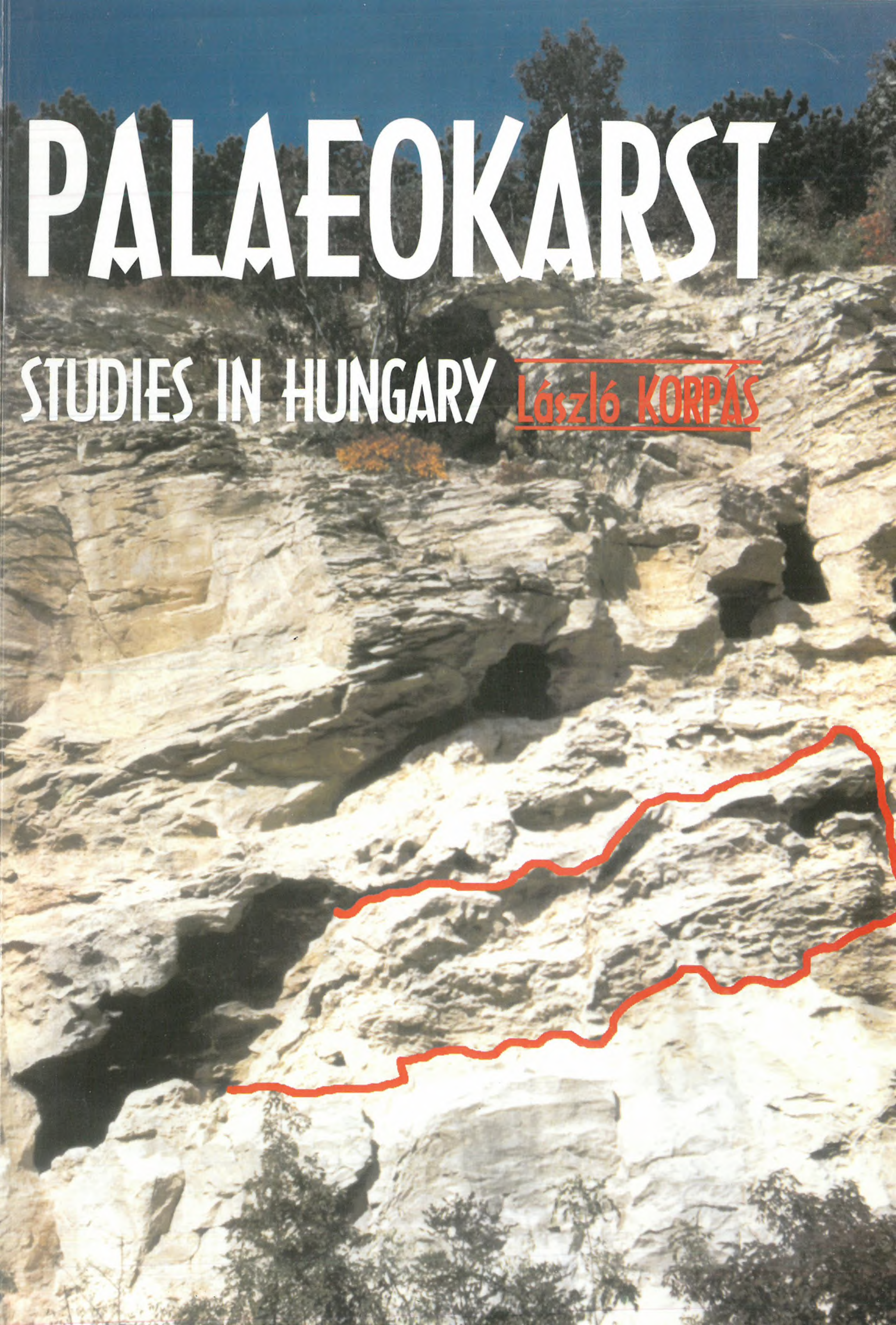
PALAEOKARST

STUDIES IN HUNGARY

László KORPÁS



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Cover sheet:
Early marine caymanite in the cavity of the Late Eocene Szépvölgy Limestone, Mátyáshegy, Buda Hills

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MAGYAR ÁLLAMI FÖLDTANI INTÉZET
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Revision by
László Jakucs Dr. Sc. (Hungary)
László Kordos Dr. Sc. (Hungary)
Dr. Arthur Satterley (United Kingdom)

Linguistic revision by
Dr. Arthur Satterley

Technical edition by
Dezső Simonyi

DTP
Ildikó Tiefenbacher

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Contents

One of today's largest provocation for society is its supply of healthy potable water. Just this phrase itself, used frequently in politics, reflects very well the day by day and more and more limited provision of this primordial human right. It seems to be noncausal, that the possession of potable water has become one of the strategic tasks in many places of the world. Potable water supply is nowadays in the focus of many national and international conflicts and it could become even more so in the future.

Hungary, because of its geological setting and favourable geographic and climatic conditions is in possession of the largest subsurface water resources in the region. The natural equilibrium of these resources with quality of potable water, has gradually broken up because of the effects of exaggerated exploitation and industrial pollution.

Therefore the welfare of future generations requires us to stop this process, already irreversible in human life-scale in order to satisfy the need for drinking water.

The surface and subsurface palaeokarst systems in Hungary represent a resource great value for the population. They give about 10% of the subsurface, partly thermal water resources, approximately 30% of the petroleum reserves, all of the bauxite, limestone and dolomite resources and a significant part of manganese ores.

The caves have extraordinary value in themselves, with their minerals, fossils and living flora and fauna. The surface occurrences of these palaeokarst systems have a specific role in the formation of the landscape and in the evolution of the biosphere. They are of increasing importance for human ecology.

Consequently the palaeokarst systems have huge natural potential. Exploitation of any elements of this potential should be realized exclusively on the account of the other ones and could result in the break up of its equilibrium. This process of deterioration was generated by the effects of mining activity, of industrialization and urbanization some ten years ago, which resulted in the acceleration of a considerable drop of the karst watertable, influenced by climatic changes too, and resulted in increasing, sometimes critical pollution of the palaeokarst systems. Among the elements of the systems the environmental factors, the quantity, quality and the state of the karst waters have become decisive. Therefore the aim of the present study is to contribute to the protection and to the rehabilitation of the palaeokarst aquifer systems.

1. *Palaeokarst systems and geological models*

1.1. Introduction

The investigation of palaeokarsts has become an outstanding area of research in the past 20 years as proved by the wide range and of the related natural and mineral resources potential (oil, bauxite, Pb–Zn ores, uranium ores, manganese ores, phosphates, nitrates, groundwater and thermal water, **Table 1**). The timeliness and significance of palaeokarst studies is enhanced by the fact that 35% of oil, 15% of bauxite, 10% of Pb–Zn and 10% of water reserves of the world are located in palaeokarst areas.

Since a considerable part of mineral resources (such as the bauxites, most karst and thermal waters in Hungary and some oil fields in SW Transdanubia, in the southern part of the Great Hungarian Plain) are connected with palaeokarst, in 1989 the Geological Institute of Hungary began a systematic study of the palaeokarst systems in Hungary.

The research aims to clarify the following points:

— time and environmental conditions necessary for the formation of palaeokarst,

— the criteria for recognition of palaeokarst,

— the role of palaeokarst in carbonate diagenesis,

— the role and significance of palaeokarst horizons in local and regional stratigraphic correlation as well as in geodynamic reconstruction,

— genetic–paragenetic relations between palaeokarsts and related natural potential,

— 3D models of palaeokarst systems.

The solution to the above mentioned problems is expected to provide an up-to-date explanation of the genesis of palaeokarst on the one hand and the conceptual renewal of the research and exploration strategies of the natural potential related to palaeokarst on the other.

1.2. What we know about the palaeokarst today

Before discussing our present knowledge of palaeokarst it is desirable to define the notion itself. This is necessary because all kinds of palaeokarsts can be classified as "karst", while the palaeokarstic origin of some of them, especially those at the surface, can be verified only with great difficulties or in no way at all. This also means that our research is limited to palaeokarst sensu stricto.

Defining palaeokarsts we used the definition of WRIGHT (1982), ESTEBAN and KLAPPA (1983), CHOQUETTE and JAMES (1988), BOSÁK et al. (1988) according to which palaeokarst was formed over geological time. The statement is evident in the case of karsts filled with younger sediments (covered karst), whereas with uncovered, relict or exhumed karsts it can be applied only occasionally. From among the general (morphological, hydrodynamic, hydrochemical, biochemical, lithologic, climatic and tectonic) conditions of palaeokarstification also specified by the above authors we could emphasize the statement by ESTEBAN and KLAPPA (1983) who point to proper "diagenetic facies" as a distinct marker.

Further on we deal with the particulars of the environment in a wider sense (facies, morphology, infillings, syn- and diagenesis).

Table 1

Examples of some raw material occurrences related to palaeokarst

Raw material type	Palaeokarstic reservoir formation				Occurrences	
	Name	Lithology	Age	Cover	Place	References
Petroleum	San Andres	Dolomite	Late Permian	Late Permian	USA/W Texas Yates field	D. H. Craig (1988)
	Hauptdolomit	Dolomite	Late Triassic	Late Cretaceous	Hungary/Nagylengyel	Gy. Bárdossy–L. Kordos (1989)
	Dachstein Limestone	Limestone				
	Ugod Limestone	Limestone	Late Cretaceous	Eocene		
	El Abra	Limestone, dolomite	Middle Cretaceous	Late Cretaceous	Mexico/Tampico	C. J. Minero (1989)
Bauxite	Szársomlyó Limestone	Limestone	Late Jurassic	Early Cretaceous	Hungary/Nagyharsány	Gy. Bárdossy– L. Kordos (1989)
	Dachstein Limestone	Limestone	Late Triassic	Middle Cretaceous	Hungary/Alsópere	
	Guajaibon	Limestone, dolomite	Middle Cretaceous	Middle Cretaceous	Cuba, Pan de Guajaibon	L. Korpás (1988)
	Hauptdolomit Kössen F. Dachstein Limestone	Dolomite, limestone	Late Triassic	Late Cretaceous	Hungary/Halimba	Gy. Bárdossy–L. Kordos (1989)
	Ugod Limestone	Limestone	Late Cretaceous	Middle Eocene	Hungary/Csabpuszta	
Mn ore deposits	Csárdahegy	Limestone	Early Jurassic	Middle Eocene	Hungary/Úrkút	Gy. Bárdossy– L. Kordos (1989)
Pb–Zn ore deposits	Knox Carbonates	Limestone	Ordovician	Ordovician	USA/E Tennessee	W. J. Mussman et al. (1988)
	Madison Limestone	Limestone	Early–Late Carboniferous	Early–Late Carboniferous	USA/Wyoming	W. J. Sando (1988)
	Leadville F.	Dolomite, limestone			USA/Colorado	R. H. De Voto (1988)
	Muschelkalk	Dolomite	Early–Middle Triassic	Middle Triassic	Poland/Krakow	K. Bogacz et al. (1970)
U ore deposits	Madison Limestone	Limestone	Early Carboniferous	Pliocene–Pleistocene	USA/Pryor-Bighorn	W. J. Sando (1988)
Water	Paget F.	Limestone	Late Pleistocene	Holocene	Bermudas	H. L. Vacher (1978)

As far as the statements made about facies are concerned, the facies of the wall rock and of the palaeokarst will be discussed separately. The examples studied (**Table 2**) suggest, that the palaeokarstification, completed in a short time by geological standards, involved shallow marine, often peritidal platform carbonates (mainly limestones, and frequently also dolomites). Taking this into account along with the considerations of ESTEBAN and KLAPPA (1983) according to which palaeokarstification appears on shallowing upward sequences which are divided by discontinuity surfaces of different order, (**Fig. 1**) the obvious conclusion to be drawn is that, there must be a relationship between the original facies of carbonate rocks and the early or initial palaeokarst facies. Evaluating the facies of palaeokarsts —in addition to the general conditions mentioned above— we emphasize its originally subaerial nature. Within this fairly "tentative" facies qualification —after DAVIS (1930) and BRETZ (1942)— it is possible to mark phreatic facies roughly equivalent to the exisiting base level as well as vadose facies situated above. To the vertical profile (**Fig. 2**) of these facies zone, (ESTEBAN and KLAPPA 1983) the following observations should be added:

— the profile statically presents a unique, idealized mature karst cycle,

— there is a permanent dynamic interaction in space and time between the separate facies and levels, in the course of which constant migration of levels and facies, their substitution for one another, superposition and repetition is to be reckoned with. The complicated result is the polycyclic karst sequence, essentially controlled by base-level variations.

Table 2

The stratigraphic position of some palaeokarst horizons

Stratigraphic position	Occurrences	References
Holocene–Pleistocene	Bahamas	K. A. Rasmussen–A. C. Neumann (1988)
Late Pleistocene	Bermudas	H. Bretz (1960)
	Florida	J. R. Dodd–C. T. Siemens (1971)
Pleistocene	Hungary	Gy. Bárdossy–L. Kordos (1989)
Pliocene–Pleistocene	Bahamas	D. K. Beach–R. N. Ginsburg (1980)
Neogene	Hungary	Gy. Bárdossy–L. Kordos (1989)
Oligocene		
Late Eocene	Hungary	S. Kraus (1988)
Paleocene–Early Eocene	Hungary	Gy. Bárdossy–L. Kordos (1989)
Late Cretaceous		
Middle Cretaceous	Mexico	C. J. Minero et al. (1988)
	Cuba	L. Korpás (1988)
	Hungary	Gy. Bárdossy–L. Kordos (1989)
Early Cretaceous	Hungary	
Early–Middle Jurassic	Spain	J. A. Vera et al. (1988)
Late Triassic	Sicily	R. Catalano et al. (1974)
Early–Middle Triassic (?)	Poland	K. Bogacz et al. (1970)
Late Permian	USA/W Texas	D. H. Craig (1988)
Early–Late Carboniferous	USA/New Mexico	W. J. Meyers (1988)
Early Carboniferous	USA/Colorado	R. H. De Voto (1988)
	England/South Wales	V. P. Wright (1982, 1988)
Silurian	USA/W Ohio	C. F. Kahle (1988)
	Sweden	L. Cherns (1982)
Ordovician–Silurian	Canada/Ontario	D. R. Kobluk (1984)
Ordovician	Canada/Quebec	A. Desrochers–N. P. James (1988)
	USA/Appalache	W. J. Mussmann et al. (1988)

STRATIGRAPHIC DISCONTINUITIES			ORDER		TIME GAP SCALE My	TIME GAP SCALE Chronostrat.	CORRESPONDING STRAT. UNIT	My
UNCONFORMITITES	COMPOSITE	E. UNCONFORMITY	1st	INTER-REGIONAL	200	ERATHEM	MEGASEQUENCE	200
		MEGA-CONFORMITY	1st		>60	SYSTEM	SUPERSEQUENCE SET (sequence of Sloss)	400
		S.U. SET	2nd		30	SERIES	SUPERSEQUENCE (synthem)	10–100
		SUPER-UNCONFORMITY	2nd		4–12	STAGE		
	SINGLE	REGIONAL UNCONFORMITIES (sequence boundaries) TYPE 1 TYPE 2	3rd	REGIONAL	~1	BIOZONE	DEPOSITIONAL SEQUENCE mesothem	1–10
CONFORMITIES		SYNTECTONIC UNCONFORMITY	3–4		0.01–1	Variable	ROTATIONAL ONLAP–OFFLAP syntectonic cum. wedge	
		BOUNDARY OF SHOALING CYCLES	4th		0.01	Not recognizable	PARASEQUENCE cyclothem	0.2–0.5
		BEDDING PLANE	5th		0.001		BED	0.01

Fig. 1. Stratigraphic discontinuities and time gaps (ESTEBAN 1991)

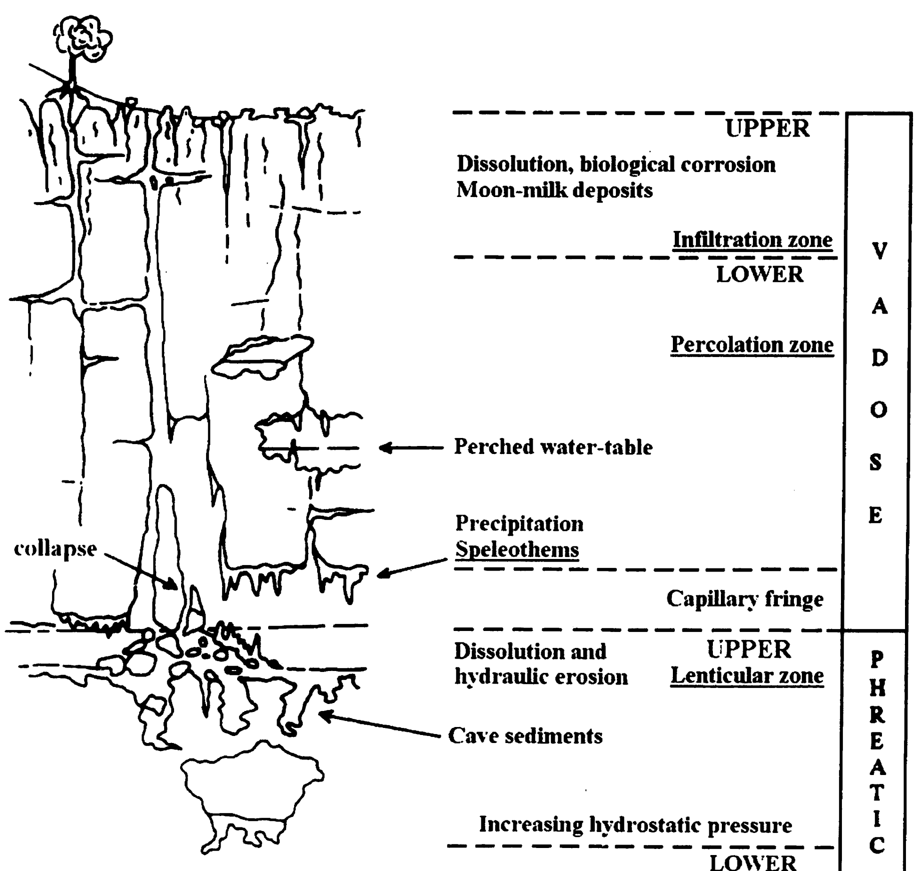


Fig. 2. Idealized authigenic karst profile, not to scale (ESTEBAN and KLAPPA 1983)

The development of the main palaeokarst stages is presented in **Fig. 3** (according to ESTEBAN 1988).

Finally, evaluating the facies of the wall rock and the palaeokarst in accordance with ESTEBAN and KLAPPA (1983) and WRIGHT (1988) we focus on the close relation between palaeokarstification and global eustatic sea level changes (**Fig. 4**).

Dealing with the morphology of palaeokarsts we concentrate on features controlled by bedding plane and by tectonism, and on their combination. Accordingly we distinguish conformable (parallel with the bedding) and disconformable (cutting the bedding) and the combination of conformable–disconformable types. By analysing the connection between morphology and facies it is possible to make certain simplifications and say that the conformable elements are characteristic primarily of the phreatic levels, while disconformable elements occur mainly in vadose levels. Since most palaeokarst systems contain both conformable and disconformable elements, they cannot, in themselves, be used mechanically in genetical and facies analysis.

The infilling sediments of palaeokarst are characterised by extreme lithological, genetic and facies variability. As a result, their general classification (**Fig. 5**) —in our opinion— is a task almost impossible to solve.

Diagnostic criteria for palaeokarsts are happened to be different clayey palaeosols and caliche profiles. A detailed description of the latter is given by ESTEBAN and KLAPPA (1983).

Syn-depositional and diagenetic phenomena are regarded as a complicated system of porosity forming and infilling processes, changing constantly in time and space (**Fig. 6**). Here we call attention again to the definition of karst as a proper "diagenetic facies" by ESTEBAN and KLAPPA (1983, p. 11). It seems to be justifiable to use their definition in a wider sense: thus we can state that the initial palaeokarst may have played a decisive role in the syngenetic and early diagenetic processes of peritidal carbonates. These processes take place in the vadose (meteoric), oscillation (transitional) and shallow phreatic zones, partly under subaerial conditions and some of them at shallow burial depths. Some diagenetic processes also operate at greater burial depths, in the deep phreatic zone. Both in the syn-diagenetic and in late diagenetic phases thermal effects cannot be excluded, and palaeokarst horizons having entered deep diagenetic zones earlier, may get into subaerial conditions again. Thus the reconstruction of diagenetic history is bound to be an extremely hard task.

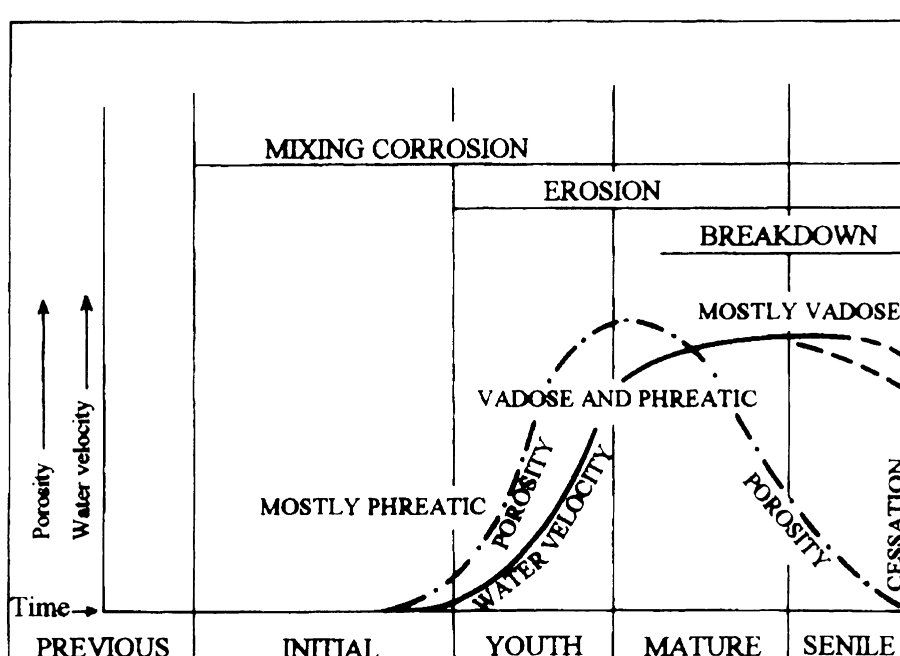


Fig. 3. Stages of karst evolution (ESTEBAN 1988)

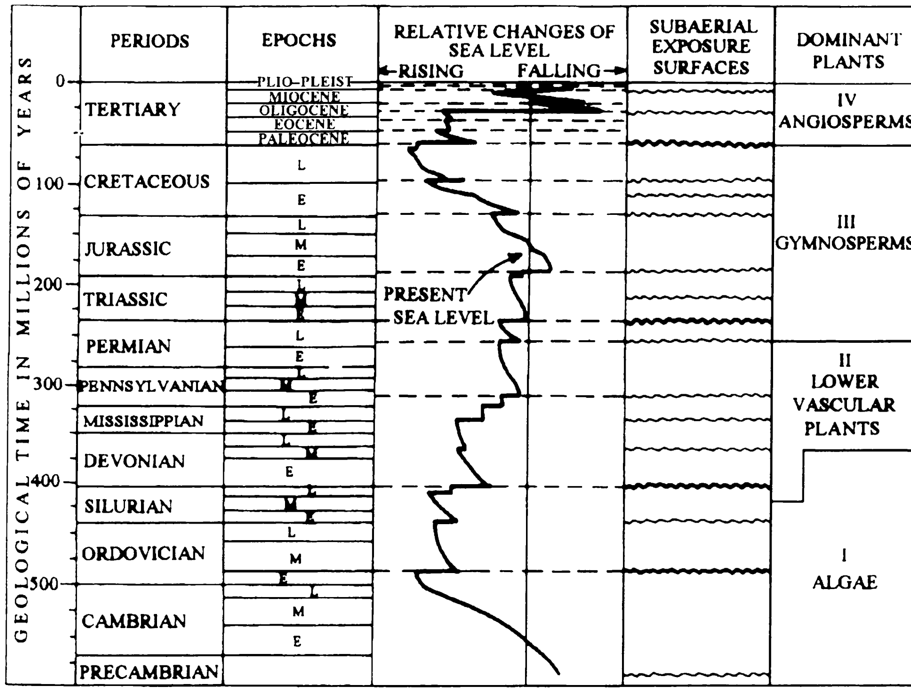


Fig. 4. Relative global sea level changes, major known subaerial exposure surfaces and dominant plant groups during Phanerozoic time (ESTEBAN and KLAPPA 1983)

ALLOCHTHONOUS CLASTIC
1) FLUVIAL — many kinds — predominantly allochthonous
2) DEJECTA CONES, COLLUVIUM and MUDFLOWS — common
3) FILTRATES — cones of fines from seepage; minor in volume
4) LACUSTRINE — rare; clays, silts and fine sands
5) MARINE — rare; beach facies at entrances
6) EOLIAN — normally minor to negligible
7) GLACIAL and GLACIO-FLUVIAL INJECTA — common in glaciated areas
8) SUBGLACIAL LACUSTRINE — often varve-like

ORGANIC
9) FLUVIAL OR EOLIAN TRANSPORTED — from tree trunks to spores
10) CAVE-USING EXTERIOR FAUNA — bones, nests and middens, faeces

AUTOCHTHONOUS CLASTIC
11) BREAKDOWN
12) FLUVIAL — derived from breakdown
13) WEATHERING RINDS and EARTHS
14) EOLIAN — derivates of all the above deposits

PRECIPITATES and EVAPORITES
More than 100 secondary minerals form in caves

PREDOMINANT ARE:
15) CALCITE — most significant
16) OTHER CARBONATES, HYDRATED CARBONATES
17) SULFATES, HYDRATED SULFATES, HALIDES
18) PHOSPHATES and NITRATES
19) SILICA and SILICATES
20) ORE ASSOCIATED MINERALS
21) ICE — as glacial injecta, glaciers, water ice, frost

ORGANICS
22) TRACKS and REMAINS CAVE ADAPTED (HYPOGEAN) and PHREATOPHYTE FAUNA

Fig. 5. Cave interior deposits (FORD 1988)

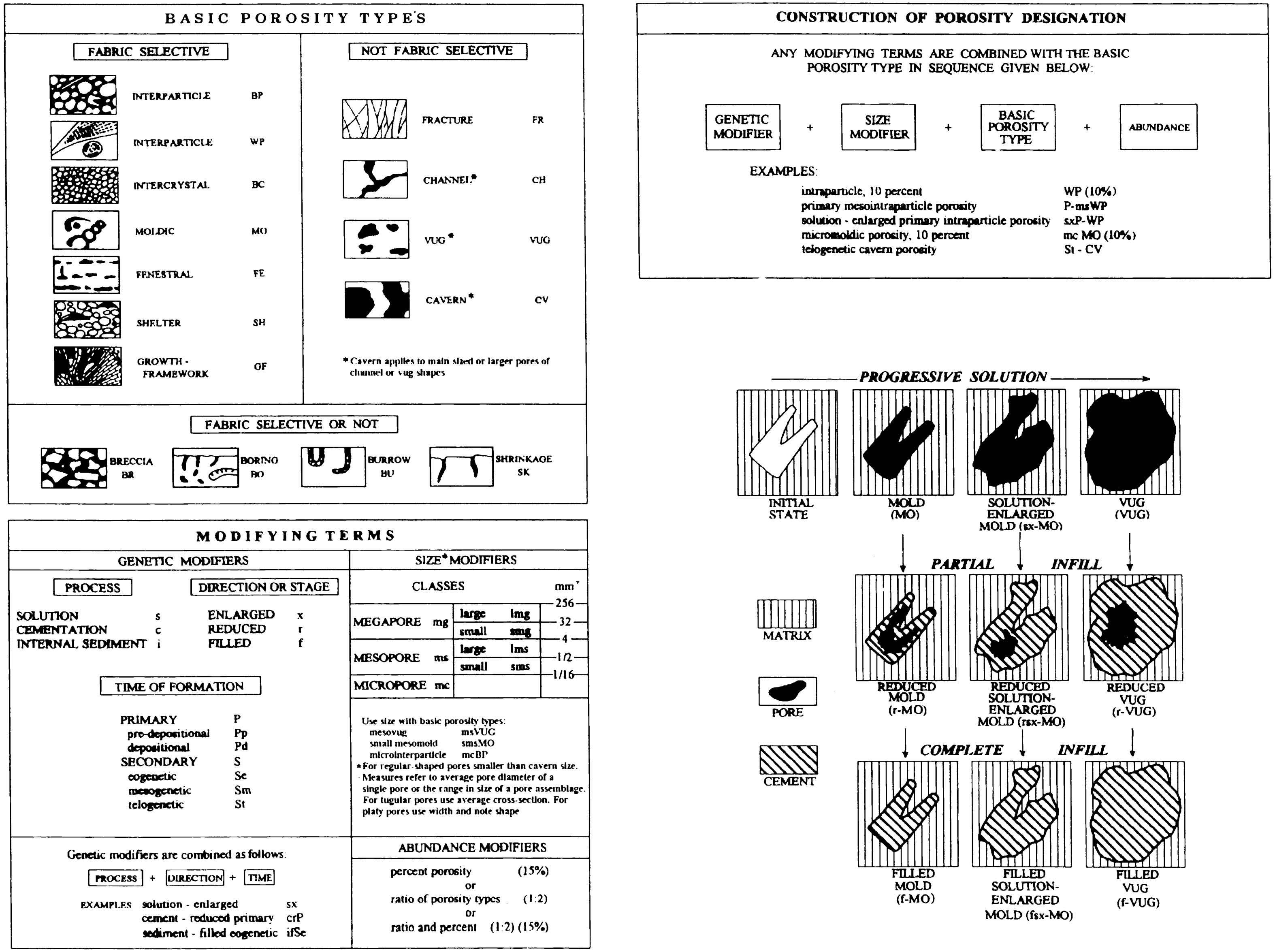


Fig. 6. Classification of pores and pore systems in carbonate rocks (CHOQUETTE and PRAY 1970)

In the classification of palaeokarsts, genetic, geographic, morphological, climatic and tectonic criteria play important roles individually or combined with each other. Some genetic classifications may be governed by authigenic or allogenic recharge (JAKUCS 1977), by stages in development (early, juvenile, mature, senile), by temperature of solutions producing karstification (e.g. thermal or hydrothermal karsts), by the relation between karstic features and the surface (e.g. covered, uncovered, exhumed karsts) and in some cases by the grade of the conservation of karsts (e.g. relict karsts). In geographic classifications the present morphological position (e.g. coastal karst, continental and alpine karst) or even geographic name of characteristic occurrences (Dinarid-karst, South-Chinese tower karst, Cuban conekarst) is important. The guiding principle of morphologic classification may be the variety of characteristic forms (e.g. conekarst, tower karst), or in some cases the landscape where the karst occurs (cockpit karst). Climatic classification is based on the present location of karst occurrences within different climatic belts (tropical, mediterranean, temperate, arctic). In a tectonic classification the global position (e.g. craton or continental platform areas, folded orogenic belts, island arcs) is a decisive factor or individual faults controlling the karst.

The description of thermal or hydrothermal karsts from Ordovician, Lower Carboniferous, Lower and Middle Triassic, and Upper Eocene formations (BOGACZ et al. 1970, BROWN 1970, ESTEBAN and KLAPPA 1983, SANGSTER 1988, MUSSMAN et al. 1988, KRAUS 1988, NÁDOR 1992a, b, KLEB et al. 1993a) suggests that they are usually formed by fluids of 100–200 °C. We suggest individual evaluation in deciding whether these are syngenetic or epigenetic.

In connection with analysis of palaeokarsts and tectonics as well as of unconformities (discontinuities) we stress again their tectonic control. The megatectonic conditions together with other (predominantly climatic) factors, are of decisive importance in the formation and development of palaeokarsts. It is an interesting phenomenon that palaeokarsts of continental platforms and of island arc areas are almost identical. In our opinion this indicates similar conditions of formation, despite the great difference in geologic–tectonic position. The important role of the linear tectonic control in the formation and development of palaeokarsts has long been known and generally accepted. In spite of that, until now there has been little attention paid to microtectonics as a diagnostic criterion in the early syn-diagenetic phase of palaeokarstification.

Unconformities (discontinuities) indicating a break in shallow marine cyclic carbonate sedimentation have been known for a long time from the work of FISCHER (1964) and their facies interpretation is universally accepted. Attention to the important role of unconformities in the formation and recognition of palaeokarsts was called by ESTEBAN and KLAPPA (1983), DESROCHERS and JAMES (1988), ESTEBAN (1988), KORPÁS and ITURRALDE VINENT (1992) JUHÁSZ et al. (1995), who stressed their close genetic relation with tectonics and eustatic sea level changes.

One of the most important practical aims in palaeokarst research is the estimation of porosity, i.e. of determining the location and grade of their reservoir potential. Porosity evolution at different palaeokarst stages, as well as in vertical profile are presented on **Figs. 3 and 7**. On the basis of these we conclude that from point of view of porosity evolution:

— the most favourable stages are found in juvenile and mature karst,

— the most favourable facies are the upper part of the phreatic zone, the transitional zone between the phreatic and vadose zones as well as the lower part of the vadose zone.

Here we would like to call attention to the case study and excellent genetic model by CRAIG (1988) which analyses the conditions of the palaeokarstic reservoir of the Upper Permian San Andres dolomite in Yates Field, West Texas.

The diagnostic criteria of palaeokarsts mostly consist of indirect, and, in smaller portion of direct elements. This is easy to understand because those should be selected from among the general criteria of karsts which prove their "palaeo" nature. From among the criteria listed by ESTEBAN and KLAPPA (1983) and CHOQUETTE and JAMES (1988) those are selected here which are specially valid for palaeokarsts:

— sequences shallowing upward with rapid wedging-outs and truncations in them,

— cyclic unconformity (discontinuity) surfaces dividing the sequences,

— authigenic/allogenic as well as monomictic/polymictic clastics connected with these,

— tepee structures and other in situ, often conformable sedimentary breccias,

— caliche facies,

— the occurrence of terra rossas and other palaeosols,

— desiccation phenomena,

— the presence of gravitational meniscus (vadose) cement,

— missing palaeontological zones.

To complete the list we would suggest to add tectonic dislocations between two discontinuity surfaces, as well as synsedimentary microtectonic phenomena producing infillings. It is an interesting fact, that no mention is made of these in the cited works.

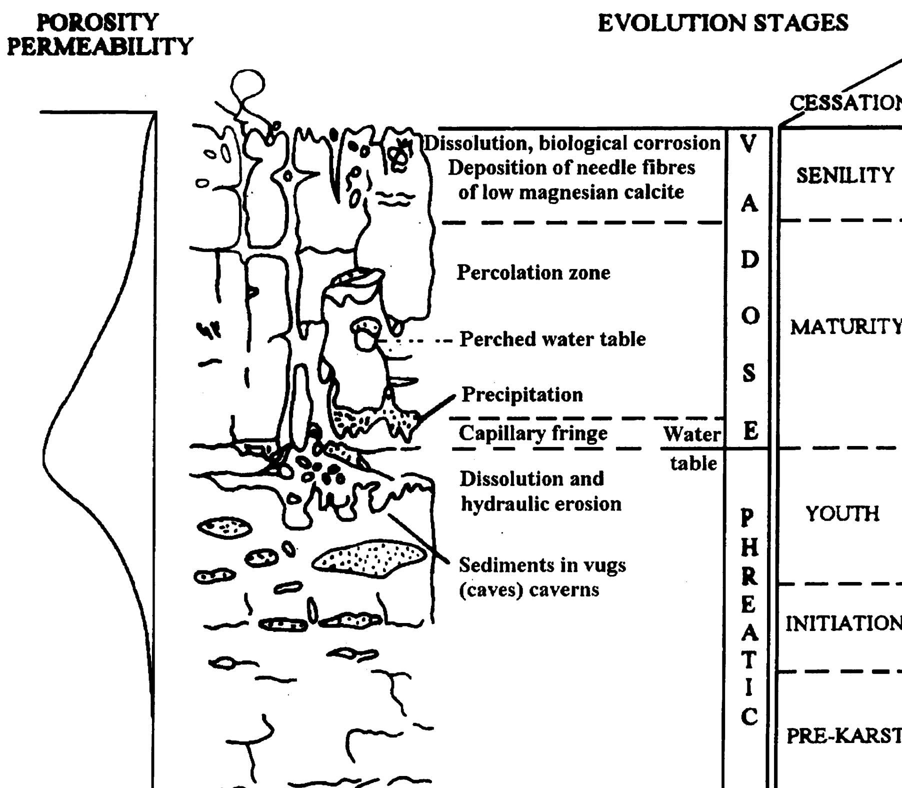


Fig. 7. Karst evolution stages and porosity–permeability in vertical section (ESTEBAN 1988)

1.3. Geological models of palaeokarsts

The theoretical model of the entire karst system as drawn up by FORD (1988) is presented in **Fig. 8**. Since in the above model palaeokarst profiles may appear anywhere, we will concentrate on those models which describe karstification near to sea level. Using the figures of ESTEBAN and KLAPPA (1983) we present the zonation of exposed coastal surfaces of recent rocky shores and sediment shores (**Fig. 9**). The further development of the above mentioned exposure surfaces in connection with sea level rise or fall is shown (**Fig. 10**). As a next step in the interpretation of the cited profiles, we analyse them as a dynamic evolution intersected by sea level changes of different order. In this way the coastal palaeokarst profiles of recent and Pliocene–Pleistocene examples were formed, which are known from the Bahamas (RASMUSSEN and NEUMANN 1988), from the Bermudas (BRETZ 1960, VACHER 1978), from Florida (DODD and SIEMENS 1971, ENOS and PERKINS 1979) and from Barbados (JAMES 1972, ALLAN and MATHEWS 1977). On the basis of this we do not consider it accidental, that the palaeokarst models drawn up in the rel-

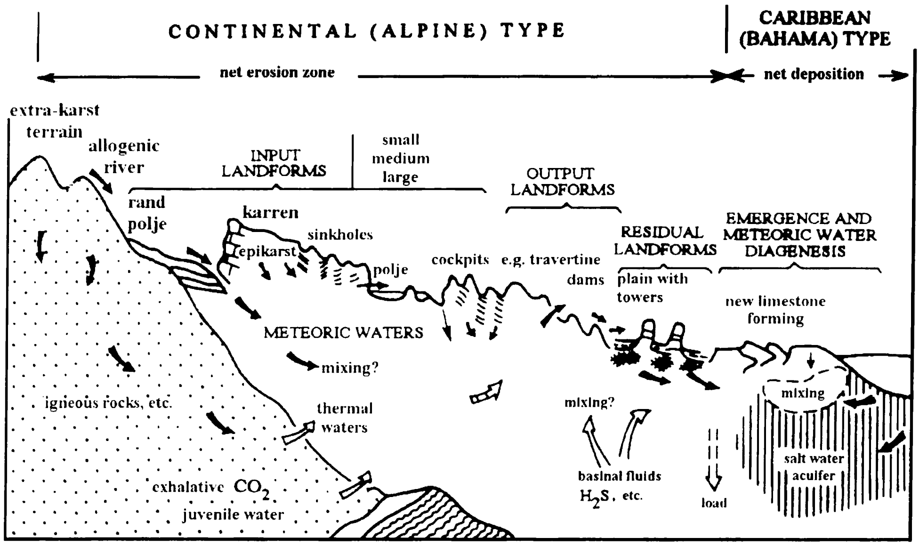


Fig. 8. The comprehensive karst system (FORD 1988)

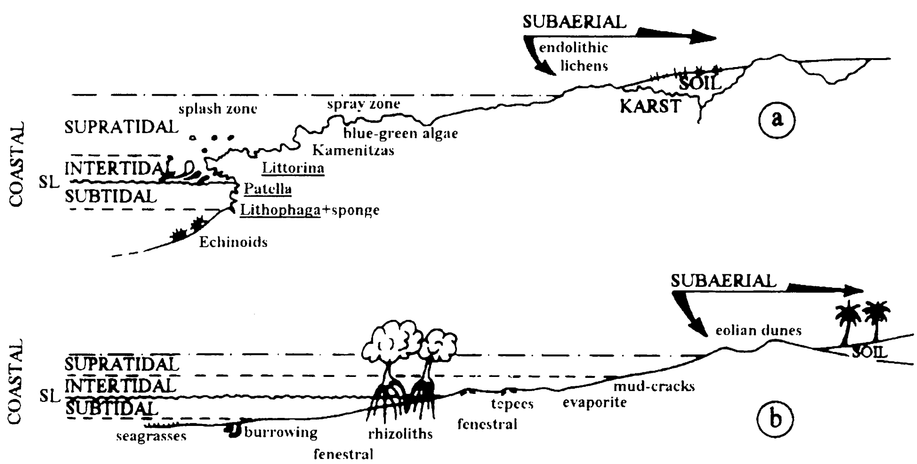


Fig. 9. Schematic cross-section of coastal surfaces (ESTEBAN and KLAPPA 1983)
a) common zonation across present-day rocky shores b) main features across present-day sedimentary shores

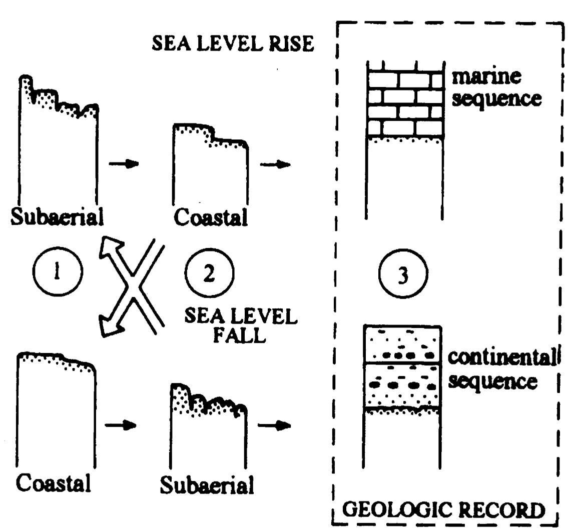


Fig. 10 Major pathways of evolution of exposure surfaces (ESTEBAN and KLAPPA 1983)

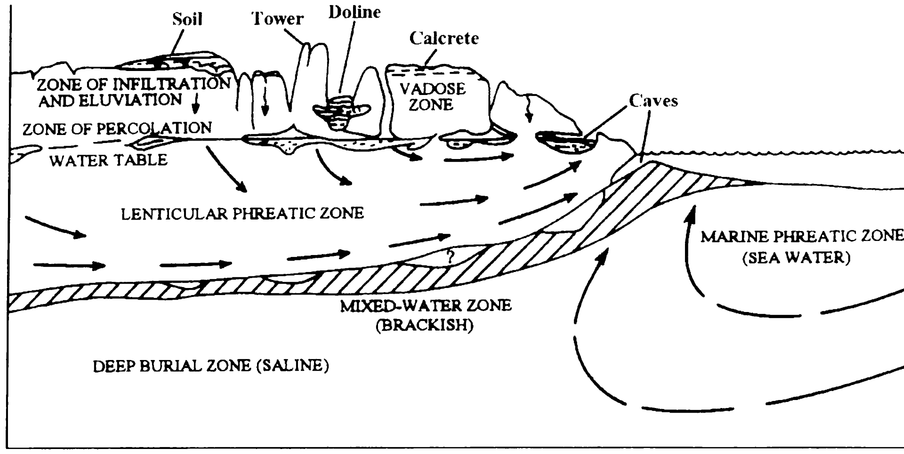


Fig. 11. General elements and hydrology of karst terrain developed along the sea (CHOQUETTE and JAMES 1988)

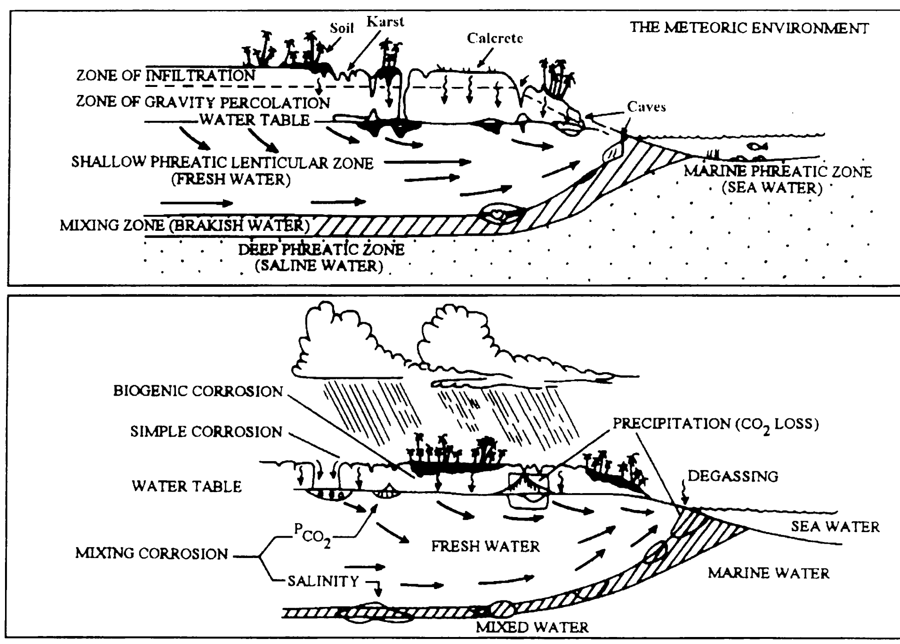


Fig. 12. Caribbean karst model (ESTEBAN 1988, after JAMES and CHOQUETTE 1984)

evant literature describe essentially present-day coastal conditions. We present more general examples, valid both for continental margins and for islands (**Figs. 11, 12**) too.

Models similar to those above were applied by CRAIG (1988, p. 360, Fig. 16/16) in the genetic analysis of the Upper Permian San Andres dolomite reservoir in West Texas, as well as by MINERO (1988, p. 393, Fig. 18/4) in the reconstruction of the palaeokarst of the Middle Cretaceous El Abra Formation in Mexico.

It is notable that although all the coastal palaeokarst models presented outline the palaeohydrologic conditions and zonation, we have come across only one example of the analysis of proper hydrologic conditions. VACHER (1978), on the Bermudas, investigated by means of a well-monitoring system the position of the Ghyben-Herzberg lens, (providing the drinking water for the inhabitants), its transitional and mixing zone, and the systematic changes of water table (p. 208, Fig. 1, p. 209, Fig. 2, p. 215, Fig. 4, p. 218, Fig. 5).

From literature and our observations (KORPÁS 1988, KORPÁS and ITURRALDE VINENT 1992, JUHÁSZ et al. 1995), we have come to the conclusion that the early phase of palaeokarstification must have taken place in a marine environment, in a position near sea level. Our opinion is that the great variety of forms characteristic of the juvenile and mature phases of palaeokarsts, in addition to the mixing corrosion in the coastal phreatic and vadose facies zone, mechanical and bioerosion have a significant role to play. Mechanical erosion —waves, tidal movements, stormtides and the earth tidal pump— has an extremely quick and intensive effect especially in places weakened by tectonics and bioerosion[1].

Conclusions 1.4.

The proposed palaeokarst model (**Fig. 13.**) contains essentially few new elements as compared to previous ones. Drawing it up we considered the basic principle that palaeokarstification, in its entirety, is the result of processes accompanied by relative sea level changes. This can be caused by eustatic sea level changes or by tectonic fac-

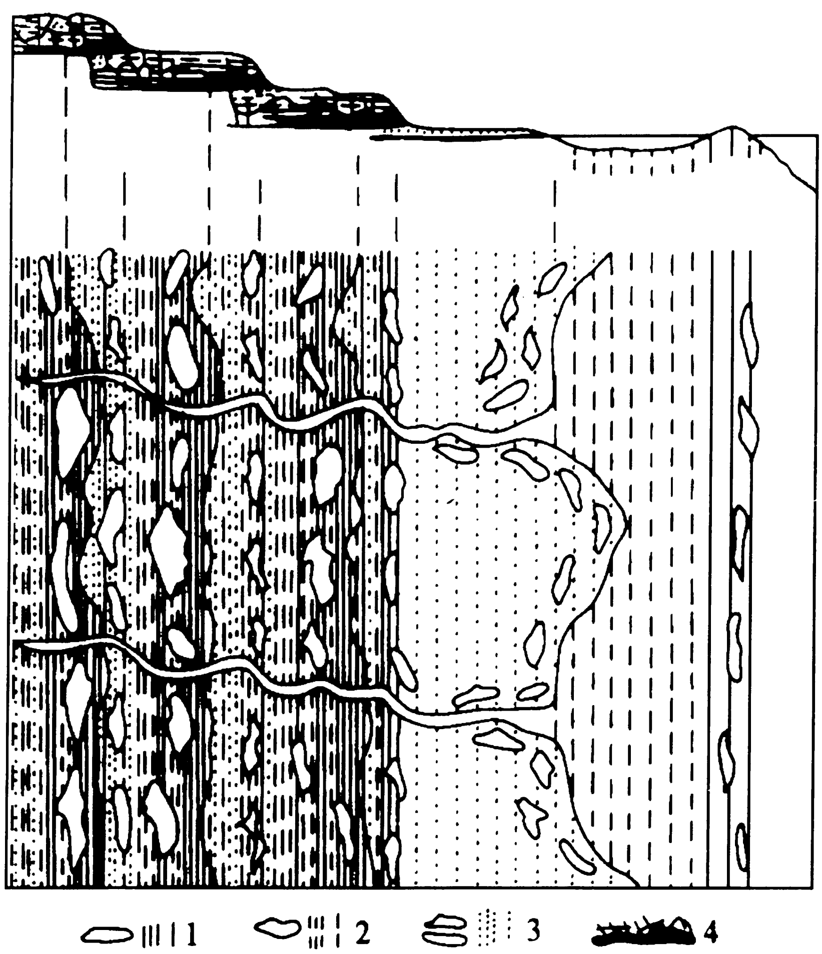


Fig. 13. Regressive terrace model of palaeokarsts (KORPÁS and JUHÁSZ 1993)
1. reef and reefal sediments, 2. lagoon sediments, 3. supratidal sediments, 4. karst levels (vadose/phreatic)

[1] The role of bioerosion in karstification is recognized by all authors quoted in this work, but they write cautiously about its extent and significance. Studies dealing with bioerosion and trace fossils (HODGKINS 1964, NEUMANN 1966 in WARME 1975, GOLUBIC et al. 1975, BROMLEY 1975, CRIMES 1975, SCHNEIDER 1976) indicate that this effect is most intense in the subtidal, tidal and supratidal zones. The significance of bioerosion lies not so much in the rates of erosion, calculated in some cases (Australia, HODGKINS 1964, 1 mm/year, Bermudas, NEUMANN 1966, 14 mm/year) as in the preparation of the substrate for mechanical erosion and mixing corrosion. CRIMES (1975, p. 118. Fig. 7/2.) publishes a synoptic facies diagram of trace fossils "participating" in bioerosion, and especially the part referring to palaeokarstification along rocky and sedimentary shore is instructive from our point of view.

tors, or by the combination of both (ESTEBAN and KLAPPA 1983, DESROCHERS and JAMES 1988, WRIGHT 1988).

Considering this, the following criteria were applied to the proposed model:

— sedimentological criteria: taken in its complexity the facies of the carbonate wall rock in palaeokarst profiles is regressive. This means that in sequences shallowing upward, regressive migration of standard carbonate facies belts is to be expected according to the trend in **Fig. 13**. The sequence boundaries coincide with the third-order discontinuity (palaeokarst) surfaces,

— tectonic and/or climatic criteria: the trend outlined above is the outcome of periodic rises and/or falls of base-level resulting from regional tectonic processes and/or sudden sea level changes,

— tectonic criteria: periodic and terrace forming emergence by all probability takes place in stages, starting from outer shelf margins and produces a littoral karst plain or tidal flat of varying width (some 100 m — some km).

The novelties of the model presented for the interpretation of palaeokarst evolution are the following:

— it associates phreatic cave levels with former sea levels,

— it attributes a significant role in the karstification of phreatic levels to mechanical erosion brought about by whirling and turbulent seawater in the system,

— it provides an acceptable explanation for the gradual wedging-outs of phreatic levels, working towards predetermined direction (windward side coasts),

— it relates the original terrace to the palaeokarst levels, partially built one upon the other.

1.5. Exploration strategies and methods of study

The exploration strategies applied to palaeokarst systems are defined by the rational exploration, use and explotation of their natural and mineral potential. Therefore, during the evaluation of this potential it seems to be primordial to estimate whether the palaeokarst systems should serve only like reservoirs and traps for this potential, or whether they are in close genetic relationship to each other. In the first case (oil, karst and thermal waters) it seems to be sufficient to determine the main conduit zones and the reservoir potential. The second possibility requires the genetic analysis of both the palaeokarst systems and their solid mineral potential. Taking into consideration that it is impossible to develop an exploration strategy aimed both for general and for specific objects, we had to stop at a point where the specific evaluation of palaeokarst systems for mineral deposits begins (**Table 3**).

Table 3

Exploration strategies (after ESTEBAN 1988)

I. Unconformities A) Their importance in exploration B) Unconformities and palaeokarsts II. Karst systems A) Modern karsts 1. Karst and lithology 2. Porosity 3. Karst facies 4. Genesis and evolution of karst systems B) Palaeokarst facies and genesis 1. Main facies 2. Characteristics of palaeokarst facies in outcrops, in drilling cores, in polished sections, in well log and in seismic profiles	3. Analysis of the palaeokarst facies (mineralogy, petrology, geochemistry, stratigraphy, palaeontology, microtectonics, physics and geophysics) 4. Palaeokarst profiles and porosity distribution 5. Genesis, evolution and 3D models III. Palaeokarst and its natural potential 1. Petroleum 2. Karst and thermal water 3. Bauxite 4. Lead and zinc ores 5. Uranium ores 6. Manganese ores 7. Phosphates 8. Nitrates

An understanding of the genesis of palaeokarst systems requires an integrated approach to their study (**Table 4**).

Table 4

Methods of study

1. Sampling (exposures, caves, mining tunels, drilling cores)
2. Methods of investigation
— Mineralogy and petrology (microfacies, mineral phases, diagenesis, fluid and gas inclusions, vitrinite reflectance)
— Geochemistry (major and indicator elements, stable isotopes of O^{18}/C^{13}, radiometric age determinations)
— Palaeontology (age and facies)
— Stratigraphy (magnetostratigraphy, sequence stratigraphy)
— Microtectonic survey
— Physical parameters and geophysical properties (porosity and permeability, well log analysis, seismic reflexion survey)

2. Worldwide examples

2.1. Bahamas

Geological setting

The Little Bahama Bank represents the second largest of the carbonate banks which constitute the Bahamas. It consists of young Mesozoic and Tertiary platform carbonates and platform margin sediments, with a thickness of 5000–6000 m covered by 30–40 m of Pleistocene and Holocene eolianites, oolite shoals and lagoonal deposits with palaeosols (HINE and NEUMANN 1977, RASMUSSEN and NEUMANN, 1988). The entire section is dissected by cave systems, and is bounded to several discontinuity surfaces, giving it a "honeycombed" architecture (BEACH and GINSBURG 1980). Two main discontinuity surfaces are exposed by boreholes and seismic surveys (**Figs. 14, 15**). The first one is located some meters below sea level, at the base of the partly soft modern sediments. The second one represents the composite, partly subaerial unconformity surface between the Holocene and Pleistocene sediments, at a depth of 20–45 m below sea level.

Palaeokarst interpretation

The profiles (**Figs. 14, 15**) represent excellent examples of windward shelf and shelf margin sections, located along the rimmed carbonate platform of the Bahamas. The deposition of the Holocene fringing reef complexes started about 10,6 ky ago (HINE and NEUMANN 1977) and continuous sea level rise has resulted in the formation of dissected reefs and reef pinnacles. The depressions between the individual reefs were filled partly by lagoonal deposits, partly by reef talus breccias. It is noteworthy, that the morphology and the sizes of these individual reefs correspond to the morphology and sizes of cone and tower karsts. The dimensions of the back reef lagoons are very similar to those of dolines, while the reef talus breccia found in depressions, using karst-terms, seems to be a collapse breccia. The entire palaeokarst profile is covered by very young sediments (3,6 ky to present in age) and is located below sea level.

2.2. Gulf of Mexico (West Florida, Campeche Bank–Mexico)

Geological setting

The 12–16 km thick platform sequence, accumulated at the southern border of the North American Craton consists of the following main lithological complexes (VINIEGRA 1971, MEYERHOFF and HATTEN 1974, SHERIDAN et al. 1981):

— Middle Triassic–Late Jurassic fluvial-delta-deltafront and prodelta clastics, deposited on platform and platform margin areas (2000–5000 m).

— Late Jurassic–Early Cretaceous evaporites (4000 m).

— Late Jurassic–Early Cretaceous platform margin carbonates (1000 m).

— Early Cretaceous–Tertiary platform carbonates, slope and intraplatform basin sediments (5000–6000 m).

In the evolution of the Early Cretaceous–Tertiary carbonate platform the Mid-Cretaceous Unconformity has an outstanding role (MEYERHOFF and HATTEN 1974). This unconformity separates the carbonate platform in the West, bordered by fringing reef complexes (Yucatan, Campeche Bank, Jordan Knoll, West Cuba, Florida), from the intraplatform basin filled with pelagic sediments in the East of the Gulf of Mexico. The Cretaceous carbonate platform sequence is the main reservoir of the Mexican onshore (Yucatan, Golden Lane, Poza Rica) and offshore (Campeche Bank) petroleum reserves. Therefore during the last two decades a great amount of seismic reflection data has been gathered throughout the Gulf of Mexico. Two eminent examples from West Florida and from the Campeche Bank are presented after SCHLAGER (1991) on **Fig. 16**. In spite of the great distance, the profiles reflect identical situations: the original boundary of a carbonate platform and that of an intraplatform basin at the base of the Albian. The carbonate platform margin, bordered by fringing reefs, and back-reef lagoonal units, the talus and the intraplatform basin, infilled by slope and pelagic sediments above the Mid-Cretaceous Unconformity are very well expressed on the right sides of both profiles.

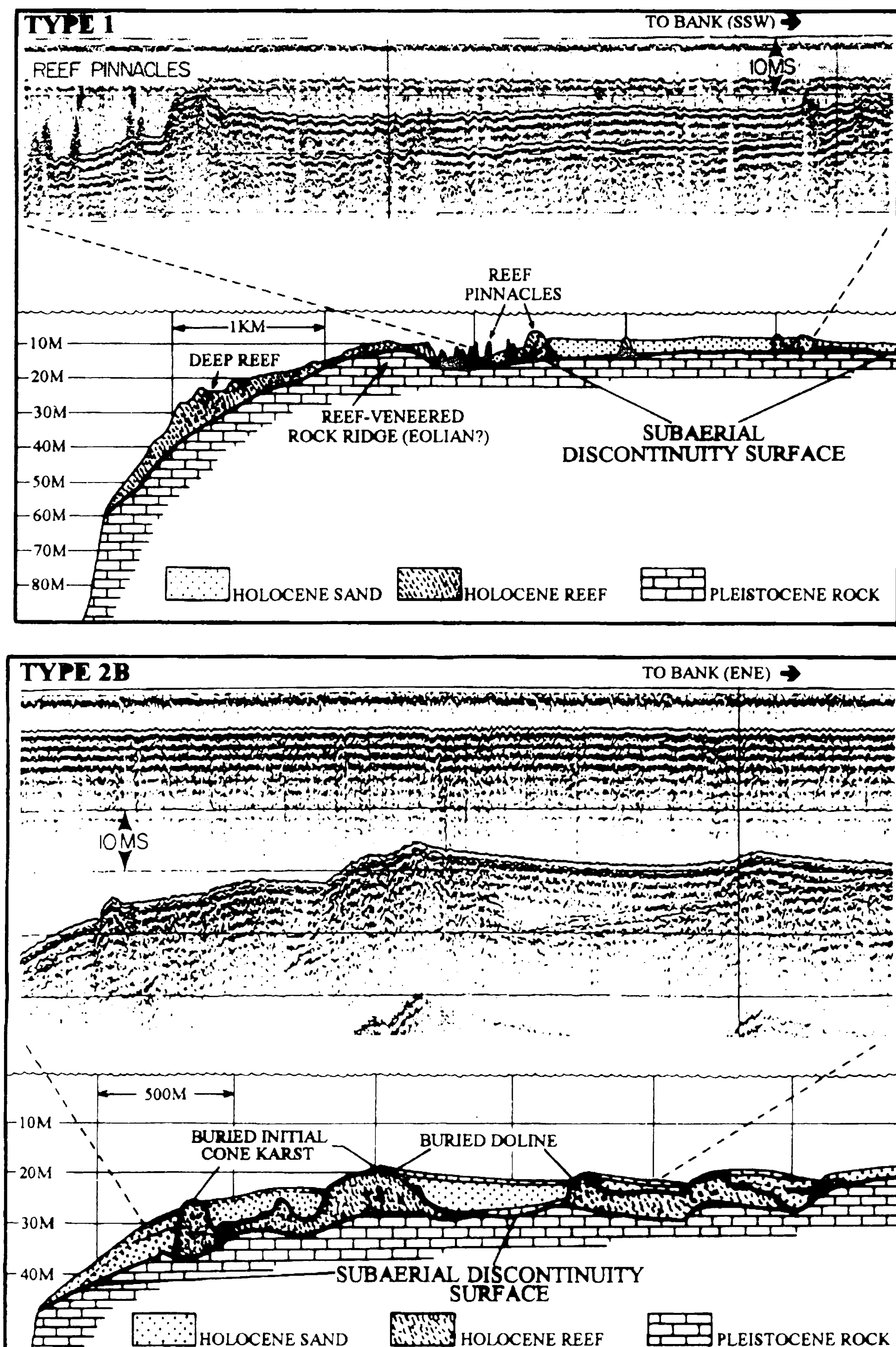


Fig. 14. Palaeokarstic interpretation of rimmed platform profiles, Little Bahama Bank (after HINE and NEUMANN 1977)

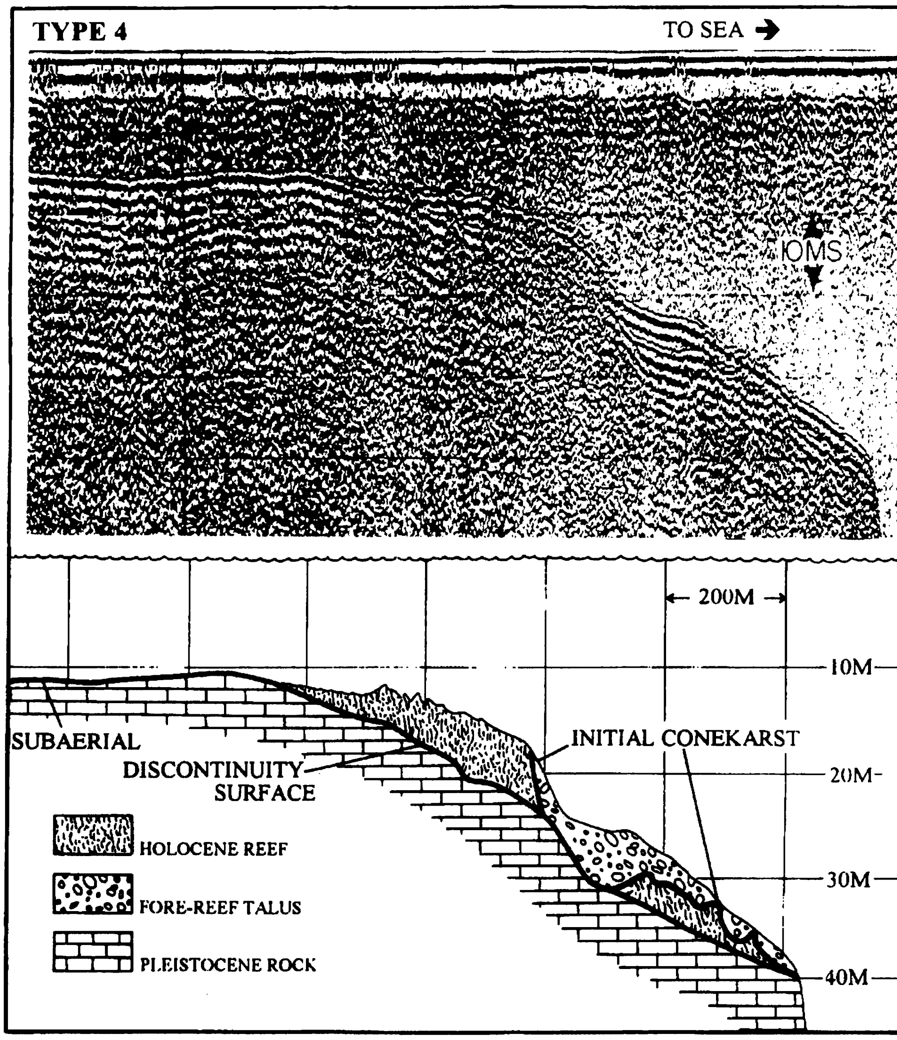


Fig. 15. Palaeokarstic interpretation of rimmed platform profiles, Little Bahama Bank (after HIŇE and NEUMANN 1977)

Palaeokarst interpretation

The seismic profiles show the internal architecture of the Cretaceous carbonate platform bordered by fringing reefs, dissected by master discontinuity surfaces and covered by pelagic sediments at the top (2.0 and 2.5 sec). The evolution and the growth of the Cretaceous carbonate platform are practically identical to recent examples of the Bahamas (**Figs. 14, 15**). The single, but great difference is in the scale. The Cretaceous profiles of West Florida and Campeche Bank represent an about 2500–3000 m thick complex, similar to those of the recent, some ten meter thick, of the Little Bahama Bank. Consequently the palaeokarstic reservoir potential of the Cretaceous platform is much greater.

2.3. Cuba

Regional geology and evolution

Cuba, as an accretionary grown neoplatform area is located at the southern border of the North American continental plate, separated from the oceanic Caribbean plate by the large scale Motagua–Cayman synistral fault (CASE et al. 1984, PINDELL 1991).

Its geological structure is illustrated after KORPÁS (1988) on **Figs. 17, 18**, briefly describing the following main phases of its geological evolution:

1. Ancient platform stage (>230 Ma).
2. Early rifting stage (?210–160 Ma) and opening of the oceanic basin (160–140 Ma).
3. First island arc stage (140–75 Ma) with its early (140–90 Ma) and late, mature phases (90–75 Ma), closed with back arc rifting (75–50 Ma).

4. Neoplatform and second island arc stages (75–45 Ma) with its early (75–55 Ma) and late, mature phases (55–45 Ma), followed by a back arc rifting (40 Ma to present).
5. Modern accretionary neoplatform stage (45 Ma to present) with the closing of the oceanic basins (45–30 Ma) and with formation of the neoplatform (30 Ma to present).

First collision and obduction (70–45 Ma). Second collision and obduction (40–25 Ma).

The tectono-stratigraphic analysis of the Cuban formations resulted in the evaluation of the palaeokarst potential of Cuba (**Fig. 18**). The estimated 15 main palaeokarstic horizons have a good correlation with the principal unconformities and coincide with the main global, eustatic sea level drops (HAQ et al. 1987).

Two examples of the Cuban palaeokarst will be described in the following (after KORPÁS and ITURRALDE VINENT 1992).

Pan de Guajaibon, West Cuba

The Upper Albian/Cenomanian diaspore bauxite deposits are a classical example of palaeokarst in Cuba (**Fig. 19**). Their detailed description is given by RAZUMOVSKIY et al. (1987) and only the major palaeokarst patterns are cited here. Within the peritidal and shallowing upward Guajaibon formation, karstification and oolitic bauxites are related to a 3rd order syntectonic unconformity. Diagnostic criteria worth mentioning are the following: shallowing upward cycles, early dolomitization in the bauxite footwall, pedogenetic features, subaerial exposure surface and

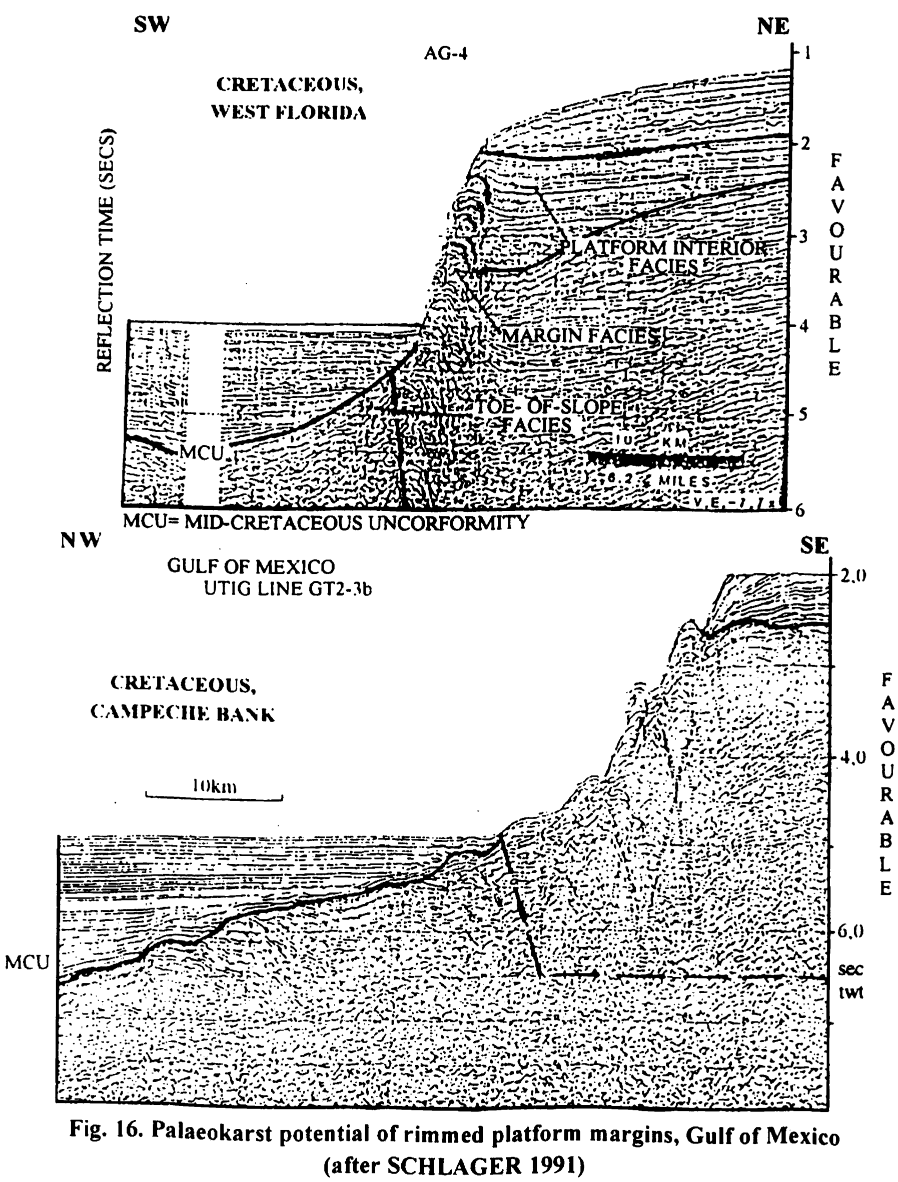


Fig. 16. Palaeokarst potential of rimmed platform margins, Gulf of Mexico (after SCHLAGER 1991)

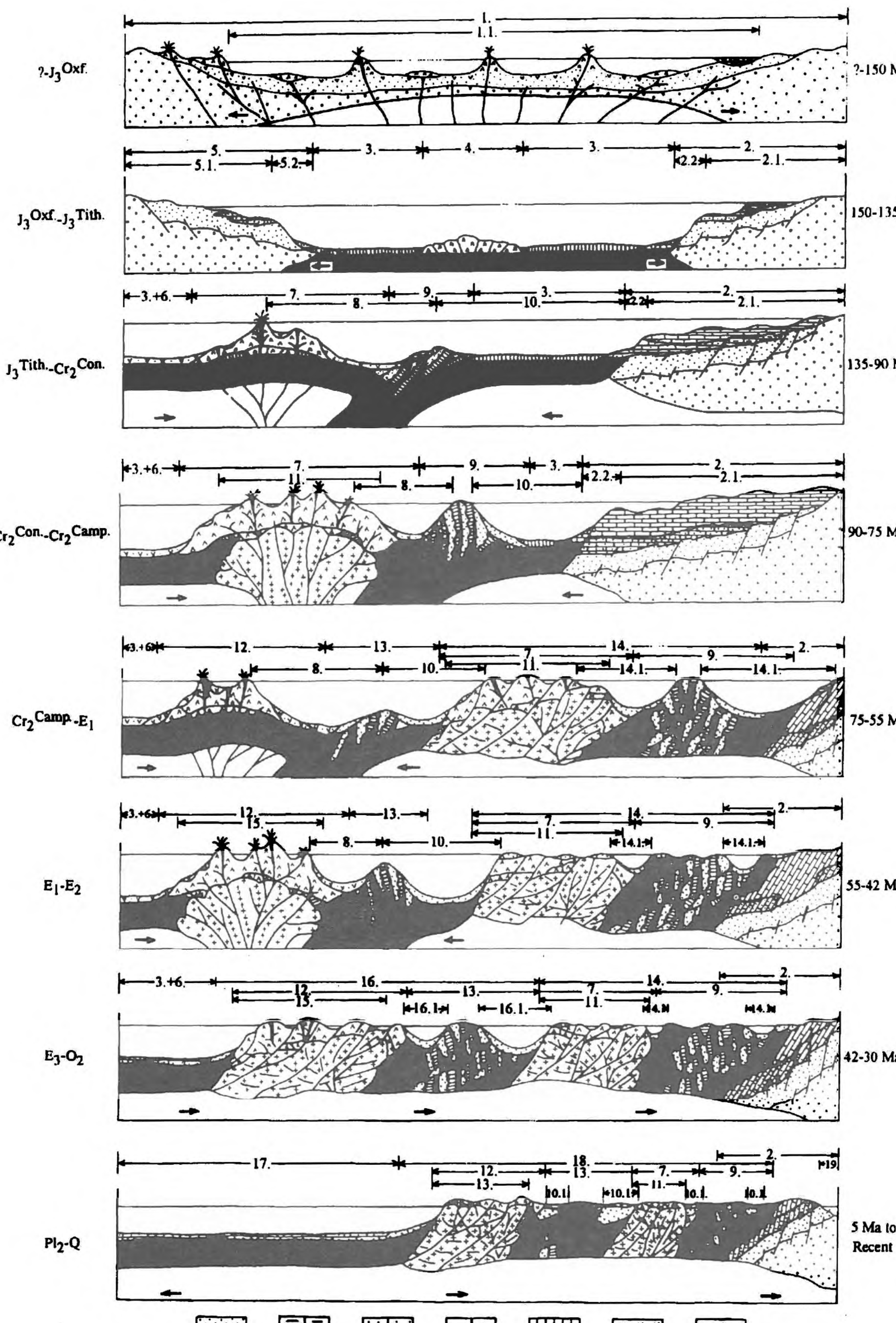


Fig. 17. Geological evolution of Cuba (KORPÁS 1988)

1. Clastics. 2. Evaporites. 3. Clastics and carbonates. 4. Carbonates. 5. Carbonates and silicites. 6. Volcanogen and sedimentary clastics. 7. Effusives. 8. Intrusives. 9. Continental crust. 10. Oceanic crust. 11. Upper mantle. 12. Relativ plate motion. 13. Faults. — 1. = Continent, 1.1. = Intracontinental sea, 2. = Continental units N, 2.1. = Bahama platform, 2.2. = Continental slope, 3. = Abyssic plain, 4. = Mid-oceanic ridge, 5. = Continental units S, 5.1. = Platform S, 5.2. = Continental slope, 6. = Back arc basin, 7. = Volcanic arc Zaza (J_3–Cr_2), 8. = Fore arc basin, 9. = Accretionary prism, 10. = Back arc basin, 11. = Cr_2 intrusives, 12. = Volcanic arc Cauto (Cr_2–E_2), 13. = Accretionary prism, 14. = Accretionary neoplatform I, 14.1. = Residual basins, 15. = E_{1-2} intrusives, 16. = Accretionary neoplatform II, 16.1. = Residual basins, 17. = Southern continental slope and abyssic plain, 18. = Composite accretionary neoplatform I+II, 18.1. = Closed residual basins, 19. = Northern continental slope and abyssic plain

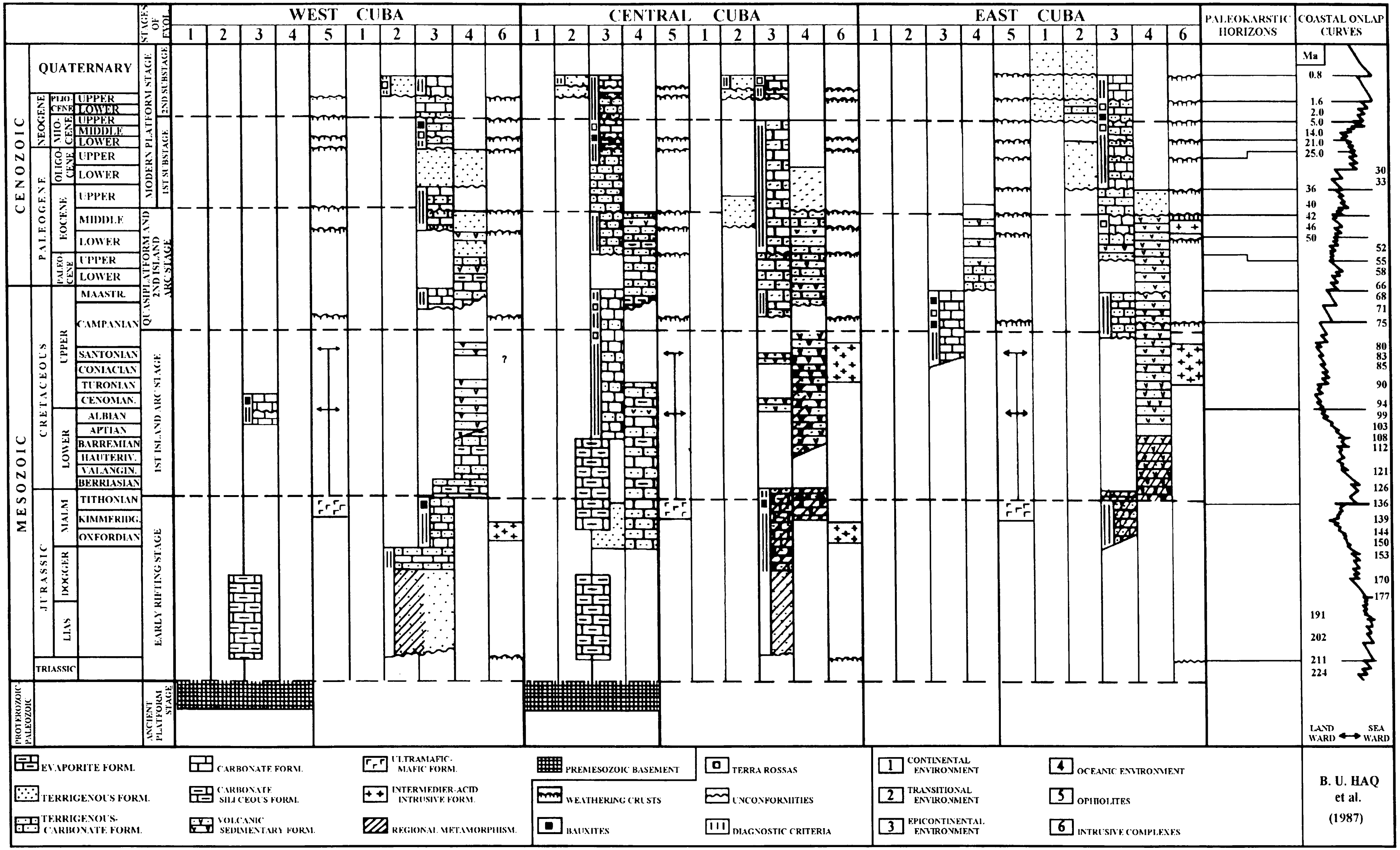


Fig. 18. Tectonostratigraphic chart of Cuba and related palaeokarst horizons

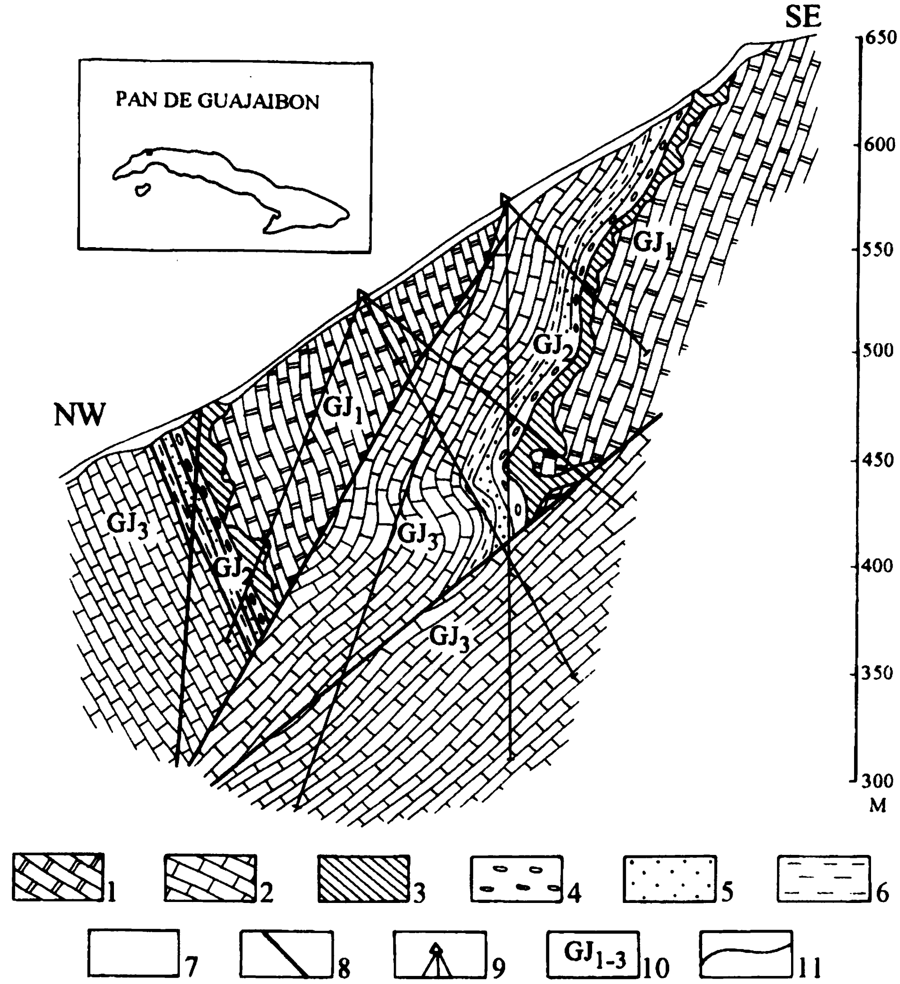


Fig. 19. Albian/Cenomanian palaeokarst horizon, Pan de Guajaibon, West Cuba (KORPÁS and ITURRALDE VINENT 1992, after RASUMOVSKIY et al. 1987)
1. Limestones, dolomites; 2. Limestones; 3. Bauxites; 4. Gravelites; 5. Sandstones; 6. Mudstones; 7. Quaternary sediments; 8. Normal and reverse faults; 9. Boreholes; 10. Guajaibon Formation; 11. Parasequence boundary

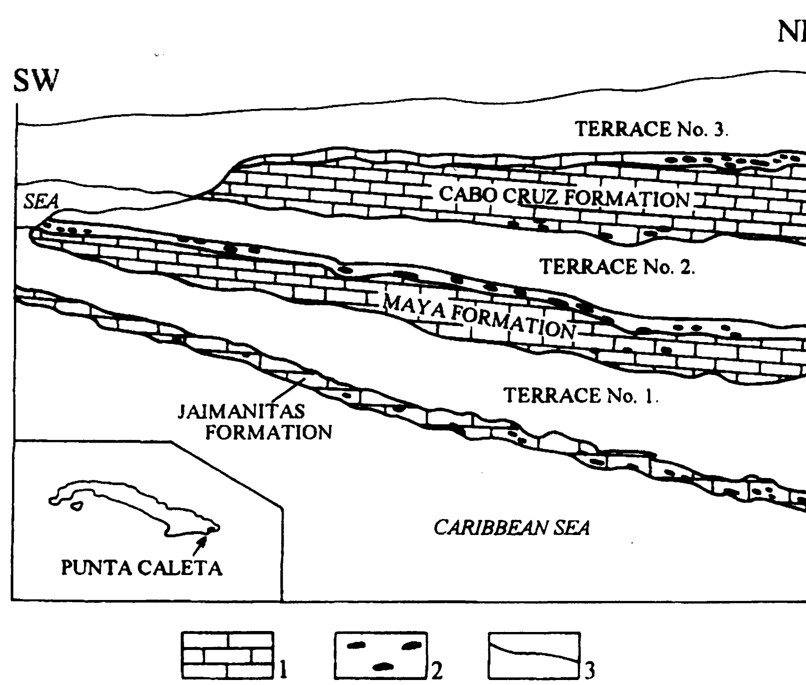


Fig. 20. Palaeokarst section in Punta Caleta, East Cuba, not to scale (KORPÁS and ITURRALDE VINENT 1992, after MARRERO 1951 and NAGY et al. 1983)

bauxite formation, monomictic–polymictic clastics with onlap contact in the hanging wall of the bauxites.

As an excellently described palaeokarst analogy the El Abra Formation in Mexico (MINERO 1988) can be cited.

Punta Maísi–Punta Caleta, East Cuba

A good example for the regressive terrace model (KORPÁS and JUHÁSZ 1993) can be found at Punta Maísi–Punta Caleta section in East Cuba, composed of 5–11 terraces and corresponding erosion surfaces (**Fig. 20**). Sporadic observations (FRANCO 1983, BREZSNYÁNSZKY et al. 1983, KORPÁS 1988) concerning diagnostic criteria for palaeokarst are the following: shows of bauxites and related palaeosols, early dolomitization and intercalations of monomictic–polymictic clastics. The shallowing upward trend can be recognized in the parasequence order of the intensively karstified peritidal limestone formations as well: Middle–Upper Miocene of Cabo Cruz, Upper Miocene–Pliocene of Maya and Quaternary to Present of the Jaimanitas. The tectonically controlled palaeokarst is overprinted by 3rd to 5th order high frequency sea level changes.

2.4. England

Geological setting

The Lower Carboniferous (Mississippian) carbonate ramp section bearing palaeokarst horizons in England (**Fig. 21**) will be discussed after WRIGHT (1988). The 35–45 m thick section of the quarry of Chipping Sodbury in the vicinity of Bristol displays the following:

— 10 m of parallel, slightly undulating bedded, oolitic grainstone (Gully oolite) rich in brachiopods, crinoids, containing rootprints and burrows. Millimetre thin joint system with red mudstone and calcite infillings and some, 3 m long, narrow V-like dissolutional pipes (**Photo 1**). The last ones are perpendicular to the bedding and they are infilled with early, marine, laminated mudstones and grainstones. The shallowing upward section is completed at the top by a bedding plane parallel and slightly irregular discontinuity surface.

— Discontinuity surface.

— 2–5 m thick, flat lens of crystalline dolomites with mud cracks. They are early, mixing zone dolomites with slight meteoric influence.

— 8–10 m of oolitic peritidal limestone and dolomites.

— Subaerial discontinuity surface.

— 12–15 m of clastic sequence: alternation of well developed, nodular, brecciated calcrete horizons (4–5 m) with medium grained, cemented, fluvial sandstones (1–2 m) The calcretes are a rhizocretion type, containing mud cracks and show a very fine joint network, infilled by dolomite and calcite.

— More than 3 m of subtidal limestones and dolomites with 5 stromatolite bearing horizons at the base.

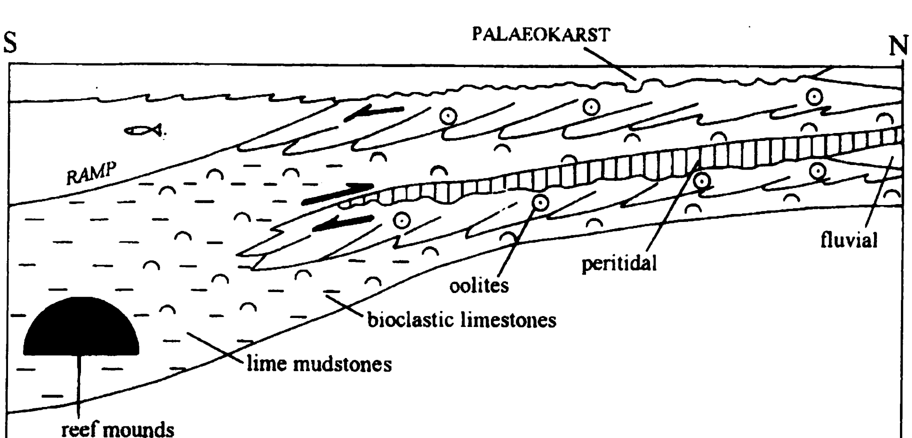


Fig. 21. Simplified cross-section of Early Carboniferous ramp in South Wales (WRIGHT 1988)

Palaeokarst interpretation

The exposed profile, as part of a homoclinal ramp sequence (WRIGHT 1988) consists of two, shallowing upward complexes, interrupted by two discontinuity surfaces and covered by subtidal carbonates at the top. Diagnostic criteria for palaeokarst: shallowing upward cycles, calcrete horizons, early mixing zone dolomites, mud cracks, dissolutional features and early marine laminite infillings, discontinuity surfaces. The first cycle of the Gully oolite represents a landward moving, shallow oolite shoal system. The sudden end of deposition of the oolites is indicated by the first discontinuity surface and by the opening of the V-like, narrow fissures. This part of the section does not show any proof for subaerial exposure and the early marine laminite infillings indicate shallow marine environment, with a maximum depth at wave base. The mixing zone dolomites with mud cracks above the first discontinuity surface show the starting of the second shallowing upward cycle in peritidal conditions. That was completed by the second, already subaerial discontinuity surface at the bottom of the first calcrete horizon. The third cycle is represented by alternation of the calcrete horizons with fluvial sediments. This means a rapid change in the depositional environment, i.e. the ramp was covered by the prograding siliciclastic lobe of river(s). The final submergence of this Caribbean type palaeokarst is indicated by the deposition of the stromatolite bearing fourth cycle of subtidal limestones and dolomites.

2.5. Spain

2.5.1. *Catalan Range*

Geological setting

The 230–300 m thick, rift related Middle Triassic Muschelkalk, as a homoclinal ramp-sequence in the Catalan Range will be discussed after CALVET et al. (1990) and CALVET and TUCKER (1995). The lower and upper units of these platform carbonates are separated by the Middle Muschelkalk evaporites and marls. The Lower Muschelkalk represents a broad, transgressive sequence, from peritidal (El Brull Formation) to lagoonal (Olesa Formation) and to sand shoal (Vilella Baixa Formation) environments. These facies are capped by a regional subaerial disconformity surface and related multiple palaeokarst horizon (**Fig. 22**). This latter one is covered by the intertidal-supratidal-sabkha dolomites of the Colldeju Formation.

The Upper Muschelkalk consists of tidal, lagoonal and outer ramp sediments (Rojals oolites, Benifallet Formation) covered by the Querol stromatolites and by the La Riba mud mound-reef complex. The deposition of the stromatolite and reef units was terminated by another regional subaerial exposure, resulting in the formation of a new palaeokarst horizon, covered completely by the Alcover dolomites.

Palaeokarst interpretation

The palaeokarst horizons as sequence boundaries indicate a relative sea-level fall, in the magnitude of 50 m, and their estimated ages are: 238 Ma and 235 Ma (CALVET et al. 1990, CALVET and TUCKER 1995).

The brief description of the composite multiple palaeokarst horizon of the Lower Muschelkalk is cited according to CALVET et al. (1990):

"The main palaeokarst level has a morphology of horizontal cavities from a few centimetres to more than 3 m in height and from a few metres to more than 100 m in length. The cavities are filled by clays and silts, red to brown in colour, with centimetre–decimetre angular blocks from cave-roof collapse. The surface morphology is smooth to angular. A vuggy porosity occurs in the carbonates close to the main palaeokarst horizons. The cavities are interpreted as caves developed in meteoric–phreatic environment."

One section of this palaeokarst horizon, suggested as a flat doline is shown on **Photo 2**.

2.5.2. *Mediterranean coast, Cap Salou*

Geological setting

One of the best examples of composite palaeokarst, related to a regional unconformity in platform sequences is described by SALAS. (1990) from the Catalan Coastal Range, near Tarragona (**Fig. 23, Photo 3**). Middle and Late Jurassic platform dolomites, with a thickness of a few 100 m developed in onshore and offshore areas and consist of massive, brecciated and bedded-laminated types of rock. The

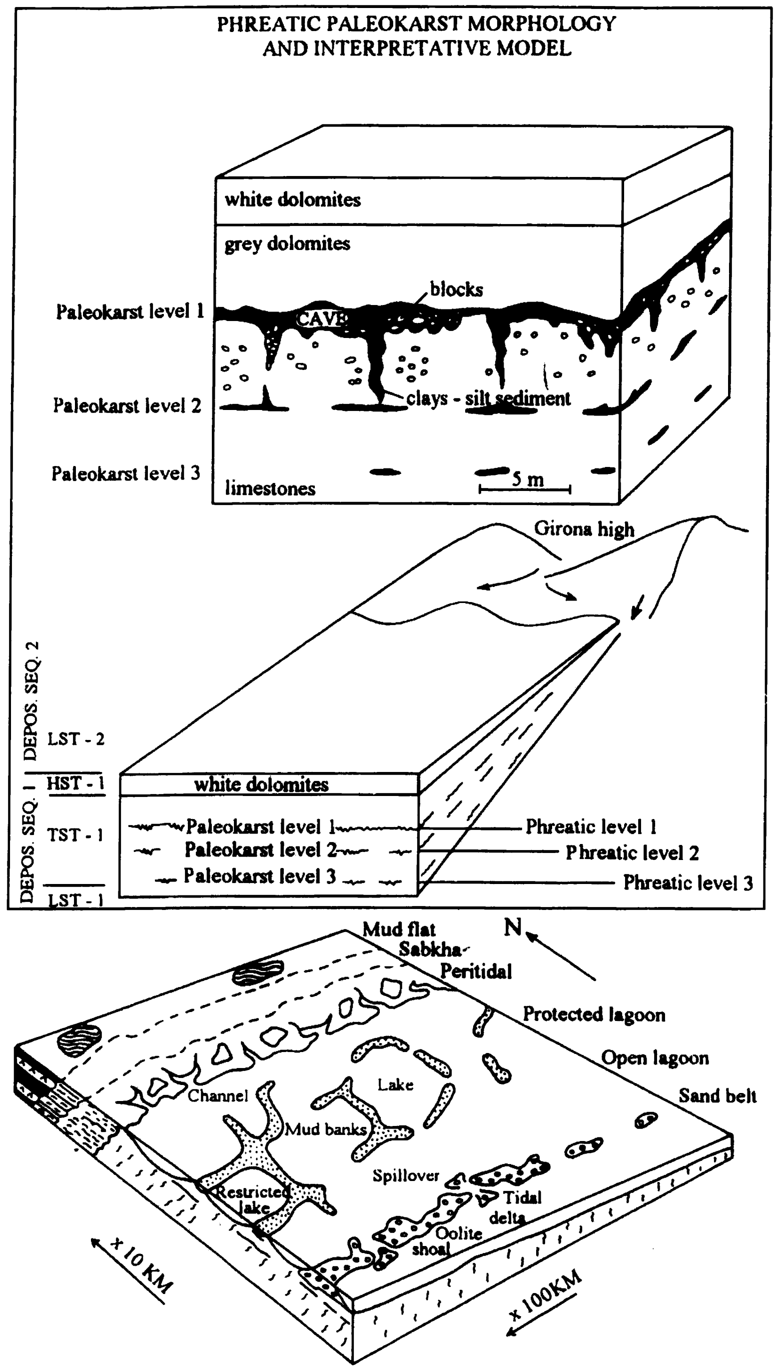


Fig. 22. Phreatic palaeokarst morphology in the Lower Muschelkalk and interpretative model related to the depositional system tracts (above). Interpretative environmental model for El Brull, Olesa and Vilella Baixa Units in the Lower Muschelkalk (below) (CALVET et al. 1990)

Callovian–Portlandian Garraf Dolomites are peritidal to supratidal dolomicrites with mud cracks. They are dissected by tidal flat breccias and thin red palaeosol layers. The top of the Garraf Formation is capped by an iron crust, connected to the subaerial exposure surface of the regional unconformity. The unconformity is covered by the Upper Barremian–Aptian shallow marine limestones of the Canteperdius and Artoles formations. The basal palaeosol layers and iron crusts alternate with frequently brecciated and dolomitized wackestones. The middle part of the section is dissected by another, internal palaeokarst surface, parallel to the bedding.

Palaeokarst interpretation

The profile described above is a classical example of composite and superimposed palaeokarst surfaces. Although the Callovian–Upper Barremian unconformity

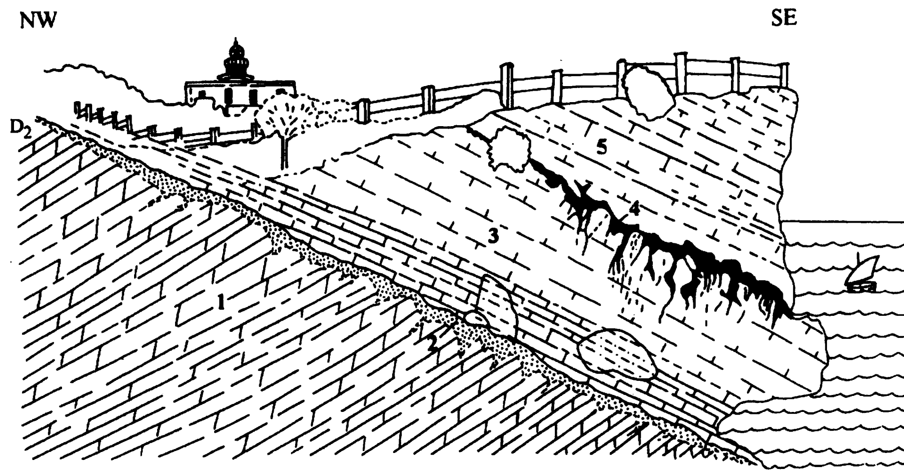


Fig. 23. Mid-Cretaceous composite palaeokarst section, Cap Salou, Spain (SALAS 1990)
1. Garraf Dolomite (Callovian–Portlandian), 2. Subaerial ironcrust, 3. Canteperdius Formation (Barremian–Aptian), 4. Palaeokarst horizon, 5. Artoles Formation (Barremian–Aptian)

is considered to be the master surface, evidence for subaerial palaeokarst can be observed below and above it too. This evidence, such as dolomitization, mud cracks, palaesol layers, iron crusts, opened joint systems with early infillings in the footwall, lateritic palaeosols, internal palaeokarst level in the hanging wall is very common. The relative succession of the superimposed palaeokarst phases is the following:

1. Middle to Late Jurassic, partly subaerial depositional palaeokarst in the Garraf Dolomites, below the unconformity.
2. Late Jurassic/Early Cretaceous subaerial nondepositional palaeokarst at the level of the unconformity.
3. Early Cretaceous depositional palaeokarst in the Cantaperdius and Artoles formations, above the unconformity.
4. Recent karstification at sea-level.

A similar palaeokarst surface, developed at the unconformity of the Liassic dolomites (Cortes del Tajuna Formation) and of the Langhian rocky shore marine sediments is shown on **Photo 4**.

2.5.3. *Mediterranean coast, Mallorca, Cap Blanc*

Geological setting

The Late Miocene Cap Blanc barrier reef, reconstructed by POMAR (1991) is one of the nicest fossil reefs in the world. The reef complex, prograded to SW is exposed in a coastal cliff, 2 km long and elevated 60–80 m above the present day sea-level (**Fig. 24**). The reef consists of the sigmoidal coset of prograding, aggrading and offlap units, corresponding to 3rd order sea-level changes. It was covered still in the Early Pliocene by the marine Son Mir Calcisiltites rich in Ammussium and planktonic foraminifera and by Pleistocene eolianites with palaeosol layers. Both the Late Miocene reef and the Pleistocene eolianites contain individual cave horizons, parallel to bedding. Some of these caves, developed in the reef-core, contain marine and planktonic foraminifera bearing sediments, corresponding to the Son Mir Calcisiltites (**Photo 5**).

Palaeokarst interpretation

The evolution of the Cap Blanc reef was terminated by emersion at about 6,3 Ma ago. This resulted in Caribbean type of karstification, with a marine phreatic cave level. Because of a later sea-level rise the whole reef was submerged and in the Early Pliocene certain parts of the caves were infilled by marine sediments. The tectonically controlled consequent sea-level falls were recorded by the Pliocene–Pleistocene uplift of this palaeokarst horizon, about 50 m above the present day sea-level.

2.6. Greece

2.6.1. *Peninsula Peloponnessos, Korfos*

The palaeokarst bearing Late Triassic Pantokrator Formation will be briefly discussed after HAAS and SKOURTSIS CORONEU (1995). The cyclic loferites of this formation (Late Norian to Rhaetian in age), correspond probably to the Dachstein Limestone. These loferites consist of supratidal A members (palaeosols, breccias), of

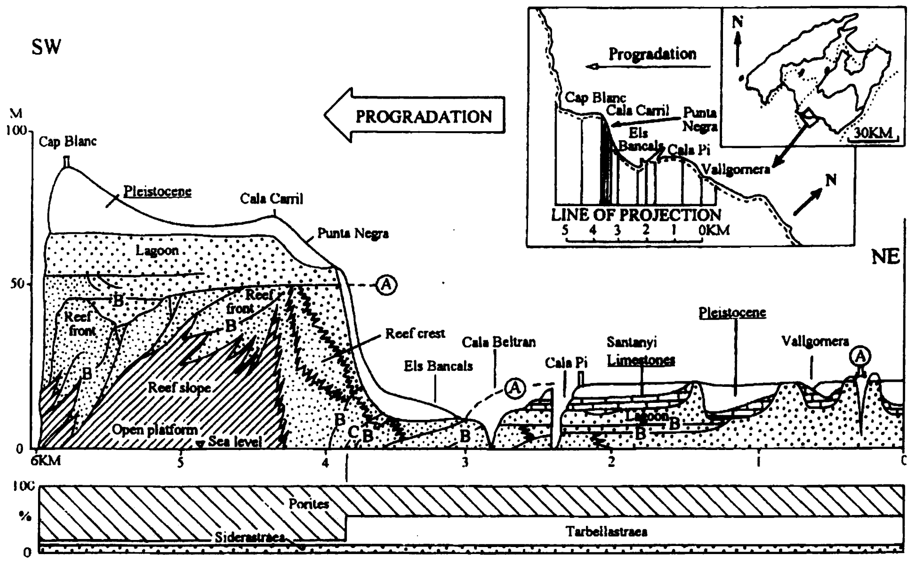


Fig. 24. Late Miocene palaeokarst section, Cap Blanc, Spain (POMAR 1991)

peritidal B members (laminated fenestral micrites with early diagenetic dissolutional features) and of subtidal C members (mudstones, packstones and grainstones). Evidence for palaeokarst are the following: 4th–5th order discontinuity surfaces and related palaeosols, parallel to the bedding; dolomitization; mud cracks; early dissolutional fabric, mouldic, fracture and vuggy porosity. Similar short term subaerial exposure periods of about 500 to 2000 years in duration were estimated in the Dachstein Limestone of the Northern Calcareous Alps, Austria (SATTERLEY, 1996) and od the Transdanubian Range, Hungary (BALOG et al. 1997). The section of the formation, bearing palaeosol horizons is presented on **Photo 6**.

3. *Palaeokarst and carbonate platforms*

3.1. Fundamentals

The foreign examples drawn from different areas of the Earth have illustratec what kind of genetic relations should be proven between the evolution of the car- bonate platforms and of the related palaeokarst systems. Among the discussed exam- ples both recent and old or fossil occurrences can be found. But their scales in time and in space are very different and the intensity of the karstification differs too. Theii single common pattern is the Caribbean type of karstification. This marine, phreatic karstification of the cited examples is proven partly by early marine infilling gener- ations and cements, and partly by the shallow marine sediments both in the hanging wall and in the footwall of the buried palaeokarst horizon. The shown examples are also illustrating, that time has no significant role in the formation and in the evolu- tion of morphologically perfect karst systems. It means that the formation of a com- plete palaeokarst level requires no more than a few 1000 years. At the same time, ir the case of composite and multiphase palaeokarst systems, frequently associated witl 1–2 order unconformities except that the time gap at the unconformity is limited to a few million years. But it does not necessarrily follow that this is the time needed foı karstification.

Consequently the new and basic element of the suggested model is the linking up of the evolution of carbonate platforms with the palaeokarst systems belonging tc them. As the cyclicity governed by a number of factors plays the decisive role in the evolution of both of them, the spatial distribution of the palaeokarst horizons inside the carbonate platforms has to reflect the periodicity of the Milankovich (12 00(years), of the precession (19–23 000 years), of the obliquity (41 000 years) and of the composite (100 000, 400 000) cycles.

3.2. The model

Although among the cited examples both rimmed platform and ramp types occur the proposed model (**Fig. 25**) was elaborated for the more attractive rimmed platforn type.

The model itself comprises a complete Milankovich cycle, starting at the relative sea-level "0 m" (s_1), with the correspondent depositional environments and karst ir the mixing and in the phreatic zones.

During the first sea-level fall of 5 m (s_2) the subaerial tidal flat areas and the relat- ed discontinuity surface (D) are increasing and the area of the lagoon is decreasing The relevant karst level (Nr. 2) is coming over to the former one and the karstifica- tion in the mixing zone is stepping gradually seaward.

The next sea-level drop of a new 5 m (s_3) has resulted in the further increase o the subaerial tidal flat areas, and the relevant karst level (Nr. 3) has developed below the former ones. So far the process is accompanied by gradual but continuous increase in porosity and correlative sediments are not to be expected in the karst sys- tem. Consequently this is a nondepositional karst period with the formation of indi- vidual and composite, mainly phreatic karst levels.

In the course of the following sea-level rise (s_4) the former depositional and kars systems fall almost completely below the sea-level. The subaerial tidal flats areas wil

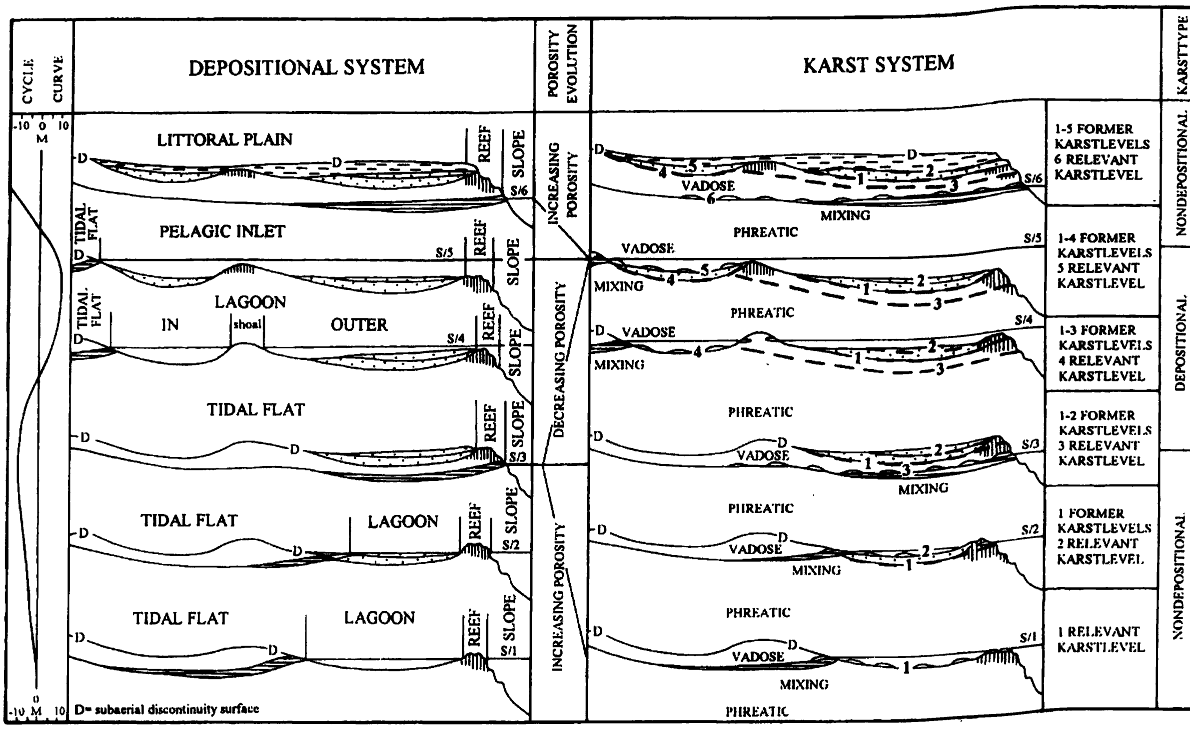


Fig. 25. Rimmed carbonate platform evolution and related karst development

be significantly reduced, while new inner and outer lagoons, separated by a shoal will be formed. This means that the former karst levels (Nr. 1–3) will be completely located below the sea-level and the relevant karst level (Nr. 4) will be situated landward.

Because of the further sea-level rise (s_5) the former depositional system will be covered by pelagic sediments and the relevant karst level (Nr. 5) will be restricted to the narrower subaerial tidal flat areas. The former karst levels (Nr. 1–4) will take their place in the shallow marine phreatic zone. These phases, accompanied by sea-level rise result in the decrease of porosity, reflected by the partial infilling of the karst system. That is the depositional period in the evolution of the karst system and proof of the highstand is given by the shallow marine infillings and cements.

Finally the cycle will be closed by rapid sea-level fall (s_6) and the former depositional system will be completely changed into a subaerial exposure surface resulting in the formation of a littoral karst plain bordered by terraces. For the evolution of the karst system this is a new, significant phase and the relevant karst level (Nr. 6) will be located below the former ones (1–5). This nondepositional period means again the increase of porosity with a new, independent karst level and the overprinting of the former ones.

Conclusions 3.3.

1. According to the explained model the evolution of the palaeokarst systems has to be in a strong correlation with the development of the carbonate platforms. The evolution of both is cyclic, but the great difference between them lies in the fact that the depositional cycles are followed by subsequent diagenetic palaeokarst cycles with a regular phase-shift in time.

2. The sea-level falls represent phases of increasing porosity and they are nondepositional karst periods.

3. The sea-level rises represent phases of decreasing porosity and they are depositional karst periods.

4. The caves, formed in these palaeokarst systems are marine phreatic in origin and the bottom surface of the individual caves is parallel to the bedding.

5. In the formation of the palaeokarst levels a greater role is attributed to mechanical abrasion, produced by the submarine earth tidal pump, than to mixing corrosion.

4. Palaeokarst systems in Hungary

4.1. Overview

The beginnings of the study and exploration of palaeokarst systems in Hungary can be traced back to the discovery of karst bauxite deposits at Halimba and at Gánt in the Transdanubian Range (GYÖRGY 1923). Since that time the most important research activity on "palaeokarsts" has been connected to continuous prospecting and exploration for bauxite, carried out in the Transdanubian Range and in the Villány Mts. These have resulted in the recognition of the main bauxite bearing "palaeokarst" horizons, tied to regional unconformities. This means that from the earliest syntheses (TELEGDI ROTH 1923, 1927a, VADÁSZ 1930, 1935, 1946, 1951) to more modern ones (BARNABÁS et al. 1957, SZANTNER and SZABÓ 1970, BÁRDOSSY 1977, 1982, SZANTNER et al. 1986) the number of bauxite bearing "palaeokarstic" horizons discovered in Hungary has increased from two (i.e. Early Cretaceous, Palaeocene–Early Eocene) to seven (i.e. Early Cretaceous, Middle Cretaceous, Late Cretaceous, Palaeocene, Early to Middle Eocene, Late Eocene, Middle to Late Miocene). An important statement on palaeokarsts was made by SCHAFARZIK and VENDL (1929), who interpreted the Late Eocene "strandwall" conglomerates, found in the Szikla chapel cave of the Gellért Hill (Budapest, Danube bank) as products of rocky shore marine karstification during the Late Eocene.

Even the term "fossil karst", corresponding to the recent use of palaeokarst was introduced by FÖLDVÁRI (1933), who discussed the pre-Eocene karst, related to the bauxite, manganese and coal deposits in the Transdanubian Range (Halimba, Gánt, Eplény, Úrkút, Dorog, Budakeszi).

A similar situation to the Gellért Hill's one was described by KRIVÁN (1959) at Csillaghegy, near Budapest, who distinguished three superimposed "fossil karst" levels. The first two, located in the Late Triassic Dachstein Limestone, was related by him to Young Mesozoic unconformities, accompanied with formation of caves and bauxites. The third, developed at the top of the earlier ones, was considered by KRIVÁN as a "syngenetic" Early Bartonian "marine fossil karst", formed on a "rocky shore" and produced by "wave turbulence".

The present day term of palaeokarst was introduced by SZABÓ (1956, 1957, 1964, 1968), who had used it for the description of the multiphase evolution of covered fossil karsts in the Midmountains of Hungary (Transdanubian Range, Mecsek, Bükk, Aggtelek). The eight fossil karst levels, estimated by him are linked to regional unconformities: 1. Early Cretaceous (Barremian), 2. Late Cretaceous to Early Eocene, 3. Early Tertiary (Eocene to Oligocene), 4–6. Late Tertiary (4=Early Miocene, 5=Sarmatian to Lower Pannonian, 6=Late Pannonian), 7. Quaternary and 8. Holocene. Later KORPÁS (1980) discovered Late Triassic bauxite shows in the Transdanubian Range, indicating subaerial exposure of the Norian–Rhaetian carbonate platform. KRAUS (1988) was the first to report early marine "watchglass" infilling sediments (laminites) of Late Eocene age in the Mátyás-hegy cave (part of the hydrothermal cave system in the Buda Hills), developed also in Late Eocene limestones.

Before the first appearance of the "diagenetic" school (ESTEBAN and KLAPPA 1983), the above mentioned had been the first pathfinders of this new karst research.

Since the end of the 80's this new approach has been introduced, developed and increasingly applied in Hungary. Systematic palaeokarst research and studies have started at the Geological Institute of Hungary (in 1989), at the Geological Departments of the Eötvös Loránd University (in 1990), at the Hungarian Hydrocarbon Institute (in 1990), at the Technical University of Budapest (in 1990, 1992) and at the Speleological Institute (in 1993).

The second comprehensive study of the palaeokarst of Hungary, using mainly the proper data and experiences of previous bauxite exploration was done by BÁRDOSSY and KORDOS (1989), estimating Early Cretaceous, Albian, Turonian–Early Senonian, Palaeocene–Early Eocene, Oligocene and Neogene to Recent independent palaeokarst horizons in Hungary, linked to regional unconformities.

The two or three phase hydrothermal evolution of the palaeokarst system in the Buda Hills was discussed by MÜLLER (1989). Systematic regional, stratigraphic, genetic studies and modelling of the multiphase and composite palaeokarst systems of Hungary started in the 90's. The topics and the scope of these studies will be briefly related in the following:

— Genetic models (KORPÁS 1990, KORPÁS and JUHÁSZ 1990, 1991, 1993).

— Genetic case studies on Middle and Late Triassic, Early Jurassic, Late Cretaceous, Late Eocene, Miocene, Pliocene and Quaternary palaeokarst systems in the Transdanubian Range (Balaton Highland, Keszthely Mts., South Bakony Mts, Gerecse Mts, Buda Hills, Naszály Hill, Csővár block) and in the Mecsek–Villány Mts. (ESTEBAN and JUHÁSZ 1990, MINDSZENTY 1990, 1992, TÖRÖK and SZABÓ-BALOG 1990, NÁDOR 1991, 1992a, b, NÁDOR and SÁSDI 1991, KORPÁS et al. 1992, KLEB et al. 1993a, b, KORPÁS and DUDKO 1993, KORPÁS et al. 1993, NÁDOR et al. 1993, SÁSDI 1993, LANTOS 1994, 1995, RÁLISCH-FELGENHAUER 1994, HAAS 1998, JOCHA-EDELÉNYI 1995, 1995, LELKES and BUDAI 1995, NAGY 1995, TÖRÖK 1997).

— Carbonate platform modelling in the Transdanubian Range (BALOG et al. 1997).

— 3D models of palaeokarst systems of the Buda Hills (KORPÁS 1994a, b, c, KORPÁS and NAGY 1994a, b).

— Palaeokarst potential of Hungary (KORPÁS 1995a, b).

4.2. Selected genetic case studies

Selected genetic case studies of Hungary will be discussed in the following. The location of the studied profiles is shown on **Fig. 26**.

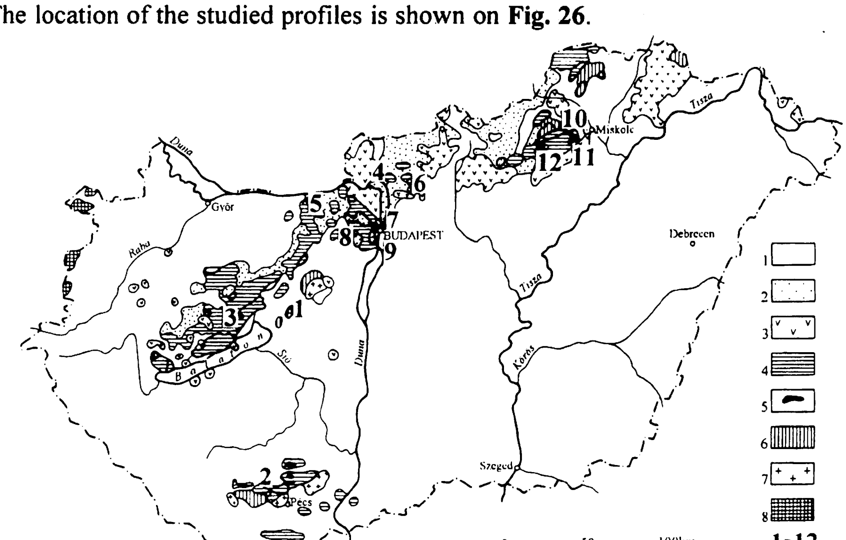


Fig. 26. Location map of the studied palaeokarst profiles in Hungary
1 = Szabadbattyán (Kőszár-hegy), 2 = Orfű (Sárkány-kút), 3 = Litér–Hajmáskér, 4 = Vác (Naszály), 5. = Pisznice cave, 6 = Csővár, 7 = Buda Hills (Rózsadomb), 8 = Buda Hills (Páty), 9 = Buda Hills (Várhegy), 10 = Bükk Mountains (Lillafüred), 11 = Bükk Mountains (Miskolc-Tapolca), 12 = Bükk Mountains (Felsőtárkány) — 1. Quaternary and Neogene sediments, 2. Paleogene sediments, 3. Tertiary volcanics, 4. Mesozoic sediments, 5. Mesozoic eruptives, 6. Palaeozoic sediments, 7. Palaeozoic intrusives, 8. Crystalline schists

4.2.1. *Szabadbattyán (Kőszár-hegy)*

Geological setting

The Middle Devonian platform sequence of the Polgárdi Limestone will be described after LELKES-FELVÁRI (1978), HORVÁTH and ÓDOR (1989) and FÜLÖP (1990). The folded crystalline limestone is composed of cyclic Lofer facies, overprinted by equigranular xenomorphic–hypidiomorphic textures metamorphic in origin. The original depositional and diagenetic features, like loferites, mud cracks, fenestrae and early dolomitization are frequently preserved despite recrystallisation. The limestone is poor in fossils, some individual corals and alga-horizons, including weakly developed stromatolites have been mentioned by FÜLÖP (1990) from the Kőszár-hegy quarry. The depositional system is interpreted by him as a shallow peritidal carbonate bank.

The limestone is cut by the narrow dikes of Early Permian quartzporphyrites and by the shallow intrusive bodies of Middle Triassic porphyritic andesites (HORVÁTH and ÓDOR 1989). Beside the diagenetic features mentioned above, hydrothermal and metasomatic alteration can be observed, such as skarns related contact metamorphism and metasomatism. The products of this alteration are: silicification, iron metasomatism with manganese, surface and subsurface galena mineralization, formation of marble, brucite–serpentinite mineral assemblages and skarns of vesuvianite-diopside-garnet type.

Palaeokarst features and interpretation

Some of the palaeokarst features of the Polgárdi Limestone have been well known for a long time (KORMOS 1911, KISS 1951, BÁRDOSSY and KORDOS 1989). KISS (1951) was the first, who has described the karstic and brecciated patterns of the hydrothermal galena mineralization discovered in the caves and cavities of the mine galleries. KORMOS (1911) published an excellent profile of the Polgárdi cave, infilled with Pliocene lacustrine sediments and eolian loess, rich in vertebrate remains (i. e. the famous fauna of Polgárdi). Similar sites were discovered and mentioned by KORDOS (in: BÁRDOSSY and KORDOS 1989), who has dated the depositional record of these infillings from the Late Miocene–Pliocene to the Quaternary.

Beside the above features the following types of infillings and generations have been observed by us in the quarry.

Vadose infilling sediments: clast supported, autoclastic, collapse breccias, encrusted by limonite in dissolutional pipes and cavities; oolitic iron laterite crust with autoclasts and waad-powder in cavities; iron rich palaesol layers, alterning with limestone; limonitic popcorn generations, precipitated on the cavity-walls.

Phreatic infilling sediments and precipiations: horizontal bedded limonitic laminites of grainstones and mudstones with collapse breccias in cavities (**Photos 7, 8**); radiaxial calcite of three generations on the cavity-walls.

It is evident that the infilling types and generations enumerated above reflect a long term palaeokarst evolution, interrupted by hydrothermal karst events. But it seems rather difficult to estimate their relative order of succession and it is even more complicated to determine their ages. Our proposed model for the relative order of succession is the following:

Phase 1 — Early, syndepositional, subaerial, coastal palaeokarst, resulting in the formation of the iron rich palaesols.

Phase 2 — Early, syndepositional, marine, phreatic palaeokarst with disconformable generations of laminites.

Phase ? — Subaerial, partly depositional palaeokarst with the formation of iron laterites, waad powder and with the limonitic popcorn generations.

Phase ? — Phreatic, marine palaeokarst with radiaxial calcites.

Phases 3–4 — Hydrothermal palaeokarst with MVT type mineralization (galena) and earlier collapse karst breccias.

Phase 5 — Subaerial palaeokarst clastic infillings of lacustrine and eolian origin, rich in vertebrate fossils.

For the age dating of this long term, multiphase and composite karst evolution we have accepted the following tie-points:

— The age of the syndepositional phases 1 and 2 is Devonian.

— The age of the hydrothermal phases 3 and 4 is supposed to be pre Middle Triassic, because the Middle Triassic andesites, according to HORVÁTH (pers. comm.) cut the MVT mineralization.

— The Late Miocene age (7.1 to 5.3 Ma, zone MN 13) of the depositional phase No. 5 is proved by biostratigraphic data (BÁRDOSSY and KORDOS 1989, HORACEK and KORDOS 1989).

The ages of the suaberial, partly depositional and phreatic, marine phases should be fitted pre-hydrothermal or subsequent to the hydrothermal palaeokarst event.

4.2.2. *Orfű (Sárkány-kút)*

The early marine palaeokarst infilling (**Fig. 27**), formed at the boundary of the Bertalanhegy Limestone and of the Dömörkapu Limestone was discovered by KONRÁD in 1993. The cavity is located in the bioclastic, brachiopod and crinoid bearing, bioturbated grainstone of the Bertalanhegy Limestone. It is elongated, parallel to the bedding and completely filled by conformable laminites of dolomitized grainstones and mudstones, formed below sea-level. Water escape structures, possibly produced by coeval earthquakes can be observed within the laminites. The palaeokarst is covered by the dolomitized mudstones of the Dömörkapu Limestone. Both limestones had been deposited in the open shelf, shelf-slope environment of a Middle Triassic carbonate ramp. The related syndepositional palaeokarst, formed at the level of the wavebase should be interpreted as submarine phreatic.

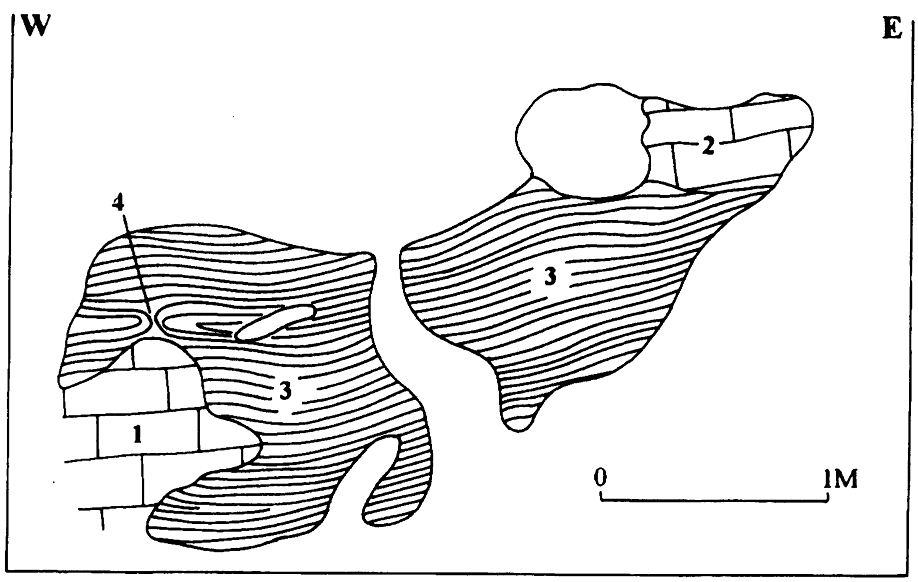


Fig. 27. Anisian early marine laminites with water-escape structures in the cavities of the Bertalanhegy Limestone, Orfű, Sárkány-kút

1. Bertalanhegy Limestone, 2. Dömörkapu Limestone, 3. Early marine laminites, 4. Water-escape structure

4.2.3. *Balaton Highland (Litér, Hajmáskér)*

Geological setting

The palaeokarst bearing Middle Triassic, disintegrated dolomite ramp sequence will be described after SZABÓ and RAVASZ (1970), BUDAI (1992), BUDAI and VÖRÖS (1992, 1993), BUDAI et al. (1993) and HAAS et al. (1993), according to the stratigraphic chart of **Fig. 28**.

Megyehegy Dolomite

Concerning its *lithology*, the major part of the formation (250 m thick) consists of massive and bedded saccharoidal, recrystallised dolosparites with scarce oncoidal–ooidic horizons. It is poor in fossils, except the rare occurrences of foraminifera, green algae, crinoids and brachiopods. The uppermost 20–25 m is com-

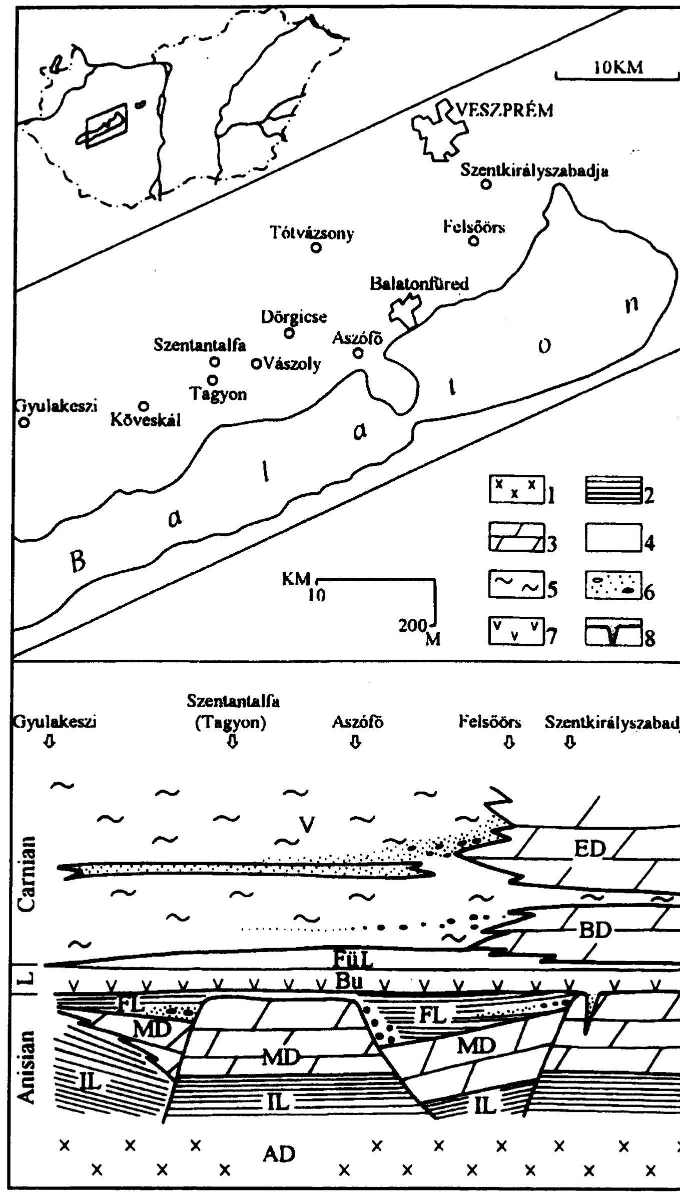


Fig. 28. Stratigraphic chart of the Middle Triassic, Balaton Highland (BUDAI et al. 1993)

1. Sabkha, 2. Restricted (periodically anoxic) basin, 3. Carbonate platform, 4. Open shelf basin, 5. Intrashelf basin with terrigenous clastics, 6. Allodapic clastics, 7. Pyroclastics, 8. Neptunian dyke; Anisian: AD=Aszófő Dolomite, IL=Iszkahegy Limestone, MD=Megyehegy Dolomite+Tagyon Limestone, FL=Felsőörs Limestone; Ladinian: Bu=Buchenstein Fm.; Carnian: FüL=Füred Limestone, V=Veszprém Marl, BD=Budaörs Dolomite, ED=Ederics Dolomite

posed of shallowing upward cycles of massive to laminated dolomites, dissected by thin volcanomictic palaeosol layers (**Figs. 29, 30, 31**).

Among the *sedimentological and early diagenetic features* the following will be outlined: sharp lower and gradual upper contact of the cycles; narrow edge stuctures, cm to dm in size, infilled with sediments and controlled by synsedimentary microfaults below the lower contact of the cycles; presence of mudcracks, intraclastic horizons, including the little fanglomerate-tongue composed of graded dolomite sand in the Litér quarry (**Fig. 29**); synsedimentary slumps, controlled by microfaults; dominantly polymodal nonplanar to planar dolomicrosparite–dolosparite textures frequently with coated grains and bioturbation; multiphase fracture system infilled mainly with dolomite and red clays, sometimes with calcite; geopetal structures of some cm in size; vuggy and mouldic porosity (50–500 μ) partly infilled with dolomite, pyrite, haematite rarely by anhydrite and gypsum.

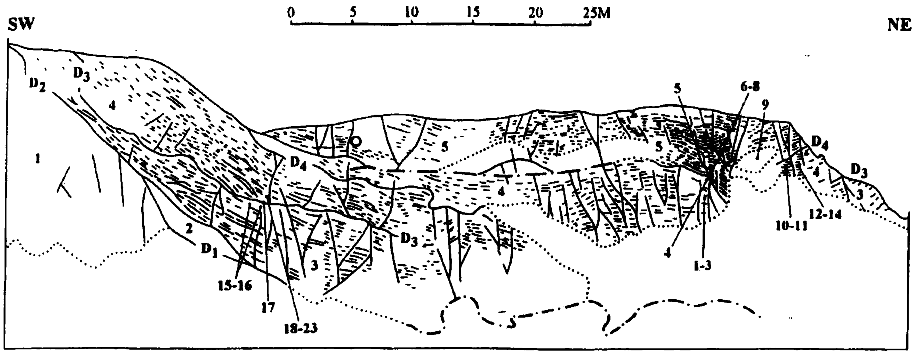


Fig. 29. Palaeokarst profile No. 1 of the Litér quarry. Buried palaeodoline in the Megyehegy Dolomite (KORPÁS and DUDKO 1993)

1. Megyehegy Dolomite, massive, 2. Fanglomerate-tongue of dolomitesand, 3–4. Megyehegy Dolomite with palaeosol layers, 5. Berekhegy Limestone, D = Discontinuity surface. ● = Site of sampling, O = Occurrences of Protrachyceras

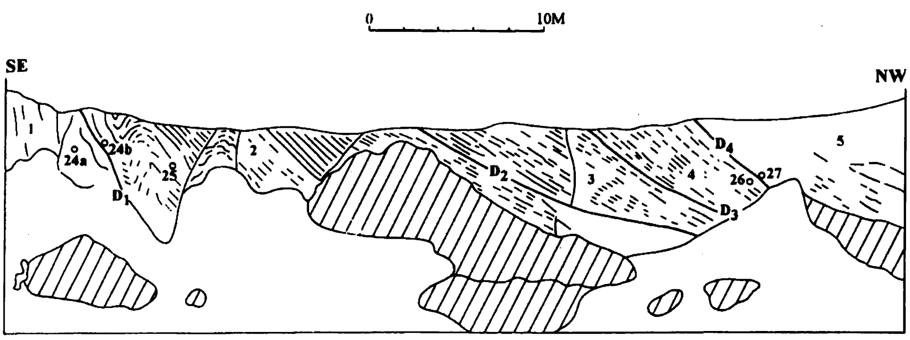


Fig. 30. Palaeokarst profile No. 2 of the Litér quarry. Transition of dolomite platform–pelagic inlet–dolomite platform (KORPÁS and DUDKO 1993)

1. Megyehegy Dolomite, massive, 2. Berekhegy Limestone, 3–4. Dolomite with palaeosol layers, 5. Massive dolomite, D=Discontinuity surface, O27 Site of sampling

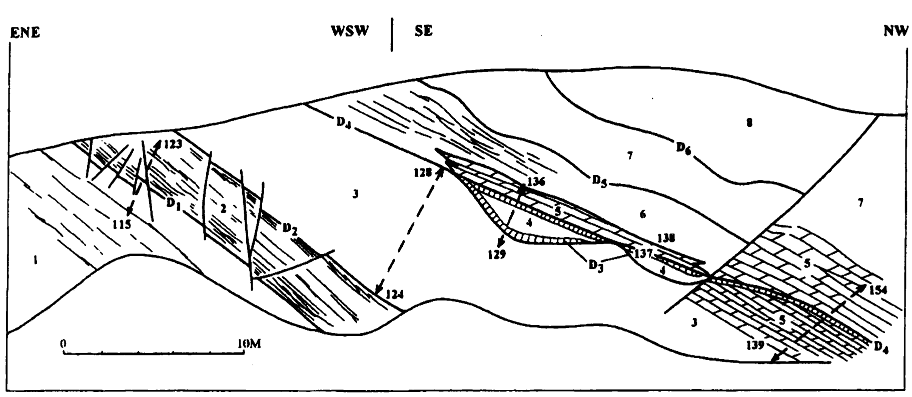


Fig. 31. Palaeokarst profile of the Hajmáskér quarry. Subaerial discontinuity surface in the Megyehegy Dolomite (KORPÁS and DUDKO 1993)

1. Massive dolomite, 2. Laminated dolomite with palaeosol layers, 3. Massive dolomite, 4. Palaeokarstic pocket infilled by dolomite and palaeosol, 5. Laminated dolomite, 6. Poorly bedded dolomite with palaeosol layers, 7. Massive dolomite, 8. Brecciated dolomite, D=Discontinuity surface, 114–128 Profile of sampling

The *depositional system* of the Megyehegy Dolomite can be characterized after the microfacies studies done by TÓTH-MAKK. The lower part of the formation was formed in slightly hypersaline environments of a shallow marine ramp. Its upper shallowing upward part represents a wide tidal flat, covered by hypersaline sabkhas, and dissected by narrow tidal channels.

Among the *palaeokarst features* the cm–dm size single and the m size composite subaerial discontinuity surfaces, exposed in both quarries are to be mentioned first (**Figs. 29, 30, 31**). The number of composite discontinuity surfaces is 4–7. They are controlled by faults and reflect internal erosion, resulting in the formation of a flat doline or karstic pockets, infilled by bauxitic palaeosols. Besides the vuggy and mouldic porosity, one can frequently observe some dm wide open joints infilled by autoclastic, well cemented dolomit-breccia. The orientation of these palaeokarstic joints is mainly of NNW–SSE coinciding with the trend of the synsedimentary faults described by DUDKO (1993). The whole section is intersected by thin layers of volcanimictic poorly developed palaeosols, rich in kaolinite and with traces of alumogoethite (**Figs. 32, 33**). These should be considered as latosols, with relative accum-

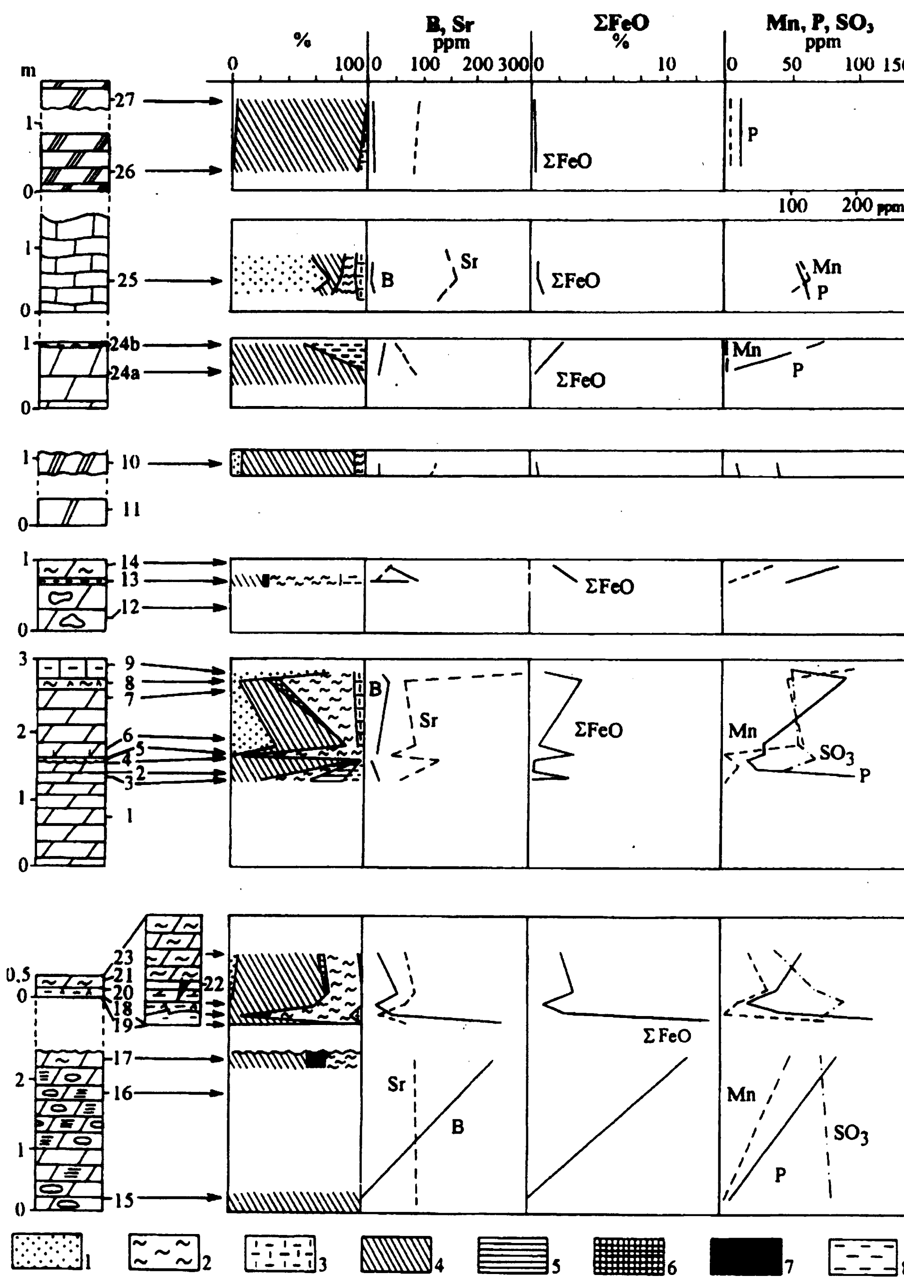


Fig. 32. Palaeokarst profiles of the Litér quarry (TÓTH MAKK in KORPÁS and DUDKO 1993)

1. Calcite. 2. Montmorillonite. 3. Quartz. 4. Dolomite. 5. Illite. 6. Feldspar. 7. Alumogoethite. 8. Kaolinite

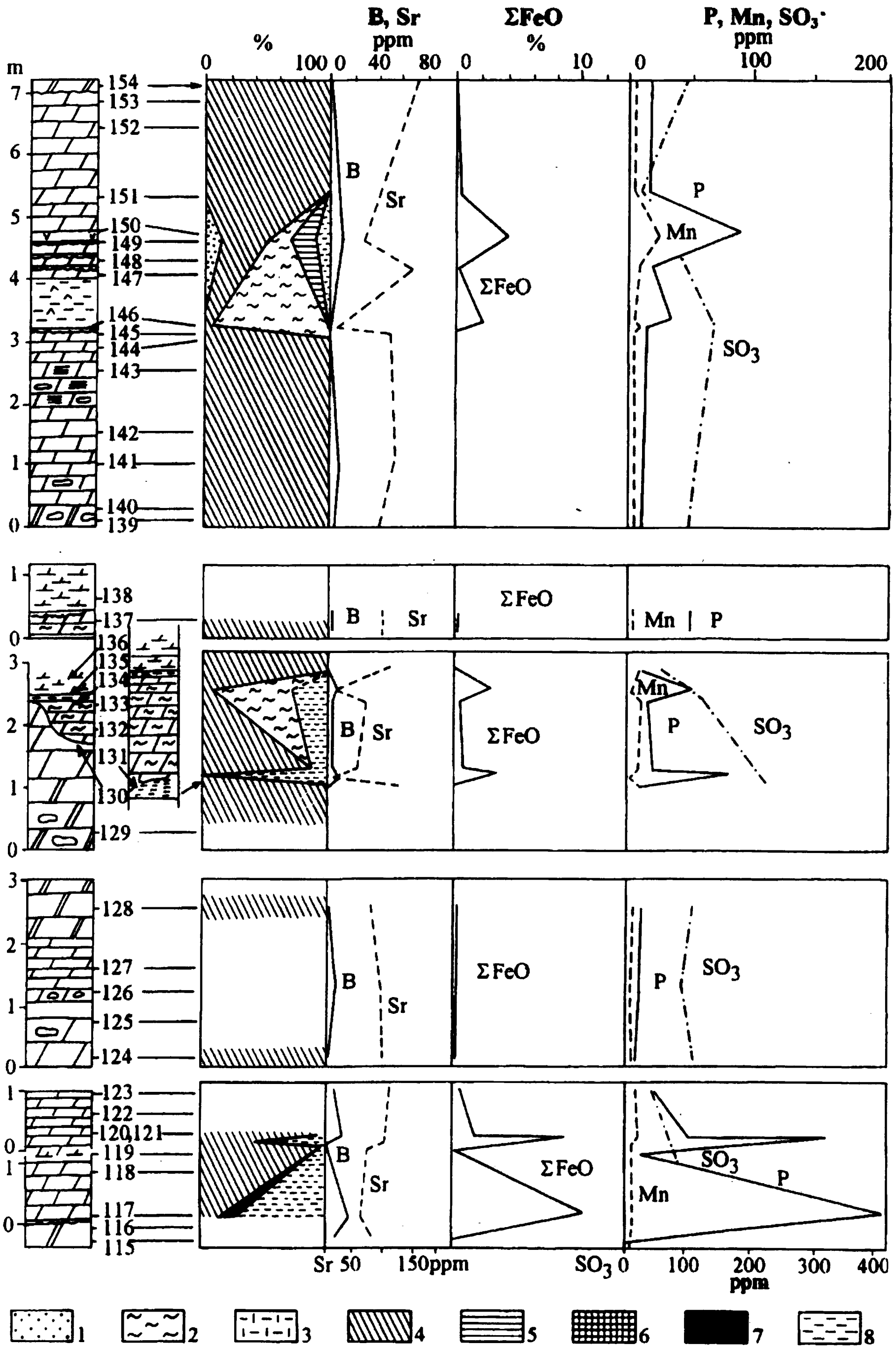


Fig. 33. Palaeokarst profile of the Hajmáskér quarry (TÓTH MAKK in KORPÁS and DUDKO 1993)

1. Calcite, 2. Montmorillonite, 3. Quartz, 4. Dolomite, 5. Illite, 6. Feldspar, 7. Alumogoethite, 8. Kaolinite

mulation of total iron (FeO: 2.9–13.0%), of alumina (Al_2O_3: 22.6–35.6%), of phosphorus (P_2O_5: 0.1–0.2%) and of boron (B: 130–270 ppm). The volcanomictic origin of these palaeosols is proven by their elevated content (11–93%) of montmorillonite (BOGNÁR) and by their high potassium range (2.0–3.0%) too. Results of more detailed X-ray and thermoanalytical studies of these palaeosols suggest that they have been deposited after eolian transport in slightly alkaline environment (with values of pH 7.4–8.15), indicating a subtropical climate with annual mean temperatures of 23–27 °C and seasonal precipitations of 750–1000 mm (KOVÁCS PÁLFFY pers. comm.). The stable isotope analysis (HERTELENDI et al. 1992, 1993) of the bed rocks and of the palaeokarstic infillings reflects mainly the marine environment with a slight meteoric influence (**Fig. 34**).

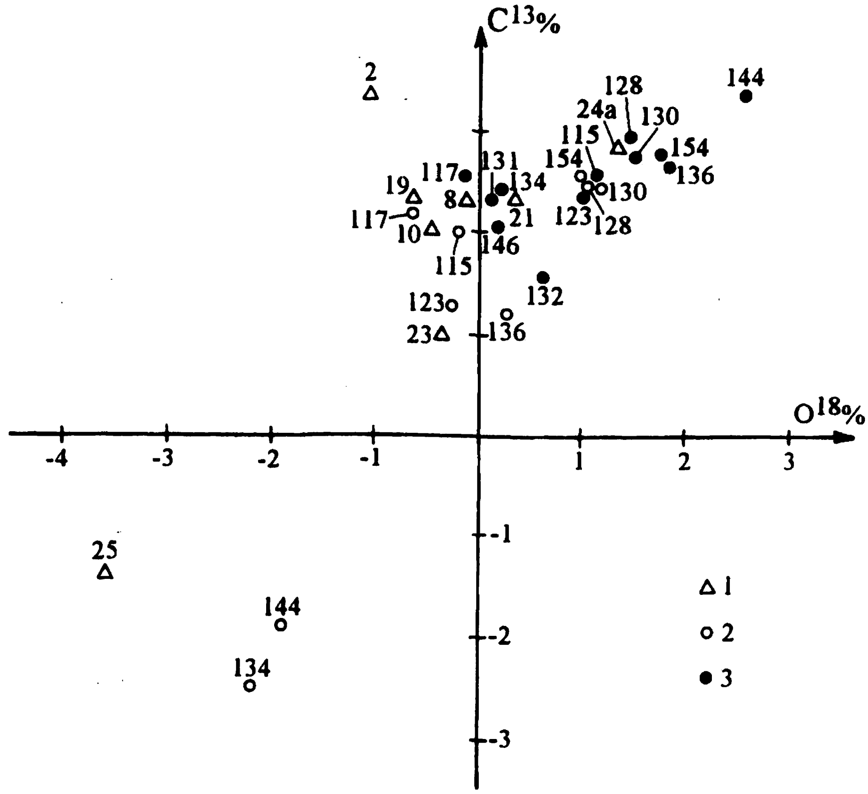


Fig. 34. Stable isotope composition of the Litér and Hajmáskér samples (KORPÁS and DUDKO 1993 after HERTELENDI et al. 1993)

1. Litér sapmles, 2. Hajmáskér samples (measured on calcite), 3. Hajmáskér samples (measured on dolomite)

Berekhegy Limestone

The *lithology* of the 10–12 m thick, folded and faulted sequence, situated at the top of the karstified Megyehegy Dolomite (**Figs. 29, 30**) consists of platy, aphanitic, nodular dolomites, dolomitized limestones and limestones, dissected by thin layers of softy marls and clays. The poor fossil ensemble is represented mainly by redeposited bioclasts (fragments of molluscs, thin-shelled bivalves, rare ostracods) less than 1.5 mm in size. A single, completely dolomitized ammonite was found in the Litér quarry (**Fig. 29**) by HAAS and identified by VÖRÖS as Protrachyceras archelaeus (LAUBE). The lower surface of the beds is sharp, irregular and presents early dissolutional features and stylolites, while their upper contacts are gradational.

The following phenomena can be related to the *sedimentological and early diagenetic features:* nodular bedding surfaces; intrabioclastic intercalations; frequent slump structures, controlled by synsedimentary faults; early stylolites and progressive dolomitization; multigenerational fracture systems infilled by white calcite; slight and poorly developed microvuggy and mouldic porosity.

The *depositional environment* is interpreted by TÓTH-MAKK as pelagic, represented by filament-bearing microsparites.

Palaeokarst features, except progressive dolomitization, like early stylolites and the slight porosity had not been recognized.

Palaeokarst interpretation

The interpretation of palaeokarst features will be illustrated with profiles, according to their scheme of correlation (**Fig. 35**).

Profile 1 (Litér quarry — **Fig. 29**): It is considered a palaeodoline, formed on the Megyehegy Dolomite platform exposed subaerially about 235 Ma ago. The flooding of this slightly asymmetric palaeodoline, 80 m in diameter and 10m in depth started by the deposition of a little dolomite–fanglomerate, followed by the alternation of hypersaline sabkha dolomites and palaeosols. The palaeodoline, becoming flatter was completely buried at 234 Ma by the pelagic Berekhegy Limestone. The palaeokarst evolution was controlled partly by 3rd order composite and superimposed discontinuity surfaces (D_1–D_4) and partly by synsedimentary faults. This tectonic control can be observed in the asymmetric flower structure of the pelagic cover too.

Profile 2 (Litér quarry — **Fig. 30**): The profile is interpreted as a transition of platform dolomites→pelagic inlet→platform dolomite, dissected by 3rd order, partly subaerial discontinuity surfaces (D_1–D_4). The Megyehegy Dolomite platform was subaerially exposed because of the sea-level fall at about 234 Ma (D_1) and covered "immediately" because of the sudden sea-level rise, producing the shallow burial of the palaeodoline. This high stand period of relatively stable water table was ended at about 232 Ma. After that datum the new shallowing upward cycle started, interrupted again by 3[rd] order discontinuity surfaces (D_{2-3}) and related palaeosols between 232–230 Ma. The progradation and stabilization of the new dolomite platform is dated by the subaerial exposure (D_4) at 230 Ma.

Profile 3 (Hajmáskér quarry — **Fig. 31**): The profile represents the shallowing upward part of the Megyehegy Dolomite platform, dissected by various simple and composite subaerial discontinuity surfaces (D_1–D_6). The D_{3-4} composite one is considered to be the master surface at about 237 Ma. This reflects a break and internal erosion, accompanied by the first steps of the process of bauxitization and episodic influence of fresh waters in the karst system.

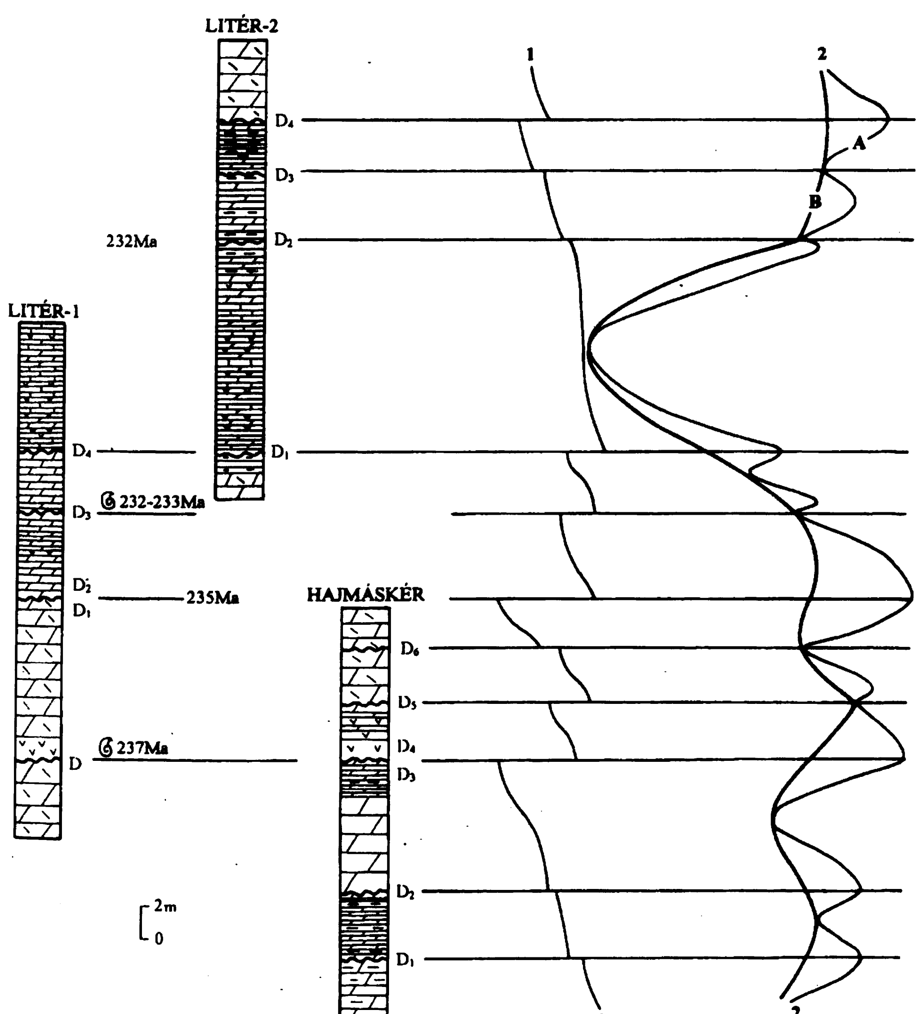


Fig. 35. Correlation of the Litér and Hajmáskér palaeokarst profiles (KORPÁS and DUDKO 1993)

1. Coastal onlap curves, 2. Sea-level curves (A: short-term, B: long-term), D=Discontinuity surface

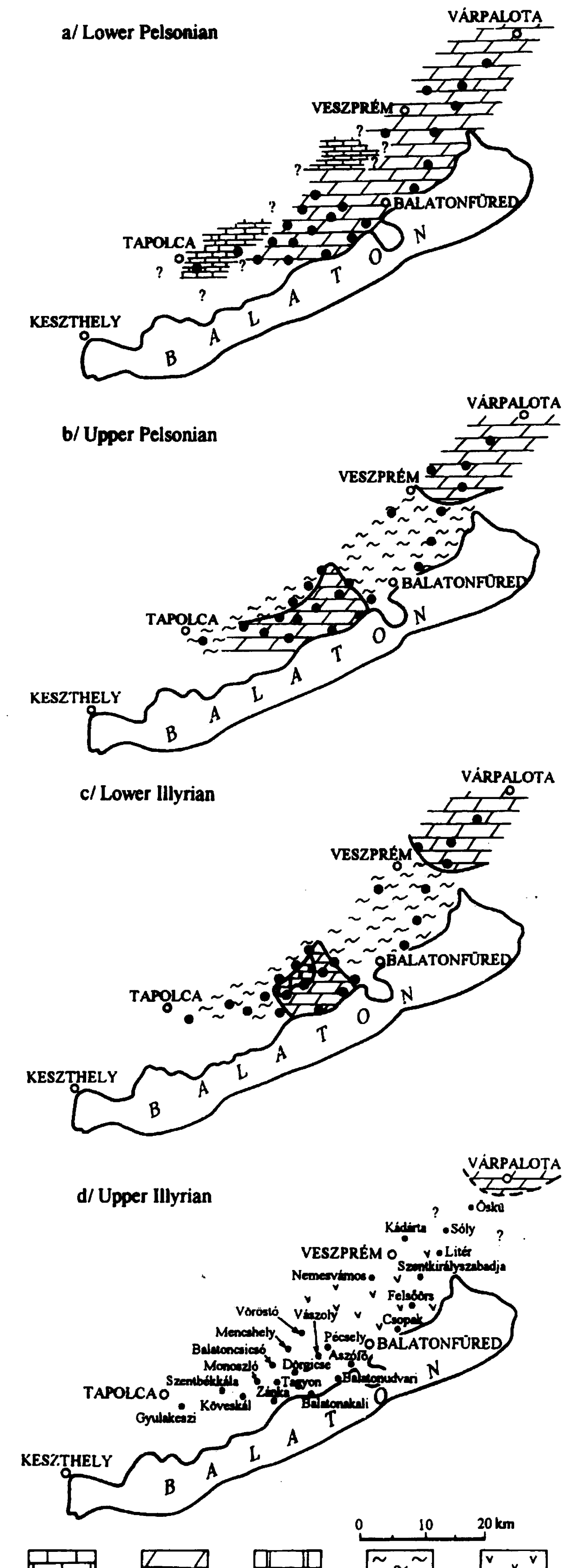


Fig. 36. Middle–Late Anisian palaeogeography, Balaton Highland (BUDAI and VÖRÖS 1992)

1. Iszkahegy Limestone, 2. Megyehegy Dolomite, 3. Tagyon Limestone, 4. Felsőörs Limest(
5. Buchenstein Formation

The history of the described profiles, at the regional scale will be outlined after BUDAI and VÖRÖS (1992), BUDAI et al. (1993) and illustrated by **Figs. 36, 37**.

Accordingly the main phases of evolution are the following:

239–238 Ma: Formation of the homoclinal ramp of the Megyehegy Dolomite.

238–237.5 Ma: Disintegration of the dolomite ramp, initial opening of the Felsőörs intraplatform basin. Platform-segments on the NE and SW border of the basin continued to exist.

237.5–237 Ma: Continuous opening of the Felsőörs basin, accompanied by the gradual subsidence of the SW platform-segment and by the starting of volcanism.

237–233 Ma: Maximum opening of the Felsőörs basin, definite subsidence of the SW platform-segment, episodic subsidence of the NE platform-segment. End of volcanism.

233–229 Ma: Closing of the Felsőörs basin, accompanied by the gradual progradation of the NE platform-segment.

The discussed palaeokarst profiles are located at the SW border of the NE platform-segment and they represent the timespan of 239–230 Ma. In their evolution governed by global and regional sea-level changes (**Figs. 35, 38**) the following cycles can be reconstructed:

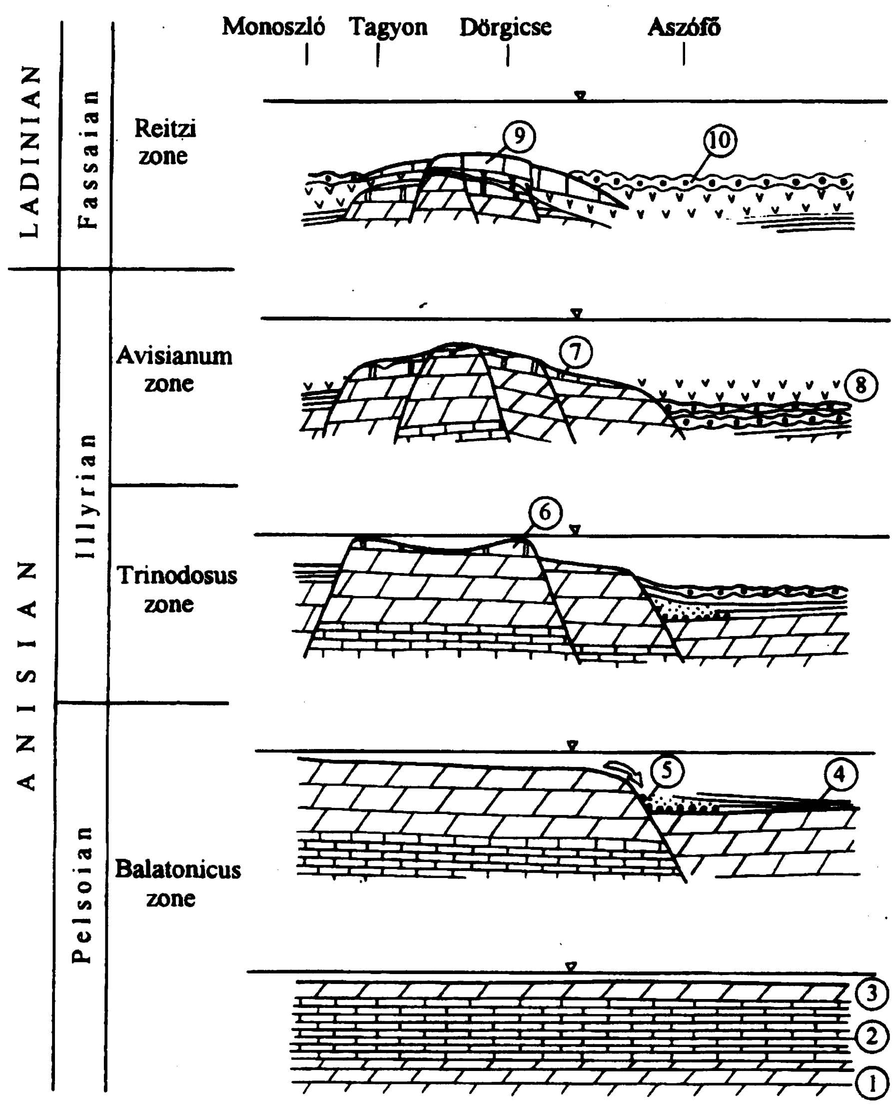


Fig. 37. Middle Triassic platform evolution, Balaton Highland (BUDAI and VÖRÖS 1992)

1. Aszófő Dolomite 2. Iszkahegy Limestone 3. Megyehegy Dolomite 4–5. Felsőörs Limestone: 4. ammonitic laminated limestone 5. brachiopodal intraclastic limestone (slump) 6. Tagyon Limestone 7–10. Buchenstein Formation: 7. crinoideal, ammonitic limestone, dolomite 8. tuff, tuffite (pietra verde) 9. "Vászoly Limestone" 10. "Nemesvámos Limestone"

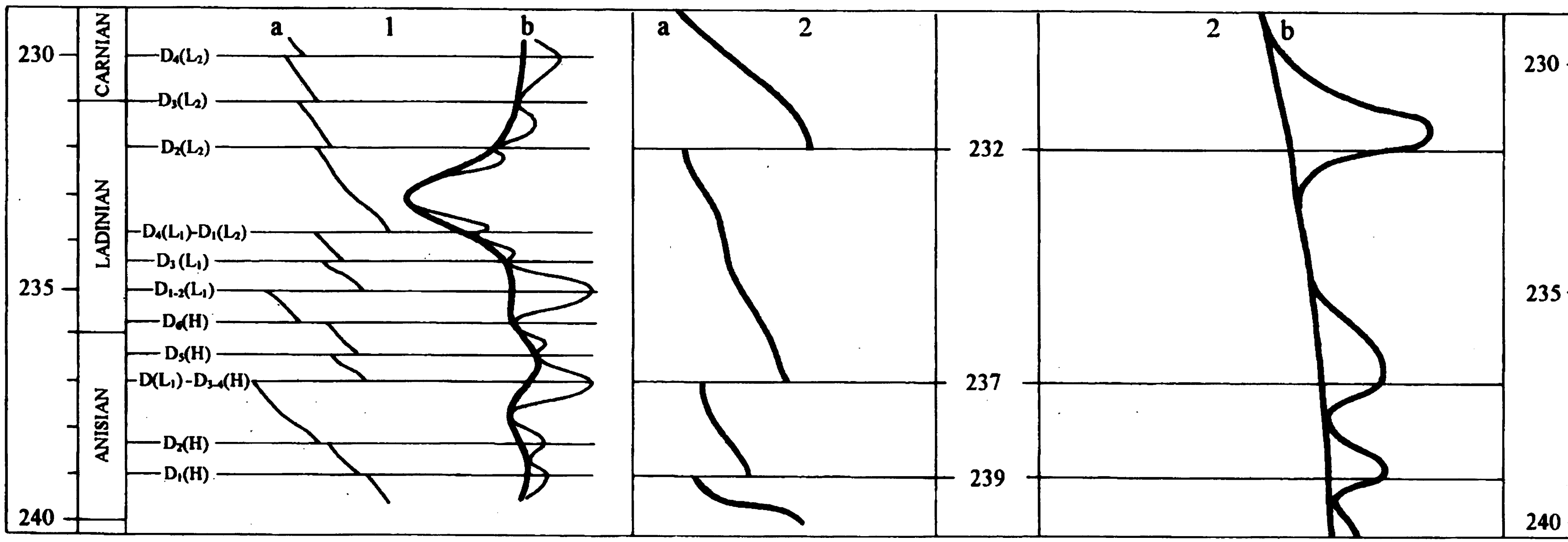


Fig. 38. Correlation of the Middle Triassic palaeokarst horizons in the Balaton Highland with the global sea-level changes (KORPÁS and DUDKO 1993)

1. Balaton Highland: a) coastal onlap curves b) sea-level curves; 2. Global changes after HAQ et al. 1987: a) coastal onlap curves; b) sea-level curves; D_1 (H)=Discontinuity surface Nr. 1, Hajmáskér; D_3 (L_1)=Discontinuity surface Nr. 3, Litér 1; D_2 (L_2)=Discontinuity surface Nr. 2, Litér 2

Cycle 1 (239–237 Ma): shallowing upward dolomites (platform).

Cycle 2 (237–235 Ma): even more shallowing upward dolomites. The datum of the first volcanic event is at 237 Ma (platform).

Cycle 3 (235–233 Ma): deepening upward pelagic sediments. Maximum flooding and termination of volcanism at 233 Ma (pelagic inlet).

Cycle 4 (233–232 Ma): even more shallowing upward pelagic sediments, followed by dolomites (platform).

Cycle 5 (232–230 Ma): shallowing upward dolomites (platform).

This means that the study area permanently maintained its platform margin position and was flooded only once, by pelagic inlet of the high-stand event between 235–232 Ma. Consequently the low-stand phases of subaerial exposure have played a definite role in the early palaeokarst evolution of the dolomite platform.

Palaeokarst evolution

The four main phases, differentiated in the evolution of the karst system will be characterized next (**Fig. 39**):

Karst phase 1 (239–237 Ma): The lower and upper stratigraphic boundaries correspond to 3rd order subaerial discontinuity surfaces at 239 Ma and 237 Ma. Depositional environment: hypersaline sabkhas with marine karst system. The poorly developed palaeosols, the insignificant microvuggy and mouldic porosity indicate a wide, flat littoral karst plain with reduced marine porewater circulation.

Karst phase 2 (237–235 Ma): The lower and upper boundaries coincide with 3rd order, subaerial discontinuity surfaces of 237 and 235 Ma. Both are low angle composite disconformities and reflect a more significant morphological dissection, accompanied by a break in deposition and by internal erosion. The depositional environment was hypersaline with sabkhas. The initial vadose karst facies was followed by a marine phreatic one. The well developed palaeosol horizons, the mainly microvuggy and mouldic porosity, the frequent cavities and fractures and the palaeokarst pockets indicate a flat, but slightly dissected littoral karst plain. The initial, intensive fresh water circulation in the vadose karst system evolved into a marine one. The NE–SW oriented extensional tectonic control can be assumed in the evolution of the karst system.

Karst phase 3 (235–234 Ma): Its lower boundary is the 3rd order subaerial, composite discontinuity surface of 235 Ma, while the upper boundary coincides with the high stand discontinuity surface of 234 Ma. The lower boundary is a low angle disconformity, accompanied by a significant break in deposition and by great internal erosion indicating the age of maximum morphological dissection. The hypersaline depositional system, dissected by sabkhas and tidal chanels was substituted at 234 Ma by a pelagic inlet, connected directly with the open sea. The poorly developed palaeosols, the intensive vuggy and mouldic porosity, the major cavities and open

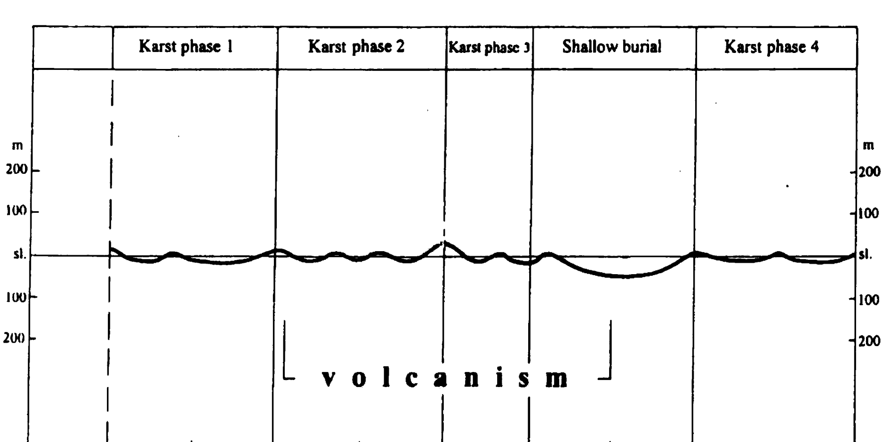


Fig. 39. Model of karst evolution of the Litér–Hajmáskér area (KORPÁS and DUDKO 1993)

joints, and a palaeodoline reflect a littoral karst plain with dissected morphology with a magnitude of 10 m. This phase can be characterized by the most intensive, mainly marine water circulation. The karst system was further controlled by NE–SW extensional tectonics.The evolution of the karst phases 1–3 was stopped by sea-level rise at 234 Ma, resulting in the shallow burial of the karst system.

Karst phase 4 (232–230 Ma): Its lower and upper boundaries coincide with the 3rd order partly subaerial discontinuity surfaces of 232 Ma and 230 Ma. It is a shallowing upward depositional system with marine karst facies. The poorly developed palaeosol horizons and the insignificant microvuggy and mouldic porosity again indicate a flat littoral karst plain with reduced circulation of marine porewaters.

Assuming those explained above can be stated that:

1. The composite marine karst system of the platform margin gradually reached its maximum dissection between 235–234 Ma.
2. This karst system was buried to very shallow depths in the timespan of 234–230 Ma and the subsequent emersion generated the new karst cycle at 232 Ma.
3. The further evolution of this composite karst system was not significantly influenced by later geological events.
4. This palaeokarst event, considering its magnitude, age and duration (238–235 Ma) is very similar to that of the Muschelkalk of the Catalan Range, described by CALVET et al. (1990), CALVET and TUCKER (1995) as discussed earlier.

4.2.4. *Vác (Naszály Hill)*

The multiphase composite palaeokarst evolution of the Naszály Hill will be discussed after JUHÁSZ et al. (1995).

Geological setting

The geological make-up of the Naszály horst, composed of Late Triassic platform carbonates surrounded by Middle Oligocene–Middle Miocene molasse and partly by Middle Miocene volcanics is illustrated on **Figs. 40, 41**. The palaeokarst phenomena are related to the Late Triassic Dachstein Limestone.

Depositional system

The Dachstein Limestone comprises stacked, cyclic platform carbonates (**Fig. 42**). The sol horizon of the Lofer cycle is absent and vadose coated grains represent the subaerial part of the cycle (BALOG and HAAS 1990). The dolomitic intertidal member is rare and poorly developed. It is characterized by fragmented, partly dolomitized laminated clasts, which resemble microbial laminites. The 5thorder cycles are about 6 m, some of them 10–15 m thick. The last ones probably represent a 4thorder amalgamated megacycle (GOLDHAMMER et al. 1987, 1990). The dominant subtidal limestone member is characterized by burrowed pellet wackestone and dasyclad-lithoclastic packstone-grainstone. Evidence of marine diagenesis includes various types of early marine cements, some filling submarine sheet-cracks. The biota is represented by foraminifera, subordinate gastropods and dasyclad algae. The deposits accumulated in a back-reef lagoon (BALOG and HAAS 1990).

Subaerial surfaces, parallel to the bedding and capping the Lofer cycles (**Photo 9**) occur throughout the succession. On these erosional surfaces vadose caps developed. They are commonly rich in dolomite microspar cement or dolomite replacement of micritized grains. Vadose pisoids, aragonite cement and concentrically cracked grains are common. Vadose diagenetic features, such as solution-enlarged vugs, with subordinate pendant and meniscus cement are common.

Palaeokarst phases

Karst processes commonly result in dissolution of the underlying rock, precipitation of speleothems from circulating fluids and brecciation and/or sedimentation in cavities. Different combinations of factors caused different sets of karst features. Differentiation and identification of the temporal succession of the karst phases is based on the concept of diagenetic overprinting.

Karst phase 1: The earliest phase of karst development is evident throughout the 320 m thick sequence investigated. It is represented by small, solutionally enlarged fractures (**Fig. 43**) and cavities that truncate skeletal fragments and early marine

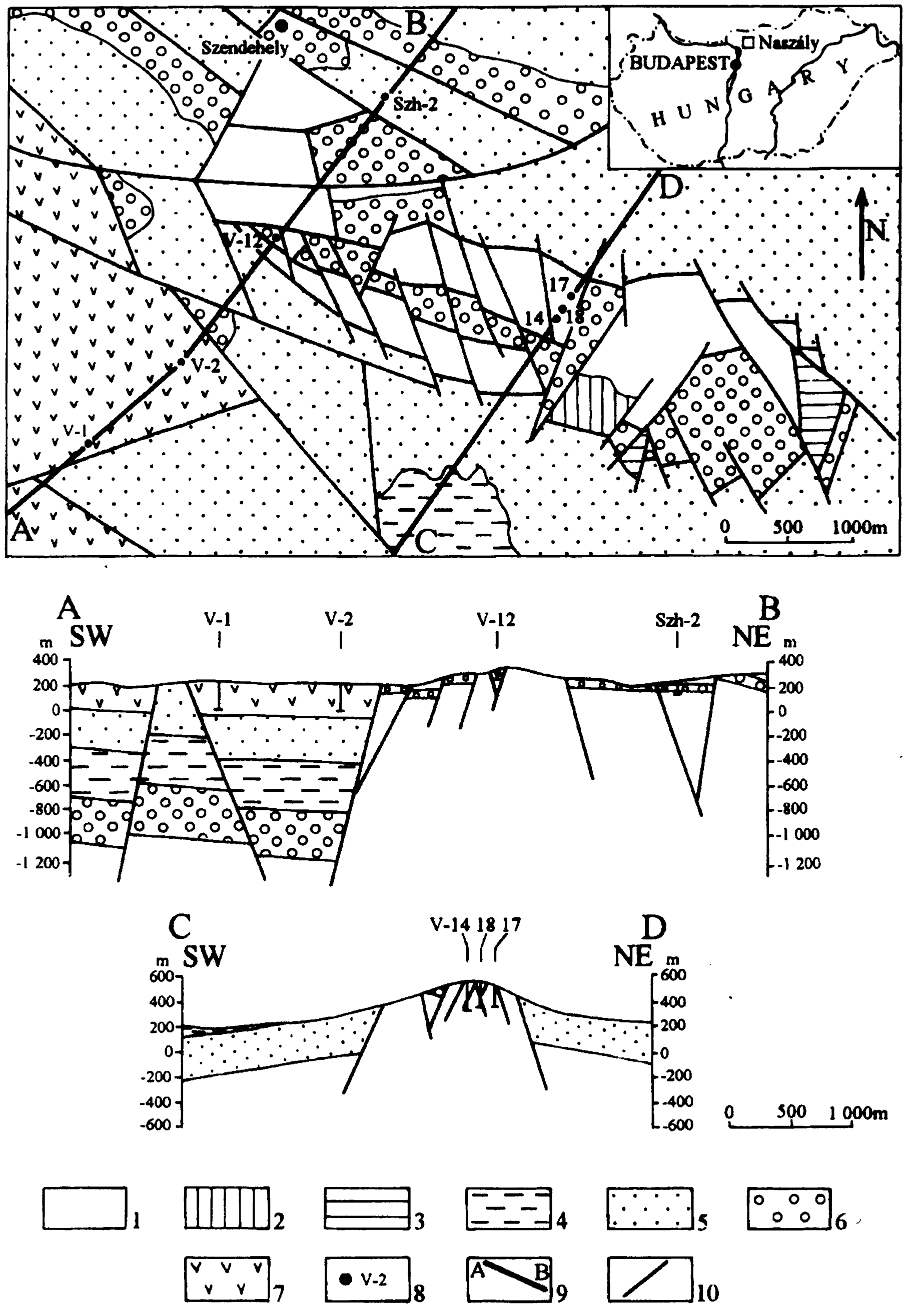


Fig. 40. Geological map and two sections of the Naszály area. (JUHÁSZ et al. 1995)
1. Upper Triassic, Dachstein Limestone, 2. Upper Triassic Main Dolomite, 3. Eocene, Szépvölgy Limestone, 4. Oligocene, Kiscell Clay, 5. Oligocene and Miocene Sand, 6. Oligocene, Máriahegy Sandstone, 7. Miocene, volcanics, 8. Borehole, 9. Profile, 10. Fault

cement. The early karst cavities are filled by oxidized microlaminated marine carbonate silt and/or clast-supported autoclastic breccia. Clasts of lithified internal sediment and marine radiaxial fibrous calcite are present. Non-luminescent marine cement and stylolites post-date brecciation. This phase post-dates submarine isopachous cementation, but pre-dates stylolite formation. Karstic features and stable isotope data (**Fig. 44**) indicate that karstification proceeded mainly in a vadose environment. The karst profile subsequently was affected by marine flooding that initiated the next Lofer cycle. Karst cavities and enlarged fractures were filled by marine carbonate silt and cement.

Karst phase 2: Features produced during this phase include lenticular caves (2–25 m wide) developed along the bedding (**Photo 9, Fig. 43**). They cut the stylolites and are filled mainly with laminated kaolinitic siltstones, claystones and brec-

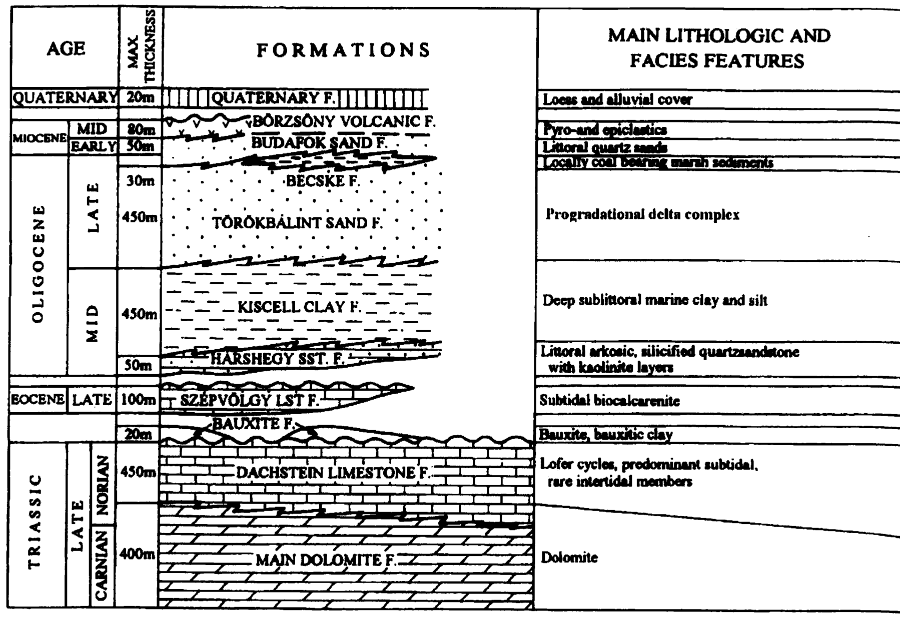


Fig. 41. Stratigraphic chart of the Naszály area and its surroundings (JUHÁSZ et al. 1995)

cia. The cave horizons also can be identified on well logs (**Figs. 42, 45**). Based on gamma logs and the cores the cave levels are seen to repeat over vertical distance of 10–70 m. Excursions on the logs indicate those levels where clay-filled caves are

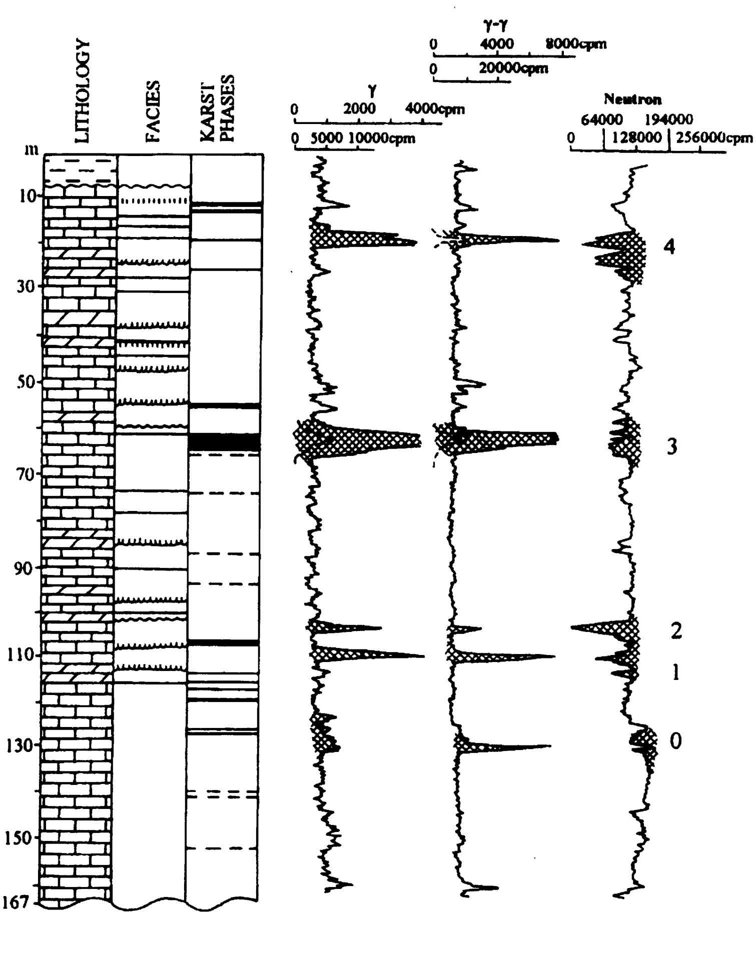


Fig. 42. A representative section of borehole Vác–3 (JUHÁSZ et al. 1995)
1. Clay, 2. Dolomite, 3. Limestone, 4. Subtidal members, 5. Intertidal members.
6. Unconformities, 7. Karst phases, 8. Cave horizons

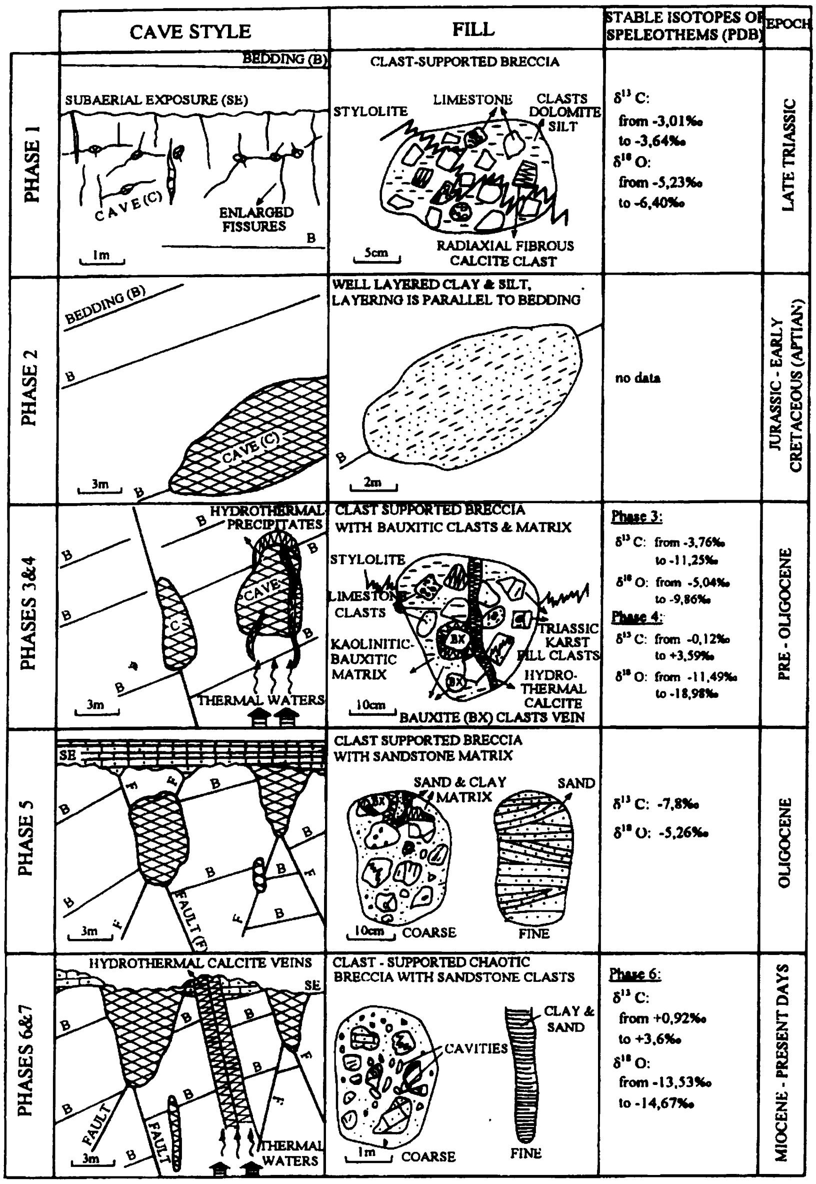


Fig. 43. Schematic illustration of the karst phases, with the main features of karstification (JUHÁSZ et al. 1995)

common. Within one major karstic horizon several caves may have developed, in the case of level 4 at least three. Between the major cave horizons there are 3–12 individual Lofer cycles (**Fig. 42**). As a result of later tectonic movements the originally horizontal cave horizons now have different declinations (**Fig. 46**). Karstic depressions and filled dolines, locally containing small marshes and swamps have formed where units of this karst phase are exposed at the surface. The second phase of karst development was related to a long term exposure in the Jurassic and Early Cretaceous. The Early Jurassic rifting caused uplift of the Naszály platform segment and the karst base-level dropped. Karstification during the intervening stages of platform stability resulted in overprinting of the earlier karst and in new generations of caves. The spacing of the cave levels suggests the Naszály area emerged and the water table dropped, causing cave levels to develop at 10–70 m intervals. Karst terraces could form in this way (MIÖTKE and PALMER 1972). Units affected by the second phase of karst development are unconformably onlapped by Eocene sediments indicating that phase 2 is of Eocene age or older. Tilting of the Triassic platform started in the Jurassic, but the major phase of deformation occurred in Mid-Cretaceous time. The fact that the layering of the cave sediments of karst phase is parallel to the bedding implies that this phase older than Mid-Cretaceous and younger than the stylolites, which are cut by products of karst phase 2. The general

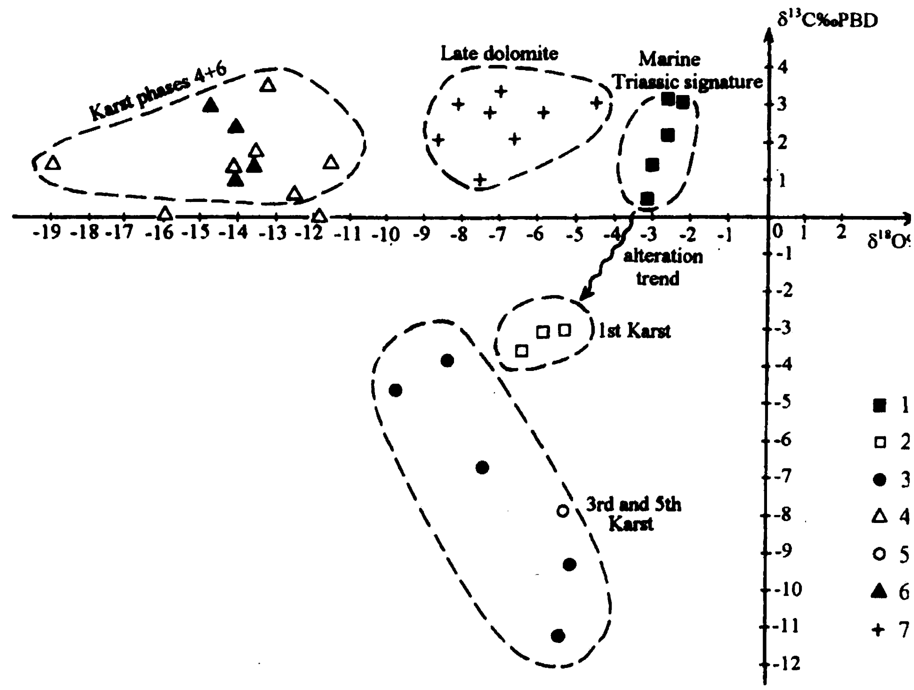
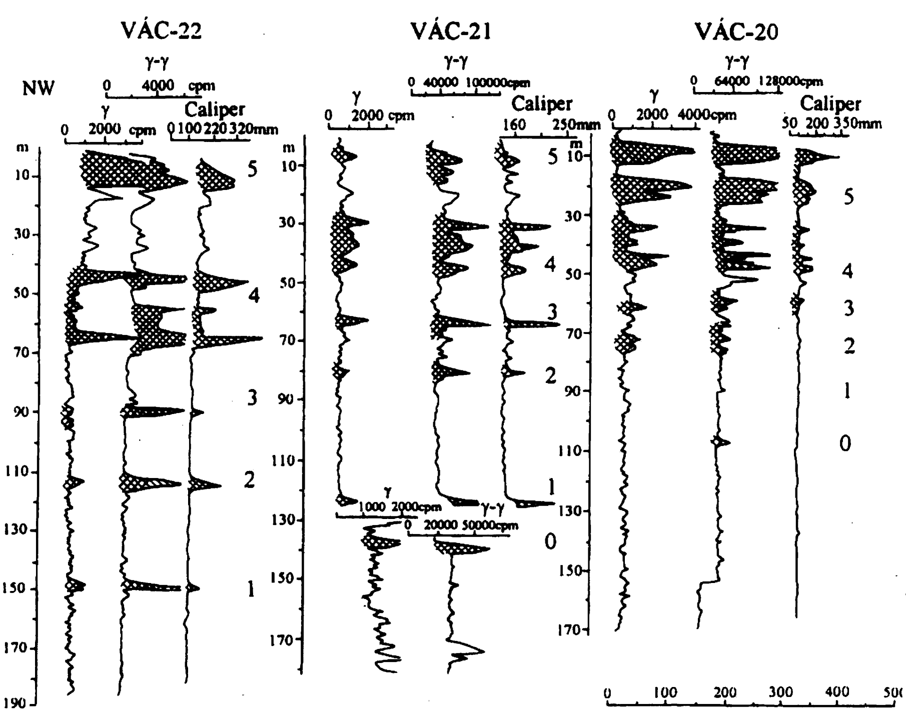


Fig. 44. Stable isotope composition of host rock and cements from the Naszály area (JUHÁSZ et al. 1995)

1. Host rock and early marine cement, 2. Karst phase 1, 3. Karst phase 3, 4. Karst phase 4, 5. Karst phase 5, 6. Karst phase 6, 7. Late dolomite

morphology of the large, elongate, lenslike caves with their filling suggests a shallow phreatic environment of formation.

Karst phase 3: This phase is characterized by non-bedding controlled karst forms. Intense dissolution along faults affected the products of karst phases 1 and 2. and caused cave formation, brecciation and sedimentation. The predominantly verti-



LEGEND: 1,2... cave horizons, γ: natural γ, γ-γ : gamma-gamma ray log Vác-22: borehole

Fig. 45. Cave horizons filled by soft siltstones and clays can be identified on well logs (JUHÁSZ et al. 1995)

cally elongated caves with rounded wall-rock surfaces are discordant to the bedding of the host rock and cut the stylolites. Clast supported breccia of this phase consists of limestone with reworked early karst deposits and cements, and a few clasts of stylolite bearing limestone, dolomite, early internal sediment and bauxite. The breccia matrix consists of bauxitic and kaolinitic clay and dolomitic silt. The fissures filled with this kind of matrix cross-cut the karst fabrics of phases 1 and 2. The stable isotope data (**Fig. 44**) suggest both vadose and phreatic environments for karstification and cementation with slight influence of hydrothermal fluids. Karst phase 3 is related to a composite pre-Oligocene unconformity which affected the earlier palaeokarsts. Naszály Hill was not completely covered by Early Eocene sediments, thus extensive karstification could re-occur during the Eocene, affecting large areas of the Triassic carbonates. The conglomerate at the base of the Eocene limestone in the Pádimentum quarry contains reworked and altered Triassic rock fragments, indicating continuous subaerial exposure over a protracted period extending from Triassic to Eocene times.

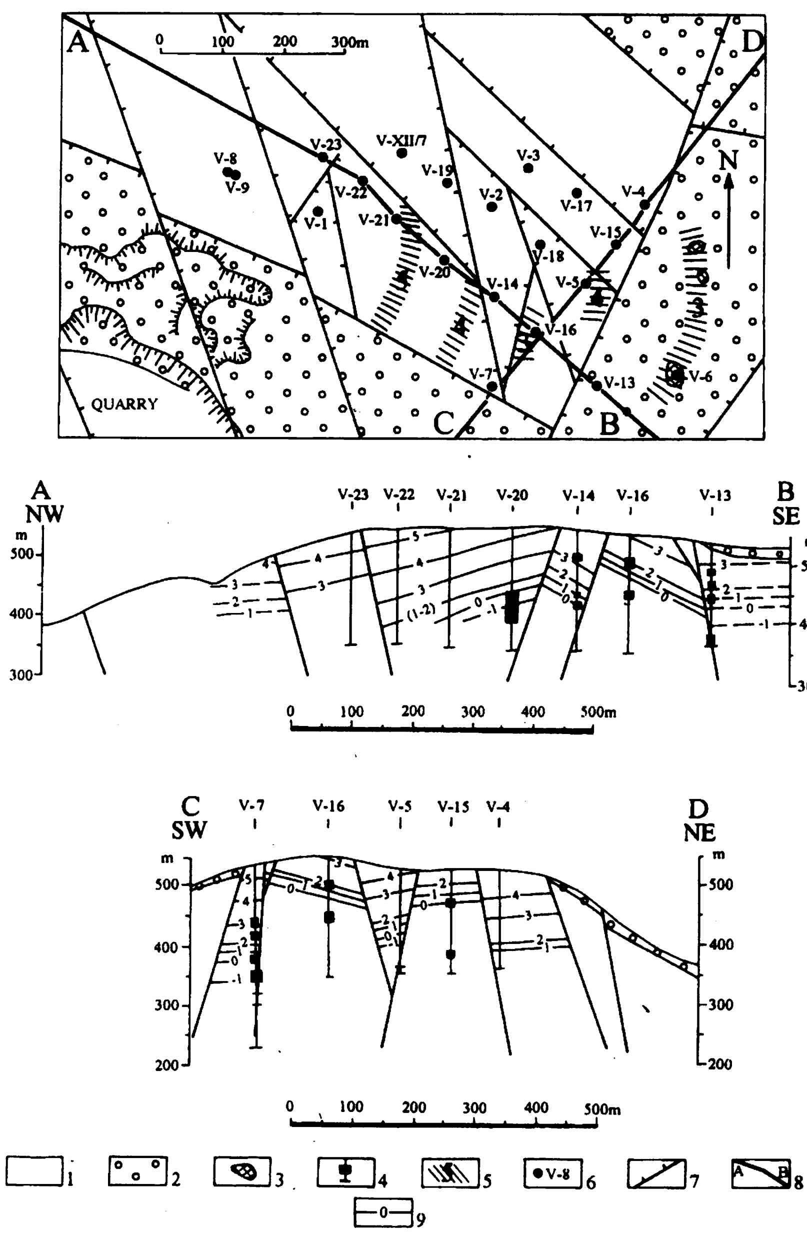


Fig. 46. Karst geological map and profiles of the Naszály Hill (JUHÁSZ et al. 1995)
1. Dachstein Limestone, 2. Oligocene, Hárshegy Sandstone, 3. Marsh, 4. "Late" dolomite bodies in borehole, 5. Karst horizons, 6. Borehole, 7. Fault, 8. Profile, 9. Cave horizon

Karst phase 4: Calcite crystals up to 5–6 mm long precipitated in the open pore space or formed small, millimetre and centimetre in size calcite veins that did not penetrate the Oligocene clastic cover. The calcite crystals are zoned, ferroan and show dull or bright luminescence. Precipitations of small calcite plates occured in the cavities. They are formed of calcite rhombohedra with micritic margins, showing bright luminescence in the centre and dull or no-luminescence on the micritic margin. Microlaminated and botryoidal, popcorn-like calcite cements with bright luminescence are also present in the cavities. These calcite generations post-date the bauxitic clast-bearing breccias and pre-date the siliciclastic filling of karst phase 5. The stable isotope data (**Fig. 44**) suggest precipitation from hydrothermal fluids. The small calcite plates of phase 4 can be interpreted as floating raft sheets precipitated in thermal waters. Karst phase 4, the first hydrothermal event is limited to pre-Oligocene karst features, consequently it is dated as pre-Oligocene.

Karst phase 5: The most characteristic feature of this karst phase is the appearance of Middle Oligocene siliciclastic Hárshegy Sandstone in the caves which is well-dated by foraminifera and nanoplankton (CSILLAG-TEPLÁNSZKY and KORPÁS 1982, BÁLDI 1983).

Karst of the Middle Oligocene phase is discordant with the bedding of the host rock, and is oriented differently to the pre-Oligocene karstic fills (**Fig. 43**). Large sinkholes up to 30 m deep contain a chaotic breccia framework supported by a matrix of yellow-brown quartz sandstone and laminated kaolinite. The sinkholes were tectonically controlled. The collapse breccia contains all the rock types of the Dachstein Limestone and the bauxite. Locally, individual clasts exhibit features typical of the earlier karst phases. Quartzitic sediments can be found among speleothems of earlier caves and conduits, having infiltrated the karst profile to depths exceeding 120 m below the present surface. In some caves the fine grained sandstone exhibits cross-stratification (**Fig. 43**). The deposits associated with this karst phase are Middle Oligocene in age and indicate that karst also developed during the Early Oligocene, probably in association with a global sea-level fall (HAQ et al. 1987). In the Naszály Hill area, similarly to the other examples of the Transdanubian Range, the Early Rupelian uplift probably intensified the magnitude of this karst phase. The former karst terrain was rejuvenated, significantly increasing fracture porosity. The rejuvenation of conduits formed during earlier karst phases, and the formation of new ones along fissures is demonstrated by the presence of the siliciclastic sediments which infiltrated them.

Naszály Hill underwent rapid subsidence in the Middle Oligocene and was covered by 350 m of marine Oligocene sediments. The onlapping Hárshegy Sandstone filled large sinkholes and caves. With increased subsidence, thick marine Kiscell Clay covered completely the karstified platform (BÁLDI 1983), acting as a regional seal and presenting further meteoric karstification.

Karst phase 6: One of the most diagnostic characteristics of this karst phase is the occurrence of the calcite veins. The veins can be traced for several hundreds of metres, even out of the area of the Naszály Hill. They cross-cut the whole section and terminate in the Oligocene cover (**Fig. 43**). Locally they crop out at the surface due to the erosion. Most of veins strike northward. Individual vein calcite crystals are large with a maximum length of 3 cm. Their colour is cloudy pale green and under cathodoluminescence they appear bright and zoned. The stable isotope composition suggests a hydrothermal origin with elevated temperatures. After the Oligocene the updoming of the Badenian volcanic centres produced an E–W oriented stress field with large-scale fractures and fissures in the vicinity of Naszály Hill (BENCE et al. 1991). The 15.2–14.4 Ma of this volcanic event is well constrained by K-Ar dating and magnetostratigraphic measurements (KORPÁS and LANG 1991, 1993). The formation of hydrothermal calcite veins in karst phase 6 is related to this post-Oligocene activity.

Karst phase 7: Presently active vertical sinkholes and vadose conduits, which formed along faults are exposed in the quarry. They contain stalactites that post-date the second hydrothermal event. Boulders and blocks of the Hárshegy Sandstone cover provide the bulk of the clast-supported chaotic breccia fills, but fragments of vein calcite of karst phase 6 also are present (**Fig. 43**). Quartz sand from the

Hárshegy Sandstone has been washed into cavities and accumulated in fine laminae. The modern karst forms of phase 7 post-date the Oligocene karstification and occurred during the post-Oligocene elevation of the Naszály horst (**Fig. 40**). This phase includes the almost complete erosion of the Oligocene marine cover and a modern vadose overprint with sinkholes, breccias and speleothems. The modern karst water table is about 120 m above sea-level and the present climate is conducive to the formation of temperate-continental karst.

Palaeokarst evolution

The model of palaeokarst evolution is summarized in **Fig. 47**. Thick platform carbonates of the Dachstein Limestone accumulated in the Late Triassic. Several episodes of subaerial exposure resulted in karstification associated with Lofer cycles at certain cycle boundaries.

At the end of the Triassic, disintegration of the platform commenced due to Tethyan rifting. The Naszály Hill platform segment was gradually uplifted and carbonate deposition ceased, following subaerial exposure and intense meteoric karstification. The bedding-parallel orientation of large phreatic caves and their infillings, proves that before the major phase of alpine uplift (Cretaceous) the Naszály Hill platform segment emerged without tilting.

Erosion during the pre-Oligocene produced composite unconformities and overprinting of earlier karst phases. During this erosional interval the platform was a low-relief area where bauxite accumulated and contributed to the karst breccias. Subsequent Late Eocene hydrothermal activity was related to local magmatism.

In the Early Oligocene (c. 36 Ma) the effects of tectonic uplift were enhanced by a large-scale eustatic sea-level fall, resulting in denudation and karstification of large portions of the Triassic platform, including Naszály Hill. Karstification produced a new generation of large sinkholes, caves and dolines. At the beginning of the Rupelian an important depositional phase started, and most of the caves were filled by marine conglomerate, sandstone and fine-grained siliciclastic sediments. Naszály Hill subsided and was covered by at least 350 m of sandstone and marl.

In the Late Oligocene and Miocene the area was uplifted again and the second phase of hydrothermal activity occurred. Most of the caprock was eroded and the carbonates repeatedly became subaerially exposed.

Pisznice cave 4.2.5.

The Pisznice cave, which had been visited in 1991 with TAKÁCS BOLNER and KRAUS is controlled by bedding and it exposes a very nice profile of sedimentary and diagenetic features (**Photos 10, 11**) developed in the Late Triassic platform carbonates of Dachstein Limestone.

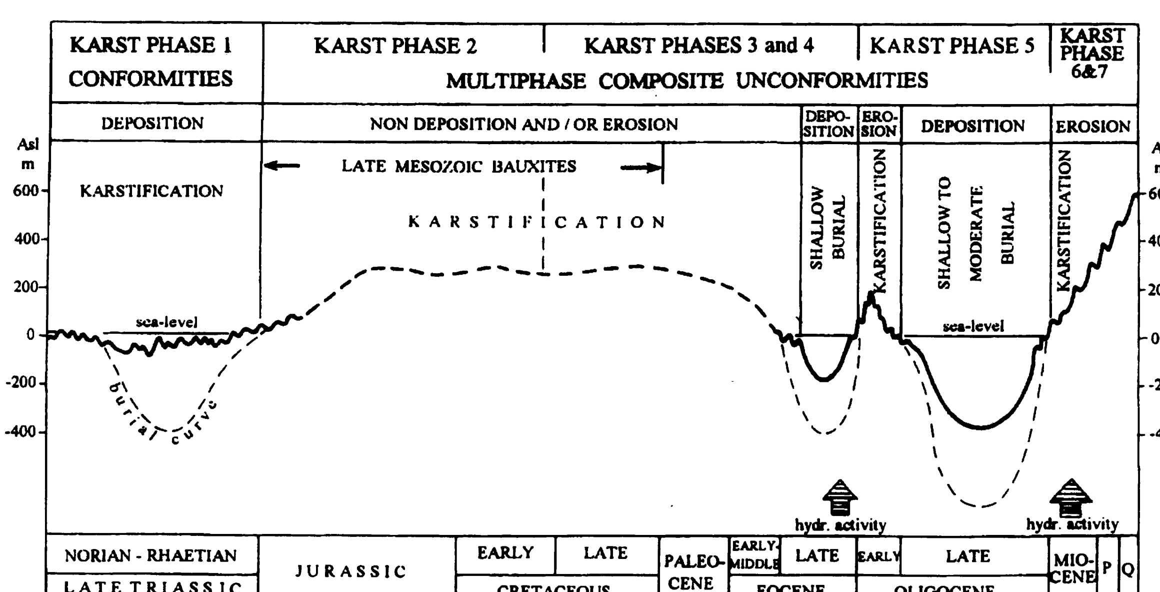


Fig. 47. Model of karst evolution, Naszály Hill (JUHÁSZ et al. 1995)

Photo 10 was taken in the Tízek-terme and illustrates the cyclic architecture of Dachstein Limestone, dissected by subaerial discontinuity surfaces (D_{1-3}), and controlled partly by synsedimentary lystric faults.

Extremely large early stylolites (**Photo 11**) were developed at the top of discontinuity surface D_3. They should be interpreted as a lapiez formed on a rocky shore and covered by the subtidal limestone member composed of autoclasts and bioclasts.

4.2.6. *Csővár (quarry and Várhegy cave)*

The palaeokarst of the upper part of the Csővár Limestone, 150 m in thickness, will be described after DETRE (1970, 1981) and KOZUR and MOCK (1991) completed by our results and by the unpublished data of HAAS and DOSZTÁLY.

Geological setting

The Csővár Limestone consists of alternating beds of grey laminated limestones, cherty limestones and marls with some silty and bituminous layers in its basal horizons. The characteristic depositional patterns are the following: abundant fine-grained deposits; presence of debris flows composed of bioclasts and lithoclasts; autoclastic talus megabreccia and olistholiths bearing horizons; gradation; internal erosional surfaces and channels, mainly parallel to the bedding; bioturbation; synsedimentary folds and slump structures. The Csővár Limestone contains a large diversity of both benthic, mainly redeposited and planktonic fossils (molluscs, foraminifera, ostracods, plant remains, algae, ammonites, sponge spicules, echinoderms, holothuroids, radiolaria, conodonts). The dominant microfacies are radiolarian-echinodermal mudstones/wackestones and in certain horizons bioclastic and lithoclastic grainstones, floatstones and rudstones. The formation is interpreted as a pelagic basinal sequence, composed mainly of distal turbidites with individual proximal ones deposited along the continental margin slope of the Late Triassic carbonate platform. Its estimated age: Norian–Rhaetian–Hettangian–(Sinemurian).

Palaeokarst features and interpretation

The palaeokarstic features of the Csővár Limestone will be discussed using **Figs. 48–53**. Two major palaeokarst horizons, parallel to the bedding are known: the first one is located in the upper part of the Pokol-völgy quarry (**Figs. 48–51**), the second one is exposed about 80 m higher in the Várhegy section (Várhegy cave, **Fig. 52** and cliff of the talus megabreccia, below the cave **Fig. 53**).

Palaeokarst horizon 1: It is a well defined level, developed at a 3rd order single discontinuity surface. The host rock is a cyclic, well-bedded to laminated bioclastic, lithoclastic, bioturbated and graded limestone alternating with cherty limestones and marls. The microfacies (LELKES) is radiolarian-echinodermal mudstone to wackestone indicating a depositional system of distal turbidites. Vuggy and fracture porosity are present. The vugs are partly infilled by marine internal sediments of laminites, representing the following generations: 1. siliciclastic clast-supported grainstones (samples Nr. 207, 218); 2. radiolarian–echinodermal, partly silicified mudstones, wackestones (samples Nr. 199, 200); 3. bioclastic mudstones, wackestones and grainstones, composed of calcite–micrite and sparite with foraminifers, ostracods, fragments of molluscs and echinoderms (samples Nr. 200, 214); 4. alternating laminae of the second and third type (samples Nr. 198, 216, 219). The fractures millimetre to some centimetres in size, and are only partially open. They are mainly filled by crystalline calcite of 2–3 generations, sometimes by micrite (**Fig. 51**). Results of stable isotope analysis (HERTELENDI et al. 1993) reflect a normal marine environment both for the host rock (^{18}O: –1.55‰→–2.36‰; ^{13}C: –0.16‰→2.18‰) and the laminated internal sediments (^{18}O: –0.91‰→–5.60‰; ^{13}C: –2.18‰→2.23‰). There is no evidence in the karst profile for subaerial exposure.

Palaeokarst horizon 2: The Csővár cave has developed along faults in the folded section of the Csővár Limestone (**Fig. 52**). The wall rock is a well bedded, grey, cherty limestone with microfacies of radiolarian–echinodermal mudstone and wackestone (LELKES). The cave itself and the fractures enlarged by dissolution are partly

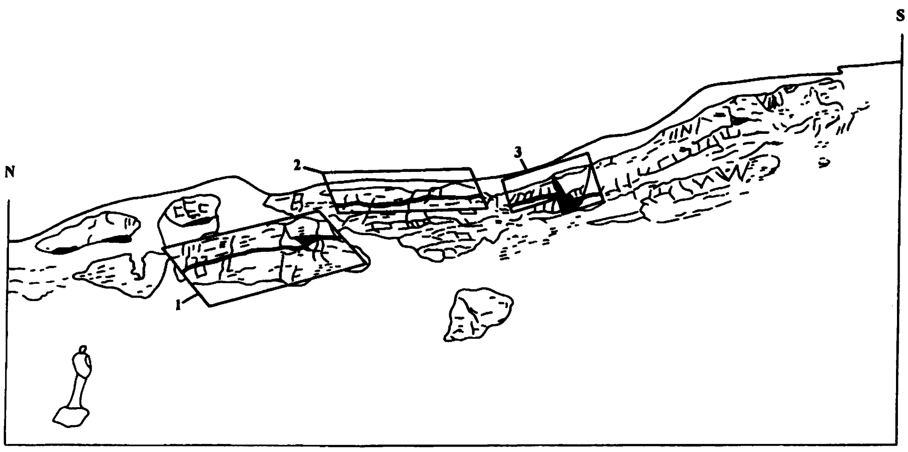


Fig. 48. Palaeokarst profile of the Csővár Limestone, Csővár quarry
1. Csővár Limestone, 2. Palaeokarst horizon, 3. Detailed palaeokarst profiles

infilled by two types of marine internal sediments. 1. Intensively altered, laminated calcarenites and clast-supported microbreccias, composed of clasts of the wall rock. They are embedded in isopachous cement of calcite and the clasts are bordered by very thin crusts of limonite (sample Nr.1, 2). 2. Massive, kaolinite–carbonate infilling composed of isopachous calcite with clasts of altered, limonitized radiolarian–echinodermal mudstones and with fragments of calcite-crystals. This cement is cut by microfissures too, infilled with radiolarian mudstone (sample Nr. 3).

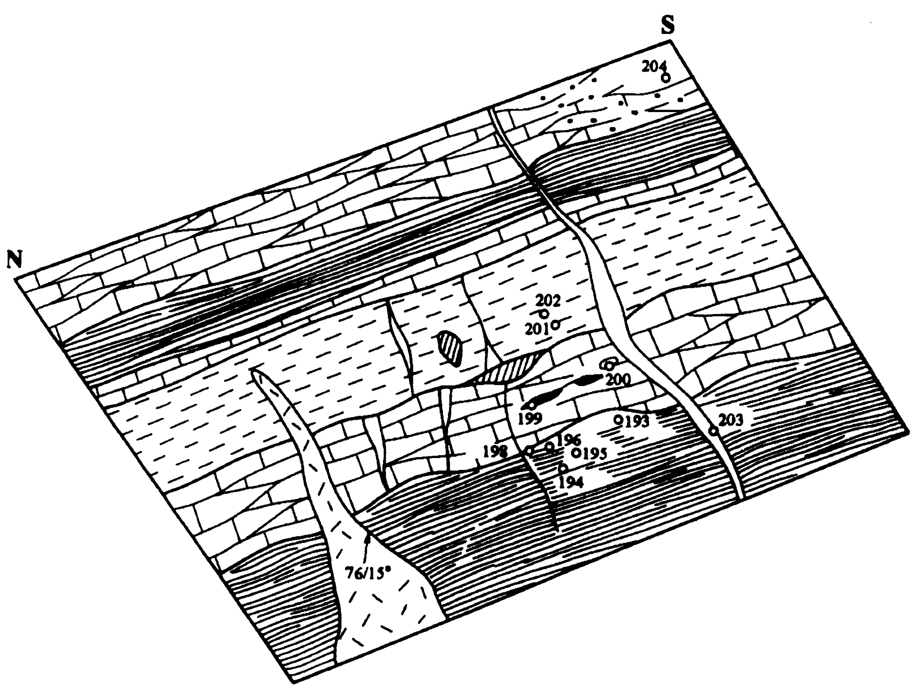


Fig. 49. Detailed palaeokarst profile 1, Csővár quarry
1. Grey laminated limestone, 2. Grey bioclastic, autoclastic limestone bank with fragments of chert, 3. Grey bioclastic laminated, graded limestone, 4. Grey well bedded limestone, 5. Debris, 6. Clasts of chert. 7. Clasts of limestone, 8. Cavities, 9. Veins of calcite, 10. Open joints, 11. Traces of bioerosion, 12. Site of sampling, 13. Dip of strata

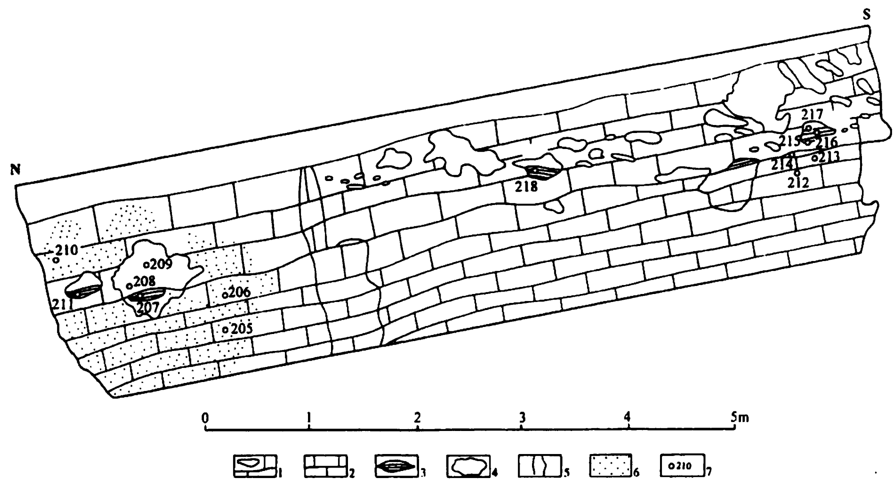


Fig. 50. Detailed palaeokarst profile No. 2, Csővár quarry
1. Grey karstic limestone bed, 2. Grey well bedded cherty limestone with marl, 3. Yellow-greyish laminites of calcarenite-infilling, 4. Cavities, 5. Open joints, 6. White crust of calcite, 7. Site of sampling.

Both infilling generations, identical to the previous ones of horizon 1 are considered as marine in origin and they do not show any signs of subaerial exposure. Results of the single stable isotope analysis on a host rock sample (HERTELENDI et al. 1993) show a hydrothermal effect in a phreatic environment with values of: ^{18}O: −11.71‰ and ^{13}C: 0.64‰.

Some metres below the Csővár cave an excellent profile of talus breccia is exposed (**Fig. 53**). Its lower boundary is a sharp irregular 3rd order composite disconformity surface reflecting internal erosion. The lens-like, poorly to medium sorted, clast-supported, autoclastic breccia is controlled by a network of extensional synsedimentary faults. The individual, mainly angular clasts show a well defined orientation, presumed as bedding, and a clear normal gradation. Below it the folded beds of the cherty Csővár Limestone are exposed with some signs of slump structures. Flat, elongated vugs and cavities, partly parallel to the bedding, partly to the disconformity surface are developed both at the boundary and below it. They are completely empty. The samples taken from the poorly developed carbonate matrix of its basal, medium and upper horizons do not contain fossils. The breccia is considered a submarine synsedimentary talus breccia, formed along the continental margin of the Late Triassic carbonate platform. As its recent analogy the talus breccia described from the Little Bahama Bank and illustrated on the **Fig. 15** should be mentioned.

Hydrothermal veins of calcite: a hydrothermal vein of calcite, light grey in color and 60 cm wide is exposed at the same stratigraphic level of the talus breccia. The vein, striking to NNE–SSW has a sharp vertical contact with the beds of the Csővár Limestone and is composed of three generations of drusy calcite. Similar veins of calcite are known in the vicinity on the Köves-domb (to SE) and on the Vas-hegy (to NW). The vein of the Köves-domb is cutting both the Csővár Limestone and its Late Eocene cover of Szépvölgy Limestone and Buda Marl. This vein 120–140 cm in thickness has the same NE–SW oriented strike and it is composed of also three generations of drusy calcite. Its fissures are partly infilled with bioclasts bearing silts of the Buda Marl, infiltrating them.The vein on the Vas-hegy has penetrated the Triassic Vashegy Dolomite and it is covered by the basal, siliciclastic conglomerate beds of the Hárshegy Sandstone, Middle Oligocene in age. It is 120–140 cm wide and strikes NNE–SSW too. The position and orientation of these veins suggest a synchronous hydrothermal event due to extensional tectonism and magmatic activity. The Late Eocene to Early Oligocene age of this hydrothermal event is evident from crosscuting.

Palaeokarst evolution

The main phases of the long-term and composite palaeokarstic evolution will be summarized in the following:

Karst phase 1: Early submarine syndepositional karst, linked to a single 3rd order discontinuity and formed in shallowing upward conditions (HAAS pers. comm.) along the continental margin slope of the Late Triassic carbonate platform. It may

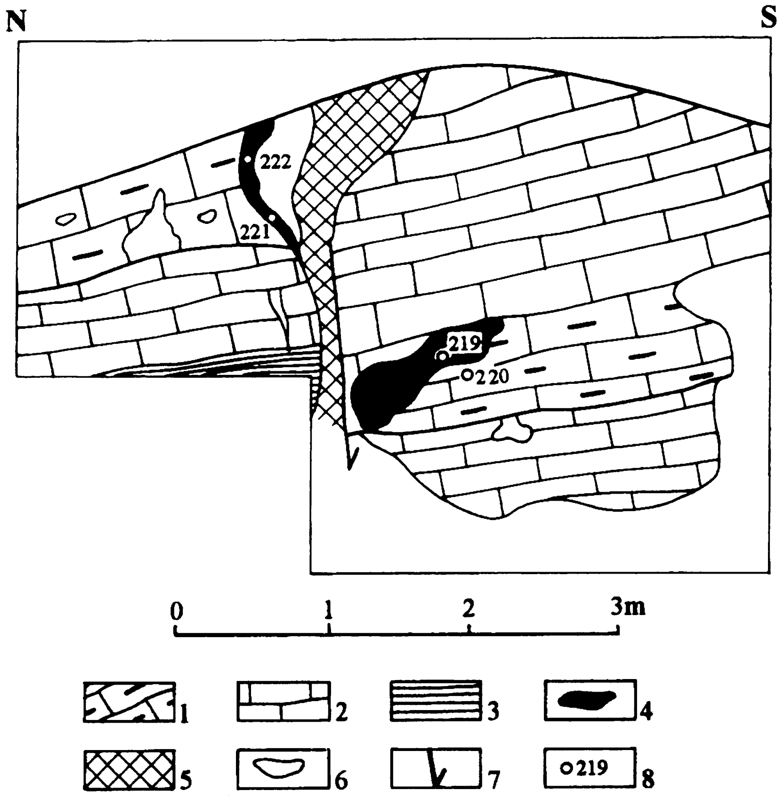


Fig. 51. Detailed palaeokarst profile No. 3, Csővár quarry

1. Grey cherty limestone, 2. Grey well bedded limestone, 3. Grey laminated limestone, 4. Yellow laminites of calcarenite-infilling, 5. Yellow friable clay, 6. Cavities, 7. Normal fault, 8. Site of sampling

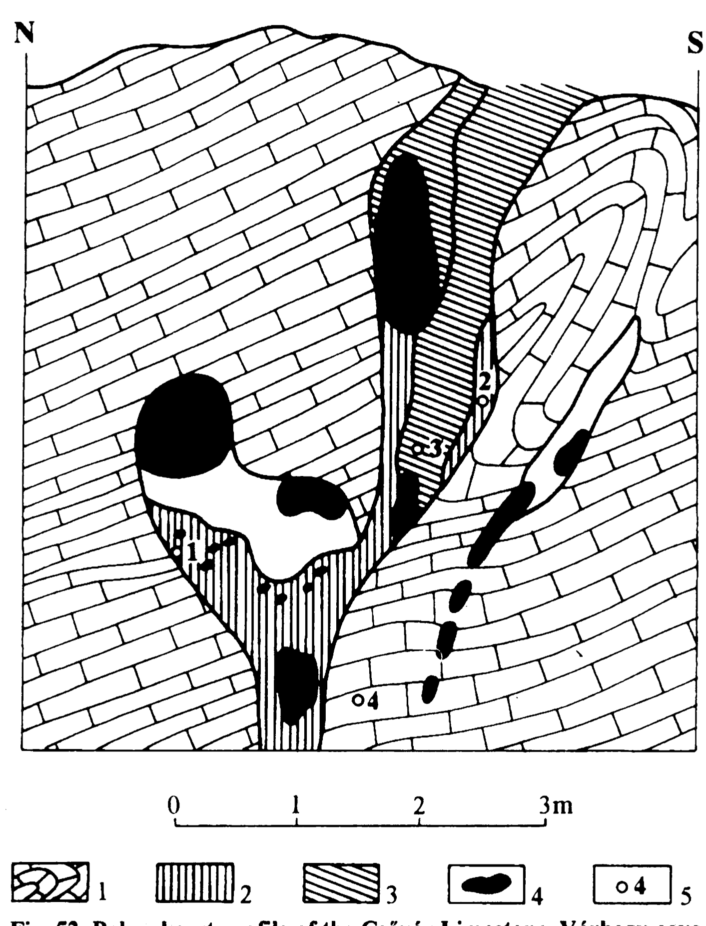


Fig. 52. Palaeokarst profile of the Csővár Limestone, Várhegy-cave

1. Grey cherty limestone, 2. Early laminitic calcarenite-infilling, 3. Late mudstone-infilling, 4. Cavities, 5. Site of sampling

Fig. 53. Palaeokarst profile of the Csővár Limestone, Várhegy. Depositional talus breccia
1. Csővár Limestone, 2. Autoclastic clast supported talus breccia, 3. Discontinuity surface, 4. Synsedimentary faults and fractures, 5. Cavities, fissures, 6. Diagenetic boundary

correspond to the global eustatic sea-level fall of 211 Ma at the Rhaetian–Hettangian boundary (HAQ et al. 1987).

Karst phase 2: Early submarine syndepositional karst related to a composite 3rd order disconformity and formed also along the continental margin slope of the Late Triassic carbonate platform. It shows a slight syndepositional overprint, causing vuggy porosity in the different infilling generations. Two options should be considered for its age estimation: the first one may coincide with the global eustatic sea-level fall supposed to be intra-Hettangian at about 205.5 Ma, while the second one could correspond to the sea-level drop of the Hettangian–Sinemurian boundary at about 202 Ma (HAQ et al. 1987).

Karst phase 3: Using regional geological and karstic evolutional analogies this phase may comprise the long-term subaerial continental karstification of Late Mesozoic to pre-Eocene age.

Karst phase 4: Includes the hydrothermal karst event in phreatic conditions, connected with Late Eocene–Early Oligocene extension and magmatism.

Karst phase 5: The composite karst system of the Csővár horst had been completely buried during Middle and Late Oligocene. The gradual uplift starting in the Early Miocene has resulted in the exhumation and in vadose overprinting of this karst system. This process was completed in the Pliocene to Quaternary.

4.2.7. *Buda Hills (Rózsadomb)*

The long-term multiphase and composite palaeokarst system will be described after KLEB et al. (1993a, b) and by KORPÁS et al. (1993). The geology of the palaeokarst bearing formations and their cover is demonstrated by the geological map of **Fig. 54** while the setting of the largest explored caves in the area is shown on **Fig. 55**.

The Late Triassic formations are located at the southern, uplifted part of the Buda Hills synform, Middle Cretaceous in age (WEIN 1977, BALLA 1988, FODOR et al. 1991a, b, 1994) and originally they formed part of a passive continental margin, bordered by an oceanic basin (KÁZMÉR and KOVÁCS 1985, HAAS 1989). The present-day distribution of the platform and intraplatform basin formations (HAAS 1989, DOSZTÁLY in KORPÁS et al. 1993) is not related to nappe structures as suggested by HORUSITZKY (1943, 1961) and by KOZUR and MOCK (1991), but they reflect more the original facies belts (WEIN 1977). The description of the Triassic formations will be given according to KORPÁS et al. (1993) and HAAS et al. (1993), using the data of RAINCSÁK-KOSÁRY et al. (1985) and of WEIN (1977).

Mátyáshegy Formation (platform margin member)

The *lithology* of the platform margin unit, 150 m in thickness consists of grey, massive, brecciated and bedded to laminated, slightly bituminous cherty dolomites, dolomitized limestones with few intercalations of marls. Dominant textures are dolomitized-silicified biociastic micrites and sparites. The bioclasts are mainly redeposited and composed of benthic foraminifera, ostracods, pellets and some oncoids, further of holothurian and echinoid fragments, radiolaria and sponges, as well as few conodonts. The typical microfacies should be considered after TÖRÖK in KLEB et al. (1993a, b) as mudstone and wackestone.

Among the *sedimentological and early diagenetic features* the following will be outlined: sharp, irregular submarine discontinuity surfaces (reflecting internal erosion) which are parallel to the bedding; synsedimentary slump structures, controlled by microfaults; redeposited, clast supported autoclastic breccias; gradation; neptunian dikes; progressive early dolomitization; multigenerational fracture system with early calcite and subsequent dolomite; early stylolites.

The *depositional system* is considered to be platform margin, representing the transition to a pelagic, intraplatform basin.

Palaeokarstic features: Six registered small caves with poor vadose and thermal precipitations; presence of submarine discontinuity surfaces, parallel to the bedding; well developed fracture porosity, mm to dm in size and infilled by early, clast supported autoclastic breccias; poorly developed mouldic and vuggy porosity in defined horizons; Late Eocene infillings of biocalcarenites in the vugs of the Csúcs-hegy outcrop and in the fissures of the Melocco quarry; veinlets of white hydrothermal calcite in the Melocco quarry.

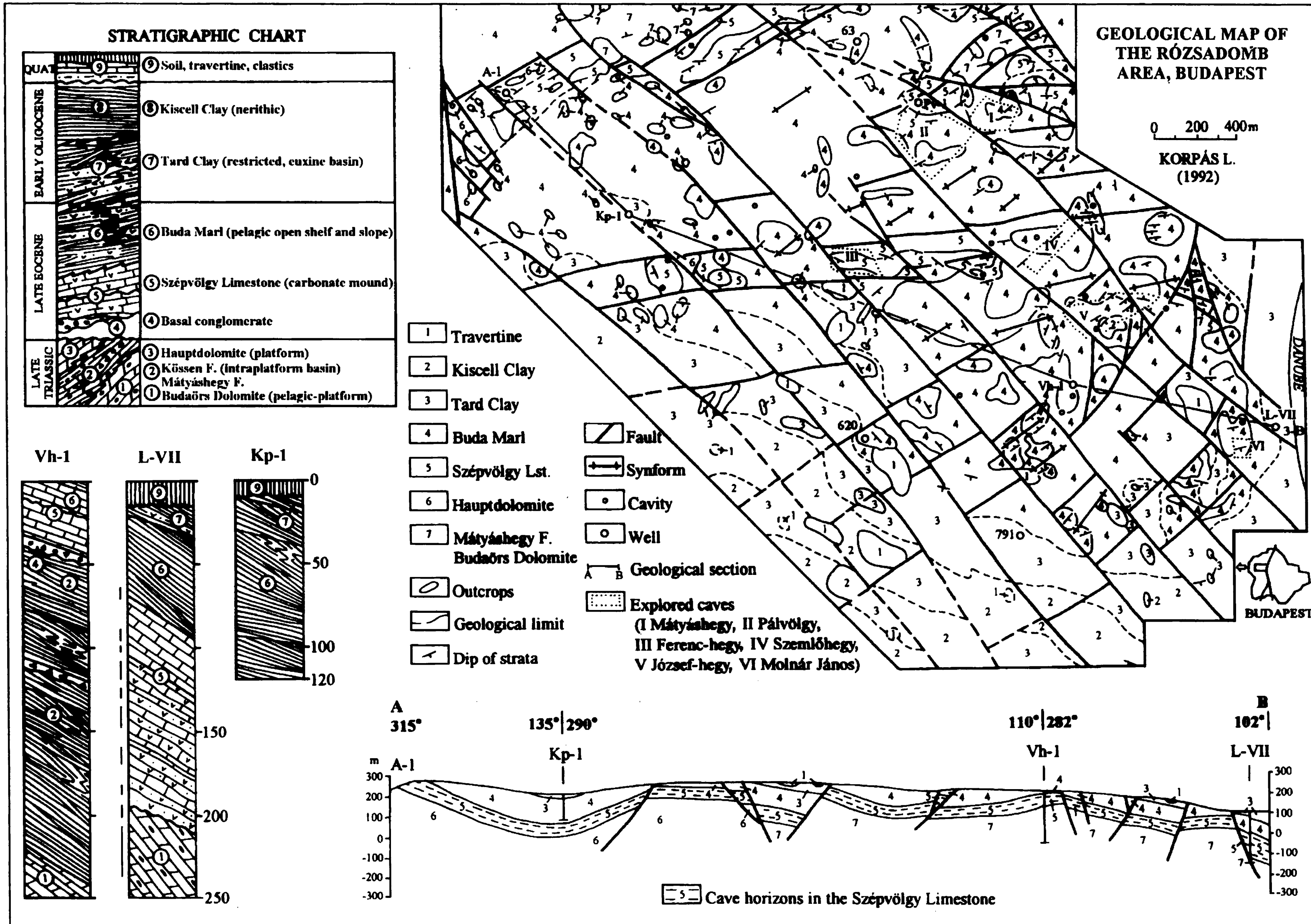
Mátyáshegy Formation (intraplatform basin member)

The *lithologic log* of this member is shown on **Fig. 56**. It consists of grey, folded, well bedded to laminated, graded, bioclastic, bituminous cherty limestones and marls dissected by a horizon of carbonate slope breccia. The lower portion of the section shows a progressive dolomitization. The rich fossil ensemble comprises mainly sporomorphas, algae, radiolaria and sponges, and subordinately holothurian fragments, conodonts and benthic organisms like foraminifera, ostracods, and brachiopods. The dominant microfacies (GÓCZÁN and ORAVECZ-SCHEFFER in KORPÁS et al. 1993) is represented by mudstones, wackestones and subordinately by packstones.

Sedimentological and early diagenetic phenomena: sharp lower and gradual upper contact of the strata; frequent slump structures and related synsedimentary faults; gradation; autoclastic and extraclastic mud supported and clast supported slope breccias; accumulation of kaolinite in well defined horizons; neptunian dikes controlled by extensional microfaults; presence of early diagenetic bacterial pyrite; early stylolites; multigenerational fracture system infilled partly by early calcite and by mud; subsequent progressive dolomitization.

The *depositional system* is considered an intraplatform anoxic basin with a dissected morphology and with a significant terrigenous input, rich in inmature, biodegraded organic carbon (HÁMOR-VIDÓ in KORPÁS et al. 1993).

This member does not show any signs of *karstification.*



STRATIGRAPHIC CHART
QUAT
EARLY OLIGOCENE
LATE EOCENE
LATE TRIASSIC
⑨ Soil, travertine, clastics
⑧ Kiscell Clay (nerithic)
⑦ Tard Clay (restricted, euxine basin)
⑥ Buda Marl (pelagic open shelf and slope)
⑤ Szépvölgy Limestone (carbonate mound)
④ Basal conglomerate
③ Hauptdolomite (platform)
② Kössen F. (intraplatform basin)
Mátyáshegy F.
① Budaörs Dolomite (pelagic-platform)
Vh-1
L-VII
Kp-1
0
50
100
120
150
200
250
GEOLOGICAL MAP OF THE RÓZSADOMB AREA, BUDAPEST
0 200 400m
KORPÁS L. (1992)
1 Travertine
2 Kiscell Clay
3 Tard Clay
4 Buda Marl
5 Szépvölgy Lst.
6 Hauptdolomite
7 Mátyáshegy F. Budaörs Dolomite
Outcrops
Geological limit
Dip of strata
Fault
Synform
Cavity
Well
Geological section
Explored caves (I Mátyáshegy, II Pálvölgy, III Ferenc-hegy, IV Szemlőhegy, V József-hegy, VI Molnár János)
A-1
Kp-1
Vh-1
L-VII
DANUBE
BUDAPEST
A
315°
135° | 290°
110° | 282°
B
102°
m
300
200
100
0
-100
-200
-300
Cave horizons in the Szépvölgy Limestone

Fig. 54.

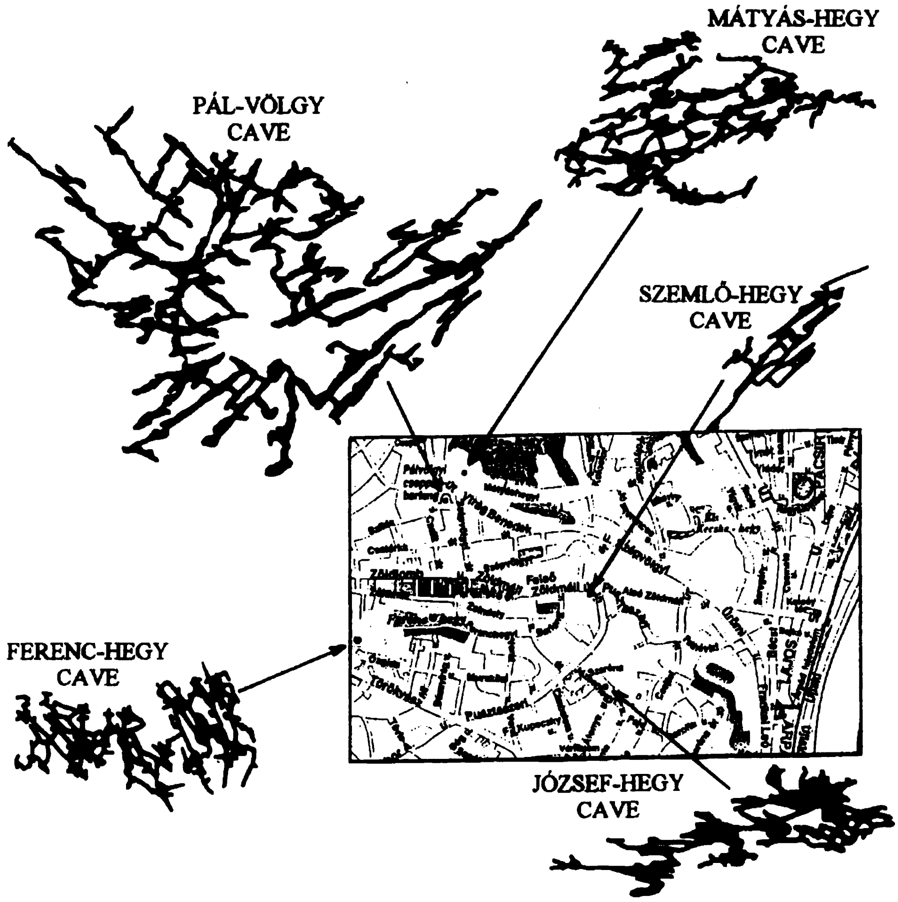


Fig. 55. The setting of the five largest caves on Rózsadomb and their plan view (KLEB et al. 1993a)

Maindolomite

The representative *lithological* section of the cyclic platform dolomite taken at the Balog-cliff with a maximum thickness of about 600 m is shown in **Fig. 57**. The mainly massive brecciated dolomite is often dissected by laminated strata of dolomarls and dolomites containing thin, very fine reddish palaeosol layers. The main textural type, according to TÖRÖK in KLEB et al. (1993a, b) and PIROS in KORPÁS et al. (1993) is intraclastic dolomicrite and dolosparite with peloidal and oolitic–oncoidal relicts. Among the few bioclasts algal plates, fragments of molluscs and at the Apáthy-cliff some impressions of Megalodus are known. A block bearing ammonites and brachiopods was mentioned by HOFMANN (1871) near the outcrop of the Apáthy-cliff.

Sedimentological and early diagenetic features: presence of clast supported autoclastic, partly polymictic breccia horizons and fissure infillings; thin palaeosol layers; mud cracks; horizons of peritidal loferites.

The depositional system is interpreted as a flat platform lagoon, broken by short subaerial exposure events and by open sea flooding periods.

Palaeokarstic features (**Figs. 58, 59, 60**): Presence of eight small cavities registered at the surface ourcrops (they are partly infilled with bauxites (Kőkapu), with basal breccias, carbonate muds and biocalcarenites, Late Eocene in age (Apáthy-cliff, Balog-cliff, Melocco quarry); signs of Late Eocene marine, rocky shore bioerosion (i.e. boring) at the Apáthy-cliff (VIGH and HORUSITZKY 1940); cm–dm wide, open dissolutional joints, infilled partly by clast supported, autoclastic breccias of dolomite; and partly by massive, altered, soft kaolinitic clay bearing clasts of dolomite; well developed vuggy porosity, parallel to the bedding; presence of palaesol layers and related 1st order discontinuity surfaces; dense fracture system, infilled by hydrothermal calcite, barite and pyrite (Apáthy-cliff).

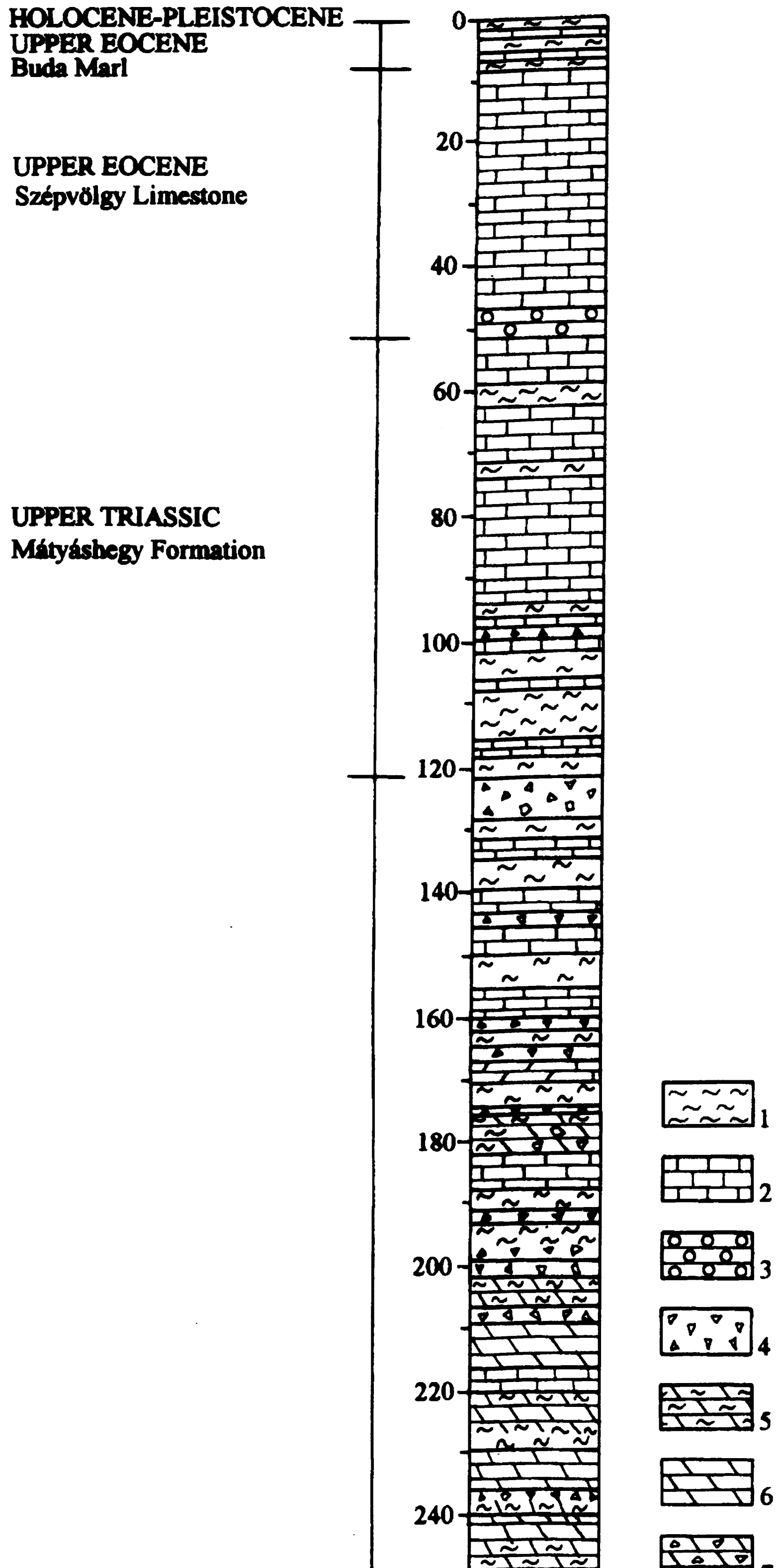


Fig. 56. Lithology and lithostratigraphy of Vérhalom-tér (Vh-1) borehole (KLEB et al. 1993a)

1. Marl. 2. Limestone. 3. Basal conglomerate. 4. Breccia. 5. Dolomitic marl, 6. dolomite. 7. Dolomite breccia

CRETACEOUS
Bauxite, red clay

Only a single outcrop is known at the Kőkapu, where this formation is overlying the karstic surface of the Late Triassic Maindolomite and is covered by the Late

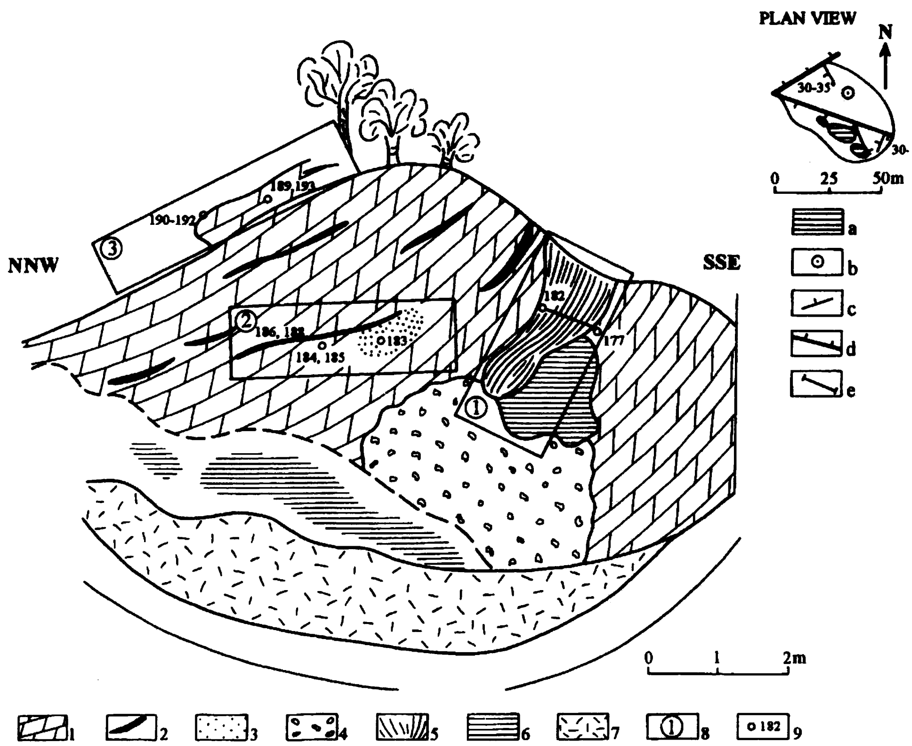


Fig. 57. Palaeokarst profile of the Maindolomite, Balog cliff (KORPÁS et al. 1993)
1. dolomite, 2. Clastic incrustations, 3. Algal mats, 4. Porous carvity infillings, 5. Infilled vertical fissure, 6. Cave, 7. Debris, 8. Sections, 9. Sampling site, a) cave, b) algal layer, c) dip of strata, d) fault, e) section line

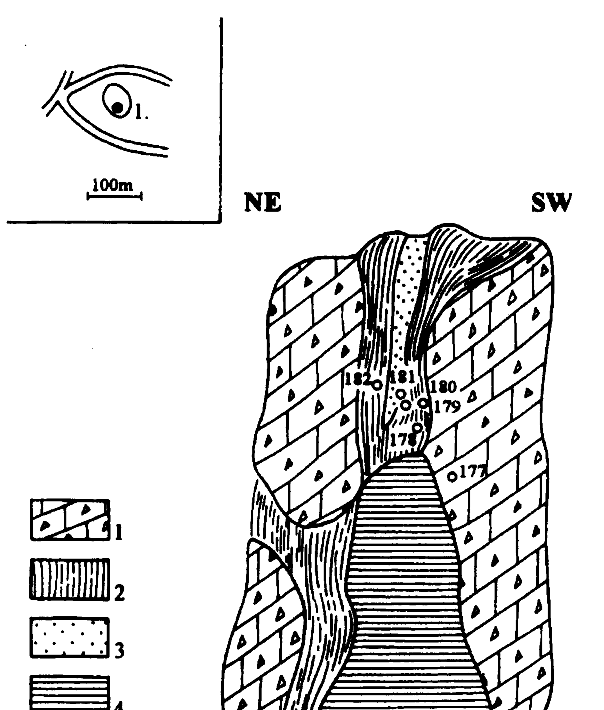


Fig. 58. Detailed palaeokarst profile No. 1, Balog cliff (KORPÁS et al. 1993)
1. Brecciated dolomite, 2. Laminitic infilling with clasts of dolomite, 3. Autoclastic breccia of dolomite, 4. Cave, 5. Site of sampling

Eocene basal conglomerates and Szépvölgy Limestone. The deposit indicates an important subaerial palaeokarst event, linked to a composite regional unconformity. Age estimation is based on regional stratigraphic evidence (WEIN 1977).

EOCENE

The Eocene formations of the study area belong to the Buda facies unit (BALÁZS et al. 1981) located SE of the Buda Line (BÁLDI and NAGYMAROSY 1976, BÁLDI 1983). This unit is characterized by a continuous marine depositional record from the Late Eocene to the Early Miocene, whereas the Early Eocene to Late Oligocene depositional record of the Bakony facies unit (BALÁZS et al. 1981) NW of the Buda Line was interrupted by a significant break and erosion, resulting in the "infraoligocene denudation" (TELEGDI ROTH 1927b).

The description of the formations is based partly on published data of WEIN (1977), BÁLDI (1984), BÁLDI et al. (1984a, b), KÁZMÉR (1985), NAGYMAROSY (1986a, b, 1992), KRAUS (1988), MÜLLER (1989, 1991), NÁDOR and SÁSDI (1991), DUBLIANSKIY (1991), FORD and TAKÁCS BOLNER (1991), NÁDOR (1992b), NÁDOR et al. (1993), KLEB et al. (1993b), KORPÁS et al. (1996), and partly on data avialable in unpublished research reports of KRIVÁN (1970), RAINCSÁK-KOSÁRY et al. (1985), NÁDOR (1992a), KLEB et al. (1993a) and KORPÁS et al. (1993).

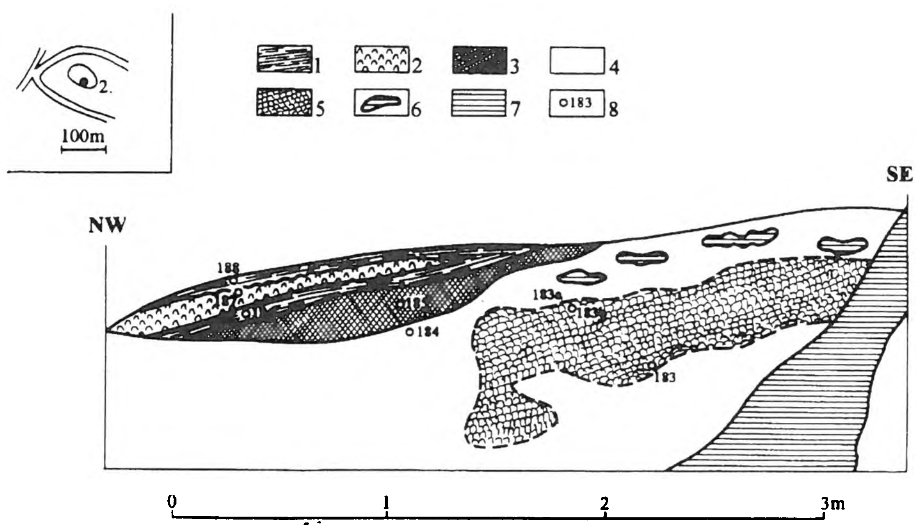


Fig. 59. Detailed palaeokarst profile No. 2, Balog cliff (KORPÁS et al. 1993)
1. Crust of dolomicrite, 2. Alga bearing horizon, 3. Brecciated dolomite, 4. Incrusted dolomite, 5. Bank of dolomite with algaes, 6. Vugs, parallel to the bedding, 7. Open fissures, 8. Site of sampling

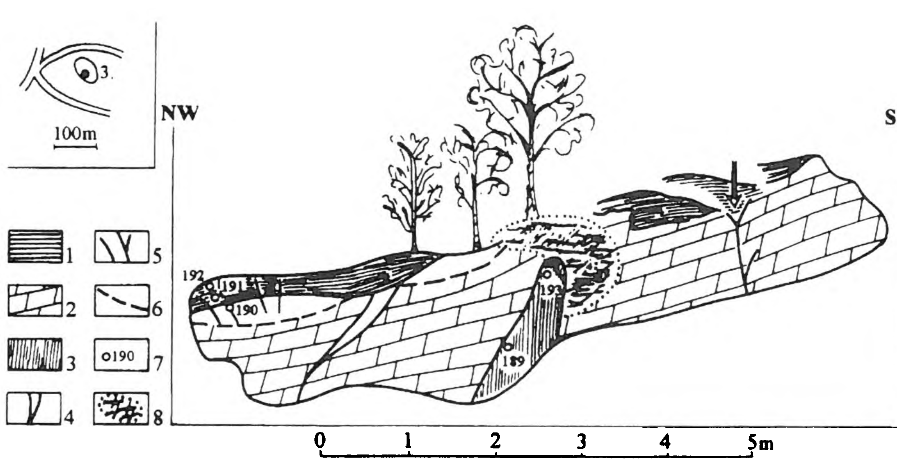


Fig. 60. Detailed palaeokarst profile No. 3, Balog cliff (KORPÁS et al. 1993)
1. Laminated doloarenite, 2. Brecciated dolomite, 3. Yellow friable clay-infilling, 4. Open fissure, 5. Fracture, 6. Bedding-surface, 7. Site of sampling, 8. Covered areas

Basal conglomerate and breccia

This basal unit, less than 10 m in thickness consist of beds and lenses of poorly to medium sorted clast or mud supported conglomerates and breccias. The clasts consist of mainly locally derived dolomite, limestone and chert from Late Triassic formations. Further characteristic components are the altered volcanic clasts of andesites and rhyolites, coming from the Wein palaeovolcano (KORPÁS and KOVÁCSVÖLGYI 1997), and quartzites. These basal beds represent a fandelta complex, deposited on the shoreface zone of the Late Eocene mobile shelf (FODOR et al. 1991b, 1994).

Szépvölgy Limestone

The surface outcrops of the Szépvölgy Limestone are concentrated in narrow W–E oriented belts (**Fig. 54**), while as host rock of the caves it is traversed by passages of some ten kilometres in length (**Fig. 55**). The best studied sections of boreholes and of quarries are demonstrated on **Figs. 56, 61, 62, 63, 64** and **65**.

Concerning its *lithology*, the formation, 50 m in thickness is well bedded and is composed of bioclasts. The main components of sand size are benthic foraminifera (Nummulites, Dyscocyclina), corallinacean algae, bryozoans, echinoids, some fragments of corals, bivalves and decapods. Planktonic fossils, such as globigerinids and radiolaria are numerically subordinate. Few extraclasts of Triassic limestone, dolomite, chert, and of altered volcanites and quartzites are accumulated in horizons of some centimetres to decimetres in thickness. Dominant microfacies types after LELKES (in KLEB et al. 1993a, b and in KORPÁS et al. 1993) are the following (**Fig. 66**): corallinacea–foraminifera–echinodermata–bryozoa packstone; foraminifera floatstone–rudstone; bryozoa–corallinacea–foraminifera packstone; coral boundstone; echinodermata grainstone–packstone. The following age constrains (**Fig. 67**) were estimated, based on biostratigraphic (BÁLDI-BEKE 1977, 1984, NAGYMAROSY 1992) and on magnetostratigraphic evidence (LANTOS in KORPÁS et al. 1993, 1996): nannoplankton zones of NP 18–19/20; normal magnetic anomaly 15 and partly the reverse one above it; 39.0–37.7 Ma of its formation, according to the timescale of BERGGREN et al. (1985).

The following *sedimentological, early diagenetic and/or hydrothermal* features are notable: signs of mass movement and redeposition; presence of submarine, partly subaerial discontinuity surfaces; marine primary intergranular and mouldic porosity; early stylolites; slight dolomitization and silicification; well developed system of hydrothermal veins and veinlets of calcite, quartz, barite, fluorite and pyrite; high temperature hydrothermal heating of <110–220 °C, confirmed by fluid and gas inclusion studies (**Fig. 68**) of VETŐ-ÁKOS (in KLEB et al. 1993a, KORPÁS et al. 1993); elevated mean vitrinite-reflectance values of 0.30–0.55 Ro% (**Fig. 69**), measured by HÁMOR-VIDÓ (in KLEB et al. 1993a, KORPÁS et al. 1993).

The depositional system of the Szépvölgy Limestone is considered by LELKES as a carbonate bank, located on a mobile dissected shelf. It represents a deepening upward sequence, interrupted by two low stand events before being drowned definitively at 37.7 Ma (KORPÁS et al. 1996).

The *tectonic style* of the formation will be outlined with reference to the microtectonic survey data (**Fig. 70**) of BENKOVICS and DUDKO (in KLEB et al. 1993a, b). The following characteristics should be noted: dominant steep dips of the strata (20–45°), oriented S–SE; dense fault system of WNW–ESE and ENE–WSW directions, composed of mainly normal, subordinately dextral and sinistral faults; signs of superimposed and multiphase dislocations along the same fault-plane; few synsedimentary faults, with an excellent example of lystric faults documented in the Pálvölgy cave (**Fig. 71**).

Palaeokarst of the Szépvölgy Limestone

The Szépvölgy Limestone hosts six major explored caves (**Figs. 54, 55**) and about 100 minor caves or cavities, described by LEÉL-ŐSSY (1995). Nowadays they are within the meteoric zone of the karst system, while only the Molnár János cave is located below it, in the phreatic zone. The limestone is dissected by three well developed cave levels, parallel to bedding (**Fig. 72**). All of them can be easily recognized in the caves and their vertical distance is oscillating between 8 and 12 m. The uppermost cave level is found at the boundary between the Szépvölgy Limestone and the Buda Marl cover. The individual cave levels are connected with each another by

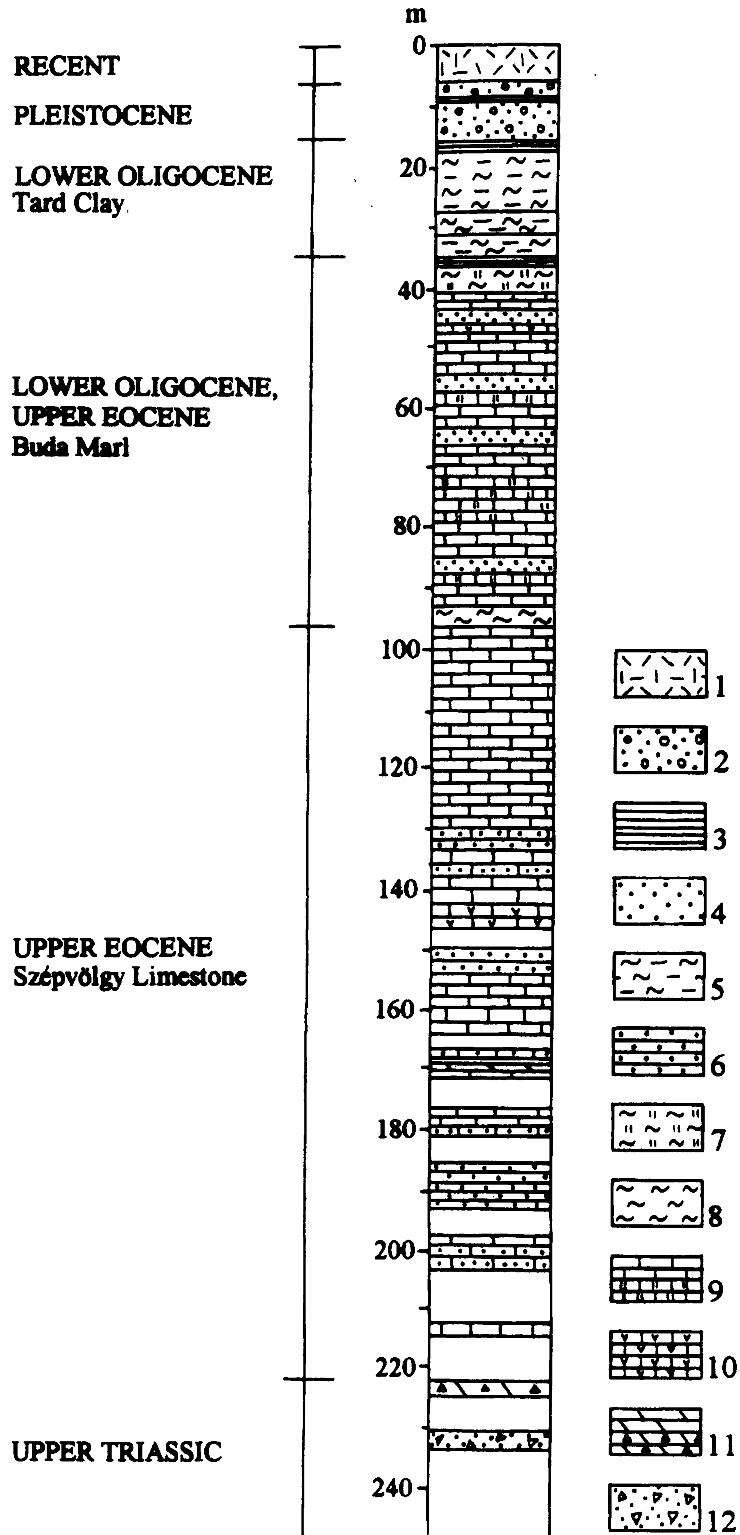


Fig. 61. Lithology and lithostratigraphy of Lukács-fürdő (L-VII) borehole (KLEB et al. 1993a)

1. Landfill, 2. Gravel, sandy gravel, 3. Clay, 4. Sand, 5. Claymarl, 6. Sandstone, 7. Silty marl, 8. Marl, 9. Limestone, silty limestone, 10. Tufaceous limestone, 11. Dolomite, dolomite breccia, 12. Rock fragments: dolomite, volcanogene and cherty sand

well developed open joint systems, roughly perpendicular to the bottom of the caves. The entire limestone section is dissected by about 10–12, mainly submarine, subordinately subaerial discontinuity surfaces (**Figs. 62, 63, 64, 65**). Two master surfaces are known: the first, partly subaerial one is located at the boundary of the lower coralgal unit and the overlying large foram-rich unit. The second, submarine one coincides with the contact of the Szépvölgy Limestone and the Buda Marl. Actually the major part of the surface and subsurface cave system is empty and its infillings can be grouped as follows: infilling sediments; hydrothermal mineral veins and silicified-kaolinitic zones; other mineral precipitations.

Infilling sediments: carbonate, siliciclastic and carbonate–siliciclastic types of infilling sediments are known in the palaeokarst system (**Figs. 64, 65, 73–80**). Their common features will be summarized in the following: rhythmic, laminated bedding; gradation; cross stratification; internal erosional surfaces; synsedimentary slump structures; presence of internal collapse breccias; accumulations of bioclasts composed of Late Eocene benthic and planktonic marine fossils; conformable and unconformable bedding with respect to the host rock.

Among the carbonate types the bioclastic early marine infillings of Late Eocene age will be outlined, known from the Pál-völgy cave, Mátyás-hegy cave and the

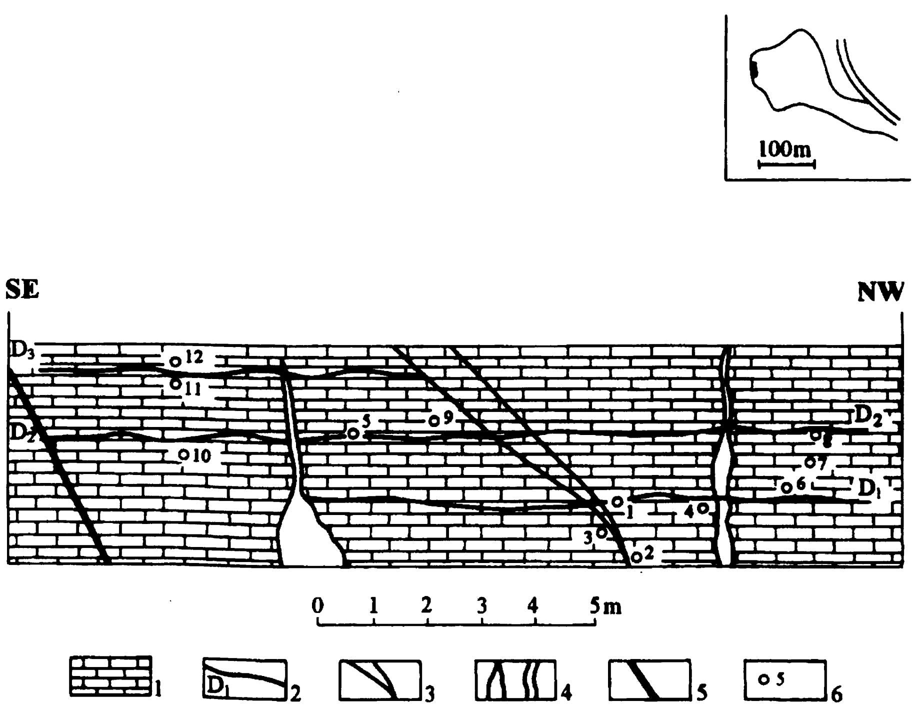


Fig. 62. Palaeokarst profile No. 1 of the Pál-völgy quarry (KORPÁS et al. 1993)
1. Szépvölgy Limestone, 2. Discontinuity surface, 3. Fractures, 4. Joints and vugs, 5. Fault, 6. Site of sampling

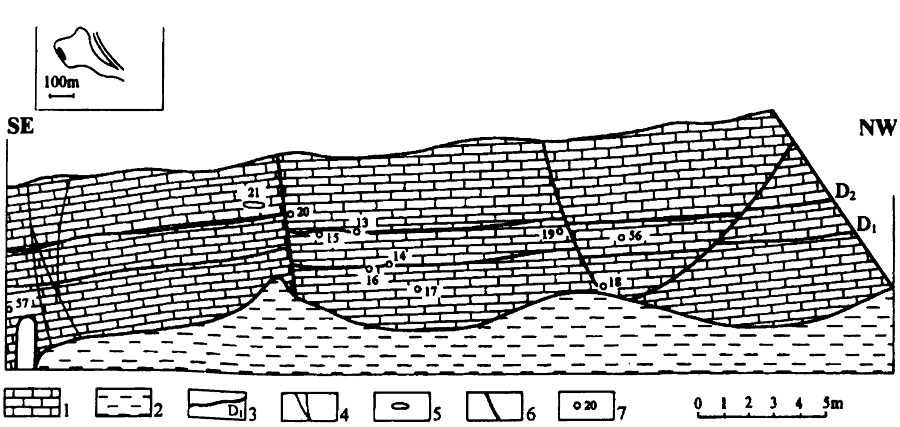


Fig. 63. Palaeokarst profile No. 2 of the Pál-völgy quarry (KORPÁS et al. 1993)
1. Szépvölgy Limestone, 2. Debris, 3. Discontinuity surface, 4. Fractures, 5. Vugs, 6. Fault, 7. Site of sampling

József-hegy cave, and further in the quarries of the Mátyás-hegy and of the Fenyőgyöngye (**Figs. 73–78**). These early marine infillings were interpreted by LELKES (in KLEB et al. 1993a, in KORPÁS et al. 1993, 1996) as rocky shore sediments, deposited at the level of wave-base or below it (**Fig. 66**). Among the diagenetic and hydrothermal effects slight dolomitization and silicification should be mentioned. Similar laminated carbonate infillings were described by KRAUS (1988) as "watch-glass" in the Mátyás-hegy cave, and by NAGYMAROSY (1986b) from the Út-hegy quarry, near Budaörs. Two, conformable and unconformable generations of

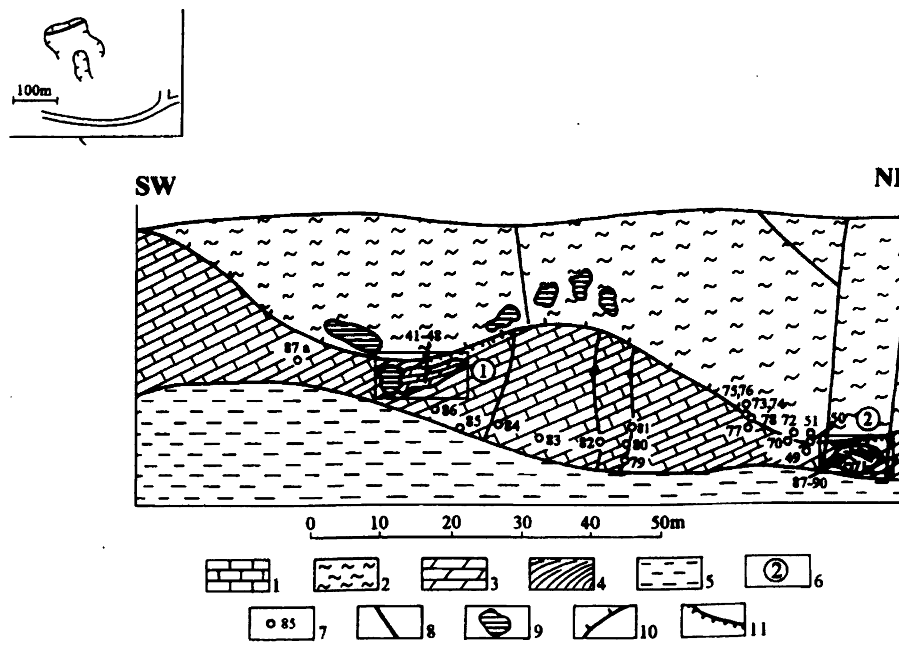


Fig. 64. Palaeokarst profile No. 1 of the Mátyás-hegy (SE) quarry, NW wall (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Buda Marl, 3. Late Eocene biocalcarenite-infilling, 4. Late Eocene laminitic-infilling, 5. Debris. 6. Detailed profile No. 1, 7. Site of sampling, 8. Fault, 9. Caves, 10. Reverse fault, 11. Erosional-tectonic contact

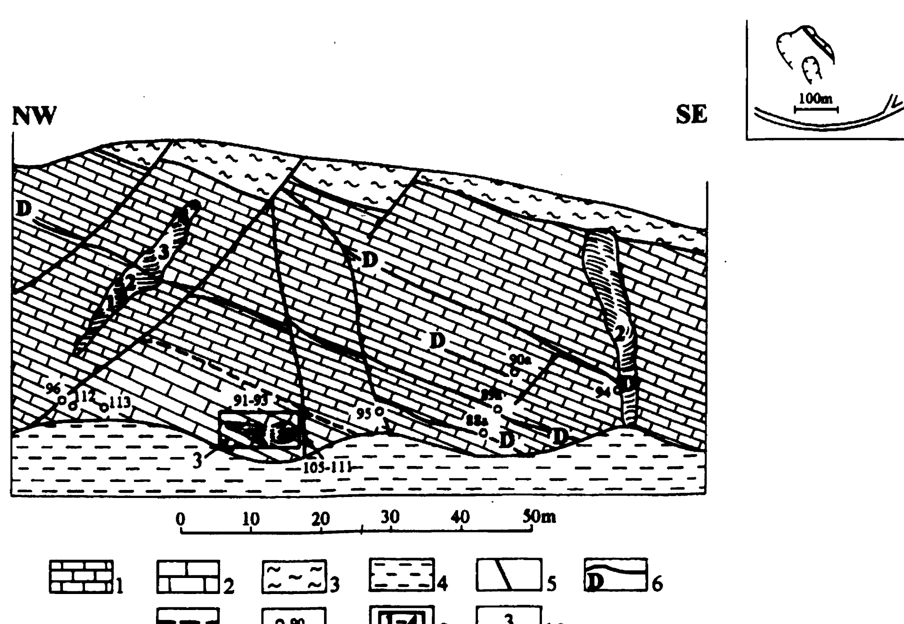


Fig. 65. Palaeokarst profile No. 2 of the Mátyás-hegy (SE) quarry, NE wall (KORPÁS et al. 1993)

1. Szépvölgy Limestone (Discocylinaean facies), 2. Szépvölgy Limestone (Algaean facies). 3. Buda Marl. 4. Debris, 5. Fault, 6. Discontinuity surface, 7. Facies boundary, 8. Site of sampling, 9/1–4. Late Eocene infilling generations, 10. Detailed profile No. 3

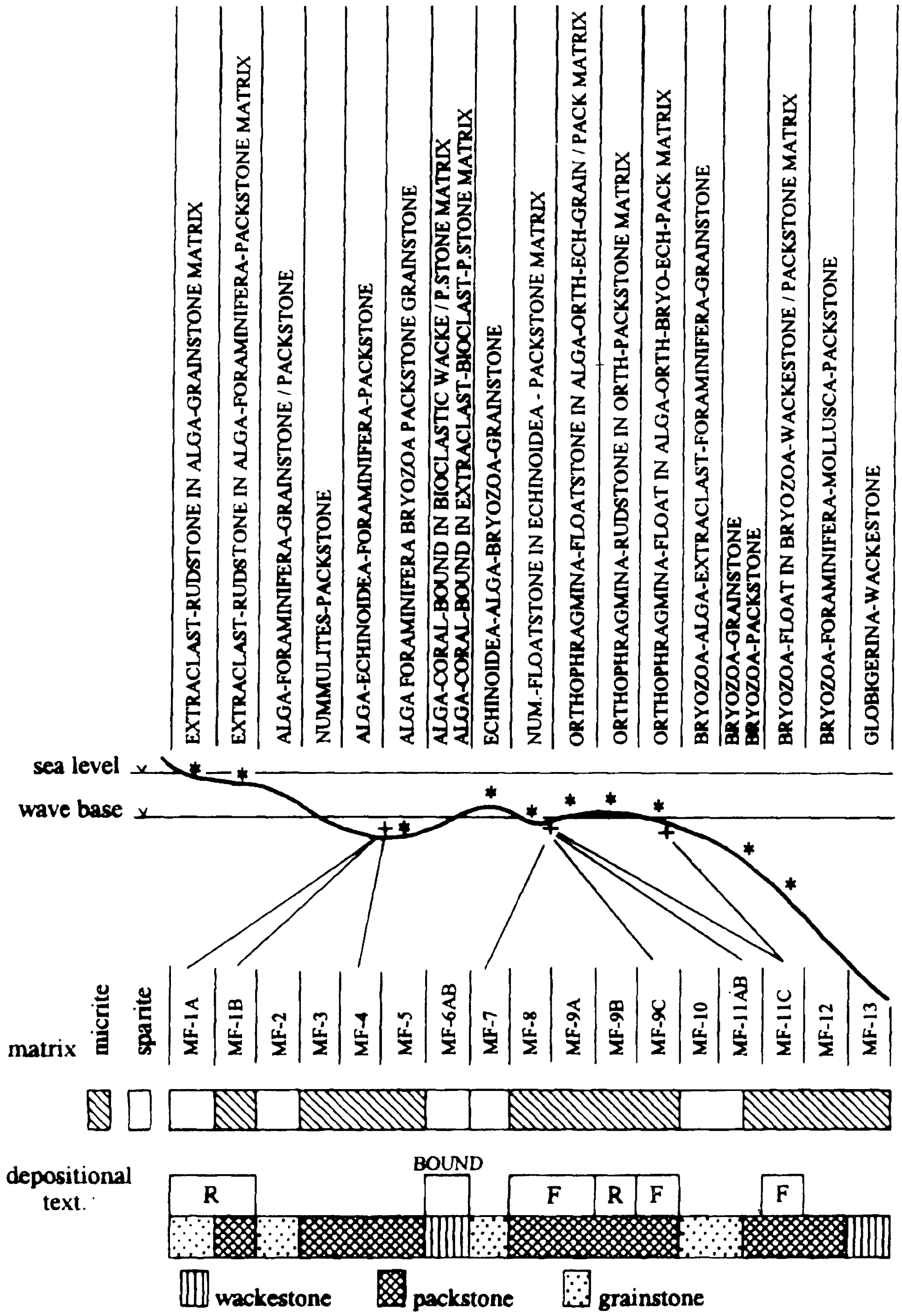


Fig. 66. Composite diagram of the Late Eocene microfacies types (KÁZMÉR 1985) indicating the major microfacies types of bedrock (*) and cave infillings (+) (KLEB et al. 1993a)

these carbonate laminites in the Út-hegy quarry are shown on **Fig. 79.**, while other occurrences of the Fenyőgyöngye quarry cited here were published by NÁDOR (1992a), NÁDOR and SÁSDI (1991) and by NÁDOR et al. (1993).

Different types of Late Eocene siliciclastic cavity fillings are known from the Mátyás-hegy quarry (Moby Dick cave) and from the Fenyőgyöngye quarry (**Figs. 80, 81**). The first one is unconformable with respect to the host-rock bedding, is an early cavity filling and is rich in benthic and planktonic foraminifera and nannofossils of Late Eocene age (KOLLÁNYI in KORPÁS et al. 1993). The Late Eocene cavity fillings of the Moby Dick cave are overlain by layers bearing quartz pebbles and fragments of Helix sp. (Quaternary in age).

The combination of carbonate-siliciclastic infilling sediments is known from the Mátyás-hegy quarry (**Figs. 64, 65, 75**) and consists of siliciclastic basal layers, rich in kaolinite and of carbonate laminites bearing Late Eocene microfossils in the upper part of the infilling.

The relative succession of the early marine infilling sediments was estimated on the basis of their conformable or unconformable bedding relative to the host rock. The formation of the first conformable generations (**Figs. 76, 77, 79**) was followed by a regional tilting of the Szépvölgy Limestone to the W and by the subsequent deposition of the unconformable infilling generations (**Figs. 73, 74, 78, 79**). Results of

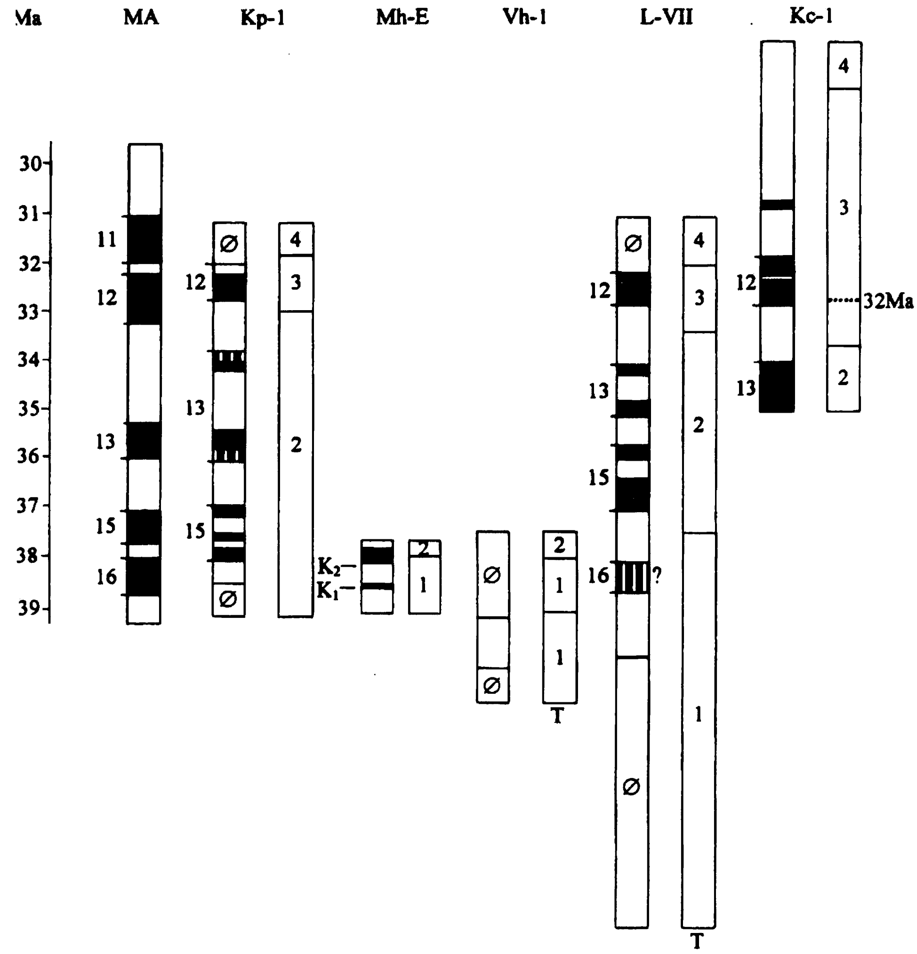


Fig. 67. Magnetostratigraphy of Late Eocene formations and related palaeokarst (after LANTOS 1994, 1995)

1. Szépvölgy Limestone, 2. Buda Marl, 3. Tard Clay, 4. Quaternary, T=Triassic, K_1=Conformable generation of early marine karstic infillings, K_2=Unconformable generation of early marine karstic infillings. Ma=Million age, MA=Magnetic anomaly, Kp-1=Borehole Kapy-1, Mh-E=Mátyás-hegy, SE quarry, Vh-1=Borehole Vérhalom-1, L-VII=Borehole Lukácsfürdő-VII, Kc-1=Borehole Kiscell-1

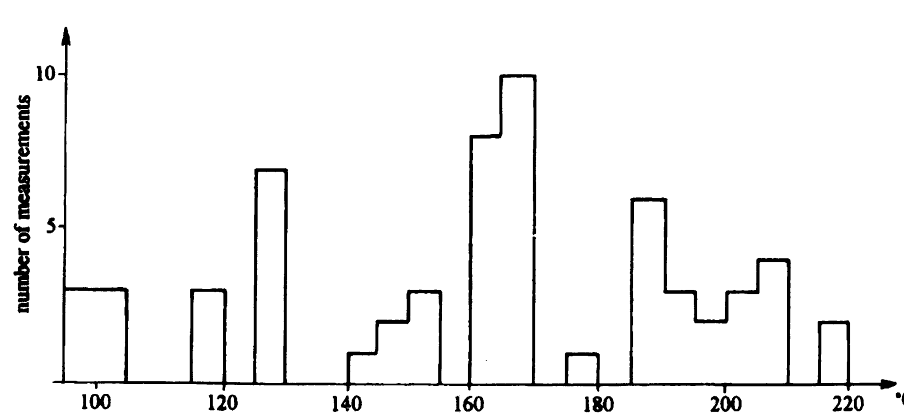


Fig. 68. Fluid and gas inclusion homogenisation temperatures of calcites (Szépvölgy Limestone, Mátyás-hegy and Pál-völgy quarries) (VETŐ ÁKOS in KORPÁS et al. 1993)

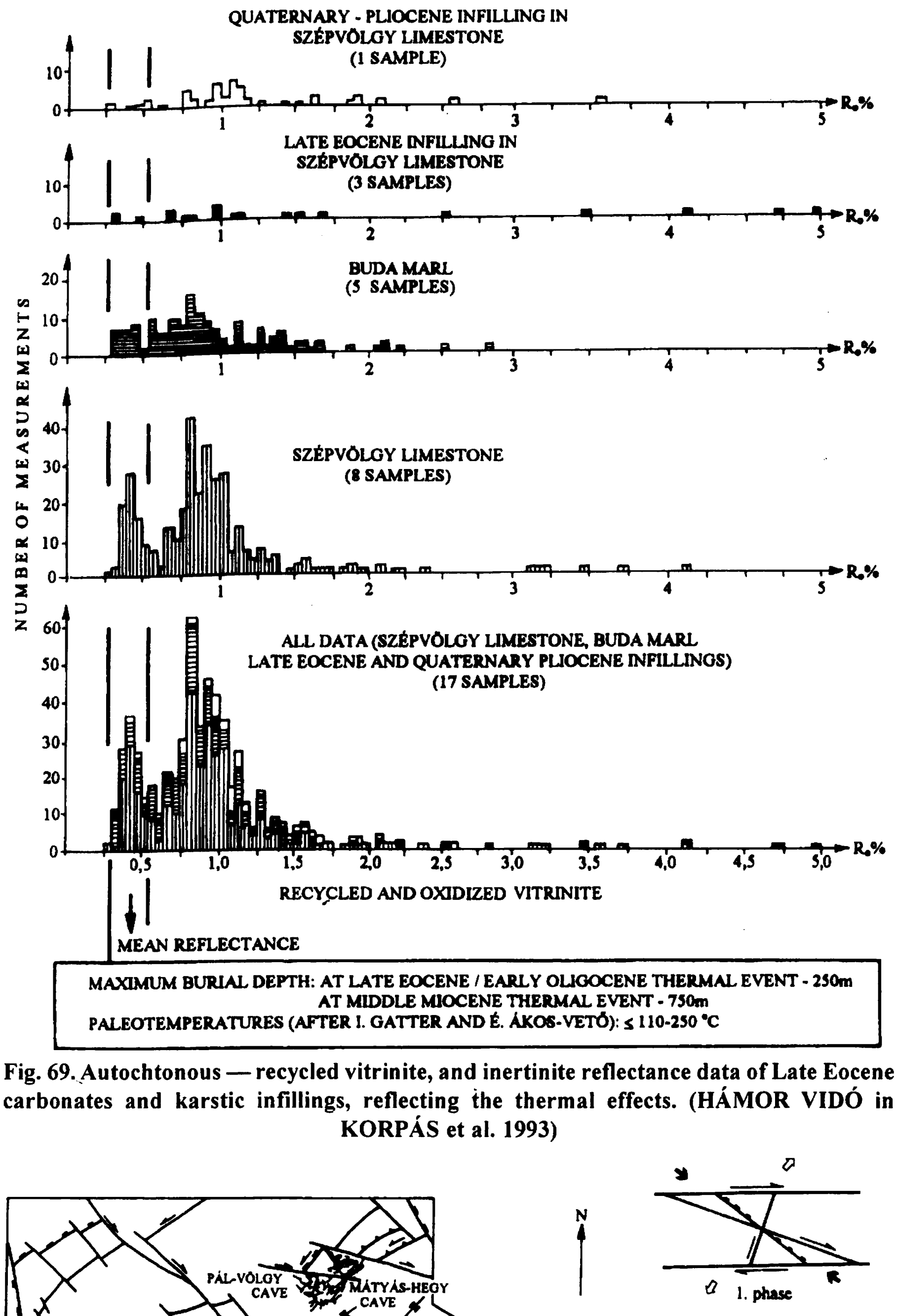


Fig. 69. Autochtonous — recycled vitrinite, and inertinite reflectance data of Late Eocene carbonates and karstic infillings, reflecting the thermal effects. (HÁMOR VIDÓ in KORPÁS et al. 1993)

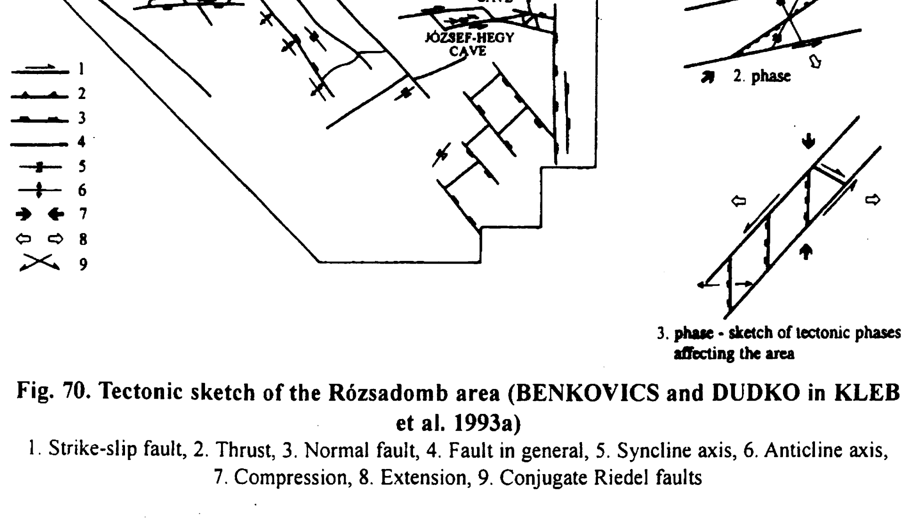


Fig. 70. Tectonic sketch of the Rózsadomb area (BENKOVICS and DUDKO in KLEB et al. 1993a)

1. Strike-slip fault, 2. Thrust, 3. Normal fault, 4. Fault in general, 5. Syncline axis, 6. Anticline axis, 7. Compression, 8. Extension, 9. Conjugate Riedel faults

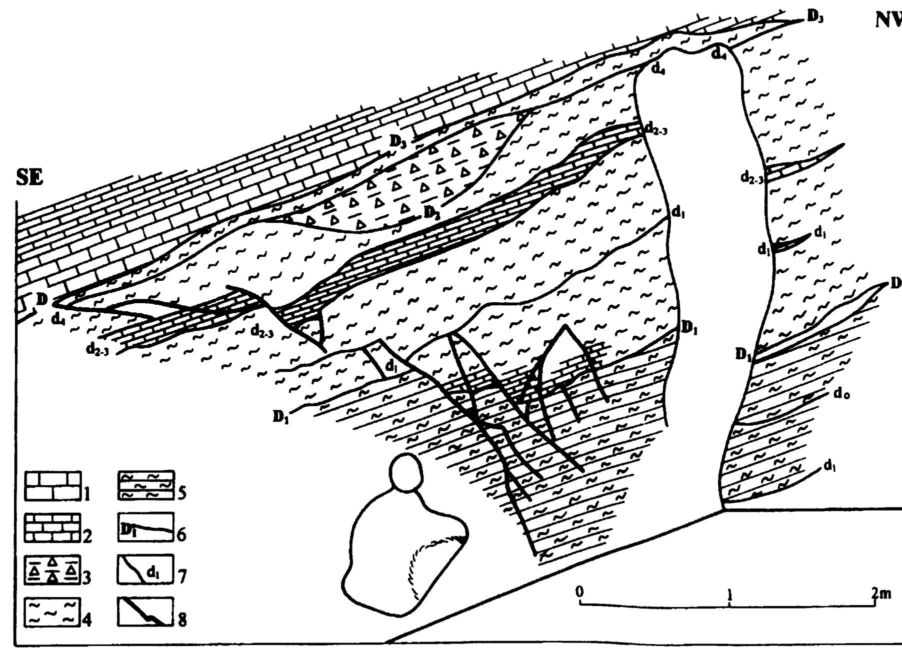


Fig. 71. Late Eocene synsedimentary lystric faults at the boundary of the Szépvölgy Limestone and the Buda Marl, Pál-völgy cave (Hajós terem) (KORPÁS et al. 1993)

1. Bedded limestone, 2. Laminated limestone, 3. Channel infilling-breccia, 4. Marl, 5. Muddy limestone. 6–7. Discontinuity surfaces, 8. Synsedimentary lystric faults

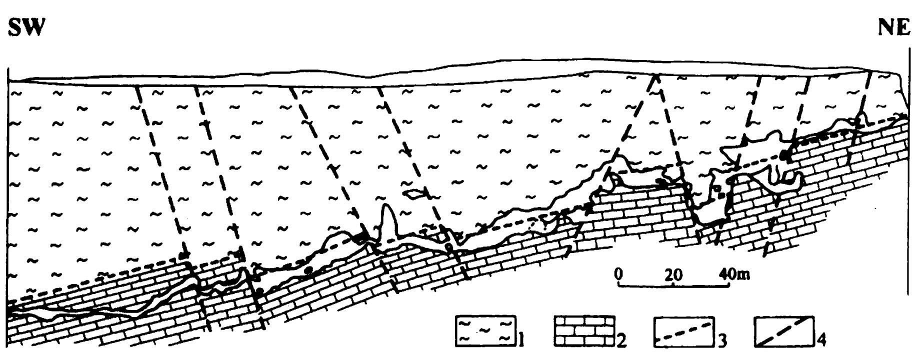


Fig. 72. Geological profile of the Pál-völgy cave, along the dip (after KOVALÓCZY 1987)

1. Buda Marl, 2. Szépvölgy Limestone, 3. Boundary of formation, 4. Normal fault

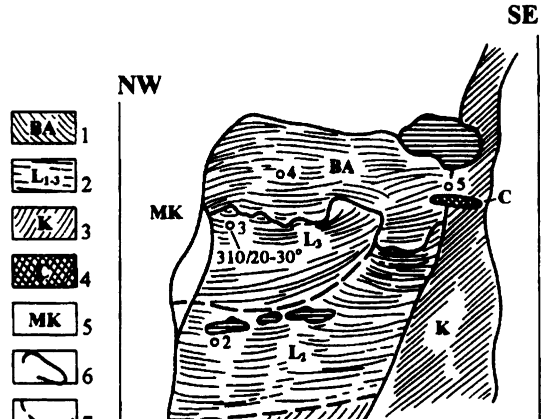


Fig. 73. Early marine carbonate-infillings of the Szépvölgy Limestone, in the Pál-völgy cave (Hajós terem — Gyöngyös) (KORPÁS et al. 1993)

1. Brecciated marl, 2. Laminated biocalcarenite, 3. Kaolinitic-silicified infillings, 4. Drusy calcite. 5. Bedrock of Szépvölgy Limestone, 6. Lithological boundary, 7. Cycle boundary, 8. Vugs, 9. Site of sampling

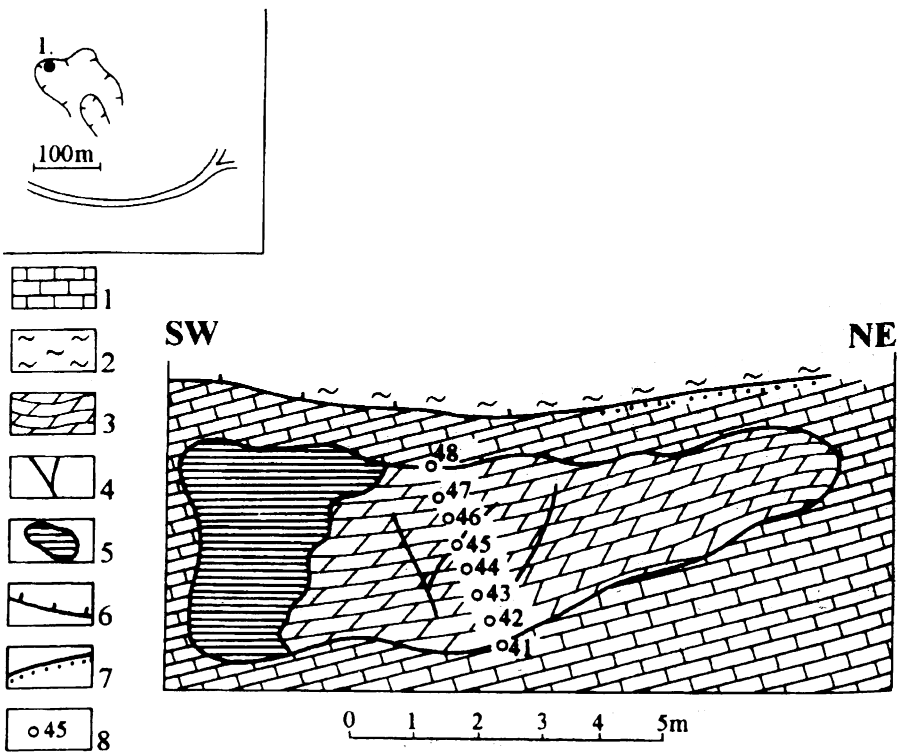


Fig. 74. Detailed palaeokarst profile No. 1 of the Szépvölgy Limestone (Mátyás-hegy, SE quarry, NW wall) (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Buda Marl, 3. Late Eocene karstic biocalcarenite-infilling, 4. Fracture, 5. Cave, 6. Reverse fault, 7. Erosional-tectonic contact, 8. Site of sampling

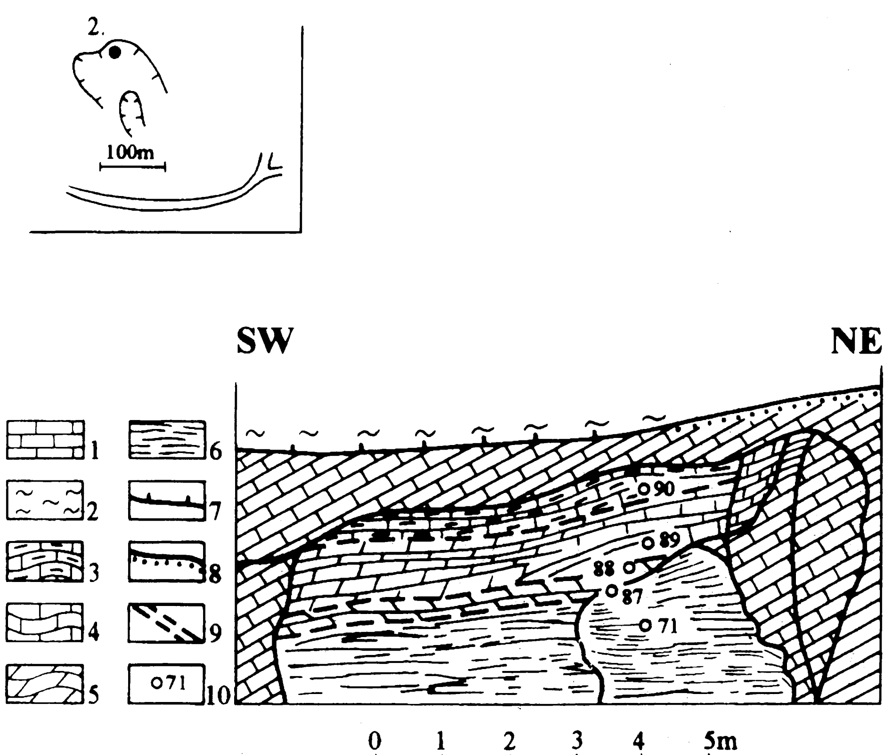


Fig. 75. Detailed palaeokarst profile No. 2 of the Szépvölgy Limestone (Mátyás-hegy, SE quarry, NW wall) (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Buda Marl, 3–6. Late Eocene infillings: 3. Muddy limestone, 4. Laminated bioclastic limestone, 5. Dolomitized limestone, 6. Laminated kaolinitic silt, 7. Reverse fault, 8. Erosional-tectonic contact, 9. Cycle boundary, 10. Site of sampling

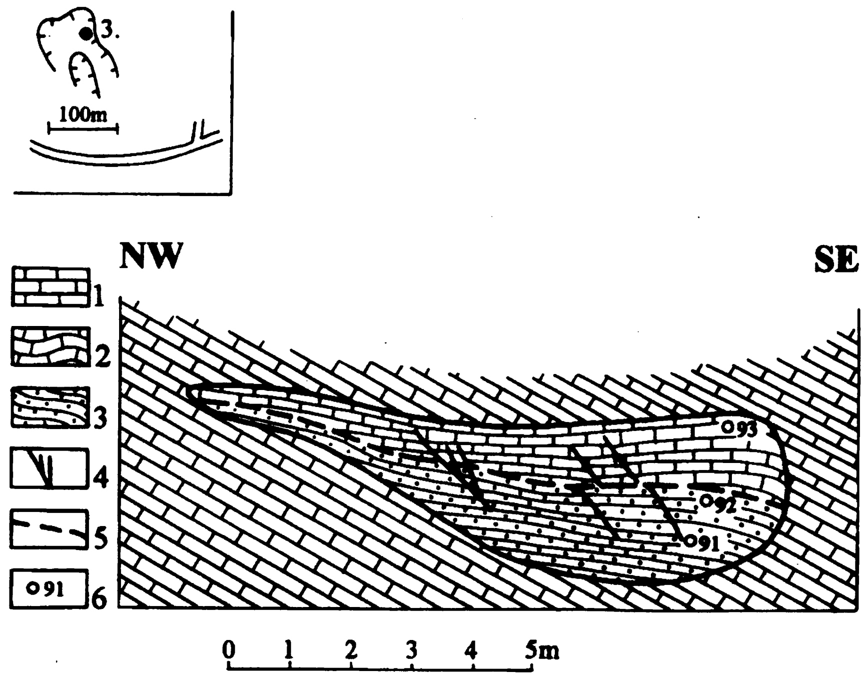


Fig. 76. Detailed palaeokarst profile No. 3 (Mátyás-hegy, SE quarry, NE wall) (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2–3. Late Eocene infillings: 2. Laminated, micritic, bioclastic limestone; 3. Laminated calcareous sandstone with extraclasts, 4. Fracture, 5. Cycle boundary, 6. Site of sampling

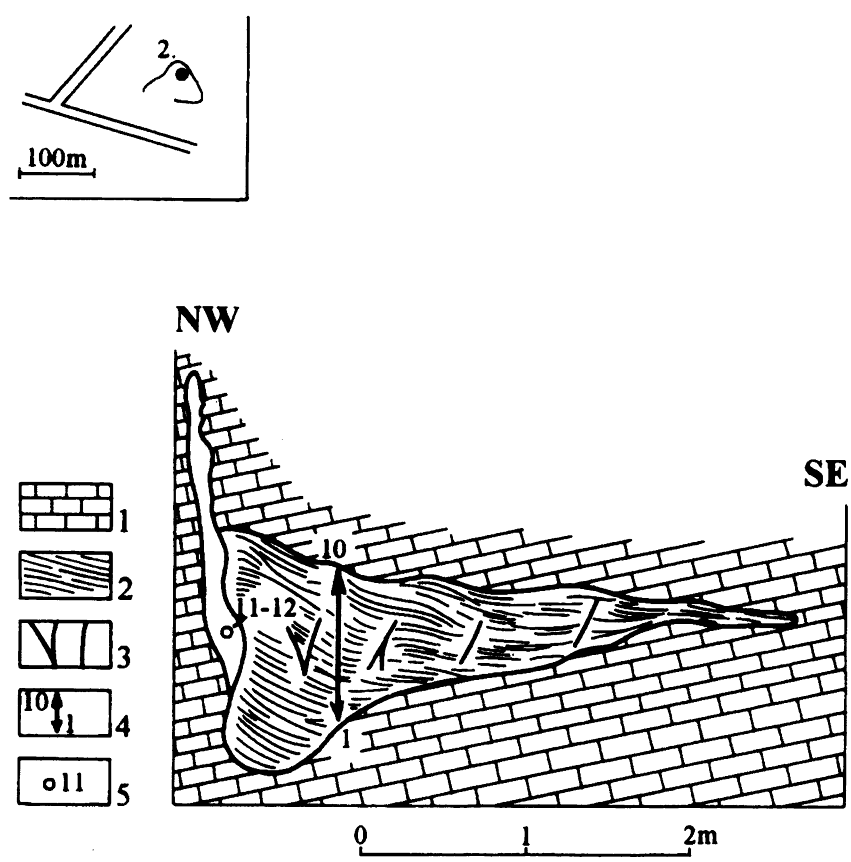
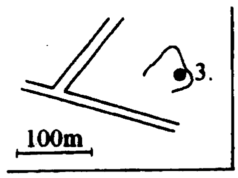


Fig. 77. Detailed palaeokarst profile No. 2, Fenyőgyöngye quarry (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Late Eocene carbonate-infilling:laminated, graded, bioclastic limestone. 3. Fracture, 4. Profile of sampling, 5. Site of sampling



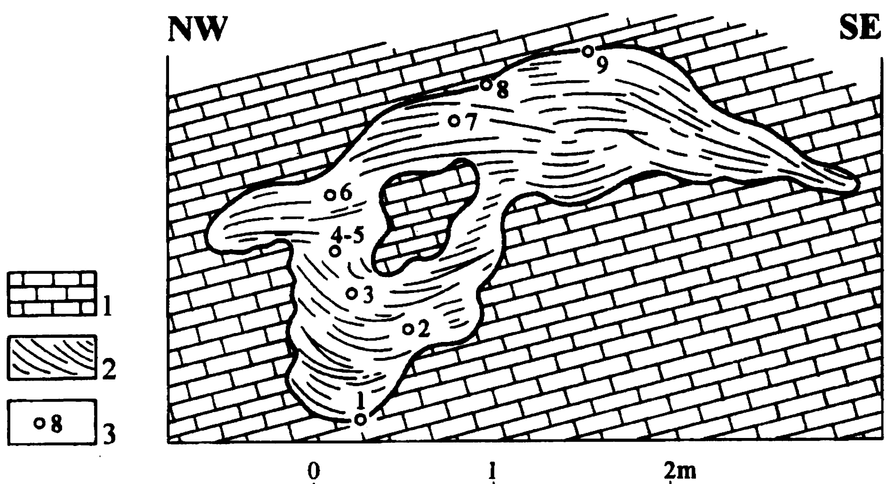


Fig. 78. Detailed palaeokarst profile No. 3, Fenyőgyöngye quarry (KORPÁS et al. 1993)
1. Szépvölgy Limestone, 2. Late Eocene carbonate-infilling:biocalcarenite, 3. Site of sampling

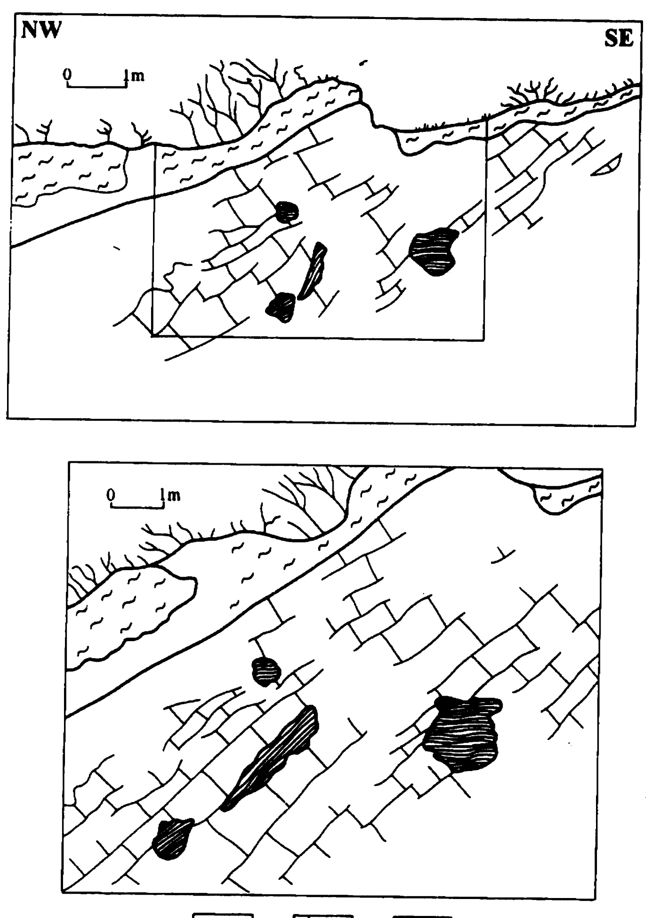


Fig. 79. Palaeokarst profile of the Út-hegy quarry, Budaörs
1. Buda Marl, 2. Szépvölgy Limestone, 3. Conformable and unconformable generations of Late Eocene marine laminated biocalcarenites

magnetostratigraphic study (LANTOS 1995, KORPÁS et al. 1996) of these infillings suggest an age of 37.5 Ma and a duration of less than 60 Ka for the formation of the conformable generations, while a timespan and duration of 36,8–36,5 Ma are set for the unconformable ones (**Fig. 67**). The westward tilting of the Late Eocene carbonate bank should be fitted between 37.5–36.8 Ma. (Absolute age determinations are referred to HAQ et al. 1987.)

Hydrothermal veins and silicified/kaolinitic zones: the silicified, argillitized joint infillings explored mainly in the caves are outlined here. The joints are some centimetres to decimetres wide, frequently connecting the individual cave levels to each other. The hydrothermal origin of some of them in the Mátyás-hegy cave has been proven by mineralogical and geochemical analysis (NÁDOR 1992a). At the same time, types of completely silicified and argillitized veins in the Pál-völgy cave were observed by us, consisting of redeposited, disoriented fragments of shells, bryozoans, echinoids and of large foraminifera in abundance. The hydrothermal vein of calcite explored in the Fenyőgyöngye quarry is 60 centimetres wide and strikes to W–E. It should be considered a special type of hydrothermal infillings. It is composed of three generations of drusy calcite and consists of well cemented collapse breccias with angular clasts of wall rock.

Other mineral precipitations after BOGNÁR (1992) and NÁDOR (1992a) are botryoidal speleothems, popcorns, floating rafts, cauliflowers, needles and dripstones. The following individual minerals can be seen: calcite, aragonite, barite, dolomite, ankerite, magnesite, gypsum, fluorite, quartz, pyrite, hematite and goethite, giving the extraordinary beauty and natural value of the caves.

The karst facies of the Szépvölgy Limestone and of its infillings should be evaluated using the stable isotope data (**Fig. 82**) of FORD and TAKÁCS BOLNER (1991), NÁDOR, (1992a) and of HERTELENDI (in KORPÁS et al. 1993). According to them it can be stated that the marine environment of the Szépvölgy Limestone and the subsequent hydrothermal effects have played the decisive role in the formation and evolution of the palaeokarst system. Vadose processes have been insignificant.

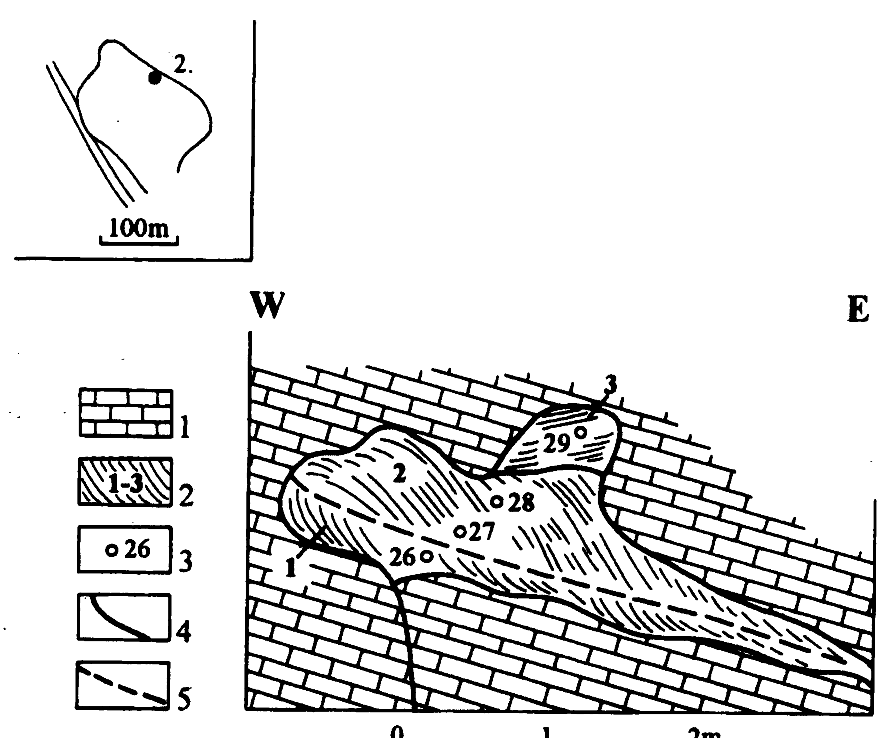


Fig. 80. Detailed palaeokarst profile No. 2. of the Mátyás-hegy W quarry (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Clastic infilling generations: 2/1. Late Eocene clastic, kaolinitic silt, 2/2. Late Eocene siltstone with lenses of sandstone, 2/3. Quaternary quartz sand, 3. Site of sampling, 4. Boundary of formation, 5. Cycle boundary

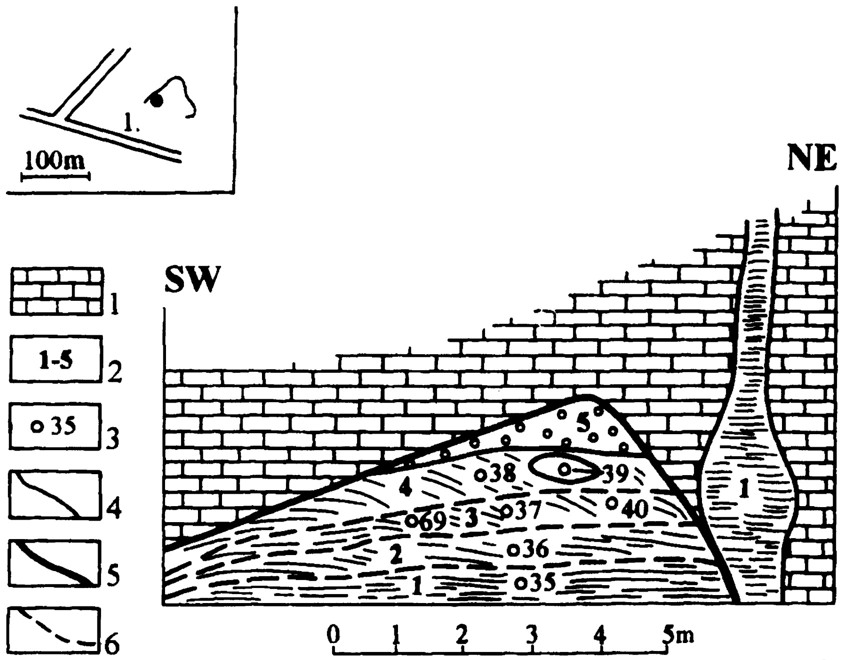


Fig. 81. Detailed palaeokarst profile No. 1 of the Fenyőgyöngye quarry (KORPÁS et al. 1993)

1. Szépvölgy Limestone, 2. Late Eocene clastic infilling generations: 2/1. Kaolinitic laminated sandstone and siltstone. 2/2. Sandstone and kaolinitic marl, 2/3. Pebble bearing sandstone, 2/4. Silicified conglomerate with extraclasts, 2/5. Autoclastic conglomerate, 3. Site of sampling, 4. Boundary of formation, 5. Fault, 6. Cycle boundary

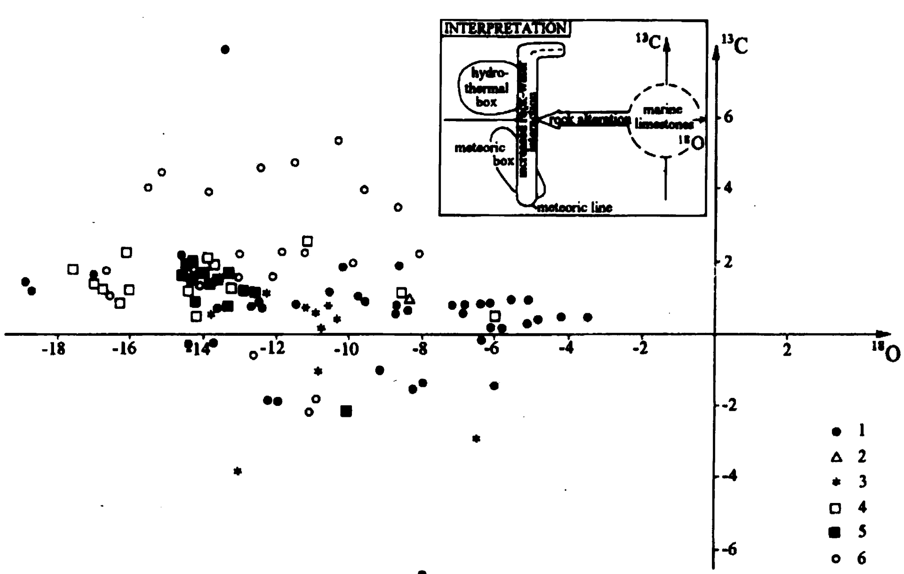


Fig. 82. Summary diagram of stable isotope analysis (KLEB et al. 1993a)

1. Szépvölgy Limestone, 2. Buda Marl, 3. Laminated cavity infilling, 4. Calcite fissures, 5. Floating rafts, 6. Botryoids

The formation covers a great part of the study area (**Fig. 54**), while it is traversed in the subsurface by extense cave systems and by many boreholes. Representative borehole logs are given in **Figs. 61, 83**.

The *lithology* of the Buda Marl, 60 to 120 m in thickness consists of flaser-bedded and laminated layers with a variable amount of clay (10–30%) and carbonate (70–90%). Its upper horizons include frequently strata of calcareous or siliciclastic sandstones. The entire profile consists of redeposited volcanomictic layers of andesite-sand. Bioclastic calcareous turbidites, described by BÁLDI et al. (1984a, b)

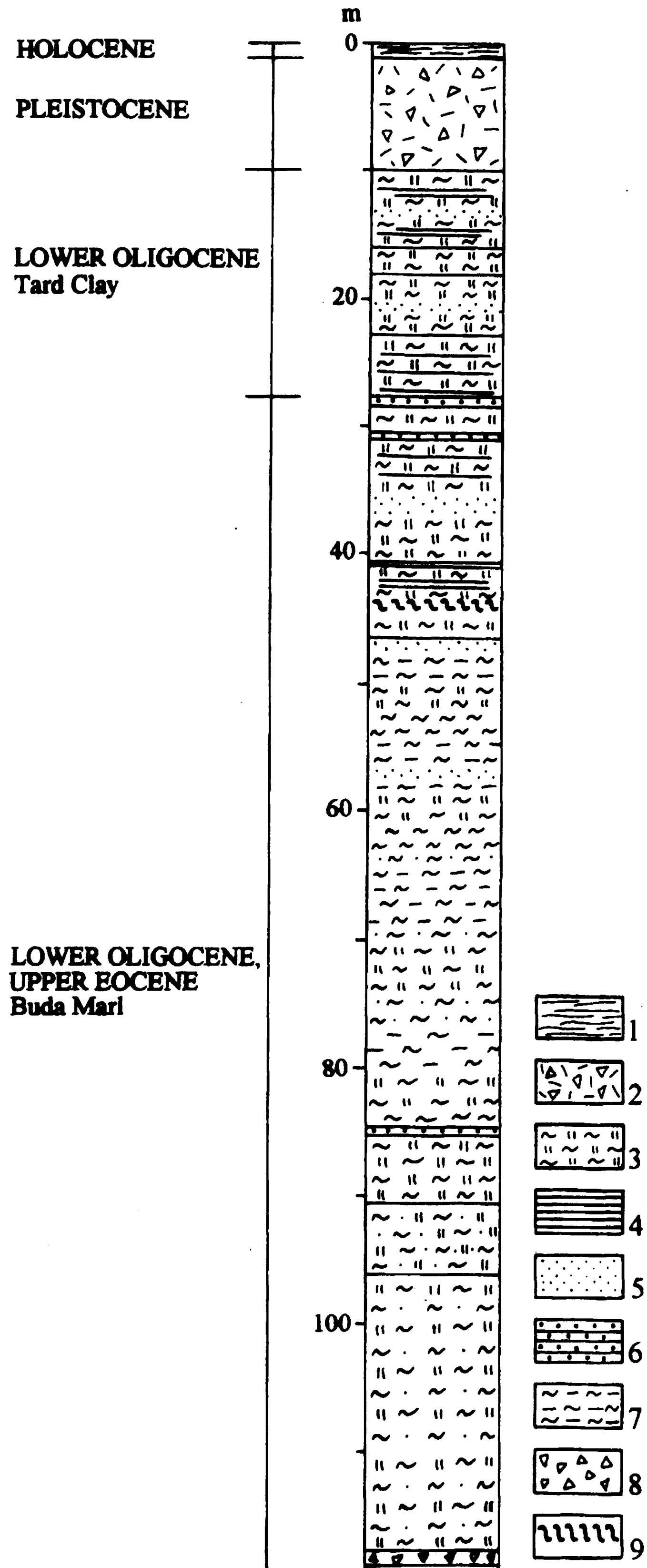


Fig. 83. Lithology and lithostratigraphy of Kapy-utca (Kp–1) borehole (KLEB et al. 1993a)
1. Soil. 2. Slope debris. 3. Silty marl. 4. Clay. 5. Sand. 6. Sandstone. 7. Claymarl. 8. Breccia. 9. Slumps

and by NAGYMAROSY (1986a) as allodapic limestones are very common in the basal horizons. The dispersed organic matter amounts to about 1% (BRUKNER-WEIN et al. 1990, VETŐ and HETÉNYI 1991), accumulated on the surface of the bedding planes, consists of carbonized fragments of plants. The formation is rich in fossils, composed of mainly planktonic foraminifera, coccoliths, sporomorphs and pollens as well as, benthic forms (a few molluscs, large foraminifera, ostracods, algae, echinoids, bryozoans) redeposited and accumulated in turbiditic layers. Dominant microfacies (TÖRÖK in KLEB et al. 1993a, b) are bryozoan packstone-floatstone in the lower portion of the formation and globigerina wackestone in its upper part. Estimated age based on nannoplankton (BÁLDI-BEKE 1977, 1984, NAGYMAROSY in KORPÁS et al. 1993), on planktonic foraminifers (HORVÁTH KOLLÁNYI in KORPÁS et al. 1993), respectively on sporomorphs and pollen (RÁKOSI in KORPÁS et al. 1993) is Late Eocene to Early Oligocene: NP zones 19/20–21. Magnetostratigraphic data of LANTOS (1994, 1995) suggest an age of 37.7–34.2 Ma for deposition.

Among the *sedimentological, early diagenetic and/or hydrothermal* phenomena the following will be outlined: sharp lower and gradational upper contacts of the strata; graded bedding; autoclastic breccias; synsedimentary slump structures; presence of allodapic limestones; synsedimentary (sometimes overturned) folds; early diagenetic pyrite; early solutional and fracture porosity; well developed fracture system with hydrothermal calcite; high mean values of vitrinite reflectance (**Fig. 69**); hydrothermal silicification and argillitization.

The *depositional system* should be considered a narrow, mobile pelagic shelf margin, deep outer shelf and slope with dissected morphology (TÖRÖK in KLEB et al. 1993a, b).

The *tectonic style* will be characterized after BENKOVICS and DUDKO (in KLEB et al. 1993a, b): highly folded formation, ordered in syncline structures of NE–SW trend (**Figs. 54, 70**); dense fault system oriented WNW–ESE and ENE—WSW; predominance of normal faults, with frequent dextral and sinistral faults and a few low angle reverse faults; signs of multiphase dislocations along the same fault plane; presence of synsedimentary, sometimes overturned folds, described by us in the Melocco quarry and in the boreholes Lukács-fürdő L-VII, Kapy utca Kp-1.

Palaeokarst of the Buda Marl

The six major caves (**Figs. 54, 55**) and the greater part of the 100 minor caves and cavities are located partly inside the Buda Marl. The uppermost passages of the caves penetrate its basal horizons with a sudden ceasing in it. The whole section of the formation is dissected by a dense, still open fracture system, with a width of some millimetres to some centimetres giving it a high porosity of 8–13% (KLEB et al. 1993a, b). The fractures are partly infilled by hydrothermal minerals, like calcite, crystalline pyrite and sometimes by quartz. Slight dissolutional phenomena were observed on the crystalline calcite in drilling cores.

OLIGOCENE
Tard Clay

Since the middle of the 70's this formation has been the target of detailed stratigraphic, sedimentological, mineralogical and geochemical studies (BÁLDI-BEKE 1977, BÁLDI 1980, 1983, 1984, KORPÁS 1981, VARGA 1985, BÁLDI and BÁLDI-BEKE 1985, BOGNÁR 1985, BRUKNER-WEIN et al. 1985a, b, 1990, NAGYMAROSY 1983, 1985, 1992, NAGYMAROSY et al. 1986, VETŐ 1987, NAGYMAROSY and BÁLDI-BEKE 1988, NAGYMAROSY in FODOR et al. 1991), therefore its characterization is based mainly on the above cited works.

Surface outcrops of the formation, 100–120 m in thickness, are concentrated in the morphologically deeper valleys of the area having a WSW, S and ESE strike (**Fig. 54**). Representative profiles of its lower portion are illustrated in **Figs. 61, 83**.

Concerning *lithology* the formation consists of rhythmic, laminated layers free of or poor in carbonate (<10%). Its basal horizons contain frequently beds of bioclastic and allodapic limestones (VARGA 1985) or by redeposited siliciclastic layers similar to the Hárshegy Sandstone. The entire section consists of volcanoclastic intercalations, andesitic in composition and produced by coeval volcanism. The formation is rich in bitumen with a content of 1.0–2.6% total organic C and with extractable bitumen of 380–1450 ppm. Content of total S oscillates between 0.98–2.37%, with pyrite bounded sulphur of 0.17–1.66% (BRUKNER-WEIN et al. 1985a, b, 1990). The source of immature kerogen is terrestrial vegetation. This is confirmed by assembleges of sporomorphs and pollens, respectively by fragments of resin and pines, described by RÁKOSI (in KORPÁS et al. 1993). The formation is rich in planktonic fossils, first of all in coccoliths, with a few pteropods, while the scarce benthic elements (bivalves, foraminifera, algae) are mainly redeposited. The uppermost levels contain well preserved, carbonized plant debris and fossil fishes in mass quantity. Age constraints based on nannoplankton (BÁLDI 1983, 1984, BÁLDI and BÁLDI-BEKE 1985, NAGYMAROSY 1985, 1992, 1993, NAGYMAROSY and BÁLDI-BEKE 1988), on radiometric K-Ar measurements (BALOGH 1985, NAGYMAROSY et al. 1986) fission track analysis (DUNKL and NAGYMAROSY 1992), magnetostratigraphic studies (MÁRTON in NAGYMAROSY 1992, LANTOS 1994, 1995) are the following: nannoplankton of NP–(21)–22–23 zones; K-Ar ages of 32.25±0.9 Ma; fission track age of 32.45±0.54 Ma; 12 and 11 magnetostratigraphic anomalies; depositional record of 34.2–30 Ma according to HAQ et al. (1987).

Sedimentological and early diagenetic features: frequent synsedimentary slumps and folds; gradation and redeposition; presence of early diagenetic bacterial pyrite; hydrothermal silicification; partly open fracture system infilled by white calcite and crystalline pyrite.

The *depositional system* is considered an anoxic, isolated bathyal basin with periodic oceanic connections.

The *tectonic style* of the formation will be characterized after WEIN (1977), BALLA and DUDKO (1990), NAGYMAROSY in FODOR et al. (1991), and FODOR et al. (1994). The highly folded strata of the Tard Clay are ordered along the same syncline structures of the Buda Marl (**Figs. 54, 70**). Some of these folds, observed mainly in drilling cores should be considered synsedimentary in origin.

The formation does not show any signs of karstification.

Kiscell Clay

The formation is 400 m in thickness, it covers the valleys, located at the W, SW and S borders of the area (**Fig. 54**). It consists of monotonous beds of silts and silty clays with a few intercalations of fine to coarse grained sandstones. The dominant clay minerals are illite, subordinately kaolinite, while the amount of carbonate is less than 25%. The bioturbated rocks of the formation are rich in molluscs, while microfossils (planktonic and benthic foraminifera, coccoliths, sporomorphs and pollen) are concentrated in certain horizons. The depositional environment was the neritic to bathyal zone of a normal marine open sea.

QUATERNARY

The Triassic and Palaeogene formations are covered by a Quaternary complex 15 m in thickness. This cover consists of alluvial gravels and sands, eolian loess and freshwater limestone.

History of evolution

The Late Triassic formations of the area formed part of a carbonate platform located at the northern passive continental margin of the African plate (HAAS 1989). This platform margin position, according to the data of HORVÁTH et al. (1983) was maintained until the Late Cretaceous. The present day boundaries of the Late Triassic formations reflect the original zonation (WEIN 1977) and their main units (KÁZMÉR and KOVÁCS 1985, HAAS 1989) are as follows:

Carbonate platform — Maindolomite, Dachstein Limestone,

Carbonate platform margin — Sashegy Dolomite, Mátyáshegy Formation (platform margin member),

Intraplatform basin — Mátyáshegy Formation (basin member),

Oceanic basin — Hallstatt Limestone.

The carbonate platform disintegrated due to the Late Triassic–Early Jurassic rifting, resulting in the definitive uplift and subaerial exposure of the platform segment of the Buda Hills. Continuous Late Triassic–Early Jurassic sedimentation should be expected only in the area of the intraplatform basin. The long-term subaerial exposure period of the uplifted carbonate platform segment started at the end of the Triassic and terminated in the Late Eocene. This has resulted in erosion and conti-

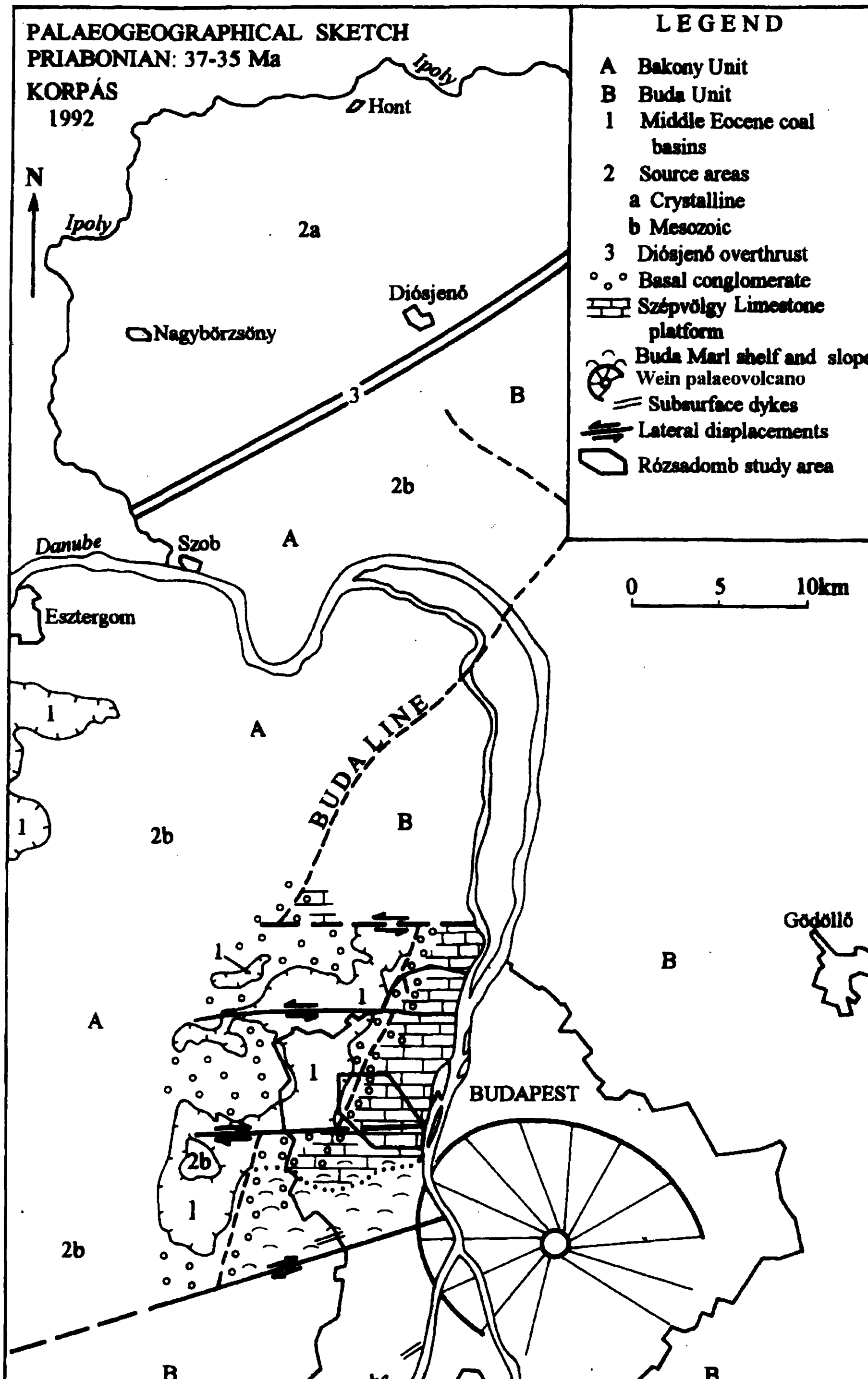


Fig. 84.

nental karstification of the flat, undissected carbonate plateau and in the deposition of some bauxites during the Cretaceous. The carbonate platform, including the Buda Hills segment separated from the African margin in the Senonian, due to a renewed continental rifting. This rifting is dated in the Buda Hills by the alkaline ultrabasic carbonatites and basalts of about 70–60 Ma (HORVÁTH et al. 1983, BALLA 1988, WÉBER 1989, EMBEY ISZTIN et al. 1989).

The Late Eocene subduction and collision in the foreland of the Southern Alps (KÁZMÉR 1984, KÁZMÉR and KOVÁCS 1985, BALLA 1988) should be considered the next important step in the evolution of the area. The Buda Line (BÁLDI and

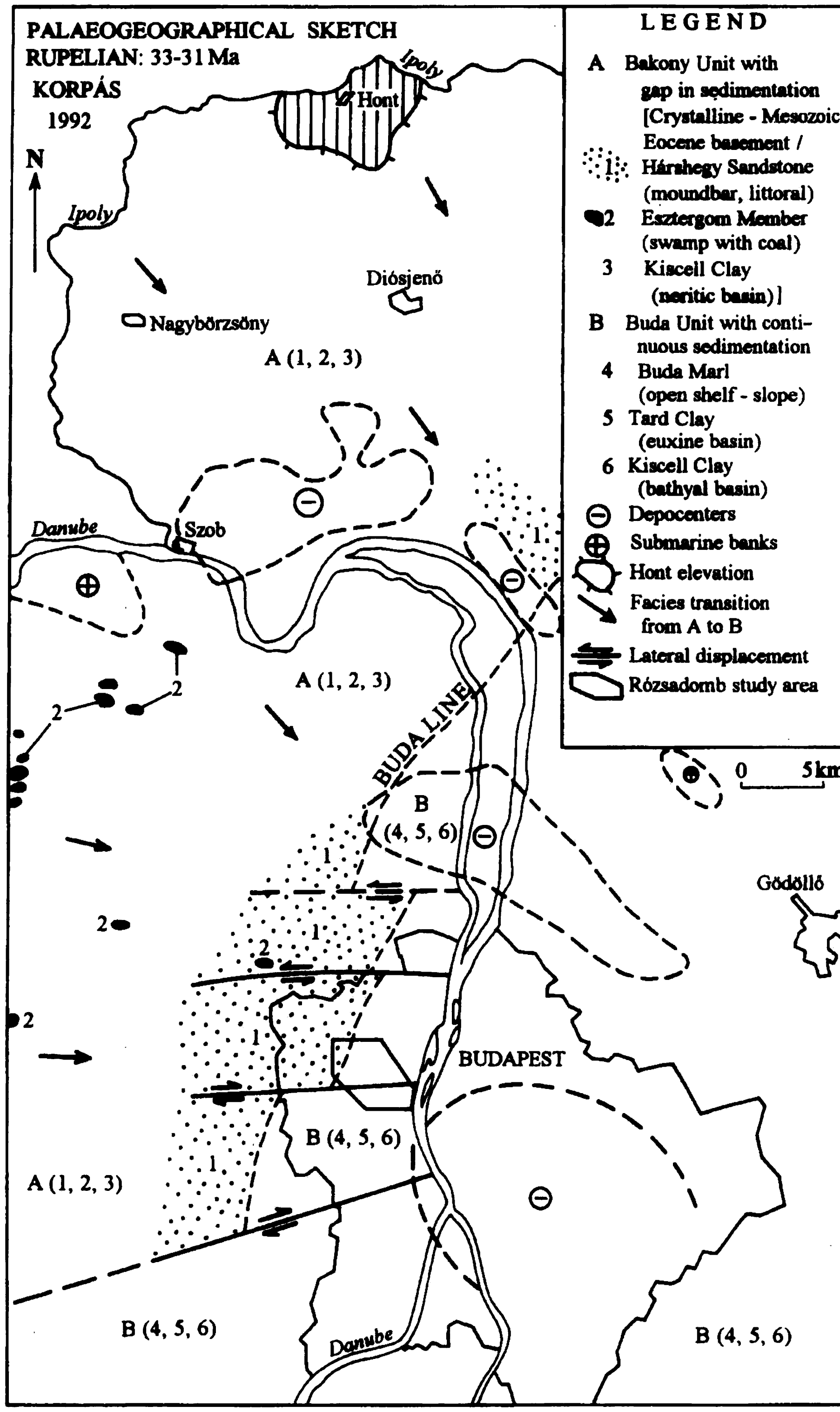


Fig. 85.

NAGYMAROSY 1976, BALÁZS et al. 1981) is the most important palaeogeographical boundary, separating the intracontinental Bakony Unit to the NW from the epicontinental Buda Unit in the SE (**Fig. 84**). The Buda Line played a decisive role in Late Eocene–Early Oligocene sedimentation (**Figs. 84, 85**).

Timespan of 37–35 Ma (**Fig. 84**): The carbonate bank of the Szépvölgy Limestone with an area of 75 km^2 formed along the narrow dissected mobile shelf, located to the SE of the Buda Line. It was bordered in the S by a composite fandelta system accumulated in the shoreface zone, and prograded to SSW and SE (MAGYARI in FODOR et al. 1991, 1994). The carbonate shelf drowned definitively at 37.7 Ma and was covered by the pelagic slope complex of the Buda Marl.

The Eocene–Oligocene volcaniclastics, the shallow intrusive igneous bodies and dikes as well as the hydrothermal alteration and related oreshows of the Buda Hills have been known for a long time (HOFMANN 1871, KOCH 1908, PÁVAY VAJNA 1912, SZÉKY-FUX and BARABÁS 1953, SZÉKY-FUX 1957, HORUSITZKY and WEIN 1962, WÉBER 1962, 1989, WEIN 1974, 1977, HORVÁTH et al. 1983, DUDKO 1984, KUBOVICS 1985, HORVÁTH and TARI 1987, EMBEY ISZTIN 1989).

They were explained by the assumption of buried volcanoes (HOFMANN 1871, KOCH 1908) or intrusive masses (HORUSITZKY and WEIN 1962, WEIN 1974, 1977) in the close vicinity of the Buda Hills or below them. The buried volcano was identified and reconstructed by KORPÁS and KOVÁCSVÖLGYI (1997) and was named the Wein-palaeovolcano (**Figs. 84, 85**). The evidence concerning its existence will be outlined in the following:

The Wein-palaeovolcano, 16–17 km in diameter and 1500–1600 m in height is situated in the SE foreland of the carbonate bank and it was the source of the Late Eocene–Early Oligocene volcaniclastics.The hydrothermal alteration and related oreshows of the Szépvölgy Limestone and of the Buda Marl can be related to the hydrothermal processes developed inside the double collapse-calderas of the palaeovolcano. The grade of the hydrothermal alteration and metasomatism increases southward in the Buda Hills, as stated by WEIN (1977) and confirmed by us. This volcanism generated earthquakes, resulting in synsedimentary tectonic phenomena, observed by FODOR et al. (1991, 1994) and by us at many places (e.g. Csillag-hegy, Pál-völgy cave, Budaörs) in the Buda Hills. Besides the above mentioned, the unconformity of about 10° between the conformable and disconformable generations of early marine infillings in the Szépvölgy Limestone (**Figs. 73–81**) could be explained by a regional updoming in the timespan of 38.2–37.7 Ma, resulting in the westward tilting of the carbonate bank. This regional updoming should be related to the implacement of an intrusive body. The tectonic evolution was controlled by WNW–ESE transpression (FODOR et al. 1991, 1994, BENKOVICS and DUDKO in KLEB et al. 1993a, b, KORPÁS et al. 1993) and its most important element is the Ferenc-hegy dextral fault (DUDKO and KORPÁS in KLEB et al. 1993a, KORPÁS et al. 1993) and the syncline–anticline structures related to it (**Figs. 54, 70, 84**).

Timespan of 33–30 Ma (**Fig. 85**): The rapid pelagic carbonate sedimentation became slower at 34,2 Ma and changed in pelitic. This was the datum of the global climatic cooling, resulting in anoxia and in the formation of the Paratethys (BÁLDI 1980, 1983, 1984). From this "moment" until 30 Ma the study area formed part of an epicontinental euxinic basin. The basin was bordered to the NW by the Buda Line, separating the prograding siliciclastic delta front of the Hárshegy Sandstone from the euxinic Tard Clay. The last volcanic event is dated at about 32.5 Ma (BALOGH 1985, NAGYMAROSY et al. 1986, DUNKL and NAGYMAROSY 1992) and its source could be identified with the Wein-palaeovolcano. The high temperature silicification of the Hárshegy Sandstone (BÁLDI and NAGYMAROSY 1976) should be related to this final volcanic event. The stress field was controlled furthermore by WNW–ESE compression, as proved by the active Ferenc-hegy dextral fault and by the syncline–anticline structures of the Tard Clay.

During the Late Oligocene (30–23.4 Ma) the area became part of a prodelta basin (KORPÁS 1981). The deposition of the siliciclastic delta front of the Hárshegy Sandstone was followed by the transgression of the Kiscell Clay, resulting in the retrogradation to the W of the delta system. This process was changed again by a new delta progradation of the Mány and Törökbálint Formations, which is dated at 25.5 Ma, following to HAQ et al. (1987).

The depositional system was gradually uplifted, completely filled, and finally entered the level of subaerial exposure by the end of the Oligocene. The maximum thickness of the sedimentary pile, deposited in the timespan of 39–23.4 Ma should be considered 700–750 m.

In the course of the Early and Middle Miocene (representative timespan of 17–15 Ma, **Fig. 86**) the area and its surroundings became land. This land, just above the sea-

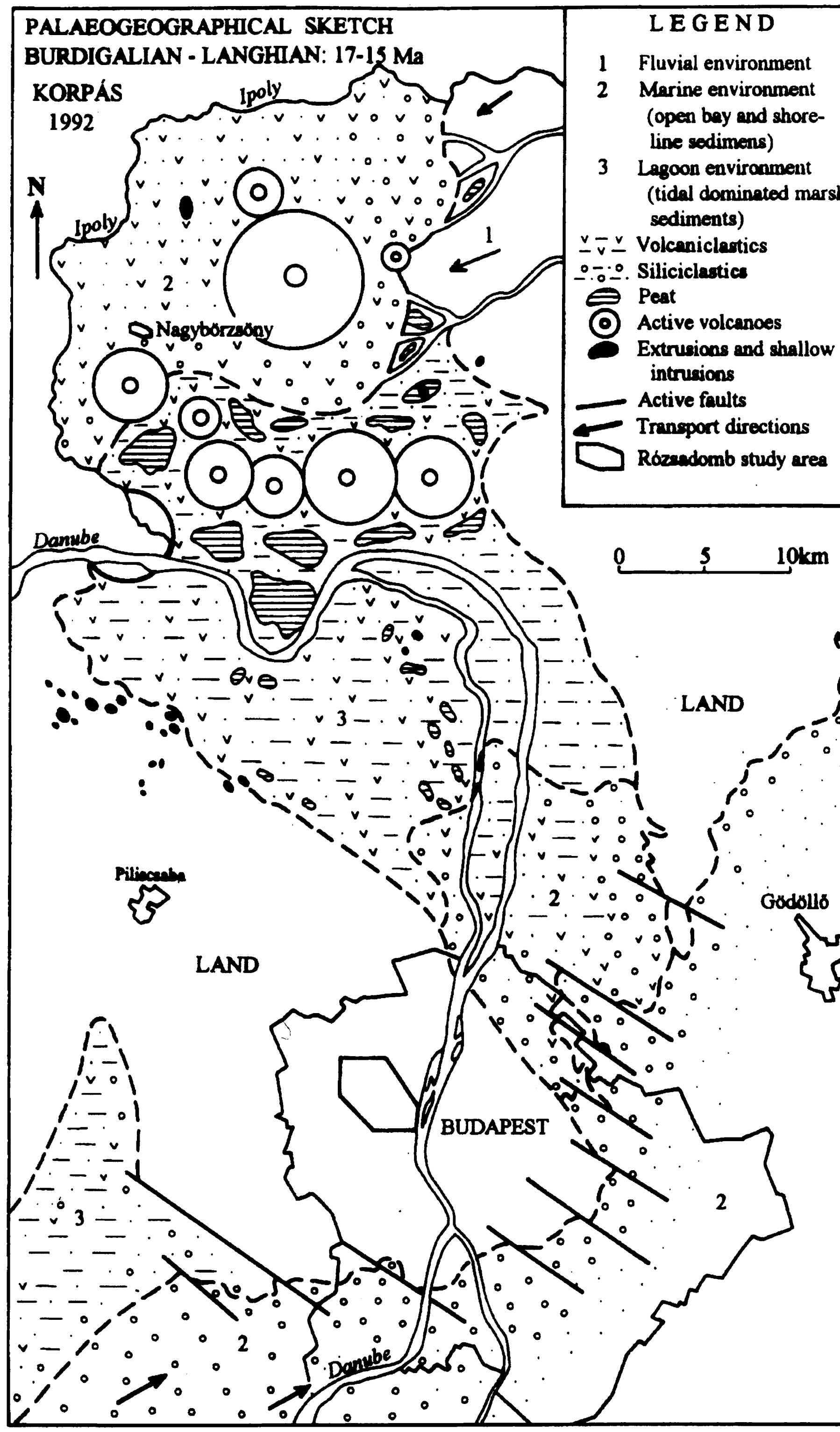


Fig. 86.

level was a flat area, surrounded by tidal dominated marshes and by marine shore-line environments (JÁMBOR 1969). At about 15 Ma a renewed volcanism started in the north, resulting in the formation of volcanoes of the Visegrád and Börzsöny Mountains (BALLA et al. 1979, CSILLAG-TEPLÁNSZKY and KORPÁS 1982, KORPÁS and LANG 1991, 1993).

In the timespan of 15–14 Ma (**Fig. 87**) the areas of land had further increased. The study area was further surrounded by tidal dominated marshes and by open marine shoreline to neritic environments. The entire Buda Hills formed part of a flat bank, near sea-level. The formation of new stratovolcanoes (BALLA 1978, CSILLAG-

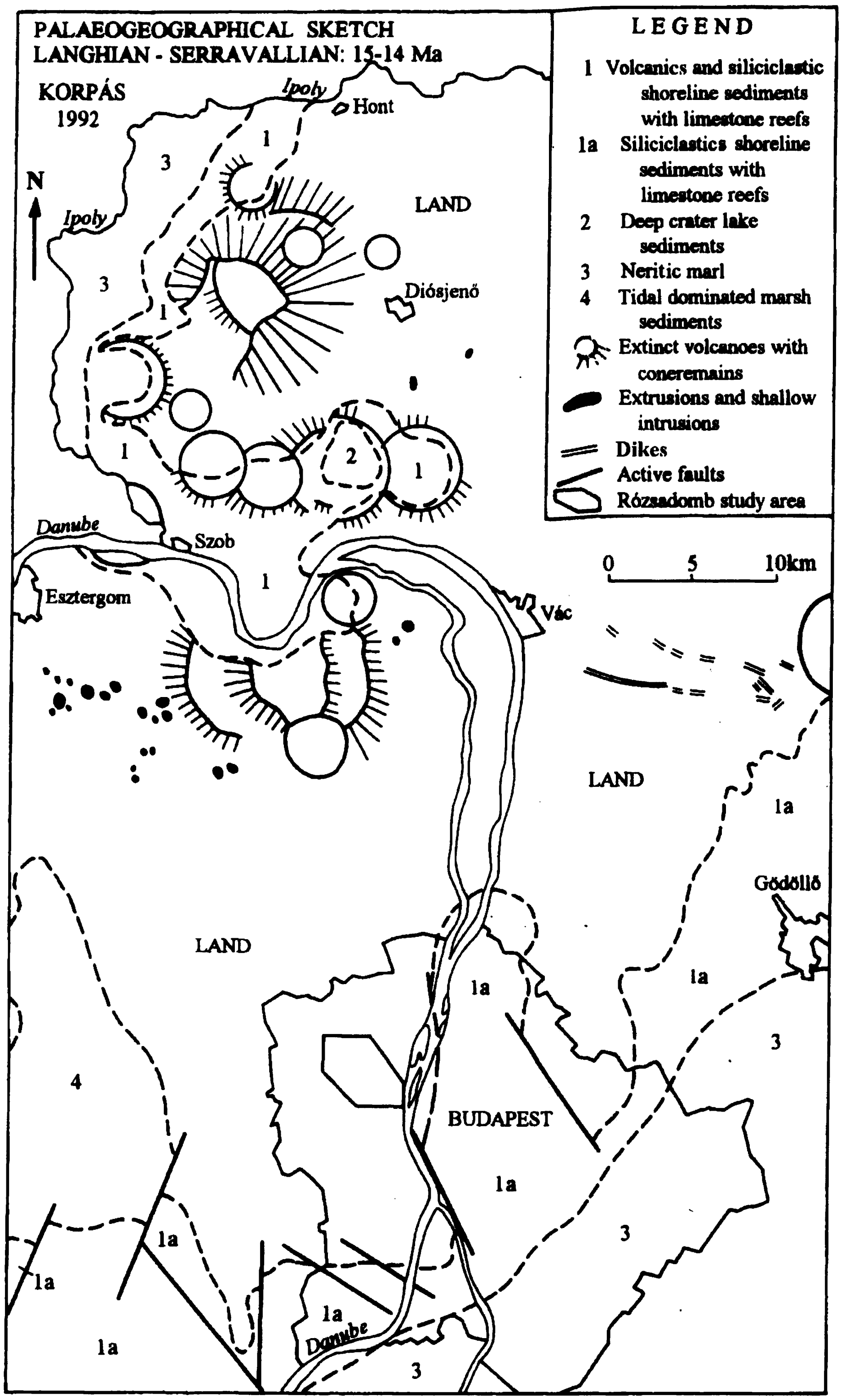


Fig. 87.

TEPLÁNSZKY and KORPÁS 1982, KORPÁS and LANG 1991, 1993) resulted in strong morphological dissection of the Visegrád and Börzsöny Mountains in the north.

Since about 13 Ma (**Fig. 88**) the area of land had increased more, including the volcanic range of the Visegrád–Börzsöny Mountains. This land was surrounded by peritidal, lagoonal and neritic environments. Because the basal strata of the deposited sediments do not contain any local clast, derived from the Triassic and Eocene formations of the Buda Hills, it can be supposed that at that time they still had not entered the level of erosion.

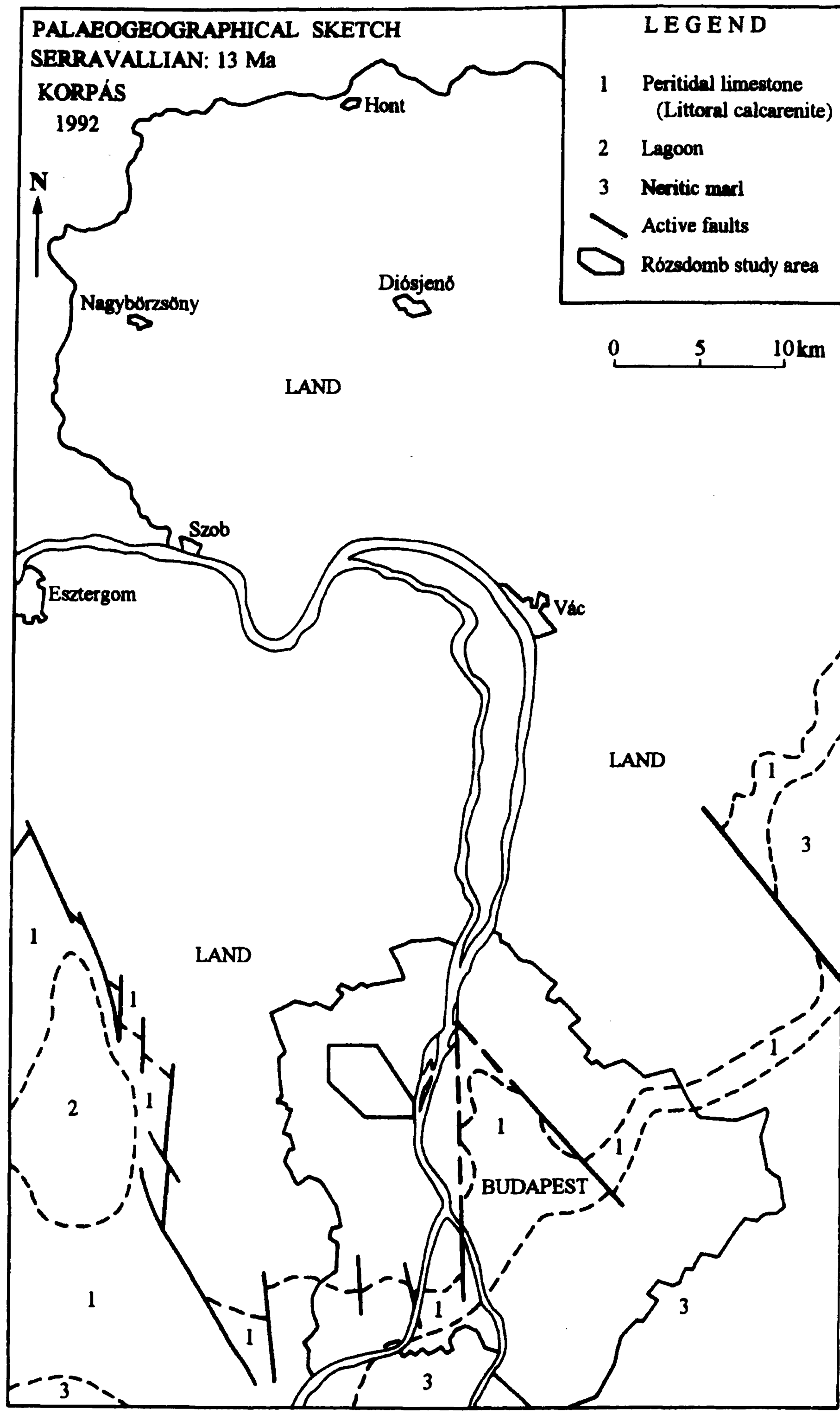


Fig. 88.

The new depositional cycle started about 10 Ma ago (**Fig. 89**), after a 15 Ma long period of nondeposition and erosion. The depositional system was controlled by the evolution of the Late Miocene Pannonian basin. The borders of this basin are represented throughout the Buda Hills and its surroundings by shoreline clastics and the oldest freshwater limestones, indicating the age of maximum flooding during the Late Miocene at about 7 Ma (FÖLDVÁRI 1932, KRETZOI 1980, KORPÁS HÓDI 1995). The study area has formed since part of a siliciclastic delta along the margin of the basin that prograded to the southeast. This depositional environment was succeded at about 3.5 Ma by an extensive lake system of freshwater limestones, fed part-

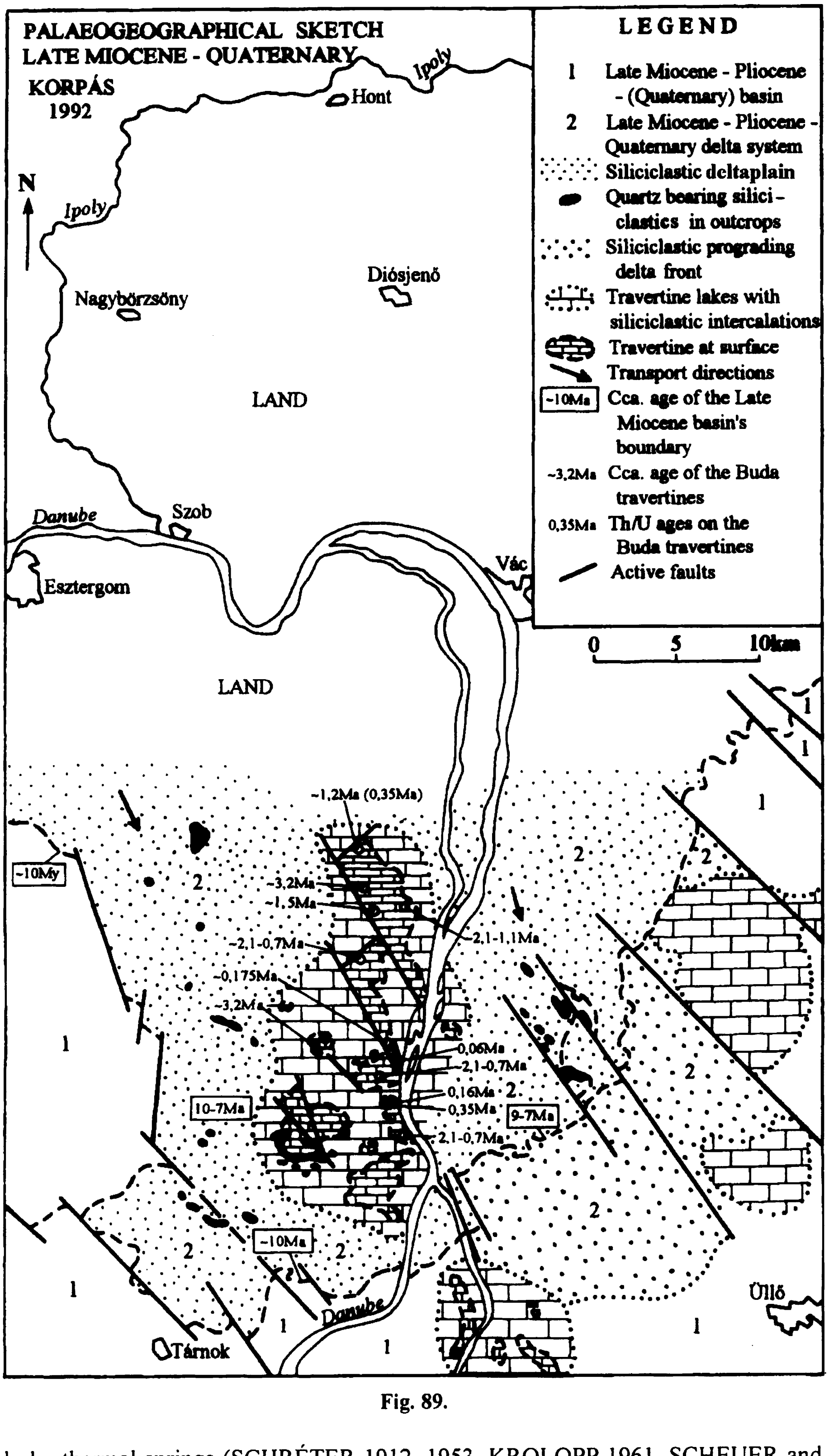


Fig. 89.

ly by thermal springs (SCHRÉTER 1912, 1953, KROLOPP 1961, SCHEUER and SCHWEITZER 1988, KOVÁCS 1995). The gradual ceasing of this delta and freshwater lake system was generated by Late Pliocene to Quaternary tectonic activity, resulting in the uplift of terraces. According to WEIN (1977) this phase is considered the decisive one in the formation of the present day landsurface and morphology. This uplift caused the exhumation of "diapiric horsts" composed of Late Triassic and Palaeogene formations together with their palaeokarst systems. The total amplitude of this uplift was estimated at 360 m by KRETZOI and PÉCSI (1979), PÉCSI et al. (1987) and SCHEUER and SCHWEITZER (1988). Concerning the tectonic evolution it can be stated after FODOR et al. (1991, 1994) and BENKOVICS and DUDKO in KLEB et al. (1993a, b) that the stress field during the Miocene was controlled not only by E–W to SE–NW extension, but by NE–SW compression too. Since 3.5 Ma the components of two stress fields have been present. The NW–SE normal faults in the wider surroundings of the Buda Hills indicate NE–SW extension (**Fig. 89**), while the N–S normal faults of the study area (**Figs. 54, 70**) show E–W extension.

Palaeokarst evolution

Before outlining the palaeokarst evolution it seems useful to recapitulate the most important facts regarding the karst system:

a) The formations of the area are dissected by two interregional composite megaunconformities in the sense of ESTEBAN (1991). The first one between the Late Triassic and Late Eocene is considered a first order megaunconformity, while the second one, below the Pliocene–Quaternary is interpreted as a second order superunconformity. The unconformities inside the Late Eocene depositional record correspond to third order syntectonic ones.

b) The study area has permanently maintained its basin-margin position during the Tertiary. The maximum burial depth at the end of the Eocene was about 250 m, and at the end of the Oligocene it was about 750 m. The mean reflectance values of vitrinites of the Late Eocene–Early Oligocene formations are between 0.30–0.55 Ro% (HÁMOR-VIDÓ in KLEB et al. 1993a, KORPÁS et al. 1993).

c) Concerning the tectonic evolution the WNW–ESE compression in the time-span of 39–30 Ma is proved by FODOR et al. (1991, 1994) and by BENKOVICS and DUKO in KLEB et al. (1993a, b). The Late Eocene synsedimentary tectonic events, i.e. the tilting of the Szépvölgy Limestone carbonate bank (between 38.2–37.7 Ma) and the graben structures in it have been documented by FODOR et al. (1991, 1994) and by KORPÁS et al. (1993). The compression phases between 15–12 Ma and since 3.5 Ma have been recognized.

d) Timing of the reconstructed volcanic events is as follows: Wein-palaeovolcano (KORPÁS and KOVÁCSVÖLGYI 1997): 38–35 Ma and 33–32 Ma; Palaeovolcanoes of the Visegrád and Börzsöny Mountains (BALLA 1978, BALLA et al. 1979, CSILLAG-TEPLÁNSZKY and KORPÁS 1982, KORPÁS and LANG, 1991, 1993): 15–14 Ma.

e) The youngest date for the high temperature hydrothermal event is about 30 Ma (BÁLDI and NAGYMAROSY 1976).

f) The karstification appears in Late Triassic (Mátyáshegy Formation, Maindolomite), in Late Eocene (Szépvölgy Limestone, Buda Marl) and in Pliocene–Quaternary (freshwater limestone) formations.

g) The karstification is clearly controlled by tectonism. The main directions are well defined in the case of the Szépvölgy Limestone and they correspond to the Late Eocene–Early Oligocene stress field.

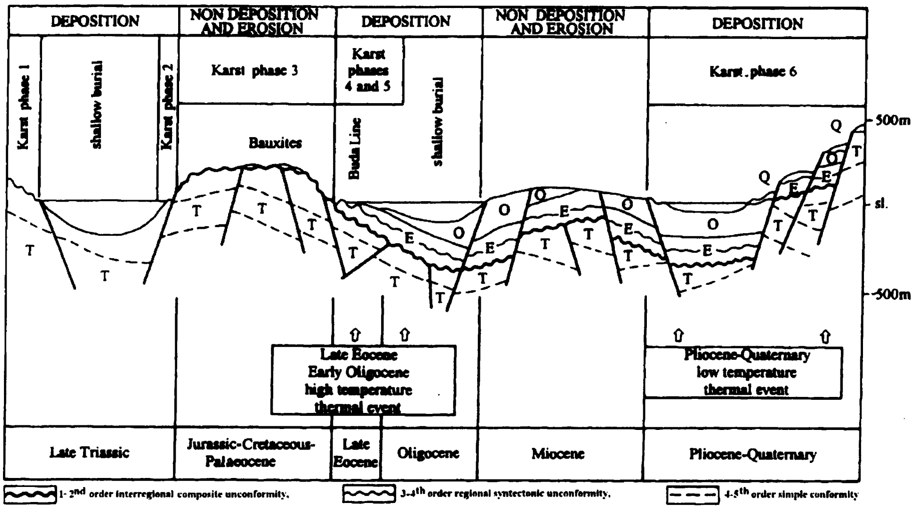
h) The cave system of the Szépvölgy Limestone consists of two, well developed levels and elements of both cold and thermal waters are detectable.

i) The following correlative sediments are known in the karst system: Maindolomite/Cretaceous bauxites (WEIN 1977); Mátyáshegy Formation and Maindolomite/Early and Late Eocene infilling sediments (WEIN 1977, KORPÁS et al. 1993); Szépvölgy Limestone/early infilling sediments (KRAUS 1988, NÁDOR and SÁSDI 1991, NÁDOR 1992a, b, KORPÁS et al. 1993, NÁDOR et al. 1993) and Pliocene–Quaternary siliciclastics, further recent soft bottom sediments.

j) Homogenization temperatures of fluid inclusions of hydrothermal calcite and barite are the following: 110–290 °C (GATTER 1984, GATTER in FODOR et al. 1991); 40–79 °C (DUBLIANSKIY 1991); 110–220 °C (VETŐ-ÁKOS in KLEB et al. 1993a, and in KORPÁS et al. 1993).

k) Minimum U/Th ages of some thermal veins and mineral precipitations in the Pál-völgy cave and Ferenc-hegy cave should be considered >350 Ka to >1.5 Ma (FORD and TAKÁCS BOLNER 1991).

l) The facies of karst system in the Szépvölgy Limestone, based on the cathodoluminescence studies of NÁDOR (1992a), JUHÁSZ and MÁTYÁS (in KLEB et al. 1993) is predominantly phreatic, subordinately vadose. Stable isotope data (FORD and TAKÁCS BOLNER 1991, NÁDOR 1992a, HERTELENDI et al. 1993



T = Triassic, E = Eocene, O = Oligocene, Q = Pliocene–Quaternary

Fig. 90. Model of karst evolution of the Rózsadomb area (KORPÁS et al. 1993)

indicate that hydrothermal processes have played an important role in the evolution of the karst system, while the effects of vadose processes might have been less significant.

The model of karst evolution, illustrated by **Fig. 90** will be discussed in the following:

Karst phases 1 and 2 (Late Triassic): Those early marine microkarsts of the Mátyáshegy Formation and of the Maindolomite of the Apáthy cliff and Balog cliff are related to these phases, which are linked to fourth and fifth order discontinuity surfaces.

Karst phase 3 (Jurassic–Cretaceous–Palaeocene): The products of this karst phase are preserved by the Late Triassic/Late Eocene megaunconformity. The main features of this long-term karst period of about 160 Ma in duration are the following:

— Definite break of sedimentation at the Triassic/Jurassic boundary, followed by a brief depositional event of bauxites during the Cretaceous.

— Long-term and continuous subaerial exposure of the flat, undissected carbonate platform.

— Polycyclic and polygenetic evolution of the continental (alpine) type karst system.

This long-term karst evolution may involve independent karst phases, like the syndepositional ones in the Late Triassic, further karst events related to the Late Triassic–Early Jurassic or to the Late Cretaceous rifting as well as to the coeval magmatism. These possibilities will not be discussed here because of the lack of data.

Karst phase 4 (39–37.7 Ma): This polycyclic karstification is related to the Late Triassic/Late Eocene first order megaunconformity and to the Late Eocene third order syntectonic unconformities. The depositional environment of basin-margin should be characterized by rapid sedimentation along a rocky shore coast with dissected morphology. The correlative, partly marine infilling sediments are present in the karst system and they are accompanied in many places by rocky shore marine bioerosion too. The evolution of the karst system was controlled and governed by a WNW–ESE compression, resulting in the formation of the tectonic network of the caves in the Szépvölgy Limestone. The development of the Szépvölgy Limestone carbonate bank was broken by two mainly submarine third order unconformities, corresponding to lowstand events. The first one is located inside the Szépvölgy Limestone, while the second one coincides with the lower boundary 37.7 Ma in age of the Buda Marl. The two phase evolution of the karst system was interrupted by the tilting to W/10° of the carbonate bank in the timespan of 38.2–37.7 Ma. The development was stopped by subsidence below the sea-level at 37.7 Ma. The karst facies was mainly marine phreatic, with few hydrothermal and vadose effects.

This depositional karst period is considered by us the main one in the evolution of the karst system, resulting in the complete formation of the known caves in the

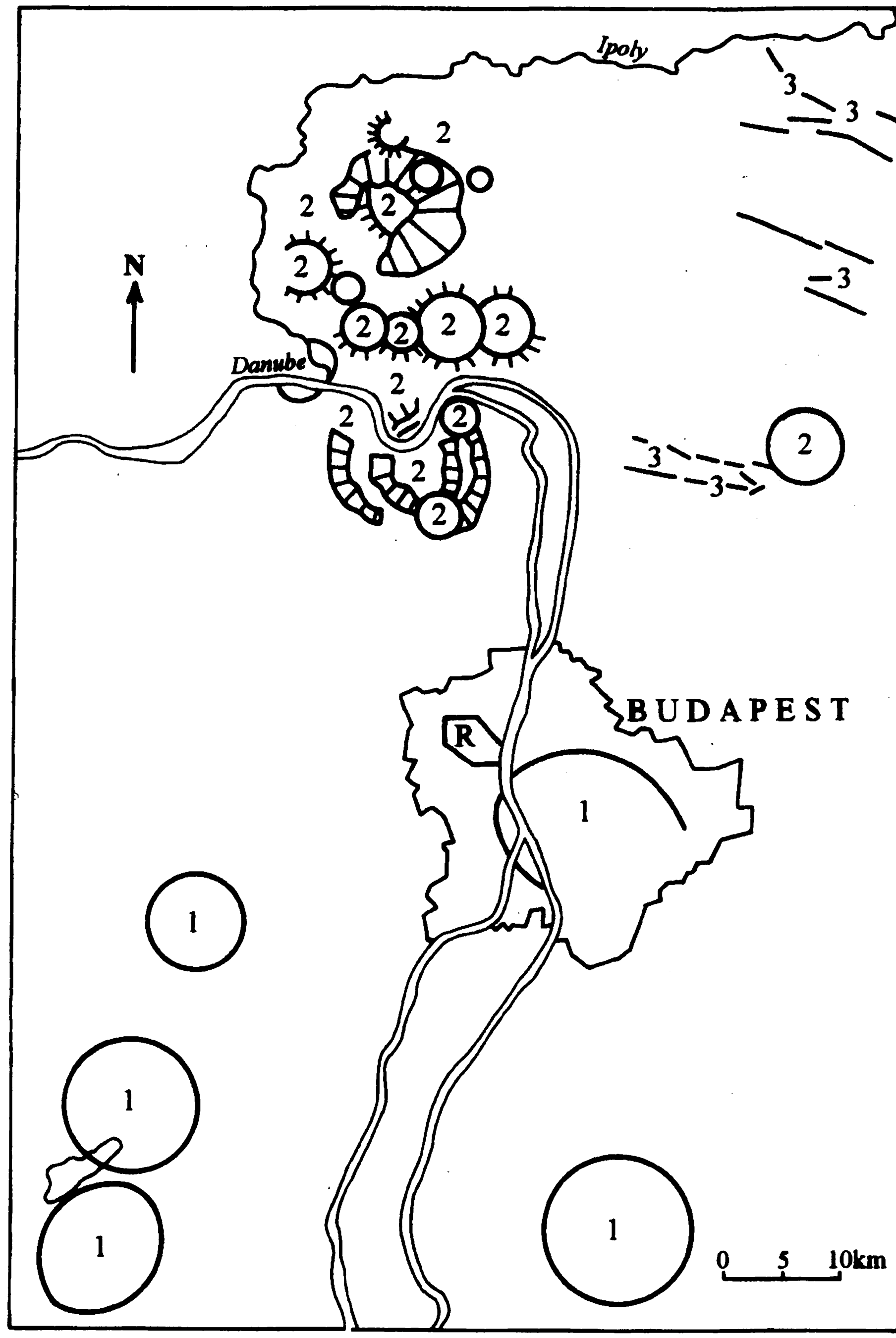


Fig. 91. Palaeovolcanoes in the surroundings of Budapest
1. 30–35 Ma palaeovolcanoes, 2. 14–15 Ma palaeovolcanoes, 3. 14–15 Ma dikes, R=Rózsadomb caves

Rózsadomb. The later phases have not provoked any more significant changes in the karst system, rather they have overprinted it.

Karst phase 5 (35–30 Ma): The timing of this high temperature hydrothermal karst event is based on the following:

a) The hydrothermal heating of the system was produced by the Wein-palaeovolcano, active in the timespans of 38–35 Ma and 33–32 Ma (KORPÁS and KOVÁCSVÖLGYI 1997).

b) Accepting the model of hydrothermal convection currents in a closed system (i.e. in burial depth), explained by MÜLLER (in BOSÁK et al. 1989) the karst system of the Szépvölgy Limestone might have been covered at least by the Buda Marl.

c) The 30 Ma of silicification of the Hárshegy Sandstone (BÁLDI and NAGYMAROSY 1976) was accepted as the youngest age-date of this karst phase.

At this time the study area formed part of an epicontinental, pelagic, partly euxinic basin. The former karst system of the Szépvölgy Limestone was buried by a sedimentary cover of 120–250 m, consequently correlative sediments are not to be expected in it. The WNW–ESE compression further controlled the evolution of the karst system (FODOR et al. 1991, 1994, BENKOVICS and DUDKO in KLEB et al. 1993a, b). The high temperature hydrothermal processes resulted in partial hydrothermal metasomatism of the Szépvölgy Limestone with its infillings, of the Buda Marl and Tard Clay, respectively of the Hárshegy Sandstone. The rich hydrothermal vein-mineralization, a great part of the mineral precipitations, further the formation of the silicified-kaolinitic zones should be related to this event. The karst facies was a hydrothermal deep phreatic one with few vadose effects. The later ones are indicating episodic open stages in the karst evolution, which can be fitted before or after the high temperature event. The source of the hydrothermal heating can be evaluated by the geographical location of the reconstructed palaeovolcanoes in the surroundings of Budapest (**Fig. 91**) and of the fluid-volcano interaction processes (**Fig. 92**).

Karst phase 6 (~3.5 Ma to recent): This partly lower temperature phase is related to a second order superunconformity, formed below the Pliocene–Quaternary cover. The lower datum of this, extent karstification may be calculated at 3.5 Ma, accordingly to SCHEUER and SCHWEITZER (1988). The evolution of the karst system was controlled and governed by its basin-margin position and by its intensive, but periodic uplift. Correlative sediments are represented by siliciclastics of the coeval delta system, by laminites of the freshwater limestones and by low temperature thermal or cold water speleothems.

The majority of researchers (VENDEL and KISHÁZI 1964, MÜLLER 1974, 1989, MÜLLER and SÁRVÁRY 1977, LIEBE and LORBERER 1978, ALFÖLDI 1979, KRAUS 1982, PÉCSI et al. 1987, SCHEUER and SCHWEITZER 1980, 1988, NÁDOR 1992a, b, KLEB et al. 1993a, b) consider this phase as the decisive one during the evolution of karst system. According to the author's opinion (KORPÁS in KLEB et al. 1993a, KORPÁS et al. 1993) this phase had only been a single stage of the polycyclic karst system which had developed mainly in the Late Eocene and was later exhumed in the Pliocene and Quaternary.

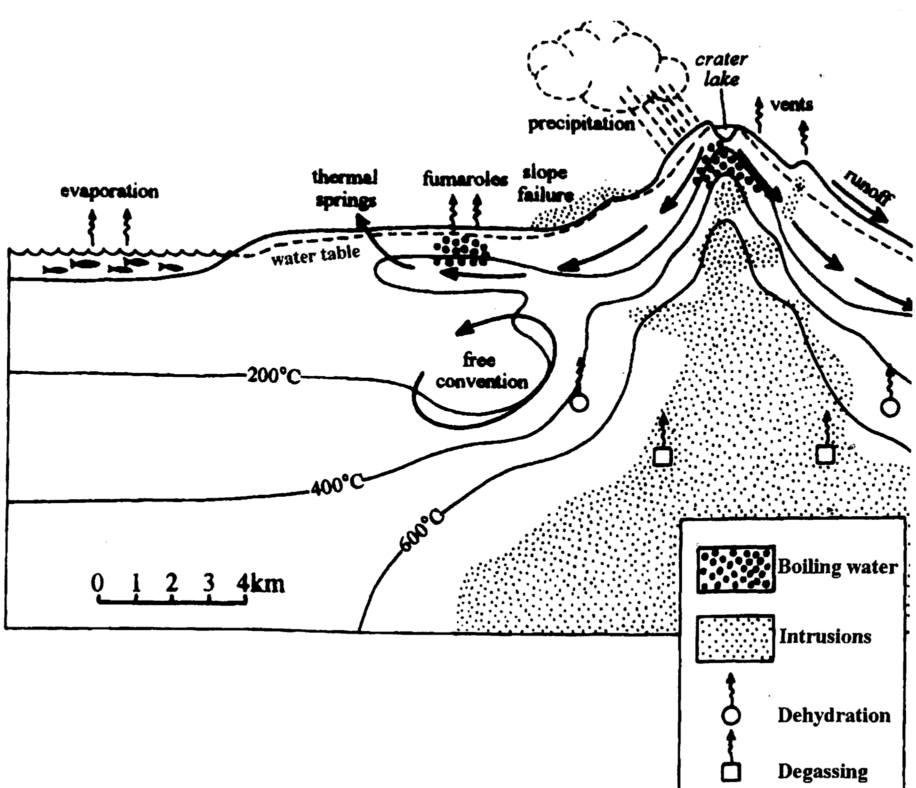


Fig. 92. Fluid-volcano interactions (Penrose Conference Report 1993)

4.2.8. *Buda Hills (Páty)*

Geological setting

The studied key section of palaeokarst in the Late Miocene Tinnye Formation with a maximum thickness of 40 m is exposed by the Mézes-hegy quarry, near Páty (**Fig. 93**). The strata of the limestone-dome (WEIN 1977) consist of well bedded oolitic bioclastic grainstones, alternating with thin lenses of intraclastic calcareous sandstones. The latter are concentrated in the upper part of the profile, and their intraclasts are composed of Triassic dolomites, Eocene limestones, Oligocene sandstones and a few quartzites. The bioclasts are represented mainly by foraminifera, less by fragments and moulds of molluscs. The typical microfacies is considered after LELKES oolitic grainstones, wackestones showing horizons of vadose pisolites and of caliche. The shallowing upward sequence is dissected by four discontinuity surfaces, parallel to the bedding (**Figs. 93–96**). The depositional system is interpreted as subtidal shoals within a flat lagoon.

Palaeokarst features and interpretation

The following criteria have been observed for the recognition of palaeokarst: shallowing upward sequence dissected by four discontinuity surfaces; intergranular and poorly developed vuggy porosity with minor cavities, mainly parallel, partly perpendicular to the bedding; slight vadose solutional phenomena on the walls of the cavities; presence of caliche; early open fissures, controlled by synsedimentary faults and partly filled by bioclastic calcareous sandstones; horizons of marine bioerosion.

The evolution of the flat lagoon was controlled and broken by four drops of sea-level resulting in its gradual and definite uplift. These sea-level falls were represented by the correspondent 4th order single discontinuity surfaces, causing brief subaerial exposure with the formation of caliche. The intergranular and slightly developed vuggy porosity, further the small cavities should be related to these short-term palaeokarstic events.

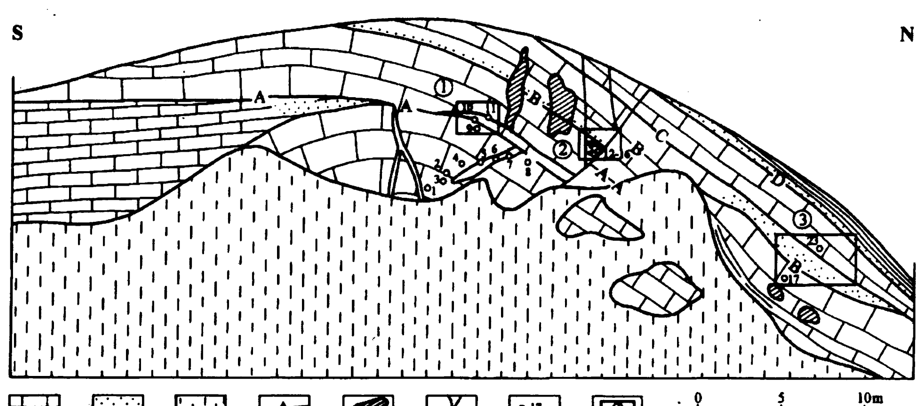


Fig. 93. Palaeokarst profile of the Late Miocene Tinnye Formation Páty, Mézes-hegy quarry

1. Oolitic limestone, 2. Calcareous sandstones with extraclasts, 3. Debris, 4. Discontinuity surface, 5. Cavities, 6. Faults, 7. Site of sampling, 8. Detailed profile No. 1

4.2.9. *Buda Hills (Várhegy)*

The palaeokarst of the freshwater limestone sequence, about 12 m in thickness of the Várhegy will be discussed after proper observations, using data of KROLOPP (1961), KROLOPP et al. (1976), SCHEUER and SCHWEITZER (1988), KLEB et al. (1993c) and KOVÁCS (1995).

Geological setting and stratigraphy

The stratigraphy and the main lithological units of the freshwater limestone are illustrated on **Fig. 97**.

Unit 1: The basal clastic strata of ≤3 m in thickness consist of fining upward alluvial gravels and sands. The matrix supported, slightly cemented gravels are derived from angular to subangular local clasts of Late Triassic dolomites, limestones and cherts, of Late Eocene Szépvölgy Limestone and Buda Marl, further of Early Oligocene Hárshegy Sandstone and of a few quartzites. They are overlain by lenses of friable, limonitic, badly sorted, medium to coarse grained sands different from

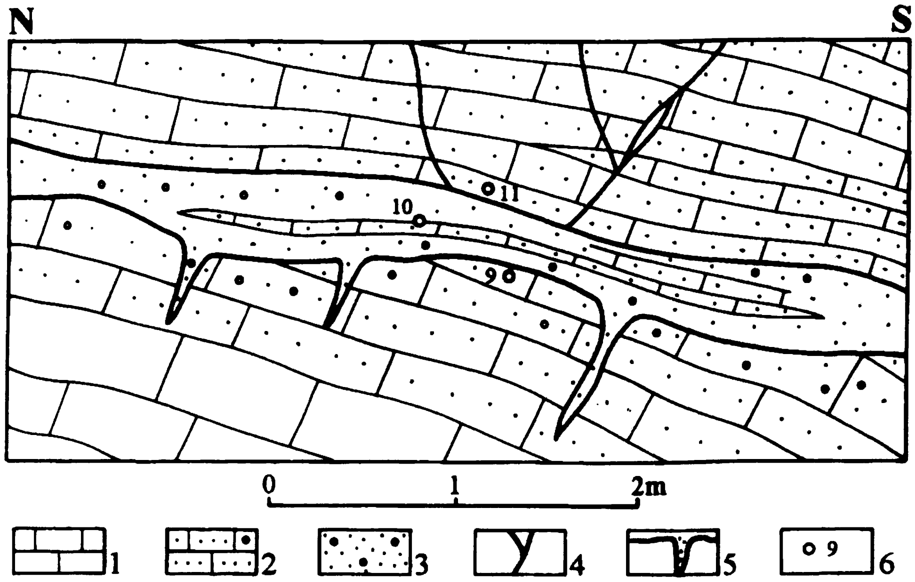


Fig. 94. Detailed palaeokarst profile No. 1 of the Mézes-hegy quarry
1. Subtidal oolitic limestone, 2. Calcareous sandstone with extraclasts, 3. Calcareous sandstone, 4. Synsedimentary normal faults, 5. Open fissures with infilling, 6. Site of sampling

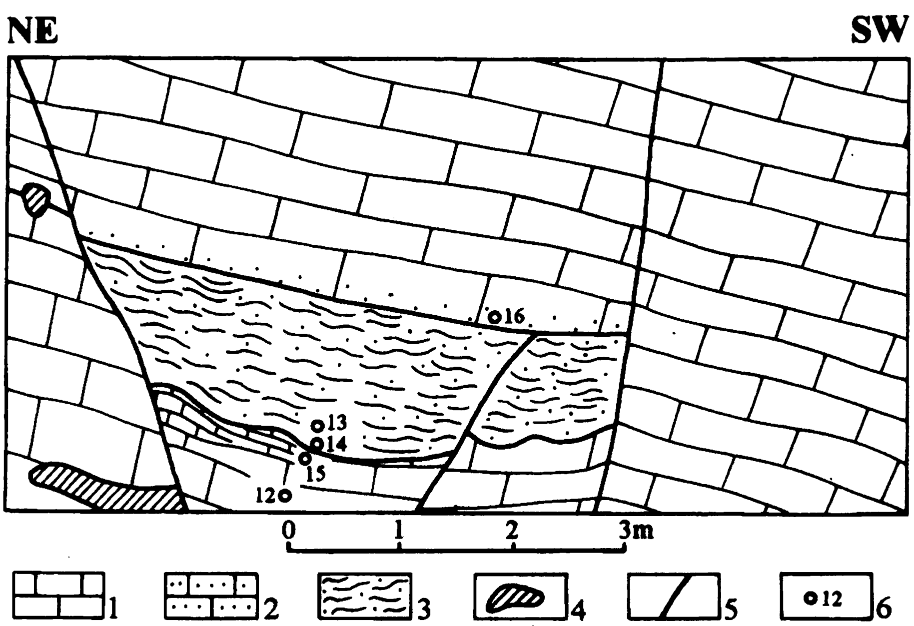


Fig. 95. Detailed palaeokarst profile No. 2 of the Mézes-hegy quarry
1. Oolitic limestone, 2. Altered mudy limestone with moulds of molluscs, 3. Calcareous sandstone, 4. Cavities, 5. Normal synsedimentary faults, 6. Site of sampling

those of the Danube in their mineralogical composition (MOLNÁR in KROLOPP et al. 1976, SZARKA in KLEB et al. 1993c). The transition to the overlying freshwater limestone (Unit 2) is represented by laminated silts and sandy clays.

Unit 2; The basal clastics are covered by laminated muddy freshwater limestone, 2 m in thickness (**Photos 12–14**). This consists of alterning laminae of soft or altered muddy algal limestones, rich in dispersed organic matter. This laminated unit is dissected by synsedimentary normal microfaults and capped by a subaerial unconformity surface reflecting a significant depsitional break and internal erosion (**Photos 13, 14**). It is the richest in fossils and consists of both terrestrial and freshwater molluscs, vertebrata, arthropods, further a great amount of algae, reed-grasses, charophytes, bryophytes, prints and detritus of plants. The dominant microfacies is considered (pel)microsparites, (pel)micrites with less stromatolite-like types of travertine and phytoclastic calcarenites (KOVÁCS 1995). The depositional system should be interpreted an open pool, fed by thermal springs. SCHEUER (in KROLOPP et al. 1976) and SCHEUER and SCHWEITZER (1988) have given the tetarata model for

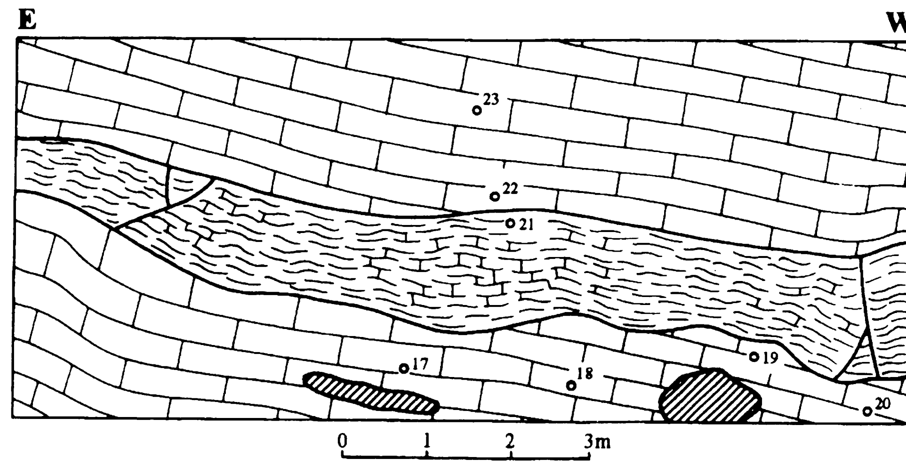


Fig. 96. Detailed palaeokarst profile No. 3 of the Mézes-hegy quarry
1. Calcareous sandstone, 2. Altered muddy limestone with moulds of molluscs, 3. Cavities, 4. Normal synsedimentary faults, 5. Site of sampling

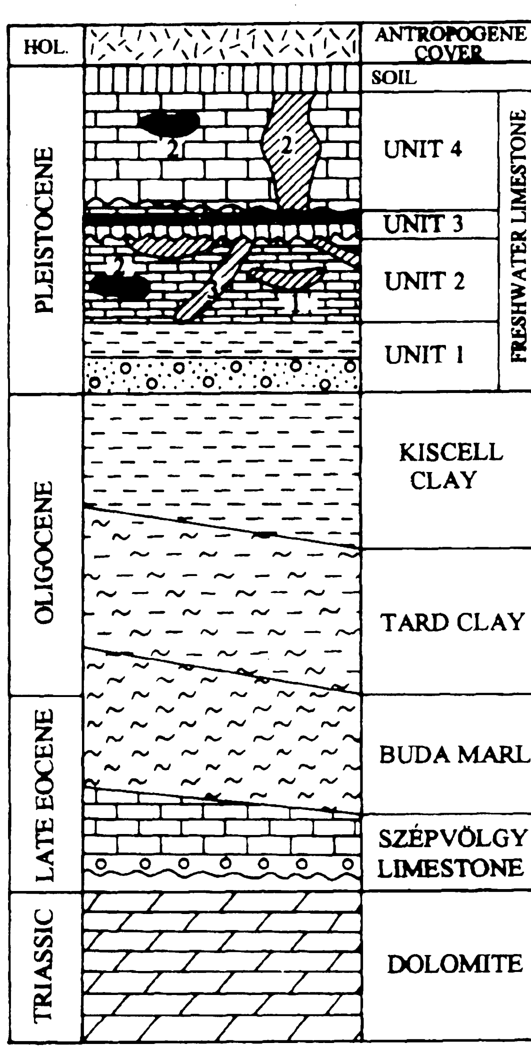


Fig. 97. Stratigraphic chart of the Várhegy (after KROLOPP et al. 1976
1. Laminitic infillings, 2. Cave infillings, 3. Fissure infillings

its formation. Ecological evaluation of fauna and flora (KROLOPP 1961, KROLOPP et al. 1976) suggests a water temperature between 20–30 °C and a gradual climatic cooling during the depositional record.

Unit 3: This is represented by a palaeosol horizon of 15–50 cm in thickness, which covers the subaerial unconformity surface and penetrates the early open joints of laminated Unit 2 (**Photos 13, 14**). Its basal layers consist of smectite-type soft clays, bearing angular clasts derived from the footwall-laminites. The main level of the massive, friable, clastic and grainy palaeosol consists of horizon "A", dark-reddish in colour and the carbonate rich horizon "B" brown in colour. Both are rich in fossils, consisting of mainly terrestrial gastropods and vertebrates (KROLOPP et al. 1976). The latter are represented frequently by redeposited and well rounded bone fragments (JÁNOSSY in KROLOPP et al. 1976). This palaeosol horizon is covered partly by thin layers of carbonate-laminites similar to the earlier ones or directly by massive crystalline freshwater limestone of Unit 4. The palaeosol unit indicates a significant climatic change, i.e. cooling of the environment (KROLOPP et al. 1976) and reflects a break and related subaerial exposure during the depositional record.

Unit 4: This unit is composed of massive, crystalline, cavernous freshwater limestone with a thickness of 7–8 m. The poor fossil ensemble consists of moulds and recrystallised shells of terrestrial and freshwater gastropods, some fragments of plants, mainly reed-grasses. The depositional environment is interpreted as a very shallow flat pool, fed furthermore by thermal springs. Maximum water temperatures were about 35 °C and the composition of the species of terrestrial molluscs indicates a new climatic change, i.e. a more humid warming up period (KROLOPP et al. 1976).

The age dating of Units 1–4 is based on the following: Vertebrata fauna — Middle Pleistocene, Biharian (Phases Tarkő and Vértesszőlős, JÁNOSSY in KROLOPP et al. 1976, SCHEUER and SCHWEITZER 1988); Th/U ages: 358 Ka, 160 Ka (SCHEUER and SCHWEITZER 1988); Magnetostratigraphy: normal polarity zone of Brunhes SCHEUER and SCHWEITZER 1988).

Palaeokarst features

The palaeokarst phenomena related to the freshwater limestone will be outlined in the following: extensive single level cave system, developed mainly in Unit 4 parallel to bedding and has a total length of passages of about 10 km; microporosity and presence of vugs and minor cavities both in Unit 2 and 4; presence of a third order subaerial unconformity surface and related palaeosol horizon between Unit 2 and 4; synsedimentary microfaults and fissures with early infilling of Unit 2 in some places (**Photos 12–14**); dm wide joint system in Unit 2, below the unconformity and infilled

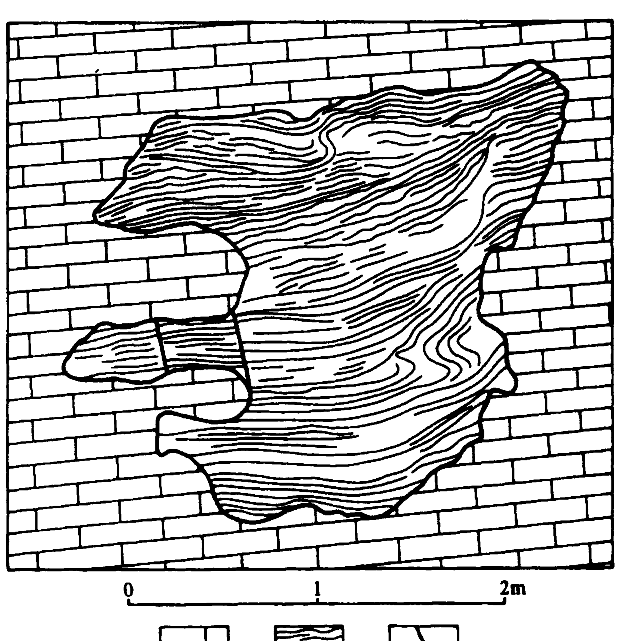


Fig. 98. Early generation of laminitic infillings in Unit 2, Várhegy, Fortuna u. 28
1. Freshwater limestone of Unit 2, 2. Laminated infilling of carbonate-mud, 3. Synsedimentary normal microfaults

by clastic palaeosols; two main generations of infilling sediments, composed of early laminites (**Fig. 98**) similar to Unit 2 and of subsequent palaeosols, rich in terrestrial fossils in Units 2 and 4 (**Photo 14**); vadose speleothems in the caves of Unit 4.

Model of karst evolution

The karst system is considered a depositional one with low temperature (20–35°C) thermal water circulation. Lack of sufficient data has not permitted the absolute timing of the karst events, therefore the 3 phase evolution of the karst system will be discussed in order of relative succession.

Karst phase 1: this depositional karst phase is related to the gradual uplift and subaerial exposure of the thermal pool of Units 1 and 2. Early diagenetic thermal convection processes and synchronous tectonic activity produced microporosity and minor fissures, and cavities, infilled by the first generations of laminites. This phase was completed by a drop of the watertable, resulting in subaerial exposure.

Karst phase 2: represents a short subaerial event of rapid continental karstification accompanied by the formation of palesol horizon (Unit 3). The subaerial karstification has penetrated the laminites of Unit 3, resulting in the formation of fissure infillings of second generation in it. This karst phase should be considered a quiet tectonic period.

Karst phase 3: the subsequent rise of the watertable started a new depositional cycle, resulting in the regeneration of the thermal, but even shallower pool of Unit 4. Sediments of this cycle covered and preserved products of karst phases 1 and 2. Common effects of renewed tectonic activity and intense solutional processes produced an extensive cave system, infilled partly by the second generation of continental infillings, i.e. of palaeosols. The gradual uplift and drop of the watertable resulted in complete and definite subaerial exposure of the whole karst system overprinting it by subsequent precipitation of younger speleothems.

Using age estimation, explained earlier, the rapid evolution of this 3 phase karst system may have been completed at about 1–1.2 Ma.

4.2.10 *Bükk Mountains*

Reconnaissance palaeokarst trips were done with GYÖRGY LESS through the Bükk Mountains in 1995. The first results of observations made in Lillafüred, Miskolc-Tapolca and Felsőtárkány will be outlined in the following. The presence of early marine infilling sediments in the karst system is considered as their main common feature.

Lillafüred

The palaeokarst bearing outcrop (**Photos 15, 16**) is located at the base of the western wall of the motor road, between the entrance of the István cave and the guide post Miskolc–Lillafüred. Overturned strata of the karstified Fehérkő Limestone, Middle Triassic in age represent the wall rock. The well bedded, peritidal, loferitic limestone is cut by a 2 m long narrow V-like fissure, infilled by the unconformable generation of early marine micritic carbonate-laminites. The upper boundary of this early, depositional karst infilling coincides with a discontinuity surface inside the Fehérkő Limestone and parallel to the bedding. The described section does not show any signs of subaerial exposure. The overturned position of this palaeokarst can be confirmed both by the internal depositional features (graded bedding, stratification) of laminites and by the reverse widening of the fissure.

The suggested model of karst is considered a depositional and submarine one. The first phase may be characterized by deposition of peritidal carbonates, followed by early fracturing and dissolution of the Fehérkő Limestone platform segment. The next event was tilting of the platform segment and the subsequent infilling of its fissures by carbonate-laminites under shallow submarine conditions. Finally interpreted as a very late event the regional overturning of the carbonate platform occurred together with its palaeokarst infillings. Stages 1 and 2 of karst evolution are considered Middle Triassic depositional karst events, while the overturning was a very late regional tectonic event of at least Early Cretaceous.

Miskolc-Tapolca

The palaeokarst section (**Photos 17, 18**) was taken at the base of the western wall (level 2) of the abandoned quarry near Miskolc-Tapolca. The karstified wall rock consists of mainly cyclic lagoonal loferites of the platform unit Berva Limestone, Middle–Late Triassic in age. The examined karstic infillings are represented by conformable and unconformable generations of early marine micritic carbonate-laminites, further by clast supported autoclastic collapse breccias and by palisade calcite on the cavity walls. The presented profile does not show phenomena related to subaerial exposure. The estimated relative order of succession of these infilling-types, based on cross-cut relations is the following:

1. Conformable generation of early marine laminites.
2. Unconformable generation of early marine laminites.
3. Clast supported autoclastic collapse breccia along the fissure and cutting both generations of early laminites.
4. Palisade calcite infilling partly the free places between the walls of the fissure and of the collapse breccia.

The karst system is considered a depositional one. The formation of the conformable laminites was followed by block tilting of the Berva Limestone. This tilting, similar to the Late Eocene one of the Szépvölgy Limestone (see the case study of Buda Hills/Rózsadomb in the chapter 4.2.7.) is proved by the subsequent unconformable laminites. The early marine laminites were cut by the opening of the fissure, infilled partly by collapse breccias and by subsequent palisade calcite. The two generations of laminites can be related to submarine depositional karst processes of Middle Triassic age.

Felsőtárkány

The palaeokarst profile (**Photos 19, 20**) was taken along the motor road in the Felsőtárkány-völgy, 200 m from its entrance, and at the base of the northern wall, near the water shaft. The karstified wall rock consists of well bedded-laminated, micritic and bioclastic Felsőtárkány Limestone of Late Triassic age. The elongated cavities, parallel to the bedding are partly infilled by conformable generation of early marine micritic carbonate-laminites. They consist of angular clasts of red radiolarites, identical with the Early to Middle Jurassic Bányahegy Radiolarites. The palaeokarst section does not show any signs of subaerial exposure. The karstification can be interpreted as a submarine depositional and Late Triassic–Early Jurassic in age.

5. The 3D model of the composite karst system, Buda Hills

5.1. Aims and methodology

The composite karst system of the Buda Hills forms part of our natural heritage due to its extraordinary beauty and its worldwide fame. This fame can be attributed both to its particular natural values and its historical culture of thermal spas. As these are to find in the zones of infiltration and mixing of the open karst system, damage caused both by climatic factors and human activity has increased during the last decades. This hazard consists of gradual and measurable pollution of the karst system (IZÁPY and SÁRVÁRY 1993) indicated frequently by sudden pollution-events, registered in the diaries of the thermal spas too. However that front of pollution is less known and documented. Downward migrating pollution has to be considered the main danger, because penetrating the mixing and phreatic zones should result in pollution of the whole karst system. From this moment the process of pollution becomes irreversible in human-life scale. How can it be predicted? The theoretical answer is very simple: by more precise determination of the geometry and distribution of infiltration and conduit zones. That was the turning-point which permitted us to see the new picture, i.e. the 3D model of the composite karst system of the Buda Hills. This 3D model surpasses the earlier traditional ones by its quality and precision and opens new ways in the process of interpretation and modelling.

The most important genetic considerations, listed in the chapter 3. Palaeokarst and carbonate platforms will be recapitulated here as a methodological base of karstgenesis:

1. The main conduit zones are the cave horizons parallel to bedding.
2. These cave horizons are practically coeval with their bedrock and they formed in well defined sections of the carbonate platform, reflecting clear regularities in their distribution.
3. The cave horizons can be detected and delineated by simple geological and geophysical methods.

Consequently the problem to solve lies in delineation of these cave horizons inside the carbonate masses of the Buda Hills. This was carried out by applying the following methods and steps:

1. Analysis of stratigraphy, evolution of the Buda Hills and its cave-register (compiled by TAKÁCS BOLNER), which resulted in the preparation of the synthetic stratigraphic chart showing the main palaeokarstic horizons (**Fig. 99**).
2. Elaboration of the regional tectonic model of the pre-Tertiary basement of the Buda Hills and its surroundings, indicating the expected location of the main palaeokarstic horizons (**Fig. 100**).
3. Construction of the geological map (**Appendix 1**) and of the pre-Tertiary basement map (**Appendix 2**) of the Buda Hills.
4. Elaboration of the 3D model of the composite karst system of the Buda Hills (**Appendix 3**).

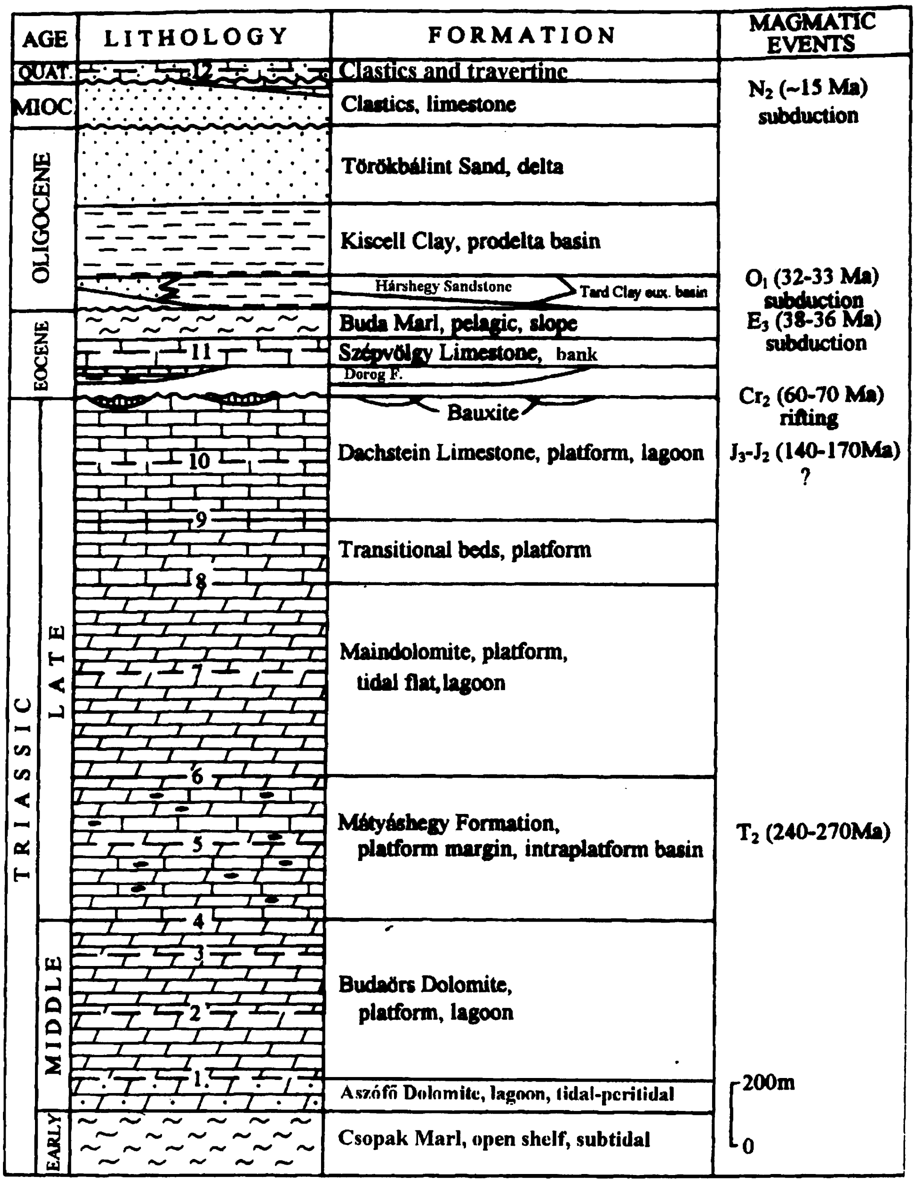


Fig. 99. Synthetic stratigraphic chart of the Buda Hills, showing the main palaeokarst horizons (1–12)

The 3D model and its importance 5.2.

Based on these results the estimated number of main palaeokarstic (i.e. cave) horizons is 12 and their stratigraphic position is the following:

Quaternary–Pliocene:	Freshwater limestone	1
Late Eocene:	Szépvölgy Limestone	1 (2 levels of caves)
Late Triassic:	Dachstein Limestone and transitional beds	2 (3–5 levels of caves)
	Dachstein Limestone/Maindolomite	1
	Maindolomite	1
	Maindolomite/Mátyáshegy Formation	1
	Mátyáshegy Formation	1 (1 level of caves)
Middle Triassic:	Budaörs Dolomite	3

Among the factors controlling the waterbalance and convection in the karst system the palaeokarstic, tectonic and morphologic ones, which will be outlined. The palaeokarstic factors (**Fig. 99, 100 and 101, Appendix 3**) give about 7–10 v% of the total porosity in the system and they are represented by the 12 cave horizons, forming a "layer-cake structure" inside the carbonate mass. Beside them a significant role should be attributed to the unconformity surfaces: first of all to the surface of the

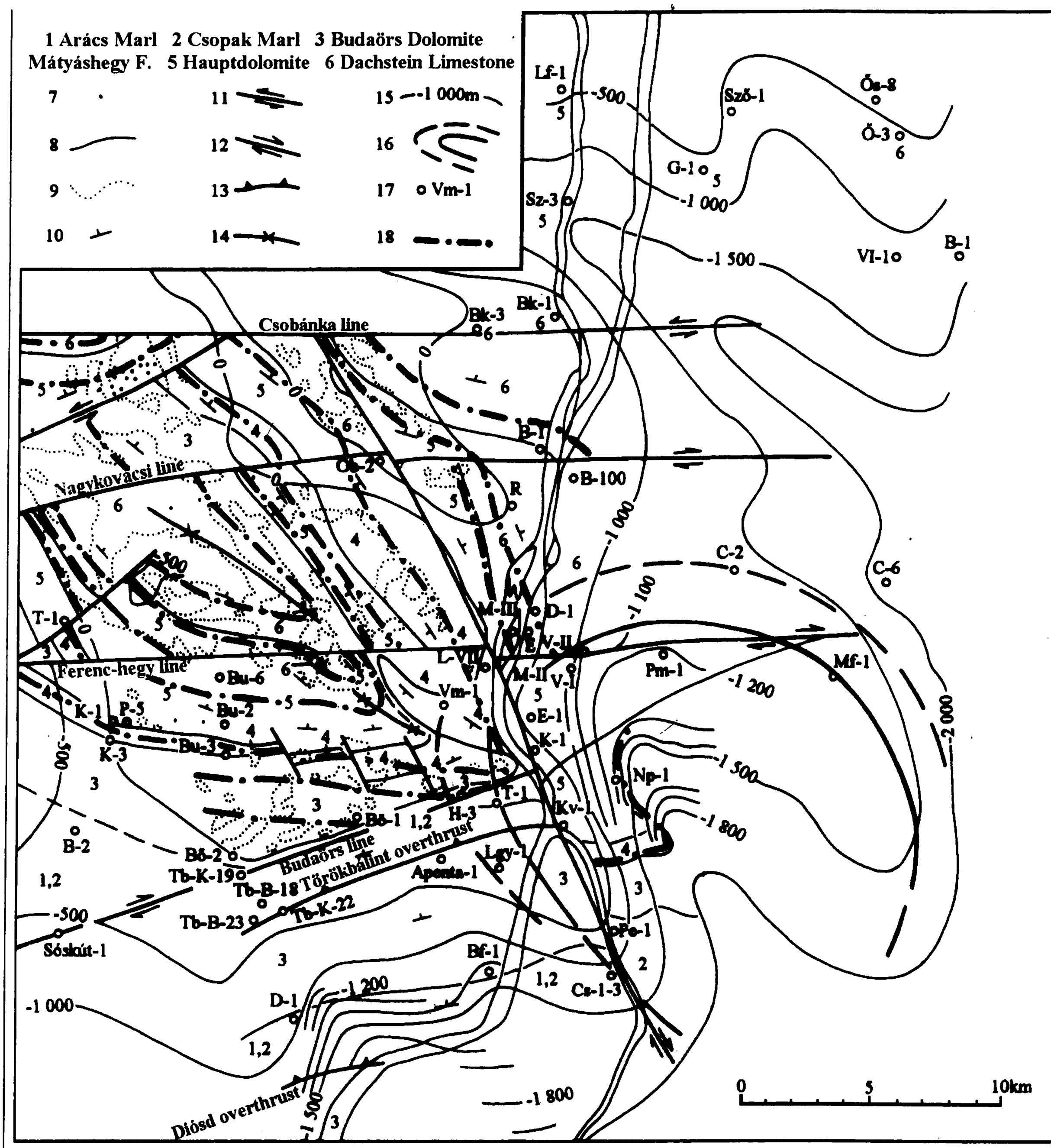


Fig. 100. Tectonic scheme of the pre-Tertiary basement and of the main palaeokarst horizons in the Buda Hills and its surroundings

Early and Middle Triassic: 1. Arács Marl, 2. Aszófő Dolomite, Middle Triassic: 3. Budaörs Dolomite, Late Triassic: 4. Mátyáshegy Formation, 5. Maindolomite, 6. Dachstein Limestone, 7. Bauxite, 8. Boundary of formation, 9. Surface outcrops, 10. Regional dip, 11. Sinistral fault, 12. Dextral fault, 13. Overthrust, 14. Axis of syncline, 15. Contourlines of the basement, 16. Wein palaeovolcano, 17. Important boreholes, 18. Distribution of the main palaeokarst horizons

composite megaunconformity between the Triassic and the Palaeogene, further to the Triassic ones, which are located at the boundary of the Triassic formations. The tectonic factors are represented by the dextral W–E fault system Late Eocene to Early Miocene in age (FODOR et al. 1991a, 1994) which gives the present day structure (**Appendix 1**). Less importance should be attributed to the NW–SE fault system formed during the Cretaceous. The role of these tectonic elements manifests more in the displacement of different geological units close to each other and probably less in the conductivity of the open joint system. The role of the morphological factors is reflected best of all on the map of the pre-Tertiary basement (**Appendix 2**), separating clearly the following main morphotectonic units from each other: the assymmetric central syncline of NW–SE trend (WEIN 1977), bordered at the NE by an anti-

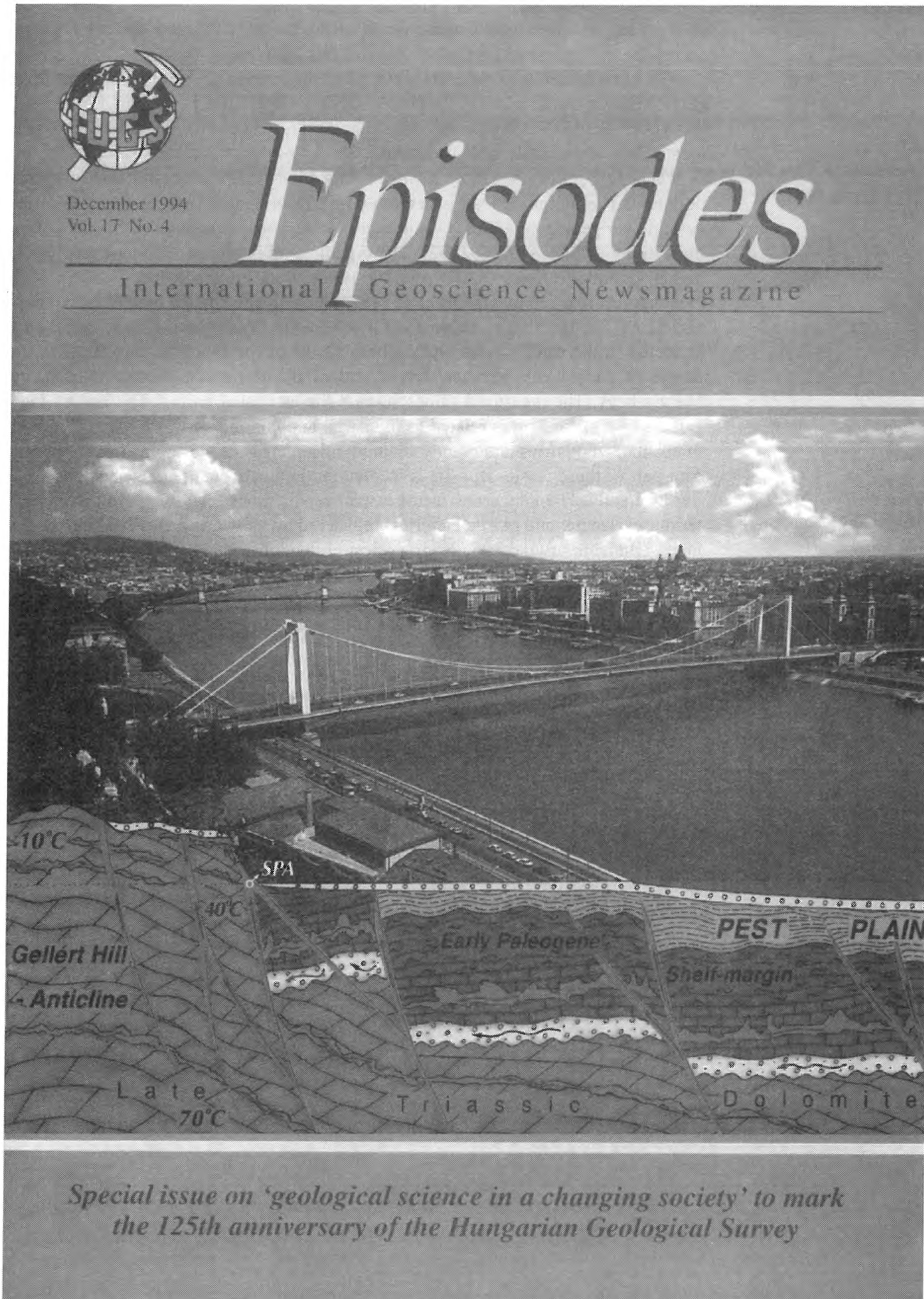


Fig. 101. Thermal karst convection system of the Buda Hills (KORPÁS 1994d)

cline, rising as a watershed in the range of the Gellért-hegy and Csúcs-hegy, and finally the monocline along the NE flank of this anticline.

The karst system is charged by water up to the levels 105–130 m above sea level (LORBERER and IZÁPY WEHOVSZKY 1992, JOCHA-EDELÉNYI and GONDÁR-SÖREGI 1994), suggesting that the greatest part of the cave horizons is located below the water table (**Appendix 3**).

One of the main sources of damage to the karst system is organic and inorganic pollution. The inorganic load is derived partly from natural sources, existing in the system over geological times (ÓDOR et al. 1994), partly provoked by human polluting activity (VERRASZTÓ 1993). The organic load is related without exception to human polluting effects (VERRASZTÓ 1993). Data on geochemical stream sediment survey, realized by ÓDOR et al. (1994) have documented a load of heavy metals like As, Cd, Pb, Sb, Zn. This load of heavy metals (hydrothermal in origin) is concentrated in the central areas and on both flanks of the syncline, as well as in the ranges of János-hegy–Széchenyi-hegy and of the Hármashatár-hegy, Vihar-hegy, Mátyás-hegy. The measured values surpass the estimated norms. The human inorganic and organic load is related to industrial and communal waste, deposited both in abandoned quarries and mainly in unurbanised areas of the Buda Hills. A significant part of these waste-deposits is located on the uncovered karst system, contacting it directly. The dangerous factories (car repair shops, battery recyclers, slaughterhouses, leather and plastic factories) registered by VERRASZTÓ (1993) in the villages (Budaörs, Budakeszi, Nagykovácsi, Pilisvörösvár, Pilisszentiván, Solymár, Pilisborosjenő, Üröm, Budakalász) produce organic and inorganic waste in a range of some hundred tonnes and cubic metres per year. Our knowledge is very poor on the present stage of this pollution front, migrating slowly downwards.

Our genetic studies have documented the common features of the evolution of carbonate platforms and related palaeokarst systems. The formation of palaeokarst systems is controlled by phases of platform evolution. Two main factors, climatic and tectonic govern this evolution. Favourable tropical–subtropical climate combined both with exposures along passive continental margins (disintegrated by rifting) and/or along active continental margins (accreted by collision) should result in the formation of palaeokarst systems. Taking into consideration the geological composition and evolution of Hungary (FÜLÖP 1990, 1994, HAAS 1994) the following main phases of platform evolution have been differentiated:

Palaeozoic: Devonian–Carboniferous — extensive carbonate platforms, related mainly to rifting.

Mesozoic: Triassic–Early Jurassic — extensive carbonate platforms, related mainly to rifting.

Late Jurassic–Early Cretaceous — reduced carbonate platforms, related mainly to collision.

Middle Cretaceous — reduced carbonate platforms, related mainly to collision.

Late Cretaceous — reduced carbonate platforms, related mainly to collision.

Tertiary: Middle-Late Eocene — small carbonate banks on platform margins, related mainly to collision.

Middle-Late Miocene — small carbonate banks, lagoons or shoals on margins of isolated basins.

Pliocene–Quaternary — small isolated pools of freshwater carbonates.

The stratigraphic chart of these carbonate formations bounded to the "platform phases" and showing the proved and predicted number of their palaeokarstic horizons is illustrated by **Fig. 102**. The map of "Palaeokarst potential of Hungary" (**Fig. 103**) demonstrates their surface or subsurface distribution, indicating the total number of their possible superpositions.

Summarizing the above it can be stated that:

1. The greatest part of the palaeokarst potential in Hungary is located in Triassic–Early Jurassic platform carbonates of the Pelsó and Tisza units. The number of individual palaeokarstic horizons varies between 2–3 (Mecsek, Villány and Aggtelek Mountains) and 10–12 (Transdanubian Range, Bükk Mountains).

2. Less significance should be attributed to the Young Mesozoic (Late Jurassic–Early Cretaceous, Middle and Late Cretaceous) platform carbonates, having palaeokarst horizons of 1–2 levels, developed locally.

3. The palaeokarst potential of Tertiary and Quaternary carbonates with 1–1 palaeokarstic levels is considered unsignificant. The Late Eocene palaeokarst, as an exceptional one has a particular role in the thermal karst system of the Buda Hills and of its surroundings.

Concerning potential resources (karst and thermal water, caves, petroleum, bauxite, fireclay, Mn ores, base metals, Carlin-type gold ores) most are genetically related to the

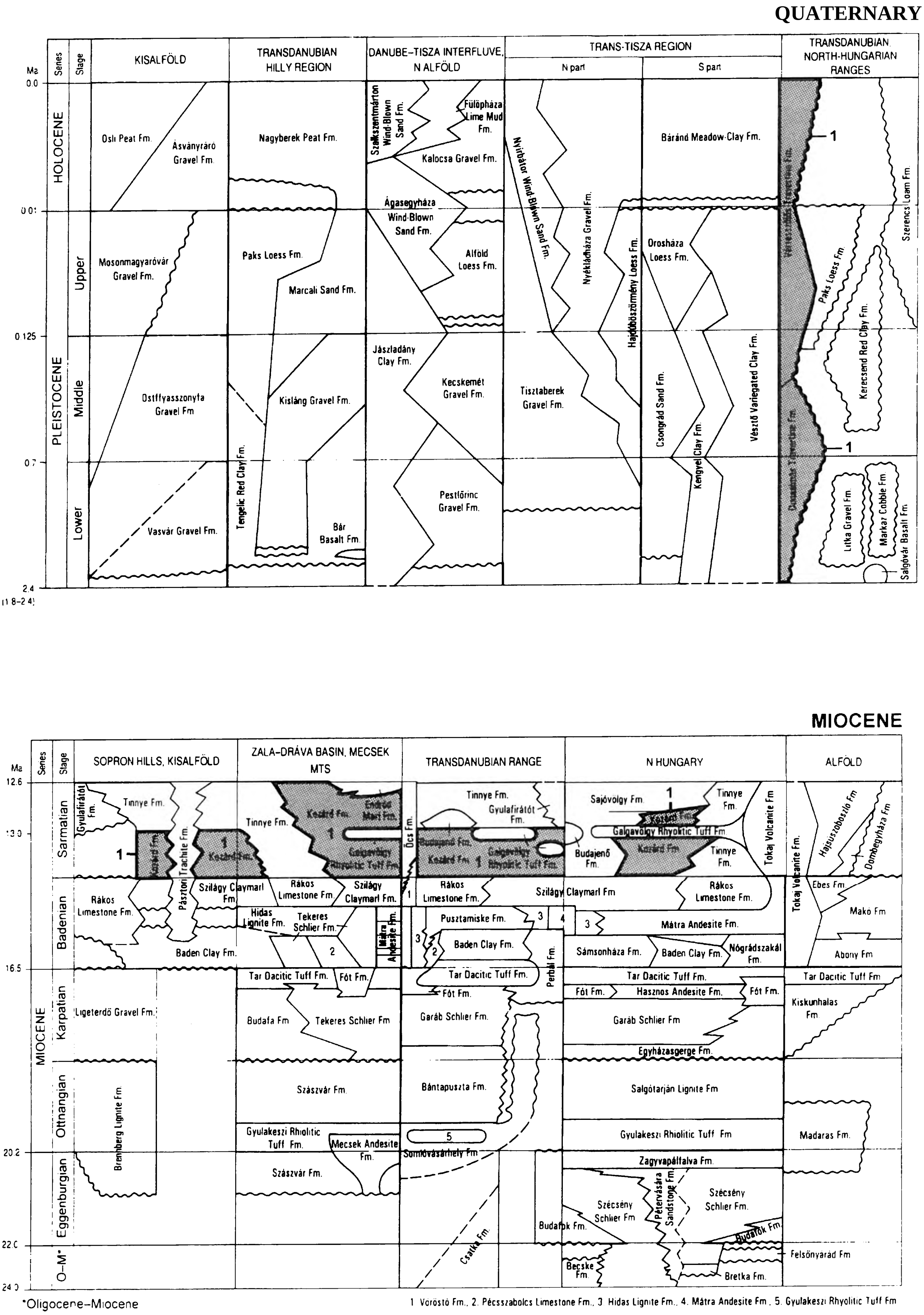
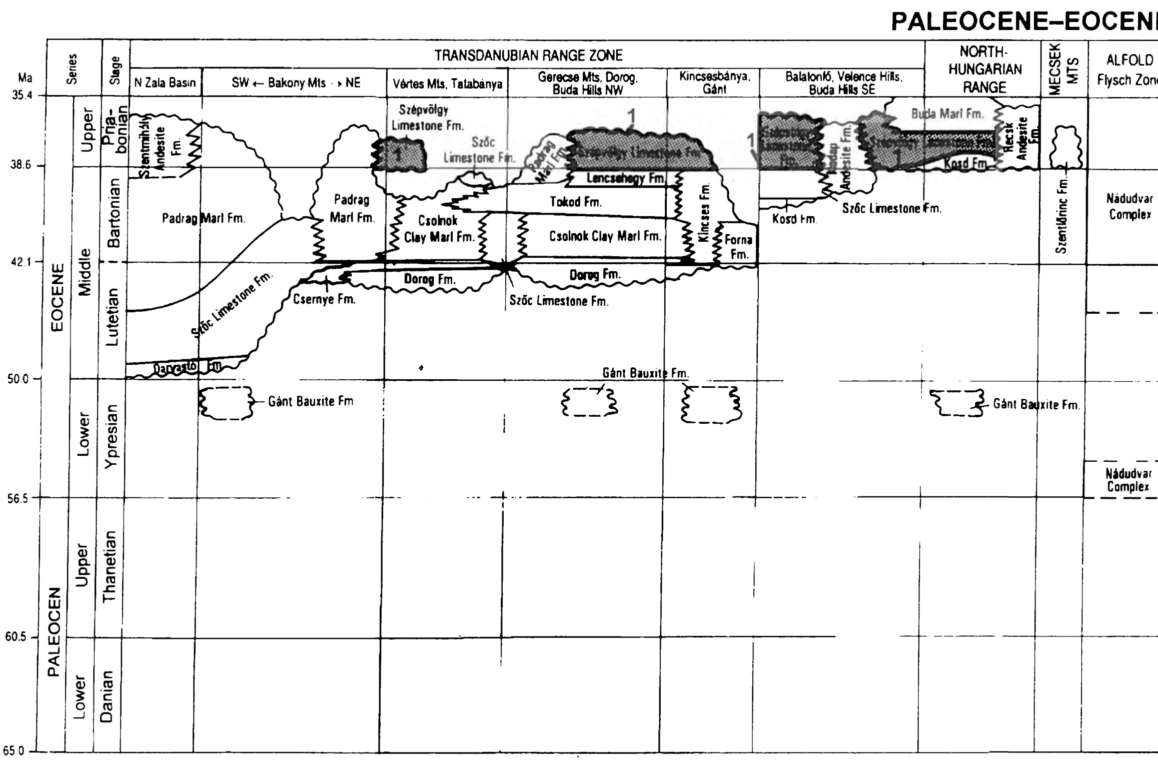
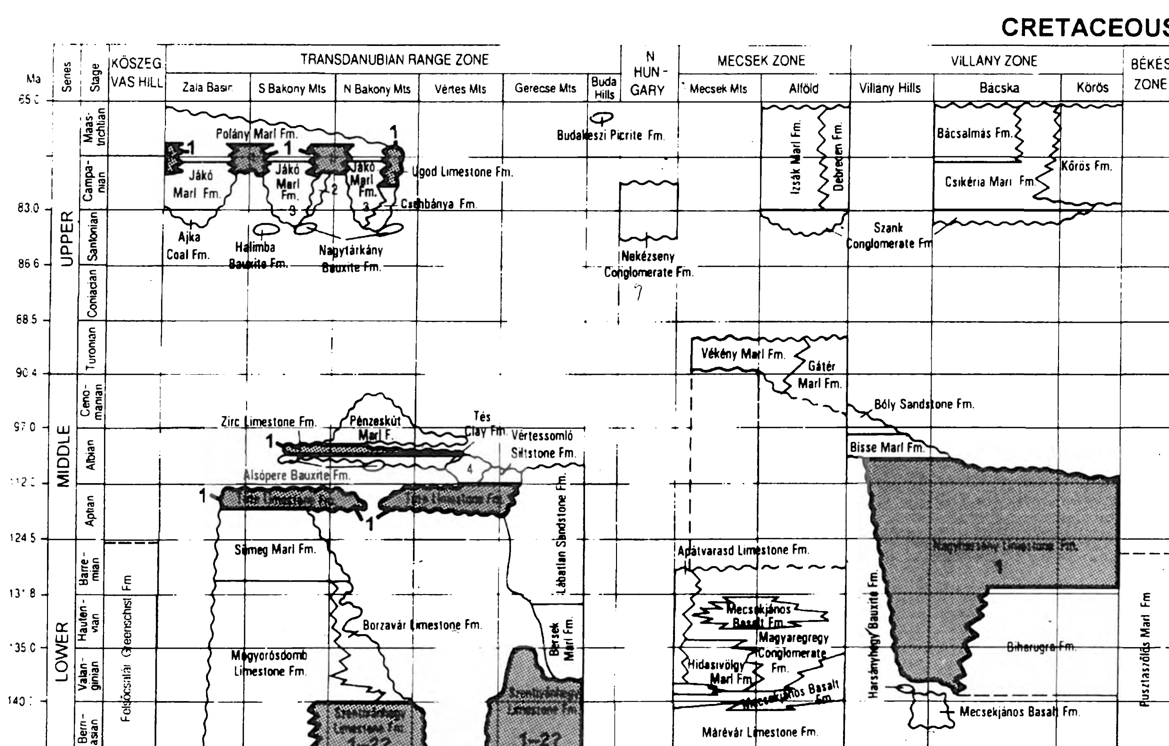


Fig. 102. Palaeokarst bearing formations of Hungary (after CSÁSZÁR 1997)

1–2. Number of the main palaeokarstic horizons. ? Unknown palaeokarstic horizons

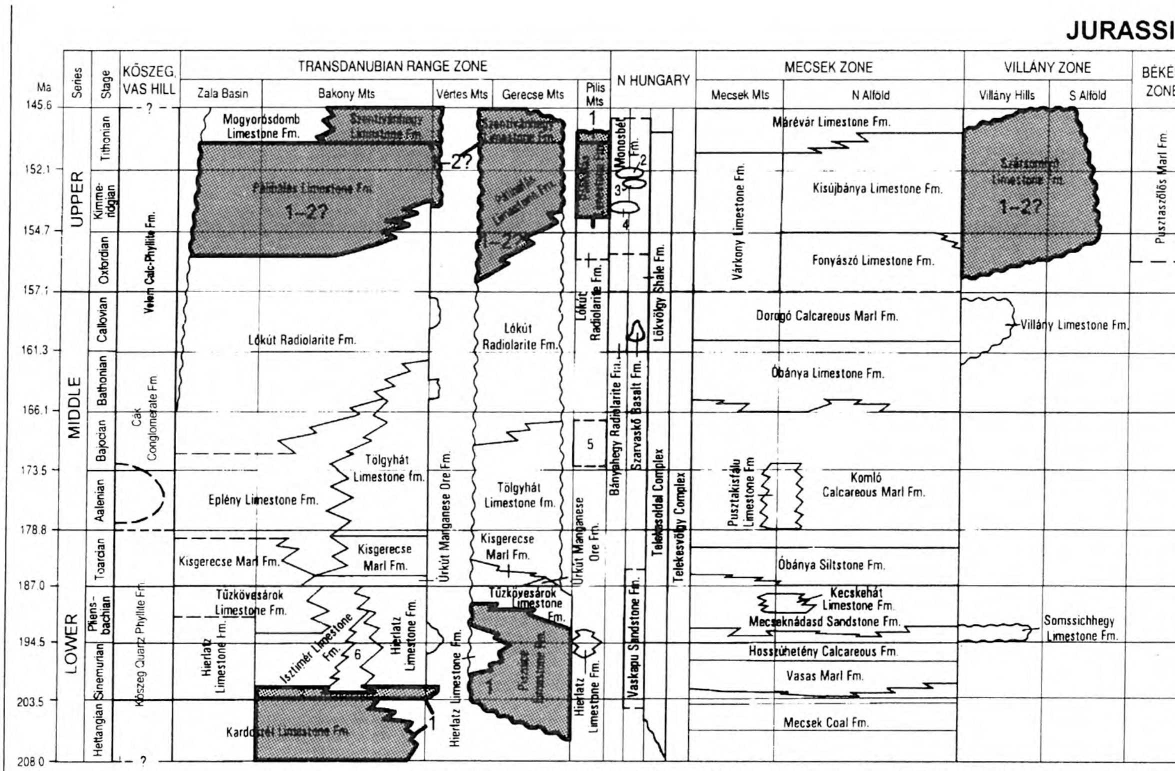


CRETACEOUS

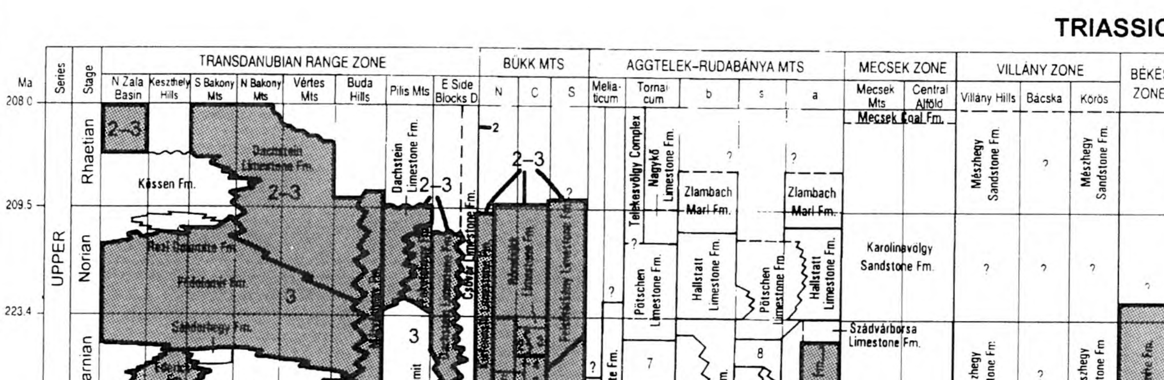


1. Ugod Limestone Fm., 2. Kozmatag Fm., 3. Ajka Coal Fm., 4. Környe Limestone Fm

Fig. 102 (continued)



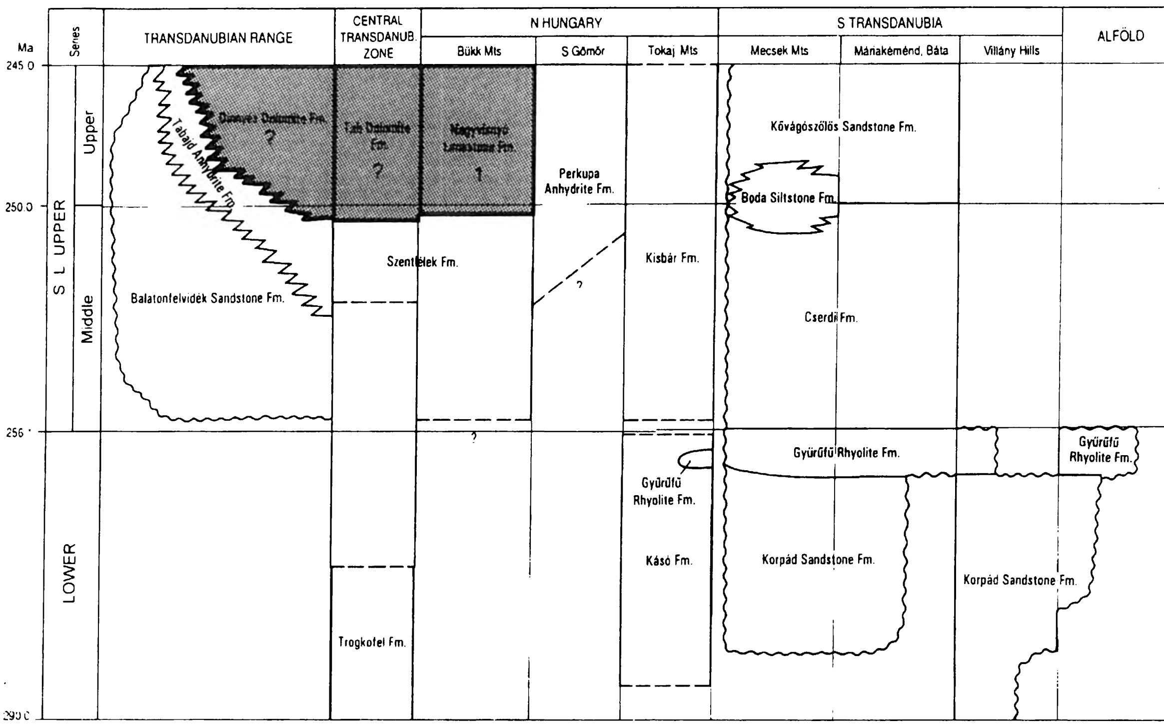
1. Szentivánhegy Limestone Fm., 2. Oldalvölgy Fm., 3. Bükkzsérc Fm., 4. Csipkéstető Radiolarite Fm., 5. Tölgyhát Limestone Fm., 6. Kishát Limestone Fm., 7. Pisznice Limestone Fm.



1. Répáshuta Limestone Fm., 2. Hollóstető Limestone Fm., 3. Bükkfennsík Limestone Fm., 4. Szinva Metabasalt Fm., 5. Vesszős Fm., 6. Hegyestető Fm., 7. Tornaszentandrás Shale Fm., 8. Szőlősardó Marl Fm.
b = Bódva Unit S = Szőlősardó Unit a = Aggtelek Unit
*Létrás Metabasalt Fm.

Fig. 102 (continued)

PERMIAN



PALEOZOIC I

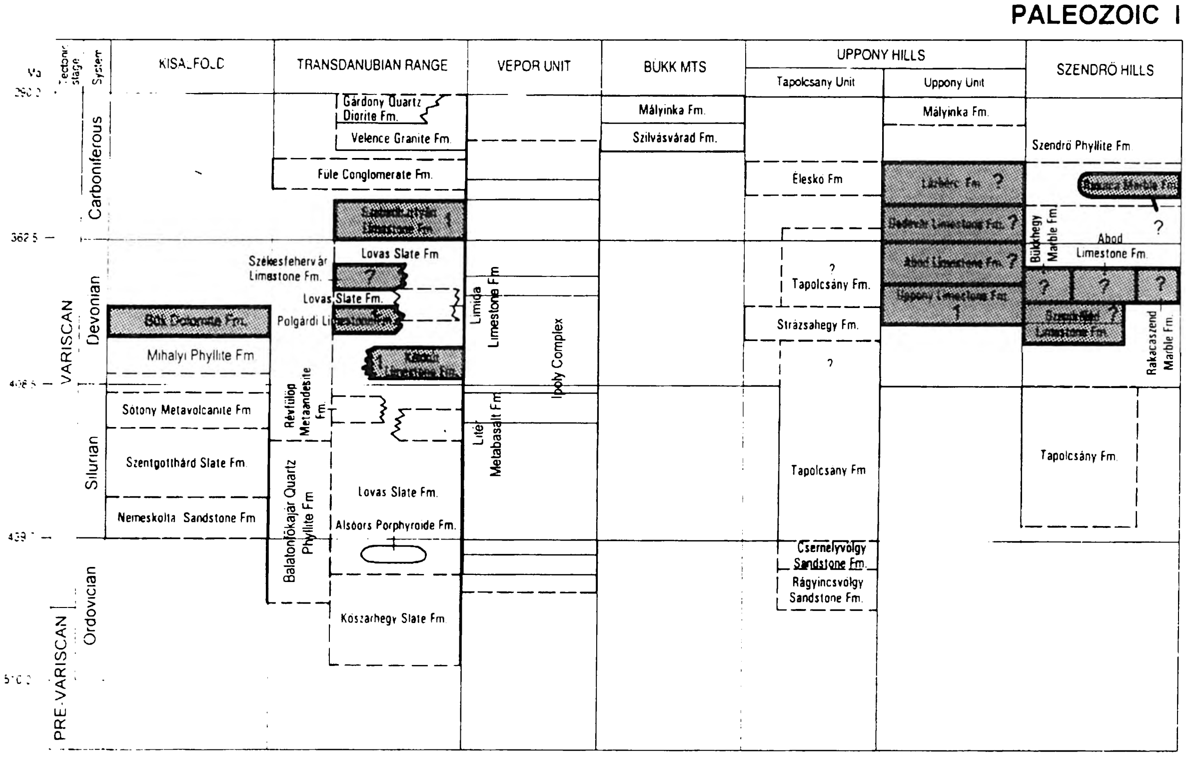


Fig. 102 (continued)

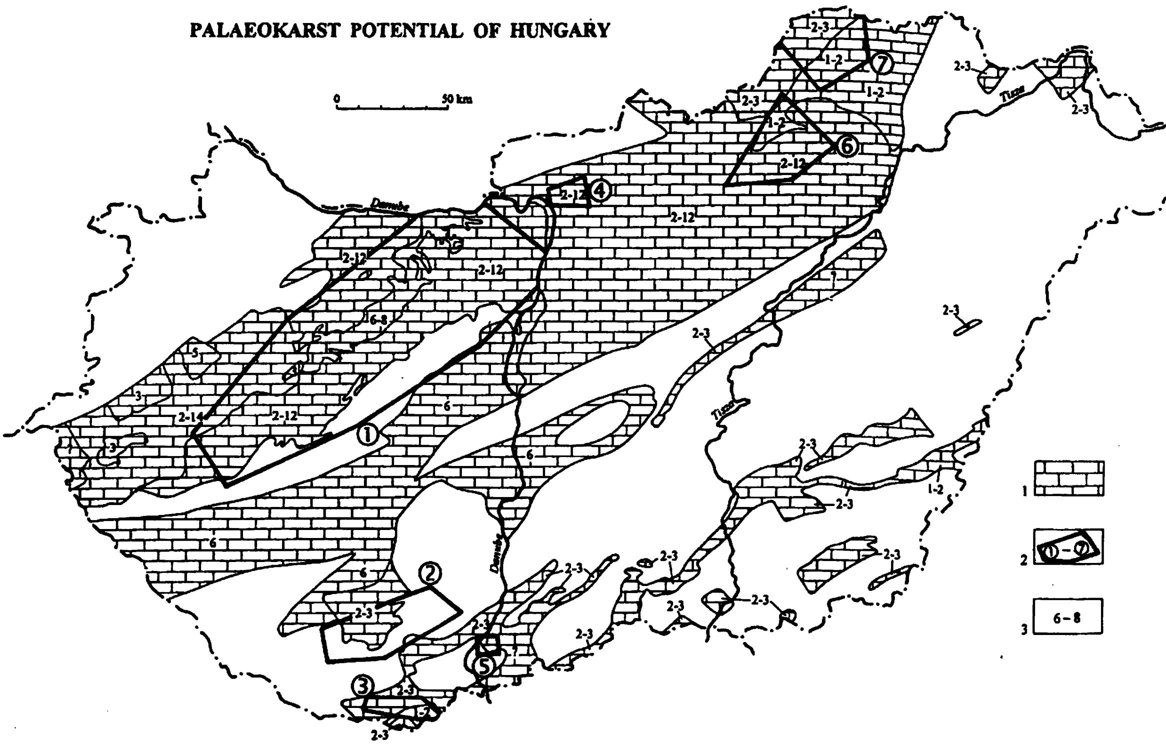


Fig. 103

1. Karstified formation of the pre-Tertiary basement, 2. Uncovered karstified formations at the surface: ① = Transdanubian Central Range, ② = Mecsek Mountains, ③ = Villány Mountains, ④ = Naszály and Csővár blocks, ⑤ = Mohács island blocks, ⑥ = Bükk and Uppony Mountains, ⑦ = Aggtelek–Rudabánya and Szendrő Mountains, 3. Number of the main palaeokarstic horizons

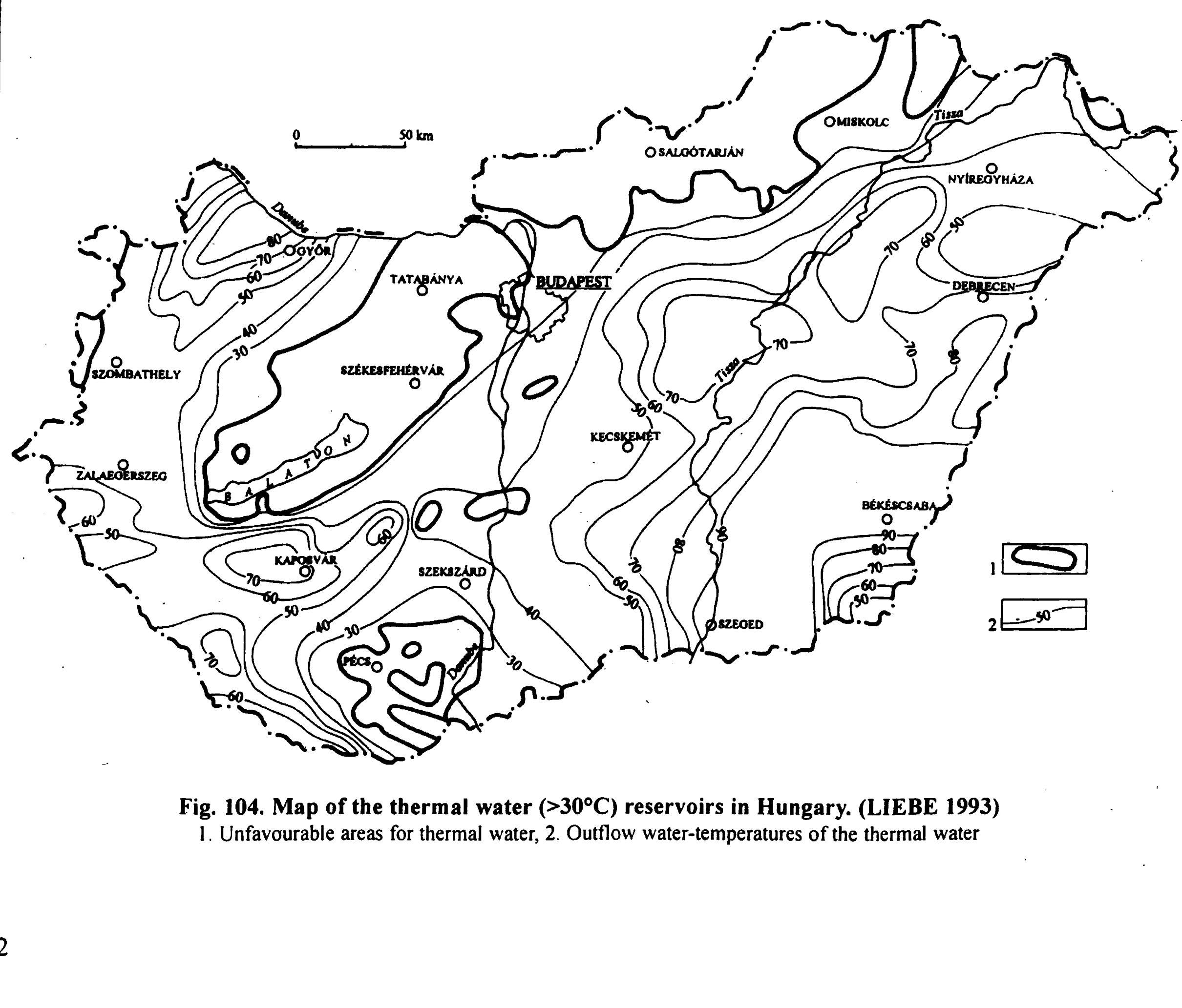


Fig. 104. Map of the thermal water (>30°C) reservoirs in Hungary. (LIEBE 1993)

1. Unfavourable areas for thermal water, 2. Outflow water-temperatures of the thermal water

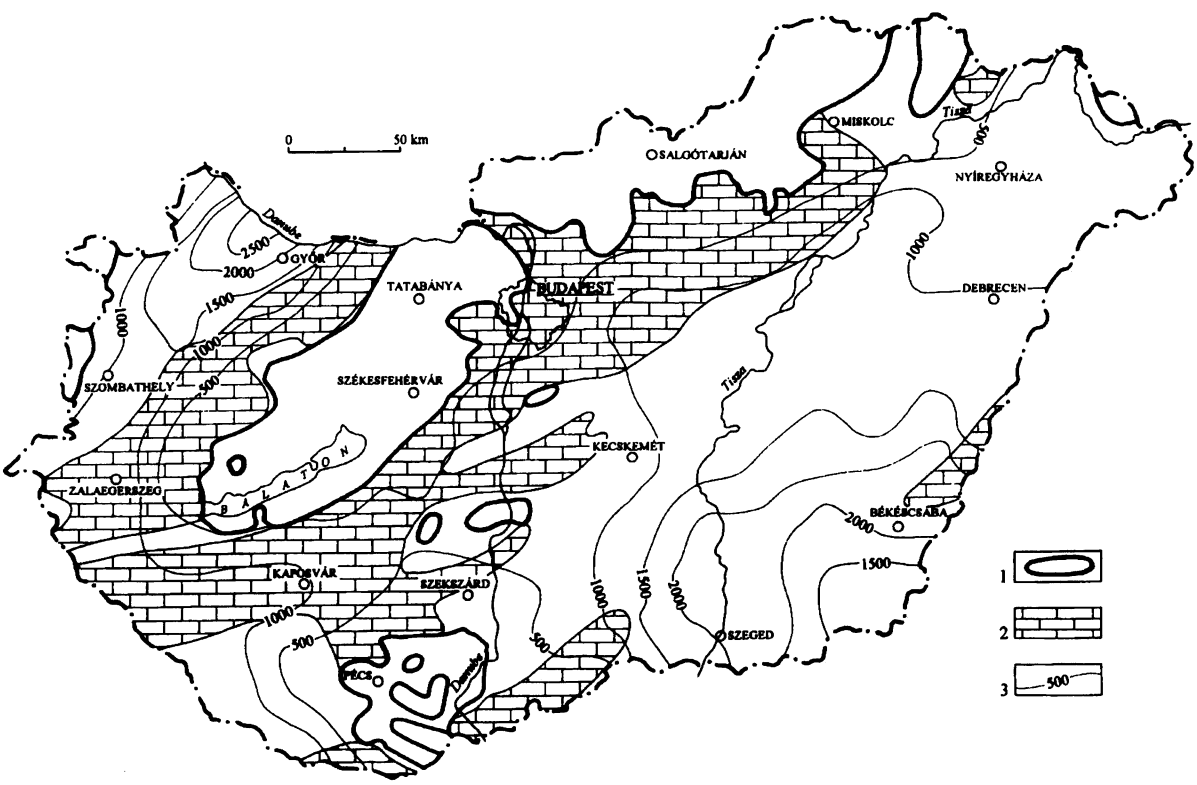


Fig. 105. Map of the karstic and nonkarstic reservoirs of thermal waters in Hungary (LIEBE 1993)
1. Unfavourable areas for thermal water, 2. Distribution of thermal karst reservoirs, 3. Depth of the Early and Late Pannonian boundary below the surface

palaeokarst systems. Let us recall again, that they yield about 10% of water reserves (**Figs. 104, 105**), approximately 30% of oil reserves (SW and Mid-Transdanubia, Danube–Tisza Interflow, Gödöllő Hills, Palaeogene basin of Northern Hungary), 100% of the bauxites and fireclays (Transdanubian Range, Danube left side horsts, Villány Mountains), as well as a considerable part of Mn ores (Bakony). An extraordinary natural value can be attributed to the 3000 known caves. Of course the relative importance of the different elements of this potential has recently changed significantly and it is still changing today. It has emphasized the role of karst and thermal waters, because of their decreasing quantity and deteriorating quality (LIEBE 1993, MINISTRY FOR ENVIRONMENT AND REGIONAL POLICY 1995). Particular attention and care have been paid to the high enthalpy karstic geothermal reservoirs (STEGENA et al. 1992, 1994, STEGENA 1994) of the deep basins in the Trans-Tisza Region (drill holes Fábiánsebestyén and Nagyszénás). More and more systematic efforts have been made to protect and rehabilitate the natural values of the caves.

In the light of the reduced use and exploitation of the traditional and nonrenewable mineral resources a stable or slightly increasing demand should be expected on petroleum deposited in palaeokarstic reservoirs. The new oil finds, located partly in palaeokarstic reservoirs should result in renewed evaluation of these reservoir systems, known in the above mentioned regions. The bauxite and fireclay deposits as well as the Mn ores located in palaeokarst and being under partial exploitation nowadays will not become of major importancer in the near future.

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Photo 1. Early marine infilling of laminite in the Gully oolite, Lower Carboniferous. Chipping Sodbury, England (KORPÁS 1992)

Photo 2. 238 Ma old flat palaeodoline in the Lower Muschelkalk. Centelles, Catalunia. (KORPÁS 1992)

Photo 3. Callovian–Barremian unconformity and related composite palaeokarst. Cap Salou, Catalunia. (KORPÁS 1992)

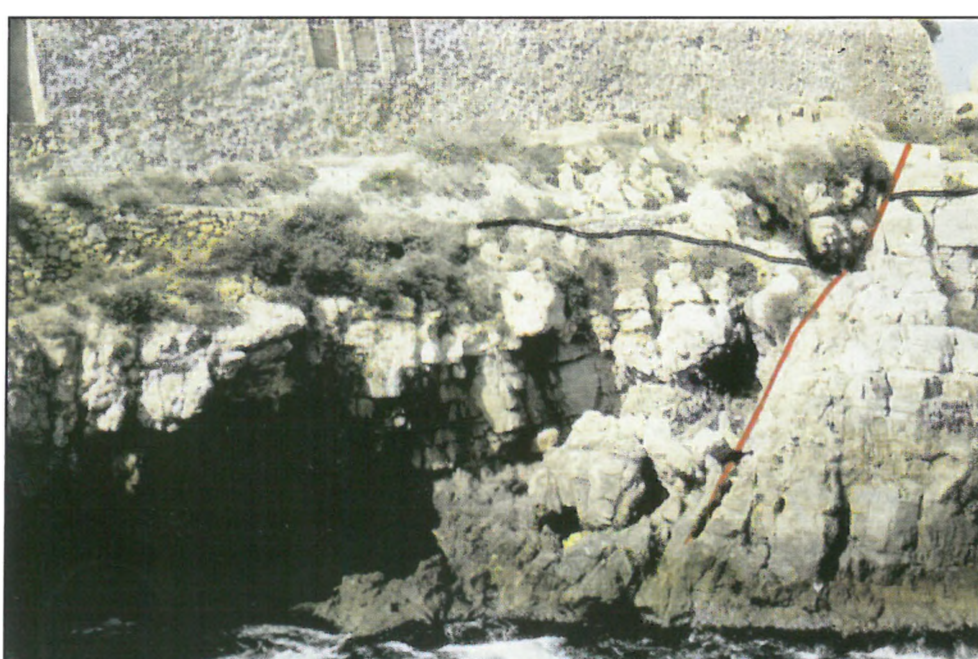


Photo 4. Liassic–Langhian unconformity and related palaeokarst. Tarragona, Catalunia. (KORPÁS 1992)

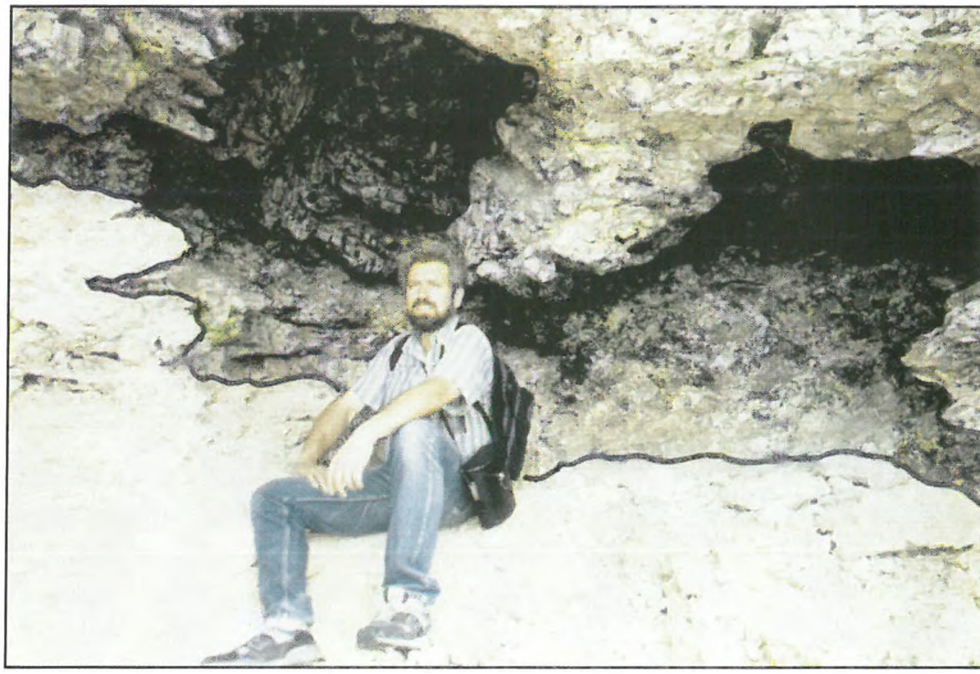
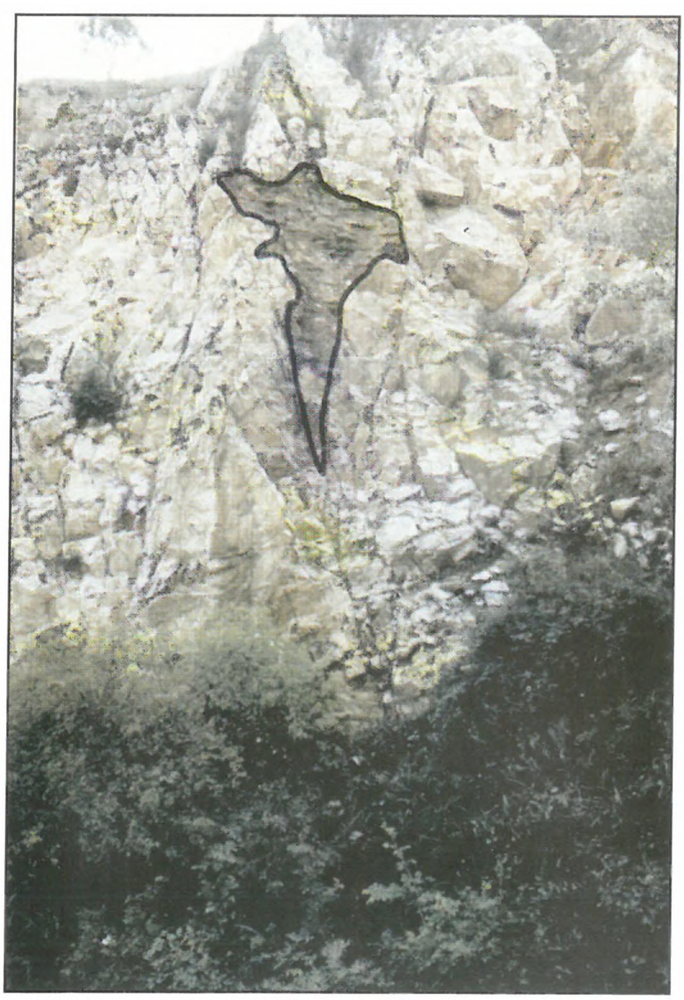


Photo 5. Late Miocene (6–7 Ma) marine infillings in the cavity of the Late Miocene (8 Ma) Cap Blanc reef. Cap Blanc, Mallorca. (KORPÁS 1993)

Photo 6. Palaeosol layers of the Pantokrator Formation, Late Triassic. Peninsula Peloponnessos, Korfos. (KORPÁS 1993)



Photos 7–8. Unconformable early marine laminites in the cavities of the Polgárdi Limestone, Devonian. Szabadbattyán, Kőszár-hegy. (KORPÁS 1994)

Photo 9. Palaeokarst horizons parallel to the bedding. Naszály Hill, Vác. (KORPÁS 1992)

Photo 10. Subaerial discontinuity surfaces and synsedimentary lystric fault in Dachstein Limestone, Late Triassic. Gerecse Mountains, Pisznice cave, Tízek terme. (TAKÁCS BOLNER 1991)

Photo 11. Early stylolites formed after lapiez at the top of a discontinuity surface in Dachstein Limestone, Late Triassic. Gerecse Mountains, Pisznice cave, Tízek terme. (TAKÁCS BOLNER 1991)

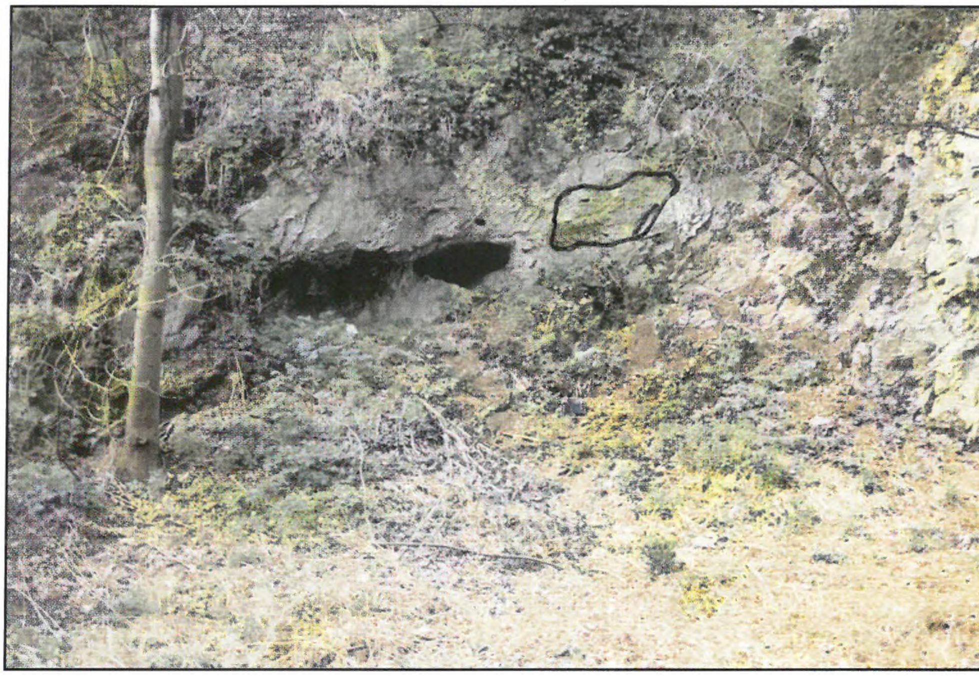
Photo 12. Normal microfaults in laminated freshwater limestone (Unit 2), Quaternary. Budapest, Várhegy, Táncsics Mihály u. 17. (KORPÁS 1994)

Photo 13. Tectonically controlled subaerial discontinuity surface in laminated freshwater limestone (Unit 2) with palaeosol horizon (Unit 3), Quaternary. Budapest, Várhegy, Táncsics Mihály u. 17. (KORPÁS 1994)

Photo 14. Open fissures in laminated freshwater limestone (Unit 2), infilled by palaeosol (Unit 3), Quaternary. Budapest, Várhegy, Táncsics Mihály u. 17. (KORPÁS 1994)

Photos 15–16. Overturned early palaeokarstic fissure with infilling of unconformable laminites in Fehérkő Limestone, Middle Triassic. Bükk Mountains, Lillafüred, István cave. (KORPÁS 1995)

Photos 17–18. Conformable (1) and unconformable (2) generations of early laminites with collapse breccia (3) in the Berva Limestone, Late Triassic. Bükk Mountains, Miskolc-Tapolca. (KORPÁS 1995)



Photos 19–20. Early marine laminites in the cavity of the Felsőtárkány Limestone, Late Triassic–Early Jurassic. Bükk Mountains, Felsőtárkány. (KORPÁS 1995)

GEOLOGICAL MAP OF THE BUDA HILLS, HUNGARY

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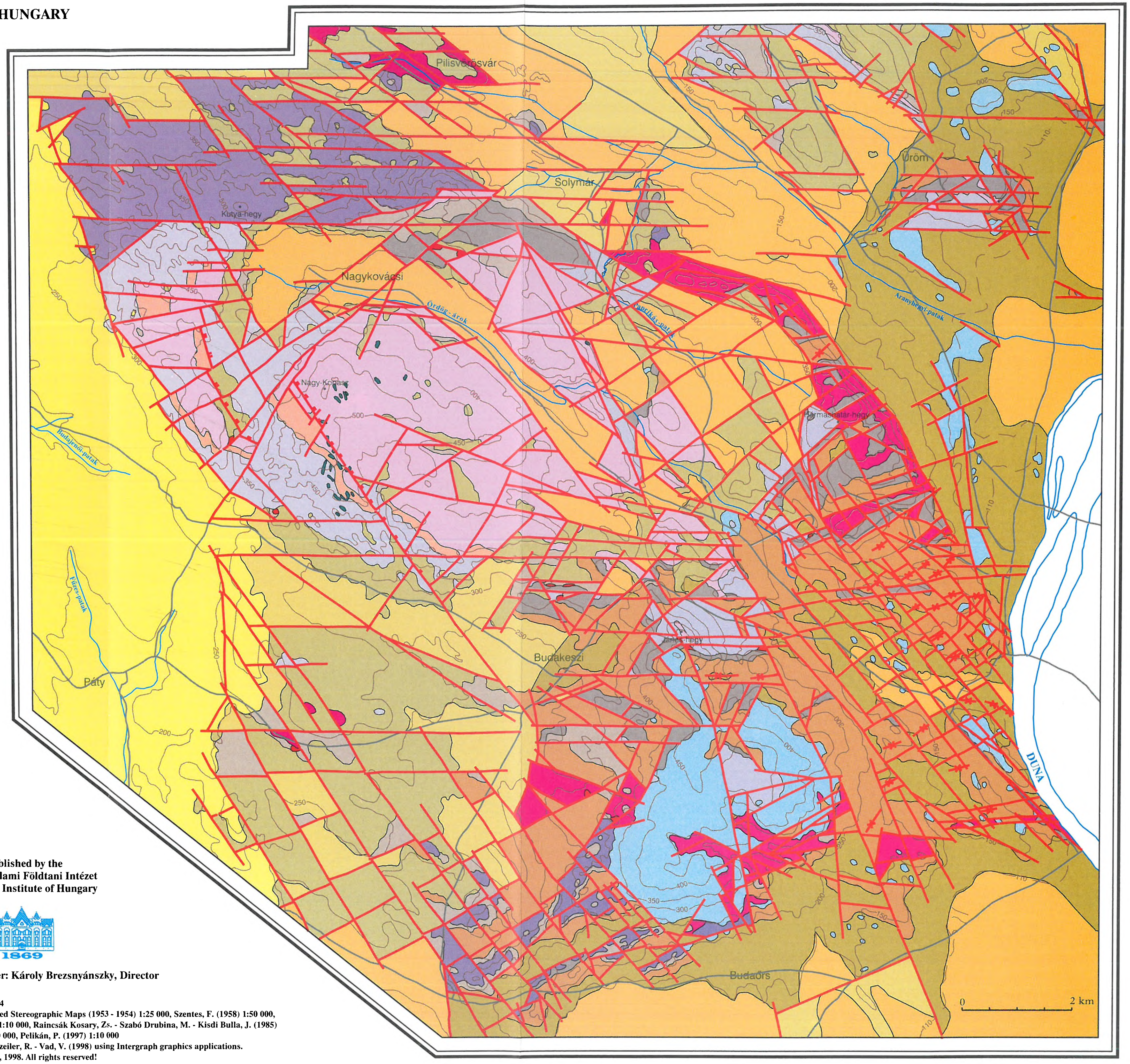

QUATERNARY - PLIOCENE

- Travertine and basal beds

LATE MIOCENE

- Undivided clastics and carbonates

LATE OLIGOCENE

- Törökbálint Sand (delta and delta front sediments)

LATE OLIGOCENE - EARLY OLIGOCENE

- Kiscell Clay (neritic to bathyal basin with prodelta sediments)

EARLY OLIGOCENE

- Hárshegy Sandstone (moundbar, littoral)
- Tard Clay (restricted basin with prodelta sediments)

EARLY OLIGOCENE - LATE EOCENE

- Buda Marl (pelagic with open shelf and slope sediments)

LATE EOCENE

- Szépvölgy Limestone (open shelf, carbonate bank)
- Basal clastics (fandelta sediments)

LATE CRETACEOUS

- Budakeszi Pikrite (continental alkaline ultrabasites)
- Bauxite and bauxitic clay

LATE TRIASSIC

- Dachstein Limestone (platform peritidal)
- Transitional Complex (platform peritidal limestones and dolomites with paleosol layers)
- Hauptdolomite (platform peritidal)
- Mátyáshegy Formation (platform and drowned platform with dolomite, chert bearing dolomite; intraplatform restricted basin with chert bearing clastics and carbonate turbidites)

MIDDLE TRIASSIC

- Budaörs Dolomite (platform peritidal with paleosol layers)

- Faults (normal, dextral and sinistral)
- Reverse faults
- Syncline

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PRE-TERTIARY BASEMENT MAP OF THE BUDA HILLS, HUNGARY

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LATE CRETACEOUS

- Budakeszi Pikrite (continental alkaline ultrabasites) at the surface
- Bauxite and bauxitic clay at the surface
- Bauxite and bauxitic clay in drillholes

LATE TRIASSIC

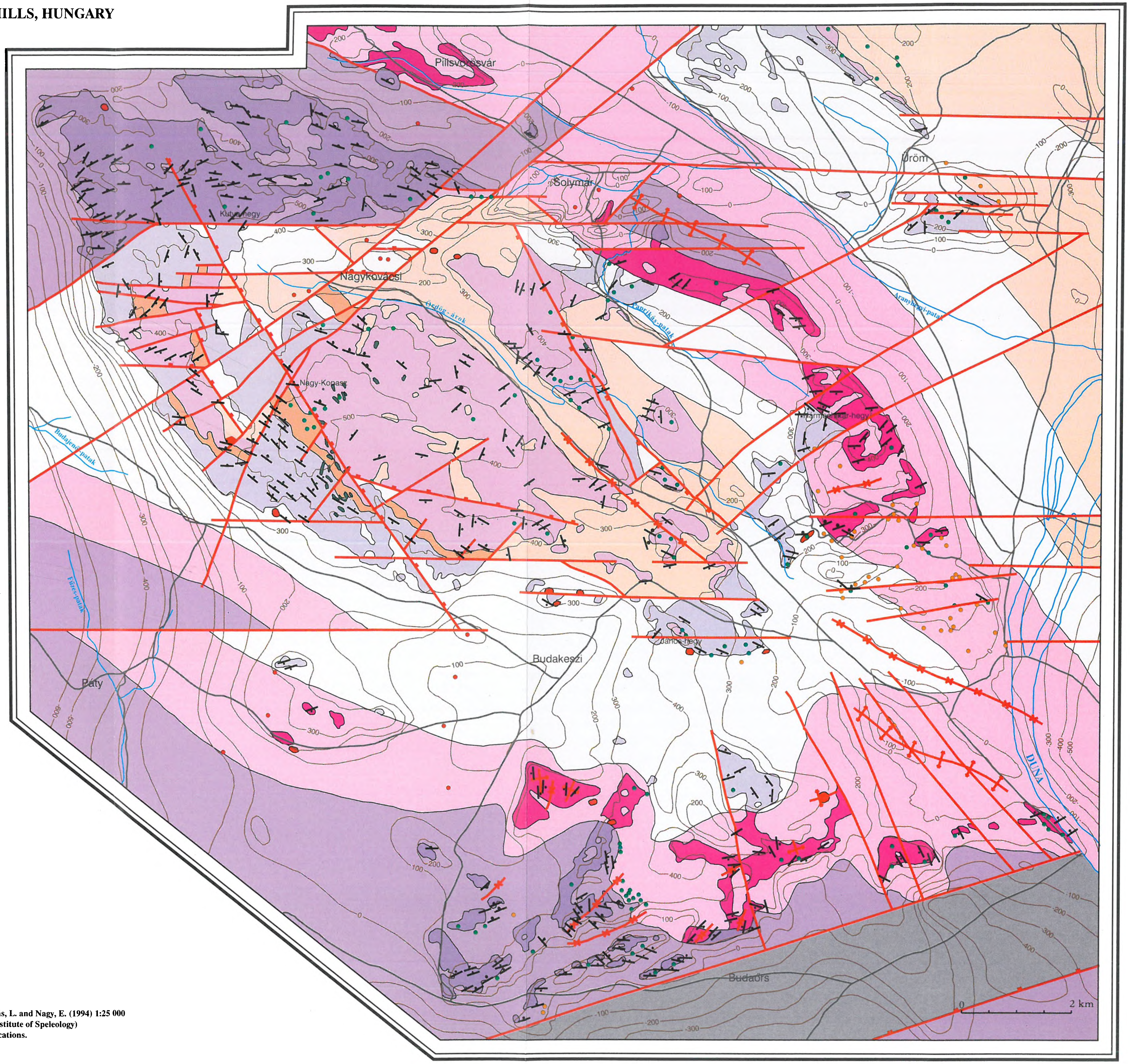
- Dachstein Limestone (platform peritidal)
- Transitional Complex (platform peritidal limestones and dolomites with paleosol layers)
- Hauptdolomite (platform peritidal)
- Mátyáshegy Formation (platform and drowned platform with dolomite, chert bearing dolomite; intraplatform restricted basin with chert bearing clastics and carbonate turbidites)

MIDDLE TRIASSIC

- Budaörs Dolomite (platform peritidal with palesol layers)

EARLY - MIDDLE TRIASSIC

- Aszófő Dolomite (lagoon, tidal-peritidal) and Csopak Marl (open shelf subtidal clastics with limestone and dolomite)
- Dip of strata
- Faults (normal, dextral and sinistral)
- Reverse faults
- Syncline
- Anticline
- 100 Basement contourlines above sealevel (m)
- Caves and cavities in Triassic bedrocks
- Dolines and sinkholes in Triassic bedrocks
- Caves and cavities in Late Eocene bedrocks



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3D MODEL OF THE COMPOSITE KARST SYSTEM, BUDA HILLS, HUNGARY

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GEOLOGY

LATE TRIASSIC

- **Dachstein Limestone and Transitional Complex** (platform peritidal limestones and dolomites with palaeosol layers in the basal Transitional Complex)
- **Hauptdolomite** (platform peritidal)
- **Mátyáshegy Formation** (platform and drowned platform with dolomite, chert bearing dolomite; intraplatform restricted basin with chert bearing clastics and carbonate turbidites)

MIDDLE TRIASSIC

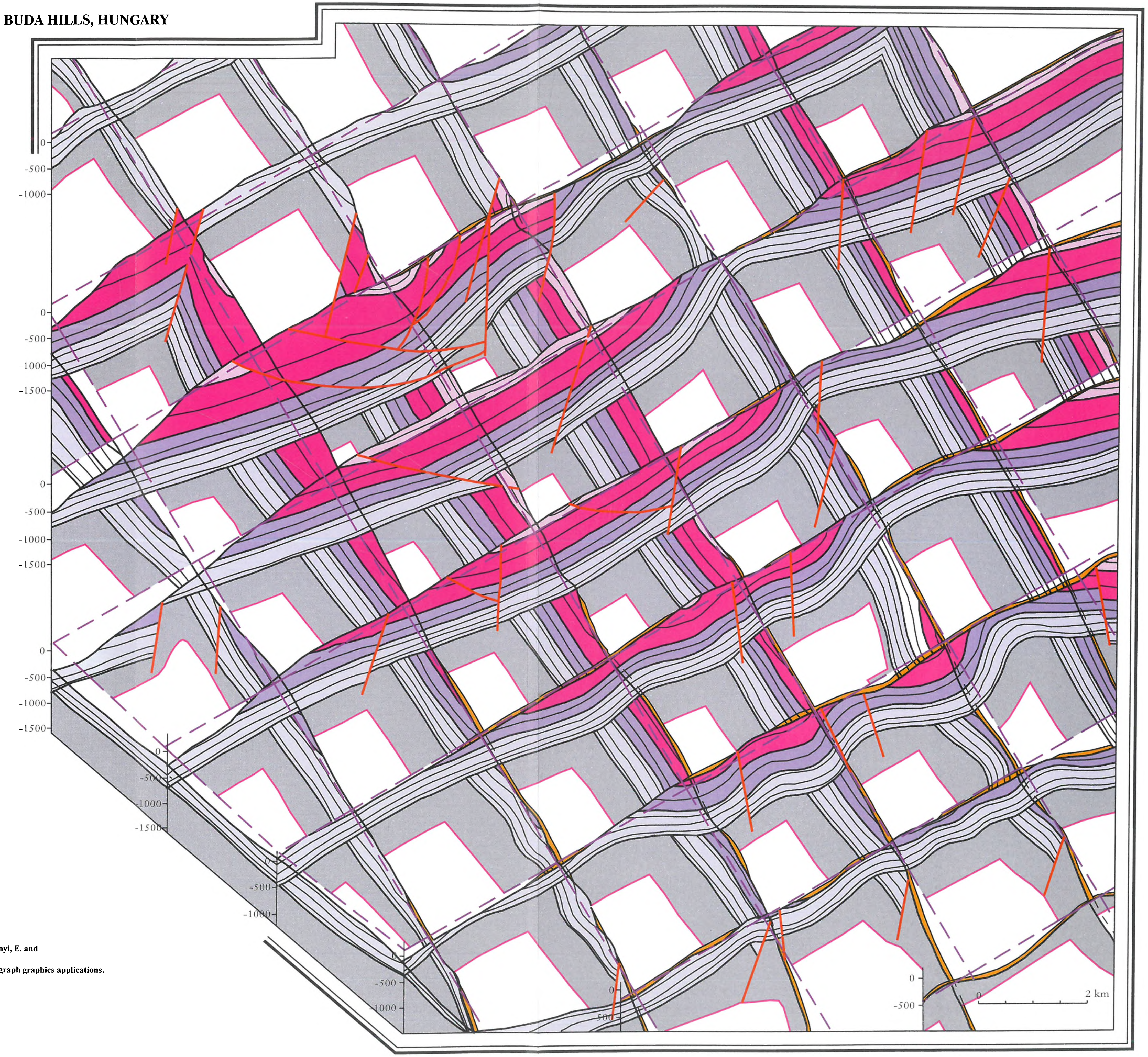
- **Budaörs Dolomite** (platform peritidal with palaeosol layers)

EARLY - MIDDLE TRIASSIC

- **Aszófő Dolomite** (lagoon, tidal-peritidal) and **Csopak Marl** (open shelf subtidal clastics with limestone and dolomite)
- Normal and reverse faults

KARST

- Main palaeokarstic horizons (composite disconformities at the boundary of the Triassic formations)
- Intraformational palaeokarstic horizons (single disconformities within the Triassic formations)
- Intraformational palaeokarstic horizon (single disconformity within the Late Eocene Szépvölgy Limestone)
- Karst water table above the sea-level (according to the state of 01. 01. 1992.)



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