QUATERNARY ENVIRONMENT IN HUNGARY

AKADÉMIAI KIADÓ · BUDAPEST

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(Studies in Geography in Hungary 26)

Edited by
MÁRTON PÉCSI and FERENC SCHWEITZER

In past decades a detailed subdivision of the Quaternary was carried out with special emphasis on climatic or environmental phases with the Last Glacial cycle of the Pleistocene. Most recently studies on global and regional scale of the Late Quaternary ecological changes have come to the fore.

This collection of papers gives an overview of the long- and short-term terrestrial records of the Middle Danube Basin (Hungary), of paleogeographical, environmental or climatic changes since the Last Interglacial, of the cycles of solar climatic types since the Riss Glacial, of the Upper Pleistocene events and of the vegetation history of the Great Hungarian Plain. Relying on complex sedimentological and radiometric investigations the Holocene evolution of the Lake Balaton is described, radiometric data are given on the Holocene deposition of the Danube, a mineralogical study for dating the thin tephra layer in Hungarian loess profiles is presented, as well as mass movements on steep loess slopes occurring on agricultural land are analysed.

The contributors of the volume meant to provide information on the recent results of their investigations for the participants of the XIIIth Congress of the International Union of Quaternary Research in Beijing.

AKADÉMIAI KIADÓ, BUDAPEST

STUDIES IN GEOGRAPHY IN HUNGARY, 26

Geographical Research Institute
Hungarian Academy of Sciences, Budapest

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Edited by
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FERENC SCHWEITZER
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PREFACE

Over the past two decades the Hungarian National Committee of INQUA has been regularly publishing collections of papers on Quaternary research for the INQUA Congresses. For the most recent XIIth Congress (Ottawa, Canada) two separate volumes on Holocene and Pleistocene environmental changes in Hungary were issued. For the XIIIth Congress in Beijing papers providing information on part of our recent results a collection entitled 'Quaternary environment in Hungary' is meant to appear.

In past decades attempts have been made at detailed subdivision of the Quaternary with special emphasis on the duration of climatic or environmental phases within the Last Glacial cycle of the Pleistocene. Most recently, the investigations of global and regional scale Late Quaternary ecological changes have come to the fore in INQUA research projects.

In this collection of papers mostly studies related to these topics appear. As an introductory paper a comprehensive overview is given of the long- and short-term terrestrial records of the Middle Danube Basin (Hungary) on the basis of analyses of subaerial profiles and deep cores. Part of the analysed profiles embrace the whole Quaternary while others even date back to the Neogene (Pécsi, M. and Schweitzer, F.). Most of the papers deal with paleogeographic, environmental or climatic changes since the Last Interglacial. N. Bariss reinterprets the cycles of solar climatic types since the Riss Glacial, based on data by Milankovitch and Bacsák. As mirrored by palynological studies, the Upper Pleistocene vegetation history for the Great Hungarian Plain is summarised by M. Járai-Komlódi. In another paper Upper Pleistocene climatic changes are described using malaco-thermometric and geochemical methods (Szöőr, Gy., Sümegi, P. and Hertelendi, E.). Also on the basis of sedimentological and geochemical analyses of young loess-paleosol sequences buried meadow and alkali soils are dated (Szöőr, Gy., Sümegi, P. and Balázs, É.). The 12, 000 years of Holocene evolution of Lake Balaton is described in more detail relying on complex sedimentological and radiometric investigations (Cserny, T., Nagy-Bodor, E. and Hajós, M.). Radiometric data are given on the Holocene deposition of the Danube by E. Hertelendi et al. A comparative mineralogical study for dating the thin tephra layer in Hungarian loess profiles is attempted by a team of Hungarian and Belgian researchers (Gábris, Gy., Horváth, E. and Juvigné, E.). Finally, mass movements on steep loess slopes occurring on agricultural land in winter are analysed (Pinczés, Z.).

The publication of the present volume of studies was supported by the Hungarian National Scientific Research Fund (OTKA).

Acknowledgements are due to all contributors (authors, translators, editors) and to the technical staff who promoted the publication of this volume.

The contributors wish success to the organizers and participants of the XIIIth INQUA Congress.

Budapest, June 1991

Márton PÉCSI and Ferenc SCHWEITZER
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SHORT- AND LONG-TERM TERRESTRIAL RECORDS OF THE MIDDLE DANUBIAN BASIN

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ABSTRACT

In Central Eastern Europe, in the Middle Danube Basin, loess and loess-like deposits cover various morphotectonic levels in ca. 150,000 km^2 total area. Basin types of various elevation and size are predominant.

Under different geomorphological or morphotectonic conditions — *over an identical time interval* — the variation in the rate of basin subsidence produces various litho- and chronostratigraphical sequences.

The subaerial sequence of basins affected by prolonged subsidence in the Quaternary is subdivided by paleosols of larger number than the loess-paleosol sequence of basins, elevated plateaus or watersheds which underwent more moderate subsidence during the Quaternary.

1. *Young loess mantles* of small intermountain basins and mountain slopes (at 150-400 m above sea level). In the mantle of slope loess maximum 3 or 4 loess/paleosols and 2 or 3 slope debris layers overlie one another. Their age is < 125,000 years B.P.

2. *Low-lying terraces and alluvial fans*, flood-plain loess. The thickness of the loess series of the subsided alluvial fan is 40-50 m; it is of Pleistocene age and partly eroded. Flood-plain loesses are 2-5 m thick, occasionally subdivided by one 1 or 2 paleosols. Their age is < 25,000 years B.P.

3. *Loess-paleosol-sand sequences on pediments* (100-150 m above sea level). In the loess-paleosol-sand sequences of 50-100 m thickness ca. 12 loess, 10 sand and silt layers and 20-30 paleosols are present. The age of these profiles with hiatuses is Pliocene-Pleistocene.

4. *Loess-paleosol-sand sequences on alluvial fans and terraces* (10-100 m relative heights). The loess sequence is 40-60 m thick and subdivided by 10-12 loess horizons, 10-12 paleosols and 5 or 6 sand layers. LTR — Pliocene-Pleistocene.

5. *Subaerial basin sediment* locally of 500-1,500 m thickness. The surface of the basin is a flood-plain at 90-100 m above sea level. In the basin sediments the number of paleosols — mostly meadow soils — may reach one hundred. In several boreholes there are 6-12 red soils or red clays between 600 m and 1,000 m. 12-16 intercalated sand layers are observed. The age of this almost complete sequence is 5.2-5.4 million years B.P.

In the most intensively subsiding basins sedimentation was almost continuous; during the Pliocene ca. 50-60 and in the Pleistocene ca. 50 soils developed.

In contrast on the non-subsiding foothills only 20-30 soils formed during the Pliocene and Pleistocene and several gaps are detected in the sequence.

INTRODUCTION

Recently, an increasing number of papers have been published on the reconstruction of paleogeographic and climatic changes over the entire Pleistocene or part of the Neogene. In several respects our days are considered the age of 'records' and thus it is no wonder that Quaternary geochronologists are also seeking to find 'long-term records' concerning sedimentary rocks. Among other achievements, the successful application of ^{18}O isotope stratigraphy of deep-sea sediments and magnetostratigraphical analyses opened up new vistas in this direction.

Among subaerial deposits the cyclical alternation of loess, loess-like deposits and paleosols also allowed good opportunities to set up 'long-term records' for the Quaternary (or perhaps extending them into the Neogene) and to make correlations with deep-sea oxygen isotope stages.

Quasi-complete loess-paleosol sequences are kown from the profiles of the *Tajik Basin, Central Asia* (Dodonov, 1982, 1984, 1987; Lazarenko, 1984; Ranov, 1980), as well as from the *Luochuan and Xifeng profiles* on the *Loess Plateau of China* studied by many (Liu, 1985, 1987; Kukla and An, 1989; Liu and Yuan, 1987; Heller et al., 1987; Sasajima and Wang, 1984).

1. On the margin of the loess-lined *Tajik Basin* the foothill slopes of 1,500-1,700 m height are dissected into interfluvial ridges. There is much resemblance between profiles, but the loess-paleosol sequences are not totally identical. The most dissected is the 180 m profile at Chasmanigar *(Fig. 1, column 21)*.

In the upper 80 m of the sequence 10 well-developed paleosols are separated by 3-10 m thick loess layers. The Brunhes/Matuyama boundary was found in the loess between the paleosols Nos. IX and X. Below paleosol No. X a very marked erosional hiatus was observed (Dodonov and Penkov, 1977; Lazarenko, 1984). Even below that there is a ca. 90 m thick subaerial series with 27 further paleosols indicated by the numbers X to XX. The layers between paleosols — mostly carbonate accumulation horizons — are very much different from true loess. For their compactness and concretion content, they are called locally 'stony loess'. According to Dodonov, the Olduvai paleomagnetic event (1.8 million years B.P.) probably occurs in the upper part of soil complex No. XX.

Below soil complex No. XX red clays and detritic loess-like deposits are found overlying Neogene conglomerate.

Consequently, the Tajik loess-paleosol sequence embraces ca. 2 million years. It has to be noted that this 'stony loess' series is not true loess.

2. In the various partial basins of the *Loess Plateau of China* of more than 1,000 m altitude the thickness of loess series and the number of paleosols show great variation (Sasajima and Wang, 1984).

The Xifeng profile differs from the Luochuan one, located 160 km to the east. Nevertheless, the sequences are generally regarded identical *(Fig. 1, column 23)*. For both type profiles in the major upper part 14 paleosols and 15 loess horizons are identified.

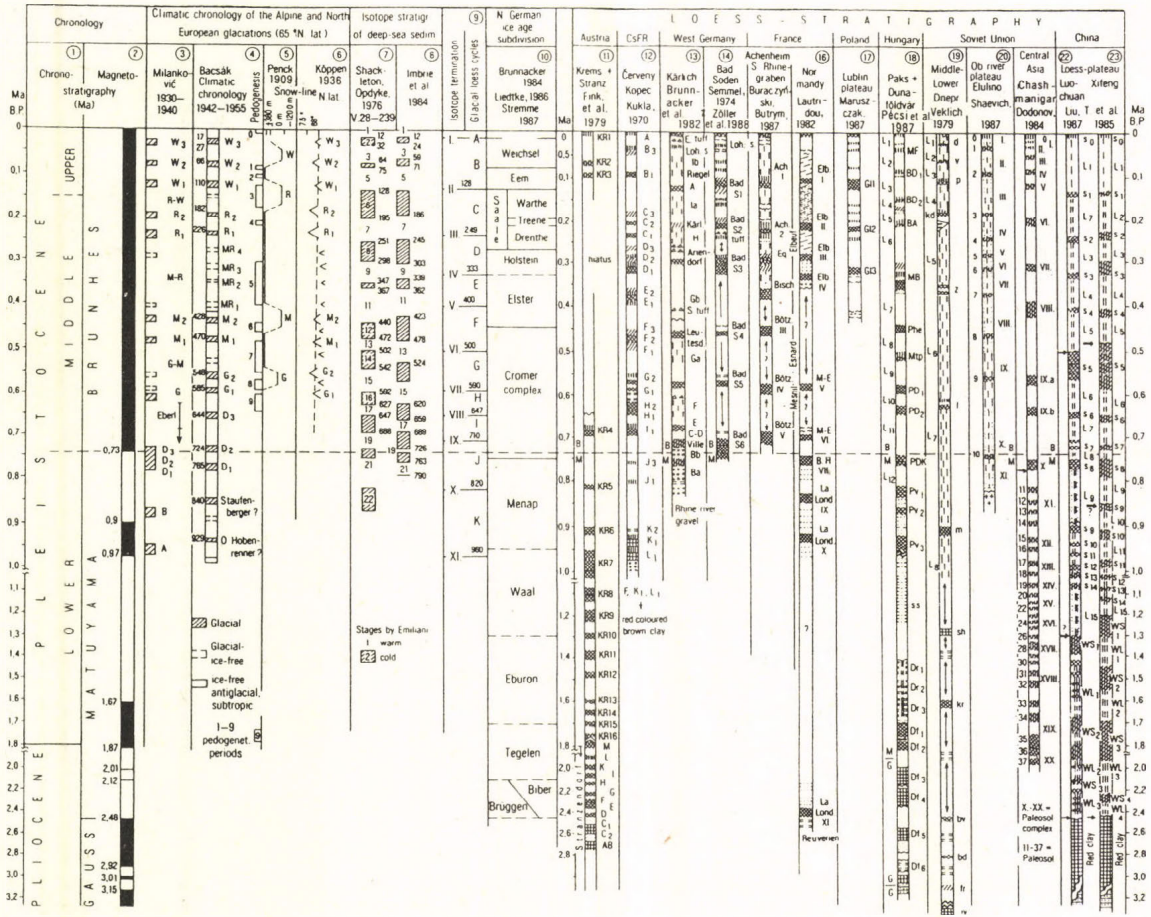


Fig. 1. Provisional correlation between Pleistocene climatic chronology, isotope stages and loess stratigraphy (compiled by M. Pécsi).

The B/M paleomagnetic boundary is placed below the paleosol S_7 in the Xifeng profile and in loess L_8 in the Luochuan profile.

The subaerial sequence cannot easily be subdivided into loess and paleosol layers below sandy loess L_{15}, but clusters of layers with weathering or soil formation of various intensity alternate. This is the so-called *Wucheng loess*, which includes 4 more pedified clusters of layers (W_{s1}-W_{s4}) and 4 clusters of loessy-loamy layers (W_{L1}-W_{L4}) in the Xifeng profile. In the Luochuan profile these divisions are only repeated three times. The paleomagnetic analyses, however, support the interpretation that in both profiles the red clays in the base of the Wucheng series belong to the Gauss epoch. The Olduvai event within the Matuyama epoch (1.67-1.87 million years B.P.) was found in the paleosol group W_{s2} in the Luochuan profile and in W_{s3} in the Xifeng profile.

In the above two key sections of the Loess Plateau of China (at Luochuan and Xifeng) 24 and 28 loess horizons and an equal number of paleosols are counted, resp. In a recently analysed profile (at Baoji) the alternation of 32 paleosols and loess horizons have been recorded (Ding et al., 1991).

From the data of paleomagnetic investigations, calculation of sedimentation rate, analysis of magnetic susceptibility curves and comparisons with oxygen isotope stages in deep-sea cores, it was established that by the evidence provided by the profiles of the Loess Plateau of China all the glacial and interglacial periods during the Pleistocene can be identified in the subaerial sequences (Liu, 1987; Liu and Yuan, 1987; Kukla and An, 1989).

The above was described in the introduction as it is doubtful whether on plateaus lying at more than 1,000 m altitude no major sedimentation gap or the removal of some layers have occurred in loose deposits over the past 2.5 million years (Pécsi, 1987). On elevated terrain with drainage network it is difficult to envisage perfect sediment traps, moreover it is known that even in deep-sea sequences hiatuses are common. In addition, the subaerial sequence of the mentioned high-lying basins are confronted with that of the Middle Danube Basin, which is locally thicker than 1,000 m and the base of the sequence lies at more than 1,100 m below sea level.

DISCUSSION

It is undoubtedly the loess-paleosol sequences that provide the best opportunities of all subaerial deposits for the reconstruction of Late Cainozoic paleogeographic-paleoclimatic changes.

The question arises: How many periods suitable for loess formation and how many favouring soil formations occurred during the Quaternary? Our experiences show that the number depended upon climatic patterns in the 'loess superzone'.

Another question which could be put is how many paleosols or loess horizons have

been preserved of all formed in a particular site? It was found that, for instance, in the Middle Danube Basin once various partial basins of different size were superimposed on one another and the rate of subsidence varied in them and resulted in subaerial sequences complete or incomplete to various degrees over identical time intervals.[1]

1. In the *small basins and foothill slopes of uplands* (at relative heights of 150-400 m) usually young loess occurs. In the mantle-like slope loess there are maximum 3 or 4 loess and paleosol layers superimposed and on the side closer to mountains 2 or 3 loess horizons with slope debris are also observed.

2. In *mountain forelands, over broad basin margins* (at 100-150 m relative heights) loess-paleosol-sand sequences of great thickness (50-100 m) occur, occasionally with 10-12 loess, 10 sand and 20-30 paleosol intercalations *(Figs. 2-5)*.

The sequence shown in Figure 3b is on a pediment enclosed by 2 low horsts, overlying Upper Miocene (Pannonian) sand. The sequence has been described in detail in previous publications (Pécsi et al., 1987), here attention is only drawn to some of the

Fig. 2. Hungarian profiles discussed in the paper

[1]Out of the possible morphotectonic units only some are treated from the viewpoint of the evolution and occurrence of the loess-paleosol sequence.

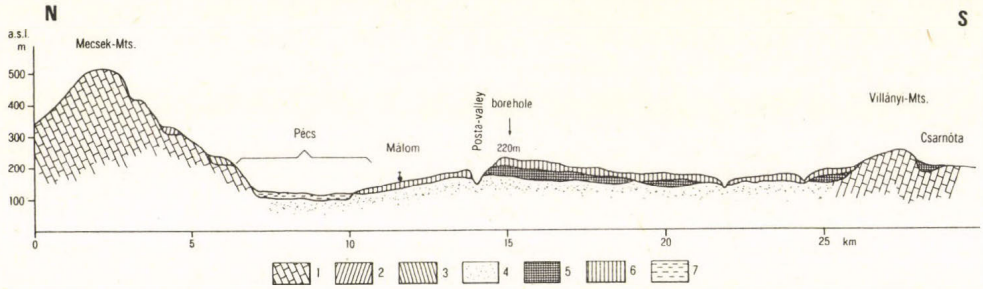


Fig. 3a. Geomorphological and geological situation of Posta valley borehole at Pécs (M. Pécsi and F. Schweitzer, 1987). 1 = Mesozoic limestone, marl, sandstone; 2 = Upper Miocene marine terrace with Sarmatian limestone; Upper Miocene marine terrace (Upper Pannonian); 4 = Upper Miocene (Pannonian) sandy formation; 5 = Pliocene reddish paleosols, red clay formation; 6 = Pleistocene loess and paleosol sequence; 7 = Upper Pleistocene-Holocene alluvial sequence

characteristics. In the ca. 60 m deep profile the upper 7 of the total 29 paleosols are dark coloured steppe soils, paleosols Nos. 7-19 are predominantly forest soils, while Nos. 20-27 and 29 are reddish clayey soils, with several subdividing sandy layers. Older red clay layers can be correlated with the classic red clays in the side of the Villány Mountains, characterized by the Pliocene Csarnotian fauna (Kretzoi and Pécsi, 1982). Here the loess-paleosol sequence — with many hiatuses (indicated by dark and light arrows) — represents the Pleistocene period.

In a basically similar morphotectonic position on elevated Upper Miocene (Pannonian) strata the loess profiles shown in Figures 4 and 5 are situated.

A characteristic feature of the subaerial sequence of the Gyöngyösvisonta profile (Fig. 6) is that it overlies an eroded pediment of Pannonian lignite-bearing sediments. Pliocene red clay is only preserved in patches. The subaerial sequence here represents the Pliocene together with the Lower and Middle Pleistocene. The Upper Pleistocene series is very incomplete (Kretzoi et al., 1982; Pécsi, 1985; Pécsi et al., 1985).

3. The loess-paleosol-sand sequences *on alluvial fans, alluvial fan terraces* (10-100 m relative height above the flood-plains) are deposited on Upper Miocene (Pannonian) surfaces, basin margins unaffected by subsidence in the Quaternary. Such a (40-60 m thick) loess series outcrops in the bluffs along the Danube at Dunaújváros, Dunaföldvár and Paks and subdivided by 10-14 loess horizons, an equal number of paleosols and 4-6 sand layers *(Fig. 7)*. In this sequence the B/M paleomagnetic boundary was identified below the 8th paleosol (PD_2) (Pécsi and Pevzner, 1974; Pécsi, 1987; Márton, 1979). The loess-paleosol sequence here is not older than the Jaramillo event (0.9 million years B.P.). However, below the 50-60 m loess series ca. 30-40 m thick 'variegated clays' and sandy silts follow. The lowermost variegated clay and red clay layers of this 'Dunaföldvár

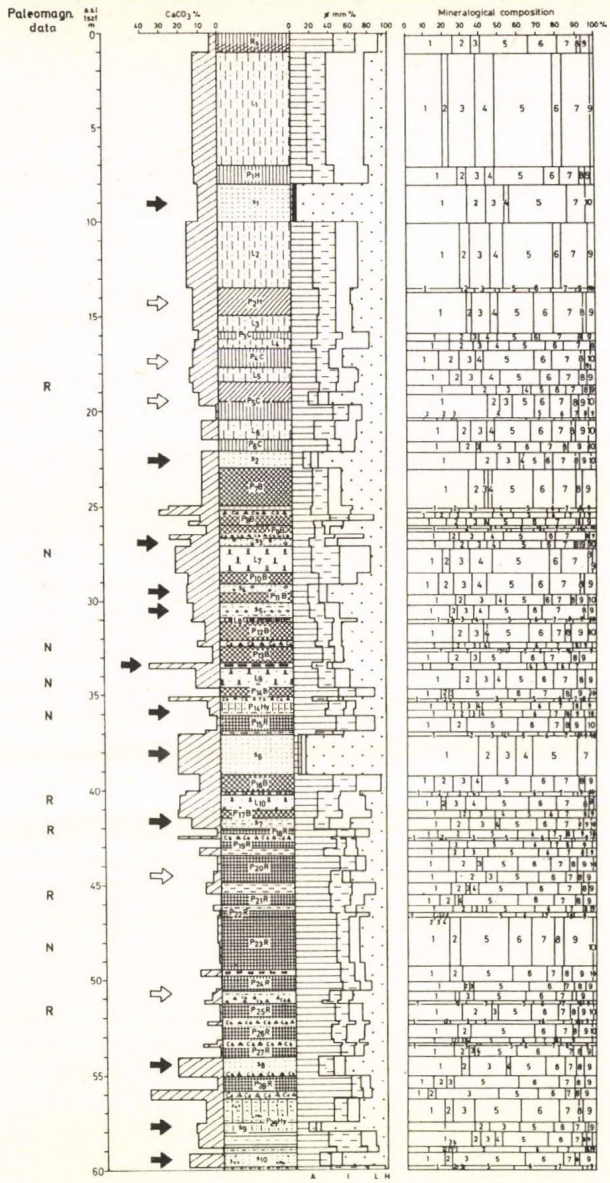


Fig. 3b. Loess-paleosol sequence of Posta valley at Pécs. (Lithological, paleopedological and mineralogical analyses made by M. Pécsi, Gy. Scheuer, F. Schweitzer, L. Gerei and M. Reményi; paleomagnetical data by P. Márton). R = reverse; N = normal polarity; L_1-L_6 = young loess; L_7-L_{10} = old loess; S_1-S_{10} = sandy layers; P_1H, P_2H = humic loess, embryonal paleosols; P_3C-P_6C = chernozem-like forest-steppe paleosols; P_8B-$P_{14}B$ = brown forest paleosols; $P_{15}R$-$P_{29}R$ = ochre-red paleosols, red clays; $P_{14}Hy$, $P_{29}Hy$ = hydromorphic meadow soils; A = clay (2-10 micron); I = fine silt (10-20 micron); L = loess (20-50 micron); H = sand (50-500 micron); 1 = quartz; 2 = felspars; 3 = calcite; 4 = dolomite; 5 = micas + hydromicas; 6 = montmorillonite; 7 = chlorite; 8 = kaolinite; 9 = interstratified minerals; 10 = Al and Fe hydroxides; ➜= significant unconformity; ⇒ = unconformity

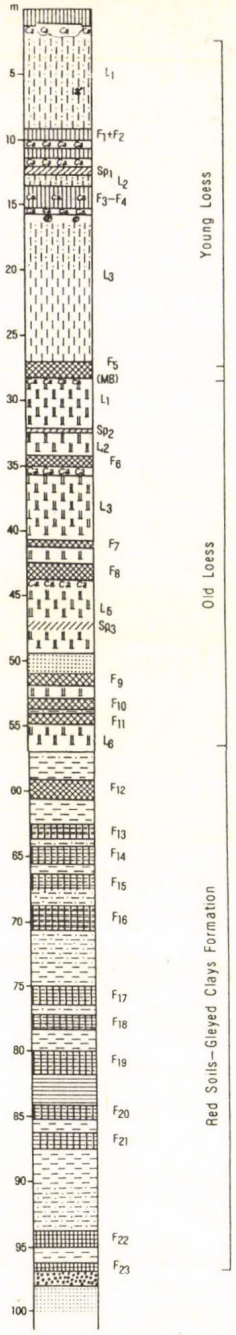


Fig. 4. Loess-paleosol and other subaerial sequences at Dunaszekcső (after F. Schweitzer and Gy. Scheuer; legend see Fig. 7)

Formation' (Pécsi, 1982) belong to the Pliocene (Pécsi, 1985). In its base also Upper Miocene (Pannonian) sediment is found. In our opinion, the 'Dunaföldvár Formation' can correlate with the 'stony loess' of Central Asia and the Wucheng Formation of the Chinese Loess Plateau and the underlying red clays (Pécsi, 1987).

Hiatuses are less frequent in the loess-paleosol sequence — with 4-6 intercalated sands — on old alluvial fans over basin margins not affected by subsidence in the Quaternary than in the mountain foreland profiles. However, the number of paleosols is similar down to the B/M boundary to the profiles of Central Asia and the Loess Plateau of China.

4. An entirely different subaerial sequence is recorded in the *parts of the basin* which underwent *gradual subsidence during the Pliocene and the Pleistocene*. The thickness of basin deposits in the Hungarian Plain locally exceeds 500-1,500 m below the flood-plains of rivers. In the most intensively subsiding basins *(Fig. 8)* red soils and red clays are repeated 6-10 times between 600 and 1,000 m depths. With the preponderant flood-plain, meadow and chernozem soils, the swamp forests also allowed the formation of lignite. The above two boreholes did not reach the sequence of the Upper Miocene (Pannonian) inland sea to 1,200 m depth. The paleomagnetic study of the cores and paleontological data cover the whole of the Quaternary and extend to most of the Pleistocene (5.25 million years B.P.; Cooke et al., 1979; Rónai, 1977, 1985b). As shown in Figures 9 and 10 in the Pliocene sequence 50-60 and in the Pliocene ca. 50 paleosol horizons are detected. Such a long Late Cainozoic geological record of subaerial sediment is only known to date from the exploration boreholes in the Hungarian Plain by Rónai (Rónai, 1985a).

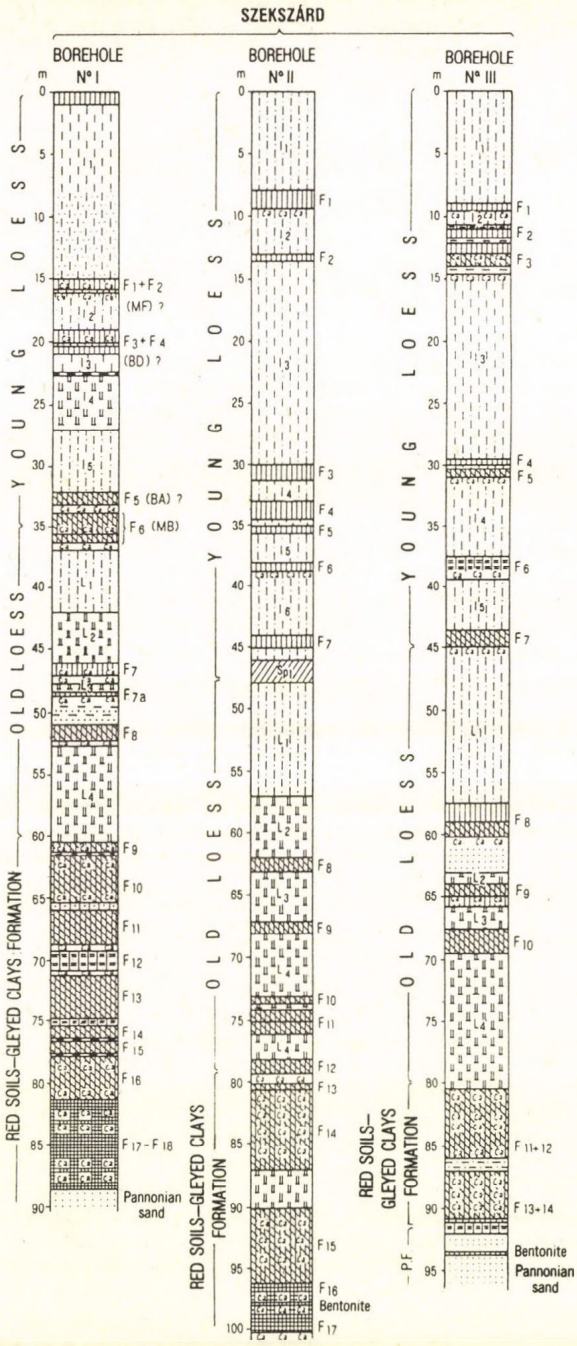


Fig. 5. Loess-paleosol sequences in the foothill zone at Szekszárd (after F. Schweitzer, Gy. Scheuer and M. Pécsi; legend see Fig. 7)

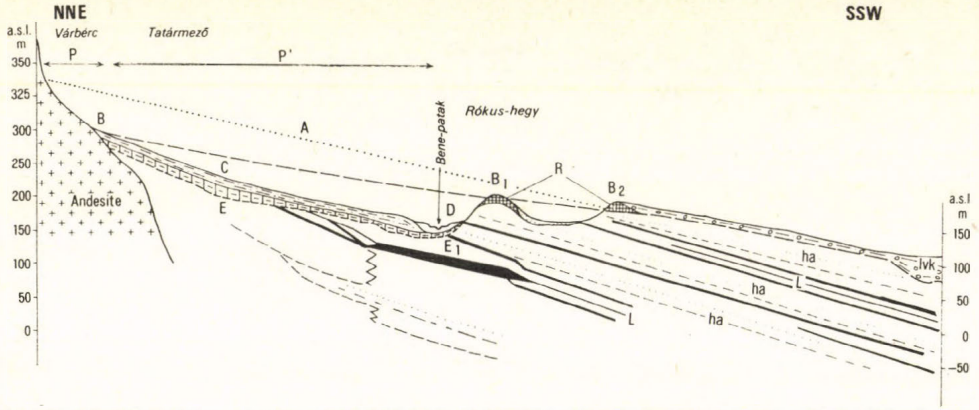
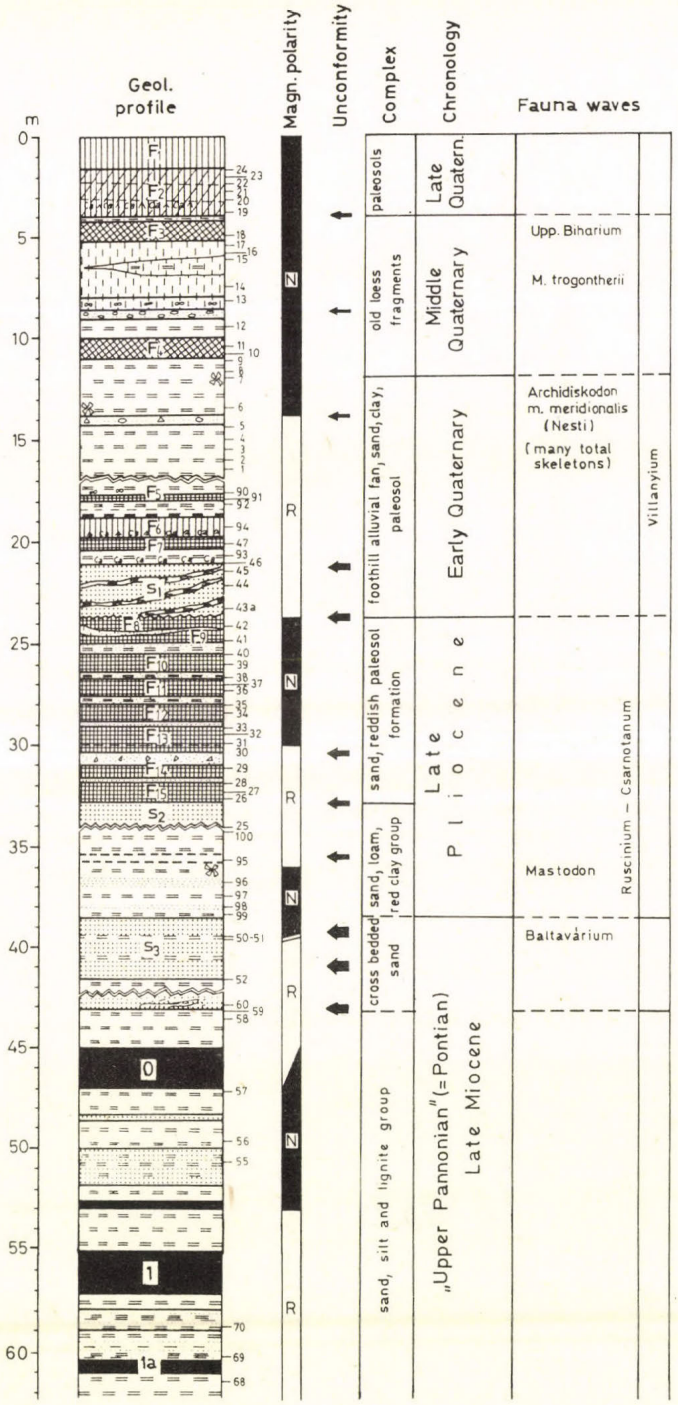


Fig. 6a. Position of the red clays at the Mátra Foothills (Abasár). A = reconstructed surface of the Lower Pannonian layers; B, B_1, B_2 = Pliocene pediment surface; C-D = Upper Pliocene-Lower Pleistocene alluvial fan; E-E_1 = strongly eroded Pannonian surface; L = lignite; h = sand; ha = sand, clay; P = pediment; P'= pediment-glacis; l,v = loess, loam; k = gravel; R = red clay; T = Upper Pleistocene terraces

Fig. 6b. Profile of the open cast lignite mine at the foot of the Mátra Mountains. (The profile was surveyed and identified by J. Balogh, P. Márton, F. Schweitzer and Gy. Szokolai under the guidance of M. Pécsi, 1982). The correlative sediments for pedimentation are the reddish purple fossil soils (F_7 to F_{15}) and clayey rock detritus between fossil soil F_6 and sand layer H_3. Profile is shown to 60 m, the lower part — 60-95 m — is only included in the description). 0-1.6 m = black meadow soil; 1.6-4.5 m = old loess with remnants of B/BC soil horizon; 4.5-5.2 m = brown forest soil; 5.2-8.1 m = loess-like material, with yellow limy sandy intercalation at the base; 8.1-8.7 m = yellow limy sand with tuff detritus; 8.7-9.2 m = sand with andesite gravel and with *Equus (Allohippus) süssenbornensis*; 9.2-9.9 m = flood plain clay soil; 9.9-11.0 m = dark-grey purplish flood plain clay soil; 11.0-12.0 m = $CaCO_3$ accumulation horizon; 12.0-13.8 m = grey clay with $CaCO_3$ concretions and tuff detritus (*Mammuthus trogontherii* and *Bison sp.* finds); 13.8-14.3 m = sand with andesite gravel (*Archidiskodon meridionalis* find); 14.3-15.3 m = clayey sand with tuff detritus; 15.3-16.7 m = grey clay soil of flood plain; 16.7-17.7 m = greyish-brown clay with tuff detritus; 18.0-18.8 m = greyish-brown clay with tuff detritus; 18.8-20.4 m = (reddish) brown clay soil; 20.4-21.1 m = limy, sandy old loess; 21.1-23.7 m = clayey sand (from 23.4 sand with tuff detritus); 23.7-25.0 m = purplish aggregated clay (with yellowish-brown sandy tuffeous-detrital wedging); 25.0-25.5 m = yellowish-brownish tuff detritus; 25.5-26.4 m = purple clay soil; 26.4-26.7 m = clayey sand with tuff detritus; 26.7-27.6 m = greyish-purplish clay soil; 27.6-28.0 m = clayey sand with tuff detritus; 28.0-28.9 m = purplish clay with tuff detritus at the base; 28.9-29.1 m = sand with tuff detritus; 29.1-30.4 m = purplish clay (from 29.9 m greyish-purplish clay); 30.4-31.0 m = yellowish-brown coarse sand with tuff detritus; 31.0-31.6 m = purplish clay; 31.6-31.8 m = clay with tuff detritus; 31.8-32.3 m = purple clay with tuff detritus; 32.3-32.7 m = purple clay; 32.7-34.0 m = yellowish-brown sand with tuff detritus; 34.0-35.8 m = sandy clay with yellowish-grey purplish sand of tuff detritus and with *Zygolophodon* find; 35.8-36.6 m = crumbled clay with yellowish-grey ferrous precipitations of purplish shade; 36.6-37.1 m = coarse-grained sand with tuff detritus; 37.1-37.9 m = ferrous sandy clay; 37.9-38.5 m = sandy clay with tuff detritus with purplish tuff detritus at the base; 38.5-41.5 m = micaceous yellow sand with thin sandy mud intercalations; 41.5-42.4 m = greyish-greenish clay; 42.4-43.1 m = ochre-yellow clayey sand; 43.1-45.0 m = grey clay; 45.0-47.0 m = lignite; 47.0-48.3 m = grey clay; 48.3-48.6 m = yellowish-grey sand; 48.6-50.0 m = grey clay; 50.0-51.8 m = yellowish sandy clay; 51.8-52.7 m = grey clay; 52.7-53.0 m = lignite; 53.0-55.3 m = grey clay; 55.3-57.2 m = lignite; 57.2-58.0 m = grey clay; 58.0-59.0 m = greyish-yellowish muddy clay being more clayey at the base; 59.0-60.5 m = grey muddy clay; 60.5-61.1 m = lignite; 61.1-62.4 m = micaceous muddy clay; 62.4-69.5 m = micaceous fine-sandy mud interwoven with grey clay bands, from which water infiltrates; 69.5-75.5 m = grey clay; 75.5-77.2 m = grey fine-sandy mud; 77.2-78.5 m = grey fine-sandy mud; 78.5-79.1 m = muddy yellowish-grey clay; 79.1-82.1 m = micaceous yellow fine sand; 82.1-84.2 m = grey fine-sandy mud; 84.2-89.6 m = grey micaceous fine sand with clay bands; 89.6-92.5 m = grey micaceous fine-sandy clayey mud being more compact at the base; 92.5-94.5 m = lignite


Geol. profile
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Upp. Biharium
M. trogontherii
old loess fragments
Middle Quaternary
Archidiskodon m. meridionalis (Nesti)
(many total skeletons)
foothill alluvial fan, sand, clay, paleosol
Early Quaternary
Villanyium
sand, reddish paleosol formation
Late Pliocene
Ruscinium – Csarnotanum
sand, loam, red clay group
Mastodon
cross bedded sand
Baltavárium
sand, silt and lignite group
„Upper Pannonian" (= Pontian)
Late Miocene

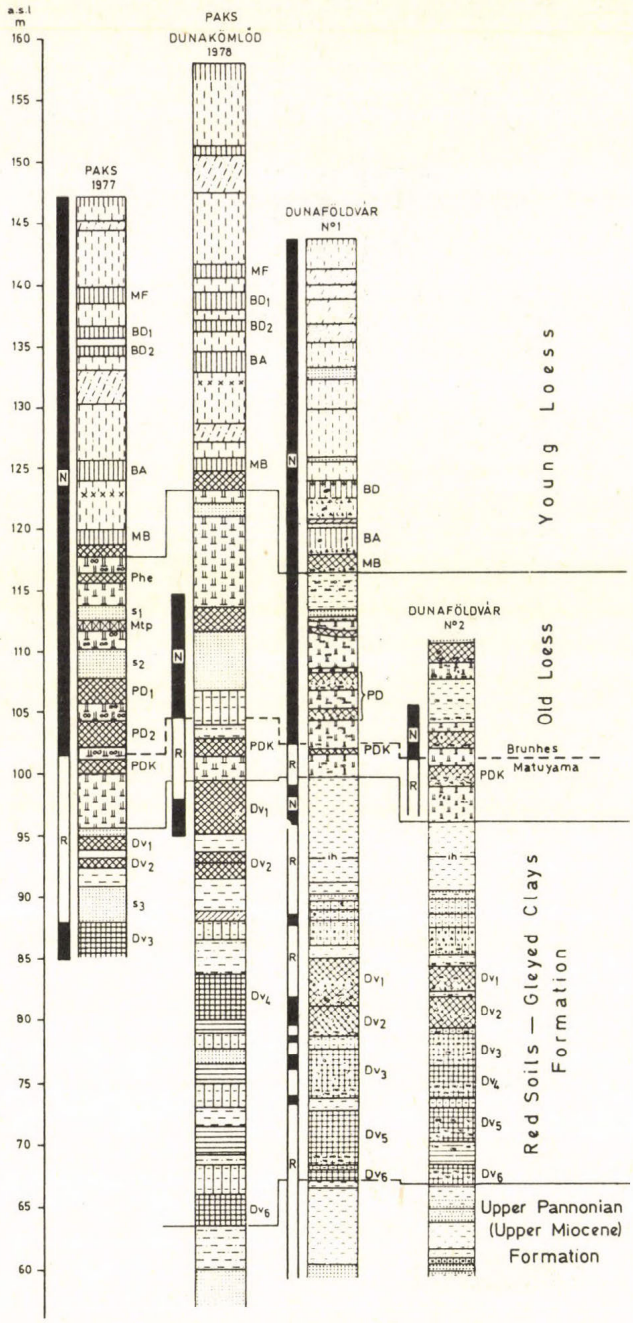


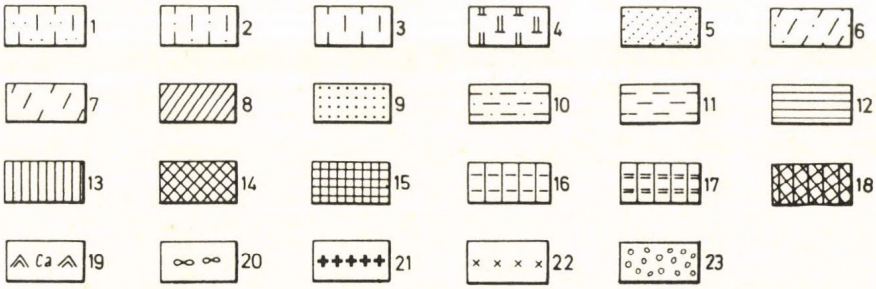
Fig. 7. Loess and paleosol sequence at Paks and Dunaföldvár (lithological, pedological analyses made by M. Pécsi, E. Szebényi and F. Schweitzer; paleomagnetical measurement made by M.A. Pevzner)

In this morphotectonic position, in the deep-subsided basin, the number of paleosols (95-115) is about three or locally fourfold exceeds that in the subaerial sequence of the non-subsiding mountain foreland or alluvial fan surfaces.

CONCLUSION

The Plio-Pleistocene terrestrial subaerial sedimentation ensuing on the Upper Miocene (Upper Pannonian) period began with red clay development on slowly subsiding basin surface and then increasing subsidence produced repeated formation of various soils and loose deposits. They are better preserved than in the sequences of the neighbouring, non-subsiding basin margins accumulated over the same period.

This approach helps to make our criticism understandable related to the interpretation of the sequence on the Loess Plateau of China as complete and full record of climatic changes over the Quaternary.



legend to *Fig. 7.*

1 = loessy sand; 2 = sandy loess; 3 = loess; 4 = old loess; 5 = slope sand; 6 = sandy slope loess; 7 = slope loess; 8 = semipedolite; 9 = fluvial-proluvial sand; 10 = silty sand; 11 = silt, gleyed silt; 12 = clay; 13 = steppe-type soil, chernozem; 14 = brown forest soil; 15 = red clay; 16 = hydromorphic soil; 17 = alluvial meadow soil; 18 = forest soil(on flood-plain); 19 = calcium carbonate accumulation; 20 = loess doll; 21 = charcoal; 22 = volcanic ash; 23 = sandy gravel; MF = "Mende Upper" forest-steppe Soil Complex (Mo. 421 29,800 years B.P., HV 27,855±599 years); BD = "Basaharc Double" forest-steppe Soil Complex; BA = "Basaharc Lower" chernozem soil; MB = "Mende Base" Soil Complex (brown forest soil + forest-steppe soil); Phe = Paks sandy forest soil; Mtp = Paks marshy soil; PD = "Paks Double" Soil Complex (brownish-red Mediterranean-type dry forest soil); PDK = Paks-Dunakömlőd brownish-red soil; Dv_1-Dv_6 = red soils (Dunaföldvár-formation); ih = silty sand; S_1-S_3 = sand

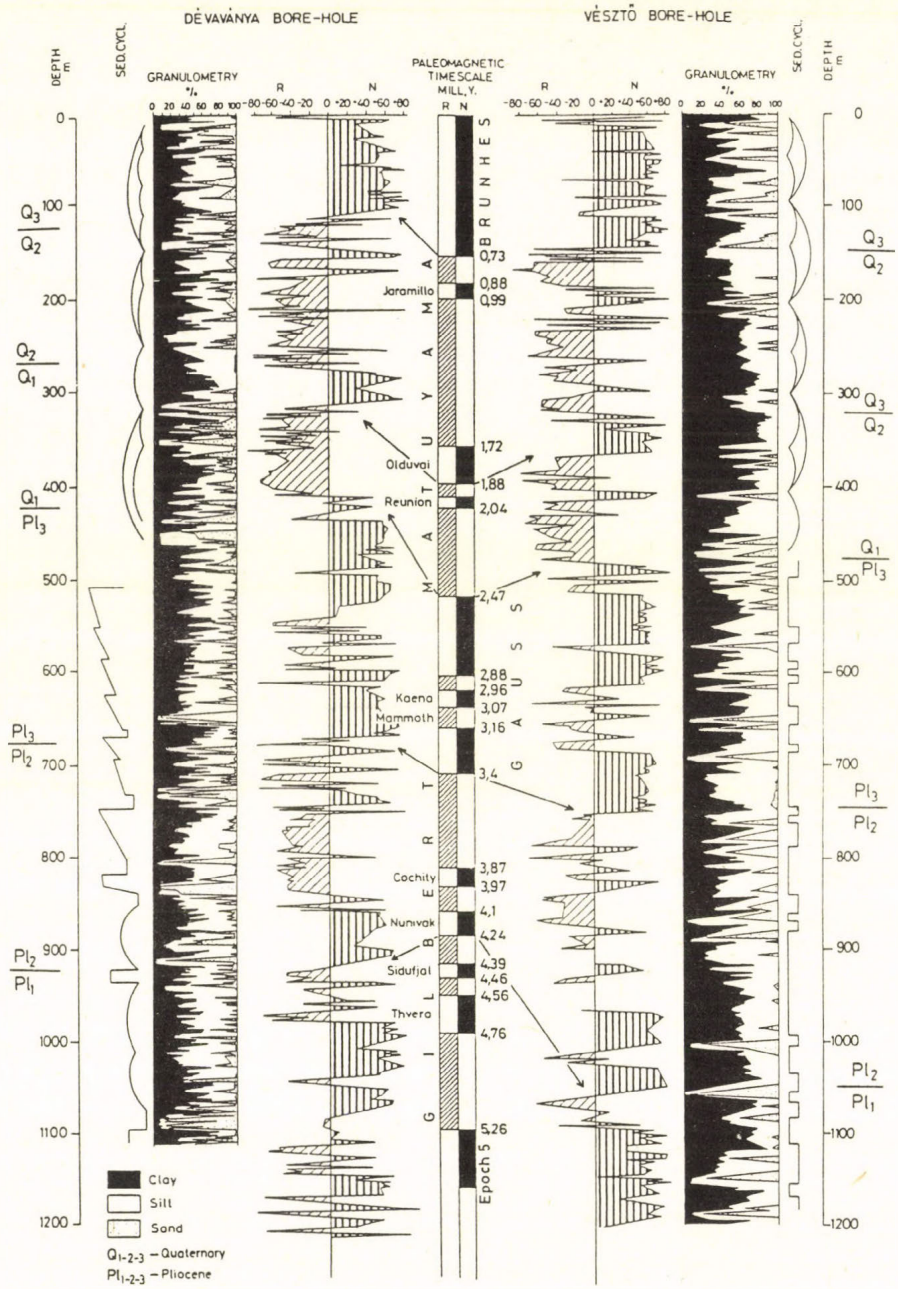


Fig. 8. Quaternary and Pliocene stratigraphy based on paleomagnetic records in Hungary (A. Rónai, 1985b); (source of paleomagnetic data: Cooke et al., 1979)

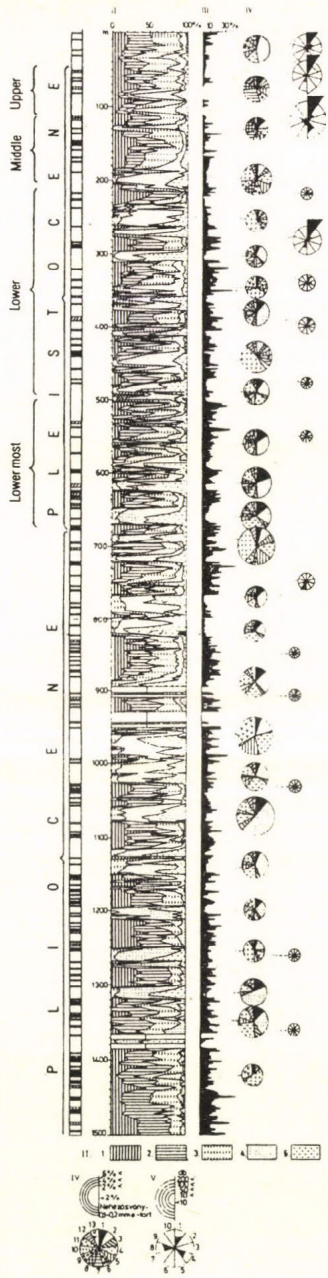


Fig. 9. Complex geological profile of the borehole of Mindszent (plotted by A. Rónai and F. Franyó). I = Paleosols in the profile: Total number of the Pleistocene series = 46; Total number of the Pliocene series = 67; for further explanation see Fig. 10

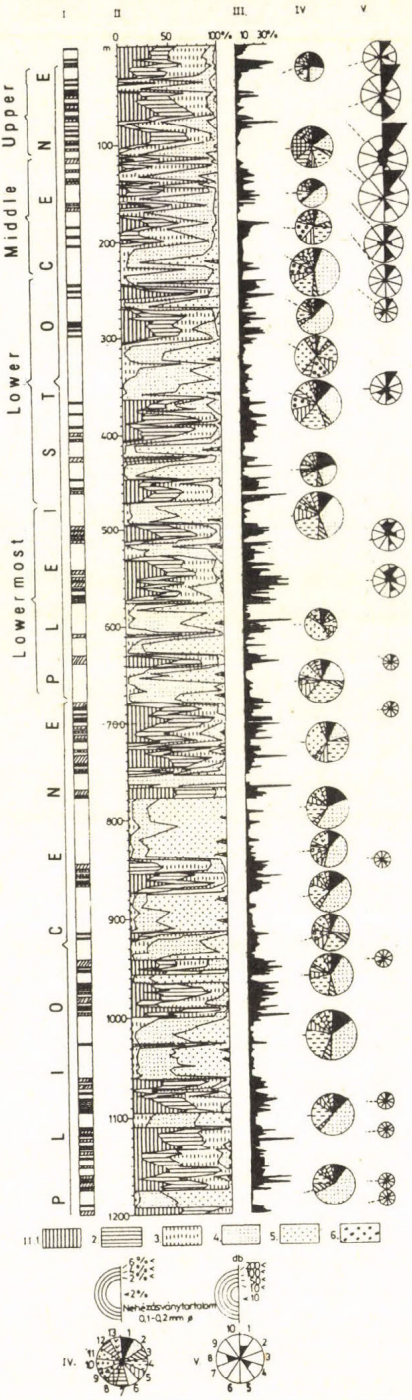


Fig. 10. Complex geological profile of the Csongrád borehole (plotted by A. Rónai and F. Franyó, A. Rónai 1985a).

I = Paleosols in the profile: Total number of the Pleistocene series = 55; Total number of the Pliocene series = 40. II = granulometry: 1 = clay; 2 = fine silt; 3 = coarse silt, sand flour; 4 = fine-grained sand; 5 = middle and coarse-grained sand; 6 = gravel. III = $CaCO_3$ content. IV = Heavy minerals: 1 = hematite, magnetite, ilmenite, leucoxene; 2 = garnet; 3 = disthene, staurolite, chloritoide; 4 = epidote, pistacite, piemontite, zoisite, clinozoisite; 5 = tremolite, actinolite, anthophyllite, glaucophane, sillimanite; 6 = green amphibolite; 7 = brown amphibole and lamprobolite; 8 = hypersthene; 9 = augite; 10 = biotite; 11 = chlorite; 12 = rutile, brookite, athanase, zircon, titanite, tourmaline, apatite; 13 = limonite, pyrite, siderite, carbonates, clay minerals. V = Ostracoda finds: 1 = *Candona parallela* G.W. MÜLLER; 2 = *Candona neglecta* G.O. SARS; 3 = *Candona rostrata* BRADY-NORM; 4 = *Candona protzi* HARTWIG; 5 = *Ilyocypris gibba* RAMDOHR; 6 = *Cyclocypris laevis* O.F. MÜLLER; 7 = *Cyclocypris huckei* TRIEBEL; 8 = *Lymnocythere inopinata* BAIRD; 9 = *Lymnocythere sanctipatricii* BRADY-ROB; 10 = *Cytherissa lacustris* G.O. SARS

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THE CHANGING CLIMATES DURING AND SINCE THE RISS/WÜRM INTERGLACIAL

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ABSTRACT

Geographers and geologists have always been interested in the reconstruction of the changing paleogeographic environment during the Pleistocene Epoch. Of particular interest in this field would be the last major glacial and interglacial cycles, that is, the time which has elapsed since the melting of the Riss glaciers. This time span is approximately 175,000 years.

In this paper an attempt is made by Bariss (1989) to retrace the paleogeographic environment along the 55° and 65° northern latitudes during the last 175,000 years by establishing the changing climate types based on the classic research of Milankovitch (1930) and Bacsák, whose original interpretations (1940, 1942) were partly modified by Bariss (1989).

Effects of the changes of the orbital elements upon the Pleistocene climates

Even before the publication of Milankovitch's classic work, Köppen and Wegener (1924) suggested that the main reason for the Pleistocene glaciations at mid latitudes was a long series of cold summers and mild but snowy winters. The cold summers prevented the complete melting of the snow from the previous winters, thereby allowing for a net surplus of snow after each summer. It was Milankovitch (1930) who pointed out that this long series of cold summers was primarily caused by the secular changes of the following orbit elements of the earth:

(1) Earth's obliquity, that is, the tilting of the earth's axis (designated by ε) with respect to the plane of ecliptic; (2) eccentricity of the earth's orbit, designated as e, and (3) length of the perihelion, Π, which determines the time when the perihelion or aphelion occurs. The specific effects of these orbital elements in the Northern Hemisphere were expressed by Milankovitch in following way:

For the summer half-year: $\varepsilon - e\sin\Pi$

For the winter half-year: $\varepsilon + e\sin\Pi$

occurred when a low value of ε coincided timewise with such a value of $e\sin\Pi$ when Π was near 90°. Because of the low annual range of temperature between the cold summers and warm winters, this climate is classified here as a *strongly oceanic* type of climate. Figure 1 demonstrates for about the last 175,000 years the changing values of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$ (Graphs 1 and 2) as well as the resulting temperature deviations of the calorific summer and winter half-years (designated as ΔU_s and ΔU_w-see Graphs 3 and 4) for the 65° and 55° northern latitudes. All of these changes and deviations were calculated with respect to their values as of 1,800 AD, which date was regarded as „present" by Milankovitch (1930).

It was the Hungarian György Bacsák who pointed out that Milankovitch recognized only one kind of interference between the maximum and minimum values of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$. The climate-type which was caused by that interference and was described above as a strongly oceanic type was named „glacial" by Bacsák because, if it lasted long enough, it could have triggered the growth of the glaciers. It is logical, however, that if one kind of interference initiates a certain type of climate, other kinds of interference will then produce other types of climates. Therefore Bacsák (1940) suggested that there are three additional possibilities for optimal interference between the courses of the curves of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$. A *strongly continental* type of climate with hot summers and very cold (and long) winters occurred when a high value of $\Delta\varepsilon$ coincided with such a value of $\Delta(e\sin\Pi)$ when the Π was near 270°. Bacsák called this kind of climate „antiglacial" because, if it lasted long enough, its hot summers could have melted the glaciers.

The strongly oceanic and continental types of climates, described above, made up the two basic climatic environments during the Pleistocene. In addition to these, there were the following two more possibilities of interference: (1) If a high value of $\Delta\varepsilon$ meets such a value of $\Delta(e\sin\Pi)$ when Π is near 90°, a *moderately oceanic* type of climate will develop with cool summers and somewhat mild winters. Bacsák called this climate „subtropical". (2) Finally, a *moderately continental* (or „subarctic", according to Bacsák) type of climate with somewhat warm summers and fairly cold winters prevailed when the $\Pi = 270°$ value coincided timewise with a low value of $\Delta\varepsilon$.

It is obvious from the above summary that essentially only two climatic oscillations alternated during the Pleistocene: an oceanic type with colder summers and warmer winters than at present (i.e., 1,800 AD) and a continental type with warmer summers and colder winters than at present. The difference between „strong" and „moderate" lies merely in the amplitudes of the curves. As indicated by Graphs 3 and 4, the degree of the deviation of the seasonal temperatures in both directions, i.e., colder or warmer, can be as much as almost ± 5 °C with respect to 1,800 AD. Graph No. 5 of Figure 1, originally proposed by Bacsák (1940, 1942) and later modified by Bariss (1989), shows the chronological distribution of the strongly or moderately oceanic or continental types of climates, plus the one with a very small temperature deviation since 1,800 AD. Since it is the summer half-year that is decisive for the growth and the melting of the glaciers, the limiting values among the classes for Graph 5 were estimated by Bariss in terms of ΔU_s. — The minimal summer cooling necessary for the triggering of the growth of

glaciers, which was 1.8 °C below the present, corresponds well to Bacsák's suggested value of 400 canonical units below the Pleistocene average.

The climate types discussed here so far represent the *solar climates* which means that the changing orbit elements of the earth affect the climate by the changing solar radiation factor according to the geographical latitudes without any regard to oceans or continents.The adjectives „oceanic" and „continental" proposed in this paper should not be taken literally; by using them, the author tried to make the temperature differences between summer and winter half-years more palpable for the reader. It should also be emphasized that once a large continental glacier starts to grow during the strongly oceanic type of climate, it creates its own glacial and periglacial climates. Consequently, the effects of the solar climates cannot be felt on and around the glacier. Glaciers continue to exist until the hot summers of the next strongly continental climate melt them. Based on the estimations of Bacsák (1942) and Bariss (1989), Graph No. 6 in Figure 1 displays the sequence and duration of the *actually glaciated and non-glaciated* periods at the vicinity of the 55° northern latitude. It should be stressed that Graphs 1 to 5 constitute a *set* with strong causative relations. This set represents the basic reason for, as well as the chronological distribution of, the solar climates. On the other hand, Graph 6 is less definitive because in addition to the changing orbital elements, other factors (such as epeirogenetic movements, changing carbon dioxide in the atmosphere, etc.) might also have influenced the growth and the melting of the glaciers. It is obvious that during glaciations the effects of the solar climates become gradually more important with increasing distance away from the glacial and periglacial regions.

The main purpose of Figure 1 is to provide a *graphic summary* for the changing climatic environment since the melting of the Riss II glacier. As stated earlier in this paper, the interferences between the high and low values of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$ (Graphs 1 and 2) were the causes for the changing solar climates (strongly or moderately oceanic or continental, or not much different from the present — see Graphs 3-5) which, in turn, were primarily, though not exclusively, responsible for the actual glaciations whose time periods are shown by Graph 6. Concerning the effects of the orbital elements, for a better illustration, the curve of $\Delta(e\sin\Pi)$ was plotted in a reverse way for Graph 2 (i.e., negative values upward). Thus, as indicated by the key on the interference possibilities in Figure 1, a *good interference* between the wave troughs of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$ initiates a strongly oceanic type of climate, whereas that between the wave crests makes a strongly continental one. Furthermore, when a wave crest of $\Delta\varepsilon$ coincides timewise with a wave trough of $\Delta(e\sin\Pi)$ a moderately oceanic climate occurs. Finally, an interference between a wave trough of $\Delta\varepsilon$ and a wave crest of $\Delta(e\sin\Pi)$ sets up a moderately continental type of climate. The „good" interference, which is particularly important for the development of a strongly oceanic or a continental type of climate, means both the sufficient amplitudes of the waves and the simultaneity of the occurrences of the culmination points. It is also obvious from Graphs 1 and 2 that the highs and lows of $\Delta(e\sin\Pi)$ alternate more frequently than those of $\Delta\varepsilon$.

Concerning the effects of the orbital elements on the deviations of the summer and winter temperatures from the 1,800 AD values, the conspicuous parallelism between the

curves of $\Delta(e\sin\Pi)$ and ΔU_s seems to indicate a strong direct relationship between them, particularly for the 55° N. latitude. A reverse relationship is true for the curve of the winter temperature, ΔU_w. Consequently, the changes of the eccentricity and the length of the perihelion appear to primarily determine that at a given time during the Pleistocene, the summer or the winter half-year was colder or warmer than those in 1,800 AD. In the case of a good interference, the role of the obliquity of the earth's axis is either enhancing or diminishing the effects of the other two orbit elements. If $\Delta\varepsilon$ is exaggerating the effects of $\Delta(e\sin\Pi)$, the result will be a strongly oceanic or continental type of climate. If, on the other hand, $\Delta\varepsilon$ is diminishing the influence of $\Delta(e\sin\Pi)$, a moderately oceanic or continental climate will occur. Except for a relatively brief time span between about 32,000 and 60,000 B.P., these relations seem to be valid for the entire 600,000 years covered by Milankovitch's (1930) calculations.

Description of the climatic environment after the Riss glaciation

(Primarily along 55° N. lat. Temperature values in degrees of Centigrade. Time is given in years before present, i.e., before 1,800 AD)

1. Major climatic oscillations based on temperature (Fig. 1)

The ice of the Riss II glacier was almost certainly melted by a *strongly continental* type of climate that culminated at about 175,000. As indicated by Graph 4, along the 55° N. latitude the temperature of the summer half-year was higher by 4.5° than in 1,800 AD.Whereas the hot summers melted the ice, the very cold winters (by 4.4° below present) probably could not produce much snow precipitation. This strongly continental climate, which lasted until approximately 167,000 was followed by *two moderately oceanic* climates and, between them, a *moderately continental* climate when the deviations from the present summer did not exceed the -1.0° and +1.9° values. From about 137,000 to 121,000 another definite continental type of climate prevailed with a peak near 128,000. The summers and the winters during this *strongly continental* climatic oscillation were similar to those around 175,000, described above. A matching of the temperature curves of Graphs 3 and 4 with those of Graphs 1 and 2 indicates that these distinctly continental climates were initiated by the relatively good interferences between the wave crests of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$. Interestingly, Soergel (1937), the originator of the first widely recognized glaciation curve, concluded that the ice of the Riss II lasted until the second strongly continental climate, which culminated at 128,000. Since the two continental types of climatic oscillations discussed above were of about equal magnitude, there is no reason why the one that peaked in 175,000 could not have melted the ice of the Riss II glaciation. Bacsák (1942) was surely correct when he suggested that the Riss/Würm interglacial was free of ice.

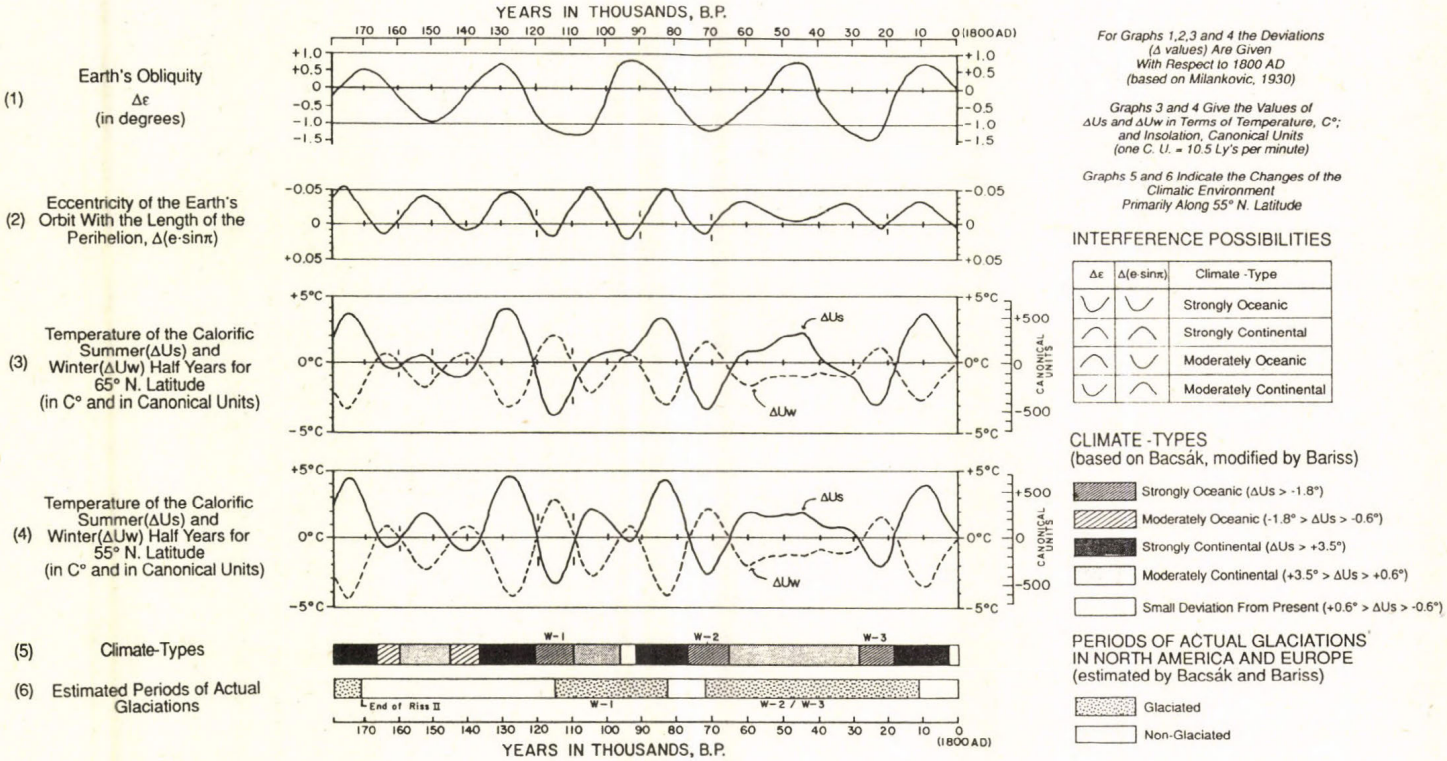

Fig. 1. Changes of selected orbit elements, temperature values, climate types and actual glaciations since the melting of Riss II glacier (approximately the last 170,000 years B.P.) for 55° and 65° northern latitudes [Compiled and modified by Bariss (1989) based on the data of Milankovitch (1930) and Bacsák (1940, 1942)] Secular changes of the earth's obliquity (Δε — see Graph 1) and eccentricity (e), together with the length of the perihelion [Π — see Δ(esinΠ) in Graph 2] with respect to their values at 1,800 AD caused changes in the temperatures of the summer and winter half-years. The secular temperature changes (shown by Graph 3 for 65° N. latitude and Graph 4 for 55° N. latitude) initiated several climate types shown by Graph 5 (according to Bacsák and Bariss). Whereas one of these climates, the strongly oceanic type, triggered the growth of the large glaciers, another one, the strongly continental type, melted them. The actual glaciation shown by Graph 6

After about 121,000 a relatively good interference between the wave troughs of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$ initiated a *strongly oceanic* type of climate that triggered the growth of the Würm I glacier. It was estimated by Bacsák (1942) that after the start of a strongly oceanic type of climatic oscillation, roughly 5,000 to 6,000 years were necessary for the growth of the glacier. Thus the starting time of the *actual Würm I glaciation*, shown by Graph 6, was presumably near the downward culmination of the curve of ΔU_s at about 116,000 when the value of ΔU_s was -3.4° with respect to 1,800 AD. After this strongly oceanic climatic period, a *moderately continental* one prevailed between 110,000 and 96,000, showing maximum summer and winter temperature deviations of +2.2° and -2.9°, respectively. Its summers, however, were probably not warm enough to melt the ice of the Würm I glacier. Consequently, this moderately continental climate, which was even weaker along the 65° N. latitude, did not affect at all the glacial and the periglacial zones where the existing ice sheet influenced the climate. During the next brief time span between 96,000 and 92,000 a very weak, „oceanic-like" climate with mild winters prevailed, one that was *not much different* from the 1,800 AD ending period of this study.

The ice of the Würm I glacier was in all likelihood melted by the hot summers of the next *strongly continental* climate that existed from 92,000 until about 77,000, with peaks around 84,000, when the summer and winter temperatures deviated by +4.1° and -4.1°, respectively, from those in 1,800 AD. The resulting *Würm I/II interstadial*, however, was short-lived because the next climatic oscillation was a *strongly oceanic* type occurring between 77,000 and 66,000, with a culmination around 72,000 when the ΔU_s was -2.8°. Consequently, the Würm II glacier became fully developed by about 72,000.

In all probability the next climate type, a long-lasting *moderately continental* one between 66,000 and about 28,000, was not able to melt the ice of the Würm II glacier because summers were not warm enough. As shown by Graphs 3 and 4, the maximum values of ΔU_s were only +1.9° at about 61,000 along 55° N. and +2.2° at 45,000 along 65° N. latitudes. It is interesting to note that, as mentioned earlier, the curves of the ΔU_s and $\Delta(e\sin\Pi)$ did not run parallel with one another between about 60,000 and 32,000. The reason for this anomaly is not clear, but such could be related to some irregularities in the perturbations within our planetary system.

Between 28,000 and 19,000 came the last *strongly oceanic* type of climatic oscillation with a relatively modest amplitude of -2.1° for the ΔU_s at its culmination around 22,000. Although this oscillation, which might be identified as Würm III, was the weakest among the ten strongly oceanic types of climates that occurred during the last 600,000 years, it did not have to create a new glacier because, as mentioned above, the ice of the Würm II „survived" the ineffective moderately continental climate between 66,000 and 28,000.

Finally, a good interference between the wave crests of $\Delta\varepsilon$ and $\Delta(e\sin\Pi)$ produced the last major climatic oscillation: a *strongly continental* type of climate from about 19,000 that culminated at roughly 10,000 with average temperature deviations of +4.0° for summer and -3.6° for winter with respect to 1,800 AD. The hot summers of this

climatic period melted the ice of the Würm II-III glacier and initiated the *Holocene*. It is important to note that Graphs 3 and 4 were constructed based upon the data of Milankovitch (1930). Because of the scale of his work, the relatively minor climatic oscillations within the Holocene cannot be detailed here.

2. *Some other aspects of the changing climatic environment*

According to many researchers, including Bacsák (1942), during the actual glaciations a *cold steppe climate* with easterly winds prevailed in the periglacial zone of Central Europe south of the ice sheet. Bacsák also suggested that the *loesses* of that region were formed under periglacial conditions. It is more likely, however, that the accumulation of the fine silt-sized particles, which later built up the loess, took place during the melting of the glacier when the outwash plains provided a rich source region for the dust. The geographical locations of the North American and European loess regions seem to support this statement. Bacsák also thought that the very weak oceanic type of climatic oscillation which culminated at 94,000 (with a ΔU_w of $+0.8^\circ$) resulted in temporary reforestation of a part of the periglacial steppe. However, it appears to be unlikely that such a weak oscillation at 94,000 could significantly have diminished the strong cooling effect of the ice sheet in the periglacial zone. Thus during the 172,000 years since the melting of the Riss II glacier, there were two time periods when the cold steppe climate prevailed in the periglacial zone of Central Europe: (1) during the Würm I glaciation for over 30,000 years and (2) in the Würm II/III glaciation for about 60,000 years, with a total over 90,000 years (see Graph 6).

During the ice-free periods of the Riss/Würm interglacial, the Würm I/II interstadial and the Holocene, with a combined time of about 80,000 years, the former periglacial region became *reforested*. These ice-free periods were (and are today) dominated by the Westerly Wind System. Bacsák stressed that it was during this „normal" climatic environment when the solar climates (summarized in Graphs 3-5) were able to control the climatic pattern of formerly periglacial Central Europe.

Major conclusions

1. It is important that researchers of the Pleistocene should always make a clear distinction between the *solar climates* (Graph 5) and the *actually glaciated versus unglaciated* time periods (Graph 6). During glaciations the solar climates could only prevail at locations far away from the glacier.

2. It was Bacsák's important conclusion that the climates during the *ice-free periods* of the Pleistocene were *heterogeneous*. In the literature it was so often stated that the climate of an interglacial time period was „cold" or „warm". Clearly this is an oversimplification.

3. The Milankovitch theory and Bacsák's supplementary works offer merely a *general guide* for the solar climates (a „climatic calendar", as Bacsák called it) of the Pleistocene. However, as stated above, although the effects of the orbital elements seem to be the main factor in initiating the growth and the melting of the glaciers, *other factors* should also be taken into consideration with regard to the origination of the Pleistocene glaciations.

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LATE PLEISTOCENE VEGETATION HISTORY IN HUNGARY SINCE THE LAST INTERGLACIAL

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ABSTRACT

The Quaternary history of flora and vegetation in the area of present-day Hungary can be reconstructed from the relict plant species which survived from various periods, from the landscapes preserving them and partly from the mega- and microfossils studied paleobotanically. Palynological investigations allow insight into the evolution of vegetation over the past 60,000-80,000 years.

INTRODUCTION

During the Quaternary glacial the whole area lay within the periglacial domain of Europe. Thus, influenced by the fundamental change in climate, the Tertiary tropical, subtropical vegetation of the Carpathian Basin has changed radically. In the glacials species shifted from N. to S., from mountains to plains and those survived in refuges spread again in the interglacials mixed and decimated. Meanwhile, some species became isolated from their former contiguous distribution area as relicts or, if also taxonomically isolated, relict-endemics. Ca. 2 % of the total vascular plant flora of Hungary (2148 species) are regarded relicts and endemics occur in almost the same number. They constitute the most strictly protected plants in Hungary. As it is doubtful to distinguish pre-Pleistocene and interglacial relicts, they are collectively considered relicts from warm periods (examples are *Calamintha thymifolia, Cheilanthes marantae, Colchicum hungaricum, Cynanchum pannonicum, Ferula sadleriana, Linum dolomiticum, Onosma tornensis, Pyrus magyarica, Silene flavescens* and *Trigonella gladiata*).

DISCUSSION

Unfortunately, we are short of finds from the older Pleistocene (Járai-Komlódi, 1971) and for lack of absolute dating their evaluation is not always possible *(Fig. 1)*.

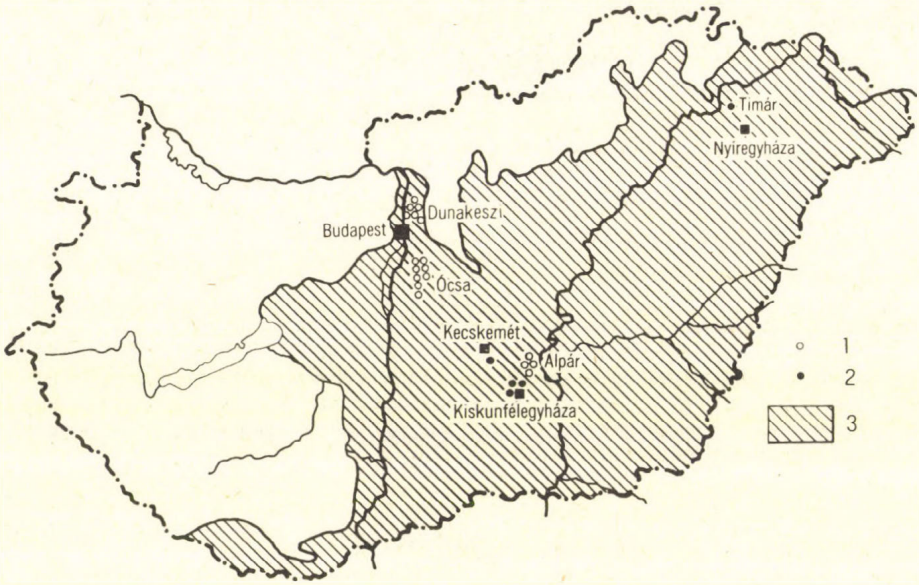

Fig. 1. Würm paleobotanical localities mentioned in the paper. 1 = Holocene and Late Glacial boreholes; 2 = Pleistocene boreholes; 3 = Great Plain

The sporadic remains permit the statement that in the plains of Hungary the European glaciations (glacials and stadials) were characterized by cold-dry periglacial loess pusztas with sparse vegetation and subalpine-subarctic steppe tundra, while in the medium-height mountains alpine vegetation probably also occurred and the refuge areas helped deciduous trees survive. From these areas coniferous forests mixed with birch or other broad-leaved trees spread over the mountains or even to the Great Hungarian Plain in the shorter interstadials, but never during climatic ameliorations of interglacial scale. From cave and open-air charcoal remnants it is known that in the maximum of the first stadial of the Lower Würm the medium-height mountains were vegetated by *Pinus silvestris* and subalpine conifers such as *Larix decidua*, *Pinus cembra* and *P. mugo*.

Last interglacial (Riss/Würm, Eem)

Cave charcoal finds (*Carpinus* sp., *Tilia* sp., *Cotinus* sp. and *Cornus* sp.) attest to the presence of thermophilous deciduous forests in the later phase of the last interglacial. They indicate a temperate, humid climate with a stronger mediterranean influence than today. At the end of the interglacial these thermophilous species were gradually replaced by *Pinus silvestris* and *Larix decidua* in the mountains.

In addition to the identification of megafossils (Sárkány, 1939; Sárkány and Stieber 1950, 1952; Stieber 1952,, 1967), pollen analysis (Zólyomi, 1952, 1987; Járai-Komlódi, 1966a, b) promotes the reconstruction of the evolution of the Hungarian landscape and vegetation over the past 60,000-80,000 years.

Early Würm

In one of the early interstadials, towards the end of the warm spell correlated with Brörup[1] scattered groves developed not only in the medium-height mountains, but also in some parts of the Plain, differing in tree composition in the Danube-Tisza interfluve *(Figs. 2 and 3)* and in the Trans-Tisza region (*Fig. 4*; Járai-Komlódi, 1966 a, b).

There are also differences in climate and vegetation between these lowland regions nowadays. The studied area on the Danube-Tisza interfluve (Kiskunfélegyháza) is one of the driest parts of the Plain as indicated by semiaridity factors (Walter, 1957; Borhidi, 1961). For climatic type this area belongs to the submediterranean forest steppe climatic zone with dry, semiarid summer. This climatic zone is characterized by mid-summer-early-autumn dry period of 1.5-3 months and early-summer rain maximum (April-June).

The present climax plant formation for this zone is steppe with oak forests, on sand Festuco-Quercetum roboris, on loess Aceri tatarico-Quercetum pubescentis-roboris associations. Today most of the area is a cultural landscape. The forests preserved are mostly flood-plain groves and oak forests of limited extension (Festuco-Quercetum roboris, Convallario-Quercetum roboris; Soó, 1940, 1964).

In contrast, the NE margin of the Great Plain — from where the attached palynological analysis is taken — belongs to the Central European climatic type semihumid throughout the year. More specifically, it lies on the border of the forest steppe zone or belongs to the closed oak forest zone. At present it is a cultural landscape with flood-plain forests. The nearest forest steppe area is more properly described as part of the eastern cool-continental forest steppe belt with Convallario-Quercetum tibiscense as its climax association. This present-day contrast in climate and vegetation must have evidently been reflected also in earlier periods.

[1]It is represented by the "Basaharc Lower" (BA) paleosol in Hungarian loess profiles (Pécsi, 1975)

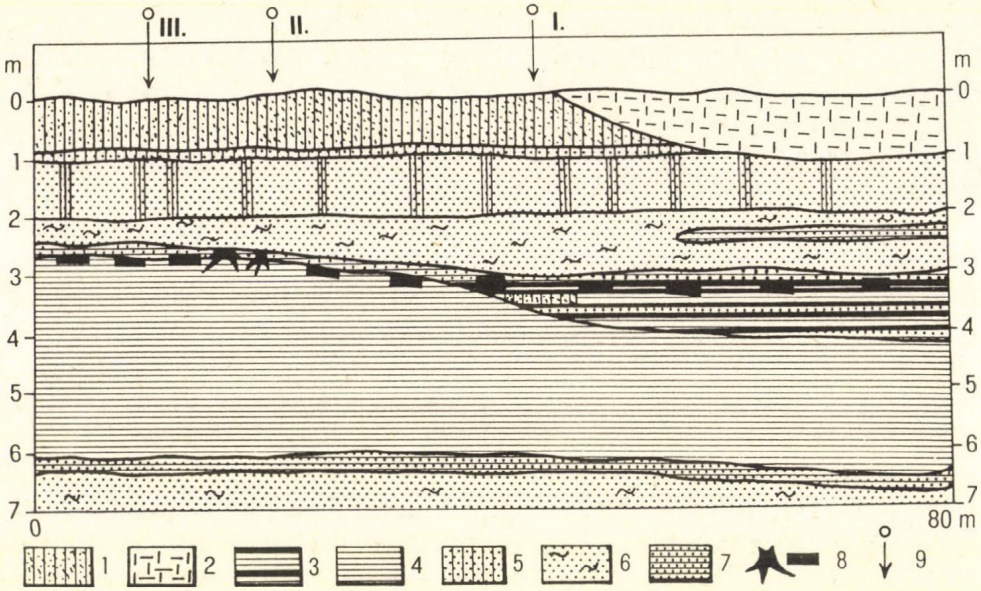

Fig. 2. Geological profile of the Upper Pleistocene locality at Kiskunfélegyháza (after Scherf, 1927, 1936). 1 = loessy silt; 2 = alkali loessy silt; 3 = peaty 'blue' clay; 4 = 'blue' clay; 5 = loessy sand; 6 = fluvial sand; 7 = sand with 'blue' clay; 8 = remnants of *Pinus cembra* trunks; 9 = borehole sites

In the second half of the warming assumed to be the Brörup interstadial, as shown by palynological data (Járai-Komlódi, 1966a), the Danube-Tisza interfluve was vegetated by subarctic birch groves mixed with pines with the predominance of *Pinus silvestris*. *Betula pendula* and *B. pubescens* and only sporadically with other pines (*Pinus cembra, Picea abies, Abies alba, Picea omoricoides, Larix decidua*) and *Alnus* sp. mostly appearing towards the end of the interstadial.

The spreading broad-leaved trees (mainly birch, locally alder or willow) indicate a temporary amelioration, but after that an undoubted cooling is shown by the abundance of pines and the reappearance of *Salix cf. berbacea*.

Simultaneously, on the NE margin of the Great Plain, in the Trans-Tisza region, conifers, primarily spruce (*Picea abies, P. omorica*) and Scotch pine, among broad-leaved trees alder gained prominence. *Pinus cembra* also appeared and formed sparse forests with *Larix decidua*. In both lowland landscapes pine-birch groves were composed of Scotch pine and *Betula pubescens* in marshlands and of Scotch pine and *B. pendula* on drier terrain. On the Danube-Tisza interfluve forest patches alternated with wet subarctic meadows with sedge as well as *Selaginella* sp. and *Pleurospermum* sp.

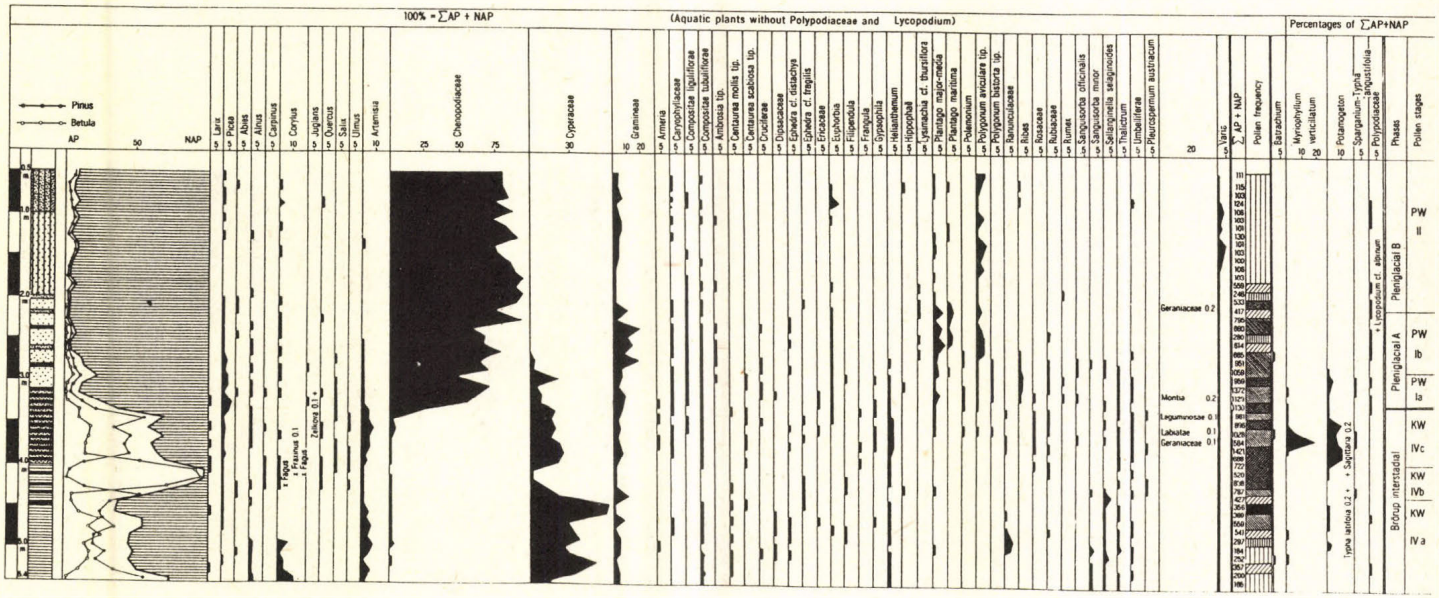

Fig. 3. Pollen diagram of the Kiskunfélegyháza profile (see Legend of *Fig. 2*)

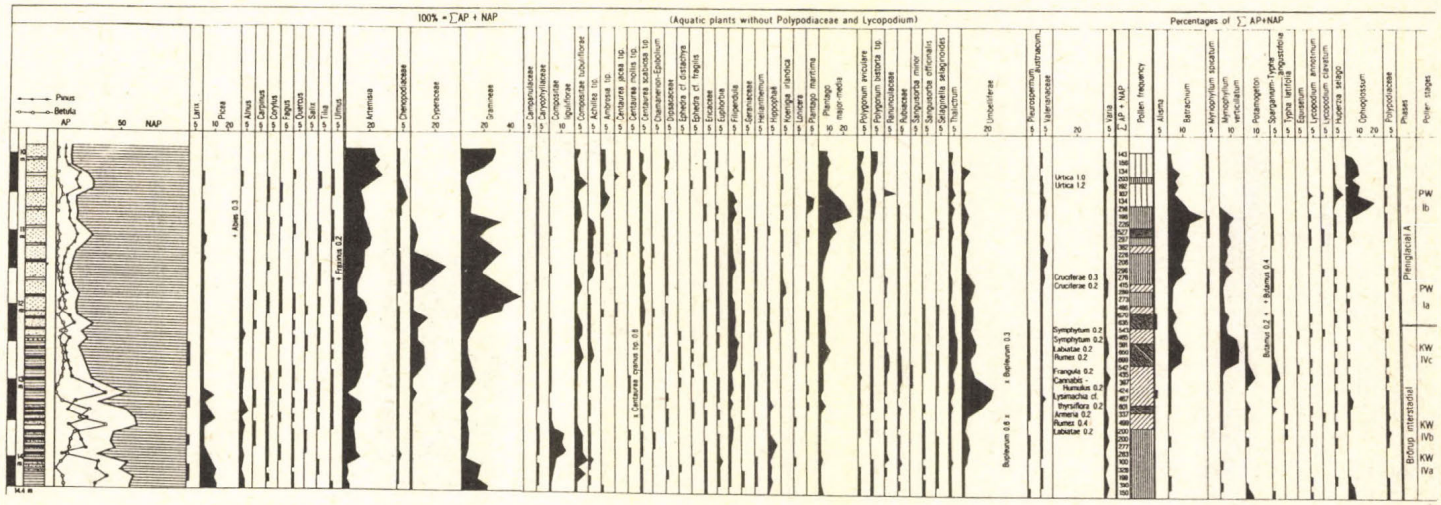

Fig. 4. Pollen diagram of the Timár profile (see Legend of *Fig. 2*)

Pollen analysis reveals the emergence of subalpine high-stalk vegetation (*Polygonum* cf. *bistorta* and *Sanguisorba officinalis*) at the end of the Brörup interstadial. This association, rich in flowering plants, was most typical of the NE margin of the Great Plain, in the lower-lying parts of the grove landscape (Járai-Komlódi, 1966a, b). Here, in addition to dicotyledonous flowering plants (*Sanguisorba officinalis*, *Filipendula* cf. *ulmaria*, *Polygonum* cf. *bistorta*, *Geranium* sp., *Thalictrum* sp., *Rumex* sp., *Epilobium* sp., *Symphytum* sp.), alpine or tundra species such as club-mosses (*Lycopodium selago*) and heather (Ericaceae) also appeared.

Along water-courses alder bushes, on shoals groves of *Hippophaë rhamnoides* grew. On the higher, flood-free terrain, treeless grassy loess pusztas with xerophilous *Artemisia* sp. prevailed in both regions, with heliophyte steppe species (such as *Sanguisorba minor, Ephedra distachya, Helianthemum* sp.).

In the interstadial of the Early Würm, the flood-plains, particularly on the Danube-Tisza interfluve, were overgrown by a very rich aquatic vegetation (*Batrachium* sp., *Myriophyllum* sp., *Sparganium* sp. and *Potamogeton* sp.). Extensive reed and bulrush beds formed as also attested by the dominance of branchiate water-snails (*Bithynia tentaculata* and *B. leachi*).

For the climatic reconstruction some climate-indicator fossils, first of all pollen, were used, keeping the principle of uniformitarianism in mind, assuming the present-day ecological conditions in the overlap of distribution areas of indicator species for the period in question (Fig. 5; Járai-Komlódi, 1969).

The Brörup interstadial flora of the Great Plain is different from that found for the W and NW parts of the Carpathian Basin. In the former *Alnus* and *Picea* are relatively more significant and some thermophilous broad-leaved trees are characteristic. The explanation of the difference may not be found in the colder climate in the Plain, but in drier and perhaps more continental conditions.

The occurrence of *Picea* is sporadic in the central part of the Plain *(Figs. 2 and 3)*, while it is more important on the margin *(Fig. 4)* and *Alnus* and other thermophilous broad-leaved trees (*Quercus* sp., *Ulmus* sp., *Tilia* sp., *Corylus* sp., *Carpinus* sp., *Fraxinus* sp. and *Fagus* sp.) are more abundant, although only reach some specimens per mille. The climate of the Brörup interstadial may not have been favourable for their spreading. According to Stieber (1952, 1967) they still occurred in the refuges of the medium-height mountains, but their extension over the Great Plain was inhibited by *Betula*, which responds more rapidly to climatic amelioration and may have occupied all the available areas. The small amount and sporadic occurrence of the pollen of thermophilous broad-leaved trees and the megafossils of uncertain age are hardly convincing evidence.

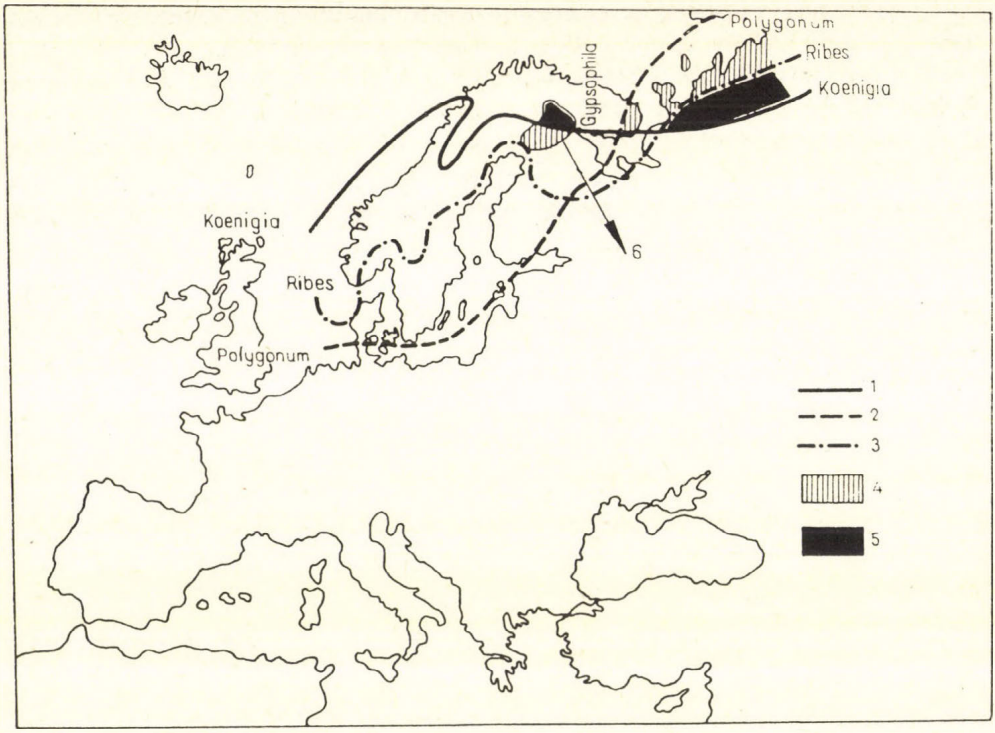

Fig. 5. Climatic reconstruction by the principle of actualism and present-day distribution area of species in the Pleistocene localities of Hungary 1 = the southern border of the area of *Koenigia islandica*; 2 = the northern border of the area of *Polygonum bistorta*; 3 = the northern border of the area of *Ribes alpinum*; 4 = the common areas of two species (*Koenigia islandica* and *Polygonum bistorta*, as well as *Ribes alpinum* and *Gysophyla fastigiata*); 5 = the common areas of three species (*Koenigia islandica, Ribes alpinum* and *Polygonum bistorta*, as well as *Koenigia islandica, Ribes alpinum* and *Gysophyla fastigiata*); 6 = the separated area of *Gysophyla fastigiata* and *Ribes alpinum*

Middle Würm

The ensuing cooling trend gradually pushed back arboreal vegetation everywhere. During the Middle Würm cooling, at least in the first phase of cold and relatively humid climate, subarctic pine forests were replaced sparse subarctic bushes of *Betula* cf. *nana*, *Salix herbacea*, *Pinus mugo*, *P. palaeomontana* and *Alnus viridis* with some *Pinus cembra* and *Larix decidua*. In addition to their pollen, megafossils (identified by Tuzson, 1929; Szepesfalvi, 1930; Scherf, 1936) also attest to subarctic arboreal vegetation.

Among the clusters of trees subalpine high-stalk vegetation, marsh meadows with grasses and sedge, rich in mosses, developed where subalpine club-mosses (*Lycopodium selago*), *Borychium* sp. and tundra species indicate relatively still humid climate. Pollen

analysis was used to identify first in Hungary true tundra species like *Koenigia islandica*, only present today beyond the Polar Circle (Járai-Komlódi, 1966a). Megafossils of cryophilous moss species were found which live today in the alpine zone of the Carpathians and on humid northern tundra marshes. The species *Scorpidium scorpioides*, *Drepanocladus exanullatus*, *D. vernicosus*, *D. fluitans*, *Hypnum hollosianum* are examples (Szepesfalvi, 1928, 1930; Boros, 1952). Besides tundra plants cryophilous and hygrophilous molluscs (*Succinea oblonga* and *Cochlicopa lubrica*) as well as eurytherm loess snails (*Vallonia costata* and *Pupilla muscorum*) also point to cold and humid environments.

During the Middle Würm cooling pine-birch groves were shrinking. To the impact of intensifying cooling, aridity and continentality most of the Great Plain loses its forest cover again. A gradual spreading of xerotherm, cold-continental loess puszta vegetation, rich in Chenopodiaceae on the Danube-Tisza interfluve and *Artemisia* sp. in the Trans-Tisza region. The high-stalk vegetation and the arctic marshes disappeared and the puszta impoverished in species *(Fig. 5)*.

The last interstadial in the Würm which saw a major soil formation is represented by the upper member of the Mende Upper Soil Complex (MF_1) dated by ^{14}C method at 26,000-29,000 years B.P.

Late Würm

In the Late Würm, over loess or sand covered surfaces the vegetation of the Carpathian Basin, with dry and cold climate, could be similar to the Inner Asian continental grasslands, treeless, with cold steppe species, generally poor in species, but rich in grasses and Chenopodiceae.

The cold and arid climate of the Late Würm did not favour aquatic vegetation either. The megafossils of *Betula* sp., *Larix* sp., *Pinus uncinata* and *P. cembra* found in loess and loess-like deposits allow conclusions for the isolated occurrence of these subarctic species and marsh plants in the Plain, along its margins and in medium-height mountains.

The peak of last glaciation was followed by minor climatic fluctuations with a gradual ameliorating trend.

Late glacial

As in the major part of Europe, also in Hungary relatively dry and cold climate prevailed with mostly tundra or tundra-like, subarctic and subalpine vegetation, similar to that typical for periglacial areas during the stadials. A slow forest development began in the stadials of the late glacial (Dryas I, II and III phases).

The minor subdivisions of late glacial (the three Dryas stadials and the three interstadials, Susaca, Bölling and Alleröd) could be only partially identified in Hungary. It is also possible that these periods, slightly more than a thousand or some hundred years ago, were not reflected so clearly in the vegetation of the present-day Hungary. To date, data are available for two late glacial cool periods (Dryas II and III) and the Bölling and Alleröd warming from the Balaton region (Zólyomi, 1952, 1987) and the Great Plain (Járai-Komlódi, 1968; Csongor and Félegyházi, 1987).

During Dryas II the typical treeless loess pusztas were enriched — as attested by pollen analysis — in light-favouring, continental steppe elements, such as *Artemisia* sp., Chenopodiaceae, *Armeria* sp., *Gypsophila* sp., *Helianthemum* sp., *Ephedra* sp. and others. In other places sparse pine-birch taiga patches interrupted by subarctic meadows with mosses and lichens and high-stalk vegetation appeared in the formerly treeless landscape. The protected peat-bogs of the Tapolca Basin subalpine-polar living fossil species (*Pinguicula alpine, Primula farinosa* and *Vaccinium oxycoccos*) and birch marshes (Nyirbátor and Bereg) — although it is not yet proved — may be preserved from this phase.

Pollen analyses from the Great Plain (Járai-Komlódi, 1968) testify that the arctic-alpine *Selaginella selaginoides*, not living in the present-day flora of Hungary, still lived in the area at that time. In the high-stalk vegetation *Epilobium* sp., *Rumex* sp. and *Sanguisorba* sp. as well as the alpine-boreal *Pleurospermum austriacum* and *Thalictrum* sp. grew. Along the rivers and on shoals *Salix, Alnus* and *Hippophaë* species formed scrubs. The fossil remnants of *Dryas octopetala*, a typical tundra plant of the late glacial in Europe, have not yet been found in the territory of the country. For this period, pollen analysis was unable to identify aquatic plants and it only confirmed the existence of few fern from this period.

The sparse *Larix decidua* and *Pinus cembra* forests of the medium-height mountains began to include more and more birch and Scotch pine.

The ensuing relatively rapid Alleröd warming of short (ca. 1,200 years) duration saw — as everywhere in Europe — the spreading of pine-birch forests (*Pinus silvestris, Betula pendula*) and broad leaved trees also appeared: in South-Germany *Populus tremuloides* and *Corylus* sp. and in Hungary *Tilia* sp., *Quercus* sp. and *Ulmus* sp. The contemporary forests may have resembled to the European taiga (S. type, pine-birch taiga mixed with broad-leaved trees in W. U.S.S.R.). Near waters willow-poplar-alder groves and on marshlands alder marsh forests with peat-fern formed. The expansion of forests is also reflected in the pollen diagrams (NAP is reduced from 45% to 8%).

The rising proportions of ferns and aquatic vegetation (*Myriophyllum* sp., *Potamogeton* sp., *Typha* sp., *Sparganium* sp. and Nymphaeraceae point to a milder and more humid climate. At the same time, over higher-lying, drier terrain, under more continental climate treeless grasslands survived.

Palynological research presents a more continental Alleröd climate for the Carpathian Basin than in NW and Northern Europe.

During the last, very short (600-800 years) cool phase of the late glacial, stadial Dryas III, there occurred only minor changes in the composition of forests, only their extension altered, as woodlands shrank in favour of the cold and dry loess pusztas rich in *Artemisia* sp. and Chenopodiaceae. The diversity of aquatic plants was also reduced, groves and marsh forests withdrew and willow-alder formed shoals groves with *Hippophaë* sp.

The short Dryas cool phases marked the end of the Pleistocene and ca. 10,000 years ago the Holocene with minor climatic oscillations but a clear climatic amelioration trend and afforestation, conceivable as another interglacial (Flandrian), began.

The main trends of Holocene vegetation history for Hungary are the same as for Central Europe. The major differences can only be detected in the behaviour of conifers, in the composition of non-arbour flora and in the formation of the steppe, culture steppe and brown forest soils as described earlier (Járai-Komlódi, 1987).

It seems certain that before human intervention (Járai-Komlódi, 1985; Bodor, 1987; Somogyi, 1987; Lóczy 1989) ca. 85% of the Hungarian Plain was covered by natural forests (mostly oak-forests). Today less than 17% of Hungary is forested and a mere 9% can be considered natural.

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SEDIMENTOLOGICAL AND GEOCHEMICAL ANALYSIS OF UPPER PLEISTOCENE PALEOSOLS OF THE HAJDÚSÁG REGION, NE HUNGARY

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ABSTRACT

On the Hajdúság region, Great Hungarian Plain, there are some shallow boreholes which disclose characteristic fossil soil zones in the Upper Pleistocene. Two chronofacies were distinguished by sedimentological and geochemical (DTA, DTG, TG, IR, X-ray, etc.) analyses and radiocarbon dating. The reddish brown steppe-like soil is an isochronous isotype, while the yellowish red steppean-sodic soil is its isochronous but heterotype equivalent for the "Mende Upper" Soil Complex (Hungary) and Stillfried B international stratigraphic horizons. It is considered the oldest known sodification process demonstrated for the area of the Carpathian Basin.

INTRODUCTION

Among the Upper Pleistocene sediments of the Hajdúság, there are characteristic layers of paleosol zones discovered during the systematical geological surveys of the Great Hungarian Plain (Rónai et al., 1963; Rónai, 1985). To date, very scanty information has been published on the genetical relation and stratigraphical position of these layers because they are typically void of fossils and have yielded no radiocarbon dates to this day. Our experience in geomorphic reconstructions suggested that the problem should be solved by a detailed litho and geochemical facies analysis. Among the methods tested, the analysis of grain size composition, the investigation of the clay mineral-carbonate-amorphous material-organic matter paragenesis and of chemical composition have proved to be most promising to reconstruct sedimentation and early diagenetical processes, fossil soil layer formation as well as the paleoclimate and paleogeography.

SAMPLES AND METHOD

The places of sampling are marked on *Figure 1*, while the stratigraphical position of the fossil soil zones is outlined on *Figure 2*. A detailed sedimentological study was performed by Sümegi (1989), serving as a basis for the present study.

Grain-size composition was determined by Papfalvy's hydrometric tests, data processed on an IBM PC/XT computer according to Gyuricza et al. (1985), taking into consideration parameters described by Inman (1952) and Folk and Ward (1957). The grain composition parameters, together with $CaCO_3$ contents were evaluated by cluster analysis, using the graphic method of Passega (1964).

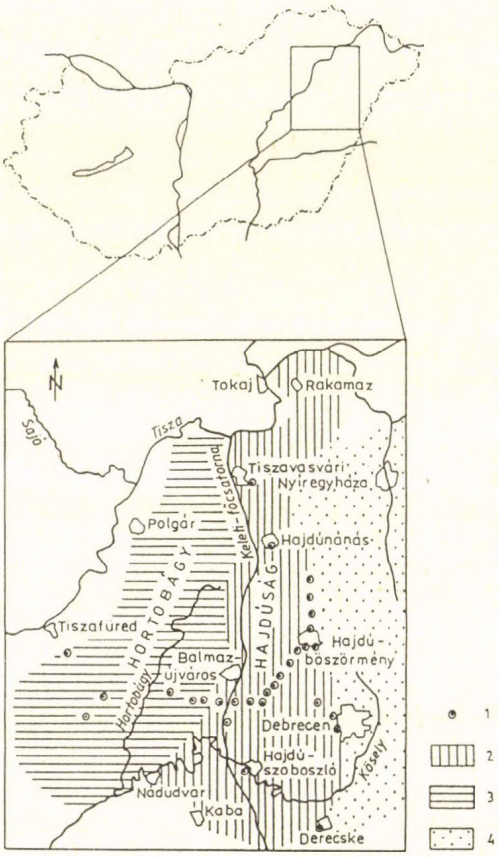

Fig. 1. Sketch map of the area. 1 = profiles; 2 = loess; 3 = flood-plain sediments; 4 = eolian sand

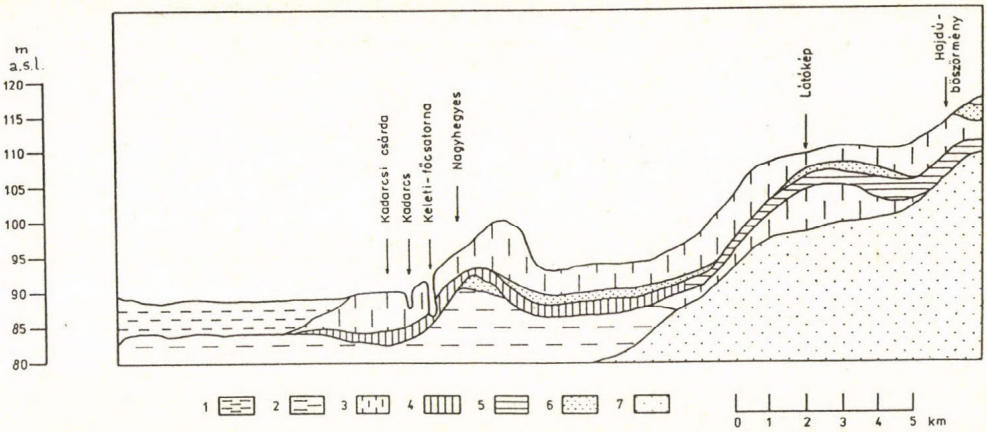

Fig. 2. Sub-surface position of the paleosol layers. 1 = riverine-riparian sediments of the Tisza period; 2 = riverine-riparian sediments preceding the Tisza period (main area of origin for sediment transportation is the northern mountain range); 3 = loess-like sediments; 4 = paleo-sodic soil formed on flood-plain sediments; 5 = steppe paleosol formed on loess and eolian sand; 6 = loessy sand and sandy loess; 7 = eolian sand

Classification of mineral parageneses was performed according to previously published thermoanalytical methods (Szöőr et al., 1978; Borsy and Szöőr, 1981; Szöőr, 1982; Szöőr et al., 1987), complemented by X-ray analysis and IR spectroscopy.

Based on previous experience, homogenised samples as well as several separated grain-size fractions were tested.

Chemical composition was measured using ICP and atomic absorption spectroscopy. Radiocarbon dates were determined by Hertelendi in the Debrecen Nuclear Research Institute.

RESULTS OF THE ANALYSES

Sedimentological analysis

The mathematical-statistical analysis of grain composition analytical parameters has effectively been used for the classification of macroscopically homogeneous loose sediments in Hungary (Bérczi, 1971; Molnár and Krolopp, 1978; Molnár and Geiger, 1981; Geiger, 1982, 1986). A method for the sedimentological analysis of three type section outcrops elaborated by authors is presented *(Figs. 3-5)*.

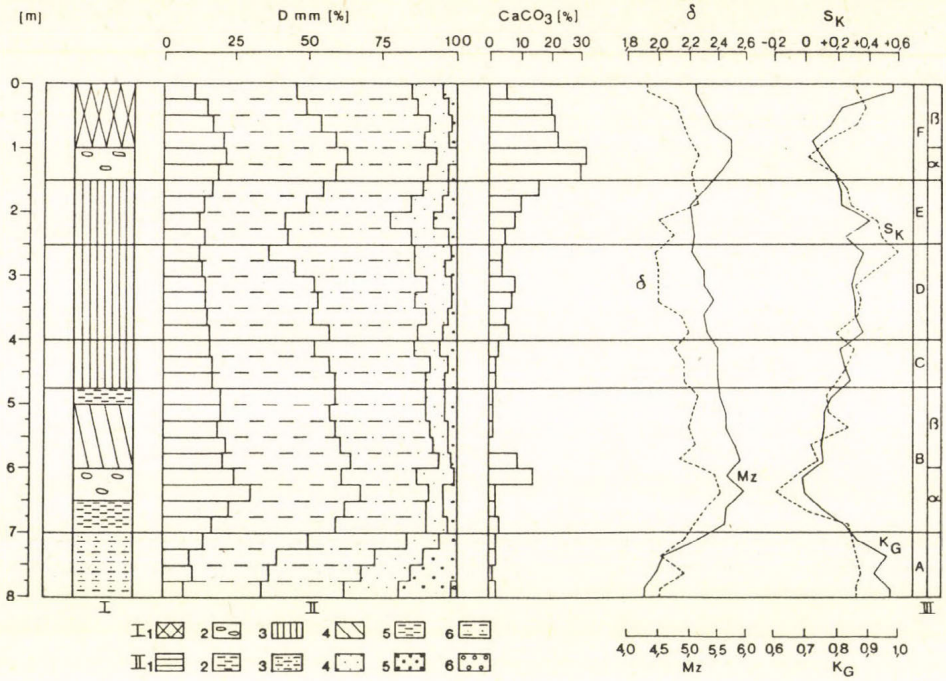

Fig. 3. Results of the sedimentological investigation of the Hajdúböszörmény section. I = Layer sequence; 1 = recent soil; 2 = horizon with concretions; 3 = loess; 4 = paleosol; 5 = clayey silt; 6 = sandy silt. II = Grain size composition: 1 = < 0.002 mm; 2 = 0.002-0.02 mm; 3 = 0.02-0.06 mm; 4 = 0.06-0.1 mm; 5 = 0.1-0.2 mm; 6 = 0.2-0.32 mm. III = Lithozones

In the present paper no detailed characterization of the sediments of the area is given. It is important to mention, however, that this sedimentological method adequately separates sediments formed on dry and wet surfaces, i.e., loess and paleosol horizons. The parent rock for paleosol formation was loess or loess-like sediments, the main difference lying in the clay mineral component, being considerably higher in most soil types *(Fig. 6)*. In Passega's CM diagram, the projection point of different sedimentological facies are grouped in fairly well demarcated domains *(Fig. 7)*. We can observe, in the case of near-surface recent soils, only the values of the coarse grain-size range (C) changes, while in the case of fossil soils, the medium values (M) are also variable. Obviously, this can be explained by different stages of decomposition and early diagenesis. The simplest explanation at hand is based on temporal differences in soil formation, but further on we can postulate that decomposition in paleosols took place under warmer and more humid climatic circumstances. These problems were adequately clarified by geochemical analyses.

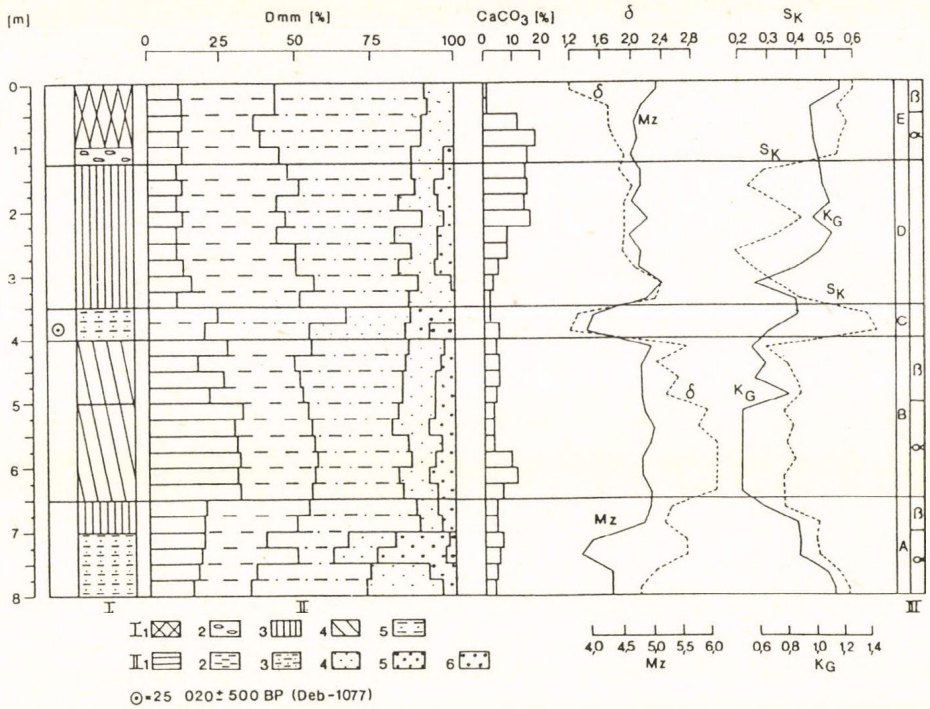

Fig. 4. Results of the sedimentological investigation of the Látókép section. I = Layer sequence: 1 = recent soil; 2 = horizon with concretions; 3 = loess; 4 = paleosol; 5 = sandy silt. II = Grain size composition: 1 = < 0.002 mm; 2 = 0.002-0.02 mm; 3 = 0.02-0.06 mm; 4 = 0.06-0.1 mm; 5 = 0.1-0.2 mm; 6 = 0.2-0.32 mm. III = Lithozones

Geochemical analysis

Our series of analysis was started on untreated and homogenised samples without silting, separation or chemical treatment in air-dry state. On *Figure 8*, DTA, DTG and TG curves (derivatograms) on soil samples of reddish colour from 4 outcrops are demonstrated. These derivatograms, produced on 'average' samples yielded very poor information on mineral paragenesis, though the typical run-off of the curves can be considered characteristic of the paleosol samples analysed. *Figure 9* is concerned with the separated grain size group analyses of the above (average) samples. Comparison can effectively be made on DTA curves fitted on equal temperature scale *(Fig. 9.)*. The mineral association indicated by thermoanalytical investigations was corroborated by complementary X-ray and IR analyses. The percentage values of the TG curves allow a quantitative comparison of the individual constituents *(Table 1)*. Results of thermal analyses can be summarized in the following:

Table 1: TG - parameters of the fossil soils (Legend see Fig. 8)

Sample		TG - parameters, Δm [%]				
		H_2O_I.	H_2O_{II}. + Org.	H_2O_{III}.	CO_2	Σ 25 - 1 000 °C
Debrecen	Rock	2.33	0.84	1.87	0.75	5.79
	> 0.06 mm	1.19	0.42	0.95	0.42	2.98
	0.06 — 0.02 mm	1.67	0.46	1.11	0.46	3.70
	0.02 — 0.002 mm	5.12	1.78	3.34	1.06	11.30
	< 0.002 mm	7.37	2.20	6.44	0.24	16.25
Hajdú-böszörmény	Rock	2.43	0.84	1.97	0.75	5.99
	> 0.06 mm	1.34	0.40	1.07	0.44	3.25
	0.06 — 0.02 mm	1.62	0.35	1.06	0.46	3.49
	0.02 — 0.002 mm	5.94	1.77	3.96	1.14	12.81
	< 0.002 mm	7.79	2.74	6.20	0.58	17.31
Látókép	Rock	2.88	1.09	2.58	1.09	7.64
	> 0.06 mm	2.59	0.89	1.94	1.13	6.55
	0.06 — 0.02 mm	2.46	0.85	2.08	1.14	6.53
	0.02 — 0.002 mm	2.69	0.96	2.30	1.15	7.10
	< 0.002 mm	4.50	1.46	3.66	1.99	11.61
Nagyhegyes	Rock	4.07	1.02	2.44	—	7.53
	> 0.06 mm	1.71	0.39	1.17	—	3.27
	0.06 — 0.02 mm	1.89	0.51	1.29	—	3.69
	0.02 — 0.002 mm	4.56	0.89	2.77	—	8.22
	< 0.002 mm	6.38	1.26	3.69	—	11.33

— Paleosol horizons comprise a considerable and measurable amount of humic matter (DTA peak temperature 300-350 °C), opposed to the underlying and covering formations, in spite of their considerably oxidized state reflected by the reddish colour.

— The carbonate content of these samples is typically low, sometimes under the detection limit. Thermal dissociation around 700 °C indicates the infiltration of Mg^{++}. The carbonate is amorphous or cryptocrystalline, concentrated mainly in the fraction below 0.002 mm.

— In the clay and fine silt fraction, apart from the clay minerals illite-montmorillonite, degraded kaolinite (Bidló, 1980) and halloysite-kaolinite (Borsy and Szöőr, 1981) was found.

The mineral paragenesis reflects soil formation on mild and humid climate. Among the sections presented here, Hajdúböszörmény *(Fig. 3)* and Látókép *(Fig. 4)* can be assigned to this type.

In the Nagyhegyes profile on the eastern margin of the Hortobágy (Fig. 5) the same reddish soil horizon was also spotted at 4.0-4.5 m; however, the mineral composition of the underlying layer (4.5-5.0 m) is totally different. The thermoanalytical character of the profile is demonstrated on Figure 10 and Table 2. Also considering X-ray and IR analyses mineral phases were identified.

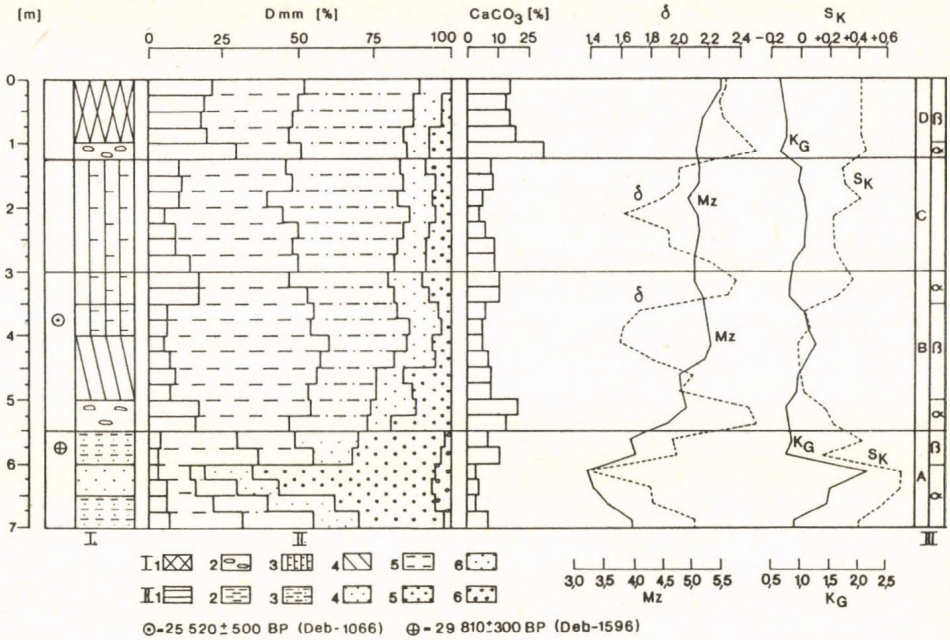

Fig. 5. Results of the sedimentological investigation of the Nagyhegyes section. I = Layer sequence: 1 = recent soil; 2 = horizon with concretions; 3 = loess-like flood-plain sediments; 4 = paleosol; 5 = sandy silt; 6 = eolian sand. II = Grain size composition: 1 = < 0.002 mm; 2 = 0.002-0.02 mm; 3 = 0.02-0.06 mm; 4 = 0.06-0.1 mm; 5 = 0.1-0.2 mm; 6 = 0.2-0.32 mm. III = Lithozones

Table 2: TG - parameters of the Nagyhegyes fossil soil section (CO_{2I} = carbon dioxide from protodolomite and kutnahorite; CO_{2II} = carbon dioxide from calcite; others see Fig. 8)

Depth [m]	TG - parameters, Δm [%]					
	H_2O_I.	H_2O_{II}. + Org.	H_2O_{III}.	CO_{2I}.	CO_{2II}.	Σ 25 - 1 000 °C
4.00 - 4.25	4.0	1.02	2.44	—	—	7.53
4.25 - 4.50	3.2	1.02	1.85	1.02	—	7.12
4.50 - 4.75	3.3	0.89	1.97	6.20	1.57	13.98
4.75 - 5.00	2.5	0.31	1.43	3.56	3.68	11.74
5.00 - 5.25	1.8	0.53	0.88	1.50	2.46	7.22
5.25 - 5.50	1.7	0.34	1.03	1.55	1.89	6.53

Typical of this latter formation at 4.5-5.0 m is the considerable quantity of amorphous polysilicates, polyaluminates (see the parameter of H_2O_I). Apart from the dominant illite (-montmorillonite), the clay mineral constituent also contains chlorite. Essential amounts of complex carbonates, calcite, protodolomite and kutnahorite also occur partly in amorphous and partly in crystalline state. X-ray analysis indicated the presence of gypsum as well. Organic matter content can be traced, but it is essentially less than in the soil layer of reddish brown colour.

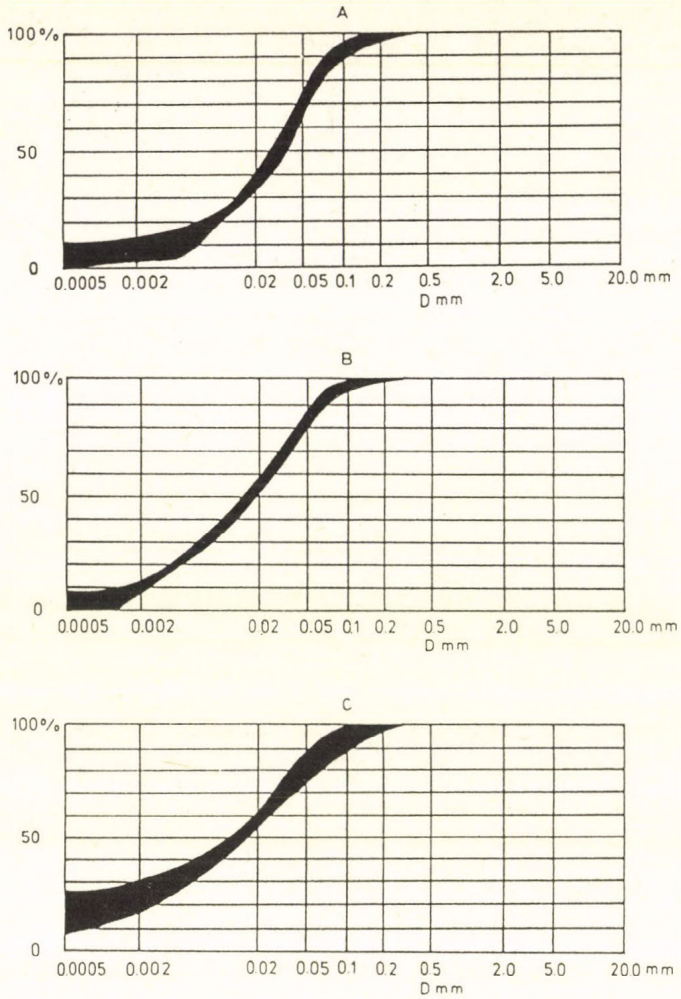

Fig. 6. Grain composition of paleosols and loess. A = loess formed on dry surface; B = infusion loess; C = paleosols

This mineral paragenesis is seemingly very similar to that of the Szabadkigyós sodic 'puszta' sediments (Szöőr et al., 1978). It is also remarkable that in course of the investigation of the Hortobágy sodic soil sequences formed on loess basis, Széky-Fux and Szepesi (1959) found similar DTA curves described as atypical, degraded structure illite-montmorillonite.

Gerei et al. (1966) emphasized, in connection with the investigation of sodic plains (both solonetz and solonchak type) the significant alteration of clay minerals, the illite-montmorillonite character and the formation of amorphous matter.

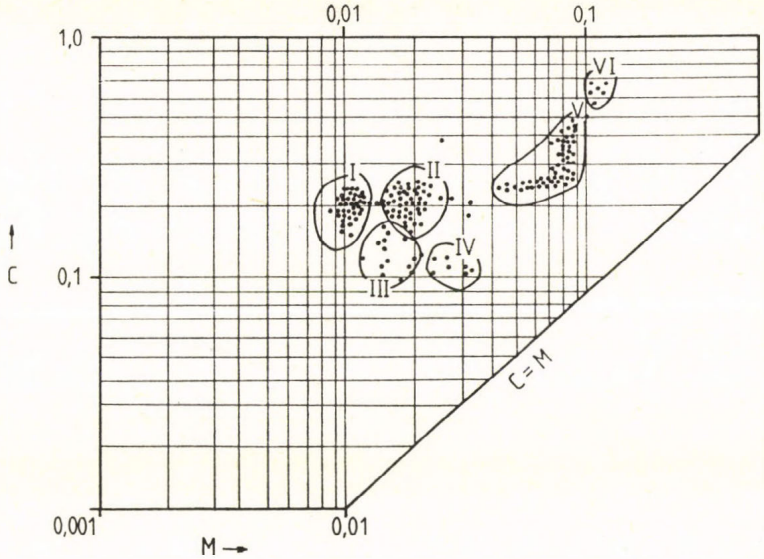

Fig. 7. Delimitation of the Hajdúság sediment types on the CM diagram. I = infusion loess; II = typical loess; III = recent soil; IV = paleosol; V = loessy sand and sandy loess; VI = eolian sand

On the basis of the above observations we could identify a steppe-like sodic plain, probably meadow solonetz type formed on temperate-arid climate.

The 'paleo-sodic' character can be further supported by chemical analyses.

The sodic character (presence of CaO polysilicates, polyaluminates) was tested by

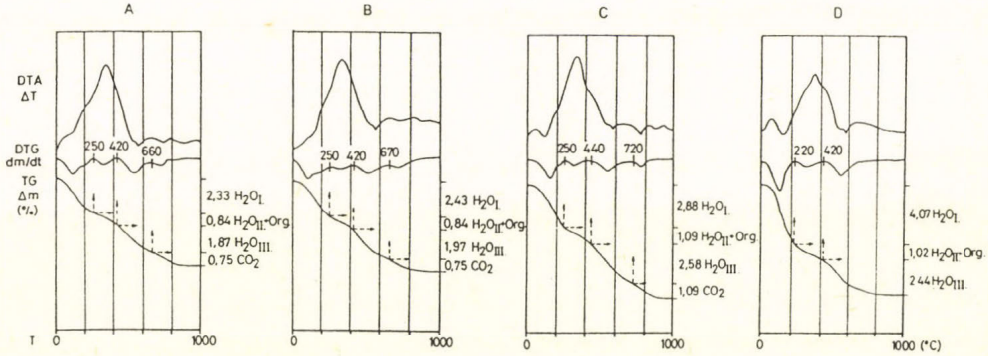

Fig. 8. DTA, DTG. TG curves of paleosols. A = Debrecen; B = Hajdúböszörmény; C = Látókép; D = Nagyhegyes (4.00-4.25 m); H_2O_I. = slightly bound water content; H_2O_{II}. + Org. = decomposition of organic matter; H_2O_{III}. = loss of structural water content of clay minerals; CO_2 = carbon dioxide from the decomposition of carbonates

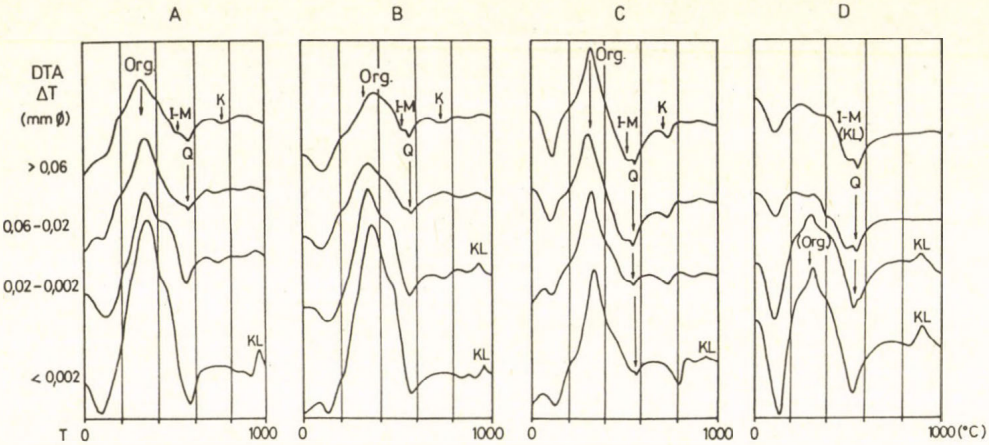

Fig. 9. DTA curves of grain size fractions from paleosol samples. A = Debrecen; B = Hajdúböszörmény; C = Látókép; D = Nagyhegyes (4.00-4.25 m); Org. = organic matter; K = carbonate; Q = quartz; KL = chlorite; I-M = illite (-montmorillonite)

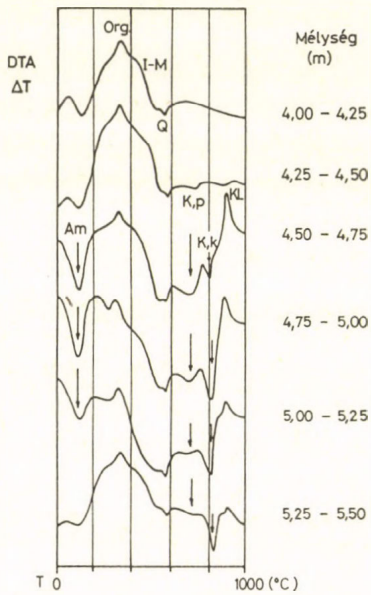

Fig. 10. Analysis of the DTA curve of the Nagyhegyes paleosol section. Org. = organic matter; Q = quartz; K,p;K,k = calcite, protodolomite; KL = chlorite; I-M = illite (-montmorillonite); Am = amorphous material

Szepesi's method for NaF pH measurements (Jolánkai, 1974). One-hour measurements in the paleo-sodic sediments resulted pH values of 10.28-10.48, while those of 24-hour measurements ranged between 10.03-10.51 pH, that stands for considerably sodic soil.

No increase in the concentration of alkaline elements was found in the paleosol *(Fig. 11)*, that can be explained by uplifting migration and chromatographic processes of the early diagenesis.

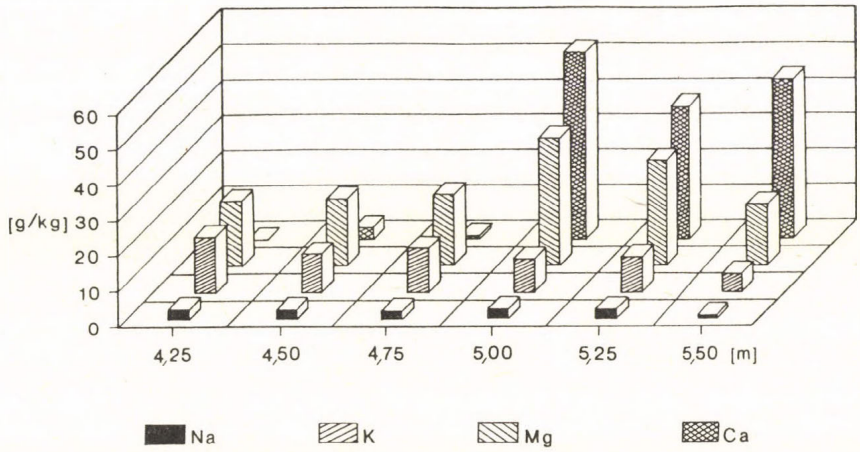

Fig. 11. Fluctuation of alkaline elements and alkaline earth elements in the Nagyhegyes section

CONCLUSIONS

Considerable amounts of eolian silt accumulated during an exactly not specified phase of the Upper Pleistocene in the Hajdúság region. In those days, the area of the Hajdúság was not yet separated, from a sedimentological point of view, from the Nyirség region. In the cool-cold but more humid Middle Würm period sandy loess accumulated in the area. Just in this period, the geomorphological and sedimentological development of the Hajdúság area started to be different than that of the Nyírség alluvial fan.

By the end of the Middle Würm, between 26,000-32,000 years B.P., soil formation took place amidst milder climatic conditions of an interstadial phase. In the Hajdúság plain, on the flooded meadows, sodic soil, while on the Hajdúhát heights chernozem formed. On the basis of radiocarbon dates, the formation of these soils was contemporaneous.

The reddish brown paleosol horizon forms a key layer of stratigraphic significance in the infusional and typical loess sections of the study area. The paleosol layers were correlated on the basis of their faciological character and stratigraphical position with the "Mende Upper" Soil Complex (Pécsi, 1965, 1972, 1975, 1977, 1979; Pécsi and Pevzner,

1974; Stefanovits, 1965; Pécsi-Donáth, 1979). The steppe-like soil of the Hajdúhát is an isochronous, isotype facies of the key section, while the sodic steppe soil on the Hajdúság is its isochronous but heterotype equivalent for the "Mende Upper" Soil Complex. This latter formation was considered contemporary, in the level of international stratigraphical horizons, the soil complex Stillfried B (Fink, 1969).

It is well known that most of the Alföld sodic steppes mainly formed in the Holocene. Spot-like occurrences of this phenomenon were traced on Pleistocene loess surfaces as well as on the Nagykunság, southern parts of the area lying to the east of the Tisza river and the Hortobágy (Geological Map of the Hungarian Geological Survey at 200,000 scale). There is practically only a very small altitudinal difference between the Pleistocene sodic spots and the neighbouring Holocene formations on the present surface.

The Nagyhegyes profile is located some 5-6 km away from the Hortobágy. It is supposed that sodification started much before the date claimed until now (Boreal phase of the Holocene — see Jakucs, 1981), already in the dry-warm period of the Pleistocene interglacial. Primary sodification in fact did not affect the Hortobágy, known to be still a flood plain at that time but marginal areas around the Late Pleistocene flood plain, such as the Hajdúság, only periodically inundated. It is considered the oldest sodification process demonstrated for the Carpathian Basin.

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MALACOLOGICAL AND ISOTOPE GEOCHEMICAL METHODS FOR TRACING UPPER QUATERNARY CLIMATIC CHANGES

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ABSTRACT

Radiocarbon ages $\delta^{18}O$ isotope ratios of mollusc finds from Quaternary fluviatile and eolian deposits are presented. Ten malacofaunistic levels and some species of chronological indicator role have been ascertained within the period 7,000-32,000 years B.P.

The paleoclimatological investigations, according to the rules of sedimentology, have been based on the oxygen isotope analysis of properly collected *Pupilla muscorum* shells. Oxygen isotope ratios of remote deposits from the same period showed good agreement. Temperatures obtained from the results of the isotopic studies and of a malaco-thermometer constructed on the basis of the dispersion and climatic demands of mollusc species are also closely correlated.

Comprehensive studies of the paleoclimatic changes, chronological, isotope geochemical data as well as bioindicative results showed the same climatic periods as found in N. and W. Europe, although the climate of the studied area was of a rather continental character at the end of the Pleistocene.

BIRTH OF CONCEPT

Several paleontological methods have been developed lately for tracing climatic changes in the Quaternary period. Some scholars aiming at a synthesis compiled climatic curves on the basis of the changes in the succession of the vertebrate fauna (Kretzoi, 1957; Kordos, 1977), palynological spectra (Járai-Komlódi, 1969; Heusser, 1973) or comparative studies on the basis of insect remains (Coope, 1975; Coope et al., 1971).

To date, nobody has risked the compilation of concrete temperature values based on microstratigraphical analysis of terrestrial molluscan (gastropode) fauna. This can be possibly understood knowing that terrestrial gastropodes were generally assigned into two extremes; individual species were treated as microclimatological indicators only and

taxa with great tolerance spread over large areas were considered absolutely useless for climatic reconstruction due to the wide temperature range typical of these species.

In our opinion, these considerations should be complemented by the following:

The micro- and macroclimatological differences within lowland regions covered by open vegetation are very small and, consequently, do not hinder reconstruction of climatic changes.

Species with large distribution area and great tolerance are active only in certain parts of the year, functioning in limited productive periods with a given temperature and humidity. During this time they are consuming food and constructing the organic-non-organic biomineralized snail-shell.

As the individuals survive the unfavourable period in an anabiotic state, our paleoclimatic deductions are necessarily valid for the contemporary active period, i.e., due to the biological rules of nutrition, the vegetation period.

On the basis of the above considerations, the ecological, paleoecological characteristics of several terrestrial Gastropoda species, persistent from the Upper Pleistocene to our days were analysed with special regard to the quality of the basement, humidity and role of the vegetation (Boycott, 1934; Frömming, 1956; Füköh, 1987; Kerney, 1966; Krolopp, 1983; Lozek, 1964; Meier, 1985; Sparks, 1964; Watts and Bright, 1968).

Moreover, based on the work of mesoclimatic research stations, information has been obtained on climate from which we could determine the optimal living circumstances for several taxa, as well as the maximum and minimum temperatures these taxa tolerate, i.e., their resistence range that was further corroborated by phenological data.

After determining the resistence range, we can compute on the basis of the equation published by Skoflek (1977) — slightly modified — vegetation climatic optimum and, consequently, the July mean temperature. Testing the mathematical validity of our hypothesis on a recent sample it has become obvious that we can reconstruct paleoclimatic changes characterized numerically by July mean temperatures relying on the malacofauna collected from loess regions.

The application of this 'malaco-thermometer' elaborated by Sümegi (1989) was tested on a detailed eco-statistical analysis of the Upper Pleistocene-Holocene malacofauna of the Alföld region (Great Hungarian Plain).

It was considered important to check the validity of this procedure against other methods. It is well-known that by the help of the shift in the ratio of stable oxygen isotopes ($\delta^{18}O$), periods of warming and cooling can be demonstrated (Emiliani, 1955, 1958; Shackleton, 1969; Harmon et al., 1978; Keith et al., 1964).

Measurements of $\delta^{18}O$ and $\delta^{13}C$ are prevailingly used recently for the paleoecological analysis of terrestrial formations, as well as tracing processes of evaporation and desalinization in natural water systems and following changes in humidity, aridity and temperature in general. Isotope geochemical studies of the freshwater shell carbonates has proved that the measurement parameters are really useful indicators of these processes

(Fritz and Poplawski, 1974; Keith et al., 1964; Siegenthaler and Eicher, 1986; Stuiver, 1970).

Attempts have been also made at palynological reconstruction (Eicher et al., 1981) as well as paleoclimatic interpretation of the changes in vertebrate succession during the Quaternary period (Deák and Kordos, 1980; Kordos, 1981). This latter also involves construction of climatic curves.

Stable isotope measurements on molluscan shells were partly aimed at solving the problem whether the values of $\delta^{18}O$ [‰] correlate with temperature values [°C] determined by malaco-thermometry. In case of a strict correlation, qualitative information obtained by stable isotope measurements were suitable for immediate determination of quantitative paleoclimatological data.

For the interpretation of the results it was considered important to calibrate the time-scale of our climatic curve with absolute chronological data.

Radiocarbon measurements were performed, parallel to the determination of stable isotope ratios, on Gastropoda-shells. Validity of isotope geochemical analyses was secured by conditions of measurements installed previously (Hertelendi et al., 1987). In the layers of vertical outcrops there are formations practically void of fossils. Their detailed lithofaciological analysis was complemented by geochemical facies analysis. By the help of a complex thermoanalytical method of clay mineral-carbonate paragenesis, fossil soil zones developed during different climatological circumstances could be identified. These places of information indirectly complemented, the climatic reconstruction based on fossils.

PALEOTEMPERATURES

In the compilation of the malaco-thermometer we had to consider that malaco-phenological data from the study area are scarce.

The area of characteristic members of the molluscan fauna is only partly overlapping, and the composition typical of loess has changed by our days. Therefore, we have adopted the approach to study the optimal mean temperature in the vegetation period at the border of the area of occurrence for species with large distribution. In the case of recent distribution it was also considered how far the species investigated intrude upwards in the mountains and which latitudes they reach (Ant, 1963; Ehrmann, 1933; Kerney, 1976; Kerney et al., 1983; Klemm, 1974; Liharev and Rammel'Meier, 1962; Ložek, 1964; Soós, 1943).

For seven gastropode taxa, optimal temperature and resistance ranges were successfully determined (Sümegi, 1989, p. 75):

Species	Optimum [°C]	Resistance range [°C]
Vallonia tenuilabris	9 ± 2	(4-13)
Columella columella	10 ± 1	(5-15)
Columella edentula	15 ± 1	(10-20)
Pupilla muscorum	16 ± 1	(10-22)
Succinea oblonga	16 ± 1	(13-19)
Pupilla triplicata	20 ± 2	(16-24)
Cepaea vindobonensis	22 ± 2	(18-26)

The validity of the above temperatures is underlined by the following:

Pupilla muscorum is a holarctic faunal element, intruding in Europe in the subarctic and mediterranean region. They can be found in mountains to 2,300 m altitude and, towards the North, to 70° latitude. Its lethal temperature range can be characterized by the 10 °C July mean temperature isotherm. Its maximal resistance temperature value is 22 °C (20-24 °C), also considering its demands for relative humidity (40-80 %; Ant, 1963).

Succinea oblonga is a taxon with Euro-Western Siberian distribution. Intrusion to altitude is determined at 2,000 m, towards the North, to latitude 61° N. The minimum temperature of existence is 13 °C July mean temperature; this isotherm is closely followed, in Britain as well as on the Scandinavian peninsula. Its eastern distribution is delimited by the Western Siberian marshes. This taxon requires a cool, temperate, humid climate. Resistance temperature is about 19 °C. This is confirmed by the fact that it is getting scarce in the Alföld, but remaining important in the NE region where, on relict marshlands July mean temperature is about 18-19 °C.

Columella edentula is of holarctic distribution, found in the Alps to 2,300 m altitude and, towards the North, to 70° latitude (same as *Pupilla muscorum*), thus the minimum temperature of existence can be considered the same (10 °C). For the other extremes, resistance temperature can be favourably studied in the Carpathian Basin. The boundary of the distribution area is running along the southern border of the northern mountain range; on the Alföld, this taxon only occurs in some marshes of cooler microclimate. Its resistance temperature in the higher range is close to that of *S. oblonga* (18-20 °C), but its resistance is better. The optimum for its existence is 14-15 °C July mean temperature.

Columella columella is an arcto-alpine element. It is known in the Alps at 2,900 m altitude. Minimum temperature of existence is marked by the 5 °C July mean temperature isotherm. The other extremes are delimited on the basis of data from Scandinavia, where it is extending to the 15 °C July mean temperature isotherm.

Vallonia tenuilabris in Europe was very important during the Pleistocene cold phases but it got completely extinct by the Holocene. Its current living place include regions of the U.S.S.R. and Mongolia in the High Altai (Ložek, 1964). In the opinion of Liharev and Rammel'Meier (1962), it is found in the Amur Basin and the Great Xinggan Mountains, China. Our personal experience is that this taxon intrudes up to the subnival region, where July mean temperature is 4 °C. In the lower-lying regions, it reaches down to where July mean temperature does not exceed 14 °C.

The species *Pupilla triplicata* is a faunal element of South-East European distribu-

tion. It can exist in the Alps to 1,500 m altitude, following the 16-18 °C isotherm. The northernmost limit of its distribution is the Krkonose Mountains (50° N. latitude).

There are isolated occurrences in Bohemia and Hungary along the southern boundary of the Transdanubian Mountains. The eastern boundary of its distribution is at the Crimean peninsula and in the Caucasus, along the 24 °C isotherm.

The species *Cepaea vindobonensis* is a ponto-mediterranean element of the fauna with an extending area. During the Pleistocene, this species appeared only during certain interglacial periods. Compared with the species *P. triplicata*, it is less resistant to aridity and requires more substantial vegetation cover. It is seldom found in the Alps over 1,200 m (18 °C isotherm), but occurs on the eastern border of the Polish Plain. The south-eastern boundary of the distribution area is running along the Caspian Sea. On the Great Hungarian Plain it lives in biotopes where the mean value of July temperature is around 22 °C.

The recent areas of the species *Succinea oblonga* and *Pupilla muscorum* overlap, but it can be observed that *S. oblonga* has an essentially more restricted temperature range, its biotope is the riparian region of lakes and running waters, marshes and bogs, while the area of *P. muscorum* is considerably larger.

The dominance of *S. oblonga* invariably indicates a cool, humid (temperate) climate, while that of the species *P. muscorum* always shows a more continental one.

In spite of the fact that specimens of *Columella edentula* and *C. columella* can occur in identical ratios within a given layer, the complementary character of the dominance of these two species is regular. Negative correlation indicates maximum cooling (stadials) or post-stadial periods which are somewhat more temperate, cool and humid.

The maximum occurrence of *C. columella* points to a stadial phase; the dominance of *C. edentula* denotes the transitional cool phase preceding and following the stadials.

The relation between the dominance of *Vallonia tenuilabris* and *Succinea oblonga* reflects a similar negative correlation.

During the analysis of the Alföld sections it was found that the climatic curves drawn by the malaco-thermometer better reflect warm than cool spells. The reason for this can be, in our opinion, that with rising climatic temperature, relative humidity is decreasing. This is a major ecological factor dramatically changing the composition of the terrestrial Gastropoda fauna and resulting in rapid extinction.

In the periods of cooling, the increase in relative humidity maintains (relatively) favourable living conditions, resulting in a slow, continuous change in the faunal succession.

The simultaneous change in temperature and humidity can lead to characteristic patterns of $\delta^{18}O$ ratios of shell carbonates. By increasing temperature and decreasing humidity, the shift in the values of $\delta^{18}O$ is positive, by decreasing temperature and increasing humidity, this value is negative.

Hundreds of analytical data of molluscan shells collected at 20 cm intervals in

various sections prove that $\delta^{18}O$ and $\delta^{13}C$ changes are essentially influenced by taxonal relations as well as environmental conditions. This taxonal determination is illustrated by analytical series on three species from various isopic facies on *Figure 1*.

Apart from taxonal relation, conditions of humidity and aridity are, as a rule, related with the actual values as manifested in the analytical results of two localities of heteropic facies *(Fig. 2)*. *Figure 3* proves that the coordinate points obtained from the specimens of one taxon, in our example, that of *Pupilla muscorum*, originating from far-lying but isochronous and isotype facies outcrops are situated in a well-defined field.

Considering these regularities it was deduced that for tracing paleotemperature changes by serial measurements we have to select an ecologically resistent taxon, widely distributed in our sections both horizontally and vertically. We selected *Pupilla muscorum* for detailed examinations and found that the tendency of the changes of $\delta^{18}O$ values corresponds to that of the malaco-thermometrical curve *(Fig. 4)*.

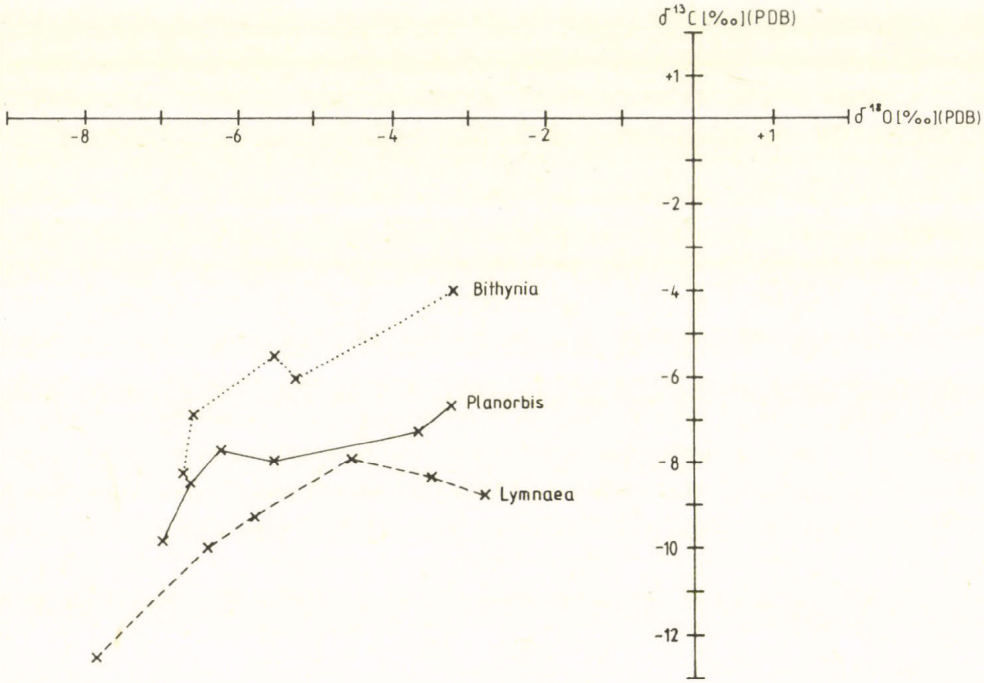

Fig. 1. Stable isotope analyses of Upper Pleistocene-Holocene gastropods from the Great Hungarian Plain

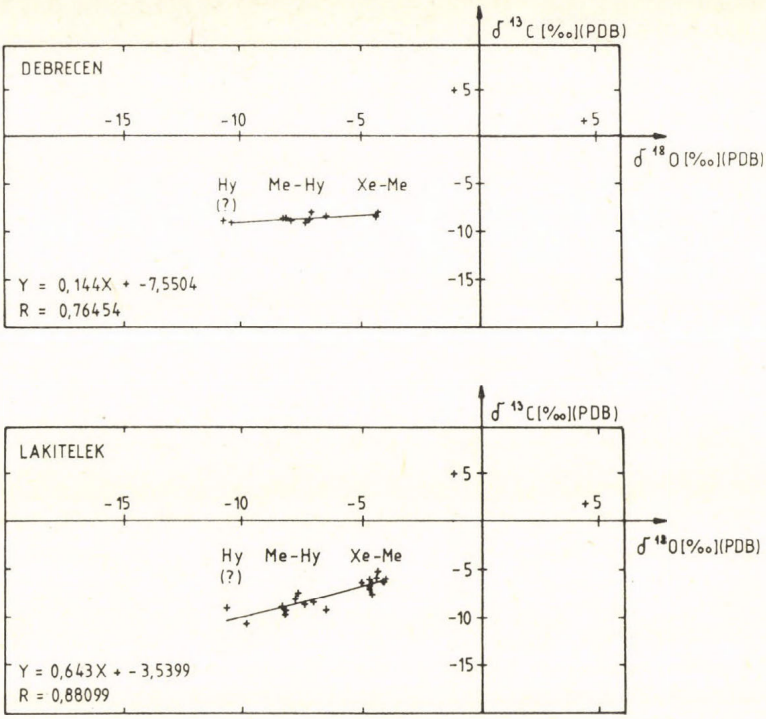

Fig. 2. Stable isotope analyses of the malacological material collected from the Upper Pleistocene heteropic facies of sections Debrecen and Lakitelek. Xe-Me = xerophilous-mesophilous species; Me-Hy = mesophilous-hygrophilous species; Hy = hygrophilous species

CLIMATIC RECONSTRUCTION FOR THE GREAT HUNGARIAN PLAIN

The investigation method described above is presented here on the instances of two sequences examined in detail used to trace Upper Pleistocene-Holocene climatic changes in the Alföld. The NE part is represented by the Debrecen Brickyard section (Fig. 4), reflecting the period between 22,000-7,000 years B.P. The following paleoclimatic changes have been registered:

25,000-22,000 years B.P.: cold, dry climate in the Great Hungarian Plain with July mean temperature about 12-14 °C. The malacofauna is dominated by *Vallonia tenuilabris*, resistant to cold and dry weather.

22,000-20,000 years B.P.: warming up with 16-17 °C July mean temperature as indicated by the dominance of the species *Pupilla muscorum*. The humic soil layer was identified as synchronous with the Dunaújváros-Tápiósüly lower paleosol (Pécsi, 1978).

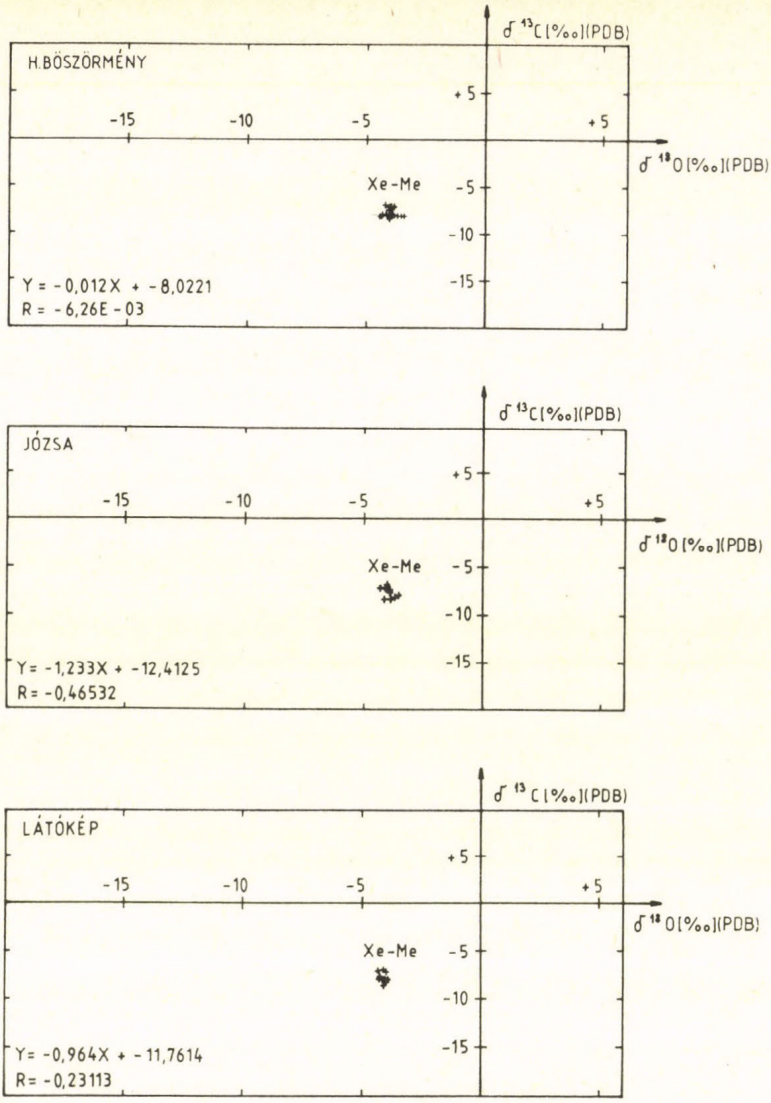

Fig. 3. Stable isotope analyses of *Pupilla muscorum* from the isochronous-isotype facies of three excavations in the Hajdúság territory

20,000-18,000 years B.P.: essential deterioration of the climate with 12 °C July mean temperature proved by the extremely high ratio of *Columella columella* and *Vallonia tenuilabris*.

18,000-16,000 years B.P.: more humid, temperate climate, indicated by the presence of the species *Vestia turgida* and *Punctum pygmaeum* (Krolopp and Sümegi, 1990). This short humid phase was followed by a gradual climatic deterioration.

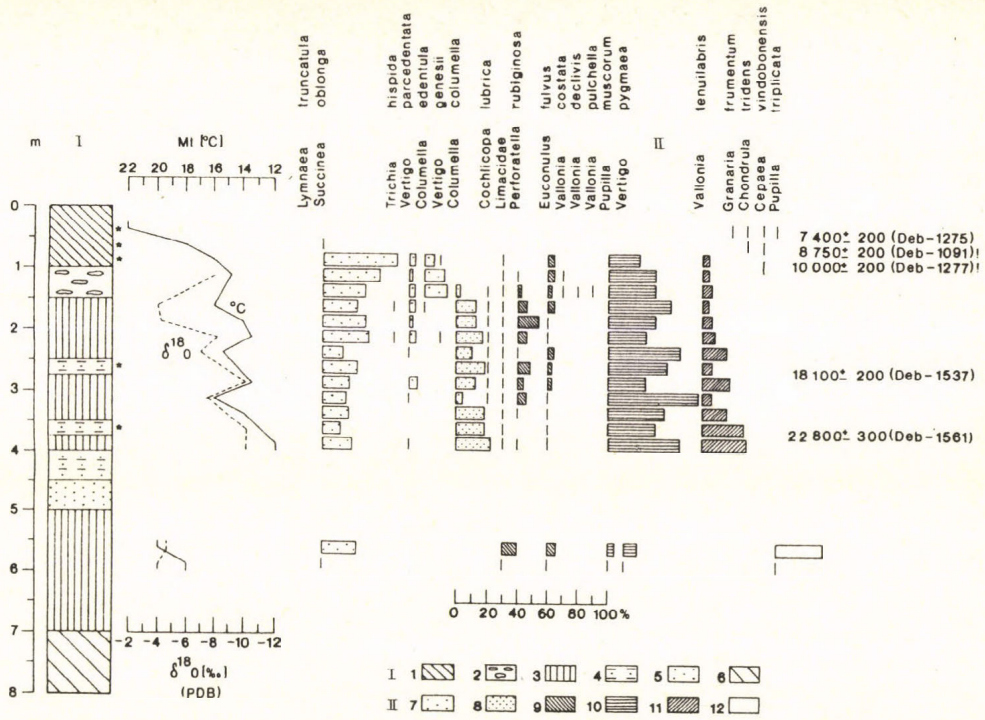

Fig. 4. Paleoclimatological analysis of the Debrecen Brickyard section. I = Genetical term for the sediments: 1 = recent soil level; 2 = layer with concretions; 3 = loess; 4 = sandy loess; 5 = eolian sand; 6 = fossil soil. II = Paleoecological groups in the Mollusca fauna: 7 = hygrophilous cold-resistant, steppe faunal elements; 8 = hygrophilous, steppe faunal elements preferring cold climate; 9 = cold-resistant, sub-hygrophilous elements requiring more extensive vegetation cover; 10 = mesophilous steppe species; 11 = elements preferring cold climate and resistant to dry climate; 12 = thermophilous steppe species resistant to dry climate; Mt = malaco-thermometer, paleoclimatic curve; * = ^{14}C sampling points; ! = ^{14}C data from the shells of *Cepaea vindobonensis*

We have to stress the dominance level of *Punctum pygmaeum*, dated at *18,000-16,000 years B.P.*. Upper Paleolithic (Gravettian) sites are abundant. The interstadial period typical of this phase can be observed both locally (Gábori-Csánk, 1978) and globally (Heusser, 1973).

16,000-14,000 years B.P.: July mean temperature could be about 13-14 °C. The species *Pupilla sterri* can be spotted for the last time in the Alföld. The dominance of cryophilous elements (*Columella columella* and *Vallonia tenuilabris*) is typical.

It is well-known, that the extent of the inland ice cover was fairly large at this time (Dryas I phase; Dreimanis and Karrow, 1972). Beginning with 14,000 years B.P., oscillations with a general rise in temperature are observed.

14,000-12,000 years B.P.: cryophilous taxa withdrawing of hygrophilous species resistent to cool climate gradully gaining dominance. In our opinion, this period can be associated with the Bölling interstadial.

12,000-10,000 years B.P.: the dominance peak of the species *Succinea oblonga* denotes a July mean temperature value of 16-17 °C.

10,000-8,500 years B.P.: July mean temperature reaches 20 °C. *Cepaea vindobonensis* appears and by 8,500-7,000 years B.P., a climate warmer and drier than today is witnessed. This climatic optimum is indicated in the Great Hungarian Plain by the dominance of *Granaria frumentum, Pupilla triplicata* and *Helicopsis striata*. During this period the youngest freshwater calcareous silt (limestone) beds of the Alföld, Balaton and Sárrét basins were formed.

CONCLUSIONS

The paleotemperatures determined by this new method are considered in global and regional contexts comparing our climatic curve with those of other authors *(Fig. 5)*.

A general agreement is found with Heusser's (1973) curve, based on palynological data calibrated with ^{14}C dates. The warming up assigned to the beginning of the Holocene as well as the Upper Pleistocene maxima and minima are clear. The differences in actual temperatures can be partly explained by the fact that the 'palyno-thermometer' of Heusser

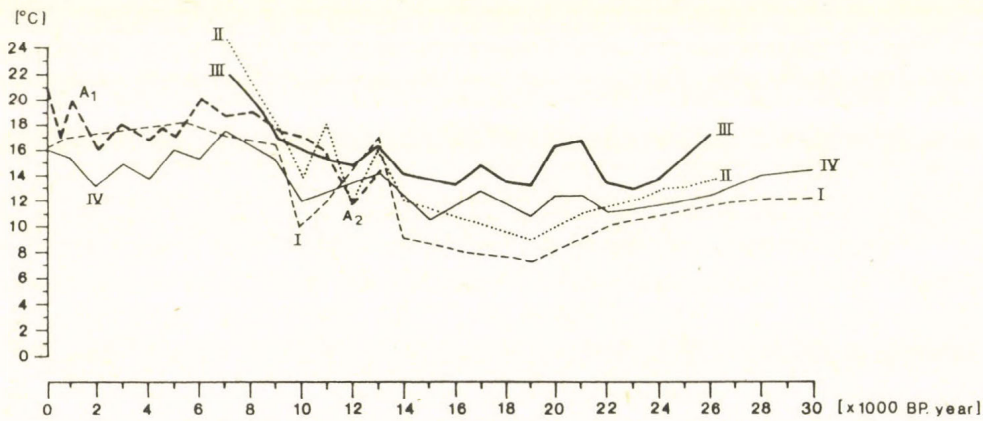

Fig. 5. Comparison of the paleoclimatic curve (malaco-thermometer) with the results of other authors. A_1 = vole-thermometer (Kordos, 1977); A_2 = vole-thermometer (Kretzoi, 1957); I = Coeloptera-thermometer (Coope, 1975); II = Pollen-thermometer (Járai-Komlódi, 1969); III = Malaco-thermometer (Sümegi, 1989); IV = Pollen-thermometer (Heusser, 1973)

was elaborated for the environs of the State of Washington in the vicinity of inland ice cover.

Our malaco-thermometer can be compared in its full range with the climatic curve of Coope (1975) based on *Coeloptera* remains from Britain, the run-off of which is very similar to that of the curve constructed by Járai-Komlódi (1969) compiles from palynological data from Hungary. This can be possibly explained by the fact that both of these works were based on the comparative study of marshes and marshy biotopes.

In the period between 30,000-14,000 years B.P., these were refuges and supply data on a special environment, therefore they are not suitable for recording detailed climatic changes. Lower temperatures can be explained by more temperate, cooler climate and, in the case of the British example, also the neighbouring ice sheet.

Authors question the validity of the 26,000-14,000 years B.P. phase on the climatic curve calibrated with ^{14}C dates constructed for the Netherlands (Zagwijn and Paepe, 1968) with a single period of cooling only (see Nilsson, 1983, p. 258). Obviously for the lowland Carpathian Basin a more differentiated climatic change is assumed.

The data of our 'malaco-thermometer' can be compared with that of the 'vole-thermometer', based on small mammal remains from the Hungarian Mountains between 13,000-7,000 years B.P. (Kretzoi, 1957; Kordos, 1977, 1981), tested by, partly, ^{14}C dates and a chronological method based on the determination of organic matter in bones (Szöőr, 1982a, 1982b). In the given phase, the two curves agree well. The differences in actual values reflect mesoclimatic deviations between lowland and hill regions.

In our opinion, the above evaluation proves the validity of the paleoclimatological method elaborated on the basis of the malacofauna as well as supports the paleoclimatological reconstruction for the Alföld region.

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ATLAS OF PALEOCLIMATES AND PALEOENVIRONMENTS OF THE NORTHERN HEMISPHERE

Late Pleistocene — Holocene

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and INQUA Commission on Loess
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CONTRIBUTIONS TO THE SEDIMENTOLOGY AND EVOLUTION HISTORY OF LAKE BALATON

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ABSTRACT

It is since 1981 that the Hungarian Geological Survey has been investigating Lake Balaton, in order to understand the development and geological history of the lake and to gather a knowledge of the diagenesis of the Holocene lacustrine calcareous muds.

As a result of geophysical and geological investigations carried out to date it can be stated that (1) Lake Balaton (recently dated to 12,000 years B.P. by radiocarbon method) was formed in the very beginning of the Holocene by the chaining of several shallow lakes with pure and cold water. As climate became warmer and more humid, the water level increased, and the barriers separating the sub-lakes gradually ceased to exist due to abrasion. The coverage of the area by water attained its maximum in the Boreal phase. Considerable ecological changes began in the Subatlantic phase when lake level dropped and the limpidity of water was deteriorating at a rapid rate; that (2) 50-70 % of the lacustrine deposits are carbonates (Mg-calcite, dolomite, calcite), mainly of allochthonous origin (inorganic segregate, products of phytoplankton metabolism, shell detritus); whereas 30-50 % is represented by siltstone, sand and clay transported by water-flows and washed in by coastal abrasion. The rate of mud deposition is 0.2-1.0 mm/year depending on subaqueous mud movement.

INTRODUCTION

Lake Balaton, the largest shallow-water lake in Central Europe is the most popular target not only for tourism and recreation but also for limnological and geological investigations. The natural ageing of the lake as well as the environmental problems caused by tourism and intensive agriculture have been studied since the turn of the 19th century (Lóczy, 1913; Zólyomi, 1952, 1987; Bulla, 1958; Szesztay et al., 1966; Bendefy and Nagy, 1969; Müller and Wagner, 1978; Marosi and Szilárd, 1981; Herodek and Máté, 1984).

It is since 1965 that the Hungarian Geological Survey has been carrying out a complex mapping in the lake catchment area (at the scale of 1:50,000) and on the shoreline zone of the lake (at the scale of 1:10,000), and an actual geological

1984; Cserny 1987; Nagy-Bodor, 1988; Cserny and Corrada, 1989, 1990; Bodor, 1987).

OBJECTIVES

The aims of the actual geological investigation launched in 1981 and expected to be completed soon are (1) to understand the development and geological history of the lake, and (2) to have a better knowledge of present-day lacustrine sedimentary processes, sedimentary features and the diagenesis of calcareous muds. The three stages of research completed are:

— *Stage 1* (to 1986): a total of 17 boreholes were drilled into the lake-bed and the results from the laboratory test of sequences taken from these boreholes were studied;

— *Stage 2* (1987-1989): the principal aim of investigation was to determine the thickness of loose lacustrine mud by continuous geophysical (seismo-acoustical and echograph-based) logging. As a result of the evaluation of reflexion logs having a total length of approx. 370 km and covering the whole lake, an isobath map of the loose mud in the lake and a seismo-stratigraphic-tectonic map of the basement have been completed at the scale of 1:50,000.

— *Stage 3* (now in progress): further 16 boreholes were drilled in 1989 in order to compile the geological map of Lake Balaton and to outline the geological history and the ecological and climatological conditions of the lake and its environs. The evaluation is in progress.

A model for the investigation is summarized in *Figure 1*, whereas *Figure 2* shows the traces of geophysical logs taken from the lake, and the localities of the boreholes. This paper gives a brief description of shallow-water Holocene deposits and, based upon data from borehole No. 24, a summary of the geological history of the lake.

HOLOCENE DEPOSITS

The Holocene mud in Lake Balaton has an average thickness of 5 m. The upper 0.5-1.5 m of this mud is in colloidal state. As shown by our measurements, the maximum mud thickness was observed at the mouth of river Zala being of greatest importance for the water recharge of the lake, whereas no mud deposition was observed in the Tihany strait. As shown by boreholes, in lacustrine deposits a few centimetres of gravel is overlain — in some cases — by a few centimetres thick peat, then by a throughout uniform calcareous mud. As for its granulometric composition, it is argillaceous siltstone with a carbonate content of 50-70 %, so actually calcareous mud is concerned. These sediments were deposited with a rather high primary porosity and the constituting carbonate

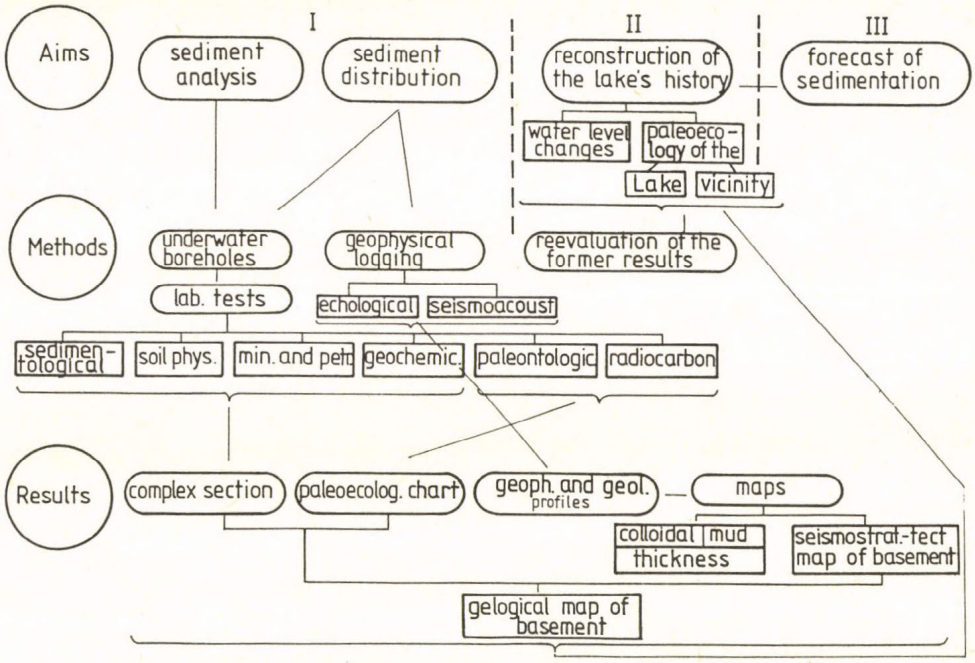

Fig. 1. Geological investigations of Lake Balaton (Hungary)

minerals are chemically rather unstable. By X-ray and DTA tests magnesium-bearing calcite, dolomite and calcite were detected. In the magnesium-bearing calcite the Ca/Mg ratio decreases with depth. The carbonate is mainly inorganic segregate and it is subordinately due to the accessorial processes of phytoplanktonic metabolism and to the remains of shell detritus. The peat encountered at the bottom of the Holocene complex is ^{14}C dated as 12,000-14,000 years B.P. With the average mud thickness values taken into consideration, the rate of mud deposition is 0.4 mm/year.

AN OUTLINED GEOLOGICAL HISTORY OF THE LAKE

To have a better understanding of the geological history of Lake Balaton, the sequence from borehole Tó-24 was subject not only to geological studies but also to palynological (Nagy-Bodor, 1988), Diatoma and Ostracoda analyses.

The aim of palynological examinations was to determine when the lake developed, to increase knowledge of the evolution history of vegetation and climate, and to identify ecological conditions and changes in water level. Diatoma analyses have allowed us to

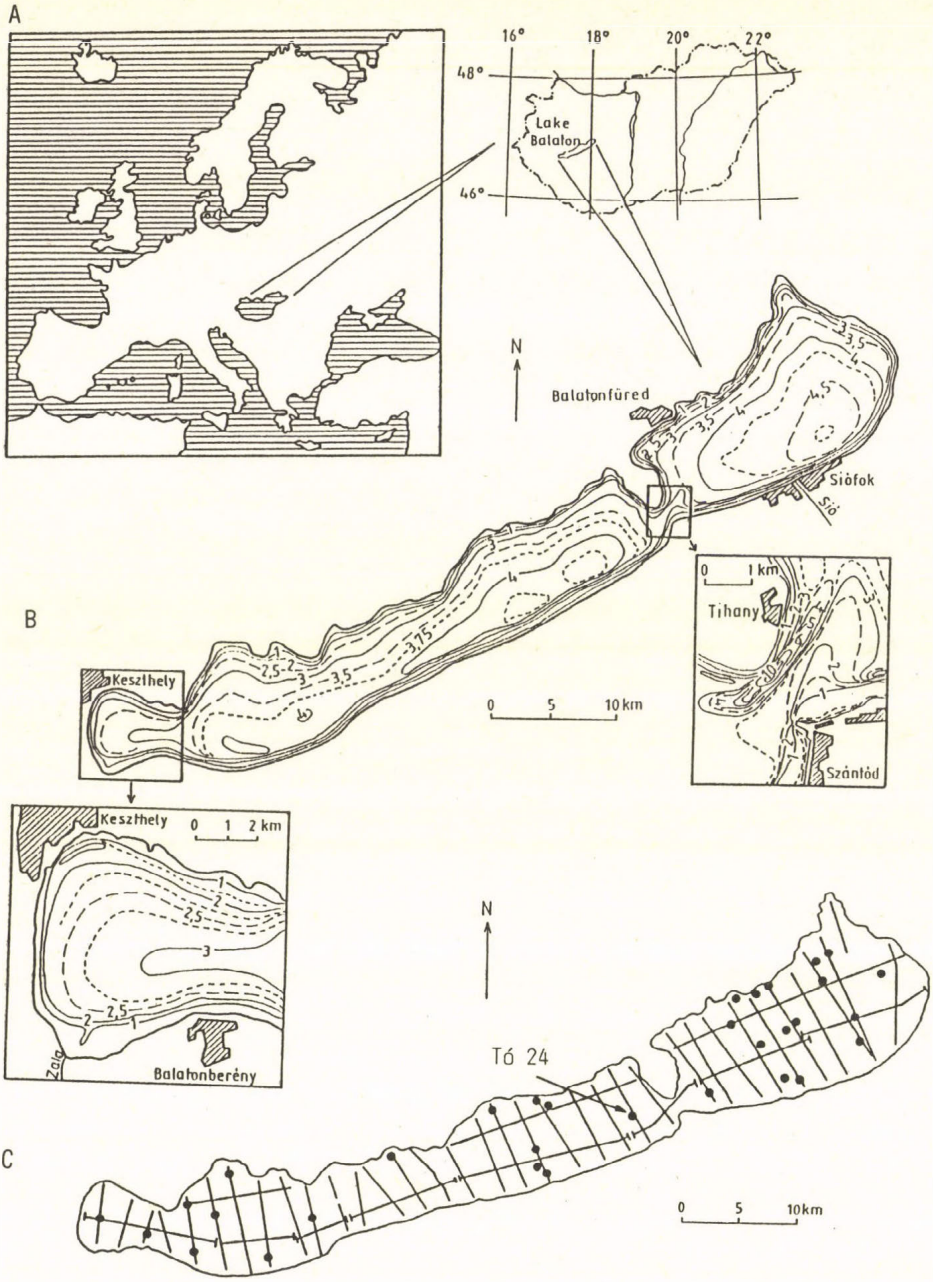

Fig. 2. Geophysical survey of Lake Balaton. A = geographical position of Lake Balaton; B = the isobathic map of the lake; C = localities of boreholes and geophysical profiles

draw conclusions mainly on the oxygen supply, limpidity and trophism of lake water, whereas the study of ostracods has allowed us to trace the changes in dissolved salt content.

In the borehole, Holocene lacustrine deposits overlay, with unconformity, an Upper Pannonian basement. No Pleistocene deposits have been observed.

The sporomorph and ostracodal faunas of the Upper Pannonian formations observed in borehole Tó-24 are characterized by a great number of species and a small number of specimens. As for Holocene deposits, this ratio is reverse. Diatoma flora has only been identified from Holocene formations. Changes in the ecological factors have revealed well-defined ecological zones for each vegetation phase.

THE BASEMENT

In the environs of the borehole the Pannonian lake was shallow. This is confirmed by the widespread occurrence of lacustrine swampy-marshy vegetation (*Myrica*, Taxodiaceae, *Typha*). At the same time, the quantitative changes of sporomorphs of plants pointing to shallow (1-3 m deep) or deeper (3-6 m deep) water exhibit frequent fluctuations which is in correspondence with the changes in the water level of the Pannonian inland lake and in the lake-shore forests.

The Ostracoda fauna shows mesohaline (*Cyprideis triangulata, Mediocytherideis kleinae*) to oligohaline (*Candona/Candona/neglecta*) conditions. The Pannonian sedimentary basin had a relatively high chloride content.

The vegetation in the environs of the lake is characterized by an assemblage consisting of *Tsuga-Cedripites-Zelkova* and *Pterocarya* which points to a climate warmer and more humid than at present.

HOLOCENE PHASES

Pine-birch (Preboreal) vegetation phase

A *Pinus-Betula* vegetation pointing to a preboreal climate has been observed in the Holocene deposits overlying, with unconformity, the Upper Pannonian formations. It was during this more humid period that the "embryonic" sub-lakes, from which subsequently the lake was formed, developed.

The Holocene flora and fauna in the log reflects a sharp change, as compared to the Pannonian.

Brackish planktonic organisms (*Paleoperidinum sp., Gonyaula sp.*) are missing

from the sporomorph assemblage, and species preferring colder climate (*Pinus-Betula*, and *Candona/ Candona/neglecta, C.C. candina, Cytherissa lacustris*) occur. In addition, ostracod species existing since the Pleistocene (*Cypridopsis vidua, Candona/Candona/hyalina*) occur. The existence of this colder period is also confirmed by species preferring oxygen-rich, cold water and proliferating in the diatom assemblage (*Melosira italica, M. subarctica*).

The *Pinus-Betula* phase is subdivided into three ecostratigraphic zones (A, B, C) which are as follows:

Zone A: The development of the lake is in initial state. Based upon the water depth assessed on the basis of the morphology of the basement and the flora and fauna of the sediment core, a few sub-lakes are presumed to have developed. These sub-lakes are likely to have had a large surface but small depth which is indicated not only by the pollen grains of shallow-water plants (*Sparganium, Stratiotes*) but also by the appearing diatom flora, the Chrysophyta cysts and the enrichment of Phytolitharia remains. In these shallow lakes the oxibiont diatoms of the algal vegetation (*Epithemia sorex*) became wide-spread but saprobionts did not yet appear. From this zone onwards, the ostracodal fauna does not indicate any change in salinity.

Gallery forests are still missing from the dry environment. In the catchment basin the pollen of deciduous birch forest and a more distant, hillside pine forest are accumulated.

Zone B: The water level in each sub-lake rapidly increases and the intensively agitated water gradually eliminates the barriers between the sub-lakes. In the aquatic environment the vegetation becomes enriched. At the same time, the amount of oxibionts decreases and saprobiont elements *(Cocconeis diminuta*) appear in a considerable amount. In the mud the broken remains of siliceous algae with thick shell are also enriched, pointing to the fact that at that time lake water was intensively agitated. In addition, the dissolved, thin shells indicate alkaline pH of the water.

During this phase gallery forests develop and mixed deciduous forest became clearly predominant in the environs of the lake.

Zone C: In the beginning of this zone the lake basin is gradually filled and by the end the geographical extent and water depth attain their maxima. The relative amount of plants preferring deeper water (*Potamogeton, Myriophyllum*) increases accordingly. This is also justified by the frequency of benthic forms (*Diploneis elliptica, Gomphonema minutum*) in the algal vegetation. The amount of oxibionts and saprobionts also increases. In the surrounding dried-up areas gallery forest and the deciduous forest became permanent, and the extent of pine forest attained a maximum.

Hazel (Boreal) vegetation phase

The subsequent steppe phase is characterized by the dominance of *Corylus*.

According to Zólyomi (1952), a local climate milder than the boreal climate in Europe was indicated not only by predominant *Corylus* but also by the more frequent occurrence of *Tilia* and *Quercus* (Bodor, 1986).

Zone D: The maximum of water level slightly drops in the middle of the zone and is restored by the end. This is also justified by a transitional decrease in the amount of deep-water plants. The proportion of oxibionts and saprobionts remains unchanged in the algal vegetation. The temporary shrinkage of the gallery forest may also be connected with the decrease in water surface.

Oak (Atlantic) vegetation phase

Zone E: Water level starts to fall gradually. The beginning of the zone is characterized by the frequent occurrence of plants preferring deep water but later the relative amount of shallow-water (*Sparganium, Typha*) and shoreline plants increases. At the same time, gallery forests and hillside pine forests also became widespread. The upper boundary of this zone is indicated by the repeated and considerable decrease in the extent of the gallery forest and by the favourable spreading of oxibionts in the algal vegetation of the lake.

Oak-beech (Subboreal) vegetation phase

Zone F: Water level continues to lower, then becomes constant at approximately the present-day level, by the time the upper third of the zone is reached. The purity of water is of medium level and the amount of aquatic plants of higher order decreases. The proportion of oxibiont algal organisms considerably increases whereas the amount of saprobiont remains constant.

In the environs of the lake, gallery forests expand whereas the hillside pine forests thin out toward the end of this zone. In more distant areas there is no remarkable change in the extent of mixed deciduous forests.

Beech (Subatlantic) vegetation phase

Zone G: After an initial drop in lake water level shallow-water conditions become characteristic, as the amount of aquatic plants preferring deeper water decreases, and in addition to permanent shallow-water ones (*Sparganium, Rorippa*) the amount of shoreline elements also increases. This is also justified by *Epithenia argus*, a siliceous alga indicating shallow-water conditions. The limpidity of water becomes deteriorated which is reflected by a decrease in the amount of oxibiont algae and a considerable enrichment of saprobiont elements.

By the end of this zone the species representing the gallery forest withdraw. The assemblage of mixed deciduous forest remains unchanged whereas hillside forests reach their present-day extension.

Zone H: is characterized by a final, considerable rise in water level and a rapid growth of the amount of saprobionts and withdrawal of oxibionts (e.g., *Epithemia sorex*). This points to the fact that the higher water level involves the accumulation of organogenic sapropel. Aquatic plants of higher order become more abundant and at the same time the *Pediastrum* green alga living on the water surface also spreads. In the environs of the lake a rapid expansion of gallery forests and a temporary one of hillside forests can be observed.

Zone I: Water level considerably falls. Oxibiont benthic diatoms are completely absent and saprobiont elements (*Cocconeis diminuta*) dominate. As compared to the previous zone, no considerable change is observed in the vegetation of the surrounding dry areas.

Vegetation phase of cultivated forests (Subatlantic)

Zone J: Human activity also influences changes in water level, limpidity and environmental vegetation of the lake.

During the earlier one-third of the zone the water level falls below today's, then rises due to planned regulation, and stabilizes at today's level during the upper one-third of the zone. The limpidity of water deteriorates, first at a low rate but later more rapidly. Shoreline vegetation (*Dryopteris sp., Chenopodiaceae sp.*) expands. *Fagus* continues to dominate the environment of the lake. Nevertheless, during the upper one-third of the zone both *Fagus* and the hillside pine forests become remarkably less dense, due to the advance of cultivated plants and fields crops.

CONCLUSIONS

1. Lake Balaton was formed in the very beginning of the Holocene and is radiocarbon dated as old as approx. 12,000 years B.P. In the beginning, a chain of several, shallow lakes with pure and cold water was formed. As temperature rose and climate became more humid, the water level increased and the barriers separating the sub-lakes were gradually obliterated by abrasion. Water surface reached a maximum in the *Corylus* vegetation phase, in the beginning and at the end of Zone D. In the oak and oak-beech vegetation phases a minor but definite decrease in water level was detected. Up to this time the water in the lake had been limpid and gallery forests were characteristic of the environs of the lake as a function of the climate.

2. In zone G during the beech phase major ecological changes took place. The water level of the lake dropped considerably and the limpidity of water became proportionally deteriorated. Human impact on the lake can be observed in zone J when the water level fell even below today's level and its saprobity and trophity was growing at an increasing rate.

3. During the development of Lake Balaton the degree and type of sedimentation of lacustrine deposits varied as a function of the depth of the lake, the extension of water surface, water quality, climate, and the vegetation cover shoreline areas. The proportion of carbonate (Mg-calcite, dolomite, calcite) is as high as 50-70 % in the composition of the lacustrine deposit. The carbonate is mainly of allochthonous (inorganic segregate, products of phytoplanktonic metabolism, shell detritus), and subordinately of autochthonous (load of water-flows, dust) origin. The remaining 30-50 % of the mud is siltstone, sand and clay transported by streams and washed in by coastal abrasion. The rate of mud deposition varied from 0.2-1.0 mm/year but this was largely influenced by the effect of subaqueous streams on the mud movement.

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RADIOCARBON AGE OF THE FORMATION IN THE PAKS-SZEKSZÁRD DEPRESSION

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ABSTRACT

The FTV Consulting-Engineering Company has conducted a comprehensive survey of the Paks-Szekszárd area. The radiocarbon dating of driftwood samples collected from boreholes — performed in the Institute of Nuclear Research — also proves the different ages of the fluviatile deposits of the Paks-Szekszárd depression. Therefore, along this section subsidence variable both in time and in space can be assumed. Of special interest is the very young age of the Szekszárd sample. It indicates that the river appeared at the foot of the Tolna Hills only about 10,000 years ago. The very intensive Holocene subsidence which drew the Danube to its present course still continues in this area.

INTRODUCTION

South of Budapest down to the state border smaller depressions were generated along the Danube due to the young (Upper Pleistocene to Holocene) movements (Ádám et al., 1955; Rónai, 1964). They were filled by the Danube with granular alluvium of 15-50 m thickness (Erdélyi, 1955; Pécsi, 1959). Among these depressions the Paks-Szekszárd one is the largest *(Fig. 1)*.

Recently, the FTV Consulting-Engineering Co. has carried out explorations of different aims (engineering-geological, soil mechanical, water acquisition) in this depression (Altnőder et al., 1988) and in certain boreholes charcoal was discovered in the fluviatile sediments (in the area of the Paks Nuclear Power Plant, in the borehole No. 881 at 20.5 m; at Szekszárd in the borehole 2/1 between 22 and 23 m; *Fig. 2*). These explored charcoal pieces are of great importance since their dating is suitable to correct the data obtained formerly by other methods.

The radiocarbon dating of the charcoal samples was carried out in the Institute of Nuclear Research.

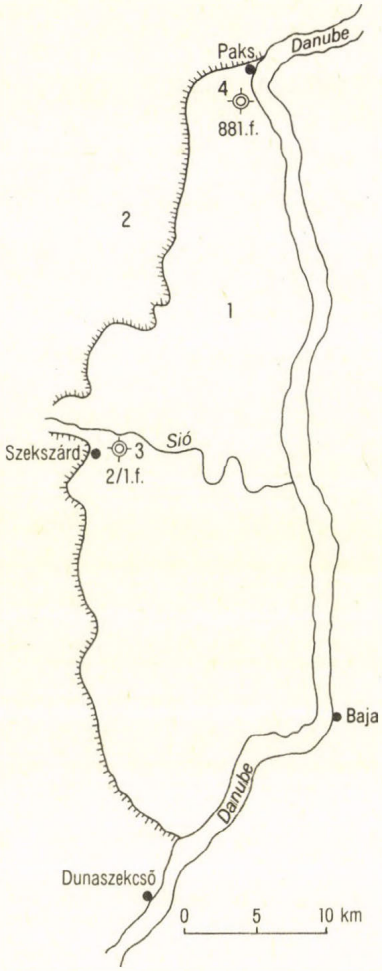

Fig. 1. Sketch of the Paks-Szekszárd depression with localities of driftwood finds. 1 = the Paks-Szekszárd depression; 2 = bluffs bordering on the depression; 3 = site of the Szekszárd borehole; 4 = site of the Paks borehole

MEASUREMENT METHOD

In the case of the charcoal samples the AAA (acid-alkali-acid) treatment was used. Subsequent to ultrasonic and physical cleaning the samples were treated by 4% hydrochloric acid, subsequently by 4 % sodium hydroxide for 24-24 hours at 80 °C. To eliminate the base traces 4 % hydrochloric acid was used again. After treatment the samples were dried.

Samples were ignited to carbon dioxide in oxygen flow and the carbon content was converted to methane (Csongor et al., 1982). Activity measurements were carried out with a low background proportional counter system (Hertelendi et al., 1989).

The age expressed in BP was calculated on the basis of the formula

$$t = -8033 \ln \frac{A_{SN}\ (in 1950)}{A_{ON}\ (in 1950)}$$

using Libby's half-life ($t_{1/2}$ = 5568 years), and where A_{ON} is the corrected activity of the NBS standard, while A_{SN} is the corrected sample activity. In case of the sample the corrected value was derived from the A_S measured activity from the formula

$$A_{SN} = A_S \left(1 - \frac{2\,(25 + \delta^{13}C)}{1000}\right)$$

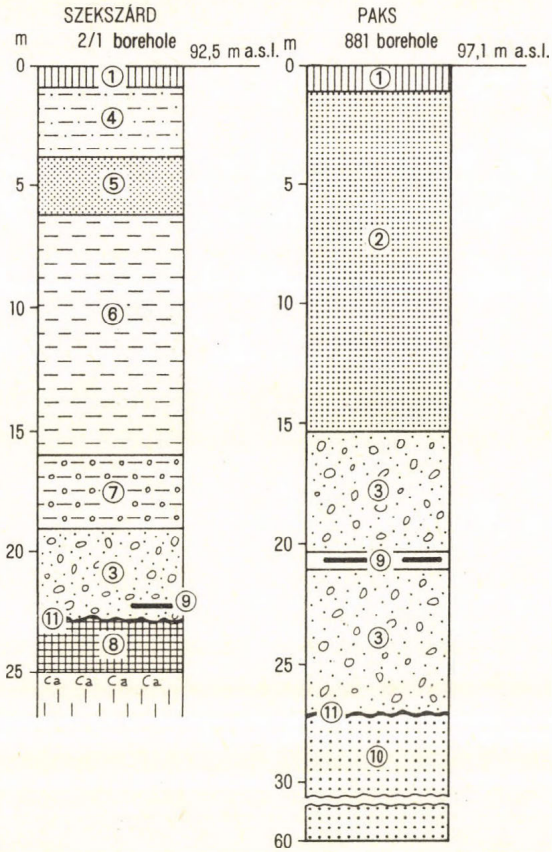

Fig. 2. Stratigraphy of boreholes with driftwood finds. 1 = soil; 2 = medium-grain sand; 3 = sandy gravel; 4 = sandy silt; 5 = fine sand; 6 = silt; 7 = gravelly silt; 8 = Upper Pliocene red clay; 9 = site of driftwood; 10 = Upper Pannonian sand; 11 = erosional unconformity

The $\delta^{13}C$ values were measured by mass spectrometer and were related to the PDB standard:

$$\delta^{13}C = \frac{\left(\frac{^{13}C}{^{12}C}\right)_{samp} - \left(\frac{^{13}C}{^{12}C}\right)_{PDB}}{\left(\frac{^{13}C}{^{12}C}\right)_{PDB}}$$

The results obtained in this way are as follows:

Code No.	Sample	$\delta^{13}C_{PDB}$ [‰]	Radiocarbon age years B.P.
Deb-953	Szekszárd borehole 2/1 at 23.3 m	-27.15	10,880 ± 150
Deb-950	Paks borehole 881 at 20.5 m	-25.59	>40,000

Since the age of the Paks sample exceeds 40,000 years B.P., concerning the granular Danube sediments, it can be stated that these are older than 40,000 years, too. The age of the Szekszárd sample on the other hand proves that the Danube appeared at the foot of the Tolna Hills in the Late Pleistocene or Early Holocene only.

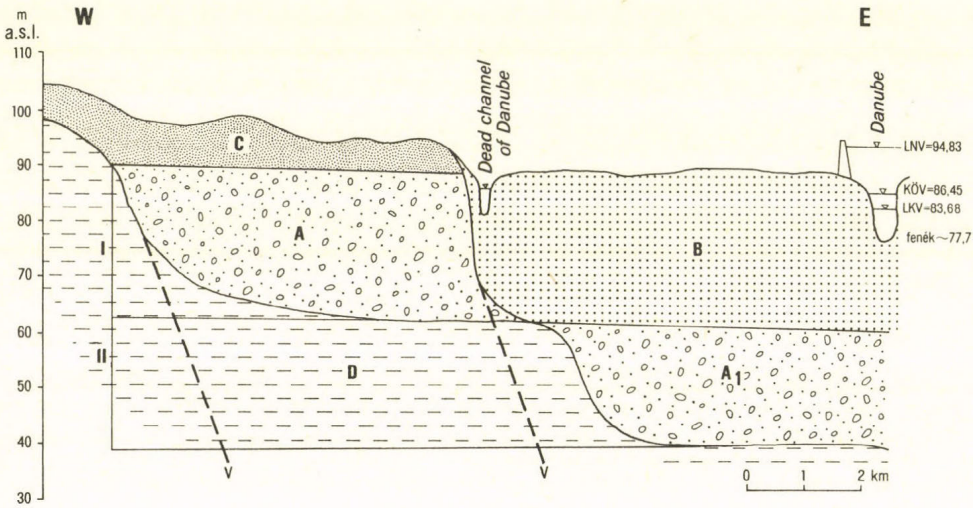

Fig. 3. Evolution of the Paks-Szekszárd depression. I = first phase of subsidence; II = second phase of subsidence; A = fluviatile deposits accumulated during the first phase and not involved in the second phase; A_1 = fluviatile deposits accumulated during the first phase and further subsided in the second phase; B = fluviatile deposits accumulated during the second phase of subsidence; C = blown sand; D = underlying Pliocene and Lower Pleistocene deposits (clay, silt, and sand); V = fault; LNV = HWL; LK KÖV; MWL; LKW = LWL; fenék = bottom of the river channel

CONCLUSIONS

1. The radiocarbon dating results support the former statements that the Danube has accumulated sediments in the area since the Upper Pleistocene (Pécsi, 1959; Borsy, 1990).

2. The radiocarbon ages also prove that the layers of the fluviatile sedimentary sequence of the Paks-Szekszárd depression are of different age. Thus, periodical subsidences with different areal extension can be assumed *(Fig. 3)*.

3. The Szekszárd sample of very young age is of special interest. This value indicates that the Danube occurred at the foot of the Tolna Hills only about 10,000 years ago. Consequently, in this area intense subsidence has proceeded to our days (Joó et al., 1985). The occurrence and sediment accumulating activity of the Danube can be derived only from these young Transdanubian movements.

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Studies in Geography in Hungary, 26 Akadémiai Kiadó, Budapest 1991, pp. 91—97.

PLEISTOCENE MARKER HORIZON IN CARPATHIAN BASIN LOESS: THE BAG TEPHRA

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ABSTRACT

Several aspects of a tephra layer present in ten loess sections in Hungary and in Czecho-Slovakia were investigated: grain size, mafic mineral suites, and chemical composition of bulk samples as well as clinopyroxenes. The tephra is identical in all localities, and the name „Bag Tephra" was given to it. This tephra allows us to propose a new stratigraphical correlation of the investigated loess sections. The tephra fall occurred after the Mindel/Riss Interglaciation. Since the closest Pleistocene volcanic fields are situated very far from the investigated localities (more than 500 km), the Bag Tephra can be used as a widespread stratigraphical marker in Central Europe. The currently available data show that the relevant volcano should be located in the East Eifel volcanic field (Germany), rather than in the Apennine (Italy) or in the East Carpathian Mountains (Romania). The correlation with any volcano in the French Central Massif or in the Aegean Sea (Greece) is excluded.

INTRODUCTION

Intercalations of volcanic ash in the Quaternary loesses of Hungary were first described in the fifties (Kriván, 1957; Kriván and Rózsavölgyi, 1964). Since then both loess stratigraphy and tephro stratigraphy have made much progress manifest in approaches as well as methods. Our investigations are aimed at re-examining the mineral and grain size compositions and the geochemical properties of the tephra and drawing conclusions for its origin and age to be used in the divisions of loess and Quaternary sequences.

SAMPLING SITES

Samples have been collected and analysed from all the tephra occurrences described by Kriván, from the laminae of volcanic origin known from literature in Slovakia (at Komjatice — Vaškovky and Karolusova, 1969), in N. Hungary (at Pásztó — Székely, 1960) and tephras to date unrecognized in loess profiles (at Basaharc, Pócsa and Kökény) have also been studied.

The about a dozen profiles with volcanic material are usually located on the E or SE facing valley sides, sheltered from western winds. The Pásztó and Pócsa sites are exceptions in this respect. Tephra thickness ranges from 1-5 cm in the profiles, but it is present at Basaharc only in nests. It is of yellowish brown or grey colour. The tephra horizons lie below paleosol complexes ranging in number from one to four *(Fig. 1)*.

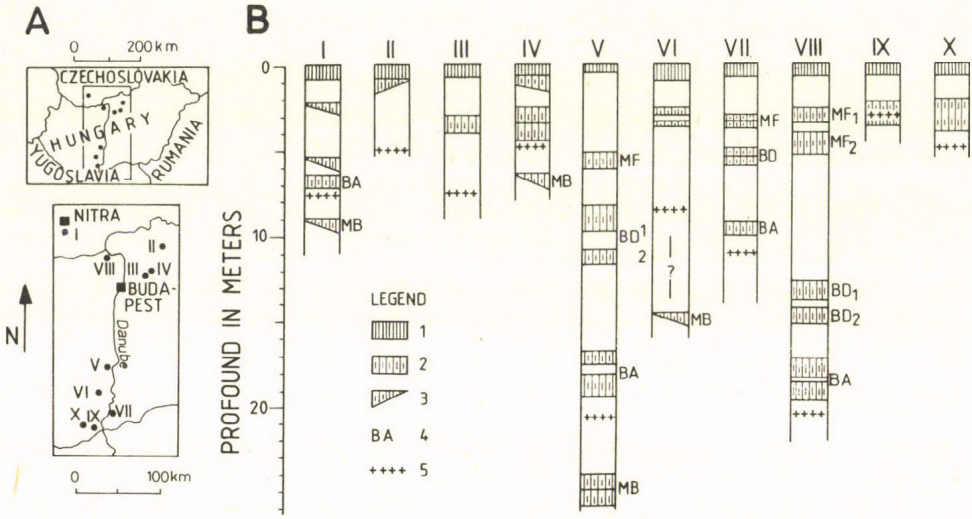

Fig. 1. Location of loess profiles and stratigraphic position of the tephra at each site. I = Komjatice; II = Pásztó; III = Bag; IV = Hévízgyörk; V = Paks; VI = Sióagárd; VII = Dunaszekcső; VIII = Basaharc; IX = Pócsa; X = Kökény; 1 = recent soil; 2 = paleosol; 3 = paleosol observed in other parts of the profile; 4 = paleosol nomenclature after Pécsi (1979); 5 = tephra

METHODS

During the *mineralogical* investigations under the polarization microscope the composition of the heavy mineral fraction above 0.063 mm grain size was determined. In every sample clinopyroxenes (monocline pyroxenes) constitute more than 90 % of

heavy minerals of volcanic origin. In addition, some per cent of brown amphibole, titanite (sphene) and olivine also occur. Apatite, biotite and green amphibole are found both in tephra and in loess and, therefore, they are grouped with minerals of non-volcanic origin.

On 10-12 *pyroxene* grains from each sample *chemical analysis* was performed and their types were determined by Morimoto's (1988) classification. On the Ca-Mg-Fe diagram the points representing the samples are located along the line, which divides diopside and unusual pyroxenes *(Fig. 2)*. All pyroxenes show high calcium and aluminium contents, but significant differences between the individual samples are not found.

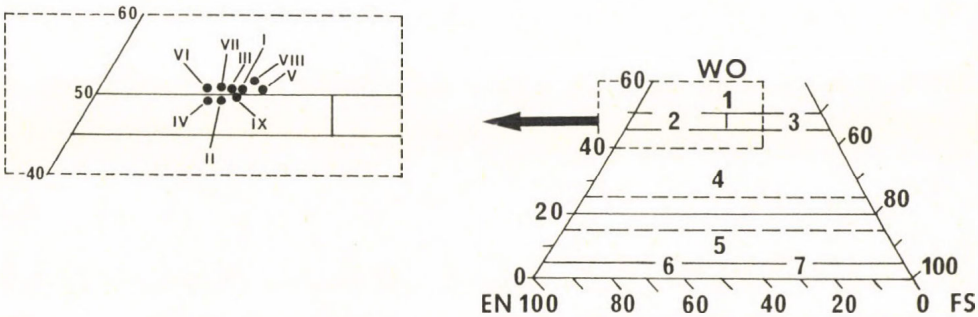

Fig. 2. Determination of pyroxenes after Morimoto's classification (1988) WO = wollastonite; EN = enstatite; FS = ferrosilite; 1 = unusual pyroxene; 2-3 = diopside-hedenbergite series; 3 = augite; 4 = pigeonite; 5-6 = enstatite-ferrosilite series, Numbers I-IX = localities according to Figure 1

During the investigation of the *general chemical composition of the tephra* several corrections had to be applied in order to eliminate the uncertainties due to alterations by weathering and leaching and to identify the type of the original rock (Juvigné et al., 1991, in press). The following conclusions can be made: the tephra could originally be of basic to intermediate nature (SiO_2 53 %), rich in iron (7 %), exceptionally rich in aluminium (22 %) and probably poor in magnesium (2.4 %). These properties contradict the definitely differentiated character of its magma.

The size of pyroxene grains, the grain size composition of samples were also determined under the microscope and grain size and its distribution was found similar in all samples. In comparison with samples from recent tuff falls it can be claimed that the tephra material analysed by us must be some hundreds of or one or two thousand kilometres distance away from the site of eruption.

DISCUSSION

The analogies in the mineral composition, the chemical properties of the pyroxenes and of the whole tephra as well as in the grain size distribution curves suggest that the studied tephra laminae are associated with *a single volcanic eruption* and, consequently, they are mentioned under the collective name of *Bag Tephra.*

During the period in question (end of Middle Pleistocene to Upper Pleistocene) the following active volcanic regions can be taken into account as source areas of the tephra: the Massif Central, the Apennines, the Aegean arc, the E-Carpathians and the Eifel. Unfortunately, there is no data in literature on localities with tephra in Quaternary deposits between the Carpathian Basin and some of the hypothetical source areas. Therefore, only the results from mineralogical, grain size composition and geochemical investigations serve as evidence to our statements.The volcanoes of the Massif Central and the Apennines supplied acidic (Cantagrel and Baubron, 1983) or semiacidic lavas (Pichler, 1970a,b), and this does not correspond to the low SiO_2 content of the Bag Tephra. The materials from the Aegean volcanoes belong to the calc-alkaline series (Fyticas et al., 1984) with usually higher SiO_2 contents (55-73 %) and much lower Al_2O_3 contents than in the Bag Tephra; consequently, they are probably unrelated to each other. The volcanoes of the E-Carpathians lie closest to the localities studied, but they mostly produced andesite, and also differ in chemical composition from the Bag Tephra (Peltz et al., 1973) and the relatively old age of the last eruptions (Peltz et al., 1987) also contradict correlation.

The Quaternary volcanic material from the Eifel is of intermediate to basic character (Simon, 1969; Frechen, 1976; Viereck, 1984; Schmincke et al., 1983). The thick tephra horizons at Rieden deserve special attention in spite of the fact that their global chemical composition is slightly different from that of the Bag Tephra (which can be explained by the inmixture of materials of unknown quantity and quality during deposition or by subsequent weathering), but in the chemical composition of their pyroxenes (Juvigné and Seidenschwann, 1989) they resemble to the samples taken in the Carpathian Basin. Another important consideration is that several tephra layers were observed (Bibus, 1974, 1976, 1980; Seidenschwann and Juvigné, 1986; Juvigné and Seidenschwann, 1989) in the loess profiles of Central and S-Germany older than Upper Pleistocene at several hundred kilometres distance from the Eifel in S, SE directions. They lie on the route of dust and ash clouds travelling from the Eifel towards Central Europe *(Fig. 3).* To our present knowledge there is no basic contradiction between the properties of the Rieden Tephra, Volcanic Eifel, and of the Bag Tephra.

The *stratigraphic position* of the Bag Tephra was reliably identified in the Paks profile, where it occurs in the loess packet between the paleosols "Basaharc Lower" (BA) and the underlying "Mende Base" (MB) (Pécsi et al., 1977). Various dating techniques have been applied to determine the ages of paleosols below and above the tephra: first of all TL dating (Borsy et al., 1980; Butrym and Maruszczak, 1984; Wintle and Packman,

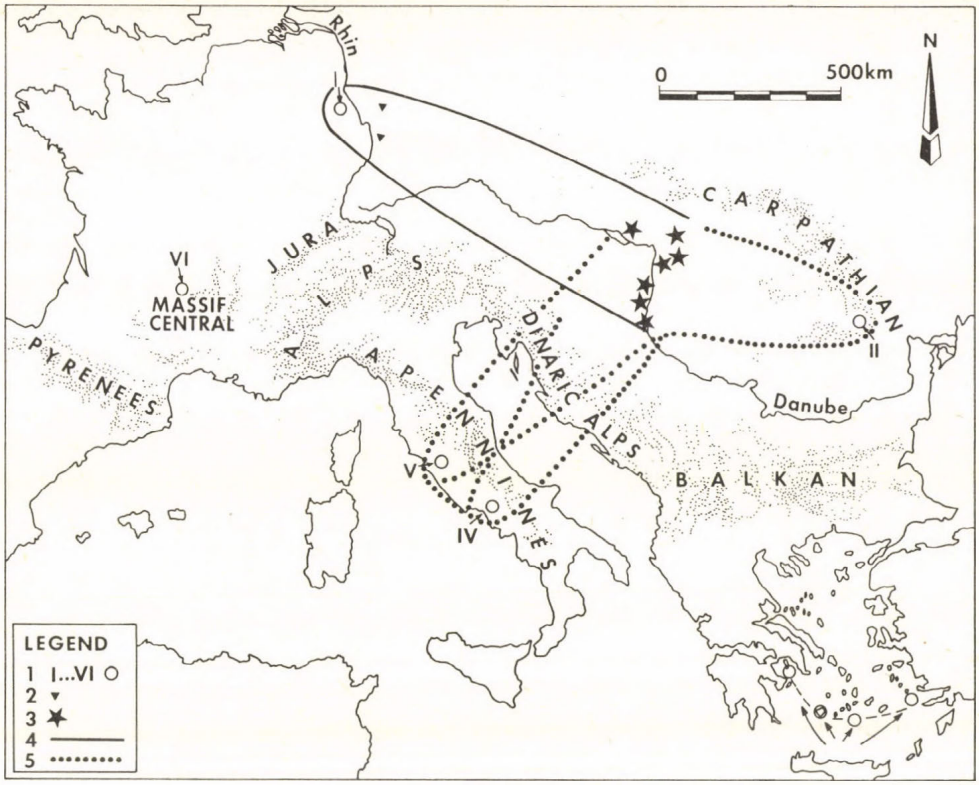

Fig. 3. Possible origins of Bag Tephra. 1 = location of explosive volcanism within the time range 730,000-380,000 years; 2 = distal occurrences of tephras from the East Eifel; 3 = investigated localities; 4 = most probable plume; 5 = less probable plumes

1988) and paleomagnetic measurements and their correlation with international time-scales (Guiot et al., 1989; Kukla, 1977; Shackleton and Opdyke, 1976). The resulting age for the intercalation of volcanic origin — on a wide range and with contradictions — can be placed between 470,000-60,000 years B.P. In our opinion the time-scale used in Hungarian loess chronology requires modification as several, yet unpublished measurements and our own data indicate. It seems obvious that the paleosol MB below the tephra represents the Mindel/Riss Interglacial. It follows from this that the tephra fall may have taken place in Early Riss, as it was described by Kriván and Rózsavölgyi (1962). The same is suggested from the date of the only tephra described from abroad (Komjatice, S. Slovakia) was also fixed immediately above the Mindel/Riss boundary (Vaškovsky, 1977). The age of the Rieden Tephra, the closest related to the Bag Tephra mineralogically-geochemically — is estimated at 400,000 years. The most probable *absolute age of the tephra* can be — according to the above considerations — in the vicinity of

300,000-400,000 years B.P., with the remark that the K/Ar isotope measurements under way on pyroxene crystals will allow more precise dating in the near future.

CONCLUSIONS

In the Carpathian Basin a tephra horizon of uniform properties and great geographical extension occurs in loess profiles and probably dates to Early Riss glaciation. Through its mineralogical and geochemical features it is bound to an intermediate basic magma rich in aluminium the major mafic mineral of which is a clinopyroxene rich in aluminium and calcium. This tephra horizon represents a new stratigraphic relationship between the loess profiles in S. Slovakia and Hungary. At the same time, it is of great significance in loess research in Hungary, since it is suitable to function as a marker layer in all the sections where the lack of some Middle or Upper Pleistocene paleosols may cause uncertainty in stratigraphy.

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Márton Pécsi

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WINTER, LATE-WINTER EROSION PROCESSES AND FEATURES IN A LOESS REGION

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ABSTRACT

Erosion on the agricultural land of Hungary is associated with showers in summer and it takes place on the thawed surface of frozen soil in winter. The magnitude of soil loss is similar in the two seasons. Soil erosion in winter takes diverse forms. In their dynamics these processes on slopes resemble to each other and represent boundary cases with the common characteristic that the thawed top soil is transported by gravitation over the frozen subsoil. The medium of transport is water, the amount of which controls the way of transport and the resulting feature. A classification is made by the amount of meltwater, and the following types were distinguished: gelifluction, soil flow, sheet wash and rill erosion.

INTRODUCTION

The extent and forms of soil degradation in Hungary are essentially determined by two main factors: relief conditions and climate. The country lies at the deepest point of the Carpathian Basin, the marginal parts are, however, occupied by mountainous and hill regions, which together constitute a significant proportion of the country's area (32 %). The hill region consists exclusively of loose marine sandy and clayey sediments, volcanic tuff and are agriculturally utilized on ploughland, in vineyards and orchards. Thus, loess due to erosion is the highest in these areas, as a consequence of tillage. On the other hand, soil degradation also takes place on flatland. Here it is mostly the sand surface which is especially damaged, primarily in spring when the poorly developed vegetation is yet unable to give protection against erosion by wind.

From the agriculturally cultivated area of the country 2,319,840 hectares are eroded, including

weekly eroded (less than 30 % of the profile removed) 37.4 %,

moderately eroded (30-70 % removed) 38.5 %,

strongly eroded (more than 70 % removed) 24.1 %.

of about 50,000,000 m^3 of soil is eroded every year from slopes. Out of this amount 40,000,000 m^3 is deposited at the feet of the slopes or the meadows and pastures lying along streams and rivers. A further 1,000,000 m^3 fills up the beds of the water courses, 8,000,000-10,000,000 m^3 accumulates in sandbars along major rivers or is transported to the sea. As experts estimate it, the lack of proper soil conservation involved an annual erosion loss in Hungarian agriculture amounts to an equivalent of 1,000,000 tons of wheat. It has been proved by calculation that with the application of the appropriate soil-conserving cultivation methods the soil loss could be reduced from 50,000,000 m^3 to ca. 12,000,000 m^3 (Stefanovits, 1963, 1964).

The other factor influencing soil erosion is climate. Hungary's climate is of transitional character, as the country lies on the border of the West-European (oceanic) and East-European (continental) provinces. This transitional character also implies that soil erosion hazard is high in both winter and summer. In winter groundfrost and meltwater, in summer heavy showers are responsible for soil erosion. Our measurements show that in the particular years the extent of erosion damage can vary with seasons according to weather conditions, whereas on the long run it is of similar scale in winter and summer.

FORMS OF SOIL EROSION IN WINTER

Erosion in winter and especially at the end of the season can have special and varied forms, which are primarily the consequences of the freezing of soil. The depth of groundfrost is influenced by altitude above sea level, exposure, the physical properties of the soil, vegetation, etc. but on the average the soil surface is frozen to 30 cm depth.

EROSION ON STEEP SLOPES

The steep slopes with no vegetation cover are highly sensitive to winter frosts if they are markedly soaked through in the autumn. Erosion due to frost action can be well observed every winter along roads, ditches, edges of silt and loess embankments. As a result of autumn and winter precipitation the walls of the above-mentioned formations wet to a few millimetres or centimetres depth. Due to groundfrost ice crystals and needles grow and separate the frozen layer from the unfrozen horizon. In consequence of the sucking effect of frost the ice needles formed continue to grow and the separated rock flake is detached from the wall. On warming up, especially on sunny early spring days the ice needles melt and the separated rock fragments fall down and pile up at the foot. Overgrowth by vegetation hinders the action of frost. The parts overgrown with roots erode slowly, whereas adjacent surfaces not protected by vegetation show a faster decline. Consequently, the surface of steep slopes (wall)

grows uneven. With the progress of denudation the steepness of the wall is reduced by the gradual accumulation of material at the foot.

MASS MOVEMENTS ON SLOPE

The common feature of mass movement types is that the thawed soil is moved over the frozen subsoil as a consequence of gravitation. The amount of water controls the way of transportation and the resulting form. On the grounds of the amount of meltwater the following groups of forms can be distinguished:

1. *Gelifluction*

Gelifluction is the slow slide of saturated loose soil material over frozen subsoil. The most important criterion of this type of movement is the presence of frozen subsoil which prevents the penetration of meltwater from the surface layer deep into the lower horizons. In consequence, the saturated top layer assumes such a state of viscosity and plasticity that it loses its stability and, as a result of gravitation slides downslope. The distribution of this process is highly limited in Hungary. It is confined, in optimal cases, to a few weeks, days or perhaps only hours at the end of winter, when conditions characteristic of periglacial climate may develop. This process can be detected from several characteristic traces: the undulating slope surface, the tilted tree trunks may be indicators of such movement. Such conditions can primarily evolve on the northerly slopes of the mountains, where there is a possibility of deep, enduring groundfrost. The moisture content of the moving material may be as high as 40 %. The movement extends over the whole slope material covering the bedrock. In this sense gelifluctional movement may also have a serious role in valley evolution.

2. *Talus creep*

This form also falls within the group of gelifluction. It evolves on fine-grain soil interwoven with plant roots. In this very slow type of movement only a thin layer is affected. For this reason no observation is possible, only tongues of soil overcreeping one another can indicate its occurrences.

3. *Mudflow*

In cases when the supply of water is so plentiful that it results in the overmoistening of soil, gelifluction gives way to a process where water is preponderant. This process is mudflow and is confined primarily to agriculturally utilized areas. The process starts, as a rule, on inflectional line of the slope or below this line, and it can arise on thick groundfrost and at the time of fast thaw (thawing can also be promoted by rain). These latter conditions can be present only on slopes with northern and north-eastern exposure. The water content of the thawed soil surface is around 50 % because of deep, impermeable groundfrost. The plastified material with high water content becomes fluent. The material which starts flowing down may assume various forms. Initially it appears as a linear

feature. A particular mudflow may reach a width of a few dozen centimetres and length of several metres or some tens of metre. In the top segment, because of the loss of material due to flow, depressions of 4-10 cm come into being. On the middle and lower segments the mudflow creeps onto the surface and continues to flow elevated by a few centimetres. Mudflow is a more rapid movement than gelifluction. Part of the runoff water, as it flows faster than dilute mud, gets free from the flowing material, and preceds it in its movement. Consequently, mud loses its water content, its motion is slowed down, the material is piled up on the slope or moves on at a lower speed as gelifluction. Linear mudflows usually appear on the slope in groups. In the course of their evolution they often adjoin one another and surface lowering becomes areal. This may result in the erosion of several centimetres of soil surface over a large area.

4. *Sheetwash*

In the evolution of this process it is essential that an excess amount of water be present at the time of thaw. In this case there is water preponderant. Sheetwash erosion plays a significant role at the beginning of the period of thaw. The frozen subsoil has not yet thawed, melting of ice only begun on the surface. Owing to the frozen subsoil the runoff meltwater cannot cut into the surface, therefore, it can run only along the surface and washes off the slope. The grains of soil freed by thaw or loosened by water are drifted away by water and transported to the feet of the slope. The process of sheetwash is limited in time. When thaw is prolonged and a larger amount of material is washed off by runoff water, sheetwash erosion gives way to mudflow (Pinczés, 1971a, 1971b, 1979).

5. *Rill erosion over frozen subsoil*

This is the form of erosion most frequently emerging at spring thaws. It occurs in areas built up of silty, loessy, clayey sediments and is completely absent in sand regions. Its evolution is also determined by weather conditions. Favourable factors are: hard winter, deeply frozen subsoil, sudden thaw. As our measurements show, the difference in erosion damage can be two to threefold in different years depending on the variation of weather. The size of the rills evolved is also influenced by the exposure of the slope. They are rare on southerly exposed slopes, since on these slopes snow melts intermittently during the winter, thus, the thickness of the snow cover has gradually decreased by the time spring comes (the depth of groundfrost is also smaller) and the meltwater of the remaining thin snow blanket does not involve a considerable erosion hazard.

The danger of erosion is highest on slopes with westerly exposure, and it is twice or three times less on easterly and northerly slopes. The thickness of the thawed soil layer determines the shape of rills. The rills deepen downwards only as far as the frozen subsoil. If the thawed layer is thin, the rills grow laterally and broaden. When a thicker layer of soil is thawed then the rills will be deeper.

Our measurements carried out over a period of several years have proved that erosional damage due to meltwater may reach 100 m^3 loss of soil per hectares. The observations also turn attention to an interesting phenomenon. The data obtained from different places in different years show that the amounts of eroded material as calculated

from the size of the rills and that of the material sedimented at the foot are not identical. The sedimented material is twice or sometimes nearly five times as much as has been calculated from the dimensions of the rills. This fact suggests that at that time of end-of-winter thaws various soil degradation processes take place simultaneously and exert a joint effect. As to our example, the sheetwash action of meltwater was much more significant than rill erosion or perhaps the erosional damage due to mudflow. The latter two eroded the surface between the rills and acted in a way that can be called „invisible erosion". In fact, there takes place, besides the well-observable linear erosion, a very strong overland erosion. Our observations testify that winter soil erosion takes place in a highly varied way. These processes frequently occur simultaneously even in a particular area, act together and may result in very marked surface denudation (Pinczés and Boros, 1967).

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