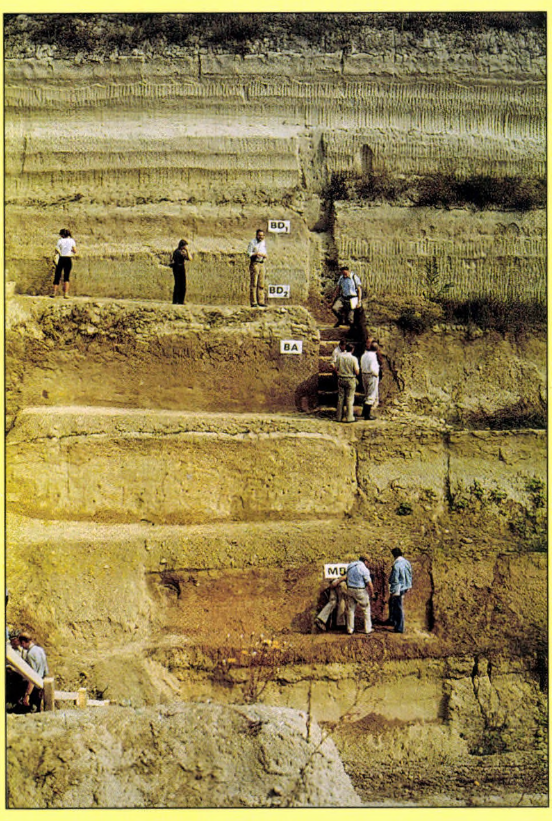
LOESS inFORM 4



In memoriam MÁRTON PÉCSI

Geographical Research Institute
Hungarian Academy of Sciences

REGIONAL STUDIES ON LOESS

LOESS inFORM 4

Recommended by the Commission on Loess
of the International Union of Quaternary Research

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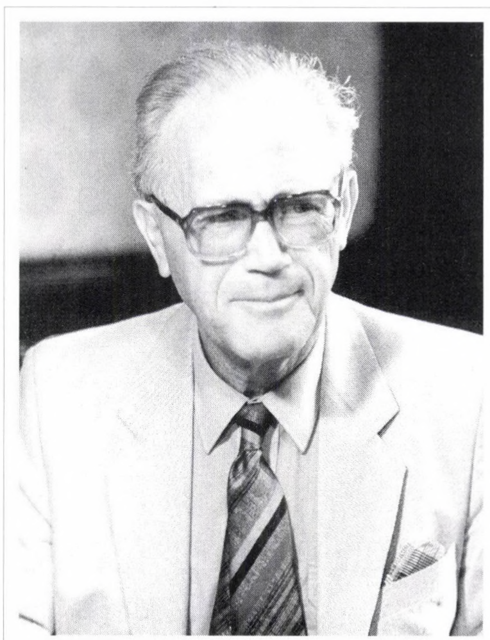
Edited by ÉVA KIS

In memoriam Márton Pécsi (1923–2003)



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Contents

Loess and Quaternary: The geomorphologist's approach

Márton Pécsi received a traditional training of a physical geographer at the University of Budapest. Though he was a scholar of versatile interest, his scientific career primarily concerned history of landform evolution, and he viewed the Earth surface as a sharp-eyed geomorphologist. Searching for an explanation of the phenomenon that many loess varieties exist, Pécsi emphasized that loess is not just the accumulation of falling dust. Types of loess and loess-like deposits might differ in basic parameters of grain size distribution, mineral composition and lithological properties. Loess properties are the result of physical, chemical, biological and pedological processes taking place under environmental conditions favourable for diagenesis (loessification), i.e. weathering of the parent material mainly in semi-arid grassland (steppe) or forest steppe.

Along his professional lifeline as a geomorphologist, Quaternary researcher and expert on loess Pécsi had the best opportunities to study the distribution of loess virtually all over the world. The fact that loess and loess-like deposits occur over a variety of landforms attracted him to different areas of the world where these sequences are encountered in extensive plains, major river basins and valleys, on plateaus and pediments. His earliest investigations in the 1950's included terrace formation in the Middle Danubian Basin, where he was challenged by the problem of rate of sedimentation. In the 1960's during his stay in the USA, when he was awarded with Ford scholarship he saw loesses in the Great Plains of the prairie belt and those in the Lower Mississippi Basin. Using another research grant in the USSR he visited a similar extensive area of loess formation, the Middle and Southern Russian Plain. Later, already as an acknowledged authority Pécsi was a permanent participant of international field symposia and was made acquainted with the plain loesses of the Lower La Plata and Lower Huanghe. He visited the Chinese and Siberian loess plateaus and one of his last field trip in the late 1990's led him to the Columbia Plateau. Loesses of Central Asia of different origin confined to pediments were the topic of a post-congress field excursion (Moscow INQUA, 1982) and the conspicuous non-typical loess varieties in New Zealand were figuring at an earlier congress (1973). In the same year he saw the 'yedoma' loess-ice complex during a field symposium in North Siberia. Naturally the valley loesses of Middle Rhine, Lower Seine and other deposits along Vistula, Oder, Elbe and Main were the occurrences he revisited during his long active period.

The origin of quartz grains of 10–50 micron size making up the bulk of loess material was a recurring theoretical problem of loess formation (frost action, glacier grinding, deposition of river load during floods, combination of glacial and desert i.e. cold and warm loess theories). The unique loess fabric is a textured loam, not merely simple sediment. By the end of his life Pécsi strongly concerned himself with the primary role of pedogenic–geochemical processes in the formation of the coarse silt fraction.

The loess–paleosol sequences undoubtedly provide the best opportunity for the reconstruction of Quaternary cyclical climatic and environmental changes and for its chronological subdivision. The basic principle is that loess strata represent cold and dry climates (glacials), whereas paleosols indicate relatively warmer and wetter paleogeographic conditions (interglacials or interstadials). When studying loess profiles Pécsi emphasised the decisive role of their geomorphological position along with bio- and lithostratigraphical composition and absolute chronological record. He paid a special attention to erosion gaps for sections in uplifted position and sediment traps in subsiding areas. In the basins the number of loess, sand and paleosol horizons could be twice as high as that of the similar units on plateaus or foothills.

Acknowledging that the loess sequence of the Loess Plateau in China exhibit a quasi-complete stratigraphic subdivision (with 24–37 loess horizons and an equal number of paleosols) down to the Quaternary–Neogene boundary 2.4 million years ago, Pécsi held that true loess has formed approximately since 1 million years ago. The deeper, older series usually consist of pink silt subseries alternating with or underlying red paleosols formed in subtropical paleoecological environment.

Chronostratigraphical correlation in a global scale has been a challenge to loess experts in general and to Márton Pécsi in particular. He made considerable efforts to reconcile various Quaternary time scales that suggest or reflect cyclic climatic and environmental change (radiation changes by Milankoviæ, $O^{18/16}$ ratio in deep-sea deposits, changes of magnetic susceptibility in sequences, fluctuations of CO_2 pressure in polar ice, pollen spectrum changes etc.). At the same time he pointed to information on certain climatic types to be gained from the traces of frost phenomena in the zones of cold loess, buried dells and river terraces and professed that the time of past events can be established from the lithostratigraphic framework with some probability.

Regional correlation was another important field of Pécsi's scientific activity. Key loess an Quaternary profiles in the Carpathian Basin: Paks, Mende, Basaharc, Gyöngyösvisonta and others have become familiar to the international community of loess experts who visited these sites frequently and took an active part in chronostratigraphic investigations, deploying novel methods of dating and laboratory analyses.

Making use of his rich experience in Quaternary geology Pécsi became a permanent contributor to mapping in the most various scales. He was a co-author of the Loess map of Europe (1977), compiled a loess map of Hungary (1982) and his map of loess of the Northern Hemisphere was included in a paleogeographic atlas of the late Pleistocene and Holocene (1992).

The past decades saw an amazingly rapid advancement of methodical approaches in loess research. Technological development has made the traditional methods (e.g. SEM analyses in biostratigraphy) more sophisticated, new methods of absolute dating (e.g. IRSL and TL analyses) came to the fore. Quantitative analyses and modelling gained impetus. These developments had prompted Pécsi to reconsider his chronological subdivision of the Quaternary. In the last decade of his active period he tended to assign older age to paleosols hitherto classified into the sequence of "young loess" i.e. formed in late Pleistocene.

Pécsi's activities as an organiser and manager in international science were far-reaching. He was president of the Commission on Loess of the International Quaternary Association (INQUA) between 1977 and 1991, worked in the Subcommission on European Stratigraphy of IGU and acted as president of the Carpatho-Balkan Geomorphological Commission. He was one of the editors-in-chief of the Atlas of Paleoclimates and Paleoenvironments of the Northern Hemisphere (Late Pleistocene–Holocene) published in international co-operation (1992).

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Paleoenvironments and climatostratigraphy of the loess–paleosol formation of Northern Eurasia

NATALIYA S. BOLIKHOVSKAYA[1]

Abstract

This paper is a summary of some chapters of the author's book – N.S. Bolikhovskaya: The Evolution of Loess-Paleosol Formation of Northern Eurasia. Moscow. Moscow State University Publishing House. 1995. 270 pp. Illustrated. (Reviewers: Prof. A.A.Velichko and Prof. V.N. Konischev), – which discusses problems of paleogeography, chronostratigraphy and genesis of loess and fossil soils.

The results of a multi-member chronostratigraphic subdivision of the loess-paleosol formation (LPF) of the Russian Plain are presented. A correlation of basic paleogeographical events of the loess areas in the Pleistocene are assessed. Landscape and climatic conditions of the epochs of the loess-paleosol formations in the East European, West European and Central Asian loess provinces are characterised. It was defined that the Matuyama-Brunhes paleomagnetic inversion is confined to the base of the Gremyach'e = Cromer II = Westerhoven interglacial sediments. Brunhes normal polarity epoch contains 8 interglacials: Gremyach'e = Westerhoven, Semiluki = Rosmalen, Muchkap = Voigtstedt = Noordbergum, Likhvin *s.str.* = Holstein, Chekalin = Kamenka = Domnitz, Cherepet' = Romny, Mikulino = Eem, Holocene and dividing them 7 glacials: Devitsa = Glacial B, Don = Glacial C, Oka = Elster, Kaluga = Borisoglebsk, Zhizdra = Orchik, Dnieper = Saale, Valdai = Weichsel. For the first time detailed reconstructions of phytocoenotic and climatic successions of the main stages, i.e. 8 interglacials and 7 glacials (including 9 interstadials and 10 stadials during Valdai = Weichsel) of the loess-paleosol formation and another sediments of the Russian Plain, various glacial-periglacial and extraglacial regions have been identified.

Introduction

When considering results of more than 160 years of investigations into loesses and interbedded fossil soils, it should be admitted that many problems related to their origin, stratigraphy, and correlation are still debated, as are environmental and climatic conditions of loess formation. The most controversial items are the genesis of loess, as well as its specific property of collapsibility, age and correlation of paleosols and loess horizons, and paleoenvironments having existed during loess formation.

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The principal method applied to these studies was the palynological approach. It permits to characterise paleoenvironments of all the stages, including loess formation, and not only those of soil formation. In this way, a whole sequence of changes in flora, vegetation, and climate can be traced within each stage of the evolution of the loess-paleosol formation (LPF).

The first attempts at palynological studies of loess and fossil soils of Northern Eurasia were made in the 1930s by V.N. Sukachev, Z.K. Dolgaya (1937), and V.P. Grichuk (1940). Later on, the studies were stopped and resumed only in the 1950s by E.T. Lomayeva. Since the 1960s a number of specialists have been engaged continuously in research on the spore and pollen of LPF (V.P. Grichuk, B. Frenzel, S.I. Parishkura-Turlo, A.T. Artyushenko, G.A. Pashkevich, N.S. Bolikhovskaya, Z.P. Gubonina, R.E. Giterman, M.M. Pakhomov, E.E. Gurtovaya, L.G. Bezus'ko, B. Urban, N.P. Gerasimenko, E.M. Malaeva, and many others). Detailed palynological studies have been conducted on loess-paleosol series of different age, on the layer-by-layer basis, (see Bolikhovskaya 1975, 1981, 1982, 1984, and others; Pashkevich 1977; Pakhomov 1983). The results including other published palynological data on LPF (summarised by the author) revealed a prevailing misconcept that loesses and loess-like deposits were formed under conditions of glacial climate exclusively, while fossil soils were products of interglacial and interstadial environments. To substantiate the conclusion, it was necessary to carry out a detailed paleobotanical analysis of the most representative sequences in loess regions, which were studied using an integrated approach.

Problems of paleogeography of the loess-paleosol formation

An overview of extensive literature on the loess genesis in Northern Eurasia shows that the cornerstone in loess formation theory is an assumption that loess originates as a result of transformation of the initial (aleuritic = loessial) fine-grained material by hypergenic processes under subaerial conditions; the material itself may be deposited by different agents (wind, slope wash, perennial streams, glacial floods, etc.). The most complicated problem of cardinal importance seems to reconstruct the climatic and environmental conditions under which the fine material was transformed into loess by hypergenic processes; it is equally important to date the loess-paleosol formation. Since the late 1960s, the author conducted palynological research on main constituents of the LPF, that is loess and fossil soils, as well as on the Late Cenozoic sediments of other genesis (glacial, alluvial, lacustrine, etc.) exposed in key sections of different loess regions; the sediments are generally in paragenetic association with loess.

The history of the LPF studies illustrates well a diversity of opinions regarding climatic and environmental conditions of the loess and paleosol formation. There is an extensive literature on the LPF of the Russian Plain. A.I. Nabokikh (1915) stated that loess could have been formed both under glacial climate (cold and dry, or cold and moderately humid) and during drier intervals of interglacials. Similar views have been advocated by I.P. Gerasimov (1939). V.D. Laskarev (1919) and G.F. Mirchink (1928) placed loess into interglacial intervals. A conclusion by D.N. Sobolev (1924) that loess (synchronous to morainic horizon) was formed in the periglacial zone at the time of the glacial maximum has been accepted and elaborated by a majority of researchers; V.I. Krokos, A.I. Moskvitin and their followers associate loess horizons with glacial epochs and attribute soils to warm intervals. The sensor character of many loess properties has brought many experts (K.I. Lukashev 1961; N.I. Kriger 1965; K.K. Markov *et al.* 1965) to a conclusion that loess and paleosol are mutually exclusive notions, because soil formation leads to degradation of properties of the underlying loess. This conclusion which is justified from lithological and geochemical viewpoints has also served as a basis for opposing loess and fossil soil horizons while solving problems of genesis, paleogeography and chronostratigraphy of the loess.

An analysis of the currently dominating schemes of the LPF chronostratigraphy and correlation on the East European platform (Velichko *et al.* 1984; Veklich *et al.* 1984) as well as a consideration of some regional schemes (Bogutsky 1975; Gozhik *et al.* 1984; Krasnenkov *et al.* 1984; Zarrina and Krasnov 1985; Shelkoplyas *et al.* 1986), together with paleogeographical data they are based on, have elucidated the most essential points of disagreement. They are the number of stratigraphically significant horizons of loess and paleosols, and reconstructions of types of the Middle and Late Pleistocene pedogenesis. Different as their views on the LPF stratigraphy and paleogeography may be, the authors of the mentioned stratigraphical schemes however are unanimous in the crucial point, that is in position of LPF horizons relative to climatic rhythms of different order. They consider the loesses as having formed under glacial climate, and the fossil soils related to interglacial and interstadial warming. The question is not so simple, however, when viewed from the palynological standpoint.

Regional aspects of the palynoindication of the loess-paleosol formation

The development of LPF within the *East European loess province* has been closely related to the ice sheet dynamics. An overwhelming majority of the LPF sections (not only those in Northern Eurasia, but all over the world) studied by the pollen method is located in the south-western sector of this province. Many of the sections have been characterised only by fragmentary pollen spectra which does not allow an unambiguous interpretation; therefore, they have been hardly mentioned in discussions of the LPF paleogeography, stratigraphy and correlation. When studied in detail, taking advantage of new data, the sections provided a new insight into the main issues related to the palynoindication of the LPF.

A.T. Artyushenko, S.I. Parishkura-Turlo, S.I. Medyanik, and other specialists (see publications of 1970 to 1992; Sirenko and Turlo 1986) noted some general features typical of the glacial epochs. Of them it should be mentioned the lack of pollen and spores of Arctic-Alpine and Arctic-Boreal plants in horizons dated to the glacial time, such as Priazovye, Tiligul, Tyasmin and others correlated with them in Ukraine and Moldova; high percentage of broad-leaved species pollen in some instances; rather rare localities with microtherms included in the Sula and Dnieper loessial palynofloras, while thermophilous and cryophilous elements are often found together and they show similar degree of preservation. Based on the above evidence the following vegetation units could be reconstructed during the glacial epochs: 1) periglacial forest steppes and steppes, locally with intrazonal broad-leaved forests; 2) steppes similar to the recent steppe communities in the south of the Russian Plain; 3) steppes and forest steppes, locally with pine-birch forests along rivers and in ravines. The latter included little admixture of broad-leaved species and cryophytes *(Betula nana, B. fruticosa, Alnaster fruticosus)* in the shrub level. A detailed analysis of such materials together with palynologic evidence of fully developed soils formed during glacial stages (see G.A. Pashkevich, N.S. Bolikhovskaya and others) suggest that some of loess horizons in Ukraine and Moldova belong to interglacials. It seems necessary to distinguish stenoperiglacial and extraglacial types of vegetation within the vast periglacial zone which existed during glacial stages of the LPF development in the Russian Plain.

Reconstructions of vegetation, climate, paleopedogenesis, compared with data on paleomagnetism and other materials, along with the analysis of geographical groups of genera in interglacial dendrofloras have led to the conclusion that the Shirokino stage in the LPF development on the Ukrainian territory (marked by the dominance of forest steppes in coastal regions of the Black and Azov seas and in the Donets Lowland) corresponded to the time of formation of the loess-paleosol series attributed to the complex Ilyinka *s.l.* interval recognised in the central regions of the Russian Plain. The Martonosha warm stage is correlated with the Muchkap interglacial of the central and southern regions of European Russia. This correlation is based on a number of indicators, such as on a wide distribution of mixed forests dominated by thermophilous and hydrophilous species almost all over the territory of Ukraine, predominance of *Taxodium, Podocarpus, Juglans, Pterocarya, Carya, Liquidambar,* etc. in the Neogene relict dendroflora (see A.T. Artyushenko, S.I. Turlo-Parishkura, and others), increased hydromorphism of the prevailing brown, lessivated and other forest soils of the Martonosha stage (see M.F. Veklich, N.A. Sirenko, J.N. Matviishina). From this interval onward, there had been a progressive increase in climatic continentality which reached its maximum at the Late Valdai time of the loess formation. The Lubny stage of the LPF development in Ukraine is correlated with the Likhvin *s.str.* interglacial. It is with less confidence that the early Zavadovka stages of pedogenesis can be attributed to the Chekalin (= Kamenka) interglacial, and soils of the late Zavadovka stages to the Cherepet' (= Romny) interglacial. There is no formation indicative of glacial climate between the Priluki and Kaidaki soils; on this ground, the period of their formation may be considered as a single interglacial rhythm with successional changes and the flora typical of the Mikulino thermochron.

The LPF of the *West European loess province* reflects a landscape evolution influenced by both ice sheets and mountain glaciers. Chronostratigraphy and reconstructions of environments of loess-paleosol formation in this region are primarily based on lithological, paleopedological, paleofaunistic, paleomagnetic data, radiocarbon and thermoluminescent dating (Pécsi 1966, 1993; Fink 1969; Haase *et al.* 1969, Smolíkova 1969; Maruszczak 1970; Somme 1977; Morozova 1981; Lautridou *et al.* 1982; Vaškovska 1985; Maruszczak 1986; Minkov *et al.* 1986; Lebret and Lautridou 1991; Sajgalik 1991 and many others), while paleobotanic evidence is much sparser. A major contribution to understanding of the LPF paleogeography and stratigraphy has been made by M. Pécsi; his monograph (1993) summarizes analytical data and reconstructions of environmental and climatic conditions of the loess and paleosol formation in Eurasia as a whole, with a special reference to Hungary.

It was B. Frenzel (1964) who laid methodical foundations for palynological studies of the loess-paleosol series in Western Europe. He developed a special technique for pollen and spores extraction from loess deposits and was the first to obtain 8 complete palynospectra for the Oberfellabrunne and Stillfried sections in the Vienna basin. The data thus obtained suggest that the Eemian interglacial forest steppes were characteristic not only of the time of the Fellabrunne pedocomplex formation but (in our opinion) also of the underlying loess. The expansion of periglacial steppes and forest steppes marked the Early Wurmian stage of initial pedogenesis. In the extraglacial (similar to interglacial) forest steppes that interval corresponds to the time when the Stillfried B pedocomplex (its studied part) and the basal part of the overlying loess were formed.

Of considerable importance for the palynology of West European LPF are the works by B. Urban (1984). She analysed horizons in a number of sections, such as Stillfried, Dolní Vestonice, Paks and Mende. The Paks section has also been studied by G.A. Pashkevich (1979) and by the author. Palynological studies on two additional sections: Lopatki in Poland and Vetovo in Bulgaria (Bolikhovskaya, 1995c) are under way. J. Heim published results of palynological studies of the Achenheim I pedocomplex from the section of the same name in Alsace (Heim *et al.* 1982; Somme *et al.* 1986). Besides, loess and paleosol horizons were characterised partially when cultural layers of Mousterian and Late Paleolithic sites were studied palynologically in Germany, Belgium, France and elsewhere (Leroi-Gourhan 1977; Renault-Miskovsky and Leroi-Gourhan 1981). As evidenced by the above facts, a majority of palynological materials from the West European LPF sections refer to the Late Pleistocene.

It should be noted that many researchers of the LPF in Western Europe (as well as those who studied other regions of Northern Eurasia) ascribe all the Wurmian fossil soils and pedosediments to interstadials. Many palynologists share this opinion, even though their own material do not agree with this unfortunately dominant "paleogeographic axiom". As an example palynospectra of the Dolní Vestonice section (on the right bank of the Morava River 50 km south of Brno) can be referred to.

As it is presented in a paper by B. Urban, the composition of the spectra suggests a few changes in the climatic-stratigraphic interpretation of paleobotanic data on a number of horizons. For example, grass communities were dominant at the time when the first post-Eemian loess was formed. The PC III chernozem soil developed under conditions of interstadial forest steppes (Amersfoort – N.B.) where steppe communities (*Artemisia, Ephedra, Helianthemum, Compositae* etc.) and patches of bare ground alternated with forest stands of *Pinus sylvestris* with admixture of *P. cembra* and, less frequently, with broad-leaved species. Both the overlying loess and degraded chernozem are attributed to the succeeding stadial, as both of them feature similar palynospectra indicative of dominating steppes and much more arid climate between the first and second stages of Wurm. The loess interlayer within PC III accumulated, in our opinion, under more humid climate of interstadial forest steppes with dominating open coniferous forests of *Pinus sylvestris* and *P. cembra* with some thermophilous species (Brörup – N.B.). A soil diagnosed as pararendzina by L. Smolíkova suggests a considerable cooling manifested in a complete absence of broad-leaved species and by an overwhelming dominance of open woodland with pine. The beginning of the subsequent loess accumulation is marked by pollen spectra indicative of forest steppes of interstadial type and more arid climate compared to the previous interstadial. One can see, therefore, that the Wurmian fossil soils of Dolní Vestonice were formed not only during interstadials, but also during the glacial stages, while the loess formation proceeded during both stadials and interstadials. Data on LPF in Austria and Czech Republic, when compared with results of pollen analysis of the corresponding sediments on the Russian Plain, show a distinct similarity between features of the Late Pleistocene LPF evolution of the Lower Morava valley and those of the LPF development in the Middle Dniester basin (located at the same latitude). Both regions reveal similar trend in the change of climate, phytocoenoses and floras.

That loess may have been formed not only under glacial and interstadial climates, but also in interglacials, has been clearly demonstrated by the reconstructions of some Early Pleistocene stages of the LPF development on the Lower Danube Plain (the Vetovo section in northern Bulgaria). It has been proved that loess 5 (that is the uppermost loess in the Ilyinka subaerial series developed in forest steppe environments) was formed at the optimum phases of the Semiluki (Cromer III) interglacial. At that time, flat interfluves and upper slopes on the right bank of the Danube were predominantly covered with herb and grass communities, with sporadic eroded ecotopes. Limited patches of forests were dominated by broad-leaved species *(Tillia tomentosa, T. platyphyllos, Quercus robur, Q. petraea* and others), stands of *Alnus glutinosa* and those of spruce, pine and birch occurred locally. Lithogenesis of the lower half of loess 4 corresponds to moderately warm and humid climate of the Muchkap (Cromer IV) interglacial; the area under study was then occupied by forests of beech, hornbeam, hazel and oak showing a considerable diversity in species composition. Among the most typical taxa of the Muchkap palynoflora in northern Bulgaria, there are *Tsuga canadensis, T. sp., Pterocarya sp., Juglans cinerea, J. re-*

gia, J. sp., Fagus sylvatica, R. orientalis, Carpinus betulus, N. orientalis, Quercus petraea, Q. robur. Tilia platyphyllos, Corylus colurna and others.

Therefore, the palynological materials obtained from the loess-paleosol sections of Western Europe reveal a marked similarity in the LPF development of some regions of the western and eastern loess provinces of Europe during the Eem = Mikulino and Würm = Weichsel = Valdai, as well as at earlier stages (Bolikhovskaya and Bolikhovsky, 1996).

Palynological studies of key sections in the *Central Asian loess province* are yet few in number. However they have provided important evidence which may be used in the hot discussion about environmental and climatic conditions about the formation of the loess and fossil soils in this region. Materials obtained by R.E. Giterman (Lazarenko *et al.* 1977), M.M. Pakhomov (1983), N.S. Bolikhovskaya (1980, 1983) and other researchers made it apparent that the loesses were formed there under rather various conditions. Loess formation extended over dry and warm interglacial steppes, locally with arid open woodlands; cold and dry periglacial steppes, extraglacial deserts and the like; cold extraglacial forest steppes, with more humid climatic conditions than the present-day ones. All the data point to the fact that loess formation proceeded during thermoxerotic phases of interglacials, as well as during cryoxerotic and cryohygrotic stages of glacial epochs. Fossil soils developed under warm and wet climate (mixed coniferous/broad-leaved forests), warm and dry climate (dry steppes), cool and relatively humid climate (pine and birch forests with admixture of broad-leaved species) characteristic of interglacials and during cryohygrotic stages of glacial epochs. No pollen spectra belonging to interglacial deserts have been found in the fossil soils.

Detailed reconstruction of phytocoenotic and climatic conditions during stages of loess-paleosol formation in the Russian Plain

The Russian Plain is considered a key stratoregion for the periodisation and correlation of the LPF in Northern Eurasia. Data obtained by detailed palynostratigraphic studies of the most comprehensive sections found in the region are in reasonable agreement with the loess and paleosol stratigraphy developed by A.A. Velichko and his co-workers (1984), and with the scheme of the subdivision of Quaternary deposits worked out for the central regions by a team led by S.M. Shik and R.V. Krasnenkov (1985). Both schemes are taken here as a basis for the LPF chronostratigraphy.

On the basis of palynological data it has been established that the period of the LPF development on the Russian Plain comprises 17 paleogeographic stages (9 interglacials and 8 glacial epochs between them) which reflect global Pleistocene climatic rhythms of the highest level *(Fig. 1)* (Bolikhovskaya 1995c). No representative paleobotanic characteristics are available which could ascertain position of

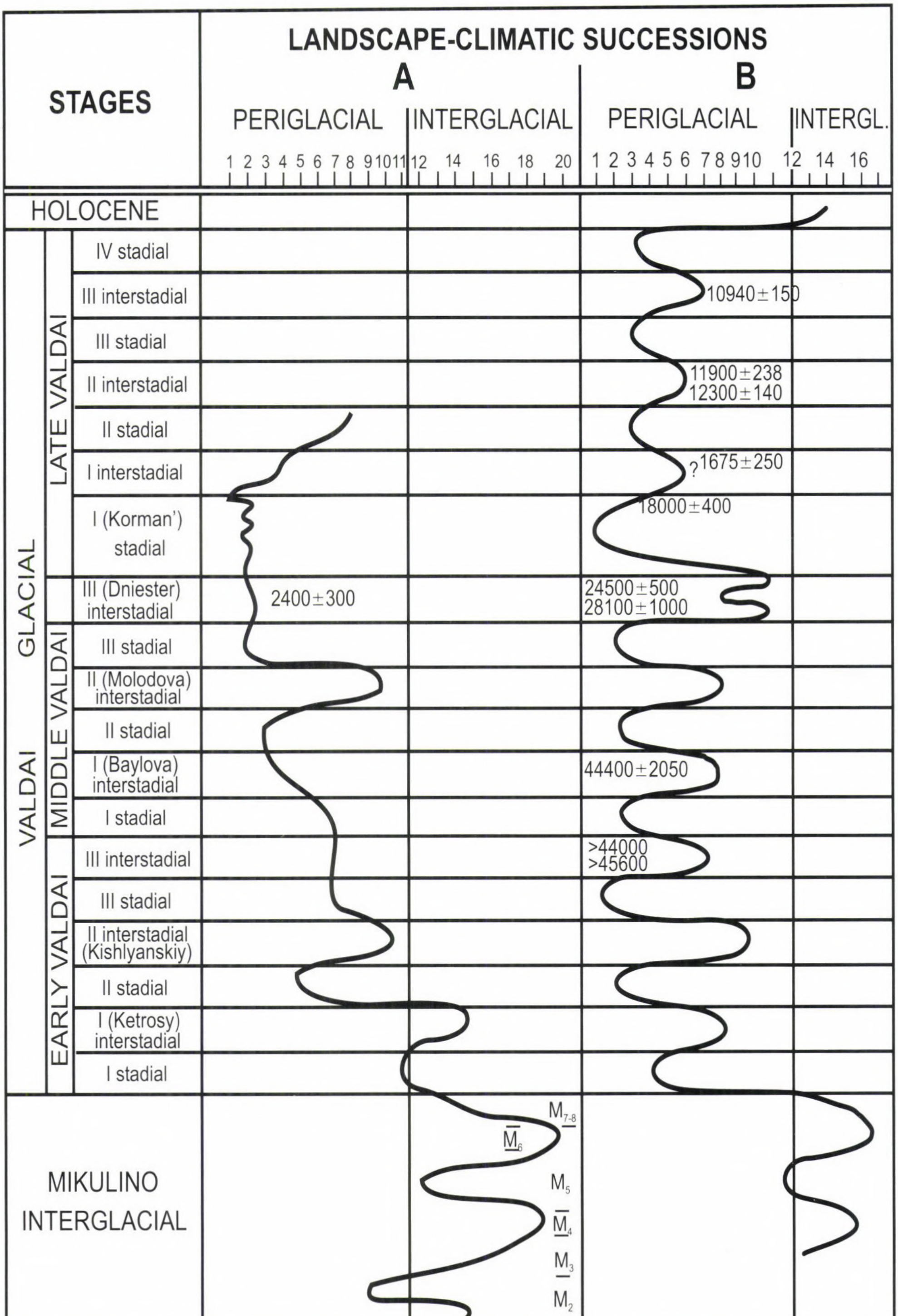
INTERREGIONAL STRATIGRAPHIC SCALE (1986)			MAGNETOST.	STAGES
PLEISTOCENE	UPPER	HOLOCENE HORIZON	BRUNHES	HOLOCENE INTERGLACIAL
		VALDAI HORIZON		VALDAI GLACIAL
		MIKULINO HORIZON		MIKULINO INTERGLACIAL
	MIDDLE	CENTRAL RUSSIAN SUPERHORIZON		DNIEPER GLACIAL
		LIKHVIN SUPERHORIZON (COMPLEX)		CHEREPET' (ROMNY) INTERGLACIAL
				ZHIZDRA (ORCHIK) GLACIAL
				CHEKALIN (KAMENKA) INTERGLACIAL
				KALUGA (BORISOGLEBSK) GLACIAL
				LIKHVIN S.STR. INTERGLACIAL
	LOWER	OKA HORIZON		OKA GLACIAL
		BELOVEZHSK (MUCHKAP) HORIZON		MUCHKAP INTERGLACIAL
		DON HORIZON		DON GLACIAL
		IL'INKA SUPERHORIZON		SEMILUKI (EARLY IL'INKA) INTERGLACIAL
				DEVITSA (MIDDLE IL'INKA) GLACIAL
			+	GREMYACH'E (LATE IL'INKA) INTERGLACIAL
		POKROVKA HORIZON	MATUYAMA	POKROVKA GLACIAL
		PETROPAVLOVKA HORIZON		PETROPAVLOVKA INTERGLACIAL

Fig. 1. Interglacial and glacial stages through the evolution of the loess-paleosol formation (LPF) in the Russian Plain

boundaries of interglacial and glacial climatic rhythms neither in the stratotypes of the Borisoglebsk, Kamenka, Orchik, and Romny horizons, nor in those of the LPF members attributed to the Ilyinka *s.l.* horizon. This being so, the author named the stages after the sections, which provided most comprehensive information (including palynological characteristics) on the corresponding stratigraphic units, so that boundaries between the latter could be located with reasonable accuracy.

The *Dniester-Prut extraglacial loess region* includes loess areas in the extreme south-west of the non-glaciated territory of the Russian Plain, that is in valleys of the Lower and Middle Dniester and Prut, and on the adjoining watersheds. The most detailed subdivision of the LPF has been performed, and climatic and environmental conditions of loess and soil formation have been reconstructed with the highest possible degree of accuracy for the Middle Dniester drainage basin. The thickest and most complete series of loess and paleosols are exposed along the scarp of the second terrace of the Dniester; fluvial deposits of about 10 m thickness are overlain by a 25 m thick mantle of loess-like eolian and deluvial deposits including 8 paleosols. Some well-known Paleolithic sites are associated with sediments of this terrace. Sections of Molodovo, Korman', Ketrosy, Kishlyansky Yar were studied by L.K. Ivanova, A.P. Chernysh, A.K. Agadjanian, N.S. Bolikhovskaya, S.V. Gubin, N.V. Rengarten and others. The results of the studies, supplemented by detailed palynological data (Pashkevich 1977; Bolikhovskaya 1981, 1986, etc.) provided information on the development of LPF, vegetation, climates and other features of paleolandscapes in the Middle Dniester stratoregion during the Mikulino interglacial and 19 stages within the Valdai epoch (9 interstadial and 10 stadial intervals, see *Fig. 2B*). It has been confirmed that material of loess horizons dated to the Late Pleistocene was accumulated both during the stadials (in tundra steppes, periglacial steppes, and forest steppes) and interstadials (in extraglacial steppes and forest steppes with patches of birch-pine forests with an admixture of hornbeam, elm etc.) of the Valdai glacial epoch, as well as during the Mikulino endothermal cooling. At the base of the section two brown forest soils are found separated with a loess layer; they developed during the thermoxerotic and thermohygrotic phases of the Mikulino interglacial. The optimum of the thermoxerotic phase was marked by zonal occurrence of forest steppes with a prevalence of oak in forests with hornbeam, elm, and ash patches; whereas the peak of the thermohygrotic phase featured hornbeam forests with an admixture of hazel, oak etc.

The Valdai paleosols formed during interstadials under extraglacial forest steppe and steppe conditions and during the Late Valdai cryohygrotic phase (the Korman' cryomorphic soil). The latter was characterised by development of tundra-forest steppes with widespread Arctic-Alpine species *(Arctous alpina, Arctostaphyllos uva-ursi, Rubus chamaemorus, Diphazium alpinum,* etc.*)*, by dominance of steppe coenoses, bush formations of *Juniperus, Betula fruticosa, B. nana, Alnaster fruticosus,* communities typical of eroded ecotopes and wetland environment, and patches of birch-pine open woodland.


STAGES
LANDSCAPE-CLIMATIC SUCCESSIONS
A
B
PERIGLACIAL
INTERGLACIAL
PERIGLACIAL
INTERGL.
HOLOCENE
VALDAI GLACIAL
LATE VALDAI
MIDDLE VALDAI
EARLY VALDAI
IV stadial
III interstadial
III stadial
II interstadial
II stadial
I interstadial
I (Korman') stadial
III (Dniester) interstadial
III stadial
II (Molodova) interstadial
II stadial
I (Baylova) interstadial
I stadial
III interstadial
III stadial
II interstadial (Kishlyanskiy)
II stadial
I (Ketrosy) interstadial
I stadial
MIKULINO INTERGLACIAL
10940±150
11900±238
12300±140
?1675±250
18000±400
2400±300
24500±500
28100±1000
44400±2050
>44000
>45600
M7-8
M6
M5
M4
M3
M2

The *Volhyno-Podolian glacial-periglacial loess region* includes loess-till and loess regions of the Volhynian Upland, north-western Podolian Upland, and the denudational-accumulational plain of the Lesser Polessye between the two uplands. The most representative loess-paleosol series are found on interfluvial plateaus and upper slopes. Their stratigraphic division is based on schemes developed by A.B. Bogutsky (1975), M.F. Veklich *et al.* (1984), A.A. Velichko *et al.* (1992). A number of sections on the Volhynian Upland, in the area of Lesser Polessye and in the north of the Podolian Upland have been studied palynologically (Bezus'ko 1981, 1989; Gurtovaya 1981; Artyushenko *et al.* 1982; and others). Even though reliable data are rather few, they give an insight into special features of interglacial and interstadial floras and phytocoenoses of the region. They also allow to calculate climatic parameters for most of the established stages of the LPF development and to date the Kaidaki, Dnieper, and Priluki horizons and the Dubny soil more precisely (Bolikhovskaya 1995c). It has been found that the pollen spectra from the Priluki and Kaidaki soils together with the underlying loess-like loam, which are attributed by R.Ya. Arap to the Dnieper glacial epoch, are of interglacial type (with *Juglans, Carpinus, Quercus, Tilia, Acer, Morus, Ulmus, Corylus, Cornus, Rhamnus, Berberis* etc.). They were probably formed within a single interglacial rhythm i.e. during the Mikulino interglacial. The Vitachev soil corresponds to the expansion of extraglacial steppes, while at the time of the early Valdai loess formation periglacial steppes were widespread, with bush formations of microtherms *(Betula fruticosa, Alnaster fruticosus,* etc.). Pollen spectra described by E.E. Gurtovaya and L.G. Bezus'ko from the Dubny (= Bryansk) soil suggest, in our opinion, that this well developed soil ho-

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Fig. 2. Detailed climatostratigraphy of the Upper Pleistocene loess-paleosol formation. Landscape-climatic successions in the Desna-Dnieper glacial-periglacial (A) and Dniester-Prut extraglacial (B) loess regions during the Late Pleistocene. – *Landscape-climatic successions:* A: 1 = cryoarid tundra; 2 = tundra; 3 = forest tundra with birch open woodland; 4 = forest tundra with coniferous open woodland; 5 = forest steppe with birch open woodland; 6 = forest steppe with coniferous open woodland; 7 = steppe; 8 = dry steppe; 9 = extraglacial steppe; 10 = extraglacial forest steppe; 11 = pine-birch open woodland; 12 = pine forests; 13 = Siberian cedar pine-spruce and birch forests; 14 = birch forests with broad-leaved arboreal species; 15 = pine-birch forests with broad-leaved arboreal species; 16 = birch-pine forests with broad-leaved arboreal species; 17 = spruce-pine forests with broad-leaved arboreal species; 18 = mixed coniferous–broad-leaved forests; = hornbeam-oak forests; 20 = oak-hornbeam forests. B: 1 = tundra-forest steppe; 2 = forest steppe; 3 = steppe; 4 = pine open woodland; 5 = pine forests; 6 = extraglacial steppe; 7 = extraglacial dry steppe; 8–11 = extraglacial forest steppe: 8 = with spots of pine forests with oak, hornbeam, elm and linden; 9 = with spots of spruce-pine forests with oak, hornbeam, elm and linden; 10 = with spots of coniferous–broad-leaved forests with predominance of oak; 11 = with spots of broad-leaved forests with oak as edificator; 12 = pine forests with rare cryophytes; 13–17 = forest steppe: 13 = with pine forests with broad-leaved arboreal species; 14= with pine–broad-leaved forests; 15 = with Siberian cedar–broad-leaved forests; 16 = with oak forests; 17 = with hornbeam forests rizon interbedding in the Valdai loess series may be of different age in various parts of the Volhynia-Podolian region. In Volhynia (the Boyanichi section) it developed during one of the Middle Valdai cold phases, while in Lesser Polessye (Podbereztsy and other sections) it spans the warmest Middle Valdai interstadial as well as the preceding and subsequent coolings. The Dubny soil in the Izyaslav section (northern Podolian Upland) was dated to the end of the warmest Middle Valdai interstadial and the subsequent stadial cooling. Material of Late Valdai loess was accumulated in this region in periglacial environments characteristic of the cryoxerotic stage of the last ice age.

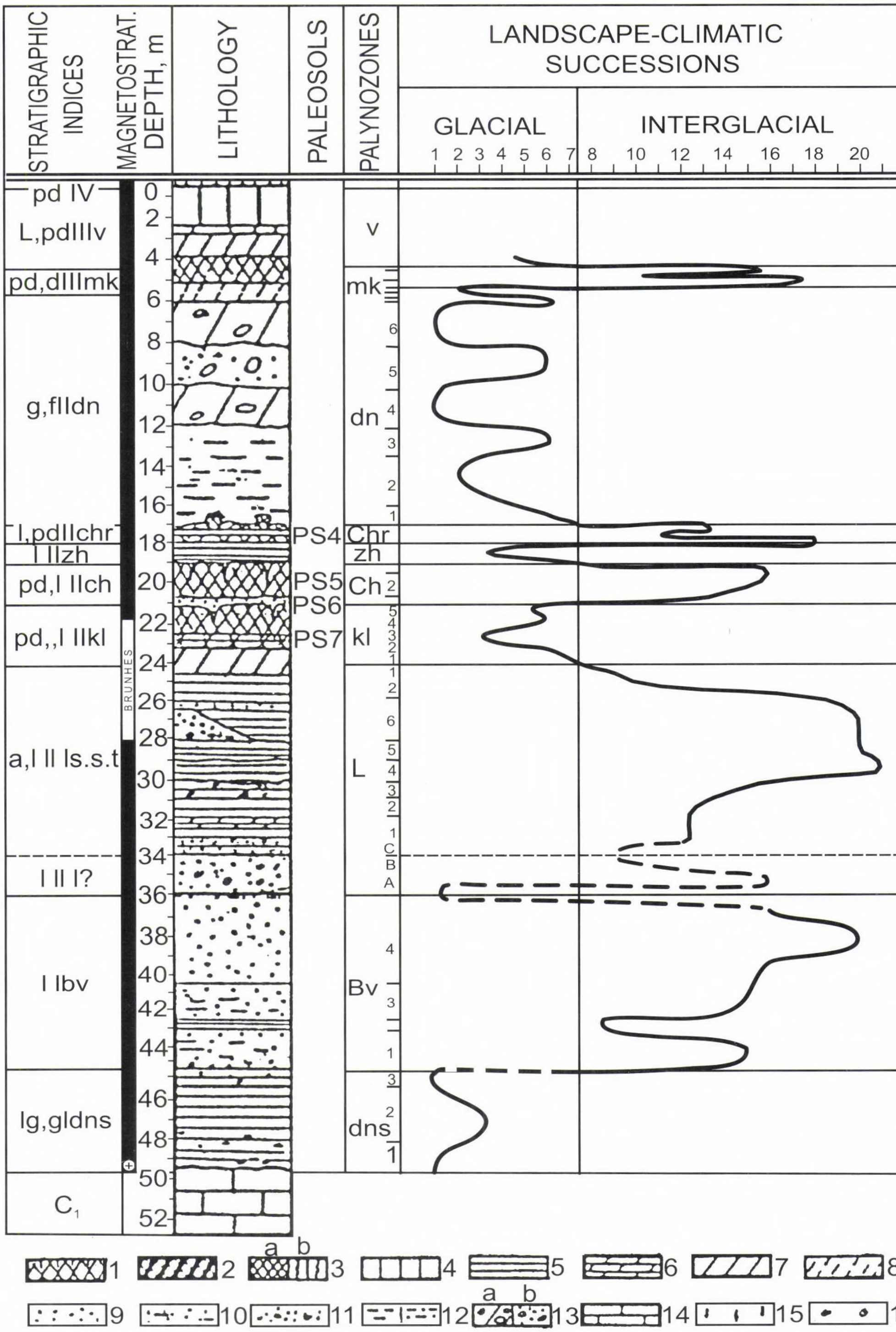
The *Desna-Dnieper glacial-periglacial loess region* is located north-east of the Dnieper Lowland, within the limits of the Dnieper ice sheet; it includes regions of the Upper Dnieper and Desna LPF. A most complex structure and considerable thickness are the characteristic features of the loess-paleosol series along the left bank of the Desna River. The author has performed a detailed analysis on the Arapovichi section which is one of most representative in the region and is located on the interfluvial plateau south-west of Novgorod-Seversky.

The section has been studied by A.A. Velichko, V.P. Grichuk, A.K. Markova, T.D. Morozova thoroughly, and by many others (see publications of 1957 to 1985), and it is one of the key sections for the stratigraphic scheme of LPF developed by the above named researchers.

Character and sequence of the palynozones identified in the section, regularities in the phytocoenotic successions, total composition of the palynoflora and individual significant species (Bolikhovskaya, 1991, 1993) testify that not only the Salyn' fossil soil belongs to the Mikulino interglacial, but also the whole series of formations overlying the till and the lower third of the Krutitsa soil (salt affected chernozem, according to T.D. Morozova 1981) *(Fig. 2A)*. Most typical of the Mikulino palynoflora in the Desna loess region are *Picea s. Omorica, P. s. Strobus, P. s. Cembra, Carpinus betulus, Fagus sylvatica, Quercus robur, Q. petraea, Q. pubescens, Tilia platyphyllos, T. tomentosa, T. cordata, Ulmus glabra, U. laevis, Celtis sp., Corylus colurna, Corylus avellana, Humulus lupulus, Lonicera,* etc. An Early Valdai interstadial has been identified within the Krutitsa soil. Another similar warming was revealed in the lower part of the Khotylevo loess horizon. Judging from palynological data, the Bryansk soil is a composite paleogeographic formation: its parent material and the genetic horizons themselves developed throughout three Middle Valdai interstadials and intermediate phases of cooling and at the beginning of the first (Ostashkov) Late Valdai stadial. The data argue against the previous belief about this soil being formed during only one interstadial. An interstadial (dated at about 16.500–15.000 yr. BP in the European sections) and three earlier warm intervals of interphase level are recorded in the sequence of the Desna and Altynovo loesses. The loess horizons in this region were accumulated in the environments of periglacial tundra and forest tundra during glacial stages, as well as in periglacial steppes and forest steppes corresponding to interphase and interstadial intervals of the Valdai epoch. The loess-like sandy loam underlying the Salyn' soil could be associated with the thermoxerotic maximum (hornbeam and oak forest phase) of the Mikulino interglacial. The fossil soils were formed under forests during both stages within the Mikulino interglacial rhythm, as well as under interstadial pine-birch forests (with oak, lime and elm) and in periglacial forest steppes. Soils also developed during interphase intervals in periglacial forest tundra, locally with open woodland composed of coniferous species and birch, and during the cryohygrotic stage of the Valdai glacial epoch when periglacial bush tundra dominated (*Fig. 2A*).

Th*e Northern Central Russian glacial-periglacial loess region* occupies the north of the Central Russian Upland within the limits of the most extensive Middle Pleistocene (Dnieper = Saale) glaciation. The composition and structure of the Quaternary sediments are represented most fully in the Likhvin section near Chekalin; the majority of paleogeographic events of the Pleistocene are recorded within the sequence. A 50 m thick sequence of loess, paleosols, tills and glacio-lacustrine, alluvial, lacustrine and bog sediments is exposed in a 2 km long scarp extending along the Oka River or in the nearby pits and boreholes. The literature on the section is quite voluminous. Most numerous are papers dealing with sediments of an oxbow lake (25 to 34 m from the top of the section) unanimously accepted as the stratotype for the Likhvin interglacial on the Russian Plain.

A comprehensive layer-by-layer characteristic of the whole sequence obtained by the author permitted its detailed subdivision and made possible the reconstruction of the diversified environmental and climatic events in the Upper Oka basin. The sequence spans the period from the Don glaciation to the Holocene (*Fig. 3),* that is six glacial epochs (Don, Oka, Kaluga, Zhizdra, Dnieper, Valdai) and six interglacials (Muchkap, Likhvin, Chekalin, Cherepet', Mikulino, and the Holocene); they are presented either as complete climatic rhythms of glacial and interglacial rank, or by considerable portions of climatic-phytocoenotic phases i.e. constituents of the rhythm (Bolikhovskaya 1995 b,c). Environments under severe climates of the periglacial tundra dominated by cryophytes *(Betula nana, B. fruticosa, Alnaster fruticosus, Dryas octopetala, Selaginella selaginoides* etc.) were characteristic of the Dnieper (= Saale) and Oka (= Elster) glaciations, with the ice sheet extending into the Upper Oka valley; they were also typical of the time when the Don glacio-lacustrine sediments accumulated. Lacustrine sediments of the Muchkap interglacial yielded palynospectra representing diversity of the interglacial flora. They include taxa characteristic of the Muchkap flora *(Tsuga canadensis, Picea s. Omorica, P. s. Eupicea, Abies sp., Pinus s. Cembra, P. s. Strobus, Larix sp., cf. Rhus sp., Carpinus betulus, C. orientalis, Fagus sylvatica, Quercus robur, Q. pubescens, Tilia platyphyllos, T. cordata, Ilex aquifolium, Ulmus laevis, U. glabra, U. campestris, Osmunda cinnamomea, O. claytoniana* etc.) along with species – indicators of this interglacial at the centre of the Russian Plain *(Cedrus sp., Tilia amurensis, Osmunda regalis, Woodsia manchuriensis, W. fragilis).* The Muchkap climatic optimum was marked by the prevalence of spruce and elm-oak-hornbeam forests. At the optimum of the Likhvin *s.str.* interglacial first oak-hornbeam forests dominated, later they had become replaced by spruce-fir and hornbeam-



STRATIGRAPHIC INDICES
MAGNETOSTRAT. DEPTH, m
LITHOLOGY
PALEOSOLS
PALYNOZONES
LANDSCAPE-CLIMATIC SUCCESSIONS
GLACIAL
INTERGLACIAL
pd IV
L,pdIIIv
pd,dIIImk
g,fIIdn
l,pdIIchr
l IIzh
pd,l IIch
pd,,l IIkl
a,l II ls.s.t
l II l?
l Ibv
lg,gldns
C1
BRUNHES
PS4
PS5
PS6
PS7
v
mk
dn
Chr
zh
Ch
kl
L
Bv
dns

beech-oak forests. Typical of the Likhvin flora are representatives of the European, Mediterranean, East Asian and North American floras, such as *Larix sp., Abies alba, Picea s. Omorica, P. excelsa, Pinus s. Cembra, P. s. Strobus, P. sylvestris, Betula s. Costatae, B. pendula, B. pubescens, Juglans regla, Carpinus betulus, Fagus sylvatica, Quercus petraea, Q. robur, Q. pubescens, Zelkova sp., Celtis sp., Ulmus propinqua, U. laevis, U. campestris, Fraxinus sp., Tilia platyphyllos, T. tomentosa, T. cordata, Acer sp., Corylus colurna, C. avellana, Alnus glutinosa, A. incana, Ligustrina amurensis, Rhododendron sp., Vitis sp., Myrica sp., Osmunda cinnamomea, Salvinia natans* and others, including species-indicators *Tsuga canadensis, Taxus baccata, Pterocarya fraxinifolia, Juglans cinerea, Castanea sativa, Ilex aquifolium, Fagus orientalis, Quercus castaneifolia, Buxus sp., Osmunda claytoniana.* During the Kaluga cool interval, in periglacial environments of forest tundra, lacustrine and fluvial sediments formed, as well as the overlying PS7 soil and parent material of the PS6 soil did, showing post-cryogenic structure. The Chekalin interglacial (its peak of heat and moisture supply was marked by the dominance of spruce-broad-leaved forests) has been recorded in the pedocomplex including paraburozem PS5 and podzolic PS6 soils. Characteristic floristic elements of this thermochron are *Picea s. Omorica, P. excelsa, Pinus s.g. Cembra, P. sibirica, P. sylvestris, Betula pendula, B. pubescens, Carpinus betulus, Quercus robur, Tilia cordata, T. platyphyllos, T. tomentosa, Acer sp., Ulmus laevis, U. glabra, U. campestris,* etc. The flora of periglacial forest-tundra of the Zhizdra cooling recovered from the lake and bog deposits includes less diversified cryophytes compared to the Kaluga flora; the latter is represented *by Larix sp., Pinus sylvestris, Betula pubescens, B. pendula, B. fruticosa, B. nana, Alnaster fruticosus, Dryas octopetala, Selaginella sibirica, Lycopodium appressum, L. pungens, Artemisia s.g. Seriphidium, Thalictrum sp.,* and others. Bog gleyed soil PS4 developed during the Cherepet' interglacial; its optimum phases are marked by hornbeam-oak and

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Fig. 3. Stratigraphic subdivision of the Likhvin key section. Phytocoenotic and climatic successions on the territory of the Upper Oka loess region in the Pleistocene (according to palynological data). *Lithology:* 1 = recent soil; 2 = Eopleistocene fossil soils and pedosediments (see *Fig. 4*); 3 = Pleistocene paleosols: a) horizon A; b) horizon B; 4 = loess and loess-like deposits; 5 = clay; 6 = foliated marl; 7 = loam; 8 = loamy sand; 9 = sand; 10 = sand with loamy interlayers; 11 = sand with gravel and pebble; 12 = fluvioglacial deposits; 13 = till; a) clayey; b) sandy; 14 = limestone; 15 = inclusions of carbonate concretions; 16 = mollusc shells. – Landscape-climatic successions: 1 = ice cover; 2 = periglacial tundra; 3 = periglacial forest tundra; 4 = periglacial steppe; 5 = periglacial forest steppe; 6 = pine-birch open woodland; 7 = larch-pine-birch open woodland; 8 = spruce forest; 9 = birch forests with broad-leaved arboreal species; 10 = pine-birch forests with broad-leaved arboreal species; Il = birch-pine forests with broad-leaved arboreal species; 12 = pine-spruce forests with broad-leaved arboreal species; 13 = birch-broad-leaved forests; 14 = pine-cembra pine-broad-leaved forests; 15 = pine-spruce-broad-leaved forests; 16 = spruce-broad-leaved forests; 17 = spruce-fir-broad-leaved forests; 18 = broad-leaved (*Quercetum mixtum*) forests; 19 = broad-leaved shady (with *Carpinus betulus* domination) forests; 20 = spruce-broad-leaved forests with subtropical taxa; 21 = broad-leaved forests with subtropical taxa

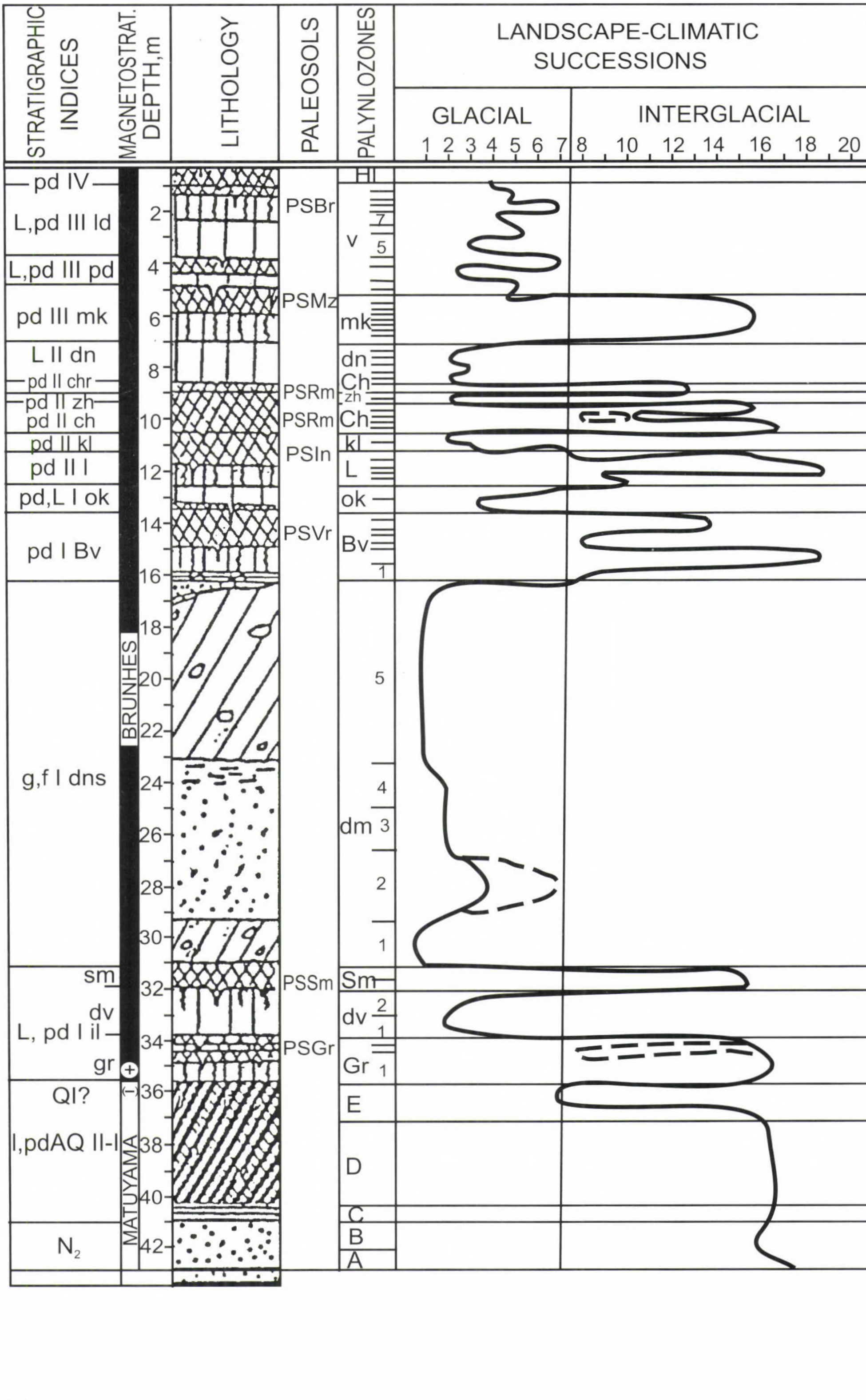
coniferous/broad-leaved forests with *Dinus s. Cembra, Pinus sylvestris, Betula pendula, B. pubescens, Carpinus betulus, C. cf. orientalis, Ostrya sp. Quercus robur, Q. cf. pubescens, Tilia cordata, T. tomentosa, Ulmus laevis, U. campestris* etc. among characteristic taxa.

During the Dnieper *s.l.* stage the following series were deposited: a) Early Dnieper glacio-fluvial silts with lemming fauna: *Dicrostonyx cf. simplicior, Lemmus sibiricus etc.* (Agadjanian 1973), and palynospectra mostly of tundra steppe type (Bolikhovskaya 1975); b) three layers of till attributed to the Dnieper and Moscow stadials and to the Dnieper-Moscow interstadial; the landscapes of the latter were dominated by open woodlands of pine, *Alnaster* and dwarf birch; c) the Late Moscow loess-like sandy loam. An Early Dnieper interstadial has been identified in the upper part of silts underlying the till; periglacial open woodlands with pine prevailed at that time. Late Moscow interstadial warming (established in ferruginous sands above the tills) is represented by a phase of periglacial birch open woodlands with *Betula fruticosa* in the shrub layer and a cover of herbs and dwarf shrubs (with *Arctous alpina, Cannabis sp., Artemisia s.g. Seriphidium, Thalictrum cf. alpinum* and others). The Mikulino interglacial (when a lessivated soil of PS2 complex formed) was marked by a dominance of forests, such as pine-birch ones at the beginning and at the end of the thermochron; elm-hornbeam-oak and birch forests at the thermoxerotic maximum; oak-elm-linden-hornbeam and spruce-pine-birch forests at the thermohygrotic maximum. The first Early Valdai cold phase established at the top of A2 horizon of forest soil was characterised by the dominance of birch forests, with *Betula fruticosa* and *B. nana* in the shrub layer.

The first Early Valdai interstadial is manifested in the sod horizon of PS2 complex by a palynozone indicative of periglacial forest steppes with steppe herb and grass communities and those of chenopods and wormwood; meadow coenoses; pine-spruce-birch forests with an admixture of oak and linden. The onset of the succeeding loess formation was marked by a sharp increase in continentality of the climate involving the expansion of tundra steppes; the periglacial flora of the latter is characterised by the presence of pollen and spores of *Pinus sylvestris, Betula pendula, B. pubescens, B. fruticosa, B. nana, Alnaster fruticosus, Cannabis sp., Ephedra strobilacea, Selaginella sibirica,* and others. Final stages of the loess formation took place during the second Early Valdai interstadial, when dominating forests were composed of *Pinus sylvestris, P. sibirica, Picea excelsa.* The overlying fossil soil PS1 (correlated unanimously by paleopedologists with the Bryansk soil), as well as the uppermost loess horizon, formed in periglacial environments dominated by birch open woodlands with *Juniperus sp., Betula fruticosa, Alnaster fruticosus* in shrub layer.

As it is evident from the foregoing, fossil soils of the Northern Central Russian glacial-periglacial loess region developed during interglacials (PS6, PS5, PS4, PS3, and PS2 forest soil) and an interstadial (sod soil of PS2 complex); some of the soils (PS7 and PS1) formed under stenoperiglacial conditions corresponding to glacial stages of glacial epochs. Loess horizons in this region formed under glacial climate only.

The *Oka-Don glacial-periglacial loess region* occupies the lowland of the same denomination and the eastern part of the Central Russian Upland within the limits of the Don ice lobe. Detailed reconstructions of landscapes and climates of the region during all the main stages of the LPF development have been worked out by the author using data obtained by studies of the Strelitsa key section located on the right bank of the Don River, 20 km from Voronezh. An almost complete sequence of the LPF is seen there in a number of quarries at the upper slope of the Devitsa River valley. Investigations carried out for many years provided detailed palynological data for the whole sequence of subaerial, lacustrine and fluvioglacial formations underlying and overlying the till deposited during Don glaciation (Bolikhovskaya 1975, 1995 b,c). Floristic, phytocoenotic and climatic successions reconstructed for all the stages of the LPF formation in the discussed region have been compared with results of lithological, paleopedological, paleofaunistic, paleomagnetic and other analyses of the exposed deposits (Krasnenkov *et al.* 1970; Agadjanian 1971; Bolikhovskaya 1984 and others; Udartsev 1980; Velichko *et al.* 1985; Zarrina and Krasnov 1985). It has turned out that in most cases there is a discrepancy between the boundaries of glacial and interglacial rhythms on the one hand, and loess and fossil soil boundaries, on the other hand *(Fig. 4)*. The submorainic loess-paleosol series superimposing the Eopleistocene red beds formed in the course of the Ilyinka interval, which comprises two interglacial rhythms and the intermediate cooling. During the early Ilyinka (Gremyach'e) interglacial steppes and forest steppes prevailed and the lower soil of this series developed at that time; during the Devitsa cooling, under conditions of periglacial tundra and forest tundra, the middle horizon of loess (separating two soils), the parent sediment of the B soil horizon, and the base of A horizon of the upper soils were formed. The overwhelming part of the thick humus horizon of the upper soil (its top has been eroded) corresponds to the Semiluki (late Ilyinka) interglacial. Palynological data on the post-Don loess-paleosol members testify that subhorizons $A1^1$ of the Vorona PC, Inzhavino, Kamenka, and Krutitsa soils correspond to cryohygrotic stages or substages of cold epochs. There is only one (Vorona) soil in the second Pre-Dnieper PC; it developed throughout the Muchkap interglacial (with coniferous/broad-leaved forests dominant at its optimum) and during the first half of the Oka glaciation. Material of the Korostylevo loess was accumulated, horizon $A1^1$ of the Vorona soil was formed and penetrated by ice wedges under conditions of the Oka periglacial environment. The upper Pre-Dnieper PC of the sequence, including the Inzhavino, Kamenka and Romny fossil soils, was formed in the course of three interglacial and two glacial climatic rhythms. Most part of the Inzhavino soil developed under forests of the Likhvin *s.str.* interglacial (pine-birch forests→birch forests with some *Carpinus betulus* and *Carpinus orientalis*→birch-cembra pine-spruce-pine forests with oak and elm→pine-fir-cembra pine-spruce forests with hemlock, hazel, hornbeam and beech→fir-spruce and beech-hornbeam-elm-oak forests→forest steppes with patches of broad-leaved forests→coniferous/broad-leaved and birch forests). The Likhvin flora includes taxa as *Tsuga canadensis, Abies sp., Picea s. Omorica, P. s.*


STRATIGRAPHIC INDICES
MAGNETOSTRAT.
DEPTH,m
LITHOLOGY
PALEOSOLS
PALYNLOZONES
LANDSCAPE-CLIMATIC SUCCESSIONS
GLACIAL
INTERGLACIAL
1 2 3 4 5 6 7 8 10 12 14 16 18 20
pd IV
L,pd III ld
L,pd III pd
pd III mk
L II dn
pd II chr
pd II zh
pd II ch
pd II kl
pd II l
pd,L I ok
pd I Bv
g,f I dns
sm
dv
L, pd I il
gr
QI?
l,pdAQ II-I
N_2
BRUNHES
MATUYAMA
PSBr
PSMz
PSRm
PSRm
PSIn
PSVr
PSSm
PSGr
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v
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ok
Bv
dm
Sm
dv
Gr
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D
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Eupicea, Pinus s. Cembra, P. s. Strobus, P. sylvestris, Larix sp., Betula s. Costatae, B. pendula, B. pubescens, Juglans regia, Carpinus betulus, C. orientalis, Quercus robur, Q. petraea, Q, pubescens, Ulmus laevis, U. carpinifolia, Humulus lupulus, Euonymus sp. In terms of taxa diversity and participation of Neogene relicts, it is on a par with that of the preceding Muchkap thermochron (the latter includes *Abies sp., Picea s. Omorica, P. s. Eupicea, Pinus s. Cembra, P. s. Strobus, P. sylvestris, Betula s. Costatae, B. pendula, B. pubescens, Zelkova sp., Carpinus caucasica, C. betulus, C. orientalis, Ostrya sp., Corylus colurna, Acer sp., Quercus petraea, Q. robur, Q. pubescens, Tilia platyphyllos, T. tomentosa, T cordata, Lonicera sp., Rhamnus* sp., *Osmunda cinnamomea* and others). The Kaluga (= Borisoglebsk) cooling was characterised by dominance of periglacial tundras and forest-tundras. The successions of the Kamenka interglacial soil formation is reconstructed as follows: forest steppes, locally with linden-hornbeam-oak and birch-pine forests→herb and grass steppes→pine-birch, oak-hornbeam and alder forests→pine-birch forests of the endothermal interval→forest sleppes. At the Zhizdra (Orchik) cooling, the periglacial steppes of the first phase were replaced by periglacial tundras. The optimum stage of the Romny pedogenesis featured hornbeam-oak forests with *Carpinus orientalis* and *Ostrya sp.*, alder forests and coniferous-birch stands. The Dnieper loess formed in cryoarid environments of periglacial tundras. Most part of the Mezin PC was associated with the Mikulino interglacial forest steppes, which were considerably reduced in area in the Upper Don drainage basin at the phase of dominance of hornbeam-oak and birch-pine forests. The first Early Valdai cooling and the subsequent interstadial can be correlated with the upper subhorizon $A1^1$ of the Krutitsa soil. 5 interstadials and 6 alternating cooling of stadial level are recorded along the interfluvial LPF profiles dated to the Early and Middle Valdai. The Late Valdai horizons are found in the cover sediments of the first and second terraces of Don. On the northern Central Russian Upland the materials obtained so far do not permit identification of interglacial loess formations. The loess horizons of the Oka-Don lowland are generations of glacial climate.

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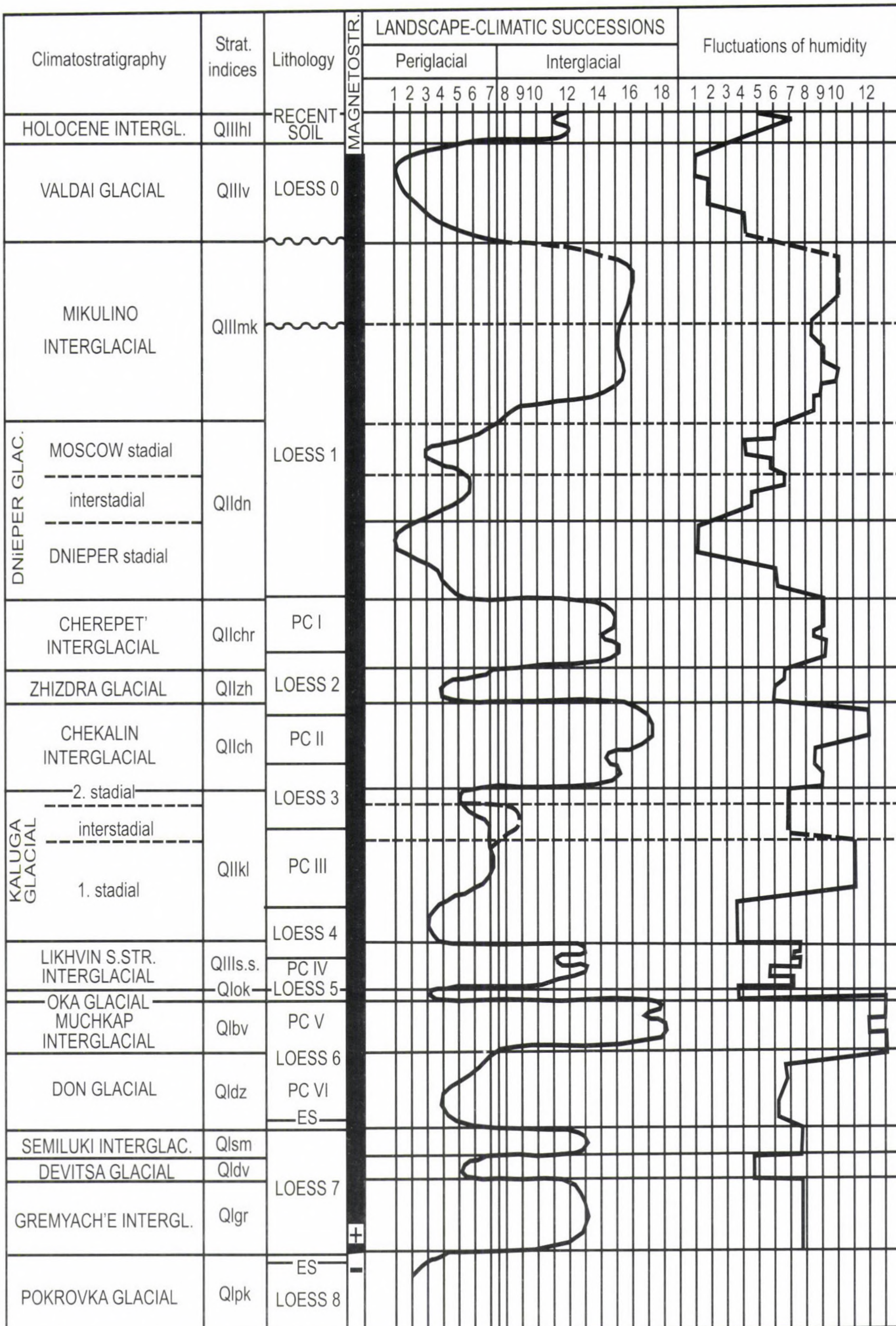
Fig. 4. Stratigraphic subdivision of the Strelitsa key section. Phytocoenotic and climatic successions on the territory of the Upper Don loess region through the Pleistocene (according to palynological data; for symbols of lithological column see *Fig. 3*). – Landscape-climatic successions: 1 = ice cover; 2 = periglacial tundra; 3 = periglacial forest tundra; 4 = periglacial steppe; 5 = periglacial forest steppe; 6 = pine-birch open woodland; 7 = extraglacial steppe; 8 = pine-birch forests; 9 = birch forests with broad-leaved arboreal species; 10 = pine-birch forests with broad-leaved arboreal species; 11 = birch-pine forests with broad-leaved arboreal species; 12 = spruce-pine-birch forests with broad-leaved arboreal species; 13 = pine-birch-broad-leaved forests; 14 = spruce-pine-birch-broad-leaved forests; 15 = broad-leaved forests; 16 = forest steppe; 17 = steppe; 18 = coniferous forests wilh rare subtropical taxa; 19 = mixed coniferous-broad-leaved forests with rare subtropical taxa; 20= broad-leaved forests with subtropical taxa


Climatostratigraphy
Strat. indices
Lithology
MAGNETOSTR.
LANDSCAPE-CLIMATIC SUCCESSIONS
Periglacial
Interglacial
Fluctuations of humidity
1 2 3 4 5 6 7 8 9 10 12 14 16 18
1 2 3 4 5 6 7 8 9 10 12
HOLOCENE INTERGL.
QIIIhl
RECENT SOIL
VALDAI GLACIAL
QIIIv
LOESS 0
MIKULINO INTERGLACIAL
QIIImk
DNIEPER GLAC.
MOSCOW stadial
interstadial
DNIEPER stadial
QIIdn
LOESS 1
CHEREPET' INTERGLACIAL
QIIchr
PC I
ZHIZDRA GLACIAL
QIIzh
LOESS 2
CHEKALIN INTERGLACIAL
QIIch
PC II
KALUGA GLACIAL
2. stadial
interstadial
1. stadial
QIIkl
LOESS 3
PC III
LOESS 4
LIKHVIN S.STR. INTERGLACIAL
QIIls.s.
PC IV
OKA GLACIAL
Qlok
LOESS 5
MUCHKAP INTERGLACIAL
Qlbv
PC V
LOESS 6
DON GLACIAL
Qldz
PC VI
ES
SEMILUKI INTERGLAC.
Qlsm
DEVITSA GLACIAL
Qldv
LOESS 7
GREMYACH'E INTERGL.
Qlgr
+
−
ES
POKROVKA GLACIAL
Qlpk
LOESS 8

The *extraglacial loess region of the East Caucasian piedmonts* is one of the most remote areas from the glaciated territories of the Russian Plain. The loess-paleosol series represent maximum thickness (70–100 m) on the European subcontinent. Previous attempts at stratigraphic subdivision of the LPF in the Caucasian piedmont and its correlation with horizons of glacial and periglacial zones of the Russian Plain were based on lithological and geochemical analyses, as well as on paleomagnetic and thermoluminescent data (Balaev and Tsarev 1964; Shelkoplyas *et al.* 1987; Fainer and Lizogubova 1987; and others). The author participated in the study of the most representative LPF sections in the extraglacial zone as a member of a group of specialists under the guidance of A.A. Velichko. A 140 meter sequence of Eopleistocene and Pleistocene deposits was studied, exposed in natural scarps and penetrated by boreholes on the interfluve and on the Kuma valley slopes and terraces near Otkaznoye village (Morozov 1989; Udartsev *et al.* 1989; Virina *et al.* 1990; Bolikhovskaya 1995a,b,c; and others). Using data obtained from studies of paleomagnetism, paleosols, small mammals, as well as palynological characteristics, the author produced reconstructions of climates and environments for all 15 climatochrons of the Brunhes epoch of normal polarity *(Fig. 5).*

For the most part of the period of more that 700 thousand years duration (from the Gremyach'e interglacial to the present days) the western part of the Terek-Kuma lowland was occupied by forest steppe interglacial landscapes or by periglacial and extraglacial forest steppes. For the first time in the Pleistocene history, steppes became dominant here during one of the phases of the thermoxerotic interval of the Likhvin interglacial. Vegetation of dry steppes and semideserts prevailed in the Middle Kuma basin at some warm Eopleistocene intervals; during the Pleistocene it as-

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Fig. 5. Phytocoenotic and climatic successions of the Middle Kuma basin during the Pleistocene (according to palynological data). – Landscape-climatic successions: 1 = periglacial semi-desert and dry steppe; 2 = periglacial steppe; 3 = periglacial forest steppe; 4 = birch and coniferous-birch open woodland; 5 = extraglacial forest steppe; 6 = extraglacial birch open woodland; 7 = extraglacial spruce and cembra pine-spruce forests; 8 = birch open woodland with broad-leaved arboreal species; 9 = birch forests with broad-leaved arboreal species; 10 = coniferous-birch and birch-coniferous forests with broad-leaved arboreal species; 11 = forest steppe; 12 = steppe; 13 = piedmont forest steppe; 14 = shrub hornbeam groves; 15 = elm-oak, oak, hornbeam-oak forests; 16 = hornbeam forests; 17 = oligo- and polydominant broad-leaved forests; 18 = polydominant broad-leaved forests with subtropical taxa. – Fluctuations of humidity curve: 1 = periglacial semi-desert and dry steppe (annual precipitation <250 mm); 2 = interglacial desert and semi-desert (250 mm and less); 3 = periglacial steppe (280–300 mm); 4 = periglacial forest steppe (300–450 mm); 5 = interglacial steppe (300–450 mm); 6 = periglacial open woodland (400–500 mm); 7 = interglacial forest steppe (400–650 mm); 8 = interglacial open woodland (600–700 mm); 9 = interglacial oak forest (550–700 mm); 10 = interglacial hornbeam forests (700–800 mm); 11 = extraglacial spruce and cembra pine-spruce forests (up to 800 mm); 12 = oligo- and polydominant broad-leaved forests under interglacial temperate climate (up to 1500 mm); 13 = polydominant broad-leaved forests under interglacial subtropical climate (>1500 mm)

sumed greater importance first at the cryoxerotic substage of the Dnieper ice age, and later at the same stage of the Valdai glaciation; the vegetation at those stages bore evidence of periglacial phytocoenoses. Dominance of forest landscapes in the region has been established with certainty for five intervals. At the Muchkap interglacial broad-leaved polydominant and oligodominant forests prevailed with a considerable share of subtropical species. Broad-leaved forests, mesophytic or xerophytic to a different degree, occupied the Middle Kuma basin during the Chekalin, Cherepet', and Mikulino interglacials. Forests of spruce and Siberian cedar pine dominated the territory at individual phases of the cryohygrotic stage of the Kaluga glaciation. The five "forest" periods mentioned above are assumed to correspond to the maximum levels of the Caspian Sea during the Pleistocene.

Automorphic fossil soils of the eastern part of the Caucasus piedmont were formed during interglacials, interstadials and cryohygrotic stages of ice ages, while loess horizons developed during glacial epochs, as well as at thermoxerotic stages and endothermic coolings of interglacials. Taphonomic and ecological-coenotic characteristics of the studied palynofloras confirmed conclusion of many researchers about primarily eolian origin of the terrigenous loess material on the Terek-Kuma Lowland.

Conclusions

1. The LPF development on the Russian Plain comprises 17 paleogeographic stages (9 interglacials and 8 glacial epochs between them). Climatic rhythms reconstructed on the basis of palynological studies of the most representative sections within glacial-periglacial and extraglacial zones were compared with paleomagnetic data from the same sections obtained by M.A. Pevzner, S.S. Faustov, A.N. Tretyak and others. It appeared that the Brunhes epoch of normal polarity spanned 8 interglacial and 7 intervening glacial climatic rhythms. The Matuyama/Brunhes reversal is located at the base of the Gremyach'e (= Westerhoven) interglacial (see *Figs 4, 5).*

2. Smaller climate-stratigraphic units are identified within climatic rhythms of the glacial and interglacial levels: those are endothermal coolings, thermoxerotic and thermohygrotic stages and substages of interglacial climatic rhythms; stadials, interstadials, interphasials, cryohygrotic and cryoxerotic stages and substages of glacial climatic rhythms. *Endothermals* have been identified in a majority of interglacials. The Gremyach'e, Muchkap, Likhvin *s.str.,* Chekalin and Cherepet' interglacials featured one endothermal each, separating thermoxerotic and thermohygrotic stages of the interglacial in concern *(Figs 3–5);* in loess-paleosol formation of the Mikulino interglacial two endothermals were recorded, one between the stages of this rhythm, and another one in the first half of the interglacial *(Figs 2–5).* The most complicated pattern of climatic rhythms has been reconstructed for four glacial stages of the LPF development: the Valdai stage features 10 stadial intervals, 9 interstadials and several interphasials; the Dnieper glacial rhythm is divided by a prolonged (Odintsovo?)

interstadial into two (Dnieper and Moscow) stages, with Early Dnieper and Late Moscow interstadials within them *(Fig. 3)*; deposits of the Don and Kaluga glacial stages have been found to contain one interstadial each *(Figs 3* and *4)*.

3. The cold stages of longest duration in the LPF formation, marked by the expansion of ice sheets, were undoubtedly the Don glacial epoch, both stages (Dnieper and Moscow ones) of the Dnieper glaciation, and the Valdai glaciation. It is strongly suggested by the composition of periglacial and glacial palynofloras and reconstructed phytocoenoses, as well as by thickness of periglacial and glacial deposits, the presence of prolonged interstadials and distinct regular alternation of stages and substages within climatic rhythms.

The reconstructions indicate a wide expansion of periglacial tundras and forest tundras in the central regions of the Russian Plain, and dominance of periglacial steppes and forest steppes (less frequently – tundra-forest steppes) in the south; they also suggest a considerable complexity of succession processes in the evolution of phytocoenoses. Ice sheets probably covered the north of the Russian Plain during all the cold stages, and occasionally penetrated the central regions of the plain. Several cold stages in the LPF development – Pokrovka, Devitsa, Kaluga and Zhizdra – which cannot be reliably correlated with till horizons on the Russian Plain resembled closely the Don, Dnieper and Valdai glaciations in the scale of climatic changes and regarding transformation of ecosystems.

4. The results of detailed studies of an almost continuous succession series of the Pleistocene interglacial and periglacial floras made necessary to revise the conclusion drawn by V.P. Grichuk (1989) who considered impossible a paleogeographic situation when on the whole territory of extra-tropical Eurasia an interglacial characterised by a flora poor in exotic elements preceded another interglacial with richer flora. A comparison has been made between series of palynofloras recovered from most complete, almost continuous Pleistocene sequences in the glacial-periglacial and extraglacial loess regions. It appeared that even though the process of depletion of exotic elements in interglacial floras had been undoubtedly in progress throughout the Cenozoic, it was interrupted occasionally during the Early and Middle Pleistocene by the appearance of floras marked by more diverse composition (at levels of species and genera) and richer in Neogene relicts as compared with preceding interglacials. A thorough analysis of the LPF sections in the extreme south-east of the East European loess province (in the middle reaches of the Kuma river) revealed that older interglacial floras – Gremyach'e (= Early Ilyinka) and Semiluki (= Late Ilyinka) – include less Neogene relicts and less taxa altogether than the younger Muchkap warm flora: in the Gremyach'e flora there are less than 50 taxa, amoug them *Nedrus sp., Picea s. Omorica, Betula s. Costatae, Fagus orientalis, Quercus robur, Q. petraea, Q. castaneifolia, Q. ilex, Carpinus caucasica, C. betulus, C. orientalis, Ostrya cf. carpinifolia, Corylus colurna, Tilia platyphyllos, T. tomentosa, T. cordata, Morus sp.* and others, while the Muchkap flora contains 90 taxa, including *Tsuga canadensis, Cedrus, Pinus s. Cembra, Pterocarya pterocarpa, Carya, Juglans cinerea,*

J. regia, Liquidambar, Castanea, Celtis, Ilex aquifolium, Fagus orientalis, F. sylvatica, Carpinus caucasica, C. betulus, C. orientalis, Hedera, Kalonymus, Staphylea, Daphne, Rhododendron, Osmunda regalis, O. claytoniana, O. cinnamomea and many others. By contrast, within the present-day forest zone (the Upper Oka region) the younger Likhvin (= Holstein) flora is by far richer than the Muchkap (= Belovezhsk) one in practically every index. The analysis of quantitative and qualitative changes of palynofloras within the East European loess province strongly suggests a progressive depletion of Neogene relicts in the Pleistocene interglacial floras along with reduction of the taxa number since the Likhvin *s.str.* interglacial in the loess regions within the limits of the modern forest zone, while in the present-day steppe and forest steppe these processes began in the Muchkap interglacial.

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Comparative studies of terraces and loess sections in the vicinity of Krasnoyarsk in the Yenisey valley

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Introduction

For several decades a recurring topic of investigations – both from theoretical and methodical aspects – have been comparative studies of characteristic valley sections of the European and Asian big rivers. An essential part of these studies were joint field surveys with the involvement of local experts of international experience focusing on the geological evolution of fluvial terraces and loess profiles.[4]

In the beginning of the 1990s at a conference of Quaternary geology and geomorphology held in Central Yakutia (centred in Yakutsk) Hungarian experts reported on the loess profiles and chronological subdivision of the Central Danube Basin; previously a similar lecture was held at an INQUA Loess Commission meeting organised on the Chinese Loess Plateau, at Xian and Luochuan (Pécsi 1987a). An agreement had been reached that specialists involved in the study of Quaternary fluvial sediments and loess-paleosol sequences in China (Xian, Beijing, Nanjing) and in Siberia (Novosibirsk, Krasnoyarsk) should study loess and terrace formations of the Danubian Basin, while Hungarian experts would be given a chance to get acquainted with similar localities along the major rivers of China and Siberia in the framework of exchange of methodologies and experience.

Of these field excursions the ones covering the Middle Yenisey Valley took place in 1992 and 1995 organised by A.F. Yamskikh, head of the Laboratory of Paleogeography, Krasnoyarsk Teachers's Training School, who had elaborated loess se-

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[4] During field excursions of international conferences or in the course of bi- or multilateral research projects we had opportunity to partake at field excursions along the representative stretches of large rivers of Central and Western Europe (Danube, Vistula, Elba, Rhine, Main, Seine...), then those of Eastern Europe and Asia (Dniester, Don, Volga, Sirdarya, Amudarya, Yenisey, Lena, Aldan, Hwang He, Ganges...), the Mississippi-Missouri (USA) and Paraná-Paraguay (Argentine).

ries of the river valleys along Yenisey and tributaries in his dissertation of doctor of sciences (Yamskikh 1992). The Hungarian party was represented by M. Pécsi and F. Schweitzer, Quaternary and loess experts from the Geographical Research Institute Hungarian Academy of Sciences (GRI HAS). Relevant type horizons of sections containing terrace material and loess sediments were studied jointly and sampled for subsequent TL analyses. M. Frechen from the Institute of Geology, University of Cologne (Germany) was requested to carry out work on dating and he undertook the task. Later Dr. Frechen at the invitation of the Siberian partner institute studied the key sections personally and collected samples for further TL analyses in 1995.

Two paleogeographers from the Krasnoyarsk institute came to Hungary in autumn 1992 and made study trips to examine key loess profiles in more detail. The first part of the analyses was finished by 1997. First and preliminary summary of ideas and results is to be attempted below.[5]

Discussion

On the Yenisey a huge dam was constructed at Divnogorsk, raising water level by 100 m, before the river leaves its mountain reach. The medium level of the dammed section (up to Abakan) is situated at 243 m a.s.l. while below the reservoir (down to Krasnoyarsk located at a 40 km distance) the medium level changes between 142 and 130 m. This way along the (300 km long) water reservoir[6] valley terraces lower than 100 m have been inundated. Natural fluvial terraces of Yenisey might be studied recently in the Krasnoyarsk Basin and along the northern foothills of the Eastern Sayan Mountains (Yamskikh 1992).

Terraces along the middle and upper reaches of Yenisey (according to the variation of mountain and basin morphostructures) differ in their relative altitudes and their number also changes. A similar pattern can be recognised along the valley sections of the Danube breaking through the Carpathians and crossing the enclosed basins (Pécsi 1959, 1971).

Several researchers studied the geological-geomorphological position, composition and age of the terraces within the mentioned sections. These investigations and results achieved by other experts were summarised recently by A.F. Yamskikh (1993, *Fig. 1* and *Table 1*).

Laboratory of Paleogeography of Krasnoyarsk Teachers' Training School predominantly has dealt with the complex task of the explanation and dating of polycyclic and polygenetic (fluvial, deluvial, eolian) evolution of lower terraces of the Yenisey valley. That is why joint field surveys were focused on the geological and

[5] Due to the considerable geographical distance a preliminary confer between the authors was considered indispensable.

[6] Between Minusinsk and Divnogorsk.

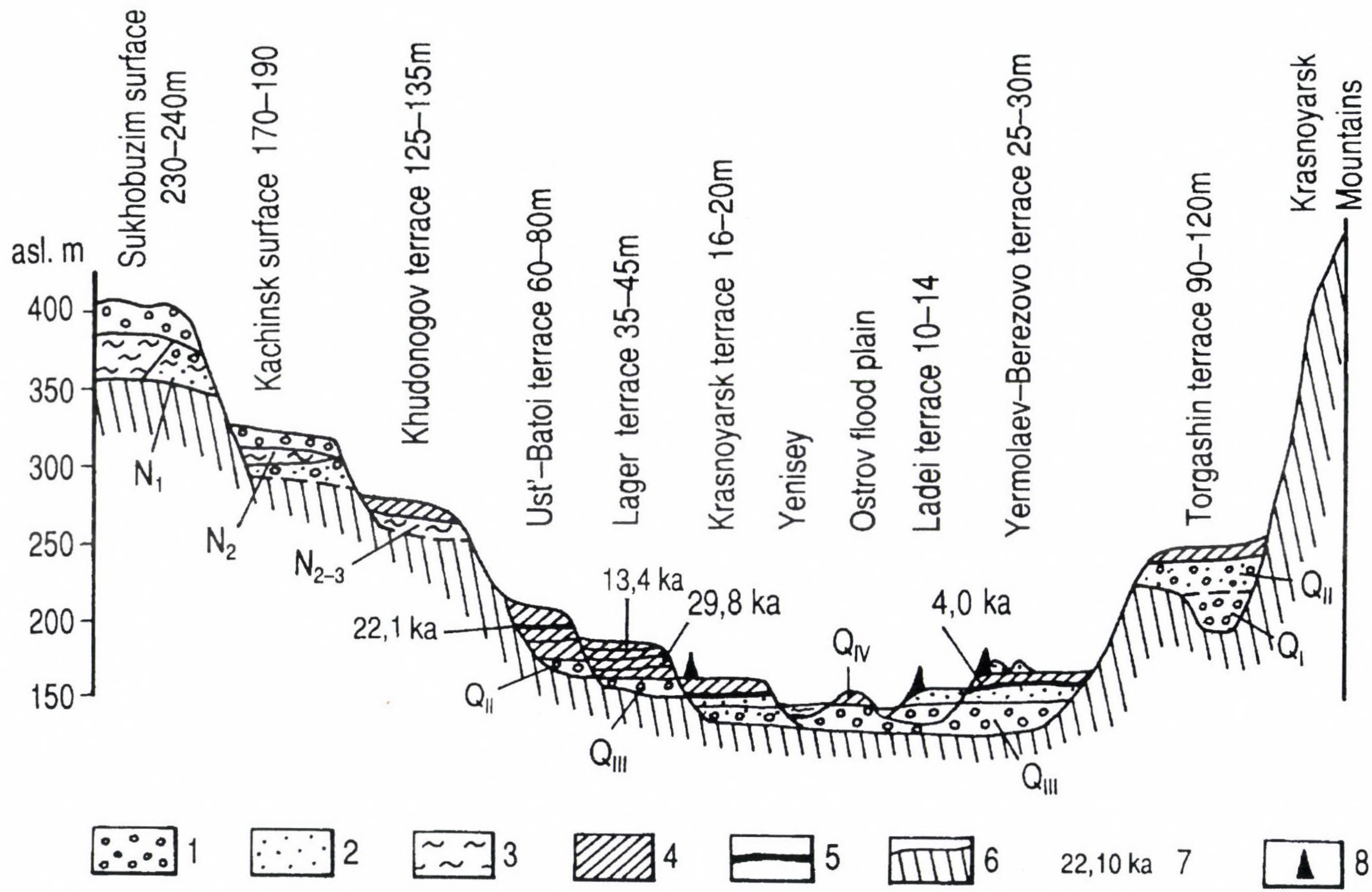


Fig. 1. Terraces of the Yenisey in the vicinity of Krasnoyarsk, ca 56° northern latitude (according to Yamskikh 1993). – 1 = gravel; 2 = sand; 3 = clay, loam; 4 = loess-like sediments; 5 = paleosols; 6 = bedrock; 7 = ^{14}C age; 8 = site of Early man

Table 1. Terraces along the middle reaches of Yenisey: a comparison between the different concepts of authors (after Yamskikh 1993)

Geomorphic levels	Yamskikh 1993	Finarov 1964	Arkhipov 1966	Borisov 1984
Flood plain level	4–7/12 m	–	–	2–6 m
Terrace I	5–12 m	8–14 m	10–12/8–11 m	4–8 m
Terrace II	14–18 m	15–25 m	15–18/12–15 m	12–15 m
Terrace III	24–30 m	30–36 m	23–27/17–25 m	15–25 m
Terrace IV	35–45	40–60 m	30–35 m	25–35 m
Terrace V	45–55–60 m	60–80 m	40–45/40–50 m	35–60 m
Terrace VI	60–80 m	100–120 m	60–65/70–80 m	60–80 m
Terrace VII	80–120 m	130–140 m	70–80/90–100 m	80–120 m
Terrace VIII	130–160	160–180 m	100–120/110–120 m	120–130 m
Terrace IX	–	–	–	–
Kachinsk surface of planation	170–190 m	200–240 m	130–140/150 m	–
Sukhobuzim surface of planation	230–240 m	–	–	–
Studied area in the Yenisey Valley	Krasnoyarsk Basin margin	Minusinsk Basin (Rakovets 1969 came to similar conclusions)	From Krasnoyarsk to the tributary of the Angara	Minusinsk Basin in the foreland of the reservoir

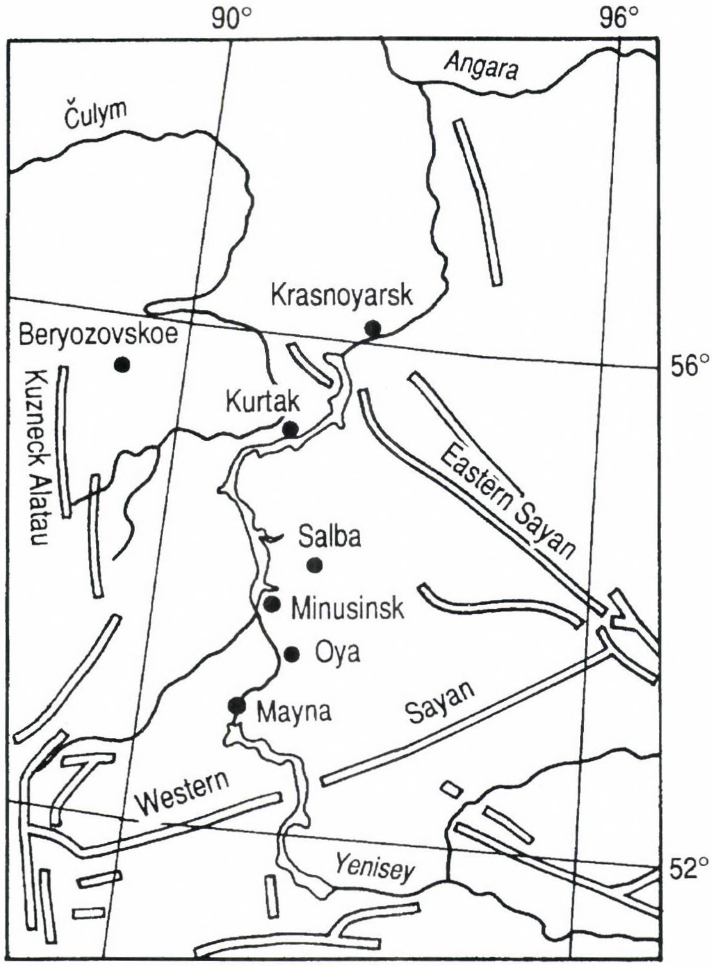


Fig. 2. Geographical setting of the Yenisey Valley

geomorphic position and phenomena typical of young terrace sediments and of the superimposing loess and slope deposits when analysing representative profiles along the river section between Krasnoyarsk and Minusinsk (*Fig. 2*).

Along with young terrace exposures young loess series (Sisim) overlying some older terraces (N°V–VI) and remnants of older degraded loess and young sand and loesses superimposing the eroded loess pillars were observed, too.

In the earlier publications there are references that in the Middle Yenisey Valley not only Quaternary terrace formations but ancient weathered rocks and Tertiary deposits could be found locally and their traces were encountered during joint field excursions. This means that some valley sections could develop well before the Quaternary, then they were partially buried and later exhumed.

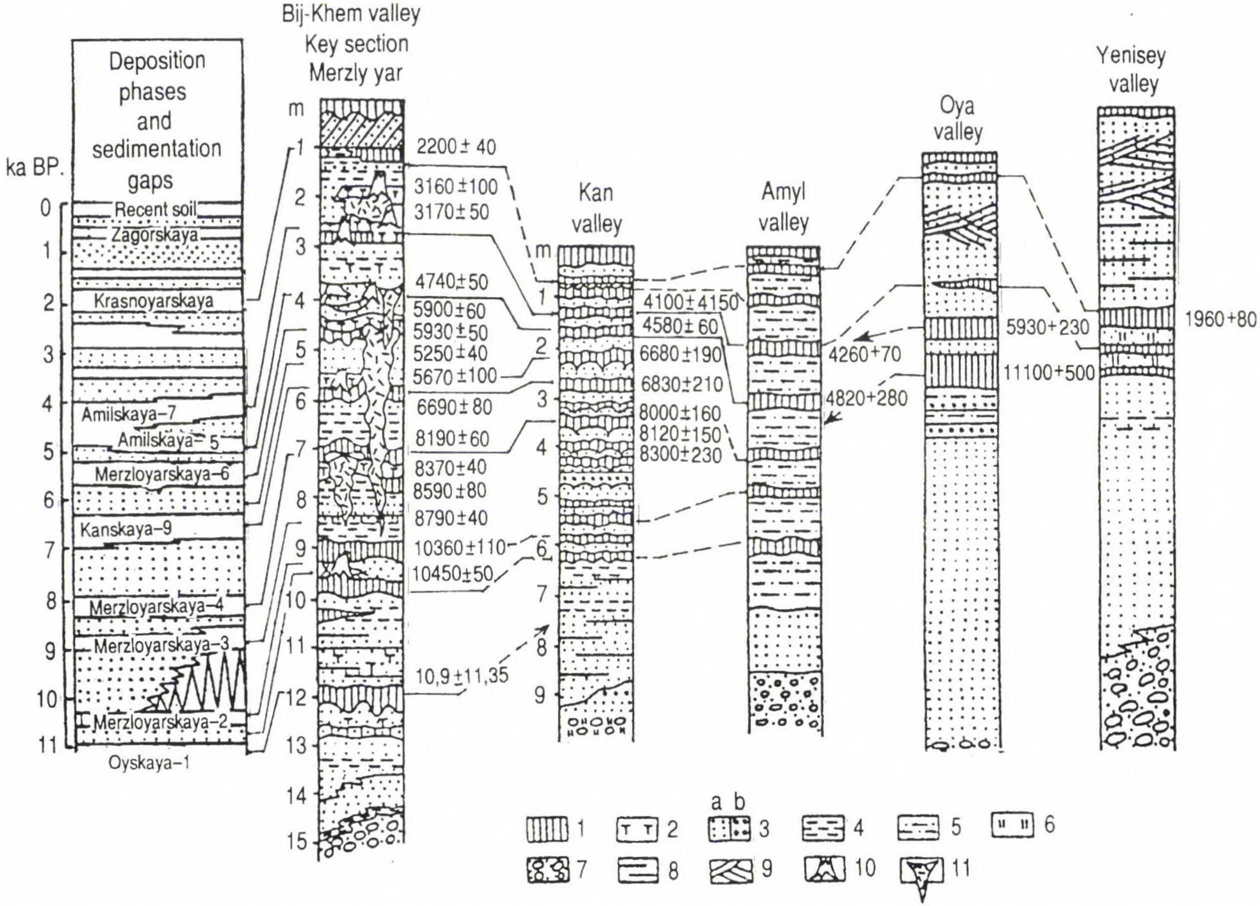


Fig. 3. Correlation of polycyclic Holocene sediments in the southern part of the Yenisey Valley. 1 = soil; 2 = peat; 3 = sand (a), gravelly sand (b); 4 = silt; 5 = loam; 6 = dusty sediment; 7 = gravel bed; 8 = horizontal layers; 9 = oblique and diagonal layers; 10 = remnants of buried forest; 11 = ice wedges

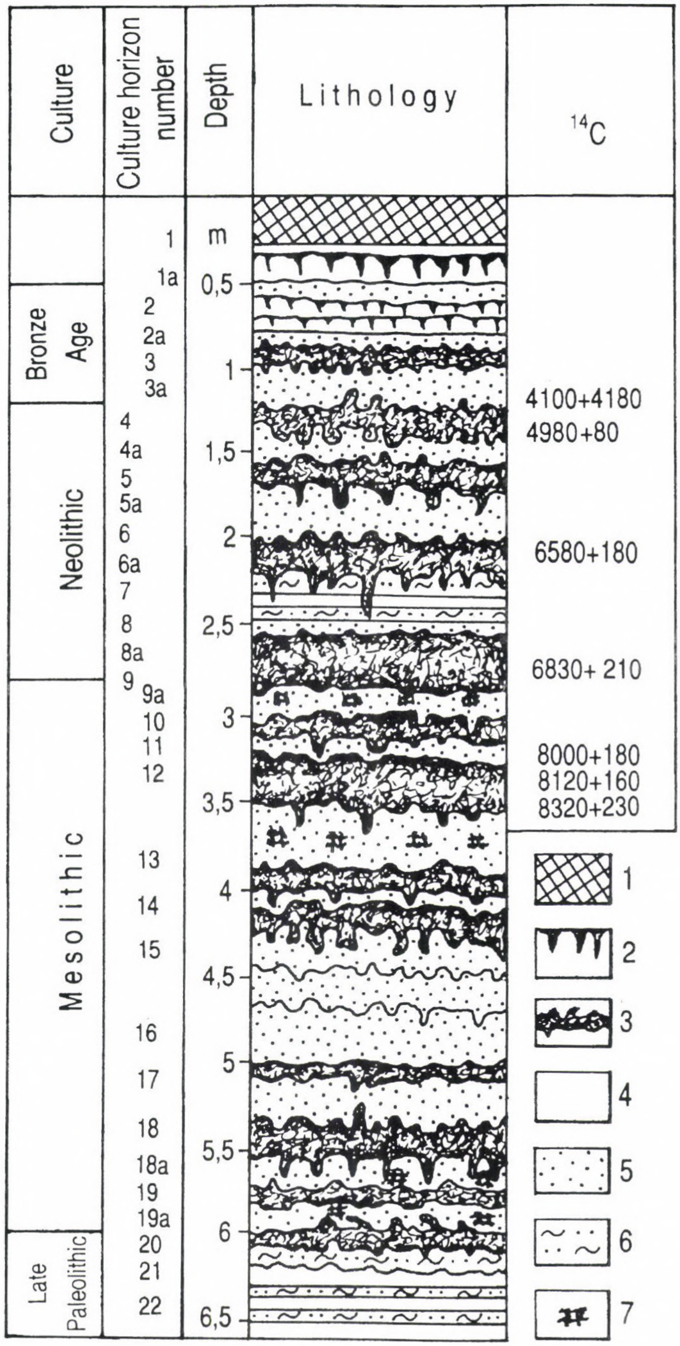
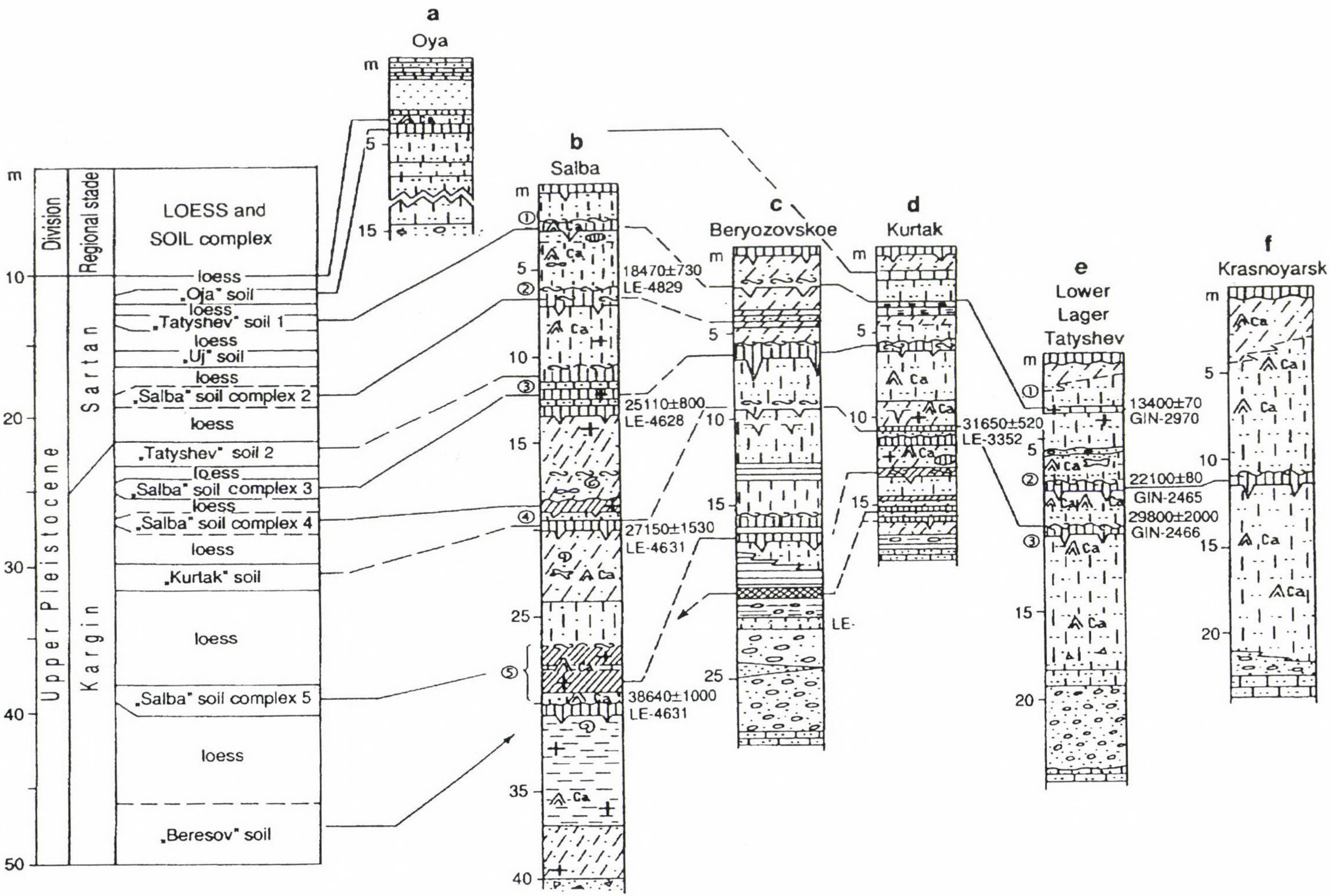


Fig. 4. Profile of polycyclic Holocene terrace in the area of multi-layered site Kazachka (on Kan River). – 1 = artificial filling; 2 = fragments of young alluvial soils; 3 = alluvial paleosols; 4 = sorted sand; 5 = non-sorted sand; 6 = loams; 7 = humous sand


a
Oya
b
Salba
c
Beryozovskoe
d
Kurtak
e
Lower
Lager
Tatyshev
f
Krasnoyarsk
Division
Regional stade
Upper Pleistocene
Sartan
Kargin
LOESS and
SOIL complex
loess
„Oja" soil
loess
„Tatyshev" soil 1
loess
„Uj" soil
loess
„Salba" soil complex 2
loess
„Tatyshev" soil 2
loess
„Salba" soil complex 3
loess
„Salba" soil complex 4
loess
„Kurtak" soil
loess
„Salba" soil complex 5
loess
„Beresov" soil
18470±730
LE-4829
25110±800
LE-4628
27150±1530
LE-4631
38640±1000
LE-4631
LE-
31650±520
LE-3352
13400±70
GIN-2970
22100±80
GIN-2465
29800±2000
GIN-2466
Ca

The Krasnoyarsk team has been involved in terrace investigations, based on complex stratigraphic analysis of profiles of the high flood plain and those of the superimposing low terraces overlying them and also on ^{14}C datings of humus and charcoal. As a result conclusions have been drawn concerning the age of flood plains and terraces (*Figs 3* through *5*).

Shaped by an extreme (seasonal and periodical) hydrological regime normal alluvial sequences were sedimented, while in some cases, e.g. during disastrous floods, lacustric-fluvial series were formed within the dammed valley sections. These sediments are much thicker and are composed by specifically layered alluvia and basin sediments. In this way periglacial and intracontinental water regimes in Southern Siberia resulted in polygenetic and polycyclic terraces. According to Yamskikh (1983) cyclic climatic change led to sedimentation phases of 21–22 ka and 7–8 ka duration in the late Pleistocene, while shorter spells (400–500 years) were characteristic for the Holocene. These fluctuations are thought to be supported by radiocarbon and palynological evidence.

For the verification of sedimentological investigations of loess-paleosol sequences and fluvial sediments in the vicinity of Krasnoyarsk summarised by Yamskikh (1992, 1993) a joint study of the profile of the so called lower Lager (Tatyshev) terrace with the overlying loess-paleosol sequence (located in Krasnoyarsk city, at October Bridge) was carried out (*Fig. 6*, Tatyshev terrace surveyed by Pécsi, Schweitzer and Yamskikh). Previously ^{14}C datings based on humus content of three soil horizons (h_1, S_1 and S_3) were performed. At a depth of 3 m the age of h_1 proved to be 13,400±70 yr, at 9 m S_1 was 22,100±80 yr and at 15 m S_3 gave an age of 29,800±2000 yr B.P.

←

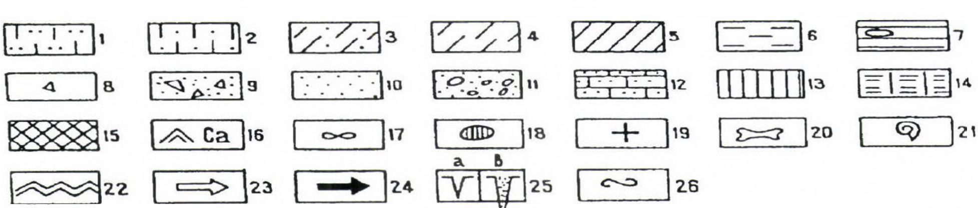


Fig. 5. Correlation of loess-paleosol profiles in the middle Yenisey Valley (after Yamskikh 1992). – 1 = loessy sand ; 2 = sandy loess; 3 = sandy slope loess; 4 = slope loess; 5 = semipedolite; 6 = gleyed silt; 7 = clay with boulders; 8 = debris; 9 = broken stone with sand; 1 = alluvial sand; 11 = alluvial boulder beds with sand; 12 = sandstone; 13 = steppe-type chernozem soil; 14 = alluvial meadow soil; 15 = grey forest soil; 16 = $CaCO_3$ accumulation ; 17 = loess doll; 18 = krotovina; 19 = charcoal; 20 = macrofauna; 21 = shells of molluscs; 22 = unconformity in the profile; 23 = traces of non-linear erosion; 24 = traces of linear erosion; 25 = pseudomorphs along ice veins; 26 = deformation due to solifluction; SL = Salba soil complex; KT = Kurtak soil complex; grain size distribution (mm): A = clay (< 0,001); I = silt (0,001–0,01); L = loess (0,01–0,05); H = sand (0,05–1,0)

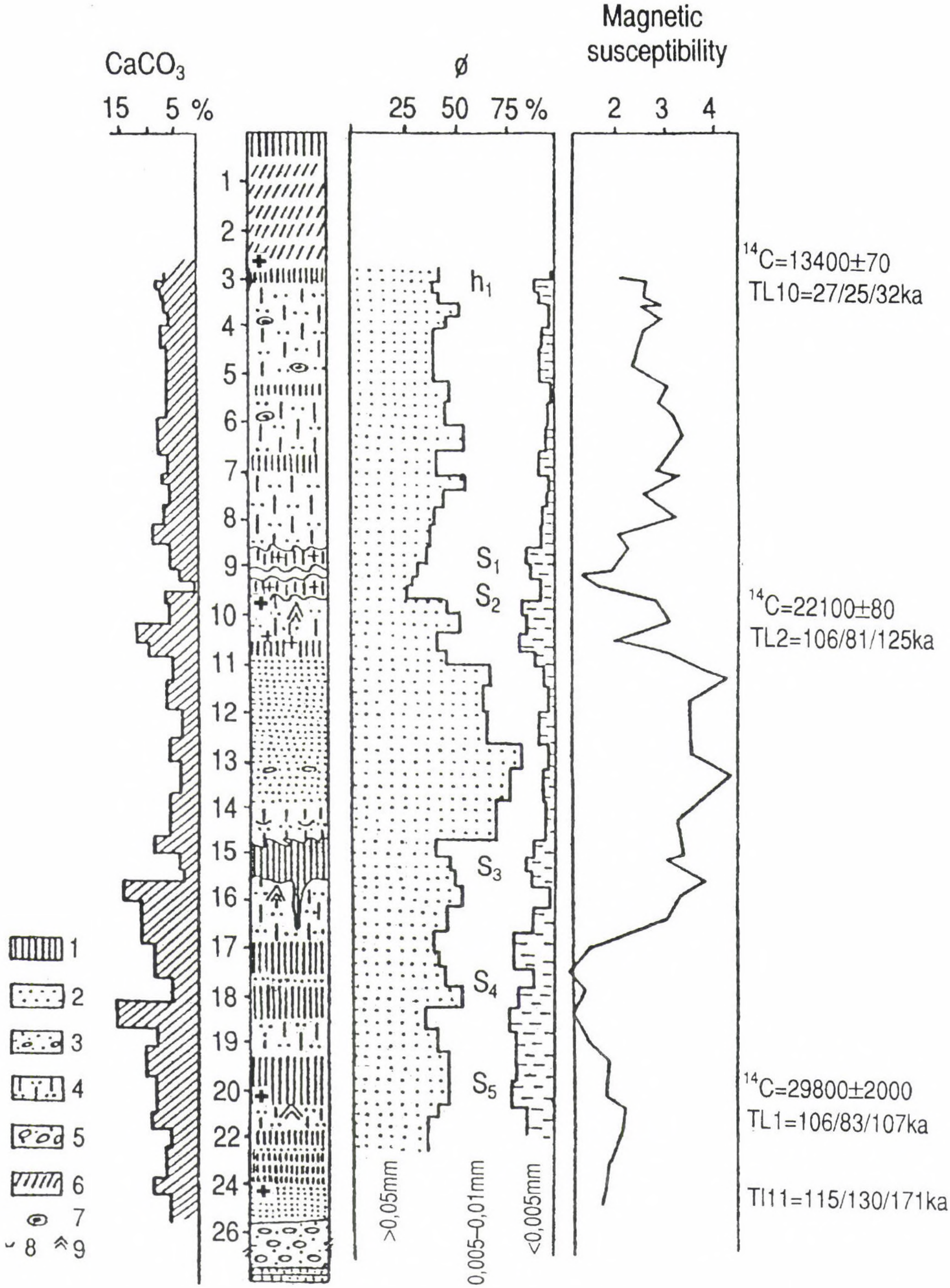


Fig. 6. Profile of the Lower Lager (Tatyshev) terrace of the Yenisey at Krasnoyarsk (according to Yamskikh 1993). – 1 = soil; 2 = sand; 3 = gravelly sand; 4 = loam; 5 = gravel; 6 = slope loess; 7 = iron concretions; 8 = ferruginisation; 9 = Ca accumulation

Young loess-paleosol-sand sequence of the lower Lager (Tatyshev) terrace at Krasnoyarsk

The Paleozoic rock basement of the lower Lager (Tatyshev) terrace is superimposed by a 24 m sequence of loess, sand and paleosols.

The basement of the rock is situated 9 m higher than the middle level of the Yenisey.

In the studied exposure it is overlain by a 1.5 m thick terrace gravel and 1–1.5 m fluvial sand with thin interbeddings of sandy clay layers. From the lower part of this layer samples were taken for TL analyses (*Fig. 6, Table 2*, Sib11).

– 22–21 m sandy silt with two embryonic, hydromorphous, slightly calcareous flood plain soils of light brown colour and of 15–30 cm thickness each;

– 21–20.30 m crumby, calcareous, sandy loess, partly with a well developed C_{Ca} level of carbonate accumulation;

– 20.30–19.30 m meadow chernozem (S_5) of high sand content; sample Sib1 was taken from here;

– 19.30–18.60 m C_{Ca} horizon of carbonate accumulation with carbonate concretions and loessy sand;

– 18.60–16.90 m dark chernozem soil (S_4) of ochre-brown colour, strongly and medium calcareous;

– 16.90–16.00 m sandy loess, strongly calcareous;

– 16.00–15.50 m horizon of carbonate accumulation(C_{Ca});

– 15.50–14.40 m chernozem soil (S_3) of high sand content and dark ochre-brown colour;

– 14.40–10.80 m loessy sand, sand;

– 10.80– 9.80 m calcareous sand, with a level of carbonate accumulation (C_{Ca}) between 10.20–10.50 m (Sib2 sample from 9.80);

– 9.80– 8.40 m double meadow soil (S_1+S_2) of black colour (with a level of carbonate accumulation (C_{Ca}) between 9.10–9.30 m);

– 8.40– 7.00 m sandy, with two intercalated tephras;

– 7.00– 6.80 m humous sandy loess (h_2);

– 6.80– 3.00 m sandy loess, loessy sand, between 5.20–5.35 m slightly humous sandy loess (h_{2sl});

– 3.00– 2.00 m two distinct layers of humous loessy sand (h_1) with a slightly calcareous accumulation, sample Sib10 was taken from here;

– 2.00– 0.00 m artificial infilling and soil developed on slope deposit covered by recent chernozem soil.

Radiocarbon data on *Fig. 5e* and *Fig. 6* (Yamskikh 1993) were based on ^{14}C datings of humus and an analogy (age of Ust'-Batoi horizon being 13.4 ka).

Experience has repeatedly shown that radiocarbon age of humus within loess-paleosol sequences is much less than that obtained by charcoal or mollusc shell analyses (see Pécsi and Hahn 1987: loess-paleosol sequence at Basaharc, *Fig. 7*).

^{14}C analyses of charcoals usually do not exceed 30–35 ka and this is to be considered as minimum age. Absolute geological age might be produced twice or several times more.

Samples collected for TL and IRSL investigations from four representative horizons (SIB11, SIB1, SIB2, SIB10, see *Fig 6* and *Table 2*) were analysed by M. Frechen, Laboratory of Quaternary, University of Cologne.

Table 2. IRSL and TL dating of loess-paleosol sequences along the middle Yenisey Valley; age in ka (analysed by M. Frechen, Laboratory of Quaternary, University of Cologne)

Sample	IRSL/REGEN	IRSL/ADD	TL/REGEN	TL/ADD
SIB 92-1	87.9±8.7	106.9±14.5	88.7±12.2	107.2±11.5
SIB 92-2	76.1±7.7	106.4±19.9	81.7±8.7	125.6±13.0
SIB 92-3	62.2±7.3	99.0±25.6	71.2±9.2	102.2±10.1
SIB 92-4	20.9±2.4	23.1±2.7	23.3±2.9	20.9± 4.6
SIB 92-5	79.8±9.5	121.9±16.9	15.7±9.8	112.6±11.6
SIB 92-6	19.0±2.3	21.9±4.4	28.4±5.2	30.3± 5.5
SIB 92-7	76.4±8.7	104.5±15.0	86.2±10.3	123.8±12.0
SIB 92-8	55.5±5.6	65.4±8.4	60.4±7.2	62.2± 6.1
SIB 92-9	75.5±8.6	96.7±11.5	90.3±10.1	110.5±13.7
SIB 92-10	22.5±2.4	27.0±5.4	25.2 ±3.1	32.2± 4.2
SIB 92-11	96.4±11.6	115.6±19.3	130.7±16.0	171.1±16.7

Places of sampling:
1 – Tatyshev terrace, Krasnoyarsk
2 – Tatyshev terrace, Krasnoyarsk
3 – Cholnokov profile in the vicinity of Krasnoyarsk
4 – Sisim loess-paleosol profile along the middle reach of Yenisey
5 – Sisim loess-paleosol profile along the middle reach of Yenisey
6 – Primorskoe 1 loess-paleosol profile along the middle reach of Yenisey
7 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
8 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
9 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
10 – Tatyshev terrace, Krasnoyarsk
11 – Tatyshev terrace, Krasnoyarsk

On the basis of the experience with studies of other fluvial terraces and superimposing loess-paleosol-sand sequences in similar geological and geomorphological position it is assumed that results of TL analyses by M. Frechen show a more reliable picture of geologic evolution as compared with those obtained through ^{14}C datings.

SIB1 and SIB11 TL datings by Frechen might be correlated with the period of development of chernozem soils within the Tatyshev profile (S_3–S_5). These soils were correlated by Yamskikh with BD_1 and BD_2 paleosols of the Hungarian loess-paleosol sequence (Yamskikh 1993, *Fig. 8*). SIB1 and SIB11 data by Frechen obtained using four different methods of calculation essentially embrace this time period of the Upper Pleistocene (*Table 2*).

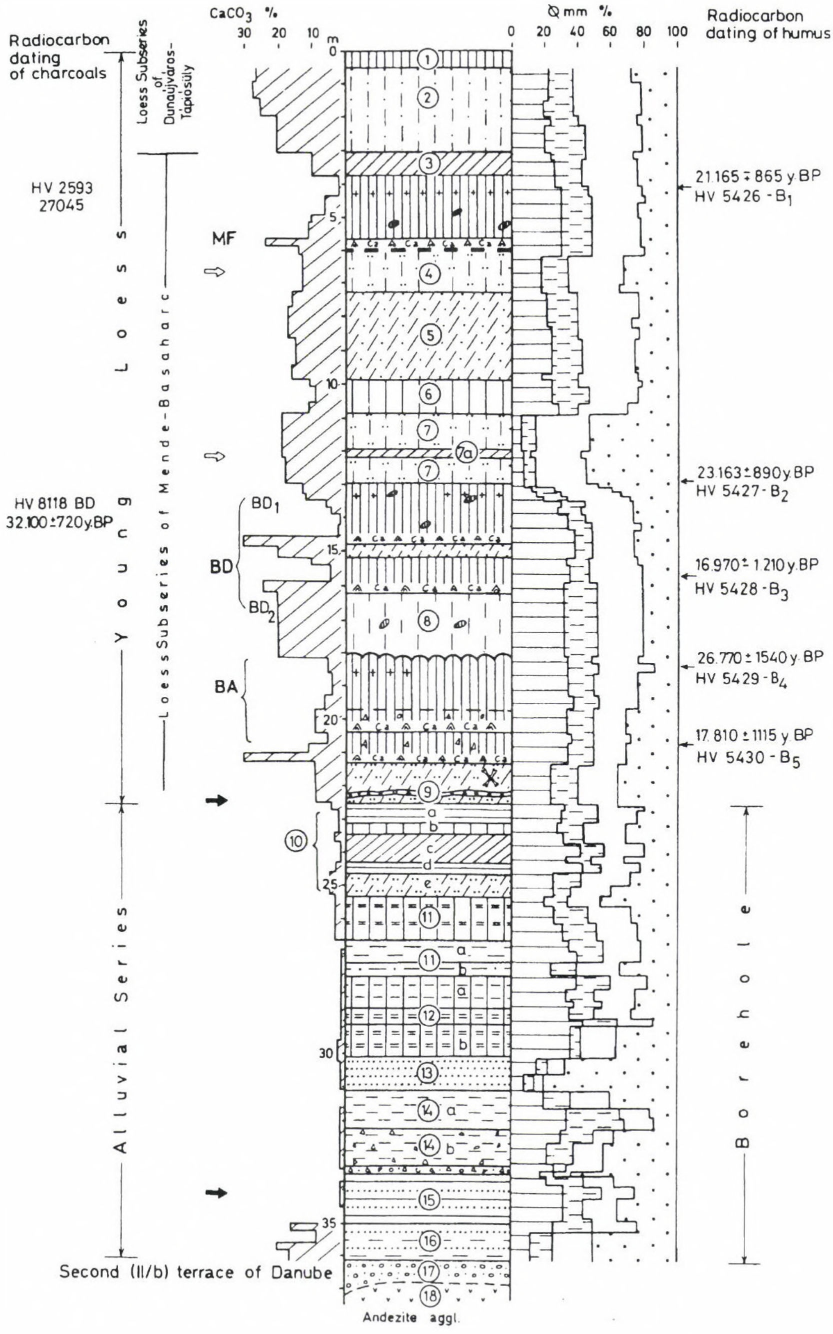


Fig. 7. Loess profile at Basaharc brickyard: stratotypes of Basaharc Double (BD1, BD2) and Basaharc Lower (BA) paleosols. – + + + = charcoal fragments; ⟶ = derasional (slope wash) unconformity; ⇒ = minor erosional unconformity

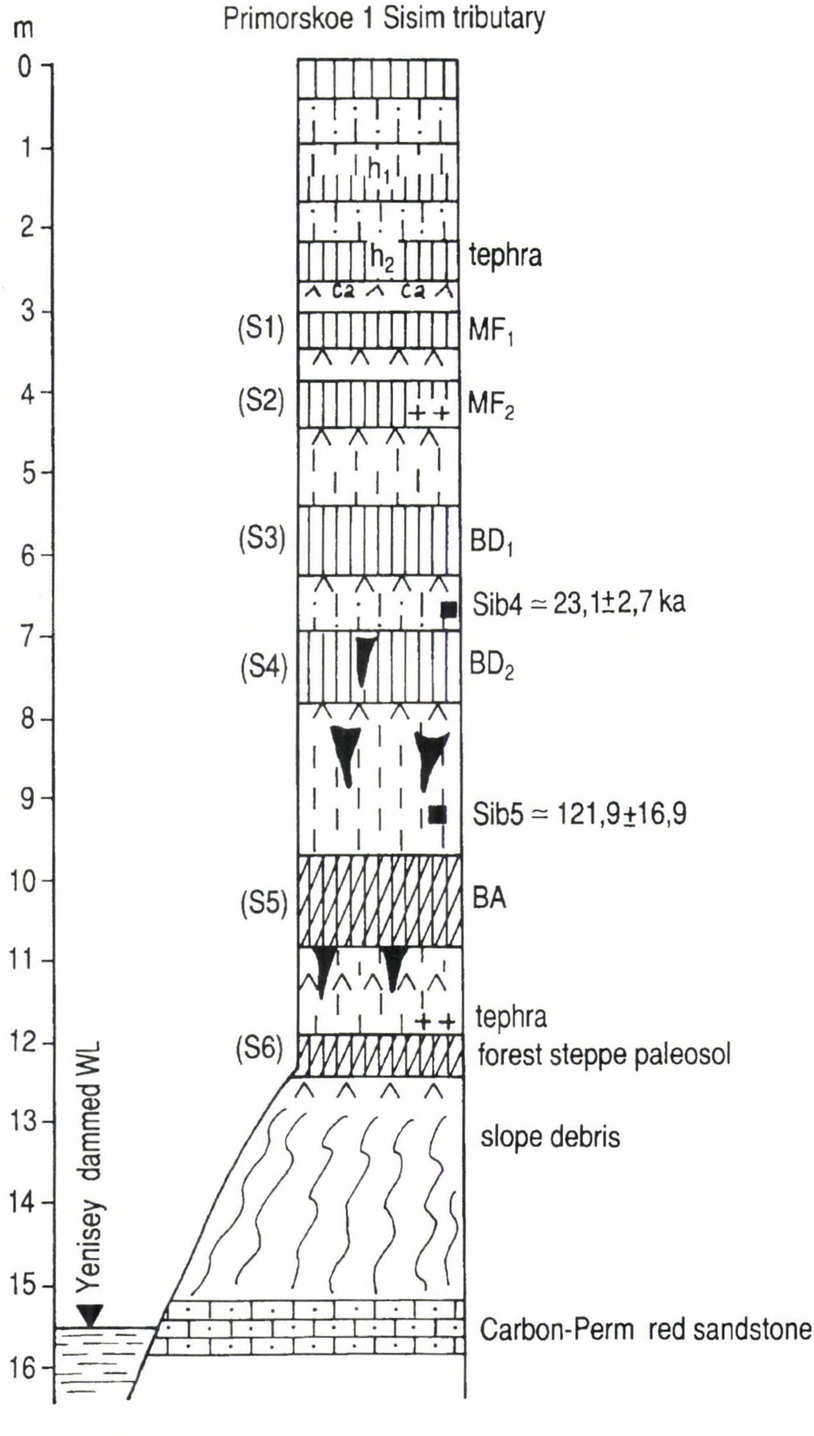
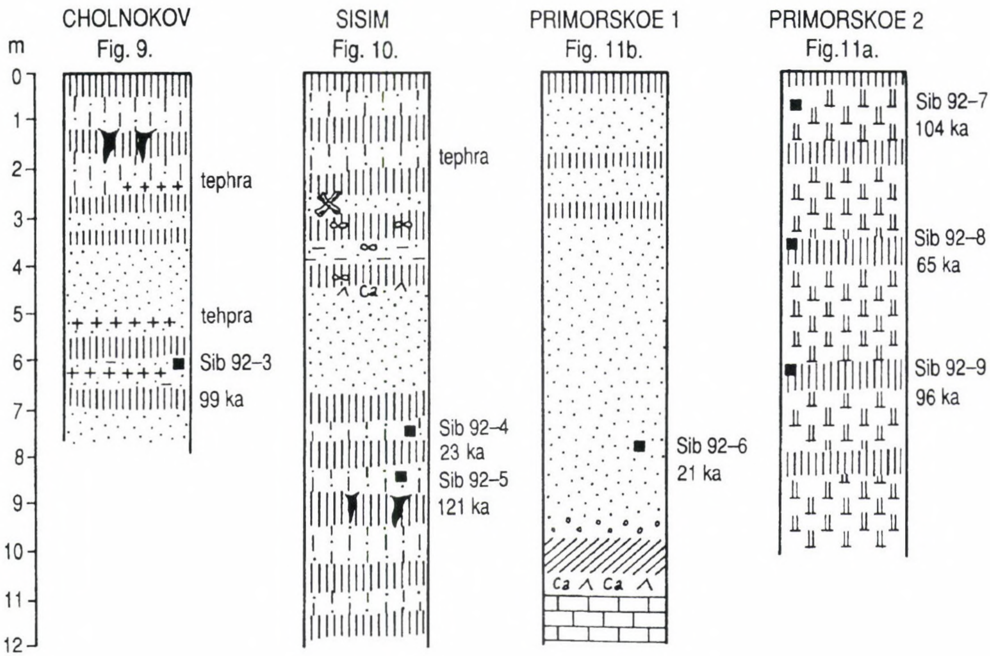


Fig. 8. Primorskoe 1 profile (Sisim – tributary of the Yenisey)



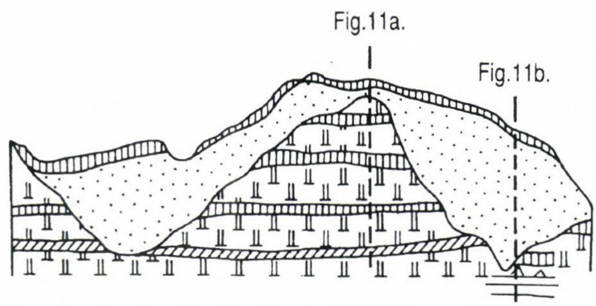


Fig. 9. Cholnokov loess-paleosol profile
Fig. 10. Sisim loess-paleosol profile
Fig. 11a and *b*. Primorskoe loess and sand bluff along the Yenisey at the Krasnoyarsk reservoir. Lithostratigraphy by Pécsi, M., Schweitzer, F. and Yamskikh, A.F., TL/IRSL dating by Frechen, M.

Young loess-paleosol series on some old terraces of the Yenisey

A young loess-paleosol sequence overlying higher terraces of the Yenisey has proven to be very similar to the one on the Tatyshev terrace (see *Fig. 9*, Cholnokov loess exposure on terrace N°V). *Fig. 10* (Sisim) also comprises a young loess series being a retreating loess bluff above the dammed water level of the Krasnoyarsk reservoir in the vicinity of Primorskoe village. The sections mentioned are presumably part of the young loesses superimposing terrace N°V (in the vicinity of the paleontological and archeological site Primorskoe 1).

In *Figs 9* and *10* white tephra interbeddings are confined to five soil horizons. Sample SIB 92-3 was taken from a chocolate brown double chernozem; its age calculated by Frechen is TL/102 ka (IRSL/ADD calculation: 99.0±25.6 ka). According to its paleopedological type and stratigraphic position this double paleosol resembles BD_1–BD_2 soil complex and is intercalated by a tephra layer of 10–15 cm thickness.[7]

Sample SIB 92-5 *(Fig. 10)* was dated 121.9±16.9 ka (IRSL/ADD). Stratigraphic position of this sample suggests a sandy loess with some carbonate accumulation between soils BD_2 and BA. Dating of sample SIB5 provides a good correlation but sample SIB 92-4 taken from between paleosols of BD_1 and BD_2 type and position shows an age of 23.1±2.7 ka. Perhaps this strong discrepancy is caused by carbonate accumulation or some other unknown effect.

In case of section Primorskoe 2 (*Figs 11a* and *11b*) SIB-6=21.9±4.4 ka; SIB-7=104.5±15.0 ka; SIB-8=65.4±8.4 ka; SIB-9=96.7±11 ka. Reliability or weakness of results might depend on the fact that in profile 11a samples originate from old loesses while in profile 11b between the remnants of the old and eroded loess there is an interbedding of (polycylic) fluvial sand and in the uppermost part thin layers of slightly humous sand occur. Similar lithostratigraphic position is often very difficult to recognise because pillars of the old loess remnants in several places are covered by young but rapidly deposited sand. This young sand layer locally underwent weak loessification.

Conclusions

Terrace and loess-paleosol profiles presented in Discussion, and their lithostratigraphic evaluation, using the traditional methods of terrace chronology and loess-paleosol stratigraphy allow an approximate and so called relative chronostratigraphic subdivision involving bio- or lithostratigraphic schemes, time scales of climate fluctuations and calculating with the rate of landform evolution and deposition. Absolute chronological methods are to complete this procedure. In our case two kinds of radiometric methods were available to assess the absolute age of certain key profiles.

Ages obtained through radiocarbon analyses considerably differ from those using TL or ISRL investigations. Radiocarbon datings of humus extraction provide

[7] Denominations BD_1 and BD_2 (Basaharc Double) were used because position and genetic type of the Siberian paleosols show strong analogies with the third and fourth (double) steppe soil complex within a sequence overlying the flood-free terrace N°II of the Danube in Hungary (Pécsi and Hahn 1987, Pécsi 1996)

young age for old soils. On the other hand ^{14}C ages represent a minimum age terminating at 25–30 ka. This method of dating in unable to determine older ages.

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Variations of eolian iron over the last 900 ka as recorded in the Xifeng loess section in China and the implications

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Abstract

Iron transported by eolian dust is thought to have significant effect on the ocean phytoplankton productivity, and thus may have played an important role in modulating global climates through influencing CO_2 uptake of the ocean. Airborne dust forming the loess-paleosol sequence in China mainly originated from the deserts in northern and northwestern China and was transported by the northwesterly winds. In this study, total Fe_2O_3 content of 852 samples collected at 10 cm intervals from the Xifeng loess section on the central Loess Plateau were analysed. The results show a significant variation of iron content over time. The values in some glacial intervals are comparable to those during interglacial periods with the exception that their fluctuation as a rule does not correspond to the long-term glacial–interglacial oscillations. Without detectable fluctuations of ~100 ka period, it rather exhibits a considerably higher frequency with a marked ~23 ka period and a clearly discernible 40 ka period. The highest iron contents were observed for the soil units S5-1 and S5-3 (marine δ^{18} stage 13 and 15). In most of the marine records, these intervals are characterised by the highest $\delta^{13}C$ values during the last 900 ka, suggesting their correspondence to climate events on global scale. The ~23 ka and 41 ka periods suggest that changes in summer insolation over the northern hemisphere have a strong effect on the airborne iron content whereas the mechanisms need further study.

Introduction

Over the last 2.5 Ma, a high amount of eolian dust has been deposited along the middle reaches of the Yellow River in northern China (*Fig. 1*) within an area referred to as Loess Plateau (Liu 1985). Complete Quaternary loess-paleosol sequences are mostly more than 150 m in thickness, and they are generally considered to represent a nearly continuous paleoclimate record, containing more than fifty intervals of soil formation interrupted by intervals of dust deposition (Guo *et al.* 1996). Major interglacial soils and glacial loess units of the last 900 ka are labeled S0, L1,

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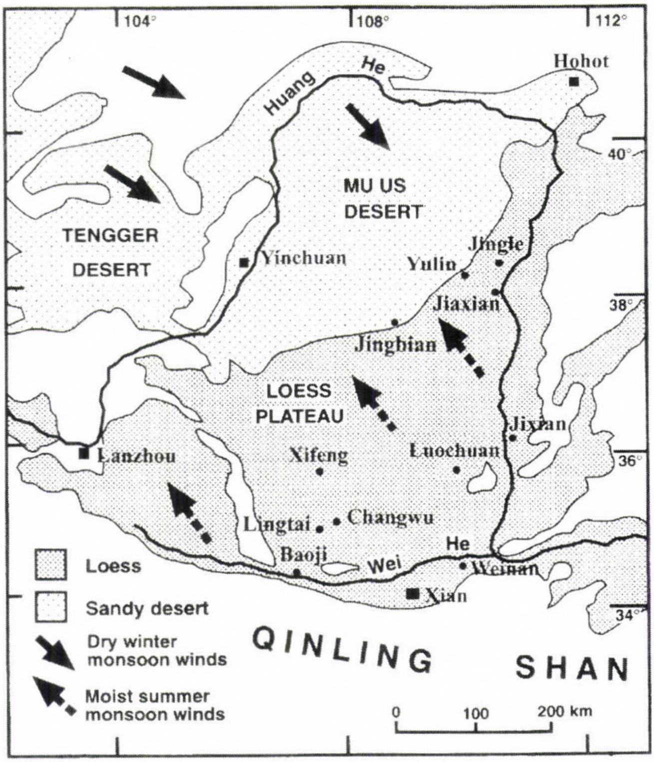

Fig. 1. Map showing the Chinese Loess Plateau, location of the Xifeng site, and the East-Asian summer and winter monsoon circulations (modified after An and Porter 1997)

S1, L2, S2... – S8, L9 from the top to the bottom (*Fig. 2*) according to the Luochuan stratotype section (Liu 1985). Eolian dust deposition and pedogenesis are in fact competing processes all the time, and the presence of a paleosol simply indicates that the latter process was predominant (An *et al.* 1991, Guo *et al.* 2000). The stratigraphy of the loess-paleosol sequence can be correlated well with that of the marine $\delta^{18}O$ record (Liu 1985). The correlation pattern for the last 900 ka has now been accepted (Liu 1985; Kukla 1987), and that for the last 500 ka was also confirmed by an eolian record in the North Pacific (Hovan *et al.* 1989), which provided a direct link between the China loess and marine $\delta^{18}O$ stratigraphy.

Iron transported by eolian dust is thought to have had a significant effect on the ocean phytoplankton productivity, and thus may have played an important role in modulating global climates through influencing the CO_2 uptake of the ocean (Martin 1985). Eolian dust forming the loess-paleosol sequence in China is mainly originated from the deserts in northern and northwestern China through the transport by the northwesterly winter monsoon winds (Liu 1985; An *et al.* 1991). Meanwhile, a significant portion of dust from the same source areas was transported by the wester-

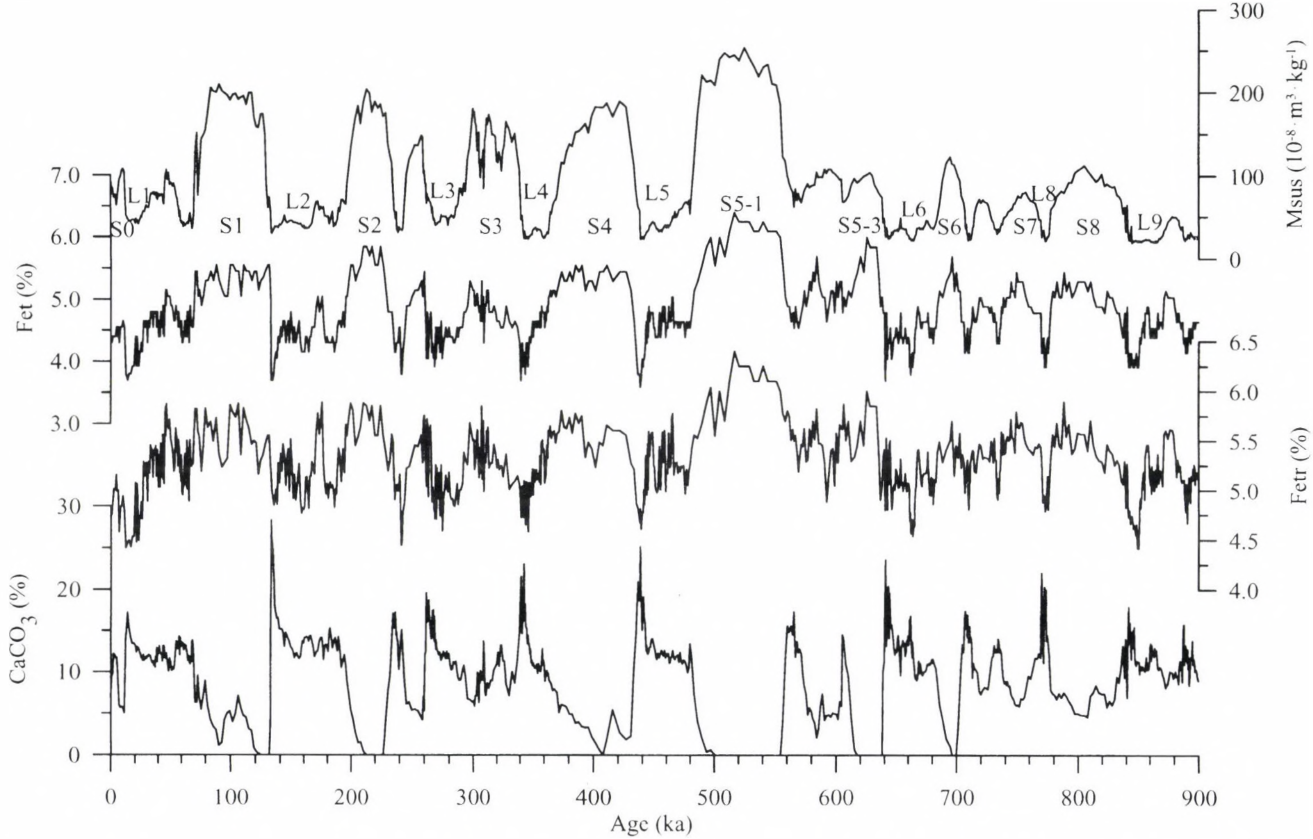

Fig. 2. Variations of magnetic susceptibility (Msus), total iron content (Fet), total iron content rectified using $CaCO_3$ content (Fetr), and $CaCO_3$ content in the Xifeng loess section. The time-scales were obtained by correlation with the susceptibility time-series from Kukla *et al.* (1990). The major soil and loess units are labeled

ly winds and deposited in the North Pacific region (Hovan *et al.* 1989). In this study, total iron content of 852 samples from the Xifeng loess section located on the central Loess Plateau was analysed to investigate how iron content of eolian origin varied over the last 900 ka, and what are likely the causes of these changes.

Materials and methods

The studied Xifeng site is located within the central Loess Plateau (*Fig. 1*) thus it can be representative for characterising the general pattern of Quaternary loess deposition in northern China. Its stratigraphy correlates well with the other stratotype loess sections (Kukla *et al.* 1990), such as with the Luochuan one (Liu 1985). Magnetic susceptibility has proven particularly useful for drawing the stratigraphic boundaries of the loess-paleosol sequences in China (Liu 1985; Kukla 1987) and the values are higher in soils than in loess. Susceptibility was measured on dry samples taken at 10 cm intervals using a Bartington susceptibility meter and expressed in SI units. The time-scale is obtained through correlation with the susceptibility curve by Kukla *et al.* (1990) from the same site. The latter is dated by a magnetic susceptibility model (Kukla *et al.* 1990) using the magnetic reversal boundaries as age control. Although this model is based on some assumptions still contentious in part (Heller *et al.* 1993; Verosub *et al.* 1993; Han *et al.* 1996), the yielded results generally are consistent with those obtained by land-sea correlation (Kukla 1987; Hovan *et al.* 1989), and thereby remains a working model for obtaining an independent time-scale.

Total iron content (Fet) was analysed on 852 samples collected at 10 cm intervals and expressed as Fe_2O_3 weight percentage. The samples were treated by acid dissolving methods and the iron content was measured using a WFD-Y2 atomic adsorption device with an analytical precision of 0.4%. The content of $CaCO_3$, more or less translocated due to the pedogenesis, was gasvolumetrically measured on the same samples for the rectification of its dilution effect on iron concentration. The rectified values therefore represent those of the non-calcareous material. Microscopic thin sections of undisturbed soil and loess samples were prepared to evaluate the possible extent of iron translocation during soil formation. Spectral analyses of the magnetic susceptibility and (rectified) Fetr time-series were performed using the Maximum Entropy method.

Results and discussions

Variations of the total Fe_2O_3 (Fet) are shown in *Fig. 2*, compared with the magnetic susceptibility curve as a stratigraphical marker. As the formation of paleosols in the studied section is characterised by significant dissolution and repeated precipitation of $CaCO_3$, which affects the relative concentration of iron oxides-hydroxides, Fet values were rectified with the $CaCO_3$ content (Fetr) as it is shown in *Fig. 2*.

Examination on microscopic thin sections of the soils revealed a mere 3–4% of clay coatings (surface estimation) in the paleosols S1, S4, S5-1 and S5-3, i.e. the best developed soils at the studied site, and clay coating is missing in the other paleosols. Significant iron translocation related to processes of clay illuviation can therefore be ruled out basically. The absence of Fe-Mn hydromorphic features also excludes the possibility of significant iron redistribution in soluble state and exchange with the overlying and underlying loess. Thus Fetr values in the studied section mainly reflect the iron content of eolian dust.

Magnetic susceptibility of the section clearly shows the alternation of loess horizons and soil units (*Fig. 2*). $CaCO_3$ content is much lower in soil units than in loess, indicating strong decalcification during soil formation. The non-rectified Fet values are generally higher in soils than in loess, showing a higher agreement with the magnetic susceptibility time-series than for the rectified values (Fetr). This is attributable to the dilution effect of $CaCO_3$, being higher in loess than in soil units.

Variations of Fetr shows that the composition of eolian dust did changed over time. Its fluctuation generally does not correspond to the changes of the magnetic susceptibility; rather they exhibit oscillations of much higher frequency, the values of which in some parts of the loess are similar to those is soil units. Highest values are observed for S5-1 and S5-3 soil units, which correspond to marine $d^{18}O$ stages 13 and 15, respectively (Liu 1985; Kukla 1987). These soils represent periods of warm extremes during the last 2.6 Ma (Guo *et al.* 1998). In marine records these intervals are characterised by higher $d^{13}C$ values (Raymo *et al.* 1997) suggesting their global significance.

To investigate the temporal aspects of eolian iron fluctuations more precisely, spectral analyses were performed on both magnetic susceptibility and Fetr, using Maximum Entropy method (Burg 1967, *Fig. 3*). Magnetic susceptibility is usually used as a proxy of the East-Asian summer monsoon (An *et al.* 1991), and therefore reflects climate changes in the Loess Plateau region. Magnetic susceptibility correlates well with the marine $d^{18}O$ record, an indicator of the global ice-volume (Shackleton and Hall 1984), suggesting that the fluctuations of the monsoon climate match well the glacial-interglacial cycles.

The spectrum for magnetic susceptibility is characterized by four discrete peaks, centered at 111 ka, 42 ka and 26 ka and 21 ka periods, with the spectral density well over the 95% confidence limit. Within the accuracy of the used time-scale, which is independent of any land–sea correlation or orbital tuning, the 111 ka and 42 ka periods are correlative with those of the orbital eccentricity (Berger and Loutre 1989) and obliquity, respectively, having appeared as the ~100 ka and 40 ka periods in geological records. The ~26 and 21-ka periods recorded by the Fetr time-series are evidently related with the ~23 and 19 ka periods of the orbital precession (Berger and Loutre 1991).

The spectrum calculated for Fetr is dominated by a strong peak at 23 ka corresponding to a precessional cycle. Its spectral density is well over the 95% confi-

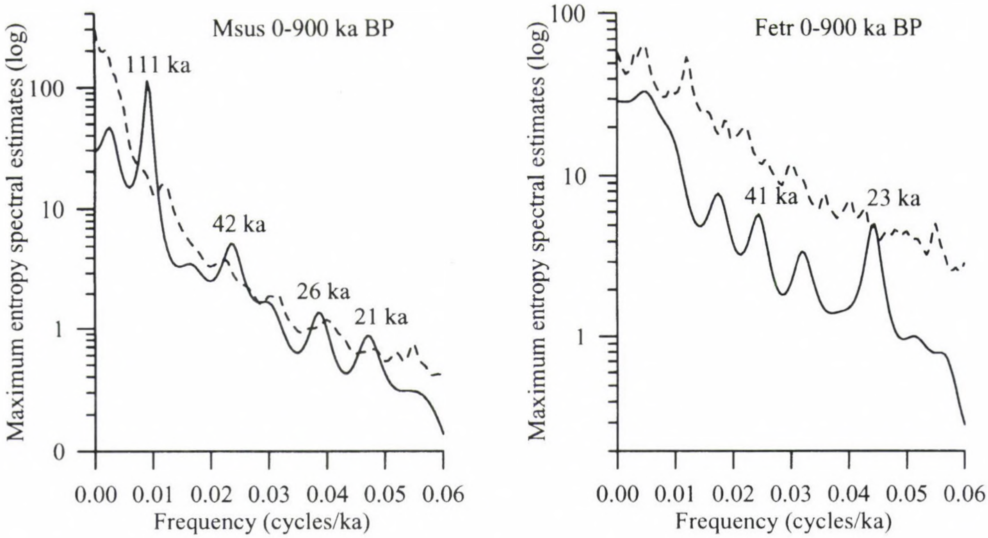

Fig. 3. Maximum Entropy spectral analyses of the Xifeng magnetic susceptibility (Msus) and eolian iron (Fetr) time-series. Continuous line (logarithmic scale) represents the spectral density and dotted line (logarithmic scale) represents the 95% confidence limit

dence limit (*Fig. 3*). The 40 ka period is clear, but the spectral density has not exceeded the 95% confidence limit. A remarkable feature is that the eccentricity period centered at 100 ka (Berger and Loutre 1991) is undetectable in the obtained spectrum.

The presence of the 111 ka, 42 ka, 26 ka and 21 ka periods in the spectrum of magnetic susceptibility indicates strong orbital control on loess deposition and soil formation in northern China over the last 900 ka, as it was reported in earlier studies (Ding *et al.* 1995). Because of the negligible effect of the eccentricity in modulating the solar insolation budget (Berger and Loutre 1991), the ~100 ka period in geological records is usually interpreted as the signal of global ice-volume variations (Imbrie *et al.* 1984). The marked 111 ka period of the magnetic time-series indicates a strong impact of glacial–interglacial cycles on Asian climate.

The most interesting feature of this study is the lack of ~100 ka period in the Fetr time-series, indicating a relative dynamic independence between the eolian iron content and the global ice-volume variations reflected by marine $\delta^{18}O$ records (Shackleton and Hall 1984). Rather the marked 23 ka period and a clear 40 ka period recorded in the Fetr time-series would suggest a strong control of solar insolation, and the well expressed 23 ka period trends to suggest an insolation-forced factor of low latitude origin.

The issue of possible mechanisms through which solar insolation affected eolian iron content in the loess-paleosol sequence in China so strongly is most puzzling. Two possibilities might be invoked to explain this phenomenon. Firstly, variations of dust source regions over time might lead to changes in dust composition. If this was the case, the higher frequency of the Fetr oscillations dominated by the 23 ka period would imply that the position of the dust deflation belt was strongly dependent on insolation changes, probably through modulating the relative position of the Asian monsoon front. Higher low-latitude insolation may lead to stronger monsoon circulation (Clemens *et al.* 1991). Consequently, the monsoon front could penetrate more deeply into the deserts in northern China and this might result in dust composition changes. Secondly, changes in eolian grain-size might also lead to changes in dust composition because mineral composition in loess varies by different grain-size fractions (Peng and Guo 2001). If this was the case, changes of Fetr of higher frequency would reflect the strength of the dust transporting winds, or/and the distance of source region from sites of dust deposition as grainsize in loess of China is usually interpreted as a proxy of wind strength or desert extension. The latter, especially its south margin, is also linked with the strength of the Asian monsoon (Guo *et al.* 2000), which is strongly forced by insolation changes.

Conclusions

This study yields a high-resolution record of eolian iron changes in the past 900 ka based on a loess-paleosol sequence in the Loess Plateau region. Since eolian iron is thought to have significant effect on the ocean phytoplankton productivity (Martin 1995), investigation on this issue would provide a basis for understanding its possible effects.

Our investigations have shown that eolian iron content is not in good agreement with the long-term glacial–interglacial oscillations. It exhibits oscillations of much higher frequency dominated by a strong ~23 ka and clear 41 ka orbital periods while the ~100 ka glacial–interglacial period is undetectable. These results suggest that eolian iron content had been strongly affected by the summer insolation changes in the northern hemisphere while the mechanisms need further study.

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Abony brickyard: IRSL and TL analysis of "infusion loess"

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Abstract

Infrared optically stimulated luminescence (IRSL) and thermoluminescence (TL) dating techniques have been applied to alluvial loess from section Abony in the Great Hungarian Plain. IRSL age estimates a deposition age between 41.700 and 55.700 yr B.P., which is much older than previously radiocarbon dated.

Introduction

More than half of the area of the Great Hungarian Plain is covered with loess-like sediments *(Fig. 1),* which have been described as "alluvial loess" or "infusion loess" (Márton *et al.* 1979; Pécsi 1965). These sediments form an almost continuous cover on the high flood plains of the Danube and particularly of the Tisza river, where they occur a few meters above the present-day flood plain on the alluvial plains. Márton *et al.* (1979) interpreted the alluvial loess in the area of interest as a result of "fluvial, flood-plain, marshy and lake sedimentation in the Upper Wurm", but formed from sandy silt.

The pits of the brickyard are located south of the settlement Abony near by the Budapest–Cegléd–Szolnok railway on the west side of the road leading to Köröstetétlen *(Fig. 2).* Geomorphically this area belongs to the low flood free plain and high flood plain depression, which is gently sloping towards the Tisza River. Abony is situated within the area of the former, with early Holocene meanders of Zagyva river, a tributary of the Tisza. The surface of the alluvial plain is covered by meadow clay soil, minor landforms are composed of alkaline-rich sediments in depressions and abandoned meanders; low sand dunes along the river bank are covered by chernozem soil.

Many of the meanders of Tisza were eliminated during the river regulation activities in the 19th century and excavations, which were carried out simultaneously, exposed remains of *Bos primigenius, Bison priscus* and locally *Elephas primigenius* from not exactly identified horizons of the infusion loess.

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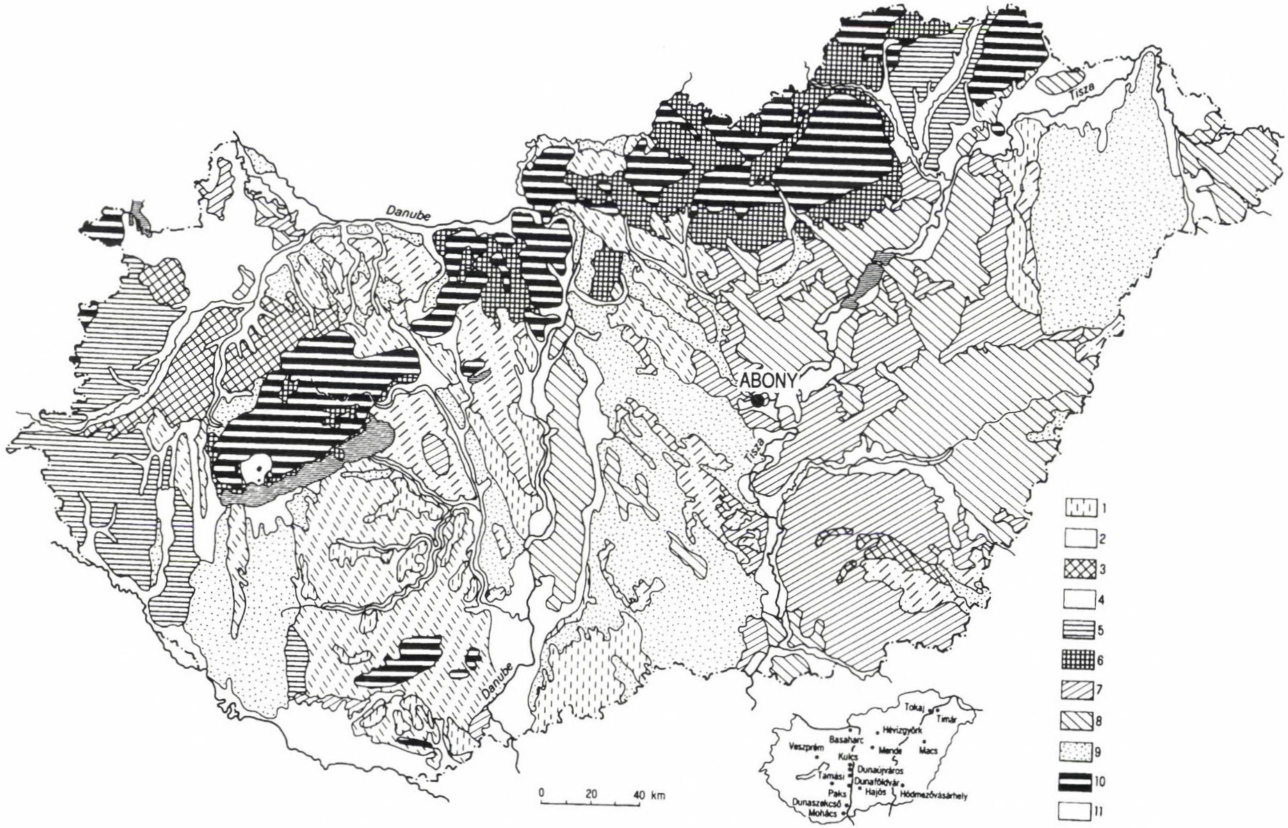

Fig. 1. Map showing the distribution of loess and loess varieties in Hungary, after Pécsi 1997). – 1 = typical loess; 2 = slope loess; 3 = brown loess, sporadic; 4 = sandy loess; 5 = brown loess, discontinuous; 6 = loess derivate, loess loam; 7 = high flood plain loess, silt; 8 = dominantly flood plain loess, silt; 9 = blown sand; 10 = medium height mountains sporadically covered by slope loess derivate; 11 = recent flood plain clay, silt, fine sand and gravel

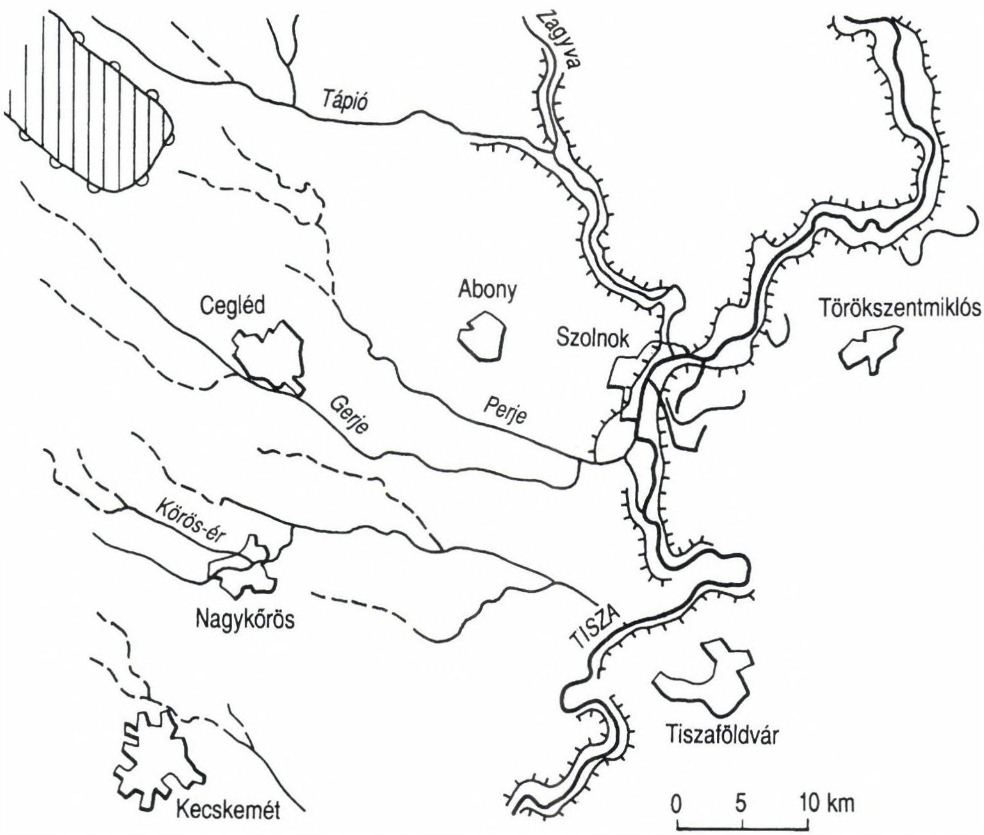

Fig. 2. Map showing the location of the section at Abony and the area around the Tisza river

In the area of Abony, brickyard pits were excavated within larger patches of the low lying solonchak pastures and of meadow soils turning more alkaline in deeper horizons. The infusion loess, which is actually used for brick manufacturing, has a thickness of 2–4 m and is superimposing fluvial sand. This latter sediment, which is locally intercalated by peat, is the first aquifer providing abundant water in the vicinity of Abony. The overlying infusion loess contains recurring fluvial sand, barely wedging out flood-plain clay soil and salt-affected soil horizons of different thickness. Similar exposures of infusion loess and fine sand were recorded along the middle stretches of Tisza in the brickyard sections of Szolnok, Törökszentmiklós, Martfű, Tiszaföldvár and at other exposures of the Tisza plain *(Fig. 2)*.

The loess sequence of the key section at Abony, where a typical loess-paleosol sequence of infusion type is exposed, has a thickness of several metres and has been studied in detail. Figures and references enclosed give an essential information and allow comparisons of the lithological properties and about the basic problems of the chronological investigations of the Middle Tisza region (Lóki *et al.* 1994; Márton *et al.* 1979; Sümegi *et al.* 1992).

Dating of mollusc shell and humus collected from the infusion loesses of the Tisza Plain was presented by Pécsi (1975), Pécsi *et al.* (1979) and Márton *et al.* (1979) resulting in radiocarbon age estimates ranging from 18 to 24 ka for the uppermost layers between 2.50 and 4.00 m below surface. However, radiocarbon dating from the 1980s and 1990s or earlier gave rise to methodological problems (Geyh 1991).

It has turned out that ^{14}C ages represent in many cases age underestimations, if the true absolute age of the samples is reaching or extending 30 ka. Radiocarbon dating results obtained through humus extraction regularly produce significantly lower values and hence age underestimations, if compared with charcoal analyses.

The chronological interpretation of these sediments is mainly based on few radiocarbon datings. in the present preliminary study thermoluminescence (TL) and infrared optically stimulated luminescence (IRSL) dating methods were applied to four samples of infusion loess from the Abony brickyard in order to test the suitability of luminescence dating, techniques and. to set up a more reliable chronological frame for the alluvial loess sequence at the Abony section.

Luminescence dating

Comprehensive reviews of the state of the art of luminescence principles, applications and limitations are provided by Aitken (1998), Duller (1996), Frechen (1998), Prescott & Robertson (1997) and Wintle (1997).

Thermoluminescence (TL) and infrared optically stimulated luminescence (IRSL) is the light emitted from crystals such as quartz, feldspar or zircon, when they are stimulated with heat or infrared light after receiving a natural or artificial dose of radiation. As a result of natural radiation in sediments, the emission of light increases with time and dose.

The equivalent dose (ED) is a measure of the past radiation absorbed and, in combination with the dose rate, yields the time passed since the last exposure to sunlight. Natural radiation results from the radioactive decay of isotopes in the decay chains of ^{235}U, ^{238}U, ^{232}Th, and decay of ^{40}K, some minor isotopes including ^{87}Rb, and cosmic rays. The luminescence age equals equivalent dose divided by dose rate.

Systematic luminescence dating studies indicate that reliable IRSL and TL age estimates can be obtained up to approximately 100.000 yr for loess and loess derivatives enabling a regional correlation of loess-paleosol sequences, a land-sea correlation, and providing a chronological framework for the last interglacial/glacial cycle, as summarized by Frechen (1998).

For older loess deposits the reliability of luminescence data is unknown owing to saturation of the signal and a lack of independent age control or dating methods (Frechen and Dodonov 1998). In the present study all measurements were carried out on the 4–11, grain size fraction, using the preparation technique, as well as the analytical and methodological approach, described by Frechen *et al.* (1997). A corrected

Table 1. Dosimetric results

Sample	U (ppm)	Th (ppm)	K (%)	α efficiency	Dose rate (Gy/ka)
AB1	2.61	8.63	1.20	0.05	2.7±0.2
AB2	2.82	9.28	1.46	0.07	3.3±0.3
AB3	2.53	9.07	1.58	0.07	3.2±0.3
AB4	2.85	10.20	1.60	0.07	3.5±0.3

Table 2. IRSL and TL equivalent dose and age results

Sample	IRSL dose (Gy)	TL dose (Gy)	IRSL age (ka)	TL age (ka)
AB1	114.5±8.7	153.0±26.5	41.7±4.6	53.3±10.2
AB2	158.6±9.5		48.2±4.9	
AB3	179.7±8.1		55.7±5.2	
AB4	154.6±6.3		44.6±4.1	

water content of 20±5% and a radioactive equilibrium of the sediments were assumed (*Table 1* and *2*). Fading tests have not been applied for the samples from Abony.

Results

Sample AB4 was taken from a grey, yellowish brown, gleyish silt horizon at a depth of 5.00 m below surface. Samples AB1–3 were taken from loess-like, clayey silt, characterized by Fe and Mn nodules, at a depth of 3.40 m, 2.50 m and 1.75 m below surface, respectively (*Fig. 3*).

The average potassium, uranium and thorium content is 1.5%, 2.7 ppm and 9.3 ppm, respectively, resulting in a dose rate ranging from 2.7 to 3.5 Gy/1000 yr. These dose rate values are comparable to those obtained from the loess sections Basaharc, Mende and Paks in earlier studies (Frechen *et al.* 1997; Wintle & Packman 1988), which ranged from 2.4 to 5.0 Gy/1000 yr. The equivalent dose (ED) values show an increase with depth for the regeneration method. The highest ED value of 180 Gy was obtained for sample AB3, the lowest one of 115 Gy for the uppermost sample AB1. The IRSL age estimates of the four samples range from 41,700±4,600 yr to 55,700±5,200 yr. B.P. The TL investigation of sample AB1 resulted in an age estimate of 53,300±10,200 yr B.P. It is likely that the sampled infusion loess was deposited between 40,000 and 56,000 yr B.P., if the average value of the age estimate is taken into consideration.

Discussion and conclusion

^{14}C age of charcoal sampled from the young loess series and analysed remarkably differ from these IRSL and TL age estimates. The IRSL and TL values have proven to be considerably older than ^{14}C values for the same aluvial loess horizon in Abony (*Table 3*).

IRSL age estimates of AB1–AB3 samples down the stratigraphical sequence were proportionally older ranging from 41,700 to 55,700 yr.

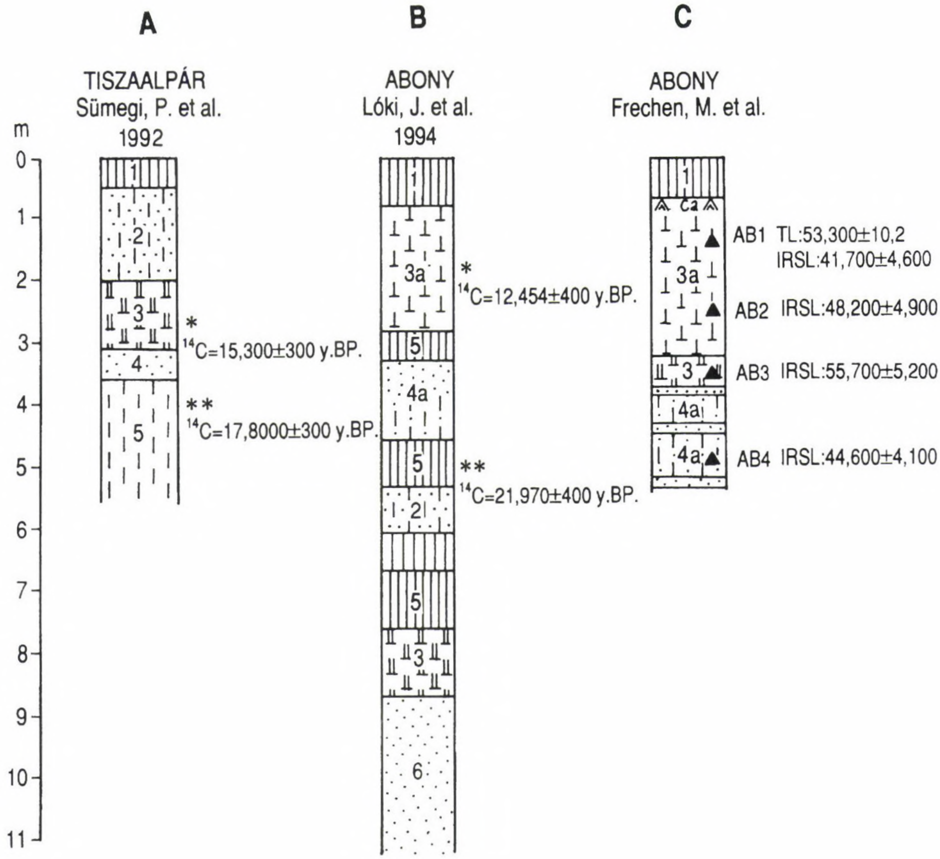

Fig. 3. Lithological and chronological comparison of the sequences from the sections at Tiszaalpár and Abony. – *A*: 1 = recent soil; 2 = loessy sand; 3 = infusion loess; 4 = eolian sand; 5 = loess; mollusc shell datings by ATOMKI, Debrecen. According to Borsy *et al.* 1991) in Bocsa, a 5 m deep borehole gave ^{14}C age of peat 21,580±300 yr B.P. *B*: 1 = recent soil; 2 = loessy sand; 3 = infusion loess; 4a = fluvial sand; 5 = loess-paleosol; 6 = fluvial sand. *C*: 1 = recent soil (salt affected meadow chernozem); 3 = infusion loess; 3a = structural flood plain loess; 4a = loessy sand; ········ = intercalated sand. Between Abony and Tószeg at the base of the riverbank dune (3.9–4.0 m) ^{14}C analysis of mollusc shells provided an age of 11,730±400 yr B.P. (ATOMKI, Debrecen, Hertelendi *et al.* 1993).

According to Márton et al. 1979, Pécsi 1982, 1993 in the Great Hungarian Plain ^{13}C and ^{14}C analyses of mollusc shells from infusion loess at 2.5–3.5 m depth by Laboratory of Quaternary, University of Helsinki provided ages as follows: Tiszaföldvár = 17,100±240 yr B.P.; Törökszentmiklós = 20,100±330 yr B.P.; Hódmezővásárhely = 24,100±360 yr B.P.; Mohács = 21,520±350 yr B.P.

Table 3. Results of chronological dating from the section at Abony

^{14}C analysis of mollusc shells sampled 1.50 m below surface	ca 12.5 ka
TL age estimate of polymineral fine grain material at 1.50 m b.s.	53.3±10.2
IRSL age estimate from the same sample at 1.50 m b.s.	41.7±4.6 k
^{14}C age estimates from mollusc shells at 5.00 m below surface	21.9±0.9 k
IRSL age estimates from polymineral fine grains at 5.00 m b.s.	44.6±4.1 k

The IRSL age estimate of sample AB4, taken from a depth of 5.00 m below surface (b. s.), is about 10,000 yr younger than that of AB3, most likely due to scattering from sample-to-sample, e.g. caused by the standard deviation of about ±10% or insufficient bleaching of the sample prior to the fluvial deposition or the flood plain soil formation.

The stratigraphic profile of the Abony brickyard and its ^{14}C age estimates are very similar to those of the Törökszentmiklós brickyard (*Figs 1* and *2*); where radiocarbon dating of mollusc shell collected at a depth between 2.20–2.60 m, resulted in age estimates of 20,100±1,300 yr. B.P.

The reasons for the apparent discrepancies between luminescence and radiocarbon age estimates could be due to:

1. Insufficient bleaching of the sediment prior to deposition which would result in an age overestimation of IRSL and TL;

2. A radioactive disequilibrium occurs for the alluvial loess (not tested), which would result in age underestimation;

3. The previous stratigraphic interpretation of the alluvial loess based on radiocarbon data is false.

However, similar discrepancies between radiocarbon and luminescence age estimates were described for the Upper Mende paleosol (MF_1) at Mende, Basaharc and Tápiósüly.

TL data obtained by Frechen *et al.* (1997), Oches and McCoy (1995), Wintle and Packman (1988), Zöller and Wagner (1990) contradict the previous radiocarbon age estimates. It is likely that the radiocarbon ages for the MF_1 paleosol are significantly underestimated.

A more systematic luminescence dating study is required to test the suitability of the methods of an analogue sequence from the area of interest using a high resolution dating approach.

However, IRSL and TL data are considered to be more reliable for sediments older than 30.000 yr. B.P. Thus, it is likely that during the late Pleistocene the alluvial loess sequence at the Abony section, was probably deposited between 40,000 and 50,000 yr B.P. The rest of the record is missing, except the recent meadow or chernozem soils.

In the closer and wider surroundings of the Abony brickyard, abandoned meanders and dunes can be found along the Zagyva river. Their geomorphologic position suggests Holocene origin.

However, the question remains why the uppermost infusion loess of the Abony brickyard profile, which is underlying the recent soil at 1.50 m below surface, is older than 40.000 yr according to the IRSL and TL data, but no traces of pseudomorphoses or cryoturbation (periglacial phenomena) or dessication cracks have survived. An extremely cold and dry climate prevailed during the late glacial between approximately 18 to 16 ka B.P.

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The Ságvár-Lascaux interstadial (Upper Weichselian) and its palaeoecological reconstruction

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Introduction

The malacological studies of the Hungarian Upper Weichselian loess deposits have recently yielded significant achievements. The quantitative studies of samples collected by standard fine stratigraphic methods have been completed and made more exact by the application of new methods. By the development of the "malacothermometer method" (Sümegi 1989, 1996) it is possible to reconstruct past July mean temperatures. There are new findings on the interdependence of the malacofauna, climate and vegetation based on the repeated analysis of former quartermalacological data (Sümegi 1995). The chronostratigraphic correlation between these features has been supported by radiocarbon datings by the method developed in the Nuclear Research Center, Debrecen (Csongor *et al.* 1982, Hertelendi *et al.* 1989, 1992).

It has been realised that local changes could be correlated with each other on a regional level and also with global climatic changes.

Based on these considerations the palaeoecological reconstruction of the Hungarian Upper Weichselian, and mainly that of loessy deposits has been accomplished (Krolopp–Sümegi 1995; Sümegi–Krolopp 1995). As a result nine shorter malacostratigraphic levels have been identified within the Upper Weichselian and Late Glacial, marking different climatic and vegetation periods.

Out of these nine periods there are seven in the Upper Weichselian, which in our opinion fall between 30,000–13,000 BP determined by radiocarbon data (Krolopp–Sümegi 1995).

Concerning the deposits between 16,000–18,000 BP (radiocarbon age) a paleoecologically remarkable period was found. This period was characterised by the occurrence of *Vestia turgida* species (Krolopp–Sümegi 1990) and the dominance of *Punctum pygmaeum* species (Krolopp–Sümegi 1991). It was separated as a Punctum pygmaeum – Vestia turgida zonula within the Semilimax kotulai subzone of the

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Bithynia leachi – Trichia hispida malacological biozone (Sümegi–Krolopp 1995). This period of the Upper Weichselian having a relatively mild climate and favourable rainfall distribution has been found to be identical with the Ságvár-Lascaux interstadial (Gábori 1965; Gábori–Gábori-Csánk 1957; Gábori-Csánk 1978).

Sediments of the Ságvár-Lascaux interstadial in Hungary

During the Ságvár-Lascaux interstadial predominantly loessy sediments were formed in Hungary. The basis for their identification and classification was the presence of *Vestia turgida* and the significant dominance value of *Punctum pygmaeum* (generally >10%, but sometimes reaching 68%, i.e. Punctum pygmaeum – Vestia turgida zonula. Besides this the following features have been taken into consideration:

- Qualitative and quantitative characteristics of the malacofauna,
- Archaeological findings,
- Vertebrate fauna and
- Radiocarbon data

Based on the above factors, sediments of the Ságvár-Lascaux interstadial have been described from 20 sites of Hungary (*Figure 1*, *Table 1*). These sites could be clustered into five groups as far as they geographical distribution is concerned.

Surroundings of the Danube Bend

Sites belonging to this area are related to archaeological excavations (Gábori-Csánk 1984; T. Dobosi 1991, 1994; T. Dobosi *et al.* 1983). There is a common malacological feature of the layers containing tools of the Gravettian culture: beside the significant dominance of *Punctum pygmaeum, Vestia turgida* was typical of all of them (Krolopp in T. Dobosi *et al.* 1983; Krolopp 1991). The radiocarbon analysis gave similar, 16,000 BP data in all the three sites (Budapest–Csillaghegy, Pilismarót–Pálrét, Esztergom–Gyurgyalag). Later (in 1994) measurements were conducted on the Budapest–Csillaghegy site (*Table 1*) using the Quaternary mollusc (*Arianta arbustorum*) shells of the archaeological excavations.

Based on the archaeological evaluation of the excavations, the stratigraphical position of the cultural layers was put into the Ságvár-Lascaux interstadial. The most characteristic site of the group is the one at Pilismarót–Pálrét (T. Dobosi *et al.* 1983), where at a depth of 1.2–1.4 m a humic cultural layer was found in the sandy loess. Almost 100 flint tools, sculptured stone pieces, Tertiary mollusc shells used as trinkets (Gábori 1969) and bones of mammals were identified within this layer. Most of the latter turned out to be reindeer (*Rangifer tarandus*) bones.

As many as 22 snail species have been found in the cultural layer and in the sandy loess above it. Species having wide ecological tolerance spectrum or preferring open forest habitats were dominant in the fauna, the ratio of the cold-indicators was low. *Punctum pygmaeum* reaches maximum of its dominance above this cultural layer. The fauna-based reconstruction of July mean temperature provided a 16°C value, the dominance of snail species living along the margin between areas with open and closed vegetation indicates a mild climate and a distribution of precipitation adequate for the formation of an incipient soil.

Northeast Hungary

Within the loessy layers deposited on the lava sheet and foothills of the Kopasz Hill at Tokaj the species composition typical of the Punctum pygmaeum – Vestia turgida zonula has been demonstrated in eight sites; were five of them supported by radiocarbon data from (*Table 1*). Because of the morphology of the hill; significant microenvironmental and microclimatic influences could be traced that may have effected the occurrence of the characteristic species of the zonula; its basic trends were recognisable (Sümegi 1996).

The most characteristic changes have been found in profile I of the Bodrogkeresztúr brickyard. Mollusc fauna found within this 7 metre deep section at a level of 1.5–2.75 m, was identified as the Punctum pygmaeum – Vestia turgida zonula. In this layer cold-indicators (*Vallonia tenuilabris, Columella columella*) are remarkably repressed, their ratio diminished from 30–35% to 9–13%, while Holarctic and Central European fauna elements preferring a milder climate and extensive vegetation cover (*Clausilia dubia, Punctum pygmaeum, Vestia turgida, Discus ruderatus, Semilimax kotulai*) became predominant. Dominance of species preferring an extensive vegetation cover exceeds 50%.

Based on the fauna composition it is probable that under the mild, cool, but not cold, relatively humid and rainy climate natural forests might have developed in the area.

Central part of the Great Hungarian Plain

Of the sites of this region the most representative sequence of series is encountered in the sand-pit at Tiszaalpár (Sümegi *et al.* 1992), revealing the sandy and loessy layers in a 6.5 m deep profile.

The Punctum pygmaeum – Vestia turgida zonula could be identified in the profile at a depth of 3.5–4.0 m. The ratio of the cold-resistant, hygrophilous forest elements *(Clausilia dubia, Arianta arbustorum, Perforatella bidentata)* is remarkable (35%) and there is an enrichment of *Discus ruderatus* (10%). Among the species living in the transitional zone between open and closed vegetation types, the dominance of *Punctum pygmaeum* is especially high, amounting to 39%. This fauna composition marks a level of forestation, where the dominant species indicate the formation of

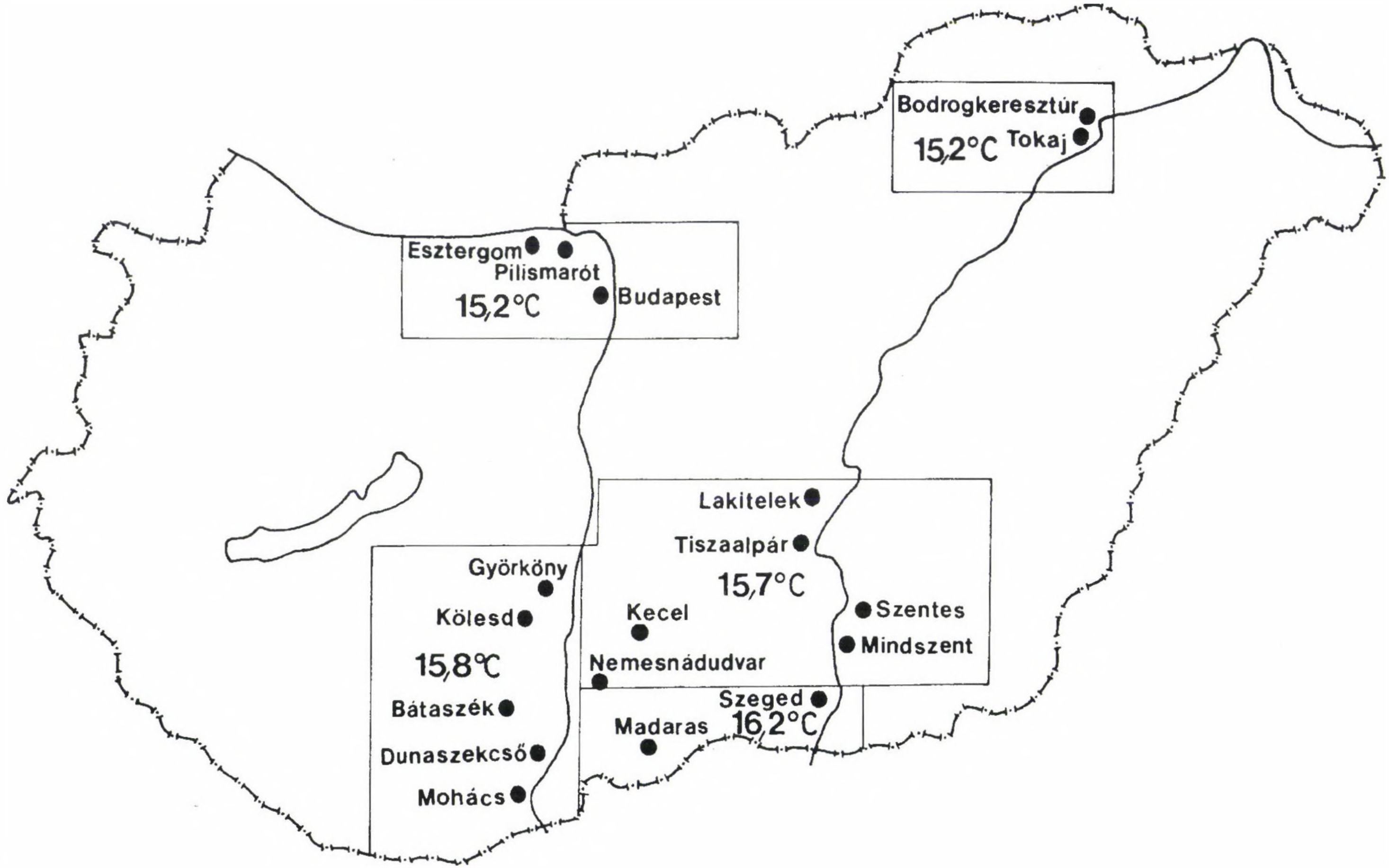

Fig. 1. Key quatermalacological loess profiles in Hungary and paleoclimatic conditions during the Ságvár-Lascaux interstadial

1. Table Chronological, paleoclimatic, archeological and paleontological data from Vestia turgida – Punctum pygmaeum zonule

Location	Deep m	Radiocarbon data (BP)	July paleo-temperature (°C)	Dominance of Punctum pygmaeum (%)	Vertebratare mains	Archeological data	Vestia turgida
Pilismarót–Pálrét	0.6–1.2	16.000 ± 200	16.0	16.4	+	+	+
Esztergom–Gyurgyalag	1.2–1.5	16.160±300	14.0	17.9	+	+	+
Budapest–Csillaghegy	1.7–2.0	15.935±142	15.6	+	+	+	+
Györköny–brickyard	1.2–3.2	–	16.0	10.7	–	–	–
Kölesd–brickyard	2.0–3.0	–	16.4	22.6	–	–	–
Bátaszék–brickyard	1.0–2.5	–	15.7	25.6	–	–	–
Mohács–birckyard	0.5–2.5	–	15.3	9.5	–	–	–
Dunaszekcső–brickyard	0.5–5.0	–	15.5	20.4	–	–	–
Kecel–birckyard	2.75–4.0	–	15.8	22.0	–	–	–
Mindszent–brickyard	0.5–1.25	–	14.2	11.6	+	–	–
Szentes–brickyard	2.0–2.75	–	15.7	13.4	–	–	–
Nemesnádudvar–brickyard	0.5–1.5	–	15.9	51.5	–	–	–
Lakitelek–brickyard	2.2–2.4	16.820±200	16.2	9.6	–	–	–
Tiszaalpár–sandpit	3.75–4.0	17.860±350	16.5	25.2	–	–	–
Madaras–brickyard	4.0–6.0	18.080±405*	15.7	47.3	+	+	–
Szeged–Öthalom. 1935	4.3–4.6	15.956±168 **	–	+	+	+	+
Szeged–Öthalom I 1995	1.0–2.75	16.000±200 16.080±150 16.323±145	16.8	24.2	+	–	+
Szeged–Öthalom II 1995	1.5–4.0	15.890±100 16.530±200 18080±200	16.2	19.3	–	–	+
Bodrogkeresztúr. brickyard I	1.75–2.0	16.850±200	15.1	16.0	–	–	+
Bodrogkeresztúr. brickyard II	2.75–3.25	17.680±200	15.9	4.25	–	–	+
Tokaj, Kereszt-mount I	1.0–1.5	17.619±170	15.6	10.4	–	–	+
Tokaj, Kereszt-mount II	1.5–2.0	–	15.1	10.7	–	–	+
Tokaj, Csorgókút-valley I	0.5–1.0	17.213±162	15.1	3.0	–	–	+
Tokaj, Csorgókút-valley II	0.75–1.0	17.504±106	15.7	36.0	–	–	+
Tokaj, Patkó-mine	2.25–2.5	16.322±162	14.0	+	–	–	+

*Radiocarbon data from the bedding loess layer of *Punctum pygmaeum* dominance level.
**A mammoth bone from archeological excavation in 1935. It was analysed by C^{14} method in 1995.

almost open forest vegetation. The high number of species (20) is comparable with that of the sites on the Great Hungarian Plain and Transdanubia. Besides the radiocarbon data the composition and development of the fauna indicate this layer to belong to the Punctum pygmaeum – Vestia turgida zonula.

Southern part of Transdanubia

There are no archaeological findings recovered from the profiles of this group of sites, and no radiocarbon data are available either. Malacological studies showing the dominance of *Punctum pygmaeum* together with the character of the fauna have proved that some layers were deposited during the Ságvár-Lascaux interstadial. A common feature of the malacofaunas studied that *Vestia turgida*, having a Carpathian distribution presently, is replaced by the forest species *Cochlodina laminata* and also *Orcula dolium*, having now a hilly distribution, occur everywhere. The dominance of *Punctum pygmaeum* gives a double-peak curve in several layer series, indicating probably a spell of deteriorating climatic conditions (Hum 1999a, 1999b).

The most characteristic sequence has been found in the site at Bátaszék, where the dominance-curve of *Punctum pygmaeum* and that of the total number of species show a Gaussian distribution, indicating that the series of deposits embraces the initial, main and final intervals of the period. The presence of groves is indicated by *Macrogastra ventricosa* and *Aegopinella ressmanni*, and species preferring the edge zone between open and closed vegetation types are predominant (Farkas 1995).

Southern part of the Great Hungarian Plain

In one of these sites the first palaeolithic finding of the Plain occurred at Szeged–Öthalom (Banner 1935), while T. Dobosi V. has excavated cultural layers with Gravettian tools in Madaras brickyard (T. Dobosi 1967, 1989).

The medium part of the Szeged–Öthalom profile – studied in detail (Krolopp *et al.* 1995) including drilling shallow boreholes – was put between 16,000–18,000 BP by radiocarbon dating. The composition of the fauna, especially the high dominance of *Punctum pygmaeum*, exceeding 30% in some layers, the occurrence of *Vestia turgida* and the malacothermometer data have made it evident that this section is identical with the Punctum pygmaeum – Vestia turgida zonula. The relatively dense vegetation cover is proved by the high dominance value (>80%) of hygrophilous and subhygrophilous species living along the contact zone between open and closed habitats. Some gallery forest species (e.g. *Perforatella bidentata*) also appear in the fauna. The recent ^{14}C measurement of the mammoth bone (*Table 1*) found in the course of the 1935 excavations (Banner 1935), confirmed that it has an identical age with the tools of the Gravettian culture.

Summary

Based on quartermalacological data between 16,000–18,000 BP in the Danube-bend, along the rim of the North Hungarian Mountains, in the southern part of Transdanubia and of the Great Hungarian Plain, species those preferring denser vegetation cover had prevailed (*Mastus venerabilis, Discus ruderatus, Punctum pygmaeum, Clausilia dubia, Vestia turgida, Macrogastra ventricosa, Aegopinella ressmanni, Semilimax semilimax, S. kotulai, Vitrina pellucida, Bradybaena fruticum,* *Arianta arbustorum*). They are also dominant in the studied profiles. Parallel with the expansion and dominance of the forest species, and of those preferring denser vegetation cover or living in the zone at the edge of open and closed habitats, the previously dominant fauna-elements of open areas (*Columella columella, Pupilla sterri, Vallonia tenuilabris*) had disappeared or their ratio decreased dramatically.

Based on data gained by the malacothermometer method the July mean temperatures increased from an earlier 12–14° to 14–17°C, while their average had risen to 15.6°C (*Table 1*). It is remarkable that for the Danube Bend and Northern Hungary a 15.2°C average has been calculated, whereas this value was 15.8°C for Southern Transdanubia and 16.2°C for the southern part of the Great Hungarian Plain (*Fig. 1*). These variations are largely similar to the recent regional differences.

Based on the dispersion process and the increasing dominance of mollusc species preferring forested, wet habitats one can state that during this 2000 year long period the amount of precipitation has also raised along with the 2–3°C increase in the July mean temperature compared with the previous climatic phase.

Based on recent analogies 16,000–18,000 years ago essentially taiga-like forests developed in the Carpathian Basin, covering large areas. At the same time malacological data indicate that there should have been areas with open vegetation, thus the vegetation cover must have shown a mosaic pattern.

The vegetation pattern reconstructed by the involvement of malacological data is supported by the analysis of the large number of charcoal samples from the sequences of similar age. Studying these samples Stieber (1967) has reconstructed a broad-leaved taiga environment in the Carpathian Basin between 16,000–18,000 BP. The new paleobotanical data suggest that there were areas covered by closed taiga forest and open coniferous forest within patches of steppe and within the taiga there were also groves of deciduous trees during the Ságvár-Lascaux interstadial period (Willis *et al.* 1995; Sümegi 1996; Rudner *et al.* 1997).

As a consequence of the emergent forests under the relatively mild, rainy climate an intense soil formation had started, as a result of which a thin, humic loessy layer, an incipient soil, the upper humic layer of the Dunaújváros–Tápiósüly loess-complex was formed (Hahn 1977; Pécsi 1975, 1993). In the bottom of the Tápiósüly profile the age of this weakly developed soil layer is around 16,000–17,000 years, thus it is identical with the development of the Punctum pygmaeum – Vestia turgida zonula.

The vertebrate fauna of the Ságvár-Lascaux interstadial is known – with a few exceptions – by findings from cultural layers, thus it is highly selected. A relatively large number of reindeer remains were found from several colonies. There is probably a correlation between the migrating direction of reindeers and the sites of human colonisation (Sturdy 1975; Vörös 1982). The distribution of Gravettian sites of Ságvár stage (Gábori–Gábori-Csánk 1957; T. Dobosi 1993, 1994; T. Dobosi–Vörös 1986, 1987; T. Dobosi *et al.* 1988) indicates a change in paleoecological conditions within the Carpathian Basin during the last phase of Weichselian. Humans of the region hunted mainly the highly mobile reindeers and wild horses (Sturdy 1975). During the Upper Pleistocene the southern boundary of the reindeer distribution stretched across the southern part of the Carpathian Basin (Vörös 1982). Based on macromammalian analyses of Upper Palaeolithic sites of the analysed region (Vörös 1982) the reindeer herds spent the winter season in the Carpathian Basin, mainly in its Transdanubian areas. The paleobotanical and malacological data of this region suggest that the areas covered by taiga and open coniferous forest were enclosed by the steppe patches and within the coniferous forest there were also pockets of deciduous trees during the microinterstadial of Ságvár stage (Stieber 1967; Willis *et al.* 1995, 1997). The nearest present-day analogue of this type of community can be seen at the southern edge of European boreal forest where many of these types are present in small pockets within the forest (Shugart *et al.* 1992).

It is therefore suggested that a special kind of open taiga environment developed in the Carpathian Basin during the microinterstadial time of Ságvár stage as the last target area of the migration of reindeer herds. The similar modern analogue to this type of reindeer migration now can be seen between taiga and tundra zones in North America and in the northern part of Eurasia. There the reindeer herds live in tundra in summer and they start migrating to the taiga zone with the end of that season. The herds spend the winter season in the taiga zone, and they return to the tundra when the winter season is over.

Based on the distribution of the archeological findspots and paleozoologial, paleobotanical and quartermalacological data a relationship seems to have existed between the distribution of the Upper Paleolithic sites and the route of reindeer migration and the ecological and vegetation conditions of the Carpathian Basin and Central Europe during the last phase of Upper Weichselian. Probably the Upper Paleolithic hunters were following herds of reindeer as they moved from the winter (ancient taiga) grounds in the Carpathian basin to summer (ancient tundra) ranges in the Alps, Bohemian Basin and German-Poland Plain during the Ságvár-Lascaux interstadial time.

Based on malacological data one can state that prior to 18,000 BP there was a warmer, but dryer climatic phase (Krolopp *et al.* 1996), followed after 16,000 BP by a colder and dryer period (Sümegi *et al.* 1991).

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Recent and Pleistocene chemical erosion in the Alpine and Dinaric karst (A case study from Slovenia)

IVAN GAMS[1]

Abstract

At present chemical erosion in the river basins is diminishing from Southern Julian Alps (with 80–110 microns annual lowering of the surface) to south-east and south (30–40 microns). During Holocene more than half of chemical erosion on the barren high Alpine karst occurred in the endokarst between the rocky surface and the subsurface water table deep below. In contrast, in the lower Dinaric karst it takes place mostly in epikarst whereas in its vadose zone the filling of voids with the flowstone prevails. In the cold stages of Pleistocene the Alpine system had expanded onto the Dinaric mountains. The density of the dolines and poljes does not show any correlation with the intensity of chemical erosion.

Introduction

Quantitative analyses of chemical erosion have been carried out only recently. Mass of the karstifying rocks ranges from sea level to an altitude of 2864 m in Slovenia. Knowing the control factors of recent erosion in the high Alpine area it is possible to estimate the intensity and effects of chemical erosion in the lower continental and submediterranean Dinaric karst during the cold phases of Pleistocene.

Recent chemical erosion in the high Alpine karst of Slovenia

Alpine karst (comprising an area of about 2237 km^2) built mostly of Mesozoic limestone (54%) and dolomite (15%) is the dominant feature in the Julian Alps (Triglav, 2864 m), Kamnik-Savinja Alps (2558 m) and in the Karavanke Mountains. Of the Karavanke (the border area of Slovenia and Austria) here only the higher carbonate crests and ridges are dealt with (*Fig. 1*).

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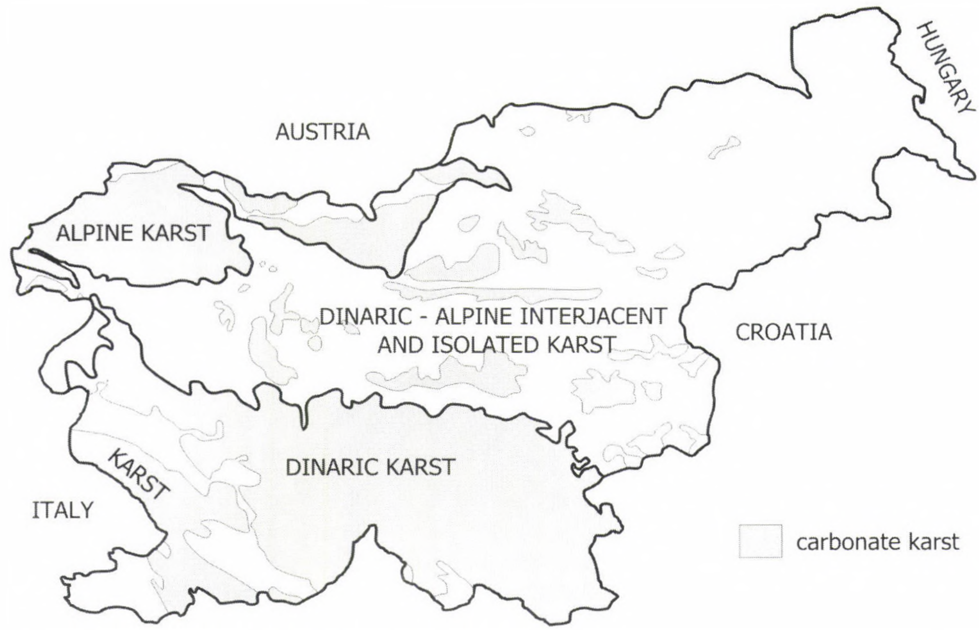


Fig. 1. The major karst regions in Slovenia: Alpine karst on its northwesten part, Dinaric karst in southern Slovenia, and Dinaric-Alpine karst between them with isolated karst. The region Karst is indicated, where the technical term karst is originated from

Measurements of water hardness on the outlet of the Triglav glacier (before it disappeared almost completely at the end of the last century) 25 m below the limestone surface found the total hardness to be 2.3 °GD (German degrees). This is a lesser amount than the saturated water at 0 °C and at the normal atmospheric pCO_2 (Ford and Williams 1989, p. 63) can dissolve. This water percolates from 2400 m through the 280 m deep Triglav pothole and then through the undiscovered cavern to the spring at the end of the valley Vrata at the altitude 1250 m a.s.l. its hardness increases to 4.3 °GD (*Fig. 2*). It means that half of the solution takes place in endokarst. After the withdrawal of the glacier many openings (ponors) in the limestone, modelled by the outlets from the glacier, had become uncovered. The water from the rubble inside of glacier has 4.0 °GD (Gams 1962 1966).

Similar low values of total hardness were recorded on the rocky limestone and dolomite of the plateau of Mt. Kanin (between 1800 and 2100 m). Until the last century three small, now melted glaciers existed on the northern side of the Kanin crest (2587 m, on the Slovenian/Italian border) in a similar exposition as in the case of the Triglav glacier. On the southern part of the Kanin plateau, in Slovenia, in the karren flowing precipitation waters and the waters from melted snowfield have 1.49 to 2.96 °GD, in the longer grooves 1.5 to 3.95 °GD and in the caves under 8–50 m deep ceilings 3.96 to 5.24 °GD. On the Kanin plateau there is a thick snow cover in winter and great differences are expected in deep depressions filled with snow. But

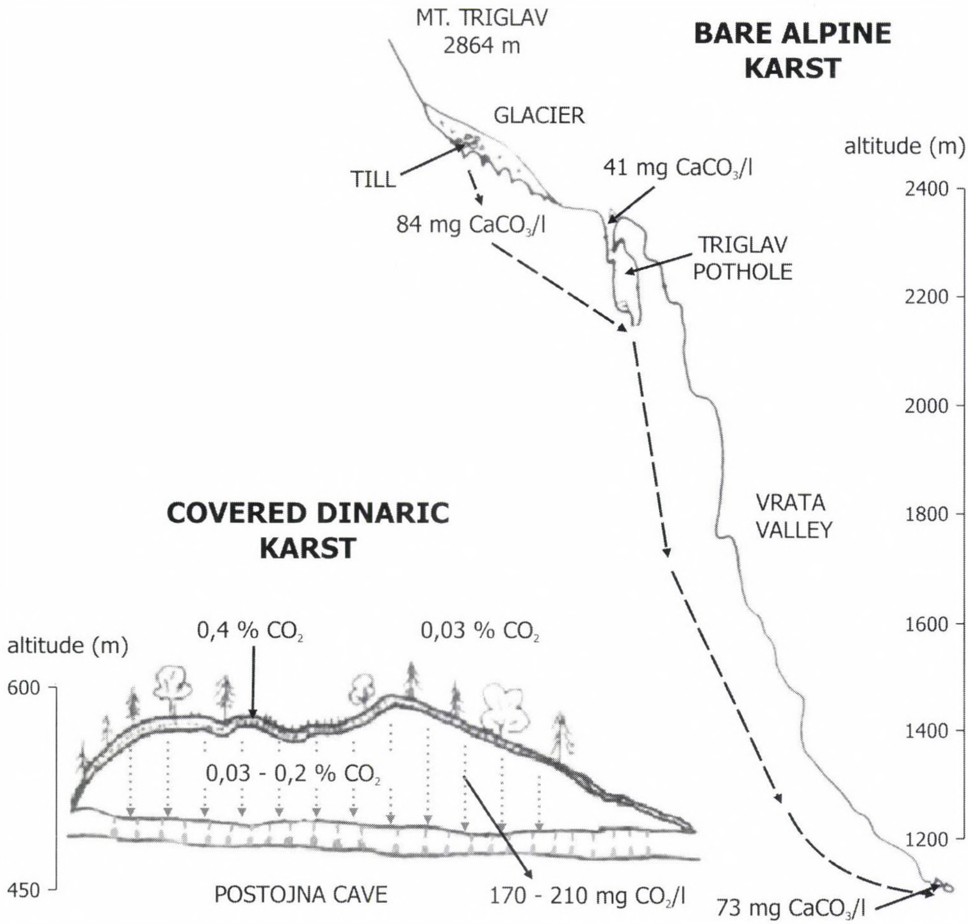


Fig. 2. The streams percolating through the vadose zone of the bare Alpine karst are aggressive down to the spring in the valley bottom (the case of stream from Triglav glacier to the spring in the valley Vrata). In the covered Dinaric karst represented by Postojna Cave the solution capability is exhausted already in the epikarst as the dripping water generally is depositing dripstone (and flowstone)

the most intense solution occurred in wide valley-like depressions. Precipitation measured by totalizator at the altitude 2080 m was found as 3418 mm annual average between 1953 and 1964. In the frame of an international research program chemical erosion was also measured with standard limestone tablets. On the tablets, exposed for three years on the Mt. Kanin at 2050 m in different sites, the solution has taken away a sheet of 0.3 to 1.6 microns per year, in the depressions more than in the convex places (Kunaver 1978, 1979).

Annual mean precipitation below the crest of Triglav (2964 m) on Kredarica (2541 m) was 1994 mm between 1961 and 1990). Measurements with limestone tablets have shown different solution rates. The highest value was found at the tablets hung 242 cm above the surface whereas the lowest one in the sandy, mostly frozen soil (Gams 1985). The calculated runoff is 95% on Kanin and 68% on Triglav.

Calculated chemical erosion in term of microns/a (i.e. meter per one million year) is given in *Table 1.*

Table 1. Chemical erosion in the Alpine karst measured in microns and water hardness in German degrees (°GD)

Location	Microns/a and °GD
Kanin* limestone tablets: on the stony surface	1.6
in soil, at a depth of 10 cm	0.3
Glijun** spring at Bovec, 480 m, °GD	5.5
River Soča, spring in Trenta valley	4.2
Kredarica***, limestone tablets, hung at + 242 cm	1.4
placed on stony surface	1.7
in soil, at a depth of 10 cm	0.3
Total hardness of water, °GD:	
stream outlet from glacier, 15 m below surface	2.3
spring outlet at the end of Valley Vrata (1250 m)	4.3
Karavanke***, Mt. Uršlja gora, 1700 m	
limestone tablets, hung at + 245 cm	1.3
placed on grass surface	0.2
in black humus rendzina, 13–15 cm	1.5

*Kunaver, J. 1978, 1979, ** Komac, B. 2001, *** Gams, I. 1985.

Measurements with limestone tablets in the barren high Alpine karst show a weak atmosphenic solution, and that on the surface and in soil patches. On the top of Mt. Uršlja gora (1700 m), in eastern Karavanke the karst is covered by grass and humus rendzina. The solution measured by limestone tablets in the air and in the soil is 0.2 to 1.5 microns per year. As the measurements at the local meteorological station show a lower amount of precipitation (1317 mm) than on Mt. Kanin, the calculated chemical erosion is less.

There are differences among singular data of measurements with limestone tablets and of total hardness of water referred to in the *Table 1* but on the high mountain stony surfaces all of them prove a weak solution, between 0.2 and 1.7 microns yearly. The total hardness of spring water in the valley below these mountains, however, is 2–3 times greater than on the high plateau thus proving that more than half of the solution takes place beneath the stony surface, i.e. in the vadose zone of about one thousand meter depth.

This, however, is valid only for the stony surface that prevails also on the slope of the once glaciated crest-valley relief. According to glacial geomorphology the typical glacial valley with steep slopes and wide bottoms is attributed to the abrasion by ice. In limestone and dolomite of the Julian and Kamnik Alps even bottoms of valley heads are wide and deeply cut. Beside glacial erosion also the melting of snow has to be taken into account falling from crests in avalanches onto the deep valley glacier, where on its flank below the snow line the water enters in contact with carbonate rock for the first time (Gams 1992). Its solution (at about zero temperature) can reach 70 mg $CaCO_3$/l (see Ford and Williams 1989, p. 639, figure 3.7 p. 63). Under the ice sheet a vivid locally accelerated solution occurs under the influence of the higher gravitational atmospheric and CO_2 pressure in the air bubbles and in the ice at the contact with valley bottom (Lismonde 2000). Typical glacial valleys in Slovenia are restricted to the compact limestone (compare Matkov kot and Logarska dolina in the Kamnik–Savinja Alps).

The southern Kanin plateau is located at an altitude between 2100 and 2350 m. One of the most abundant springs in the Bovec basin is Glijun (480 m, *Fig. 3*), situated in foothill position, near the settlement Bovec. Monthly oscillation of its hardness (fluctuating between 5 and 5.5 °GD), temperature and discharge of the water originating from high Alpine karst (Komac 2001) is modest in comparison to its counterparts in the Dinaric karst. The lowest values of calcium, total and carbonate hardness are typical of summer; this is the time of the highest discharge and low water temperature. Based on data of the meteorological station at Bovec (452 m a.s.l.) 22.8 per cent of annual precipitation falls down in summer, 21.4 per cent in winter, 24.9 per cent in spring and 31.6 per cent in autumn. The most intense snowmelt on the Mt. Kanin occurs between April and July; moisture and rainwater increase the discharge and reduce the hardness.

The total hardness of Glijun exceeds by about 1 °GD the hardness equilibrated with pCO_2 in the free atmosphere (at 5°C). An overwhelming part of this surplus

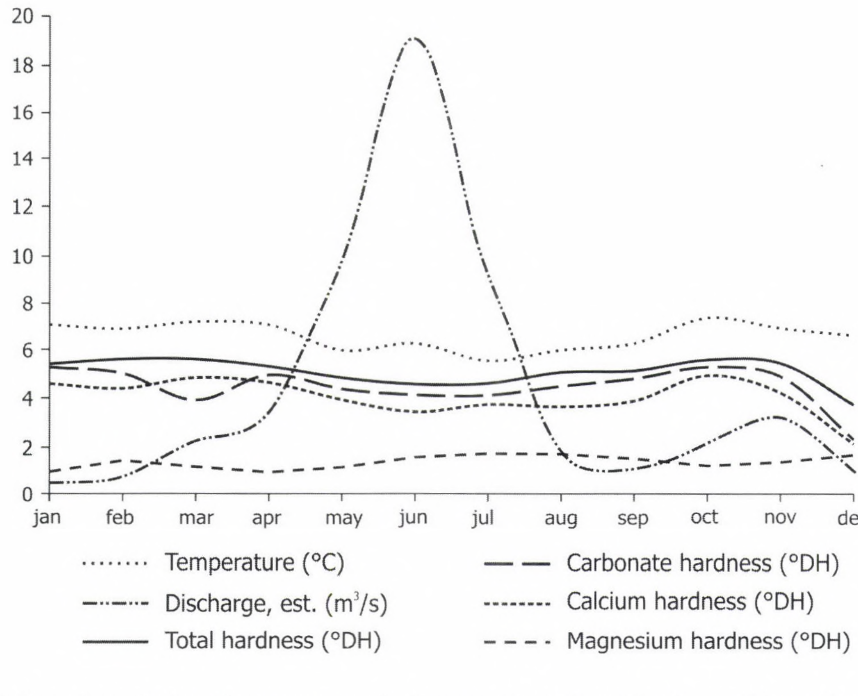


Fig. 3. Annual oscillation of temperature, discharge (estimated) and hardness of the spring Glijun draining the barren Alpine karst plateau of Kanin (Komac 2001)

presumably derives from the solution of the semipermeable dolomite, which lies in the base of the Dachstein (lower Triassic) limestone caprock of about 1000 m thickness. Over the 50 square km surface of the Kanin plateau more than 800 potholes are registered in the cave inventory of Slovenia.

Eleven of these potholes are deeper than 500 m and five of them are deeper than 1000 m, including the deepest one (Čehi II, 1553 m). They prove the intensive recent solution in the endokarst down to the semipermeable dolomitic base where the channels of minor steepness end.

Runoff, total hardness and chemical erosion of the rivers draining the high and stony Alpine karst of Julian and Kamnik Alps and their forested foreland are listed in *Table 2*. These basins in the Julian Alps display the highest precipitation and karst denudation values. Runoff data figuring in *Fig. 3* is based on records by gauges (1961–1990, Površinski 1998).

Table 2. Annual chemical erosion of three main rivers draining the Julian and Kamnik Alps

River name	Run-off l/s/km^2	Total hardness in mg $CaCO_3$ /l	Calculated chemical in erosion micron/a
Soča, at Kobarid	77.8	118	92.9
Tolminka, Tolmin*	78.0	79	78.1
Sava Bohinjka , Sv. Janez*	90.0	120	108.9
Kamniška Bistrica**, Kamnik	38.0	141	53.6

*The size of the drainage area is not certain
**About half of the drainage area is situated outside of the carbonate high Kamnik Alps

The greater chemical erosion of the Sava Bohinjka derives from the higher share of the forested area within its watershed. In the eastern valley slope of the river Sava Bohinjka, near the village Bohinjska Bela at an altitude of 1000 m is the entrance to the old touristic cave pod Babjim zobom.

Its speleothem was dated ca. 40,335 years (Franke and Geyh 1971). The cave was not yet open during the Wurmian glaciation still it preserved its speleothem in spite of the low temperature during the Wurmian glacial and nearness of the top of the Bohinj glacier.

Local intensification of solution by streams running down the stony slopes toward the covered karst was confirmed at Velo Polje (1680 m). Along the southern slope of the Mt. Triglav, namely in the Veljska dolina (valley), some impermeable interbeddings in the limestone force the water to flow as torrent to the bottom of Velo polje where it penetrates through the alluvial soil thus becoming more aggressive.

Locally accelerated solution was measured also in the nearby Malo polje along the torrents from the rocky Mt. Tošc (2275 m) while meandering on the marshy plain. Total hardness increased from 1.1 to 2 °GD and, in another stream, from 4.8 to 6.8 °GD (Gams 1963). Locally intensified solution under similar conditions is postulated for many poljes in the lower Dinaric karst during cold stages of the Pleistocene when temperatures decreased by ca 10 °C (Frenzel, Pécsi and Velichko 1992).

The present karst denudation in the Slovenian part of the Dinaric karst

Karst denudation in the main drainage areas of the Dinaric karst in Slovenia (with ca 5,000 square km surface of limestone and dolomite, *Fig. 4*) is calculated on the basis of measured calcium, magnesium and total hardness multiplied by the discharge of the rivers, and finally it is expressed in m^3CaCO_3+$MgCO_3/km^2/a$. If all the solvents originated from the karst surface, the extent of downwearing could be assumed as lowering of the surface equal to one meter in one million year. This is more or less valid for the karst covered by soil and vegetation that is characteristic for the Dinaric karst and lower part of the Alpine karst in Slovenia. Sources for the calculated discharge are publications of the Hydrometeorological Service of Slovenia (Površinske, 1998). The applied mean hardness of river water is the average of the summer measurements at mean stage. This method was preferred to the repeated measurements in different seasons.

Hardness of water in the rivers of the Slovenian Alpine and Dinaric karst tends to increase southward and eastward with the lowering runoff and altitude and growth of water temperature (*Fig. 4*), soil thickness and annual evapotranspiration. Runoff is the main factor for the chemical erosion in Slovenia, too. In the high Slovenian Alpine karst the water hardness is low and runoff is high as a result of low evapotranspiration and abundant precipitation. The highest values of karst denudation (above 100 $m^3/km^2/a$) are typical of the higher Dinaric and lower Alpine plateaus covered by soil and forest (the river basins of the Soča at Kobarid, the Tolminka at Tolmin and of the upper Kolpa). The maximum hardness and minimum karst denudation occur under submediterranean climate with the lowest runoff and the thickest soil in the lowest-lying southern part of the peninsula Istria, where annual evapotranspiration nearly equals precipitation (800 mm, Gams 1966). Annual evapotranspiration in Slovenia ranges from less than 200 mm (in the Alpine stony karst) to more than 750 mm (Površinski 1998).

The close relationship between the total hardness of the deep springwater and a low seasonal oscillation of its temperature have not been explained yet enough (*Fig. 5*). Biological activity obviously has an important role to play here.

There is a belt of lower Triassic dolomite along the northern border of the Dinaric karst in Slovenia, where the dissolved $MgCO_3$/l usually constitutes 33 to 45 per cent of the total hardness. This is a fluviokarst with valleys and ridges, an area almost without dolines, poljes and uvalas and with watercourses on the surface. There are only few caves and the slopes are covered with a thin layer of soil. Nevertheless, the degree of downwearing is somewhat higher here than in the neighbouring river basins with similar conditions, i.e. those built of limestone.

This fact seems to be in contradiction with the theory on the less solubility of dolomite at temperatures between 8–12 °C. The seasonal oscillation of calcium, magnesium and total hardness of the rivers with Triassic dolomite bedrock is minor in comparison with that of the rivers flowing from limestone. This is evident from

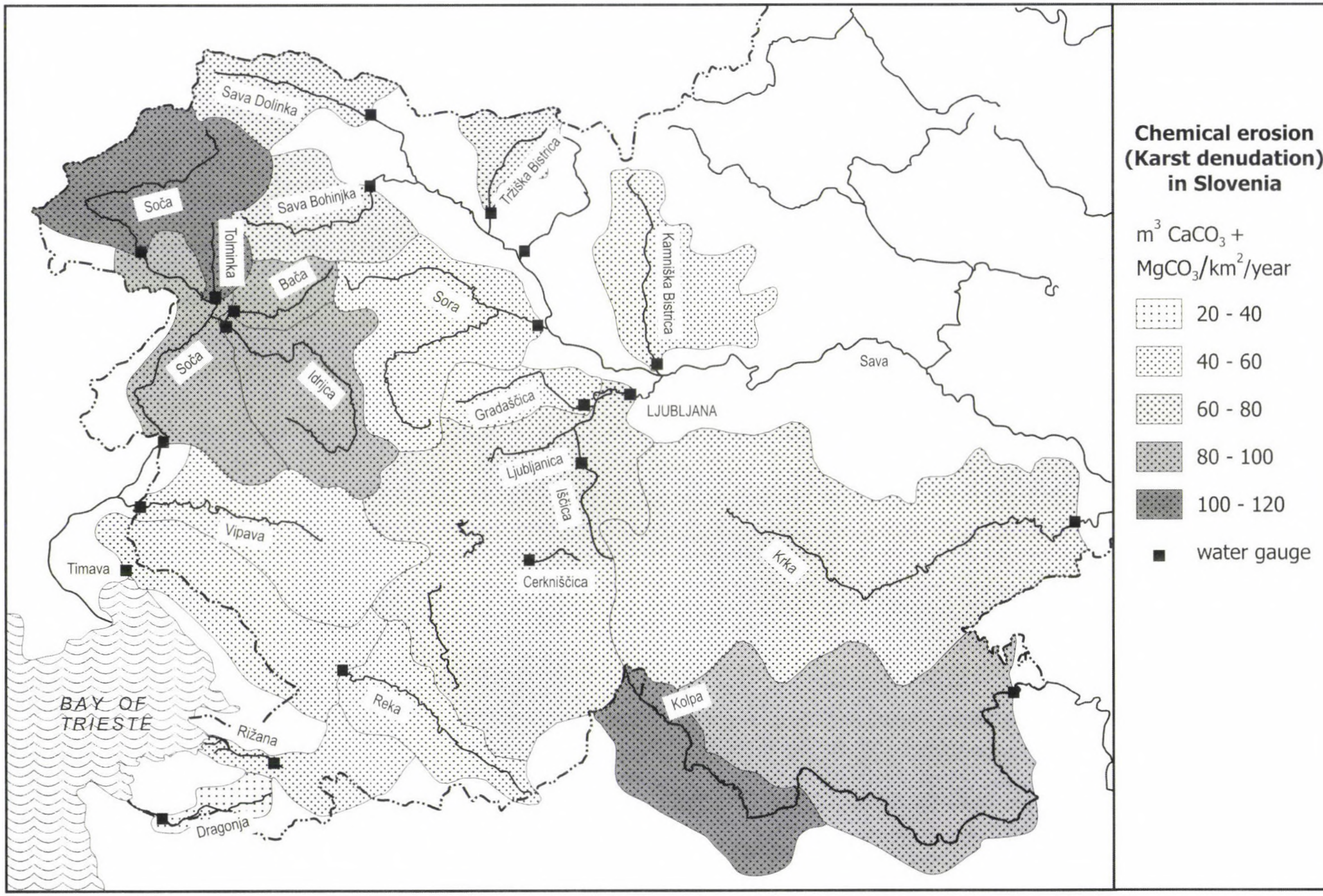


Fig. 4. Correlation of temperature (in °C) and carbonate hardness (°GD) of the main karst springs in Slovenia and in Croatian part of Istria. Below 6 °GD are Alpine springs and above 12 °GD springs in Istria

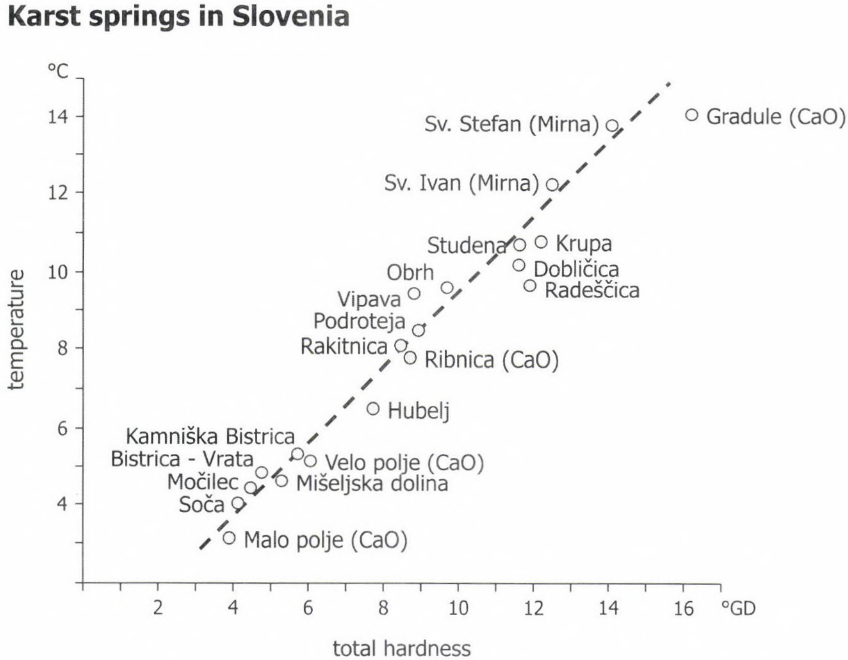


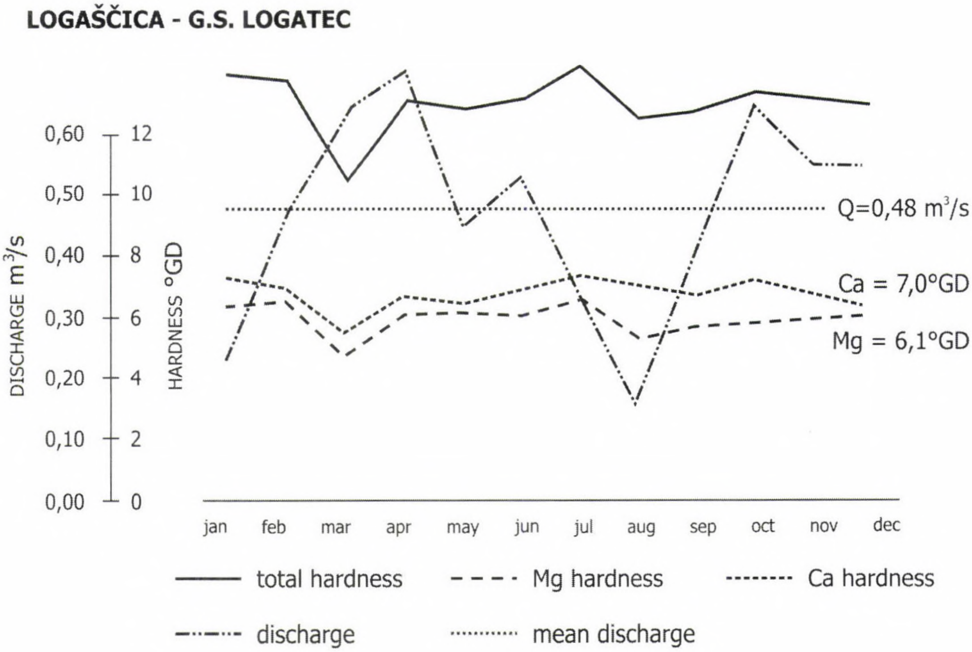
Fig. 5. Annual oscillation of discharge and hardness of the river Logaščica draining the fluviokarst built of Triassic dolomite (upper diagram). Below is the spring Veliki Obrh draining the prevalent limestone river basin (Kolbezen 1976).

the comparison of the Logaščica draining the dolomite with the spring in limestone Veliki Obrh (Kolbezen 1976, *Fig. 6*). A most reasonable explanation is a less diminished hardness of water after its percolation in the vadose zone. In the limestone there are larger voids and it has a better connection with surface atmosphere.

This water is therefore loses pCO_2 faster and deposits flowstone. During the Holocene the Dinaric karst covered by soil has been in stage of general filling of caves that formed mostly in the glacial epochs. Under the 10–120 m thick limestone nearly all of dripping waters in the 6 km long touristic channels of the Postojnska cave deposit dripstone.

However, only some of them conduct aggressive water, mostly at the end of the side channel Pisani rov where the surface of the collapsed doline (Jeršanove doline) is near (*Fig. 2*). The average total hardness of the dripping water on the ceiling in the Postojna cave is 170–210 mg $CaCO_3$/l, and in the region of Karst it is even higher.

The total hardness of the spring water draining the Triassic dolomite (with the same amount of atmospheric precipitation) is a little higher than that within the limestone, where there is a greater number and denser network of karst depressions on the surface. This proves that the general intensity of chemical erosion has no direct effect on the karst "dissection" (i.e. on the density of depressions).



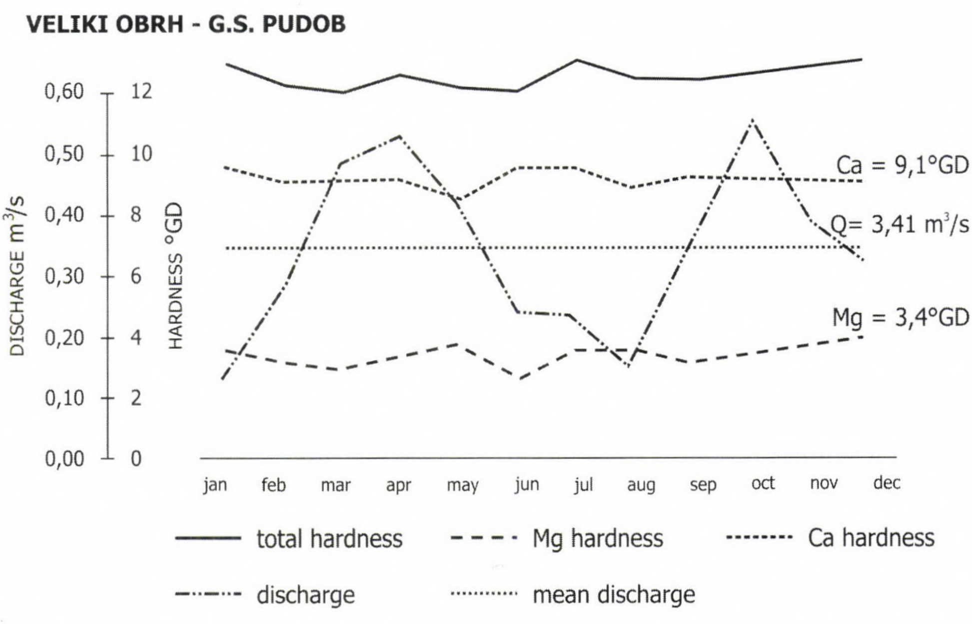


Fig. 6. The streams percolating through the vadose zone of the bare Alpine karst are aggressive down to the spring in the valley bottom (case of stream from Triglav glacier to the spring in the valley Vrata). But in the covered Dinaric karst represented by Postojna Cave the solution capability is exhausted already in the epikarst as the dripping water is generally depositing dripstone (and flowstone)

The present snowline on the Mt. Triglav lies at about 2700 m a.s.l. Based on the data recorded at the nearby meteorological station Kredarica (2514 m) the local annual temperature is estimated at -1.9 °C. Wurmian glaciers of limited extension in the Dinaric karst covered only the southern slopes of Mt. Golaki (1485 m) and V. Snežnik (1796 m); with snow line presumably stretching between 1350 and 1550 m, i.e. 1150–1350 m lower than at present on the Triglav. In the central Julian Alps the tree line appears at 1800 m a.s.l. in the Kamnik Alps it is at 1700 m and on the Mt. Snežnik at 1650 m. July temperature (1961–1990) at the tree line on the Krvavec (Kamnik Alps) is 11.3 °C and on Mt. Uršlja gora (Karavanke) it is 10.6 °C. If the Wurmian tree line was to a similar extent lower as the snow line, the forest grew then in the littoral karst up to the altitudes of 200–230 m. July temperature recorded at the meteorological station Komen (289 m) now is 20.9°C. On the other hand, palynological research has found Wurmian tundra park vegetation all over the lowland of Slovenia. Trees had retreated into microrefuges, from where the coniferous forests and later (under warmer climates) the deciduous ones spread (Šercelj 1996). In the coldest Wurmian glaciations the fauna of the continental and littoral karst was composed by cave bear, wolf, fox, cave lion, hyena, red deer, Pleistocene bovides, locally marmot, mountain hare, steppe wissent and reindeer (Pohar 1995). This arctic Alpine fauna is typical for areas covered by grass vegetation and permafrost disrupted in summer. Under cold Pleistocene climates the soil cover was getting thinner, but the soil below grass still contained enough CO_2 to protect the dripstone in the caves from decay. The mentioned forest refuge areas had been in the thermal zone with higher daily minimum temperatures, at present they are to be found between 30 and 200 m above the bottom of basins and valleys.

The cold Pleistocene phases presumably experienced a reduced chemical erosion in the Dinaric karst as a consequence of the 750–1000 mm lower precipitation at the maximum cooling of the last glaciation (Frenzel, Pécsi and Velichko 1992, p. 45) and owing to scarce vegetation. There were better conditions, however, for forming potholes similar to the present state in the barren Alpine karst (*Fig. 6*). Inherited from the Pleistocene with cold climate there are more potholes than horizontal caves in the Slovenian karst, formed mostly by allogenic sinking rivers. As in the cold Pleistocene phases the surface of the semipermeable dolomite, impermeable Eocene flysch and older impermeable sediments was frozen for longer time then at present, a higher amount of water was flowing to the more permeable limestone, depositing in the contact karst caused more river transport to occur (as recently at Velo polje at the altitude 1680 m), forming the alluvial bottom of the uvalas and poljes. During the Pleistocene snow of longer persistence also provided better conditions for doline formation.

As the Holocene has lasted for only about 1/200 part of the Quaternary, the present larger depressions in the Dinaric karst are considered to have formed during the cold Pleistocene epochs, slightly modified by the warm interglacials. Under conditions of reduced precipitation and sparser vegetation cover in the cold Pleistocene epochs chemical erosion also reduced. Still an assumed surface lowering of 20–60 m over the past two million years could result in the emergence of the shallow poljes, uvalas and dolines.

Conclusion

In the Holocene more than half of the total chemical erosion in the barren high Alpine karst of Slovenia has occurred in the vadose zone between the stony surface and the phreatic zone. It has created favourable conditions for the deep pothole formation by an aggressive impact of waters deriving from atmospheric precipitation, melted snow and ice. In the Dinaric karst of Slovenia the solution capability of the water from precipitation percolating subsequently through the vegetation and soil cover, is mostly exhausted in epikarst, that is in the contact area of soil and carbonate rock. In the vadose zone the filling of the voids by flowstone deposition is common. In the Pleistocene cold phases the predominant solution in the vadose zone has expanded also onto the higher Dinaric plateaus where the number of potholes many times exceeds that of horizontal caves. The most decisive factor for karst denudation (downwearing) is the amount of precipitation.

The latter being lower in the cold Pleistocene climates the chemical erosion was less intense, too. Intensity of chemical erosion (30–120 micron/a in Slovenia) calculated from water hardness and run-off in river basins is not in correlation with the density of karst depressions on the surface. The best conditions for local accelerated solution and formation of depressions (dolines, uvalas, poljes) during the Holocene evolved in the transition alpine zone from the rocky slope to karst covered by soil. In the Pleistocene glacial stages similar conditions prevailed in the overwhelming part of the higher Dinaric Karst in Notranjsko and Dolenjsko where there is a dense network of dolines, poljes and uvalas.

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Terrace formation versus slope evolution in the incised Carpathian valleys during the Quaternary

LESZEK STARKEL[1]

Within the fluvial system two independent groups of landforms exist: slopes and valley floors. The complex of denudational processes changes the configuration of slopes (or valley sides) through the delivery of debris and fine material to the valley floors. On the floor the transversal transport is converted into longitudinal one by the flowing water. But these landforms, slopes and valley floors are in the fact closely interconnected. The fossilisation of terraces is an outcome of this interrelationship.

In the tectonically active Carpathians river incision causes a steady lowering of the valley floor, formation of terraces and – by undercutting of valley sides – rejuvenation of the slope surface. This undercutting makes the valley sides steeper and shorter. As a result slope processes, and especially the gravitational ones are reactivated and sediments deriving from the slopes are distributed on the valley floor.

The incision and formation of terrace steps is realised during the glacial – interglacial cycle. Every terrace step is recognised as the product of this cycle (Dziewañski, Starkel 1962, Starkel in print). In the case of most Pleistocene terraces the erosional bottom cut in the bedrock is widened by lateral erosion during the interglacial – glacial transition when the river starts to carry more debris and changes its channel pattern from the meandering type to a braided one *(Fig.1)*.

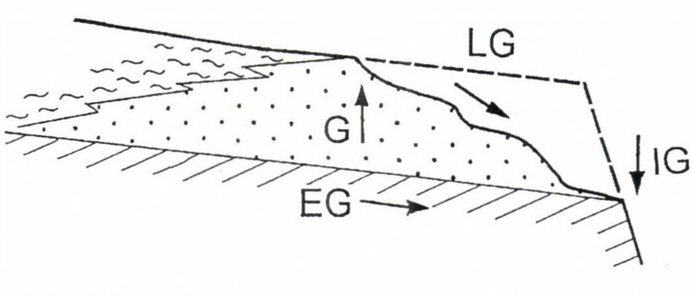

Fig. 1. Simplified model of formation of one terrace step in the Carpathian valley (after Dziewański and Starkel 1962, Starkel 1965). – EG = early glacial; G = glacial (pleniglacial); LG = late glacial; IG = interglacial

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During the pleniglacial the delivery of sediments from the slopes by periglacial processes is massive enough to provoke aggradation. Fluvial and slope sediments are interfingering frequently (Klimaszewski 1958, 1971, Starkel 1960, 1965) and depositional glacis is formed at the base of the slope. The increasing aridisation of climate during the sinking phase of pleniglacial stage causes the turn from aggradation to erosion, still accompanied by active delivery of slope sediments. The following interglacial is expressed in the incision of the river channel into the bedrock and in weakening of slope processes.

The following cold stage reactivates the gravitational and slope wash processes along the whole slope, but mainly in its lower steeper part. These sediments start to be deposited over the abandoned terrace level. Its fossilisation begins with the redeposition of coarser material by solifluction processes and by slope wash below the gradient of 2–4°. A typical example of such a fossilised terrace step was described from the environs of Solina–Zabrodzie in the upper San valley, where the pre-Eemian terrace was fossilised during the last cold stage (Dziewański, Starkel 1967; *Fig. 2*). The catchment of the upper San valley abounds in remarkable examples of terrace fossilisation (Starkel 1965, 1969).

The degree of fossilisation of Pleistocene terraces depends on different factors such as the lithology (rock resistance), total length of slope, extent of underwashing of the lower slope segment and the width of terrace surface. Slope length and gradient control the redeposition of material by gravitational processes. In the highly resistant rocks the fossilisation of terraces is a very rare phenomenon. In contrast, on the less resistant flysch sandstones with interbedded shales it is very common. Narrow terrace benches can be fossilised easily for the slope sediments may reach the lower terrace or the flood plain where they are deposited interfingering with

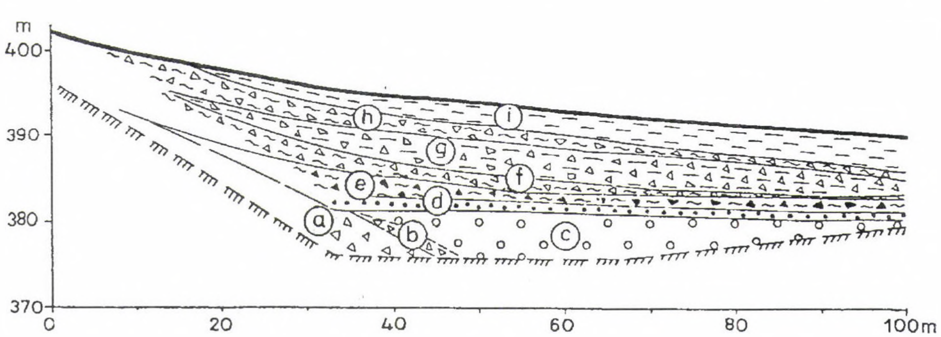

Fig. 2. Compiled section of fluvial and slope deposits fossilising the middle terrace at Solina-Zabrodzie in upper San valley (after Dziewański and Starkel 1967). – Sediments from Middle Polish Glaciation: a = talus debris; b = mixed slope and fluvial deposits; c = fluvial gravels; d = sandy silts (fluvial); e = slope covers (decalcified), sediments from Vistulian Glaciation: f, h = solifluctional members with debris; g = solifluctional-deluvial (slope wash) member with rare occurrence of debris; i = top deluvial (silty, sandy) member

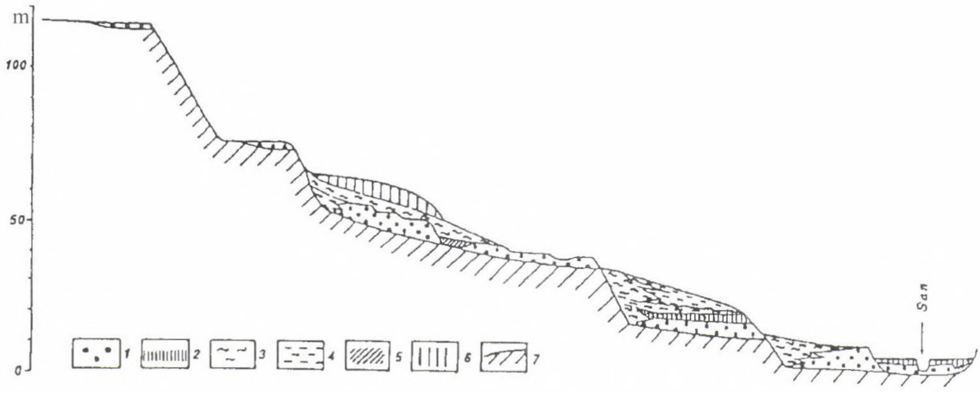

Fig. 3. Quaternary terraces in the Upper San valley (after Dziewański and Starkel 1962, Starkel 1965, Gerlach *et.al.* 1997 with later modifications). A 40–55 m high terrace is correlated with Sanian (Elsterian) glaciation (500–450 ka BP). – 1 = channel facies; 2 = overbank facies; 3 = solifluction deposits; 4 = slope-wash deposits; 5 = paleochannel fill with interglacial flora; 6 = loess-like deposits; 7 = bedrock

fluvial sediments. Wider terraces *(*see *Fig. 2)* facilitate the differentiation of sediments into solifluctional steeper glacis (10–4°) and slope-wash flat glacis (2–4°).

A staircase of terraces is the final product of simultaneous tectonic activity and climatic cyclicity during the Quaternary. On the way to fossilisation it presents a multisegmental profile with several alternating convex and concave sections *(Fig. 3).* This staircase is better expressed if the river displays a long-term tendency to a uniform lateral shift. Therefore the higher (older) terraces are to be fossilised first. Then we may speak about the extension of slope from the top down to the valley floor (Starkel 1987; *Fig. 4a*). In the narrow tributary valleys where only one single terrace step has developed this fossilisation is reflected in a long concave slope segment, which reaches the incised river channel.

In a larger valley with well developed terrace staircase such as the Dunajec valley (Zuchiewicz 1983) the fossilised higher terraces start with time to play the role of the middle slope segments and the aggradation is replaced by degradation *(Fig. 4b).* This causes the lateral cutting of fossilised higher terraces and therefore only the shallow fossilised erosional benches with the remains of terrace gravels may be preserved on the gentle slope surfaces.

Additional factor playing important role in the extension of fossilised Pleistocene terrace is the deposition of allochtonous loess, which accelerates the fossilisation and even leads to formation of convex upper slope sections (Alexandrowicz, Łanczont 1995, Łanczont 1997, Zuchiewicz and Butrym 1990, Gerlach *et al.*1997). This eolian component is also present in the thick sequences of solifluctional–deluvial deposits, which during the last cold stage reached 10–20 m thickness (Sobolewska *et al.* 1964).

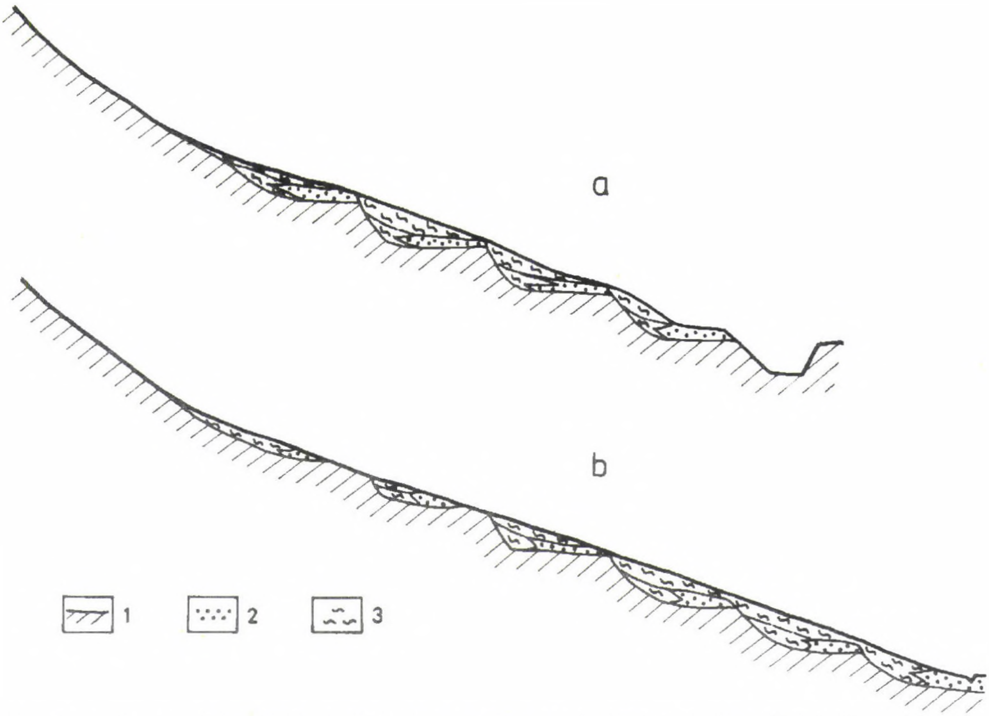

Fig. 4a,b. Stages of fossilisation of Quaternary terraces in the San valley (flysch Carpathians) – a = fossilisation of higher terraces; b = degradation of higher fossilised terraces and fossilisation of lower terraces. – 1 = bedrock; 2 = alluvium; 3 = slope deposits

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MOMENTS OF A LIFE

J. Fink (with a baseball cap on), M. Pécsi's predecessor in the presidency of the Commission on Loess, International Union for Quaternary Research (INQUA) at the key loess section of Paks (above) and participants of the DEUQUA field symposium in the same place (below), autumn 1978

Scientific session opening the symposium of the INQUA Commission on Loess held at the Hungarian Academy of Sciences (Budapest, August 1979). From left: A. Dodonov, N. Opdyke, M. Pécsi, O. Fränzle (speaking), A. Rónai, H.B. Cook, M. Kretzoi (above), and M. Pécsi delivering explanation during the field trip at Szeged (below)

Field symposium of the INQUA Commission on Loess (August 1979). M. Pécsi's explanation at the Mende exposure. Among the participants: A. Bronger, B. Urban and D.H. Yaalon (above) and a general picture of the audience (below)

Confer with Liu Tungsheng and M. Pevzner at the high bluff of the Danube, (Dunaföldvár, autumn 1985)

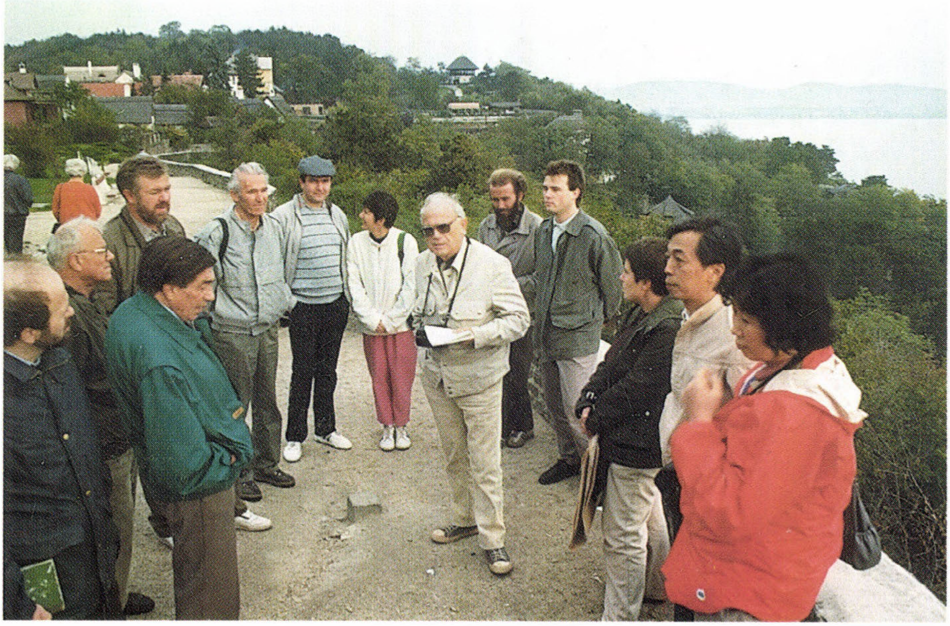
Field symposium to study the Paks key loess section (September, 1986). Among the foreign participants: A. Mójski, B. Grabowska-Olszewska, T.D. Morozova, B. Frenzel, A. Velichko, A. Markova, I. Spasskaya. Hungarian experts: L. Gerei, S. Marosi, P. Csorba

At the Vértesszőlős site of Early Man (VII. Polish-Hungarian Geographical Seminar, October 1986). Explanation at the pavilion (above) and on terrace morphology (below). From left: L. Starkel, F. Schweitzer, O. Morozova, A. Kotarba, T. Kalicki, E. Krolopp

With K. Brunnäcker and F. Schweitzer during a post-congress (Moscow, 1982) field symposium in Central Asia

Studying the Gyöngyösvisonta pit with Quaternary experts from Siberia (S. A. Arkhipov and V. S. Zykina) and F. Schweitzer

Meeting of the editorial board and scientific symposium on the Atlas of Paleoclimates and Paleoenvironments of the Northern Hemisphere (Late Pleistocene–Holocene). From left: M. Pécsi, I. Spasskaya, B. Frenzel, A. Velichko (lecturing)



A group of international experts on geomorphology at Tihany on Lake Balaton, Hungary (among the participants: H. Mensching, I. Gams and An Zhisheng) in the early 1990's.

M. Pécsi's with colleagues and disciples at the 15th INQUA Congress, Durban, South Africa in 1999

Three generations meet. M. Pécsi surrounded by Hungarian geomorphologists of middle age (Gy. Hahn, M. Balogh-di Gleria, A. Nemerkényi, Á. Kertész) and youngsters (autumn 1999)