf (Fig. 4).

There are several fragmentary descriptions of such imbrications in the literature but the described natural cross-sections are never more than several hundreds of meters long. To better understand peculiarities of epidermal tectonism in West Siberia it seems more productive to consider the structure of the much longer cross-section of well exposed Atlym disturbances of Cenozoic rocks on the Lower Ob river (Fig. 2) which presents an outstandingly instructive case.

Atlym disturbances

A strip of alpine-type disturbances of soft Paleogene and Quaternary sediments up to 25 km wide has long been known in bluffs of the north-eastern bank of the Ob river near Maly Atlym settlement (Atlym in Fig. 2) (Li & Kravchenko, 1959; Nalivkin, 1960). The strip of Atlym disturbances is of considerable interest because of its large size embracing all major types of local Cenozoic formations from Eocene marine clay to Pleistocene diamictons and fluvial sands. This is the reason why they were studied by a profile of mapping boreholes (Fig. 7) which penetrated the disturbed zone down to the horizontally laying strata. The disturbed rocks upstream on the river Ob proceed southwards to the drift limit over undisturbed Paleogene strata. The subsequent study of the Atlym disturbances shed a new light on their origin and on general glaciotectonic situation in the West Siberian Basin (paper 8 in Ch. II).

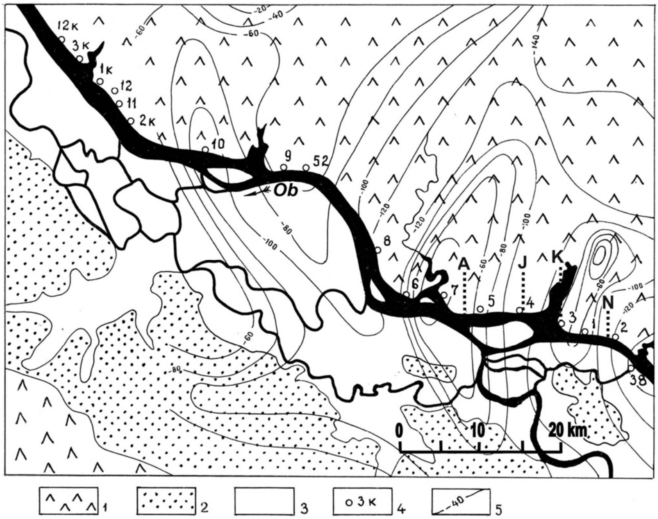

Fig. 7. Location map of Atlym area in river Ob valley (see Fig. 2 for geological background). 1 – interfluvial plain with altitudes 120 to 180 m, 2 – terraces of Ob river; 3 – floodplain, 4 – borehole of Glavtyumengeologia Corporation and its number, 5 – isolines of roof of Tavda clay Formation ($Pg_{2\ 3}tv$); dotted lines indicate end points of detailed profiles A – J (Fig. 8) and K – N (Fig. 9).

The detailed profiles in cross-section (Fig. 8, 9) have not been published previously although they contain a large number of clear indications of alpine-type glaciotectonism endemic for northern West Siberia. The profiles were composed by measuring attitude of contacts and fault planes in the bluffs 15 km long on the northern bank of the river Ob. The measurements were controlled by continuous photography from a boat which was a regular practice before the era of GPS. The measured contacts of the local sedimentary formations have been extrapolated below the river level using cores of boreholes drilled by Glavtyumengeologia Corporation for geological mapping. The resultant generalized profile shows a complicated zone of deep crumpling of the Paleogene strata ca 35 km wide between wells 2 and 10 (Fig. 10).

The most salient features of the cross-section pictured in Fig. 8 and 9 is the dominant western inclination of the structural elements with increase of the dip angles and ratio of shear strain eastwards. The western dips are recorded for overturned folds, thrust planes, imbricated slices and clay injections. The most common features identified by the detailed structural survey are numerous listric faults growing steeper and more tightly spaced upstream, i.e. eastwards. In the westward (upglacier) direction the structure gradually becomes simpler, with gentle slightly asymmetric folds prevailing and only occasional upthrusts and clayey injections breaking through the overlying Oligocene sand (Fig. 10).

In the downglacier direction simple slightly asymmetric folds change into tight recumbent folds with numerous zones of mylonisation and injections of the Eocene clay. At the very end of the detailed profile the intensely deformed sub-vertically dipping and heavily foliated Paleogene rocks are suddenly replaced by subhorizontal strata lying in the normal stratigraphic order upon the Eocene clay which is positioned 250 m below the surface (station 18.4–18.8 km in Fig. 9). About 67° E (right of Fig. 10) the zone of alpine folding abruptly terminates, and farther upstream, i.e. eastwards, the river bluffs display only horizontally layered Upper Oligocene–Lower Miocene sands and rhythmites covered by Quaternary sands and diamicts about 12 m thick. These relations indicate that the increasing downglacier shear strain in incompetent rocks was met by a solid competent dam (paper 8 in Ch. II).

The mean thickness of the Tavda Eocene-Upper Oligocene clay as traced in boreholes along the general cross-section is 170 m (borehole 36). However, upglacier it is getting abnormally thin (~ 100 m) whereas at the downglacier its thickness increases up to 200-250 m (boreholes 1 to 6).

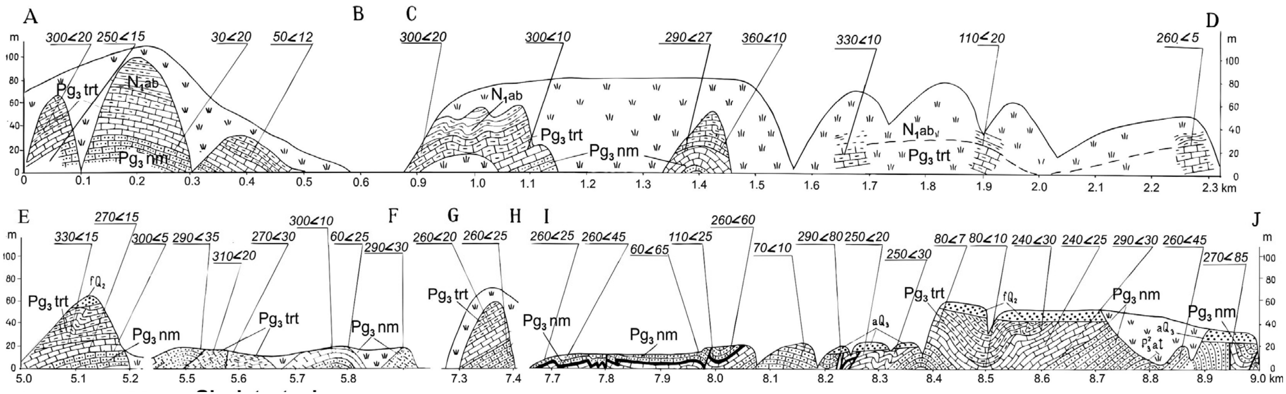

Fig. 8. Detailed geological profiles of right bank of river Ob between 66°46.5′ and 67°4.5′ E. Fig. 8 (A–J) is west of Maly Atlym creek, Fig. 9 (K–N) is east of it (location in Fig. 7). Distances are from point A, altitudes are from low river Ob at 15 a.s.l. Attitudes of structural planes are indicated above the sections as dip azimuth (1st number) and dip angle (2nd number). Symbols: 1 – Quaternary glaciofluvial and alluvial sands, 2 – diamicton of Middle Pleistocene till, 3 – silty rhythmite of Abrosimovka Formation (N_1 ab), 4 – silt-clay rhythmite of Turtas Formation (Pg_3 trt), 5 – silty clay interbedded with sand and lignite of upper Novomikhailovka Formation (Pg_3 nm), 6 – sand with clay interlayers of lower Novomikhailovka Formation (Pg_3 nm), 7 – sand of Atlym Formation (Pg_3 at), 8 – shaly clay with siderite lenses of Tavda Formation ($Pg_{2\ 3}$ tv), 9 – lignite marker, 10 – faults: a – measured, b –inferred.

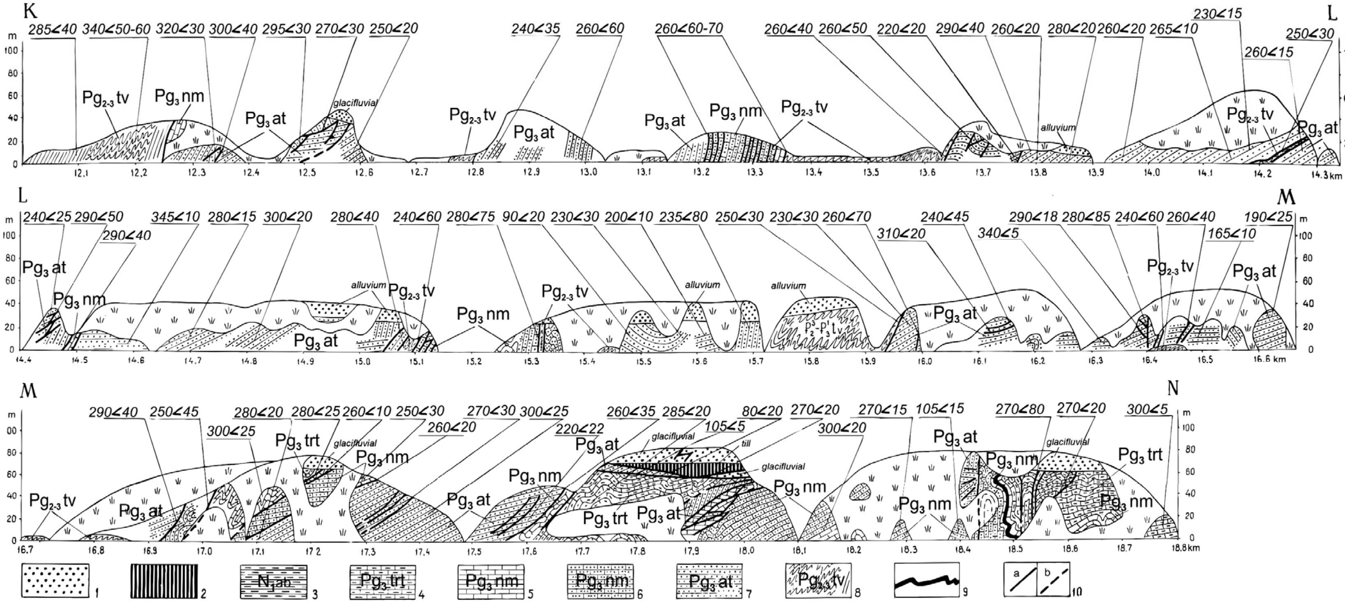

Fig. 9. Explanations are in Fig. 8.

This feature looks as a result of extension upglacier and injection of pressurized clay downglacier. In borehole 3 the deformed Eocene clay with slickensides was observed down to the depth of 310 m below the river level (Fig. 10). Therefore, the entire zone of the disturbed soft rocks is ~ 400 m thick which is probably the world record for epidermal glaciotectonism. The underlying Paleocene and Mesozoic strata beds down to the Paleozoic basement, which is 2.8 km deep, in seismic profiles are basically horizontal with only minor undulations (Li & Kravchenko, 1959).

Discussion

The epidermal alpine-type deformations of West Siberia on account of their size have been perceived by some geologists as rootless neotectonic structures and manifestations of deep-seated tectonism (e.g. Krapivner, 1986). However, after many drilling projects and seismic profiling of 1960s-80s it became clear that these arches reflect disharmonic folds and thrusts accompanied by large detached blocks of soft rocks and usually cannot be traced deeper than 200-300 m (e.g. Arkhipov et al., 1976, 1986; Astakhov, 1986, 2011).

The disharmonic disturbances of West Siberia are structurally very similar to ice-pushed ridges or thrust moraines in glaciated sedimentary basins of northern Europe and America where they are related to tangential glaciotectonism, i.e. to subglacial deformation of soft rocks (Levkov, 1980; Aber & Ber, 2007). The most revealing is their orientation which is discordant to the Mesozoic-Paleogene structural pattern but quite consistent with the periphery of reconstructed ice sheets in the arctic lowlands and on the Kara Sea shelf (Astakhov, 1977, 1986). Thus, the glacial origin of epidermal disturbances in Cenozoic rocks of northern West Siberia seems highly probable.

The largest and best exposed Atlym disturbances can shed more light on glaciotectonism in West Siberia. Originally the geologists described the structure as basically folded with complications of minor faults. They explained the Atlym disturbances either by unspecified glaciotectonism (Li & Kravchenko, 1959) or were cautious about genetic implications (Nalivkin, 1960). A dissenting view imagined this crumpling of Cenozoic rocks as produced by a deep-seated slip-strike fault (Krapivner, 1986). The latter idea was not adopted by local geologists because numerous seismic exploration profiles clearly demonstrated that the deeper Jurassic-Cretaceous sedimentary strata were horizontally lying upon the Paleozoic basement with very rare faults (Kontorovich et al., 1975).

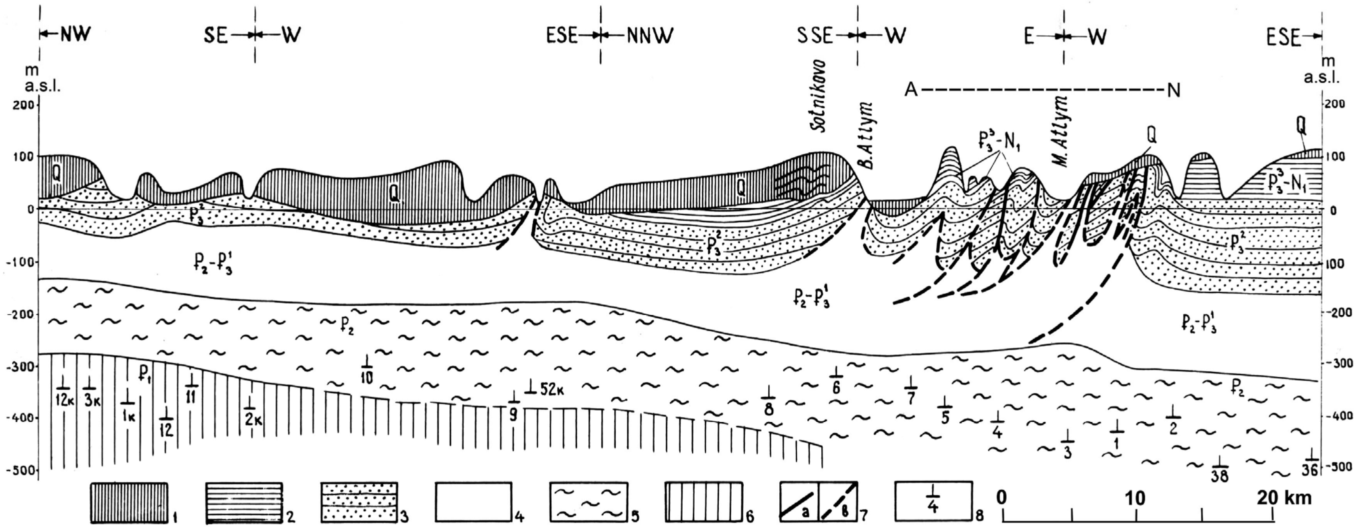

Fig. 10. Synthetic profile of Cenozoic formations along the Lower Ob river based on Fig. 8 and 9 cross-sections controlled by boreholes between 65°40′ E/62°40′ N and 67°15′ E/62°04′ N (modified after Astakhov et al., 1996). Terrestrial formations: 1 – Quaternary deposits, 2 – Upper Oligocene–Lower Miocene silt-clay rhythmites, 3 – Middle Oligocene sands with silty clay and lignite interlayers. Marine formations: 4 – Upper Eocene-Lower Oligocene shaly clay with siderite lenses, 5 – Eocene opokas, diatomites and clays, 6 – Paleocene clay. 7 – faults: a – observable, b – inferred, 8 – bottoms of numbered boreholes. Broken line A–N indicates location of detailed profiles.

The crucial feature for understanding the origin of the Atlym disturbances is the difference of the deformation styles of the mechanically diverse Paleogene rocks which is evident in minor structural complications of the maximally deformed downglacier part of the Atlym profile. The Middle Oligocene Atlym sand displays only brittle deformations without minor folds or crenulation or any other ductile structures. Commonly it is broken into sharp-edged blocks divided by listric upthrusts, sometimes stacked into a pile of `chips` (Fig. 11, i).

The extreme brittle deformation is observable in typical fault breccia consisting of small angular splinters (Fig. 11, ii). Such deformations normally develop in competent rocks such as cemented sandstones whereas presently the Oligocene sand is completely loose. In this case the only possible cement during the deformation was ice. This is also suggested by the 120 m long wholesome raft of the Middle Oligocene sand at the below the basal till exposed upon the distorted Upper Oligocene rhythmites (station 17.8 km in Fig. 9). The presently thawed sand raft, piggyback riding on the younger sediments, shows no strong internal deformation and was probably incorporated into glacial ice prior to the maximum compression.

More ductile deformations are visible within the overlying Novomikhailovka sand, silt, clay and lignite interbedded together with Turtas and Abrosimovka silt/clay rhythmites. In the downstream, i.e. upglacier end of the profile (Fig. 8) these rocks commonly occur in harmonic, slightly asymmetric folds eastwards changing into overturned folds (Fig. 11, iii). Farther eastwards subhorizontal stress obviously increases to produce recumbent folds with crenulated bedding and listric upthrusts (Fig. 11, iv), or even glaciotectonic mélange (Fig. 11, v). These formations also must have been frozen as indicated by their position atop of the perennially frozen Middle Oligocene sand (paper 8 in Ch. II).

At last, the extreme deformations are displayed by the Eocene montmorillonite marine clay of the Tavda Formation which is everywhere exposed in tabular injections up to 200 m thick (station 12-12.2 km in fig. 9) or clastic dikes 0.3-1 m thick (station 14.2 km). The Tavda clay pierces all younger Oligocene formations which is visible from colour difference between the greenish clay with brown ciderite lenses and white sand of the overlying Middle Oligocene Atlym Formation (Fig. 11, vi).

Such protrusions could develop also in thawed clays of which the thin apophyses seemingly indicate. But in this case it is unlikely because the internal structure of the injected clay is not just flow foliation but also crenulation cleavage. Second, the clay dikes contact with the surrounding sand by flat slickensides with practically no sand xenoliths incorporated into the protrusions. And third, the pulled apart siderite bands in several places are bent into small tight folds which are normally formed not in liquid but in relatively competent medium to translate the stress necessary for loaded flexion of hard rock (paper 8 in Ch. II).

These and other similar features tell us that the Tavda clay was also frozen when it pierced through the overlying hard-frozen sandy sediments. Water saturated clayey sediments can retain their plasticity in a wide range of negative temperatures due to preservation of combined water in liquid state. The elastoviscous behaviour of the frozen clay is normal because it always retains interstitial liquid water. Its shear strength more time-dependent that that of crystalline ice and is reduced by an order when prolonged stress is applied. Numerous experiments indicate that frozen clay at least up to -7°C shows dynamic viscosity lower than pure ice, i.e. after initial resistance is overcome the stressed frozen clay would creep faster than pure ice (Tsytovich, 1973; Williams & Smith, 1989). The implication is that the maximum deformation of the glacier/frozen clay couplet would occur below the ice/rock interface with resultant squeezing out of the clay upwards and with the glacier floating upon a thick clay pillow (paper 8 in Ch. II).

Fig. 11 (next page). Deformation structures of Tertiary sediments in northern bank of river Ob near Maly Atlym settlement; one metre shovel for scale. i) Upthrusted chips of the Middle Oligocene Atlym sand; station 12.5 km in Fig. 9. ii) Fault friction breccia of silt splinters in the footwall block under a listric upthrust in Atlym Middle Oligocene sand; station 12.5 km in Fig. 9. iii) Middle Oligocene Novomikhailovka silt and sand interbedded in disharmonic fold overturned eastwards; station 1 km in Fig. 8; section ~12 m wide. iv)**.** Middle Oligocene Novomikhailovka Formation (A) with lignite bed (a) upthrusted along a listric fault over Upper Oligocene Turtas rhythmites (B) with a drag fold in the footwall block (b); 5.5 km in Fig. 8. v) Glaciotectonic mélange of the folded Novomikhailovka sands, silts and clays interbedded; station 9 km in Fig. 8. vi) Sub-vertical clastic dyke of the Eocene Tavda clay with siderite lenses (A) piercing overlying Middle Oligocene sand of Atlym Formation (B); C are parallel clay apophyses, station 15.3 km in Fig. 9. Top right is person for scale.

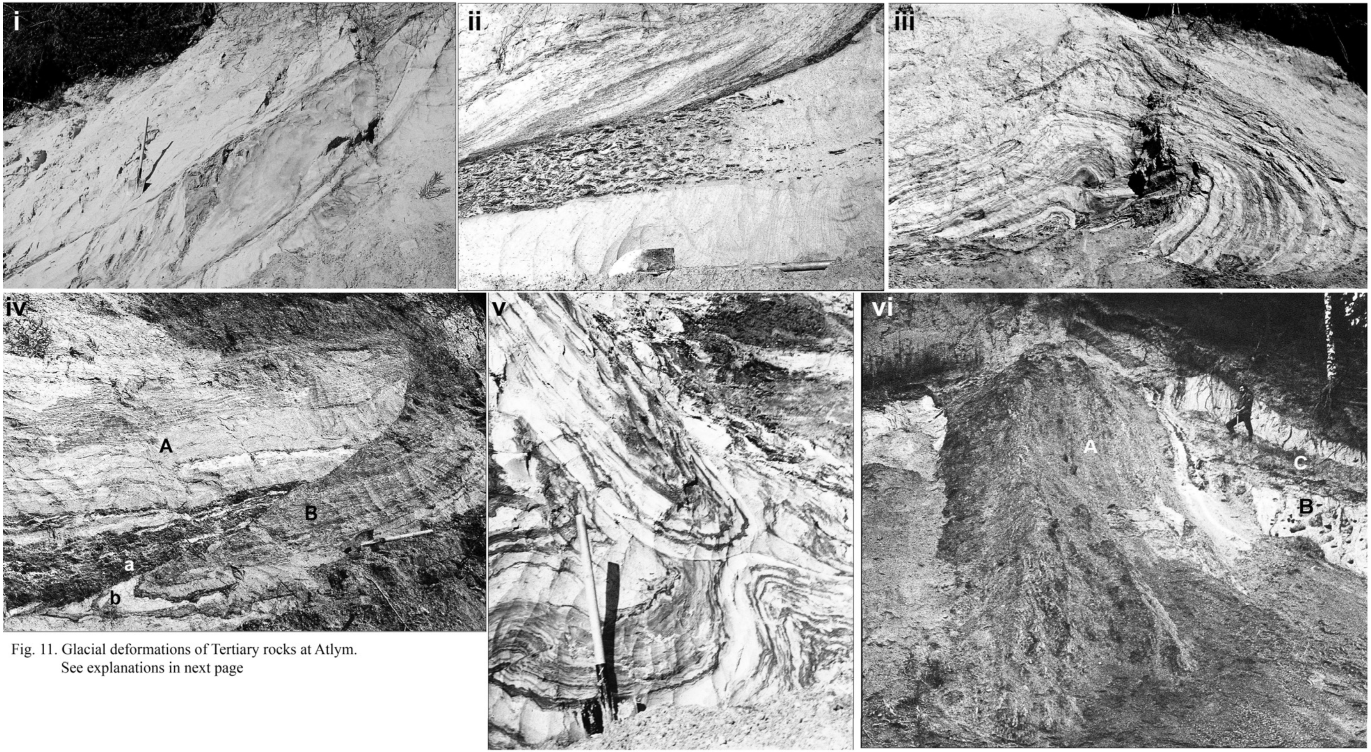

Fig. 11. Glacial deformations of Tertiary rocks at Atlym.
See explanations in next page

See centrefold for this image in colour.

Model of Atlym disturbance

The observable mode of deformation of the Paleogene sediments in the Ob valley is markedly different from what might be expected from their mechanical properties in the present thawed state. Therefore the glacier bed at Atlym must have been frozen during the deformation. This means former perennially frozen substrate ~300 m thick under a thick Middle Pleistocene glacier ~300 km up-ice of the drift limit. The development of the Atlym disturbances with possible sedimentological and structural consequences are discussed previously (paper 8 in Ch. II) are briefly listed below using the model from this paper (Astakhov et al., 1996) (Fig. 12).

Stage 1 pictures a glacier advance over perennially frozen Paleogene sedimentary substrate initially supporting not very thick ice. At *stage 2* with ice thickening the shear resistance limit is first overcome in less competent clay formations below the base of the ice. The *dynamic sole* of the glacier – a subhorizontal zone in which the maximum deformation occurs – is established in this thick clay formation. This means that the sliding component of ice advance plunges into the frozen clay which is capable of faster deformation than ice itself. The monocline of the Oligocene frozen sand provides two different situations: i) upglacier where the thinner and brittle sand over extending clay and under the increasing ice load breaks to produce slabs and ii) downglacier where the flow of compressed clay meets a competent dam in the form of the thicker sand.

Stage 3 accounts for the maximum ice thickness when the dynamic sole was already firmly established within the underlying frozen clay. The viscoelastic creep of the sheared frozen clay takes most of the basal stress and supports movement of the glacier with the frozen sand attached to the base. The breakage of the competent sand roof terminates in the eastern end of the profile (Fig. 10) where the east-dipping sand formation is thick enough to resist any glacial stress. This implies a sudden reduction of the effective velocity of the clay creep and its pillow-like thickening against the sand obstacle with steep listric thrusts penetrating into the ice/sand couple. Kinematically it would mean a leap of the dynamic sole upwards into media with another low shear strength, i.e. into pure solid ice. The underground dynamic sole in the form of the creeping clay transforms into a plane of décollement in the glacier ice, over which the thick glacier slides as a whole block. Thus, the dynamic sole is split into two levels: subglacial, within the frozen clay, and englacial, within the ice above.

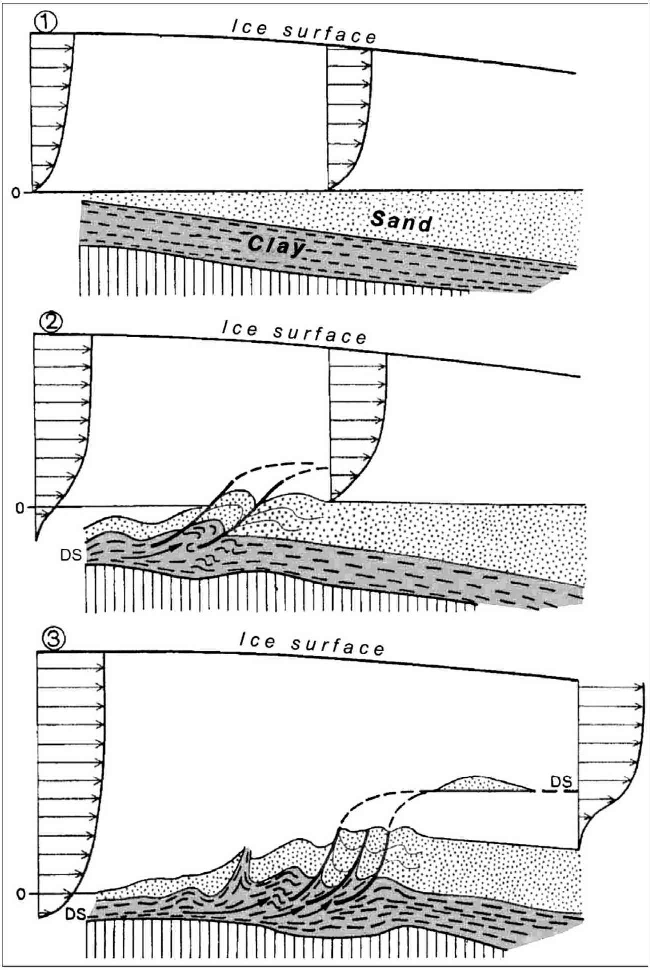

Fig. 12. Idealised stages of deformations of frozen sand and clay under thickening ice sheet as inferred from the Atlym disturbances. Parallel arrows show vertical distribution of summary velocities within ice/permafrost couplet. DS is dynamic sole of the glacier (Astakhov et al., 1996).

A likely result of the higher position of the dynamic sole is englacial sliding of large rafts of frozen sand. Gradually descending, such rafts would eventually join the basal debris-laden ice closer to the drift limit resulting in their stagnation and deposition together with frozen diamicts, according to Lavrushin`s model (1976). Rafts of Paleogene sand are found atop of diamict formations on the Irtysh river ~200 km to the SE (Kaplyanskaya & Tarnogradsky, 1974).

The purpose of the model in Fig. 12 is to explain major glaciotectonic disturbances within the thick subglacial permafrost which are positioned far upglacier. The disturbances of the Atlym type, namely the great width and the record depth of the distorted zone, demand high hydrostatic pressure of ice at least 1 km thick. In this case rugged subglacial topography is not necessary for large-scale deformations of the substrate and entrainment of large erratic blocks. What really necessary is downglacier decrease of ice thickness providing deviatoric stress at the ice base plus rheologically divergent sedimentary formations which are available everywhere in the West Siberian Basin. In the Atlym case an important role was played by the monoclinal structure of the Tertiary rocks, which provided the downglacier thickening of the competent sandy sediments and termination of clay creep by a frozen sand dam.

In general, within the realm of cold-based glaciers the instability of the dynamic sole of the glacier caused by diverse rheologies of the substrate seems the main reason for detachment and downglacier transportation of large slabs of frozen sediments. Thus, a dry-based glacier can erode and tectonise large volumes of sedimentary rocks by splitting its dynamic sole which may vaccilate between its englacial, basal, and subglacial positions (paper 8 in Ch. II). Judging by the numerous hill-hole pairs of West Siberia the mechanism of glaciotectonic excavation and upthrusting exemplified by the Atlym disturbances can readily account for the wide stratigraphic gap with the lack of Neogene and Upper Paleogene formations (Fig. 2) eroded in Arctic and Subarctic West Siberia and transformed into thick sheets of fine-grained diamictons.

Conclusions

– Ice Age tectonism is the final and important structural stage in the history of a great sedimentary basin and its marginal highlands.

– Glaciotectonism is a principal force constructing major landforms of the West Siberian Plain and its borderlands.

– Glaciotectonism by cold-based ice sheets applied to perennially frozen substrate is the main erosive agency responsible for the stratigraphic gap by removal of large volumes of Tertiary sedimentary rocks from the northern part of the Basin.

– Glaciotectonism is a crucial factor of the recent redistribution of oil and natural gas reserves of the northern part of the Basin.

Acknowledgements

This study was supported by the St. Petersburg State University. The original field data were acquired by the author during his work for the Geological Survey of the USSR. The digital relief map is kindly provided by the Centre for Remote Sensing of Russian Geological Research Institute (VSEGEI), St. Petersburg.

SUMMARY

The research performed over several decades is presented in this book in order to elucidate dim points of the glacial history of northern Russia. The conclusions herein are based on the long-term integrated study of diverse phenomena such as glacial and aeolian sediments and landforms, old permafrost with buried ice bodies, epidermal deformations, distribution of petroleum reserves, organic fossils, Paleolithic sites, etc. All these phenomena are recorded in the thick polymict formation of the northern Pleistocene. The most meaningful inferences can be found in papers 5, 8, 21, 20, 22, 23, 26. They boil down to the following.

1) The Eurasian margin outside the limits of Fennoscandian glaciations was repeatedly invaded by Quaternary inland glaciers up to 3.5-4 km thick. They advanced from the arctic shelves, where the feeding ice domes resided, down to 59-60°N. Their behaviour only in the west of European Russia was similar to the classical mode of wet-based fast-moving glaciers of Europe and North America. Most of northern Russia featured a peculiar, Siberian type of inland glaciation conditioned by the thick, historically stable permafrost and invariably continental climate.

2) The western influence was felt mostly during two last short interglaciations when Atlantic sea water penetrated far into Central Siberia via deep glacioisostatic depressions along the arctic continental margin. These events, called the Northern and Boreal transgressions in European Russia, Kazantsevo and Karginsky transgressions in Siberia are correlated with marine isotope stages 7 and 5 respectively.

3) However, the persistent continental climate played a decisive role guarding the great stock of cold accumulated in the lithosphere by the ice ages which could not dissolve during interglacial episodes. The basically cryoarid environment facilitated not frontal ice retreat but areal deglaciation when thick sluggish glaciers turned into stagnant ice masses. Only the initial ice disintegration was rapid due to the rise of sea level. The following slow glaciokarst degradation of stagnant ice sheets was impeded by the cold climate and subglacial permafrost which effectively protected stagnant glacial ice from basal melting.

4) As a result, stagnant ice buried by glacial debris and partly stored in the perennially frozen glacial sediments did not melt out for tens of thousands years. The typical subglacial deposits in arctic Siberia are not

tills of Atlantic type but ice/diamicton couplet with detached sedimentary blocks. Fossil glacier ice persisting in the present permafrost is an obvious indication of a slow tempo of environmental changes in northern Eurasia where many features are inherited from the Ice Age.

5) The Siberian type of glaciation was especially evident in the last glacial cycle which started in the east of the High Arctic to produce the Early Weichselian ice sheet on the Kara Sea shelf (90-80 ka BP by optically stimulated luminescence dating). With the growing aridity the ice accumulation zone moved southwestwards in the form of the Middle Weichselian ice sheets (60-50 ka BP by OSL dates). During the final, driest episode (33 to 13 OSL ka BP) the precipitation level was sufficient only in the far west where the Barents Sea ice sheet coalesced with the Late Weichselian ice sheet of Fennoscandia. East of that zone the snow accumulation was so meager that noticeable glaciers could not develop even in the mountains of the Urals and Central Siberia.

6) The northward drainage of the great Siberian rivers was completely blocked in the early glacial episodes but was not obstructed during the coldest part of the glacial cycle contemporaneous with the last glaciation of Fennoscandia and MIS 2.

7) During the last glacial cycle very thick permafrost with cloudless skies, short warm summers and dominant aeolian activity reigned all over the Russian subcontinent. The aridity of the environment was growing to reach the maximum during the last and coldest episode roughly synchronous with the Late Weichselian of Europe. The paleoenvironment was beneficial for cryoxeric biota of the mammoth tundra-steppe and for initial migration of *Homo sapiens* northeastwards (40–28 thousand radiocarbon years BP). Later the climate became even colder and drier but still tolerable for the mammoth megafauna which in places survived up to the late Holocene.

8) The Pleistocene glaciation made up the last tectonic stage in the history of arctic and subarctic terrains. The continental ice masses fused together with perennially frozen substrate moved slowly but were able to exert great pressure on the soft substrate to produce unusually large and deep glacial disturbances eroding and transporting large volumes of subglacial sediments. The thick ice sheets of arctic origin profoundly influenced the geological structure leaving large tracts of alpine-type distrurbances 300-400 m deep in the sedimentary basins and isostatic troughs with compensation uplifts in the form of the northern Urals and Central Siberian uplands. The intermittent load of inland glaciers kilometres thick played the role of the `ice pump` in vertical and lateral redistribution of petroleum deposits (paper 26).

Colour Centrefold

Paper 18

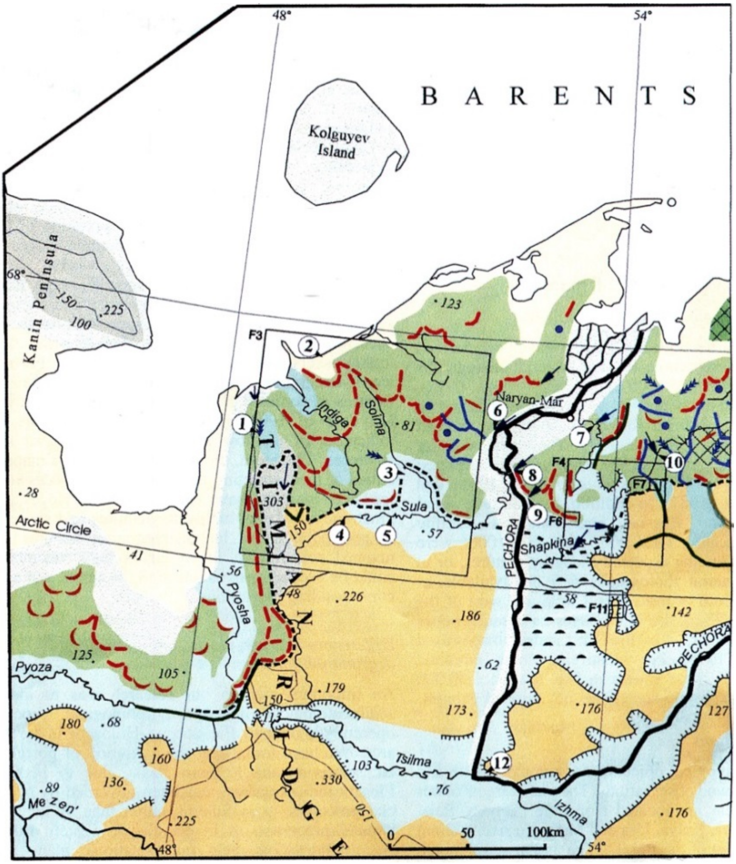

Fig. 2. Map of glacial and periglacial features. Non-studied areas are shown in grey shades, depending on altitudes. Rectangles with numerals relate to figures in the text. The broken line – inferred ice-sheet limit named the Markhida Line, i.e. the southern boundary of hummock-and-lake landscapes of Markhida, Harbei and Halmer types. Black arrows with circled numerals – key sections in Mangerud et al. (1999): 1 – Harius Lakes; 2 – Timan Beach; 3 – Urdyuzhskaya Viska; 4 – Sula sect. 7; 5 – Sula sect. 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: 21 – Byzovaya; 22 – Pymva-Shor; 23 – Mamontovaya Kurya. Sites on the Yamal Peninsula according to Gataullin and Forman (1997) and Gataullin et al. (1998): 24 – Marresale; 25 – Mutny Mys.

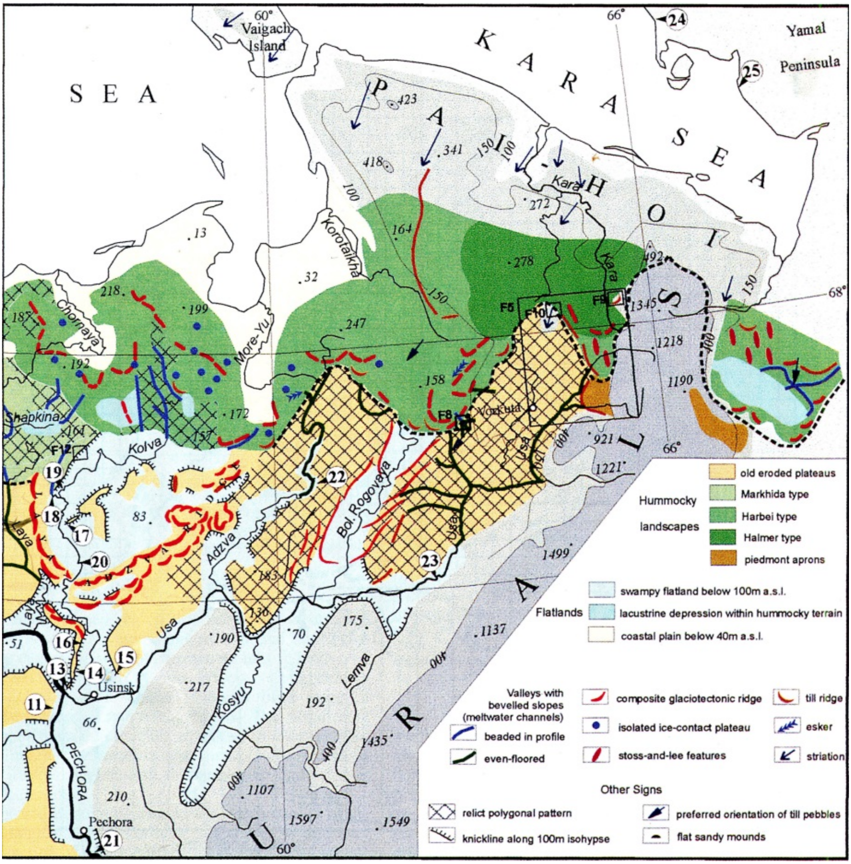

Fig. 2. Continued

Paper 20

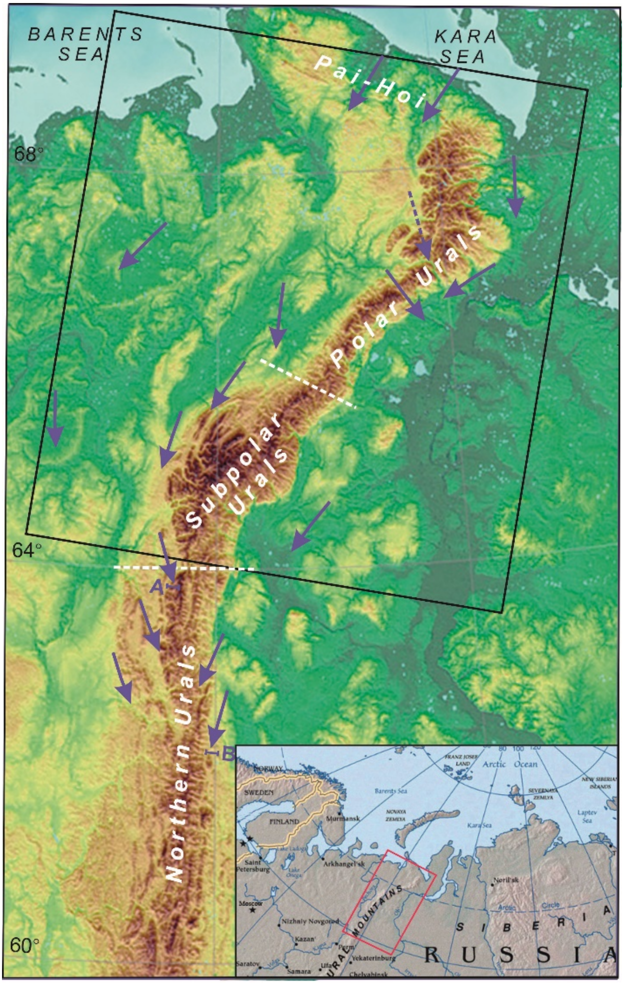

Fig. 1. Orographic subdivisions of the northern Urals with main directions of Middle Pleistocene ice motion by various authors. Broken arrows are ice flows inferred from geomorphic directional features; solid arrows – boulder trains. Short bars A and B indicate profiles in Fig. 2.

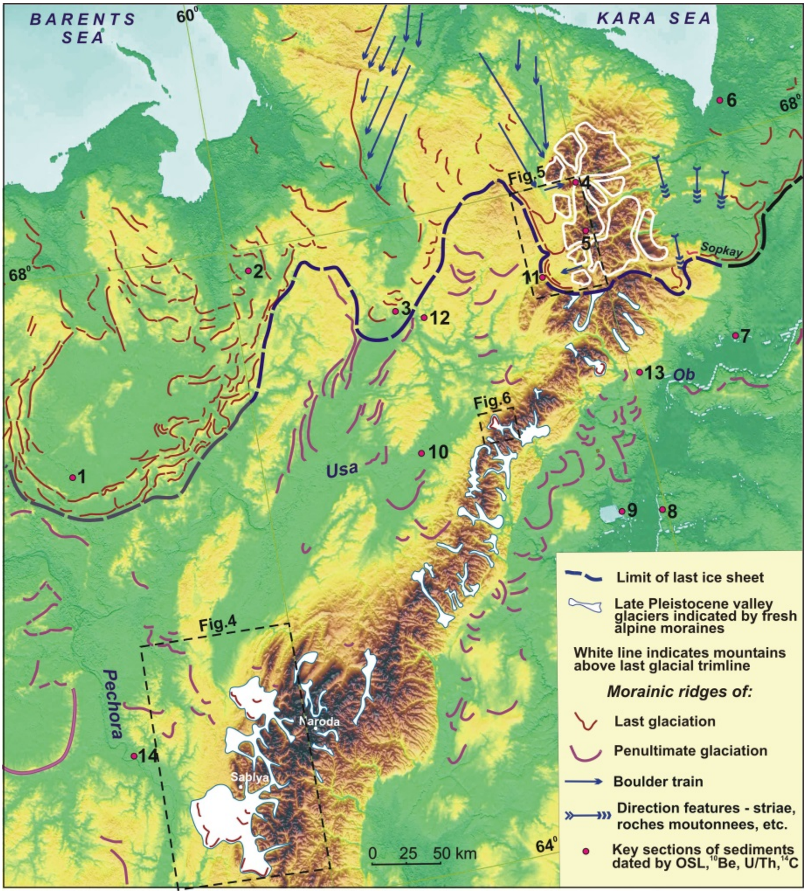

Fig. 3. Principal Pleistocene glacial features of the Polar and Subpolar Urals based on interpretation of aerial and satellite images and ground truthing by the Russian-Norwegian PECHORA project (modified after Astakhov, 2011 and maps at http://booksite.elsevier.com/9780444534477). Numbered stratigraphic sites: 1 – Podkova (Mangerud et al., 1999); 2 – river More-Yu (paper 19, Ch. IV); 3 – river Ileymusyur (Mangerud et al., 2004); 4 – river Bol. Kara (Nazarov et al., 2009); 5 – Chernov glacier (Mangerud et al., 2008); 6 – river Yerkata (Mangerud et al., 2004; Astakhov, 2006a); 7 – Sangompan and 8 – Pitlyar (paper 14, Ch. III); 9 – Shur peat (paper 12, Ch. III); 10 – Meskashor and 11 – river Bol. Usa (Svendsen et al., 2014); 12 – river Seyda (paper 3, Ch. I); 13 – river Hanmei (paper 14, Ch. III), 14 – Byzovaya Paleolithic site (Heggen et al., 2010). See site 11 also in Figs 5 and 7.

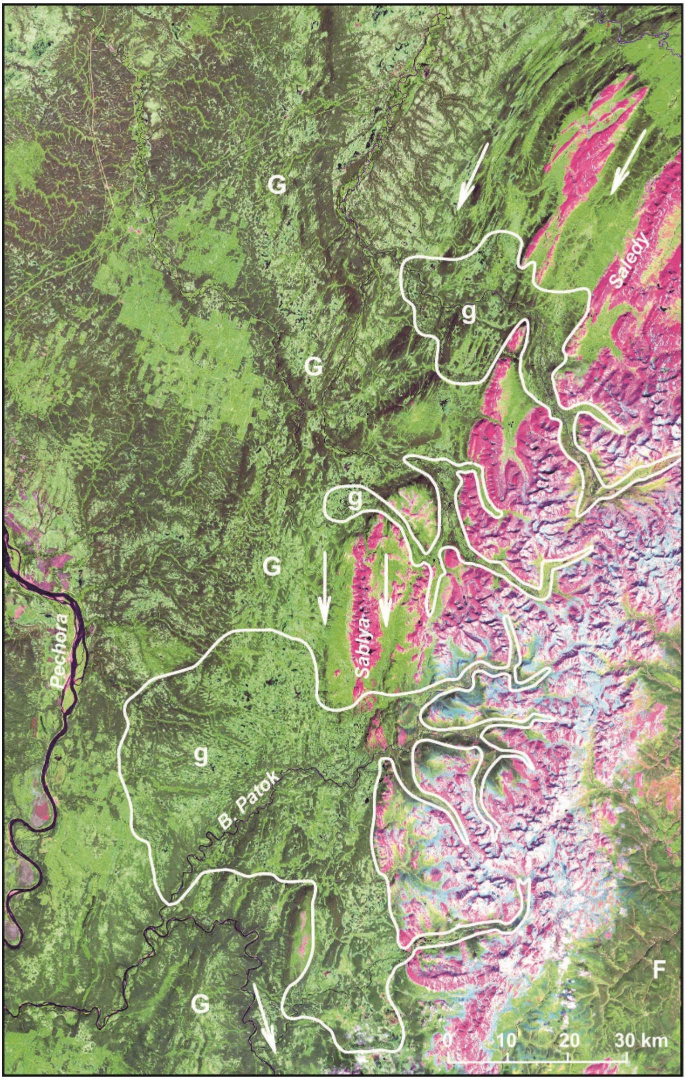

Fig. 4. Subpolar Urals in Landsat image. Forested slopes are dark green. G – older generation of hummocky moraines; g – fresh morainic landscape of smaller morainic hummocks and ridges; F – heavily dissected landscape of fluvial erosion without visible glacial features.

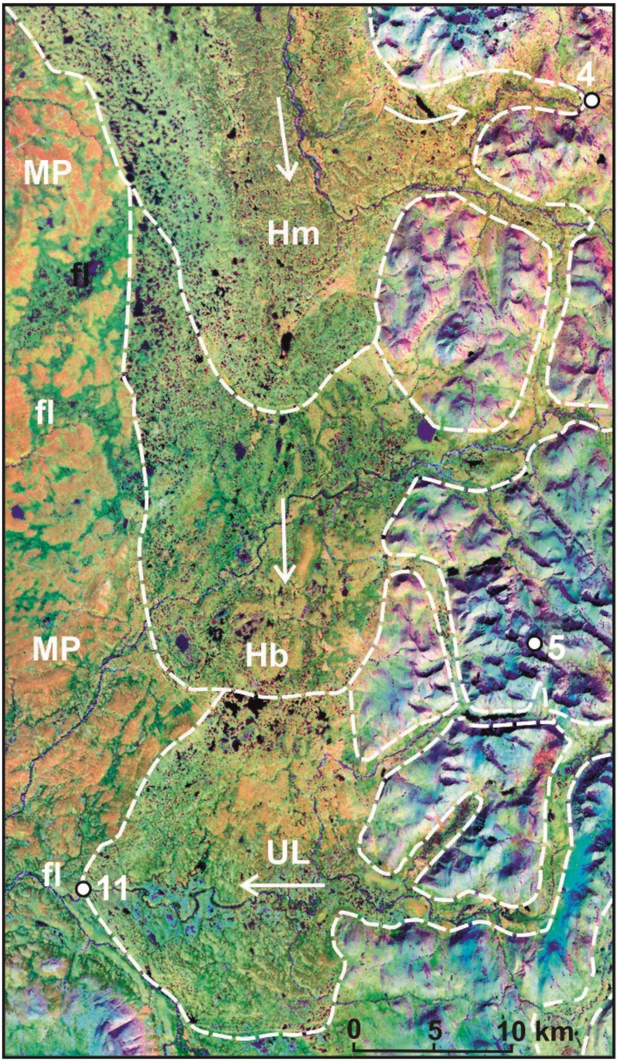

Fig. 5. Moraines of main ice lobes of the last glaciation (outlined by broken lines) in the western piedmont of the Polar Urals north of 67° N as seen in Landsat satellite images. Hm – fine-gravel hummocks of Halmer ice lobe; Hb – gravel moraines of Harbei lobe; UL – boulder moraines of Usa lobe; MP – Middle Pleistocene moraines; fl – outwash (modified after paper 18, Ch. IV). Arrows indicate inferred ice flow directions. 4, 5 and 11 are dated sections in Fig. 3.

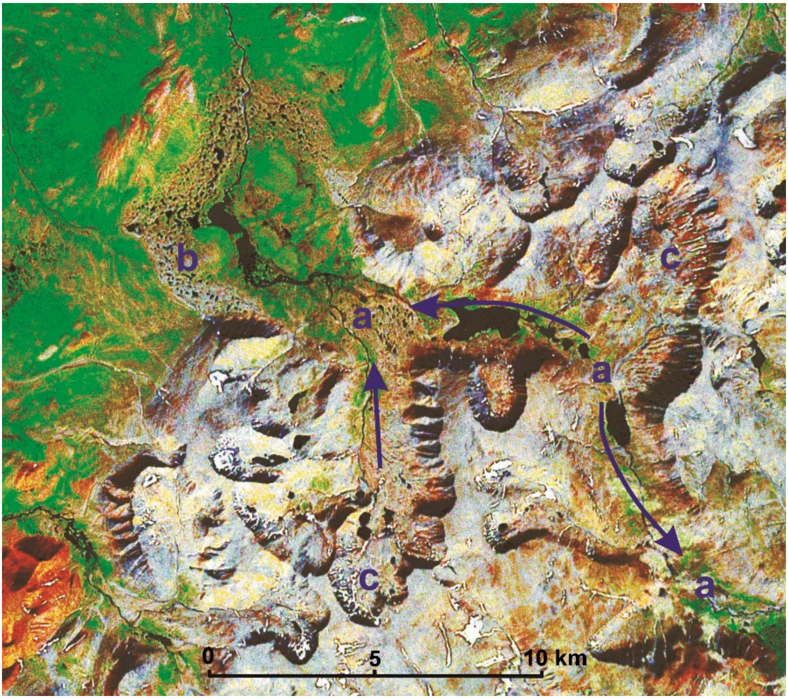

Fig. 6. Satellite view of Hoila transfluent glacial trough (a–a) terminating at the morainic loop (b) on the western piedmont, Arctic Circle. Ice-flow directions are marked by arrows, feeding cirques by (c). See Fig. 3 for location.

Paper 22

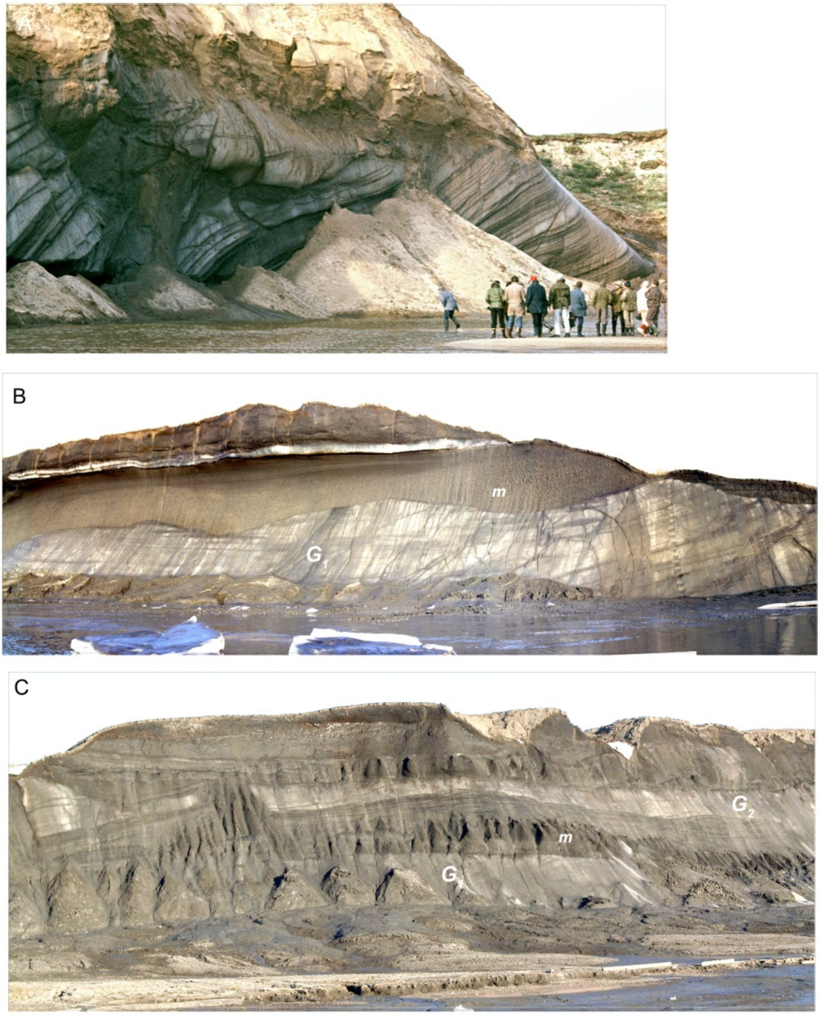

Fig. 6. Subglacial deposits in arctic Siberia. A. Deformed fossil ice with dirt bands, small faults and thrusts covered by melt-out silt in the Gydan Peninsula, 70° N/75° E (Yu in Fig. 1). B and C. Easternmost and oldest fossil glaciers exposed in coastal cliffs of the New Siberia Island, MB in Fig. 1. B. Glacier ice of Middle Pleistocene age (G_1) overlain by marine clay (m) with distinct thaw contact. C. Same members as in (B), overlain by another sheet of fossil glacial ice (G_2). Bluffs are 35 m high. Photos are by V. Tumskoy.

Paper 23

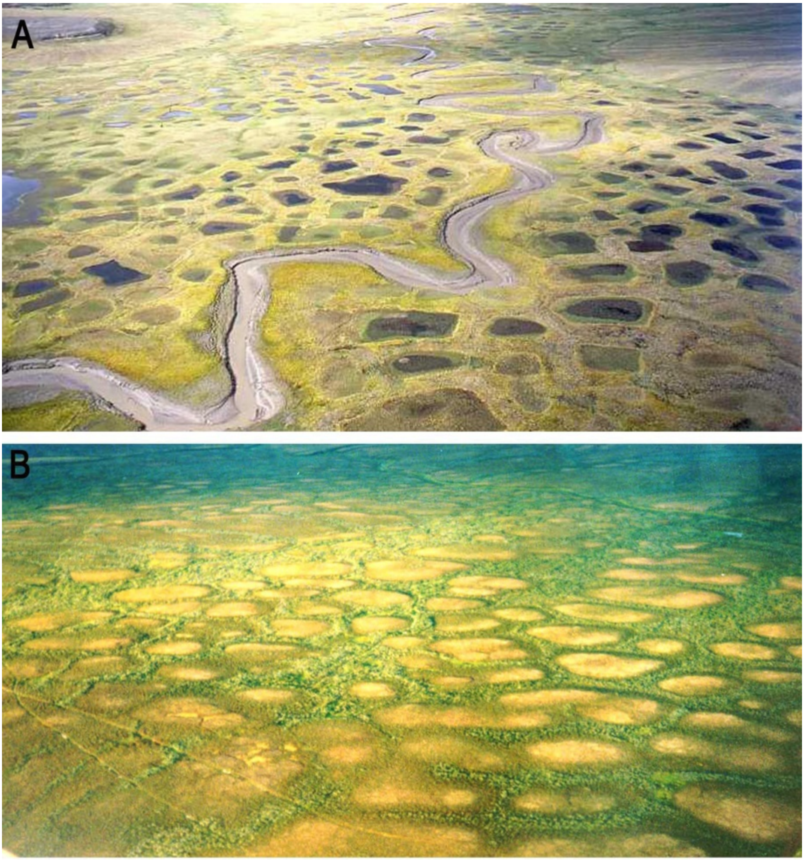

Fig. 3. Helicopter view of tundra polygons bounded by ice wedges. A. Concave active polygons in the Yamal Peninsula (photo by Ó. Ingólfsson https://notendur.hi.is/oi/ siberia_photos.htm). B. Convex relict polygons south-west of the Vorkuta city, Cisuralia. Green stripes are shrubs along the bounding relict ice wedges (Astakhov & Svendsen, 2011).

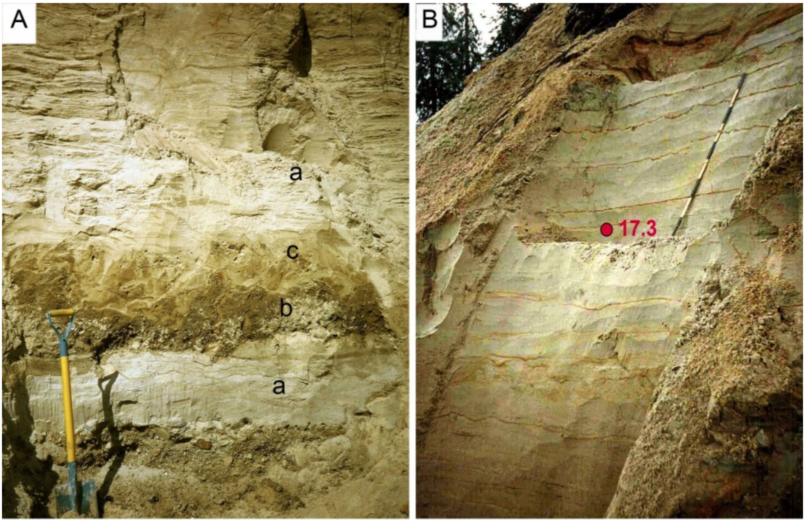

Fig. 7. Coarser subaerial sediments of the Pechora Lowland. A. Subaerial sequence at Akis (A in Fig. 1): aeolian sand (a) containing a wisp of soliflucted diamict (b) with arid soil (c) on top (photo by author). B. Aeolian medium to fine sand on river Vonda (V in Fig. 1). Parallel rusty bands are casts of former veins of segregated ice formed at the base of each subsequent active layer syngenetically with accretion of sand. Red number is OSL age, ka BP. 1 m ruler for scale (photo by O. Nikolskaya).

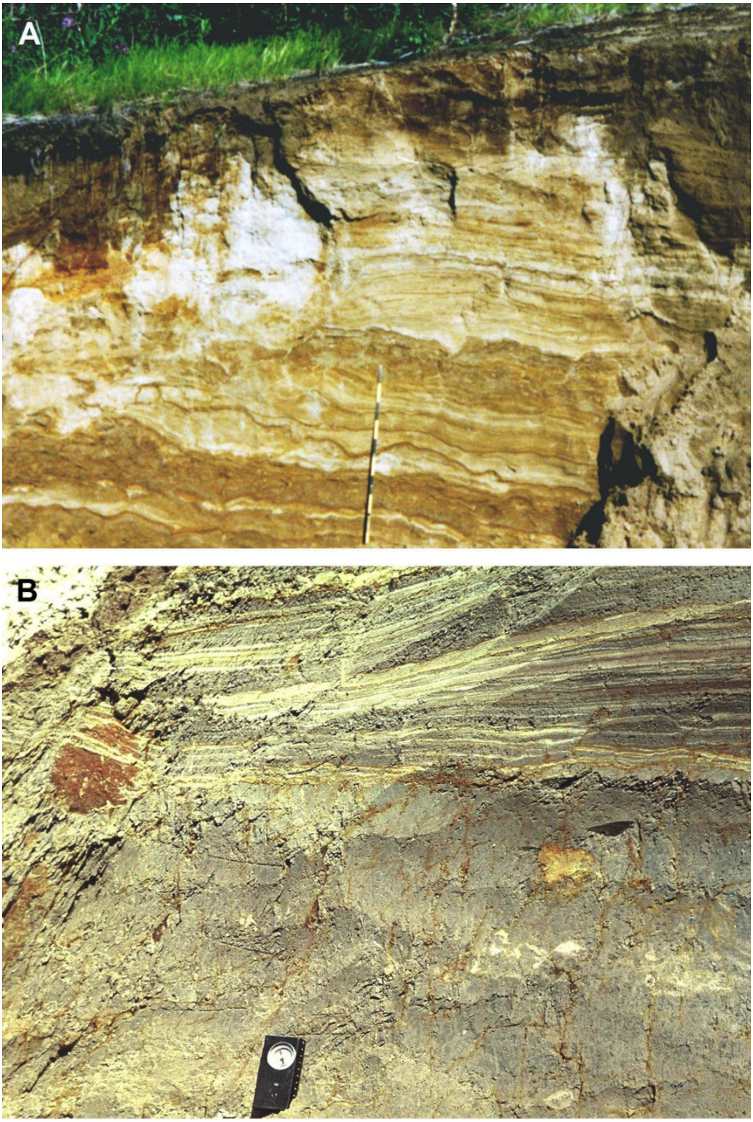

Fig. 8. Fine-grained subaerial sediments in northern flatlands. A. Niveo-aeolian coversand with wavy lamination on a high terrace of river Kolva, Pechora Lowland, 1 m stick for scale (Astakhov & Svendsen, 2011). B. Loess-like silt (bottom) with coversand (top) at Kiryas on the River Ob (Fig. 13, LKM in Fig. 1).

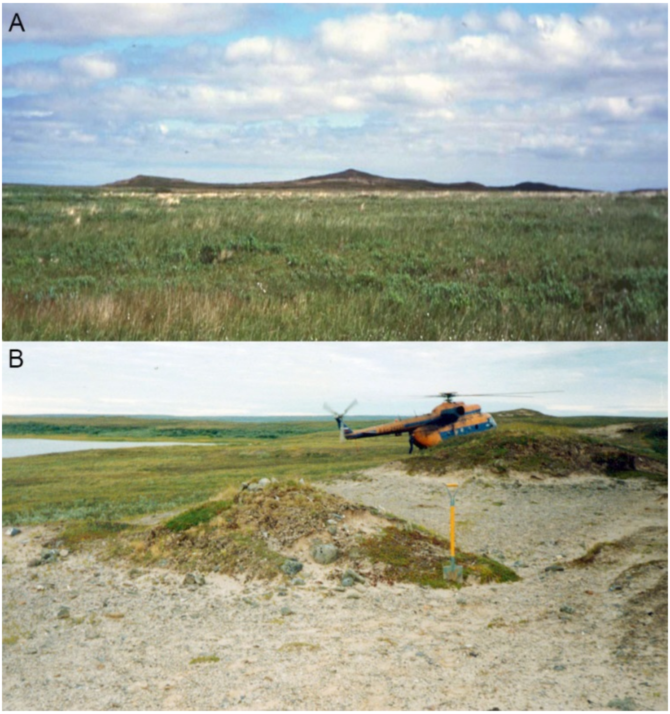

Fig. 11. Deflation residuals in arctic European Russia (Astakhov and Svendsen, 2011). A. Isolated conical hillocks west of the Pechora mouth (kames by Lavrov & Potapenko, 2005). B. Small conical knolls armoured by cobbles and gravels along the crest of the glaciotectonic ridge at Vashutkiny Lakes, 67°57´ N/61°40´ E.

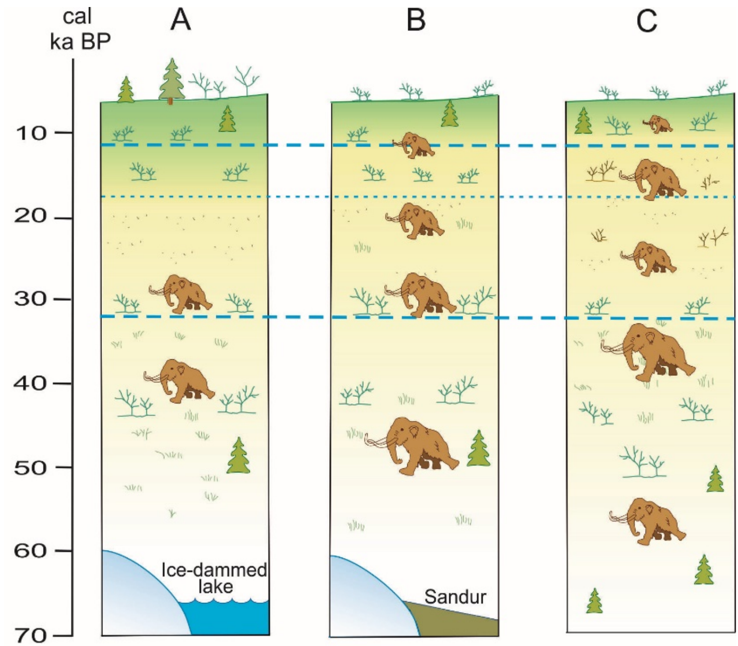

Fig. 16. Sketch of evolution of postglacial environments in northeastern European Russia (A), northern West Siberia (B) and northern East Siberia (C). Density of yellow colours depicts increasing aridity of periglacial landscapes; greenish shades designate relatively mesic environments. Blue lines are best dated correlation levels.

Paper 25

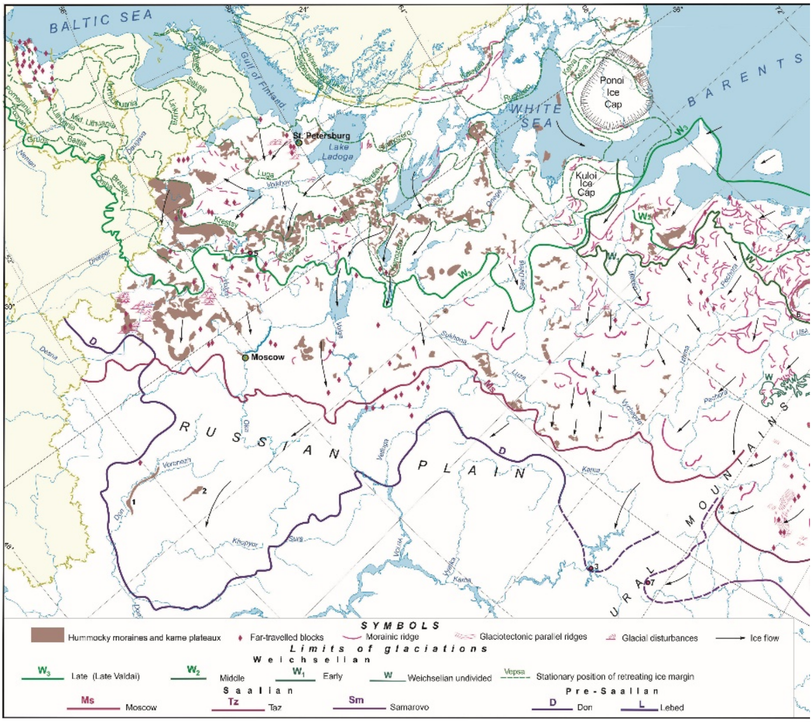

Fig. 1. Glaciomorphological Map of the Russian Federation (continued in next page). Localities referred to in the text: 1 – glaciofluvial plateau upon the Don Till, 2 – Surena glaciofluvial ridge, 3– southernmost exposures of lowland tills in the western Uralian foothills, 4 –Rogovaya ridges, 5 – the highest point of the Valdai Upland, 346 m a.s.l., 6 – Laya-Adzva double ridge, 7 – southernmost sections of lowland till in the Urals, 8 – Yugan megablock of Jurassic sediments in Samarovo Till, 9 – Agan morainic ridge with detached blocks of Paleogene sand, 10 – glaciotectonic parallel ridges in Fig. 4, 11 – morainic ridges up to 130 m high, 12 – Sopkay moraines, 13 – Bol. Tundrovaya inselberg 618 m a.s.l. covered by tills with lowland erratics, 14 – Mokoritto morainic loops, 15 – Norilsk end moraines.

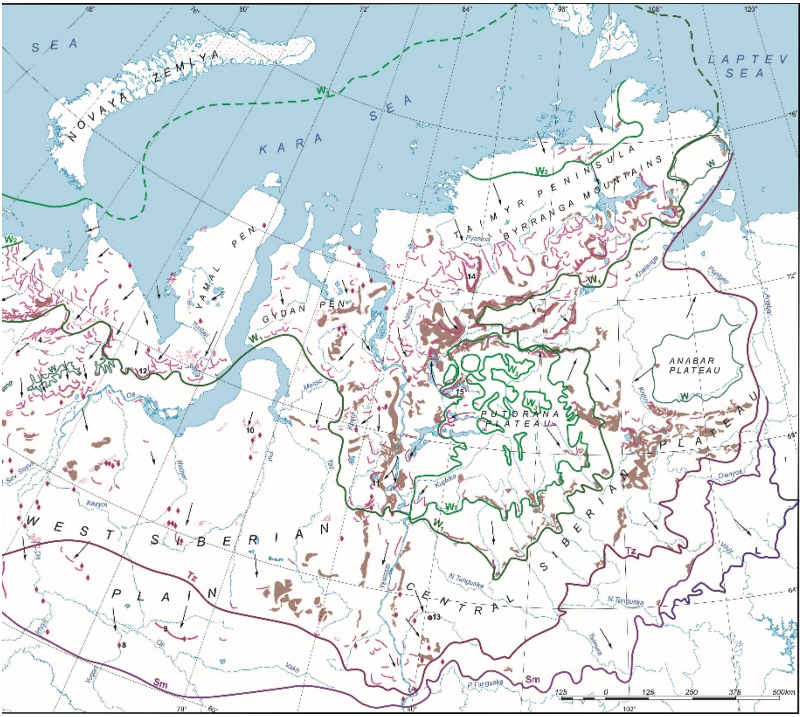

Fig. 1. Continued

Paper 26

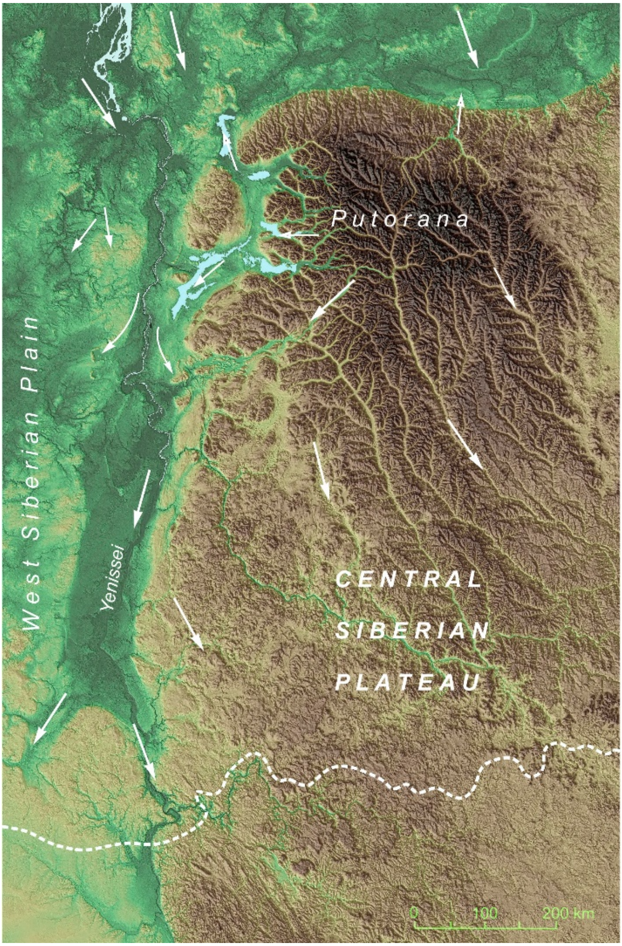

Fig. 3. Digital model of eastern West Siberia and Central Siberian margin. Broken line is the glacial drift limit, arrows indicate former ice flows. Change of light shades of boreal forests in the south to dark colours of montane tundra in the Arctic reflect gradual increase of elevations from 300 to 1300-1700 m contrary to the deepening of the Paleozoic-Triassic basin.

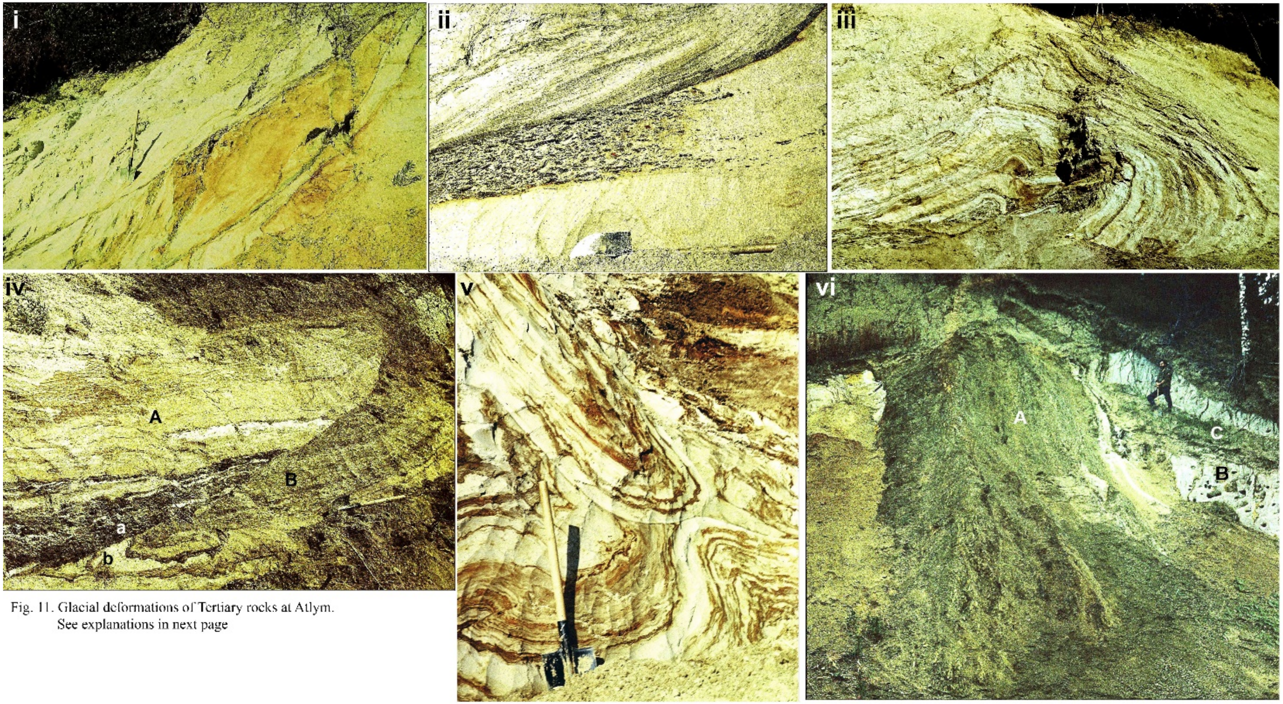

Fig. 11. Glacial deformations of Tertiary rocks at Atlym.
See explanations in next page

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