PALEOGEOGRAPHY AND LOESS

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PLEISTOCENE CLIMATIC AND ENVIRONMENTAL RECONSTRUCTIONS

AKADÉMIAI KIADÓ · BUDAPEST

PALEOGEOGRAPHY AND LOESS

(Studies in Geography in Hungary, 21.)

Edited by
M. PÉCSI and A. A. VELICHKO

The INQUA Commission on Loess and Commision on Paleogeographic Atlas discussed and fixed the contents of the Paleogeographic Atlas of the Northern Hemisphere: a series of maps registering global paleoenvironmental changes during the Upper Pieistocene. The lectures presented at the joint session of the two commissions are published in this volume. Papers are mostly concerned with Late Quaternary environmental changes and climates in Europe relying on the analyses of loesses, paleosols, moraines of the succession of biogenic phenomena and of fossil animal and plant finds and, last but not least, of climatic and rellef changes.

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Geographical Research Institute
Hungarian Academy of Sciences

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PREFACE

The reconstruction of the changes of the physical, human and transformed environment during the youngest geological period, the Quaternary, falls within the scope of numerous disciplines. Quaternary research attempts to follow in a retrospective way the evolution of life and Earth history during the last two million years with special regard to the last hundred thousand years. One of the goals of this study is to determine the impact of the cyclically changing climate and paleoenvironments of the Quaternary on the present environmental conditions. The paleogeographic indications of global climatic and environmental changes can be especially well traced in glacial forms, terrestrial sediments, particularly in loess and paleosol series. In various loess layers the traces of fossil plants, animals and early man with his tools as well as physical processes, for instance, remnants of landforms developed under permafrost conditions, have also been preserved. At the same time, based on the recognition of the genetic types of paleosols intercalated between loess layers the warmer and more humid ecological conditions similar to but not entirely identical with the present-day environment, can be reconstructed.

Thus, the identification of changes in a loess series and the results of loess stratigraphy are closely related to the interpretation of Quaternary events and changes in the paleoenvironment. The age and nature of a characteristic paleoenvironment can be reconstructed through the interdisciplinary investigations of paleosols intercalated in loess and of the enclosed paleontological and archeological finds.

With the aid of these methods, in the joint meeting held at Budapest, two committees of the INQUA (the Commission on Loess and Commission on Paleogeographic Atlas) discussed and fixed the contents of an Atlas of Paleogeography, a series of maps registering global paleoenvironmental changes in the Northern Hemisphere during the Upper Pleistocene.

The lectures presented at the joint session of the two commissions are published in a volume entitled 'Paleogeography and Loess' in the English-language series of 'Studies in Geography in Hungary' edited in the Geographical Research Institute of the Hungarian Academy of Sciences. Papers are mostly concerned with Late Quaternary environmental changes and climates in Europe relying on the analyses of loesses, paleosols, moraines, of the succession of biogenic phenomena and of fossil animals, plants and paleoarchaeological finds, and last but not least, of climatic and relief changes.

The publication is dedicated to the Congress of the International Union for Quaternary Research to be held in Ottawa, in order to demonstrate the close corporation and interdependence in subject and methods between loess and paleogeography studies. We believe the papers on the Late Quaternary paleoenvironments to be useful contributions to the UNESCO Project 'Global Change' as well as to the interpretation of the maps in the Atlas of Paleogeography.

Budapest, May 1987

Márton Pécsi

and

Andrey Velichko

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RELATIONSHIP OF CLIMATIC CHANGES IN HIGH AND LOW LATITUDES OF THE EARTH DURING THE LATE PLEISTOCENE AND HOLOCENE

A.A. VELICHKO

ABSTRACT

Classical Quaternary paleogeography distinguishes the following global latitudinal belts: glacial, periglacial and arid-pluvial. Chronological correlation of natural events within these belts is an indispensable precondition to global climatic reconstructions important for the general conception of climatic changes.

At present, research workers of the Department of Paleogeography Institute of Geography Academy of Sciences USSR have compiled a series of maps, representing reconstructions of paleoclimates on the basis of special paleogeographic quantitative methods. The series includes temperature maps for July, January, mean annual temperature and that of total annual precipitation for the extratropical northern hemisphere; the maps refer to the last interglacial optimum Mikulino (Eem) isotopic stage 5e, 30,000-120,000 B.P.) and the last glacial maximum (isotopic stage 2, 19,000-3000 B.P.). The quantitative paleoclimatic data indicate an unambiguous maximum amplitude of thermal fluctuations suggested for high latitudes.

During the last interglacial optimum (stage 5e) arctic seasonal temperatures were about 10 °C higher than the present day values. Within subtropical and tropical zones the anomalies were small (about 1 to 2 °C) but an alternation of belts of weakly positive and negative anomalies occurred. A narrow strip with positive anomalies was situated at the equator. Annual precipitation increased in most of extratropical Eurasia in the last interglacial maximum.

During the last glacial maximum (stage 2) the temperatures decreased not only in high latitudes but also in the medium ones. At this time the lower latitudes experiences aridization.

Solution of a fundamental problem is necessary to understand the causes of global climatic changes: whether the phases of climatic parameters (temperatures and precipitation) coincide in the northern and southern hemispheres. Global longitudinal paleogeographic profiles argue for a synchronous character of changes.

*

INTRODUCTION

Global paleoclimatic reconstructions permit us to analyze fluctuations in temperatures and precipitation which occurred in various natural terrestrial zones in the past. These data indicate drastic changes in the general circulation of the earth's atmosphere. The efficacy of this approach has been demonstrated by environmental reconstructions of the last glacial maximum. Soviet researchers have also obtained similar data for the last interglacial optimum - i.e. for a period of high-temperatures similar to those expected to occur at the beginning of the next century given the technogenic growth of CO_2 contents in the atmosphere.

The availability of these data allows to compare paleoclimates of the high and low latitudes extant during the two counterphases of a climatic macrocycle.

EEM INTERGLACIAL AND HOLOCENE

The last interglacial optimum (Eem, Mikulino, or Riss-Würm) is a basic chronological segment which occurred about 130,000 years ago. It corresponds to the M-6 zone in palynological diagrams of continental deposits and to stage 5e of the deep-sea core record. We derived information on terrestrial climates by using paleobotanical methods for estimating past temperatures and precipitation (climagrams and areagrams) developed by SHAFER, V. and IVERSEN, J. and elaborated by GRICHUK, V.P. (1973, 1982). Temperature estimates obtained by these methods are accurate to ± 1 °C and estimates for annual precipitation to ± 50 mm. GRICHUK, V.P., GURTOVAYA, E.E., and ZELIKSON, E.M. analyzed the published pollen data from 50 sites which had the necessary species definitions. The sites satisfying these requirements were primarily located in Europe. There were considerably fewer of them in extratropical Asia and only three in North America.

BARASH, M.S., BLUM, N.S., and NIKOLAEV, V.I. derived surface water temperatures of the Atlantic and the Pacific from extant literature. Their work took into account the distribution of planktonic Foraminifera groups in the deep-sea cores (BARASH, M.S. 1974), as well as paleotemperatures reconstructed on radiolaria, coccoliths, and oxygen-isotope data. Appropriate tables (STROKINA, 1982) were then used to transform these water temperatures into air temperatures.

Data on hand point out two main features of the Mikulino (Eem) interglacial climate (VELICHKO, A.A. et al. 1982, 1984). The climate of the interglacial optimum was, at least in the eastern hemisphere, warmer and more humid than today. There was also a considerable latitudinal uniformity in temperatures and in precipitation. Both deviated from mean latitudinal continental values much less than they do today.

Mean annual temperatures as well as those for July and January months, together with estimates for the amount of annual precipitation were reconstructed and mapped for extratropical areas of the northern hemisphere for the Eem interglacial optimum. Air temperatures over low latitude oceans were also reconstructed. This permitted us to interpolate air temperatures over adjacent continental regions as well. Furthermore, we were able to reconstruct the mean temperatures of the near surface air layer in the northern hemisphere through special calculations done on the latitudinal belts using grids with angles of 5° latutide-longitude. Their value was 1,7°-2° C warmer than today.

There is little reliable information on the low latitudes in the eastern hemisphere during the last interglacial optimum. Studies done at the Shati lake in Lybia (PETIT-MAIRE, M. et al. 1982), for example, indicate that maximal water levels were reached 130,000 years ago (dated by the Th/U method). This corresponds to the time of the interglacial optimum. On the whole, the pluvial in Africa north of the equator is comparable in time to the Eem Interglacial. Direct quantitative values for annual precipitation, to date, have been obtained only for Europe northern Africa, and for extratropical Asia *(Fig. 1)*.

The existence of extensive chronometric data permits us to securely outline the changes in climatic events during the late glacial period and the Holocene, as well as to correlate changes observed at different latitudes. This can be done by comparing data received on the fluctuations in temperatures and precipitation in high and middle latitudes during the last 12,000 years (KLIMANOV, V.A.--ELINA, G.A. 1984; KLIMANOV, V.A.-BEZUSKO, V.G. 1981) with those on arid and pluvial phases in low latutides (based on research in the Sahara and in equatorial Africa) (BUTZER, K.W. et al., 1972; WILLIAMS, M. 1977; PETIT-MAIRE, M. et al., 1982; STREET, F.A.--GROVE, A.T. 1976, et al.).

This comparison is further justified because the studied regions lie in a single meridional belt of about 30° E. L., and thus form a paleogeographical profile. The results of this comparison are as follows *(Fig. 2, Table 1)*.

The development of the two warm phases in high latitudes is quasi synchronous with that of the two humid phases in low latitudes. The correlation between them is not direct however. In the tropical belt the main phase of humid climate occurred about 9,000 to 8,000 years ago. In the high and temperate latitudes that was the time of the first rather well pronounced, but not the principal, rise in temperatures. The humid phase in the low latitudes which corresponds to the time of the highest temperatures in the high latitudes (the so-called thermal optimum, about 5,500 to 5,000 years ago) was not so pronounced.

Since the Holocene optimum was intermediate in heat supply between the Eem Interglacial and the climate of today, we can suggest that the mean global temperature during the main Holocene optimum was 1 °C higher than today. Paleoclimatic evaluations (HANSEN, I. et al., 1981) suggest the same conclusion.

If we combine the data sets on the interlatitudinal changes in the availability of heat and of moisture during the Eem and the Holocene, we can note the following *(Fig. 3)*.

The high latitudes (from 60° to 65°) featured the greatest rise in temperatures. In some places on continents temperatures in the summer were 6° to 8° and in the winter 10° to 12° higher than today. The warming, however, was not worldwide and depended on local features of atmospheric circulation.

Table 1 Phases of fluctuations in heat and moisture supply in the eastern hemisphere during the Holocene

	High and low latitudes	Low latitudes
Late glacial, 12,000-10,000 years ago	The growth of heat supply	Beginning of humidifying
Early Holocene thermal optimum, 9,500-5,000 (8,000) years ago	During some phases (8,500; 7,800 years ago) heat supply higher than at present	The main phase of humidity
Insignificant falls in temperature within the Middle Holocene, 7,500 (8,000)-6,000 years ago	Heat supply drops to the level close to the present-day one	Humidity somewhat reduced
The main climatic optimum, Middle Holocene, 6,000-4,000 (3,000) years ago	Heat supply is the highest	High humidity
Late Holocene, 3,000 (3,500) years ago to the present; regular fall in temperatures	Heat supply decreases, approaching, with fluctuations, the present level	Aridization

In the middle latitudes (up to 45°-50°) there was also a rise of temperature. Here, however, the rise was not so significant. The value of positive temperature deviations here gradually, north to south, dropped to zero (compared to present-day ones).

In the low latitudes temperatures decreased only slightly. At the same time, it seems probable that local increases in temperatures (compared to present-day ones) may have occurred during the time of general cooling. These rises were quite insignificant (not more than 1° to 2° C). This conclusion is suggested by temperature data from the Atlantic which indicate that one such belt of slightly increased temperatures was located at 20°-15° N. A second one was positioned near the equator.

The 2° mean annual rise in temperatures which took place during the Eem (Mikulino) was accompanied in all extratropical areas of Eurasia by an increase in precipitation. This increase in all areas (including in high and middle latitudes) was up to 50 to 70 per cent. The increase (in absolute values) was considerably greater in the west than in the east of Eurasia.

A considerable increase in annual precipitation (up to 100 per cent) was typical for the zone located south of the 45°-50° latutides. This area today features a deficit of moisture.

During the Holocene optimum mean annual temperature rose by 1° C (i.e. an increase half as large as during the Mikulino optimum) in the high and, partially, in the temperature lat-

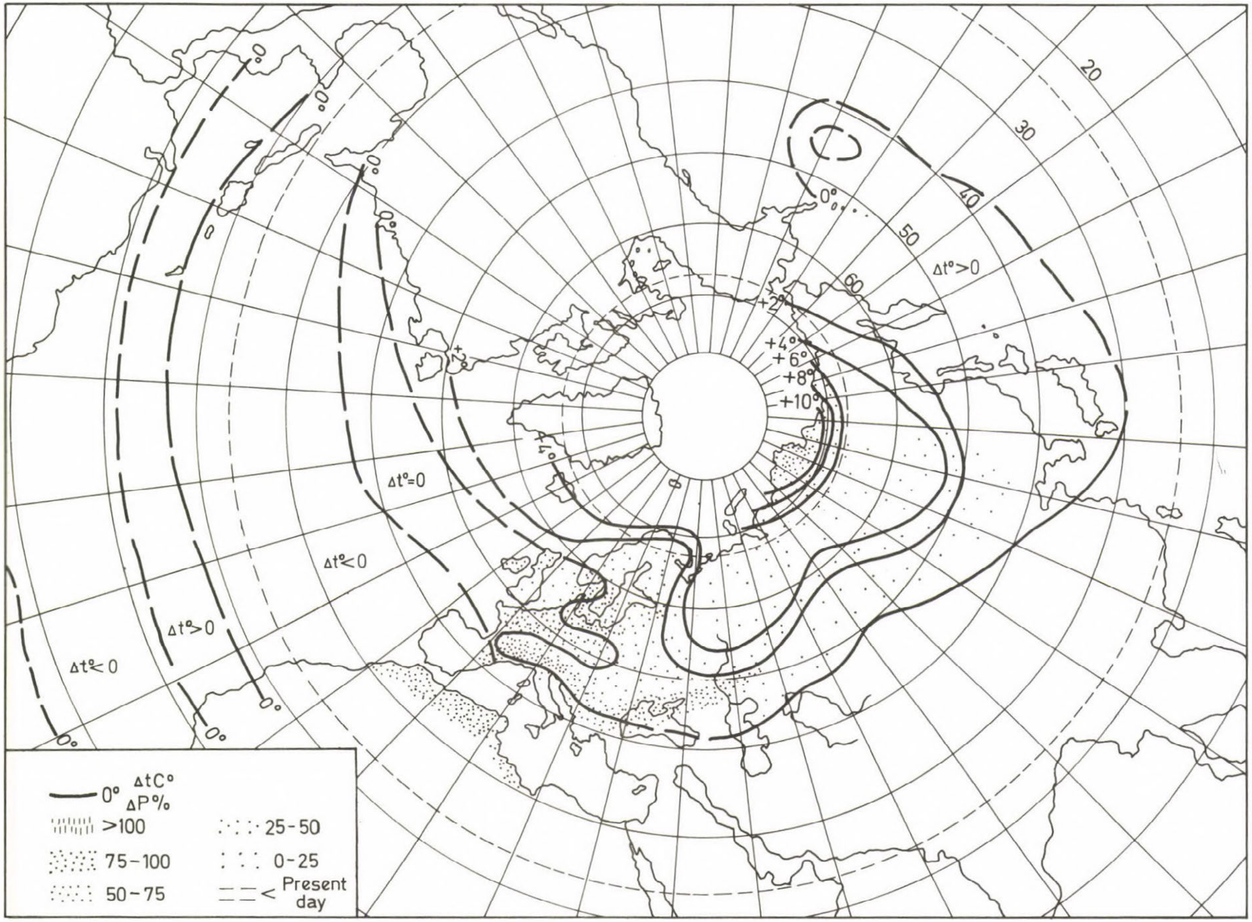

Fig. 1a. Deviations in mean annual temperatures (Δ t° C) and precipitation(Δ P, mm) from present-day values in the Eem (Mikulino) interglacial optimum

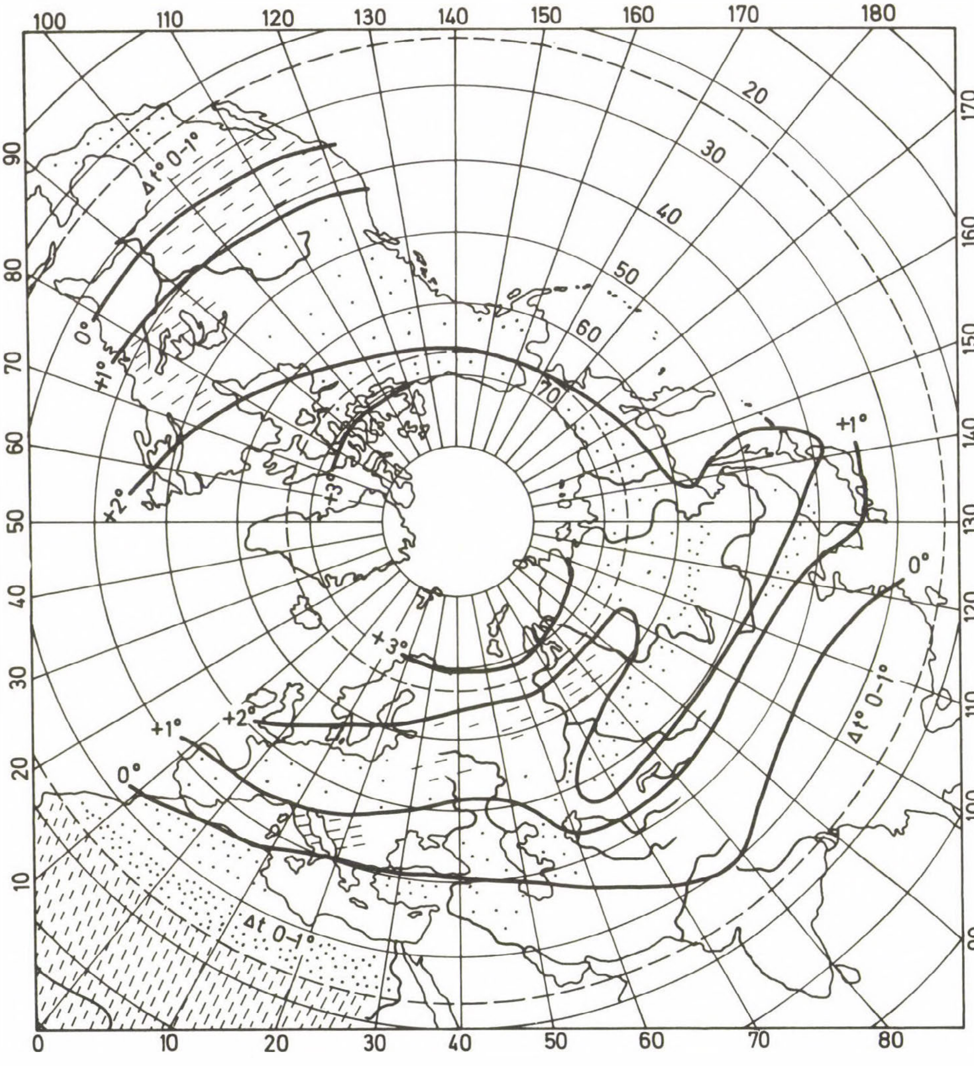

Fig. 1b. Deviations in mean annual temperatures (Δ t° C) and precipitation (Δ P, mm) from present-day values in the Holocene optimum

itudes. Although precipiation was also somewhat higher, this increase was noticeably less than during the Mikulino optimum, and did not exceed 100 or 150 mm per year. Some sites in the south of the forest zone indicate that precipitation not only did not increase, but was 25 to 50 mm per year lower than at present (KLIMANOV, V.A.--ELINA, G.A. 1984; KHOTINSKY, N.A.-SAVINA, S.S. 1985). At the same time, the extratropical low latitudes (including areas with semi-arid and extra-arid climates today) witnessed a sharp increase in humidity during the main warm phase of the Holocene. Judging from lake sediments data from Africa and western India, the most pronounced increase in precipitation (more than 300 mm per year) occurred about 9,000 to 8,000 years ago (KUTZBACH, J.E. 1981).

It is not difficult to notice that, on the whole, the intertropical areas with increased precipitation during the Holocene were those which experienced small decrease in temperatures during the Eem optimum. It is worth noting that at least the northern part of the same zone of decreased temperatures (the subtropics) featured a considerable increase in precipitation during the Eem interglacial optimum. In still lower latitudes, this time is comparable to the pluvial.

The above observations permit us to draw the following conclusions. On continents, a considerable global increase in temperatures (a mean annual rise of not less than 2°) was accompained by a significant global rise in precipitation. In the high and middle latitudes, an increase in precipitation was accompained by a rise in temperatures. In the lower latitudes this increase was associated with localized small decreases in temperatures (compared to present-day ones). This decrease in temperatures occurred as a result of increased evaporation in more humid climates, rather than as a result of changes in the input of heat. Under conditions of lower rank global warming (global temperature rise about 1 °C), the increase in precipitation was not as pronounced in the high and middle latitudes of the eastern hemisphere. There, in some cases, precipitation even decreased locally.

The correlation between heat and humidity was essentially different on the continents of the western hemispheres. Here during the Holocene thermal optimum, annual precipitation rates increased considerably (locally by more than 200 mm) in most parts of the temperature and subtropical zones (WEBB, T.--CUSHING, E.J.--WRIGHT, H.E. 1983).

THE EPOCH OF THE LAST GLACIAL MAXIMUM

The relationship between climatic events in high and low latitudes extant during the last glacial maximum (20,000 to 17,000 years ago) has been discussed in great detail in the literature, and several global maps published which depict the state of the natural components at that time (McINTRYRE, A. et al. 1976; BOWEN, D.D. 1979).

This period saw a fall in temperatures which resulted in the expansion of glaciers, ice sheets, and of permafrost on

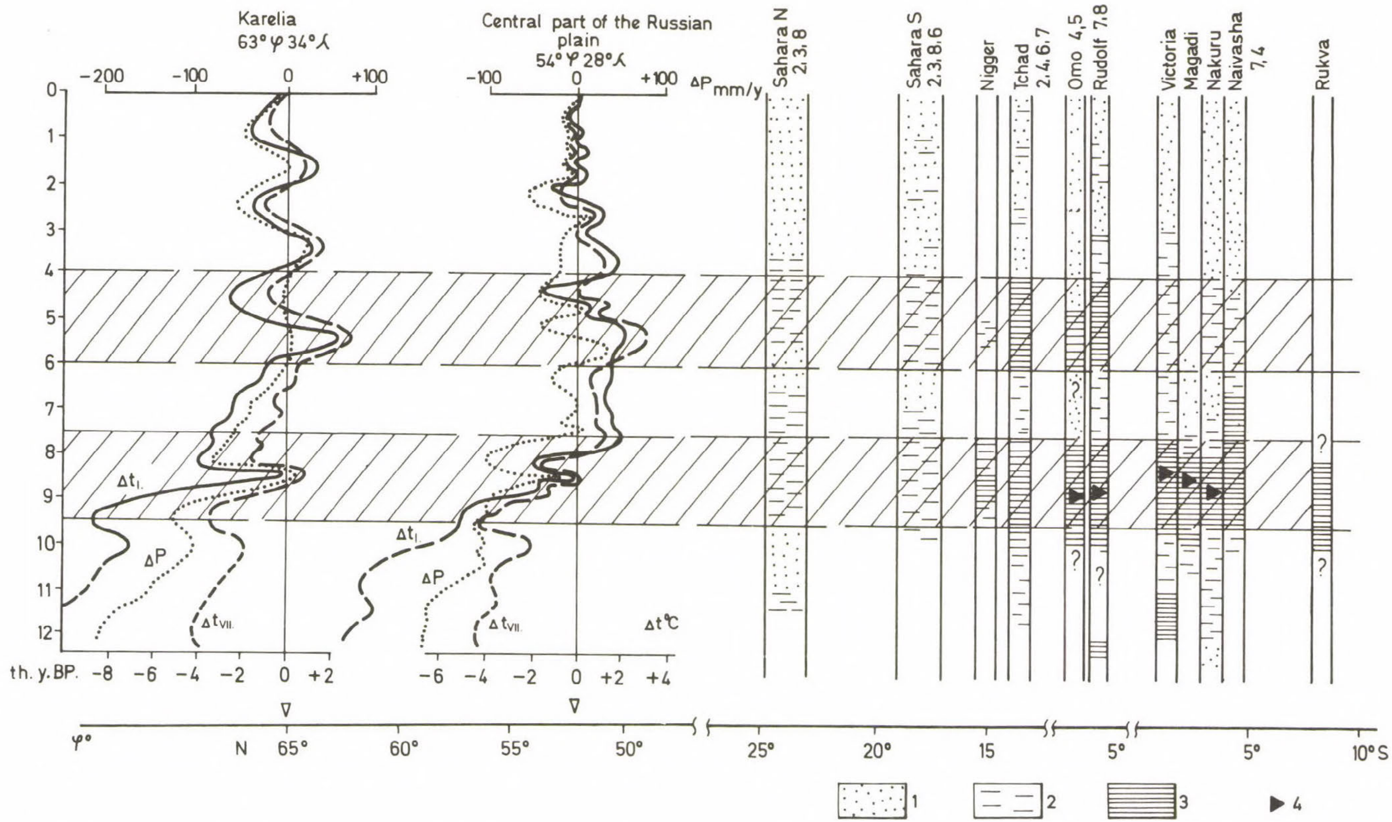

Fig. 2 Correlation of heat and moisture supply in high and low latitudes during the Holocene (after KLIMANOV, V.A.--BEZUSKO, V.G. 1981; KLIMANOV, V.A.--ELINA, G.A. 1984; BUTZER, K.W.--ISAAC, G.L. et al. 1972; FAURE, H.-WILLIAMS, M. 1977; GOUDIE, A. 1977; HEINE, K. 1978; PETIT-MAIRE, M. et al. 1980, and the author's data). Lake basins stands: 1 = low; 2 = middle; 3 = high, 4 = state of overflow

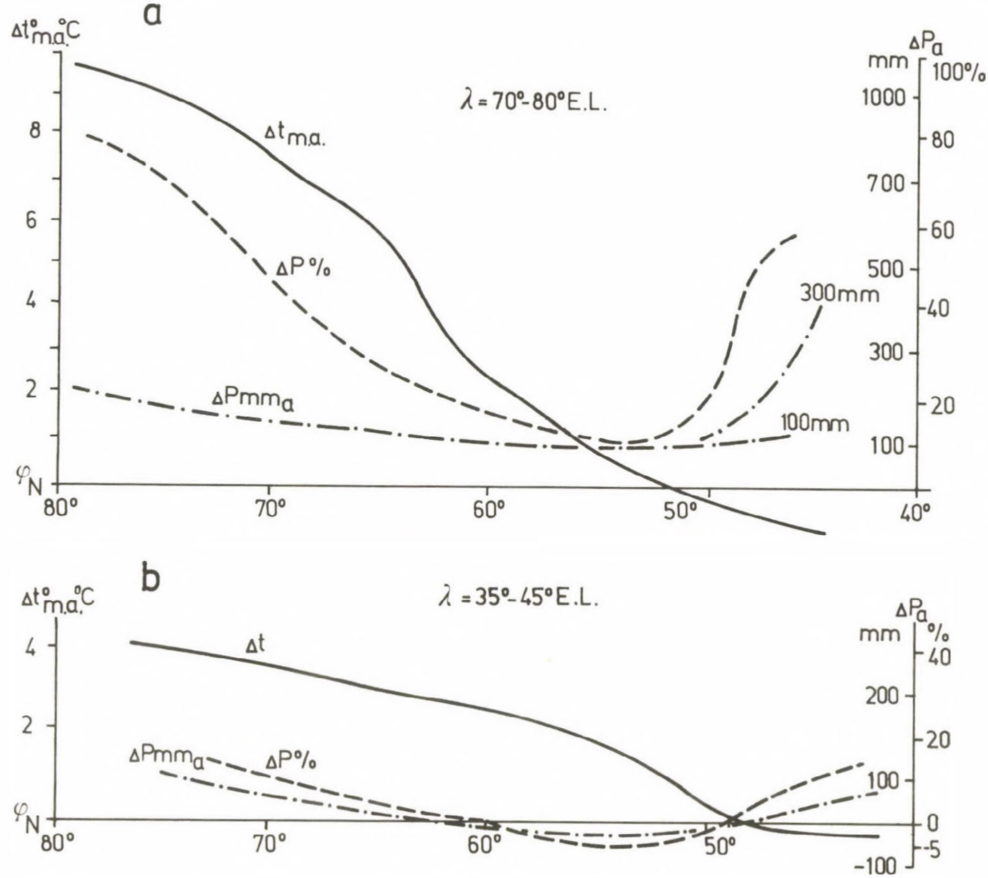

Fig. 3 Meridional profile of deviations from present-day values of mean annual temperatures ($\Delta t°$ C) and precipitation (ΔP mm year): a) during the Mikulino optimum in Western Siberia and in Kazakhstan by 70°-80° E; b) during the Holocene on the Russian Plain by 35-45° E

the continents, as well as an expansion of the sea ice in the direction of the equator (up to 35° or 40° N.L. in the northern hemisphere). A sharp global decrease in precipitation was an equally important second feature of this glacial maximum *(Fig. 4)*. This decrease, which occurred during the maximum of the last glacial epoch, resulted in the destruction of the zonal components not only in the area of the tropical forests but also in the bipolar forest zones of the northern and southern temperate belts. The latter two are global indicators of latitudinal belts of stable humidity. The situation, in reality, was even more complex. The woodland territories in temperate latitudes of North America differed considerably from this arid scenario. We have considered the reasons for this elsewhere (VELICHKO, A.A. 1980). Increases in humidity were also recorded in some regions of South America and of Africa - in South Kalahari, in particular (HEINE, K. 1980). The existence of these wetter regions, however, does not warrant replacing our current concept of a predominantly arid hyperzones with one which emphasizes migrating belts of humidity in both hemispheres. During the last glacial maximum the following three natural belts formed in the eastern hemisphere under the impact of decreased temperatures and increased aridity: the glacial one (in high and middle latitudes), the cryoarid one (in middle and high latitudes). The zonal structure of the earth simplified considerably at this time and was replaced by a hyperzone (VELICHKO, A.A. 1973).

Data on hand permit us to evaluate both the overall climate as well as inter-latitudinal changes in climate extant during the last glacial maximum as follows.

The global level of heat supply was much lower than at present. Some researchers (HANSEN, I. et al. 1981), estimate the mean global fall in temperatures during the period of the greatest cooling as 5,3° for the continental regions and 2.3° for ocean surfaces. Global cooling was about 3° C and primarily resulted from the fall in temperatures during the cold seasons.

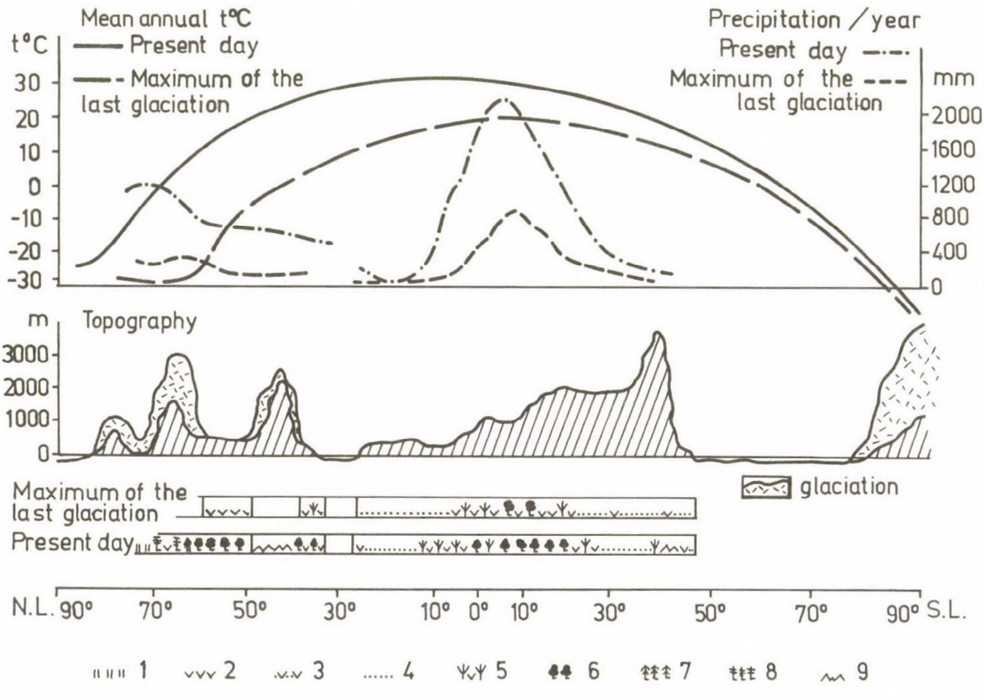

Fig. 4 Meridional landscape-climatic profile of the earth during the last glacial maximum

Vegetation: 1 = tundra; 2 = steppe; 3 = semi-desert; 4 = desert; 5 = broad-leaved forest; 7 = coniferous forest; 8 = deciduous small-leaved forest; 9 = alpine vegetation

In the northern hemisphere, the greatest fall in temperatures occurred in high latitudes (over 60°) in the glacial and north of the cryoarid belts (VELICHKO, A.A. 1980). Mean annual temperatures over the Scandinavian ice sheet dome approached -35°. At present, the mean annual temperature here is close to 0°-2°. At the same latitudes, temperatures in permafrost regions were somewhat higher than those over the glacier: -25° to -27° (today they are -16°). In the periglacial belt there was also a considerable fall in mean annual temperatures. Values close to the margin of the Scandinavian glacier approached -10° to -12° (at present they are +6°). They were 14° to 18° lower near the southern margin of the permafrost area (i.e. in the latitudes 50° to 45° N.L.).

Although sparse, the available data (in particular on South America, Australia, and South Africa) on the mean annual temperatures in lower latitudes (20°-30°) indicate a drop of 8° to 10° (BOWLER, J.M. 1976; HAGEDORN, J. 1982 and others).

Finally, a fall in mean annual temperatures also occurred within the equatorial belt. In Columbia, for example, they fell by 6° (VAN GEEL, B.--VAN DER HAMMEN, T. 1973). In this belt, a fall in temperatures is also observed in some oceanic regions. For example, the equatorial water temperatures in the eastern half of the Atlantic ocean dropped by 2° (McINTYRE, A. et al. 1976).

The decrease in the amount of precipitation was also global. The sum of annual precipitation over glaciers in high latitudes (the Scandinavian ice sheet, for example) decreased by at least 55% to 60%. These values are derived without taking into account the increased amount of precipitation in mountains today. In the cryoglacial belt (as estimated for Eastern Europe) in high and middle latitudes, precipitation diminished by at least 40%. Precipitation also decreased in intertropical areas. This occurred later in regions where deserts prevail today and in areas with present-day monsoon climates.

The strongest aridization took place in the equatorial belt. Annual precipitation there decreased by 60% to 75% and tropical forests were replaced by savannas (VAN DER HAMMEN, T. 1974). In general then, comparisons of inter-latitudinal data indicate a uniform but lowered level of precipitation.

These general climatic features of the glacial maximum were expressed differently in the two hemispheres. In the western hemisphere: *1)* aridization was weaker, *2)* glaciers were more extended (aside from the Wisconsin ice sheet), *3)* some elements of forests persisted in the periglacial zone, and according to some *4)* pluvial lakes in the American southwest reached their highest stands.

COMPARISON OF INTER-LATITUDINAL CHANGES IN CLIMATE DURING THE LAST INTERGLACIAL OPTIMUM AND THE LAST GLACIAL MAXIMUM

Taking the Late Pleistocene macrocycle as an example, we can compare climatic conditions of the two opposite phases *(Table 2)*. These phases had the following characteristics in comparison to conditions today *(Fig. 3)*.

Table 2 Comparison of climatic events in different latitudes

Latitudes	Interglacial optimum (stage 5e)		Last glacial maximum (stage 2)	
	January temperatures	amount of annual precipitation	January temperatures	amount of annual precipitation
High	sharp rise (> 10° C)	increase in some places 50%	sharp fall (> 20° C)	decrease (> 50%)
Middle	rise (< 10° C, in the south- about 0°)	increase (in the south up to 100%)	fall (> 10 °C up to 20° C)	decrease in some places
Low	alternation of belts of weak positive and negative anomalies	increase (in present desert areas 100%)	weak fall (from 1° to 3°)	considerable decrease (in some places, where now forests are common, over 50% to 75%)

Temperature deviations were much more intensive during the glacial epoch than during the interglacial one. Based on mean global air temperature values, we can suggest that the earth's surface cooled off much more intensively during the glacial maximum than it warmed up during the interglacial optimum (in comparison to temperatures today). Meridional profiles of both temperature and precipitation deviations were asymmetrical.

The greatest change in temperatures during both phases (the glacial and interglacial ones) took place in the high latitudes. These regions were responsible for the fact that glacial cooling was far more pronounced than was interglacial warming.

At the same time it should be underscored that deviations from this pattern (namely, temperature rise during the interglacial and fall during the glacial periods) existed even at the equator.

Patterns of change in annual precipitation in high and partly in middle latitudes, during both glacial and interglacial epochs were similar to those observed for temperatures. Their trend during the last interglacial optimum, however, in the latitude 45° to 50° N.L. was ambiguous. During the interglacial a slight fall in temperatures (compared to present day values) was accompained by a sharp rise in precipitation. This increase was probably most acute in latitudes up to 10°-15°. This area also experienced increasing aridization during the glacial epoch. The greatest relative decrease in precipitation, however, occurred in the equatorial belt.

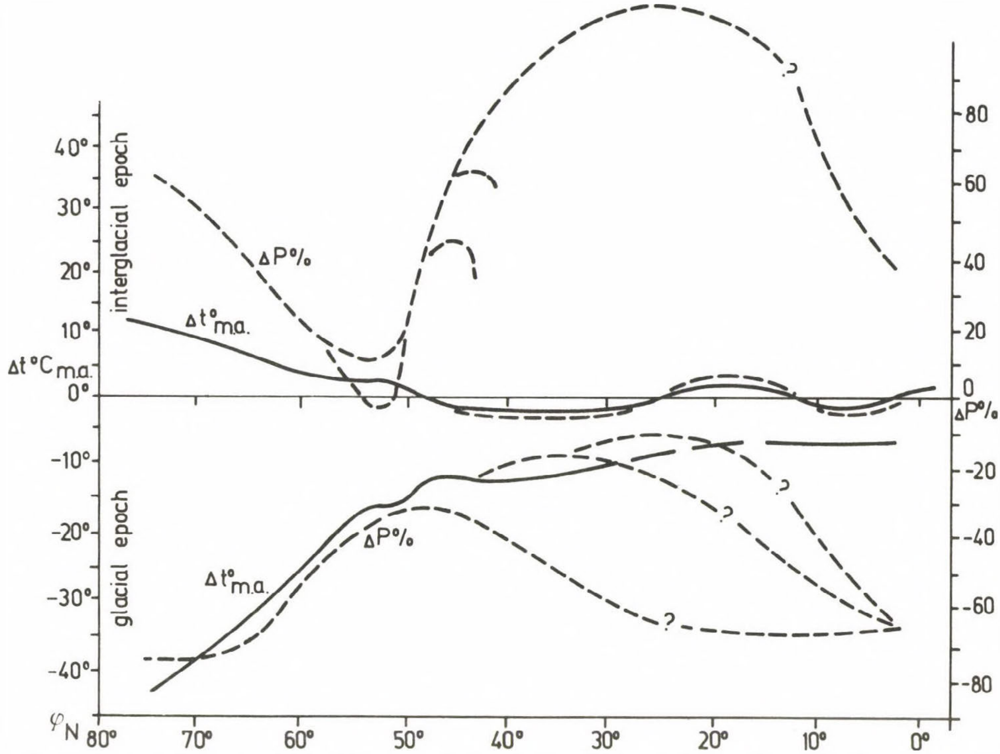

Fig. 5 Generalized scheme of changes in climatic parameters in high and low latitudes at the time of interglacial optimum and the glacial maximum

(Δ t m.a. and Δ P - deviations of mean annual temperatures and precipitation from the present-day values)

The climatic changes discussed above reflected changes in both the general circulation of the earth's atmosphere, as well as changes in the earth's heat balance. While numerous studies exist on general air circulation during the glacial period, fewer deal with atmospheric circulation during the interglacial optimum - a question considered in earlier articles (VELICHKO, A.A. 1980).

Here we will deal with only some ideas pertaining to general circulation. In extra-tropical regions, the interglacial optimum saw a considerable increase in the transfer of moisture from Atlantic to the continents (i.e. from west to east). In high latitudes, this process coincided with the intensification of the North Atlantic warm stream. The Gulf stream penetrated far to the east. This resulted in the warming of the whole Eurasian Arctic region (especially in its Asiatic part). At the same time, there was a decrease in the area affected by the North Asiatic anticyclone. This was especially true during the winter months. This decrease brought about a more frequent penetration of the area by western cyclones, and resulted in the rise of both temperatures and of precipitation in the middle latitudes, especially in inter-continental regions.

This increased influence of the western transfer was, however, relative rather than absolute. Its importance became more obvious because of the reduction in the meridional circulation pattern.

It is important to consider more closely the factors responsible for the more stable manifestation of the western transfer extant during the interglacial optimum (i.e. under conditions of higher global thermal regime). This phenomenon, which is reliably documented by the available data, contradicts climatological theory which holds that the western transfer should be weakened when high latitudes undergo warming and the thermal gradient (pole - equator) is decreased.

One possible explanation for this may be the delay in the response of atmospheric and oceanic circulation systems to the warming. The latter recreated high thermal regimes near the equator as nearly as the late glacial, while glacial systems in high latitudes dissipated more slowly and conserved low temperatures longer. McINTYRE, RUDDIMAN, MANGERUD, and VELICHKO in a series of publications (e.g. VELICHKO, A.A. 1980) have suggested that this asymmetry can account for some phenomena which occurred between the warm oceans and cold land masses. Holocene research indicates that the same low temperatures existed in the high latitudes as late as the beginning of the interglacial. Thus, the equator-pole temperature gradient was higher at that time than during any other period of the whole climatic macrocycle (glacial-interglacial age). This brought about a more active western transfer and an acceleration of the warm Gulf stream. Due to the inertia of the system, however, the stream reached its maximum somewhat later than the peak of the temperature gradient. This occurred during the first part of the interglacial, and soon decreased due to the absence of "feeding mechanism". Thus, it is not coincidental that oceanological data indicate that the Eem warming was extremely short-lived in the equatorial Atlantic.

These observations indicate that the thermal "circulation" optimum must have occurred close to the beginning of the interglacial - a conclusion supported by Holocene data as well. They also explain the existence of an older thermal optimum (dated between 9,000 and 7,000 years ago) which featured high humidity. This humidity was a result of active latitudinal circulation present everywhere, including in such arid regions as the Sahara, and the Middle East and others. This optimum may also correspond to the short global phase when the thermal regime was quasi-stable. The meridional gradient was lessened by then - which, probably, accounts for the observed lack of increase in precipitation in the middle latitudes.

If the above arguments are correct, then we can assume that the same double optimum existed during the Mikulino (Eem) Interglacial as well. Data from the Barbados and other regions which document several phases in the rise of ocean levels, as well as pollen information from the Grand Pile sections studied by G. WOILLARD-ROCOUR, are in accord with such an assumption.

The fundamental differences in the mechanisms which produced the two thermal optima and the resulting differences in the precipitation regimes are very significant and should be considered when using paleoclimatic data to model future climatic scenarios of man-induced warming. Specific attention should be paid to past worldwide increases in precipitation. It is possible that these increases were partly due to the "circulation optima". Since man-induced warming cannot produce this effect, precipitation will increase less in the future than it did in the past. Furthermore, this increase in precipitation will be mostly due to an increase in evaporation in the ice-free oceans and the subsequent transfer of this moisture to the continents. This pattern has been already confirmed by specialists on modern precipitation regimes (M. I. BUDYKO, O.A. DROZDOV, etc.).

At the same time, data on hand indicate that surface water temperatures near the equator were higher during the interglaical optimum than recorded at present. This suggests that higher thermal states of the troposphere existed at those latitudes and resulted in a considerable lowering of the tropopause. This would have favoured *1)* the preservation of the pole-equator gradient, *2)* the western transfer during the climatic optimum, and *3)* the transportation of moisture from the Atlantic to interior Eurasia.

The period of maximal cooling during the glacial epoch saw a general weakening in the centers of activity of atmospheric circulation. This feature has been extensively considered in the literature (FAIRBRIDGE, R. 1972; GATER, L. 1976, etc.). Anticyclonic systems predominated in atmospheric circulation of regions outside of the tropics. In the south they merged with the belt of the high pressure. During the late glacial maximum the sea within the present-day areas of low pressure (in particular the Icelandic minimum) became covered by ice so that activity of the atmospheric centres decreased and consequently intensity of inter-latitudinal exchange reduced. The joint action of these phenomena created states prerequisite for a general reduction in precipitation over the continents. This trend towards aridization, however, was less pronounced in the western hemisphere (in North America) than in the eastern one.

We can see some regional differences in hydrothermal regimes during the warm and cold epochs. These differences are more evident when events in the two hemispheres are compared to each other. The noted differences in atmospheric circulation have already been discussed elsewhere (VELICHKO, A.A. 1980).

There was a considerable decrease in both temperatures and precipitation in the equatorial belt. This indicates that the zone of the intratropical convergence was not as active as today, and suggests that radical changes occurred in the atmospheric circulation pattern.

Finally, fluctuations in the thermal patterns of equatorial belts also raise the eternal question about the causes of climatic changes. The cooling of these belts during the glacial maxima and their warming during interglacial optima, together with the synchronized thermal fluctuations observed in both hemispheres do not favour arguments about changes in the earth's orbit advocated by supporters of the Milankovich theory. Rather, they favour the hypothesis of fluctuations in the total heat income of the earth as a whole.

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CHRONOLOGIC CORRELATION OF LOESSES AND GLACIAL DEPOSITS IN POLAND

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ABSTRACT

Theoretical base of age correlation between loess and glacial deposits in Poland is outlined. A stratigraphic subdivision of Polish Quaternary based on chronologic data is proposed with special reference to the stratigraphy and chronology of loesses. The main problems of chronologic correlation of glacial deposits and loesses in Poland are discussed.

*

INTRODUCTION

During the last decade considerable number of studies concerning the stratigraphic subdivision and dating of Quaternary deposits were published in Poland (e.g. LINDNER, L. 1982, MARUSZCZAK, H. 1980, 1984, 1985, 1986; MOJSKI, J.E. 1982a, 1982b, 1985, 1986 in print; ROŻYCKI, S.Z. 1980; WOJTANOWICZ, J. 1983). The stratigraphic subdivision is usually compared with so-called absolute age but with different accuracy. The dating of deposits is performed by means of the thermoluminescence (TL) method and comprises the main Pleistocene facies including loess and recently also tills. Methodical premises of the thermoluminescence dating are as widely known as the doubts revealed mainly by physicists. At least one thousand TL data were obtained in Polish laboratories but only some of these have been published, and only few have been evidenced properly by geological data. These data with all their controversies allow to correlate the index deposits of Polish Pleistocene series, i.e. loesses and glacigene sediments in time. Both of the facies are very common, moreover these were formed under similar climatic conditions. This confirms the opinion on the approximately synchronous forming of the succeeding series and makes possible the attempt of correlation.

In Poland the time relations between the periods of deposition of loess and the periods of origin of glacigene sediments as well as their relationship with the type of climate have been the subject of investigations for a long time. These concerned the position of loess in the glacial cycle (development, maximum extent and decay of an ice sheet in accord with the evolution of climate). The opinion on loess formed mainly during the maximum extent of the ice-sheet (SAWICKI, L. 1932; JERSAK, L. 1977) as well as during the beginning of its thawing prevails (JAHN, A. 1950, 1956). The opinion of RÓŻYCKI, S.Z. (1961, 1972) on the loess dust deposition at the beginning of a glacial cycle seems to be rather isolated. The idea of cyclic accumulation of loess is very old but was introduced into the Polish literature by JERSAK, J. only in 1977. At the same time, the research carried out in the Polish Lowland required the adaptation of the "glacial cycle" (GALON, R. 1981) in the analysis of glacial series in the Lower Vistula River valley. The above mentioned studies were confirmed by the results of investigations on the time of accumulation and alteration of the loess cover (MARUSZCZAK, H. 1980, 1985, 1986). Finally, these have led to the conclusion that a glacial cycle should not be identified with the entire glacial period, i.e. with a stage in the stratigraphic sense, because it is a lower rank unit, usually stadial (substage), and it can occur many times during a cold glacial stage.

STRATIGRAPHY AND CHRONOLOGY OF GLACIAL DEPOSITS

The present correlation is based on the subdivision of the Polish Pleistocene proposed by the author in recent papers (MOJSKI, J.E. 1986, in print). In case of loesses the correlation comprises the last 400,000 years including two interglacials: the Mazovian and Eemian ones, and three corresponding glacial stages i.e. Wilganian, Odranian and Vistulian. In Poland no older loesses were recognized in details. However, there are some notices about their occurrence in the southeastern part of the Polish Lowlands within the San Glaciation deposits (RZECHOWSKI, J. 1967).

Two of these glaciations are stratigraphically divided into subunits (stadials and interstadials), and at least nine of these occur within the Odra and Vistula units. This subdivision is shown in *Fig. 1* illustrating both the duration and the nomenclature of subunits. It is worthy of mention that warm stages and substages (interglacials and interstadials) were comparatively short. Each of these lasted not longer than 20 ka. On the other hand, cool (cold) stages and substages were long and lasted over 100 and 20 ka respectively. Moreover, the warm units are defined according to the palaeobotanical evidences comprising a characteristic evolution of flora during an interstadial (interglacial) period.

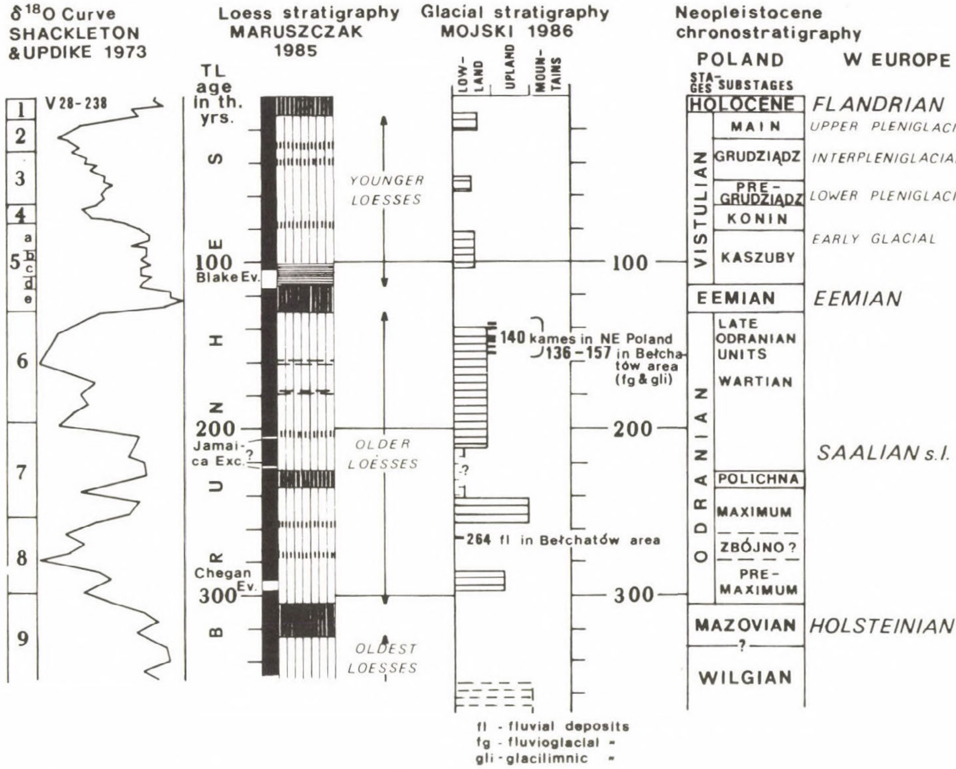

Fig. 1 Chronologic correlation of loess and glacigene deposits in Poland

Some problems, however, remain unsolved, e.g. the long-lasting Warta unit (between 210 and 140 ka). Although at that time a few marginal belts were formed, especially in the Podlasie Plain, there are insufficient evidences of lithostratigraphic classification in the recognized geological sections. Paleontological evidences are also not strong enough to determine the stratigraphic position of the time period just before the Warta unit. Thus, in an adequate column a question mark is inserted. It is also worthy of mention that according to the recent investigations a cool substage was separated from the Maximum Stadial (Substage) by the Zbójno unit (LINDNER, L.-BRYKCZYNSKA, E. 1980) which displays the characteristics of a warm interstadial (interglacial?).

STRATIGRAPHY AND CHRONOLOGY OF LOESSES

Stratigraphic and chronologic data concerning loesses derive from the studies by MARUSZCZAK, H. (1976, 1980, 1984, 1985, 1986) and their major part is shown in *Fig. 1*. It follows from these data that during the last 400,000 years two interglacial resound appeared in the loess profiles in Poland, i.e. the older (330-300 ka BP) and the younger one (130-110 ka BP). Both of these are expressed by grey brown forest lessive soils which occur in loess profiles. Each soil bears particular, typological features and contains typical epigenetic periglacial (frost) structures which facilitate their recognition in the field. The features mentioned above are exemplified in the vicinity of Hrubieszow in Eastern Poland *(Fig. 2)*. These two soils divide the entire Polish loess profile into three parts which relate to three cold stages. The Oldest Loess (nomenclature after MARUSZCZAK, H.) corresponds to the Wilga Stage, and the Old and Young ones correspond to the Odra and Vistula stages, respectively, whereas the TL time intervals are 350-330 ka BP, 300-130 ka BP and 110-12 ka BP. In Old and Young Loess beds there are some sedimentary gaps caused by considerable decrease of the rate of loess deposition. MARUSZCZAK, H. has distinguished six such gaps in the Odranian loess (Old Loess). The third gap from the

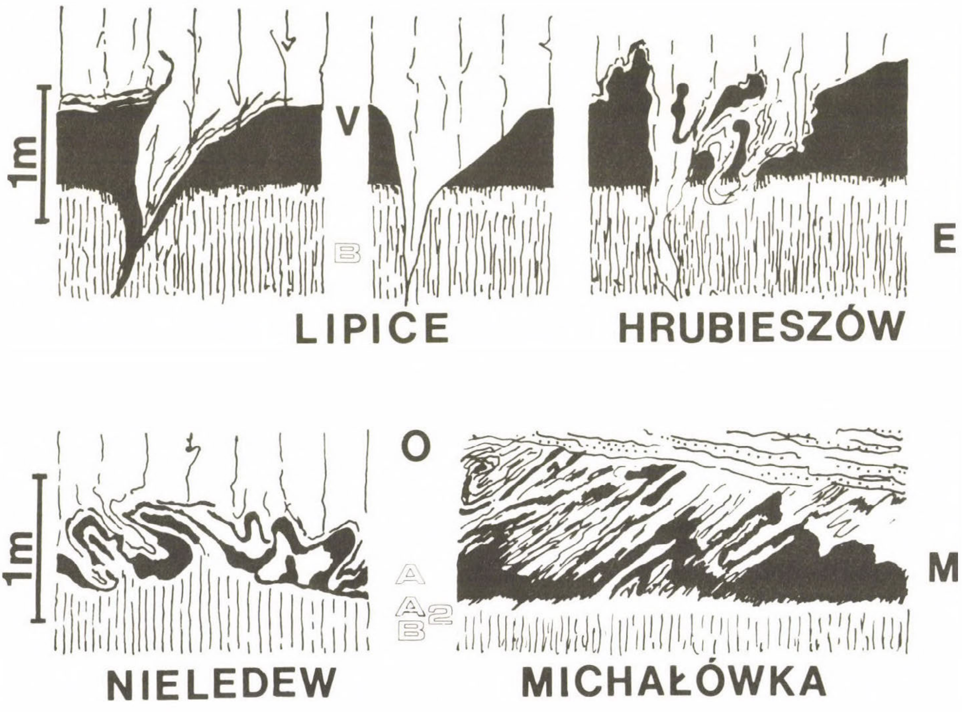

Fig. 2 Fossil soils typology of Eemian and Mazovian age in loesses of Poland

E = Eemian; O = Odranian; M = Mazovian; V = Vistulian; A, A_2, B = genetic soil horizons bottom (335-325 ka BP) is marked by a well-developed forest soil that possibly evidences the interglacial climate conditions of short duration. All the remaining soils enclosed in the Odranian loess are poorly developed (initial gley soils or chernozems) as these were formed during periods of a few thousand years.

There are three soils (initial subarctic brown soils and chernozems) distinguished in the Vistulian (Young) loess. These were formed in 80-75 ka BP, 42-37 ka BP and 32-28 ka BP intervals, respectively. The older one evidences a considerable warming, whereas both of the younger soils evidence the complicated evolution of climate which preceded the maximum extent of an ice-sheet in Europe during the Late Vistulian. The above soils divide the Vistulian loess into four horizons which correspond to cool substages. Out of these substages only the last but one can be questioned because of the small thickness of its local loess bed.

CHRONOLOGIC CORRELATION OF GLACIGENE DEPOSITS AND LOESSES IN POLAND

A proposal of chronologic correlation of glacigene deposits and loesses in Poland is presented in *Fig. 1*.

As far as the subdivision is concerned the Oldest Loess corresponds to the Wilgian, and the Old and Young Loesses correspond to the Odranian and Vistulian, respectively. Each substage has its equivalent also within the loess cover. In the Odranian a Pre-Maximum Substage of this stage with well-developed till horizon (300-280 ka BP) is related to the Lowermost Old Loess (according to the terminology of MARUSZCZAK, H.) the age being estimated to 310/300-280/270 ka BP. The Maximum Substage (Stadial) corresponds to the Lower Old Loess (280/270-260/255 ka BP) and to the Middle Old Loess (255-230/225 ka BP) together with a poorly developed gley soil between them. At Nieledew an initial chernozem occurs in the place of this soil. There are no data so far about an interstadial period in glacigene sections.

The Middle Old Loess (MARUSZCZAK, H. 1985) is not weathered and based on its typical development it can be distinguished from other Old Loess horizons. Thus, it evidences the full development of periglacial conditions that predominated within a narrow extraglacial belt in front of the maximum limit of Odranian ice sheet.

Higher up there is a hiatus (235-225 ka BP) expressed by a forest soil. There are no reliable evidences in the glacial series concerning this period but in some sections the interstadial flora was recorded. In one of these sites, in the Lublin Upland, at Polichna village, the age of deposits was determined by means of the TL method. This particular site is situated in the marginal belt of all the Pleistocene glaciations and from its name the unit is called the Polichna Substage (according to the investigations carried out by LINDNER, L.--MARUSZCZAK, H.--WOJTANOWICZ, J. 1985).

The upper section of the Old Loess can be divided into four parts. It belongs to the time interval between 225 and 130 ka BP (MARUSZCZAK, H. 1985) which is related to the Warta unit and probably to the younger chronostratigraphic units of the Odranian. The latter are distinguished by some authors based mainly on geomorphological criteria (ROŻYCKI, S.Z. 1967; LINDNER, L. 1984). However, these lithostratigraphic and morphostratigraphic units embodied by the upper section of the Old Loess correspond to the same till horizon in the Polish Lowlands which originated during the longlasting thawing of the Odranian ice sheet, terminated between 160 and 135 ka BP *(Fig. 1)*. The above TL data were calculated for the youngest glacigene deposits in the area of Widawka lobe, at Belchatów, south of Łódź (BUTRYM, J.--BARANIECKA, M.D. et al. 1982)..

MARUSZCZAK, H. has divided the Young Loess into four periods. Three of these (the Lowermost Young Loess, 100-75 ka BP, the Lower Young Loess, 75-42 ka BP and the Upper Young Loess 28-15 ka BP) are related to the three cold substages which are represented by glacigene deposits of Kaszuby, Pre-Grudziads and Main (Leszno) Substages (Stadials). All of these (especially the deposits of the two younger stadials) were dated by means of the TL method in numerous outcrops and drillings. However, the Middle Young Loess (40/37-30/28 ka BP) corresponds probably to the cool oscillation during the Grudziads Interstadial. It has been recognized also in the Russian Plain area within the Megainterstadial of the Middle Valdai (ARSLANOV, C.A. 1982) as two short cool periods (Bugrov and Shenskh ones).

Thus in Poland the chronologic correlation of loess covers and glacigene deposits is much more reliable in case of the last cold stage (Vistulian) than in the Odranian. However, the position of the Middle Young Loess refers to its interstadial accumulation, when the ice cover was lacking not only in Poland but probably in the major part of Scandinavia. It seems to confirm the old opinions among others DYLIK's J. 1960, 1966 on the lack of casual relationship between the occurrence of glacial and periglacial environments at the same time (i.e. during the Pleistocene in Poland). Thus, it supports the thesis, that not every periglacial period corresponds to the period of development (advance) of the Scandinavian ice sheet.

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TYPE-LOCALITY OF YOUNG LOESS IN HUNGARY AT MENDE

M. PÉCSI

ABSTRACT

The sequence of loess and paleosols can be traced at Mende without major interruptions, hence it is ideal for subdiving litho- and chronostratigraphically the Upper Pleistocene loesses and for reconstructing the cyclic changes of paleoenvironment in Hungary. Among the four soil complexes of the Mende profile (MF, BD, BA, MB), the "Mende-Upper" Soil Complex (MF) and the "Mende-Base" Soil Complex (MB) have been interpreted as significant stratotypes of fossil soils in the young loess. The MF Soil Complex is a stratotype that separates the Middle Würm from the Upper Würm. The MB fossil Soil Complex is a stratotype that marks probably the last interglacial.

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INTRODUCTION

The loess exposure in Mende brickyard has been considered since the 1960's as one of the most important type profiles of the Upper Pleistocene loess series in Hungary and in the Carpathian Basin (PÉCSI, M. 1965, 1966). A number of loess exposures (Basaharc, Dunaújváros, Tápiósüly etc.) have been examined in an attempt to record the stratigraphical sequences of young loess, and to correlate the paleosol horizons. The loess profile at Mende proved to be the most typical *(Figs. 1, 2, 3)*. The sequence of the stratigraphical series known as the young loess is present in this profile.

Although the profile in the Basaharc brickyard near the town of Esztergom is also fairly complete, in the Mende exposure only insignificant erosion hiatuses are observed between some layers in the form of buried and filled dells.

In the past twenty years several Hungarian loess profiles have been analysed and their stratigraphical sequences were correlated. Based on their specific characteristics, loess and paleosol complexes have been defined in different type locali-

Fig. 1 Map showing loess profiles in Hungary which had been analysed lithologically and pedologically

1 = type profiles studied in detail; 2 = profiles referred to in literature

Fig. 2 Cross-section of the loess exposures in Mende brickyard (after HAHN, Gy. 1965)

A-K and E_1, E_2 are humus horizons and paleosols; f = borehole

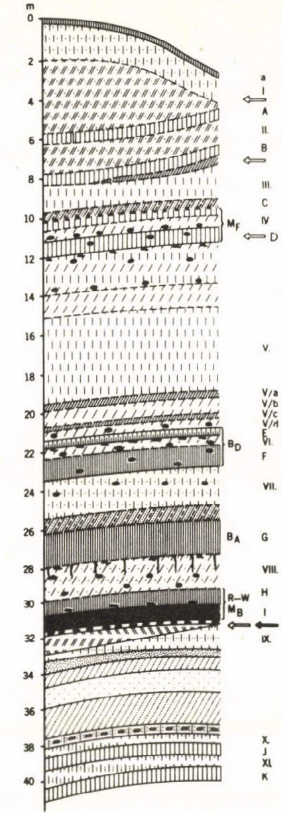

Fig. 3 Generalized profile of the loess exposure in Mende brickyard (PÉCSI, M. 1965)

a = recent chernozem; A-K = paleosols; I-IX = serial number of loess and loess-like formation; ⇒=traces of derasional processes; →=traces of erosional processes

ties or marked according to the position they occupy in the stratigraphical sequence (PÉCSI, M. 1965, 1966, 1975, 1982; PÉCSI, M.--PEVZNER, M.A. 1974; PÉCSI, M.--HAHN, Gy. 1970 and PÉCSI, M. et al. 1977). The loess bodies of coherent lithology repated in similar form in the profiles with typical paleosol complexes are called loess series.

Young loess in Hungary has an average thickness of 20-25 m and is subdivided lithologically into two subseries.

a) *The Dunaújváros--Tápiósüly* subseries of young loess comprises the uppermost 5-10 m of very loose, more or less sandy loess layers.

b) The *Mende--Basaharc* subseries includes the 10-15 m thick sequence of 3 loess horizons and 4 paleosols *(Figs. 3, 4)*.

THE UPPER PART OF YOUNG LOESS

(Dunaújváros--Tápiósüly subseries)

The 10 m thick upper series of young loess contains two poorly developed humus soils (h_1, h_2 horizons) in both of which charcoal remains have been found. The upper humus horizon in the Tápiósüly profile (located nearby) was so rich in charcoal fragments *(Betula, Pinus cembra* and *Larix picea)* that they were sufficient for radiocarbon analysis. The age of the charcoal was fixed (Hv 1615) 16 750 ± 400 years B.P. In the two humus horizons and in the overlying loess strata many bone fragments of *Rangifer tarandus* shovels have been discovered. The second humus horizon of the Mende profile is located in a similar stratigraphical position in several other loess profiles in Hungary (Tápiósüly, Dunaújváros, Dunaszekcsö and Balatonszabadi--Sóstó). The radiocarbon age of charcoal remains is 20,000 years B.P. in this stratigraphic horizon. In the so called "Dunaújváros-Tápiósüly" subseries the humus horizons usually contain mollusc fauna that prefer cold and humid climatic conditions. In the sandy loess and loess layers those molluscs prevail that thrive in cold, dry climatic phases (WAGNER, M. 1979). Thus the upper 10 m thick sequence of the Mende profile had accumulated during very cold and dry climatic conditions, occasionally interrupted by shorter spells of cold and relatively humid climates. Two phases of both dell erosion or accumulation (infilling) have been recorded. As attested by rhythmic sedimentation, microstratification and sandy material in dells, accumulation took place under cold arid climate on a sparsely vegetated surface. The whole skeleton of a young mammoth has been collected from

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Fig. 4 1 = recent chernozem, locally chernozem and brown forest soil (two storey profile); 2 = sandy loess (l_1), in the l_2 loess a whole skeleton of a young Mammoth was found; 3 = weak humus horizon (H_1, H_2); 4 stratified loessy sand (l'_1) at the lower part reindeer bone remnants occur; 5 = stratified sandy slope loess; 6 = stratotype of Mende Upper (MF) soil complex, it is a two storey profile of forest-steppe soil. In its ipper part (MF_1) there are many charchoal fragments (Picea, Larix, Pinus cembra). The Cca horizon of MF_2 is rich in lime and carbonate concretions; 7 = typical loess, but in the lower part there is a little more sandy loess with tusk and mandible of Elephas primigenius; 8 = "Basaharc double" soil complex (forest-steppe-like paleosol soil); l_5 = sandy loess and remnants of Elephas primigenius; 9 = "Basaharc lower" fossil soil (BA) locally the uppermost soil sediment; l_6 = stratified slope loess and remnants of Elephas sp., Equus sp.; 10 = stratotype of Mende Base (MB) soil complex, the upper part (MB_1)is a dark steppe-like (chernozem-like) paleosol; the lower part (MB_2) is a well-developed brown forest soil; 11 = alluvial, proluvial sand; the Base soil complex developed probably during the second half of the Riss-Würm interglacial, because the alluvial sand below that cca 125,000 years old according to the thermoluminescence data (BORSY, Z.--FÉLEGYHÁZI, J.--SZABÓ, P.P. 1979)

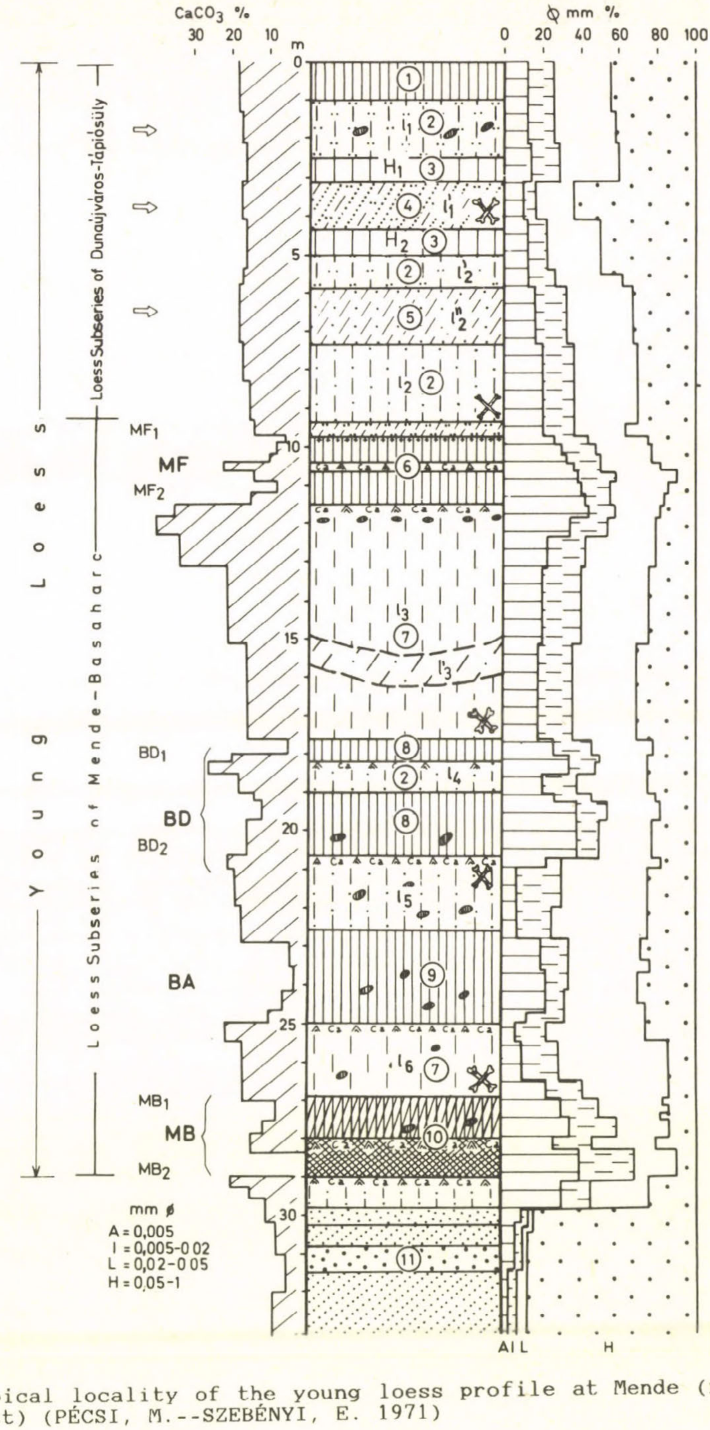

Fig. 4 Typical locality of the young loess profile at Mende (near Budapest) (PÉCSI, M.--SZEBÉNYI, E. 1971)

the typical (true) loess sequence (l_2) between 8-10 in the Mende profile. We are of the opinion that the Dunaújváros--Tápiósüly subseries of young loss had developed during the cold maximum of the last glacial stage. The cold, dry arctic loess-tundra climatic phases were interrupted by 2-3 shorter cold and humid phases during which sparse taiga forests could grow *(Fig. 5).*

THE LOWER PART OF YOUNG LOESS IN THE MENDE BRICKYARD

(Mende-Basaharc subseries)

The Mende-Basaharc subseries of young loess consists of four complexes of paleosols that enclose three significant loess packets *(Fig. 4).* No suitable explanations are available on the reasons of the formation of double soils (MF_1 and MF_2, and BD_1 and BD_2, respectively) and of the intercalated thin loesses. It can be presumed but is not proved yet that a hidden erosion gap exists between the double soils.

The "Mende-Upper" Soil Complex

The first developed double fossil soil in the Mende profil is situated between 10-12 m. The two horizons of this forest-steppe type soil can be recognized with ease in several other loess profiles in the Carpathian Basin. The upper part (MF_1) of the soil complex is poorly developed chernozem-like soil with krotovinas and charcoal, dated consistently by three laboratories 27-28,000 years B.P. (PÉCSI, M. 1965, 1975; SEPPÄLÄ, M. 1971)[1]

The lower part of the "Mende-Upper" Soil Complex (MF_2) is a well developed forest-steppe type chernozem-like paleosol. Its pedological characteristics are shown in *Fig. 6* and *Table 1* (PÉCSI, M. et al. 1977). Soil formations of similar age like the "Mende-Upper" Soil Complex have been described in several sections in Europe. Its local names include *Stillfried B* in Austria, *Kesselt* in Belgium, France and the FRG, *Gleina-Böden* in the GDR, *PKl* Czechoslovakia and Romania, and *Vitachev* and *Bryansk* in the USSR.

The "Basaharc-Double" Soil Complex (BD) in the Mende profile

Underlying the MF soil complex there is a 6 m thick almost homogenous loess packet *(Fig. 4, Table 2).* The underlying forest-steppe type double soil is remarkably well developed at Mende. This soil complex is conspicuously present in many loess exposures in Hungary and contains charcoals of coniferous trees.

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1 29,800 ± 600 Lb.No.Mo 422 (PÉCSI, M. 1966); 27,200 ± 1400 Lb.No.I. 3130 (SEPPÄLÄ, M. 1971; 27,855 ± 1589 Lb.No.Mo HV 5422 (PÉCSI, M. 1975).

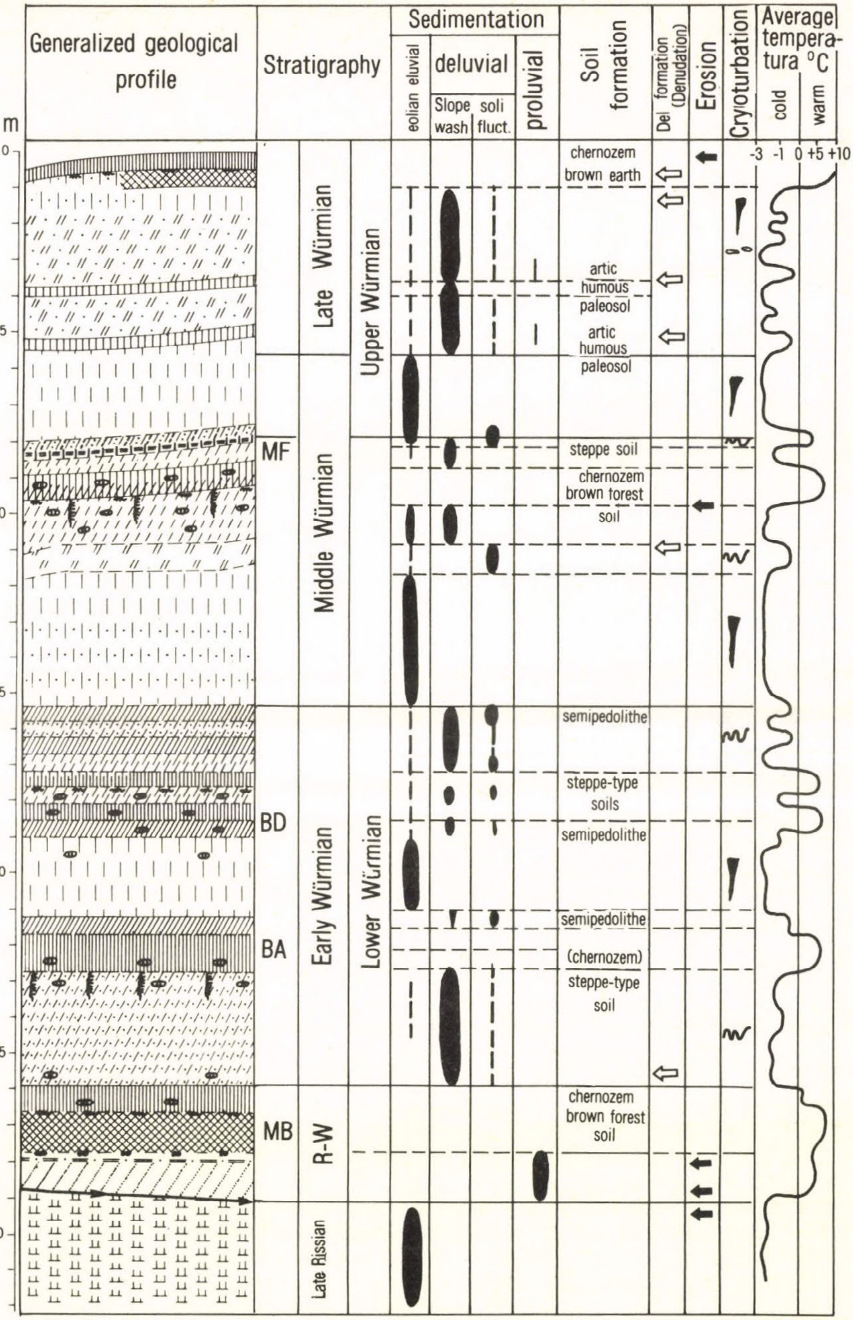

Fig. 5 Paleogeographical profile of the young loess series, generalized mainly from the Mende brickyard (PÉCSI, M.)

Table 1 Results of pedological analysis of the "Mende-Upper" Soil Complex (SZEBÉNYI, E. 1968) in a profile at Mende brickyard

Depth, m	$CaCO_3$ %	humus %	hy %	clay A	silt I	loess L	sand H	Ca	Mg	Colour
				mmØ, gr %				mg equiv/100 g		
5.85–9.75	15.9	0.27	1.24	19.1	15.8	32.8	32.3	4.20	0.0	2.5YR 5/4 yellow loess
9.75–9.90	6.3	0.86	1.65	25.7	19.0	31.6	23.6	3.70	3.75	2.5YR 5/4 MF_1A_1 horizon
9.90–10.15	7.6	1.29	1.97	29.4	18.1	30.5	22.2	9.60	7.63	10YR 5/3 A_2 horizon
10.15–10.35	10.1	1.23	1.82	31.9	18.1	27.0	22.3	7.50	9.82	10YR 5/3 AC horizon
10.35–10.60	22.0	0.55	1.50	33.7	16.7	31.1	18.1	3.00	7.63	2,5YR 6/4 C horizon
10.60–10.90	13.9	0.86	2.42	38.3	20.0	30.7	10.0	10.00	6.54	10YR 5/4 MF_2A_1 horizon
10.90–11.25	7.6	1.24	2.83	39.9	15.4	30.2	14.0	11.80	8.73	10YR 4/3 A_2 horizon
11.25–11.50	15.2	0.86	2.41	41.7	15.2	26.5	16.2	8.60	8.73	10YR 4/4 AC horizon
11.50–11.80	36.1	0.43	1.26	44.7	14.1	23.2	18.3	7.70	5.45	10YR 7/3 C horizon
11.80–12.30	39.7	0.27	1.10	33.2	19.8	26.7	20.1			10YR 8/4 C horizon
12.30–13.10	34.2	0.27	0.95	21.1	20.3	38.9	19.7			2.5YR 7/4 yellow loess
13.10–15.10	21.1	0.17	1.08	20.2	20.0	35.0	24.8	3.39	3.27	2.5YR 6/4 yellow sandy loess

It was first described in the loess exposure of the Basaharc brickyard in the Danube Bend and introduced in literature as Basaharc Double Soil Complex (PÉCSI, M. 1965). This paleosol complex is also considered a Middle Würm formation.

In Basaharc we had collected charcoal samples from the BD_1 soil. The charcoal remains are older than 32,100 ± 720 years B.P. (Hannover, 8116). According to our earlier investigations the rate of sedimentation of young loess in Hungary was 1 m/2000 years. In the knowledge of these calculations if we add the time needed for the formation of fossils soils in the profile we would suggest that the "Basaharc-Double" Soil Complex is probably 42,000-45,000 years old (PÉCSI, M. 1972). The pedological characteristics of the "Basaharc-Double" Soil at Mende are shown in *Fig. 7* and *Table 2*. From the slightly sandy loess layer of 2 m thickness underlying the BD soil bone fragments, molars and pieces of an *Elephas primigenius*'s tusk have been recovered, the radiocarbon dating of the bone indicates a minimum of 30,000 years B.P. BUTRYM, J. and MARUSZCZAK, H. (1984) found the age of the soils BD_1 and BD_2 in the Paks brickyard 37,000 and 41,000 years B.P., respectively, by TL analyses.

Table 2 Results of pedological analysis of the "Basaharc-Double" Soil Complex in the Mende brickyard exposure (SZEBÉNYI, E. 1968-1976)

Depth, m	Thickness of strata	$CaCO_3$ %	humus %	hy %	clay mm Ø	silt gr	loess %	sand	Ca	Mg	Colour
									mg, equiv/100 g		
15.10—17.60	2.50	16.5	0.27	1.12	18.4	16.8	36.3	29.0	14.20	6.54	2.5YR 6/4 yellow loess
17.60—18.00	0.40	5.5	0.62	1.79	25.8	19.5	32.5	21.6	14.90	6.54	10YR 6/3 A horizon
18.00—18.20	0.20	19.8	0.62	1.61	34.7	15.7	24.0	24.8	10.90	6.54	10YR 7/3 A/C horizon
18.20—18.45	0.25	26.1	0.27	1.28	33.2	14.0	28.3	24.3	6.00	5.45	10YR 7/3 C horizon
18.45—19.00		18.1	0.21	1.12	20.7	17.2	38.1	23.3	7.40	3.27	5YR 7/3 yellow loess
19.00—19.20		16.6	0.21	1.13	19.4	20.8	30.6	29.8			10YR 6/3 yellow loess
19.20—19.40		16.4	0.21	1.09	19.5	19.4	30.1	31.1			10YR 6/3 yellow loess
19.40—19.60		16.7	0.21	1.08	18.5	19.0	30.8	32.7			10YR 6/3 yellow loess
19.60—19.80		17.0	0.21	1.00	17.4	19.0	30.5	33.0			10YR 6/3 yellow loess
19.80—20.00	1.55	16.5	0.21	1.08	17.8	21.2	30.0	31.2			10YR 6/3 yellow loess
20.00—20.30		13.8	0.32	1.13	18.1	19.7	30.6	32.5	5.32	2.06	10YR 6/3 A1 horizon
20.30—20.60		8.4	1.21	1.29	18.4	18.9	32.8	28.9	8.80	2.06	10YR 6/3 A1 horizon
20.60—20.75		3.8	0.21	1.40	23.1	21.2	30.2	26.0	12.32	2.06	10YR 6/3 A2 horizon
20.75—20.90		4.2	0.32	1.72	29.6	21.3	27.0	24.3	11.87	1.03	10YR 5/4 A2 horizon
20.90—21.05		5.8	0.32	1.79	31.7	13.7	27.7	26.7	11.87	1.03	10YR 5/4 A2 horizon
21.05—21.15	1.15	5.4	0.32	1.83	30.9	16.3	25.7	29.2	11.87	1.03	10YR 5/4 A2 horizon
21.15—21.30		12.9	0.21	1.49	30.3	17.7	25.2	27.6	9.80	5.68	10YR 6/4 AC horizon
21.30—21.50		12.5	0.21	1.33	30.8	17.5	22.4	30.1	9.80	5.68	10YR 6/4 AC horizon
21.50—21.65		10.4	0.21	1.36	27.5	15.4	26.0	30.8	9.80	5.68	10YR 6/3 AC horizon
21.65—21.83	0.78	13.4	0.21	1.56	23.5	17.2	29.2	30.8	5.60	3.10	10YR 6/3 AC horizon
21.83—22.03	0.20	17.1	0.21	1.17	20.7	19.0	29.7	30.5	5.60	3.10	10YR 6/3 C horizon
22.03—22.18	0.15	17.9	0.21	1.05	18.4	20.0	31.9	31.6	7.77	1.03	10YR 7/3 yellow loess

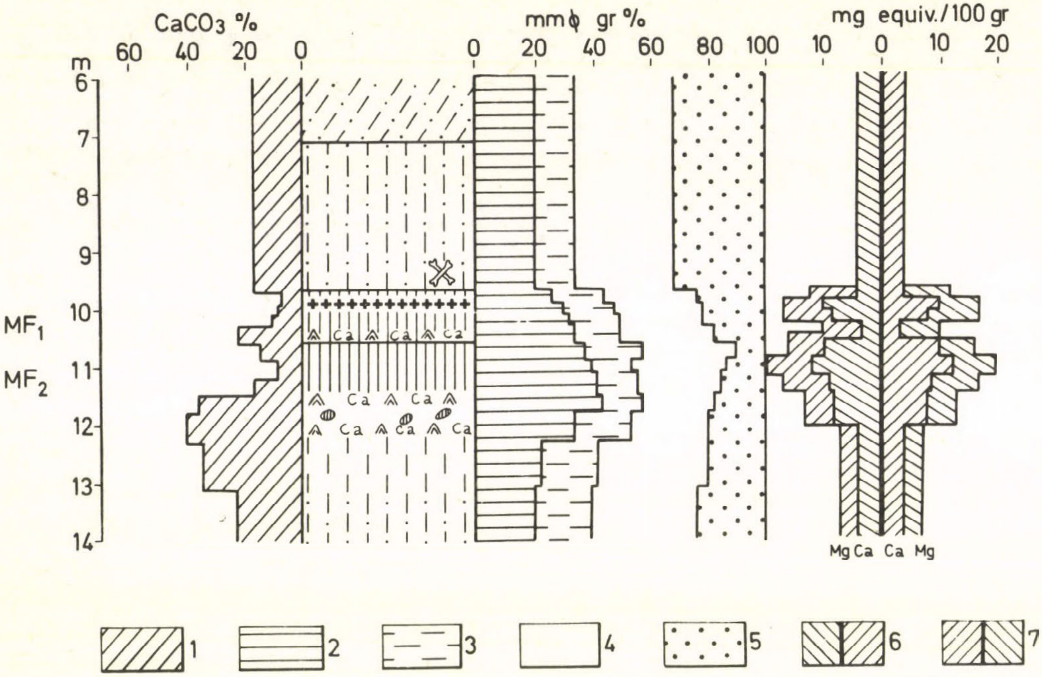

Fig. 6 Pedological section of the "Mende-Upper" Soil Complex in the profile at Mende brickyard

1 = $CaCO_3$ content; 2 = clay fraction (< 0.005 mm ø); 3 = silt fraction (0.02-0.05 mm ø); 5 = sand fraction (greater than 0.05 mm ø); 6 =exchangeable Ca mg equiv/100 gr; 7 = exchangeable Mg mg equiv/100 gr; MF_1 = upper soil of the "Mende-Upper" Soil Complex; MF_2 = lower soil of the "Mende-Upper" Soil Complex

The "Basaharc-Base" Soil (BA) in the Mende Profile

The BA soil in the Mende profile is a remarkably well developed dark coloured compact chernozem-type (chernozem-meadow) soil *(Fig. 4)*. It is mostly rich in krotovinas. Its pedological characteristics are shown in *Fig. 8* and *Table 3* (PÉCSI, M. et al. 1977). Direct evidence about the absolute age of this soil is not available at present. Relying on our calculations about the rate of sedimentation and (fossil) steppe soil formation we may estimate the age of the BA soil to be cca 65,000 years. The BA soil of the Paks loess exposure was dated by the TL examinations of BUTRYM, J. and MARUSZCZAK, H. (1984) 81,000 years B.P. In the Mende profile below the Basaharc-Base Soil there is a 2 m thick somewhat stratified loess stratum, the lower part of which is solifluction loess *(Fig. 3)*. Teeth of *Equus* sp. have been found here which are most likely of Würmian type, determinated by M. KRETZOI.

Table 3 Results of pedological analysis of the "Basaharc-Lower" Soil Complex in a profile at Mende brickyard (SZEBÉNYI, E. 1976)

Depth, m	$CaCO_3$ %	humus %	hy %	clay	silt	loess	sand	Ca	Mg	Colour
				A	I	L	H			
				mm∅, gr%				mg equivG/100 g		
22.18—22.31	15.6	0.32	1.18	21.1	20.7	28.9	29.1	5.10	3.71	10YR 6/3 sandy yellow loess
22.31—22.40	15.0	0.43	1.35	25.4	19.3	26.5	28.9	5.10	3.71	10YR 6/3 sandy yellow loess
22.40—22.49	15.0	0.43	1.40	26.3	19.0	27.8	26.6	12.90	4.18	10YR 6/3 sandy yellow loess
22.49—22.61	12.9	0.43	1.70	32.1	17.3	27.3	23.9	12.90	4.18	10YR 5/4 A horizon
22.61—22.72	11.3	0.43	1.94	34.8	15.1	27.0	23.3	12.90	4.18	10YR 5/2 A horizon
22.72—22.89	11.6	0.43	2.02	32.9	16.7	26.2	24.1	12.90	4.18	10YR 5/2 A horizon
22.89—23.10	11.6	0.43	1.72	34.6	17.8	27.9	20.2	5.16	8.77	10YR 5/3 A horizon
23.10—23.45	10.4	0.43	2.02	34.4	19.0	27.0	17.4	5.16	8.77	10YR 4/4 A horizon
23.45—23.65	11.2	0.43	2.09	36.2	15.8	29.4	19.2	5.16	8.77	10YR 4/4 A horizon
23.65—23.85	12.5	0.43	1.95	34.3	18.7	27.8	19.1	11.35	2.58	10YR 4/4 A horizon
23.85—24.05	13.7	0.43	1.91	35.2	18.5	27.2	18.1	11.35	2.58	10YR 4/3 A_1 horizon
24.05—24.25	9.6	0.43	1.88	35.9	15.0	26.3	7.96	22.5	5.16	10YR 4/4 A_2 horizon
24.25—24.45	8.7	0.43	2.04	36.2	15.8	24.0	22.8	8.56	5.16	10YR 4/3 A_2 horizon
24.45—24.65	8.7	0.43	2.14	36.8	17.6	23.6	22.8	8.56	5.16	10YR 4/3 A horizon
24.65—24.85	8.3	0.43	2.13	35.6	17.1	23.8	23.9	8.56	5.16	10YR 4/3 A horizon
24.85—25.05	7.5	0.43	2.07	36.7	18.0	20.3	25.0	6.71	5.16	10YR 4/4 A horizon
25.05—25.25	8.3	0.43	2.06	35.1	14.6	21.6	27.2	6.71	5.16	10YR 4/3 A_3 horizon
25.25—25.45	7.5	0.43	1.96	34.4	20.5	26.5	18.6	8.56	4.13	10YR 4/3 A_3 horizon
25.45—25.65	8.3	0.43	1.86	33.0	20.2	28.2	18.7	6.71	3.10	10YR 4/4 AC horizon
25.65—25.80	11.2	0.43	1.48	35.3	20.8	26.8	17.3	8.77	1.55	10YR 4/3 AC horizon
25.80—25.95	9.5	0.43	1.50	35.7	20.6	26.9	17.2	8.77	1.55	10YR 4/3 AC horizon
25.95—26.10	14.6	0.21	1.35	33.7	21.2	26.7	17.5	8.77	1.55	10YR 4/4 C horizon
26.10—26.25	13.3	0.21	1.32	31.0	22.0	27.7	18.4	6.10	1.55	10YR 5/3 C horizon
26.25—26.40	12.5	0.21	1.07	26.8	23.0	30.6	19.1	6.30	1.55	2.5Y 6/4 C horizon
26.40—26.55	12.5	0.21	1.12	26.3	23.3	30.4	30.0	6.30	1.55	2.5S 6/4 C horizon
26.55—26.75	11.2	0.21	1.05	26.3	21.6	31.6	20.4	6.30	1.55	2.5Y 5/4 C horizon
26.75—26.95	10.0	0.21	1.05	25.8	21.4	32.2	20.5			2.5Y 5/4 C horizon
26.95—27.15	10.4	0.21	1.16	26.6	21.2	30.3	21.7			2.5Y 5/4 yellow loess
27.15—27.30	7.9	0.21	1.22	26.0	23.5	29.8	20.7			2.5Y 5/4 loess
27.30—27.45	9.1	0.21	0.99	25.9	23.2	30.4	22.0			10YR 6/4 loess
27.45—27.65	8.3	0.21	1.15	24.9	22.0	30.2	23.2			loess

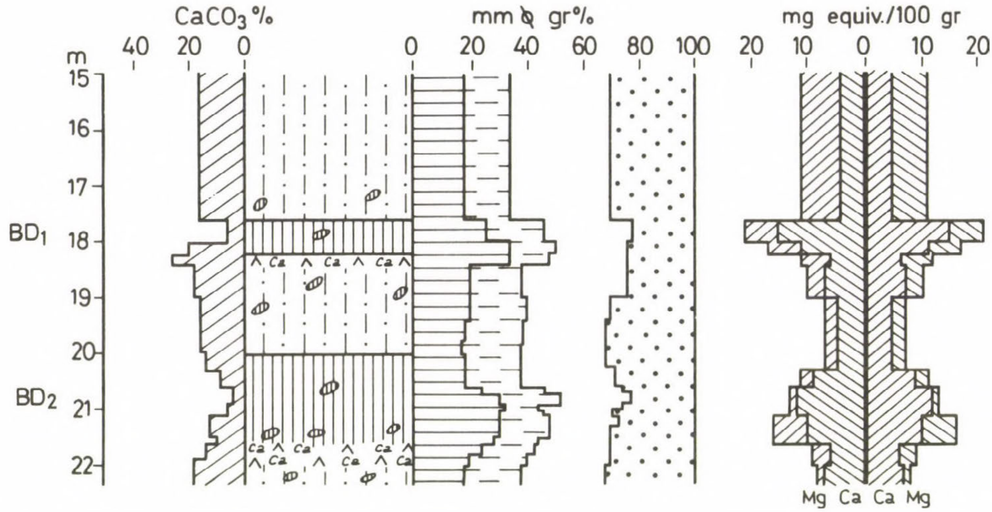

Fig. 7 Pedological profile of the "Basaharc-Double" Soil Complex (after PÉCSI, M.--SZEBÉNYI, E.) in the profile at Mende brickyard (1968-1976). (For legend see Fig. 6)

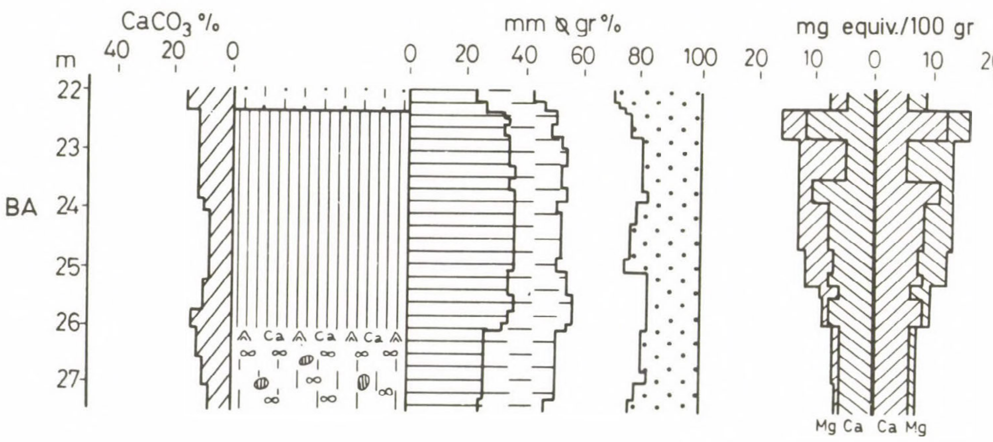

Fig. 8 Pedological profile of the "Basaharc-Lower" Soil Complex (after PÉCSI, M.--SZEBÉNYI, E.) in a profile at Mende brickyard (1976). (For legend see Fig. 6)

The "Mende-Base" Soil Complex (MB)

This double soil consists of a brown forest soil and a forest-steppe-type chernozem soil. The upper unit, the forest-steppe-type soil (MB_1) directly overlies the lower, a reddish brown forest soil (MB_2; *Fig. 9, Table 4*). The stratigraphical position of the "Mende-Base" Soil Complex in Hungary and in the Carpathian Basin is such, that it may be regarded as a stratotype that separates the young loess from the old loess. This was first described by M. PÉCSI (1965) and the pedological analysis was done by P. STEFANOVITS (1965).

M. PÉCSI suggested that this soil complex had probably formed during the second half of the last interglacial (R-W). From the overlying young loess sequence the vertebrate and mollusc fauna and the floral remains recovered were all formations of the last glacial, while the series of chernozem-

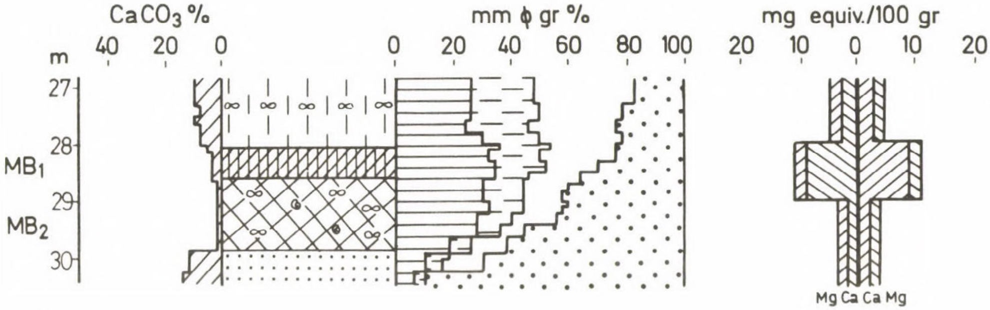

Fig. 9 Pedological profile of the "Mende-Base" Soil Complex (after PÉCSI M.--SZEBÉNYI, E.) in a profile at Mende brickyard (1976). (For legend see Fig. 6)

type soil-horizons (MF, BD and BA) interbedded in the young loess were considered the indications of warmer intervals within the Würm glacial. There is little doubt that the MB soil is an interglacial formation, but was it indeed formed during the last interglacial?

ABSOLUTE AGE OF THE MENDE BASE SOIL COMPLEX

The duration of the Würm and Riss-Würm interglacial are dated differently on the absolute time-scale by various authors. The climatic optimum of the Riss-Würm interglacial was dated as between 80,000-90,000 years B.P. by B. EVANS (1972), based on maximum solar radiation during this time. C. EMILIANI (1969) and others examined deep sea sediments and consider the climatic optimum to have been reached between 90,000-100,000 years B.P. A. DREIMANIS and A. RAUKAS (1975) put this phase between 110,000-130,000 years B.P., while W.S. BROECKER and J. DONK (1970) after analysing the cyclic changes in the O^{18}/O^{16} isotope ratios

Table 4 Result of pedological profile analysis of the "Mende-Base" soil complex in the Mende brickyard exposure (SZEBÉNYI, E. 1976)

Depth, m	$CaCO_3$ %	humus %	hy %	clay A	silt I	loess L	sand H	Ca	Mg	Colour
				mmØ, gr%				mg equivG/100		
27.65—27.80	7.50	0.—	1.30	30.4	20.9	26.4	22.2	3.71	2.13	10YR 5/2 yellow loess
27.80—27.95	6.25	0.—	1.50	35.6	18.8	21.8	24.1	8.77	2.13	7.5YR 5/4 yellow loess
27.95—28.10	4.17	0.—	1.63	32.6	17.1	25.6	24.8	8.77	2.13	7.5YR 4/4 MB soil
28.10—28.25	5.00	0.—	1.63	34.6	16.7	17.9	30.8	8.17	2.13	7.5YR 4/4
28.25—28.40	5.00	0.21	1.61	33.4	13.5	17.0	35.6	8.17	2.13	5YR 4/6
28.40—28.55	0.83	0.21	1.50	30.3	14.1	18.7	36.3	8.77	2.13	5YR 4/3
28.55—28.60	1.25	0.—	1.36	29.2	13.0	18.7	39.5	8.77	2.13	5YR 4/3
28.60—28.70	1.25	0.—	1.88	29.4	12.9	15.8	42.4	8.58	2.03	5YR 4/4
28.70—28.80	0.83	0.—	2.03	29.3	12.8	16.5	42.3	8.58	2.03	5YR 5/4
28.80—28.90	0.83	0.—	2.40	32.8	11.5	15.2	40.5	2.06	2.03	5YR 4/4 B horizon
28.90—28.95	2.92	0.—	2.55	33.2	11.5	14.5	40.3	2.58	2.03	5YR 5/4 B horizon
28.95—29.05	0.83	0.—	2.60	32.3	11.0	14.8	41.9	2.58	2.03	7.5YR 5/6 B horizon
29.05—29.20	2.08	0.—	2.35	27.1	12.4	15.0	44.5	2.58	2.03	7.5YR 5/6 BC horizon yellow loess
29.20—29.40	0.83	0.—	2.21	28.6	8.0	13.6	50.0			
29.40—29.60	2.50	0.—	1.77	18.3	7.3	11.4	62.7			

in the deep sea sediments dated the optimum of the Riss-Würm interglacial as 127,000 ± 6000 years B.P. The age of the Blake Event in the Upper Pleistocene sediments was paleomagnetically determined as 107,000 years B.P.

FINK, J. and KUKLA, G.J. (1977) have suggested that we must look for the evidence of the Blake Event in the young loess, in the upper part of the Riss-Würm soil formation and in the overlying loess strata. In order to determine wether the Blake Event is represented, we carried out paleomagnetic analyses of the whole loess profile.

In the past year two geophysical laboratories have analysed independently by different methods some critical sections of the Mende loess profile, its specific stratotypes. M.A. PEVZNER has completed the paleomagnetic investigation of the lower part of the profile in the Geophysical Laboratory of the Geological Institute of the Soviet Academy of Sciences (PÉCSI, M.--SZEBÉNYI, E.--PEVZNER, M.A. 1979). Subsequently or parallel to these investigations P. MÁRTON of the Geophysical Department of the Eötvös Loránd University, Budapest has also carried out serial analyses (MÁRTON, P. 1979).

In spite of the very careful and detailed examination (samples were taken from the MB soil at 5 cm depth intervals) repeated several times, the Blake Event could not be found. All the samples presented normal polarity. Therefore, among others, the assumption was made that the Blake Event should probably precede the formation of the MB Soil Complex and, consequently, it cannot be found in the Mende Profile.

Even in a previous paper, we calculated the age of the MB Soil Complex cca 110,000-120,000 years B.P. (PÉCSI, M.-

PEVZNER, M.A. 1974) and, since no paleomagnetic data were found to support it, other methods of justification were sought.

Great help was given by the group of scientists in Debrecen engaged in thermoluminescence dating (BORSY, Z.--FÉLSZERFALVI, J.--SZABÓ, P.P. 1979), who undertook the absolute dating of the paleosol.

They obtained 105,000 ± 17,000 years for the absolute age of the MB Soil Complex, while in the Paks exposure the age of the loess underlying the MB soil was found 125,000 ± 20,000 years (BORSY, Z.--FÉLSZERFALVI, J.--SZABÓ, P.P. 1979). The investigations were followed by the TL analyses of two Polish experts. BUTRYM, J. and MARUSZCZAK, H. (1984) determined the age of the MB_1 soil of the Paks exposure as 121,000 ± 16,000 years and of the MB_2 soil as 124,000 ± 17,000 years.

Consequently, according to the above TL analyses the formation of the MB Soil Complex took place between 105,000 and 124,000 years B.P. The time interval determined by instrumented measurements supported and justified our previous estimation or calculation concerning the last interglacial age of the MB soil though recently the different TL laboratories provide rather scattering absolute ages on certain older fossil soil. Thus, the absolute dating of the MB paleosol cannot be regarded as closed.

CONCLUSION

The best type-locality for young loess in Hungary is the Mende brickyard section. In the upper part of the profile (8-10 m) two thick loess packets (l_1 and l_2), two embrionic humus soils (h_1 and h_2) and a subarctic soil occur. The humus horizons h_1 and h_2 were dated 16,000-20,000 years. They may represent taiga grove environment. Immediately above the humus horizons remnants of reindeer antlers and bones regularly occur *(Rangifer tarandus* marker horizons). From the loess l_1 above paleosol MF_1 an almost complete skeleton a young *Elephas primigenius* was recovered. In this latter paleosol coniferous charcoal remnants also occur, their age is 27,000-29,000 years. The soil complex Mende Upper (MF_1 and MF_2) in young loess represents the interstadial with forest steppe, climatic amelioration between the Middle and Upper Würm.

The upper part of young loess formed cca 28,000-12,000 years B.P., when mostly dry cold loess steppe and cyclically cool and humid subarctic taiga groves alternated on two or three occasions. Mollusc and pollen analyses also indicate the coldest environments within the Würm glacial (WAGNER, M. 1979, URBAN, B. 1984).

In the lower part of young loess (15 m) three major thick loess layers and three chernozem type forest steppe soils (BD_1, BD_2, and BA) occur. There are also frequent charcoal remnants of coniferous trees. Remains of elephant teeth and tusks were all recovered from intercalated loess horizons. They are mostly

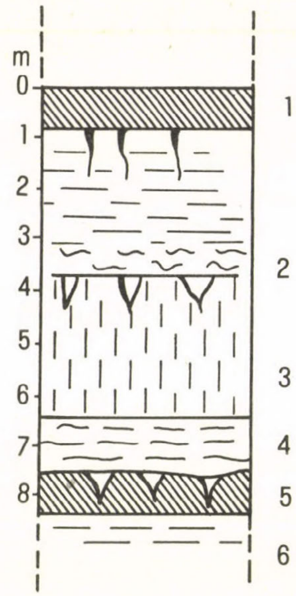

Fig. 10 Accumulation phases of slope deposits between two paleosoils during the last glacial. A schematic profile (PÉCSI, M. 1962)

1 = fossil steppe soil (chernozem). A polygon network of $CaCO_3$ filled cracks stretches from its base to 1.5-2 m depth; 2 = slope loess or sandy loess of rhythmic stratification. Sheet wash accumulations. Little amount of rainfall in early summer, considerable accumulation of snow; seasonally frozen subsoil; 3 = microstratified slope loess with weak cryoturbation structure. Humid and cold climate (solifluction and sheet wash) during formation; 4 = unstratified, 'typical' loess complex. Accumulated in the coldest and driest continental climatic phase, mainly of eolian dust; 5 = stratified slope loess, often substituted by other stratified slope deposits (loess-like semipedolites). Deposits were accumulated by pluvionival slope wash in the early stadial; 6 = fossil forest soil (or forest-steppe soil). Conditions of formation: warm and humid forest climate. The forest is eliminateed by climatic deterioration; locally cryoturbation phenomena occur in the top of the forest soil.

found directly above paleosols. These paleosols attest to the three interstadials within the Middle and Lower Würm.

In some parts of the profile the semipedolites in soils indicate the slow decay of taiga forest steppe, solifluction, and soil redeposition. Subsequently, upon the semipedolites stratified slope loess and unstratified true loess were deposited pointing to a colder and drier climate. Among the infraloess paleosols the driest climatic stage is represented by unstratified and mostly sandy loess layers. In these stages vegetation was sparse and wind became the most effective geomorphic agent.

A change to a slightly more humid climate is shown by the superposition of solifluction and then loess reworked during snowmelt upon the unstratified loess. The sediment cycle between two paleosols in young loess ends by interstadial steppe climate, i.e. formation of chernozem-like soils *(Fig. 10)*. The best preserved examples of this series are found between the soil complexes MF and BD in the Mende profile *(Fig. 5)*. In contrast, between BA and MB only stratified soliflectional loess formed and it indicates cool or cold-humid anaglacial climatic conditions during the Early Würm stadial.

The lower member of soil complex (MB_2) is a brown forest soil with numerous $CaCO_3$ concretions and pollen of conifers. Thus, it may represent interglacial climate with characteristic dry and humid seasons. MB_1 soil was formed under forest steppe and contains much humus and numerous krotovinas. The transition to A horizon is gradual. It indicates slow climatic deterioration and drier conditions with prevailing forest steppe.

With the exception of the forest soil MB_2, in young loess only chernozem-like forest steppe soils (BA, BD_1, BD_2, MF_1 and MF_2) or subarctic humus soils (h_1 and h_2) occur. In Hungary this non-forest paleosols are characteristic within the young loess where recently brown forest soil is formed. Consequently, these indicate similar climate fluctuations within one glacial cycle. The MB_2 brown forest soil formation considerably differs from these. The decision whether the MB soil complex was formed during the last or the penultimate interglacial stage needs further investigations. Similarly, the paleogeographic significance of the erosion hiatuses within the loesses and of evolution of the double soils should also be elucidated.

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COMPOSITION OF LATE PLEISTOCENE LOESSSES OF THE EUROPEAN USSR

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ABSTRACT

In the whole northern hemisphere Late Pleistocene loess is most widespread within the European USSR. Here it forms a single loess belt from the western boundaries of the USSR to the Trans-Volga regions. In the north, loess does not occur along the margin of the glacier. In the south, the loess belt reaches up to the coasts of the Black and Azov seas.
The material composition of loesses (granulometric, mineralogical) regularly changes latitudinally and meridionally. Minerals are characterised by a homogeneous composition. Yet the Volyno-Podolia, Dnieper-Oka, and Don provinces are distinguished. The differences concern morphology of grains (colouring, presence or absence of inclusions, dual character, etc.). Late Pleistocene loesses are characterized by quite low weathering indices (K_I = 0.19-0.35).

MAIN UNITS OF LOESS SERIES

During the whole Pleistocene, the European part of the USSR was an area of intensive loess formation. Loess deposits, typical loesses, and loess-like formations accumulated here during Late Pleistocene cold intervals which followed the Mikulino (Eem) Interglacial. The loess cover reached its maximum dimensions during the Late Pleistocene over the whole European loess area including the European part of the USSR (VELICHKO, A.A.-KHALCHEVA, T.A. 1982). Loesses of this period formed an indivisible loess belt in the middle and western sectors of the Russian plain which occupied most of the extraglacial territory and covered three main "loess" regions: the Volyno-Podolia in the west, the Dnieper basin in the center, and the Don and Oka basins in the eastern parts of the plain.

Detailed studies of Late Pleistocene loesses and of alternating fossil soils by means of palynological, paleofaunistic, paleocryologic, paleopedological, and mineralogical analyses, together with absolute dating, permitted us to establish a chronostratigraphic scheme for the loess-soil series for the study areas (VELICHKO, A.A.--MOROZOVA, T.D.--UDARTSEV, V.P. 1985).

The stratigraphic scheme *(Table 1)* compiled for the European part of the USSR, allowed us to make correlation of loess horizons which corresponded to the different epochs of loess accumulation, and to compare lithological data (the granulometric and mineralogical composition, in particular) from different regions to obtain information on the conditions of loess formation in them.

Table 1 Late Quaternary glacial and interglacial events in the European USSR

Link	Overhorizon	Horizon
Holocene	Holocene	Holocene
Late Pleistocene	Valdai Glacial Epoch	Loess III (Altynovo)
		Yaroslavl cryogenic horizon
		Gleyey level (Trubchevsk soil)
		Loess II (Desna)
		Vladimir cryogenic horizon
		Bryansk soil
		Smolensk cryogenic horizon (phase "b")
		Krutitsa soil of the Mezin soil complex
		Thin intramezin (Sevsk) loess
		Smolensk cryogenic horizon (phase "a")
	Mikulino Interglacial	Salyn' soil of the Mezin soil complex

The Late Pleistocene Valdai glacial epoch of loess accumulation is subdivided into two stages which differ in time and in paleogeographic conditions, and which are separated by a somewhat warmer period during which the interstadial Bryansk fossil soil was formed.

In the Mezin soil complex, at the boundary between the Salyn' and Krutitsa phases, there are remains of the Sevsk loess horizon. This thin deposit is not preserved everywhere because it was reworked by soil formation processes of the superimposed Krutitsa soil (VELICHKO, A.A.--MOROZOVA, T.D. 1972). Because of this we did not analyze the substance composition of this loess.

Loess I (Khotylevo) deposited during the pre-Bryansk stage of the Late Pleistocene, differs from the post-Bryansk loess in having larger argillaceous content and lower velocity of accumulation of the initial material. This is a widely distributed, stable in thickness, loess horizon. On the Volyno-Podolian upland it is 0.8-1.0 m thick, in the Dnieper basin 1.0-1.5 m, and on the Oka-Don plain 0.5-1.0 m thick.

The post-Bryansk stage of loess formation featured the accumulation of two loess horizons: loess II (Desna) and loess III (Altynovo) which are divided by an interstadial cryomorphic fossil soil (Trubchevsk).

Loess II (Desna) is the most pronounced and the thickest of Late Pleistocene loesses. Its thickness, however, varies from region to region. Its maximal thickness of 3-4 m is observed in the mid-Dnieper basin. On the Oka-Don plain and on the Volyno-Podolian upland its thickness does not exceed 2-3 m.

REGIONAL CHARACTERISTICS OF LOESS ACCUMULATION

Thus, thickness of loess and II and III are maximal in the middle Dnieper basin. This region is the most representative for post-Bryansk loesses and reflects the most favourable past conditions for loess formation.

The existence of optimal conditions for loess accumulation during the Late Pleistocene in the middle Dnieper basin are also indicated by data from granulometric analysis *(Fig. 1)*.

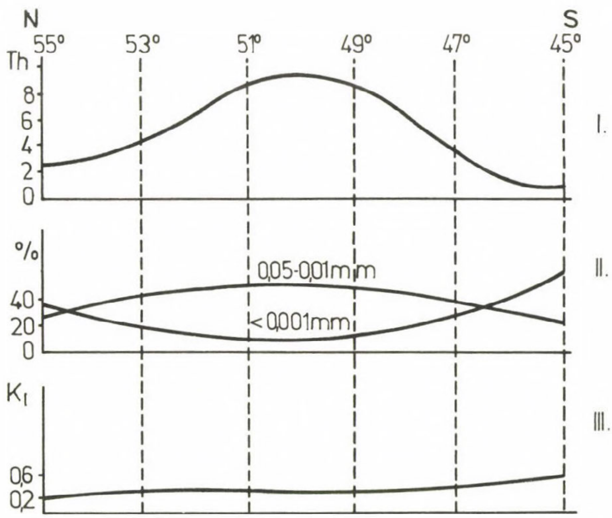

Fig. 1 Change of properties of Late Pleistocene loesses along the meridonal profile across the Russian plain

I = thickness; II = granulometric composition; III = weathering coefficient

They indicate, that together with the prevalence of a silt-stone fraction common to all "loess" regions, the content of the sand fraction in the pre-Bryansk loess I in the Dnieper basin amounts to 5-6%, and of clay up to 10-12%.. Southward and south-eastward, the composition becomes more argillaceous, whereas westward and eastward it becomes more sandy (on the Oka-Don plain: sand - 10-15%, clay - 20-40%, the Volyno-Podolian upland: sand - 10-20%, clay - 20-30%) Eastward, the post-Bryansk loess horizons contain increasing amounts of sand and clay (15-20% of sand, 30-40% of clay). In the west (Volyno-Podolia), the granulometric composition in comparison to the central regions (the Dnieper basin) practically does not change. Silt fraction here amounts to 60-60% (VELICHKO, A.A.--MOROZOVA,.T.D.-UDARTSEV, V.P. 1985).

Changes in Late Valdai loess properties are observed also from north to south *(Fig. 2)*. The area of the maximal accumulation of post-Bryansk loesses, which lies 200-300 km south of the boundary of the Valdai glaciation, forms a specific "loess esker" which occupies the Dnieper basin. It continues to the east of this region (the Oka-Don plain) as well, but in a less prononunced way.

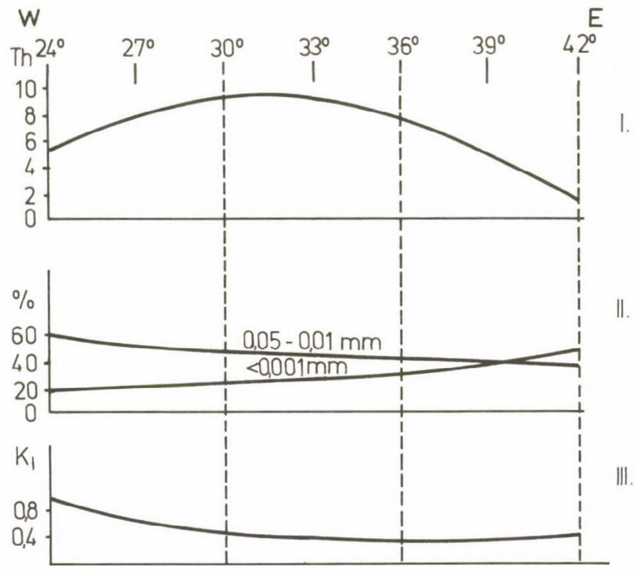

Fig. 2 Change of properties of Late Pleistocene loesses in latitudinal direction across the Russian plain

I = thickness; II = granulometric composition: III = weathering coefficient

The absence of the fraction of coarse and medium sand (more than 0.25 mm) and a similar content (about 30-40%) of the three main fractions - fine sand, silt, and clay- is typical for the zone with the thickest and most homogeneous accumulation of loesses (51-52° N). Northward of the area of the maximal accumulation a small content of fine and coarse sand fraction is observed. The correlation of silt and clay fractions here is close to that typical for loesses in the area of maximal acccumulation.

SPECIAL FEATURES OF MINERAL COMPOSITION AND WEATHERING

For the study of the mineral composition of loesses, we chose the silt fraction (0.1-0.01 mm) since it prevailed in its mass and was the most informative one. The analysis of this fraction showed that the mineral composition of Valdai loesses changes little in all the studied regions of the Russian plain. The heavy fraction everywhere amounts to less than 0.1% and includes minerals of the ore group and of the titanium containing group (transparent), the latter contain sphene, rutile, brookite, anatase, tourmaline, zircon and garnet. The epidote group, which is also present, contains epidote and zoisite; the amphibole group is mainly represented by standard hornblende and biotite. Disthen, staurolite, apatite, and picotite are present is smaller amounts.

The main mass of the light fraction consists of quartz. The amounts of feldspar (mainly orthoclase, more seldom plagioclase and microclin), of calcium carbonates, muscovite, and of argillaceous-micaceous aggregates are considerably lower. Opal and chalcedony as well as glauconite and volcanic glass occur in single grains.

The similarity in the mineral composition of the Valdai loesses from the different regions (from the Volyno-Podolian upland, the basin of the Dnieper, and the Oka-Don plain) allows us to assume that, at the time these deposits accumulated, there was no single strictly localized source of removal. In addition , the observed similarity in the loess mineral masses also partially resulted from numerous previous episodes of redeposition of loess minerals from older horizons.

The mineralogical composition of loesses from the Volyno-Podolia, the Dnieper, and the Oka and Don basins shows also some differences associated with the differential distribution of various rock forming minerals. These differences primarily occur in the heavy fraction minerals.

Minerals of the epidote group prevail in all three loess provinces. Zoisite - the most resistant to the processes of chemical weathering among minerals in this group - is prevalent in the Volyno-Podolia area. In the basins of the Dnieper, Oka, and Don rivers, epidote predominates. The percentage of differentially weathered epidote increases in these areas.

The Volyno-Podolia area contains higher amounts of such stable minerals as zircon and garnet. Its mineralogical province, on the whole, can be characterized as that of zoisite-garnet-zircon.

The middle Dnieper basin contains a considerable quantity of hornblende together with a high percentage of minerals of the epidote group. Rutile prevails among the stable minerals. Thus, here we have an epidote-hornblende-rutile province.

The mineral composition of loesses from the Oka and the Don plain differ from each other. The Oka basin is more akin to the middle Dnieper basin, and can be characterized as an epidote-hornblende-garnet province. In the middle Don basin minerals of the mica, epidote and rutile groups predominate. Thus, the mica-epidote-rutile province is represented in the east of the study area.

The delinated mineralogical provinces in Late Pleistocene loesses evidently came about because of geographic differences, as well as differences in the specifics of climate which occurred during the accumulation of loesses in the Late Pleistocene in European USSR *(Fig. 3)*.

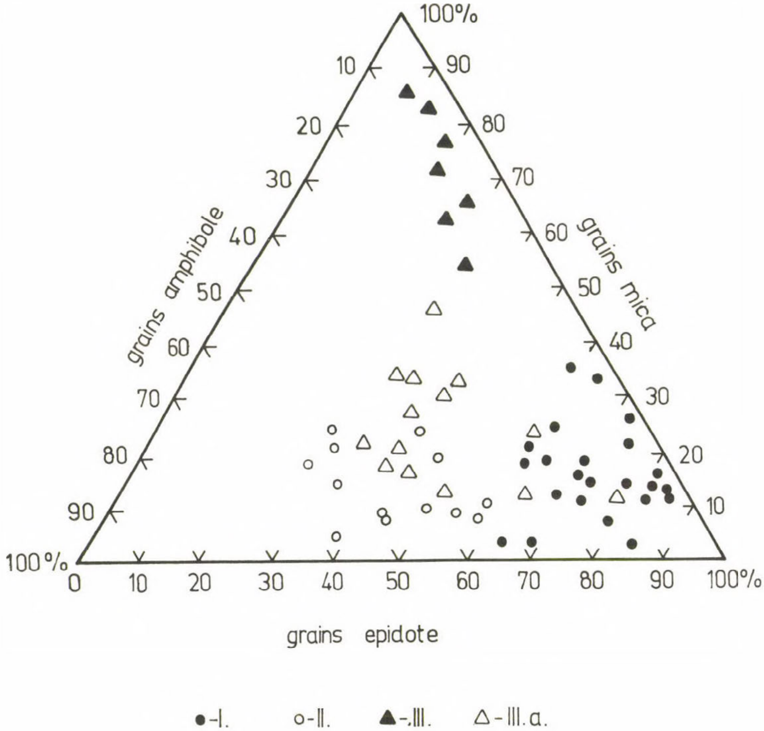

Fig. 3 Diagram of the distribution of rock-forming minerals of Late Pleistocene loesses in three loess provinces

I = the Volyno-Podolia, II = the Dnieper basin; III = the Don basin; IIIa = the Oka basin

The basic principle of mineral stability can be used to account for the mineralogical differences observed in the Valdai loesses

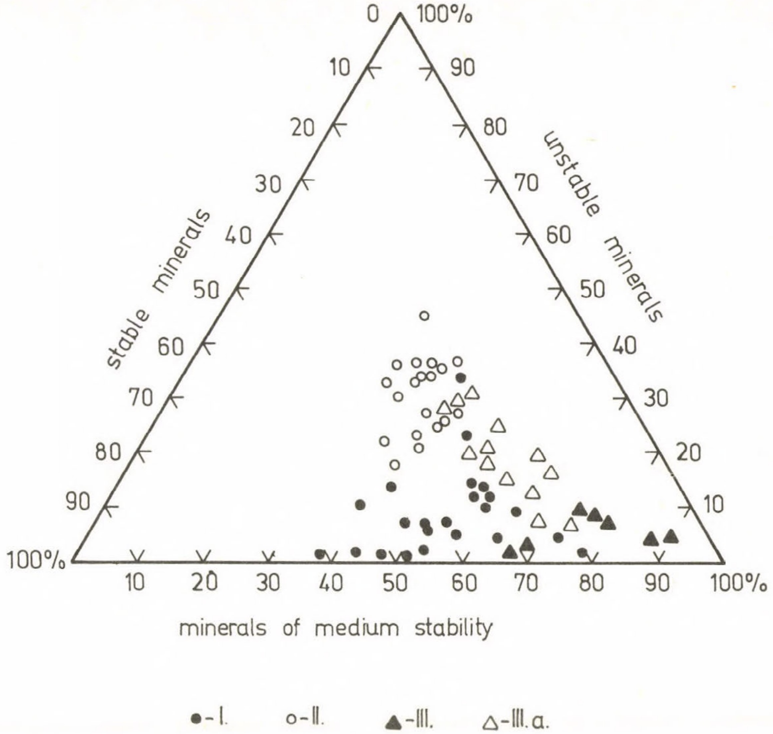

Fig. 4 Diagram of the distribution of mineralogical composition of Late Pleistocene loesses by the extent of their stability against weathering in the three loess provinces

I = the Volino-Podolia; II = the Dnieper basin; III = the Don basin; IIIa = the Oka basin

of different regions. In addition to other methods, the degree to which the layers have weathered (defined by the morphological properties of the grains, as well as by the correlation between mineral groups differing in stability) serves as an important lithological indicator of conditions extant when the deposits were formed.

Comparisons between the three "loess" regions, going from west to east, made on such variables as the extent of weathering of their mineral matter, their weathering coefficient values, and the degree of surface corrosion in mineral grains indicate the following reularities *(Fig. 4)*. The weathering of Valdai loesses in the three studied areas, on the whole, is not great (KHALCHEVA, T.A.--CHIKOLINI, N.I. 1982). The material is fresh everywhere, but the grains with clear crystallographic outlines occur rather seldomly. Traces of weathering are observed only on the grains of unstable mineral components: hornblendes, epidote and felspars. In general, only very stable minerals, such as zircon, tourmaline, garnet, and the titanium containing ones, are not affected by weathering. The values for weathering coefficients (K_I) - which express the correlation between rather stable mineral components (zircon and tourmaline) and unstable (hornblendes and pyroxenes) ones - are low. They average to 1.0 in the Volyno-Podolia, 0.32 in the Dnieper basin, and 0.45 on the Oka-Don plain.

It is evident that the pre-Bryansk and post-Bryansk stages of loess accumulation differed not only in the velocity but also in the extent of weathering. In the Dnieper basin, for example, where loesses are the most pronounced and the most typical, the K_I for the Khotylevo loess is 0.5 while average values for the Desna and Altynovo loesses are 0.32.

A north to south change in the extent of loess weathering was traced most thoroughly in the Dnieper basin *(Fig. 2)*. Here the degree of weathering in Late Pleistocene loesses shows little change latitudinally. This observation confirms the hypothesis that hyperzonal conditions existed in the whole extraglacial area during the period of Valdai glaciation (VELICHKO, A.A. 1973). A certain variation in the degree of weathering of loess mineral masses can, however, be observed in this broad hyperzone. These differences are manifest not only in the increase in erosional coefficients (K_I in the north is equal to 0.19, in the central part to 0.32, and 0.50 in the south), but also in both changes in grain surfaces and in traces of secondary alterations in the grains. The above differences indicate that insignificant climatic fluctuations existed within the limits of the hyperzone, and that they were associated with the differences in latitude.

In the Oka and Don basins, weathering remained approximately on the level typical for the area of maximal loess accumulation (the basin of the middle Dnieper). This is indicated by good preservation of the mineral materials, the absence of traces of weathering on grains, and by low coefficients of the latter (K_I = 0.45).

The extent of weathering of Late Pleistocene loesses increases to 1 in the western part of the loess area (on the Volyno-Podolian upland). Grains of unstable mineral components become thinner at the edges, and an accumulation of the weathered clay materials is observed along the cracks of cleavages as well as on the surfaces. These changes make their use for diagnoses difficult.

CONCLUSION

The above observations permit us to make the following conclusions: the loess area of the European part of the USSR reflects the existence of a wide hyperzone during the Valdai glaciation (VELICHKO, A.A.--KHALCHEVA, T.A. 1982). Both the mechanisms of loess accumulation and extant paleoclimatic conditions contributed to the formation of deposits with very similar sets of minerals and with poor weathering of the mineral masses.

The prevalence of specific mineral distinguish these regions and characterize the differences in the condition extant during periods of loess accumulation.
On the whole, the low degree of weathering of the mineral masses (which, in a way, is a result of paleoclimatic conditions during loess formation) varies in relation to the geographic position of the region. The data show that some regional differences existed within a singly hyperzone including somewhat higher humidity in the west (Volyno-Podolia) and increased aridity in the east (the Dnieper, Oka and Don basins). Furthermore, we have shown that the Oka-Don loess region is not homogenous in mineralogical contents. The Oka basin is closer to the Dnieper basin than to the Don. Finally, our mineralogical studies indicate the existence of three loess regions in European USSR: the Volyno-Podolia, the Dnieper and Oka basins, and the Don basin.

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GEOCHRONOLOGY OF LOESS IN CENTRAL ASIA AND QUATERNARY EVENTS

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ABSTRACT

The paper shows the relation of geotectonic, paleoglacial and paleoclimatic events with the loess forming processes. This connection has been confirmed by the available geochronological data. Specific position of Paleolithic finds in loess-paleosol sequences permits to elucidate some features of ancient human migration and reflect paleoenvironmental changes.

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INTRODUCTION

The problem of loess origin has always been in the focus of attention of research into the mountains and deserts of Central Asia. The works of F. RICHTHOFEN, P.V. MUSHKETOV and V.V. OBRUCHEV were particularly significant for developing ideas on loess formation. In their opinion, shared by their followers, typical loess was formed under arid and cool climatic conditions with the eolian factor controlling the accumulation of dust material. Loess and loess-like mantles in Central Asia differ in physical properties, thickness and age. The absence of consensus on the origin of subaerial mantles gives rise to heated debates on the significance and length of geological processes controlling the loess formation. Recognition of "warm" loess facies sometimes results in attempts by paleo-reconstructions which render it impossible to correlate directly loess-forming processes and glacial events. It follows from the above that the significance of geochronological data obtained from loess studies in Central Asia during the last decades is steadily growing. The regions under study include primarily the USSR Central Asia, the Kashmir Valley in Northern India, the loess areas of the Potwar Plateau in Northern Pakistan and the Loess Plateau in Northern China.

The available data on the age of loess mantles allow to perform preliminary interregional correlations and to relate the main stages of loess formation to Quaternary geological events which caused or accompanied loess-forming processes.

GEOTECTONIC EVENTS, GLACIATION AND LOESS

The latest 3-4 m.y. in Central Asia are known to witness continentalization and aridization of paleoclimate. These events were considerably influenced by the development of mountain structures and increased isolation of inner Asian areas from the free access of moist air masses from seas and oceans. The analysis of the recent tectonic movements and the identification of tectogenetic phases in Central Asia permit the following boundaries to be drawn. The tectonic phases in Soviet Central Asia are distinctly confined to the turn of the early and late Pliocene (ca. 4-3.5 m.y.), Late Pliocene/Eopleistocene (ca. 2-1.8 m.y.), Eopleistocene/Pleistocene (ca. 0.8 m.y.)[1]. Reactivation of tectonic movements in also evident in the late Pleistocene (the last 130 th.y. - DODONOV, A.E. 1978, 1986). Available geological evidence suggests correlation of the phases of diastrophism in Northern India and Northern Pakistan with similar ones in Soviet Central Asia. According to the data presented by BURBANK, D.W. and JOHNSON, G.P. (1982, 1983), the phases of tectogenesis have been traced in Kashmir at stratigraphic levels of 4 and 1.8 m.y. B.P. and the active orogeny has been recognized during the last 350 th.y. A phase of diastrophism is confidently identified in Northern Pakistan at the boundary of 2 m.y. B.P. (BURBANK, D.W.--RAYNOLDS, R.G.H. 1984). Phases of reactivation of tectonic movements have been manifested in Northern China at the boundaries of 3.5, 2.4, and 1.2 m.y. (WU Xihao, 1983).

The growth of mountain structures resulted in the development of high-mountain glaciation and the appearance of periglacial zone. This preconditioned the accumulation of large amounts of thin material in nival zone, which was then transported fluvially to foothill plains where it was deflated and carried by wind to piedmont slopes and watersheds.

The presence of upper Pliocene moraines in Eastern Pamir is evidenced from paleomagnetic data (PENKOV, A.V. et al. 1976). Ancient glaciations comparable in time with the formation of the Tatrot, Pinjor and Boulder Conglomerate sequences have been assumed in the Himalayas (DE TERRA, H.--PATERSON, T.T. 1939). The Tibet uplifting for not less than 3000 m inferred during the late Pliocene and Pleistocene (LIU Tung-sheng, 1986), has also permitted mountain glaciation and enlarged the scopes of physical weathering, which contributed to the formation of large amounts of silt material redeposited in the form of eolian dust at the foothills of Kunlun and in the Huang Ho valley.

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[1] The stratigraphic scale compiled by NIKIFOROVA, K.V. et al. (1984) is used for the last 4 m.y.

It is natural to expect the paleoclimatic and paleogeographical environment and especially the loess-forming stages to undergo considerable changes which coincided with the events mentioned above. Taking into account the age of the oldest loess horizons, a relative synchronism can be traced in tectogenesis, mountain glaciation and loess formation.

The age of stone loesses[1] in most complete sections in South Tajikistan is assumed ca. 2.0-2.4 m.y. (DODONOV, A.E. 1986). In Northern China, according to the data of Chinese scientists (ZHANG Zonghu, 1984; LIU Tung-sheng, 1986), the age of the oldest loesses is estimated at about 2.4 m.y. In Kashmir valley the age of the loess mantles which accommodate up to 10 paleosols (AGRAWAL, D.P. 1982) appears to stay within the Brunhes paleomagnetic epoch. In the Potwar Plateau of Northern Pakistan, with reference to the data presented by DE TERRA, H. and PATERSON, T.G. (1939), the loesses as thick as 60 m occur, while their geological setting suggest their origin in the Brunhes epoch. In southeastern West Siberian Plain - the region adjacent to the Soviet Central Asia - the oldest loess horizons are dated at about 0.7-0.6 m.y. (ARKHIPOV, S.A. et al. 1982; VOLKOV, I.A.--ZYKINA, V.S. 1982). The boundary marking the start of loess formation in each of the above regions is assumed to manifest critical desiccation and cooling. If the earliest boundary of loess formation in Northern China and the Soviet Central Asia is tentatively drawn at ca. 2.4-2. m.y. B.P., it most probably coincides with the phase of tectogenesis dated at ca. 2.4 m.y. in Northern China and at 2 m.y. - in Central Asia. It can also be correlated with the beginning of the cold Pretiglian stage in Western Europe (2.3 m.y.). The geochronological boundary drawn at ca. 0.8-0.7 m.y., which is associated in some regions with the beginning or intensification of loess material accumulation, manifests the significant phase of tectogenesis at the same stratigraphic level. The analysis of data on the origination of loess formation in different areas of Central Asia shows that the diachronism of its initial stage amounts to more than 1 m.y., that is, the stages of development of adequate Quaternary geological processes occurring in different belts and landscape zones were evidently heterochronus.

LOESS GEOCHRONOLOGY, CORRELATION AND SOME PALEOCLIMATIC EVENTS

Now we shall turn to a more detailed stratigraphy of loess horizons and buried soils. At present the loess areas in Central Asia lack the generally accepted regional loess-paleosol stratigraphic scales, though there already appear certain premises for their compilation.

The uppper Pliocene-Eopleistocene subaerial sections of Soviet Central Asia, in southern Tajikistan and the Tashkent area feature a significant number of red, red-brown and brown soils. Up to 10 red and red-brown paleosols have been recognized

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[1] Pink coloured sandy silt and loam locally intercalated with gravel.

close to each other in some sections (Karamaydan, Khonako) below the Olduvai paleomagnetic event. The Eopleistocene - from the base of the Olduvai event to the level of 0.75 m.y. - represents 25 horizons of red-brown and brown soils (the Chasmanigar section), conventionally referred into nine pedocomplexes, from the XI to the XIX (DODONOV, A.E.--PENKOV, A.V. 1977). Pleistocene loess-paleosol sequences include 10 pedocomplexes and ten loess horizons. The Xth (750 th.y.), VIIth (ca. 300 th.y.), VIth (ca. 200 th.y.), Vth (ca. 120 th.y.) and IInd (25 th.y.) pedocomplexes serve as regional marker horizons. Their dating is based on paleomagnetic, thermoluminescence as well as archaeological evidence. Relatively well-studied Paleolithic monuments belonging to the VIth (Karatau) and Vth (Lakhuti) pedocomplexes correspond to the epoch of Premousterian pebble culture, thus dating these pedocomplexes to earlier than 70 th.y.[1]).

The Wucheng subaerial formation is recognized in Northern China within the range of the Matuyama paleomagnetic epoch, below the Jaramillo event (HELLER, F.--LIU Tung-sheng, 1982). It is represented by alternating red and red-brown paleosols and loess horizons containing carbonate concretions. The structure and paleomagnetic characteristics of the Wucheng subaerial sequence allow its correlation with the Eopleistocene loess-soil formations of Southern Tajikistan. The younger Lishi and Malan sequences of the Loess Plateau in China are dated by Chinese scientists to the Middle and Late Pleistocene (after the European scale), resp. The transition from the Wucheng to Lishi loesses traced slightly below the Jaramillo event at approximately 1.1 m.y., has been adopted as the boundary between the Lower and Middle Pleistocene (HELLER, F.--LIU Tung-sheng, 1982). The Lishi loess comprises distinct brown fossil soils alternating with loess horizons. The Matuyama-Brunhes inversion serves as the significant datum mark when correlating loess and paleosols horizons in the sections of South Tajikistan and the Loess Plateau in China. Its position in the section permits to determine the number of paleosols (or pedocomplexes) formed during the last 0.73 m.y.

The Louchuan core penetrating the loess-paleosol section determines the Matuyama-Brunhes inversion at the level of the 8th buried soil (HELLER, F.--LIU Tung-sheng, 1982). Other sections of the Loess Plateau (Pingliang, Xifeng, Lanzhou) show not less than 9 buried soil horizons comprised within the Brunhes epoch (ZHANG Zonghu, 1984; BURBANK, P.W.--LI Jijun, 1985). It is noteworthy that the Lanzhou section in the western Loess Plateau shows the Matuyama-Brunhes inversion at 180 m depth, that is the loess sequence existing within the Brunhes epoch is two times as thick as the analogous one in central parts of the Loess Plateau and in Soviet Central Asia.

The Malan loess on top of the subaerial section of the Loess Plateau is usually referred to the interval between the Ist buried soil and the recent soil. The black soil in the middle part of the Malan loess divides it into lower and upper horizons. Dating of the Malan loess is still debated. It is

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[1] According to V.A. RANOV, the boundary between the Mousterian and final Acheulian in Central Asia can be drawn at the level of 70-80 m.y. (DODONOV, A.E.--RANOV, V.A. 1984).

considerably controlled by the dating of the Ist fossil soil. In case the Ist paleosol of the Louchuan section is correlated with the 5th stage of the oxygen-isotope scale (HELLER, F.--LIU Tung-sheng, 1982; LIU Tung-sheng et al. 1985), the Malan loess should be dated to the interval from ca. 75 th.y. B.P. to the Holocene. According to another version, the 4th, 3rd and 2nd paleosols in the Louchuan section are correlated with the 5th stage of the oxygen-isotope scale (LIU Zechun, 1982). This suggests correlation of the Ist paleosol with the interstadial of the last glaciation, thus assuming the Malan loess to be still younger. We agree with the second version. It is possibly the Malan loess (the interval between the Ist buried soil and the recent soil) that corresponds to the last glacial maximum (18 th.y. B.P.).

The last glacial maximum in Soviet Central Asia saw the formation of a loess horizon up to 15 m thick, separating the IInd pedocomplex and the recent soil. The Ist pedocomplex of grey immature soils in the sections of South Tajikistan probably corresponds to the black soil horizon in the Malan loess. The youngest loess mantle is the most extensive in the foothills of Middle Asia. The accumulation rate of dust material at the end of the Late Pleistocene amounted here to 1-2 mm per year.

Notable are the easily-compared data pointing to the fact that the wind activity and the accumulation of dust material in the desert-loess belt of southern and northern hemispheres were intensified during the last glacial maximum (18 th.y.). The data of J.M. BOWLER has shown that the period of intensified eolian activity and loess-forming processes in Central and Southern Australia at ca. 25-14 th.y. B.P. coincided with those in China and New Zealand (The Report of Workshop held in Australia, 1980). The formation of the loess horizon in New Zealand dated at 25-10 th.y. B.P. is referred to the same interval (MILNE, I.P.G.--SMALLEY, I.J. 1979; PALMER, A.S. 1982). According to the data of Chinese scientists the sharp change from mild to more severe paleogeographic environment occurred in Northern China at the boundary of 26-23 th.y. B.P. Meanwhile, anticyclone air masses gained force and at the same time were shifted southward, while the influence of the summer monsoon on the continent decreased. As the deserts moved to the south, loesses extended to vast areas of Northern China simultaneously with the advance of high-mountain glaciers (LIU Tung-sheng et al., 1985). The last glacial maximum witnessed considerable extension of sand deserts, their advance to savannas and tropical forests as well as their movement towards the poles (SARNTHEIN, M. 1978). The paper of D.W. PARKIN and N.J. SHACKLETON (1973) illustrates the relationship between the intensified eolian activity in nearland atmospheric layers and the deterioration of the climate. It shows that the amount of dust carried over from the Sahara towards the Atlantic increased during the intervals corresponding to the "cold" stages of the oxygen-isotope scale.

In the light of regional and global comparison the correlation of loess-paleosol sections in Northern China, Central Asia, Europe and North America with the oxygen-isotope scale is of considerable interest (KUKLA, G. 1978; HELLER, F.--LIU Tung-sheng, 1982; ZHENG Honghan, 1982; DODONOV, A.E.--RANOV, V.A. 1984). Such correlation confirms the complicated structure of global climatic changes within the last 1.8 m.y. and multiple glacials and interglacials. Up to 20 cold and warm intervals have been recognized during the last 0.7 m.y. The last 100 th.y. are assumed to show stadials and interstadials of the Würm (Vistulian, Wisconsian) glaciation. The time span of 700-100 th.y. B.P. was characterized by the alternation of glacials and interglacials, the latter being often represented by several optimums.

Correlation of loess-soil scales and the oxigen-isotope curves is mainly possible due to such paleomagnetic datum marks as the Olduvai and Jaramillo events and the Matuyama-Brunhes inversion. It should be noted that correlational reconstructions are insufficiently controlled geochronologically, especially those performed for the loess-paleosol sequences within the range of 700-100 th.y. The stages of the oxygen-isotope scale has been dated from calculating deep-sea sedimentation rates on condition that radiogeochronological control of the youngest stages and the paleomagnetic reference points at the level of 730 th.y. were provided. Such age estimation is difficult for the units of the loess-paleosol scale due to irregularity of loess accumulation, since soil-forming epochs showed the minimum accumulation rate of thin dust particles. Besides, the rates of loess-formation tended to increase through the Eopleistocene to the end of the Pleistocene reaching their maximum during the extreme cooling in the Late Pleistocene.

The structure of loess mantles and the Paleolithic archaeology have reflected the events related to environmental changes and distribution of early man. We are reminded that the absence of finds characteristic of the earliest studies of Quaternary geology in Central Asia made some scientist reject the possibility of Paleolithic in the desert-loess belt due to severe climatic conditions in spite of the fact that the Neolithic sites are known from Northern China and Mongolia. At present the Paleolithic sites in Hihoudou, Lantian, Gunvalin, Kehe, Dingcun, Shuindungou, Sjara-Osso-Gol, etc. are known in the basin of the middle Huang He in loess and underlying deposits. They characterize the stages of distribution of ancient man in the desert-loess belt from 1-0.7 m.y. up to the Neolithic. Up to the 70's the Paleolithic monuments in Soviet Central Asia have been discovered on terrace surfaces (the Kayrakkum, Ak-Jar, Kara-Bura ones), in caves (the Teshik-Tash, Obi-Rakhmat and Oghi-Kichik ones) and in thin subaerial mantles developed on terraces (the Kulbulak, Shugnou and Samarkand sites) (RANOV, V.A.--NESMEYANOV, S.A. 1973). Their age stays within 50-60 th.y., by stratigraphic data. However, in the 70's a number of Paleolithic finds of a new type have been discovered in South Tajikistan. These were confined exclusively to paleosol. in thick loess sequences. These finds shifted the start of the Paleolithic in Soviet Central Asia to at least 0.8 m.y. B.P. The well-studied newly-discovered Paleolithic sites include the Karatau site (finds from the VIth pedocomplex - about 200 th.y.), Lakhuti (the Vth pedocomplex -about 130 th.y.), Khonako (the VIth and Vth pedocomplexes) and Kuldara (XIth pedocomplex - about 800 th.y.) (LAZARENKO, A.A.--RANOV, V.A. 1977; DODONOV, A.E.--RANOV, V.A. 1976; DODONOV, A.E. et al., 1978; RANOV, V.A. 1982). These Paleolithic monuments are characterized by stone artifacts possessing well-expressed features of East-Asian pebble cultures (RANOV, V.A. 1980). It is typical of all pebble tools found in paleosols to form no clear cultural horizons. The stone material is distributed over the whole profile, being confined to the optimal part of paleosols. Attention should be paid to the fact that while studying the loess-paleosol sections of Soviet Central Asia no Paleolithic finds have been recovered in proper loess horizons. It brings us back to the assumption that in the desert-loess belt at the early stages of Paleolithic man distribution the climate of loess-forming epochs was hardly favourable, while the epochs of soil formation were characterized by optimal climatic environment suggested by the stone artifacts associated with paleosols. The periods of loess formation featured poorer faunal and floral kingdoms and worse climatic conditions (desiccation, severe winters, frequent dust storms) as compared to those of soil formation. This probably resulted in ancient people migrating from loess areas and populating other regions. Accounting for frequent climatic fluctuations during the Pleistocene one may infer such migrations to occur repeatedly.

CONCLUSIONS

Phases of tectogenesis which caused the growth of mountain ranges in Central Asia contributed to the continentalization of its inner areas. Major neotectonic phases date back to ca. 4-3.5 m.y., ca. 2.4-1.8 m.y., ca. 0.8 m.y. and the last 0.3-0.1 m.y. B.P. The data analysis on the start of loess formation in different region shows its considerable heterochronism, reaching more than 1 m.y. The distribution of Paleolithic sites of the early stages of the ancient man in Soviet Central Asia were evidently confined to the soil forming intervals, i.e. to optimum paleoclimatic conditions.

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PALEOGEOGRAPHY OF BLOWN SAND IN HUNGARY

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ABSTRACT

Blown sand surfaces amount to almost 20 per cent of the area of Hungary. Most of them accumulated on alluvial fans. Blown sand movements primarily produced wind furrows, oval-shaped sand hummocks and residual ridges. Parabolic dunes developed, first of all, in the Nyírség. Until the seventies the age of dunes could be estimated from stratigraphical evidence and palynological investigations. ^{14}C dating allowed us to conclude that the Upper Pleniglacial was the period of maximum blown sand movement. In smaller areas considerable movements took place in the Older and Younger Dryas. In the Boreal phase of the Holocene a much lesser area was affected by movements than it was assumed earlier. During the investigation of stratigraphy in the Great Hungarian Plain, several layers of blown sand were found in the Danube-Tisza interfluve. They indicate sand movements even prior to the Upper Pleniglacial in the Great Plain (locally even in the Lower Pleistocene).

INTRODUCTION

Related to the area of the country, Hungary has much blown sand. Blown sand surfaces amount to almost 20 per cent of the area *(Fig. 1)*. This fact itself explains the long tradition of blown sand research in Hungary. J. CHOLNOKY (1902, 1907) achieved remarkable results in the investigation and genetic classification of blown sand landforms as early as the first decades of this century. Later too he laid emphasis on the study of blown sand areas, as this is proved by his work issued in 1940 (CHOLNOKY, J. 1940).

In the 30's a new impetus was given to this research by the activity of L. KÁDÁR (1934, 1938). Among others, it was him who called attention to parabolic dunes and described them from Hungary (KÁDÁR, L. 1951).

Beginning in the early 50's research activites also intensified in blown sand regions. They first only involved the environs of Budapest, the Danube-Tisza interfluve, the Nyírség and the Mezöföld plains, but were later extended to all the

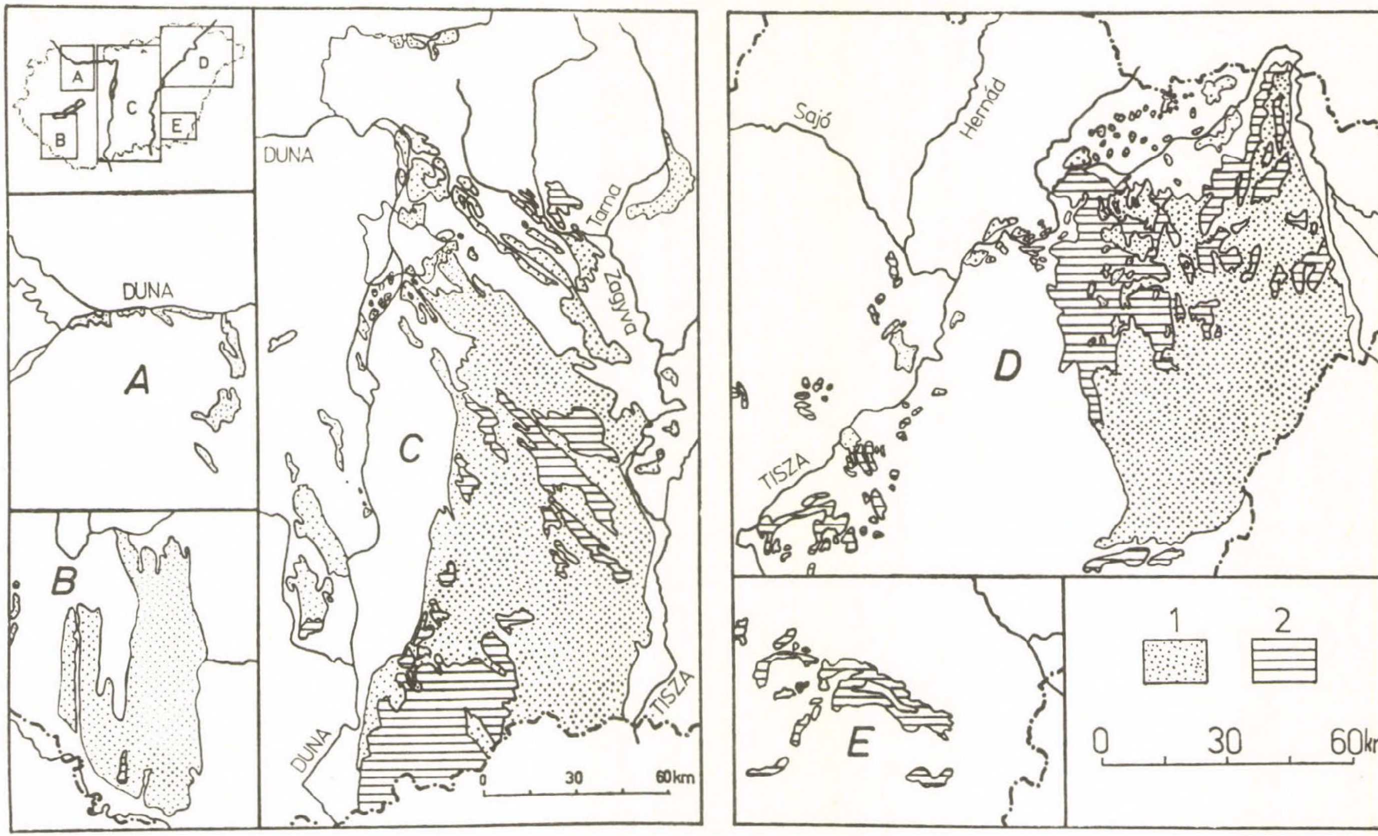

Fig. 1 Hungarian blown-sand territories

1 = loose blown-sand; 2 = blown-sand covered with loessy blanket

the blown sand areas. (BULLA, B. 1951; KÁDÁR, L. 1951, 1956, 1966; MAROSI, S. 1955, 1958, 1967, 1970; PÉCSI, M. 1960, 1967; BORSY, Z. 1961, 1966, 1968a, 1968b).

Numerous new data were added to the knowledge on the origin of blown sand areas in Hungary and useful information for practice. In some issues, however, conflicting opinins arose and intense debates were provoked. The debates also had their merit in encouraging further research.

A new comparative investigation of blown sand regions (BORSY, Z. 1974, 1977a; LÓKI, J. 1981) enabled us to provide data, more precise than ever, on the evolution history of landforms.

It became evident as early as the early 50's that most of the blown sand in Hungary formed on Quaternary alluvial fans. Although blown sand occurs on river terraces too (mainly along the Danube) it has a limited extension.

Stratigraphical and palynological analyses helped us to determine the age of blown sands: most of them were formed during the last (Würm) glacial (Upper Pleniglacial). The dates of subsequent sand movements, however, were not established for certain. There had been an enduring view that the Boreal phase saw intensive sand movements in several parts.

During the major sand movements of the Upper Pleniglacial primarily wind furrows, oval-shaped sand hummocks and residual ridges appeared. Parabolic dunes only covered large areas in the Nyírség and in the Danube-Tisza interfluve. However, in the latter area they have a lesser role in the assemblage of landforms.

The above underline the different nature of Hungarian blown sand areas from those in the German-Polish Plain, Denmark, or the Netherlands as the latter are mostly characterized by parabolic and coastal dunes.

MAIN PERIODS OF SAND MOVEMENT DURING THE UPPER PLENIGLACIAL

The lack of radiometric data concerning the age of dunes caused more and more difficulties in our research during the seventies. Although paleosols were frequently found in dunes, they did not contain any charcoal. The first dune with charcoal remnants suitable for 14C analysis was found in the Bodrogköz in 1978 *(Fig. 2)*. Subsequently, our efforts were successful in several parts of the north-eastern Great Hungarian Plain *(Fig. 3)* and an ever more precise picture was formed of the origin of dunes on the alluvial fans in Hungary.

During the Upper Pleniglacial (35,000-28,000 years BP, after the Stillfried interstadial) climate turned more arid and colder again. Under this climate the loose deposits of alluvial fans and terraces were only covered by sparse, cold steppe vegetation. This vegetation cover was unable to protect the surface from strong NW, N, and NE winds; sand movements resulted and a variety of blown sand features developed.

The first major wave of sand movement lasted for cca 5000 years and radically transformed alluvial fan surfaces. The

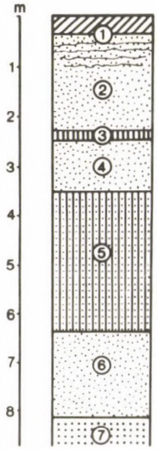

Fig. 2 Segment of the exposure SW of Vajdácska

1 = brown forest soil with iron-pan layers;
2 = blown-sand from the older Dryas;
3 = fossil soil from the Bölling interstadial;
4 = blown sand from the second half of Upper Pleniglacial period;
6 = blown-sand evolved in the first part of the Upper Pleniglacial;
7 = fluvial sand

most intensive sand movement producing features of largest dimensions affected the Nyírség and some parts of the Danube-Tisza interfluve. Over most of the interfluve and other alluvial fans sand movement was modest and the resulting blown sand surfaces display limited relief.

The first cold peak of the Upper Pleniglacial was followed by cca 1000-1500 years of milder and more humid climate. The dune surfaces and flat blow-outs of smaller relief were more efficiently protected by vegetation and the development of falling-dust mantles began in the Nyírség, the North Great Plain alluvial fan series, the Danube-Tisza interfluve, the Mezöföld, the Inner-Somogy and some patches of the Maros alluvial fan. They were getting thicker during the second half of the Upper Pleniglacial (disregarding the Lascaux optimum) since under the cold and dry climate of this phase much dust was blown from the alluvial fan sands, still in motion, and from alluvia. (In mountainous and hill regions frost weathering and the loosening affect of frost on soils produced much dust, which was carried by wind to great distances.) Under periglacial climate the dust fallen upon the dunes of the alluvial fans was diagenetized into loess or loess-like mantles. Its thickness amounted to 3-5 m by the end of the Upper Pleniglacial.

The loess mantle has preserved the dunes of Upper Pleniglacial age to our days. The strike of loess-mantled dunes clearly reflects the wind conditions prevailing in Hungary during the Upper Pleniglacial.

SAND MOVEMENTS IN THE LATE GLACIAL

Sand movement was considerably reduced in the blown sand areas of Hungary by the late Upper Pleniglacial. In the Bölling interstadial climate became milder and a little more humid. The

Fig. 3 Blown-sand areas of the NE part of the Great Plain

1 = blown-sand; 2 = blown-sand covered by loessy blanket; 3 = fossil soils from the Bölling interstadial; 4 = fossil soils from the Alleröd interstadial; 5 = fossil soils of Late Glacial age (Bölling, Alleröd); 6 = fossil soils buried in the early Subboreal

steppe or forest-steppe vegetation of dune surfaces became denser. Under cold continental steppe or forest-steppe conditions soils of chernozem dynamics formed.

During the only 500-600 years of the cold and dry Older Dryas this soil formation was interrupted in several places of the higher lying sections of the Danube-Tisza interfluve, the Nyírség, Bodrogköz (BORSY, Z.--CSONGOR, É.--LÓKI, J.--SZABÓ, I. 1985), and the Tarna alluvial fan and sand movements resumed. It is evidenced by the soil horizons of Bölling interstadial age *(Figs 2* and *4)* found in several places recently. The age of soils was estimated by the ^{14}C analyses of the incorporated charcoal remnants.

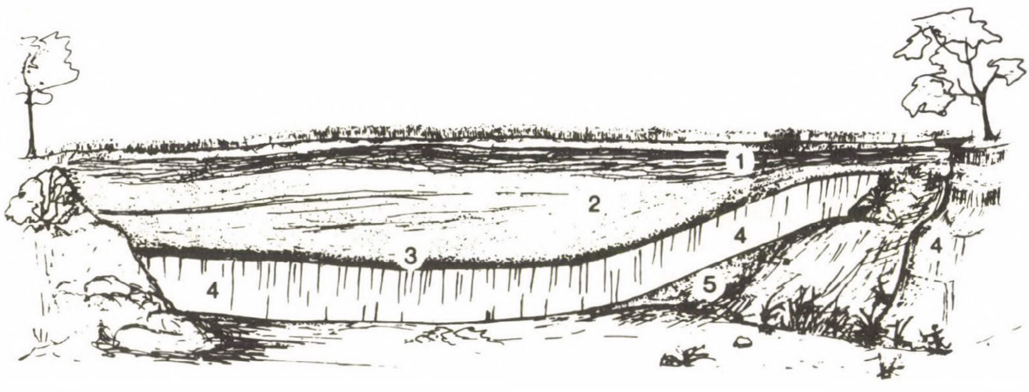

Fig. 4 Exposure W of Aranyosapáti

1 = brown forest soil with iron-pan layers; 2 = blown-sand evolved in the older Dryas; 3 = fossil soil from the Bölling interstadial; 4 = loess; 5 = blown-sand evolved in the first part of the Upper Pleniglacial

In this respect, one of the best exposures of the Great Plain is at Aranyosapáti (BORSY, Z.--CSONGOR, É.--SÁRKÁNY, S.--SZABÓ, I. 1982). It is clear there how the young blown sand of variable relief was deposited upon the previous loess-mantled dune surface in the Older Dryas.

The Older Dryas sand movement came to an end by the Alleröd interstadial. Under the milder and more humid climate of the Alleröd a more closed steppe vegetation developed on the loose blown sand too and sand movement practically ceased. During the Alleröd alluvial fans were covered by forest-steppe or steppe vegetation. In the steppes of the Great Plain chernozem-like soils formed on loess-mantled dunes, while steppe sand soils developed on sand.

The amelioration of the Alleröd interstadial lasted for cca 900 years. In the Younger Dryas another climatic deterioration took place and drier conditions prevailed. On higher-lying dunes or where groundwater was deeper, vegetation became sparser and sand movement started anew.

The sand in motion deposited upon Upper Pleniglacial features in several parts and buried them. Sand movement is attested by the Alleröd soil horizons buried under 4-10 m thick Younger Dryas blown sand *(Fig. 5)*.

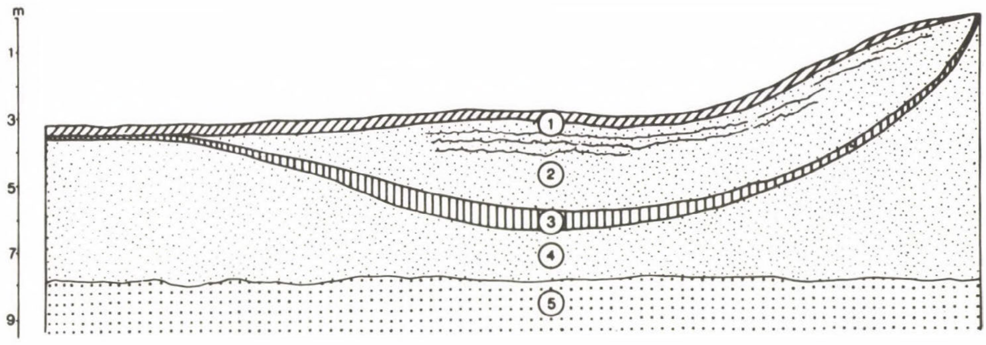

Fig. 5 Exposure at Bodroghalom

1 = brown forest soil with iron-pan layers; 2 = blown-sand from the younger Dryas; 3 = fossil soil from the Alleröd period; 4 = blown-sand evolved in the Upper Pleniglacial; 5 = fluvial sand

As a matter of course, the sand movements of the Late Glacial extended over much smaller areas than those in the Upper Pleniglacial. In the Late Glacial the overwhelming majority of loess-mantled dunes were stabilized. In blown sand regions movement were able to reactivize on drier surfaces, particularly on the elevated accumulations.

Another difference between the Late Glacial and the Upper Pleniglacial is that during the former sand surfaces were more protected and extended sand features could not come about. For instance, flat blow-outs and accumulated sand fields could not develop any more.

In the areas of major Older or Younger Dryas sand movements the identification of forms had been difficult as from below the Late Glacial sand mantle Upper Pleniglacial dunes may frequently outcrop.

The radiometric dating of paleosols occurring in the sand dunes was considerable help in the more precise subdivision of the Late Glacial in Hungary. The investigations provided unambiguous evidence to at least four major climatic changes in Hungary. Previously no data had been available on sand movements in the Late Glacial. The view was held that the paleosols in the dunes formed in the Preboreal and the overlying blown sand was dated Boreal.

The Preboreal phase followed in the Late Glacial by milder climate. Sand surfaces became increasingly stabilized and sand movement practically ceased. Although in the Boreal phase small-scale movement did take place, this did not alter the topography of sand-mantled alluvial fans.

Due to anthropogenic influence, sand movement began in the early Subboreal phase along the Danube and the part of the Nyírség bordering the Érmellék. However, it only affected limited areas. (Holocene sand movements are also evidenced by a thin buried soil horizon or pedified horizons. In some dunes two or three pedified horizons are observed. Since the archaeological finds are not suitable to date to paleosols precisely and no radiometric dating is available, the correspondence between paleosols of distant areas has not been accomplished.)

Considerable sand movements have not even been induced by deforestations in the Middle Ages. The situation changed from the 18th century, when clearances extended to dune regions of high relief. From these surfaces soil mantle was easily removed, particularly from loose sands. Subsequently, wind erosion intensified and sand movement began. This phase was characterized by the dissection of previously accumulated features. Intensive wind erosion turned the landscape into wilderness in some parts, especially in the southern Nyírség and the sand surfaces of higher relief in the Danube-Tisza interfluve. Blown sand has been stabilized by forests and shelter tree belts since the late 19th century.

It is striking that the dimensions of features formed in the 18-19th centuries are much less than those of sand features shaped during the Upper Pleniglacial of the Late Glacial.

Recently, data has also been collected for sand movements in Hungary prior to the Upper Pleniglacial. S. MAROSI (1965) mentioned it more than twenty years age that in Inner-Somogy sand movement was possible in the early Würm. Z. BORSY (1974) pointed out that in the southern Nyírség a core borehole traversed blown sand between the depths of 70 and 80 m.

SUBSURFACE BLOWN SAND LAYERS UNDER THE GREAT HUNGARIAN PLAIN

The electron microscopic analyses of sands from the Great Plain cores in the Körös region (BORSY, Z.--FÉLSZERFALVI, J.--LÓKI, J. 1982) and the Maros alluvial fan did not indicate the presence of blown sand (BORSY, Z.--FÉLSZERFALVI, J.--LÓKI, J. 1982; BORSY, Z.--FÉLSZERFALVI, J.--FRANYÓ, F.--LÓKI, J. 1985). In the Danube-Tisza interfluve, however, blown sand layers at depth are found in each borehole (BORSY, Z.--FÉLSZERFALVI, J.--FRANYÓ, F.--LÓKI, J. 1987). When interpreting electron microscopic analyses, the largest amount of blown sand was revealed below surface in the environs of Kecskemét *(Figs 6* and *7)*. Even in the early Würm this area was affected by eolian processes. This is reflected in the occurrence of blown sand and loess layers down to 38 m depth.

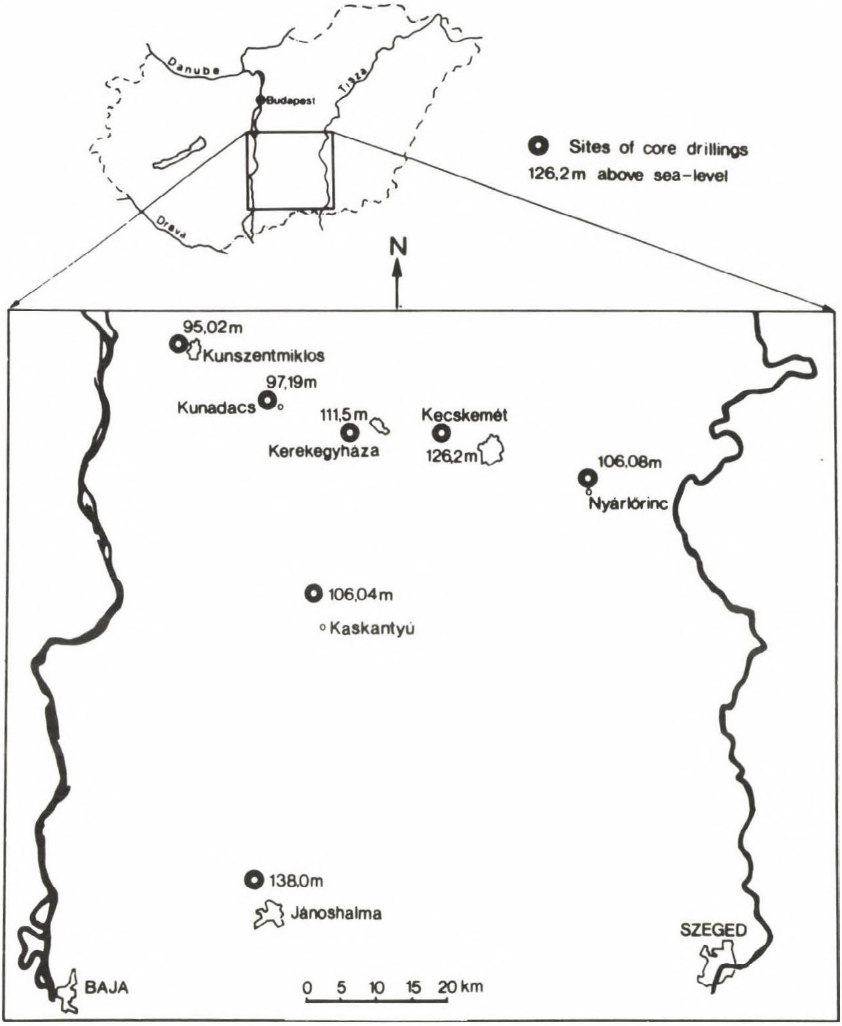

Fig. 6 Sites of the core drillings in the Danube-Tisza Interfluve region

In the Nyárlörinc core it is striking that blown sand is present there even in Lower Pleistocene layers. At Jánoshalma Lower Pleistocene layers are absent, but in the 130.3 m of Quaternary sequence thin blown sand layers repeatedly occur.

The subsurface sand layers in the cores of the Danube-Tisza interfluve attest to the fact that some parts of the Danubian alluvial fan temporarily became flood-free. During

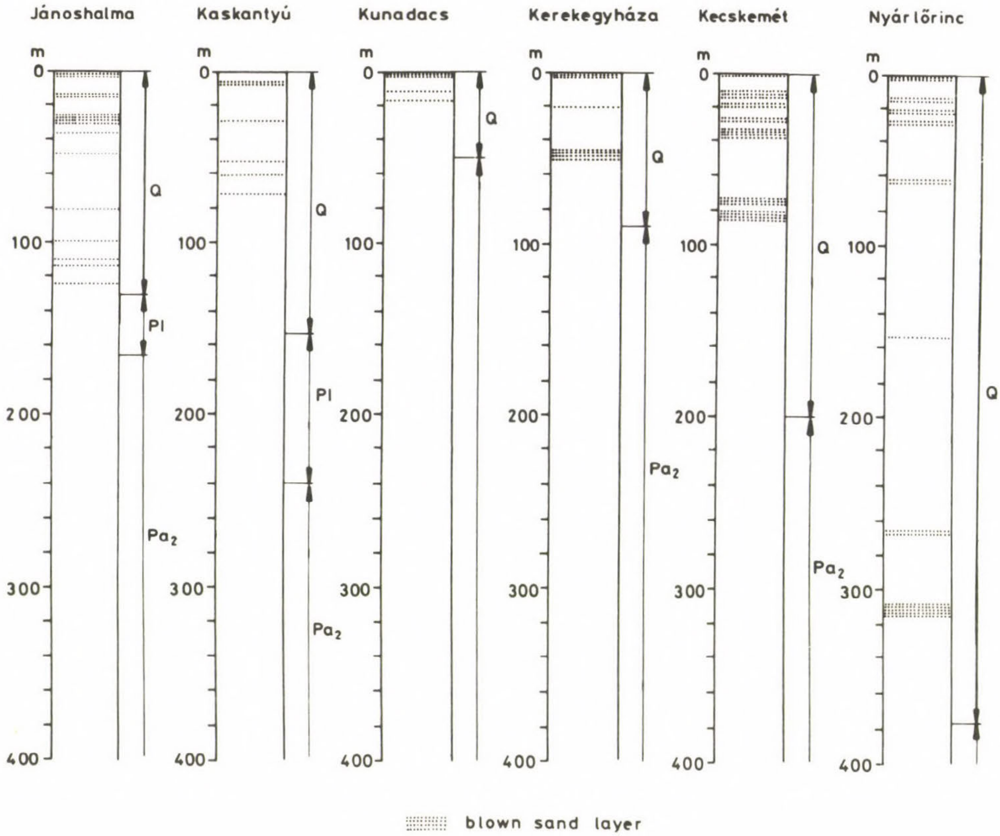

Fig. 7 Occurrence of eolian sand beds with sampling in core drillings set in the Danube-Tisza Interfluve

Q = Quaternary; Pl = Pleistocene; Pa_2 = Upper Pannonian

dry periods, for instance, in the glacials, surficial fluvial material was redeposited by strong winds and blown sand mantles of variable thickness formed.

Parallel to the above-mentioned investigations, an electron microscopic analysis was carried out for the sand layers intercalated in the Paks loess (BORSY, Z.--FÉLSZERFALVI, J.--LÓKI, J. 1984) and it was observed that repeated sand movements took place there too in the Würm and Riss.

CONCLUSION

The above achievements provide evidence for the presence of blown sand below the surface of the Great Hungarian Plain in several places and several layers. However, these layers are much thinner than, for instance, the surficial accumulative sand fields or the sand of parabolic dunes of up to 16-20 m height. This indicates that the most favourable conditions for blown sand accumulation existed during the Upper Pleniglacial.

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PALEOCLIMATIC RECONSTRUCTION OF THE LATE PLEISTOCENE IN THE UPPER AND MIDDLE DNIEPER REGIONS AND IN BYELORUSSIA FROM DATA ON SMALL MAMMALS

A.K. MARKOVA

ABSTRACT

Climatic reconstructions are made for some stages of the Late Pleistocene Mikulino Interglacial, Early Valdai, Bryansk Interstadial, and Late Valdai in the Russian Plain based on data of small mammals. The method of climatological diagram and microtheriological data were employed. The Mikulino Interglacial was characterised by quite high July and January temperatures similar to present values. Total annual rainfall was not much different from that today.
During the Early Valdai there was a sharp fall in winter temperatures, especially in Byelorussia. July temperatures dropped by nearly 5 °C. The amount of annual rainfall and solid precipitation considerably decreased. During the Bryansk Interstadial the natural and climatic conditions did not change considerably in comparison with the Early Valdai. In Byelorussia and the Dnieper region in the Late Valdai period severe continental conditions prevailed. Microtheriological data allow to detect the main climatic changes that took place during the Late Pleistocene in the central part of the Russian Plain.

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INTRODUCTION

Many climatic factors profoundly affect living organisms. For mammals these factors, first and foremost, include temperature values. Considerable rises or falls in temperatures may lead to the death of animals as well as to such a reduction in their activity that they are not able to reproduce and survive. Second, the amount of precipitation is also extremely important for living organisms since this conditions the state of the vegetation cover, the moisture capacity of plant fodder, the availability of functioning water basins, etc. Third, the depth of the snow cover is very important for the existence of big and small mammals since, to a large extent, it determines their possibilities to migrate and to obtain food (for large mammals). The snow cover often plays a protective role for small mammals - sheltering them against low temperatures, sharp temperature fluctations, strong winds, etc. Among many mammals of the tundra and forest-tundra zone even reproduction is partially performed under the protection of the snow cover.

Critical values for the depth of snow cover for different species of ungulates in the USSR are summarized by A.A. NASIMOVICH (1955) and A.N. FORMOZOV (1976). These authors concluded that the critical depth of snow cover is 80 or 90 cm for reindeer, 90 or 100 cm for elk, 50 or 60 cm for red deer, 30-40 0- for European roe, 30-40 cm for sika and wild boar, 40 cm 40 horses, and 20 cm for saiga.

The presence of a frozen snow crust during thaws is a considerable obstacle for many species of mammals which often leads to their death and hinders their migration and food extraction. Furthermore, it prevents small mammals from getting out from under the snow.

S.S. SHWARTS and his colleagues have shown the extent of adaptive responses in organisms to life in various environments (e.g. tundra, steppe, etc.) (SHWARTS, S.S. /ed./ 1976; SHWARTS, S.S. 1980; SHWARTS, S.S.--PYASTOLOVA, O.A. 1971 and others). Their studies of exchange processes is mammals prove that mammalian adaptation to extreme tundra conditions occurs not only through the development of exterior adaptational abilities (such as a dense fur cover, transformation of the extremities, etc.), but also primarily through 1) improvement in physical thermoregulation, 2) economizing of matter exchange, 3) an increased ability to accumulate energy, 4) a special pattern of vitamin exchange, and 5) through an increased ability to change physiological functions in response to sharp changes of living conditions (i.e. a profound change in tissue metabolism).

These features are characteristic for the endemic taxa of tundra, steppe, and mountain areas. Widespread species (such, for instance, as the narrow-skulled vole, the tundra vole, and the water vole) choose a different way of adapting - by migration to areas where microclimatic conditions correspond to the requirements of their organisms.

Thus, every natural zone contains taxa which are adapted to a specific environment not only by their exterior adaptational properties but also by the specificity of their metabolisms which are directed to the most economic exchange of substance under given conditions of life.

In tundras, these processes are primarily directed to economizing the organisms' matter exchange. In steppes, semideserts, and deserts, they are expressed in an increased resistance of tissues to dehydration. Typical small mammals of the steppes *(Lagurus lagurus, Microtus socialis, Microtus brandti)* do not hibernate, are closely associated with zonal types of vegetation, are well adapted to feeding on moisture deficient xerophytic types of vegetation, and are able to exist without water for a long time (FORMOZOV, A.N. 1976). Ungulates of the steppe zone (saiga, Persian gazelle, onager) have good eyesight and senses of smell. They can run at high speeds which is an adaptational reaction to open landscapes. There are many other similar adaptational reactions which include also those on the issue level.

Adaptations of animals to life in forests include, on the one hand, adjustments to feeding on specific forest products, and, on the other, the development of numerous ethological abilities. For example, the existence of animals in the family of dormice is controlled by the availability of nut and fruit trees and of bushes. The common field mouse and the yellow-necked mouse are adapted to feeding on acorns and seeds of arboreal plants. The latter have learned to climb trees to get their favourite food, etc.

The above examples show that interrelationships between mammals, natural environment, and climate are, on the one hand, very complicated and, on the other, very diagnostic.

Many mammals who are highly specialized for certain environments can serve as good indicators of these environments, and of the climatic conditions in particular.

The discovery of such indicator species of mammals at sites can help us reconstruct both the natural environments and, in part, even the climate extant at the time the site was formed.

Some species of small mammals are especially diagnostic and were considered in making our climatic reconstructions of some Late Pleistocene stages.

DISCUSSION

In this study we examined four Late Pleistocene stages on the Russian Plain: 1) the Mikulino Interglacial; 2) the epoch of the Early Valdai; 3) the Bryansk Interstadial (30,000 to 25,000 years ago), and 4) the Late Valdai *(Fig. 1)*.

We use both our own materials as well as extant literature on fossil small mammals to reconstruct these sections. The distribution of the available data for the study area is very irregular. Because of this we based our climatic reconstructions on the following two key regions: Middle and Upper Dnieper region (small mammal faunas of this region have been extensively studied by the author who augmented them with data from the literature) and Byelorussia. Microtheriological faunas from Byelorussian Late Pleistocene sites were studied by A.N. MOTUZKO (1981, 1985) and P.F. KALINOVSKY (1979, 1980, 1983).

Climatic reconstructions were done by combining climatological diagrams with data on microtheriological faunas. The method used in making climatological diagram consists of compiling climatic areas for every species discovered in the deposits of microtheriological fauna and of determining the centers of their co-occurrence. These centers of concentration illustrate the most probable climatic features extant during specific Late Pleistocene stages. Such a method of compiling climagrams has been suggested by J. IVERSEN (1944) and is widely used in palynology. Recently it has been increasingly used for interpreting theriological materials as well (MARKOVA, A.K. 1975, 1982; AGADZHANYAN, A.K. 1982).

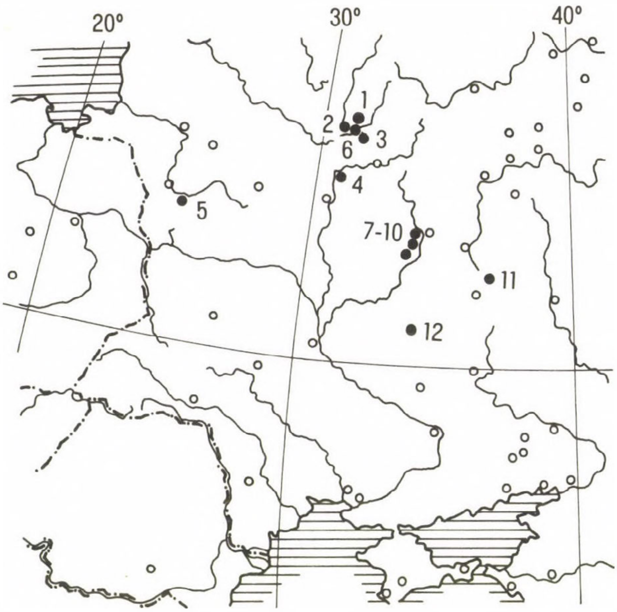

Fig 1 Late Pleistocene small mammal sites

1 = Drychaluki; 2 = Borisova Gora; 3 = Diskeninovo; 4 = Pashino; 5 = Timoshkovichi; 6 = Gralevo-2; 7 = Betovo; 8 = Arapovichi; 9 = Chulatovo; 10 = Novgorod-Seversky; 12 = Malutino; 12 = Gadyach

Climatic areas were outlined for the most representative strictly specialized species - endemics of certain geographical zones. Data on widespread intrazonal taxa were not used. In most cases we used reconstructed areas for mammals. These indicate the distribution of the different species which existed prior to anthropogenic impact.

The climagrams permitted us to calculate the parameters for the four Late Pleistocene epochs presented in *Table 1*. The species composition of mammals from the relevant deposits is given in *Table 2*.

Data on the occurrence of microtheriological faunas in the Middle and Upper Dnieper region and in Byelorussia suggest that the Mikulino Interglacial was characterized by quite high July temperatures similar to those recorded today *(Table 1)*. Theriological data from the Malyutino site (basin of the river Seim) indicate that mean January temperatures also did not differ from modern values.

The same is true for January temperatures in Byelorussia based on microtheriological data from the deposit at Borisova Gora (basin of Zapadnaya Dvina). This fauna was studied by P.F. KALINOVSKY (1983) and A.N. MOTUZKO (1985). Materials from the Timoskovichi site (Grodno Oblast') suggest some decrease in

Table 2

Species compositon of Late Pleistocene small mammals from the sites of Byelorussia and of the Upper and Lower Dnieper region

Species \ Sites	Malyutino	Borisova Gora	Timoskevichi	Gadyach +	Betovo	Gralevo-2	Arapovichi	Khotylevo II	Drychaluki	Chulatovo I	Diskeninovo	Pashino	Novgorod-Seversky
Insectivora:													
Sorex sp.			+			+							
Soricidae gen.		+										+	
Desmana moschata L.		+	+										
Lagomorpha:													
Lepus sp.					+								
Ochotona sp.		+				+			+		+	+	
Ochotona pusilla Pall.				+	+								
Rodentia:													
Citellus sp.	+			+		+	+						+
Citellus birulai J.Gromov					+								
Marmota bobac Mull.					+		+			+			+
Allactaga jaculus Pall.							+			+			+
Cricetus cricetus L.		+			+								
Cricetulus migratorius Pall.				+									
Glis sp.		+	+										
Apodemus (Silvimus) flavicollis Melchior	+												
Mus musculus L.		+											
Ellobius talpinus Pall.	+												
Clethrionomys glareolus Pall.		+	+									+	
Lemmus sibiricus Kerr.		+			+	+		+	+	+	+	+	+
Myopus vel Lemmus			+										
Dicrostonyx aff.simplicior Fejfar						+							
D.ex gr.gulielmi Sanford					+		+	+	+	+	+	+	+
Dicrostonyx sp.		+											
Arvicola terrestris L.						+						+	
Lagurus aff.lagurus Pall.	+			+			+			+			+
Eolagurus aff.luteus Eversm.	+												
Pitymys aff.subterraneus DeSelys-Longchamps		+											
Microtus (Stenocranius) gregalis Pall.	+			+	+	+	+	+	+	+	+	+	+
M.(Microtus) arvalis Pall.	+	+	+		+								+
M.(Microtus) agrestis L.	+	+	+	+				+	+		+	+	
M.(Microtus) oeconomus Pall.	+	+										+	+
M.(Microtus) ex gr. middendorfi Poljakov-hyperboreus Vinogradov									+		+		

Table 1 Paleoclimatic parameters of the Late Pleistocene (by theriological materials from the deposits of the Upper and Middle Dnieper region and Byelorussia)

Geochronology			Deposits of the Upper and Middle Dnieper region	Climagrams: temperatures, °C		Climagrams: precipitation, mm		Sites of Byelorussia	Climagrams: temperatures, °C		Climagrams: precipitation, mm	
				January	July	annual	solid		January	July	annual	solid
Valdai glaciation		Late Valdai	Novgorod-Seversky Chulatovo I Khotylevo II 23 360±270	-16 - -14 - -12 -18 - -15 - -12 -26 - -20 - -15	+22 - +23 +16 - +19 - +22 +15 - +19 - +23	180-280-380 120-260-400 300-350-400	30-75-120 30-75-120 60-90-120	Pashino Diskeninovo Drychaluki 23 630±370	-27 - -23.5 - -20 -50 - -35 - -20 -24 - -19 - -15	+13 - +14 - +15 +13 - +15 - +17 +14 - +17 - +20	280-360-440 120-270-420 320-370-420	40- 80-120 20- 85-150 50-115-150
Valdai glaciation		Bryansk interstadial	Arapovichi 24 000±300 31 500±140	-18 - -17 - -16	+18 - +19 - +20	150-275-400	20-70-120					
Valdai glaciation	Early Valdai	Cold stage of the Early Valdai	Betovo 36 100±550	-17 - -16 - -15	+15 - +16 - +17	200-350-500	50-110-170	Gralevo-2	-30 - -23.5 - -17	+14 - +15.5 - +17	130-265-400	50- 90-130
Valdai glaciation	Early Valdai	Krutitsky interstadial (?)	Gadyach	-15 - -14 - -13	+19 - +20 - +21	300-420-540	60-100-140					
		Mikulino interglacial	Malyutino	-12 - -9 - -6	+19 - +21 - +23	350	120	Borisova Gora Timoshkovichi	-12 - - 9 - - 6 -23 - -14.5 - - 5	+19 - +21 - +23 +13 - +18 - +23	400-500-600 300-450-600	80-130-180 90-170-200

winter temperatures (KALINOVSKY, P.F. 1980; MOTUZKO, A.N. 1985). It should be noted, however, that only a limited number of mammal species have been identified in this deposit. This, undoubtedly, affects the accuracy of paleoclimatic parameters.

An analysis of climagrams constructed on microtheriological data from the sites mentioned indicates that the total annual rainfall was close to present-day values *(Table 1)*. The values for solid precipitation, likewise, did not differ much from those recorded today.

In general, then, paleoclimatic data obtained from microtheriological materials indicate rather warm and close to present-day conditions for the Mikulino Interglacial. A similar interpretation is suggested by the analysis of the species composition of mammals at the sites. The discovery of a small number of remains of Siberian lemmings at Borisova Gora do not contradict this interpretation. This deposit, according to A.N. MOTUZKO (1985), may date to the initial stage of development of Mikulino faunas. Forest mammals with a certain admixture of tundra and steppe species predominated in the species composition of these faunas. Lemmings are absent from the later Mikulino faunas of Byelorussia - e.g. from Timoshkovichi fauna. This fauna belongs to the period of the interglacial optimum.

Faunal remains form the Malyutino site are quite different. Tundra species are absent from this assemblage and forest forms are few in number. The steppe species serve as a background here.

Thus, the northernmost faunas (Borisova Gora and Timoshkovichi) indicate the existence of a forest zone, while further to the south (at 50° to 52° N) microtheriological data suggest the presence of forest-steppe conditions. Overall, Mikulino Interglacial microtheriofauna from the center of the Russian Plain point to a zonal structure of the environment.

Materials from the Gadyach site most probably date to the Krutitsky Interstadial of the Early Valdai. Faunal remains here were discovered in molehills in the Mezin soil complex (which is known to have a complex structure). The first and oldest phase of this complex formed during the Mikulino Interglacial. The more recent soil formed during one of the early interstadials of the Valdai. In this section humus horizon materials fill in the molehills. This, indirectly, indicates, that the fauna belongs to a more recent stage in the formation of the Mezin soil complex.

The climagram method allowed us to determine the following paleoclimatic parameters for the Gadyach site: the mean temperature of the warmest month reached 20 °C and did not differ from present-day values. January temperatures, on the other hand, in comparison to present-day values, were lower by 6° to 8° C. The total annual rainfall was somewhat lower than at present, the amount of solid precipitation was smaller as well *(Table 1)*.

In the Early Valdai epoch there was a sharp fall in winter temperatures (in Byelorussia - by almost 20° C, in the Dnieper region - by 10° C). Winter temperatures at this time decreased from west to south-east. July temperatures dropped by nearly 5° C. The amount of rainfall decreased considerably, and solid precipitation diminished somewhat as well *(Table 1)*. During this time the southern boundary of the reindeer area shifted considerably to the south, while that of the saiga to the north-west. These facts suggest that the depth of the snow cover was not over 20 to 30 cm (a critical depth of snow cover for the above animals considered here) in these regions. Overall, the early Valdai mammalian faunas reflect a sharp deterioration of environmental conditions (fall in temperatures and decrease in the amount of rainfall) which resulted in a drastic restructuring of the animal world. Tundra species penetrate far to the south and steppe ones to the west. Forest mammals clearly played a minor role at this time. They were more numerous in the west of the Russian Plain, in particular in Byelorussia. In the east, they were practically absent *(Table 2)*.

Environmental conditions did not change very much during the Bryansk interstadial. Tundra species, in particular *Lemmus* and *Dicrostonyx*, were widespread as before. Microtheriological data from the sites of Arapovichi (basin of the Desna river) and of Troitsa II (basin of the Oka river) indicate that low winter temperatures reaching -20° C persisted during this period. Summer temperatures were somewhat lower than present-day ones. The amount of precipitation (both annual and solid) was significantly lower than at present as well *(Table 1)*.

Materials from Drychaluki, and Pashino (KALINOVSKY, P.F. 1983; MOTUZKO, A.N. 1985) indicate that harsh continental conditions persisted in Byelorussian in Late Valdai. January temperatures reached -30° C, July ones were between 12° and 16° C, and annual precipitation was reduced by about a half *(Table 1.)*

A similar climatic situation existed in the Middle Dnieper region as well. Winter temperatures were over 10° C lower than at present while summer temperatures differed little from those of today. The less drastic reduction in winter and summer temperatures observed in the Middle and Upper Dnieper region can be explained by the more southern geographical location of this region.

In conclusion, microtheriological data permit us to detect the main climatic changes which took place on the center of the Russian Plain during the Late Pleistocene.

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VEGETATION RECONSTRUCTION FROM RELICT LOCALITIES OF MODERN PLANTS IN THE CAUCASUS

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ABSTRACT

During the coldest time of the Late Pleistocene-Holocene - 18-20 thousand years ago - maximum displacement of altitudinal vegetation belts took place. For the reconstruction of its scale and, consequently, the spatial distribution of vegetation, important data may be extracted analysing relict localities of modern plants - species growth sites in the distance from the main area of their distribution under alien ecological conditions. Problems connected with the choice of these localities from dot maps of modern plants areals of the Caucasus and their use for the mapping of the mentioned chronosection vegetation have been considered. An example of such application in the reconstruction of alpine and subalpine belt boundaries is given for the highest part of the Greater Caucasus.

INTRODUCTION

The reconstruction of natural conditions of the last glacial epoch in the Caucasus has remained one of the cardinal problems of Quaternary paleogeography that requires thorough study. In spite of the more than a century long history of investigations of the problems associated with the last glaciation, the extreme complexity of this region, its exclusive spatial heterogeneity have presented the formation of a clear picture, reflecting the nature of the Caucasus of this stage.

We have faced a task of constructing a map of vegetational cover of the Caucasus for a chronological level of 18-20 thousand years B.P., corresponding to the maximum of the last glaciation. It seems obvious that the bases for such reconstruction should be the materials of paleobotanical investigations. However, the analysis of the available data showed that current paleobotanical knowledge cannot provide the necessary information beacuse of the circumstances below. First, only a few palynologically studies sections are available for the time interval under consideration. Further, a special study of spore and pollen spectra formation carried out in the mountain areas of the Caucasus has shown (KLOPOTOVSKAYA, N.B. 1973; 1980) that in most cases the latter reflects the natural vegetation of a small orographically isolated area. At the same time, mountain territories often exhibit displacement of vegetational belts, different kind of disturbances of the natural distribution of vegetation types, adjacent mountain slopes are frequently occupied by forests, taking different hypsometric position in the altitudinal zonation of the vegetational cover of the region. All these do not allow wide-scale extrapolations based on palynological data. Still more local conditions could be described by macrofossils, the findings of which are practically missing for the period of interest.

The insufficiency of the actual material for mapping the vegetational cover of the Caucasus at the maximum cooling of the last glaciation turned us to take an analysis of the distribution of modern plants over this region, with the aim of finding relict localities. A site or sites of a species' growth, located outside the main area of its distribution were taken as localities of this type *(Fig. 1)*. The localities were chosen

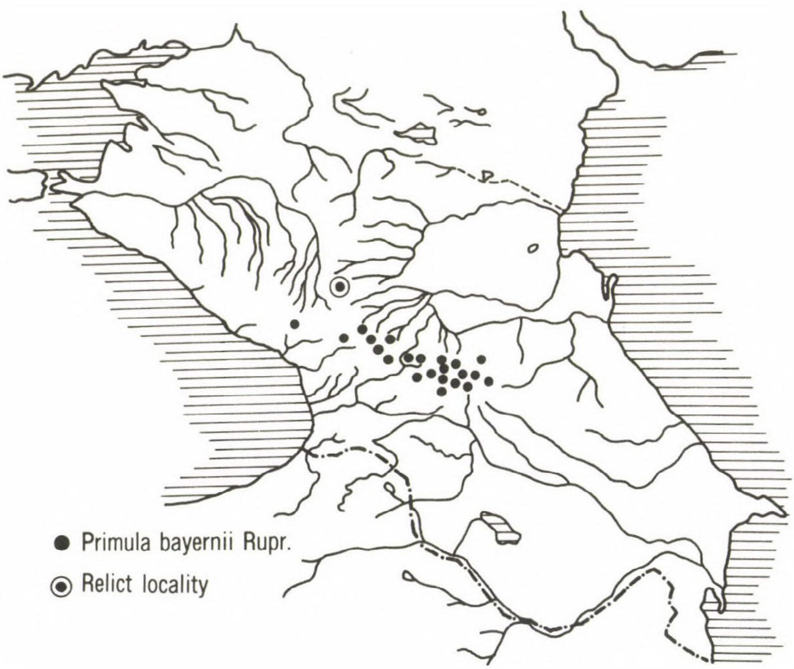

Fig. 1 The dot (distribution) map of Primula bayernii Rupr. on the the territory of the Caucasus

from the dot maps of areas, given in seven volumes of the "Flora of the Caucasus" compiled by A.A. GROSSHEIM (1939, 1940, 1945, 1950, 1952, 1962, 1967).

DISCUSSION

As an initial prerequisite we have adopted the principle that the chronological section in question - maximum cooling of the last glaciation - corresponds to the most severe time on a planetary scale in the Pleistocene and Holocene. Therefore, the displacement of the altitudinal landscape belts, and of the altitudinal belts of vegetational cover to the lowest hypsometric levels is associated with this chronological level; the response of the vegetational cover to the subsequent course of climatic changes was also manifested in the dynamics of the altitudinal belts, though the degree of expressiveness of the latter, their displacement in cold phases were correspondingly smaller.

It is generally accepted that in the Pleistocene the largest expansion of Caucasian glaciers was not in the last (Würm) but in the preceding (Riss) glaciation, which for this reason should be considered as maximum. V.Z. GULISASHVILI (1961), an outstanding expert of vegetation of the Caucasus, proceeding from the fact that *Pinus hamata* (Stev.) D.Sosn., *Betula verrucosa* Ehrh., *B.litwinowii* A.Dol. and *Salix caprea* L. grow directly at the edge of modern glaciers, attempted to mark the lower boundary of the Riss glaciers in different basins by pine and birch remnants preserved at the lowest level. These euryoecic species are represented on underdeveloped soils, rocks, where no other tree species compete with them and this fact determined their existence here for so a long time. V.Z. GULISASHVILI observed such assamblages also at frontal moraines of the last glaciation. Coincide of sites of the frontal moraines of the maximum, as well as of the last glaciation with the growth sites of the tree species mentioned above allowed V.Z. GULISASHVILI to suggest "green moraines" on the boundaries of ancient glaciations on the basis of the existence of pine and birch stands. The same is presumed by him in the basins, where glacial deposits were not preserved.

It is obvious that the evidence for the last glaciation were preserved in the Caucasus, similarly to other regions, much better than those of the preceding one (Riss). This is well illustrated by the degree of moraine preservation, relating to these glacial stages. It is natural to draw analogy with the relict localities of individual species.

It should be mentioned that pine and birch trees located directly at the edge of Riss glaciers on hypsometric marks lower than similar assemblages, fixing the glacier edge in the last glacial stage should be regarded as a particular case. Displacement of vegetation belts caused by maximum cooling is naturally evidenced in modern vegetational cover much more distinctly than the shift of belts in a more remote and less cold Riss epoch, though representing the maximum of glaciation.

Having thoroughly analyzed the distribution of growth sites of modern plants on the locality maps of the areas and having traced the correspondence of eco-coenotic conditions in the growth sites to the nature of the given species, we have distinguished 400 species which are encountered in 981 case within the range of development of other ecological groups; localities of alpine plants in mid-mountain and low-mountain areas, forest elements - in the conditions of steppes or steppe deserts (semi-deserts).*

It is to be understood, that only a part of the isolated sites could be used while mapping vegetational cover of the 18-20 thousand years chronological section. These sites on the whole fix different time deviations from the modern distribution of vegetation and floral superpositions of different age. It has become necessary to carry out eco-coenotic analyses and to choose those species from the combination of species, that correspond to the greatest degree to the severe climatic conditions in each individual site. Just for this reason the growth sites of alpine species were primarily dealt with.

The volume of this paper does not allow any detailed consideration of relict localities indentified by us for the whole of the Caucasus in order to demonstrate their highly informative value for the reconstruction of vegetation. We shall make an attempt to show the expediency of their use for the reconstruction of the areal distribution of alpine and subalpine belts only, and as an example we take the territory between 42° and 44° EL and 46° and 42° NL *(Fig. 2)*. It includes the most elevated part of the Greater Caucasus with its highest summit Elbrus, its northern gently including macroslope, the greater part of the Stavropol upland, the southern branches of the Greater Caucasus within the borders of Western Transcaucasus, the eastern end of the Kolkhida lowland, and northern foothills of the Little Caucasus. This is the scene of modern glaciers that occupy here a more extended area than in other parts of the Greater Caucasus. Here the alpine, subalpine, forest, forest-steppe and steppe zonal vegetation types are represented.

On the northern macroslope of the Greater Caucasus we have identified 19 relict localities of 16 alpine species *(Table 1)*. The position of these localities** showed that, as it was expected, these have a regular distribution pattern clearly outlining in their totality the concrete area. These are observed at the lower part of the distribution zone of subalpine meadows, which in the Caucasus are usually found in combination with

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* Pursuing a concrete object we did not focus our attention on the localities, fixed hysometrically above the main area of these or other less numerous species. It is obvious that they reflect time interval characterized by warmer than present-day climatic conditions. The problem - whether these localities are associated with interglacial or any thermal optimum in a postglacial period, is outside the scope of the problem in question.

** In Fig. 2 the relict localities of different species often coincide in this and other cases.

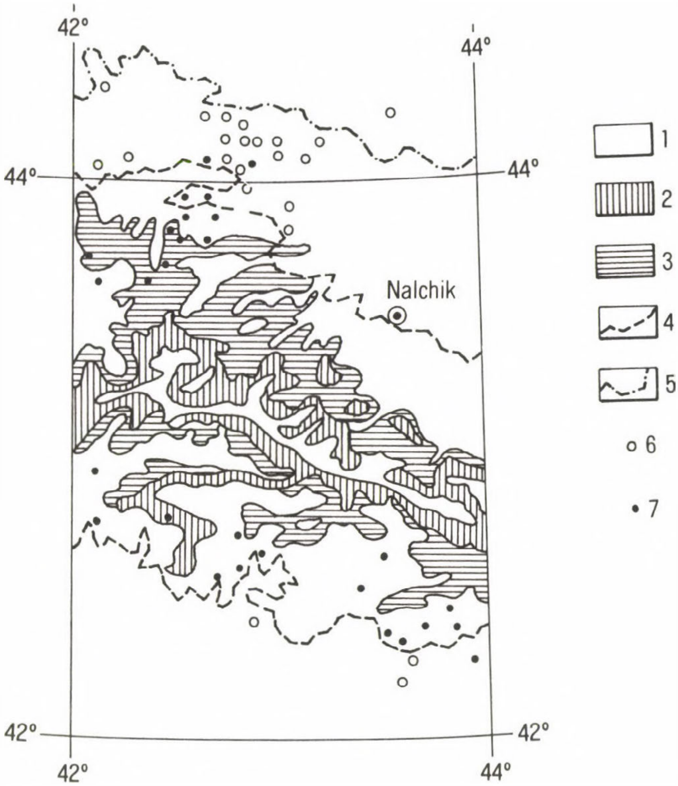

Fig. 2 Relict localities of alpine and subalpine species and assumed boundaries of corresponding belts at the maximum cooling of last glaciation on the territory between 42° and 44° EL within the Greater Caucasus

1 = Glaciers; 2 = Carpet-like alpine meadows and alpine meadows; 3 = Subalpine meadows usually found in combination with rose bay shrublets, elfin woodland and open woodland; 4 = Lower boundary of alpine belt at the maximum cooling of last glaciation; 5 = Lower boundary of subalpine belt at the maximum cooling of last glaciation; 6 = Relict locality of alpine species; 7 = Relict locality of subalpine species

rose bay shrublets, elfin woodland and open woodland, in mid-mountain steppe meadows, in low-mountain meadow steppes and in steppe meadows, in herb and grass steppes (with *Festuca* and *Stipa*). The great number of localities, their pattern distribution in harmony with the relief, enable us to consider it to be the zone of the past extension of the alpine belt. Such a conclusion is also supported by the fact that the species presented in *Table 1* are the plants associated solely with the

Table 1 List of alpine species, the relict localilites of which are fixed on the northern macroslope of the Greater Caucasus

Species	Number of relict localities
Thesium alpinum L.	1
Cerastium cerastoides (L.) Britt.	2
Cerastium undulatifolium S.et L.	2
Gypsophilla tenuifolia M.B.	1
Thalictrum alpinum L.	1
Saxifraga flagellaris Willd.	1
Saxifraga adenophora C. Koch	1
Potentilla crantzii (Cr.) Beck.	1
Dryas caucasica Juz.	1
Alchimilla languida Bus.	1
Vicia alpestris Stev.	1
Chamaesciadium acaule (M.B.) Boiss.	1
Primula bayernii Rupr.	1
Gentiana caucasica M.B.	2
Gentiana djimilensis C. Koch	1
Veronica monticola Trautv.	1

Table 2 List of subalpine species, the relict localities of which are fixed on the northern macroslope of the Greater Caucasus

Species	Number of relict localities
Lycopodium selago L.	1
Juniperus sabina L.	2
Briza elatior Sibth. et Sm.	1
Colpodium versicolor (Stev.) Schamlh.	1
Zerna variegata (M.B.) Nevski	3
Muscari pallens Fisch.	2
Orchis globosa L.	1
Orchis sphaerica M.B.	1
Herminium monorchis (L.) R. Br.	2
Polygonum carneum C.Koch	1
Stellaria anagalloides C.A.M.	1
Silene linearifolia Otth.	1
Anemone fasciculata L.	2
Sempervivum caucasicum Rupr.	1
Sorbus caucasigena Kom.	1
Polygala alpicola (C.A.M..) Rupr.	1
Epilobium algidium M.B.	1
Gentiana aquatica L.	1
Gentiana angulosa M.B.	1
Nepeta grandiflora M.B.	2
Lamium tomentosum Wild.	1

alpine belt (localities of the species growing in the alpine and subalpine belts, are considered together with localities of the subalpine species). The frequency of numerous relict localities within a specific territory enables us to consider these to be the reflection of vegetation that existed in a concrete epoch, i.e. during the maximum cooling of the last glaciation. The same material shows the former position of the lower boundary of the alpine belt. On the other hand, this material cannot be used for determining the upper boundary. In this case, the use of the data on the location of the corresponding paleoglacial relief forms should be justified.

The next stage is the selection of relict localities of subalpine plants from the territory, which naturally does not include the area of the alpine belt.

28 relict localities of 21 species have been recorded on the northern slope *(Table 2)*. They are scattered over a large area, the northernmost points are at the latitude of the settlement Essentuki. Most species are found in the area of meadow steppes and steppe meadows, as well as of herb and grass steppes (with *Festuca* and *Stipa*). A greater diversity of their locations is observed here, the density of relict localities is more pronounced. It should be pointed out that the lower limit of their distribution is very clearly expressed: the boundary passes the latitude of the settlement Mineralnye Vody, although the phytocoenoses, in which these species are disseminated, are widely spread further to the North.

CONCLUSIONS

The analysis of distribution of relict localities of the alpine and subalpine species enables us to suggest that in this part of the northern slope of the Greater Caucasus the boundary between the alpine and subalpine belts in the maximum cooling of the last glaciation passed within the range of absolute heights of approximately 800-1200 m and the lower boundary of the subalpine belt at an absolute altitude of 500-800 m. Now, these boundaries are at altitudes of 2300 and 2200-2250 m, respectively (GULISASHVILI, V.Z.--MAKHATADZE, L.B.--PRILIPKO, L.I. 1975).

The position of the alpine belt on the steeper southern macroslope of the Greater Caucasus could be inferred in the chronological section from the data on the growth of 16 alpine species in 22 relict localities *(Table 3)*, most of which are associated with beech forests. As for the subalpine belt, only 4 species in 4 sites have been distinguished here *(Table 4)*. To reconstruct the hypsometric position of its lower boundary, it is necessary to extrapolate, using material from the adjacent territory. In this part of Western Transcaucasus the boundary between the alpine and subalpine belts is drawn within the absolute heights of 500-800 m. Recently it passes here at the altitude of 2100-2200 m. Based on the distribution of relict

Table 3 List of alpine species, the relict localities of which are fixed on the southern macroslope of the Greater Caucasus

Species	Number of relict localities
Salix hastata L.	1
Cerastium undulatifolium S. et L.	2
Cerastium oreades B. Schischk.	1
Arenaria lychnidea M.B.	2
Gypsophila tenuifolia M.B.	1
Saxifraga scleropoda Somm. et Lev.	2
Potentilla crantzii (Cr.) Beck.	1
Dryas caucasica Juz.	1
Alchimilla sericea Willd.	1
Trifolium polyphyllum C.A.M.	1
Astragalus levieri Freyn.	2
Geranium renardii Trautv.	1
Carum alpinum (M.B.)	1
Chamaesciadium acaule (M.B.) Boiss.	2
Gentiana caucasica M.B.	2
Omphalodes lojkae Somm. et Lev.	1

Table 4 List of subalpine species, the relict localities of which are fixed on the southern macroslope of the Greater Caucasus

Species	Number of relict localities
Geranium psilostemon Lbd.	1
Daphne glomerata Lam.	1
Daphne alboviana G.Wor.	1
Pedicularis atropurpurea Nordm.	1

localities of the subalpine species, on their small number and on the character of the relief, it is possible to suggest that the vertical range of the subalpine belt was restricted.

As a conclusion it should be pointed out that the wide use of the data on relict localities does not exclude the necessity of using the data obtained by other methods, first of all by the palynological method. On the contrary, this will only provide an opportunity for data comparison and their mutual control will support the validity of the reconstruction.

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METHODICAL ASPECTS OF GENETIC DIAGNOSTICS OF FOSSIL SOILS

T.D. MOROZOVA

ABSTRACT

Fossil soils are polygenetic formations bearing features of pedometamorphic transformation (diagenesis) which developed after pedogenesis was completed. These specific features of fossil soils should be taken into account when identifying their genesis. Therefore, representative data on ancient soil formation can be obtained employing a set of analytical techniques. The most stable features of pedogenetic processes imprinted in the mineral and organic patterns of soils are revealed through them. The most efficient is the micromorphological method based on studies in thin sections prepared from samples of undisturbed structure.

*

Fossil soils are widely distributed in almost all loess regions of the world. They are the most important component of loess-paleosol formation.

It is difficult to overestimate the significance of fossil soils for reconstructing paleoenvironments and for marking and dating components in the chronostratigraphic subdivision of loess sequences. The accuracy of such constructions, however, to a great extent depends on the accuracy of genetic diagnostics of fossil soils *(Fig. 1)*.

The fully developed fossil soils, in contrast to modern Holocene soils, reflect the result of completed soil-formation during warm (interglacial) semicycles of the natural-climatic macrocycle.

Soil profiles form as a result of continuous effects of soil-formation factors during different climatic stages of the interglacial epoch. Significant differences exist between buried soils and present-day ones. Fossil soils contain much more complex information on processes which occurred at different times and which affected not only the formation of the soil profiles, as such, but also their preservation.

We can distinguish the following main processes responsible for the formation of the buried soils observed in loess-paleosol formations:

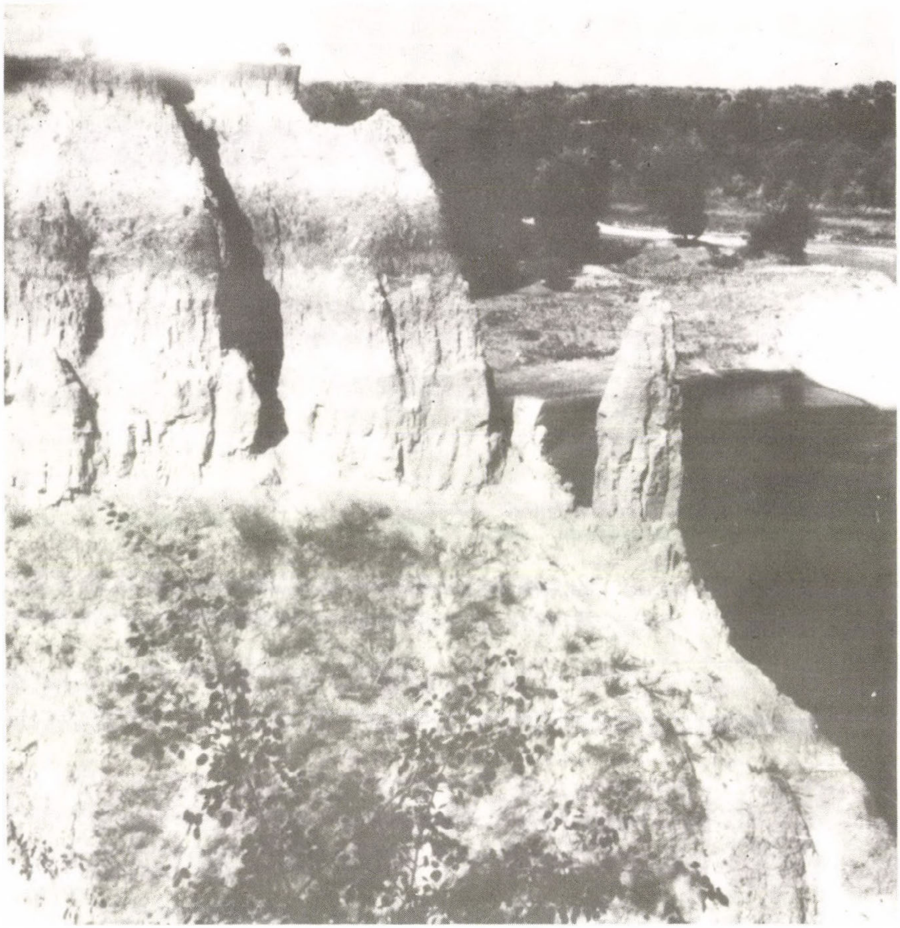


Fig. 1 Cross-section near Gunki village (central part of the Russian Plain)

1. Processes of soil formation which took place during periods of intensive pedogenesis in warm semicycles. Their direction and intensitiy changed in response to different environmental-climatic conditions.

2. Processes specific to transitional stages to cold semicycles - in particular, to glacial ones. These processes, which took place at the end of or at the completion of soil formation during warm semicycles, greatly affected the preservation of soil profiles, especially their upper parts marginal to the over-

lapping loess horizons. They included: cryogenic processes which contributed to turbation of soil horizons, 2) processes of sedimentation intensification when intermediate deposits transitional from fossil soils to loess horizons were formed between loess and soil, and 3) removal and redeposition processes which affected materials in the upper parts of fossil soils located on slopes (formation of pedosediments).

3) Pedometamorphic (after GERASIMOV, I.P. 1971) or diagenetic processes which transformed the buried fossil soils. As a result of these processes fossil soils lost their indices of low stability and acquired new secondary ones.

In studying the genesis of fossil soils, the greatest attention is, naturally, paid to indices of ancient pedogenesis which reflect conditions extant when the soils proper were formed. These processes lie at the base of paleopedological reconstructions. The study of the different forms of disturbances of the soil profile and of pedometamorphic changes in their properties is also very important. This permist us to understand the essence of natural processes which followed the cessation of soil formation, and to evalute the reliability of paleopedological reconstructions.

Because of this data on the stability of soil indices and on the factors and processes affecting soil preservation are of utmost importance. It is crucial that we be able to distinguish the soil indices proper from those which occurred as a result of diagenetic and pedometamorphic processes. The need to do so was emphasized many times by I.P. GERASIMOV (1961, 1971) and others. Our ability to do this, in turn, determines our possibilities for using modern methods of research when studying buried soils.

The set of analytic methods which we use in our studies depends on the preservation of the most stable indices of soil formation (MOROZOVA, T.D. 1981). They include the following:

Analytical methods for studying the mineral components of the soil

Data from chemical and physical-chemical analyses, including mechanical and gross composition analysis, and on the composition of clay minerals, can be used for diagnostic purposes.

In a number of cases, however, when fossil soils were subjected to the impact of secondary cryogenesis and fracturing, thier genetic horizons may appear to be mixed, and the data of the distribution of mineral components along their profiles may be greatly distorted.

Analytical methods for investigating the organic matter

The composition and characteristics of the organic matter in the soil are the most important diagnostic criteria of soil formation processes. Soils show differences in their humus content depending on natural climatic conditions. The fractional composition of soils, the properties of their humic acids, and the regularities of humus distribution along this profile all change as well.

When the soil is buried, in spite of the fact that it preserves its dark color, the amount of its organic matter decreases noticeably. Analyses of radiocarbon data on the ages of the different fractions of humus indicate that fractions of the humic acids proper and the humic acids of humins are the most stable and the least subject to pedometamorphic transformations (GERASIMOV, I.P.--CHICHAGOVA, O.A. 1971). Numerous studies have confirmed the hypothesis, first advanced by I.P. GERASIMOV (1961), that the correlation of the main groups of humic matters (humic, fulvoacids, humins) in fossil soils generally reflect two main directions in the formation of paleosols: of the forest or of the steppe type.

The data on some properties of humic acids (optical density, stability to electrolites) are also used as diagnostics of old pedogenesis. This was demonstrated for the first time in the study of humic acids in soils of the Bryansk interval and of the Mikulino Interglacial (MOROZOVA, T.D.--CHICHAGOVA, O.A. 1968).

The study of the characteristics of organic matter in relation to the genesis and diagenesis of fossil soils of the different Pleistocene ages permits us to draw definite conclusions about the stability of the organic matter and about the way it is transformed in fossil state.

Studies by O.A. CHICHAGOVA and A.E. CHERKINSKY (1979) pointed out two groups of specific properties of organic matter in buried soils. The first group consists of conservative properties used in defining the essence of old soil formation, and includes: 1) the group composition of humus and the ratio of humic acid carbon (C_{ha}) to that of fulvoacid carbon (C_{fa}) as an indicator of humus type, 2) the extent of humification of the organic matter as characterized by a relative content of C_{ha} compared total soil carbon (oral communication) and 3) physical-chemical properties of HA (HA optical density).

The second group includes properties associated with the transformation of the organic matter in time, as well as those resulting from diagenesis: 1) the total content of the organic matter in hydrolyzed compounds, and 2) the fractional composition of humus.

The low humus content, absence of hydrolyzed compounds and mobile HA forms, an increase in humin fraction content, and the prevalence of fractions of humic acids associated with calcium - all features common in Pleistocene soils - are a consequence of humus transformation due to decomposition, mineralization, and irreversible coagulation and dehydration. All of them took place when the soils were buried in carbonate loesses.

Micromorphological method

The micromorphological method is the most effective one for studying fossil soils. It is based on the study of microsections from soil samples with natural and undisturbed structures. This method permits us to observe manifestation of elementary soil processes in greater detail. In particular, it permits the study of such things as humus accumulation, the motion and redistribution of different components within a soil profile, the weathering of soils and neogenesis of minerals, the formation of neogenetic salts and of ferrum hydroxide in soils, and so on *(Figs. 2a, b, c)*.

This method, developed by KUBIENA (KUBIENA, W.L. 1970, etc.), is widely used for the study of fossil soils in different countries (MATVISHINA, Zh.N. 1982; MOROZOVA, T.D. 1981; ZYKINA, V.S. et al. 1981; SMOLIKOVÁ, L. 1972, 1975 etc.; BRONGER, A. 1972, 1976 etc.; KONECKA-BETLEY, K. 1976, 1979 etc.).

The micromorphological method permits us to study those details of the soil profile which contain the basic information both on processes of old pedogenesis and on those which resulted from soil fossilization and which distorted its overall structure.

Micromorphological indicators in soils can be divided into the three groups on the basis of their mutability through time and in response to diagenesis (MOROZOVA, T.D. 1981):

1. stable - microstructure of the main mass, aggregates, forms of pores, microstructure of optically oriented clays, argillaceous cutanes;
2. of low stability - humus.
3. unstable - neogenetic salts.

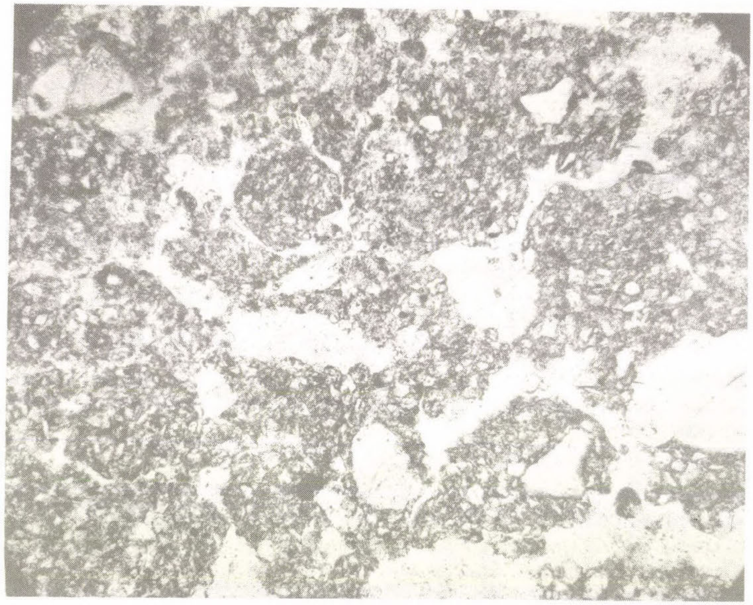
Diagnostic transformations in the microstructure of fossil soils, as well as the primary composition of these soils, depend on such factors as the natural aging of the soil matter. This is primarily reflected in the properties of the humus and of the organic remains. The pressure of the overlying deposits contributes to the consolidation of microstructures, to the reduction of the area of interaggregate pores, and to the convergence of the aggregates and their certain deformation. Secondary gleying, which destroys aggregates, leads to their guttering and to the decoloration of the soil plasma. The diagenetic processing of soils with salt solutions contributes to their general consolidation and to secondary deposition of newly formed carbonates, gypsums, and other minerals. The study of these fossil soil diagnostics permits us to reconstruct the evolution of Pleistocene soil formation processes as reflections of changes in global climate.

Reconstructions of the ancient soil cover are of special interest because they point to geographic regularities in the distribution of soils which existed during ancient epochs of soil formation

Detailed studies of long sequences at key sites (in the walls of quarries and ravines, on river and sea shores) permit us to trace changes in soil structure by analyzing paleomeso- and microrelief and studying soils in Catena. This permits us to compile lists of automorphic features of the zones of paleosols, and to characterize the hydromorphic soils which accompanied them under subordinate conditions of relief.

Small scale mapping of these data permitted us to reconstruct the soil cover extant over most of the European territory during two Late Pleistocene epochs of soil formation - the Bryansk interval (24-30 thousand years ago) and the Mikulino (Eem) Interglacial (MOROZOVA, T.D. 1981; GERASIMOV, I.P.--VELICHKO, A.A. eds. 1982).

a



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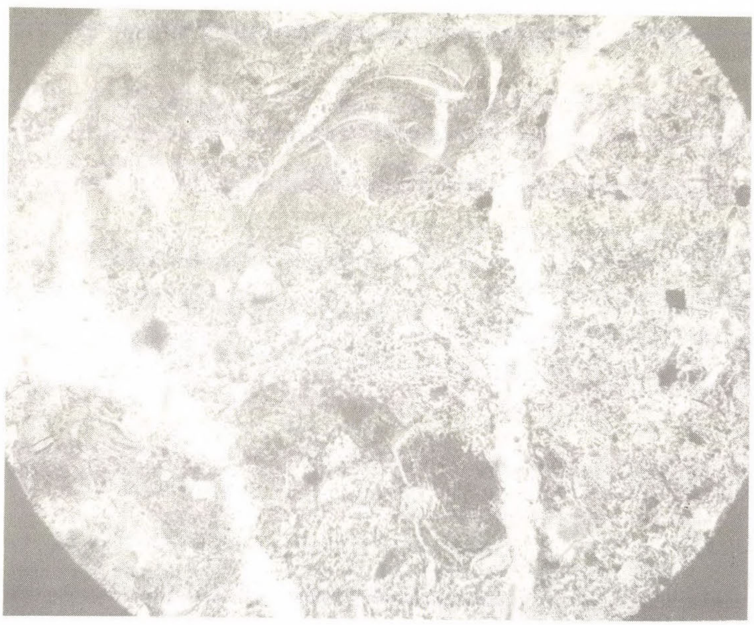

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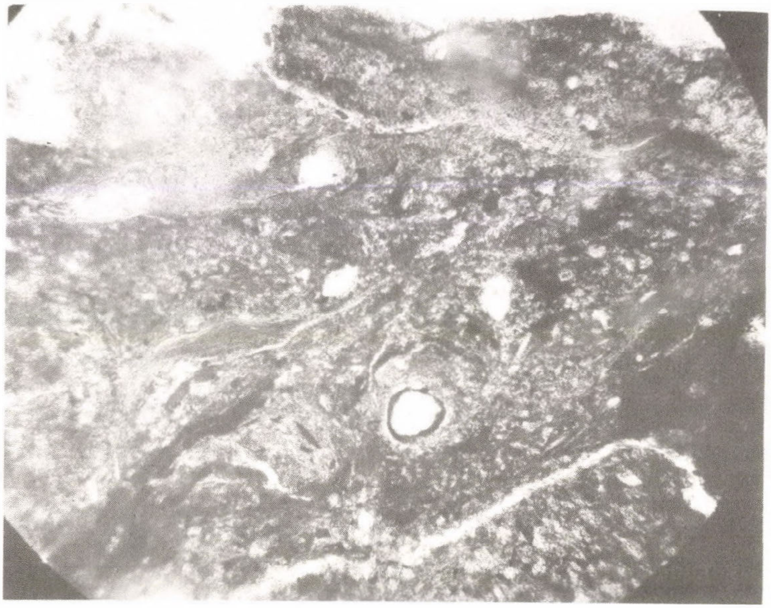

Fig. 2 Microstructure of fossil soils

a) Aggregates in the humus horizon of the Mikulino paleosol. Konstantiny cross-section (central part of the Russian Plain)

b) Argillaceous cutanes in the Bt horizon of pseudopodzol soil of Mikulino Interglacial. Konstantiny cross-section (central part of the Russian Plain)

c) Argillaceous-ferruginous cutanes in the Btg pseudogley horizon of a Riss-Würm interglacial soil. Straubing cross-section (FRG)

In summary, the main diagnostic indicators of the ancient pedogenesis are: 1) the reconstructed structure of soil profiles; 2) combination of the revealed properties; 3) microstructures in the genetic horizons and in soil profiles; 4) the distribution of stable mineral components along the profile, 5) the type of humus and optical properties of humic acids, 6) the combinations of soils in the Catena, and 7) the reconstructed soil cover.

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MORPHOLOGICAL FEATURES OF PALEOSOLS FROM PAKS WITH REGARD TO THEIR PALEOECOLOGICAL INTERPRETATION

O. MOROZOVA

ABSTRACT

During the comprehensive geological, pedological and geographical research of the Paks loess profile two paleosols were analyzed micromorphologically in 1985.
The paleosols selected were 1) A paleosol (Basaharc Lower - BA) in young loess at a depth 21-23 m; 2) A double paleosols (Mende Base Soil Complex - MB) on the boundary of young and old loess at 27,5-29 m.
Thin sections were prepared of samples collected from genetical horizons and analyzed using petrological microscope. On the basis of micromorphological properties relevant differences in the genesis of these soils were shown to be produced by various environmental conditions such as gradients of atmospheric temperature and precipitation, factors of the soil moisture regime, and influences from vegetation cover and duration of the soil formation.

*

INTRODUCTION

The micromorphological method is the most informative and widest-spread means of paleosol studies (BIRKELAND, P.W. 1984; BREWER, R. 1972; BULLOCK, P. 1985; DOBROVOLSKY, G.V. et al. 1974; FEDOROFF, N. 1971; MOROZOVA, T.D. 1964; MÜCHER, H.J.--MOROZOVA, T.D. 1983). The reasons supporting this method are evident in the final goal of the geographical study of paleosols i.e. reconstructing the ecological conditions of soil formation. It is possible to apply this approach since the paleosols reflect the range and mechanism of ecological changes in their properties.

As soil formation mostly involves the spatial rearrangement of substrate (in the present case: of loess material) rather than changes in composition, the results are manifest essentially in pedomorphological features.

Morphological analyses represent fundamental tasks in any pedological research. While classical pedology concerned with recent soils and the given environments and morphological studies are of descriptive nature, in paleopedology morphological properties become predominant in reconstructing the environments of soil formation.

Within morphological investigations three levels can be distinguished: macro-, micro- and submicro-levels.

Transition on microscopic level produces not only quantitative change in investigations (increase in the number of the studied objects) but also leads to qualitative change due to application of techniques based on the optical properties of materials. Optical properties are not less characteristic and selective than their behaviour in chemical reactions, and allow the soil composition to be studied while retaining the natural spatial arrangement of soil material.

Attempts were made earlier at analyzing the Paks paleosols micromorphologically (BRONGER, A. 1970a, 1972, 1976; STEFANOVITS, P.--RÓZSAVÖLGYI, J. 1962, 1965). However, the related publications either allude at some micromorphological features without describing them in detail (BRONGER) or refer to research made long time before in which out-of-date methods of preparation and description were used (STEFANOVITS, P. and RÓZSAVÖLGYI, J.) and this lack of information necessitated a new investigation.

MATERIALS AND METHODS

When selecting the paleosols the following aspects were considered:

- the Paks profile is the key loess section studied in greatest detail of all Hungarian profiles; it was mentioned as early as the 1930s (BULLA, B. 1934, 1937/38) and the study covering its Quaternary complex started in the early 1960s in international cooperation led by M. PÉCSI.(ÁDÁM, L. et al. 1954; BORSY, Z. et al. 1979; GEREI, L. et al. 1985; MÁRTON, P. 1980; PASKEVICH, G.A. 1980; PÉCSI-DONÁTH, É. 1980; SZEBÉNYI, E. 1980; WAGNER, M. 1980).
- the selected buried soils are strikingly marked within the loess series,
- previous analyses and field observations attest to significant differences in genesis of these two soils,
- similar soils are also common in other loess profiles of Hungary.

The paleosols selected were:

1. Paleosol intercalated in young loess at a depth of 21-23 m (Basaharc Lower - BA)
2. Paleosol situated on the boundary of young and old loess at a depth of 27.5-29 m (Mende Base Soil Complex - MB).

A sketch of loess profile with the indication of buried soils and the results of complex investigations were published previously (PÉCSI, M. 1977, 1980; PÉCSI, M.--SZEBÉNYI, E. 1971).

For micromorphological investigations samples were taken from each genetical horizon and from the cover layer transitional to loess, one sample from each.

Thin sections were prepared in Dokuchaev Pedological Institute, Moscow. For impregnation natural resin was applied. Some additional thin sections of paleosol MB were prepared in Soil Department INA (Paris-Grignon) using synthetic resins and they were also considered in the analysis.

Thin sections were analyzed using a petrological microscope.

RESULTS AND DISCUSSION

Some main analytical data for the profiles BA and MB are given in *Table 1* (after PÉCSI, M. et al. 1977).

Table 1 Main analytical data

Profile Hor/sec N	Particle-size analysis (%) Depth (m)	Clay 2µm	Silt 20-2µm	Fine-sand 20-200 µm	Sand 200-2000 µm	$CaCO_3$ (%)	Humus (%)
Basaharc Lower Cover layer sect.1.	21.10-21.35	24.3	21.C	40.0	21.0	22.9	0-
A horizon sect.2.	21.90-22.05	40.6	14.2	26.0	18.8	11.7	0.43
B horizon sect.3.	22.40-22.60	38.7	15.3	23.2	18.8	18.0	0.65
Mende Base Lower Cover layer sect.4.	26.50-27.50	27.4	27.5	31.8	12.9	16.72	0-
A horizon sect.5.	27.90-28.05	36.3	20.6	25.2	16.9	11.70	0.21
B horizon sect.6.	28.55-28.75	36.6	18.1	25.3	19.7	6.27	0-

When interpreting the data acquired, the micromorphological features of soils are treated in relation to 'elementary' pedological processes and typology of paleosols was consciously avoided such conclusions would have reached behind the evaluation of micromorphological data. Also, establishing types of paleosols cannot be attained by the analogy of recent soils.

Drawing analogies is hindered by the following:

- A paleosol is the manifestation of a 'completed cycle' of soil formation, while a recent soil represents a certain stage in its evolution,

- During a 'completed cycle' of soil formation which is accepted to subdivide into phases: initial (pre-optimum), optimal and final (post-optimum) a qualitative change in character and type of soil formation processes is taking place. For example in case when cessation of soil formation was brought about by cooling of the climate, in the final phase along with decline of soil formation process its change to boreal type must have occurred and morphological features produced by the latter must have been added to those formed during the phase of optimum.

The above suggest that the analogous form of a paleosol is not some concrete recent soil but a series of recent soils which reflect a range of ecological changes in space (thousand kilometres) in the same manner as they took place in a given site over a period of time (several thousand years). In addition, important issues are diagenesis and pedometamorphic processes in paleosols (GERASIMOV, I.P. 1971; FENWICK, I.M. 1985; JENKINS, D.A. 1985; VALENTINE, K.W.G.--DALRYMPLE, J.B. 1976; VELICHKO, A.A.--MOROZOVA, T.D. 1976; YAALON, D.H. 1971).

BASAHARC LOWER SOIL (BA)

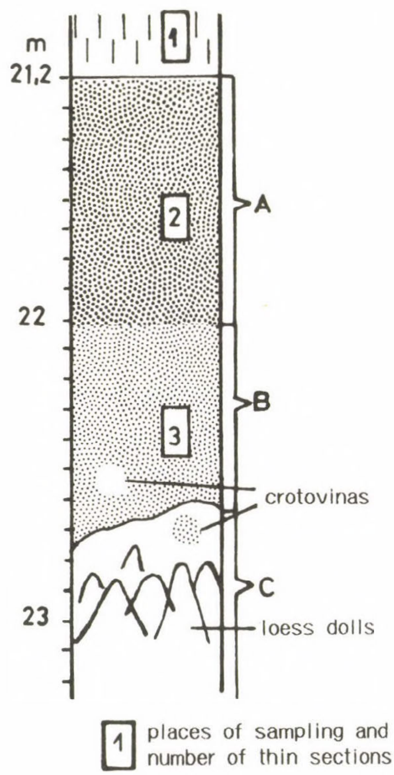

Fig. 1 Soil profile BA

Description of soil profile

Being the most developed paleosol in the Upper Pleistocene loess sequence, Basaharc Lower Soil (BA) is found a 21-23 metres depth from the recent surface (Fig. 1).

The humified layer has a thickness of 1.6 m, maximum humus content is 0.6% (weight percentage). The colour changes from 10 YR 4/4 to 4/6 brown. Sharply mottled. The spots from castings have diameter of 0.005 m, they are lighter in colour than the soil matrix in the upper part of the profile and inverse in the lower one.

The soil mass is intensively reworked by earthworms (spots) and larger animals (crotovinas $d \approx 0.05$ m in the A and C horizons). Clay-loam texture, weakly developed structure. Its type is medium crumb biogene structure. The consistency of the dry soil changes with depth from sligthly hard to hard.

Within the profile two horizons can be distinguished (A and B being different in colour and consistency).

Transition to the C horizon with sharp, uneven boundary. Horizon C_{Ca} is enriched in carbonate. In the bottom of the soil profile a layer of loess dolls with the mammilated peaks tending towards the soil profile is found.

The paleosol has a gradual transition in colour and texture to loess. It is a dull yellow-orange 10 YR 6/3, silt loam with hard consistency and cemented by carbonates.

Micromorphological Data

Thin section 1. Cover layer transitional to loess (Photo Ia)

A cover silt loam of massive structure. Predominating voids are chambers with the size of 0.002-0.004 m and vertical channels, some horizontal planes (60-200 µm), porosity is about 10%. The skeleton consists of carbonate, quartz, micas grains with size from coarse to fine sand. Grains of medium size 200-500 µm are rare. Random distribution pattern. Porphyro-skelic distribution between the plasma and skeleton grains. Small content of plasma. Isotropic clayey-carbonate plasma in the fleck. The basic structure can be determined as a Channelled Porphyroskelic Calciasepic s-matrix.

Special feature: carbonate separation (Calcitans) around the pores.

Characteristics: Abundant calcareous material, plenty of grains of primary carbonate in the skeleton, small plasma content.

Thin section 2. A horizon (Photo Ib, c)

The structure is similar to section 1. The predominating voids are channels (d ≈ 100-300 µm, l ≈ 2000 µm), porosity is 3-5%. The ratio of S/P is about 1. In the skeleton more grains of medium sand are found than in section 1. Mineral composition: quartz, feldspars, micas and rarely primary carbonates. The carbonate-clayey plasma commonly has a flecked orientation pattern but plasma separations with striated orientation occur subcutanically to the surfaces of skeleton grains (Photo Ib, c). Thus it has a Skelsepic plasmic fabric.

Special features: mainly associated with voids. These are plant remains, excreta partly ferri-carbonized (Photo Ic), occasionally ferruginous micronodules (d ≈ 1000 µm), carbonate separations around pores are very rare.

Characteristics: moderate carbonate content, heterogenous s-matrix, zones (spots) differ in colour and carbonate content .

Thin section 3. B horizon (Photo Id, e, f)

Structure and porosity are similar to section 2, S/P ratio differs. More plasma in comparison with the above horizon. The skeleton has more sandy character. Plenty of medium and coarse sand grains. In the mineral composition quartz is predominant. Clayey-carbonate plasma has a darker colour than in section 2 because of the presence of humus. Very much enriched in carbonates and perfectly aggregated in flecks. Calciasepic plasmic fabric. Widely exhibited features of secondary carbonates: cloud-like clayey-carbonate concentrations in the ground mass (Photo Id).

Special features: Calcite pseudomorph of plant remains, (roots) in the voids (Photo Ie). Carbonate separations around the pores (Photo If).

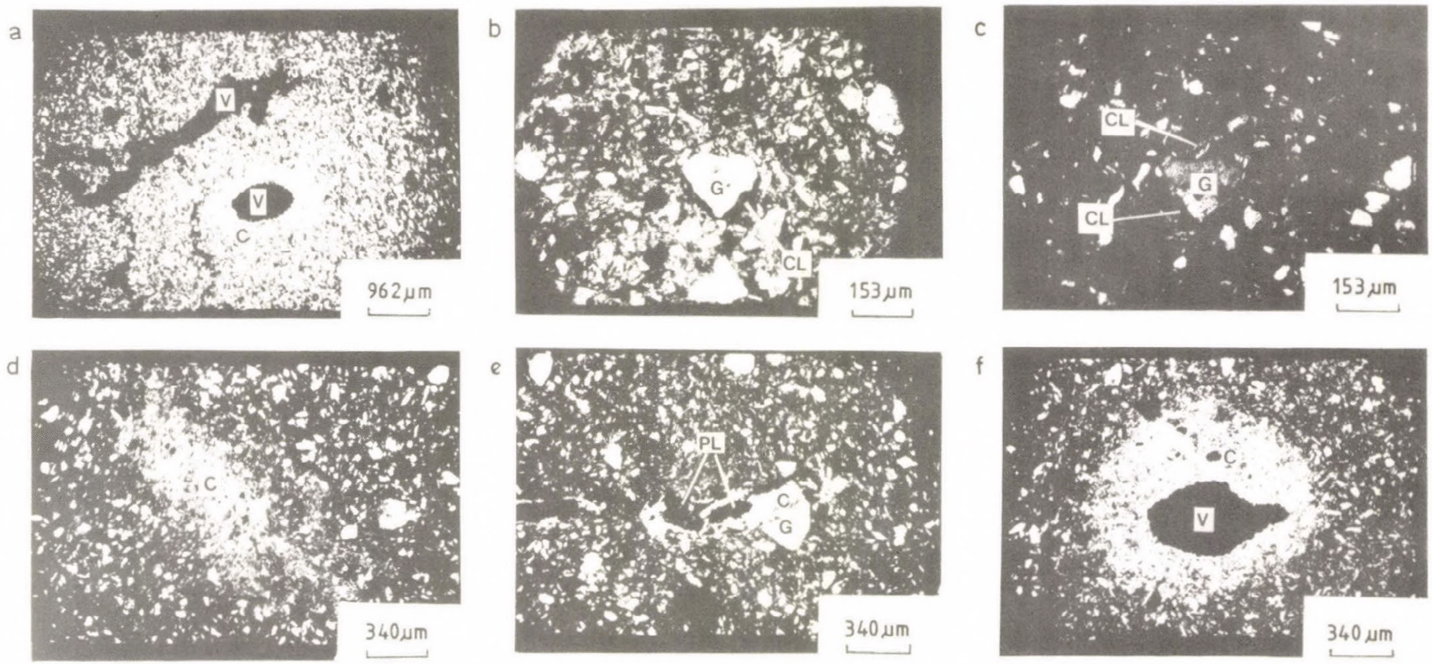

Photo I. Fragments of thin sections from soil BA

a) Cover layer transitional to loess. Voids - channels (V), calcitan (C); XPL

b) A horizon. Coarse angular mineral grain (G) with clay coating - pendent (CL); PPL

c) as b) but XPL

Calcitic pedofeatures in B horizon (d, e, f)

d) Impregnation of s-matrix by calcite (C); XPL

e) Calcite coating pseudomorphosing (C) plant tissue (PL), angular mineral grain (G); XPL

f) Calcitan (C) around biogene channel (V); XPL

Characteristics: Abundant secondary forms of carbonates. Very much humified plasma.

DISCUSSION

When investigating the conditions of Basaharc Lower soil formation the *direct role of the biogenic factor* was striking. It is manifest in the following:
- accumulation of organic matter in the soil. It is attested by dark colour and humus content data *(Table 1)*, and, on micro-level the uniform soaking of soil plasma with dark humus. The distribution and form of organic matter indicate that it was formed within the soil and transformed as a result of micro-organism activity. The conclusion can be drawn that organic matter was produced from subsurface roots and the conditions for microfaunal life were favourable,
- the favourable conditions for microfauna are also proved by the fact that the activity of soil fauna was the main factor in the formation of soil macrostructure, while on micro-level soil structure has retained the loessy aggregation typical of the parent material.

Mechanical mixing of soil material by burrowing animals also gives evidence to active soil fauna (soil material is sharply mottled: spots $d \approx 0.005$ m from earthworms and crotovinas $d \approx 0.05$ m). On micro-level it is manifest in zones variously soaked by humus.

The coarse mixing of the soils substrate and its spotted appearance point to *soil formation of relatively short duration*.
- active carbonate migration leading to their accumulation at base of profile. It cannot be imagined without the catalityc influence of biotic activity.
- Activity of fauna and lack of oversaturation traces allow the conclusion that soil formation took place under the *conditions of balanced water regime* where downward water motion was predominant.

Moderate leaching also played a significant role in soil formation and resulted in the impoverishment of the A horizon in comparison with B in contents of humus and carbonates attended by their accumulation in the deeper soil layers. Micromorphologically it is expressed by the darker colour of plasma, abundance of secondary forms of carbonates and Calciasepic plasmic fabric in soil material from horizon B by the state of plasma and ferruginous pedofeatures in soil material from A horizon.

It is assumed that the mentioned leaching indicates the final stage of soil formation, a transition to cold cycle.

MENDE BASE SOIL COMPLEX (MB)

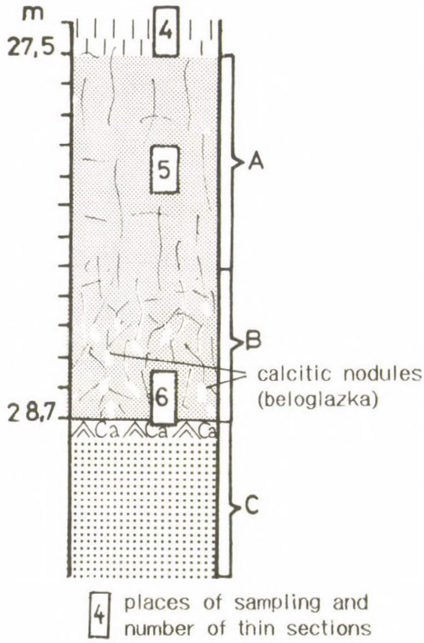

Fig. 2 Soil profile MB

Description of soil profile

The Mende Base Soil Complex (Fig. 2) is situated at a depth between 27 and 29 m at the boundary of old and young loess sequence. The thickness of the profile is 1.20 m, and consists of two parts which can be sharply distinct from each other.

The upper part has a massive structure with rare thin vertical cracks. Very coarse prismatic structure. The consistency of the dry soil is very hard. The colour is bright brown, small (d ≈ 1000 µm) Fe-Mn spots occur. The transition to the lower horizon is clear in structure.

The lower horizon has angular blocky structure. The skew distribution pattern of interpedal voids unlike to the vertical cracks above horizon. Appearance of loose carbonated segregation in cavities between secondary peds.

Transition ot the C_{Ca} horizon with a very sharp stright boundary. The C-horizon is represented by carbonate accumulation. It is loose and seems is no question of breccia. The colour as breccia. But in this case there is light yellow-orange (7.5 YR 8/3.).

Micromorphological Data

Thin section 4. Cover layer transitional to loess (Photo IIa)

Massive structure with short (max. 0.003 m and average 0.002 m long) channels (d ≈ 100-300 µm). Mineral composition of the skeleton consists of carbonates, feldspars, micas (biotite, muscovite), chlorite, amphibole, olivine, primary carbonate, granate, epidote, zoisite. Very angular grains dominated by fine and medium sand.
The skeleton has a random distribution pattern although long-shaped grains have locally often subparallel orientation. The clayey-carbonate plasma is well aggregated. Microcrystalline plasmic fabric.

Special features: Fresh plant remains in the voids, occasionally ferriginous mottle (d ≈ 300 µm).

Characteristics: Inhomogeneity of soil material, particularly evident to the naked eye or at low magnification with oblique incident light, due to zones d ≈ 500 µm which have more ferruginous and clayey s-matrix.

Thin section 5. A horizon (Photo IIb, c)

Channelled massive structure. Porosity: 3-5%, ratio of S/P ≈ 1.

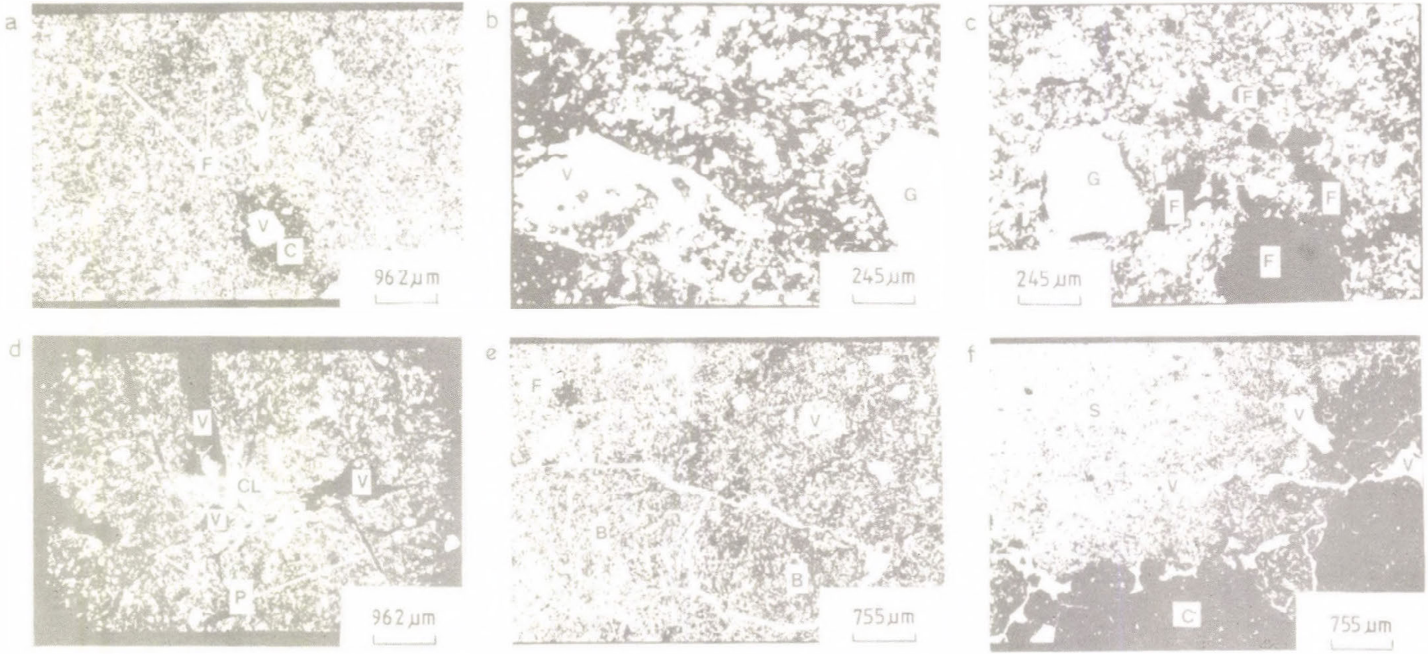

Photo II. Fragments of thin sections from soil MB

a) Cover layer transitional to loess. Channels (V), calcitan (C), Fe-Mn mottles (F); PPL

b) A horizon. Coarse rounded quartz grain (G), large biogenic void with excremental infilling (V); PPL

c) A horizon. Coarse rounded quartz grain (G), Fe-Mn aggregate nodules (F); PPL

B horizon (d, e, f)

d) Subangular structure. Clay separations with striated orientation within s-matrix and associated with planes (P) chambers (V); XPL

e) Crescent or bow-like fabric pedofeatures (B), Fe-Mn mottle (F), void (channel) with calcite coating (V); PPL

f) Sharp boundary B/C_{Ca}. Micritic calcite nodules (C), cracks and chambers (V), s-matrix (S); PPL

The skeleton is much coarser in the cover layer. Grains with size of 400-500 μm (medium sand) are frequent. Mineral composition: quartz, feldspars, less amount of carbonates. Strongly weathered mineral grains occur. The ferruginous-carbonatic-clayey plasma is moderately humified.

Special features: Carbonate separations around channels (calcitans). Cloud like impregnative calcitic pedofeatures in the plasma. Plenty of nucleic and amiboidal Fe-Mn nodules with a size of 400 μm or less (Photo IIc). Charcoal fragments. In colour and organization of the s-matrix rounded (d $\approx$ 5000 μm) zones can be observed. Excremental infillings of large biogenic voids subject to high biogenic activity (Photo II b).

Thin section 6. B horizon and boundary B/C_{Ca} (Photo IId, e, f)

Subangular-blocky structure with chambers (size about 300 μm). Medium macrovoids, porosity 20-25%. The S/P ratio $\approx$ 1. Skeleton is similar to the above horizon (see thin section 5) but less of rounded coarse skeleton grains occur, ferruginous-clayey plasma has Skell-vo-masepic plasmic fabric.

Special features: Fe-Mn concretions dominate in form of small (d 1000 μm and less) amiboidal nodules (ferrugionus mottle). Charcoal fragments in the s-matrix are not frequent. Fabric pedofeatures: crescent or bow-like (Photo IIe), passage features and slickensides (Photo IId). Carbonates are present in form of nucleic micritic calcite nodules with sharp boundary, completely isolated from s-matrix in cracks between peds and compose C_{Ca} horizon and also form coatings within voids in the immediate vicinity ($\approx$ 0,03 m) of boundary B/C_{Ca} (Photo IIf).

Characteristics: plenty of clay separations with striated orientation occur as zones within the s-matrix in form of bow-like arrangement subject to high biological activity and are associated with the walls - slickensides (Photo IId) as evidence of differential movement under pressure.

DISCUSSION

The morphological features of Mende Base Soil Complex show *the deep transformation of the substrate* during soil formation process. Substrate changes are manifest in:

- profound changes in the arrangement of material. For instance, B horizon totally lost the arrangement of material characteristics of loess and became Skell-vo-masepic plasmic fabric at microlevel and subangular-blocky structure at macrolevel.
- the apppearance of striated clay features embedded in the decalcified soil matrix and the phenomenon of slickensides along planes.
- intensive weathering of material expressed by the occurrence of strongly weathered mineral grains and bright brown colour of the soil due to presence of free Fe oxides.

The soil profile indicates a formation of *intricate genesis* and micromorphological data support the view of the *paleosol MB as a soil complex.*

The upper and lower part of the profile, the horizons A and B differ basically and the processes occurring in them are, in some respect, of opposite direction.

This is manifest, first of all, in the migration and distribution of carbonates.

In the upper part of soil profile (cover layer and A horizon) s-matrix is enriched in carbonates as indicated by the abundance of calcitic pedofeatures and calcareous plasma. On the contrary, in the lower part (B horizon) s-matrix is almost completely refined from carbonates which are segregated in form of isolated nucleic nitritic calcite nodules in cracks between peds and in C_{Ca} horizon which is completely composed by them.

Such a distribution of carbonates can be explained if a change of soil formation type is assumed in a way that the type which led to refining from carbonates was the initial one.

There is no evidence that the type of soil formation changed through mechanical deformation (deposition).

Content and state of organic matter are evaluated as follows: analytical determinations of total humus content and morphological data indicate *minimal humus content and normal distribution* in the profile. Micromorphologically it is confirmed by the fact that carbonatic plasma is moderately humified in A horizon and charcoal fragments occur in the whole depth of the profile. The latter makes us assume that conditions within soil profile were not favourable for microbial transformation (of organic matter into humus) as in the case of the Basaharc Lower paleosol).

Low humus content, concurring with evidence of a very high activity of soil fauna and occurrence of charcoal fragments support the conclusion that the accumulation and decomposition of organic matter took place not within but on the surface of soil profile like in case of forest litter. In this case not the organic matter itself is incorporated into the soil but the agressive products of its decomposition, which, among other factors promote the profound changes mentioned above.

Plenty of ferruginous and manganiferous pedofeatures indicate *conditions of oversaturation* in soil.

The profound changes affecting the structure of soil material described in the paper cannot be imagined without considering the role of the fluid phase of soil manifested in the alternation of intensive wetting and drying.

SUMMARY

As stated above, the ultimate objective of studying paleosols is the reconstruction of the paleoecological environment. Finally, comparison is made between the environments of formation of the two soils, to the extent it is made possible by the micromorphological data obtained in the investigation.

The formation of the Basaharc Lower paleosol took place under condition of 'moderate' and balanced water regime with equal temporal distribution of precipitation, which was not too abundant. Temperatures, precipitation and, as a consequence, vegetation conditions favoured the development of soils benefitial to soil fauna activity (rodents, earthworms and micro-organisms). Their activity had an impact on soil structure and mixing of soil material. Soil fauna was the main structure forming factor in soil BA. During soil formation loess material did not undergo deep transformations and preserved on microlevel typical for loess structural organization. Nature of decomposition of organic matter, its distribution along and transformation within the profile indicate the predominance of steppe vegetation at that time.

Compared with the previous soil the Mende Base Soil Complex reflects a wider range of changes in ecological conditions and a longer duration of soil formation. It is indicated by deeper transformation of loess material in soil MB in comparison with BA as a consequence of the higher gradient of air temperatures and atmospheric precipitation during the formation of soil MB. In its formation together with the biogenic factor fundamental role was played by the physico-chemical processes of interaction between the solid an liquid phases of the soil.

Morphological features of soil MB give basis to assume predominance of arborescent vegetation (decidious forests) during optimum stage of soil formation and later a transition to steppe phase.

These conclusions, drawn on the basis of micromorphological features of studied soils, agree with M. PÉCSI (PÉCSI, M. et al. 1977): the conception that Basaharc Lower Soil (BA) is a well developed compact chernozem-type soil, while Mende Base Soil complex (MB) is a double paleosols, upper member of which is a chernozem-like soil and the lower one a Parabraunerde that can be classified as a reddish-brown mediterranean-type forest soil.

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RELIEF-FORMING PROCESSES DURING THE LAST ICE AGE

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ABSTRACT

On the basis of certain groups of geomorphic processes controlled by climate several zonal types of morphogenesis are distinguished, including glacial, periglacial, humid temperate, semi-arid, arid temperate, arid tropical, semi-humid tropical, subtropical, humid tropical and equatorial types. A short description of the main processes is given for each type. Areal distribution of the main types of morphogenesis within the extratropical northern hemisphere at present and during the last glacial maximum is shown at small-scale maps.

INTRODUCTION

Geographic research today is increasingly focusing on the evolution of landscapes during the natural macrocycle (from about 120,000 to 100,000 years ago) which embraced the last interglacial (Eem, Mikulino, or Sangamon) the last glaciation (Würm, Valdai, or Wisconsin), and the recent interglacial - Holocene. This interest is quite understandable because most of the prognostic concepts are based on paleogeographic data used to model the range in landscape variations under conditions of both natural and human impact on the environment.

Because of the greater importance of endogenic factors, present-day topography, at least its large forms, seems to be less climate-controlled than other landscape components. If, however, we turn our attention to the dynamics of the relief forming processes (this aspect is now in the limelight because of increasing deterioration induced by man) we see that they, too, proceed under climatic control.

The correlation between climate and relief can be approached from different perspectives. The first is based on an elaboration of systems of morphoclimatic zonation or types of exogenic morphogenesis originally developed by J. TRICART (1957), L. WILSON (1969), J. BÜDEL (1977), and H. WILHELMY (1975). A.P. DEDKOV and his colleagues (DEDKOV, A.P. et al. 1977) developed one of the most detailed and elaborate classification schemes for the relief-forming processes in various landscape zones. They definitively demonstrated that different zones are affected by different processes.

On the other hand, individual relief-forming processes and their genetic groups appear to be restricted to certain climatic and landscape conditions. Surface slope wash, for example, is a prevalent process on the semi-arid landscape, and to a lesser extent - on dry periglacial ones (in both cases due to scarcity of the vegetation cover). Eolian processes act in hot or cold deserts, while gelifluction occurs mostly in permafrost regions. Traces of such processes which are fixed in landforms and deposits, can, therefore, be interpreted as evidences for paleoclimates and paleoenvironments. Some of these processes act within such narrow limits of climatic parameters, that the resulting landforms can be directly interpreted in terms of past climates. They include cryogenic forms (which indicate ground temperatures), fluvial landforms (hydrological regime, river discharge), proluvial fans (discharge of flash floods), valley landforms (created by floating ice), many eolian formations (wind regime), etc. Since they were considered in detail elsewhere (SPASSKAYA, I.I. 1985), here we will just briefly discuss the main genetic groups of relief-forming processes, and outline the most favourable environmental conditions for each of them.

PRESENT-DAY GEOMORPHIC PROCESSES CONTROLLED BY CLIMATE

The most general groups can be listed as follows:

1) Glacial processes (due to the action of moving glaciers). In this group we include the action of meltwater at the ice surface, at the ice-bedrock interface, within the ice mass, and, finally, near the ice margin. At present, these processes are rather limited in their distribution and confined to the high arctic.

2) Cryogenic processes which result from the impact of low temperatures on soil, especially in cases where temperatures fluctuate above and below 0° C. This group includes various processes of frost sorting, frost creep (downslope movement of debris due to needle ice), frost heaving, ice wedging, altiplanation, etc.

3) The large group of slope processes which, strictly speaking, includes all the processes of downslope movement of the weathered material. This group can be subdivided into:

a) gravitative (in the narrow sense) movements, such as rockfalls, slumps, rock avalanches, debris flow, etc. These are primarily controlled by geologic and topographic factors.

b) various kinds of creep-processes. These are nearly universal and result from a great variety of factors including cycles of freezing and thawing, wetting and drying, heating and cooling, as well as from some biogenic activities (burrowing animals, plant roots), and even from chemical processes (e.g.

the growth of gypsum crystals). They are most typical for the humid temperate zone, but other climatic conditions can give rise to some specific kinds of creep (thermal creep in deserts, frost creep in tundra) as well.

c) Solifluction type of viscous-fluid flow. This requires that the soil have a considerable water content and mostly occurs in the permafrost zone (within the active layer) and in the wet tropics.

4) Another large group of processes are produced by water flow over land surfaces. This group includes sheet and rill wash on slopes insufficiently protected by vegetation, splash erosion, the action of small ephemeral watercourses, as well as perennial rivers forming well developed valleys. Sufficient precipitation to induce surficial runoff is a general pre-requisite for this whole group of processes. This pre-requisite is met in several zones.

5) Eolian processes (deflation, the formation of hollows or of desert pavement, sand transport by air flow, and sedimentation of fine particles from the air) are restricted mostly to arid areas as well as to some regions in the periglacial zone which have open vegetation and sufficiently strong winds.

In this overview, we intentionally limited ourselves to the groups of processes which we consider to be primary and did not take into consideration a number of others, such as karst processes and wave action. It should be noted that the zonal distribution of processes depicted in *Fig. 1,* are shown for so-called "natural" environment without human impact.

Given the above, we can outline the following main types of morphogenesis controlled by climatic-landscape factors:

glacial type,
periglacial (present-day polar desert and tundra),
humid temperate (including most parts of the forest and steppe zones),
semi-arid (steppes and semi-deserts),
arid temperate (including extratropical deserts with cold winters),
arid tropical ("hot" deserts),
semi-humid tropical and subtropical (monsoon, seasonal wet tropics),
humid tropical and equatorial.

Table 1 gives a short description of the predominant present-day processes operating in various zones together with the typical landform indicators for each complex of processes. The two can be used for reconstructing morphogenesis of past epochs.

In general, the morphogenetic types outlined above have much in common with those previously indentified by SHCHUKIN, I.S. (1960) and BÜDEL, J. (1977). We do not agree, however, with the latter in ascribing active processes of valley formation to the periglacial tundra. Based on the research of Soviet and other specialists, we argue that fluvial processes in tundras (especially in the continental Siberian one) are rather limited in scope. This is due to the short frost-free period and small contents of debris in channels.

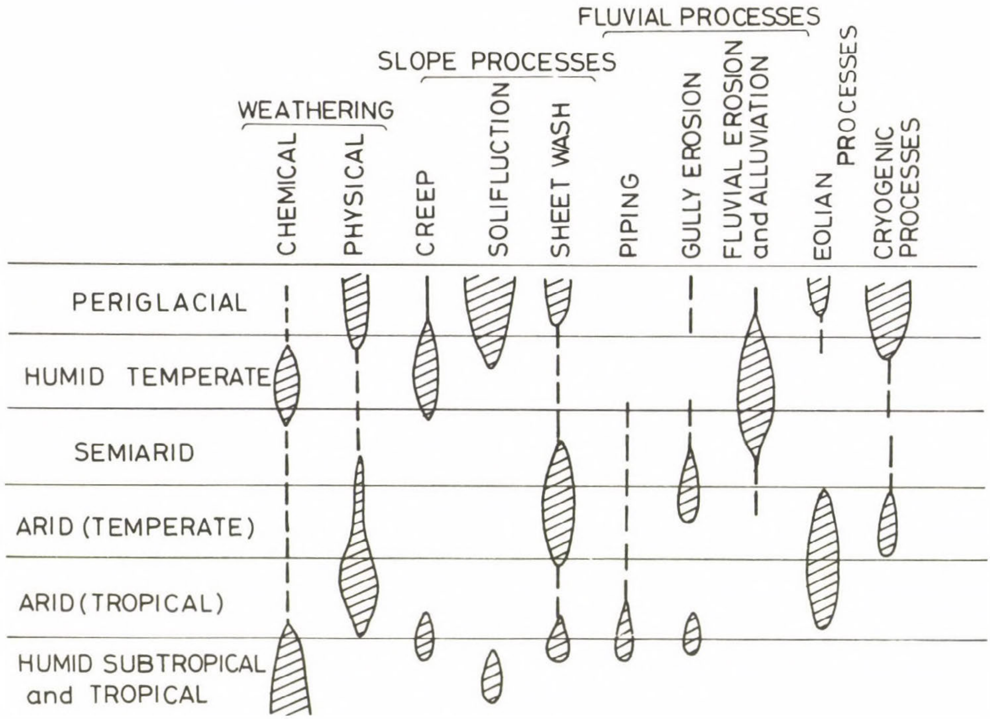

Fig. 1 Predominating morphogenetic processes pertaining to various landscape zones

The "sectoral" pattern of the processes within the zones, which overall reflect the degree of the continentality of the climate, should also be considered when looking at morphoclimatic zonality. For example, regional differences are quite discernible in the present-day humid temperate zone (EMBLETON, C. 1984). In Western Europe, which now has a humid mild climate, river discharge is regular, and river valleys are wide and have graded profiles. Valley slopes there are long and graded, and slow creep is the predominant process. Eastern Europe features a greater degree of erosional dissection. This is because more severe winters here produce soil freezing and a fluctuation in river discharge. Winter discharge is low and is followed by spring flooding. Under "natural" conditions, relief in this area undergoes slow evolution because of the protective effect of vegetation. Human activity here, however, accentuates some erosional processes such as slope wash and gully erosion, which are tpyical of more arid regions.

Numerous data from the Russian Plain indicate that fluctuations in discharge through the year also result in active evolution of valley floors. Thus, in regard to this humid zone, we can hardly agree with BÜDEL, J. (1977) who considers the whole zone an example of "abated valley formation" ('ektropische Zone retardierter Talbildung').

Table 1 Relief-forming processes in the main present-day landscape zones

Zones	Prevailing processes	Landforms
Periglacial	Physical weathering, frost cracking, solifluction, block streams, ice wedging, frost upheaval; in drier regions - frost creep, nival-eolian processes, altiplanation, deflation	Patterned ground, pingo, deflation pavements
Humid temperate	Physical and chemical weathering, fluvial processes (permanent flows), creep, in more continental areas on permafrost-solifluction	Well-developed valleys, covered karst
Semi-arid	Physical and chemical weathering, sheet and rill wash, gully erosion; locally-cryogenic processes	Dense erosional dissection, pediments
Arid	Mostly physical (including salt and thermal) weathering, eolian processes, occasionally-ephemeral stream erosion, pedimentation, in deserts with cold winter - frost cracking, frost sorting etc.	Complex of eolian landforms
Transitional from arid to humid tropics	Re-activation of erosion processes, most active piping and gully erosion	
Humid tropical	Chemical weathering, tropical solifluction, karst, chemical erosion	Planation surfaces with thick weathering crust

Vertical formation makes the outline of morphogenetic types in mountains difficult. In general, it is possible to distinguish "periglacial mountains" with well pronounced cryogenic processes and small local glacial formations from "temperate mountains", covered with forests, where fluvial and non-cryogenous slope processes dominate. In addition, a rather large area is occupied by mountains with mixed semi-arid and humid morphogenesis. Here the local pattern of processes is controlled by topography. Purely arid mountains are not very numerous, but they do contain a spectacular complex of processes ("desert" weathering including exfoliation, thermal creep of debris, all kinds of desiccation cracks, etc.).

GEOMORPHIC PROCESSES OF THE LAST GLACIAL MAXIMUM: A RECONSTRUCTION

Figures 2 and *3* schematically show the areal distribution of the noted processes extant in the extratropical northern hemi-

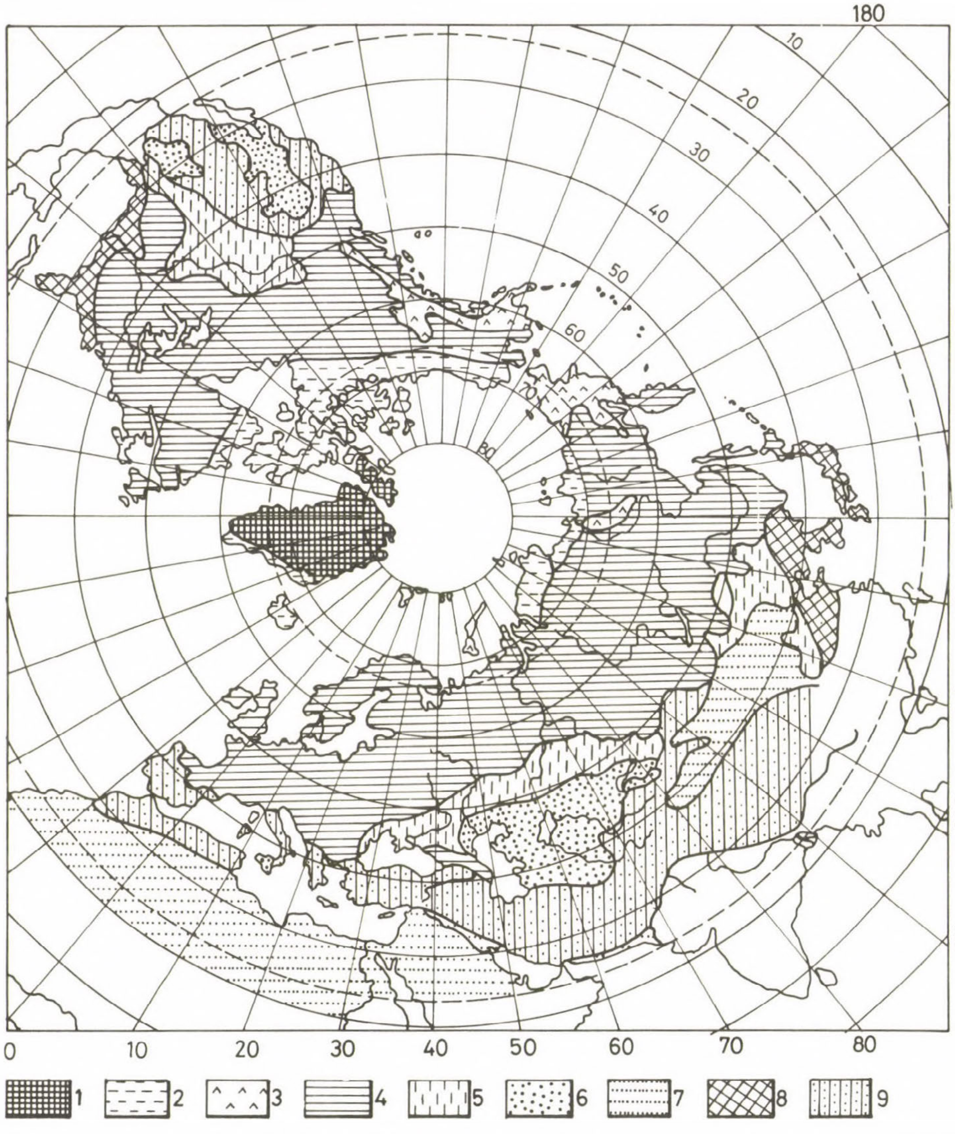

Fig. 2 Present-day distribution of zonal types of morphogenesis

1 = glacial; 2 = periglacial (plains); 3 = periglacial (mountains); 4 = humid temperate; 5 = semiarid; 6 = arid temperate; 7 = arid tropical; 8 = semihumid subtropical (seasonally wet); 9 = mixed semiarid and semihumid in mountains

sphere today and during the last glacial epoch. The reconstruction of the relief-forming processes during the glacial period are based on numerous summaries of paleogeographic and Quaternary studies including "Paleogeography of Europe during the

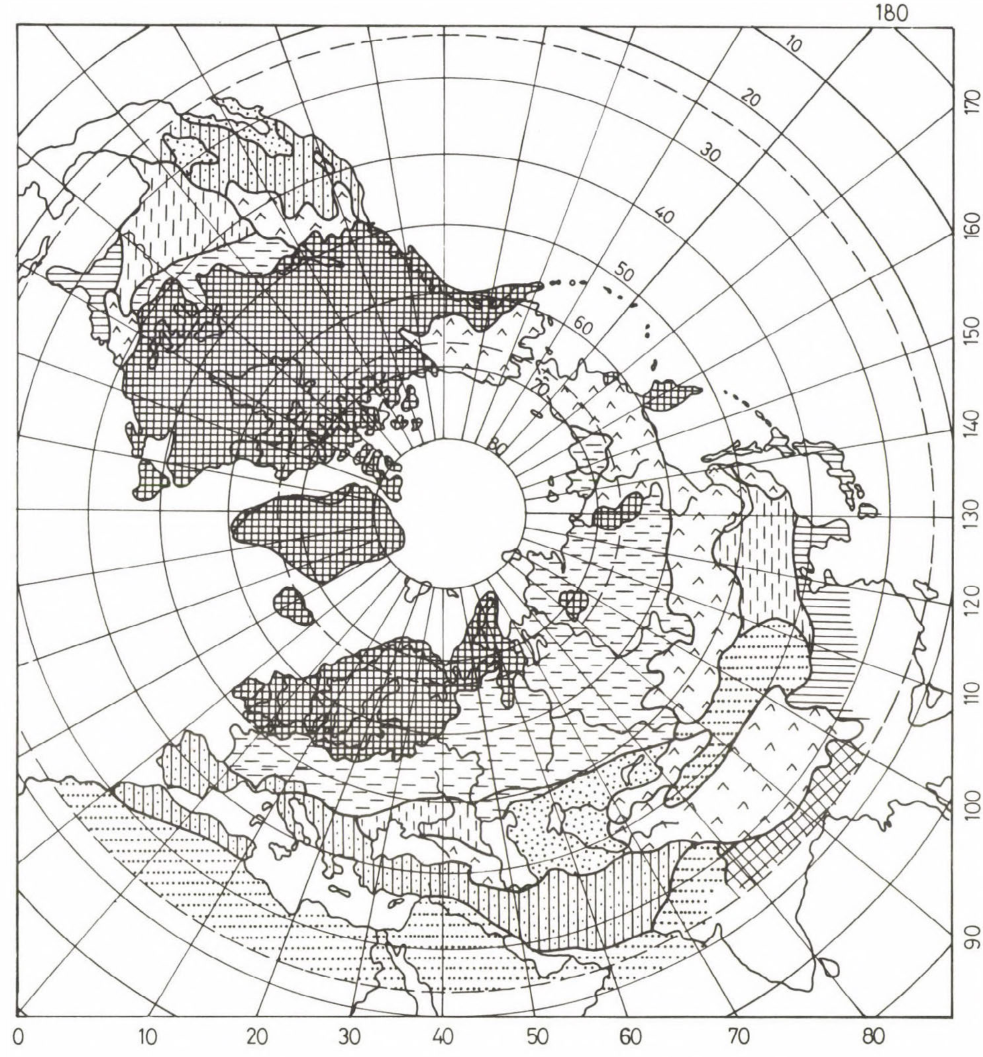

Fig. 3 Distribution of the main zonal types of morphogenesis during the Late Pleistocene glaciation (For legend see Fig. 2)

last 100,000 years" (GERASIMOV, I.P.--VELICHKO, A.A. eds., 1982), "Late Quaternary environments of the United States" (WRIGHT, H.E. ed. 1983), as well as on materials published by the working group of the IGCP project "Quaternary glaciations of the Northern hemisphere" and by CLIMAP project members (1976)

The resulting reconstruction, which is by no means complete, shows that morphogenesis of the glacial period was drastically different from that observed today. These differences were produced by colder and drier climate (paleoclimatic data indicate global cooling by 3° C). This cooling brought about an increase in the areas occupied by glaciers and by low-temperatures permafrost, and resulted in active glacial and periglacial morphogenesis. In the high and middle latitudes vast areas were covered by glaciers. Many types of glacial landforms and deposits have been extensively outlined for these areas in the literature and need not be discussed here. Although many mountains in the periglacial zone were covered by alpine glaciers, *figure 3* shows only continental ice sheets and mountain glacial complexes (like the Cordilleran in North America).

The Pleistocene periglacial zone which was widespread in Europe also contained some regional differences (like the present humid temperate zone). These resulted from the different degree of continentality in the different regions. The more "maritime" climate of Western Europe was conducive to cryoturbations, involutions, and solifluction, while the Russian Plain and Siberia preserve numerous traces of more arid and severe environment which A.A. VELICHKO described (1973) as "tundra-steppe". Within this zone morphogenesis was primarily controlled by low winter temperatures, high continentality, relative aridity, low intensity of rainfalls, and by the scarcity of vegetation. Under those conditions cryogenous processes were most important. Those, however, which required high moisture content in soils, developed only locally. Frost fissures formation seems to have been prevalent. It created regular polygonal nets and preconditioned posterior processes of surface modelling. Under conditions of low rainfall intensity, the surface runoff could not have been very active. Rain water, however, may have created deep gully-like rills on valley slopes insufficiently protected by vegetation.

On watersheds, sheet wash accounted for some short-distance transport of loose material (that, for example, which was deposited by wind and not yet fixed by vegetation). Surface runoff increased in different localities as a result of an accumulation of wind-driven snow and its subsequent melting in the spring. Under very low winter temperatures the snow of micro-crystalline structure is very loose. It can be easily carried by wind and can accumulate rapidly in sheltered places. Meltwater results not only in sheet wash, but in solifluction as well. This induce a further development of linear hollows. This process facilitates a greater accumulation of snow during the following winters, which, in turn, intensifies denudation. This ongoing process represents a clear example of positive feedback. Spatially differentiated denudations proceeds at the same time that eolian dust is being deposited and forming the loess mantle.

The existence of arid conditions during morphogenesis is also indicated by the widespread presence of fossil dunes, by traces of wind erosion (numerous ventifacts and desert pavements in present-day Siberian taiga), and by the loess mantle. Aridity also persisted in Soviet Central Asia in the present-day areas of steppes and semi-deserts. Only in the Far East, where the climate was wetter due to the influence of the Pacific, did solifluction create thick deposits at the foot of the slopes.

It is important to note that both climatic features of the time (i.e. low temperatures and aridity) slowed down the rate of the denudation processes. The lack of moisture in the soils lessened the rate of fluid-plastic movement which transferred the weathered materials from the watersheds into fluvial arteries. The frozen state of regolith during most of the year lowered by an order of magnitude the rate of denudation. Because of this, watersheds were often excluded from the denudation systems and were preserved almost intact. The same reduction was typical for the low mountains. Small rivers here (often completely frozen in winter) could not transport the debris supplied by physical weathering, and it accumulated in the internal basins. Denudation processes controlled by topography were continuously active only in the high mountains.

The reconstruction of the relief-forming processes extant during the Late Pleistocene glaciation is of interest because it presents an important environmental component at one of the "extremes" of the natural-climatic cycle. When these data are supplemented by similar reconstructions for the warmest phases of the cycle, we will have information on both the range of fluctuations in the rates of these processes and on their areal distribution. This information is significant for our models of future development of these processes.

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EVOLUTION OF THE LATE PLEISTOCENE ENVIRONMENT IN THE CAUCASUS

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ABSTRACT

Rhythmical climatic changes in the Caucasus during the Late Pleistocene are reflected by the Black an Caspian sea level changes (phases and stages) in shifts of mountain glaciers and in archaeological evidence.
An attempt is made at the correlation of changes of environmental conditions in the Caucasus against the background of climatic oscillations within the Late Pleistocene macrocycle. A chronostratigraphic correlation scheme is suggested.

EVOLUTION OF THE LATE PLEISTOCENE ENVIRONMENT IN THE CAUCASUS

The reproduction of a possibly complete picture of the stage-by-stage development of the natural conditions in the Caucasus during the Late Pleistocene is of paleogeographic significance, since in geological-geomorphological respect by means of the example of this complex it is possible to reveal the main regularities of the development of natural phenomena in the adjacent areas of the Alpine orogenic belt, as well.

As a result of a prolonged effect of endogenous and exogenous processes on the relief, in the Late Pleistocene the territory of the Caucasus had the appearance of a mountain region close to the modern one with a well-developed hydrographic network and altitudinal landscape zonation.

At present, it is generally accepted that in the Pleistocene the glaciation of the Earth was caused by the climatic changes of planetary scale. The analysis of the actual data on paleogeography of the northern hemisphere and on the ancient glaciation enables to conclude that the origin of glaciers at the beginning of the Quaternary on the highest massifs of the Caucasus and in the glaciation centres of the Russian Plain (in Scandinavia and Novaya Zemlya) occurred almost simultaneously. Later the growth and dynamics of the ice cover of plains played a major role in the development of glaciation in the Caucasus *(Table 1)*. Glaciers in the mountains in contrast to those on the plains continued to exist even under significant climatic oscillations manifested in the Pleistocene. While, as a result of deglaciation, in the interglacial the glaciers almost disappeared in the plains, in the mountains they only decreased, and began to advance again with the onset of the next cooling.

Thus, the Late Pleistocene glaciation in the Caucasus as a modern one had an inherited character. It can be described more comprehensively than other Quaternary glaciations due to the good preservation of paleoglacial features, as well as of the glacial and lacustrine-glacial deposits.

LATE PLEISTOCENE GLACIAL EVENTS IN HIGH MOUNTAINS

In the Caucasus the authentic traces of the Late Pleistocene glaciation in the form of frontal and lateral moraines, as well as troughs, riegels, corries, horns are encountered almost in all valleys, where present-day glaciers are developed. They are located also in areas where the latter are missing.

A detailed analysis of the available material and the author's observations in the Little Caucasus enable to distinguish two phases of cooling and the advance of glaciers, and two phases of warming in the climatic macrocycle of the Late Pleistocene of the Caucasus. The early phase of warming (corresponding to Mikulino Interglacial) represents a mesocycle in the system of natural changes, while the late phase could correspond to the microcycle and is an interstadial.

Having studied the mountainous part of South Ossetia, L.A. VARDANYANTS pointed already in 1932 to the existence of two phases of Late Pleistocene glaciation in the Greater Caucasus. In the central part of the Greater Caucasus (northern slope) E.E. MILANOVSKY (1966) distinguished the early and late Bezingian phases of glaciation, while the early and late Nenskrian glaciation phases were distinguished on the southern slope (TSERETELI, D.V. 1966), and in the Southeastern Caucasus - the Shahnabadian glaciation with two phases, i.e. Lazinian and Shahdjuzinian were determined (DUMITRASHKO, N.V.--BUDAGOV, B.A. 1985; BUDAGOV, B.A. 1964). Cooling determined two advances of glaciers and caused the displacement of altitudinal landscape belts towards the intermontane depressions. These phases were separated by the period of general warming, being manifested by the formation of lacustrine-fluvioglacial deposits in the retreat zone of glaciers.

The extent of glaciation on the northern slope of the Greater Caucasus considerably exceeded that on the southern slope. The glaciation in the south-eastern part of the Caucasus was more limited due to extreme aridity and continentalitiy.

The traces of glacial deposits of the early Nenskrian (Kalininian on the Russian Plain) glacial phase can be detected in the form of frontal and lateral moraines in the southern slope of the western and central part of the Greater Caucasus between 620 and 1600 m (TSERETELI, D.V. 1966). In the south-

Table 1 Stratigraphic correlation scheme of the Upper Pleistocene in the Caucasus

Subdivision	Regional scale: European part of USSR, Super-horizon	Regional scale: European part of USSR, Horizon	Regional scale: Caucasus, Super-horizon	Regional scale: Caucasus, Horizon	Black Sea basin	Black Sea basin	Absolute age, thousand years	Caspian Sea basin	Caspian Sea basin	Absolute age, thousand years	Anticaucasus	Archaeological age	Alpine stratigraphy
Holocene					Chernomorian			Novocaspian			Holocene	Bronze Eneolithic Neolithic	Holocene
UPPER PLEISTOCENE	VALDAI	Ostashkovian	Bezingian (Nenskrian)	Late Bezingian (Late Nenskrian)	Novoeuxinian	Novoeuxinian III regression (Novoeuxinian)	11.6 15.0 23.0	Khvalynian	Upper Khvalynian transgression	11.3 13.0 15.8	Mountain glaciation	Mesolithic Upper Paleolithic	W_2
									Enotaevian regression	25.6 29.2			
		Leningradian (Mologo-Shexnian)		Aktoprak (Bechoian) interstadial		Novoeuxinian II transgression (Surogh?)	29.2 33.0 41.2 49.0		Lower Khvalynian transgression	33.0 37.0 42.0 47.0	Beshumian interstadial	Late Mousterian	W_{1-2}
		Kalininian		Lower Bezingian (Lower Nenskrian)		Novoeuxinian I regression (postkarangatian)	55.0 60.0 74.0		Athelian regression	60.0 75.0	Mountain glaciation	Early Mousterian	W_1
		Mikulino		Interglacial		Karangatian transgression	80.0 88.0 95.2 110		Upper Khazarian	80.0 88.0 91.0 120	Interglacial	Late Acheulian	R-W

eastern part of the Greater Caucasus the glaciers reached the height of 2600 m in the Lazinian phase of glaciation (DUMITRASHKO, N.V--BUDAGOV, B.A. 1958; BUDAGOV, B.A. 1964). At that time snow line was located lower by 700-800 m in the Western Caucasus, by 900-1100 m in the Central Caucasus, by 600-700 m in the Eastern Caucasus and by 1100-1300 m in the South-eastern Caucasus.

In the Transcaucasus characterized by lower absolute altitudes of the mountain structures and by drier climate, the scale of the Late Pleistocene glaciation was significantly more modest (BALYAN, S.P.--DUMITRASHKO, N.V. 1962; DUMITRASHKO, N.V. 1982).

In the western areas frontal moraines are observed at altitudes of 1500-1550 m, while in the Transcaucasus upland at 2000-2200 m. In the west the snow line was at the absolute altitude of 2000-2400 m, and in the East at 2600-2700 m; the descent of the snow line was equal to 400-600 m (MARUASHVILI, L.I. 1956; MAISURADZE, G.M.--KLOPOTOVSKAYA, N.B. 1969; MAISURADZE, G.M. 1973; 1982).

Glacial deposits of the late Nenskrian (Ostashkovian for the Russian Plain) glaciation are located 300-500 m above the early Nenskrian moraines. In spite of the maximum severity of climatic conditions, the glaciers of the second phase were smaller in size.

The results of the comprehensive investigations, studies of the Pleistocene fauna, flora, the lithology of some deposits, and the cryogenic phenomena indicate that in the Caucasus there was an abrupt cooling and aridization of climate in the late Nenskrian phase (SOLECKI, R.S. 1964; NEUSTADT, M.I. et al., 1965; VELICHKO, A.A. 1968; GRICHUK, V.P. et al. 1970; MAISURADZE, G.M. et al. 1975; MARUASHVILI, L.I. 1975). The location of moraines of this phase by several hundred metres higher than of the preceding ones enables to suggest that in spite of a significant cooling in the late Nenskrian phase, the scale of glaciation was smaller than that in the early Nenskrian. In Transcaucasia (in the area of Beshumi settlement, Arsiani range) the discovery of intermoraine lacustrine-fluvioglacial deposits deformed by cryoturbation points to the cryothermal conditions of that time. The results of palynological studies confirm a significant downward displacement of vegetational belts (MAISURADZE, G.M.--KLOPOTOVSKAYA, N.B. 1969).

The generalization of the actual material allows to suggest that the warming within the Nenskrian glaciation (Bechoian phase, Aktoprakian phase, Leningradian for the Russian Plains) was insignificant and was of an interstadial character. Lacustrine-fluvioglacial deposits are associated with this period. According to the calculation of annual layers in varved lacustrine clays at Aktoprak settlement (northern slope of the Central part of the Greater Caucasus) the duration of the interstadial phase was 15-20 thousand years (MILANOVSKY, E.E. 1966) It was stated that the pollen of broad-leaved trees prevailed at the altitude of about 1500 m on the southern slope of the Greater Caucasus in the intermoraine lake deposits of the Mulkhra river valley (Upper Svaneti). The accumulation of lake deposits took place under warmer and more humid conditions than in the periglacial (TSERETELI, D.V. 1966).

In the Caucasus the rhythm of changes of the natural conditions in the interval of the geological history considered left its imprints in the facies succession of the Ponto-Caspian nearshore marine deposits, as well as in the archaeological evidences of multi-layer mountain caves.

HISTORY OF THE BLACK SEA BASIN

In the Caucasus at the beginning of the Late Pleistocene the transgressions of the Ponto-Caspian sea basins took place: the well-known Karangatian transgression of the Black Sea and the Upper Khazarian of the Caspian Sea. At present, it is generally accepted that the Karangatian glacio-eustatic transgression is correlated with the Mikulino (Riss-Würm) Interglacial. This was the time of a warm and probably even arid climate (FEDOROV, P.V. 1978).

The Karangatian sea basin was rich in fauna of Mediterranean species. The small discharge of the rivers flowing into the Black Sea and the transfer of waters of the transgressing Mediterranean Sea brought about an increase of salinity in the basin up to 30%. In the Karangatian the natural conditions were aggravated by oscillations. They had a cyclic character and were manifest by the changes of climatic situation within the interglacial (warming and cooling). Recent data obtained by A.P. SHCHEGLOV (1986) point to three warm climatic microcycles and high level of the Karangatian sea, respectively. It is expressed by the formation of three marine terraces lying in 7-9 m, 14-16 m and 25-27 m a.s.l. in the northwestern regions of the Caucasian Black Sea coast. Weathering crusts dated by the TL method to 87, 111.1 and 120.8 thousand years, respectively, are fixed in subaerial deposits located on marine terraces. Transgressive warm microcycles alternated with regressive cold microcycles, which are reflected by the existence of incised river valleys and by the accumulation of poorly rounded "cold" basal alluvium on river terraces of the corresponding age.

In Kolkhida the Karangatian marine terraces are encountered at the altitude of 12-15 m, and in the area of Sochi - up to 30 m. Synchronous marine deposits with the Mediterranean fauna were detected by wells in Kolkhida in a depth of 95-105 m (DJANELIDZE, Ch.P.--MIKADZE, I.S. 1975).

In the central part of Kolkhida (the mouth of the Rioni river) the marine Karangat is overlapped by a regressive series of lagoonal-deltaic and alluvial deposits in the depth interval of 80-95 m. The regressive (Novoeuxinian I) phase subsequent to the Karangatian transgression is contemporaneous with the beginning of the early Nenskrian glaciation, the fact being proved by the incising of river valleys in the down-stream and mid-channel and by the accumulation of "cold" basal rubbly alluvium in the river beds.

The traces of overdeepening up to 60-80 m are detected along the Caucasian shelf, where alluvial-deltaic sands were drifted to. The renewal of volcanism in the Greater Caucasus and in the Transcaucasian volcanic upland is also related to this chronological section.

The discovery of marine layers of the glacio-eustatic transgressive phase (Novoeuxinian II) in Kolkhida in a depth of 52-80 m (DJANELIDZE, Ch.P.--MIKADZE, I.S. 1975) was important. Marine clays and sands with brackish-water fauna alternate here with peat beds. The age of peat fixed by a well in a depth of 64 m is 31,300 years (Tb-56). Marine clays and clayey sandstones have an ingressive character of bedding. They contain fauna of the desalted basin and are fixed by wells in the central part of Kolkhida in the radius of 12-15 km from the mouth of the Rioni river (LALIEV, A.G. 1957; TSERETELI, D.V. 1966). Novoeuxinian II marine deposits are also observed in the northwestern part of the Black Sea shelf in a depth of -20 to --25 m. These are Tarkhankutian layers with brackish-water fauna (NEVESSKAYA, L.A. 1965). The analysis of the material shows that the Novoeuxinian II transgression did not reach the recent sea level: in the peak of transgression -30 to -25 m is indicated. Radiological data obtained to date fit in the time frames corresponding to the Middle Würm transgression of the world ocean, i.e. 50-22 thousand years ago (KIND, N.N. 1963; SEREBRYANNY, L.P. 1963; GUILCHER, A. 1974; MÖRNER, N.A. 1974; EMILIANI, C. 1964, etc.).

The Novoeuxinian II transgressive phase or climatic microcycle can be parallelized with the intra-Nenskrian interstadial. At that time a retreat of glaciers and accumulation of lacustrine-fluvioglacial deposits occurred in the mountains of the Greater and Little Caucasus.

In the Novoeuxinian III phase, in connection with a sharp cooling, the Black Sea basin underwent a regression, at which the level lowered by 50-60 m below the present one. In the interval of 37-52 m of the Kolkhida wells lagoonal-deltaic deposits were discovered, which are replaced further inland with terrigene-alluvial deposits. The regression caused the renewed incision of large river valleys and accumulation of basal "cold" boulders. It is supposed that the regression attained its maximum 20-18 thousand years ago.

HISTORY OF THE CASPIAN SEA BASIN

The rhythmic changes of natural conditions in the Late Pleistocene are reflected also in the Caspian area, where transgressions and regressions of the Caspian lake-sea were controlled by climatic factors (RYCHAGOV, G.I. 1977; FEDOROV, P.V. 1978).

The Late Pleistocene of the Caspian basin begins with the late Khazarian transgression prolonged and complicated by oscillations. The area of the transgressing sea was considerably smaller than that of the preceding transgressions, which was due to the activity of differential neotectonic shifts. An ingressive bay was formed in the Mtkvari depression, where clayey and sandy-argillaceous deposits of several dozens of metres thickness accumulated. Abrasion and abrasion-aggradation marine terraces developed in the foothills, on the slopes of anticlinal morphostructures and in the Apsheron peninsula at absolute altitudes of 40-45, 60 and 75-80 m that can be assigned to three climatic microcycles within the late Khazarian mesocycle (ALI-ZADE, S.A. et al. 1978).

In the first half of the late Khazarian moderately warm climatic conditions prevailed. In the second half warming took place, causing the slow increase of evaporation of the Caspian water masses. As a result, the salinity of the sea water increased from 14‰ to 17‰. The Binagadian (Apsheron peninsula) fossil fauna and flora, correlated with the Mikulino Interglacial, indicate the warming and partial aridization at the end of the late Khazarian.

Subsequently, the Athelian regression followed, when a vast area of the Caspian Sea bottom emerged (RYCHAGOV, G.I. 1977). Its coastal line was by some 20-25 m below the present one. Intense erosion developed in the foothills and valley incision took place. At the same time, the accumulation of subaerial deposits increased in the coastal zone. During the Athelian regression the climate was cool (FEDOROV, P.V. 1978), which could be due to the early Bezingian (Lazinian) glaciation.

At the final stage of the Late Pleistocene the vast Khvalynian (early Khvalynian) transgression proceeded in the Caspian Sea basin with a water level up to absolute height of 45-50 m, i.e. exceeding by 100 m that of the preceding regression. The early Khvalynian sea penetrated far into the synclinal depressions and intermountain areas, reaching the Mingechaurian meridian in the Mtkvari river depression. The marine facies are represented by the alternation of sands and clays, less frequently with intercalations of shingle, coquina and volcanic ash. The existence of three marine terraces at absolute heights of 10-15 m, 20-25 m and 40-45 m is the indication of several sea levels and accordingly of the oscillations of climate. The lower Khvalynian fauna complex points to desalting of the basin (FEDOROV, P.V. 1978). Most probably the maximum early Khvalynian transgression was confined to the interstadial within the Bezingian glaciation. In most of the cases the available radiological data (TL dates in the range of 37-70 thousand years) confirm to above said fact.

The early Khvalynian transgression was replaced by a considerable regression (Enotaevian) with the level up to -45 m. The activation of erosional and denudation processes, the accumulation of terrigeneous material in the coastal zone, the erosion dissection of the lower Khvalynian deposits followed (RYCHAGOV, G.I. 1977). The break between the lower and upper Khvalynian deposits is also confirmed by the existence of a sharp abrasion bench, separating the lower and upper Khvalynian terraces (FEDOROV, P.V. 1978).

The short-term Enotevian regression was succeeded by the late Khvalynian transgression, in the maximum of which the level was located at the mark of 0 m. It was disturbed by oscillations, evidenced by the levels of marine terraces at minus marks 2, 11-12 and 16-17 m. The gradual decrease of basin area and the increase of salinity were caused by the more arid climate.

TL data are within the range of 11-22 thousand years, which enables to correlate the late Khvalynian transgression with the Novoeuxinian III climatic microcycle in the Black Sea basin and with the late Bezingian (Shahdjuzinian) glaciation in the Greater Caucasus.

ARCHAEOLOGICAL EVIDENCE OF PALEOCLIMATES

The changes of natural conditions, associated with considerable climatic oscillations, are fixed by archaeological evidence: multi-layer sites in mountain caves of South Ossetia, e.g. Kudaro I, Kudaro III, Tsona, Kvedi etc. (LYUBIN, V.P. 1950, 1974; KALANDADZE, A.N. 1965; TSERETELI, D.V. 1970).

Early Acheulian and the Late Mousterian industries, as well as the Neolithic ones are found in the Kudaro I and Kudaro III caves. The Late Acheulian, Early Mousterian and Upper Paleolithic are missing in the cave deposits. Similar picture can also be observed in the Tsona cave at the altitude of 2100-2150 m and in other mountain caves. The location of the Early Acheulian and Late Mousterian cultures in mountain caves of the Caucasus was determined by favourable climatic conditions of that period, while the absence of traces of the Late Acheulian, Early Mousterian and Upper Paleolithic cultures could be explained by considerably cooling, which forced man to abandon the mountain regions.

The cultures being not represented in cold epochs in high mountains continued to develop in the foothills and intermountain depressions. Mousterian and Upper Paleolithic cultural layers of the Akhshtyr, Navalishin and Kepshin caves (GRICHUK, V.P. et al. 1970; LYUBIN, V.P. 1974) contain pollen and fauna, indicating the cold climate to prevail at that time. In these caves the Mousterian culture is most probably related to the early stage of its development. The remarkable descend of vegetation belts (in the range of 1200-1400 m in the Mzymta river basin the Akhshtyr cave), is probably confined to the maximum phase of glacier development in the Caucasus (early Bezingian glaciation), when the descend of the snow line reached 800 to 1000 m in the Western Caucasus.

The Late Mousterian is dated to 47,000-52,000 years by ^{14}C (IVANOVA, I.K. 1965), for in the Tsona and Kudaro caves to 58,000+15,000 and 65,000+10,000 years respectively by I_o/U, and in the Shanidar cave (Iraq) to 46,000-48,000 years by ^{14}C (SOLECKI, R.S. 1964). As to the Upper Paleolithic, it covers the time interval of 22,000-15,000 years.

The Late Pleistocene oscillations of climate are evidenced in lithologic characteristics of the cave deposits. The existence of coarse clastic material in the Mousterian horizon of the Kepshin cave and Kudaro III (LYUBIN, V.P. 1974), in the Upper Paleolithic horizon of Bronze cave of the Tsutskhvati multistage karst cave system (MARUASHVILI, L.I. 1975), as well as in the Akhshtyr and Navalishin caves, where the Upper Paleolithic is dated by ^{14}C to 19,500+500 years (horizon 2_2) and

22,000 years (horizon 3), should be linked with the second phase of glaciation in the Caucasus. The lithofacies change of deposits, the fauna composition and the settling pattern of Upper Paleolithic man undoubtedly points to a sharp cooling and aridization. Remnants of boreal fauna were found in the Upper Paleolithic sites of Imeretia and Abkhazia. Cold forest-steppe with rare pine stands and almost complete absence of thermophilous elements were developed in the extraglacial areas of Eastern Georgia in the interval of 20,000-14,000 years (TUMADJANOV, I.I.--GOGICHAISHVILI, L.K. 1969).

CONCLUSION

It follows from the above said facts that the Upper Paleolithic can be connected with the late Nenskrian phase of glaciation (Late Würm), and the Late Mousterian with the intra-Nenskrian interstadial. The Early Mousterian covers the epoch of the Early Nenskrian glaciation (Early Würm) and a part of the preceding interglacial (Riss-Würm) epoch.

In the Late Pleistocene the changes of natural conditions in the Caucasus were caused also by tectonic movements, which have determined the further development of the existing macro- and mesostructures - the rise of ranges and sinking of intermountain areas (Sub-Khvalynian neotectonic cycle); the bottom of the Black and Caspian Sea was involved in tectonic shifts, which together with glaciation and climatic oscillations played an important role in the development of regressive phases in the Ponto-Caspian basin and in the deformation of marine terraces.

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The majority of the present papers were delivered as lectures at a seminar organized by the INQUA Hungarian National Committee and Section X. (Geo- and Mining Sciences) of the Hungarian Academy of Sciences in Budapest, October 1984. Since the early 1980s the achievements of loess research in China have been internationally recognized.

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The intimate link between periglacial geomorphology and Quaternary studies aimed at paleogeographical reconstruction is well illustrated by the problems presented by loess and loessic deposits. As a consequence, both the INQUA Commission on Loess and the IGU Commission on the Significance of Periglacial Phenomena welcomed the opportunity of sponsoring a joint field meeting in Normandy, Jersey and Britany in August, 1986. The objectives were to examine loess from the points of view of stratigraphy and sedimentology, to specify research methods for identifying loess, periglacial deposits and their characteristics, as well as to assess the paleogeographic implications of their occurrence.

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