UNIVERSITY OF ALBERTA

PLEISTOCENE CLIMATIC HISTORY IN THE MINUSINSK BASIN, SOUTHERN SIBERIA, RUSSIA

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JIRI CHLACHULA ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance a thesis entitled PLEISTOCENE CLIMATIC HISTORY IN THE MINUSINSK BASIN, SOUTHERN SIBERIA, RUSSIA here submitted by JIRI CHLACHULA in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY.

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ABSTRACT

Climatic evolution in Siberia during the Pleistocene is still insufficiently known. A complete, high-resolution palaeoclimatic record for the last two glacial - interglacial cycles in southern Siberia has been studied at Kurtak in the Northern Minusinsk Basin. The unique, 34 m thick Late Quaternary sequence of loess and palaeosols (Section 29) with a differential degree of pedogenic development provides evidence of a strongly fluctuating, but patterned climatic change, with an increasing deviation amplitude during the Late Pleistocene cycle (130-10 ka). Mineralogical composition and grain morphology of the undifferentiated (aeolian and colluviated) loess indicates a local provenance and a short transport of the silt derived by glacial erosion of the granitic and metamorphic bedrock in the Kuznetskiy Alatau and the Eastern Sayan Mountains, and subsequently subaerially redeposited from the alluvial plain in the Yenisey River valley.

Recorded palaeoclimatic regularities manifested by a succession of Chernozemic, Brunisolic and Gleyed Regosolic soils are consistently traced by magnetic susceptibility characterized by markedly increased values during cold intervals, and decreasing values for warm intervals. This specific pattern (with the magnetic susceptibility minima in the most weathered interglacial soils) radically differs from the Chinese loess record, where the ferromagnetic concentration in palaeosols correlates directly with the enhanced degree of pedogenesis. A process of ferromagnetic mineral depletion during soil development under periodic oxidation and reduction conditions is proposed to account for the magnetic susceptibility decline because of leaching of less stable iron hydroxides. Increased amounts of unweathered iron minerals (magnetite, titanomagnetite and ilmenite) in the Siberian loess inherited from the glacially-eroded primary geological sources, coupled with the more intense aeolian activity during cold climatic intervals, are believed to be the main controlling factors.

The loess - palaeosol record from the Kurtak area, well-correlated with the marine Ox[illegible]e Stages 1-5 provides evidence of a pronounced climatic change in the Minusins[illegible] during the Late Pleistocene, with very cold and dry glacial stages, and cool and more humid mid-last glacial interstadials. The last interglacial *sensu lato* (130-73 ka) includes several cold and warm substages with a strongly continental trend during its first half. The palaeoclimate evolution is manifested by shifts in the major ecotones, with southern taiga established during the early interglacial and early interstadial stages under more humid and warmer conditions, succeeded by semiarid steppe and tundra-steppe, respectively. A gradual transition into glacial stages is evidenced by expansion of boreal forest and tundra-forest, eventually replaced by cold periglacial tundra-steppe with intensive loess sedimentation during glacial maxima. A cold-adapted arboreal vegetation appeared during warmer oscillations. Fossil fauna and cultural remains suggest analogous Late Pleistocene environments throughout southern Siberia.

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INTRODUCTION

Present as well as past climatic change has become an important area of investigation during the last twenty years. Several natural sciences, including climatology and atmospherics, evolutionary biology, terrestrial and marine palaeontology, soil sciences, geomorphology and geology, among others, have become involved in a series of projects aimed at reconstructing pathways and the rate of palaeoclimatic change on a local, regional, as well as global scale. More recently, computerized simulation programs of past environments in specific geographical areas and time periods have been introduced to assess the effects of orbital variations on seasonal and latitudinal distribution of solar radiation and atmospheric circulation patterns, and consequent changes in regional temperatures, precipitation and moisture balance (Kutzbach 1987; Kutzbach and Wright 1985; Kutzbach *et al.* 1991; Harrison *et al* 1991). Postulated long-term models, though theoretical and unverifiable in essence, are believed to provide some means of prediction for future climatic evolution in the context of the overall climate history of the Earth, and to assess the potential global environmental impact of the human factor.

The palaeoclimate studies focus on the Quaternary Period, *i.e.,* the time span of the last 2.5 Ma. Due to the nature of the problem, Quaternary studies have become increasingly interdisciplinary, integrating Quaternary geology, glaciology, sedimentology, palaeogeomorphology, palaeopedology, palaeontology, palaeobotany, Palaeolithic archaeology, and many other fields. Recent developments in dating methods, improving their reliability and accuracy (*e.g.,* the AMS, ESR dating, amino-acid racemination, thermoluminescence, optical luminescence, fission-track, isotopes studies and magnetostratigraphy) have significantly contributed to establishment of more detailed chronostratigraphic frameworks and to refinement of existing palaeoclimatic sequences. Except for long-term climatic records of marine and terrestrial deposits (Shackleton and Opdyke 1973; Martinson *et al.* 1987; Kukla 1987, Rutter *et al.* 1991a, 1991b, 1991c), studies on high-resolution loess and palaeosol records is the key focus of the current trend. Although dynamically evolving, the investigations are still in the initial stage with many problems to resolve.

Among the continental deposits, loess / löss (*i.e.,* a silty aeolian sediment in its primary contexts) has attracted most attention because of its high environmental susceptibility and long-term geostratigraphic records. The accumulation of loess is generally correlated with cold and dry stages, when silty particles were carried by wind from alluvial plains in ice-marginal areas and other periglacial deflation surfaces, and deposited in extraglacial areas as fine aeolian dust. During the following warm, non-glacial intervals, soils developed on the silty substratum because of surface stability and increased temperature, both leading to progressive weathering. Loess deposits proved to be a rich source of fossil micro- and macro-fauna, and early cultural remains. Chronology of the climatic events is provided by the radiocarbon technique for the time span of the last 40 ka and by the thermoluminescence method with reliable results to *ca.* 300 ka and possibly until 800 ka. Palaeomagnetism is widely used for a relative and absolute chronostratigraphy for the entire Quaternary, as well as earlier periods. Most data on loess came from several more intensively investigated areas in Central and Eastern Europe, Central Asia and north-central China (Pécsi 1991). Relatively few studies have been done in other parts of the world with extensive loess deposits.

Siberia, possibly because of its vast territory and geographical isolation, remains one of the least geologically known areas in the Northern Hemisphere, and in the World in general. Geographically, the overall Siberian territory comprises the northern part of Asia, an area of about 14 million km^2 characterized by highly diverse topographic relief with vast lowlands extending in the eastern and north-central regions, and broad highlands and high mountain ranges in the south (the Altay and the Sayan Mountains) and the east (the Yablonovyy Khrebet, the Stanovoy, the Verkhoyansk, the Dzhugduzuhur and the Koryak Mountains). Climatologically, Siberia is characterized by extreme seasonal climatic fluctuations with the lowest temperatures measured on Earth in central Yakutia (below -80°C). Many of the world's largest rivers (Ob, Yenisey, Angara and Lena) flow though its territory with a north-south-oriented drainage pattern emptying into the Arctic Ocean across all vegetation zones from the southern Siberian mountains through semideserts, steppes, southern taiga, boreal forests, forest-tundra, steppe-tundra and arctic periglacial tundra in the north. Because of the specific zonal distribution coupled with the pronounced climatic continentality of the territory, the Siberian loess record, especially in the southern areas, provides an excellent source of high-resolution proxy data for palaeoclimatic studies of a regional, as well as global significance.

Although more discontinuously and less extensively distributed than the loess in the Russian Plain farther west, or on the north-central China Plateau to the southeast, the loess and related deposits in southern Siberia reach up to a 150 m thickness (the Ob Plateau), and represent the most significant climatic record for this part of northern Asia, with a complete chronostratigraphic coverage at least since the late Middle

Pleistocene. Moreover, because of the geographical location between the major Central Asian (Himalayan) mountain systems and the Arctic Ocean, predisposing a strong continentality of the area, the Siberian loess may provide more detailed information on high-resolution climatic shifts, which are not detected in thick, but more uniform Chinese loess - palaeosol sequences. The Siberian loess record is also of major importance for the establishment of a palaeoclimatic correlation framework with the European, the Central Asian, and the Chinese loess provinces. Eventually, the Siberia-based loess research may significantly contribute to a better understanding of global climatic change and its mechanism during the Quaternary Period.

The present study deals with poorly-known loess deposits in the upper Yenisey River Valley, which have become exposed after large-scale slope erosion triggered by a rise of water level in the Krasnoyarsk lake. A particular focus is on the last interglacial - glacial cycle, *i.e.*, 127-10 ka (St-Onge 1987), for which there is a continuous and qualitatively exceptional geological proxy record. Geographically, the larger study area lies in the transitional subarctic continental zone between the north Siberian lowlands south of the Arctic Ocean and the south Siberian mountain system north of the Gobi Desert. Analyzed data were collected during the 1993 and 1994 field investigations in the Kurtak area in the Northern Minusinsk Basin (the upper Yenisey River valley). Published materials (mostly in Russian) from adjacent regions in the main sedimentary basins in the upper Ob River and Angara River valleys are included for a chrono-stratigraphic comparison and a regional correlation of the loess-palaeosol record.

The goal of the study is the reconstruction of the paleoclimatic evolution and related palaeoenvironmental changes in southern Siberia during the Quaternary Period, with a focus on the Late Pleistocene (130-10 ka). Recorded past climatic shifts in terms of chronostratigraphy, periodicity, intensity and a possible impact on the terrestrial ecology may be eventually incorporated into a general framework of the global Late Pleistocene climate history in the Northern Hemisphere. The unique location of the principal study area close to the geographical centre of Asia has particular significance for mapping changes of the arctic air-mass circulation above eastern Eurasia during the Pleistocene Epoch, as reflected in the textural and compositional patterns of aeolian deposits, and for assessing influence of the tectonically rising Himalaya Range that increased the continentality of the north-central Asian climatic regime between the Tibetan Plateau and the Arctic Ocean. Palaeoecological implications may be potentially applied for modeling adaptive patterns and responses in past biotic communities, and the main preconditions for early human occupation in the northern part of the Asian continent. Because of the nature of proxy data, the project entails an interdisciplinary approach. A long-term continuation of the initiated field investigations is envisaged.

Because of the similarity of physiographic (terrestrial continental) settings, the chronostratigraphic framework used in the study follows the concept of St-Onge (1987) on the Late Pleistocene stratigraphy in Canada. Accordingly, the last interglacial (Sangamonian) is correlated with the time interval 127-73 ka, the early last glacial (Early Wisconsinan) 73-62 ka, the mid-last glacial (Mid-Wisconsinan) 62-25 ka, and the late last glacial (Late Wisconsinan) 25-10 ka. This differs from the "short" last interglacial chronologies of Martinson *et al.* 1987 and Mangerud 1989, based on the deep sea oxygen isotope records, with the last interglacial *(sensu stricto)*, being the equivalent of the marine Oxygen Isotope Stage 5e (130-122/120 ka), and the last glacial corresponding to the Oxygen Isotope Stage 5d-1 (122/120-12/10 ka).

The study is structured in the following way. After the introductory part, natural environments of the larger study area are discussed in terms of the geographic setting, the recent climatic conditions, and the vegetation zonation (Chapters 1 and 2). A regional review of the pre-Quaternary and Quaternary geology is included. More background information is provided in the subsequent chapter, describing topography, geology and recent environments in the principal study area at Kurtak in the northern Minusinsk Basin (Chapter 3). This is supplemented by a summary of the existing Russian chronostratigraphic Quaternary geology system. The main part of the study (Chapter 4) focuses on description and analysis of a most complete loess-palaeosol section in the area (Kurtak Section 29). This includes a stratigraphic subdivision of the geological record; mineralogy of loess deposits; classification and interpretation of buried soil horizons based on field studies, and laboratory analysis of thin sections. The implied palaeoclimatic reconstruction follows (Chapter 5). The geochronological framework (Chapter 6), supported by the radiocarbon and relative palaeomagnetic (remanence) dating, is correlated with the magnetic susceptibility curve, which proved to be indicative of climatic change and is correlated with the deep-sea ∂^{18} oxygen isotope record (Chapter 7). Palaeoclimatic implications are integrated with other environmental (palaeontological and palynological) record from other localities in the Kurtak area (Chapter 8). Present archaeological evidence derived from the geological (loess and alluvial) deposits in the area is included for completeness. Eventually, all the above contextual proxy data are applied to reconstruct a climatic history of the region in the broader geographical framework of Central Siberia as in integral part of the Pleistocene climatic evolution of the continental North Asia (Chapters 9-10).

1. GEOGRAPHY AND NATURAL ENVIRONMENTS OF SOUTHERN SIBERIA

The southern part of Central Siberia is a broad territory geographically delimited by the Ob River basin in the west and the Angara River basin in the east. From the south, it is bordered by the province of eastern Kazakhstan (the Kazakh Republic) and western Mongolia. Administratively, the territory includes the Novosibirsk Oblast' (region), the Kemerovo Oblast' and the Altay Kray (district) in the west; the Gorno Altay and the Tuva Republics in the south; the Khakassia Republic and the southern section of the Krasnoyarsk Kray in the central part; and the western parts of the Irkutsk Oblast' and the Buryatia Republic in the east. Altogether, the whole area includes a little less than 10 % of the existing Russian Federation (Figure 1).

Topographically, southern Central Siberia is very diverse with relief gradually rising southward up the continental slope from the Northern Siberian Plain over the Central Siberian Plateau and the Yeniseyskiy Kriazh (range) to the Altay - Sayan Mountain System in the south. The West-East / NW-SE oriented southern Siberian mountain ranges form a natural barrier between the northern subarctic uplands and Arctic lowlands, and the southern steppes and deserts in Mongolia and northwestern China. The Altay in the southwest is the highest mountain range in the area (Belukha Mt., 4506 m). To the east, it is connected with the Tannu Ola Range and the Western Sayan Mountains (2930 m) separated by the Tuva Depression. To the north, the Khakasso-Minusinsk Depression is formed between the Kuznetskiy Alatau (2178 m) and the Eastern Sayan Mountains (3491 m). Westward, the Kuzbas Depression on the other side of the Kuznetskiy Alatau is connected with the Western Siberian Lowland in the north. Eastward, the area is delimited by the southern margin of the Siberian Plateau and the adjacent Irkutsk Basin bordered by the Eastern Sayan from the south and the west, and the Baykal Range from the east. The central part of the basin comprises the Lena - Angara River Plateau (maximum elevations about 800 m).

The principal physiogeographical features of larger Central Siberia (*i.e.*, the Siberian Plateau, the Northern and Southern Siberian Lowlands, the Taimyr and Sayan-Altay Mountains) were formed during the Eocene and Oligocene by the breakup of the Palaeogene Siberian Platform. Palaeogene and Neogene deposits are widely distributed in the main sedimentary basins above Cambrian, Jurassic and Cretaceous formations (Kriger 1963). The early neotectonic activity continued during the Miocene. Orogenic processes became increasingly active during the Pliocene (after *ca.* 3.5 Ma.), initiating the Baykal Rift System (currently 1620 m deep below the Baykal Lake level), and rising in to high elevations the Baykal Range (2588 m) and the Eastern Sayan Mountains (3491 m) (Figure 1).

Formation of the southern Siberian mountains had a direct impact on atmospheric circulation. Free passage of warm and humid air masses from the southeast became increasingly obstructed by the rising orographic barrier, so cold arctic streams became dominant. In the process, the present strongly continental regime was established. Most significant and geologically relatively recent topographical changes occurred during the Pleistocene, particularly the later stages, when intensive denudation processes modified the uplifted areas, and thick deposits released from the orogenically active margins filled the deepening depressions forming large sedimentary basins. Extensive aeolian (sandy and loessic) sediments were subsequently accumulated in the main valley depressions and within the transitional zone between the southern Siberian mountains and the northern lowlands (Volkov and Zykina 1983).

Hydrologically, most of the study area is drained by the Yenisey River emptying into the Arctic Ocean. From its origin in the Eastern Sayan to the Yeniseyskiy Bay in the Kara Sea it is 3354 km long. The river flows through all main geographical and vegetational zones of northern Asia, from the mountains in the south through the mixed southern taiga and steppe, boreal forest, the northern subarctic lowlands and tundra to the Arctic coast in the north. The major tributary in the eastern upper reaches is the Angara River draining the Baykal Lake. The western part of southern Central Siberia is drained by the Ob River, with the Chulym River as the main tributary.

The present climate is strongly continental with marked seasonal and diurnal / nocturnal temperature fluctuations, and a moderate amount of precipitation due to the geographic location in the middle of Asia (Figure 1), as well as orographic isolation from the monsoon system. Current atmospheric conditions are controlled by the maximum intensity of the Siberian Anticyclone moving a cold, dry arctic air (the Arctic Front) during winter, and moderating regional air-mass temperatures increased during summer. Precipitation brought by the western cyclone circulation (the warm Atlantic Stream) as well as the eastern air masses from the Pacific coast are considerably weakened by mountain ranges and increasing distance from the oceans. Seasonality is very pronounced with a short and windy spring and autumn. Active permafrost characterizes most of the territory, especially in the northern areas, where it reaches up to several hundred meters in thickness. Mountain glaciers are present in higher elevations in the Altay and the Eastern Sayan Mountains (above 3000 m). Climatic conditions in the southern part of the Krasnoyarsk Kray are influenced by the Asiatic Anticyclone with a stationary centre above Mongolia, and the Western Cyclone (the warm and humid Atlantic Stream). During winter, the Polar Front moves southward and becomes established above the central part of the continent, causing a significant drop in temperatures and increased aridity. Strengthened atmospheric circulation from the west causes mixing of the polar and Atlantic air-masses to form cloudiness (Golovin 1972:31).

In the southern Krasnoyarsk area, maximum precipitation is received on the western slopes of the Eastern and the Western Sayan Mountains (1000 mm and 1200 mm), whereas a dry climate prevails in the depressions, with a minimum of 240 mm in the southern Minusinsk Basin. Average temperatures are -16/-18°C in January and +18/+21°C in July. Mean annual temperature in the city of Krasnoyarsk is +1°C (Gromov and Lbova 1961:12; Likhanov and Khaistova 1961:40). Summer precipitation and glacial ablation in the mountains provide most of the river discharge. In the southern Irkutsk area, maximum annual precipitation is at the margin of the Irkutsk Basin near the Eastern Sayan Mountains (700-800 mm), with 350-400 mm in the central part of the basin and only 250-300 mm in the river valleys. Average July temperature is +14/+17°C; the average January temperature is -20/-25°C. The mean annual temperature fluctuates between -1°C and -4°C, being partly influenced by the warming effect of Baykal Lake (Medvedev *et al.* 1990:6-7).

Due to the latitudinal and physiographic distribution predisposes a marked biotic zonation is expected. Mixed boreal forests with fir, birch and willow are established on the southern Siberian uplands and plateaus; dominantly coniferous forests cover the southern mountains with alpine vegetation in high elevations (above 1700-2000 m). A typical southern taiga (above 700 m asl.) is characterized by spruce, larch and birch at lower altitudes, and pine and cedar at higher altitudes mainly on northern, more moist slopes. The upper tree line lies at 1500-1700 m asl. Steppe and forest-steppe with parkland communities occupy lowlands.

2. PHYSICAL SETTING OF THE STUDY AREA

2.1. TOPOGRAPHY, CLIMATE AND VEGETATION

The main study area lies in the southern part of the Krasnoyarsk Region along the upper reaches of the Yenisey River in the northern part of the Minusinsk Basin ([illegible] 56°N and 89-94°E) (Figures 1-2).The local geomorphological setting is rather diverse. The basin is one of a series of tectonic depressions in the eastern part of the West-East oriented Caledonian - Hercynian Altay - Sayan Mountain System. It is bordered on the west by the Kuznetskiy Alatau (maximum elevation 2178 m asl.), and the western foothills of the Eastern Sayan Mountains on the east (maximum elevation in the study area 1778 m). Orographically, the Minusinsk Basin is divided into the Northern (Chulymo-Yeniseyskaya), Central (Sydino-Erdinskaya) and Southern (Abakano-Minusinskaya) Depressions with average altitude in the central part between 300-400 m, and 450-500 m in marginal areas. Because of the marked topographic gradient, the central part of the depression drops from 500-700 m in the south to 150-300 m in the north. The northern part of the Basin is delimited by the Solgonskiy and Batenevskiy Kriazh Ranges with altitudes below 900 m. The central part is structurally controlled by a zone of tectonic breaks running in a north-south direction across the Batenevskiy Kriazh Range. In the northwest, the Basin is connected by the Nazarovskaya Depression with the broad Western Siberian Lowland. Mountain valleys manifest an apparent asymmetry due to differential erosional processes, with steep and rocky northern sides, and gradual southern slopes covered by talus deposits.

The broader geographical area is drained by the Yenisey River. In the southern part, the central river valley lies at an altitude of 250-300 m, rising to 400-600 m in the marginal areas. Most discharge is supplied by its right tributaries from the Eastern Sayan Mountains (the Koma, Ubey, Sisim, Dervina, Kazyr and Kizir Rivers) and the Western Sayan Mountains (the Abakan River); less discharge from the Kuznetskiy Alatau, which belongs to the hydrological system of the Chulym River. The absolute elevations of the divide, formed by an old Paleogene surface, are 440-500 m asl (80-130 m above the Yenisey River level). Because of a blocked drainage and low precipitation values, saline lakes and over 50 mineral (sulfate, chloride and sodium-manganese) lakes were formed in earlier, now fragmented river valley located mainly in Khakassia, in the western part of the Minusinsk Basin. The Yenisey River, in places deeply cut into the rocky bedrock, was artificially dammed in the early 1970s to form the 400 km long Krasnoyarsk Lake, one of the world's ten largest freshwater reservoirs. Due to the river regulation, the water level was raised to 40-100 m above its original stand depending on the proximity of the dam.

The present climate is generally dry and strongly continental with cold and dry winters with little snow cover, and warm to hot summers. Mean annual temperature in the Northern Minusinsk Depression is -0.5°C (average January temperature -18.1°C, July temperature +17.6 °C). In the central part of the Southern Minusinsk Depression, the continentality is even more pronounced, with mean annual temperature -0.6°C (average January temperature -20/-21°C, July temperature +19/20°C), and maximum temperature deviations of -56°C and +37°C. The maximum regional solar radiation is in July; maximum diurnal/nocturnal temperature fluctuations are in August (+24°C / -4°C). Annual precipitation in the central steppe zone is 250-300 mm, most of which (80-90%) falls between June and October, but mainly during early summer. May is the driest month. In the forest-steppe zone in the marginal areas, particularly in the northwestern part of the Minusinsk Basin, continentality and the temperature contrast decreases, with about a 50 % (in the mountains more than 100 %) increase of precipitation to 350-1000 mm *per annum.* Prevailing winds and humid air-masses come from the west (Drozdov *et al.* 1990a:15; Derevianko *et al.* 1992a:14). In the spring as well as during drier summers, wind storms frequently occur.

The vegetation cover is characterized by grasslands in the interior basin in the southwest; forest-steppe is established in the northeastern part of the depression and along the foothills, and mixed southern taiga prevails in the mountains. Specific environmental zonation defines the general ecosystem pattern, with dry, semi-dry and wet open steppe settings, and the latitudinal distribution of the larch-dominated southern and birch-dominated northern taiga. In the Eastern Sayan foothills, the larch-pine-birch forest in lower elevations gives way to the dark spruce-fir-pine forest in upper elevations, which is eventually replaced by alpine vegetation above the 1700 m altitude. Except for the elevation factor, the mosaic nature of the landscape is determined by slope orientation and pedo-substrate differences. The transitional zone between the (forest-)steppe and taiga is rather narrow, sometimes only a few kilometers broad.

In the study area in the northwestern part of the Minusinsk Basin, the recent vegetation spectrum based on pollen records in the forest-steppe zone includes from 40 to 60% tree taxa (*Pinus silvestris* 29-70%, *Betula* sp. 14-60%, *Alnus* sp. 5-6%, *Salix* sp. 2-4%, *Picea* sp. 1-7%, *Abies* sp. 1%), 30-50% grasses and sedges(*Artemisia* sp.

20-25%, Cyperaceae 11-16%, Poaceae 7-12%, *Ephedra* sp. 0-4%, other 27-37%), and 20-25% spores (Polypodiaceae 52-67%, Bryales 16-22%, *Sphagnum* 7-16%, *Lycopodium* 1-9%). The composition from the parkland steppe zone in the Kurtak area includes from 83 to 94% arboreal taxa (*Betula* sp. 17-87%, *Pinus silvestris* 11-30%), grasses and herbaceous plants 4-16% (Asteraceae 25%, Fabaceae 25%, Cyperaceae 7%, Poaceae 4.5%, other 11%), and 0.6-2.3% spores (Polypodiaceae, *Lycopodium*) (Drozdov *et al.* 1990a:53).

Well-developed Chernozemic soils, Luvisolic soils and Podzols correspond with the vegetation zones. Specifically, Black and Brown Chernozems (Kastanozems) and Solonetzic soils are distributed in steppe areas; Gray Chernozems are developed under forest-steppe; Luvisols, Brunisols and Gleysols under forest (taiga). Analogous buried soils (palaeosols) are found on silty loessic and fine-grained colluviated substrata, suggesting broadly similar past environmental conditions. In the southern steppes, Brown Chernozems occur on sandy/silty deposits in the most arid areas with average annual precipitation of 200 mm (Yepokhina 1961:148). In the foothills, black/brown forest-steppe soils are formed under a parkland vegetation with birch and spruce to about 700 m altitude; in the high elevations with increased moisture regime, pine (*Pinus silvestris*) and the Siberian pine (*Pinus sibirica*), generally called "kedr" *i.e.*, "cedar" in Siberia, grow on Podzolic soils.

The southern Siberian Chernozems are, in general, analogous to those originally described from the Russian Plains in the European part of Russia west of the Ural Mountains (*e.g.*, Velichko *et al.* 1992b). Although the Canadian soil classification system is used throughout this study, particularly in the analytical palaeopedological section, the Canadian Chernozems are classed in the American classification under Borolls, *i.e.*, cold soils.

2.2. GEOLOGICAL HISTORY

Both relief and geological structure of the southern part of the Krasnoyarsk region manifest intensive geomorphological processes acting in the area in the past. The Cambrian and Proterozoic volcanism disturbed and dislocated the original pre-Cambrian crust composed of igneous (mainly granitic) and metamorphic rocks. During the Lower and Middle Palaeozoic, tectonic activity carved the former landscape to create a system of mountain ranges separated by deep depressions, in places major synclines. Igneous andesitic rocks and metamorphic quartztose, carbonate and gneissic rocks are the main lithologic components (Dodin 1961). The mountain ranges (the Altay, the Kuznetskiy Alatau, the Eastern Sayan) constitute the western part of the Caledonian and Baykal structural complexes of southern Siberia. Later Hercynian tectonics and the Oligocene orogenic activity modified the area to form the configuration of the depressions in the Minusinsk Basin, which were subsequently filled by Devonian, Carboniferous, Jurassic and Paleogene volcanic, lacustrine and alluvial deposits. Specifically, they include Devonian lacustrine sediments, Cretaceous tuffaceous deposits, coal, conglomerates, sandstones, siltstones, limestones and claystones, Jurassic sands and loams, and various Paleogene alluvia (Drozdov *et al.* 1990a:17). The Mesozoic formations (terrigenous and coal deposits) are mainly distributed in the northern part of the Minusinsk Basin, whereas Cambrian andesitic and carbonate rocks and Caledonian granites outcrop in the Eastern Sayan Mountain area (Figure 3). Locally, Cambrian terrigenous deposits are exposed in the western as well as eastern margin of the Minusinsk Basin (Dodin 1961). Increased sedimentary activity and weathering rate, evidenced by the thick clastic accumulations, suggest a more pronounced continental regime established during and after the Cretaceous Period. Extensive proluvia near the mountain fronts, and lacustrine / alluvial formations in the main sedimentary basins were deposited throughout the Miocene and Pliocene.

During the Late Pliocene and the Early Pleistocene, an early fluvial system was established, accompanied by ongoing progressive tectonic uplift. Alluvial fans and relics of terraces about 200 m above the present valley bottom date to this period. Progressive erosion of the mainly Jurassic and Paleogene fine-grained sediments contributed to the high degree of a loamy component within the Quaternary alluvial deposits. A neotectonic movement during the early Middle Pleistocene further more divided the Minusinsk Basin into the northern and the southern parts. The former was subsequently filled by a complex, polygenetic 10-100 m thick series of alluvial, lacustrine, proluvial and aeolian deposits of the Koiskaya Formation assigned to the Penultimate Glacial Period (the Samarov / Riss 1 / Illinoian Glaciation) and the previous interglacial (the Tobolsk / Mindel-Riss / Yarmouth Interglacial) (Drozdov *et al.* 1990a:18-19). A more recent uplift of the Western Sayan Mountains during the second half of the Late Pleistocene caused a partial transformation of the former river system.

In the upper Yenisey River valley, a series of alluvial terraces is discontinuously developed, but relatively distinctively preserved. The lower terraces of the Yenisey River (8-12 m, 18-20 m, 30-40 m) are now flooded by the Krasnoyarsk Lake. The high terraces (70-90 m, 110-130 m, 150-170 m) are preserved mostly as relics.

Chronology of these early alluvia is poorly known. Some seem to have been formed gradually over a longer time span, with several cycles of deposition and erosion, as is evidenced by a complex stratigraphy with numerous truncation surfaces, and differences in lithology of sandy / gravelly materials. Based on fossil faunal evidence, the 60-80 m (70-90 m) terrace is believed to comprise the Tobolsk - Samarov interglacial/glacial cycle; formation of the 30-40 m terrace is correlated with the last interglacial (Kazantsev / Riss-Würm / Sangamonian) and the early part of the last glacial (Zyryansk / Early Würm / Early Wisconsinan); the 18-20/25 m terrace with the Karginsk / Mid-Würm / Mid-Wisconsinan Interstadial (Non-Glacial Interval); and the 8-12/15 m with the Sartan / Late Würm / Late Wisconsian Glaciation. Early Holocene terraces are elevated 5-6 m and 2-4 m above the Yenisey River (Drozdov *et al* 1990a:20, 39). In a geomorphological synthesis of the northern Minusinsk area provided by Laukhin (1979), the lowest terrace I (8-11 m) is assigned to the Holocene; the complex of lower terraces II-IV (12-15 m, 17-25 m, 30-35 m) to the Upper Pleistocene; the middle terraces V-VII (40-50 m, 70-80 m, 90-100 m) to the Middle Pleistocene; the upper terraces VIII-IX (100-120 m, 150 m) to the Lower Pleistocene, although the highest may be already Pliocene in age. The highest and heavily weathered alluvia situated at a relative altitude of 200-240 m below proluvial and colluvial loamy and sandy gravels and loess deposits are presumably relics of the former Yenisey - Chulym River system (Drozdov *et al.* 1990a:39). Loess deposits, usually 10-20 m thick, are distributed over most of the landscape, but mainly on valley slopes where they are up to 35 m thick. They are presumably of local origin, and include primary air-born loess, as well as secondarily colluviated loess-like deposits (discussed below).

During glacial periods, large parts of the Altay and the Eastern and Western Sayan Mountains were covered by ice with montane glaciers advancing into the upper reaches of the main river valleys. At the same time, loess, derived from alluvial plains, was deposited on the slopes of periglacial basins. Surface stabilization and progressive weathering during interglacials partly slowed down the rate of acting geomorphologic processes. The insufficiently known chronology of the alluvial and subaerial deposits is caused by a lack of radiometrically or palaeomagnetically dated units, and absence of a well-established correlative chronostratigraphic framework on a broader regional scale.

The adjacent area along the lake has been continuously modified by landslides and wave undercutting of unconsolidated loessic deposits, forming the surrounding slopes of the valley, particularly from the western side. A gradual sediment saturation at the base causes the principal rotational slide mechanism, accelerated by a partial lost of a supporting (top-consolidating) vegetation cover. The slope-retreat rate is highly variable, depending on loess thickness, underlying geological structure and morphology of the deposits. On the highest sections, rather narrow, but 2-7 m deep and up to 30 m long gullies form at the top of the eroded landscape surface. The following gradual slope disintegration of the main sedimentary structure significantly accelerates the process of slope retreat with a rate of up to 10 m in one year. The process of erosion along the rocky hills, mostly distributed on the forested eastern side of the Yenisey valley, is less dramatic. A feedback effect of the Krasnoyarsk reservoir is well manifested by the rapid process of landscape destruction, with unconsolidated silty materials filling the inundated valley bottom. This, in turn, causes a relative lake-level rise that accelerates the process of erosion. A disclosure of a highly interesting geological record is a scientific benefit at the expense of the altered natural environment.

My field investigations in the Kurtak area were carried out in July-August 1993 and 1994. Most of the time I spent working in the Berezhekovo sector (Figure 2) with the most extensive loess sections, located about 1 km south of the central camp. Except for the principal Quaternary research, geo-archaeological studies were pursued at other sites (Kamennyy Log, Sukhoy Log and Razlog). In the following year, additional loess-palaeosol magnetic susceptibility data and palaeosol samples were collected from the key section (Kurtak 29). Subsequent (rescue) investigations (August 1994) focused on a unique Middle Pleistocene Early Palaeolithic occupation site at Ust'-Izhul', discovered in 1993, and increasingly exposed to river erosion (Figure 2). My research schedule was largely independent of activities of the fieldwork-organizing institution (Archaeological and Palaeoecological Laboratory of the Russian Academy of Sciences, Siberian Branch, Krasnoyarsk), pursuing a long-term archaeological research program in the area. About 50-80 people, mostly students, lived in the main camp. Five of them helped me to clean the investigated loess sections prior to sampling, which was a very welcome initiative. A smaller tent camp (10-15 people) was established at Ust'-Izhul', located about 7 km north of the main camp. Transport from and back to Krasnoyarsk was by ex-military lorries; commuting in the study area was mostly on foot, and by boat between the sites along the lake shore and across the Krasnoyarsk Reservoir. Occasionally, I took horses for trips into interior accompanied by local "cowboys". Horseback and helicopter transport were indispensable in the Altay Mountains, where I spent about a month. Weather was very good most of the time, and combined with beauty of the countryside, friendliness of local people and substantial help of Russian colleagues made my stay both seasons very enjoyable and scientifically prolific.

	Siberia	Central Europe	North America
Late Pleistocene			
22-10 ka	Sartan Glacial	Würm 3	Late Wisconsinan
55-22 ka	Karginsk Non-Glacial	Würm 2	Mid-Wisconsinan
110-55 ka	Zyryansk Glacial	Würm 1	Early Wisconsinan
130-110 ka	Kazantsev Interglacial	Riss-Würm	Sangamonian
Middle Pleistocene			
180-130 ka	Tazov Glacial	Riss 2	Late Illinoian
200-180 ka	Shirtinsk Interglacial	Riss 1-2	Middle Illinoian
270-200 ka	Samarov Glacial	Riss 1	Early Illinoian
390-270 ka	Tobol'sk Interglacial	Mindel-Riss	Yarmouth
Early Pleistocene			
510-390 ka	Nizhiansk Glacial	Mindel 2	Kansan
560-510 ka	Til'timsk Interglacial	Mindel 1-2	-
600-560 ka	Azov Glacial	Mindel 1	-
670-600 ka	Talagaykin Interglacial	Günz-Mindel	-
720-670 ka	Mansinsk Glacial	-	-
Eopleistocene			
>720 ka	Gornofilinsk Interglacial	-	-

Table 1. Chronostratigraphic correlation of Siberia, Central Europe and North America (modified and summarized according to Arkhipov 1983; Drozdov *et al.* 1990a:19, tab. 1; Volkov and Zykina 1982:tab. 1). *N.B.* The Pleistocene chronology respects the Russian classification system with glacial and interglacial "horizons" and the short chronology of the last (Kazantsev) interglacial (130-110 ka) *versus* the Sangamonian Interglacial (130-73 ka). The Karginsk Non-Glacial Horizon includes three interstadials dated to 55-45 ka; 43-33 ka (Malakhetskoe IS); 30-22 ka (Lipov-Novoselovo IS). The Sartan Glacial Horizon includes three interstadials dated to 16-15.5 ka; 13-12.2 ka (Kokorevo IS); 10.8-10.4 ka (Taimyr IS) (King 1974, cited by Tseitlin 1979:14).

3. THE KURTAK ARCHAEOLOGICAL REGION

3.1. GEOGRAPHICAL LOCATION

The principal study area, referred to as the Kurtak Archaeological Region formally established in 1988 (Drozdov *et al.* 1990a, 1990b, 1990c), is located in the upper Yenisey River valley in the Northern Minusinsk Depression at 91° W longitude and 55° N latitude (Figure 2). The area lies in the steppe zone at the Kuznetskiy Alatau Mountains on the western side of the Krasnoyarsk reservoir between the Trifonovka Bay in the south and the Ust'- Izhul' Bay in the north. On the east side of the lake, foothills of the Eastern Sayan Mountains rise directly from the shore. The Yenisey River is here about 7 km wide (the valley with the upper terraces extends 10-12 km), inundating the Kurtak Syncline. The parallel, northwards-oriented Izhul Syncline stretches west in the Chulym River valley. North and south of the Kurtak area, the Yenisey valley narrows to a 2-3 km width cutting through the Novoselovo and Daurskoe Paleozoic Anticlines (Drozdov *et al.* 1990a:36).

The Kurtak Archaeological Region (named from a nearby farming village translocated from the Yenisey valley before flooding) is about 20 km long and is characterized by a series of exposed sections stretching along the lake shore recently formed by wave erosion (Figure 2). The area has been formally divided into five sectors in respect to local geomorphology:

1. The Trifonovskiy Uchastok (sector), located over a distance of *ca.* 6 km between the Bezguza Bay and the Kamennyy Log (gully) Site at the foot of a 70-80 m terrace relic, is covered by 12-15 m of Quaternary sediments in the eroding part.
2. The Kurtakskiy Uchastok, extending over 2 km between the Kamennyy Log and the Sukhoy Log Sites, has locally exposed Carboniferous sandstone 5-7 m above the lake level, and 7-8 m of loess and colluvial deposits in the exposed slope walls.
3. The Sukhoy Uchastok, extends about 1.5 km between the Sukhoy Log and the Berezhekovo Sites, and has exposed bedrock up to 10 m high covered on top by thin alluvial deposits and overlain by loess sediments.
4. The Berezhekovskiy Uchastok, extending 2.2 km south of Chany Bay, has partly exposed relics of two terraces (70-80 m and 90-100 m) covered by a 20-35 m thick loess.
5. The Izhulskiy Uchastok, extends for about 5 km north of Chany Bay. It has a more complex geological structure with alluvial fan deposits (2-20 m high) locally cut into up

to 30 m of exposed Carboniferous bedrock, and capped with colluvial and aeolian loess cover (1-10 m). A distorted 60 m terrace is adjacent to the lake shore.

Research was initiated in the upper part of the Yenisey River valley after progressive erosion of unconsolidated aeolian and slope deposits from water-table fluctuation of the Krasnoyarsk reservoir after dam construction in 1971 had flooded the valley to 65 m above the original river level at 247 m asl. The twenty-five year lake-shore erosion had caused a gradual 0.5-4 km lateral slope retreat accompanied by large-scale landslides of the unconsolidated deposits (consisting mostly of loess), and other mass-wasting processes triggered by distortion of the lateral slope support. The 10-40 m high, steep, and continuously eroding sections exposed a nearly complete Late Quaternary geological record containing a series of Palaeolithic stone industries and a rich Middle and Upper Pleistocene fauna (Drozdov *et al.* 1990a; Derevianko *et al.* 1992a). Relics of the old (Middle and Early Pleistocene) Yenisey River terraces (70-80 m and 150 m) are locally preserved in the form of a thin alluvium. Occasional extensive alluvial fan deposits built by erosion of the Early Pleistocene and Pliocene terraces fill the Carboniferous bedrock depressions (*e.g.*, the Razlog Site) (Figure 34-A).

Until the ponding of the Yenisey River, Quaternary geology of the northern Minusinsk area was poorly known (Finarov 1963; Arkhipov 1971; Laukhin 1979). The previous field studies were largely limited to neotectonics and geomorphological mapping of the old river terraces, most of which are today inundated by the Krasnoyarsk reservoir. The ongoing lake erosion has revealed a more complex geological structure than was previously assumed. Systematic geological studies have been carried out since the late 1980s, largely in connection with archaeological discoveries of early cultural remains in eroded sections at the lake shore. The main focus of the ongoing investigations has been reconstruction of the Pleistocene history of the area and establishment of a detailed geochronological framework in order to determine the age of the cultural material and contextual palaeoenvironmental conditions. Regardless of the archaeological work, the large-scale sections with thick loess-palaeosol sequences provide an excellent opportunity for an independent Quaternary palaeoclimate research, which was the principal subject of this study.

3.2. QUATERNARY GEOLOGY

Following the initial geo-archaeological investigations of the Pleistocene sections at Kurtak in 1988-1989, four basic stratigraphic units, assigned to the Upper Pleistocene, have been recognized (Drozdov *et al.* 1990b; 1990c; Laukhin *et al.* 1990a). The underlying stratigraphic record is rather fragmentary and still poorly studied. Although largely Middle Pleistocene in age, some basal colluviated gravelly strata overlying the 70-80 m terrace in the Berezhekovo sector are believed to be Early Pleistocene lying on a Pliocene terrace relic because of reversed magnetic polarity (V.P. Chekha, personal communication 1993).

The four well distinguished and broadly laterally traceable stratigraphic units include (from the bottom): the Pedocomplex of the Kamenolozhskaya - Sukholozhskaya soils assigned to the last interglacial (Kazantsev / Sangamonian) *(sensu lato);* the Tchaninskaya Tolscha of 0.5-17 m thick loess deposits of the early last glacial (Zyryansk / Early Wisconsinan); the Kurtak Pedocomplex of the Karginsk / Mid-Wisconsinan Interval; the Bezguzinskaya Tolscha of 1-10 m thick loess deposits of the late last glacial (Sartan / Late Wisconsinan). Stratotypes of the above units have been described at the Berezhekovo Site.

A complete, composite stratigraphic profile in the Kurtak area includes, according to Drozdov (1992), and Chekha and Laukhin (1992):

the Kochkov Horizon (Early Pleistocene; >780 ka.), formed by up to 10 m of alluvial and colluvial deposits on the 60 m terrace relict or the Carboniferous bedrock and incorporating a fossil fauna (rodents) of the Tamansk Complex of Western Siberia;

the Lebedskiy Glacial Horizon (the early Middle Pleistocene; 600-380 ka), formed by up to 50 m (?) of colluvial deposits at the 110-130 m terrace platform (the Berezhekovo Site), and associated with a large fauna of the Tiraspol Complex of Western Siberia (*e.g., Equus* ex. *germanicus sanmeniens)*;

the Berezhekovo Series (the Tobolsk Interglacial; 380-260 ka), formed by up to 15 m of sandy gravelly alluvial deposits (the Razlog Site);

the Bakhtinskiy Glacial Horizon (late Middle Pleistocene; 260-130 ka.), formed by periglacial alluvia and colluvia, largely secondary deposits on the 60 m terrace, and palaeontologically associated with *Mammuthus chosaricus* and an early *Equus* sp;

the Kurtak Series (the Kazantsev Interglacial; 130-100 ka), including the Kamenolozhskaya soil (2 m) developed on a diverse substratum (the Carboniferous sandstones, periglacial alluvium, loess);

the Muruktinsk Glacial Horizon (the early Zyryanks Glacial; 100/74-50 ka), including the Sukholozhskiy Pedocomplex of several soil horizons of a total thickness of 2-3.5 m; the Tchaninskaya Tolscha of loess deposits (up to 10-17 m), and the Berezhekovo Tolscha.

the Kurtak Pedocomplex (the Karginsk Non-Glacial Interval; 50-22 ka.), including two soil horizons (with ^{14}C dates of >30,000-29,400 B.P., and 25,000-22,000 B.P.) separated by cryoturbation of the Konoshelskoe Stadial;

the Bezguzinskaya Tolscha (the late Zyryansk / Sartan Glacial; 22-10 ka.), formed by 1-10 m of loess deposits with a well-marked frost wedge horizon dated to 20,000-18,000 B.P.

the recent (Holocene; <10 ka.) aeolian (0.5 m) and colluvial (maximum 7 m) sandy deposits.

The above chronostratigraphic units have been tentatively correlated with fluvial terraces in the lower (submerged) part of the Yenisey Valley - Terrace I (8-12 m; late Sartan), Terrace II (15-20 m; early Sartan/Karginsk), Terrace III (30-40 m; Zyryansk/Kazantsev). The upper terraces - Terrace IV (60 m; late Middle Pleistocene), Terrace V (70-90 m; early Middle Pleistocene), Terrace VI (110-130 m; Early Pleistocene), Terrace VII (150-170 m; Pliocene/Miocene?) are preserved in relics above the lake level (Derevianko *et al.* 1992a:18). Because of the different chronostratigraphic classification of the Quaternary Period between the Western and the Russian systems (see below), the upper terraces (above 60 m) are referred to as Neogene in age (Drozdov 1992:16).

Despite the significant amount of fieldwork done so far, there is disagreement about the nature of the loessic sediments, whether they are subaerial and/or subaqueous in origin (discussed below). Also the number, exact age, correlation and environmental significance of buried and diverse soil horizons needs further clarification and in-depth study. A further study focus on the most complete and representative sections, and their spatial correlation and structural (sedimentological and pedological) comparison in terms of palaeo-geomorphology and parent material differences would provide a more coherent and conclusive picture on the past environments and related natural processes.

The existing chronology of the loess-palaeosol sequences is based on radiocarbon, thermoluminescence and palaeomagnetic dating, and the micro- and macro-faunal chronostratigraphy. All data, however, are to be considered as preliminary. The most reliable temporal stratigraphic assignment has been provided by the ^{14}C method. A detailed review and a critical examination on the Quaternary history, chronostratigraphy and palaeoecology of the Kurtak area is discussed below in the context of particular problems of the study.

The following chapter is a summary of stratigraphic, sedimentological, palaeo-pedological and geochronological investigations carried out at the most complete currently exposed section (Kurtak 29) investigated so far. Correlation with other sections in the study area is included. The loess-palaeosol sequence, chronologically comprising the second half of the Middle Pleistocene and most of the Late Pleistocene, provides the principal, high-resolution proxy data for reconstruction of the Late Quaternary environments and the palaeoclimatic history of the Northern Minusinsk Basin.

4. THE BEREZHEKOVO SITE (KURTAK SECTION 29)

4.1. TOPOGRAPHY AND STRUCTURAL GEOLOGY

The Berezhekovo Site is situated in the central part of the Kurtak Archaeological Region containing the most complete Quaternary sections (Figure 2). The top surface is formed by a plateau of the highest flanks of the Yenisey valley, gradually rising westwards towards the Kuznetskiy Alatau foothills. Open steppe stretches over most of the adjacent landscape, with isolated parkland communities of deciduous trees. In the shore zone, relics of former river terraces and alluvial fans, are locally exposed within the steeply eroded sections in a relative elevation of 70-80 m and 90-100 m. Higher terraces (110-130 m and 150-160 m) are well developed in the landscape relief, and sealed above by 10-15 m of aeolian and colluvial cover deposits. The fluctuation of the Krasnoyarsk reservoir prior to the present water level stabilization (234 m asl.) caused intensive undercutting of the slopes formed by loess deposits, and revealed a series of buried soils. A rich fossil fauna and archaeological record have been eroded from the basal alluvial gravels beneath up to 35 m thick loessic deposits. In the southern part of the site, the Carboniferous sandstone bedrock is horizontally exposed at an elevation of 72-75 m above the former Yenisey River, and in places covered by 1-1.5 m of gravelly alluvium.

Because of the unconsolidated nature of the slope-forming sediments, ongoing erosional processes rapidly modify the section relief. The present sections are semi-recessive to prominent, temporarily stabilized under an angle of 50-70°, although high walls with an angle over 80° form steep ramparts at some places. The characteristic mechanism of the slope retreat is gradual slumping of loess in large, 2-5 m high blocks along structural joints in the upper part that develop into massive blocks in the middle and lower parts is the characteristic mechanism of slope retreat. Removal of the lateral support and liquification of the basal part during periods of higher lake-water level are the main erosional agents, in addition to rain and wind erosion.

The earliest Quaternary deposits above the Carboniferous bedrock are exposed only locally in the southern part of the Berezhekovo Site. They include poorly sorted and weathered alluvium, presumably relics of a Pliocene terrace overlain by thin Early Pleistocene gravels (Excavation 1, Strata 8-10) (Drozdov *et al.* 1990a:44-49). The Pleistocene stratigraphy is best documented in the high sections in the middle part of the site, with up to 25-35 m of loess deposits incorporating a series of palaeosol horizons

(Figure 4-A, Figure 5), which are better preserved than in the archaeological sections in the southern part of the site, where the cover deposits are thinner, the buried soils are largely disturbed by solifluction and some hiatuses are evident. Although comprising only a small part of the Quaternary period (about the second half of the Middle and most of the Upper Pleistocene), the central part of the Berezhekovo Site, about 1.2 km south of Chany Bay, includes almost a complete and unique climatostratigraphic record for the last glacial - interglacial cycle in northern Eurasia.

4.2. STRATIGRAPHY

The key section (Kurtak 29) is located in the middle of the Berezhekovo shore zone with the thickest and best exposed loess deposits. The section is separated by a gradually rising 1-5 m high sandy-silty platform terrace (the formerly submerged beach) extending about 35 m off the lake shore. A composite profile is mapped in three adjacent and stratigraphically superimposed sections I-III (16.00 m, 12.50 m and 6.60 m high) cut into a 70-90° steep slope wall (Figure 4-B). The 34 m section includes 31.5 m of the exposed stratigraphic sequence above the present beach in the prominent cliff (Sections I-II), and 2.5 m of deposits partly excavated at the base into the lake shore platform sealing the original slope structure, with the deepest part 2-3 m above the lake level (Section III). The natural face of the slope was vertically cut by a series of 1.5 m high, 1.2 m wide and 1-3 m deep (*i.e.*, horizontally incised into the cliff wall) segments to expose an undisturbed and complete stratigraphic profile, and to facilitate access for subsequent stratigraphic, sedimentological, palaeopedological and palaeomagnetic studies and sampling.

Structurally, the section is formed by aeolian loess, colluviated loess deposits and palaeosol horizons displaying a varying degree of pedogenesis. The base of the loessic unconsolidated strata is not exposed. The contact to the bedrock or the basal alluvium has not been reached in the lowermost excavated section (III) which lies about 3-5 m deeper (information based on field observations prior to the site inundation; V.P. Chekha, personal communication 1994). Except for the major sedimentary facies and developed buried soil horizons, the visibility of other stratigraphic units is relatively limited in a naturally exposed profile. Detailed documentation was carried out in the freshly cleaned walls, and repeated a few days later when they dried up in order to trace calcic (BCca) soil horizons and to retrieve other sedimentary and pedogenic features.

A total of 55 stratigraphic units has been recognized within the Kurtak 29 Section. These include 20 aeolian loess units, 3 colluviated loess units, and composite soil horizons of 32 variably developed soils. A composite stratigraphic profile is summarized in Figure 6. The lateral stratigraphic correlation is provided by a Brunisolic (Bmkgj) soil horizon (Sk_{18}) for Sections I and II at 16.50-17.00 m below the surface, and by a Chernozemic (Ahy-Bmk-BCca) soil horizon (Sk_{30}) for Sections II and III at - 27.20-28.10 m m below the surface. Except for erosional surfaces between most of the loess - palaeosol units, no stratigraphic hiatuses in the form of major disconformities have been identified. Loess facies analysis, methodology, criteria and hierarchy of soil classification are specified in the following section. A detailed loess - palaeosol stratigraphic description is provided in Appendix A. The below discussed grain-morphological, sedimentological, and mineralogical analyses were provided to clarify the geological origin and geographical provenance, the mode of transport, and depositional and postdepositional processes of loess in the study area. Related palaeopedological studies focused on structural and mineralogical characteristics of buried fossil soil horizons in terms of loess weathering processes and climatic conditions during the geochemical and geophysical alteration of the parent material.

4.3. LOESS ANALYSIS

Loess constitutes the bulk of the structural material in the section. However, as stated earlier, the nature, origin and mode of accumulation of the deposit is subject to discussion (also see below). In this section, basic characteristics of the Kurtak loess, in terms of (micro)morphology of particular sedimentary units, their mineralogical composition, and grain texture attributes, are summarized for clarification.

4.3.1. Structure and Texture

A typical loess (löss) is defined as a silt-sized, pale yellow, highly porous (45-50%), solid (if dry) and cohesive (if wet), non-stratified deposit, dominated by 0.02-0.05 mm grain-fractions, and mineralogically characterized by quartz (50-70 %) and carbonates (5-30%) (Pécsi 1987:82). A loess substratum subsequently pedogenically altered into a soil horizon may be primary, *i.e.*, derived by aeolian activity during cold climatic periods, or secondary, *i.e.*, redeposited by fluvial and/or colluvial processes.

Both principal loess / loessic facies are present at the Berezhekovo Site, as well as the larger study area. The first loess facies can be characterized as a massive, unstratified to weakly stratified light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m) silty deposit of a massive, single-grain structure; non-sticky (w), very friable (m), soft (d); and with many, micro-vesicular pores. A certain degree of mottling expressed by a varying amount, size and distinctiveness of ferric and manganese oxides is present in several levels throughout the section as a result of water table fluctuation and colloid mineral migration associated with initial oxidation-reduction soil formation process [illegible] ation). Loess forming the upper part of the section is largely pale brown (1[illegible] ellowish brown (10YR 5/6m); in the middle part light brownish gray (2.5[illegible] olive gray (5Y 6/2d) / dark grayish brown (2.5Y 4/2m); and in the lower pa[illegible] pale brown (10YR 7/3-6 /3d) / pale to dark brown (10YR 6/3-4/3m). Average pH value is 7.4.

Despite a general uniformity of the loess facies, some variations are apparent throughout the stratigraphic profile. A conformable gradational horizontal bedding is manifested by 1-5 cm thick brownish / greenish / grayish layers within thick loess units. The loess sediment forms most of the upper half of the section, where it occurs in 0.3-3.0 m thick packages intercalated by more or less developed palaeosol horizons (Figure 6). On the other hand, its appearance in the lower half of the section is limited to silty loamy and brownish less distinct strata. A fine-grained matrix supporting large angular to subangular quartz grains (30-50 μm), and numerous small voids within the silty material characterized the microstructure of the sediment (Figure 13-A).

The second loess facies is distinguished by a fine, irregularly stratified structure formed by interstratified, light brownish gray (2.5Y 6/2d) to light olive gray (5Y 6/2d) / dark grayish brown (2.5Y 4/2m); pale brown (10YR 6/3d) / brown (10YR 5/3-4/3m); and brown (10YR 5/3d) / dark brown (10YR 3/3m), and undulating 0.2-1 cm thick silty to silty-loamy layers. The sediment is slightly sticky (w), friable (m) and hard (d). Faint to distinct mottles are occasionally present, as well as minute (1-2 mm) white (10YR 8/2d) calcium carbonate precipitates. Average pH value (7.6) is slightly higher than at the first loess facies.

Irregular, tapering laminae are either horizontally superimposed, or variably dip at 1-5° angles. Contrary to the first loess facies with smooth, gradual and diffused contacts, the layered structure of the second facies is distinct and sharp. Minor (0.5-2 cm) frost wedge casts and involutions are frequently present. Thickness of these loessic packages varies from 1.50-3.20 m in the lower part of the section to 0.2-1.0 m in the upper part. The thinner and more compact brownish strata, however, may in fact represent redeposited (colluviated) soils, because of an apparent increase of a loamy component and mineral and calcium carbonate migration shown by mottling. Except for deposition rate, thickness of the accumulated loess was dependent on the local relief. The silty fraction is dominant; thin sand lenses rarely occur, indicating periods of minor loess erosion and/or non-deposition. Micromophology of the loessic deposit is characterized by a more compact structure than the first facies, with an increased textural variability in the form of grain clustering and rearrangements (Figure 14-A, B). The size and shape of grains is about the same in both loess facies.

Genetic interpretation of the above loess and loessic (loess-like) deposits is summarized in the following discussion.

4.3.2. Mineralogy

From a mineralogical point of view, the partly stratified 0.2-5 cm thick loess beds in the Kurtak sections include quartz (52-61 %), local bedrock materials (15-23 %), biotite, chlorite and calcite (up to 10 % combined), muscovite and amphibole (5 %), and some heavy minerals (apatite, zircon, tourmaline) (Drozdov *et al.* 1990c:33-35, 37). A greater percentage of chlorite is apparent in samples from the middle part of the section, accounting for the the gray-greenish colour of the sediment. In terms of structural and lithological correlation of the loess units throughout the stratigraphic sequence, quartz and biotite characterize the second, thinly stratified (colluviated) loess facies, whereas quartz and chlorite are the major mineral constituents of the first loess facies. The changing gray and greenish color of the individual sedimentary bands within the loess deposits apparently reflects the percentage variation of biotite and chlorite, respectively.

The specific mineralogical composition has been analyzed from 10 major loess units at 2.5 m, 9.4 m, 14.0 m, 15.5 m, 17.5 m, 18.8 m, 21.5, 24.0 m, 31.0 m, 32.5 m depth; and five soil horizons (Sk_9, Sk_{12}, Sk_{18}, Sk_{24}, Sk_{30}), some of them with distinctive lithologies characterized by change in colour and textural variability. The mineralogical (X-ray diffraction) analysis was provided by the Rigaku - Geigerflex graphic monochrometer (Department of Geology, University of Alberta), and compared with the mineral distribution in thin sections. The results showed about the same proportion of the main minerals. In all samples, quartz is the dominant component, with a percentage range of 54-72%, followed by feldspar (10-23%), including plagioclase and orthoclase, calcite (4-7%), chlorite (3-6%), biotite (2-5%), amphibole, muscovite and magnetite (1-7%), and some sporadically represented (mainly heavy) minerals (<1%). The maximum proportion of quartz (72%) occurs in the 30-50 cm loess unit between the two Chernozemic soils (Sk_{22} and Sk_{24}) in the last interglacial pedocomplex. A local origin of the sediment derived from more weathered deposits between two periods of intensive pedogenesis most likely accounts for the reduction of the less stable mineral components. Advanced mineral weathering of biotite into chlorite is apparent by the colour (greenish-grayish-brownish) variation of individual laminae. Similarly, the increase of quartz in the massive colluviated loess unit in the lower (between Sk_{26} and Sk_{27}), as well as the analogous unit in the basal part of the sequence (below Sk_{32}) correspond to the nature of these deposits with signs of pedogenic alteration. Lower values of calcite and a slightly increased percentage of amphibole is documented in the loess stratum above the niveo-aeolian / colluviated loess deposit (below Sk_{25}). A higher amount of orthoclase (11 %) is recorded in the most recent loess (between Sk_2 and Sk_3), dating to the last glacial stage. The variation in the mineral percentage distribution, however, is not considered to be particularly significant. Except for a general percentage increase of quartz grains, no other major mineralogical difference has been observed between the loess and the soil horizons, as documented in the most representative loess and palaeosol samples (Figures 8-10; Tables 3-6).

A more detailed compositional study has been provided by the EDX SEM (Energy Dispersive Scanning Electron) Analysis (SEM laboratory, Department of Metallurgy and Mining Engineering) based on calculation of elemental weight-percent distribution, and applied to the most representative loess and palaeosol samples. The documented data include a typical aeolian loess from the late last glacial climatic minimum (Table 3:1), an analogous, partly colluviated deposit from the early last glacial (Table 3:2), and a sample of a Chernozem (A horizon) correlated with the last interglacial climatic optimum (Table 3:3). The compositional element percentage values show an apparent increase of silica, iron and aluminium (42.1%; 20.2%; 12.1%) in the interglacial soil compared to the colluviated loess (41.1%; 17.6%; 9.9%) and the pure aeolian loess (37.3%; 17.0%; 10.6%). The percentage amount of calcium, however, is considerably higher in the loess samples (13.0%; 16.8%) than in the humic A horizon of the interglacial soil (6.3%). The mineral variation is evidently related to pedogenesis causing a breakdown of some compositional minerals releasing aluminium, iron, silica and other unspecified elements, and their secondary accumulation in weathered horizons, accompanied by a gradual depletion of calcium from the altered soil horizons and its migration into lower parts of the soil sequence. The above trend has been also recorded consistently in the additional seven loess and palaeosol (Regosolic, Brunisolic and Chernozemic) samples. The quantitative distribution of the principal mineral oxides corroborates the elemental analysis, with the highest percentage values of SiO_2, Al_2O_3 and Fe_2O_3, and the lowest values of CaO in the last interglacial Chernozem (Figures 7-9; Tables 4-6).

The subsequent mineralogical analysis under the ARL SEMQ electron microprobe (SEMQ laboratory, Department of Geology, University of Alberta) was carried out to specify the major ferromagnetic components and their proportions that were not detected by the X-ray diffraction analysis, because of marked magnetic susceptibility variation in particular samples. A series of iron and titanium compound minerals was recorded with typical magnetite, titanium-rich magnetite, titano-magnetite and ilmenite being the most frequent ones (Figures 10-11). Other heavy minerals include sphene, zircon, rutile, augite, clinopyroxene, garnet and epidote among others. From the analyzed samples, the highest proportion and widest range of ferromagnetic minerals was found in unweathered loess sediments (below SK_2). A lower amount was observed in initially developed soils (Sk_9, Sk_{18}), and the lowest frequencies in well-developed Brunisolic and Chernozemic soil horizons (Sk_{12}, Sk_{24}, Sk_{30}).

The recorded susceptibility variation resulting from the mineral composition of loess and palaeosol samples could potentially reflect: a) a differential amount of larger ferromagnetic grains in respect to proportional variation of the supporting matrix; b) an overall mineralogical composition; or c) a differential alteration degree of the original mineral components. Because no significant compositional and textural differences have been recorded in the loess facies and the buried soil horizons within the study section, the laboratory scrutiny has focused on the nature of present ferromagnetic minerals present.

ARL SEMQ analysis that attempted to identify the principal factor behind the magnetic susceptibility fluctuation has been performed on 10 selected and most typical loess and palaeosol samples. In a loess sample from the last glacial stage with the maximum amount of the ferromagnetic minerals, the mineral series are characterized by a gradual increase of titanium and a simultaneous decrease of iron, which can be followed from a typical magnetite (with the highest proportion of the elemental iron) through titano-magnetite into ilmenite (with the highest proportion of titanium) (Figure 10:1-3). In a Chernozemic A horizon of a well-developed soil from the penultimate interglacial, the continuation of the series can be followed through rutile and ferroan rutile, distinguished on the basis of the differential amount of iron (Figure 11). Similarly as in other palaeosol samples, a gradual increase of other elements such as calcium, aluminium, magnesium, but mainly silica has been observed in the particular ferromagnetic minerals at the expense of iron and titanium. The compositional element differences between titano-magnetite grains from the above loess (Figure 10:3) and soil (Figure 11:3) samples are apparent. This fact corroborates the above results of the EDX SEM analysis, as well as the thin-section analysis (page 34), showing an increased amount of clay fraction (composed principally of Si, Al and Mg) and a secondary concentration of the leached calcium, as a result of advanced weathering processes affecting the original parent material (*i.e.* loess). The gradual pedogenic modification of the silty deposit is best monitored by the ferromagnetic susceptibility reduction. Therefore, geophysical and geochemical changes accompanying soil-formation processes can be regarded as the principal controlling agent behind the magnetic susceptibility reduction. The significance of this phenomenon is discussed in the section on magnetic susceptibility of the loess-palaeosol record.

4.3.3. Mineral Grain Morphology

Surface morphology of fine sand and silt grains, using high level magnification (100-1500x) under the scanning electron microscope, provides additional information about formation processes, mode of transport, and post-depositional processes. Most of the grains, 30-50 μm on average, exhibit angular shapes and clear, unweathered surface morphology typical of local transport from a nearby source (Figure 12). The routine analysis confirmed the assumption that most of the material was derived from the mountain areas surrounding the basin by glacial erosion of local granitic and metamorphic rocks. Interestingly, structure, texture and composition of 30 μm thin sections of the typical loess do not differ much from sections on parent rocks. A limited proportion of more mature, subangular to subrounded grains in some loess samples may indicate a longer-distance transport, but also secondary deposition from eroded older geological formations in the study area. The latter explanation would be supported by the presence of mature, well-rounded quartz grains identical to those from locally exposed Carboniferous sandstone bedrock, that were most likely released during past periods of intensive aeolian activity in the Minusinsk Basin.

4.3.4. Discussion and Conclusion

As mentioned above, disagreement persists about the nature and the mode of deposition of the loess. Primary aeolian (subaerial) sedimentation is assumed (Arkhipov *et al.* 1992:10; Zykina 1992:102), as well as colluvial and/or alluvial accumulation. A colluviated nature for the loess deposits has been suggested at the Kamennyy Log Site (Drozdov *et al.* 1990a:26), and a subaqueous (glaciofluvial / glaciolacustrine) depositional environment for the 1-30 (?) m thick Tschaninskaya Formation at Berezhekovo, with presumably a more dynamic aggradation process in the lower part (Drozdov *et al.* 1990c:33-35, 37). It has been suggested that the silty materials were derived during a period of increased glacier ablation from the Kuznetskiy Alatau Mountains, when glacial water may have been at 20-30 m above the present reservoir level, inundating the Upper Yenisey Basin and extending over the divide into the Chulym River valley creating a large glacial lake. Following the final drainage, the silty sediments are thought to have been subsequently colluviated by slope processes (Drozdov *et al.* 1990a:61). However, clear evidence in support of this scenario is lacking.

Although some loess deposits within the study area indicate a more complex and non-uniform formation process, most of the silty materials, especially those in upper Section I, were clearly subaerially derived (the Bezguzinskaya Tolschta). This inference is based on the "disorganized" original microstructure (Figure 13), as well as the massive, uniformly bedded aggradational structure of the silty deposit with slightly undulating silty laminae without apparent bounding surfaces that are features characteristic of aeolian settings. The above pattern differs from structurally more variable (clayey, silty, sandy) glaciolacustrine deposits with a more distinctive layering. Absence of extensive horizontal bedding, poor sorting, vertical sedimentary facies variability and the lack of stream-directional patterns combine to exclude a (glacio)fluvial origin of these deposits.

A niveo-aeolian and colluvial origin is assumed for the more compact and finely interstratified silty-loamy deposits that form a significant part of the lower Sections II-III. Increased humidity during earlier phases of loess deposition is also indicated by the segregated and vertically reoriented mineral fabric in this colluviated loess facies

(Figure 14). Presence of small subrounded quartz grains in both (unmodified aeolian and colluviated) facies may indicate more distant transport from glacial environments in the upper reaches of the Yenisey River. Some coarser sediments may also have been subaerially eroded from the local Carboniferous sandstone bedrock. Overall, the fine clastic material was apparently introduced from different sources during cold and dry climatic intervals, and redeposited onto the nearby slopes by wind. This implies a partly mixed origin of the loess deposit.

Besides the provenance of the loess and the means of its (re)deposition, an additional climate-indicative factor is inherent in the geological record. Thermally induced fracturing of grains along decompression fracture lines in silt grains has been observed particularly in the later aeolian loess found in the ice-wedge sediment fill. The fragmentation of grains in frost-wedge casts in a Chernozemic soil is documented in Figure 15. Although other causes for disintegration of quartz grains may have been involved (*e.g.*, unloading in the original geological formations), extreme cold and/or fluctuating climatic conditions with microscopic capillary water infiltrations into the mineral grain fissures is believed to have been the principal agent. This issue is beyond the scope of this study so is not pursued in detail, but may be a subject for further scrutiny. In any case, the envisaged climatic factors behind the observed clast-shattering processes, the high angularity of most grains, the increased proportion of the less stable minerals, as well as the poor sorting of the silty sediment adds further support to the idea of a dominantly subaerial accumulation of the Kurtak loess.

In sum, an aeolian origin is assumed for the first loess facies, whereas a niveo-aeolian formation accompanied by solifluction and secondary periglacial deformation phenomena (documented by minor frost wedge casts and involutions) is indicated for the second facies. In both cases, the specific laminated structure reflects cyclic (seasonal ?) patterns of local loess sedimentation. The light (grayish and greenish) and thicker silty layers presumably accumulated under colder and arid conditions; the darker (brownish) and thinner silty and silty-loamy layers were deposited during cool and more humid climatic oscillations. This issue is discussed in more detail in the section on palaeoclimatic reconstruction.

4.4. PALAEOSOL ANALYSIS

4.4.1. Soil Formation Processes: Principles and Objectives of Study

Five major factors are usually stated to have a direct bearing on soil development -- climate, relief, organisms, parent material and time (Fenwick 1985; Wright 1986). Position of the water table recently has been suggested as another significant variable (Catt 1990:73). Soil formation is largely influenced by specific climatic zones, although azonal soils also occur (*e.g.*, Gleysols in waterlogged areas). The initial soil formation process is manifested by root traces and a minor accumulation of organic matter in the uppermost part of the otherwise unaltered parent material. Incipient clayey peds and formation of calcareous, sesquioxide and humic horizons specify weakly developed soils. Well-developed clayey skins and a distinct soil profile with diagnostic A-B-C horizons are characteristic for moderately developed soils. Finally, strongly weathered soils show thick brown/red B-horizons with a well-organized structure.

In the context of Quaternary studies, the basic pedostratigraphic unit is defined as a palaeosol having undergone pedological processes similar to, or stronger than the Holocene soil in the particular area (Ding *et al.* 1991:172). Palaeosols can be classified according to several criteria. Frequency of soil formation cycles separates *monophase soils* (*i.e.*, those without subsequent structural modification due to climatic stability and unchanged vegetational cover) from *polyphase soils* (*i.e.*, those showing a climatic variability in repeated superposition of pedogenic elements) and *polycyclic soils* (*i.e.*, those with various, stratigraphically superimposed horizons) (Fedoroff and Courty 1987:257). Climatic influences are recorded in the B horizon, undergoing physical and chemical alteration processes, the most important of which is clay illuviation (argillification), carbonate leaching and rubification/brunification.

In terms of geological position, palaeosols can be preserved as *relict soils* (always polygenetic), being exposed on the present surface, or as *buried soils* (mono- or polygenetic), largely unaffected by current weathering processes. The former can be used to estimate age of soils, periods of landscape stability, timing of faulting, *etc.*; the latter as stratigraphic and palaeoenvironmental markers (Fedoroff *et al.* 1990:653). Post-burial diagenetic processes are expressed by compaction of the clayey structure; cementation by precipitation of mineral components (Fe, Mn, Al) from ground water in voids and fissures; changes in geochemistry, and alteration of organic material (Catt

1990:66). Preservation rate of an individual soil varies considerably. Rarely, however, is a palaeosol preserved in a fully developed original state. Usually, the upper part of the soil profile is truncated, so that only the most stable and resistant B-horizons are more or less preserved, though still in a geochemically and biologically altered condition from which the climatic information may be difficult to retrace. Except for texture and mineralogical characteristics of fossil soils, loess malacofauna proves to be a very helpful means of chronostratigraphic correlations, allowing a more detailed subdivision of early glacial periods, which may be difficult to compare solely on the basis of weakly developed palaeosol horizons, especially in the middle latitudes (Lozek 1990). Biological evidence (molluscs, phytoliths, charcoal, pollen, spores, macrobotanical and faunal residues, *etc.*) may provide a useful information on duration of pedogenesis, the type and amount of humus at the beginning of soil formation, the dominant climate and vegetation, or the nature of substratum.

In sum, the principal objectives of palaeopedological studies used in Russia include: 1) soil formation processes under warm and/or humid conditions, 2) soil transformation processes in transitional periods towards cold stages (*e.g.*, cryoturbation, solifluction in the form of pedosediments), and 3) diagenetic transformation processes of buried soil (*e.g.*, modification of the original soil structure, acquisition of new features because of a low soil stability). Palaeosols, therefore, must be regarded as an open environmental system and reconstruction provided in a sequence by combination of soil properties (Morozova 1987). In the context of this study, palaeosols are considered as all pedogenically changed horizons that show *any* pedological alteration of parent material that is absent in the pure loess. This approach was predetermined by the specific nature of the investigated loess - palaeosol record characterized by a gradual transition throughout several palaeoclimatically significant types of buried soil horizons.

The main objectives of the palaeopedological studies in the Kurtak area were: 1) identification of soil-formation phenomena and their separation from purely sedimentary structures; 2) specification of nature of the pedogenic parent material transformation *in situ* or in a colluviated state; 3) a retrospective assessment of the loess sedimentation rate; 4) establishment of a hierarchy of pedogenic traits and processes; and 5) reconstruction of environmental conditions prior, during and after the soil formation in terms of *protoevent*, *synevent* and *postevent* (Velichko and Morozova 1987).

4.4.2. Methods

Field pedological studies at Kurtak Section 29 included recognition of buried soils, their description and preliminary classification, and sampling for a subsequent laboratory study with thin sections. The applied analytical methodology follows the Russian system (Morozova 1987:31). The description of soils (the Canadian System of Soil Classification) is provided in Appendix A. Biological samples (molluscs, charcoal, pollen and faunal remains) were collected during field investigations.

The amount of organic matter accumulation; degree of clay accumulation (argillification) expressed by clay coatings and intensity of braunification; $CaCO_3$ content and secondary carbonate precipitates (neoforms), periglacial phenomena (solifluction, cryoturbation, frost disturbances), and magnetic susceptibility are considered as the principal proxy indicators of the past climatic change in the micro- and/or macro-structure of preserved humic (A) and mineral (B/BCca) horizons. Therefore, a particular attention was paid to the above aspects. The specific field and laboratory analytical procedures are described below in the particular study sections. The outlined pedological subdivision of the palaeosol record is eventually discussed in terms of a general palaeoclimatic evolution in the Kurtak area.

4.4.3. Field Recognition Criteria

Previous palaeopedological investigation in the Yenisey River basin and throughout southern Siberia have focused on well-preserved, diagnostic and broadly distributed Late Pleistocene soils (Demidenko 1990b; Dergacheva *et al.* 1984; Dergacheva *et al.* 1992). In the Kurtak area, three groups of soils have been specified in terms of visibility and preservation. These are intact fossil soils; disturbed and partly redeposited soils from periglacial processes (solifluction, injections, involutions, cryoturbation, frost wedge casts); and poorly developed (incipient) soils (Chekha 1990). Strato-pedological correlation has been provided on the basis of soil morphology, substratum lithology, stratigraphic position of specific horizons, their palynological and paleontological content, and radiocarbon and palaeomagnetic dating (Dergacheva *et al.* 1992; Zykina 1992, Morozova 1990).

The recognition and classification of fossil soils within the study section is based on a set of physical criteria, including color change, structure, consistency, porosity, degree of mottling, and secondary calcium carbonate content, compared with the pedogenically unaltered parent material (loess). A sum of the above attributes was used for a conclusive identification of pedogenically altered loess horizons (Fenwick 1985). Well-developed buried soils were recognized and subsequently classified (see below) without major difficulties, especially where horizons A horizons are preserved. Certain problems were encountered by differentiation of initial regolithic soil horizons from brownish finely stratified loess layers, and by separation of several pedogenic events in composite and/or colluviated soils with partly superimposed mineral horizons (B horizons). In the latter cases, pedogenic horizons were recognized by diffused lower boundaries and a more pronounced brownish colouring towards the top, combined with an increased consistency and a presence of some iron mottles. Nevertheless, identification of soil forming processes was provided by thin section analysis, and after comparison with the palaeomagnetic susceptibility curve discerned to be specifically diagnostic for pure and pedologically altered loess deposits (discussed below).

Overall, the recorded palaeosols are spatially more or less continuous. The most expressive units can be well correlated laterally for a long distance between the sites (up to several kilometers), if exposed. Stratigraphic position, morphological characteristics, secondary periglacial deformations, and radiocarbon and palaeomagnetic dating have been used to provide a chronostratigraphic control framework. The general stratigraphy is consistent with other Kurtak loess sections, although a different degree and nature of past morpho-topographical processes, related factors and other cover disturbances caused a certain pedostratigraphic variation. The original landscape topography is evidenced by a lateral thickness variation of individual soil horizons, becoming thicker in places of shallow depression and thinner on formerly exposed palaeosurfaces, and partly or fully disturbed, truncated, or redeposited by solifluction in less stable places, such as gullies and steep slopes.

4.4.4. Laboratory Analysis

Laboratory studies were carried out in order to obtain more information about the nature and forms of past pedogenic processes, and to provide a control for the field palaeosol classification. Study of the sampled palaeosols in thin sections was the main analytical technique. Soil micromorphology studies were carried out for classification of particular soil horizons, determination of the former vegetation cover and the principal soil forming processes, as well as later pedogenic, biogenic and cryogenic disturbances (Mücher and Morozova 1983; Fedoroff and Courty 1987; Fedoroff *et al.* 1990; Bronger and Heinkele 1989, 1990).

For the preparation of thin sections, 3.5 x 5.0 x 2.0 cm block samples were collected from undisturbed buried soil structure at 30 locations of the freshly cleaned stratigraphic profile. The dry samples were impregnated by a clear resin (Petroproxy 154) and cut to a 2 mm thin section, ground and eventually polished to the standard 30 µm thickness. Visual analysis was provided by a petrographic microscope (Leitz Type HM-POL) under a cross-polarized light and the 25x and 100x magnification level, and under a microscope with a reflected light source. *Handbook for Soil Thin Section Description* (Bullock *et al.* 1985) was used for formal visual comparison of particular soil-diagnostic micromorphological attributes (structure, texture and composition) in the analyzed soil samples. Photographic documentation on a 36 mm film was provided in the process of microscopic investigation.

In general, the genetic concept of soil micromorphology addresses in detail the soil structure (*i.e.*, the size, shape and arrangement of primary particles and voids in both aggregated and non-aggregated materials, and the size, shape and arrangement of any aggregates present), fabric (*i.e.*, the overall organization of constituent particles), and pedofeatures (*i.e.*, the discrete fabric units). For the purpose of the present study, and due to weakly developed micromorphological features of most palaeosols and/or their secondary distortion, the analysis was limited to the basic aspects of soil structure (*i.e.*, the size, shape and arrangement of primary particles and voids in both aggregated and non-aggregated materials, and the size, shape and arrangement of any aggregates present) as expressed by grade of peds, as this is the most diagnostic for the degree of pedogenic development used for a general palaeosol classification (Morozova 1987). Descriptive reference to specific soil macro-structure patterns and final soil classification was done according to the *Canadian System of Soil Classification* (1987).

Periglacial deformations of most fossil soil horizons were easily recognized in the field. Relict microfeatures caused by frost action have been identified in palaeosol horizons in thin-sections (van Vliet-Lanoë 1985). Microscopic horizontally-aligned structural patterns in the second loess facies suggest thaw cycles (Figure 14-A); present frost cracks in large quartz grains in ice-wedge fill, and minor involutions in sedimentary laminae indicate existence of seasonal frost (Figure 15-A, 15-B). Periodic processes of freezing and thawing are also documented in the Chinese loess (Cao 1990). In the Russian plains and Siberia, cryogenic features in loess deposits and buried soils are evidenced, except for well-developed ice- and loess-wedge casts, by non-illuviated grain argillans and humus-clayey aggregations characteristic for slower and discontinuous pedogenic development (Mücher and Morozova 1983:175; Morozova 1990). Nevertheless, in the study of past soil formation processes in the Kurtak area, the main problem concerns diagenetic changes and distinguishing characteristics of relic soils related to two or more periods of soil development.

4.4.5. Palaeosol Classification

Thirty-two buried, pedogenically altered soils displaying various levels of pedogenesis have been identified in Kurtak Section 29. Genetically, these include (in the level of soil orders) 7 Chernozems (including one forming the present surface, and one transformed into Cryosol), 17 Brunisols and 8 Regosols. The number of well-defined soils is, however, fairly limited, and most of the palaeosols are represented by weakly developed (initial) soil horizons, mostly of the Brunisolic and Regosolic Order (Figure 6).

4.4.5.1. Chernozemic Soils

Chernozemic soils are the most distinctive palaeosol units (Sk_{11}, Sk_{12}, Sk_{21}, Sk_{22}, Sk_{24}, Sk_{30}), located at -9 m, -10 m, -18 m, -18.5 m, -19 m and -28 m depth below the present surface (as determined by a cumulative thickness of the A, B, BCca and Cca horizons). They are distinguished by a dark brown (10YR 3/3d) / dark grayish brown (10YR 4/2d) and very dark grayish brown (10YR 3/1m) mineral (Ah) horizon characterized by a weakly developed fine, granular structure and decomposed, organic matter accumulation (Figure 17-A). In all soils, silty clay is the dominant matrix, being sticky if wet and becoming very hard when dry. The mineral horizons (Bmk) are dark brown (10YR 4/3d), very dark grayish brown (10YR 3/2m), with a silt loam, weak, medium subangular blocky structure and a few distinct, strong brown (7.5YR 4/6m) mottles. Brown (10YR 5/3d), dark brown (10YR 3/3m) colour and massive structure are characteristic for the transitional BCca horizons defined by increased accumulation of calcium carbonate exceeding its original amount in the parent material (loess). Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m) silty material with common, fine to medium, white (10YR 8/2d) calcium carbonate mottles within a massive porous and friable structure represents the Cca horizons above the unweathered loess.

In the A and B horizons, abundant krotovinas and charcoal are present. Except for the present-day *Black Chernozem* (Sk_1), all other pedogenically equivalent soils have been classified as *Orthic Dark Brown Chernozems.* Nevertheless, it is likely that the original soils would have been classified as Chernozems, had they not undergone diagenetic changes after burial, including a partial bleaching of the humic horizon. Basically, all Chernozemic palaeosols are severely distorted by cryogenic processes, especially the thickest interglacial soils. Also two poorly preserved and altered soils are assigned to the Chernozemic Order from early mid-last glacial interstadials, with the lower soil (Sk_{12}) largely truncated and the upper soil (Sk_{11}) cryoturbated by numerous involutions and injections. All Chernozems are distinguished by strongly calcareous BCca horizons and mineral Bmk horizons with pH values of 6.0-6.4 for the last interglacial soils, 6.8 for the penultimate interglacial soil, and 7.2-8.0 for the truncated mid-last glacial horizon.

Micromorphology of the interglacial Chernozems is characterized by a few large and mostly disintegrating quartz grains, and by increased amount of the fine clayey matrix in B horizons as a result of progressed weathering processes (Figure 17-A). Organic matter accumulation in the form of humus aggregates is apparent in the Ah horizons (Figure 17-B). Overall, microstructure of the Chernozemic soils (especially of Bmk horizons) is not clearly observed, likely because of an insufficient development and/or because of compaction and secondary physical alteration by the thick overburden filling original peds.

4.4.5.2. Brunisolic Soils

Brunisolic soils are most frequently represented palaeosols in the section showing a different degree of pedogenesis. They have been separated into two categories on the level of subgroups within the group of *Eutric Brunisol.* The classification is based on the colour differences, soil structure grade, secondary calcium carbonate content, and frequency and degree of ferric and manganese oxide mottling.

Orthic Eutric Brunisols, being the most common (Sk_2, Sk_4, Sk_{10}, Sk_{19}, Sk_{20}, Sk_{23}, Sk_{25}, Sk_{29}, Sk_{31}, Sk_{32}), are distinguished by pale brown (10YR 6/3d), brown (10YR 5/3m) mineral horizons (Bmk) with a weakly developed, medium subangular blocky structure in a silt loamy, slightly sticky (w) to slightly hard (d) material (Figure 16-A). Usually, the upper part is colluviated; presumably thin A horizons are not preserved and were truncated if developed at all. Accumulation of organic matter is evident in more developed Brunisols (Figure 17-B) with some charcoal occasionally present. Transitional horizons (BCk and BCca), if present, have lighter yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4/m) silt loam, massive structure with some prominent or distinct yellowish red (5YR 5/8m) mottles, and secondary calcium carbonate accumulation in the BCca horizons. Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m) silty and massive, single-grain structure is the Ck horizon below.

Gleyed Eutric Brunisols (Sk_6, Sk_7, Sk_8, Sk_9, Sk_{13}, Sk_{14}, Sk_{18}) are characterized by a pale brown (10YR 6/3d) / brown (10YR 5/3m) to light yellowish brown (10YR 6/4d) / yellowish brown (10YR 5/4m) mineral (Bmkgj) horizon and a very weak to massive structure in a slightly sticky (w) / slightly hard (d) silt loam with common, fine to medium, prominent, yellowish red (5YR 5/8m) mottles indicating intensive gleying. Gleying also effects to a lesser measure the underlying pedogenically largely unaffected white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m) silty horizon (Ckgj) with few, fine and medium, olive yellow (2.5Y 6/8m) mottles, and a massive, single-grain structure. pH values vary from 6.8 to 8.0.

Microstructure of the incipient (less developed) Brunisolic mineral horizons is distinguished by a non-uniform texture with a dispersed grain arrangement (mainly of quartz), some increase of clayey matter, and fewer voids comparing to the parent material (Figure 16-B); and by large ferric mottles characteristic for initial oxidation processes, but without any apparent or only minor organic matter accumulation (Figure 16-A). More mature Brunisolic palaeosols show a further increase of illuviated organic matter, a more advanced oxidation in mineral horizons that approach in some extent typical Chernozems (Figure 17-B), as well as secondary accumulation of leached calcium carbonates. No A horizons has been clearly identified in any of the Brunisolic soils, most likely because of truncation. Some horizons were apparently redeposited by solifluction (and preserved as pedosediments), especially in the upper part of the section. Despite this fact, their classification to the Brunisolic soils is based on their original, climatically-determinant characteristics which are absent in unaltered loess.

Stratigraphically, Brunisols are generally located in the section between Chernozems and Gleyed Regosols. In comparison with Chernozems, Brunisols are much less disturbed by cryogenic processes, with only minor frost wedge casts penetrating their surfaces, as well as by fewer krotovinas, suggesting a reduced biotic activity and less severe climatic fluctuations.

4.4.5.3. Regosolic Soils

Seven pedostratigraphic units (Sk_3, Sk_5, Sk_{15}, Sk_{16}, Sk_{17}, Sk_{26}, Sk_{27}, Sk_{28}) described as *Gleyed Regosols* are distinguished by weakly developed pedogenic (Ckg) horizons discernible only on the basis of common, fine to medium, strong brown (7.5YR 5/8m) ferric mottles, and faint to distinct, black (2.5Y 2/0m) mottles formed within a brownish gray (2.5Y 6/2d) / dark grayish brown (2.5Y 4/2m) silty non-sticky (w) and soft (d) material of a single-grain loessic structure. The substratum is also partly mottled (Ckgj). Iron mottles are mainly distributed along vertical channel-like pores formed by grass rootlets. pH values vary from 7.2 to 7.8. There is no major difference in micromorphology of Gleyed Regosols comparing with unweathered loess, retaining the high porosity of the parent material, but with signs of incipient oxidation processes (*i.e.,* gleying) in the form of ferruginous cuttans. No cryogenetic processes have been observed in the Regosolic horizons.

Stratigraphically, Gleyed Regosols are located between aeolian or colluviated loess deposits and (Gleyed) Brunisols.

4.4.5.4. Cryosolic Soil

There is only one, but laterally easily-traceable and physically marked *Orthic Turbic Cryosol* characterized by involutions and cryoturbations (Sk_{11}). It has a dark grayish brown (10YR 4/2d) / very dark grayish brown (10YR 3/2m) Ahy horizon and a very weak, fine, subangular blocky structure in sticky (m), hard (d) silt loam. Abundant charcoal is present. The light brownish gray (10YR 6/2d) / grayish brown (10YR 5/2m) Bmgy horizon is formed by a sticky (m) and firm (d) silty material with massive structure and common, medium to large, distinct, strong brown (7.5YR 4/6m) mottles. Both A and B horizons are largely secondarily distorted by cryogenesis in the form of 10-50 cm x 2-15 cm lenses. The soil is developed on a light gray (2.5Y 7/2d) / light brownish gray (2.5Y 6/2m) silty massive loessic substratum (Ck horizon) with secondarily superimposed common, prominent, yellowish red (5YR 5/8m) to red (10R 4/8m) mottles and ferric, 1-3 cm large aggregations. Prior to the periglacial deformation, the original soil was a Chernozemic (Gleyed Dark Brown Chernozem). Micromorphology of an undisturbed part of the soil horizon is broadly analogous to other Chernozems, although structure (peds) is even less apparent than in the more developed interglacial soils.

4.5.6. Discussion and Conclusion

The palaeosol sequence comprises a series of qualitatively different soil units. Dark Brown Chernozems are the most distinctive units with a fully-developed soil horizon (Ah-Bmk-BCca-Cca-Ck) sequence; (Gleyed) Orthic Eutric Brunisols are characterized by mineral Bmk(gj) horizons; Gleyed Regosols are distinguished by mottled Ckg horizons.

In terms of soil structure morphology, weak to moderate fine granular structure with non-accommodating grains separated by larger compound voids is characteristic for Chernozems (*e.g.*, Bullock et al. 1985: fig. 47c). Most Brunisols are distinguished by a weakly to very weakly developed subangular blocky structure, sometimes difficult to discern from a massive, single-grain structural arrangement of the parent loessic material. In the most developed, although still initial Brunisols, the solid parent material is divided into subangular aggregates (peds) separated by short planar voids on all or most of their sides (*op.cit.*:fig. 47d). Rarely, they show a platy structure, presumably of initial eluviated mineral (Ae) horizons with horizontally elongated peds separated by planar voids (*op.cit.*: fig. 47f). However, this determination is not conclusive as the horizontal layering may also have resulted from subsequent frost action. Gleyed Regosols are characterized by a channel structure with short irregular vertical pores within the parent material matrix without any prominent pedogenic aggregates (*op.cit.*: fig. 46j). The progressive pedological development from the unaltered parent material into mature Chernozems starts with the initial gleyzation of loess expressed by brownish ferric mottles, but without apparent structural changes of the silty matter. Increased compaction under periodic reduction and oxidation conditions led to secondary mineral precipitation, clay formation. Advanced weathering is also evidenced by a lower angularity of most grains. A gradual organic matter accumulation, secondary calcium carbonate leaching and formation of advanced pedogenic horizons reflect more intensive loess weathering processes affecting the stabilized surface under prolonged climatically favorable conditions.

The gradual pedogenic alteration is well evidenced in the palaeosol series, with a generally increasing amplitude of weathering that can be specifically summarized in the following sequence:

loess --- Gleyed Regosol --- Gleyed Brunisol --- Brunisol --- Chernozem.

The initial alteration (reduction) of parent material is expressed by incipient gleying in a cold and humid environment and, and seasonally waterlogged setting. A progressive leaching of calcium carbonates from the original loessic substratum into the lower part of the pedogenic profile accompanied by organic matter accumulation in A horizons and brunification of B horizons reflects a gradual increase of summer temperatures and surface stability that contributed both to a prolonged weathering process and formation of Brunisolic soils (subsequently partly colluviated) and, under more continental conditions, of Chernozemic soils. Nevertheless, some Chernozems (*e.g.*, the gleyed mid-last glacial Cryosol) are less "mature" than some well-developed interglacial Brunisols characterized by a higher degree of mineral weathering, but a lower amount of organic matter accumulation. Inferred palaeoenvironmental reconstruction and a general palaeoclimatic evolution are discussed in the following section.

5. PALAEOCLIMATIC RECONSTRUCTION

In terms of palaeoclimatic implications and palaeoenvironmental reconstruction of the section Kurtak 29, several basic variables were taken into consideration in a uniform methodological approach. These include the specific, climate-diagnostic characteristics of particular palaeosol units (*i.e.*, the degree of soil weathering, the thickness and character of mineral horizons), the degree and forms of periglacial deformation (*i.e.*, cryoturbation, involutions, frost wedge casts, seasonal permafrost depth), and the stratigraphic distribution and the character of loess relative to pedogenically altered palaeosol units. Additional palaeoecological evidence including palaeomagnetic, palynological and palaeontological data is provided in the following chapters. The interpretation of the geological record follows the natural stratigraphic order starting with the basal deposits.

As the lowermost part of the section is not exposed, the earliest deposits following the deposition of the 60 m terrace are unknown or only poorly documented (Figure 5). Evidence for older (Early Pleistocene) deposits from other parts of the Berezhekovo Sector is fragmentary and does not provide a coherent, spatially traced picture. The first recorded palaeosol unit in the section, preserved as a light brown Bmkg horizon of a Brunisolic soil (Sk_{32}) (see Appendix A), is developed on a calcareous substratum of a colluviated loess. The high pH status of the parent material and the characteristic soil morphological features suggest forest or shrub vegetation under humid and cool climatic conditions, suggesting a relatively well-drained site as indicated by activated calcium carbonates illuviated into the BCca horizon. Occasional ferric mottles indicate a seasonal saturation of the subsurface layers. A drop in temperatures is assumed to account for a total rearrangement of the soil structure, presumably by periodic frost action under a humid climate, into a horizontally inclined (30°), uniformly 0.5 cm-bedded mass. A conformable, gradual transition of the overlying unstratified aeolian loess into a progressively layered and intercalated fine brownish-grayish laminated (colluviated loess) unit indicates a fluctuating climatic trend with humid oscillations. Distinct contacts between the interbedded light and dark laminae in the middle part of the deposit suggest a seasonal (?) climatic fluctuation of short-term cold and dry environmental conditions, and more moderate and humid conditions, respectively (see the niveo-aeolian loess facies; page 23-24).

Following a cold interval evidenced by massive loess deposition, more warming occurred with a decreased loess sedimentation rate, leading to formation of a

Brunisolic soil (Sk_{31}), indicating a forest / shrub cover. The brownish Bmk and the BCca horizons indicate a broadly similar condition as during the formation of the previous palaeosol. Subsequent periglacial activity is documented by involutions disturbing the original soil surface and possibly eroding the A horizon.

After deposition of the overlying loess cover, a next warm but drier and prolonged climatic interval relates to formation of the lower Chernozem (Sk_{30}) with a thick, dark brown Ahy horizon and well-developed mineral horizons (Bmk, BCca), marking prolonged pedogenesis interval and surface stability in an open steppe setting. As in the lower Brunisol, krotovinas and detrital charcoal are common, showing a high rate of biogenic activity and frequent grassland fires. The shift towards an increasingly continental semiarid climate is indicated by the dark colour of the humic horizon with organic material introduced by the root system of forbs and grasses that suggest warm summers, but cold winters.

The first recorded climatic optimum (defined by the highest summer temperatures and represented by the above Chernozem) was followed by dramatic cooling, evidenced by deep frost wedge casts which disrupted the original Chernozemic soil sequence into a flame structure with wavy tongues filled with organic material of the severly disturbed Ah horizon. Cooler temperatures persisted after this event as evidenced by a reactivated loess deposition accompanied by ongoing pedogenesis. The thick brown forest soil (Sk_{29}) with more shallow frost wedge casts in the lower part of the mineral (Bmky) horizon (compared to the underlying Chernozem) was developed under a more humid but still cold climate. The lack of a well-defined soil structure most likely reflects rapid pedogenic formation (Figure 18).

The subsequent cooling and accelerated loess deposition was temporarily slowed by two warming episodes, leading to formation of a Gleyed Regosol (Sk_{28}) distinguished by common, brown ferric mottles and black mottles precipitated in a light gray loess matrix. Dark brown iron oxides distributed along 1-2 cm vertical tubular channels of former rootlets indicate an arctic steppe-tundra with forbs and shrub vegetation in a periglacial setting with a ground water percolation and periodic surface saturation.

After a prolonged and/or more intensive period of loess sedimentation represented by about a 1m thick and unweathered aeolian loess unit, defining the first well-marked climatic minimum, the study area experienced a long series of sharp and short oscillations of cool / humid intervals, and cold / arid intervals. This relatively long period of short climatic oscillations is evidenced by the typical layered structure of the second loess facies with intercalated brownish silty-loamy and grayish to olive silty undulating and inclined laminae, a deposit more than 3 m thick, overlying Regosolic soil (Sk_{27}). The sharp bedding contacts, the distinctive nature of the 1-10 mm individual layers and presence of numerous, minor (0.5-3 cm) frost wedge casts throughout the deposit support the idea of very marked climatic fluctuations, possibly on a seasonal basis. The fine layered structure is assumed to have resulted from a periodic accumulation of the loess dust on a subsequently melted snow cover, and later on a barren seasonally saturated surface. Such a niveo-aeolian process is proposed to account for the colluviated nature of the deposit, largely analogous to the stratified loess in the basal part of the section (Figure 6). Another Gleyed Regosol (Sk_{26}), evidenced by a Ckg horizon, is developed on top of the colluviated loess unit before a new cooling interval, marking the second climatic minimum evidenced by a aeolian loess accumulation into a weakly stratified to massive unit.

Following a new cold interval of aeolian loess sedimentation, a climatic warming took place indicated by the formation of a Brunisol (Sk_{25}) with a light brown Bmk horizon and calcareous BCca horizon, which likely developed under a parkland / forest vegetation at a relatively well-drained site. A further decrease of loess deposition related to establishment of more continental conditions with higher seasonal temperature variation marks formation of the second Chernozemic soil recorded in the section (Sk_{24}). The thick, dark brown Ahy horizon is characteristic of a cold semiarid climate with mesophytic grasses and forbs. The rather weak fine granular structure suggests an early stage of soil formation process. As in the lower Chernozem, intensive biological activity is indicated by numerous krotovinas (rodent holes). The lack of well-developed structure in these steppe soils has been suggested to result from a more rapid pedogenesis than under brown forest soils with prismatic and subangular blocky structures (Dergacheva *et al.* 1984:18). A period of cold climate is evidenced by prominent, up to 1 m deep and 30 cm wide frost wedge casts vertically dissecting the Chernozemic soil down to the parent material (Figure 19-A). Such well-developed frost wedge casts would imply a time interval of several hundreds or even a few thousands of years (C. Schweger, personal communication 1995).

A subsequent pedogenically favorable climatic interval is revealed by a superimposed Brown Chernozem or a mature Brunisol (Sk_{23}) preserved in the form of a mixed, intensively bioturbated, yellowish brown to dark brown Bmk horizon. Later climatic deterioration is evidenced by involutions that preceded as well as postdated deposition of the overlying layer of pure loess. Marked frost wedge casts cutting from the loess surface are related to next episode of extreme cold with no vegetation present. This is followed by next interval of advanced pedogenesis under moderate, but still generally cold climatic conditions, as documented by later Dark Brown Chernozem (Sk_{22}) with severely bioturbated and subsequently cryoturbated, though still well-recognizable soil profile with the Ah-Bmk-BCca-Cca horizons. The environmental setting was probably similar to that of the lower Chernozem developed under a cold semiarid steppe, but evidently within a shorter time span as evidenced by a reduced thickness of the soil horizon. Pedogenic development renewed after surface stabilization following a short interval of loess sedimentation. Nevertheless, a tendency toward decreased continentality and a reduced biotic activity is evidenced by a thinner Ah horizon of the superimposed Dark Brown Chernozemic soil (Sk_{21}). This trend is also indicated by frost wedge casts limited to the upper part of the soil, and less intensive subsequent cryoturbation than found in the lower Chernozems (Figure 19-A).

A later weathering interval led to formation of a light yellowish brown Brunisol (Sk_{20}), suggesting a more humid and less seasonal cool climate with invasion of boreal forest. Ongoing loess deposition of increasing magnitude is expressed by weakening of the poorly developed mineral horizon (Bmk). Colluviation of the uppermost part of the soil corroborates the cooling trend with reduced evaporation, surface runoff (likely from snow melting) and initiation of solifluction processes. The superimposed 10 cm thick loess cover indicates a short cold and dry episode.

The subsequent surface stabilization and moderate temperature increase is evidenced by the formation of a Brunisol (Sk_{19}). The decreased thickness, the darker colour and the weakly developed subangular blocky structure of the Bmk horizon suggest a cold and semiarid parkland / steppe vegetation. As in the previous Brunisol, solifluction processes became active after the end of this pedogenic interval.

The following loess deposition took place under simultaneous, but reduced weathering processes in a cold environment. The development of a pale brown Gleyed Brunisol (Sk_{18}) is indicative of an ephemeral cold and semiarid forest-tundra / steppe-tundra vegetation established on seasonally saturated ground with a high water table. The superimposed Gleyed Regosol (Sk_{17}) suggests a further cooling and a waterlogged surface expressed by abundant mottling of the loessic substratum. Absence of the A and B horizons, the vertical iron oxides precipitated along rootlets, and horizontal rusty layer superimposed over the former Brunisol indicate a sparse arctic tundra vegetation on top of a thin, seasonally melted subsurface layer approximately 50 cm above permafrost (Figure 19-B).

The overlying series of light brownish gray to light olive gray loess intercalated with partly colluviated Gleyed Brunisolic Bmkgj / Gleyed Regosolic horizons (Sk_{16}) is correlated with the third recorded climatic minimum. The massive, single-grain and weekly laminated aeolian loess structure shows progressive loess accretion under a very cold and hyperarid climate. The incipient soil horizons, distinguished by a brownish colour and varying degree of mottling, suggest a short warmer interval. Occasional krotovinas in the incipient B horizons in this (Tschaninskaya) loess filled by a darker humic material suggests presence of thin A soil horizons. An indicator of prolonged surface saturation (waterlogging) under a cold climate is another Gleyed Regosol (Sk_{15}) overlying the second brownish colluviated unit prior to deposition of the superimposed loess cover.

A general warming trend in the upper part of the loess section is evidenced by the appearance of a series of Brunisols intercalated with light brownish gray beds of loess without sharp upper and lower boundaries, suggesting a differential rate of loess sedimentation and pedogenesis. The accompanying decrease of ground saturation and lowering of the water table due to progressive lowering of the permafrost table is indicated by a gradual pedogenic development from a Gleyed Brunisol (Sk_{14}) with an intensively mottled Bmkgj horizon and a massive structure, through a pale brown to brown, largely colluviated series of three Brunisolic horizons (Sk_{13}) characterized by partly decalcified Bmkj horizons with a few iron mottles in the basal part and a weakly developed subangular blocky structure, into a mature Brunisol (Sk_{12}) with a yellowish brown Bmk horizon underlain by a well-developed BCca horizon enriched by secondary carbonates. In terms of vegetation change, this pedogenic pattern can be interpreted as a shift from arctic tundra steppe through tundra forest into boreal forest or a mixed shrub-forest. The gradual increase of soil permeability from the basal waterlogged Gleyed Regosol into the well-drained uppermost Brunisol is also reflected in the re-appearance of biological activity present in the later unit in the form of krotovinas and abundant charcoal. Although the latter soil (Sk_{12}) is classified as *Orthic Brunisol,* it may in fact represent the lower part of a truncated (Orthic) Chernozem. If so, a moderately cold semiarid, transitional forest-grassland vegetation with some trees, shrubs, forbs and grasses can be suggested. This interpretation is supported by the strongly developed BCca horizon, identical with those observed in the lower, completely preserved Dark Brown Chernozems. The carbonate horizon indicates better drainage and advanced pedogenesis. In any case, this unit clearly defines the third climatic optimum recorded in the section.

The following cold and arid episode with loess deposition preceded another, but likely a short and/or less intensive pedogenic interval evidenced by the ***Dark Brown Chernozem*** (Sk_{11}) with a distinctive dark brown Ah horizon. The specific characteristics of the soil with abundant charcoal indicates a cold semiarid climate with parkland vegetation during the time of soil formation. A decrease of annual temperatures and evaporation led to a rise of the permafrost table, causing intensive cryoturbation with involutions, followed by a marked gleyzation and a partial bleaching of the soil, which was eventually developed into an *Orthic Turbic Cryosol,* possibly in a forest tundra setting. A wood of a pine tree was recovered from this soil in the northern part of the Berezhekovo Site dated to about 30 ka (see below). The Cryosol shows a varying amount of humus. Numerous iron concretions are present as a result of precipitation of oxidized ferric minerals (Figure 20).

A prolonged, stable period of pedogenesis with simultaneous, relatively moderate loess deposition under a generally cool and semiarid to subhumid climate is evidenced by formation of a thick Brunisol (Sk_{10}) with a yellowish brown Bmk horizon and a weak subangular blocky structure, probably because of the ongoing loess accumulation. A thin layer of brownish colluviated loess separates the soil horizon from a pale brown Gleyed Brunisol (Sk_9) with a mottled and colluviated Bmkgj horizon that indicates seasonal ground saturation, presumably in a boreal forest or tundra-forest setting. A general trend of climatic deterioration is recorded in a subsequent series of three weakly developed Gleyed Brunisols (Sk_8, Sk_7, Sk_6), morphologically identical with the previous palaeosol, and a Gleyed Regosol (Sk_5), all intercalated by beds of light gray to white loess, increasing in thickness in the upper part of the sequence. The following Brunisol (Sk_4), with a better developed medium subangular blocky structure in the Bmk horizon (than the previous Brunisols) with krotovinas, indicates a temporal warming interval with surface stabilization and forest re-establishment at a relatively better-drained site. The overlying Gleyed Regosol (Sk_3) signals the beginning of a climatic deterioration and a rise of the permafrost table.

Gradual decrease of annual temperatures, pronounced aridity and reactivated aeolian activity relates to progressive aggradation of loess into a nearly 3 m thick sedimentary sequence. The increasing thickness of individual strata towards the upper part of the unit with a gradual colour change from light brownish gray to light gray marks the fourth climatic minimum in the stratigraphic record. The pale brown stratum at the top of the section interpreted as a Brunisolic Bmk horizon suggests invading vegetation. A short cold interval with deposition of loess precedes a final period of intensive weathering that produced the recent Black Chernozemic soil at the top. Its formation relates to the present-day strongly continental semiarid climate, with short, warm summers and dry, cold winters with little snow cover, and a seasonal distribution of precipitation.

In terms of palaeoclimatic evolution, three major cycles can be recognized showing a patterned succession of loess deposition followed by soil formation. Although with variations, some reoccurring palaeoclimatic patterns are evident (Figure 18). The first (lower) cycle (C_1) can be determined by the Brunisol (Sk_{31}) and the loess below analogous Brunisol (Sk_{25}); the second cycle (C_2) by the Brunisol (Sk_{25}) and the loess unit below the mature Brunisol / truncated Chernozem (Sk_{12}); the third cycle (C_3) by the latter Brunisol/Chernozem (Sk_{12}) and the loess below an incipient Brunisol (Sk_2). Particular environmental trends can be consistently followed from the defined intervals throughout the climatic minima and optima. Onset of the fourth cycle (C_4) is evidenced by the latter Brunisol below the present-day Chernozem. A segment of another cycle from the basal part of the section is not included in the discussion because of its short geological record.

In sum, the three defined and fully-recorded cycles start with formation of diagnostic Brunisols (Sk_{31}, Sk_{25}, Sk_{12}) in the loessic substratum. These are distinguished by well-developed Bmk horizons with krotovinas (C_{1-3}) and moderately to well-developed BCca horizons. Separated by a thin cover of aeolian loess, the Brunisols are followed by Chernozemic soils (Sk_{30}, $Sk_{24-22-21}$, Sk_{11}). All other Chernozems are well developed with diagnostic Ahy horizons. The first four are disrupted by deep frost wedge casts; the fifth Chernozem is cryoturbated and involuted into a Cryosol (Figures 21-23). The last (present-day) Chernozem is still intact, although minor frost wedge casts are already apparent. In the longest warm cycle (the beginning of C_2), possibly another Chernozem is present, equally disturbed by frost action. The later development is uniform for all three cycles, starting with formation of thick Brunisols, which occur either as a single unit (Sk_{29}, Sk_{23}), or in a series (Sk_{20-19}, Sk_{10-9}), succeeded by initial mineral brown soil horizons with a less-advanced degree of pedogenesis (Sk_{18}), which may be interbedded by thin, partly colluviated loess layers (Sk_{8-6}, Sk_4), and finally superimposed by Gleyed Regosols (Sk_{28}, Sk_{17}, Sk_5, Sk_3).

The following periods of loess accumulation mark the intervals of climatic minima. In the first cycle (C_1), most of the loess is colluviated and only its lower and upper parts preserved the original aeolian nature of the silty sediment. The transitional development of progressive warming between the cycles either follows the opposite Gleyed Regosol - Brunisol trend (C_2), or shows a dramatic temperature increase leading to fully developed Brunisols.

The general climatostratigraphic development based on the palaeopedological record can be subdivided in four intervals. The initial interval for all cycles relates to a gradual establishment of strong continental conditions within a semiarid environment, with highly fluctuating seasonal temperatures (*i.e.*, warm summers and cold and dry winters), and reduced aeolian activity. Less seasonal temperature variation with increased humidity and higher ground watertables, with gradual surface saturation under cold climatic conditions due to progressive permafrost buildup, characterize the second interval. This is followed by the third interval with extremely cold temperatures, hyperarid climate and intense loess aggradation. The last interval marks a warming trend leading to establishment of the climatic optima. The overall amplitude of climatic change is less pronounced in the first cycle (C_1), where most of the loess is colluviated, thus suggesting more balanced long-term temperatures with frequent humid oscillations and snowfall. In the later cycles (C_2, C_3), the climatic minima as well as maxima are more pronounced, marked by prolonged loess sedimentation as well as soil formation periods. In the first part of the second cycle, gradual decrease of continentality is clearly evidenced by the lower magnitude of both pedogenesis (as marked by the series of Chernozemic and Brunisolic soils) and cryogenesis (manifested by frost wedge casts and involutions). Nevertheless, the assumed shortening of pedogenic phases during the second half of the Late Pleistocene (Zykina 1992:106), has not been observed. During the mid-last glacial optimum, correlated with the Brunisol / Chernozem (Sk_{12}), the climate was apparently warmer and more humid than at the present time. An analogous palaeoenvironmental pattern has been also recognized from the Angara River valley (Drozdov 1992:14).

The same palaeoclimatic succession has been found in western Siberia in the Late Pleistocene record, with cyclic processes of loess deposition during dry and cold periods, soil formation under warm and humid conditions, and subsequent cryoturbation due to climatic cooling (Volkov and Zykina 1991:41). A similar sequence has also been suggested for the Mezin Pedocomplex of the last (Mikulino) interglacial in the Russian Plains, evidenced by three palaeoclimatic phases (Velichko and Morozova 1985; Velichko *et al.* 1985). These include the *protoevent*, characterized by texture and lithological properties of the parent material (loess), and cryogenic and other features inherited from the previous glacial stage after the loess deposition; the *synevent,* including the active pedogenic processes (*i.e.,* organic matter accumulation, mineral migration, lessivage, *etc.*); and *postevent,* represented by secondary periglacial deformations.

Therefore, the palaeoclimatic cycle documented in southern Central Siberia as well as other parts of northern Eurasia is characterized by loess deposition under dry and cold climatic conditions, with increased warming and surface stability leading to progressive pedogenesis, followed by intensive solifluction and cryoturbation during a cold but humid periglacial climate. Based on the palaeopedological evidence, this general pattern corroborates the vegetation change at the Kurtak Site with barren arctic tundra correlated with intensive loess deposition periods, replaced by steppe-tundra and forest-tundra during the initial warming intervals. Expanded boreal forest was later replaced by continental shrub steppe and forest steppe during the climatic optima. Increased humidity and less marked seasonality provided conditions for recolonization of the boreal forest. A further decrease of temperatures and humidity led to establishment of forest-tundra and arctic meadow.

The above climatostratigraphically and palaeopedologically specified cycles are correlated with the last glacial and interglacial period (C_{3-2}), and most of the penultimate glacial / interglacial period (C_1). Relative and absolute chronology of the geological record is provided in the following chapter.

6. CHRONOLOGY OF THE KURTAK LOESS-PALAEOSOL RECORD

The radiocarbon chronology of the Kurtak loess-palaeosol provides the most complete absolute chronological record for the last 40-35,000 years (Orlova *et al.* 1990). A series of radiocarbon dates has been obtained from the upper loess sections of the Kurtak sites (Table 2). Most of the dates are on charcoal detritus dispersed within archaeological horizons. In most cases, the dates are chronostratigraphically consistent and can be, therefore, considered as reliable. Exceptions are three radiocarbon dates on charcoal from the well developed Chernozem (the Kamenolozhskaya Soil) -- two from Kurtak IV (horizon 7 and 9) (34,800±1,200 B.P., 35,600±1,200 B.P.), and one from Berezhekovo 21 (horizon 6) (38,300±1000 B.P.). These, however, must be considered as dates on secondarily intruded charcoal as they contradict the well-established last interglacial stratigraphic position of the Chernozemic soil. In general, ^{14}C dates provide the key chronostratigraphic control for the more recent palaeomagnetic susceptibility and remanence records, which are discussed below.

The loess - palaeosol sequence in Section 29 can be correlated with other sections in the Kurtak area. The existing radiocarbon dates from other sites, mainly derived from archaeological strata, can be used to establish a preliminary chronological framework allowing an approximate temporal fixing of particular fossil soil and loess horizons. The following is a short summary of the available data, assessing the Kurtak loess-palaeosol chronostratigraphy, starting from the more recent record.

At the Kamennyy Log Site (Figure 2), five unevenly developed pedogenic horizons (including the recent soil) with krotovinas and about the same number of solifluction episodes have been identified throughout a 15 m thick stratigraphic sequence, with the last interglacial Chernozemic soil at the base. A radiocarbon date of 19,000±1,500 B.P. was obtained from the second fossil soil, classified as Kastanozem (Drozdov *et al.* 1990b:26-29). The profile documents the cyclic nature of a fluctuating climate disrupted by short periods of deposition of secondarily colluviated loess. A series of gleyzation horizons, identical with those from the upper part of the Kurtak Section 29 have been recognized in Excavation I (Strata 2-15) within the upper 5-5.7 m thick loess deposits above the interglacial soil. In agreement with the present interpretation, these are believed to indicate humid oscillations under generally cold climatic conditions (Drozdov *et al.* 1990a:79-83; Drozdov *et al.* 1990b:54; Dergacheva *et al.* 1992). Except for the morphostratigraphic, visually recognized soil horizons, an additional seven humic buried soils are believed to have been identified, partly in superimposed positions, on the basis of organic carbon maxima and other variables (calcium carbonate contents, pH, magnetic properties, clay coating, humus chemistry and ratio of humic acids) (M.I. Dergacheva, unpublished data; personal communication 1993). In the Russian terminology, these soils are described as steppe and forest-steppe Chernozems from the last interglacial; southern Chernozem from the early last glacial (Zyryansk) interstadials; Chernozem-like soils and gray forest soils from the mid-last glacial interstadials (Karginsk 2 and Karginsk 1, respectively) (Zykina 1992:105).

A similar initial pedogenic pattern has been observed at the Kurtak IV Site, with three radiocarbon dates on charcoal between about 23,500 and 24,200 B.P. from two brownish loamy horizons in the upper part of the section, corroborating dates produced from the Kamennyy Log Site (Table 2). Although an exact stratigraphic correlation of these units among the individual sections cannot be positively provided, the ^{14}C dating supports an early Sartan (early Late Wisconsinan) age. Intensive gleyzation processes due to warming oscillations between periods of loess deposition are indicated by Gleysols which preserved as initial B horizons (Demidenko 1990a:42). Similarly, two Luvisolic (?) palaeosols from the late Sartan deposits at the Berezhekovo I Site with Bmk and BC horizon have been reported (*op.cit.*:44-45). Analogous, poorly developed (Brunisolic and Gleysolic) soils were widely distributed in western Siberia during the Late Pleistocene (Dergacheva *et al.* 1984:18).

The series of B horizons in the middle part of the key Section 29/I, assigned to the Kurtak Pedocomplex (Figure 20), is differentially preserved and visible in the Kurtak area. The Cryosol (Sk_{11}) in the upper part of the pedocomplex, represents the best chronologically fixed and well-defined chronostratigraphic marker in the area. It has been dated by radiocarbon on charcoal to 29,410±310 B.P. at Berezhekovo Section 21 (horizon 3), to 29,400±400 B.P. at the Kashtanka Site (horizon 11) and on wood from the southern part of the Berezhekovo Site to 30,400±700 B.P. The lower soil (Sk_{12}) was dated at > 30,000 B.P. on charcoal, also from the nearby Berezhekovo Section 21 (horizon 3) (Table 2). This soil is tentatively compared with a buried brown soil on top of the 15-18 m terrace (II) in the Angara River valley, dated to 39-30 ka (Drozdov 1992:14).

In the broader geographical context, the Kurtak Pedocomplex can be correlated with the Iskitimskiy Pedocomplex of Western Siberia (Zykina *et al.* 1981; Volkov and Zykina 1984). Palaeoclimatologically, it has been assigned to the Lipanovsko-Novoselovo Interstadial (30-22 ka) of the late Karginsk (late Mid-Wisconsinan) with formation of the initial Black Chernozemic soil, subsequently disturbed by early Sartan solifluction (Demidenko 1990a:39), *i.e.*, manifesting the same palaeoclimatic pattern as at Kurtak.

The thick loess deposits with Gleyed Regosols and Gleyed Brunisols below the Kurtak Pedocomplex remain undated as these are beyond the ^{14}C -dating limits. The single date of > 35,000 B.P. on charcoal from loess below the Kurtak Pedocomplex at the Kamennyy Log Site (Derevianko *et al.* 1992a:57) supports the assumption that these sediments were accumulated during the Zyryansk (Early Wisconsinan) Glacial Stage.

The underlying series of buried soils showing a progressive pedogenic development down the stratigraphic profile represents the so called Sukholozhskiy Pedocomplex recognized in the Kurtak area (Drozdov *et al.* 1990a). Although no absolute dates are available from this section, it is presumably late last interglacial *(sensu lato)* (in the Russian classification system early last glacial in age; *i.e., ca.* 100-75 ka) based on its stratigraphic position and a degree of weathering processes. In the most complete Kurtak sections (Berezhekovo and Kamennyy Log) the pedocomplex is over 2 m thick, disrupted by frost wedge casts about 1-5 cm wide and 10-50 cm deep. Three cycles have been observed in the Kamennyy Log Section 9, with loess deposition followed by formation of a Kastanozem distorted by later solifluction and periglacial deformation (Chekha 1990:28). Two fossil soils with little differentiated A-BC horizons have been described (Zykina 1992:104) from the site with the lower 0.5-0.8 m thick loess substratum (Derevianko *et al.* 1992a:57).

The best-developed Chernozem in the study section -- the Kamenolozhskaya Soil -- (Sk_{24}) is correlated with the last interglacial climatic optimum and was originally described at the Kamennyy Log Site, where it is formed on silty-sandy deposits. The problematic nature of the two dates of 34,800 B.P. and 35,600 B.P. produced on charcoal from the upper part of the Ah horizon has been mentioned. These are considered to be invalid because of the assumed secondary intrusion of organic material by later solifluction. The dates may be too young, as they approach the technical potential of the conventional radiocarbon dating method as evidenced by the high range of error (Table 2). The dates are also not chronologically consistent with incorporated rodent taxa (*Myospalax myospalax* Laxmani) and the early form of mammoth *(Mammuthus primigenius)*, as well as morphology of the associated, typical Middle Palaeolithic ("Mousterian") stone tools recorded on top and within the Chernozemic soil at that site (Derevianko *et al.* 1992a:58). In view of the strongly developed pedogenic profile of this soil, pedomorphologically identical with the present-day soil, the

Kamenolozhskaya Soil was definitely formed during the last interglacial *(sensu stricto)*. It correlates well with the Berdskaya Soil of Western Siberia (the eponymous site at Berdsk, *ca.* 80 km south of Novosibirsk), assigned to the Kazantsev (Sangamonian) climatic optimum, *ca.* 125 ka (Volkov and Zykina 1982, 1984). The two superimposed Chernozemic soils (Sk_{22}, Sk_{21}) above the Kamenolozhskaya Soil, recognized in Kurtak 29 section are believed to have formed at a later part of the same interglacial period. The underlying Brunisol (Sk_{25}) is accordingly assigned to the early stage of the last interglacial.

The second well-developed and nearly identical Dark Brown Chernozem (Sk_{30}) with two Brunisolic horizons (Sk_{31}, Sk_{29}) in the lower part of the Kurtak Section 29 below the Kamenolozhskaya Soil, separated by a 3.5 m thick package of a colluviated loess and a series of Regosolic and Brunisolic soils, may be correlated with the penultimate interglacial (the Tobolsk - Mindel/Riss - Yarmouth Interglacial). The massive largely colluviated loess deposit between the two interglacial soils would therefore represent the Riss / Illinoian Glacial Period. The analogous niveoaeolian colluviated deposits below the second interglacial soil, with a Brunisolic soil near the base of the section, are the earliest recorded units of the stratigraphic profile. Because of the normal palaeomagnetic polarity within the Brunhes Chron, they are likely late Middle Pleistocene (*ca.* < 300 ka).

Chronology of the earlier Quaternary record in the Kurtak area is insufficiently known. At Berezhekovo, two loess formations, including two pedocomplexes each, are presumably separated by a palaeomagnetic reversal (V.P. Chekha, personal communication 1993), and stratigraphically expressed by an erosional contact that cuts under an angle of 50° the underlying silty-clayey formation. The first (basal) pedocomplex of the lower formation includes rodents of the Tiraspol Fauna assigned to the Middle Pleistocene. The upper pedocomplex of the lower formation was dated by thermoluminescence to 450±42 ka and is correlated with the Belovo Pedocomplex on the Ob Plateau of Western Siberia (TL 510±30 ka) (Arkhipov *et al.* 1992:11-12). The lower pedocomplex of the upper formation dated to 130±10 ka at Kurtak is correlated with the Berdsk PK of Western Siberia (140±14 ka). The uppermost pedocomplex includes two palaeosols (located above the palaeomagnetic Blake Excursion, 117-111 ka; Løvlie 1989b:6) is correlated with the Iskitimskiy Pedocomplex (Arkhipov *et al.* 1992). On the basis of the TL dating, Strata 5-7 from Excavation 1 at Berezhekovo are assigned to the late Middle Pleistocene, to a time span chronologically estimated to between 270-130 ka (Drozdov *et al.* 1990a:63).

Regardless of certain chronostratigraphic consistence, the TL dates cannot be considered as totally reliable because of technical problems of the method. They therefore can provide only an approximate chronological control. In the process of investigations, a series of 25 TL samples was taken from selected loess and palaeosol horizons in the section Kurtak 29. Unfortunately, no laboratory measurements have been carried out yet because of the lack of funding. The dates should help establish the chronostratigraphic framework for the high-resolution loess-palaeosol record.

In spite of the chronological shortcomings, the loess - palaeosol record from the Kurtak 29 section is the longest and most complete of all stratigraphic sequences from the Kurtak area, comprising the last glacial and interglacial cycle, and a significant part (if not all) of the previous (penultimate) glacial and interglacial cycle. Because of the colluviated nature of loess in the lower part of the section, some temporal hiatuses cannot be fully excluded, because of sharp, although conformable contacts of the main loess unit. Nevertheless, a temporal continuum of the loess-palaeosol record in Section 29 at least since the late Middle Pleistocene is sufficiently supported. Chronostratigraphic correlation with other palaeosols from southern Siberia is provided below (page 90).

Date (B.P.)	*Site / Section (Horizon)*	*Material*	*Context*
14,300±100 (LE-1457)	Kurtak III / R_2[5]	charcoal	loess
14,390±100 (LE-1456)	Kurtak III / R_1[5]	charcoal	
14,600±200 (GIN-2101)	Kurtak III / R_2[5]	charcoal	
16,900±700 (GIN-2102)	Kurtak III / R_1[5]	charcoal	
19,000±1,500 (IGAN-1046)	Kamennyy Log I (hor. 13)[1]	charcoal	
20,800±600	Kashtanka (hor. 9)[6]	charcoal	Kurtak PK
21,800±2,000 (IGAN-1049)	Kamennyy Log I[1]	charcoal	
23,830±2,000 (IGAN-1050)	Kamennyy Log I[1]	charcoal	
23,470±200 (LE-2833a)	Kurtak IV (hor. 11)[5]	charcoal	
23,800±900 (LE-4155)	Kurtak IV (hor. 11)[5]	charcoal	
24,000±2,950 (LE-4156)	Kurtak IV (hor. 11)[5]	bone	
24,170±230 (LE-3351)	Kurtak IV (hor. 11)[5]	charcoal	
24,400±1,500 (IGAN-1048)	Kamennyy Log I[1]	charcoal	
24,800±400 (GIN-5350)	Kurtak IV (hor. 11)[5]	charcoal	
24,805±425	Kashtanka (hor. 9)[6]	charcoal	Kurtak PK
24,890±670 (LE-3357)	Kurtak IV (hor. 11)[5]	bone	
27,470±200 (LE-2833)	Kurtak IV (hor. 12)[5]	charcoal	
29,400±400 (GIN-6999)	Kashtanka (hor. 11)[6]	charcoal	Kurtak PK
29,410±310 (SOAN-2806)	Berezhekovo 21 (hor. 3)[2]	charcoal	Kurtak PK
> 30,000 (SOAN-2807)	Berezhekovo 21 (hor. 3)[2]	charcoal	Kurtak PK
30,400±700 (AECV-1938C)	Berezhekovo - North[7]	wood	Kurtak PK
31,650±520 (LE-3352)	Kurtak IV (hor. 17)[5]	charcoal	
32,380±280 (LE-3638)	Kurtak IV (hor. 17)[5]	bone	
34,800±1,200 (GIN-6090)	Kurtak IV / 9[3]	charcoal*	Km. Soil[8]
35,600±1,200 (GIN-6089)	Kurtak IV / 7[3]	charcoal*	Km. Soil
38,300±1000 (GIN 6088)	Berezhekovo 21 (hor. 6)[4]	charcoal*	Km. Soil
> 35,000 (SOAN-2805)	Kurtak IV[2]	charcoal	loess
> 42,100 (AECV-1939C)	Ust'-Izhul'[7]	bone	colluvium
> 40,050 (AECV-2033C)	Ust'-Izhul'[7]	charcoal	colluvium
> 41,810 (AECV-2032C)	Ust'-Izhul'[7]	charcoal	colluvium
> 42,190 (AECV-2034C)	Ust'-Izhul'[7]	charcoal	colluvium

Table 2. Radiocarbon dates from the Kurtak area ([1]C[illegible]kinskiy *et al.* 1990; [2]Drozdov *et al.* 1990a; [3]Drozdov *et al.* 1990b; [4]Drozdov *et al.* 1990c; [5]Svezhentsev *et al.* 1992; [6]Derevianko *et al.* 1992a; [7]Chlachula, this volume; [8] Kamenolozhskaya Soil. *Contaminated stratum (secondary intrusion of dated organic matter). For site location see Figure 2.

7. MAGNETOSTRATIGRAPHY OF THE KURTAK LOESS-PALAEOSOL RECORD

In spite of intensive geo-archaeological investigations in the Kurtak area, the chronostratigraphic framework is still imperfectly established and the temporal scale is poorly known, especially in the lower parts of the loess sections. Magnetostratigraphy of the key investigated section was carried out for relative dating and reconstruction of the past climatic change in the area. Absolute dates provided by the radiocarbon technique are limited to the upper part of the geological sequence with most of the earlier loess - palaeosol record having little if any direct chronological control. The current studies have therefore focused on expanding the existing chronological scale by applying palaeomagnetic (remanence and susceptibility) dating techniques to provide a relative assessment of the time and rate involved in the stratigraphic formation. Magnetic susceptibility also proved to be the most exact and informative high-resolution palaeoclimatic proxy method applied in conjunction with the palaeopedological data.

7.1. MAGNETIC REMANENCE

7.1.1. Principles and Application

Palaeomagnetic remanence studies were undertaken for the purpose of relative chronostratigraphic dating. Except for the long-term palaeomagnetic reversals, magnetic remanence variations in the form of short, high-amplitude excursions are believed to be useful chronostratigraphic markers especially in loess and other relatively thick, fine-grained deposits with an age beyond the limits of the radiocarbon dating method (Barendregt 1984; Løvlie 1989a, 1989b). Because of the possible Early Pleistocene age of the 65 m terrace below the investigated section, the last magnetic reversal defined by the Matuyama / Brunhes boundary (0.78 Ma) was not anticipated. On the other hand, the great thickness of the loess with multiple palaeosols was believed to provide some potential for recording high resolution palaeomagnetic anomalies. Such excursions are high amplitude directional variations of the palaeomagnetic signal on the order of 5-10,000 years recorded in a sediment in the process of deposition in a previous magnetic field. Excursions, chronologically fixed by the ^{14}C, TL, $^{238}U/^{235}Th$ and $^{40}K/^{40}Ar$ dating methods, have been successfully used as chronostratigraphic markers in geology, as well as geo-archaeology since the 1970s (Tarling 1983).

Palaeomagnetic remanence studies were first carried out during initial field investigations in the Kurtak area in the late 1980s, specifically in the excavated sections at the southern margin of the Berezhekovo Site (Drozdov *et al.* 1990a; Sidoras 1990). At that stage, a technique developed by the Russians was applied using cubic samples 4x4x4 cm impregnated by silicate glues or consolidated by paraffin, and taken at 0.2-0.5 m vertical intervals. According to Sidoras (1990:56), seven to eight palaeomagnetic episodes / excursions, and one polarity reversal correlated with the Matuyama / Brunhes reversal were found (with no primary data presented). These are believed to have been identified as natural anomalous palaeomagnetic inclination deviations (10-30°), all within a 2.5 m thick loess section! At Excavation 1, all palaeomagnetic anomalies are presumably recognized within the second part of the Middle Pleistocene and the Upper Pleistocene. In Berezhekovo Section 21, one excursion correlated with the Laschamp Excursion (34±4 ka.) is thought to have been located in the upper part of the Kurtak Pedocomplex dated by radiocarbon to 29,400 B.P. Other assumed excursions from Sections 21, 22 and Excavation 2 (horizon 5) were tentatively correlated with the Lake Mungo Excursion (26/31 ka.) and the Biwa I Excursion (180 ka.) (Drozdov *et al.* 1990a:63-64).

The reported results, however, cannot be viewed as objective because the large sampling intervals are unlikely to record any short-term high-resolution palaeomagnetic deviations. Some of the assumed excursions may in fact represent post-depositional structural disturbances of loess and truncation intervals. The claimed magnetic excursion around 30 ka may in fact reflect an anomaly resulting from an original distortion of the sample taken from a heavily cryoturbated soil. Also, the reported palaeomagnetic reversal between Stratum 5 and Stratum 6 in the Berezhekovo Excavation 2 (Drozdov *et al.*1990c:40) requires further verification. It should also be mentioned that a total of 12 palaeomagnetic episodes and excursions allegedly have been found in western Siberia within the Brunhes Epoch (Zubakov 1983) and 10 from the southern Ural Mountain area (Suleimanova *et al.* 1990). Authenticity of all the reported magnetic anomalies, however, does not inspire too much confidence because of their unusually high numbers within the local Quaternary deposits.

7.1.2. Methods

For the palaeomagnetic remanence investigations, a freshly cleaned vertical wall was prepared and a total 340 samples was taken by a 10 cm interval throughout the 34 m section (Kurtak 29) in cubic 2x2x2 cm (8 cm^3) plastic boxes (Appendix B). Laboratory analyses (Palaeomagnetic Laboratory, Department of Physics, University of Alberta) included measurement of natural remanent magnetization (NRM) and its routine demagnetization in an alternating field (AF) to remove unwanted "overprints." Individual inclination and declination values for each stratigraphic interval were plotted on diagrams following the stratigraphic sequence.

7.1.3. Results

The NRM and AF-cleaned magnetic data indicate normal polarity throughout the entire section (NRM: D=0°, I=58°, k=85, a_{95}=0.8°; 10mT: D=352°, I=62°, k=82, a_{95}=0.9°) (standard calculations according to Fisher 1953). However, in the upper part of the section, two ~50 cm-thick intervals containing systematic directional deviations are found, principally in the declination stratigraphic record. These are located at 6.8-7.2 m and 9.2-9.6 m below the present surface, centred at 7 m and 9.4 m, respectively (Figure 22-A). The peak of the upper anomaly lies at the contact of the Ckgj horizon of a Gleyed Brunisol (Sk_7) and the Bmkgj horizon of the underlying Gleyed Brunisol (Sk_8). The lower recorded anomaly is located within a loessic Ckg horizon below the involuted Orthic Turbic Cryosol (Sk_{11}). Both magnetic features lie, in geological contexts, within consistently aggradational sequences of silty deposits.

Declination and inclination data in the particular intervals display in both cases a coherent sequential pattern of magnetic vectors. In the upper interval a clockwise loop is traced out, in the lower one a more linear anti-clockwise feature is recorded (Figure 23). These high-resolution palaeomagnetic remanence patterns evidently reflect real features of the Earth's magnetic field in the study area. This conclusion is supported by the fact that the present inclination in the Northern Minusinsk Basin is 73° N, whereas the recorded inclination in the collected samples is 62° N. These facts indicate that two palaeomagnetic anomalies are well-established.

In view of the ^{14}C chronostratigraphy in the Kurtak area, and the radiocarbon date of 30,400±700 B.P. (AECV-1938c) on spruce wood from the A horizon of the Cryosol (Sk_{11}) located between the two magnetic deviations, the upper magnetic anomaly could correspond to the (later) Lake Mungo Excursion (26-31 ka), also reported from Horizon 5 of Excavation 2 at Berezhekovo (Drozdov *et al.* 1990a), or the most recent Laschamp-Olby Excursion (26.4 ka). The lower anomaly may correspond to an earlier Laschamp-Olby Excursion (35.4 ka) (dates according to Løvlie, table 1).

Subsequent palaeomagnetic measurements of the previously studied Section 22 confirmed the presence of reversed polarity in the basal part of a series of loessic deposits (Figure 5). In view of the normal magnetic signal below the magnetic reversal (Matuyama/Brunhes = 0.78 Ma), the recorded interval in the middle of the sampled profile (around the 250 cm depth level) may well represent the Jaramillo Event (0.90-0.98 Ma) (Figure 21). Because the upper part of the section shows a discontinuous record with major hiatuses, it was not subjected to closer palaeomagnetic scrutiny. The palaeomagnetic samples were taken at large (30 cm) intervals, because a stratigraphic discontinuity was apparent in the section, with the Early Pleistocene record truncated and covered by Late Pleistocene (late last glacial) loess sediments. The results support the assumed Early Quaternary age of the basal loess, and suggest an earlier (Early Pleistocene / Late Pliocene) age for the underlying alluvial terrace. This chronology would also corroborate the presence of a Late Tertiary / Early Quaternary fossil microfauna (molluscs) found in the alluvium directly overlying the Carboniferous bedrock (V.P. Chekha, personal communication 1994).

7.2. MAGNETIC SUSCEPTIBILITY

7.2.1. Principles and Application

Palaeomagnetic susceptibility studies were carried out at the Kurtak section because magnetic susceptibility of loess / palaeosols may be indicative of climatic change (Heller and Liu 1982, 1984; Kukla 1987; Kukla *et al.* 1988; Kukla and An 1989; Wang *et al.* 1990; Han *et al.* 1991; An *et al.* 1991; Ding *et al.* 1991; Liu *et al.* 1991; Liu *et al.* 1993; Rutter *et al.* 1991a-c). Since the first comprehensive palaeomagnetic study on loess (Heller and Liu 1982), it has been assumed that the inherited mineralogical composition of parent material and pedogenic maturation of soil horizons have a direct effect on magnetic-susceptibility enhancement in palaeosols.

After recognition of the direct relationship between magnetic susceptibility variations in loess and climatic change (Heller and Liu 1982), two main theories emerged about the mode by which the loess - palaeosol record acquires the magnetic signal. The first hypothesis assumes a constant input of magnetic minerals as a result of influences of volcanic activity and due to extraterrestrial magnetic materials in the form of meteorites. It has been suggested that the decreased susceptibility in loess is due primarily to dilution processes (Kukla *et al.* 1988). The alternative hypothesis proposes a pedogenic mechanism of the increased magnetism of soils as a result of an *in situ* formation of magnetic minerals in the process of parent material weathering (Zhou *et al.* 1990; Maher and Thompson 1991; Evans and Heller 1994). Recently, *in situ* pedogenic origin of bacterial magnetite in the form of magnetofossils has been suggested to be the principal agent of magnetic susceptibility enhancement in Chinese palaeosols (Evans and Heller 1994). Carbonate and iron-bearing silicate leaching may also be involved (Han *et al.* 1991). Other climatically indicative variables include the total calcium carbonate content, Si/Al, F/Cl and Fe(free)/Fe(total) ratios, and total amount of free iron oxides (Liu *et al.* 1991:65).

7.2.2. Methods

In the first stage of investigations (summer of 1993), the same palaeomagnetic samples used for the palaeomagnetic remanence analysis (taken at 10 cm intervals and providing a total of 340 samples) were used for magnetic susceptibility (Appendix B). After the analyses aimed at detection of palaeomagnetic secular variations, these were subsequently measured in the Palaeomagnetic Laboratory at the Department of Physics, University of Alberta, for the purpose of the magnetic susceptibility studies. In the second stage (summer of 1994), high-resolution palaeomagnetic susceptibility data were collected by 3.5 cm intervals, focusing explicitly on the last interglacial, 6 m long stratigraphic section (Figure 31), the analysis of which is in progress.

7.2.3. Results

The first susceptibility analysis showed a marked fluctuation of magnetic values throughout the stratigraphic record with a sinuous curve, suggesting a definite pattern (Figure 24). From a palaeoclimatic point of view, the susceptibility maxima correlate exactly with the intervals of the most intensive loess deposition, whereas minima correlate with the occurrence of the best developed pedogenic horizons. The relationship between the climatic change and magnetic susceptibility fluctuation is evident (Figures 24-26).

In the basal part of the Kurtak Section 29, the first susceptibility minimum (-32 m) corresponds to the presence of a Bmk horizon of a Brunisol or a truncated (?) Chernozemic soil (Sk_{32}). The following lower values (-31/-29 m) reflect the partly colluviated nature of the loess with some indications of incipient pedogenesis. The second susceptibility minimum (-28/-27 m) correlates with the first stage of intensive loess weathering with two soil formation intervals, the last of which includes the lower Chernozemic soil (Sk_{30}). Following a temporary susceptibility increase because of climatic cooling documented by deep frost wedge casts in the Chernozem, a short interval of stable magnetic values is shown by the overlying Brunisolic soil (Sk_{29}). Following a Gleyed Regosol (Sk_{28}), suggesting a further decrease of temperatures, the sharp increase of susceptibility in the superimposed aeolian loess (-26 m) correlates with the first climatic minimum. The subsequent, rather irregular magnetic susceptibility pattern (-25/-22 m) is associated with accumulation of the massive colluviated loess deposit bounded by two initial gleyzation horizons (Sk_{27}, Sk_{26}). The assumed niveo-aeolian formation of the finely laminated 3 m thick unit over a long period characterized by marked seasonal temperature and humidity fluctuations agrees with the recorded high-deviation palaeomagnetic curve. The Gleyed Regosols monitored by stabilization of the medium susceptibility values are in the field expressed solely by intensive mottling without any other pedogenic evidence. A new prominent susceptibility peak in the overlying loess (-21,5 m) relates to the second climatic minimum.

The beginning of the second palaeoclimatic cycle, as defined by the palaeosol record, is evidenced by a dramatic drop in magnetic susceptibility (-21 m). This, however, starts already in the upper part of the aeolian loess with gradually increasing signs of pedogenesis near the top in the form of a relatively thick, but less pronounced, Bmk and BCca horizons of a Brunisolic soil (Sk_{25}). It must be noted that no truncation surfaces are evident in this particular interval, which could alternatively account for the marked drop of the magnetic susceptibility values. The following enhancement of magnetic susceptibility (-20 m) reflects a short-lived interval of loess deposition due to climatic cooling, followed by formation of a Dark Brown Chernozem (Sk_{24}) and the overlying, largely disturbed, Brunisol (Sk_{23}), both magnetostratigraphically recorded by a progressive decrease of susceptibility values within the interval of the second climatic optimum (-19.5/-19 m). A new episode of loess sedimentation associated with another susceptibility peak (-19 m) indicates a marked cooling, also evidenced by two generations of partly superimposed frost wedge casts separated by the loess stratum

(Figure 19A). The following magnetic reduction to the lowest recorded susceptibility values (-18.3 m) relates to two warm and arid interglacial intervals characterized by formation of two superimposed Chernozemic soils (Sk_{22-21}). The susceptibility minimum was succeeded by a gradual rise of palaeomagnetic susceptibility directly correlated by weakening of the pedogenic process well-recorded in a series of three Brunisols (Sk_{20-18}), topped by a Gleyed Regosol (Sk_{17}), indicating a progressive shift towards a colder and more humid climate (-18/-16 m). The subsequent period of reactivated loess sedimentation is documented by the ongoing rise of magnetic susceptibility to its recorded maximum, followed by a series of peaks with decreasing magnitude (-16/-14 m). The following trend is characterized by a continuous reduction of magnetic susceptibility accompanied by short progressive deviations towards the lower values, associated with formation of initial pedogenic horizons (Gleyed Regosols and Gleyed Brunisols; Sk_{16-15}; Sk_{14-13}) within aeolian loess deposits (-14/-11 m).

After the second minimum magnetic susceptibility level (-10 m) associated with formation of a (truncated) Chernozem, or a well-developed Brunisol (Sk_{12}), the original A horizon was eroded and subsequently covered by a thin loess layer clearly recorded by the increase of susceptibility signal. The following formation of the regolithic, cryoturbated Chernozem (Sk_{11}) correlates with the return to slightly lower magnetic susceptibility, with a minor positive deviation consistent with the interval of soil deformation due to cryoturbation related to a colder oscillation (-9 m). Return to more stable conditions with advanced pedogenesis in the form of two Brunisols (Sk_{10-9}) again corresponds to two deviations with reduced values (-8/-7.5 m). The following interval of gradual cooling is reflected by a consistent rise of susceptibility (-7.5/-4 m) with minor negative oscillations related to formation of weakly developed gleyed Brunisolic horizons (Sk_{8-6}, Sk_4) and one Gleyed Regosol (Sk_5). As in the previous cycle, increased loess deposition without any indication of pedogenic processes is evidenced by the highest susceptibility values with the peak defining the last climatic minimum (-2 m). This is eventually followed by a short, marked interval of susceptibility decline associated with the appearance of a Brunisolic soil (Sk_2). Finally, a temporary cooling evidenced by a small palaeomagnetic susceptibility peak related to a new episode of loess sedimentation only slowed down the general warming trend leading to establishment of present-day conditions and formation of the Holocene Chernozem (Sk_1) characterized by a lower magnetic signal, analogously to the buried Chernozemic soils.

Overall, the recorded magnetic susceptibility changes match the palaeoclimatic evolution as recorded by the palaeosol sequence. The interglacial and warm interstadial Chernozemic and mature Brunisolic soils are characterized by the lowest susceptibility values (90-150 x 10^{-5} S.I.). Increased susceptibility is found in weakly developed Brunisols (150-250 x 10^{-5} S.I.). Slightly higher, but still comparable values were measured in the colluviated loess facies (160-410 x 10^{-5} S.I.). This fact would corroborate the assumption of ongoing pedogenic processes following niveo-aeolian loess accumulation. High magnetic susceptibility was recorded in Gleyed Regosols (190-390 x 10^{-5} S.I.). Finally, the magnetic susceptibility peaks clearly correlate with periods of loess deposition (350-580 x 10^{-5} S.I.). Interestingly, even minor climatic oscillations are indicated by negative or positive deviations of the magnetic signal. For example, a progressive mottling of the loess corresponds to decreasing susceptibility values. Also, periglacial deformation by frost action and cryoturbation of the Ah and Bmk Chernozemic soil horizons are evidenced by sudden shifts towards higher values in the magnetic record following episodes of less than 10 cm of loess sedimentation.

Conclusively, the paleomagnetic susceptibility record in the Kurtak Section 29 monitors the past climate change in the area during the last two glacial - interglacial cycles, with increasing, high-resolution amplitude during the Late Pleistocene cycle.

7.2.4. Discussion

The susceptibility curve in the Kurtak Section 29 provide definite evidence of climate-dependent magnetic variations correlated with deposition of loess and formation of palaeosols. The pattern, however, does not corroborate the view derived from the Chinese loess, which suggests that the enhanced magnetic susceptibility relates to secondary precipitation of ferromagnetic minerals because of pedogenesis. Accordingly, the susceptibility peaks are there located in the most developed palaeosols, where the magnetic signal is about 3-4 times higher than in loess (Heller and Liu 1982) (Figure 27), whereas, susceptibility of the Siberian loess from the Minusinsk Basin is 5-25 times higher than some of the Chinese loess with average values of 20-70 x 10^{-5} S.I. (An *et al.* 1991:fig. 1; Liu *et al.* 1991:fig 12). A reverse magnetic susceptibility pattern has also been reported from Alaskan loess deposits with high susceptibility values measured in unweathered aeolian sediments and low values in pedogenic horizons (Begét and Hawkins 1989).

The cause of this difference is not yet fully understood. It the case of the Alaska loess, velocity and competence of wind carrying the small silt-sized particles has been suggested to be the primary factor. Variations in wind velocities are believed to be reflected by fluctuations in the dispersal of magnetic minerals. During a single loess transportation event, higher susceptibilities are assumed to be characteristic of proximal deposits, while loess distributed for longer distances is depleted in magnetic minerals (Begét and Hawkins 1989:151; 1990:41). This model would corroborate the magnetic susceptibility pattern in the Kurtak loess, implying deposition of larger silt grains (including larger ferromagnetic minerals) during cold periods characterized by increased aeolian activity, and thus contributing to the total magnetic susceptibility enhancement relative to buried soils. Local provenance of the loess deposits is evidenced by a limited grain abrasion in both the loess and palaeosol units, identical mineralogy and the lack of mineral weathering. These aspects indicate that there was no apparent difference in the distance of loess transport, which could account for differential susceptibility values. The comparable grain size in the unweathered loess and well-developed palaeosols, however, also suggests other than purely transport variables involved, likely related to a post-depositional parent material (loess) alteration during warming climatic intervals.

The element SEM and the ARL SEMQ electron microprobe mineralogical analyses showed a higher content of unweathered ferromagnetic minerals (magnetite, titanomagnetite and ilmenite) in loess than in well-developed soils. This fact suggests that differences in the quantitative distribution of detrital ferromagnetic minerals within the Kurtak section may account for the recorded pattern of susceptibility variation. A partial depletion of magnetic elements from the original ferromagnetic minerals due to weathering processes of the parent material by alteration into less susceptible and stable iron oxides (goethite) due to periodic oxidation and reduction processes, subsequently removed by leaching, would explain the decreased magnetic signal in palaeosols. Alternatively, it can be suggested that the total magnetic capacity of a soil horizon is not only a function of weathering intensity and time, but may *a priori* depend upon the quantity and quality of primary magnetic minerals within the unaltered parent material inherited from original geological sources. At present, these are only working hypotheses which require further verification and testing.

In any case, the Siberian and Alaskan data show that the model assuming magnetic peaks correlating with well-developed palaeosols due to landscape stability, increased temperatures and intensive weathering processes is not universally valid. Therefore neither the pedogenic origin due to transformation of loess under humid and warm interglacial conditions (Heller and Liu 1984), nor the dilution theory, assuming an airborne origin of the ultrafine magnetic dust (Kukla *et al.* 1988), can be generally applicable. Other variables, possibly related to stability of ferromagnetic minerals are apparently involved.

8. PLEISTOCENE ECOLOGY OF THE KURTAK AREA

In addition to the loess - palaeosol record, palynological and palaeontological evidence provide sources of proxy data for reconstruction of the palaeoenvironmental history in the Kurtak area. Nevertheless, because both the Middle and Late Pleistocene biotic data are still fragmentary, the following review provides only general information about the nature of past biological communities in the northern Minusinsk Basin. A more precise temporal correlation cannot be pursued at the present time because of the lack of a well-established regional chronostratigraphic framework.

8.1. FLORA

Most of the palaeobotanical evidence consists of pollen data, as only rarely has fossil wood has been found in the Kurtak loess sections. Palynological studies have been carried out chiefly at the two principal localities - Berezhekovo and Kamennyy Log, largely in conjunction with buried cultural occupation surfaces. Unfortunately, no additional palynological data have been provided by the author from section Kurtak 29, because of small samples, and because of low pollen concentrations in loess and most palaeosol horizons. At present, the longest and so far the most complete record comes from the Berezhekovo Section 1 (Drozdov *et al.* 1990a:55-56). Nevertheless, as with most other pollen evidence, chronology is only tentatively detennined.

The basal part of the Berezhekovo sequence formed of a thin Middle Pleistocene alluvium (Stratum 8) yielded pollen of Siberian "cedar" *(Pinus sibirica),* spruce, and fir, suggesting presence of southern taiga in the area. A more informative assemblage came from a Chernozemic / Luvisolic soil (Stratum 7) variously assigned to the Tobolsk Interglacial / the Samarov Interstadial (*op.cit.*:62) or the last (Kazantsev) Interglacial (Derevianko *et al.* 1992a:80). A mosaic steppe-parkland with pine and birch is documented by a pollen record (Complex I) with 35-98% of grasses including herbaceous plants - Asteraceae (43-58 %), Chenopodiaceae (0.6-1.9 %); 1-13 % of arboreal taxa - *Pinus silvestris* (34-58 %), *Pinus sibirica* (7-10 %), *Betula* sp. (27-31%), *Betula nana* (2-3 %), *Salix* (6.5 %), *Alnaster fruticosus* (3.4 %); and 1-52% of spores, dominated by *Botrychium* (84 %).

The overlying loess with a Gleysol horizon at the base (Strata 6-4) includes pollen (Complex II) indicative of a humid boreal mixed forest with birch, fir and spruce, suggesting a colder climate than at present, likely of the early last glacial in age.

Arboreal taxa (44-58%) prevail in the record with *Picea* sp. (24-51 %), *Pinus silvestris, Abies sibirica* and *Betula* cf. *albae* (5-17 % each), *Pinus sibirica* (< 8 %), followed by spores (33-40 %) with *Botrichium* and Bryales, and less grasses (3-29 %). A marked fluctuation of arboreal (mostly coniferous) and non-arboreal taxa among the particular strata is apparent. In Stratum 6 (the Berezhekovo Series), spores (Polypodiaceae, *Botrichium)* prevail (Derevianko *et al.* 1992a:81). Increase of conifers is particularly manifested in Stratum 5, with 51 % of pollen of *Picea,* whereas grasses are represented only by 3 %. In Stratum 4, on the other hand, grasses and spores dominate, and about 12 % accounts for *Abies* (Drozdov *et al.* 1990a:63).

Drier conditions than in the underlying horizons are indicated by pollen data from an incipient Brunisolic soil (Stratum 3) typical of a mixed forest and forest steppe with pine and birch (Complex III). Arboreal taxa include *Pinus silvestris* (33-42 %), *Betula* sp. (35 %); *Picea* sp. (11 %); grasses (27 %) with Cyperaceae, Asteraceae, *Chenopodium;* and some spores (9 %). Cold steppe with pine and birch characterizes the overlying initial soil (Stratum 2), with pollen (Complex IV) dominated by grasses (59 %) - Chenopodiaceae, Cichoriaceae, Asteraceae (altogether 85.7 %); with fewer trees (34 %) and spores (6 %) *(op. cit.).*

Finally, the pollen record (Complex V) from the lower part of the Holocene soil (Stratum 1) includes grasses (74 %), trees (16 %) with *Betula albae* (68-98 %) and *Pinus silvestris* (2-32 %), and spores (10 %) with *Bryales* and Polypodiaceae, typical of a dry continental climate. The upper part of the present Chernozem produced pollen of trees (66 %) with *Pinus silvestris* (58-72 %), *Betula albae* (22-38 %), *Picea obovata* (3-7 %); grasses (23 %) with Asteraceae (30-62 %), Cyperaceae (5-10 %) and Chenopodiaceae (5-12 %); and spores (11 %) with Polypodiaceae (52-79 %), *Lycopodium, Bryales* and *Sphagnum,* characterizing altogether the present parkland steppe in the surrounding area *(op. cit.).*

An analogous pollen record, covering the Late Pleistocene, has been derived from the Kamennyy Log Excavation 1 (Drozdov *et al.* 1990a: 84-87). A total of seven climatic phases have been specified in terms of a general palaeoclimatic evolution. Phase I, assigned to the last interglacial, is characterized by southern taiga-steppe vegetation, with coniferous forest taxa dominated by spruce, pine and cedar, and open parkland communities with pine and birch. Expansion of boreal forest is assumed for Phase II with a distribution increase of cedar *(Pinus silvestris).* A cooling trend during the Phase III caused a shift towards a forest-tundra setting, followed by warming during Phase IV leading to re-establishment of southern taiga. A new climatic deterioration with a more pronounced continentality and return to a cold tundra environment is associated with Phase V. The succeeding mixed boreal forest with pine and birch (Phase VI), and a cool humid steppe (Phase VII) relates to warming and cooling stages, respectively. Phase VII marks the recent climatic conditions within a southern parkland-steppe (Drozdov *et al.* 1990b).

At the same site, pollen of *Larix, Pinus, Picea, Abies* and *Betula* has been also found in (interstadial?) colluviated loess deposits (Derevianko *et al.* 1992a:58). Finally, at the Kurtak IV Site, a cold (arboreal) forest-steppe with birch, pine, larch and spruce, and some spores *(Lycopodium alpinum, Selaginella sibirica)* has been dated to the late Karginsk Interstadial (*ca.* 24-23 ka).

In sum, the palynological data document shifts in the vegetation patterns between southern taiga and forest steppe in the Minusinsk area, also inferred by a more regional pollen analysis (Kol'tsova 1990), and corroborating the palaeosol record from the Kurtak Section 29. During warmer Late Pleistocene interstadials, conditions similar to the present were established with steppe and forest-steppe replaced by forest-tundra and steppe-tundra during colder glacial intervals.

8.2. FAUNA

In comparison with the palaeobotanical data, the palaeontological record has been largely derived from secondary and/or poorly documented stratigraphic contexts. Most of the bones and teeth, either well-preserved or in fragments, have been eroded onto the present beach (Figure 33-A). Provenance of some of the faunal material can be determined from the stratigraphic position of exposed fossil-bearing deposits. Much less data come from controlled excavations of more limited extent.

So far the most complete, but only roughly chronologically determined faunal record, comes from the Berezhekovo Excavation 2 (Drozdov *et al.* 1990a:58). There, remains of *Ursus* sp., *Bison priscus* Boj., *Citellus* sp., *Myospalax* sp., and a variety of rodents (*Microtus gregalis* Pall., *Microtus oeconomus* Pall., *Lemmus obensis* Brants, *Arvicola terrestris* L. and *Sorex* sp.) were found in a Chernozemic 0.7-1.0 m thick soil disturbed by solifluction (Stratum 5). An early form of *Mimomys* sp. and *Prolagurus* sp. are believed to indicate an Eopleistocene (*i.e.*, Early Pleistocene) age in the Russian chronostratigraphic terminology, as supported by the reversed magnetic polarity of the incorporating deposits (Drozdov *et al* 1990c:40). However, this assignment does not fit the occurrence of a large Middle Pleistocene fauna (< 730 ka), including *Coelodonta antiquitatis* Blum., *Equus hemionus* Pall., and *Bison priscus* Boj., recorded in the underlying sandy gravelly 0.4-0.7 m thick alluvium (Stratum 6). Further research is needed to clarify this issue.

At the Berezhekovo Excavation 1, wooly rhinoceros *(Coelodonta antiquitatis)* was found in the lower part of the Tchaninskaya Loess Series (Stratum 3) of the early last glacial stage. A typical cold loessic molluscan fauna, indicative of a dry climate, has been recorded from the aeolian deposits (Stratum 5), represented by *Succinea oblonga* Drap., *Pupilla muscorum* L., *Pupilla turcmenica, Vertigo pseudostriata* (Lzk.), *Columella columella* (Mrt.), *Vallonia aff. enniensis* (Gredl.) and *Eoconulus fulvus* (Müll.). A rich fauna, including Ursus sp., *Equus* sp., *Coelodonta antiquitatis* Blum., *Capreolus capreolus* L., *Megaloceros giganteus* Blum., *Rangifer tarandus* L., *Bison priscus* Boh., *Ovis* sp., *Lepus* sp., *Alactata* sp., *Myospalax* sp., *Lagurus lagurus* Pall., *Arvicola* sp.; and some mollusc species (*Pupilla muscorum* L., *Pupilla turcmenica* Bttg., *Vertige pseudostriata* Lzk and *Columella columella* Mrt.) came from the 3 m thick Berezhekovskaya Loess Series (Stratum 6) with advanced gleyzation and some krotovinas in the lower part (Derevianko *et al.* 1992a:78). A Middle Pleistocene fauna with *Equus mosbachensis - germanicus, Coelodonta antiquitatis, Megalocerus gigantheus* Blum., *Bison priscus* Boj., *Saiga tatarica, Lagurus lagurus* Pall., and rodents *Mimomys oeconomus* Pall., *Myospalax* sp., *Mustela* sp. were found in the lower, partly redeposited 0.8 m thick interglacial soil (Stratum 7). Some Upper Pleistocene rodent taxa as *Microtus gregalis* and *Arvicola terrestris* L. would not contradict the earlier chronology, as the later species most likely have burrowed into and became incorporated within the deeper deposits. *Equus* sp., *Equus* cf. *hemionus, Rangifer tarandus* L., *Sorex sp.*, *Citellus* sp., *Lemmus obensis* Brants, *Microtus* sp. from the underlying 0.2-0.5 m alluvium (Stratum 8) characterize an earlier, probably periglacial faunal community. Finally, *Mimomys* sp., *Prolagurus* sp., *Allophajomys* sp. represent the earliest fossil rodent remains from the 0.4-1.1 m relic of presumably Early Pleistocene (Pliocene in Russian terminology) alluvium in the basal part of the section (Stratum 9) (Derevianko *et al.* 1992a:76-78).

At the Kamennyy Log Site I, a mixed late Middle Pleistocene and Upper Pleistocene fauna collected along the present lake shore, but eroded from primary geo-contexts, includes an early form of *Mammuthus primigenius* Blum., *Equus mosbachensis - germanicus, Equus* cf *przewalski* Pol., *Equus* off. *hydruntinus* Reg., *Coelodonta antiquitatis, Alces alces* L., *Rangifer tarandus, Bison priscus* Boj.

(Drozdov *et al.* 1990a:93). Some rodents (*Myospalax myospalax* Laxm., *Citellus undulatus* Pall.) came from a more intact geological position from the Sukholozhskiy Pedocomplex assigned to the early last glacial stage (Derevianko *et al.* 1992a:57). An early form of *Mammuthus primigenius* and *Equus mosbachensis - germanicus* was also recorded in the 3-4 m thick periglacial alluvium below the first last interglacial Chernozem (*i.e.*, the Kamenolozhskaya Soil) (*op.cit.*:59).

Fossil remains have been directly dated only at the Kurtak Site IV from a 24,000-23,000 B.P. cultural horizon. *Ursus* sp., *Panthera spelaea* Doldf. and *Alces alces* represent a typical interstadial (Karginsk / Mid-Wisconsinan) periglacial assemblage. Fauna (*Equus caballus, Megaloceros giganteus, Bison priscus* and *Coelodonta antiquitatis)* from the Chernozemic soil below (>35,000 B.P.) is most likely of last interglacial in age (Drozdov *et al* 1990a:106-107). A (late) Middle Pleistocene fauna has been collected from a 12 m platform below a 160 m terrace from the Verkhniy Kamen Site, mainly represented by early forms of mammoth (*Mammuthus chosaricus* Dubr.) and horse *(Equus* sp.), presumably from the Samarov (Illinoian) Glacial (Drozdov *et al* 1990a:111-112).

In the summer of 1993, a unique faunal assemblage was recovered in the shore zone at the Ust'-Izhul' Site 10 km north of Kurtak on a 65 m Early Pleistocene terrace of the Yenisey River (Drozdov *et al.* 1995). The remains were incorporated in the B horizon of a secondarily gleyed reddish soil, the upper part of which was colluviated and interstratified by yellow-grey-greenish silty deposits, above a compact Early (?) / Middle Pleistocene Chernozemic Ah horizon The fossil fauna present at the site included an early form of *Mammuthus primigenius* Blum., *Coelodonta antiquitatis, Bison priscus, Equus* sp., *Cervus elaphus;* and *Marmota* cf. *barbacina* and *Meles meles* (N.I. Ovodov, personal communication, 1994). The most abundant species -- mammoth -- is represented by mostly juvenile remains of at least 12 individuals, parts of which were recorded in anatomical order. A rich malacofauna and rodent fossil remains (so far undetermined taxonomically) are also present. The assumed (late) Middle Pleistocene age of the site is based on the taxonomical classification of the mammoth remains, the stratigraphical position of the find sealed below the last interglacial pedocomplex (Figure 35), and by a radiocarbon date of >42,100 yr B.P. (AECV 1939c) obtained from a mammoth phalanga.

In sum, in despite of the less cohesive picture of palaeoclimates and problematic dating of the earlier finds, the fossil record clearly indicates a wide range of large as well as small animals occupying the Northern Minusinsk Basin at various stages during the Middle and Late Pleistocene. Presence of cold-adapted species suggests a certain ecological potential of periglacial environments for diverse biotic communities that expanded during stadial intervals. The high number of rodent species from warmer interstadial and interglacial soils document their specific natural habitat within dry and cold steppe and forest-steppe settings.

8.3. EARLY HUMAN OCCUPATION

Systematic geological and archaeological investigations since the late 1980s in the northern Minusinsk Basin have produced unexpected evidence of early human occupation of this part of Siberia dating at least to the late Middle Pleistocene. Research was initiated in the upper part of the Yenisey River valley after progressive erosion of unconsolidated aeolian and slope deposits by water-table fluctuation of the Krasnoyarsk reservoir exposed the 10-40 m high sections. The erosion revealed a nearly complete Late Quaternary geological record containing a series of Early, Middle and Late Palaeolithic stone tool collections and a rich Middle and Upper Pleistocene fauna (Drozdov *et al.* 1990; Derevianko *et al.* 1992a). Until the ponding of the Yenisey River, the oldest archaeological remains known in the region dated to about 30,000 years B.P. (Praslov 1984). The main focus of the current geo-archaeological investigations is to reconstruct the Quaternary history of the Palaeolithic settlement of the area and establish the necessary geochronological framework. Correlation of the principal Palaeolithic and Quaternary periods is provided below (Table 7).

The Middle and Upper Palaeolithic cultural horizons have been recorded within the Late Pleistocene loessic deposits. Several sites have been recognized in the Kurtak area, with older Early Palaeolithic stone industries largely redeposited by past as well as present hydrologic processes. After the first finds of an archaic lithic industry (stone tool assemblage) at the Berezhekovo Site in 1987, four other major locations with Early Palaeolithic lithic artifact occurrences have been recognized (*i.e.*, the Kamennyy Log, Sukhoy Log, Verkhniy Kamen and Razlog sites). At all places, artifactually flaked cobbles have been found on the surface of the eroded 60-65 m high terrace remnant exposed by waves undercutting the slope. Most artifacts have been washed from their original geological context onto the present beach, occasionally together with fossil fauna and later Palaeolithic stone tools derived from loess deposits above the terrace. Ongoing wave action has caused sorting of the flaked lithics as well as other clastic materials. This phenomenon is particularly apparent at the Kamennyy Log Site, extending 2-4 km south of Berezhekovo, with large flaked cobbles dispersed in the southern part of the site, in places of the most intense wave energy. Most of the small-sized artifacts and lithic fragments are concentrated in the northern part, where more quiet waters seasonally inundate the adjacent beach. At the Sukhoy Log and Verkhniy Kamen Sites, most of the coarse pebbly alluvium has been washed away, and the macrolithic industry with large fossil bones (mainly of early forms of mammoth and horse) is found directly on the Carboniferous sandstone bedrock that forms prominent ramparts elevated 1-7 m above the lake. At the latter site, the presence of *Mammuthus chosaricus* Dubr. implies a glacial (Samarov / Illinoian) age for the Early Palaeolithic cultural assemblage, provided that both records are contemporaneous (Drozdov *et al.* 1990a:111-112).

A more complex situation is found at the Razlog Site situated 3 km north of Berezhekovo. Up to 25 m thick alluvial fan deposits filled a depression between two bedrock ramparts 90-120 m high (Figure 34-A). Abundant and variable lithic artifacts (Figure 33-B) are found with large mammal bones among cobble gravels on the shore at the foot of an eroding 20 m high alluvial fan capped by about 10 m of Late Pleistocene aeolian and colluvial loess sediments. Except for artifacts from the eroded 60 m terrace, most of the modified lithics have been introduced from the slope colluvium, including weathered gravels secondarily derived from the highest (130-150 m) terraces above the site. The Tobolsk (Yarmouth) Interglacial has been the estimated minimum age for the lithic assemblages (>250 ka) (Drozdov *et al.* 1990a:93; Drozdov 1992:28).

A large scale excavation carried out at the Berezhekovo Site has exposed about 60 m^2 at Excavation 1 and about 80 m^2 at Excavation 2 (Drozdov *et al.* 1990a; Derevianko *et al.* 1992a). At the first location, Upper Pleistocene (Middle Palaeolithic - "Mousterian-like") stone artifacts are found in the colluviated Chernozemic soil (Stratum 7) of the last interglacial. Earlier flaked lithics are secondarily distributed in a 70-80 m Middle Pleistocene alluvium (Stratum 8a), also incorporating fossil remains of *Mammuthus primigenius* Blum. and *Equus mosbachensis,* assigned to the end of the Tobolsk Interglacial. The 0.7-1.0 m colluvium is presumably formed by redeposited and poorly sorted sandy gravels of a 90-100 m terrace above the section (Drozdov *et al.* 1990a:66). At the second location, a small collection of rudimentarily modified and secondarily rolled artifacts was found in the 0.5-0.7 m thick Early Pleistocene / early (?) Middle Pleistocene alluvium/colluvium (Stratum 6). The suggested age of the lithic industry is chronostratigraphically inferred in respect to the established geo-archaeological framework of the Berezhekovo Site. Early and Middle Palaeolithic artifacts are dispersed in large numbers on the present lake shore (Figure 37).

At all sites, cobble-sized clastic raw materials (mostly quartzite, less vein quartz and basalt) collected from the fluvial/alluvial gravels, were used for stone tool production. At least two Early Paleolithic industries can be recognized on the basis of degree of corrosion and differential patination (Chlachula *et al.* 1994). The first (older) series, comprising part of the collections from the alluvial fan at the Razlog Site and possibly some artifacts from the neighboring Verkhniy Kamen Site, is characterized by weathered, minerally stained and heavily rolled quartzite tools with a uniform dark brown or cinnamon patina covering both the unmodified as well as flaked faces of particular specimens (Figures 38-39). The second (younger) series, including most of the stone artifacts from other sites, is distinguished by a lesser degree of abrasion and white unpatinated flake scars on cobbles with yellowish cortex (Figure 33-B). Only occasionally, a reddish patina occurs on the flaked cobble faces. Because of the same (secondary) geological provenance of both industries, an original distribution of the older series near Razlog on the high 100-120 m terrace is assumed. These early artifacts are assumed to have been subsequently redeposited in the process of erosion of the Early Pleistocene / Pliocene-age terrace and colluviated with other weathered and patinated clastics within the formed alluvial fan at the foot of the terrace. Formation of the fan was a gradual process, with a periodic appearance of a dry sandy-gravelly surface in a shifting channel environment, presumably occupied by early people leaving behind the more recent, unpatinated and less abraded stone artifacts.

There are no apparent technological or typological differences between the two industries. All artifacts, forming a very rudimentarily modified "pebble industry," are produced by hard hammer, direct percussion technique. Unifacial choppers and simple cobble cores, both characterized by only a few flake removals, are the most frequent forms. A specific artifact type with a protruding laterally flaked distal edge is often encountered in the archaic (older-series) assemblage at Razlog. More elaborate flaking, present on some bifacial choppers, is rare. Most flakes exhibit limited unifacial or alternate retouch, although a few better-produced side scrapers occur (Figure 37).

Apart of the exceptional Ust'-Izhul' Site, being the only Early Palaeolithic site recorded in the primary context, and allowing a more exact age-assessment (determined by stratigraphy, radiocarbon dating, palaeomagnetism and palaeontological taxonomy), chronology of most of the Early Palaeolithic record in the Kurtak Archaeological Region is based on the relative chronostratigraphy of the industry-incorporating deposits within a larger geomorphological setting. A (late) Middle Pleistocene age for the less-weathered Early Palaeolithic stone artifact collections (Series 2) is supported by their stratigraphic position beneath a complex of palaeosols, including one or two well-developed interglacial Chernozemic soils. As stated above, greater antiquity is assumed for the more corroded lithic industry (Series 1) from the Razlog Site. Nevertheless, so far the oldest artifacts with a better chronostratigraphic control come from the excavated Section 1 at the Berezhekovo Site in the upper part of a 0.3-0.5 m thick bed (Stratum 8) (Drozdov *et al.* 1990, fig. 34). The associated remains of large fauna (early forms of *Equus* sp.) and rodent taxa suggest an early (?) Middle Pleistocene or even Early Pleistocene age (>0.78-0.5 Ma). A very early age (>1 Ma) may be implied for occasional artifacts eroding from the 65 m terrace (Figure 36). This is covered by 30 m thick loess deposits, in the basal part of which two palaeomagnetic reversals has been recorded, with the latter (characterized by a normal polarity) correlated tentatively with the Jaramillo Event (0.98-0.9 Ma) (Figure 21). Confirming data and location of artifacts *in situ,* however, are required as this rudimentarily flaked lithic assemblage would be the earliest cultural find reported from southern Siberia and one of the oldest in Eurasia. Overall, the cultural evidence from the Minusinsk Basin indicates that parts of southern Siberia in the main river valleys, including the upper Angara River Basin (Medvedev *et al.* 1990), were occupied at several stages during the Middle Pleistocene.

The only Palaeolithic site in the study area found *in situ (sensu stricto)* is the Ust'-Izhul' Site discovered in august 1993 after exposure by wave erosion of the river bank, and investigated in summer of 1994. A unique and perfectly "fresh" stone tool assemblage was recorded during the subsequent excavations, together with an abundant fossil fauna record (see above), as well as some other cultural features (as fireplaces). Lithic artifacts (mostly represented by unmodified flakes), made from various clastic rocks selected from river gravels, were flaked and used directly at the site (Figure 41). Modified (flaked, cut and scraped) bones of rhinoceros, bison; bones and tusks of mammoth; and a flaked caribou antler were found. The spatial distribution of the skeletal remains suggests that the large animals were hunted by Palaeolithic people in the nearby area and transported in pieces (dissected parts) to the site.

This interpretation is supported by finds of three mammoth juvenile skulls, two left mammoth shoulderblades, a left femur still attached to tibia and a bison skull. The fossil remains, recorded largely in anatomical order, were concentrated in an area of *ca.* 6x8 m. Close to 200 artifacts mostly flaked from several large quartzite, quartz, basalt and silicate-rock cobbles, collected by the early people from the Yenisey River terrace, were dispersed among the bones within an interstratified silty-clayey matrix. The exceptional nature of the discovery is also documented by three fireplaces, one of which (F_3) was found between a large mammoth tusk and a broken caribou antler, and evidenced by a dispersed charcoal over an area of about 4 m^2, with a radiocarbon date >42,190 yr B.P. (AECV 2034C). Two burnt stone fragments were removed at the second fireplace (F_2), 1-1.5 m in diameter (>40,050 yr B.P.; AECV 2033C), situated about 5 m from the first one. The third fireplace (F_1) (>41,810 yr B.P.; AECV 2032C) was located about 70 m north of the main occupation site, with a small set of artifacts, including a well-produced unifacial chopper, found nearby. At all fireplaces, coniferous wood *(Abies* sp.) was used as fuel as identified from charcoal samples (analyzed by B. LePage, Department of Botany, University of Alberta, 1995). Associated malacofauna (*Pupilla muscorum* Lzk.; *Pupilla loessica* Lzk.; *Succinea oblonga elongata* Sand.; *Vallonia tenuilabris* Braun), all being typical loessic mollusc species characteristic of xerotheric continental grassland environment. The (late) Middle Pleistocene age of the site is evidenced by the stratigraphic geological position below the last interglacial pedocomplex (Figure 41). Regardless of its antiquity, the site is without parallel in all Siberia in its taphonomical and contextual completeness (Laukhin *et al.* 1995).

The Middle Palaeolithic record characterized by a "Mousterian-like" Tradition, distinguished by the Levallois (*i.e.*, prepared core) technology, has been found at the Kamennyy Log II (Sections 9-11, Excavations 2-4) and Berezhekovo Sites (Excavations 1-2) in the last interglacial Kamenolozhskaya Soil and the Sukholozhskiy Pedocomplex of the early last glacial interstadials. Typical Levallois specimens also have been found in the shore zone at other sites in the Kurtak area, but without clear geological contexts (Figure 40). At present, there is no consensus that the Middle Palaeolithic lithic industry can be really associated with the classical European and Near Eastern Mousterian, although Mousterian influences in the Altay cave sites are evident. Also, the Middle Palaeolithic technology of stone flaking, especially the Levallois technique, survives for some time in the later southern Siberian Palaeolithic assemblages.

Late Palaeolithic stone industries have been recorded in the Sartan (Late Wisconsinan) loess and intercalated, weakly developed interstadial soils. At the Kashtanka Site, two cultural horizons dated to 21,800-24,800 B.P. are located in the upper part of Stratum 11 near the top of the Kurtak Pedocomplex within a 3-3.5 m loess deposited on a 70-80 m terrace platform (Drozdov *et al.* 1990a:117-119). At the Kamennyy Log Site I, analogous tools have been recovered from two cultural layers (Stratum 11) in humic loamy sediments dated between 23,500 and 24,900 B.P. A Late Palaeolithic site has been discovered at Ust'-Izhul', close to the Middle Pleistocene "mammoth site". Finally, the most recent Palaeolithic find comes from the Listvenka Site with radiocarbon dates of 13,590±350 B.P. and 12,750±140 B.P. (Table 2).

Changing Pleistocene climates in the study area during and after episodes of early human occupation is also evidenced by a varying degree of patina and/or aeolian abrasion. A dark red-brown patina found on the archaic series of artifacts from Razlog and Verkhniy Kamen Sites indicates a warm and relatively humid climate, despite the fact that no other palaeopedological evidence in the form of red soil was recorded in the sections. Intensified aeolian activity during early cold climatic intervals had exposed and abraded abandoned archaeological remains prior to their (re)burial under loess cover. This phenomenon can be particularly well recognized in the Middle Palaeolithic assemblages, including an intensively polished Levallois core (Figure 40:2), but also present are fresh specimens identifiable as Levalloisian, indicating their rapid burial prior to the main aeolian episode (Figure 40:1).

Overall, the Palaeolithic record from the Kurtak sites provides evidence that the Minusinsk Basin was repeatedly occupied by early people during the Pleistocene, including colder stages with periglacial conditions. The abundant cultural remains can be used as another, although indirect and still poorly dated, nevertheless important source of proxy data for reconstructing the Quaternary environments and palaeoclimatic history of the region. Although still fragmentary, the biotic (floral and faunal) data combined with the Palaeolithic record provide additional evidence of climatic evolution in the upper Yenisey River valley during the Pleistocene.

Time Period	*Quaternary Stratigraphy*	*Palaeolithic Chronology*
15-10 ka	Final Pleistocene	Final Palaeolithic
35-15 ka	~late Late Pleistocene	Upper Palaeolithic
130-35 ka	~early and middle Late Pleistocene	Middle Palaeolithic
>130 ka	Middle Pleistocene	Early Palaeolithic

Table 7. Schematic correlation of the Quaternary stratigraphy and the principal Palaeolithic chronology (according to the author).

8.4. SUMMARY

The loess - palaeosol stratigraphic sequence from the Kurtak area provides evidence of complex and highly fluctuating climatic changes in the northern part of the Minusinsk Basin during the Pleistocene, and southern Siberia in general. Although temporally limited to more recent stages of the Quaternary period, the key section (Kurtak 29) comprises a unique, continuous high-resolution record for the last glacial - interglacial cycle, and most, if not all of the previous late Middle Pleistocene cycle.

Mineralogical composition of the loess, predominantly quartz, plus lesser amounts by feldspar, chlorite, calcite, and other less stable minerals as amphibole, mica and albite, indicate a predominantly local provenance of the silty deposit. Limited, short-distance transport is evidenced by the fresh surface morphology of the fine silt fraction, and by overall low to moderate abrasion of the grains. The larger, angular to subangular, generally less resistant minerals most likely originate from igneous and metamorphic rocks eroded by glaciers in the nearby Eastern Sayan Mountains, and transported by ablation discharge into the river valley. Fluvial abrasion on a significant proportion of quartz grains may imply a more distant origin, for example, from glacial environments in the upper reaches of the Yenisey River, but also partially due to redeposition from the local Carboniferous sandstone bedrock. The fine clastic materials, apparently introduced from different sources and periodically accumulated on the alluvial plain of the Yenisey River during cold and dry climatic intervals, were redeposited by wind onto the eastern slopes of the Kuznetskiy Alatau Mountains, thus indicating a prevailing wind direction from the (south)east. Some loess deposits were subsequently colluviated; others may have been obliterated by erosional processes, although no major hiatuses are evident in the stratigraphic record.

Loess colluviation took place on steeper slope sections, presumably largely unprotected by vegetation cover, under more humid cooler conditions. Distortion of the deposits, however, was very limited. The secondarily modified (colluviated) loess facies indicates a seasonal, periodic sedimentation of the silty dust onto a thin, subsequently melted layer of snow, or onto barren land, resulting in the characteristic finely and irregularly laminated structure with intercalated darker and lighter bands subsequently distorted by minor frost wedge casts and periglacial involutions. Such a niveoaeolian loess formation process, implying a relatively high amount of winter precipitation and moderately cold climate, would account for a significant amount of the subaerial deposit in the earlier stages of loess sedimentation. The accelerated rate of loess sedimentation prevented establishment of a more permanent vegetation cover, so only an incipient grassy vegetation is assumed. The varying thickness of individual laminae indicates fluctuating intensity of loess deposition.

Initial pedogenesis in the form of gleying occurred during intervals of decreased loess sedimentation, but still under cold climatic conditions and within an arctic tundra-steppe setting with periodically saturated ground. Gradual increase of annual temperatures and better surface drainage led to formation of Gleyed Regosols, Gleyed Brunisols and eventually well-developed Brunisols; corresponding to establishment of periglacial steppe-tundra, forest-tundra and boreal forest, respectively. Subsequent increase of summer temperatures caused a shift towards a semiarid, strongly continental climate that led to expansion of grasslands and retreat of forest into higher, more humid locations. The rate of soil formation processes accelerated somewhat, but because of the shortened annual vegetation cycle, due to prolonged and very cold winters, the characteristic Chernozemic soils have a less-developed structure than mature Brunisols formed under more humid and less continental conditions in a parkland / forest setting. The stratigraphic succession of particular buried soil horizons supports the assumption of brown forest soils development during the first part of warmer interglacials under a more humid and moderate climate, and their succession by Dark Brown Chernozemic soils indicate a drier and more continental environment (Dergacheva *et al.* 1984:25). Deep ice wedge casts in the Chernozemic soils attest to cold episodes during the last two interglacial cycles, followed by less-developed Brunisols than those from earlier interglacial intervals. Periglacial activity during interstadials were limited to small ice-wedge development and cryoturbation, both indicative of a less continental, humid climate, with more stable annual temperatures lacking major seasonal and short-term fluctuations.

Palaeomagnetic susceptibility has proven to be a very sensitive, high-resolution indicator of climate change in the area. A definite correlation exists between intervals of climate deterioration with loess deposition and increased magnetic susceptibility; and between intervals of climatic amelioration with progressive soil formation and a gradual lowering of susceptibility values. The cause of this palaeomagnetic patterning, reversed to the Chinese as well as Central Asian and European loess, is not fully understood (Figure 27). A differential amount of ferromagnetic minerals, originally derived from rock formations in loess during colder intervals, and/or their gradual depletion and modification into less magnetic iron oxides by weathering processes during warm intervals with surface stabilization and advanced pedogenesis, is assumed to account for the magnetic susceptibility fluctuation. This assumption, however, requires further verification, and the palaeomagnetic record requires closer geophysical scrutiny.

In terms of a general palaeoclimatic evolution, both the loess - palaeosol record and the palaeomagnetic susceptibility curve indicate a changing climatic pathway with an increasing climatic amplitude (Figure 26). In comparison with the deep-sea oxygen isotope record, some similarities are apparent (Figure 28). After the first interglacial / warm interstadial of the late Middle Pleistocene in the lower part of the stratigraphic section, a significant drop in temperatures ended a longer period of relative environmental stability tentatively correlated with the Oxygen Isotope Stage 7 (with possible substages 7e-7a). The following cold period (Stage 6) experienced significant short-term climatic fluctuations, contributing to intensive loess colluviation between two cold and arid intervals with aeolian loess sedimentation. Dramatic climatic amelioration after the termination of the first climatic cycle led to a renewed and long-term landscape stabilization accompanied by intensified soil formation processes, but disrupted by short intervals of loess deposition and periglacial surface deformation. Following formation of the first Brunisolic soil, the trend towards more pronounced continentality is manifested by formation of Chernozemic soils subsequently dissected by deep frost wedges, indicative of an increasingly continental climate with warm summers and very cold winters. The three well-developed magnetic susceptibility peaks corresponding to a series of palaeosols exhibiting a decreasing degree of pedogenesis, as well as cryogenic disturbances, would correlate with the Isotope Substages 5e, 5c, 5a of the last interglacial *(sensu lato)* (Figures 29-30). A sudden palaeoclimatic shift toward a progressive warming at the beginning of the interglacial pedocomplex is also recorded in the oxygen isotope curve defining Termination II (*i.e.,* the base of Stage 5, *ca.* 130/127 ka) (Figure 29). The only difference between the two records is in the consistently declining tendency of $\partial^{18}O$ in the oceanic record, whereas a gradual increase of magnetic susceptibility characterizes the palaeoclimatic evolution evidenced by a series of Chernozemic soils with the lowest values in about the middle of the last interglacial (Figure 26). The last interglacial climatic optimum (Stage 5e), defined as the first peak in the oxygen isotope record, however, would correspond to two warming intervals associated with the formation of the basal Brunisol and the overlying thickest Chernozemic soil at a -21/-20 m depth in the studied section (Figures 26, 30).

A new prolonged interval of loess sedimentation relates to progressive cooling of the early stage of the Last Glacial Period (Stage 4) culminating in several cold and hyperarid stadials represented by the first magnetic susceptibility maxima. After a gradual climatic amelioration with periodic appearance of initial Brunisolic horizons, a new interval of low magnetic susceptibility correlates with the mid-last glacial interstadials (Stage 3), separated by episodes of minor loess deposition and surface cryoturbation. The first interstadial (associated with a mature Brunisol or a partly truncated Chernozem, formed during the mid-last glacial climatic optimum) was likely just as warm, but more humid than the present interglacial. Gradual cooling, with higher humidity evidenced by a series of Gleyed Brunisols and Gleyed Regosols, indicates that the transition towards the last glacial stage (Stage 2) was less dramatic than during the previous glacial interval (Stage 4). A marked drop of temperatures associated with intensive loess deposition about the time of the last glacial maximum corroborates the second magnetic susceptibility peak. A progressive warming towards the present interglacial defines the final climatic stage (Stage 1) in the area.

In summary, following the last interglacial, the climate in the Minusinsk basin became more pronounced with cold and dry conditions during the glacial stages interspersed with moderate and more humid mid-glacial interstadials. Both the last and the present interglacials show an increasing trend towards a strongly continental climate with warm summers and cold winters, and a low amount of precipitation. Climatic change during the late Middle Pleistocene was apparently more balanced with longer and more moderate interglacial intervals and less cold and arid glacial intervals.

In terms of general environmental evolution, the recorded palaeoclimatic pattern documents a shift in major ecozones; with southern taiga, established during a more humid and warmer early last interglacial and early interstadial stages, replaced by steppe and tundra-steppe, respectively, as a result of progressed continentality, which was eventually succeeded by boreal forest and tundra forest expansion due to gradual cooling and increase of relative humidity. Eventually, a periglacial tundra-steppe with intensive loess sedimentation and some initial arctic tundra-shrub vegetation developed during warmer oscillations of otherwise very cold and hyperarid glacial stages.

Responding to even minor, short-term atmospheric fluctuations, the continental palaeoclimatic sequence from the Minusinsk Basin is believed to represent a far more accurate, high-resolution record for the particular area, as well as southern Siberia in general, than the "globally-indicative" deep-sea oxygen isotope record. This is because of its geographical isolation in the centre of Asia and far from the world's oceans, which significantly affect local climates along continental margins. Conclusively, the loess - palaeosol sequence at Kurtak (Section 29) represents the most complete and continuous Late Quaternary palaeoclimatic record known in Siberia.

Period/Stage $\partial^{18}O$	Setting	Soil/PK	Fauna	Flora	Early Sites
Holocene	Parkland	Chernozem	*Ursus* sp.	*Betula albae*	N/A
(<10 ka)	- Steppe	Luvisol	*Lepus* sp.	*Pinus silvestris*	
($\partial^{18}O$ Stage 1)			*Marmota* sp.	*Pinus obovata*	
(Kurtak Sk_{2-1})				*Alnus* sp., *Salix* sp.	
				Asteraceae, Fabaceae	
				Cyperaceae, Bryales	
				Polypodiaceae	
LATE PLEISTOCENE					
Sartan Glacial	Tundra-Steppe	Gleyed Regosol	*Mammuthus*	*Betula nana*	*Blade Industries*
(22-10 ka)		Gleysol	*primige[illegible]*	*[illegible]lix* sp.	Kamennyy Log
($\partial^{18}O$ Stage 2)			*Pupilla lo[illegible]*	*[illegible]temisia*	Ust'-Izhul' 2
(KS 29 Sk_{3})			*P. muscor[illegible]*		
			Succinea ob[illegible]		
Karginsk NonG	Forest-Steppe	Brunisol	*Ursus* sp.	*Pinus silvestris*	Kamennyy Log
(55-22 ka)	Boreal Forest	Gleyed Brunisol	*Panthera spelaea*	*Picea* sp., *Abies*	(hor. 11)
($\partial^{18}O$ Stage 3)	(Taiga)	Chernozem	*Alces alces*	*Betula, Larix* sp.	Kashtanka
(KS 29 Sk_{13-4})		Cryosol		*Lycopodium alpin.*	(hor. 11)
				Selaginella sibirica	
Zyryansk Glacial	Tundra-Steppe	Gleyed Regosol	*Bison priscus*	*Artemisia*	
(110/73-55 ka)	Forest-Tundra	Gleyed Brunisol	*Alces alces*	*Chenopodium*	
($\partial^{18}O$ Stage 4)			*Coelodonta ant.*	*Pinus silvestris*	
(KS 29 Sk_{17-14})			*Equus przewalski*		
			E. hydruntinus		
			Succinea oblonga		
			Pupilla muscorum		
			P. turcmenica		
			Vertigo pseudostriata		
			Columella columella		
			Vallonia enniensis		
Kazantsev InterG	Forest-Tundra	Gleyed Brunisol	*Equus caballus*	*Picea* sp.,	*Mousterian*
(130-110/73 ka)	Forest-Steppe	Brunisol	*Bison priscus*	*Pinus sibirica*	Berezhekovo
($\partial^{18}O$ Stage 5)	Steppe	Chernozem	*Coelodonta ant.*	*Pinus silvestris*	(hor. 7)
(KS 29 Sk_{25-18})			*Megaloceros*	*Betula albae*	Kamennyy Log
			gigantheus	Cyperaceae	Ust'-Izhul'
			Ursus sp.	Asteraceae	
			Myospalas m.	Polypodiaceae	
			Citellus sp.	Cichoriaceae	
			Microtus greg.		
			M. oeconomus		
			Lemmus obensis		
			Arvicola terrestris		
			Sorex sp.		

Table 8. Quaternary environments in the Kurtak area: a summary of the pedological, biotic and cultural evidence.

Period/Stage $\partial^{18}O$	Setting	Soil/PK	Fauna	Flora	Early Sites
MIDDLE PLEISTOCENE					
Tazov Glacial	Tundra-Steppe	Gleyed Regosol	*Mammuthus*	*Artemisia*	*Early Levallois*
(180-130 ka)	Forest-Tundra	Gleysol	*primigenius*	*Betula nana*	Ust'-Izhul' 1
($\partial^{18}O$ Stage 6)		Colluviated	*M. chosaricus*	*Pinus silvestris*	
(KS 29 Sk_{28-26})		loess	*Coelodonta*	*Abies sibirica*	*Pebble Industries*
			antiquitatis	*Chenopodium*	Kamennyy Log
			Equus hemionus		Razlog
		Pupilla muscor.	*E. mosbachensis-germanicus*		
		P. turcmenica	*Rangifer tarandus*		
		P. loessica	*Megaloceros gigantheus*		
		Columella col.	*Ovis* sp., *Lepus* sp.		
		Vertigo pseudos.	*Lagurus lagurus*		
		Succinea oblonga	*Cervus elaphus*		
		Vallonia	*Mar rbacina*		
		tenuilabris	*Me*		
Tobol'sk InterG	Forest-Steppe	Brunisol	*E. mosbachensis*	*Pinus silvestris*	*Pebble Industries*
(390-270 ka)	Parkland -Steppe	Luvisol	*Coelodonta ant.*	*Pinus sibirica*	(60-65 m terrace)
($\partial^{18}O$ Stage 7)	Steppe	Chernozem	*Megaroceros g.*	*Betula* sp.	Kamennyy Log
(KS 29 Sk_{32-29})			*Bos primigenius*	*Salix* sp.	Sukhoy Log
			Saiga tatarica	*Alnus* sp.	Razlog: II series
			Lagurus lagurus	Asteraceae	Verkhniy Kamen
			M. oeconomus	Botrichium	Berezekovo 1
			Myospalax sp.		(hor. 8a)
			Mustela sp.		
EARLY PLEISTOCENE					
(>780ka) Glacial			*Equus* sp.		Berezhekovo 2
			E. cf. *hemionus*		(hor. 6)
			Rangifer tarandus		Berezhekovo
			Sorex sp., *Citellus* sp.		(65 m terrace)
			Lemmus obensis		Razlog: I series
Interglacial			*Mimomys* sp.		(alluvial fan)
			Prolagurus sp.		
			Allophajomys sp.		

Table 8. Quaternary environments in the Kurtak area: a summary of the pedological, biotic and cultural evidence (cont').

9. THE QUATERNARY ENVIRONMENTS OF CENTRAL SIBERIA

The following is a regional review of Quaternary environments and related palaeoclimates of Central Siberia. Glacial evidence, palaeontological and palynological data, and the earliest Palaeolithic record are summarized as proxy data. The focus is on Late Pleistocene evolutionary pathways that are correlated with the high-resolution palaeoclimatic stratigraphic framework established in the Kurtak area. Geographically, the region under discussion is delimited by the Altay-Sayan Mountain system in the south, and the Central Siberian Plateau in the north (Figure 42). References to adjacent territories of northern and Central Asia, including western and eastern Siberia, Mongolia, northwestern China, and the eastern regions of the Russian Plain are provided. The geological history and palaeoenvironmental record are treated separately for the Early, Middle and Late Pleistocene.

Because of significant differences between the Western and the Russian systems in the chronostratigraphic terminology, as well as chronological scales, the geological time limits and nomenclature has been adjusted to meet international standards. Most Russian scholars still consider the date of 1.8 Ma as the Pliocene/Pleistocene boundary, although a date of 2.4 Ma has become more widely accepted in accordance with the INQUA chronostratigraphic commission. The term *Eopleistocene* used in the Russian literature does not have an equivalent in the western Quaternary terminology; it refers to the first part of the Pleistocene following the Tertiary Period, broadly corresponding to the Early Pleistocene with an upper chronological limit around 0.78 Ma (the Matuyama / Brunhes palaeomagnetic boundary). The Russian *Lower Pleistocene* includes approximately the first half of the European / American Middle Pleistocene (ca. 0.78-0.4 Ma). The Russian *Upper Pleistocene* comprises the last interglacial and glacial cycle in congruence with the INQUA system. Accordingly, the *Middle Pleistocene* as used by Russian scholars, is significantly shorter (*ca.* 0.4-0.1 Ma). The denomination of glacial/interglacial and stadial/interstadial periods and stages is referred to by the original Russian names, mostly derived from eponymous localities and principal study areas of Central Siberia (Table 1, page 14). The European / North American equivalents, if applicable, are provided in parentheses.

9.1. GLACIAL HISTORY

9.1.1. Continental Glaciations

Climatic change during the Quaternary Period in Central Siberia and the surrounding areas of North-Central Asia is not equally well documented in all areas. Compared with glacial evidence, particularly in mountain areas, interglacial deposits are less known because commonly they are buried or have been eroded. The best studied Pleistocene sections are located in the southern Novosibirsk area (the Ob Loess Plateau) and the adjacent part of the Altay Kray district; in the eastern part of the Kulunda region; in the Kuznetsk Basin; and on the right bank of the Irtysh River (Zykina 1986:115). During the Quaternary glaciation, a wide periglacial zone free of continental ice extended between the partly glaciated Altay and Sayan Mountains in the south and the Arctic ice sheet north of the Podkamennaya Tunguska River in the north. In the west, the periglacial "superzone" (Arkhipov 1991b:63) extended from the Ural Mountains in the west as far as the Verkhoyanskiy Khrebet Range in the east.

During glacial periods, intensive loess accumulation took place adjacent to the continental ice and in foothills. During glacial retreats, glaciolacustrine basins formed in the major continental depressions and river valleys (Arkhipov 1989, 1991b). In the southern part of the glaciated Central Siberian Plateau, changing environments are evidenced by thick interstratified successions of alluvial, lacustrine, aeolian and glacial deposits well-preserved in river valleys and continental depressions (*e.g.,* the Muruktinsk Depression). In the southern periglacial zone, river terraces and loess deposits document the cyclic pattern of interglacial and glacial stages, respectively.

Several major glacial centres are assumed to have existed in Siberia during the Pleistocene. The western Siberian centre was located in the Ural Mountains; the northern on the Kara Sea shelf and the surrounding islands; the central on the Central Siberian Ploskogorye (Plateau), the adjacent Taimyr Peninsula and the Putorana Plateau; and the eastern in the Verkhoyanskiy Khrebet Range. The earliest glaciations are assumed to have started as early as in the Late Pliocene as evidenced by old weathered moraines from the Chukotka and Kamchatka Peninsulas (Laukhin *et al.* 1990b).

Exact timing of long-term climatic change over such a large territory is still problematic, particularly for the earlier periods. In general, about four major regional glaciations are believed to be recognized in the Siberian Quaternary record, dated/estimated to *ca.* 2.5-2.4 Ma, 1.65-1.6 Ma, 0.9 Ma, and 0.75-0.7 Ma (Nikiforova 1989:39). However, the first well-documented glacial event took place after the Matuyama/Brunhes magnetic boundary (Arkhipov 1991a:27). Three major Pleistocene glaciations have been suggested for the southern part of Siberia on the basis of the evidence from the Western Siberian Lowland, the Altay and the Sayan Mountains, and the Transbaykal area, tentatively correlated with the European Alpine glaciations (Arkhipov 1971, 1983; Arkhipov *et al.* 1982:150).

In Central Siberia, the first marked glacial event dates to the Early Pleistocene, correlated with the Lebedskoe Glaciation, evidenced by two moraines separated by marine interglacial sediments in the north, and alluvial deposits in the southern regions (Isaeva *et al.* 1986). Two or three glacial periods are assumed to have occurred during the Middle Pleistocene (Arkhipov *et al.* 1982), although an exact geographical extent of those glaciations is problematic, particularly in southern Central Siberia (Fainer and Fainer 1990). Increasing climatic continentality led to the maximum Pleistocene glaciation believed to have occurred during the late Middle Pleistocene (the Samarov / Illinoian Glaciation). During this time, Central Siberia was covered by an ice sheet of about 85,000 km^2, with the centre located on the Putorana Plateau after merging with the Taimyr Peninsula glaciers (Sukhorukova 1991:81).

The Late Pleistocene glacial history is best known and well documented. Relatively complete sections are found in the Northern Siberian Lowland above the last interglacial marine sediments. The maximum transgression in the area defined by the Kazantsevskaya Svita Formation is dated to about 135 ka, *i.e.*, earlier than the generally accepted last interglacial optimum (*ca.* 125 ka) (Velichko 1993:28). During the early stage of the last Glaciation, four ice centres formed in the Siberian North - the northwestern centre on the Kara Sea shelf, the Putorana and Anabar centres in north-central Siberia, and the east Siberian centre in the Verkhoyansk Mountains (Isaeva 1984). Between 70-50 ka, the Arctic ice sheet expanded southward onto the Putorana Plateau and into the lower Tunguska and Kolyma Basins, and south to about 63° N in the middle Yenisey Basin as evidenced by the Yermakovskaya Moraine (Velichko 1993:49).

The following mid-last glacial Karginsk interval dated to 50-23 ka (Isaeva 1984:21) includes three interstadials and two stadials recognized in continental, as well as marine sequences (Velichko 1993:51). During the final stage of the last glacial, one major glaciation of the Central Siberian North is associated with two major ice-lobes on the Taimyr Peninsula and the Putorana Plateau, but most of the vast territory remained unglaciated. The ice, having advanced farther south than during the previous early last glacial (Early Wisconsinan) stage, is evidenced by three terminal moraines. Two stades of mountain glaciations are assumed for the Urals during the last glacial stage, with the thickest ice accumulated in the northern and western part of the mountains (Velichko 1993:25). In eastern Siberia, the maximum last glaciation is assumed to have occurred prior to 20,000-18,000 B.P. in the Verkhoyansk Khrebet Range (*op.cit.* :56). In the adjacent unglaciated central part of the Siberian lowlands, large glaciofluvial basins were formed by discharge from the major rivers.

9.1.2. Mountain Glaciations

The Pleistocene glacial history of the southern Siberian mountains is insufficiently known because of a lack of systematic, large-scale research. Field mapping has concentrated mostly in better accessible and spatially limited areas. The earliest presumably Middle Pleistocene mountain glaciation is evidenced by erratics correlated with cryogenesis and intensive solifluction of unconsolidated sediments in the foothills (G.Ya. Baryshnikov, personal communication 1994).

Early last glacial ice advances are evidenced by up to 100 m high terminal moraines in the Western Sayan Mountains and 400 m high in the Northern Baykal Range (Figure 1), suggesting considerable alpine glacier activity (Velichko 1993:50). Significant ice thickness has been estimated from the Bol'shoy Sayan Mountains close to the Mongolian border and the Kropotkina Range, possibly reaching up to 700 m on slopes in places of active ice advances or even surges (Rezanov and Nemchinov 1990), whereas only small alpine glaciers are locally distributed on the highest mountain ranges today. Glacial activity became much reduced during the Karginsk (Mid-Wisconsinan) Non-Glacial Interval. An extensive glaciation of the Eastern and Western Sayans occurred during the last glacial maximum (the Chul'khinskoe and Bashkhenskoe Glaciations) correlated with formation of the lower (I) terrace in the Yenisey Valley. Although unglaciated at the present time, numerous glacial landforms (cirques, kames, moraines, erratics) provide evidence for glaciation in the Kuznetskiy Alatau Mountains during the Last Glacial Period (Likhanov and Khaistova 1961:43). This was presumably the main source area of the Kurtak loess.

Very little is known about the Pleistocene glacial history of the Altay, particularly about the earlier alpine glaciations (Figure 42). The transitional zone between the high southern mountains and northern lowlands has not experienced major geographic changes of the landscape since the Upper Pliocene, when the main topographic features were formed. In general, about five major glaciations in the adjacent mounta[illegible] been suggested, with the last two assigned to the Late Pleistocene (Baryshn[illegible] 1992). Those, however, may have had very little effect on the local natural environments in the frontal foothills area (Maloletko 1963:179). In the Gorno Altay, the Early Pleistocene evidence was largely obliterated by erosional processes of subsequent glaciations. Two moraine complexes correlated with 200 m terraces in the foothill valleys are believed to be Middle Pleistocene in age, and presumably represent the maximum glaciation (G.Ya. Baryshnikov, personal communication 1993). The last interglacial deposits (alluvial and alluvio-lacustrine) in the Altay and the Sayans are poorly preserved. Sporadically, relics of 50-60 m terraces are located on older Middle Pleistocene alluvial accumulations. The lower complex terraces are associated with the Late Pleistocene glacial stages (Rakovec and Shmidt 1963). A series of high terraces is developed on the upper Biy River which are believed to have formed during a catastrophic drainage of the Telecke Ozero Lake located in a 900 m deep tectonic graben (analogous to the Baykal Lake) filled by glacial meltwater from the surrounding Western Sayan and Altay Mountains. In the lower reaches of the valley near the city of Biysk, up to 150 m thick loess deposits are exposed above the river. Most of the subaerial sediments are thought to have accumulated during the Late Pleistocene above truncated Early (?) / Middle Pleistocene formations (G.Ya. Baryshnikov, personal communication 1994). An extensive loess cover is also known in the area of the Rudno Altay (30-50 m thick). An epigenetic (*in situ* -) origin of the Altay loess has been suggested (Kriger 1963:144), but the actual formation mechanism remains unclear. Comparing to the major river basins, the mountain loess cover is more discontinuous and variably preserved.

In western Mongolia, the area adjacent to the southern Krasnoyarsk Kray, a major orogenetic period, that began during the Late Miocene and continued until the early Middle Pleistocene, formed a system of mountain ranges separated by deep depressions subsequently filled by large lakes. The neotectonic movement, accelerated during the Early Pliocene, caused significant uplift of the Mongolian Altay, the Gobi Altay, the Khangai, the Kheitei Mountains and the Plateau Dariganga. Simultaneously, a gradual deepening of the basin depressions (the Great Lake Basin, the Valley of Lakes, the Northeastern Mongolian Basin) took place, followed by intensive denudation and river erosion to form 300-500 m deep valleys with up to 250-300 m high terraces (Derevianko *et al.* 1990a:73). The first glaciations ar· ass· iated with the occurrence of large erratic boulders high in the mountains. Dur lacial periods, ice covered parts of he Mongolian Altay and the Khar gai Mountains, h the largest Kobgo River valley glacier re i g 100 km (*op.cit.*:80).

9 1.3. P iglacial Zone

The most detailed geomorphological and geological Late Quaternary record is documented from the main valley systems of the Yenisey and Angara Rivers in the periglacial zone north of the Altay and Sayan Mountair (Figure 1). In the upper Yenisey River valley, Early Pleistocene river incision almost to the present level, relates to formation of the 80-100 m high river terraces. After neotectonic activity, presumably during the early Middle Pleistocene, these terraces were eroded to a thin alluvium at the 60-65 m elevation, and later buried by 10-50 m thick aeolian loess and partly colluviated loessic deposits. In the Angara River valley during this time, up to 25 m of the 70-80 m gravel terraces were formed. Maximum glaciation occurred during the second part of the Middle Pleistocene, corresponding to the Samarov and Tazov (Illinoian) Glaciations, separated by the Shirtinsk Interglacial. In more protected southern areas, climatic change is evidenced by expansion of forest-tundra and southern taiga (Drozdov 1992). Overall, however, the Early and Middle Pleistocene records are fragmentary.

The Late Pleistocene period elapsed without significant neotectonism. The Late Quaternary deposits are thus better preserved, and therefore better understood than earlier ones. Increased river erosion during the last interglacial correlates with accumulation of the 45-35 m and 30-35 m terraces in the Yenisey and Angara River valleys. Pollen records indicate about the same vegetation as at the present time (Laukhin 1986). Following the Zyryansk Glacial, the 15-18/25 m terrace of the early Karginsk Interstadial (43-30 ka), and the 10-12 m terrace of the late Karginsk Interstadial (30-25 ka) were cut into the earlier glaciofluvial deposits. The last terrace (6-8 m) in the Yenisey Valley correlates with the final Sartan glacial stage, and assigned to about 13-10 ka, with two interstadial (Kokorevo and Taimyr) soils on top dated to 12,700-10,800 B.P. (Drozdov 1992:15). Both areas are crucial for correlation of the glacial / interglacial records from the northern Siberian lowlands and the southern Siberian mountains. Geological evidence of the last Late Pleistocene glacial event includes cryogenic structures, loess deposition, intensive aeolian weathering, broad deflation surfaces and deposition of limestone rubble in mountain caves.

In the northern periglacial areas, a harsh and strongly continental climate prevailed during most of the Last Glacial Period, with increasing nivation activity on the Central Siberian Plateau, accelerated erosional processes in the southern river valleys, denudation in the unglaciated mountains foothills and extensive loess and sand sedimentation in smaller inter-mountain , as well as larger continental depressions. Thick aeolian sands and loess were accumulated in Yakutia from deflation surfaces on the major waterdivides during the final part of the Last Glacial Period. Initial hydromorphous soils, developed within the subaerial deposits, indicate syngenetic pedogenesis (Gubin 1990). In northern Eurasia in general, the maximum loess sedimentation of the late last glacial stage occurred between 23-17 ka and [illegible] 12 ka, separated by a short interval of pedogenesis (17-16 ka). In an[illegible]ria, at least three warming oscillations have been recorded between 15 ka and [illegible].5 ka, characterized by forest expansion as indicated by pollen data (Velichko *et al.* 1990:30-32). On the European Russian Plains, the maximum last glacial cooling related to an intensified loess sedimentation is dated to 23-17 ka, stratigraphically above the Bryansk Soil (Velichko 1993:19).

During the last glacial maximum, permafrost expanded to a depth of 400-600 m in the south, and 600-800 m in the unglaciated north, with a maximum of more than 1000 m beyond the ice centre on the Putorana Plateau in north-central Siberia (Velichko and Nechaev 1992). However, the initial permafrost build-up occurred during the early Quaternary (Baulin and Danilova 1984). Permafrost was not preserved during the last interglacial in the western part of Siberia, whereas it remained widely distributed farther east. Syngenetic and epigenetic frost-cracking in the last interglacial soils in the Yenisey and the Angara River basins indicates much more severe ground freezing conditions than at present. A high permafrost table is recorded for the Karginsk Interval, as evidenced by intensive involutions of soil horizons. Ice-wedge polygons caused by frost action formed in the Minusinsk Basin and the surrounding mountains during the Sartan Glaciation (Baulin and Danilova 1984).

9.2. PALAEOPEDOLOGICAL EVIDENCE

Past climatic change in Siberia is best elucidated by the palaeopedological record, although there are persisting problems in correlation of equivalent units on local, as well as regional scales (Volkov 1983).

In the eastern part of southern Siberia during the Late Pliocene, reddish alluvial and aeolian sediments weathered into brown soils suggest warm and humid conditions, also indicated by mixed forest-steppe pollen with pine, birch and oak. Modern types of soils appeared for the first time towards the end of the Pliocene as a result of progressive continental cooling, as evidenced by the appearance of fossil periglacial features (as involutions, ice wedge casts). Climatic fluctuations became more pronounced at the beginning of the Early Pleistocene recorded by a greater variety of soils formed (Chernozems, Kastanozems, Brunisols) typical of diverse natural settings and a mosaic landscape. The first marked climatic shifts occurred late during the Early Pleistocene, causing a complete zonal palaeogeographic rearrangement. Warm and humid conditions with forest expansion and formation of present-day soils characterize the Early and Middle Pleistocene interglacials (Medvedev *et al.* 1990).

In the Kuznetsk Depression west of the Minusinsk Basin on the other side of the Kuznetskiy Alatau Mountains, a spectacular Quaternary record is provided by a 30-60 m thick series of loess and lacustrine deposits intercalated by up to 16 well-developed Early, Middle and Late Pleistocene palaeosols above the Pliocene red soils (Zubin *et al.* 1982). A series of buried deeply-weathered Early and Middle Pleistocene soils are also documented in the Rudno Altay area with physical characteristics of a more pronounced pedogenesis than encountered in arid areas of Central Asia, possibly because of more intensive interglacial weathering processes in high mountain areas due to increased solar radiation (Kriger 1963). "Subtropical" red soils developed on a calcareous substratum have been recently described at the Ust'-Karakol Section, in the Gorno Altay, and assigned, on the basis of a TL dating, to the Lower Pleistocene (the Middle Pleistocene according to the INQUA chronostratigraphy) (Derevianko *et al.* 1993) (Figure 42).

The last glacial cycle was most climatically divergent in terms of glacial and interglacial / interstadial maxima, corroborating the evidence from the upper Yenisey River region. In the western part of the Irkutsk region, a key section is documented at the Igetey Site on the banks of the Bratsk Reservoir along the upper Angara River, with up to a 10 m high sequence of loess deposits, including about five well-developed palaeosols on the Cambrian bedrock (Figure 42).

Along the upper Angara, deep brown forest and Chernozemic forest-steppe soils developed during the last interglacial in response to favorable natural conditions and a higher concentration of biomass than at the present time. *Larix* and some deciduous arboreal taxa, including *Quercus* sp. (oak) and *Tilia* sp. (lime),which do not occur in southern Siberia today, were widely distributed in the area (Medvedev *et al.* 1990:12). A variety of soils, developed during the last interglacial, documents the changing ecology of particular areas. At the Ust'-Koba Site on the Angara River (Figure 42), forest-tundra is indicated by tundra Gleysols, northern taiga by Gleyed Podzolic soils, central taiga by Podzolic soils, southern taiga by brown soils, and dark coniferous forest by dark brown forest soils (Demidenko 1990b:140). Previously, Podzolic soils of central taiga, gray forest soils of southern taiga and Chernozems of forest-steppe have also been identified on the southern Yenisey Plateau. There, Chernozem-like parkland soils are succeeded by steppe Chernozems, and eventually by brown and gray forest soils by the end of the interglacial *(op.cit.*:141). Taiga and steppe-taiga were the dominant habitats in the Krasnoyarsk region, whereas steppe and mosaic forest-steppe prevailed in the Chulym Basin (Demidenko 1992:67). Despite the zonal differences in the vegetation pattern on the Angara Plateau with more forest and forest-steppe communities than in more open settings in the Chulym-Balakhtinsk Depression, some general similarities in the palaeoenvironmental conditions are evident (Demidenko 1990b:table 1).

A temporary warming trend, following the loess sedimentation of the early last glacial stage, is pedologically evidenced by the Baygansk Soil, succeeded by a new interval of loess deposition and subsequent solifluction towards the end of the mid-glacial stage. Periglacial forest-steppe and forest-tundra were present in the area during the early stage of the last Glaciation (the Zyryansk Glaciation). The Malokhetskoe Interstadial of the early Karginsk (Mid-Wisconsinan) Interval is represented by a variety of soils similar to those of the last interglacial with permafrost soils of forest-tundra, Podzolic and brown soils of *Pinus - Betula* parklands, and forest-steppe Chernozem-like soils. However, the latter reflect local conditions in more protected southern areas, than a regional climatic pattern *(op.cit.*:142). In the Krasnoyarsk Kray today, analogous arctic tundra soils are distributed north of 75° N latitude; Gleyed Podzolic soils and Gleysols of forest-tundra between 65° and 70° N; Podzols of northern taiga with *Pinus silvestris* and *P. sibirica* between 60° and 65° N; and gray and brown forest soils of southern taiga north of 56° N (Yepokhina 1961).

As in the Minusinsk Basin, brown forest soils of the Osinsk Pedocomplex of the Angara Valley, formed during the Karginsk Interstadial (50-23 ka), indicate an analogous warmer and humid climate. In accordance with the Kurtak record, the increased humidity continued throughout the early part of the last glacial stage until about 21 ka, accompanied by solifluction processes and evidenced by a series of colluviated Brunisols. A return to the previous environmental zonation is documented during for the later Lipovsko-Novoselovo Interstadial (30-22 ka). Humid and cold oscillations during the early last glacial were gradually succeeded by cold and dry phases, corroborating the evidence from Kurtak (Demidenko 1992:68). The maximum loess sedimentation (up to 6 m) occurred before 17 ka, followed by a temporary permafrost disintegration, a progressive gleyzation and initial soil formation due to establishment of forest-tundra with *Pinus cembra, Pinus* sp. *(Dyploxylon), Pinus sibirica* and *Betula* sp. (Medvedev *et al.* 1990:14). Only incipient soil horizons developed during warmer oscillations of the last glacial stage between intervals of maximum loess sedimentation. In the upper Angara River region, increased aridity during the following Konoshelskoe Stadial contributed to expansion of forest-steppe with secondary podzolization and gleyzation in more humid and ground-saturated settings. In the Kurtak area, such brownish, loamy zonal soils have been identified at the Berezhekovo and Listvenka Sites, and correlated with the colder Trubchev Interstadial (17-16 ka.) with boreal forest-steppe, and the following humid Kokorevo and the drier Taimyr Interstadials characterized by forest and steppe expansion, respectively (Deminenko 1990b:144; 1992:67). A new episode of pedogenesis occurred on the Angara between 14.0-10.5 ka after a period of loess sedimentation (Medvedev *et al.* 1990:14). During the early postglacial, solifluction was particularly active, although did not reach the intensity of the early last glacial stage prior to the Karginsk Non-Glacial Interval. A similar climatological and palaeopedological pattern has been documented in adjacent Siberian regions.

Overall, the loess - palaeosol record from the southern Angara River region is very similar to the stratigraphic sequence and pedogenic developments in the Minusinsk Basin, suggesting more or less the same climatic conditions throughout the transitional zone between the high southern Siberian mountains and the northern lowlands and plateaus during the Late Pleistocene. Although not as deep (only 8-10 m) as the Kurtak loess, the Igetey sequence shows identical palaeoenvironmental shifts in terms of loess sedimentation, soil formation and periglacial disturbance. The only apparent difference consists in the absence of any pedogenic hiatus in the 2-2.5 m thick last interglacial Igetey Pedocomplex, which is shown by a thin layer of loess between two Chernozems in the Kurtak Pedocomplex and the Berdsk Pedocomplex of western Siberia. A buried soil of the Osinsk Pedocomplex dated to 24,400±400 B.P. (GIN-5327) (Medvedev *et al.* 1990:28) stratigraphically and radiometrically corresponds well to a Brunisolic horizon at Kurtak IV (Table 2). The lower degree of mineral weathering of loess, largely composed of quartz (55-75 %), plagioclase (10-20 %) and gneiss (2-8 %) (*op. cit.*:46), indicates a less pronounced weathering process in this area, and possibly throughout southern Siberia during the Late Pleistocene. Consistently, several periods of sedimentation, pedogenesis, cryoturbation and solifluction are recorded in the Kurtak and Igetey sections. Overall, a climatostratigraphic uniformity of Late Pleistocene environments is well evident in the foothill zone north of the Altay - Sayan Mountains of southern Siberia.

The interglacial and interstadial soil complexes recognized in the Kurtak area have also been recorded in other parts of northern Eurasia. The lower well-preserved dark brown Chernozem from Kurtak Section 29 may correlate with any of the late Middle Pleistocene soils from the Kuznetsk area (Volkov and Zykina 1991:44), or the Koinikhinsk (Riss 1/2) or the Shipunov (Mindel/Riss) Pedocomplexes from the Novosibirsk Loess Plateau (Volkov and Zykina 1983). The most developed last interglacial palaeosol sequence chronostratigraphically corresponds with the Berdsk Pedocomplex from the Novosibirsk Priobie (Dergacheva and Zykina 1978). On the European Russian Plains, the analogous and very distinctive Mezinsk Pedocomplex traceable throughout much of the territory follows the late Middle Pleistocene (Riss 1/2 ?) Kamenskiy Pedocomplex characterized by developed brown forest soils (Velichko *et al.* 1992b:135).

On the Ob Plateau (the Novosibirsk Priobie), the eastern part of Kulunda, and on the Altay Plains near Biysk in the southeastern part of western Siberia, loess deposits up to 150 m thick accumulated during the Late Pleistocene during several stages of aeolian reactivation of fine silt derived from large (up to 100 m deep) deflation surfaces in lowland areas (Volkov and Zykina 1983:19-20). The early Quaternary record is little known because of the lack of good sections (Foronova 1990). The series of palaeosols from the Novosibirsk Loess Plateau includes the Evsinsk Pedocomplex with two strongly weathered Luvisolic soils located below the Matuyama / Brunhes palaeomagnetic boundary; the Shardikhinsk Pedocomplex with a brown forest and Chernozemic soils characteristic of a pronounced continentality of climate; the Shipunovsk Pedocomplex formed by three forest-steppe luvisolic Chernozems assigned to the Mindel/Riss (Yarmouth) Interglacial; the Kamikhovskiy Pedocomplex with two Chernozems indicating a milder continental climate during the warmer Riss (Illinois) 1/2 interstadial; the Berdsk Pedocomplex with two semi-arid steppe Chernozems of the last interglacial; and the Iskitimskiy Pedocomplex formed by two secondarily colluviated Chernozems dated to 34-24 ka (Zykina 1986) (Figure 42).

Loess sedimentation is believed to have taken place during the second half of each glacial period, with palaeosol formation during the succeeding interglacials, followed by cryoturbation during the first half of the each glacial (Volkov and Zykina 1977). Climatic fluctuations in periglacial areas are recorded by solifluction episodes between periods of soil formation and loess deposition on the Novosibirsk Loess Plateau (Volkov and Zykina 1982:27). A moderate runoff in the river valleys and on slopes during the last interglacial optimum accelerated towards the end of the interglacial, but reached its maximum during the early last glacial stage and the early and late parts of the mid-last glacial (Karginsk) interval (Volkov and Zykina 1984:124). In vast areas of western Siberia, the buried soils and loess deposits have been extensively used for pedo-bio-stratigraphic correlation of past climatic cycles (*e.g.*, Arkhipov *et al.* 1982; Arkhipov 1989). During the Early and Middle Pleistocene, palaeosols in southern Central Siberia were mostly developed on lacustrine sediments, but loess was the main soil substratum during the Late Pleistocene (Volkova 1990).

In terms of a broader regional correlation of the Kurtak sequence with the well-known western Siberian loess - palaeosol record, the basal Chernozem could correspond any of the late Middle Pleistocene Chernozemic soils; the colluviated loess to the Suzunskiy Horizon which has been correlated with the Oxygen Isotope Stage 6; the last interglacial pedocomplex to the Berdsk Pedocomplex (Stage 5), including two cryogenetically disturbed Chernozems separated by a loess horizon; the early last glacial loess to the Tulinskiy Horizon (Stage 4), the mid-last glacial pedocomplex to the Iskitimskiy Pedocomplex with two soils and a loess horizon (Stage 3); the late last glacial loess to the El'tsovkiy Horizon (Stage 2), and the present Chernozem to the Cyminskiy Pedocomplex (Stage 1) (Volkov and Zykina 1991). As in the upper Yenisey loess record, sharp transitions between cold and warm intervals have been observed (Velichko 1993:36).

Temperatures during the last interglacial optimum are believed to have been higher than at present (Volkova 1990), while climatic conditions were broadly similar during the warmer interstadials with forest-steppe and steppe vegetation (Zykina 1990:33). Approximately the same average July temperature, a slight (*ca.* 1°C) increase of the MAT, and 100 mm increase in annual precipitation is assumed for the last interglacial optimum in the western part of southern Siberia (Velichko *et al.* 1991:207-209). During the climatic optimum of the last (Kazantsev) Interglacial, vegetation zones in western Siberia are believed to have shifted by 500-700 km to the north due to increased summer temperatures by +4/+5°C relative to the present values (Volkov and Zykina 1991:44). On the Russian Plain, an increase of January temperatures by up to 8°C to an average of -3°C has been assumed, whereas July temperatures were similar to those of the present (Velichko 1993:13). Increased precipitation and higher annual temperatures contributed to northward expansion of deciduous forests about 500-600 km beyond their present distribution limits.

The Berdsk Pedocomplex, 2.5 m thick, developed during the last interglacial (the Kazantsev Interglacial) and the early last glacial (the Krutinskoe Interstadial), and consists of two Chernozemic soils, the latter of which was disturbed by up to 3 m deep frost wedge casts. A 0.5-1.0 m loess horizon is interbedded between the soils (Velichko 1993:32). The lower Chernozem exhibits longer pedogenic development under a warm and humid climate in the forest-steppe zone, whereas the lower less thick forest-steppe Chernozem has been correlated with the Brørup Interstadial of Europe (Volkov and Zykina 1984:121). Both soils were subsequently modified under humid and cold conditions by cryogenic processes to form a network of fissures filled by humic loam (*op.cit.*:fig. 12-3). A similar pattern is recorded at Kurtak.

During the following early last glacial stage, gleyzation processes took place within the 1.5-2.5 m Tulinskiy (Tula) Loess Horizon. The mid-last glacial interval (the Karginsk Interstadial) is characterized by the Iskitimskiy Pedocomplex, including two soil horizons separated by a 0.7 m loess. The lower, more mature brown Luvisolic soil (?) / Chernozem with an illuvial Bt horizon, dated at 35-31 ka / 33-29 ka, was formed during the interstadial climatic optimum may well correlate with the Kurtak truncated Chernozem/Brunisol. The upper Podzolic (?) or weakly developed Chernozemic soil has been dated at 26-24 ka / 22-19.5 ka (Velichko 1993:33-35; Volkov and Zykina 1984:122). The earlier soil is thought to have formed under climatic conditions like those of the present. Both soil units are secondarily disturbed by cryogenic processes. Correlation with the Kurtak record is unclear, as at least three well-developed soil horizons are present at Kurtak, and the chronostratigraphy does not quite correspond to the western Siberian sequence (with the first soil from the Ob Plateau and the second soil in the Minusinsk Basin both dated to ca. 30 ka).

Weakly developed late interstadial soils of the Bryansk / Karginsk Pedocomplex of the Russian Plain and Siberia, dated at 31-24 ka, are similar in degree of pedogenesis to Humic Gleysols of central Yakutia, formed under cold climatic conditions with average January temperature less than -27°C and low precipitation (300-400 mm) (Velichko 1985; Morozova 1981). A very low degree of weathering is also characteristic for interstadial soil horizons of the last (Vaidal) glaciation in the Russian Plains. Limited solifluction, reflecting reduced snow cover suggest strongly continental climate in the region (Velichko 1993:19).

The maximum aridization at the western margin of southern Siberia correlates with the Eltsovskiy Loess Horizon, dated to 19-14 ka, and chronologically corresponds to the time of formation of large, ice-marginal glaciolacustrine basins in the Irtysh and Ob River valleys. Finally, following the Suminskiy (Suma) Pedocomplex with several incipient soil horizons, the Baganskiy Loess accumulated during the final Pleistocene and the early Holocene (Velichko 1993:32-35).

In sum, the palaeopedological evidence from Siberia and adjacent territories shows a pattern of considerable uniformity of pedogenic processes over vast areas throughout the Pleistocene in spite of the morphostructural differences between particular geographical settings. The presence of almost identical palaeosol records in the main river valleys (the Ob, Yenisey and Angara River basins) in southern Siberia suggests broadly similar climatic conditions over particular time intervals throughout the territory.

9.3. BIOTIC (PLANT AND POLLEN) EVIDENCE

Pollen and plant macrofossil data provide important evidence on Pleistocene ecology of western Siberia and Mongolia, particularly from the natural depressions in the transitional zone between the high mountain ranges and the northern lowlands.

A significant cooling in Siberia has been recorded for the Pliocene with pollen taxa indicative of tundra and forest-tundra (Volkova and Baranova 1980). Recently, a classification of long-term climatic fluctuations between the Late Pliocene and the Late Pleistocene has been provided on the basis of variations in specific pollen concentrations (Volkova 1991). The warmest interglacials (1.2 Ma, 130 ka, 5.5 ka) were presumably characterized by an increase of temperatures by 3-6°C in July and 2-7°C in January allowing some subtropical vegetation to establish *(Azola interglacialica)*. A mild boreal climate is assumed for interglacials (800 ka, 290 ka, 190 ka) with expansion of dark coniferous forests of pine *(Pinus silvestris)*, spruce *(Picea glauca)*, fir (*Abies* sp.) and Siberian cedar *(Pinus sibirica)*, suggesting about the same average July temperatures (+18°C), but a rise of average January temperatures by 7-9°C (to -10/-12°C). Moderately cold boreal interstadials (650 ka, 70 ka) marked by an increased spruce distribution would imply a 3-4°C temperature decrease in January to about -25°C, and approximately the same negative deviation in July to about +15°C. In western Siberia in general, forest-steppe vegetation expanded during the early interglacial periods, whereas periglacial steppe vegetation prevailed during the later interglacial stages.

Three vegetation chronozones have been suggested from the pollen record for the last interglacial from northern Asia: a) the early stage, with coniferous forests dominated by pine and spruce, indicating approximately the present-day climate; b) the middle stage, with expansion of spruce and Siberian pine ("cedar") during the climatic optimum with increased annual temperatures and humidity then now, and c) the late stage, with pine and birch, suggesting colder and drier conditions (Velichko 1993:49). During the interglacial climatic optima, dark coniferous taiga with cedar, fir and spruce became broadly distributed on the Central Siberian Plateau; mixed forests with a significant proportion of warm deciduous trees became established in the southern Angara and Yenisey River valleys (Velichko 1993:47). Thermophilous taxa as lime also appeared in the Northern Baykal Range. Studies from the key Yelogui Section in the middle Yenisey River valley suggest a 14°C rise of the average July temperature during the early last interglacial to +18°C during the last climatic optimum. Average January temperature of -18°C during that time dropped to -26°C by the end of the interglacial. Mean annual precipitation is estimated to have increased by 100% from 250 mm during the early interglacial to about 500 mm during the climatic optimum (Gurtovaya 1987).

During glacial periods, only birch *(Betula sibirica)* in tundra-steppes, and pine *(Pinus* sp.) with spruce (*Picea* sp.) in river valleys were the principal arboreal species that survived under cold periglacial conditions. During the penultimate glaciation, vegetation was reduced to more resistant subarctic taxa *(Betula nana, Salix polaris, Picea obovata, Larix sibirica)*, indicating an annual temperature drop of 9-10°C to average January temperature of -34/-40°C, and +14/+16°C in July. During the last glacial maximum, being the coldest glacial interval during the Pleistocene, polar tundra covered most of Siberia with only an ephemeral vegetation present *(Betula nana, B. verrucosa, Artemisia vulgaris, Lycopodium appressum, Polygonum amphidium)*. Pollen of birch and pine occasionally occurs in the early last glacial interstadial deposits from southern as well as northern Siberia (Velichko 1993:50). Only Polypodiaceae *(Lycopodium pungens, L. alpinum)* with some *Alnus* sp. and *Betula nanae* were present under early last glacial conditions. Forest-tundra trees became more widely distributed during the later phase due to progressive warming and increased humidity.

In the North Siberian Lowland, the warming trend during the Karginsk (Mid-Wisconsinan) Non-Glacial Interval is evidenced by pollen of pine, spruce, larch and the Siberian "cedar" during warm interstadials, and birch during cold stadials. A similar distribution is documented in the Central Siberian Plateau (Velichko 1993:51). In the southern area of Central Siberia, dark coniferous forests (southern taiga) dominated by *Pinus sibirica* (20-50%) and pine (10-30%), with spruce, fir and larch, experienced a significantly broader geographical distribution. During the climatic optimum, arboreal taxa represent 60-90% of the pollen records (*op.cit.*:53).

In western Siberia, Early Quaternary ("Eopleistocene") deposits dated to 1.8-0.7 Ma are known from the extraglacial steppe-forest zone at the southeastern margin of the Western Siberian Lowland (the Kochkov Horizon). The deposits are represented by fluvial sands and gravels accumulated under moderate climatic conditions, followed by lacustrine clays laid down during colder intervals. Palynologically, this change is also documented by replacement of a forest vegetation dominated by birch with the Siberian "cedar" and pine, by a more cold-adapted forest-tundra and arctic tundra taxa, including *Saxifraga* and *Rubus*. The southward shift in zonal flora distribution is estimated to have been about 9-10°. The onset of cold climatic conditions tentatively was correlated with the Alpine (Günz) glaciation (Volkova 1979). However, it is evident that the geological record is largely incomplete and the inferred climatic history more complex.

The early Middle Pleistocene (0.7-0.4 Ma) is represented by two periods - the Talagaikin Interglacial and the Shaitansk Glacial. The climate during the interglacial is assumed to have been comparable to the present. Two glacial events of the subsequent glacial period (correlated with the Mindel Glaciation) are evidenced by a southwards migration of forest-tundra to about 58° and 55° N., respectively. Arctic and alpine plant taxa spread over much of the southern portion of Siberia, suggesting a mean annual temperature drop of about 7-8°C below the present value (Volkova 1979:248-249).

The late Middle Pleistocene in western Siberia (0.4-0.1 Ma) includes four major periods with two glacial - interglacial cycles. The Tobolsk Interglacial, represented by marine, fluvial and lacustrine sediments, is thought to be equivalent to the Mindel/Riss Interglacial. During its climatic optimum, the forest boundary presumably moved 300-500 km farther north, and the dominant arboreal taxa (pine and spruce) became distributed over much of Siberia. Ostracoda and molluscs indicate a significantly warmer climate than today. A progressive cooling at the end of the interglacial is documented palynologically by the increase of steppe taxa at the expense of coniferous trees, correlating with a temperature fall of about 3-4 °C. During the onset of the Samarov Glaciation, arctic and tundra zones moved southwards, perhaps as much as 1000-1200 km. Newly established frigiphilous vegetation, with dwarf birch *(Betula nana)*, polar willow *(Salix polaris)* and herbs with grasses *(Lycopodium pungens, Juncus arcticus, Ranunculus hyperboreus)* indicates very dry and cold conditions. A warming trend of the Shirta Interstadial caused a temporary climatic amelioration before the onset of a new glacial period (the Tazov / Riss / Illinoian Glacial). The MAT is assumed to have been about 9-10 °C lower than at present. The landscape was characterized by glacial and periglacial vegetation dominated by grass and sedges of the Chenopodiaceae-Gramineae group with some aquatic plants and mosses in boggy areas; and dwarf birch and spruce in open steppes. The forest boundary may have shifted about 1000 km to the south *(op. cit.)*.

The Late Pleistocene (125-10 ka) starts with the Kazantsev Interglacial (130-110 ka) during which coniferous forests spread far to the north and became established over large parts of the Siberian territory, including the modern tundra and forest-tundra zones. In central Siberia, the forest line is assumed to have shifted 2-4° to north (Baulin and Danilova 1984). Presence of thermophilous plant and animal species indicates that during the climatic optimum temperatures were about 4-5 °C higher than present. During the subsequent cooling related to the Zyryansk Glaciation (110/100-55 ka), forest-tundra and steppe-tundra became re-established. The vegetation patterns presumably changed insignificantly during the following Karginsk Interstadial Interval (55-24 ka), although forests experienced some northward expansion. The last climatic deterioration occurred during the Sartan Glaciation (24-10 ka), which was probably the coldest glacial stage during the Pleistocene, similar to the Samarov (early Illinoian) Glaciation. The severe climate caused periglacial tundra to migrate south to about 55° N (Arkhipov *et al.* 1977; Volkova 1979).

In southern Siberia, informative but isolated records come from the mountain valleys and the foothill depressions. Warm subtropical flora of the late Early Pleistocene, including lime, chestnut, and thermophilous malacofauna were recorded in a deeply weathered terra rossa soil dated by TL at 770 ka at the Chernyy Annui / Ust'-Karakol Site in the northwestern part of Gorno Altay (Derevianko *et al.* 1992c) (Figure 42). Fluviolacustrine sediments with abundant fossil faunal remains, plant macrofossils and pollen from buried valleys in the central part of the southwestern Altay provide evidence of changing natural conditions towards the Middle Pleistocene, characterized by increased aridity and low temperatures (Deviatkin 1963). Approximately present-day conditions existed in the Altay Foothills area during the early Middle Pleistocene with grasses (30-70%) and some trees (birch, pine and spruce) (the Lower Kochkov Horizon). An increased amount of pine pollen indicates colder (glacial?) conditions during the following period (the Upper Kochkov Horizon). In general, a changing vegetation pattern of forest-steppe and periglacial steppe may have characterized the climatic conditions during interglacial and glacial stages, respectively (Maloletko 1963:181).

A Middle Pleistocene pollen record from the upper Yenisey region, from a 60 m terrace in the Khamsara River valley, shows a cold periglacial vegetation established in the area. The present arboreal and grass taxa, including *Pinus silvestris* (19.3%), *Pinus* sp. (16.3%), *Betula nana* (4%), *Picea* sp. (3%), Cyperaceae (26.5%), Chenopodiaceae (6.1%) and *Artemisia* (1%) chronostratigraphically correlate well with the fossil remains and the Early Palaeolithic stone industries derived from river terraces at the same elevation in the Kurtak area. An Upper Pleistocene palynological assemblage from a 40 m terrace indicates a similar climate, although the pollen spectrum partly differs, with *Pinus* sp. (53.9%), *Picea* sp. (7.8 %), *Abies* (1.8%), *Alnus* (1.3%), *Ephedra* (10.4%), Chenopodiaceae (7.8%), Cyperaceae (7.8%), *Artemisia* (1.3%) (Yamskikh 1972:63-64, 74).

During the last interglacial optimum, a belt of mixed broad-leafed forest with dark conifers formed in southern Siberia (Grichuk 1992a). Dark coniferous forest dominated by larch, with average January temperatures of -22°C and July temperatures of +16°C, and mean annual pr ıpitation of 385 mm, were established in the Altay Foothills. On the eastern slop ne Abakan Range on the southern margin of the Minusinsk Basin, the presence of a larch-pine forest and forest-steppe suggests average January temperatures of -18°C, July temperatures of +18°C close to the present average, and annual precipitation of 450 mm about 100 mm more than present-day values (Velichko *et al.* 1991:195, 207-209). For the northern part of the basin, average January temperatures at about 120 ka are estimated to have been 4-6°C more than present values, the February temperatures increased by 3°C, July temperatures by 1°C, and August temperatures by >2°C, the MAT by >2°C, and the mean annual precipitation increased by 100-200 mm (Velichko *et al.* 1992a). A key pollen record for the last interglacial close to the study area is known from the northeastern foothills of the Kuznetskiy Alatau Range. A gradual transition within the cryoxerotic stage of the late penultimate glaciation is suggested by periglacial *Artemisia* tundra and insular spruce forest that was replaced by dark coniferous forest and forest-steppe at the end of the glacial period. The zonal succession during the following thermoxerotic stage of the last interglacial is manifested by expansion of dark coniferous forests *(Picea, Pinus sibirica* and *Abies)* during the early interglacial, enriched by some broadleafed taxa *(Alnus, Quercus, Ulmus* and *Tilia)* during the interglacial climatic optimum, and reduction to more cold-adapted taxa during the late interglacial, including birch, alder, "cedar" *(Pinus sibirica)* and pine *(Pinus silvestris)* (Grichuk 1984:165).

Dated pollen sequences from the Karginsk non-glacial (interstadial) deposits provide evidence for a variety of settings in southern parts of western Siberia. Forest-steppe with birch and fir became established at Krasnyy Yar in the Novosibirsk area between 30,870-28,200 B.P.; coniferous forest dominated by spruce from the Lipovka Site in the Tovol River valley, 100 km SW of Tobolsk, dated around 30 ka.; and steppe from the Ob River valley with radiocarbon dates of 33,900-32,300 B.P. (Nikolaeva *et al.* 1989). In southern Central Siberia in general, average February as well as August temperatures are thought to have dropped by about -4/-6°C compared to the present, the MAT by -2/-4°C, and the mean annual precipitation by 0-100 mm for the late mid-last glacial interstadials between 35-25 ka (Frenzel 1992a). At cave sites in the Altay, climatic amelioration during the Karginsk / Mid-Wisconsinan Interval is indicated by a marked increase of coniferous trees, including spruce, fir and larch, but largely dominated by pine *(Pinus silvestris)*, replacing the colder steppe-tundra stadial vegetation with *Artemisia, Asteraceae, Chenopodium* and *Salix* (Derevianko and Markin 1992).

A significant retreat of boreal forest during the following Sartan (Late Wisconsinan) Glacial reflects a considerable drop in temperatures and a shift towards a harsh, arid and extremely cold climate (Vorobieva 1992:48). During the last glacial maximum, *i.e.*, between 21,000–18,000 years B.P., average February temperatures are assumed to have decreased by -12/- 4°C, August temperatures by -6/-8°C, the MAT by -8°C, and the mean annual precipitation by 250 mm (all minimal deviations from present-day values) (Frenzel 1992b). Steppe communities with drought-resistant taxa *(Artemisia)* and some locally distributed patches of spruce and birch forest covered periglacial lowland areas of southern Siberia (Frenzel 1992c), with boreal pine and spruce forest in foothills (Grichuk 1984:174; 1992b). Regional temperatures raised again during the Holocene climatic optimum with positive average deviations of 2-3°C during January and 1°C during July (Klimanov and Velichko 1992).

In western Mongolia, a warm and humid subtropical Pliocene vegetation flourished in the upland areas during early periods of reduced tectonic activity and stable climate, receiving, in now arid lowlands, up to 1000 mm of precipitation (compared to 400 mm today). Dark coniferous forests with hemlock, pine, spruce and fir have survived in certain locations, mainly in river valleys until the present. The Late Pliocene and Early Pleistocene pollen data indicate a broadleafed deciduous forest established in the region with winter temperatures 10-12°C above the present values (Deviatkin and Malaeva 1990a:179). Increased aridity during the Late Pliocene caused an expansion of forest-steppe and retreat of coniferous taiga. Some deciduous taxa, including *Acer, Buxus, Tilia* and *Quercus* indicate that warmer and less continental climate persisted in more humid river valley refugia with estimated mean annual precipitation of >700 mm (Derevianko *et al.* 1990a:92).

Significant shifts in the climate due to regional aridization during the Early Pleistocene caused a total transformation of biotic zones. Increased continentality during the early Middle Pleistocene (after 0.7 Ma) is monitored by the presence of recent arboreal taxa *(Larix, Picea, Pinus sibirica, Betula)*. Forest expanded from the mountains into the former steppe zone during warmer and humid, but still generally cool periods, but retreated during cold climatic oscillations (Deviatkin and Malaeva 1990b:135). Mixed parklands prevailed during most of the Pleistocene, with average July temperatures during colder periods of 10-17°C, and a low amount of precipitation (400-500 mm). Deciduous forests covered large areas of the more humid mountainerous parts of western Mongolia during the last interglacial. In the adjacent southern Transbaykal area, thermolphilous taxa *(e.g., Azola)* survived until the end of the Middle Pleistocene (Yerdikhinskiy 1990:200).

The continentality over much of Mongolia and the adjacent regions become stronger during the Late Pleistocene, although relatively warm summer temperatures persisted during the early last glacial (with an average of 17-19°C), leading to establishment of open grasslands because of increasing aridity. During the mid-last glacial stage, relative humidity increased to allow some trees temporarily to recolonize (mainly birch), before being replaced by arid periglacial steppes during the last glacial stage (Deviatkin and Malaeva 1990b). Warm and arid conditions in lowland southern Mongolia, and cold and more humid conditions in the northern mountains characterize the present climate.

In sum, the palaeoclimatic history of southern Siberia and the surrounding areas shows a changing pattern of forest expansion during interglacial stages, and tundra-steppe expansion during cold intervals. Nevertheless, the vegetation zonation was locally more complex according to geographical and topographical variables. Also, the interglacial ecology does not include a single natural setting, but shows a gradual transition from particular ecotones before and after the climatic optimum. Similar gradual changes can be followed in the palaeoenvironmental evolution during the glacial stages.

9.4. FOSSIL FAUNAL EVIDENCE

Fossil faunal records are rather limited, and no comprehensive study is available at present. The best fossiliferous sections with late Tertiary and early Quaternary (Pliocene and Early Pleistocene) palaeontological remains are located in western Mongolia, although even those are still very poorly studied (Derevianko *et al.* 1990a). The first significant faunal changes occurred after the Jaramillo palaeomagnetic event (0.98-0.9 Ma) in response to gradual cooling with increasing amplitude during the late stage of the Early Pleistocene (*i.e.*, > 0.78 Ma) (Arkhipov 1991a:27). Most of the palaeontological finds, however, date to the Middle and Late Pleistocene, with mammoth *(Mammuthus primigenius)*, aurochs *(Bos priscus longicornis)*, horse *(Equus caballus)* and wooly rhinoceros *(Coelodonta antiquitatis)* being the most frequently encountered megafaunal forms. The corresponding faunal record from the upper Yenisey Valley is summarized in the palaeoecology section of the Kurtak area. There, the fauna record is until now the only means of dating the high river terraces, with *Equus ex germanicus samneniensis* from the 100 m (Early Pleistocene) terrace, *Mammuthus chosaricus* from the 45-55 m, and *Bison priscus longicornis* from the 31-33 m terrace, both assigned to the Middle Pleistocene (Tseitlin 1979:95).

During the last interglacial (Kazantsev), the Arctic Ocean transgression filled Agapa and Piasyna River basins, and the lower Yenisey Basin with marine deposits. Both molluscs and foraminifera, dominated by boreal and arctic-boreal species, indicate a warmer climate near the Arctic shore than at present (Velichko 1993:47).

Faunal remains from the early last glacial are mostly found in periglacial alluvia in the southern parts of Siberia. Among others, wooly rhinoceros *(Coelodonta antiquitatis)*, wolf *(Vulpes vulpes)*, horse *(Equus caballus)*, and mammoth *(Mammuthus primigenius)* have been taken from Terrace III in the Yenisey and Angara River valleys (Laukhin 1982). Remains of aurochs *(Bos primigenius)* and red deer *(Cervus elaphus)* have also been found in the Khamsara River gravels of a 40 m terrace (Yamskikh 1972:73). The early last glacial (Zyryansk Glacial) faunal record, including bank and water voles, and several lemming species (including collared lemming, Siberian brown lemming, and steppe lemming), has been documented from the Kuznetsk Basin close to the main study area. Large fauna, also found from the western slopes of the Altay Mountains, is represented by cave lion, cave hyena, Asiatic wild ass, argali sheep, Siberian ibex, yak and aurochs (Vereshchagin and Kuz'mina 1984:219). Other, widely distributed periglacial steppe mammals include wolf, red and arctic fox, brown bear, sable, wolverine, northern pika, Eurasian beaver, woolly mammoth, woolly rhinoceros, Siberian roe deer, red deer, moose, reindeer and steppe bison *(op.cit.)*.

Palaeontological evidence from the following Karginsk Non-Glacial Interval suggests a fluctuating, but generally moderate climate. A cold loessic malakofauna from the early Konoshelskoe (Mid-Wisconsinan) Stadial from the Ust'-Koba Site in the Upper Angara Valley, including *Pupilla muscorum, Succinea oblonga* and *Vallonia tenuilabris,* is typical of a cold tundra environment (Vasil'evskiy *et al.* 1988). The interstadial large fauna include broadly distributed Upper Pleistocene periglacial species such as mammoth, bison and wapiti.

A similar faunal composition characterizes the last glacial (Late Wisconsinan) stage associated with expansion of periglacial grasslands and tundra steppes in the boreal and subarctic zone of Siberia. The tundra-steppe mammal assemblage around the last glacial maximum included *Mammuthus primigenius, Coelodonta antiquitatis, Lepus sibiricus, Ochotona, Marmota, Spermophilus* sp., *Stenocranius, Alopex, Putorius Gulo gulo, Rangifer tarandus, Bison priscus,* and *Saiga tatarica. Cuon, Uncia* and *Capra* lived in the unglaciated parts of the Altay and Sayan Mountains (Baryshnikov and Markova 1992).

In general, the large, highly mobile and broadly dispersed terrestrial fauna is less palaeoclimatically diagnostic than the marine microfauna or the continental loessic malakofauna.

9.5. PALAEOLITHIC OCCUPATION

Early cultural evidence documented by stone and/or bone artifacts provides another source of proxy data which can be used to present a chronostratigraphic reconstruction of past environments and climate change. Until about five to ten years ago (depending on the particular geographical area), the earliest cultural record from Siberia, or north Asia in general, was assumed/dated to be at most 30-40 ka old (Tseitlin 1979; Praslov 1984), and a spatio-temporal correlation of palaeoenvironmental archaeological proxy data was limited. During the past several years, systematic field investigations have accelerated because of large-scale surface cover disturbances due to construction of dams on the largest Siberian rivers, and other extensive industrial field operations (*e.g.*, coal mines, road construction) several hundred early sites, some of them of a very high antiquity, have been located (Larichev *et al.* 1987). The number of the archaeological sites predating the last interglacial is rising and their potential for a local as well as regional reconstruction of past natural settings is increasing.

The earliest occupation of Siberia is speculated to have taken place as early as 800 ka but certainly during some of the later Middle Pleistocene interglacials prior to the last interglacial (Ranov 1990). Usually, the Tobolsk Interglacial (390-270 ka) is assumed to have been the most favorable time to colonized vast territories of Siberia, when the MAT was by 3-4°C higher than at present (Arkhipov 1991b:67). Early people are assumed to have migrated from Mongolia as far north as 60° N latitude repeatedly during warmer Pleistocene periods (Derevianko 1990:13). The main colonization of southern Siberia is visualized to have occurred during interglacials, whereas only local movements of Palaeolithic groups are envisaged during cold intervals, particularly the earlier glacial stages (Yamskikh 1992:91). Approximately at the same time (*i.e.*, during the late Middle Pleistocene) the Early Palaeolithic occupation is believed to have established in southern regions of the Russian Plains in the area north of the Black and Azov Seas (Velichko and Kurenkova 1990). Nevertheless, the initial settlement is likely to have occurred well before that time.

In more southern areas of Central Asia, a few Early Palaeolithic sites from loess deposits in Tadjikistan (the Kuldara Site) predate the Matuyama/Brunhes palaeomagnetic reversal (>0.78 Ma) (Ranov 1988; Lazarenko 1992). The most ancient claim from Eurasia, however, has been recently made from Yakutia from the middle Lena River basin at the Diring Yuriakh Site, situated on a 105 m bedrock platform. The original, problematic dating of the recorded "pebble" industry to 3.2-1.8 Ma based on the palaeomagnetic remanence would split the origin of humanity between tropical Africa and subarctic Northern Asia (Mochanov 1992). Lately, however, the site has been redated by the TL method to about 400 ka (M. Watters, personal communication 1994), still a very respectable age for a cultural assemblage for this part of the Old World.

In Western Siberia, the oldest stone tools are scattered on high river terraces, but in part also within the gravelly alluvia. According to their stratigraphic position and the associated malakofauna, an Early Pleistocene age has been suggested. Specifically, this concerns the Narkh-el-Kebir and Sitt Markho Sites located on 130 m Ob River terraces which are presumably as much as a million years-old (Deviatkin *et al* 1992). Other sites on 50-60 m terraces date to the Middle Pleistocene (*e.g.*, Latamna with lithic artifacts derived about 8 m below the surface and dated by thermoluminescence to 567-324 ka.). The earliest dates coincide with archaeological finds from adjacent parts of Central Asia referred to as the "Loess Palaeolithic" that started around 0.8 Ma (Lazarenko 1992).

In the Altay foothill area, the Ulalinka Site, discovered in 1960s in alluvial deposits, has been variously estimated to be 700-90 ka old, with the more recent dates (*ca.* 100-150 ka) being more likely (A.P. Derevianko and G.Ya. Baryshnikov 1993, personal communications). Other very early (Middle Pleistocene) sites are located at Filimoshki and Mokhovo near Krasnogorsk in the Kuznetsk Depression (Derevianko 1990). The Mokhovo Site is of particular interest, as this is deeply buried (-30 m) in the Bachatsk Crater in the Kuznetsk coalmine area (Figure 42). The age of the recorded stone tools, referred to the Acheulian Tradition, was originally assigned to the last interglacial because of presumably Upper Pleistocene fauna *(Equus caballus, Coelodonta antiquitatis, Bison priscus* and *Alces alces),* which, however, are also known from the previous period (Derevianko *et al.* 1990b). Recently, a thermoluminescence date of *ca.* 430 ka has been obtained from the overlying horizon, corroborating the chronostratigraphy of the associated Middle Pleistocene interglacial malakofauna assemblage from the artifact-bearing deposits (Nikolaev and Markin 1990:244; S.V. Markin, personal communication 1993).

In southern Siberia, four major archaeological regions have been recognized during the last several years. This include the Southern Minusinsk Basin and the Western Sayan Foothills, the Central and Northern Minusinsk Basin with the adjacent

Eastern Sayan and the Kuznetskiy Alatau Foothills, the Krasnoyarsk - Kanskaya forest-steppe, and the upper and middle Angara River valley with tributaries (Drozdov and Chekha 1992:92; Vasilevskiy *et al.* 1988). In the upper Yenisey region, the earliest cultural record is assumed to date at least to the second half of the Middle Pleistocene during the period of maximum glacial activity in the surrounding mountains (Chekha and Kol'tsova 1992). At that time, glaciers in the Western Sayan Mountains expanded down into the foothills to about 300-400 m altitude preceded by a down-slope migration of dark coniferous forest. In the Minusinsk Basin, protected by mountains from the west and east, and separated by high ranges from the frigid northern tundra, favorable (although periglacial) conditions are believed to have persisted in this area during that time. This assumption would corroborate the colluviated nature of the older Kurtak loess, indicative of a relatively humid and moderately cold climate with minor oscillations. Abundant fossils from the earlier periglacial alluvia give further support to this scenario.

The local Palaeolithic record has been assigned to four chronological stages. The earliest known cultural traces (Stage 1), from the Kamennyy Log and Verkhniy Kamen sites on the 60 m terrace, are dated to a pre-Samarovo (Mindel/Riss) interglacial (Chekha and Laukhin 1992:262). Culturally, the sites are represented by pebble tools comparable to the Soan Culture in northern India and Pakistan (Drozdov and Chekha 1992:93). The earliest Palaeolithic occupation of the middle Angara River region is documented at the Igetey Site chronostratigraphically assigned into the Samarov (Riss/Illinoian) Glaciation, *i.e.*, prior to 130 ka (Medvedev *et al.* 1990:141). The characteristic core and flake industries are analogous to those from the Kurtak area.

The second stage includes a "Mousterian-like" Tradition in the Minusinsk Basin (the Kamennyy Log and Berezhekovo Sites), found within the last interglacial soil and the early last glacial pedocomplex (the Sukholozhskiy PK). Expansion of the occupation habitat into the mountain area is likely to have occurred after the last interglacial during warm early last glacial interstadials when manifestations of the Mousterian industry, with the characteristic Levallois technology, appeared at sites in marginal areas. A temporal hiatus is assumed for the early last glacial maximum when cultural remains are absent. Mousterian sites are also known from Khakassia, the Sayan Foothills and the upper Angara River basin (Praslov 1984:314).

Broad parts of southern Siberia were reoccupied during the warmer mid-last glacial interstadials correlated with transitional Middle/Upper Palaeolithic industries (Stage 3), distinguished by archaic traits and a high percentage of archaic "Lower" and

"Middle" Palaeolithic tool types. In the Minusinsk Basin, this occupation has been recorded in the lower part of the Kurtak Pedocomplex (the Kashtanka and Kurtak IV Sites) with ^{14}C dates between 24,900 and 21,800 B.P. (Chekha and Laukhin 1992:262). The settlement area was gradually vacated towards the end of the Karginsk (Mid-Wisconsinan) Interstadial before total abandonment around the last glacial maximum (20-18 ka). In the Western Sayan Mountains, analogous Upper Palaeolithic sites are distributed on 23-25 m river terraces (Vasil'ev 1990); and in the Angara Valley that contains the well-known Ust'-Koba Site (Drozdov 1992) (Figure 42). It is mostly assumed that the main settlement in Siberia and the Russian Plain occurred during this time, *i.e.*, in the later part of the mid-last glacial interval (*ca.* 30,000-24,000 B.P.).

The last Pleistocene colonization stage includes the late Upper Palaeolithic stone industries best known from the upper Yenisey and Angara River regions. The most intensive concentration of occupation sites relates to a warm interstadial between 16-13 ka, prior to the final cold oscillation. In the Angara Valley, the famous Afontova Gora and Kokorevo cultures appeared.

Another north Asian Palaeolithic centre with possible early cultural manifestations is located in southern and western Mongolia. The Pliocene environments there were more favorable for early human settlement than in later periods because of rich and variable fauna, and warm mid-latitude and subtropical forests. Mosaic landscape with mountain forest-steppe, saline steppe, and alpine steppe provided a variable potential for early Palaeolithic occupation. During the Pleistocene, the most suitable conditions are believed to have existed during more humid early glacial stages with up to a 50-100% precipitation increase compared to interglacial and late glacial stages. The earliest documented Palaeolithic sites are usually deeply buried in alluvial fan deposits, but also dispersed on the high river terraces located in the former transitional forest / steppe zone (Derevianko *et al.* 1990a:108, 450).

More than twenty Early Palaeolithic localities have been recently recognized on the southern slopes of the Altay and Khangar Mountains, and the central Gobi Desert, with the earliest "pre-Acheulian" sites in the Narin-Gol River valley (Deviatkin and Malaeva 1990a). The Early Palaeolithic occupation is assumed to have concentrated in the foothills during interglacial periods, and shifted into more open plain settings during glacial periods. Climatic fluctuations and differential amounts of precipitation are believed to have been the principal factors behind this geographical distribution. Parkland forest-steppe in the foothill areas may have been the most favorable locations for occupation (Deviatkin and Malaeva 1990b:138). Geographically, the early sites are concentrated in southern Mongolia, whereas the later sites are found in the more hilly central and northern parts of the country as a result of climatic shifts towards continental climate, and more pronounced regional aridization (Derevianko *et al.* 1992b).

The Middle Palaeolithic settlement of southern Siberian mountains, correlated with the last interglacial and the first half of the last glacial *(ca.* 130-40 ka) is anthropologically characterized by an early form of *Homo sapiens* and by *Homo sapiens neanderthalensis.* The Neanderthal man is traditionally believed to have been the bear of the Mousterian Tradition characterized by the prepared core technology. A concentration of the Mousterian occupation is in the Russian Altay with numerous cave sites dated prior 40-50 ka (*e.g.*, the Denisova, Okladnikova, Strashkaya, Ust'-Kanskaya and Kaminnaya Caves) (Derevianko and Markin 1992; Shun'kov 1992). The identical geographical distribution of later sites and time-transgressive stone tool technologies suggest a possible local cultural continuity into the later Upper Palaeolithic sites, and a gradual transition of the Neanderthal man into fully modern *Homo sapiens sapiens,* as envisaged for Central Europe and the Near East. The cultural record from this period (the Karginsk / Mid-Wisconsinan Interstadial) was presumably obliterated by erosion of glacial meltwater from the Altay Mountains stored in large proglacial lakes and released as catastrophic floods, which are thought to have destroyed most of the older open-air sites on the river terraces in the foothills (Baryshnikov 1990:58).

In terms of general cultural evolution, two separate Upper Palaeolithic traditions in southern Siberia have been suggested -- a more progressive tradition with a blade technology derived from the local Middle Palaeolithic Levallois (*i.e.*, the prepared core) technique; and a more conservative tradition, largely retaining the Early Palaeolithic core and flake (*i.e.*, the pebble tool) stone flaking (Drozdov 1992:50-51; Larichev *et al.* 1990).

A successful adaptation of the Late Palaeolithic people to Siberian periglacial environments is explicitly evidenced by open-air sites reported from the Arctic. Some sites were presumably occupied during the last glacial maximum on the Novosibirskiye Ostrova (the New Siberian Islands) and on the Wrangel Island, when the continental shelf expanded and the present islands were connected with mainland Asia sometime between 20-18 ka (Pitulko 1992). Late Palaeolithic arctic sites are also known from the

European part of Russia. The Byzovaya Site in the middle Pechora River, dated by radiocarbon to 28-25 ka is located 175 km south of the Polar Circle (Ivanova *et al.* 1989). Palaeolithic settlement also reached other northern areas including the Oka and Lesna River basins (Velichko *et al.* 1990:81). Nevertheless, the northernmost parts of Eurasia were likely occupied at least since the mid-last glacial interstadials, as documented by a possible cultural evidence from the west-central Chukotka Peninsula (Laukhin 1990; Laukhin and Drozdov 1991a, 1991b), and the Nepa Site in the upper Tunguska River region (Zadonin and Semin 1990). However, the archaeological record from the Diring-Yuriakh Site in the middle Lena River basin (Figure 42) shows that most of the Siberian territory was colonized (at least temporarily during interglacial periods) much earlier, sometime during the Middle Pleistocene or even before that time (Mochanov 1992). On the other hand, vast glacial lakes in the interior western Siberia during the late last glacial stage inhibited the Upper Palaeolithic settlement of this area between 22,000 and 17,000 years B.P. (Arkhipov and Volkov 1980).

Conclusively, the spatial distribution of early cultural remains documents climatic instability over large parts of Siberia during the Pleistocene. Particular geographical locations of archaeological sites indicate that environmental conditions during the early periods were generally more favorable for Palaeolithic occupation than during later periods. The northern subarctic and the southern desert areas had a reduced biological potential than at present because of low annual temperatures (*e.g.*, Yakutia), extreme seasonality contrasts (*e.g.*, the Arctic coast), and/or low precipitation values (*e.g.*, southern Mongolia). Consequently, some regions were probably more densely and/or frequently occupied during the Pleistocene than they are now. Therefore, even if relatively fragmentary, the cultural remains may prove to be a very significant indicator of past climatic change in a particular area.

The evolutional process of the Palaeolithic colonization of Siberia is still poorly known, although every year archaeological investigations supply new, sometime rather unexpected and surprizing evidence about the early human prehistory of the territory. The traditional views, assuming a very late (Late Pleistocene) inhabitation of most of northern Eurasia, have been definitely challenged. Current data show that people were present long before that time. Several recent archaeological discoveries disprove the long-held assumption of a late penetration by Upper Palaeolithic people into the middle and high latitudes of northern Asia. Instead, glacial-interglacial and stadial/interstadial cycles may have regulated periodic movement of early people northwards in response to climatic fluctuations, predetermining the inhabitability of large geographical areas. Nevertheless, during glacial maxima, most of Siberia seems to have been vacated, especially during the earlier periods, because of the expansion of continental glaciers, and presence of harsh and inhospitable environments in periglacial areas. Gradual adaptation to cold natural habitats accelerated during the Late Pleistocene in connection with the cultural and biological evolutionary adjustment, enabling people to establish permanently in the vast Siberian territory.

9.6. SUMMARY

The geological and palaeoecological evidence from southern Siberia and neighboring regions shows patterned climatic changes during the Quaternary period leading to formation of the present environment. Trends toward a strongly continental climate characterized by increased aridity and high seasonal fluctuations can be followed in the geological record since the late Pliocene. The subsequent development during the Quaternary is evidenced by zonal shifts in vegetation distribution, with expansion of a taiga forest northward during interglacial periods and warmer interstadials, succeeded by subarctic periglacial forest-steppe, and tundra-steppe during cold glacial intervals, when the tree cover periodically became much reduced and spatially limited to refugia in the southernmost portions of Siberia. Complex ecological changes are evident from both pollen and fossil faunal remains recorded from stratified alluvial and loess deposits. However, the key information has come from the loess - palaeosol record, recovered mainly in areas of active archaeological investigations.

In western and southern Siberia, loess with intercalated palaeosol horizons consistently shows a periodic, patterned climatic cycle of loess deposition, followed by soil formation, with superimposed cryogenic deformations preceding subsequent loess accumulations. This pattern indicates a broad uniformity of climatic change in extraglacial areas of Siberia. The pre-Late Pleistocene history is not as well known because of the lack of long-term correlative regional sequences. The most complete picture emerges for the Quaternary environments of the last glacial - interglacial cycle. During the Kazantsev Interglacial, broad areas of southern Siberia were covered by coniferous or mixed forests, with forest-steppe distributed in lower elevations and near the river valleys, and steppe in the southern interior basins. On the Angara River Plateau and in the upper Yenisey Valley, a variety of soil types shows a mosaic character of vegetation in respect to particular topographical locations. Coniferous northern-taiga forests associated with Brunisolic and Podzolic soils expanded during the first part of the last interglacial. Later, pronounced continentality of climate is reflected by the occurrence of Chernozemic steppeland soils, followed by dark brown soils formed under mixed southern-taiga forests, indicating shifts to cold and humid conditions. During the early last glacial stage (the Zyryansk Glaciation), periglacial forest-steppe expanded to cover most of the Angara Plateau and the upper Yenisey Basin. Following a period of intensive loess deposition, a renewed pulsation of pedogenic processes during the Karginsk Non-Glacial Interval led to formation of zonal soils as a result of warm climatic oscillations. The interstadial mosaic landscape with parkland vegetation was transformed into periglacial tundra during the late last glacial stage (the Sartan / Late Wisconsinan Glaciation), with warm interstadial oscillations initiating formation of zonal soil horizons during the final stage. This palaeoclimatic evolution pathway can be documented over much of southern Siberia.

Abundant Early Palaeolithic artifacts made on local stone raw materials provide explicit evidence of climatically favorable conditions for human occupation of the Minusinsk Basin and other parts of southern Siberia prior to the last interglacial. The variability of some lithic assemblages, the differential degree of patination and aeolian abrasion of individual artifacts indicate repeated occupation of the vast territory during several stages. The fossil fauna associated with the cultural records in the main river valleys and depressions suggests that early people were either well-adapted to cold periglacial environments, or the climatic conditions were not as harsh as usually assumed for the Middle Pleistocene middle / higher latitudes. Stages of climatic deterioration and onset of full glacial conditions are monitored by the absence of any archaeological finds, what is best documented in the Kurtak area during the early and late last glacial maxima. In sum, palaeontological and Palaeolithic records, though less climatostratigraphically specific, provide a further source of palaeoenvironmental proxy data complementing the biotic and palaeopedological evidence.

10. THE TERTIARY - QUATERNARY CLIMATIC EVOLUTION IN NORTH-CENTRAL ASIA

The following is a review of long-term climatic change during the late Tertiary and early Quaternary periods in a larger territory of central Asia, with a specific focus on correlation with the southern Siberia record. Although significant progress has been done in understanding palaeoclimatic shifts in the Russian (formally Soviet) part of the Northern Hemisphere, there is disagreement among Russian scholars about establishing a general climato-stratigraphic correlation during the Quaternary. One view assumes increased precipitation during glacial and less during interglacial periods (Nikonov *et al.* 1989; Derevianko *et al.* 1990a). Another view suggests that inter-glacials were relatively humid, and pronounced aridity existed only in orographically isolated areas (Velichko and Lebedeva 1974). Generally, terrestrial records from intra-continental depressions are of utmost significance.

Climatic evolution in northern Asia is characterized by gradual aridization beginning in the Late Miocene. In western Siberia, this trend is best documented in the Pavlodar Formation dated by palaeomagnetism between 5.9 Ma and 5.5 Ma with a palaeoclimatic carbonate record indicating an increase in MAT by about 16°C in comparison with present values. The gradual trend towards a marked continentality of climate is also indicated by accumulation of carbonate crusts in western Mongolia, and by expansion of steppes in the Transbaykal area (Zykin 1991). Significant cooling during the Early Pliocene (5.4-3.2 Ma) is documented by pollen records suggesting average temperature values of -24°C in January and +18/+20°C in July, *i.e.*, a colder climate than in western Siberia at the present time. A similar trend is indicated by malacofauna with a 45% increase of Arctic species (Zykin *et al.* 1991). Progressive cooling continued throughout the Late Pliocene (3.2-2.5 Ma), marking the onset of full glacial conditions in the Northern Hemisphere possibly as a result of the tectonic uplift of central Asia (Derbyshire 1991), causing, among others, a change of the climatic circulation pattern and disruption of faunal exchanges between Siberia and Sinoindia.

The Plio-Pleistocene chronology of stratigraphic sequences is based on the Siberian palaeomagnetic record. The earliest glaciations are dated to about 5.4 Ma, 3.2 Ma, 2.5 Ma and 1,8 Ma ago. The most intensive cooling is thought to have occurred about 3.2 Ma ago, which significantly effected the further climatic development and eventual overall biotic evolution in northern Asia (Zykin 1991:14). In the Northern Siberian Lowlands, permafrost started to form during the Late Pliocene around 4.5-4

Ma, and a tundra vegetation spread in northern subpolar areas (Laukhin 1993). In the southern part of Western Siberia, significant cooling of the Late Pliocene climate is documented by malacofauna from the Seletinsk Period (3.2 Ma), and the Irtysh Period (2.5 Ma) (Zykin *et al.* 1990). During the Early Pleistocene, Siberia experienced pronounced climatic fluctuations with colder as well as warmer periods than at the present time (Laukhin *et al.* 1990b). The maximum Pleistocene glaciation is correlated with the Samarov / Illinoian Glacial Period.

Intensification of aeolian activity during the Early Pleistocene contributed to accumulation of loess in parts of southern Siberia from local as well as more extensive southern deflation surfaces in Mongolia and Central Asia by prevailing (south)eastern winds. In the western areas, in the upper Ob River basin, loess deposits attain over 100 m (Volkov and Zykina 1982; Zykina 1986); in the upper Yenisey Valley a maximum of 50 m (Drozdov *et al.* 1990a), and a little less in the Angara Valley (Medvedev *et al.* 1990). Farther east, in the Irkutsk area, the thickness of the loess cover decreases to a maximum of 20-30 m. There, the silty sediment is derived from various (alluvial, proluvial and lacustrine) sources. In addition to aeolian transport, a "lithogenetic" *in situ* loess formation process has been suggested as a result of cryogenic parent material fragmentation in hyperarid environments (Ryaschenko 1990).

In Mongolia, this climatic development leading to establishment of more continental conditions is correlated with a period of intensified aeolian activity excavating large deflation basins in the Gobi Desert. Up to 300 m of loess interspersed with extensive palaeosol sequences was formed on the neighboring north-central China Loess Plateau above the Late Pliocene red soils. The most complete climatostratigraphic record at Baoji includes the longest series of 37 palaeosols corresponding to a time span of about 2.5 Ma, *i.e.,* covering the whole Pleistocene (Rutter *et al.* 1991b). Recognition and study of the pedogenic phenomena associated with biological activity, and rubification and decalcification of parent material provides basic proxy data for reconstruction of the local North Chinese palaeoclimatic history (Guo *et al.* 1991). The best developed Chinese palaeosols characterized, by thick clay coatings and strongly developed Bt-horizons, date to the second half of the Early Pleistocene (1.2-0.85 Ma) (Ding *et at.* 1991). Weaker pedogenesis and rubification during the later periods may be explained by a continuing aeolian-dust supply during the pedogenic process, and/or because of a colder and drier climate. Studies of variation and composition of palaeosols from the Late Pleistocene Malan Loess suggest that the past environment was not much different than the present one, with steppe areas in the northwest and forests in the southeast (Wen *et al.* 1987). A progressive climatic transition from the last interglacial with arboreal vegetation trending towards colder and drier conditions is evidenced by accumulation of thick loess deposits. The parent silty material was simultaneously pedogenically altered in the process of sedimentation to form thick soil horizons. Analogous trends have been observed in Europe, as well as North America (Bronger and Heinkele 1990), as well as in the study area in southern Siberia.

In Central (a former Soviet) Asia, Late Pliocene glaciations have been recorded in western Kazakhstan in the Dzhungarskiy Alatau, and in the Tian-Shan and Pamir Mountains. Progressive Late Tertiary and Quaternary orogenesis caused intensive erosion of pre-existing deposits in high mountains, and aeolian, colluvial, fluvial and lacustrine sedimentation in river valleys and lowland basins (*e.g.*, the Fergana and Chuyskaya Basins). Loess accumulation occurred along the foothills, most intensively in the Southern Tadjik Depression and the Tashkent region (Velichko 1993:93). The initial accumulation of loess in Tadjikistan occurred during the Late Pliocene (Dodonov 1986), and the total thickness eventually reached up to 200 m (Lazarenko 1984). Some early loessic deposits are now located as high as 2.5 km because of the orogenic uplift. Most of those loesses were later colluviated and redeposited into valleys, as in other parts of Central Asia with a higher regional topography (Lange 1990). Rather intensive pedogenic processes are documented in loess deposits in western Turkmenistan, where about 80% of the total thickness of the 40-45 m sections is formed by more than 20 well-developed palaeosols with up to 2 m thick horizons. Presumably 8-9 magnetic excursions have been found in the loess - palaeosol record (Lazarenko *et al.* 1990:119). During the Late Pleistocene, loess accumulation is believed to have continued during both cold and humid, as well as warm and dry stages, with sedimentation maxima about 75 ka and 120 ka, respectively (Deviatkin 1990:176). In the Kazakh and other Central Asian steppes, maximum aeolian activity now occurs during summer and early fall with an average loess sedimentation rate of 0.2 mm per year (Lomov and Finaev 1992:166).

The increasing climatic amplitude since the Pliocene throughout the Pleistocene has also been documented in the geological record from the European Russian Plains (Velichko 1990; 1993:11). A marked change in microfaunal assemblages with a transition to cold-adapted species is best documented from the Early and Middle

Pleistocene loess deposits in the upper Don River area (Iosifova 1990). It has been suggested that loess was accumulating on the Russian Plain throughout the Pleistocene, with decreased sedimentation during warmer intervals. Loess accumulated in spatially most bio-geographical zones, including steppe, forest-steppe, tundra-steppe and tundra (Bolikhovskaya 1990). One of the most complete loess - palaeosol records with more than 100 m thick sections is located in the southern part of the plain, where 10-12 m of loess accumulated during the last (Vaidal / Late Wisconsinan) glacial stage (Velichko *et al.* 1985:73). Very thick (up to 140 m) loess deposits, mostly colluvial in origin, are also present in the Transkaukasia, where the Matuyama / Brunhes palaeomagnetic boundary has been identified in the basal part (Virina *et al.* 1990).

As evident from the above review, loess-palaesol records from Eurasia show a consistent pattern of climatic change with a gradual cooling trend from the late Tertiary throughout the Quaternary Periods. The Pleistocene palaeosols in most loess sections show a similar degree of pedogenic processes, allowing a global correlation of major interglacial horizons (Bronger and Heinkele 1989). For example, both in Europe and Eastern Asia, it has been concluded that the penultimate (Holstein) interglacial was substantially warmer and more humid than now, while the last interglacial (Eem) was comparable to or only slightly warmer than the present Holocene interglacial. In Central Europe, as much as eight major depositional loess cycles, separated by interglacial periods with temperate forests, were found within the Brunhes Chron (Fink and Kukla 1977). A similar record is documented in the Chinese loess (Rutter *et al.* 1991c).

A broader continental palaeoclimatic correlation of the Siberian loess record with other Eurasian loess provinces has not been attempted because of a lack of comparable, high-resolution data. Nevertheless, some general analogies exist. A loess-palaeosol profile from southern Poland, for example, shows a comparable pedostratigraphic succession characterized by a decreasing pedogenesis of individual palaeosols, with the Blake Excursion located in a 0.5 m loess unit above an interglacial soil that is followed by a series of (late last interglacial) mostly weakly-developed soil horizons (Maruszczak *et al.* 1986:figure 2). A Late Pleistocene loess/palaeosol record, also sharing some similarities with the Kurtak sequence (Figure 32), is from Nußloch near Heidelberg, Germany, with the Mosbach interglacial soil followed by a series of interstadial soils of the early and middle stages of the Last Glacial Period, and a series of initial (Cryosolic), partly colluviated soils from the late last glacial (Late Würmian) stage (Bente and Löscher 1987; Pécsi 1992:fig. 11). This fact may suggest that climatic patterns were broadly analogous throughout mid-latitude Eurasia despite the great distances. The results from the Minusinsk Basin further confirm the assumption that high-resolution records may be anticipated in a continental transect running across the Asian continent, particularly in the areas north and south of the Himalayas. Also, because of its particular location close to the geographical centre of Asia, the palaeoclimatic (palaeopedological and magnetic susceptibility) record from Kurtak shows a higher resolutions for the last glacial-interglacial cycle than is observed in the thickest Chinese loess section (Figure 27).

Besides the terrestrial loess-palaeosol sequences, a correlation of the deep-sea oxygen isotope record with the southern Siberian loess has been attempted. The magnetic susceptibility curve from Kurtak Section 29 exhibits some general similarities with the orbitally-tuned stacked oxygen isotope SPECMAP curve extending over a time span of about 300 ka (Pisias *et al.* 1984; Martinson *et al.* 1987) (Figure 28). The Siberian loess-palaeosol sequence largely corroborates the Late Quaternary climatic curve recognized in the Pacific deep sea cores (Shackleton and Opdyke 1973; Hays, Imbrie and Shackleton 1976), but again provides a higher resolution signal through the Late Pleistocene Period than documented in the marine records (Figures 29-30).

Specifically, the Oxygen Isotope Stage 7 is tentatively correlated with the series of Brunisolic soils with a Chernozem of the Penultimate (late Middle Pleistocene) Interglacial (Pedocomplex 3) in the basal part of Kurtak Section 29. The following cold interval, ending the period of environmental stability (Stage 6), experienced significant short-term fluctuations with niveo-aeolian (colluviated) loess deposition between two cold and arid phases of loess sedimentation. Rapid climatic amelioration after the termination of the first climatic cycle led to landscape stabilization accompanied by intensified soil formation processes, disrupted by short intervals of loess sedimentation and periglacial surface deformation. Oxygen Isotope Stage 5 of the last interglacial *(sensu lato)* is correlated with a series of palaeosols exhibiting a decreasing degree of pedogenesis (Pedocomplex 2) correlating with three well-developed magnetic susceptibility minima (tentatively identified as $\partial^{18}O$ Substages 5e, 5c, 5a). A trend towards more pronounced continentality around the mid-last interglacial *(sensu lato)* is manifested by formation of three Chernozemic soils severely distorted by deep frost wedge casts filled with shattered silt grains indicating extremely cold and/or fluctuating climatic conditions.

A prolonged interval of aeolian loess sedimentation relates to the progressive cooling of the early stage of the last glacial (Stage 4) culminating in several very cold and arid oscillations represented by magnetic susceptibility maxima. Following a gradual climatic amelioration with the periodic appearance of Gleyed Regosolic and later Brunisolic horizons, a new interval of low magnetic susceptibility correlates with the mid-last glacial interstadials documented by a series of palaeosols (Pedocomplex 1), separated by episodes of minor loess deposition and surface cryoturbation. During the mid-last glacial optimum (Stage 3), evidenced by a truncated Chernozem / a mature Brunisol, the climate in the area was likely as warm or even warmer, but more humid than now. A radiocarbon date of 30 400±700 B.P. (AECV-1938c) on wood was obtained from the superimposed Cryosol. Gradual cooling and increased humidity is evidenced by a series of Gleyed Brunisols suggesting that the transition towards the final last glacial stage (Stage 2) was less dramatic than during the previous early last glacial interval. A marked drop of temperatures associated with loess deposition during the last glacial maximum is reflected by the second magnetic susceptibility peak. Progressive warming towards the present interglacial characterizes the final stage of the last climatic cycle.

Altogether, the Kurtak loess-palaeosol stratigraphic sequence shows a comparable, high resolution palaeoclimatic record for the late Quaternary cycles as found in the marine oxygen isotope records. Nevertheless, the last interglacial interval in the Siberian section is characterized by more complex climatic patterning with several warm and cold intervals not found in the deep sea cores (Figure 30). Also, the climatic development during the following glacial does not follow the gradual cooling trend evidenced by the oxygen isotope curve. Instead, the Kurtak record shows a marked period of climatic amelioration during the mid-last glacial (Mid-Wisconsinan) stage analogous to the last interglacial in the studied Section 29, but of lesser amplitude. On the other hand, the sudden onset of the last interglacial, and, on the contrary, a progressive, short-term transition from the last interglacial into the early last glacial, and the rapid shifts between the warm and cold intervals within the last interglacial (Oxygen Isotope Stage 5; 130-74 ka) are equally recorded in the Siberian loess-palaeosol sequences, in the deep-sea cores, as well as in marine and glacigenic deposits in Greenland (Funder *et al.* 1991:107). This consistency corroborates the general assumption of a global nature of climatic change during the Pleistocene. The high-resolution magnetic susceptibility signals from the Kurtak loess suggest that they may be indicative not only for local, but also broader regional palaeoclimatic evolution in southern Siberia, and the general trend may be possibly extrapolated to other continental areas of North and Central Asia.

Chronostratigraphic palaeoclimatic discrepancies have been pointed out between the deep sea core stratotype in tropical Pacific Ocean (Shackleton and Opdyke 1973) and continental (loess) records. For example, in the most complete loess-palaeosol section at Baoji on the north-central China Loess Plateau some of the best developed soils (suggesting warmer climatic conditions than today) do not corroborate particular oxygen isotope stages indicative of cold climatic fluctuations (*e.g.*, Stage 11) (Rutter *et al.* 1991c:988 [illegible] [illegible]arly, it has been pointed out that in the Carpathian Basin the information de[illegible] [illegible]rticular geological setting is more detailed than the deep-sea record (Bror[illegible] [illegible]ele 1989). In view of these facts, it is apparent that specific local conc[illegible] [illegible]ed in various parts of the world, and general validity of climate-indicative patt[illegible] found in the marine proxy records onto the continents may therefore be simplistic. Instead, it may be suggested that independent palaeoclimatic records should be used for terrestrial and oceanic environments, as each may reflect a differing amplitude and scale of past atmospheric changes over a particular region of the globe. Because of the nature of palaeoclimatic proxy data, loess-palaeosol sequences are believed to provide a better background for high-resolution, continental palaeoclimate histories.

The current Holocene climate in the Eurasian steppes is controlled by the Atlantic, Arctic and Pacific high-pressure centres (Khotinskiy 1984). The regional humidity regime is considered to be the main environmental factor. In southern Siberia and adjacent areas, two arid/semiarid steppe zones were formed. The first, including northern Kazakhstan, the Gorno Altay, Khakassia and the Transbaykal area, lies within the ephemeral reach of the waning Atlantic cyclone; the second zone in the southeast including the Tuva Depression, the Gobi and the Takla-Makan Deserts, the Ordos Mountains and western Heilongjiana (old Manchuria) receives air masses from the Pacific Ocean and northern Siberia. In general, the actual position and strength of the Icelandic Minimum and the Azov Maximum determines the rate and location of climatic changes over northern Eurasia (Maloletko 1992:176). The recent continentality of climate in the central part of north Asia is controlled by the Siberian Anticyclone locally reinforced by geomorphic changes related to the early Late Pleistocene neotectonics in the Transbaykal area (Yerdikhinskiy 1990).

By extrapolation from the present atmospheric circulation pattern, it has been suggested that during interstadials the Icelandic Minimum moved more towards the north, causing increased expansion of warm and dry air masses eastwards across Europe and Western Siberia into Central Siberia (Maloletko 1992:177). In Eastern Siberia, however, weakening of the Siberian Anticyclone during winters presumably contributed to increased snow precipitation, and summer rainfall brought by monsoons from the Pacific Ocean because of northward retreat of the Polar Front. During stadials, the eastern cyclones shifted southward, causing humid air-masses to flow over the southern arid and semiarid continental steppes, thus enhancing their ecological productivity. In Siberia, the southeastern monsoon system was strengthened at the expense of the Polar Front, bringing more precipitation and decreasing the annual seasonality extremes, particularly marked during mid-interglacial and interstadial stages *(op.cit.)*.

Compared to the present values, during the last interglacial optimum, an annual temperature rise of 2-5°C in the northern continental interiors has been assumed to be due to a 12-13% increase in summer solar radiation, and a slight reduction of winter radiation, when perihelion occurred in the northern summer around 126 ka (Kutzbach *et al.* 1991:223). In northern Eurasia, however, no major climatic differences are believed to have existed between the last interglacial and the present (Holocene) interglacial, with both stages characterized by similar patterns of climatic successions (Velichko *et al.* 1991:191). During the climatic optimum, increased summer and winter temperatures as well as precipitation has been documented in eastern Europe and Siberia as a result of expansion of the Atlantic stream across the Russian Plain into Siberia (Velichko *et al.* 1985:71). Arctic air masses were confined to more northerly locations and the Icelandic minimum became more significant. The Siberian Anticyclonic circulation was reduced considerably, whereas the southeast Asian monsoonal activity became more prominent (Velichko *et al.* 1991:191). At present, the north Asian anticyclone occupies vast areas and merges with the central Asian high-pressure centre. The maximum positive values of winter and summer temperature deviations during the last interglacial presumably did not coincide in time with shorter winter optima following the summer insolation maxima because of the delayed heat transfer mechanism from oceans to continents (*op.cit.*:198). Increased heat supply caused more prominent latitudinal zonation. The highest contrasts have been documented in the coastal Arctic regions because of the influx of warm Atlantic waters, with an average 12°C increase of winter temperatures; on the Taimyr Peninsula summer temperature increased by 6-8°C (Velichko 1984; 1993:206). In the southern part of central Siberia north of the Minusinsk Basin, a high-pressure zone became established as one of the regional autonomous centres (Velichko 1984:fig. 25-10), causing more pronounced continentality with increased summer temperatures, but colder winters.

In sum, in respect to palaeoenvironmental data from the Northern Minusinsk Basin correlated with palaeoclimatic proxy records from other areas, and extrapolating the recognized long-term atmospheric trend, the future climate in this part of north-central Asia can be predicted to become over the following millennia gradually cooler, more humid, and eventually less continental with more balanced seasonal temperature fluctuations than at present.

CONCLUSION

Studies on climate and climatic change in southern Siberia during the Pleistocene Period are still in the initial stages compared with other, more intensively researched areas in the Northern Hemisphere. Despite this fact, the loess - palaeosol record from Kurtak provides evidence of pronounced, short-term climatic fluctuations in the Minusinsk basin during the Late Quaternary. A dominantly subaerial process of loess deposition is indicated by grain-morphology, sedimentological and structural characteristics of the silty material. Fresh surface of the silt fraction, limited abrasion of individual grains, and a low degree of mineral weathering combine to indicate a local provenance for the loess, originally glacio-fluvially derived from the eroded igneous and metamorphic bedrock of the surrounding Kuznetskiy Alatau and the Eastern Sayan Mountains.

The high-resolution palaeoclimatic record is documented by loess-palaeosol magnetic susceptibility curve suggesting a pattern of increasing amplitude during the Late Pleistocene (130/127-10 ka). A marked and patterned fluctuation of the susceptibility signal, correlated with the $\partial^{18}O$ deep-sea stages, is observed throughout the stratigraphic record, with a consistent trend of magnetic susceptibility maxima corresponding with intervals of the most intense loess deposition, and the minima with the occurrence of the most advanced palaeosol horizons. The recorded magnetic susceptibility of the Siberian loess is opposite to what is found in Chinese and European loess sequences. In the Kurtak loess, Chernozems and mature Brunisols are characterized by the lowest susceptibility values (90-150 x 10^{-5} S.I.). Increased susceptibility is found in weakly developed Brunisols and anologous pedosediments (*i.e.*, colluviated soils) (150-250 x 10^{-5} S.I.), Gleyed Regosols (190-390 x 10^{-5} S.I.), and the second (colluviated) loess facies (160-410 x 10^{-2} S.I.). Finally, the aeolian loess registers the highest magnetic susceptibility (300-590 x 10^{-2} S.I.), with peaks (500-580 x 10^{-2} S.I.) coinciding with periods of the most intensive loess sedimentation during glacial maxima. The magnetic susceptibility pattern inversely correlates with the ∂^{18} O deep sea records, thus suggesting a global significance of the Kurtak sequence.

The greater amount of unweathered ferromagnetic minerals (magnetite, titano-magnetite and ilmenite) in the loess relative to the palaeosols seems to be the controlling factor. The marked decrease of ferromagnetic minerals in weathered loess horizons is clearly proportional to the degree of pedogenic weathering. Advanced oxidation and reduction processes of iron-rich minerals in the loessic parent material and subsequent transformation and alteration of less stable ferric components (*e.g.*, haematite to goethite) is considered to be the main controlling mechanism for reduction of the magnetic susceptibility signal. Increased aeolian deposition capacity and introduction of larger ferromagnetic minerals during the cold intervals may also have been involved.

In terms of soil formation processes, gradual pedogenic alteration is particularly well evidenced in the palaeosol series. The incipient alteration (leaching) of the calcareous parent material is evidenced by mottles as a result of gleying, indicative of cold and humid environments, and a seasonally waterlogged setting. Progressive leaching of calcium carbonates from the loessic substratum into lower parts of the pedogenic profile accompanied by accumulation of decomposed organic matter in A horizons, and brunification of B horizons reflects a gradual increase of summer temperatures and surface stability that contributed to a prolonged weathering process and the eventual formation of Brunisolic and Chernozemic soils.

From a palaeoecological perspective, the onset of pedogenesis occurred during cold intervals of decreased loess sedimentation in arctic tundra characterized by periodically saturated ground and intensified gleyzation processes. Increase of annual temperatures and better surface drainage contributed to formation of Gleyed Brunisols and eventually well-developed Brunisols, and the establishment of periglacial steppe, and boreal forest, respectively. Further rise of summer temperatures caused a shift towards a semiarid, strongly continental climate, that led to expansion of grasslands and retreat of forests into higher, more humid mountain locations. The rate of soil formation processes relatively accelerated, but because of the shortened annual growth cycle due to prolonged and very cold winters, the characteristic Chernozemic soils have a less developed structure than mature Brunisols that formed at the beginning of the last two interglacials under more humid and less continental conditions in a parkland / forest setting. The overlying series of Dark Brown Chernozemic soils is indicative of drier and more continental conditions. Deep frost/ice wedge casts on top of the Chernozems suggest cold (~10^3 yr) episodes during the interglacial cycles.

Relatively mild, moist climate with a summer insolation deficit is believed to have prevailed during the interstadials prior to the initial last glacial (Early Wisconsinan) maximum. Periglacial deformations during the mid-last glacial (Mid-Wisconsinan) interstadials were limited to small ice-wedge casts and cryoturbation, suggesting that the climate was less continental, with more stable humidity, short melting seasons, and poor drainage under continuous permafrost. Periodic ground saturation is evident by gleyed and colluviated Brunisols after 30 ka until the onset of the late last glacial (Late

Wisconsinan) stage about 21 ka. A similar climatic pattern during the last glacial-interglacial cycle exhibited by both geological and biotic data has been recognized in other parts of southern Siberia, suggesting that there was a relative environmental uniformity in the Ob, Yenisey and Angara River valleys north of the Altay-Sayan Mountain system.

Quaternary studies in Siberia have witnessed major developments in understanding Pleistocene climatic evolution. Thick loess sections in the main river valleys on the Ob Plateau and in the upper Yenisey area present the best potential for further palaeoclimatic studies. The marked progress during recent years has been accelerated by intensified palaeoenvironmental, and related archaeological studies on local as well as regional scales, and by correlation of the key geological records by application of various field techniques and analytical methods. Complementary palaeopedological, palaeobotanical, palaeontological and archaeological data provide a more exact and complete body of information on the nature, periodicity, duration, and amplitude of climatic change and the overall character of natural environments than a geological record alone. An interdisciplinary approach is therefore prerequisite for producing a more accurate reconstruction of local, regional as well as territorial palaeoclimatic histories, and for assessing the climatic impact on biotic communities and past atmospheric circulation patterns. It also has a direct bearing on comprehending the early human prehistory.

In conclusion, it is suggested that a comprehensive study of loess-palaeosol sequences in extensive stratified sections, combined with comparison and correlation of well-described stratigraphic profiles elsewhere may provide an informative data base about past atmospheric changes in northern continental Asia in terms of frequency, periodicity and intensity of fluctuating climatic events. Future magnetic susceptibility studies on loess and palaeosols coupled with other palaeoenvironmental (biotic and cultural) proxy data from broader geographical areas may provide a more complete picture of the Pleistocene climatic evolution in the southern Siberian sector of the Northern Hemisphere.

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APPENDIX A

KURTAK SECTION 29: PEDOSTRATIGRAPHIC DESCRIPTION

Sk$_1$ *Orthic Black Chernozem*

0-25 cm **Ah**: Very dark gray (10YR 3/1d), black (10YR 2/1m); silty clay; strong, fine to medium granular; slightly sticky (w), friable (m), slightly hard (d); abundant, fine and medium inped/exped roots; common, vertical, fine and medium pores; krotovinas present; gradual, irregular boundary; 20-25 cm thick.

25-50 cm **Bm**: Very pale brown (10YR 7/4d), light yellowish brown (10YR 6/4m); silt loam; few fine and medium, faint, dark yellowish brown (10YR 4/6) mottles[1]; moderate, medium subangular blocky; slightly sticky (w), friable (m), slightly hard to soft (d); common, fine inped/exped roots; few, vertical, fine pores; krotovinas present; clear, wavy boundary; 20-30 cm thick.

50-65 cm **BCca**: White (10YR 8/2d), light gray (10YR 7/2m); silt; massive; few, fine roots; very few, fine vertical pores; strongly effervescent; clear, smooth boundary; 10-15 cm thick.

65-80 cm **Ck**: Very pale brown (10YR 7/3d), pale brown (10YR 6/3m) to light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m) in the basal part; silt; massive; non sticky (w), very friable (m), slightly hard (d); many, micro, vesicular pores; gradual, smooth boundary; 15-25 cm thick[2]. *Loess.*

Sk$_2$ *Orthic Eutric Brunisol*

80-100 cm **Bmk**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; weak to moderate, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); some charcoal present; moderately effervescent; gradual, smooth boundary; 20 cm thick.

100-110 cm **BCk**: Yellowish brown (10YR 5/8d), dark yellowish brown (10YR 4/6m); silt loam; few, small, prominent, yellowish red (5YR 5/8m) mottles; very weak, medium, subangular blocky; non sticky (w), friable (m), slightly hard (d); strongly effervescent; clear, smooth boundary; 10-15 cm thick.

110-370 cm **Ck**: Light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m); silt; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 290 cm thick. The unit consists of pale brown (10YR 6/3d), yellowish brown (10YR 5/6m) horizontally bedded layers 1-3 cm thick in the lower part of the unit and 1-5 cm in the upper part; 260 cm thick. *Loess.*

Sk$_3$ *Gleyed Regosol*

370-390 cm **Ckg**: Light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m); silt; common, medium, prominent, yellowish red (5YR 4/6m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 20 cm thick.

390-410 cm **Ckgj**: Light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m); silt; few, medium, prominent, yellowish red (5YR 4/6m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 20 cm thick. *Loess.*

Sk$_4$ *Orthic Eutric Brunisol*

410-460 cm **Bmk**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; some charcoal present; clear, smooth boundary; 50 cm thick; the upper 20 cm colluviated.

460-480 cm **BCkg:** Yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4/m); silt loam; few, medium, prominent, yellowish red (5YR 5/8m) mottles; white (10YR 8/2), 1-2 cm thick calcareous layers; massive; slightly sticky (w), friable (m), slightly hard (d); some krotovinas; strongly efervescent; gradual, smooth boundary; 20 cm thick.

480-520 cm **Ck:** Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 40 cm thick. *Loess.*

Sk$_5$ *Gleyed Regosol*

520-530 cm **Ckg:** Very pale brown (10YR 8/3d, 7/3m); silt loam; common, fine, distinct, strong brown (7.5YR 5/8m) mottles along vertical rootlets; massive; sticky (w), friable (m), slightly hard (d); gradual, smooth boundary; 10 cm thick.

530-540 cm **Ckgj:** Very pale brown (10YR 8/3d, 7/3m); silt loam; few, fine, distinct, strong brown (7.5YR 5/8m) mottles; massive; sticky (w), friable (m), soft (d); gradual, smooth boundary; 10 cm thick. *Loess.*

Sk$_6$ *Gleyed Eutric Brunisol*

540-580 cm **Bmkgj:** Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; common, fine, distinct, strong brown (7.5YR 5/8m) mottles along vertical rootlets; massive; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; gradual, smooth boundary; 40 cm thick.

580-630 cm **Ckgj:** Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine, distinct, strong brown (7.5YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 50 cm thick. *Loess.*

Sk$_7$ *Gleyed Eutric Brunisol*

630-660 cm **Bmkgj**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; common (in the basal part), fine, distinct, strong brown (7.5YR 5/8m) mottles; very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; clear, smooth boundary; 30 cm thick.

660-700 cm **Ckgj**: Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine and medium, strong brown (7.5YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 40 cm thick. *Loess.*

Sk$_8$ ***Gleyed Eutric Brunisol***

700-730 cm **Bmkgj**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; common, fine to medium, prominent, yellowish red (5YR 5/8m) mottles; very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; gradual, smooth boundary; 30 cm thick.

730-760 cm **Ckgj**: White (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine and medium, olive yellow (2.5Y 6/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 30 cm thick. *Loess.*

Sk$_9$ ***Gleyed Eutric Brunisol***

760-800 cm **Bmkgj**: Light yellowish brown (10YR 6/4d), yellowish brown (10YR 5/4m); silt loam; common, fine and medium, prominent, yellowish red (5YR 5/8m) to red (10R 4/8m) mottles; weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; gradual, smooth boundary; 40 cm thick; the upper 10 cm colluviated.

800-820 cm **Ckgj**: White (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine and medium, olive yellow (2.5Y 6/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 20 cm thick.

Sk$_{10}$ ***Orthic Eutric Brunisol*** **(Kurtak PK - 1 hor.; 22-24 ka)**

820-870 cm **Bmk**: Yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4m); ; silt loam; few, fine and medium, prominent, yellowish red (5YR 5/8m) to red (10R 4/8m) mottles in the basal part; very weak, medium, subangular blocky; sticky (w), firm (m), slightly hard (d); strongly effervescent; abrupt, broken (involuted) boundary; 50 cm thick.

870-890 cm **BCk**: Light yellowish brown (10YR 6/4d), yellowish brown (10YR 5/4m); silt; common, fine and medium, distinct, strong brown (7.5YR 4/6m) mottles; massive; slightly sticky (w), very friable (m), soft (d); gradual, smooth boundary; 20 cm thick.

Sk$_{11}$ ***Gleyed Dark Brown Chernozem --- Orthic Turbic Cryosol.***

890-920 cm **Ahy**: Dark grayish brown (10YR 4/2d), very dark grayish brown (10YR 3/2m) silt loam; very weak, fine, subangular blocky; sticky, firm, hard; abundant charcoal present; abrupt, broken (involuted) boundary; 30 cm thick.

920-940 cm **Bmgy**: Light brownish gray (10YR 6/2d), grayish brown (10YR 5/2m); silt; common, medium to large, distint, strong brown (7.5YR 4/6m) mottles; massive; slightly sticky (w), friable (m), firm to slightly hard (d); charcoal present; Ahy and Bmgy horizons distorted in 10-50 cm x 2-15 cm lenses; 20 cm thick.

930-950 cm **Ckgy**: Light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m); silt; secondarily superimposed common, fine and medium, prominent, yellowish red (5YR 5/8m) to red (10R 4/8m) mottles and 1-3 cm large iron concretions; massive; non sticky (w), loose (m), soft (d); many, micro, vesicular pores; clear, smooth and irregular boundary; 20 cm thick. *Loess.*

950-990 cm **Ck:** Light gray (2.5Y 7/2d), light brownish gray (2.5Y 6/2m); silt; non sticky (w), loose (m), soft (d); many, micro, vesicular pores; clear, smooth and irregular boundary; 30 cm thick. *Loess.*

Sk$_{12}$ *Orthic Eutric Brunisol / truncated Dark Brown Chernozem (?)*

990-1020 cm **Bmk(g)**: Yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4m); silty clay; secondarily superimposed common, fine and medium, prominent, yellowish red (5YR 5/8m) mottles (g), and a 1 cm thick ferric horizontal layer in the basal part; very weak, subangular blocky; sticky (w), friable/firm (m), hard (d); krotovinas and charcoal present; moderately effervescent; diffuse, smooth boundary; 30 cm thick; the upper part truncated.

1020-1050 cm **BCca**: Very pale brown (10YR 8/3d - 7/3m); silt; few, fine, prominent, yellowish red (5YR 5/8m) mottles; massive; slightly sticky (w), friable (m), hard to very hard (d); many, micro, vesicular pores; very strongly effervescent; abrupt, iregular (flame) structure; 30 cm

Sk$_{13}$ *Gleyed Eutric Brunisol*

1050-1160 cm **Bmkgj**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; few, medium, prominent, yellowish red (5YR 4/6m) mottles; massive to very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); krotovinas and charcoal present; moderately effervescent; gradual, smooth boundary; 110 cm thick; (several /3?/ colluviated Bmkj horizons).

1160-1200 cm **Ckgj**: Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; common, fine and medium, prominent, strong brown (7.5YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 40 cm thick. *Loess.*

Sk$_{14}$ *Gleyed Eutric Brunisol*

1200-1230 cm **Bmkgj**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; common medium, prominent, yellowish red (5YR 5/8m) mottles in the basal part; very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; gradual, smooth boundary; 30 cm thick.

1230-1350 cm **Ckgj**: Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine and medium, prominent, strong brown (7.5YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 120 cm thick. *Loess.*

Sk$_{15}$ *Gleyed Regosol*

1350-1370 cm **Ckg**: Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; common, fine and medium, prominent, strong brown (7.5YR

5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 20 cm thick.

1370-1480 cm **Ckgj:** Light gray (2.5Y 7/2d) to white (2.5Y 8/2d), light brownish gray (2.5Y 6/2m); silt; few, fine and medium, prominent, strong brown (7.5YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 110 cm thick. *Loess.*

Sk_{16} *Gleyed Eutric Brunisol / Gleyed Regosol*

1480-1540 cm **Bmkgj**: Very pale brown (10YR 7/3d), pale brown (10YR 6/3m); silt loam; common, fine and medium, faint and distinct, yellowish brown (10YR 5/8m) and strong brown (7.5YR 5/8m); massive to very weak, medium, subangular blocky; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; clear, smooth boundary; 60 cm thick; (two colluviated Bmkgj horizons).

1540-1630 cm **Ckgj**: Light brownish gray (2.5Y 6/2d) to light olive gray (5Y 6/2d), dark grayish brown (2.5Y 4/2m); silt; few, fine and medium, faint and distinct, yellowish brown (10YR 5/8m) and strong brown (7.5YR 5/8m), and few, fine, faint to distinct, black (2.5Y 2/0m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; few, 1-2 cm large calcium carbonate concretions; clear, smooth boundary; 90 cm thick. *Loess*

Sk_{17} *Gleyed Regosol*

1630-1650 cm **Ckg**: Light brownish gray (2.5Y 6/2d) to light olive gray (5Y 6/2d), dark grayish brown (2.5Y 4/2m); silt; common, fine and medium, faint and distinct, yellowish brown (10YR 5/8m) and strong brown (7.5YR 5/8m), and few, fine, faint to distinct, black (2.5Y 2/0m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, irregular boundary; 20 cm thick.

Sk_{18} *Gleyed Eutric Brunisol* (Sukholozhskiy PK - hor. 1)

1650-1700/1720 cm **Bmkgj**: Pale brown (10YR 6/3d), brown (10YR 5/3m); silt loam; common (partly superimposed) fine and medium, faint, yellowish brown (10YR 5/8m) mottles (mostly in the upper part), and few, fine, faint, black (10YR 2/0m) mottles; very weak, medium, subangular blocky; slightly sticky (w), very friable to friable (m), soft to slightly hard (d); few, 1-2 cm calcium carbonate concretions; moderately effervescent; gradual, smooth to wavy boundary; 50 cm thick.

1720-1740 cm **Ckgj**: Light brownish gray (2.5Y 6/2d), dark grayish brown (2.5Y 4/2m); silt; few, fine and medium, distinct, yellowish brown (10YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; clear, smooth boundary; 20 cm thick. *Loess.*

Sk_{19} *Orthic Eutric Brunisol* (Sukholozhskiy PK - hor. 2)

1740-1770 cm **Bmk**: Brown (10YR 5/3d), dark brown (10YR 3/3-4/3m); silty clay; few, medium, distinct, strong brown (7.5YR 5/8m) mottles; very weak, fine to medium subangular blocky; slightly sticky (w), very friable (m), slightly hard to hard (d); calcium carbonate concentration in the basal part; weakly (in the upper part) to moderately effervescent; some charcoal present; gradual, wavy boundary; 30 cm thick, the upper 20 cm colluviated.

1770-1790 cm **Ck**: Light brownish gray (2.5Y 6/2d), dark grayish brown (2.5Y 4/2m); silt; few, medium, distinct, yellowish brown (10YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; abrupt, smooth boundary; 5-15 cm thick. *Loess.*

Sk$_{20}$ *Orthic Eutric Brunisol* (Sukholozhskiy PK - hor. 3)

1790-1820 cm **Bmk**: Light yellowish brown (10YR 6/4d), dark yellowish brown (10YR 3/4m); silty clay; few, fine, faint, reddish yellow (10YR 5/8m) mottles; weak, medium, subangular blocky; slightly sticky (w), very friable (m), slightly hard to hard (d); moderately effervescent; gradual, smooth boundary; 30-40 cm thick; upper 10 cm colluviated.

Sk$_{21}$ *Orthic Dark Brown Chernozem* (Sukholozhskiy PK - hor. 4)

1820-1830 cm **Ahy**: Dark brown (10YR 3/3d), very dark grayish brown (10YR 3/1m); silty clay; few, fine, distinct, strong brown (7.5YR 4/6m) mottles; weak, fine, granular to subangular blocky; sticky (w), firm (m), very hard (d); krotovinas and charcoal present; abrupt, irregular boundary with tongues into underlying layers; 10-15 cm thick; frost wedge casts (10-20 cm deep).

1830-1840 cm **Bmy**: Yellowish brown (10YR 5/4), dark brown (10YR 3/3); silty clay; few, fine, faint, strong brown (10YR 6/4m) mottles; weak, medium, subangular blocky; slightly sticky (w), very friable (m), slightly hard to hard (d); some charcoal present; abrupt, irregular (cryoturbated) boundary; 10-15 cm thick.

1840-1850 cm **BCca:** Brown (10YR 5/3d), dark brown (10YR 3/3m); silt; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; massive; slightly sticky (w), firm (m), hard (d); calcium carbonate concentration; strongly effervescent; gradual, smooth boundary; 10-15 cm thick.

1850-1860 cm **Cca**: Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m); silt; common, fine to medium, white (10YR 8/2d) calcium carbonate mottles; massive; slightly sticky (w), friable (m), hard (d); many, micro, vesicular pores; diffuse, smooth boundary; 20 cm thick..

Sk$_{22}$ *Orthic Dark Brown Chernozem* (Sukholozhskiy PK - hor. 5)

1860-1870 cm **Ahy:** Dark brown (10YR 3/3d), very dark grayish brown (10YR 3/1m); silty clay; few, fine, distinct, strong brown (7.5YR 4/6m) mottles; weak, fine, granular to subangular blocky; sticky (w), firm (m), very hard (d); krotovinas and charcoal present; abrupt, irregular boundary with tongues into underlying layers; 10-15 cm thick; frost wedge casts (10-50 cm deep).

1870-1880 cm **Bmy**: Yellowish brown (10YR 5/4), dark brown (10YR 3/3); silty clay; few, fine, faint, strong brown (10YR 6/4m) mottles; weak, medium, subangular blocky; slightly sticky (w), very friable (m), slightly hard to hard (d); some charcoal present; abrupt, irregular (cryoturbated) boundary; 20 cm thick.

1880-1890 cm **BCca:** Brown (10YR 5/3d), dark brown (10YR 3/3m); silt; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; massive; slightly sticky (w), firm (m), hard (d); calcium carbonate concentration; strongly effervescent; gradual, smooth boundary; 10-25 cm thick.

1890-1900/1930 cm **Ck**: Pale brown (10YR 6/3d), brown (10YR 4/3m); silt; few, fine, faint, brown (10YR 5/8m) mottles; massive, single-grain; non sticky (w), very friable (m), soft (d); many, micro, vesicular pores; krotovinas; abrupt, irregular boundary; 20-40 cm thick; frost wedge casts (20-40 cm). *Loess.*

Sk_{23} *Orthic Eutric Brunisol* (Sukholozhskiy PK - hor. 6)

1900-1920/1940 cm **Bmky**: Yellowish brown (10YR 5/4), dark brown (10YR 3/3); silty clay; few, fine, faint, strong brown (10YR 6/4m) and abundant, fine, distinct, black (7.5YR 2/1m) mottles; moderate, medium, subangular clocky slightly sticky (w), very friable (m), slightly hard to hard (d); calcium carbonate concentration near the base; strongly effervescent; many krotovinas; some charcoal present; abrupt, irregular (involuted) boundary; 10-30 cm thick.

Sk_{24} *Orthic Dark Brown Chernozem* (Kamenolozhskaya Soil)

1920/1940-1960 cm **Ahy**: Dark brown (10YR 3/3d) to dark grayish brown (10YR 4/2d), very dark grayish brown (10YR 3/1m); silty clay; few, fine, distinct, strong brown (7.5YR 4/6m) mottles; moderate, fine, granular; sticky (w), firm (m), very hard (d); many krotovinas; charcoal present; abrupt, irregular boundary with tongues into underlying layers; 20-30 cm thick; frost wedge casts (70-100 cm deep).

1960-1980 cm **Bmy**: Dark brown (10YR 4/3d), very dark grayish brown (10YR 3/2m); silt loam; few, fine, distinct, strong brown (7.5YR 4/6m) mottles; weak, fine to medium subangular blocky; sticky (w), friable to firm (m), hard (d); few, small (max 1 cm) quartz pebbles; krotovinas; charcoal present; abrupt, wavy boundary; 20-40 cm thick.

1980-2010 cm **BCca**: Brown (10YR 5/3d), dark brown (10YR 3/3m); silt; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; massive; slightly sticky (w), firm (m), hard (d); gradual, smooth boundary; 30 cm thick; some charcoal; coarse sand lenses; strongly effervescent; 30 cm thick.

2010-2020 cm **Cca**: Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m); silt; common, fine to medium, white (10YR 8/2d) calcium carbonate mottles; massive; slightly sticky (w), friable (m), hard (d); many, micro, vesicular pores; few, small (max. 0.5 cm) carbonate concretions; diffuse, smooth boundary; 10 cm thick..

Sk_{25} *Orthic Eutric Brunisol*

2020-2060 cm **Bmk:** Light yellowish brown (10YR 6/4d), yellowish brown (10YR 5/4m); silt loam; few, fine, distinct, strong brown (7.5YR 4/6m) mottles; very weak, medium subangular blocky; sticky (w), friable to firm (m), hard (d); strongly effervescent; krotovinas; some charcoal present; gradual smooth boundary; 40 cm thick.

2060-2080/2120 cm **BCca:** Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m); silt; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; common, fine to medium, white (10YR 8/2d) calcium carbonate mottles; massive; slightly sticky (w), firm (m), hard (d); very strongly effervescent; gradual, smooth boundary; 30 cm thick.

2080/2120-2180 cm **Ck:** Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m); silt; few, fine, faint, yellowish brown (10YR 5/8m), and common, fine, fine, faint, black (10YR 2/1m) mottles (at the base); massive, single-grain; non sticky (w), friable (m), slightly hard (d); many, micro, vesicular pores; gradual, smooth boundary; 80-90 cm thick. *Loess.*

Sk$_{26}$ *Gleyed Regosol*

2180-2200 cm **Ckg:** Pale brown (10YR 6/3d), brown (10YR 5/3-4/3m); silt; common, fine, fine, faint, black (10YR 2/1m) mottles; massive, single-grain; non sticky (w), friable (m), slightly hard (d); many, micro, vesicular pores; gradual, smooth boundary; 30 cm thick.

2200-2490 cm **Bmkgj/Cky**: Light brownish gray (2.5Y 6/2d) to light olive gray (5Y 6/2d), dark grayish brown (2.5Y 4/2m); pale brown (10YR 6/3d), brown (10YR 5/3-4/3m), and brown (10YR 5/3d), dark brown (10YR 3/3m), 0.2-1 cm thick interstratified layers; silt; few, fine, distinct to faint black (10YR 2/2d; 7.5YR 2/2) mottles; few, fine to medium, white (10YR 8/2d) calcium carbonate 1-2 mm concretions (near the top or the unit); laminated irregular structure with minor frost wedge casts and involutions, dipping at *ca.* 5° towards north; slightly sticky (w), friable (m), hard (d); clear, wavy boundary; 280-290 cm thick; *Niveoaeolian colluviated loess.*

Sk$_{27}$ *Gleyed Regosol*

2490-2520 cm **Ckg:** Pale brown (10YR 6/3d), dark brown (10YR 4/3m); silt; common, fine, faint; yellowish brown (10YR 5/8m) ferric mottles along vertical rootlets; black (10YR 2/2m) mottles; massive, single-grain; non sticky (w), loose (m), soft (d); diffuse, smooth boundary; 30 cm thick.

2520-2650 cm **Ckgj:** Pale brown (10YR 6/3d), dark brown (10YR 4/3m); silt; few, fine, faint; yellowish brown (10YR 5/8m), and black (10YR 2/2m) mottles; few, fine to medium, white (10YR 8/2d) calcium carbonate 1-2 mm concretions; massive, single-grain; non sticky (w), loose (m), soft (d); diffuse, smooth boundary; 130 cm thick. *Loess.*

Sk$_{28}$ *Gleyed Regosol*

2650-2670 cm **Ckg:** White (10YR 8/2d), light gray (10YR 7/2m); silt; few, fine and medium, distinct, dark brown (7.5YR 3/4m) mottles along vertical rootlets, and common, fine, distinct, very dark gray (7.5YR 3/0m) mottles; massive, single-grain; slightly sticky (w), friable (m), slightly hard (d); gradual, smooth boundary; 20 cm thick.

Sk_{29} *Orthic Eutric Brunisol*

2670-2720 cm **Bmk(gj)y**: Brown (10YR 5/3d), dark brown (10YR 3/3m); silt loam; few, fine and medium, distinct, strong brown (7.5YR 5/8m, and common, fine, distinct, very dark gray (7.5YR 3/0m) mottles (superimposed from the above Ckg horizon); weakly stratified massive, single-grain near the top, weak, medium, subangular blocky in the lower part; slightly sticky (w), friable (m), slightly hard (d); moderately effervescent; krotovinas near the base; clear, irregular boundary disrupted by frost wedge casts (10-30 cm deep); 50 cm thick.

Sk_{30} *Orthic Dark Brown Chernozem*

2720-2740 cm **Ahy**: Very dark grayish brown (10YR 3/2d), very dark brown to black (10YR 2/2-2/1m); silty clay; weak, fine granular; sticky (w), firm (m), hard to very hard (d); many krotovinas; abundant charcoal present; abrupt, irregular boundary with tongues into underlying layers; 20-25 cm thick (70 cm with frost wedge casts); frost wedge casts (20-50 cm deep).

2740-2780 cm **Bmy**: Dark brown (10YR 4/3d - 3/3m); silt loam; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; few, fine, white (10YR 8/2d) calcium carbonate 1-2 mm concretions; very weak, medium subangular blocky; sticky (w), friable to firm (m), hard (d); moderately effervescent; krotovinas and some charcoal present; abrupt, wavy boundary; 20-40 cm thick.

2780-2810 cm **BCca**: Very pale brown (10YR 7/3d), pale brown (10YR 6/3m); silt; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; common, fine, white (10YR 8/2d) calcium carbonate 1-3 mm concretions; massive; slightly sticky (w), firm (m), hard (d); strongly effervescent; krotovinas present; abrupt, irregular (involuted) boundary; 30-40 cm thick.

2810-2830 cm **Cca**: Light yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4m); silt; common, fine to medium, white (10YR 8/2d) calcium carbonate concreations; massive; slightly sticky (w), friable (m), hard (d); many, micro, vesicular pores; strongly effervescent; diffuse, smooth boundary; 20 cm thick.

Sk_{31} *Orthic Eutric Brunisol*

2830-2860 cm **Bmk:** Light brownish gray (10YR 6/2d), grayish brown (10YR 5/2m); silt loam; fine, few, distinct, strong brown (7.5YR 4/6m) mottles; very weak, medium subangular blocky; sticky (w), friable to firm (m), hard (d); strongly effervescent; krotovinas and some charcoal present; abrupt, wavy / broken (involuted) boundary; 20-30 cm thick.

2860-2880 cm **BCca:** Pale brown (10YR 6/3d), brown (10YR 5/3m); silt; common, fine, white (10YR 8/2d) calcium carbonate 1-2 mm concretions; massive and platty; slightly sticky (w), firm (m), hard (d); very strongly effervescent; krotovinas present; smooth, gradual boundary; 20 cm thick.

2880-2930 cm **Cca:** Light yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4m); silt; common, fine to medium, white (10YR 8/2d) calcium carbonate 1-2 mm conreations; massive; slightly sticky (w), friable (m), hard (d); many, micro, vesicular pores; few, small (max. 0.5 cm) carbonate concretions; diffuse, smooth boundary; 40-50 cm thick.

2930-3040 cm **Bmkgj/Cky**: Light brownish gray (2.5Y 6/2d) to light olive gray (5Y 6/2d), dark grayish brown (2.5Y 4/2m); pale brown (10YR 6/3d), brown (10YR 5/3-4/3m), and brown (10YR 5/3d), dark brown (10YR 3/3m), 0.2-2 cm thick interstratified layers; silt; few, fine, faint black (10YR 2/2d) mottles; few, fine, white (10YR 8/2d) 1-2 mm calcium carbonate concretions; laminated irregular structure with minor frost wedge casts and involutions, dipping at *ca.* 3° towards north; slightly sticky (w), friable (m), hard (d); clear, wavy boundary; 100-110 cm thick. *Niveoaeolian colluviated loess.*

3040-3120 cm **Ck**: Very pale brown (10YR 7/3d), pale brown (10YR 6/3m); massive, single-grain; non sticky (w), friable (m), soft (d); abrupt, smooth boundary; 100-120 cm thick. *Loess.*

Sk$_{32}$ ***Orthic Eutric Brunisol***

3120-3150 cm **Bmky**: Light brown (7.5YR 6/4d), dark brown (7.5YR 4/4m); silt loam; few, fine, faint, strong brown (7.5YR 5/8m), and common, fine, faint, black (7.5YR 2/0m) mottles; common, fine, white (7.5YR 8/2d) calcium carbonate concretions; very weak, medium, subangular blocky; secondarily modified by frost actions into coarse platty; sticky (w), firm (m), hard (d); strongly effervescent; krotovinas present; abrupt, wavy (involuted) boundary; 30 cm thick.

3150-3190 cm **BCca**: Light yellowish brown (10YR 5/4d), dark yellowish brown (10YR 4/4m); silt; few, fine, faint, strong brown (7.5YR 5/8m) mottles; common, fine, white (7.5YR 8/2d) calcium carbonate concretions; massive; slightly sticky (w), firm (m), hard (d); strongly effervescent; abrupt, irregular (involuted) boundary; 40 cm thick.

3190-3310+ cm **Cca**: Yellowish brown (10YR 5/4d), dark yellowish brown (10YR 3/4m); silt; few, fine, white (7.5YR 8/2d) calcium carbonate concretions; massive, weekly stratified; slightly sticky (w), firm (h), hard (d); some detrital organic matter present; 120+ cm thick.*Colluviated loess?*

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1) In mineral horizons of Chernozemic soils (Bm) and Brunisolic soils (Bmk), the suffix *g (gj)*was not attached because of a negligible amount of mottles. The formula Bmkgj defines Gleyed Eutric Brunisols distinguished by more intensive mottling.

2) All loess is uniformly highly calcareous. Hydrochlorid acid reaction in terms of effervescence was specified only for mineral Bmk and BCca horizons.

APPENDIX B

KURTAK SECTION 29: PALAEOMAGNETIC DATA

No.	*depth*	*strike*	*dip*	*roll*	*susceptibility*
JC00	0	0	0	0	140
JC000	10	0	0	0	220
JC001	20	56	-1	1	281
JC002	30	56	-1	0	290
JC003	40	57	-3	1	307
JC004	50	56	-2	0	309
JC005	60	53	3	1	269
JC006	70	59	0	2	241
JC007	80	56	-1	2	224
JC008	90	55	-1	1	244
JC009	100	55	-1	-2	295
JC010	110	56	-4	-2	315
JC011	120	58	0	-3	373
JC012	130	59	-4	-2	375
JC013	140	60	0	2	397
JC014	150	66	1	-1	405
JC015	160	66	0	0	467
JC016	170	66	1	3	443
JC017	180	65	1	1	425
JC018	190	66	0	2	431
JC019	200	67	-3	2	452
JC020	210	64	1	1	577
JC021	220	64	1	2	488
JC022	230	65	0	1	430
JC023	240	64	3	0	385
JC024	250	64	0	2	331
JC025	260	63	1	3	394
JC026	270	59	0	2	397
JC027	280	64	0	0	339
JC028	290	61	-1	3	396
JC029	300	60	3	1	381
JC030	310	61	1	0	409
JC031	320	62	4	2	400
JC032	330	61	4	0	390
JC033	340	62	4	3	332
JC034	350	61	1	-2	376
JC035	360	62	4	3	419
JC036	370	63	0	0	208
JC037	380	64	0	-1	356
JC038	390	61	-2	3	343
JC039	400	61	-3	-3	409
JC040	410	60	0	-3	353
JC041	420	59	1	1	323
JC042	430	59	1	0	237
JC043	440	60	2	0	246
JC044	450	60	2	3	275
JC045	460	66	1	-3	239
JC046	470	64	-2	-1	312
JC047	480	64	1	-2	275
JC048	490	64	-2	-2	310

JC049	500	60	4	-1	307
JC050	510	62	0	0	297
JC051	520	60	1	-4	295
JC052	530	60	2	1	273
JC053	540	60	0	0	286
JC054	550	61	0	-2	247
JC055	560	58	4	0	233
JC056	570	58	-3	3	265
JC057	580	61	0	1	250
JC058	590	60	0	1	312
JC059	600	61	1	-1	320
JC060	610	58	3	-1	262
JC061	620	59	1	0	277
JC062	630	54	2	-1	228
JC063	640	56	-2	-1	260
JC064	650	58	3	1	214
JC065	660	57	4	0	240
JC066	670	56	0	-3	284
JC067	680	60	1	-3	267
JC068	690	59	3	2	258
JC069	700	53	2	0	199
JC070	710	56	2	1	238
JC071	720	55	3	2	227
JC072	730	56	-3	4	196
JC073	740	57	-1	3	224
JC074	750	55	0	3	248
JC075	760	60	0	-4	176
JC076	770	60	4	-1	150
JC077	780	60	5	0	172
JC078	790	59	-4	-3	128
JC079	800	61	-3	1	119
JC080	810	58	0	3	115
JC081	820	60	0	-1	126
JC082	830	57	-1	1	145
JC083	840	60	2	-1	127
JC084	850	60	3	-1	154
JC085	860	60	2	-2	141
JC086	870	58	0	3	117
JC087	880	58	-2	0	142
JC088	890	58	-4	3	185
JC089	900	67	0	-2	167
JC090	910	61	1	3	147
JC091	920	63	1	4	145
JC092	930	64	3	1	143
JC093	940	64	4	4	127
JC094	950	62	3	5	136
JC095	960	62	4	-2	172
JC096	970	62	2	2	199
JC097	980	64	-1	4	186
JC098	990	65	0	-1	113
JC099	1000	60	-3	0	100
JC100	1010	65	0	4	125
JC101	1020	62	0	2	173

JC102	1030	65	0	0	187
JC103	1040	62	-1	-2	178
JC104	1050	61	2	-2	173
JC105	1060	64	1	1	168
JC106	1070	64	4	0	193
JC107	1080	65	3	0	216
JC108	1090	66	1	-1	229
JC109	1100	65	3	-1	261
JC110	1110	65	1	0	212
JC111	1120	65	1	-2	381
JC112	1130	58	-1	1	271
JC113	1140	62	0	0	336
JC114	1150	64	-3	1	195
JC115	1160	64	-3	4	267
JC116	1170	70	0	-1	276
JC117	1180	68	-1	-4	300
JC118	1190	70	2	1	326
JC119	1200	74	1	-2	266
JC120	1210	70	-1	-4	326
JC121	1220	70	5	-3	274
JC122	1230	71	5	-2	226
JC123	1240	73	0	1	331
JC124	1250	72	2	-1	311
JC125	1260	73	0	0	356
JC126	1270	72	0	1	317
JC127	1280	70	-5	0	346
JC128	1290	74	-2	-3	326
JC129	1300	64	-4	-1	455
JC130	1310	70	-3	-2	383
JC131	1320	71	3	-1	316
JC132	1330	74	2	0	334
JC133	1340	73	-1	-3	337
JC134	1350	73	1	3	321
JC135	1360	72	-4	3	320
JC136	1370	75	5	2	310
JC137	1380	76	4	4	366
JC138	1390	67	1	2	395
JC139	1400	65	0	0	353
JC140	1410	68	3	2	490
JC141	1420	71	2	-2	419
JC142	1430	73	1	0	432
JC143	1440	70	-2	-1	388
JC144	1450	70	2	-3	542
JC145	1460	66	0	-1	454
JC146	1470	68	4	0	424
JC147	1480	60	5	-1	380
JC148	1490	65	1	-3	350
JC149	1500	67	-1	0	405
JC150	1510	66	0	2	539
JC151	1520	66	1	0	384
JC152	1530	69	2	-4	381
JC153	1540	70	4	-2	428
JC154	1550	60	3	-2	578

JC155	1560	65	2	0	459
JC156	1570	69	4	-1	442
JC157	1580	65	1	0	402
JC158	1590	68	2	-3	430
JC159	1600	64	-4	-3	367
JC160	1610	65	-3	-2	411
JC161	1620	65	1	-2	382
JC162	1630	60	0	3	326
JC163	1640	61	4	-3	294
JC164	1650	60	-1	-2	237
JC165	1660	63	-1	-2	292
JC166	1670	66	-3	-4	297
JC167	1680	64	-2	-5	261
JC168	1690	67	2	-2	274
JC169	1700	69	5	-5	276
JC170	1710	68	4	1	215
JC171	1720	64	0	0	240
JC172	1730	64	4	1	239
JC173	1740	63	5	1	222
JC174	1750	70	4	2	224
JC175	1755	66	-3	2	244
JC176	1760	68	-3	1	267
JC177	1770	66	3	3	226
JC178	1780	66	5	-1	186
JC179	1790	66	-3	-1	151
JC180	1800	64	2	-3	137
JC181	1805	63	2	-2	145
JC182	1810	70	0	-1	122
JC183	1820	69	0	0	093
JC184	1830	70	5	0	076
JC185	1840	64	3	0	092
JC186	1850	67	3	0	137
JC188	1870	69	5	3	139
JC189	1880	63	2	-2	194
JC193	1890	69	1	2	262
JC194	1900	68	2	-3	195
JC195	1910	70	1	-2	098
JC196	1920	69	4	1	095
JC197	1930	70	5	2	115
JC206	1940	61	5	4	157
JC198	1950	68	2	2	207
JC199	1960	69	1	1	228
JC205	1970	74	-1	1	223
JC200	1980	69	4	-3	215
JC201	1990	60	-3	-5	147
JC202	2000	65	-3	1	183
JC203	2010	62	0	-3	184
JC207	2020	62	-1	0	145
JC208	2030	61	4	0	125
JC209	2040	60	4	-2	164
JC210	2050	63	3	-3	130
JC211	2060	62	2	-3	127
JC212	2070	60	-1	-3	185

JC213	2080	61	-4	-3	160
JC214	2090	62	0	-5	202
JC215	2100	63	1	-4	431
JC216	2110	61	1	-4	282
JC217	2120	63	-1	-4	335
JC218	2130	64	0	-4	412
JC219	2140	68	-1	-4	401
JC220	2150	61	-3	-4	405
JC221	2160	71	-2	3	342
JC222	2170	62	5	-1	324
JC223	2180	60	1	1	292
JC224	2190	63	0	1	299
JC225	2200	61	-1	1	246
JC226	2210	62	2	0	192
JC227	2220	70	-3	-1	159
JC228	2230	70	0	4	250
JC229	2240	68	-1	-3	369
JC230	2250	71	1	-2	405
JC231	2260	68	0	1	177
JC232	2270	70	2	-2	185
JC233	2280	68	0	-1	327
JC234	2290	70	1	-4	226
JC235	2300	71	-1	2	240
JC236	2310	70	2	-3	242
JC237	2320	64	2	-1	235
JC238	2330	62	-3	-2	317
JC239	2340	61	-3	5	204
JC240	2350	62	4	0	280
JC241	2360	70	1	3	263
JC242	2370	65	-1	2	240
JC243	2380	63	0	-1	354
JC244	2390	66	0	-1	275
JC245	2400	65	1	0	266
JC246	2410	66	0	0	173
JC247	2420	65	0	1	228
JC248	2430	65	2	2	267
JC249	2440	60	3	2	223
JC250	2450	64	0	-1	203
JC251	2460	70	4	-1	355
JC252	2470	66	-4	2	315
JC253	2480	70	-3	0	329
JC254	2490	68	-3	3	214
JC255	2500	69	0	1	271
JC256	2510	67	-1	0	242
JC257	2520	68	2	4	283
JC258	2530	69	2	1	296
JC259	2540	66	1	4	318
JC260	2550	66	0	-2	350
JC261	2560	70	1	-4	347
JC262	2570	69	0	-4	339
JC263	2580	68	-4	1	330
JC264	2590	71	-1	-1	337
JC265	2600	73	-1	-3	324

JC266	2610	77	2	-4	328
JC267	2620	73	0	-4	320
JC268	2630	68	4	-3	341
JC269	2640	70	5	-3	348
JC270	2650	67	3	3	293
JC271	2660	69	5	3	234
JC272	2670	68	4	0	228
JC273	2680	65	5	0	182
JC274	2690	70	5	-1	204
JC275	2700	71	3	4	205
JC276	2710	70	-1	-1	201
JC277	2720	66	2	0	178
JC278	2730	64	-5	4	143
JC279	2740	62	5	4	137
JC280	2750	61	-2	3	175
JC281	2760	67	-2	1	163
JC286	2770	59	-4	1	152
JC287	2780	55	5	2	142
JC288	2790	54	5	-3	194
JC289	2800	54	-3	-1	170
JC290	2810	65	4	1	180
JC291	2820	59	1	-1	182
JC292	2830	57	-1	1	153
JC293	2840	63	0	-1	128
JC294	2850	61	3	-2	116
JC295	2860	58	1	1	172
JC296	2870	59	2	1	185
JC297	2880	61	5	1	191
JC298	2890	61	0	3	217
JC299	2900	60	1	-3	222
JC300	2910	60	0	1	213
JC301	2920	62	-4	2	206
JC302	2930	61	-4	3	180
JC303	2940	61	-5	2	185
JC304	2950	59	-1	1	176
JC305	2960	60	-4	4	160
JC306	2970	60	-2	4	199
JC307	2980	62	2	-3	195
JC308	2990	60	1	-2	187
JC309	3000	60	0	3	157
JC310	3010	63	1	-3	186
JC311	3020	65	-4	-3	216
JC312	3030	60	0	-3	210
JC313	3040	60	2	-1	176
JC314	3050	62	-3	1	221
JC315	3060	60	-2	-1	230
JC316	3070	59	0	0	220
JC317	3080	61	0	-1	264
JC318	3090	59	-1	-3	229
JC319	3100	56	1	4	218
JC320	3110	54	5	0	209
JC321	3120	64	-3	0	160
JC322	3130	65	-3	-3	141

JC323	3140	65	0	-4	124
JC324	3150	61	1	-1	161
JC325	3160	62	-1	0	153
JC326	3170	63	5	0	146
JC327	3180	61	3	0	144
JC328	3190	61	5	1	136
JC329	3100	66	3	1	178
JC330	3210	60	1	2	177
JC331	3220	59	2	3	168
JC332	3230	62	0	4	179
JC333	3240	58	0	2	185
JC334	3250	62	2	-1	187
JC335	3260	58	3	3	190
JC336	3270	63	5	-3	166
JC337	3280	62	4	-1	168
JC338	3290	60	3	-2	183
JC339	3300	64	5	-2	196
JC340	3310	60	-3	1	203

KURTAK SECTION 22

No.	*depth*	*strike*	*dip*	*roll*	*susceptibility*
JC501	030	55	2	-5	154
JC502	060	53	-5	-3	168
JC503	090	56	-2	-5	057
JC504	120	62	3	-3	090
JC505	150	55	-2	-5	075
JC506	180	44	-1	-3	182
JC507	210	70	-3	-1	212
JC508	240	62	-2	0	201
JC509	270	60	0	0	156
JC510	300	61	-4	-3	224
JC511	330	63	-5	-1	186
JC512	360	59	-4	4	163
JC513	390	52	-2	1	156
JC514	420	51	-5	2	235
JC515	450	60	-2	0	108
JC516	480	53	-6	-3	153
JC517	510	48	-5	3	225
JC518	540	53	-3	0	108
JC519	570	52	-4	1	161
JC520	600	56	1	-3	140
JC521	610	56	4	0	120

FIGURES

Figure 1. Geographical location of the study area.

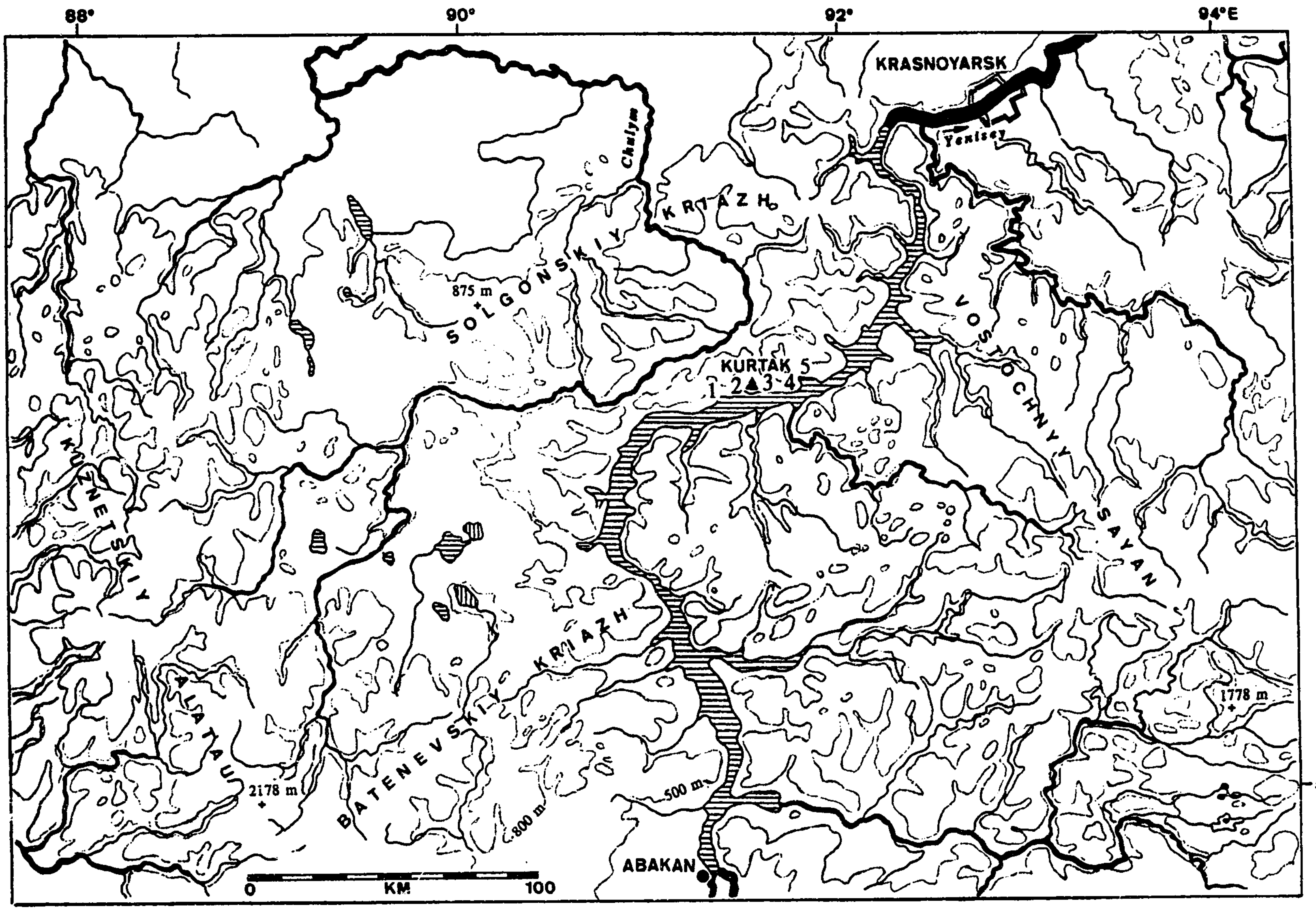


Figure 2. Topographic Setting of the Minusinsk Basin with location of the Kurtak sites along the Krasnoyarsk Reservoir shore. 1 - Kashtanka, 2 - Kamennyy Log & Sukhoy Log, 3 - Berezhekovo, 4 - Razlog & Verkhniy Kamen, 5 - Ust'-Izhul' (triangle - Kurtak village).

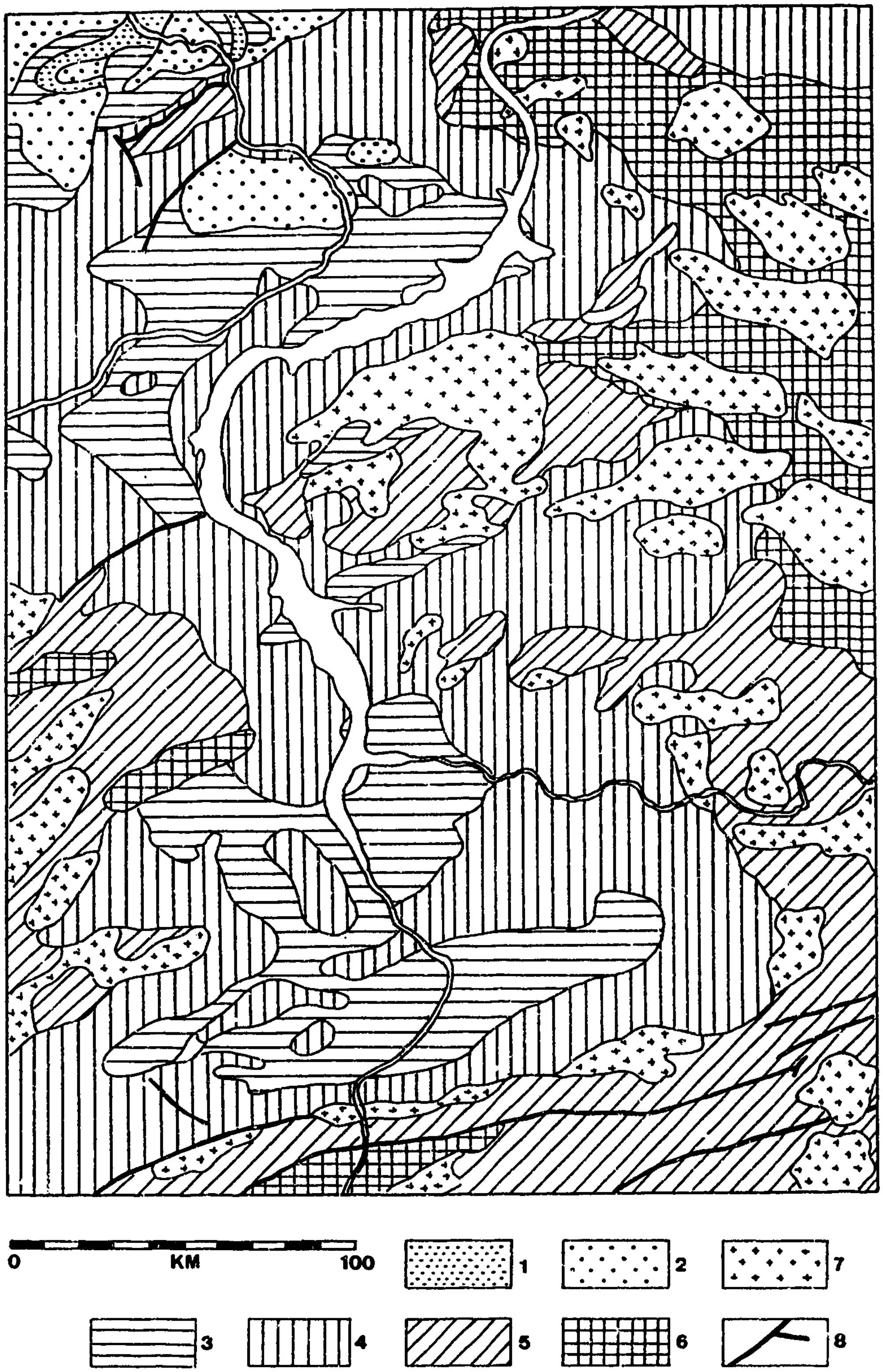
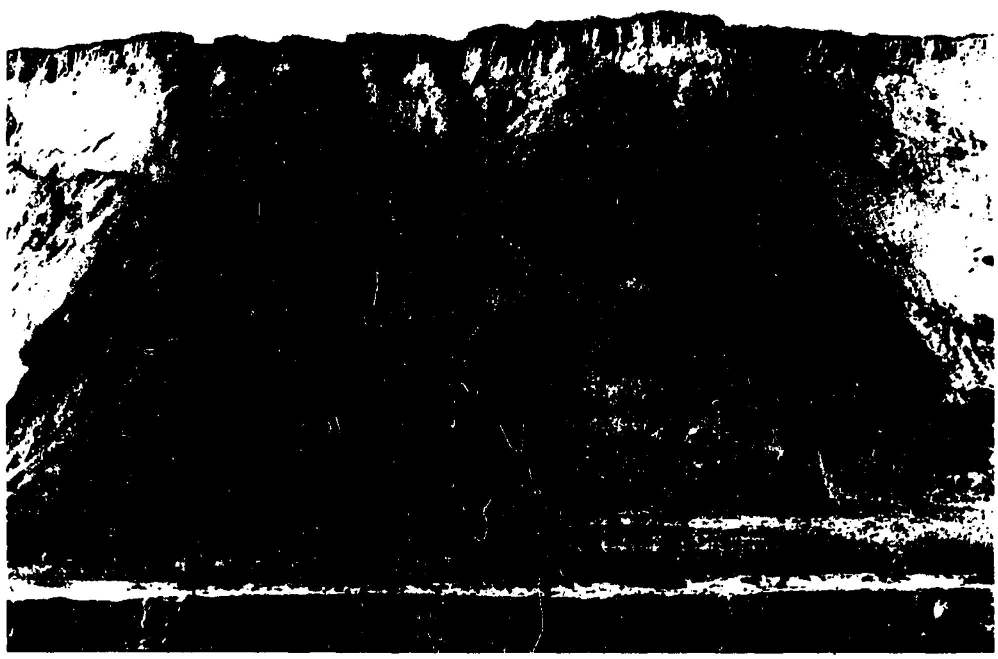


Figure 3. Structural Geology of the study area. Legend: 1 - Quaternary alluvial sands and gravels; 2 - Palaeocene and Jurassic alluvial sands; 3 - Carboniferous alluvial sands and lacustrine silts (cemented); 4 - Devonian lacustrine sands, silts and clays; 5 - Cambrian volcanic deposits; 6 - Proterozoic volcanic deposits; 7 - Palaeozoic igneous (mostly granitic) rocks; 8 - Principal tectonic contacts. (According to the Geological Map of the Krasnoyarsk Kray 1966, 1:2 500 000).

A



B

Figure 4. (A) Central part of the Berezhekovo Site with semi-prominent 30 m high cliffs. The exposed last interglacial pedocomplex is indicated by the arrow. (B) View of Kurtak Section 29. Three major pedocomplexes (indicated by the arrows) are assigned (from the bottom) to the penultimate interglacial, the last interglacial and the mid-last glacial interstadial. Height of the cliff is 32 m.

BEREZHEKOVO SITE

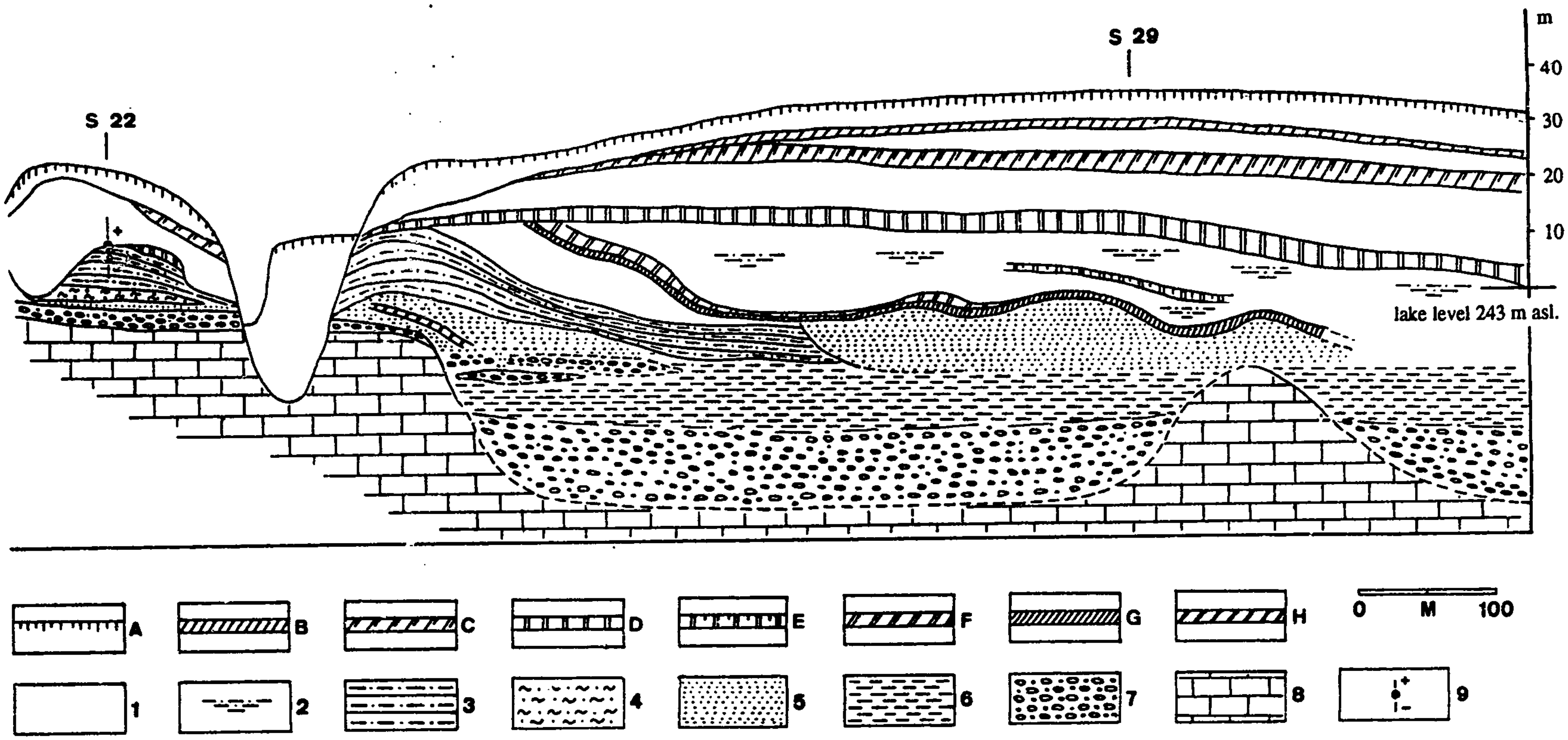


Figure 5. Stratigraphic profile of the central part of the Berezhekovo Site with Kurtak Section 22 and Section 29. Legend *Soils:* A - recent (Holocene) Chernozem; B - Sartan (Late Wisconsinan) Pedocomplex; C - Kurtak (Mid-Wisconsinan) Pedocomplex; D - Last Interglacial Pedocomplex; E - Penultimate Interglacial Pedocomplex; F-H - Early Pleistocene soils. *Structural Geology:* 1 - aeolian loess; 2 - colluviated loess; 3 - subaerial sands and silts (the Tazov / late Illinoian Glacial); 4 - periglacial alluvium (the Samarov / early Illinoian Glacial); 5 - subaqueous sands (Early Pleistocene); 6 - subaqueous silts and clays (Early Pleistocene); 7 - sands and gravels of the Early Pleistocene 70/90-100 m terrace; 8 - sandstone bedrock (Carboniferous); 9 - palaeomagnetic polarity boundary.

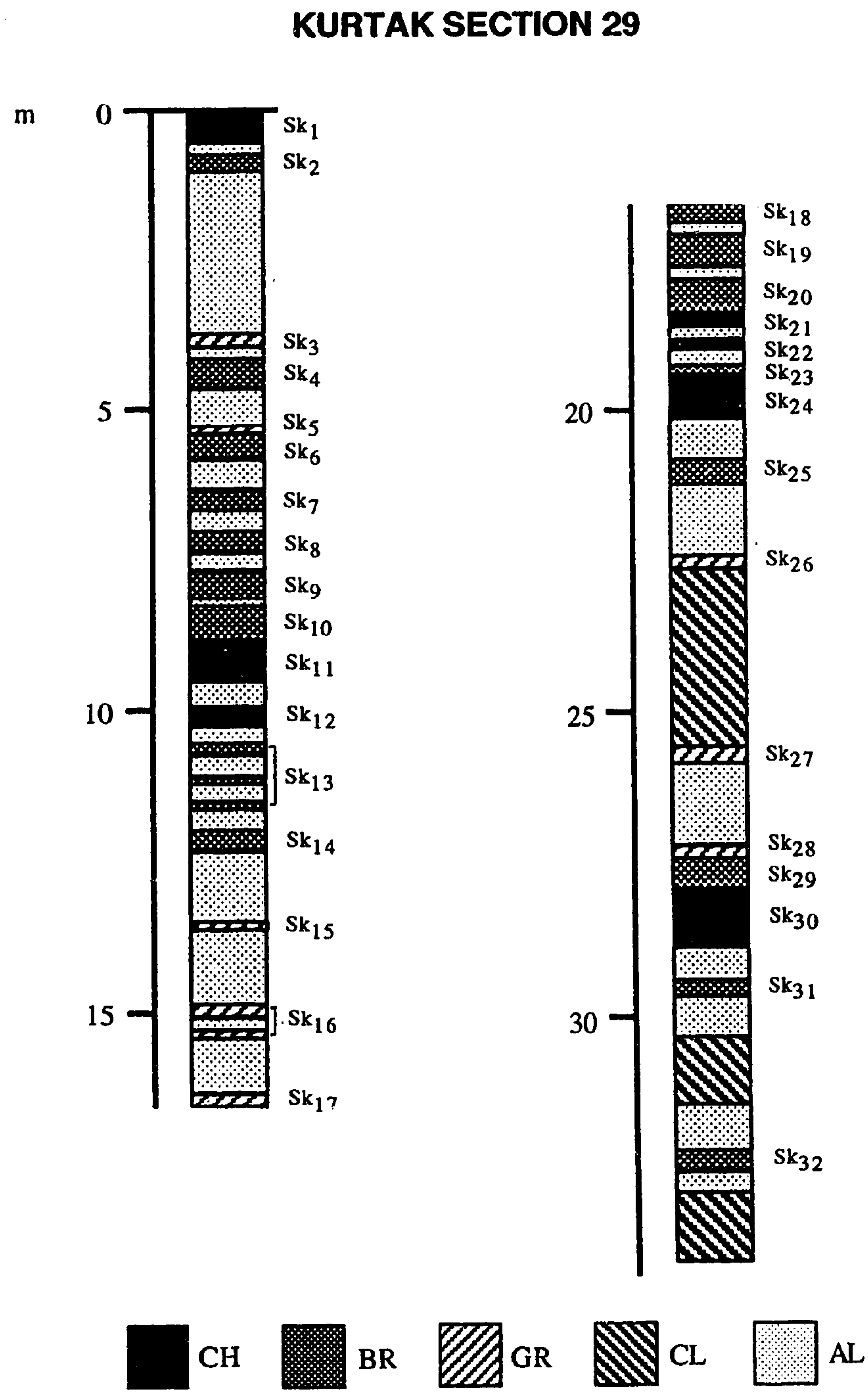


Figure 6. Composite stratigraphic profile of Kurtak Section 29. Legend: CH - Chernozems; BR - Brunisols; GR - Gleyed Regosols; CL - colluviated loess; AL - aeolian loess.

Spectrum: LOESS S2-3

All elmts analysed,NORMALISED
TOTAL AREA= 233173

ELMT		ZAF	%ELMT	+-	Error	ATOM.%
O K :	0	.378	9.385	+-	.707	17.780
NaK :	0	.674	1.004	+-	.322	1.324
MgK :	0	.686	2.145	+-	.228	2.675
AlK :	0	.768	10.587	+-	.238	11.894
SiK :	0	.761	37.304	+-	.289	40.255
K K :	0	.951	3.854	+-	.160	2.988
CaK :	0	.923	16.757	+-	.241	12.673
TiK :	0	.791	1.317	+-	.172	.833
FeK :	0	.844	17.647	+-	.441	9.578
TOTAL			100.001			100.000

1

Spectrum: LOESS W1

All elmts analysed,NORMALISED
TOTAL AREA= 218064

ELMT		ZAF	%ELMT	+-	Error	ATOM.%
O K :	0	.401	10.203	+-	.675	18.941
NaK :	0	.691	1.356	+-	.310	1.752
MgK :	0	.696	1.805	+-	.216	2.205
AlK :	0	.781	9.890	+-	.231	10.888
SiK :	0	.774	41.116	+-	.300	43.475
K K :	0	.936	4.097	+-	.160	3.112
CaK :	0	.914	13.011	+-	.224	9.642
TiK :	0	.795	1.521	+-	.167	.943
FeK :	0	.844	17.001	+-	.442	9.041
TOTAL			100.001			100.000

2

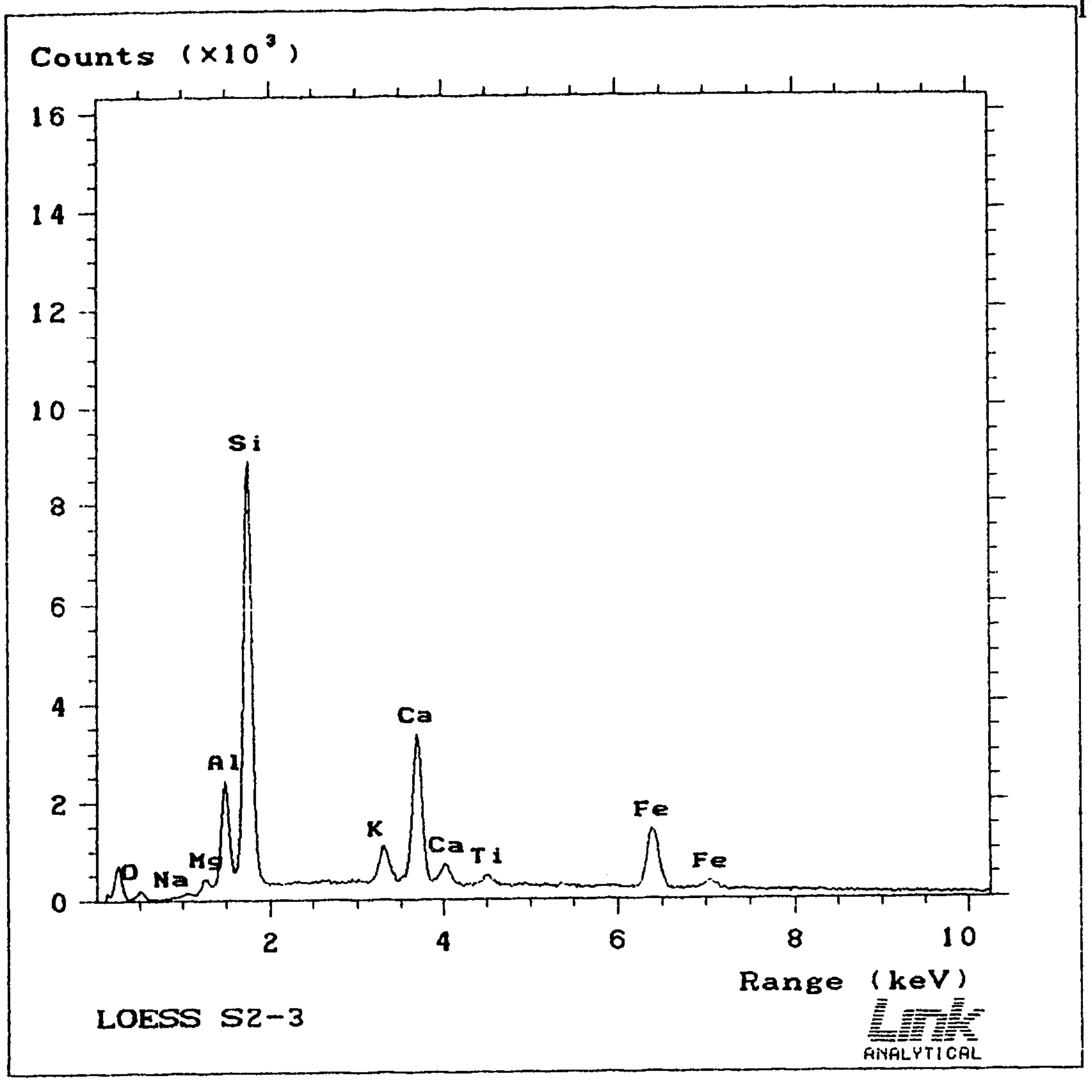
Spectrum: SOIL R-W A-HORIZON

All elmts analysed,NORMALISED
TOTAL AREA= 220668

ELMT		ZAF	%ELMT	+-	Error	ATOM.%
O K :	0	.470	9.807	+-	.614	18.186
NaK :	0	.689	1.266	+-	.318	1.634
MgK :	0	.695	2.123	+-	.229	2.591
AlK :	0	.775	12.110	+-	.250	13.319
SiK :	0	.749	42.138	+-	.313	44.511
K K :	0	.922	4.444	+-	.167	3.372
CaK :	0	.907	6.327	+-	.189	4.684
TiK :	0	.814	1.590	+-	.173	.985
FeK :	0	.851	20.175	+-	.467	10.719
TOTAL			99.980			100.000

3

Table 3. Relative percentages of element distribution in the late last glacial loess (1), early last glacial loess (2) and the A horizon of the last interglacial Chernozem (3). EDX SEM analysis.

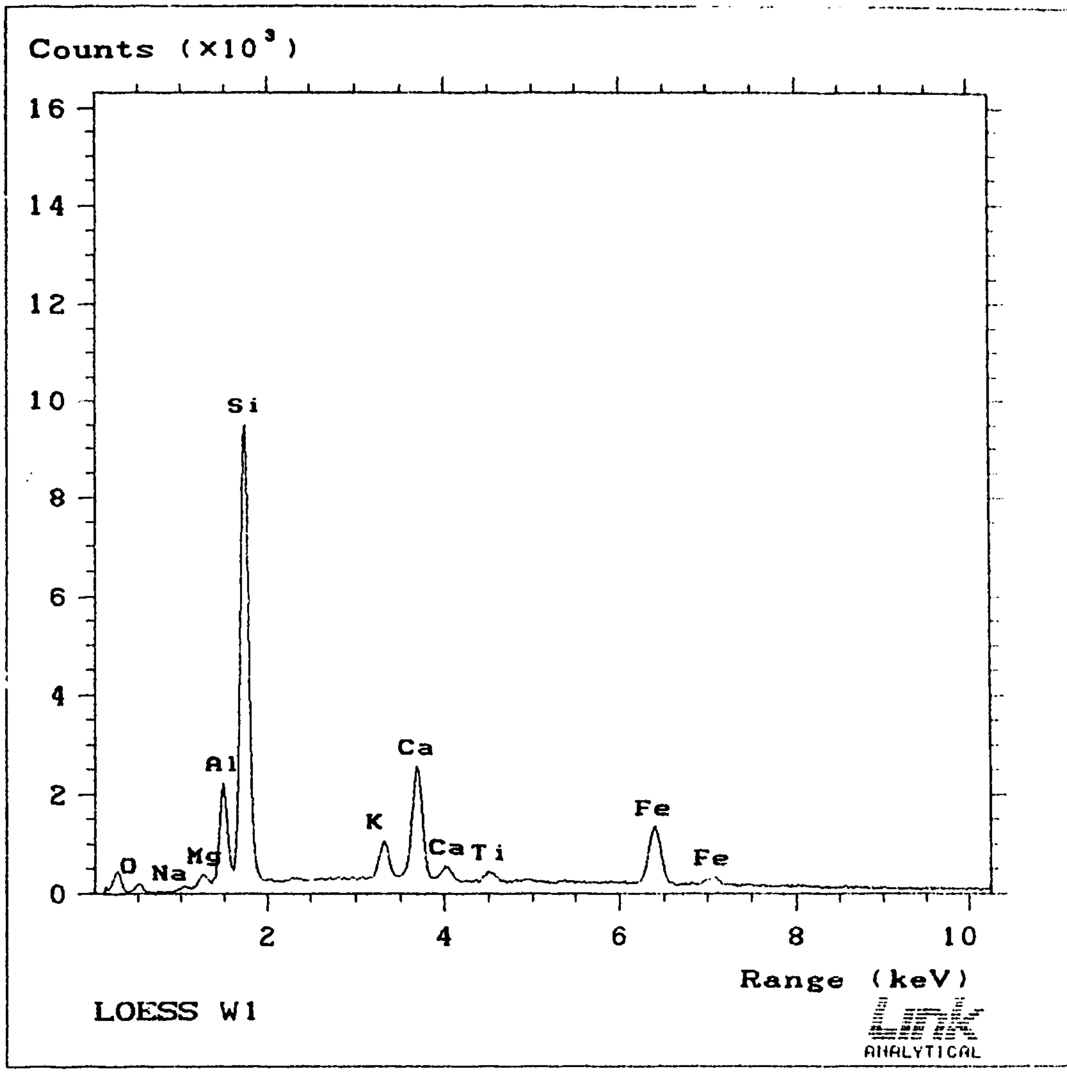


Spectrum: LOESS S2-3 MIN/MET/PET ENG

Last elmt by STOICH.,NORMALISED

ELMT		ZAF	%ELMT	+-	Error	ATOM.%		%OXIDE
NaK :	0	.597	.709	+-	.226	.698	Na2O	.956
MgK :	0	.629	1.453	+-	.155	1.353	MgO	2.410
AlK :	0	.726	6.953	+-	.156	5.833	Al2O3	13.137
SiK :	0	.761	23.158	+-	.180	18.661	SiO2	49.540
K K :	0	.982	2.319	+-	.096	1.343	K2O	2.794
CaK :	0	.951	10.104	+-	.145	5.706	CaO	14.138
TiK :	0	.815	.794	+-	.103	.375	TiO2	1.325
FeK :	0	.842	10.982	+-	.274	4.451	Fe2O3	15.702
O K :	0	.000	43.528			61.579		
TOTAL			100.002			100.000		100.002

Figure 7/ Table 4. Relative percentages of element and mineral distribution in the late last glacial (Sartan / Late Wisconsinan) loess. EDX SEM analysis.

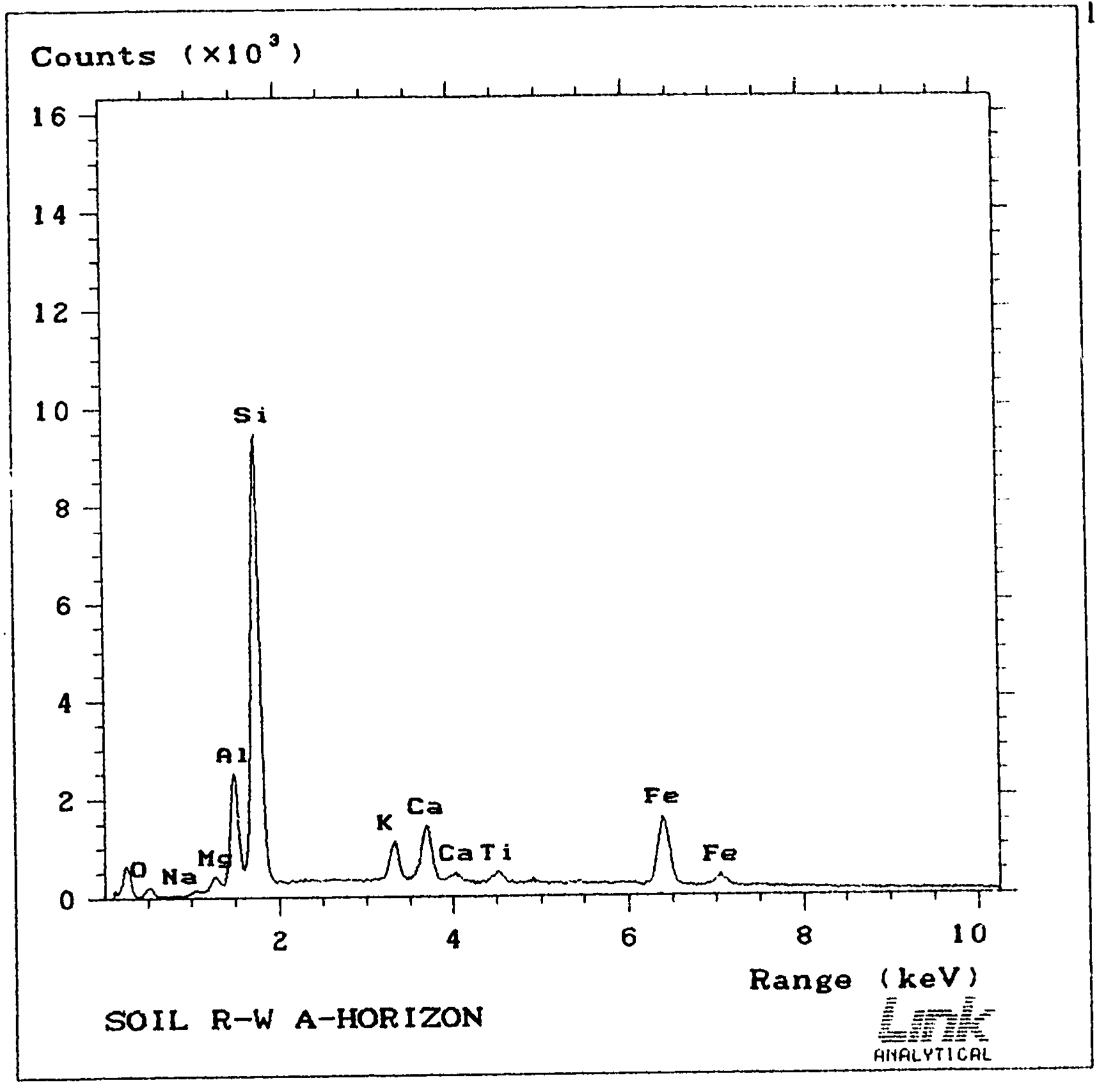


Spectrum: LOESS W1 MIN/MET/PET ENG

Last elmt by STOICH.,NORMALISED

ELMT		ZAF	%ELMT	+-	Error	ATOM.%		%OXIDE
NaK	: 0	.606	.955	+-	.217	.929	Na2O	1.288
MgK	: 0	.635	1.217	+-	.145	1.119	MgO	2.018
AlK	: 0	.734	6.463	+-	.151	5.353	Al2O3	12.213
SiK	: 0	.771	25.374	+-	.185	20.186	SiO2	54.280
K K	: 0	.972	2.425	+-	.094	1.386	K2O	2.921
CaK	: 0	.945	7.726	+-	.133	4.307	CaO	10.810
TiK	: 0	.818	.908	+-	.100	.424	TiO2	1.515
FeK	: 0	.843	10.462	+-	.272	4.186	Fe2O3	14.957
O K	: 0	.000	44.471			62.111		
TOTAL			100.002			100.000		100.002

Figure 8/ Table 5. Relative percentages of element and mineral distribution in the early last glacial (Zyryansk / Early Wisconsinan) loess. EDX SEM analysis.



Spectrum: SOIL R-W A-HORIZON MIN/MET/PET ENG

Last elmt by STOICH.,NORMALISED

ELMT		ZAF	%ELMT	+-	Error	ATOM.%		%OXIDE
NaK	: 0	.602	.884	+-	.221	.855	Na2O	1.192
MgK	: 0	.632	1.415	+-	.153	1.294	MgO	2.347
AlK	: 0	.729	7.797	+-	.161	6.422	Al2O3	14.734
SiK	: 0	.754	25.389	+-	.189	20.083	SiO2	54.312
K K	: 0	.963	2.579	+-	.097	1.465	K2O	3.106
CaK	: 0	.942	3.692	+-	.110	2.047	CaO	5.166
TiK	: 0	.831	.943	+-	.103	.438	TiO2	1.573
FeK	: 0	.847	12.290	+-	.284	4.890	Fe2O3	17.572
O K	: 0	.000	45.012			62.508		
TOTAL			100.002			100.000		100.002

Figure 9 / Table 6. Relative percentages of element and mineral distribution in the last interglacial Chernozem (A horizon). EDX SEM analysis.

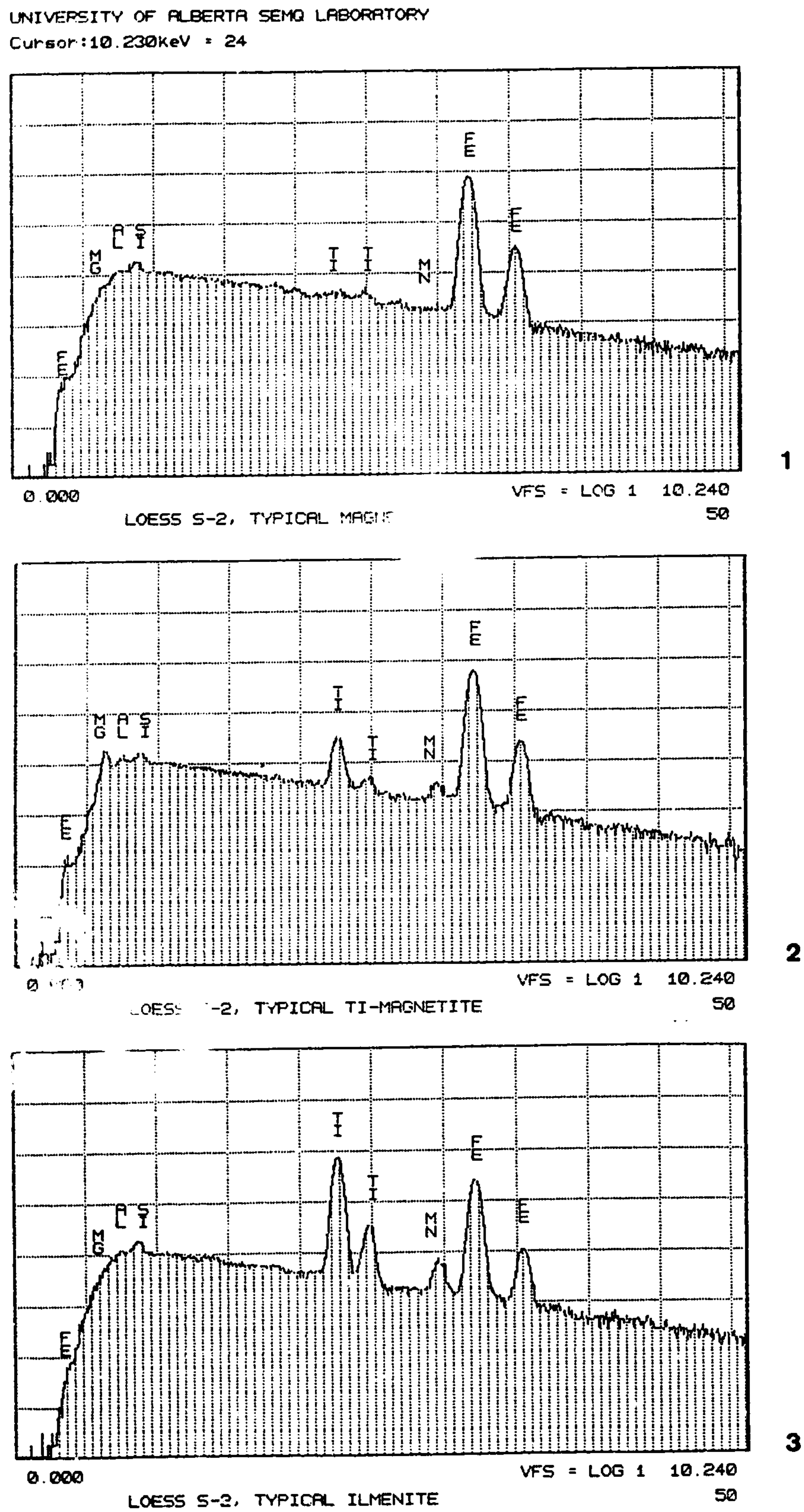


Figure 10. Element composition of typical magnetite (1), titano-magnetite (2), and ilmenite (3) grains from the late last glacial (Late Wisconsinan) loess. ARL SEMQ electron microprobe analysis.

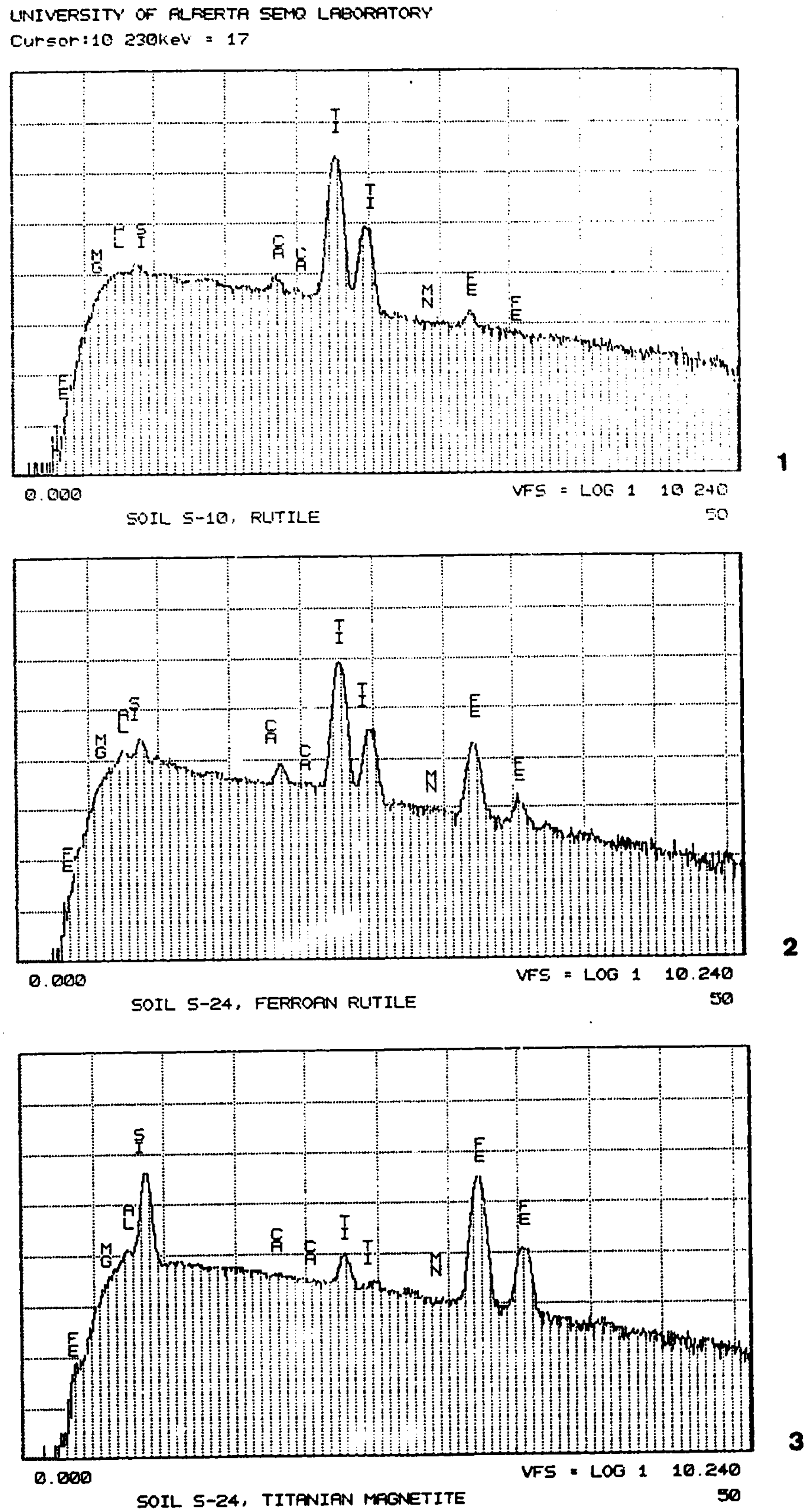
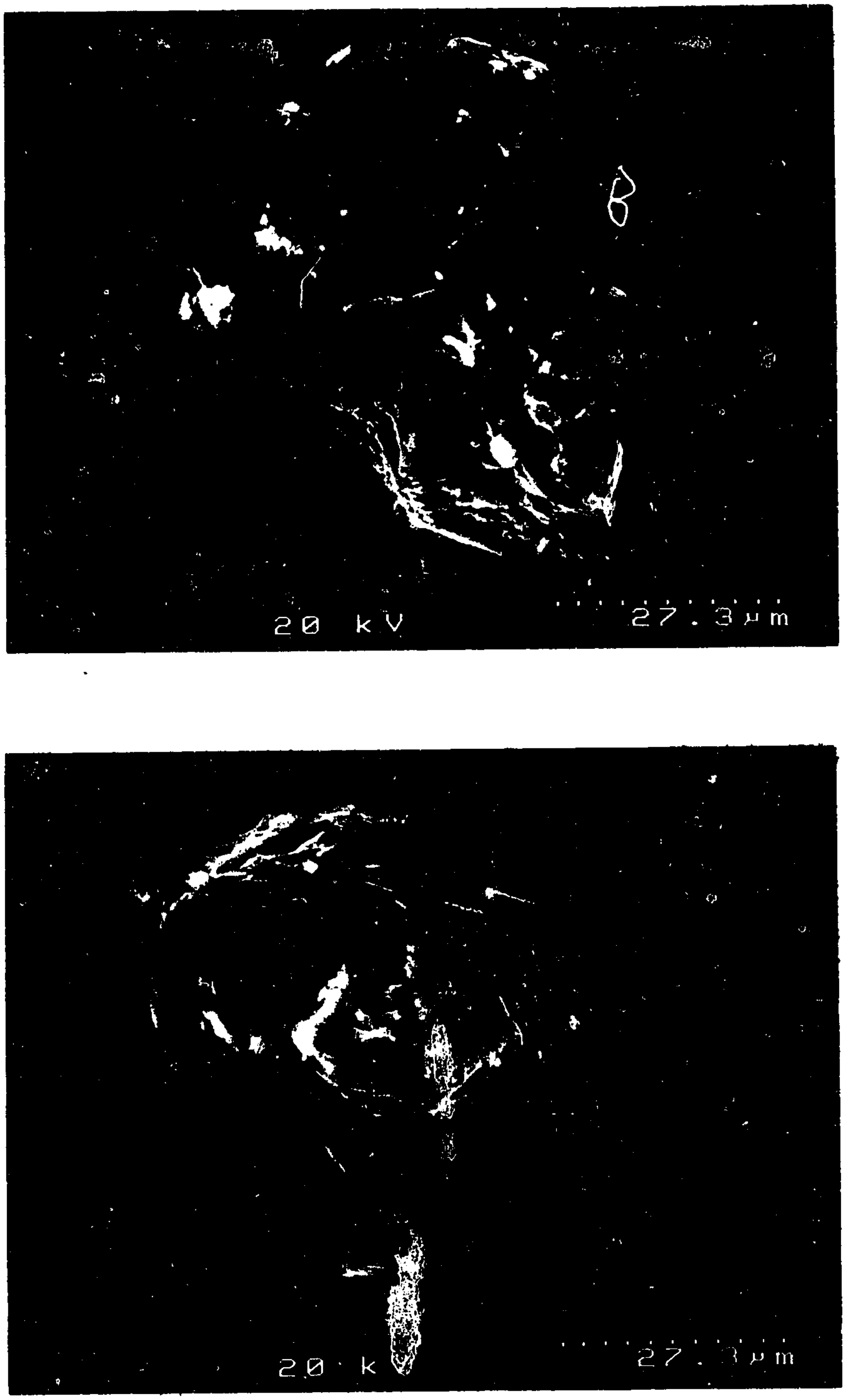


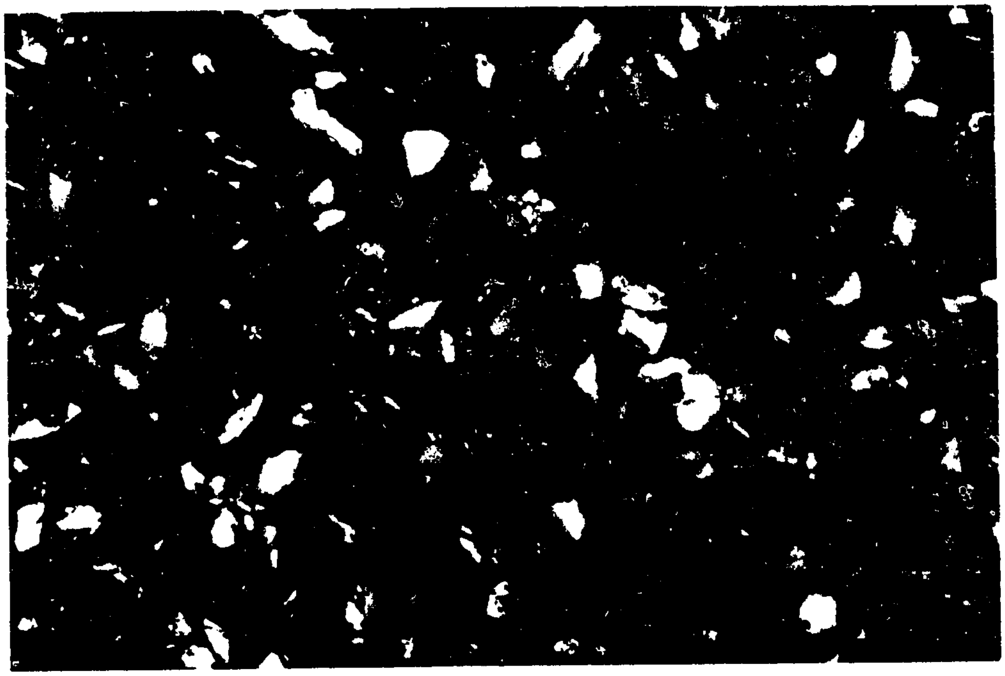
Figure 11. Element composition of typical rutile (1), ferroan rutile (2), and titano-magnetite (3) grains from a mid-last glacial (Mid-Wisconsinan) Brunisolic soil (1), and a Chernozem from the penultimate interglacial (2-3). ARL SEMQ electron microprobe analysis.



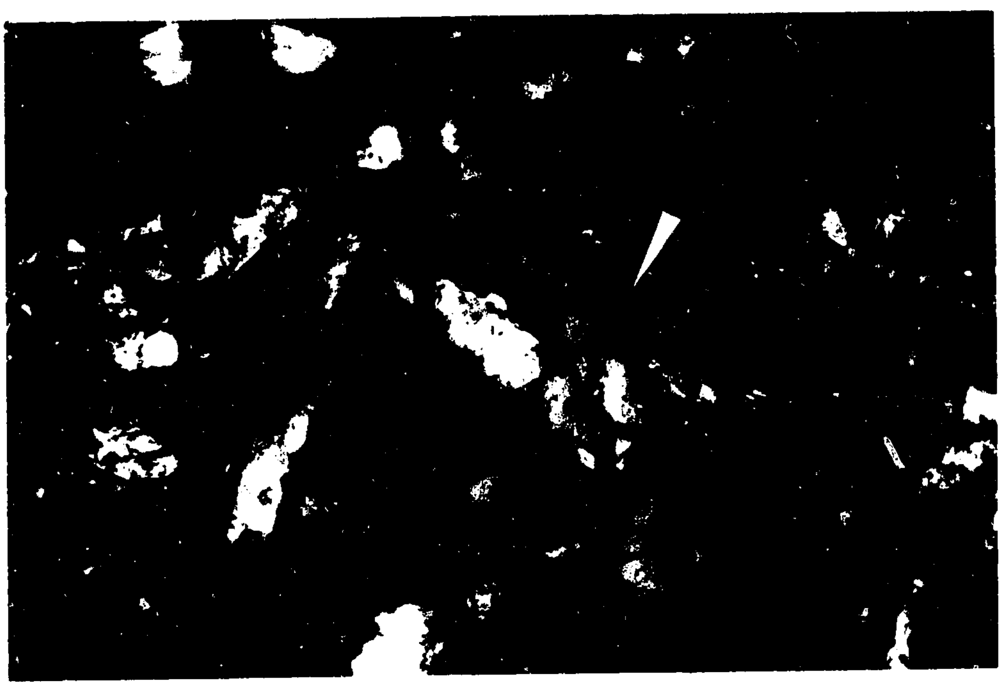
A

B

Figure 12. SEM photographs of typical quartz (A) and iron mineral (B) grains from the early last glacial loess. Presence of smaller subrounded quartz grains seen in the background of the upper photograph suggests a mixed origin of the loess, predominantly formed by local materials. The limited surface abrasion of both grains indicates short-distance transport of the sediment. Magnification 1100x.

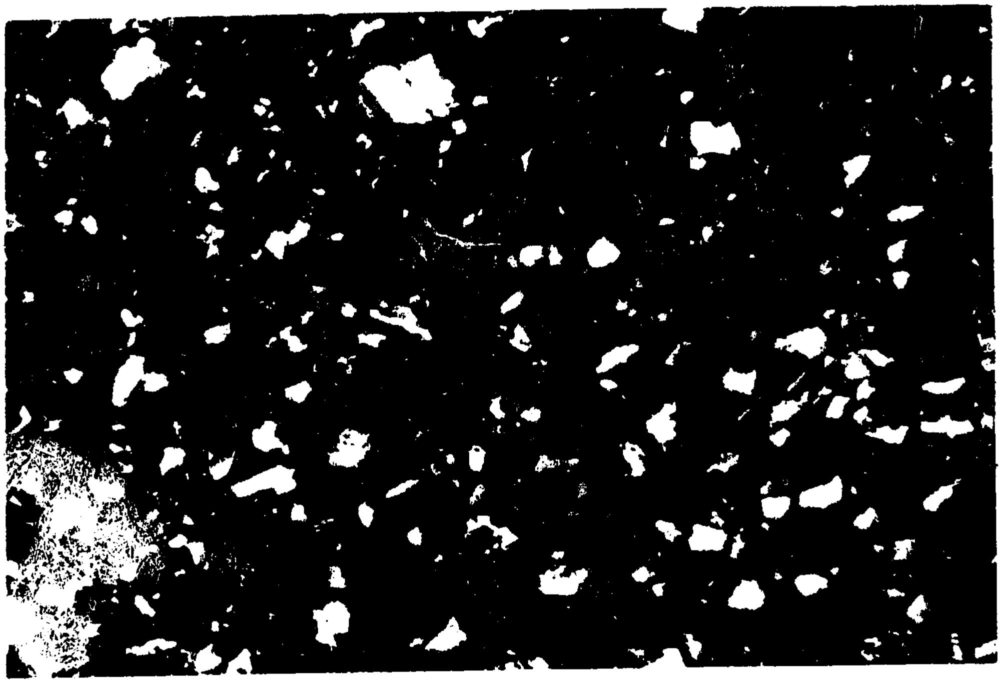


A



B

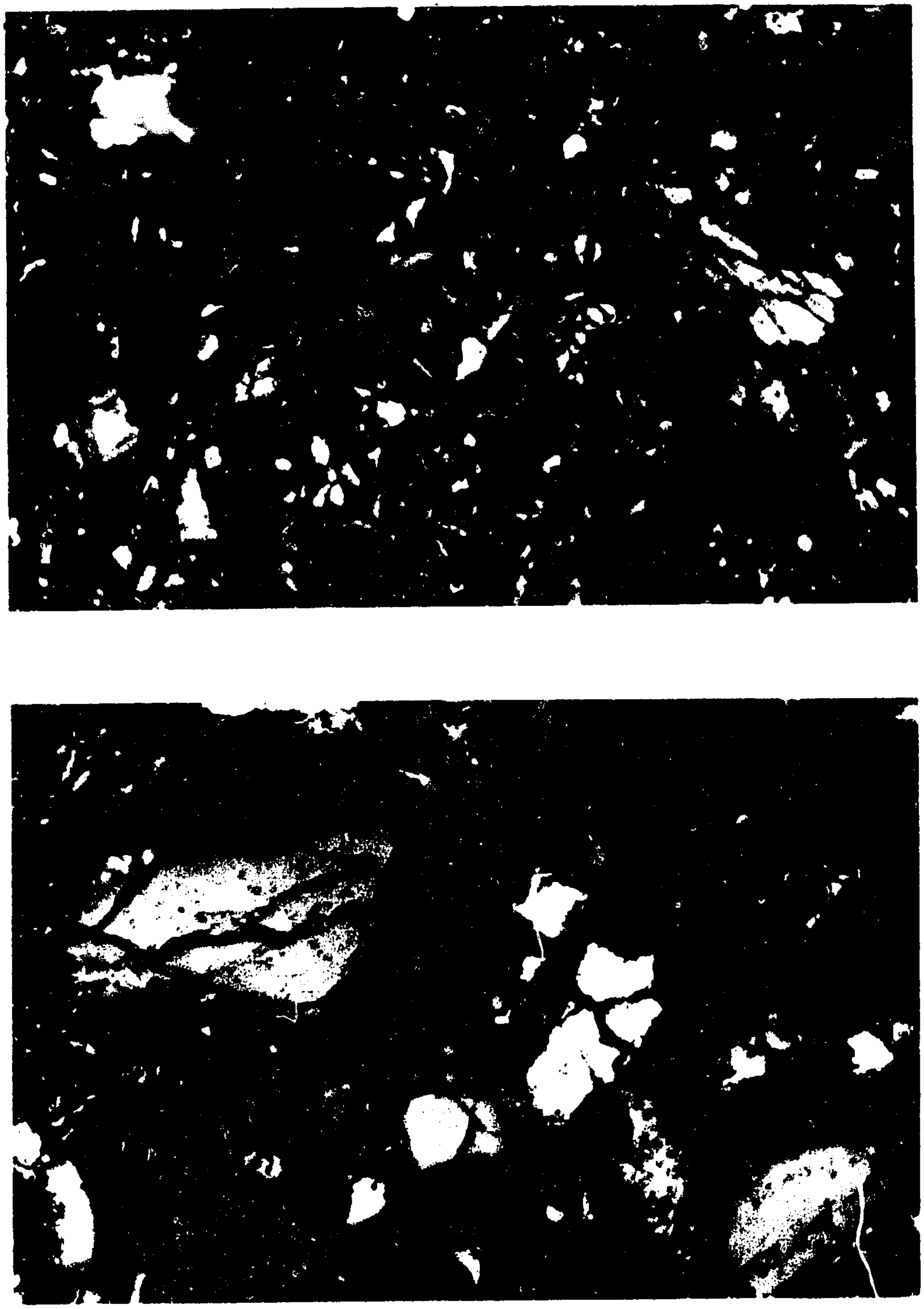
Figure 13. Thin-section photographs of the typical loess deposited during the last glacial maximum, showing rather fresh non-abraded and unweathered mineral grains, including the less resistant ones as amphibole (indicated by the arrow). Magnification: (A) 62.5x = 1.2 mm wide section view; (B) 156x = 0.64 mm wide section view.



A

B

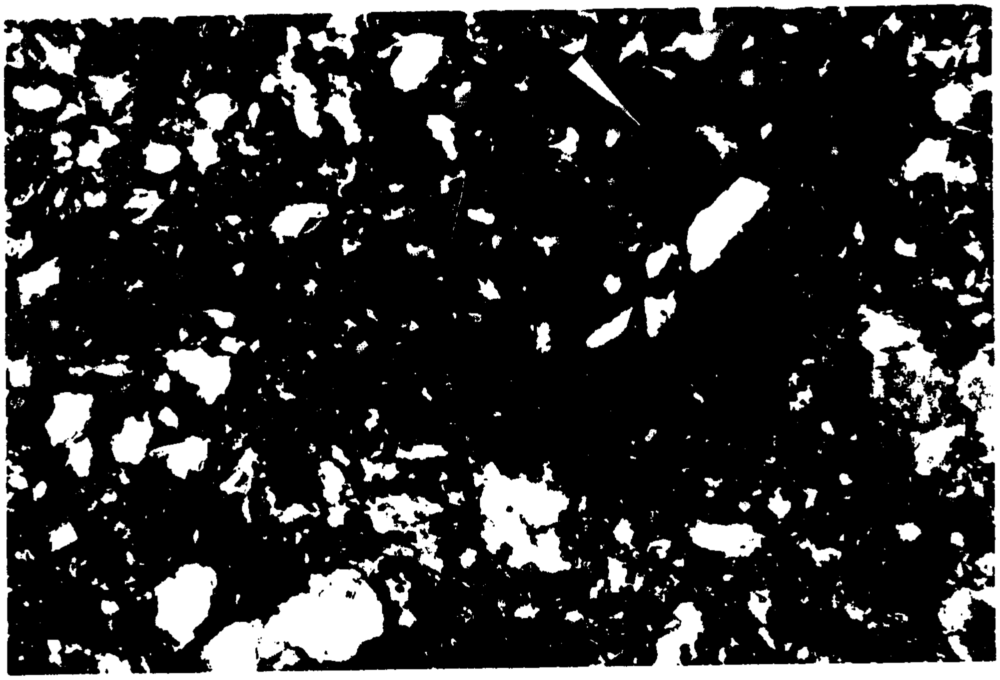
Figure 14. Thin-section photographs of the niveoaeolian (colluviated) loess facies from the lower part of the section between two interglacial pedocomplexes. A vertical rearrangement of grains (A) and grain clustering (B) suggest processes following deposition of the loessic dust on a subsequently melted show cover. Magnification: (A) 62.5x; (B) 156x.



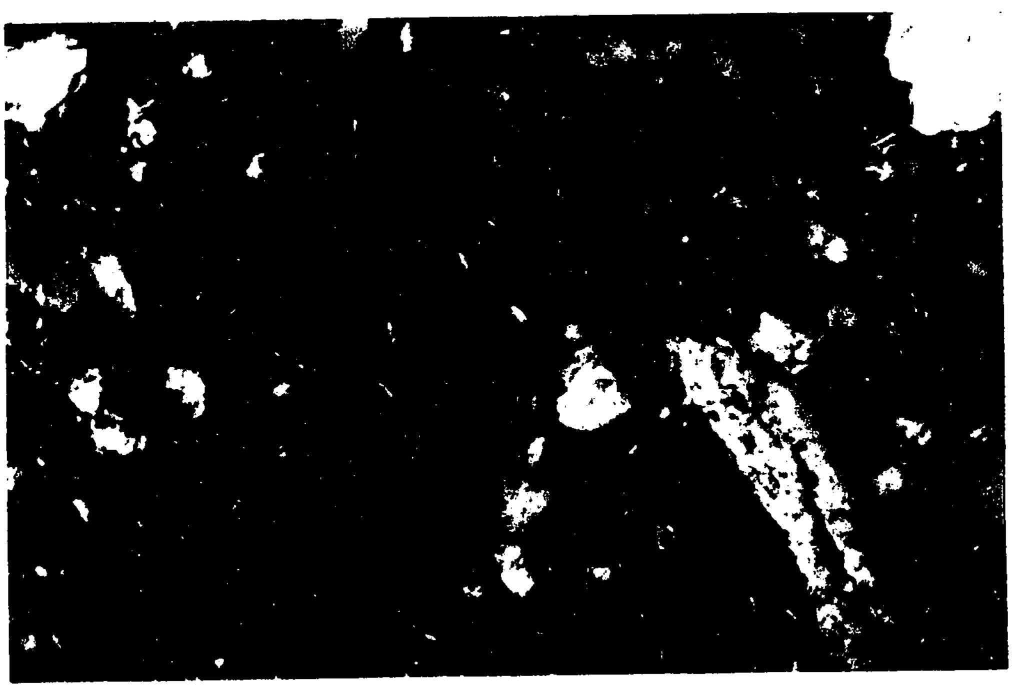
A

B

Figure 15. Thin-section photographs of frost-shattered grains (mainly quartz) from frost-wedge casts in the A horizon of a Chernozemic last interglacial soil. Fragmentation due to freezing of capillary water followed the original decompression lines inherited from the eroded igneous and metamorphic bedrock. Magnification: (A) 62.5x; (B) 156x.

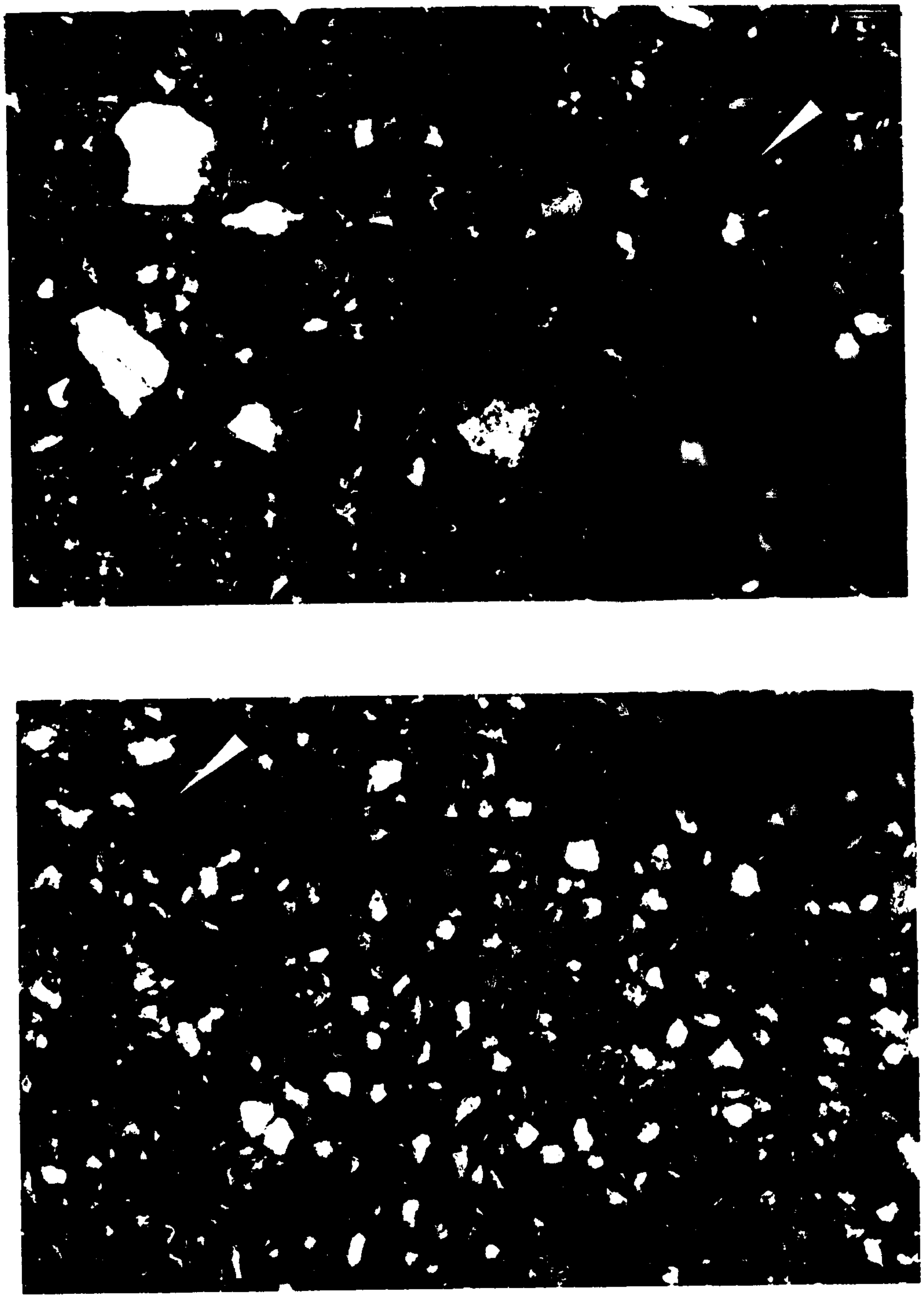


A



B

Figure 16. Thin-section photographs of a Brunisolic Bmkgj horizon of the late mid-last glacial interstadial interval (*ca.* 28-25 ka). (A) Presence of ferric mottles (indicated by the arrow) suggests periodic oxidation and reduction processes due to seasonal ground saturation (62.5x). (B) Detailed view of grains with fresh surface morphology (156x).



A

B

Figure 17. Thin-section photographs of the A horizons of the most developed Chernozemic soil of the last interglacial (A) and a Brunisolic soil from the late penultimate interglacial (B), with concentrations of organic matter (the black mass indicated by the arrow). Note the different size of mineral grains. Magnification: 62.5x.

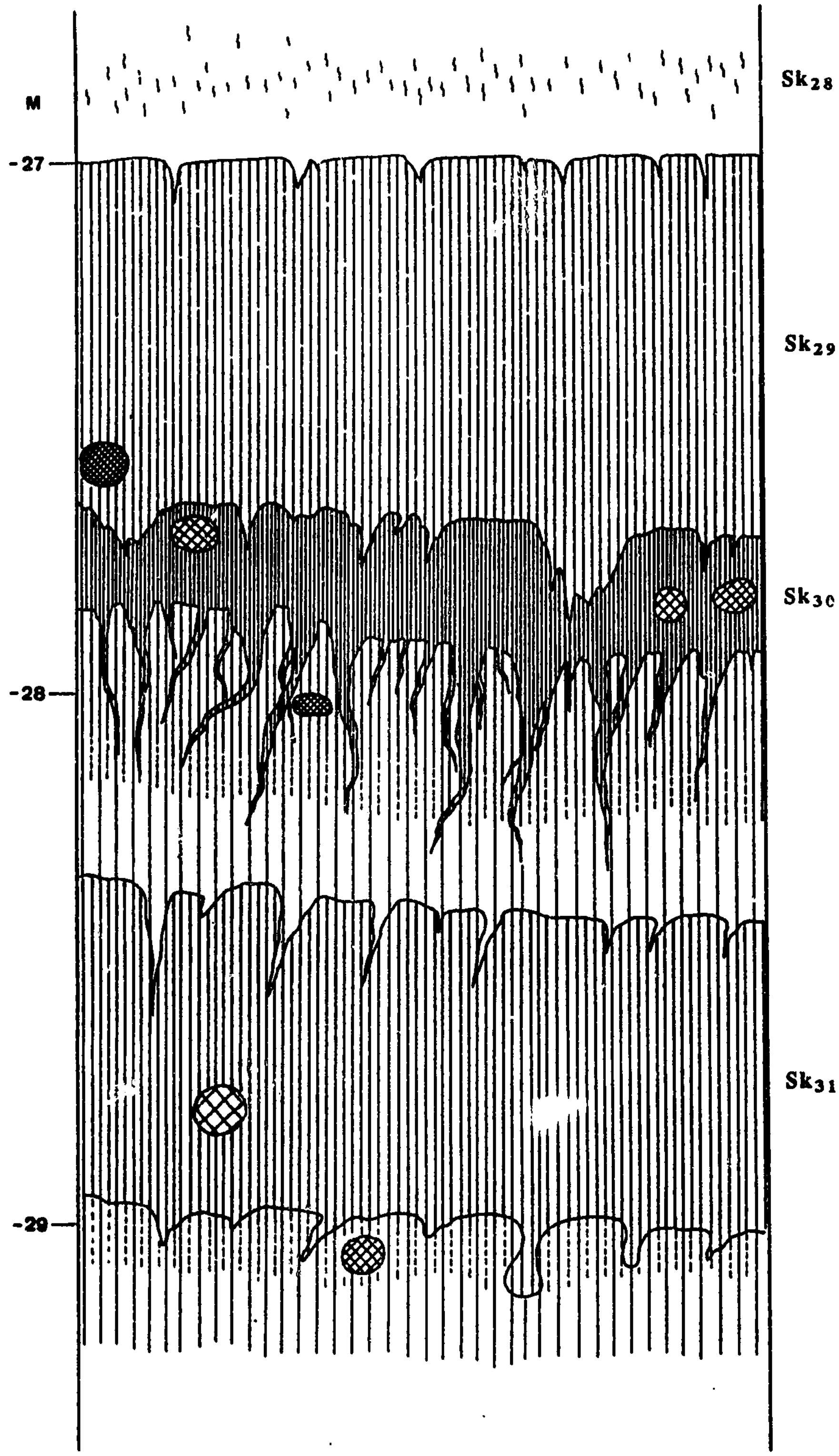


Figure 18. Kurtak Section 29. The penultimate (late Mid-Pleistocene) interglacial pedocomplex (Oxygen Isotope Stage 7). For legend see Figure 19B.

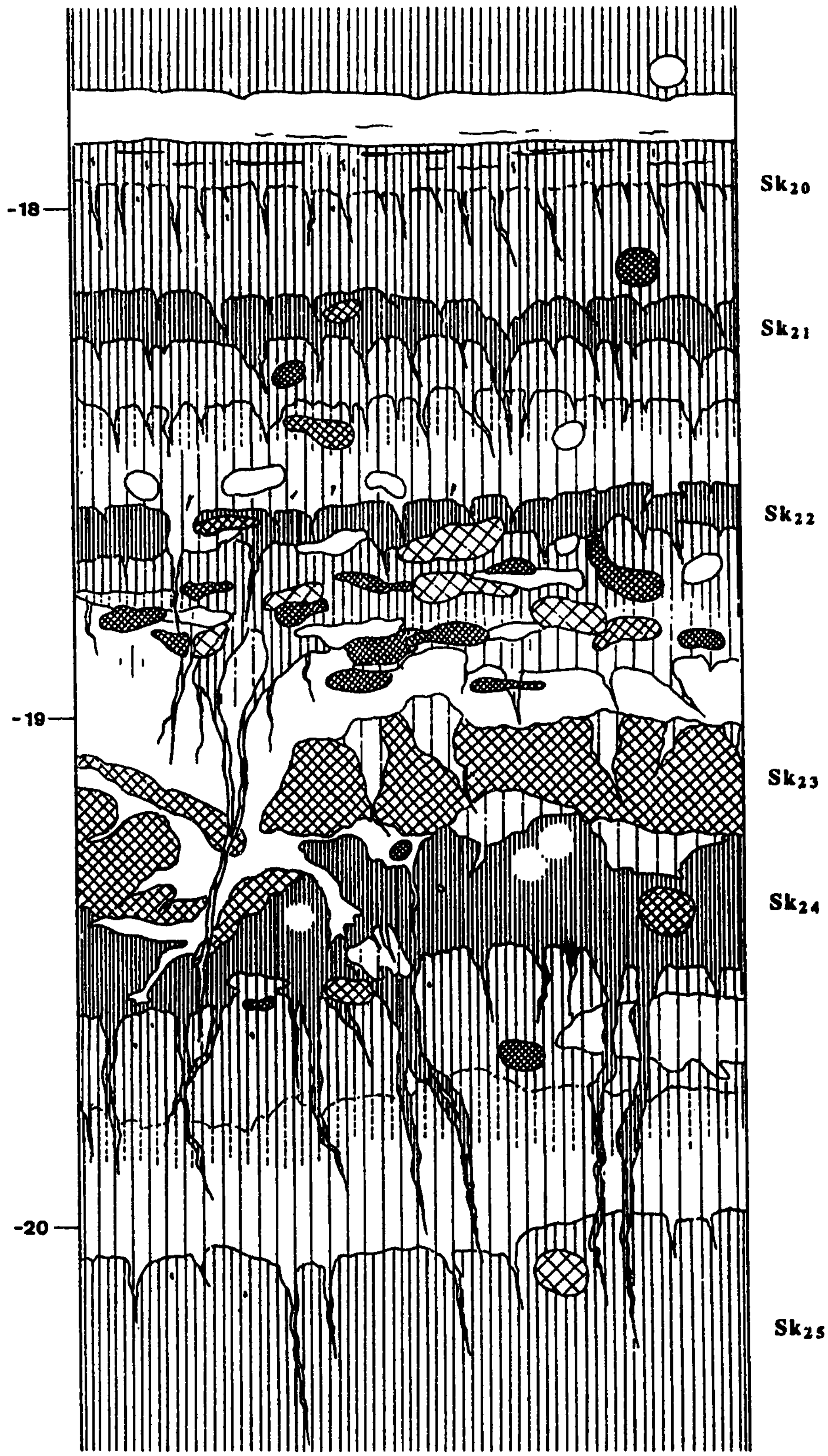
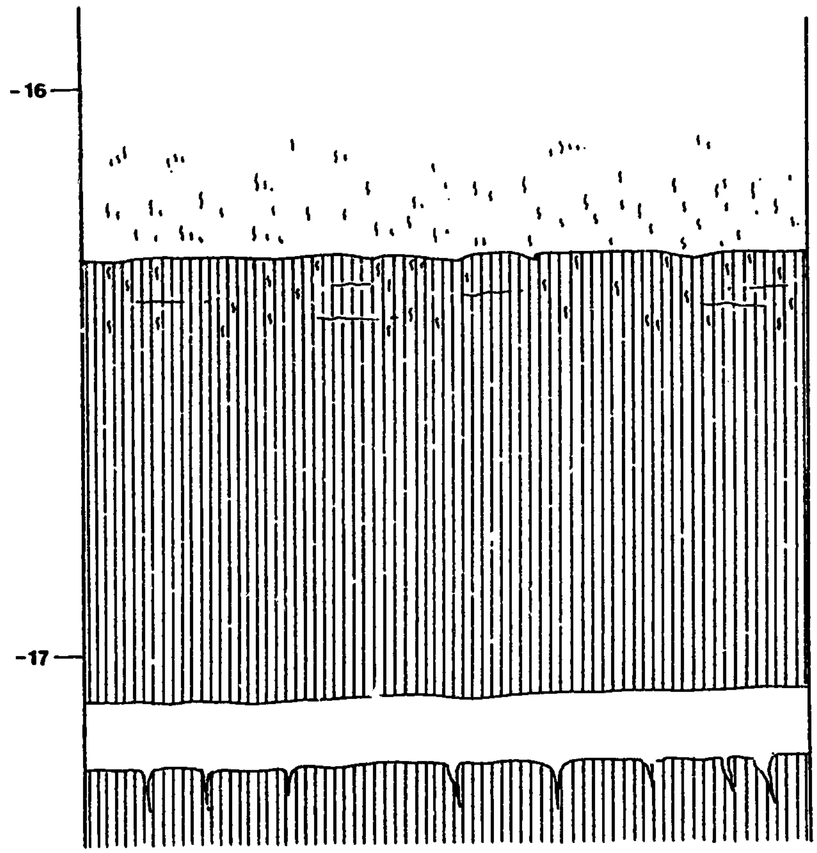


Figure 19A. Kurtak Section 29. The last interglacial pedocomplex (Oxygen Isotope Stage 5; 130-74 ka). For legend see Figure 19B.



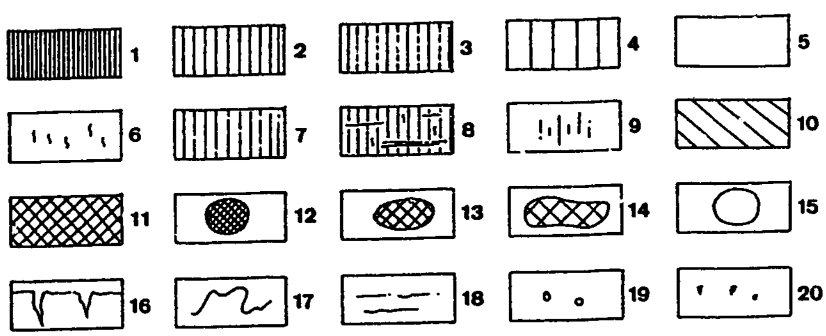


Figure 19B. Kurtak Section 29. The last interglacial pedocomplex (Oxygen Isotope Stage 5; 127-73 ka); con't. Legend: 1 - Chernozemic Ah horizon; 2 - Chernozemic/Brunisolic Bmk horizon; 3 - Chernozemic /Brunisolic BCca horizon; 4 - Chernozemic/Brunisolic Cca horizon; 5 - loess (CK horizon); 6 - gleyed Regosolic Ckg/Ckgj horizon; 7 - gleyed Brunisolic (Bmkgj) horizon; 8 - partly colluviated Brunisolic B horizon; 9 - fragmented Chernozemic B horizon; 10 - undifferentiated infill of truncated surfaces; 11 - pedo- / bioturbated Chernozemic/Brunisolic A/B horizon; 12-15 - krotovinas filled by matter of Ah horizons (12), Bmk horizons (13), BCca/Cca horizons (14) and by pure aeolian loess (15); 16 - frost wedges in Ahy and Bmky horizons; 17 - cryoturbations; 18 - minor solifluctions; 19 - small pebbles; 20 - charcoal.

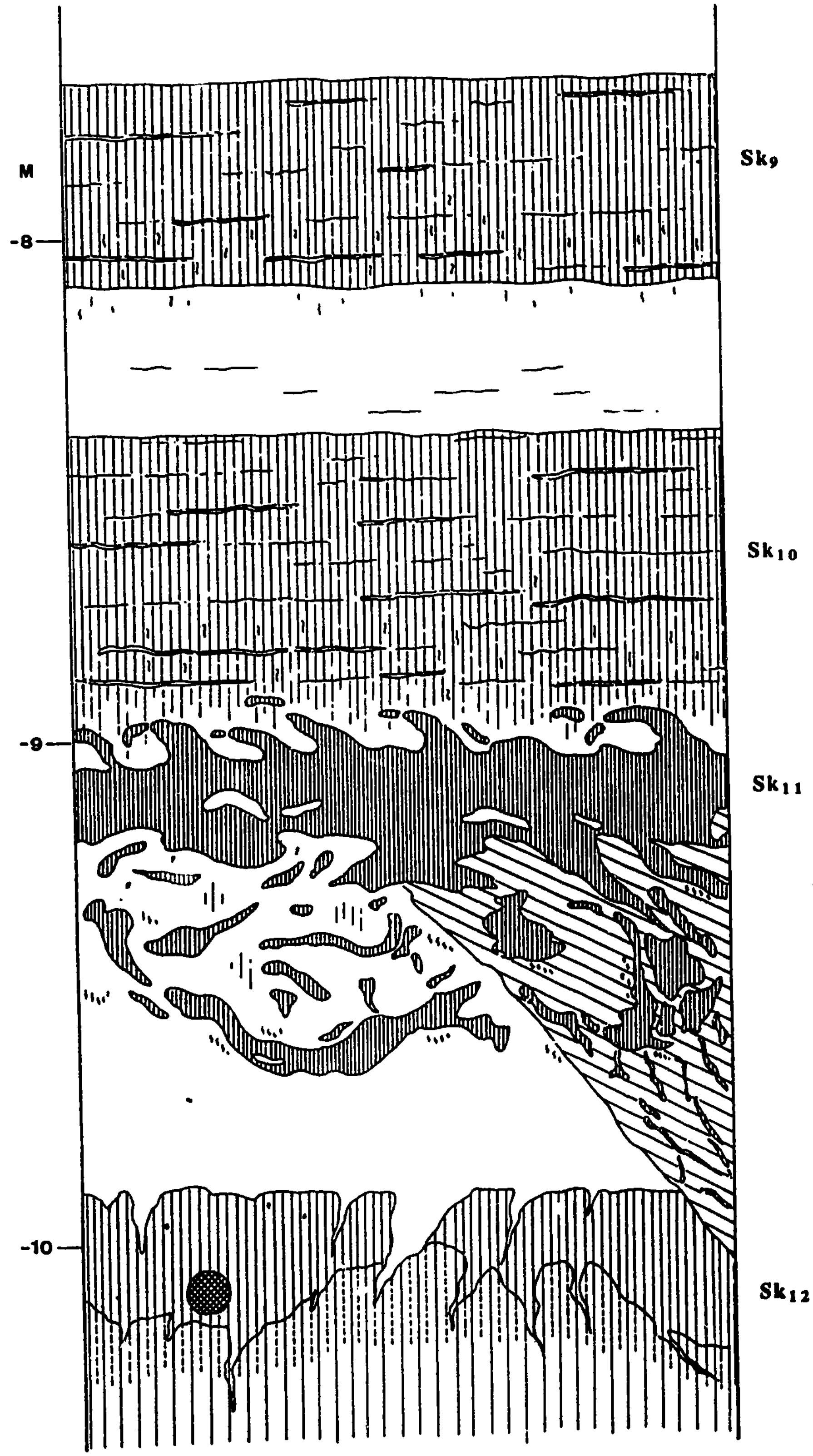


Figure 20. Kurtak Section 29. The Karginsk (Mid-Wisconsinan) pedocomplex (Oxygen Isotope Stage 3; 62-25 ka). For legend see Figure 19B.

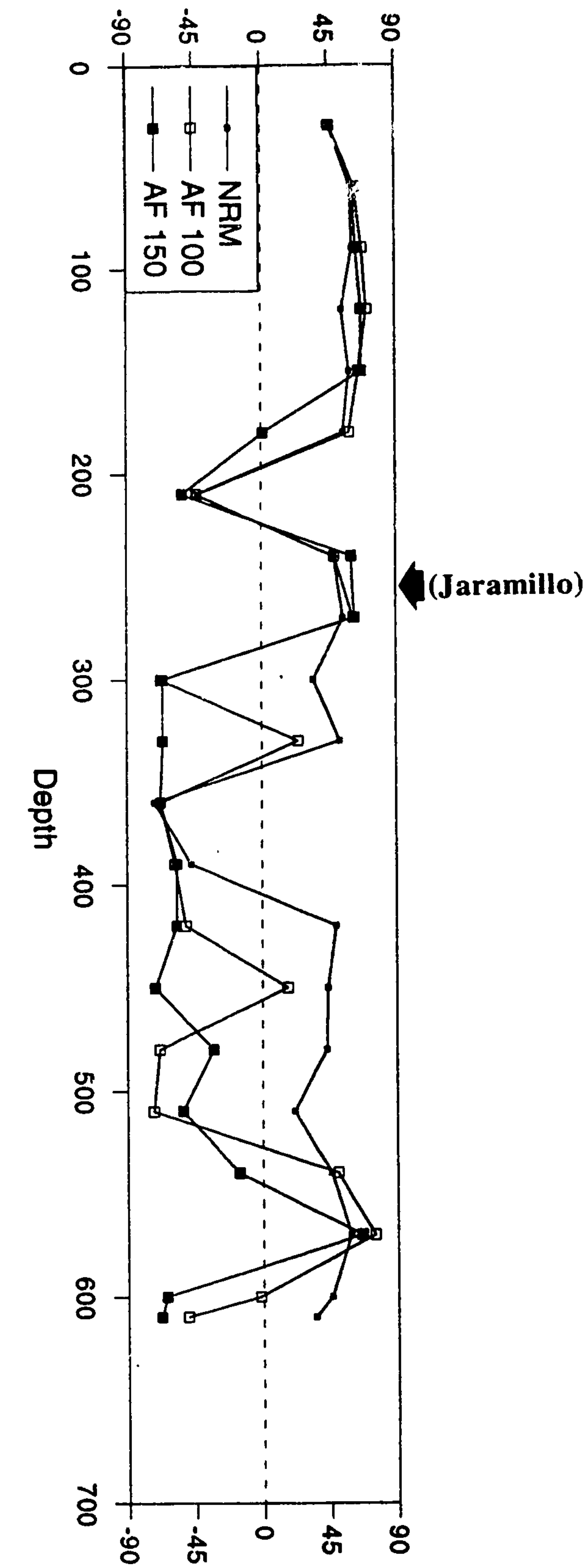


Figure 21. Kurtak Section 22. Palaeomagnetic (remanence) curve with a normal polarity interval of the Jaramillo Event (0.90-0.98 Ma) around the 250 cm depth level (indicated by the arrow) within the Matuyama Chron (defined by the polarity reversal at the 200 cm depth level).

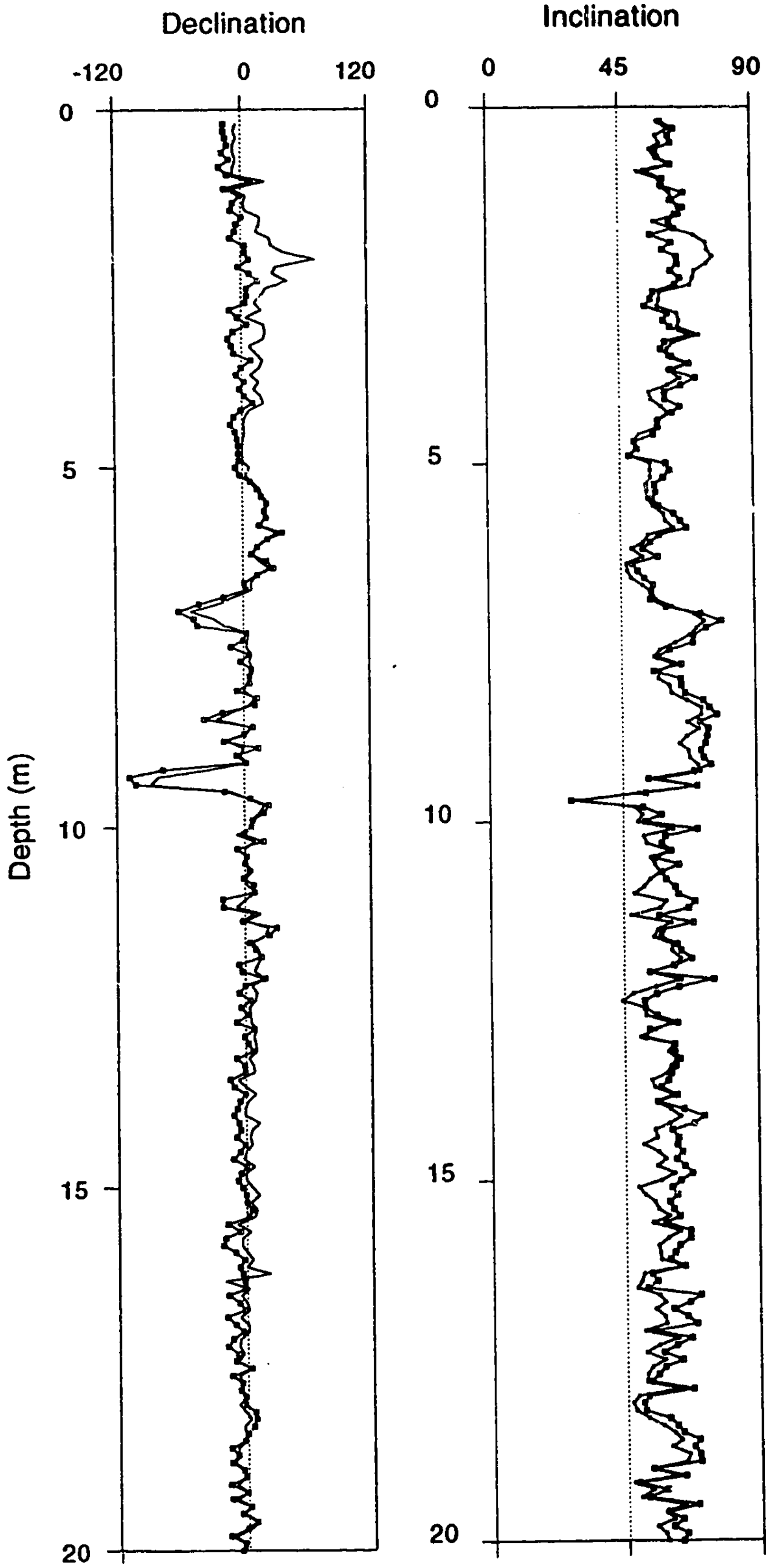


Figure 22A. Kurtak Section 29. Palaeomagnetic (remanence) curve with two recorded magnetic anomalies within the 5-10 m stratigraphic interval.

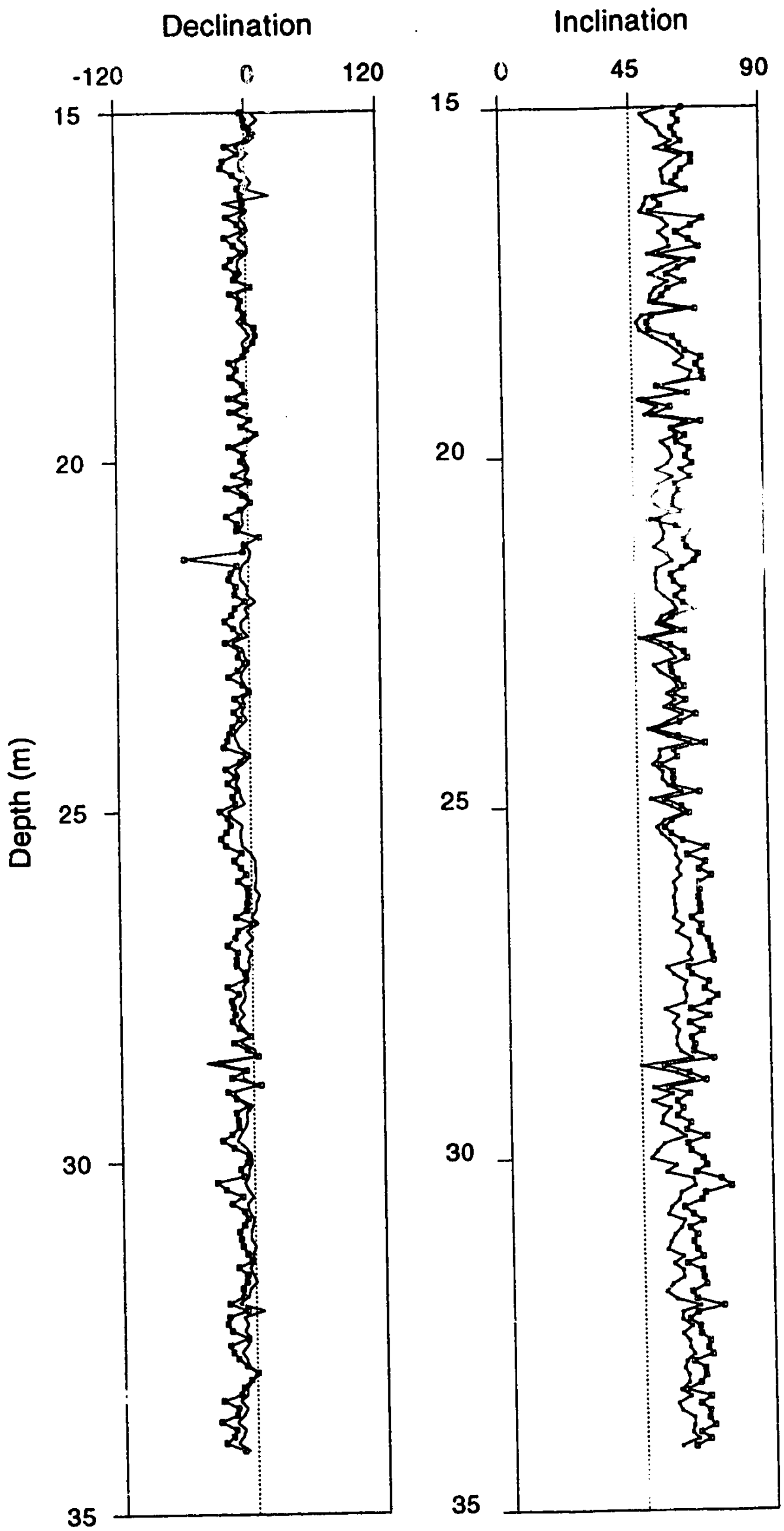
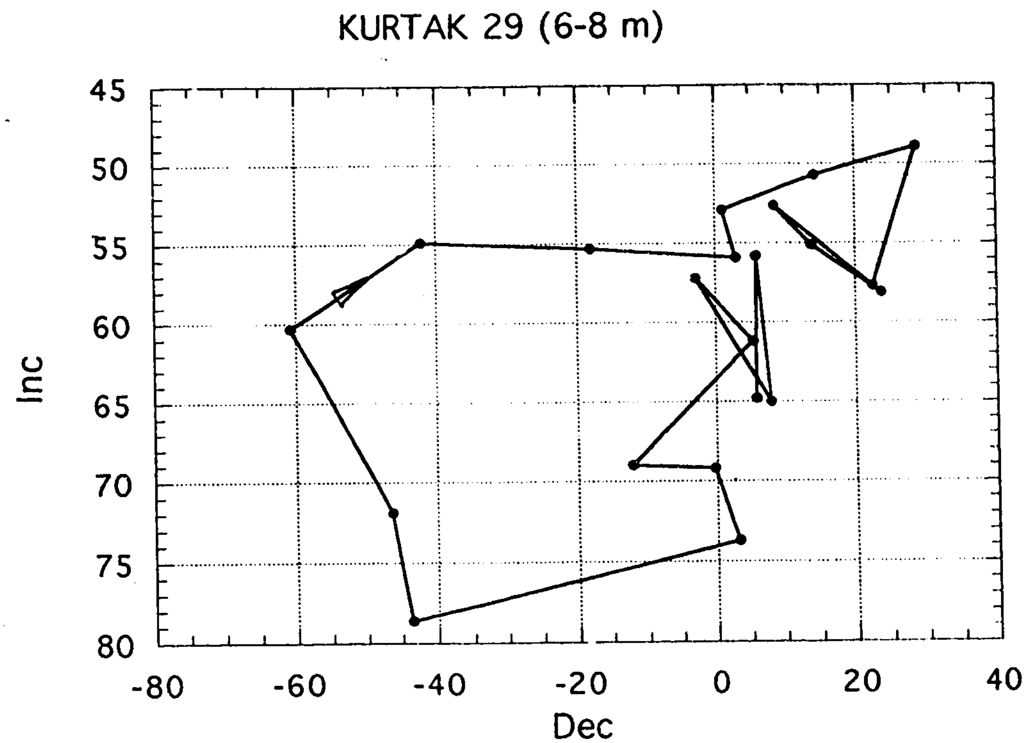
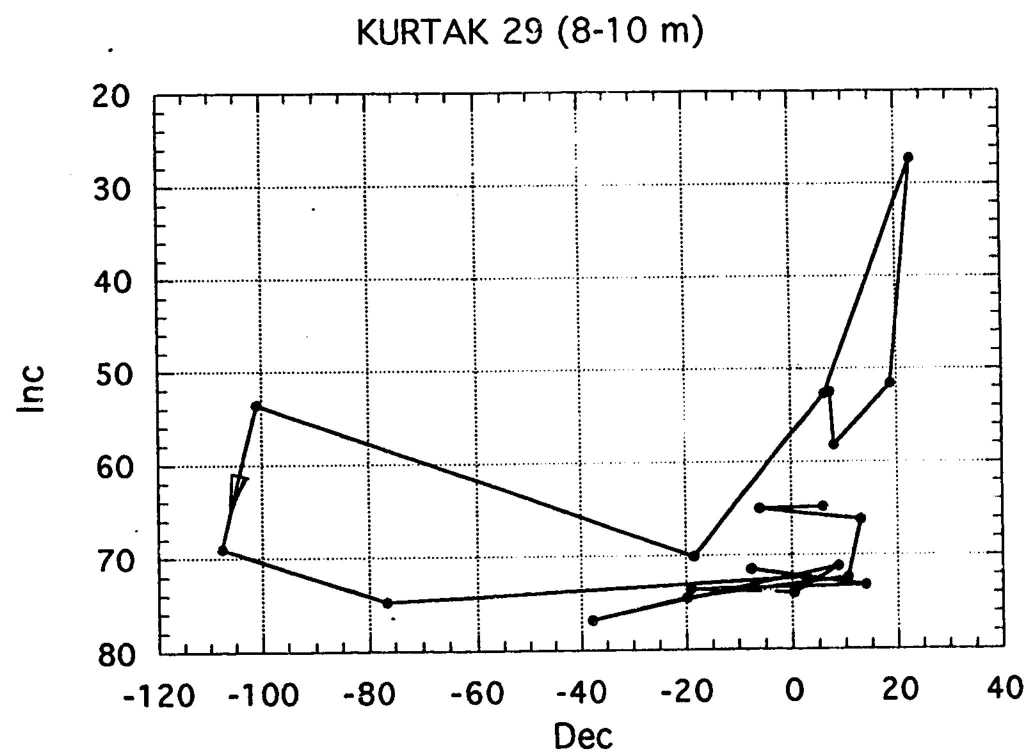


Figure 22B. Kurtak Section 29. Palaeomagnetic (remanence) curve (cont').



A



B

Figure 23. Kurtak Section 29. Magnetic vectors after 10 mT AF cleaning of the upper anomaly (A) and the lower anomaly (B) recorded at -7 m and -9.5 m, respectively. Arrows indicate orientation of the directional pattern of declination and inclination values in respect to the present magnetic field.

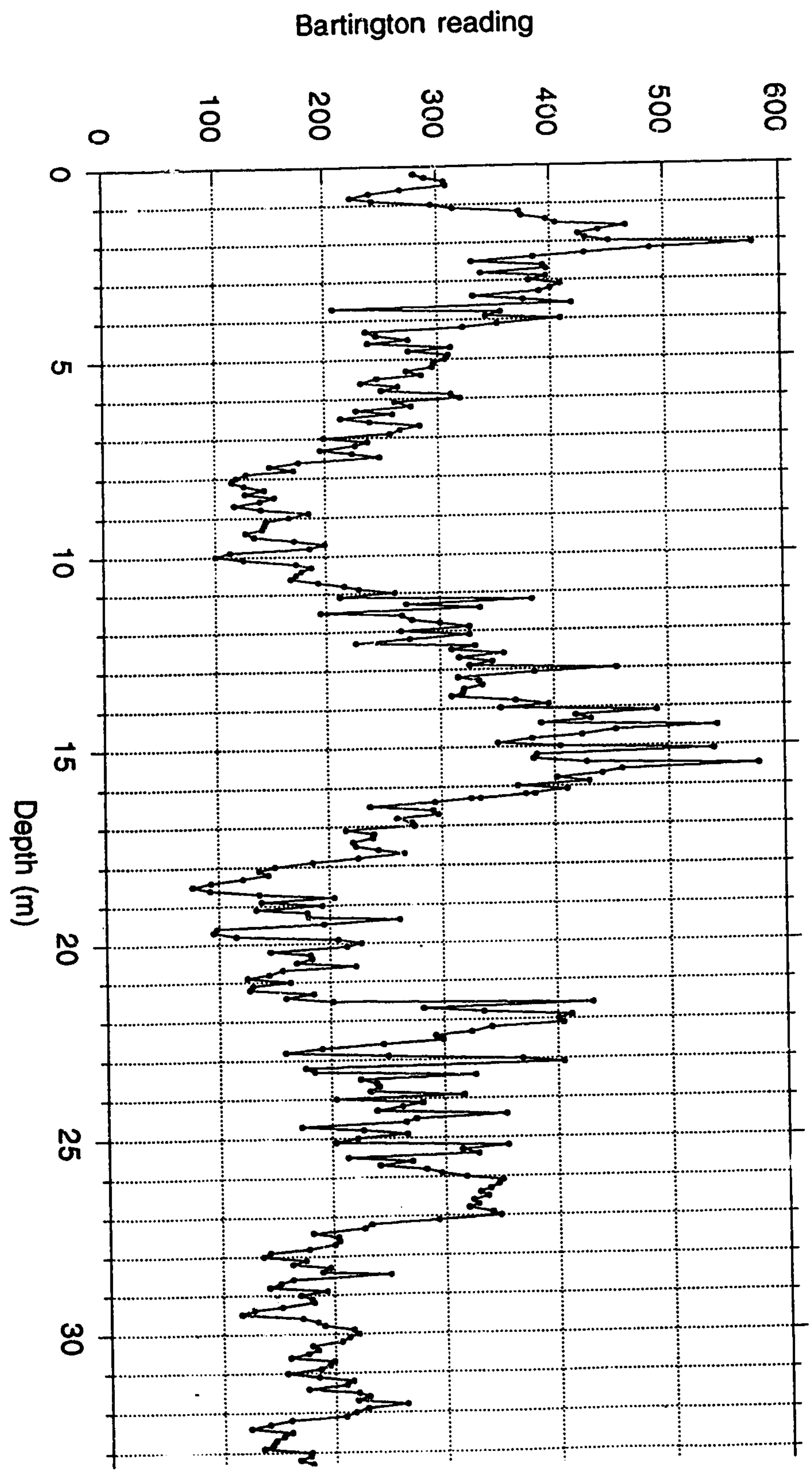


Figure 24. Kurtak Section 29. Magnetic susceptibility curve of the loess-palaeosol record.

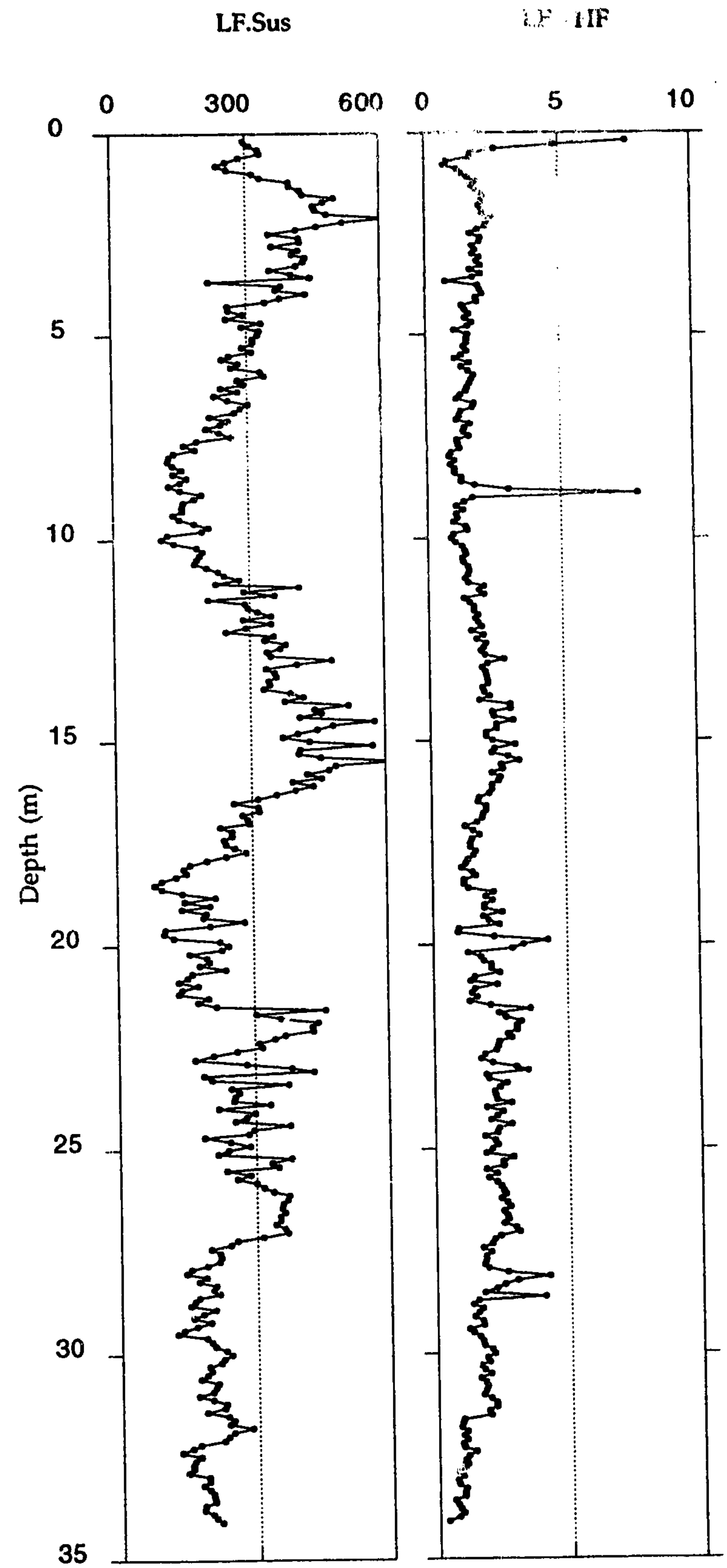
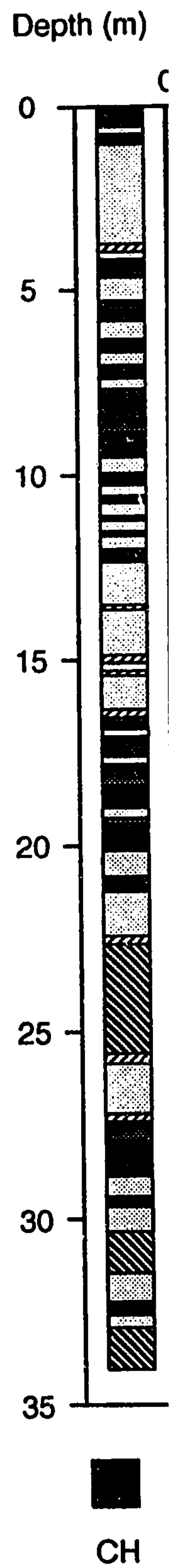
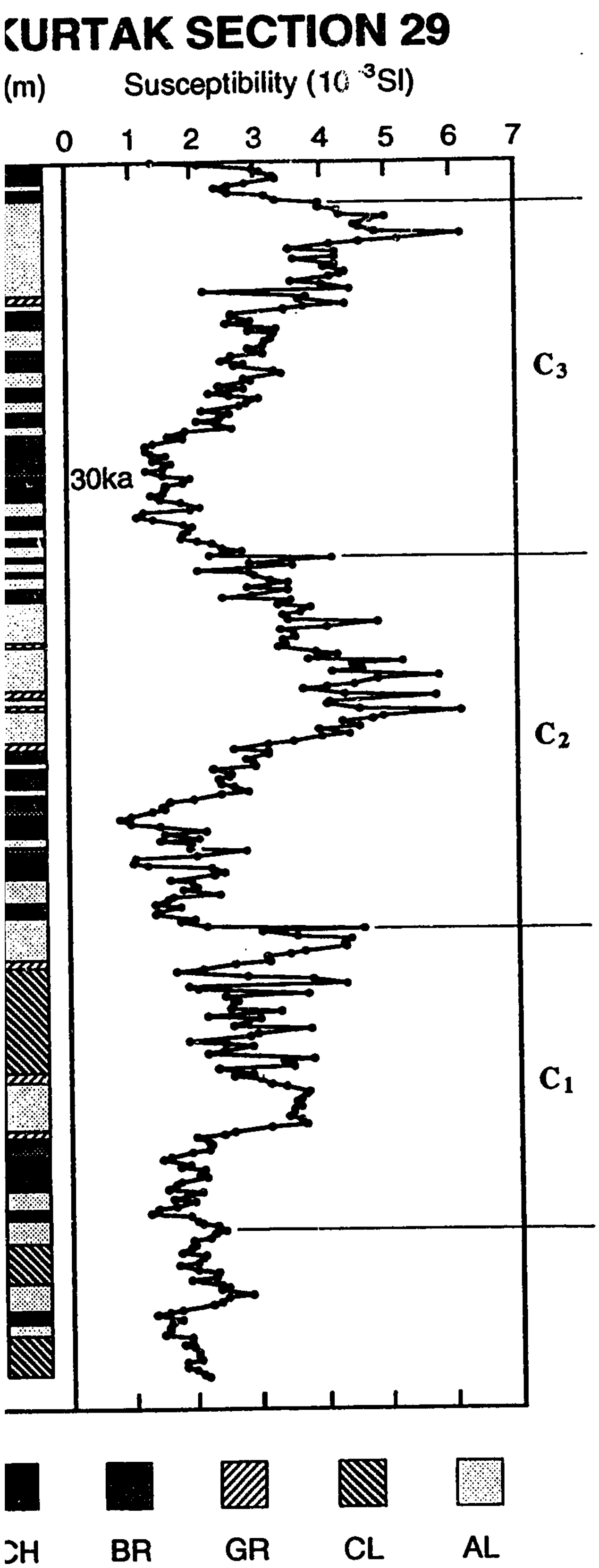


Figure 25. Kurtak Section 29. Bifrequential distribution of magnetic susceptibility values.

KUI



Section 29. Correl
e. Legend: CH - C
AL - aeolian loess



Correlation of the loess-palaeosol record and the corresponding magnetic
CH - Chernozems; BR - Brunisols; GR - Gleyed Regosols; CL -
loess. C_1, C_2, C_3 - palaeoclimatic cycles.

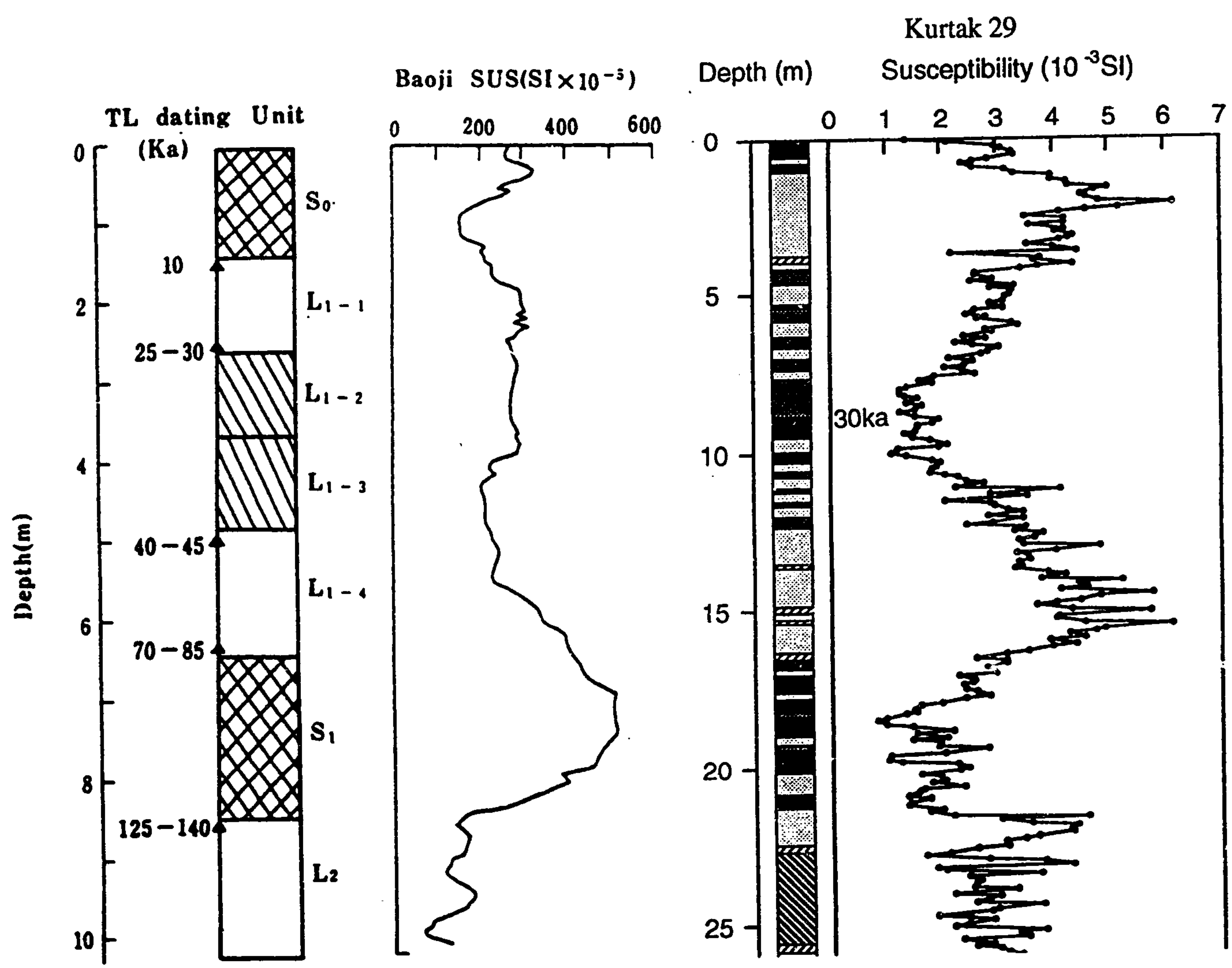


Figure 27. Comparison of the loess-palaeosol records and the magnetic susceptibility curves in the Baoji Section, China (Imbrie *et al.* 1984; Lu *et al.* 1987) and the Kurtak Section 29, for about the last 150 ka.

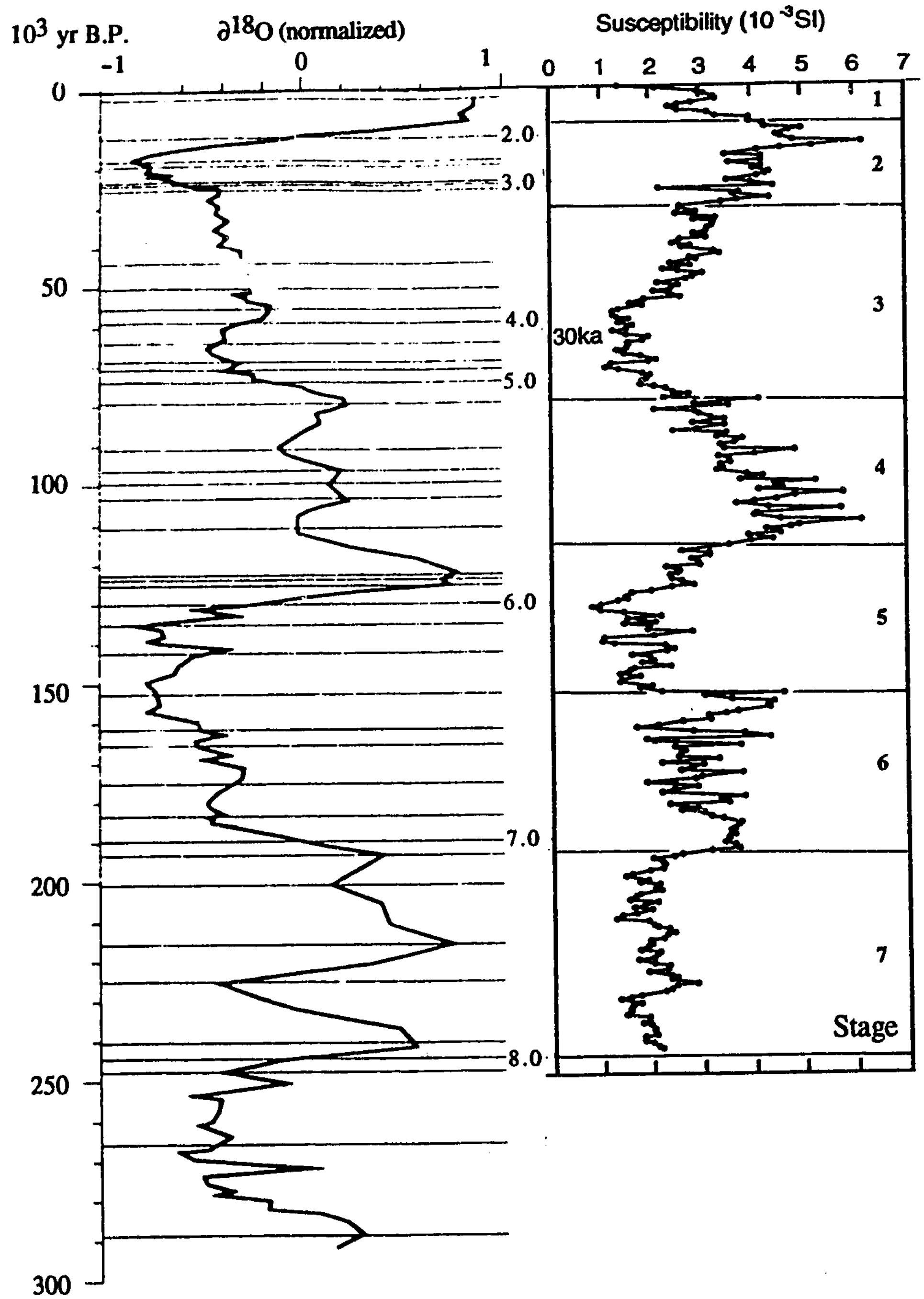


Figure 28. Correlation of the orbitally based chronostratigraphy - stacked oxygen isotype deep-sea record (Pisias *et al.* 1984, modified by Martinson *et al.* 1987) with the Kurtak Section 29 loess-palaeosol record.

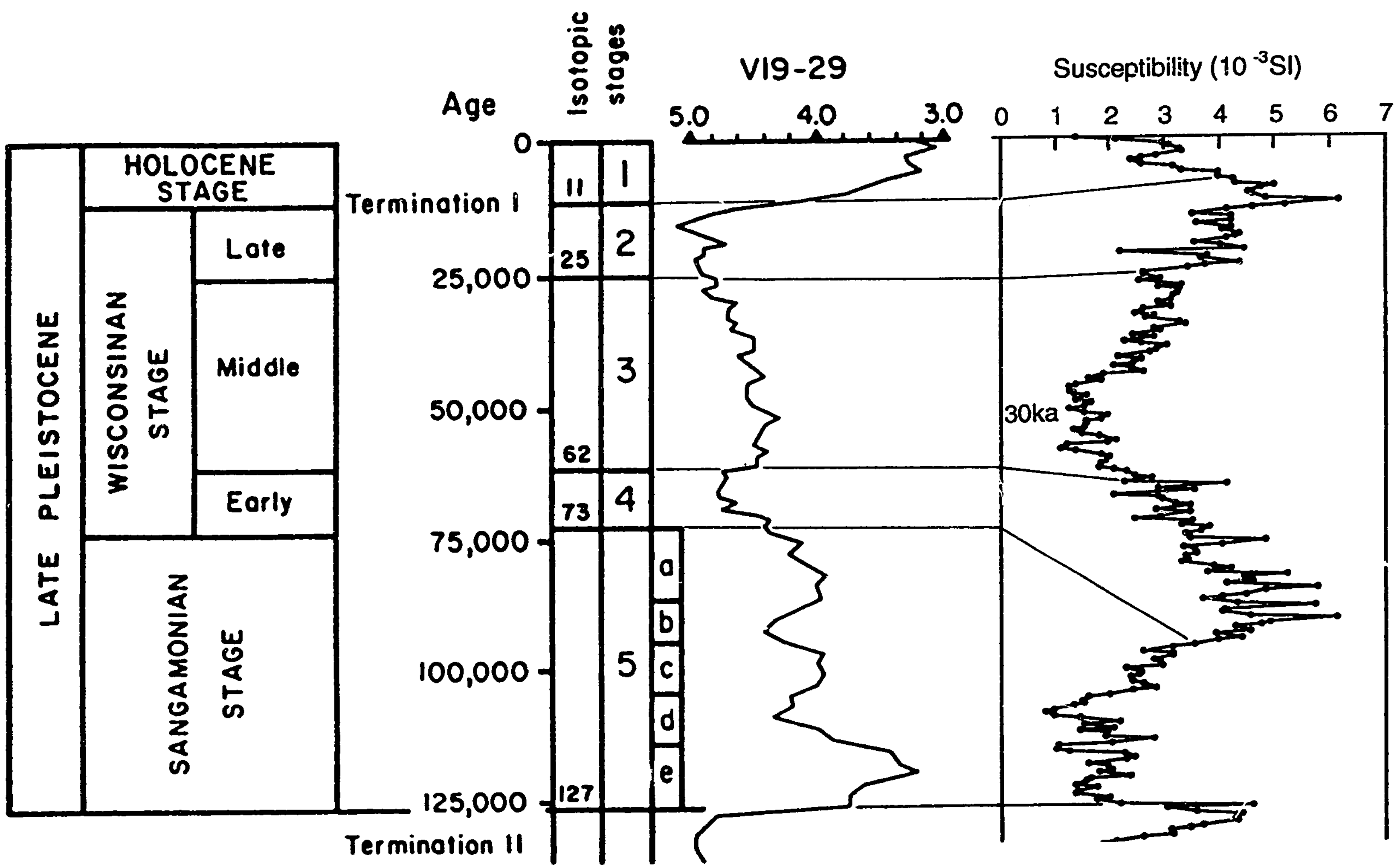


Figure 29. Correlation of the Late Pleistocene continental stratigraphy in Canada (St. Onge 1987) and the marine oxygen isotope stratigraphy of normalized $\partial^{18}O$ stages (Martinson *et al.* 1987, modified by Mangerud 1989) with the Kurtak Section 29 loess-palaeosol record.

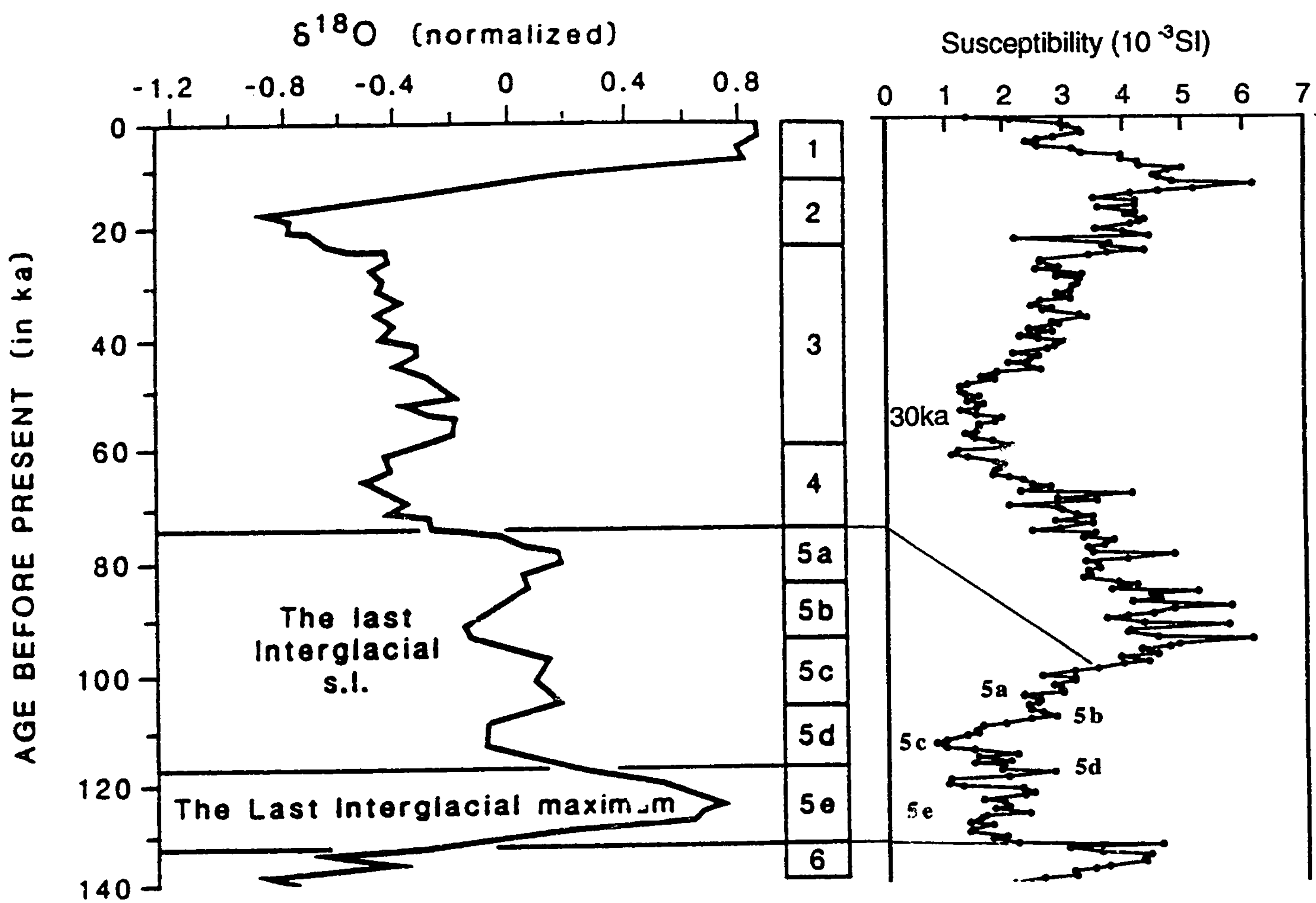


Figure 30. Correlation of the Oxygen Isotope Substages 5e-5a (Martinson *et al.* 1987, modified by Anderson *et al.* 1987) with the last interglacial loess-palaeosol record at Kurtak, Section 29.

Figure 31. Kurtak Section 29. View of the last interglacial loess-palaeosol record in the Kurtak Section 29 sampled by 3 cm intervals for a refined, high-resolution magnetic susceptibility analysis.

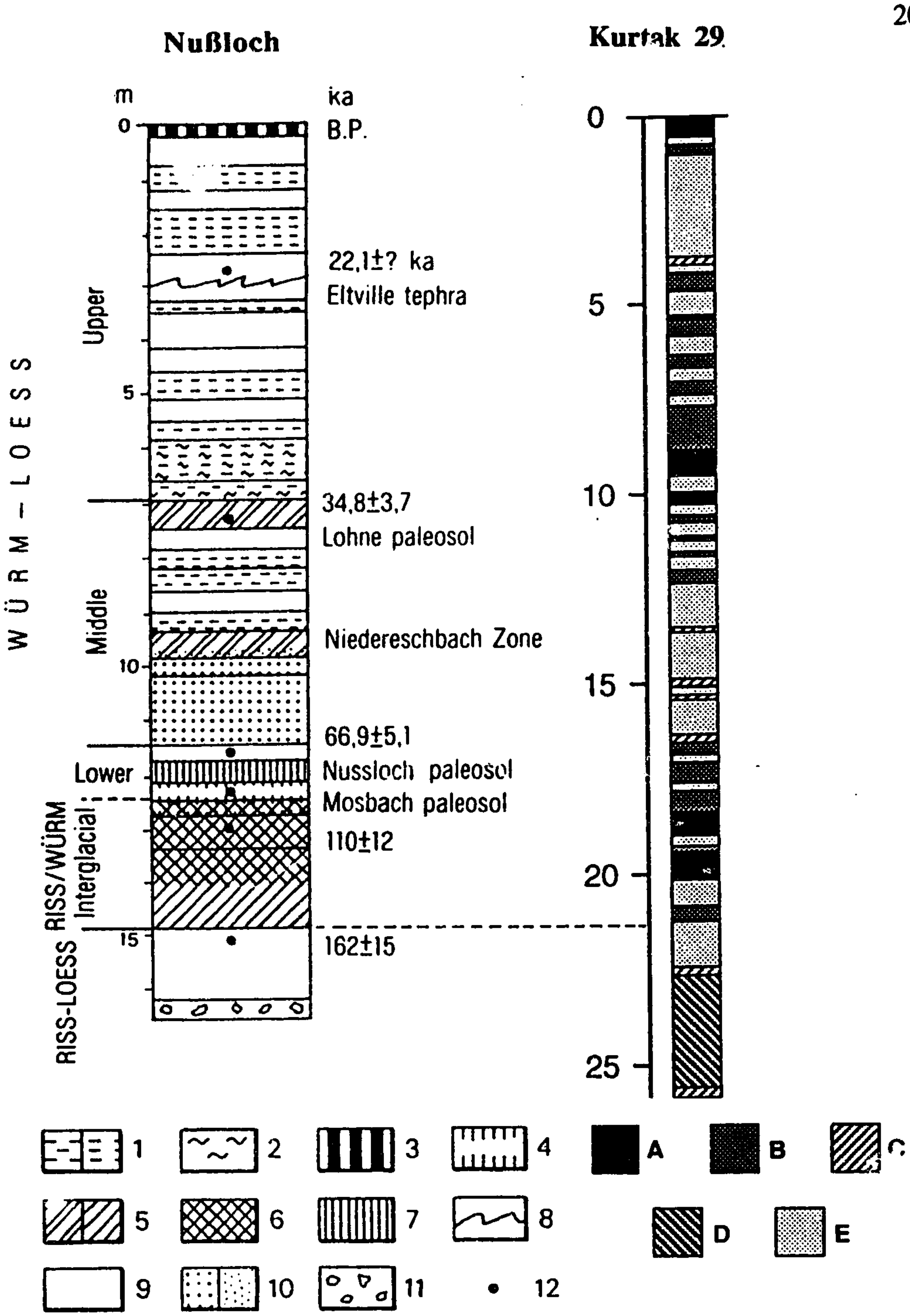
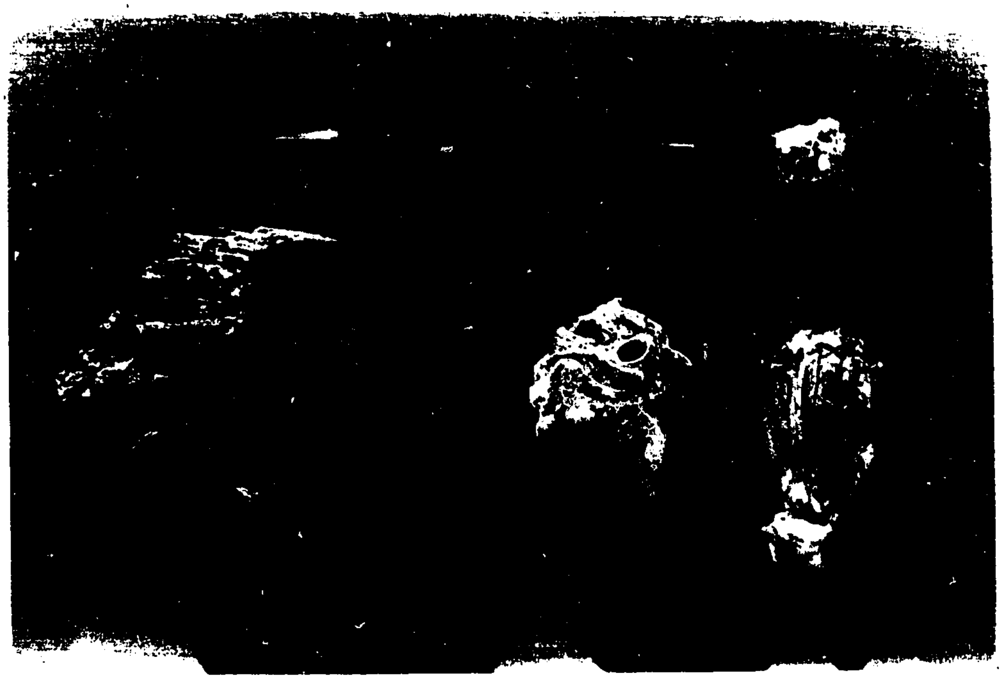


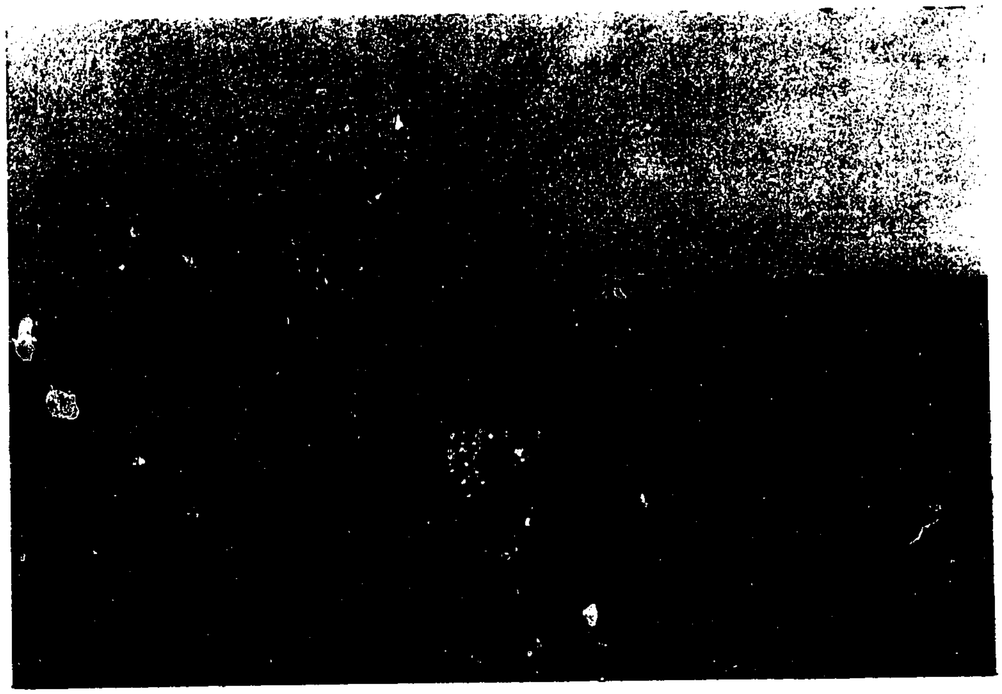
Figure 32. Comparison of the loess-palaeosol records from Kurtak Section 29 with the Nußloch Section in the Rhine Valley, Germany (from Pécsi 1992:fig. 11). Legend *Nußloch:* 1 - tundra gley; 2 - solifluction; 3 - Ap horizon; 4 - Al horizon; 5 - Bv horizon; 6 - Bt horizon; 7 - humic zone; 8 - Eltville volcanic ash; 9 - loess; 10 - sand/sandy loess; 11 - limestone debris; 12 - sites of TL sampling. *Kurtak:* A - Chernozem; B - Brunisol; C - Gleyed Regosol; D - colluviated loess; E - aeolian loess.



A

B

Figure 33. (A) Berezhekovo Site. Fossil fauna remains eroded along the lake shore from the 65 m terrace relic. From left: an early form of mammoth *(Mammuthus trogontherii / Mammuthus antiquus)*, wooly rhinoceros *(Coelodonta antiquitatis)*, horse *(Equus sp.)*, red deer *(Cervus elaphus)*. (B) Razlog Site. Middle Pleistocene (Early Palaeolithic) quartzite artifacts distributed on the 60-65 m terrace and within the above alluvial fan.



A

B

Figure 34. (A) Razlog Site. Eroded distal part of the alluvial fan below a 150 m terrace platform. (B) Ust'-Izhul Site. View of the late Early Pleistocene site prior to excavation, with fossil skeletal remains (mainly of an early *Mammuthus primigenius*).

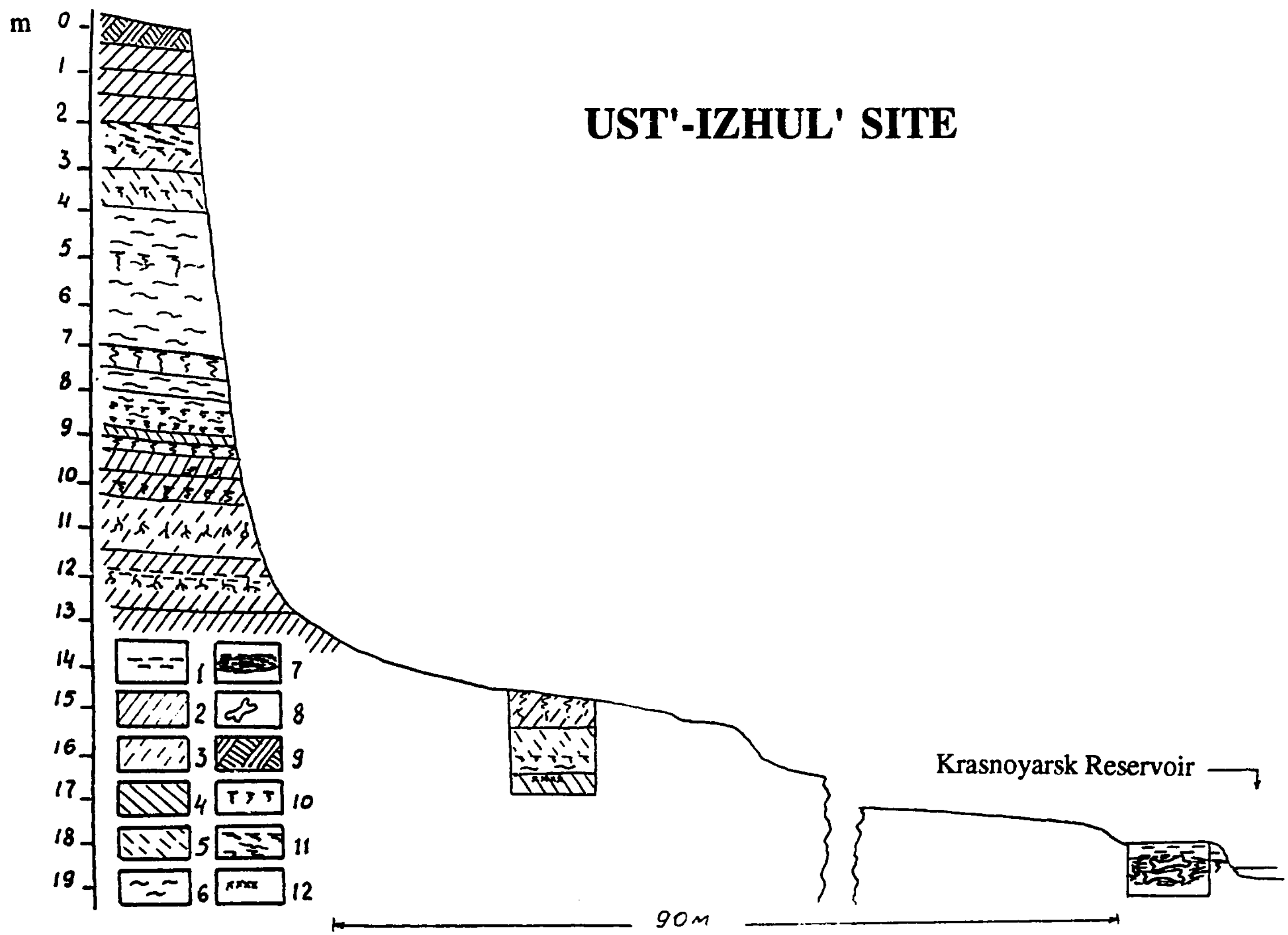


Figure 35. Ust'-Izhul'. Stratigraphic profile of the site. Legend: 1 - clays, 2-3 - clayey sediments, 4-5 - silty-sandy sediments, 6 - silt, 7 - gleying related to an eroded palaeosol, 8 - faunal remains, 9 - Ah horizon of the present Chernozemic soil, 10 - buried palaeosols, 11- the Kurtak soil (colluviated), 12 - silt with organic detritus, 13 - calcium carbonate neoforms along rootlets (after Laukhin *et al.* 1995).

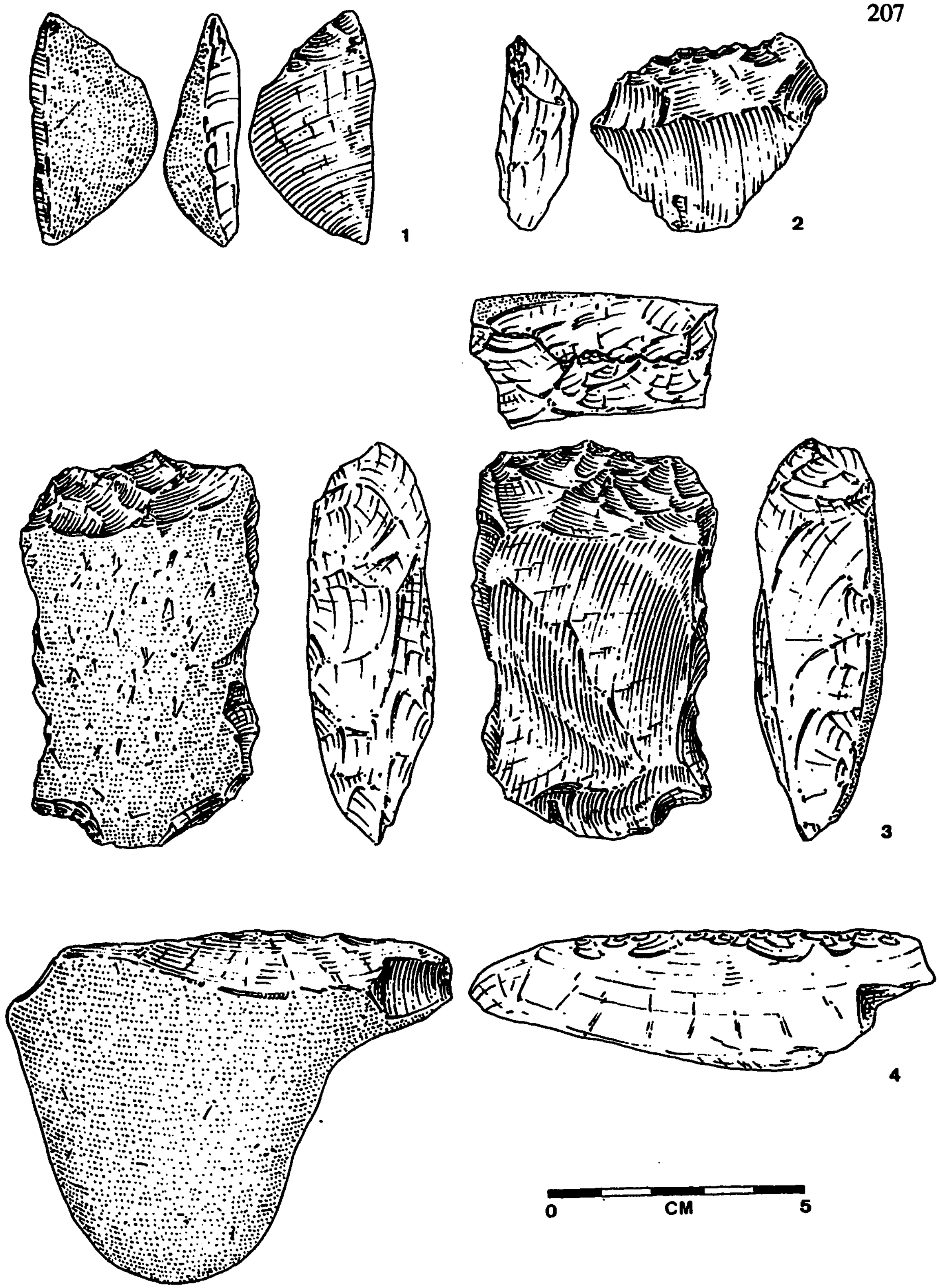


Figure 36 Berezhekovo Site. Early Palaeolithic (Early Pleistocene?) stone artifacts from the 60/65 m terrace (1-2, 4 quartzite; 3 siliceous rock).

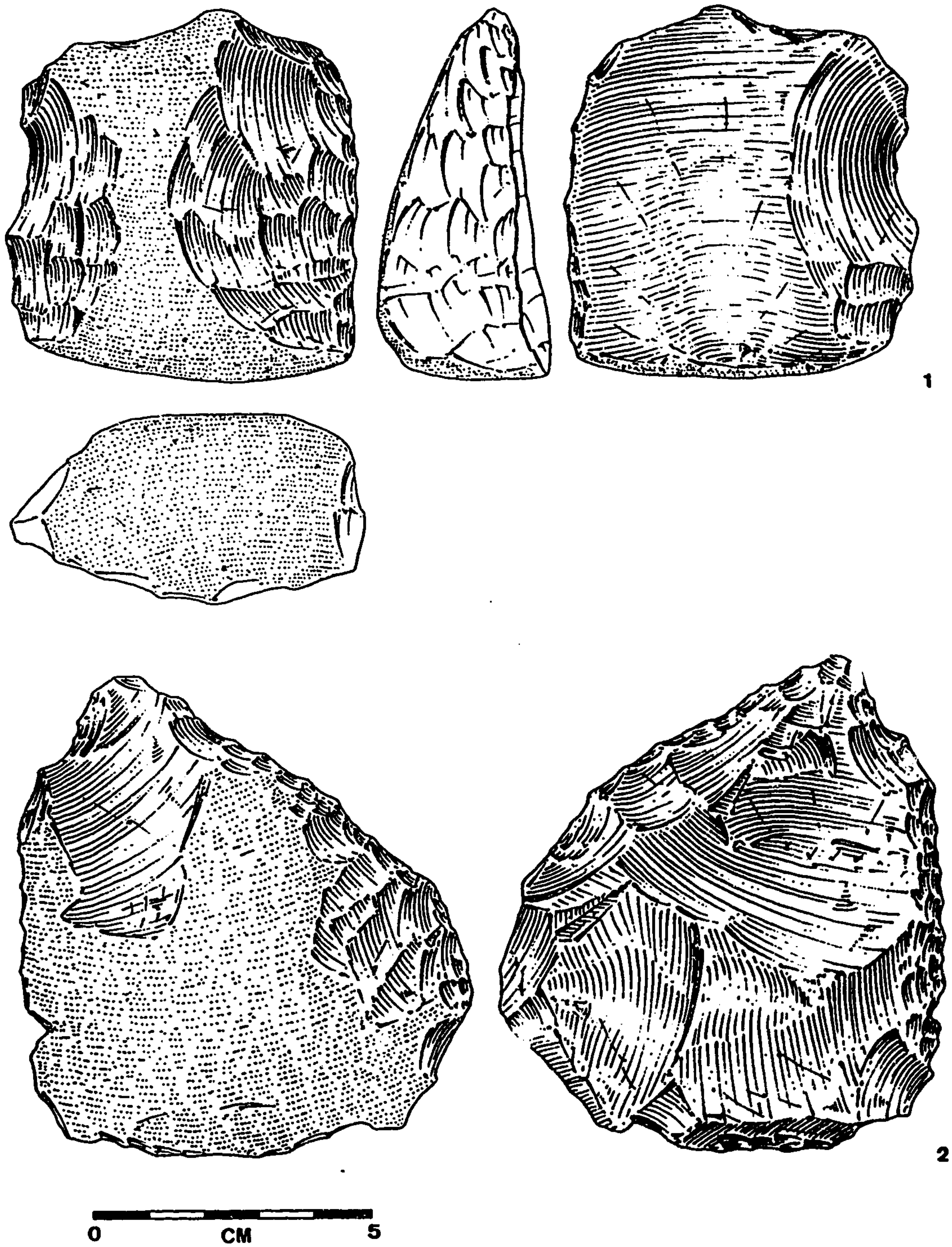


Figure 37. Berezhekovo Site. Early Palaeolithic (Middle Pleistocene) stone artifacts (side scrapers) retouched on quartzite flakes found in association with fossil fauna (Figure 33A).

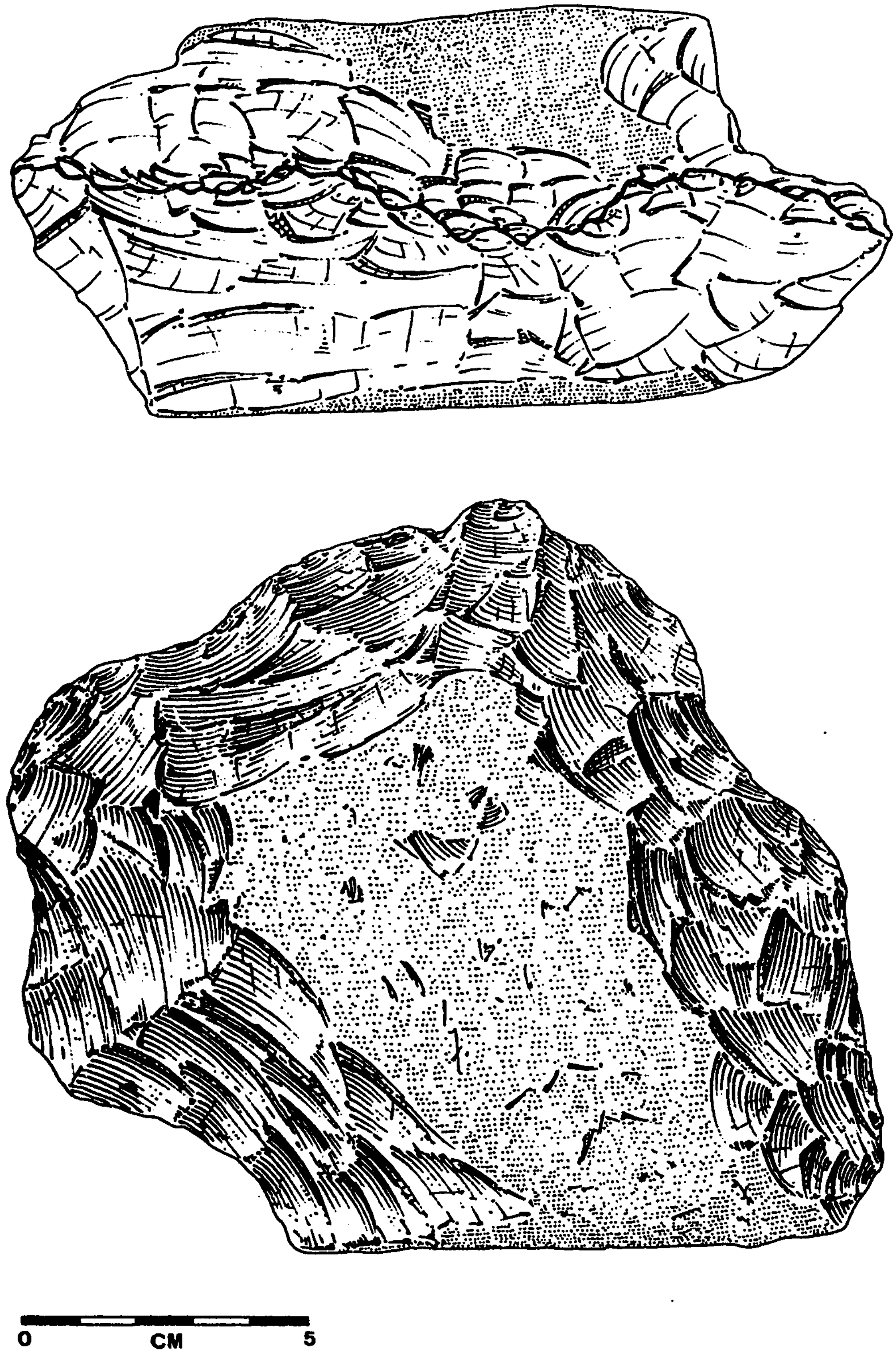


Figure 38. Razlog Site. Early Palaeolithic (Early Pleistocene?) bifacial chopper / core on a quartzite cobble.

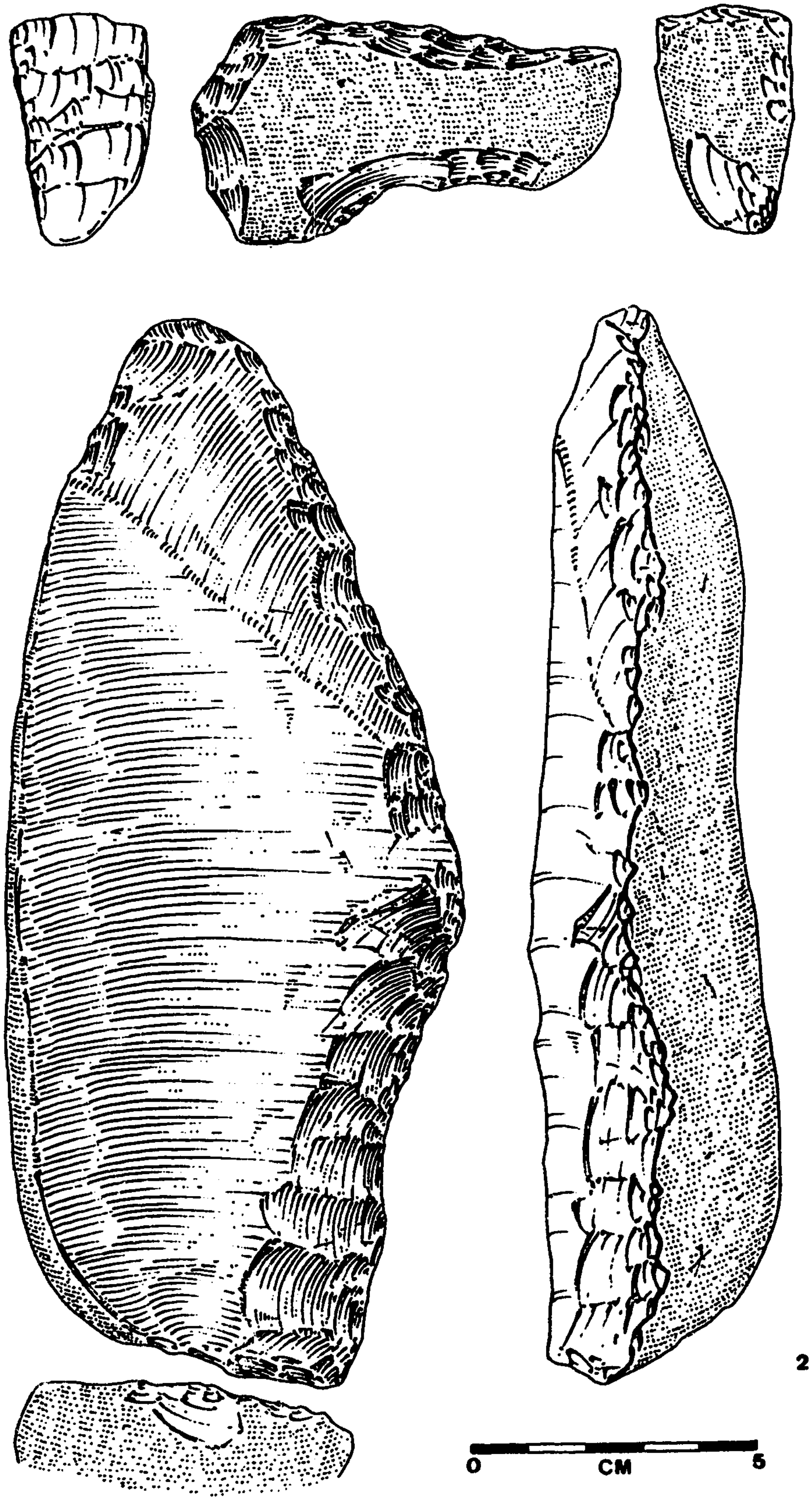


Figure 39. Verkhniy Kamen Site. Early Palaeolithic (Middle Pleistocene) stone artifacts (side scrapers on quartzite flakes) from the 65 m terrace.

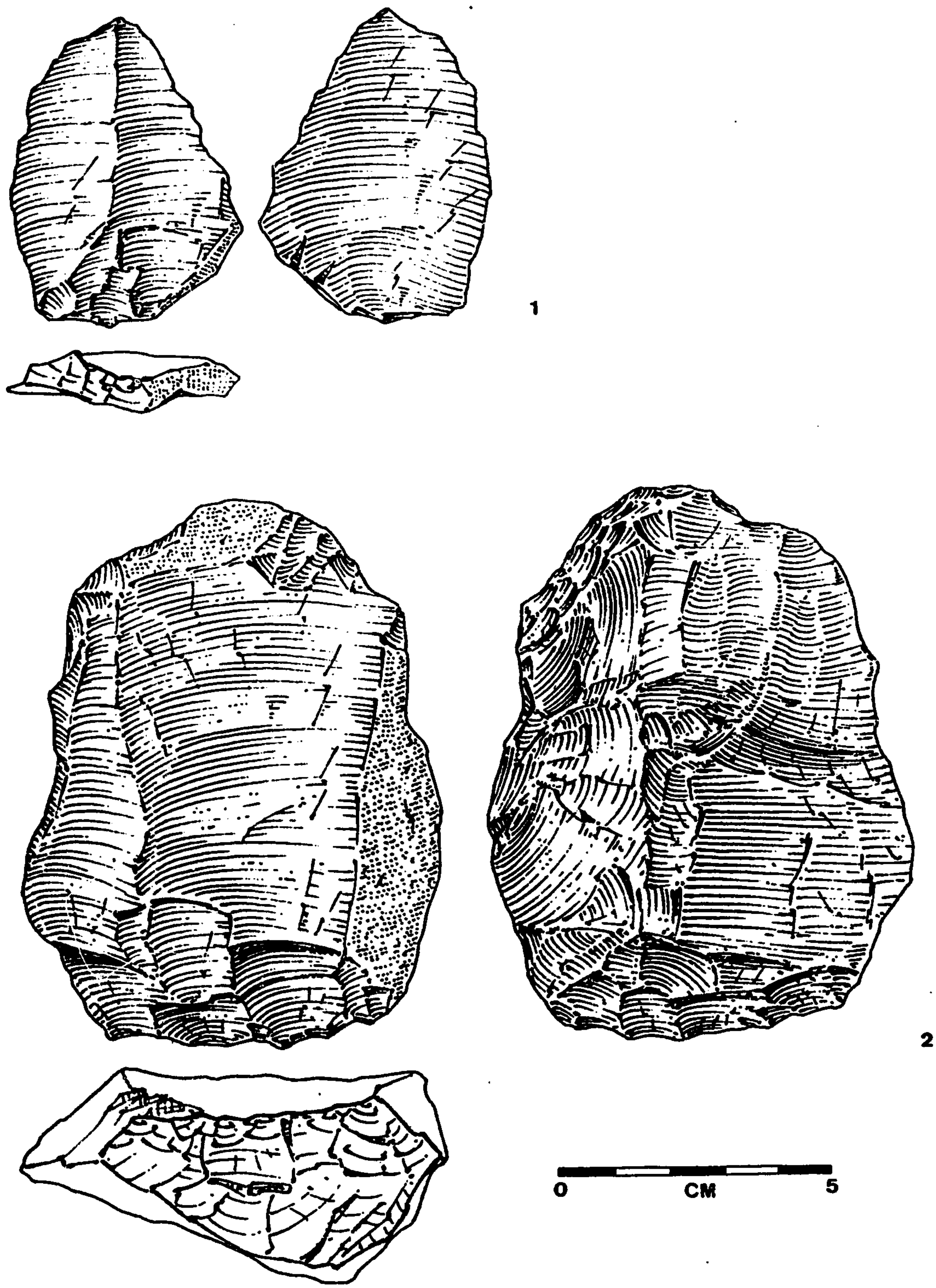


Figure 40. Verkhniy Kamen Site. Radiolarite Levallois point (1) and Levallois core (2). Middle/Upper Palaeolithic (?).

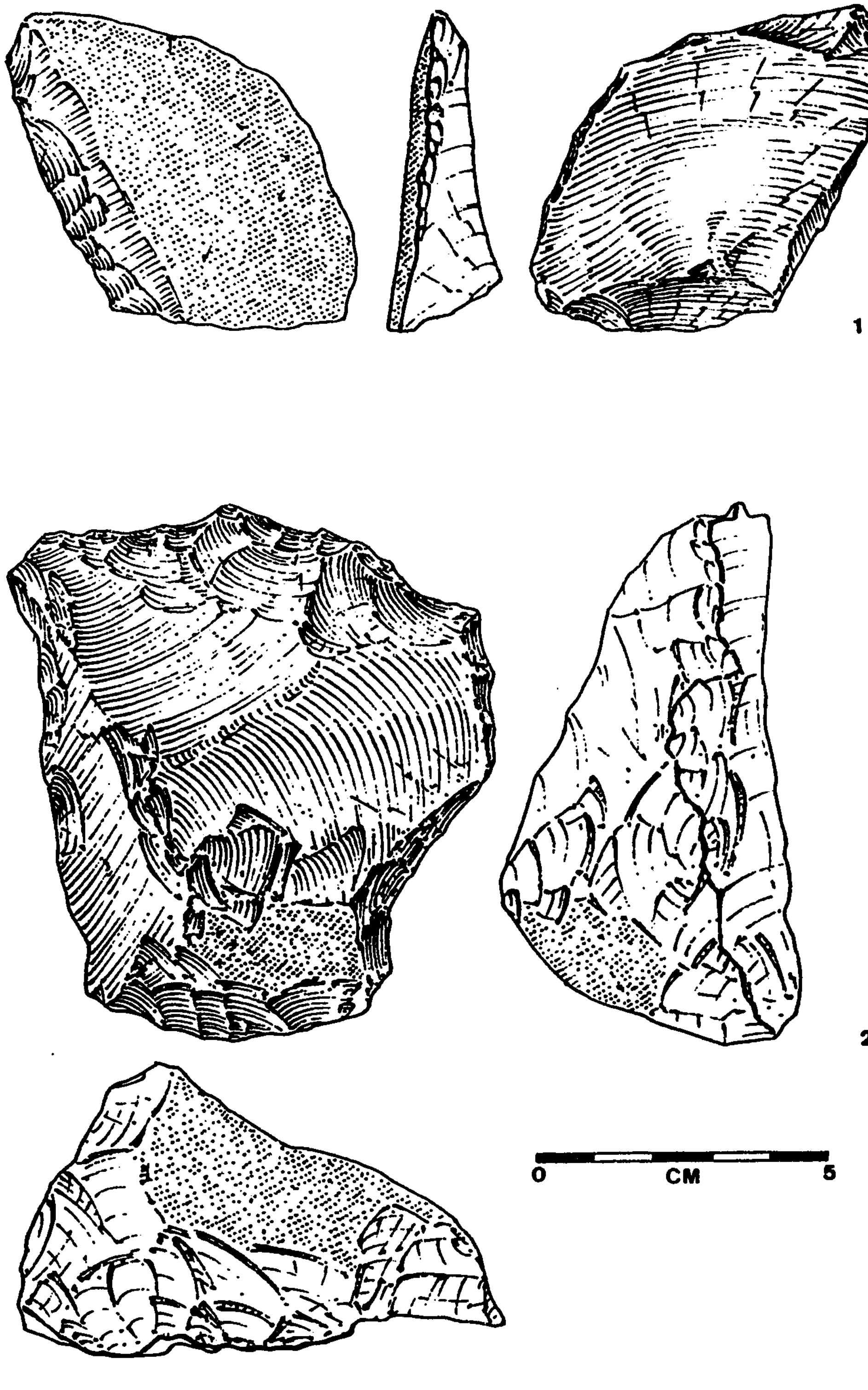


Figure 41. Ust'-Izhul Site. Quartz artifacts (side scrapers) found in association with the fossil faunal remains at the Early Palaeolithic occupation site.

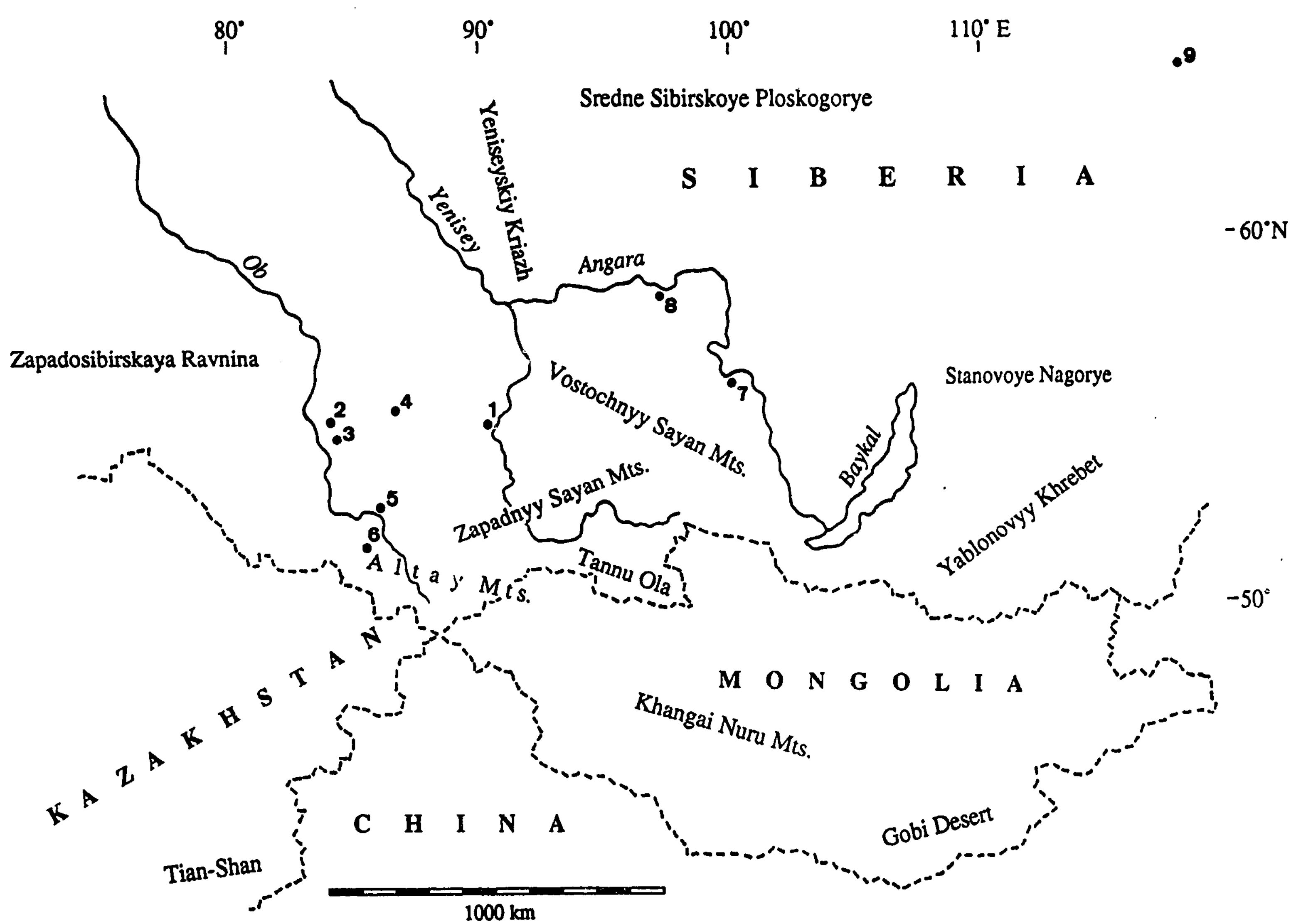


Figure 42. Location map of southern Siberia with the principal physiographic features (khrebet / range; ploskogorye / plateau; ravnina / lowland) and sites discussed in the text. Site legend: 1 - Kurtak; 2 - Berdsk; 3 - Iskitim; 4 - Mokhovo; 5 - Biysk; 6 - Ust'-Karakol; 7 - Igetey; 8 - Ust'-Koba; 9 - Diring.