GEOLOGICA CARPATHICA, 53, 6, BRATISLAVA, DECEMBER 2002
351–367

LOWER TRIASSIC SHALLOW MARINE SUCCESSION IN THE BÜKK MOUNTAINS, NE HUNGARY

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(Manuscript received December 7, 2001; accepted in revised form June 18, 2002)

Abstract: This paper presents the Lower Triassic sequence in the Bükk Mountains (NE Hungary). On the basis of the revised macrofossil-collections and previous conodont studies, a biostratigraphic correlation and a chronostratigraphic subdivision of the formations and members is given. *Hindeodus parvus* (Kozur et Pjatakova), *Isarcicella isarcica* (Hukriede), *Claraia aurita* (Hauer), *Eumorphotis* gr. *multiformis* (Bittner), *Eumorphotis* cf. *hinnitidea* (Bittner), *Costatoria subrotunda* (Bittner), *Eumorphotis kittli* (Bittner), *Tirolites illyricus* Mojsisovics, *Tirolites seminudus* Mojsisovics, *Costatoria costata* (Zenker), *Rectocornuspira kalhori* Brönnimann, Zaninetti et Bozorgnia, *Cyclogyra? mahajeri* Brönnimann, Zaninetti et Bozorgnia, and *Meandrospira pusilla* (Ho) are the main index fossils. Two sections, which represent the entire Lower Triassic succession, were studied from a sedimentological point of view. These are the Gerennavár section for the Gerennavár Limestone Formation, and the Lillafüred section for the Ablakoskővölgy Formation, respectively. The sedimentological data and facies interpretations are summarized in facies models. During Early Triassic time the depositional area of the Bükk Mountains were situated in the epeiric shelf of the Western Tethys. The model of the Gerennavár Limestone Formation reflects the high-energy, tide- and wave-dominated shallow shelf, characterized by a protected–stabilized muddy sand flat and a high-energy sand belt. The model of the siliciclastic, lower part of the Ablakoskővölgy Formation indicates a mixed, dominantly siliciclastic shallow shelf with deposition in the coastal, shoreface and transitional zones. The model of the carbonate, upper part of the Ablakoskővölgy Formation refers a storm-controlled shelf with facies representing the whole spectrum from the peritidal–shallow shoals to the low-energy deeper subtidal zone below the storm wave-base. Lithologic and facies comparison of the Lower Triassic succession of the Bükk Mountains to other sequences of the Western Tethyan depositional area reveals many differences, and fewer similarities, which suggests local controls on depositions, that is locally different terrigenous siliciclastic input, different subsidence rate, antecedent topography, and dominance of the local physical regime.

Key words: Lower Triassic, NE Hungary, Bükk Mountains, stratigraphy, facies interpretation, facies model, fossils.

Introduction

Reinvestigations of the Mesozoic succession in the Bükk Mountains have given rise to much controversy for the last two decades. It seems that the only exception is the Lower Triassic part since the basic stratigraphic subdivision by Schréter (1935, 1953, 1954) and Balogh (1964) is commonly accepted. The lithostratigraphic subdivision of Lower Triassic deposits was established by Balogh (1980) and was refined by Pelikán (1985a, 1995) as a result of a mapping program of the Hungarian Geological Institute starting in 1979.

The aim of this paper is to summarize the present-day knowledge of the Lower Triassic sequence in the Bükk Mountains and to present a general outline of its stratigraphy and sedimentology. A review of the fossil collections of the Hungarian Geological Institute provides a possibility for comparison of the succession to the Western Tethyan biozonation, and for the chronostratigraphic subdivision. Sedimentological studies of the Gerennavár and Lillafüred section have resulted in more detailed facies interpretations and a better understanding of the facies successions of the Lower Triassic formations in the Bükk Mountains.

Geological setting

The Bükk Mountains are situated in Northern Hungary, south of the Inner Western Carpathians. It is a part of the Bükk Composite Unit (Bükkia Composite Terrane) which belongs to the Pelso Megaunit (Pelsonia Composite Terrane) (Kovács et al. 2000) (Fig. 1). However, Paleozoic–Mesozoic sequences were deposited in the northwestern neighbourhood of the Inner Dinarides according to the paleogeographical reconstruction by Protić et al. (2000), and Filipović et al. (in press). The block of Paleozoic–Mesozoic sequences of the Bükk Mountains was displaced northeastward to its recent position along a large-scale, regional dextral lateral fault-zone (Mid-Hungarian Lineament) during the Tertiary (summary in Fodor & Csontos 1998).

The succession of the region comprising the Bükk Mountains is composed of intensely deformed, anchimetamorphic Mesozoic rocks, surrounded by a non-metamorphic Paleogene-Neogene cover (Árkai 1973, 1983; Csontos 1999). The Lower Triassic formations are exposed only in the Northern anticline (details in Less et al. in press). Because of low-grade metamorphism and strong ductile deformation of the sedi-

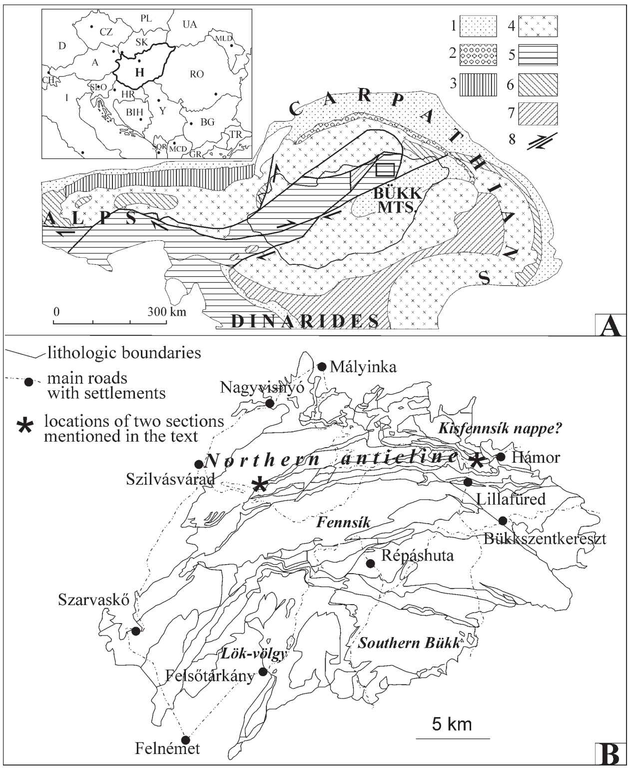

Fig. 1. A — Schematic terrane map of the Circum Pannonian Region (Kovács et al. 2000); **1** — Flysch Belt, **2** — Klippen Belt, **3** — Northern Calcareous Alps, **4** — Early Alpine unit related to European continental margin, **5** — Early Alpine shelf sequences related to the Apulian (Southern Alps and Outer Dinarides) continental margin, **6** — Ophiolites of the Penninic Ocean, **7** — Ophiolites of the Vardar Ocean, **8** — Major strike-slip zones; **B** — Position of the Northern anticline in the structure of the Bükk Mountains (after Csontos 2000).

mentary rocks, the original macroscopic depositional structures are only scarcely visible, which makes facies interpretations more difficult.

The Lower Triassic sequence (Fig. 2) is more completely represented by the Bálvány section (lower part of the Gerennavár Limestone), the Gerennavár section (Gerennavár Limestone Formation, Fig. 3), and an overturned but continuous section at Lillafüred (Ablakoskővölgy Formation, Fig. 4). The two latter sections were studied in detail from a sedimentological point of view and were amplified by observations from other outcrops.

Fossil assemblages and stratigraphy

A review of the available fossils in the collection of the Hungarian Geological Institute is presented here. Most of the macrofossils were collected by Legányi and Schréter (Schréter 1935, 1953, 1954) and Balogh (1964). They have collected and described relatively rich faunal assemblages from a few locations. Especially, the surroundings of Bogdány-tető (at the non-metamorphosed northern limb of the Northern anticline) yielded 'numerous' fossils. According to Less et al. (in press), the Ablakoskővölgy Formation crops out there, however, the original stratigraphic position of the members was disturbed by complicated tectonic displacements. Otherwise, there are only scanty data on the occurrence of the marker species. Based on the descriptions of Schréter (1935, 1953, 1954) and Balogh (1964), and the results of mapping (Less et al. in press), the major part of the index fossils could be assigned to lithostratigraphic units, established by Balogh (1980) and Pelikán (1985a, 1995). However, precise positions of the markers within the formations or members could not always be reconstructed. Systematic collection of macrofossils from the Scythian part has not been carried out since Schréter's and Balogh's collections. The preservation of the fossils is generally not good.

Foraminifers were determined by Bérczi-Makk (Bérczi-Makk 1986, 1987; Bérczi-Makk et al. 1995; Pelikán 1995) and Oravecz-Scheffer (Pelikán 1985a). Kozur (1985, 1988) studied the conodont and ostracod assemblages near the Permian/Triassic boundary.

Biostratigraphic zonation of the Western Tethyan sections is mostly based on molluscs (Krystyn 1974; Broglio Loriga et al. 1983; Neri & Posenato 1985; Broglio Loriga & Mirabella 1986; Broglio Loriga & Posenato 1986; Broglio Loriga et al. 1990; Posenato 1992) and conodonts (Perri 1991; Yin 1993; Yin et al. 1996; Zhang et al. 1996; Orchard 2001), and partly foraminifers (Salaj et al. 1983; Broglio Loriga et al. 1990), but it seems that some taxa of the foraminifers are only facies-indicators (see Hips 1996). However, as in many parts of the Western Tethyan domain, within the Lower Triassic the boundary between the Induan and Olenekian stages cannot be determined due to the simple, low-diversity fauna. A more detailed chronostratigraphic subdivision of the formations and their members in the Bükk Mountains can be made only on the basis of the three-fold subdivision (Griesbachian, Nammalian, Spathian), as in other Western Tethyan sequences. The paper does not deal with the problem of the Permian/Triassic boundary.

Gerennavár Limestone Formation

The underlying 'Leptodus member' of the Upper Permian Nagyvisnyó Limestone Formation is composed of black, thin-bedded limestones rich in micro- and macrofossils (Fülöp 1994). The lowermost beds of the Gerennavár Limestone Formation indicate a sharp lithological change.

Lithostratigraphic definition. The Gerennavár Limestone Formation consists of predominantly grey, thick-bedded oolites, and subordinately of thin-bedded mudstones. The lower part of the formation is subdivided into two characteristic horizons. The 'basal bedset' of the Gerennavár Limestone Formation (Figs. 2–3) is approximately 1 m in thickness, and consists of dark grey clayey marls, followed by sandstones, calcareous sandstones, sandy clayey marls, and in a few sections sandy dolomites. It is overlain by the 'transitional bedset' ca. 6–7 m in thickness, which is made up of thin- to thick-bedded, dark grey, fine laminated limestones. Above them thin- to thick-bedded mudstones are recorded, and upwards thick-bedded oolite limestones form the bulk of the formation ca. 100–130 m in thickness (Figs. 2–3). Locally coarse crystalline, late diagenetic dolomite lenses occur.

Fossils. From the 'basal bedset', Csontos-Kis (in Pelikán & Csontos-Kis 1990) reported the occurrence of bivalves and

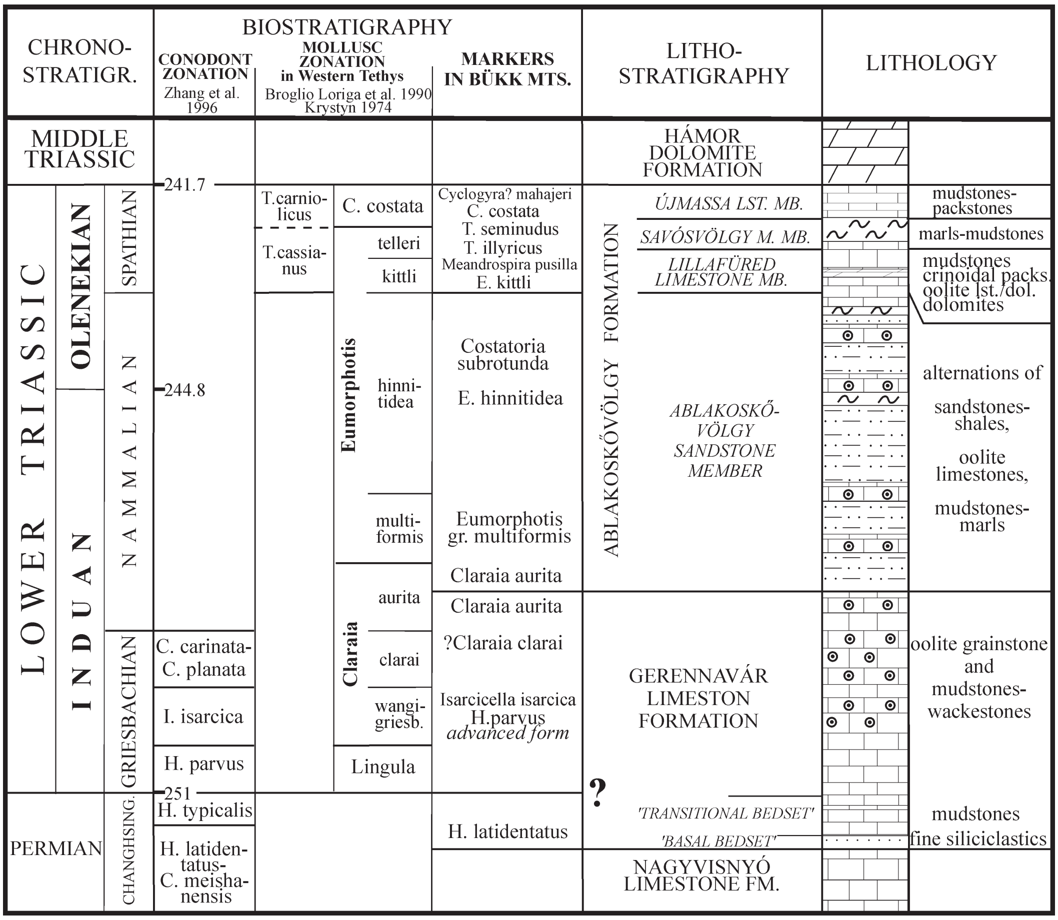

Fig. 2. Stratigraphic subdivision, and biostratigraphic markers of the Lower Triassic sequence in the Bükk Mountains. For comparison conodont biostratigraphy (only in the upper Changhsingian and Griesbachian) and mollusc biozonation in the Western Tethys are also indicated. Vertical subdivision is time proportional (radiometric data of the Permian/Triassic boundary is the proposal of the Permian/Triassic Boundary Working Group 1999, based on Zhang et al. 1992; Claoué-Long et al. 1991; Renne et al. 1995; Bowring et al. 1998; Metcalfe et al. 1999, two other data are compiled by Gradstein et al. 1994).

brachiopods, but some of the species are not valid, because the species proposed by Schréter have never been published. This rich Upper Permian (Changhsingian) macrofauna is characterized by the abundance of Pectinoid bivalves, while brachiopods and gastropods are rarer.

From the 'transitional bedset' only microfossils of Changhsingian age were found, that is foraminifers: *Earlandia dunningtoni* (Elliott), *E. tintinniformis* (Mišik), *E. deformis* Bérczi-Makk, *Neotuberitina reitlingerae* (Mikl. Maklay), *Globivalvulina graeca* Reichel, *Geinitzia* sp., *Ammodiscus* sp., *Pachyphloia* sp. (Bérczi-Makk 1986, 1987; Bérczi-Makk et al. 1995), *Agathammina pusilla* (Geinitz), *Paraglobivalvulina mira* Reitlinger (Kozur 1988), ostracods: *Indivisia buekkensis* Kozur, *Goranella* sp., *Judahella bogschi* Kozur; holothuridea: *Theelia dzulfaensis* Mostler et Rahimi-Yazd; and conodont: *Ellisonia transita* Kozur et Mostler (Pelikán 1985b; Kozur 1985, 1988; pers. commun. in Fülöp 1994).

Kozur (1988) reported the occurrence of the conodont *Hindeodus latidentatus* (Kozur; Mostler et Rahimi-Yazd), which is identical with *H. praeparvus* Kozur (Kozur 1996), from the 'transitional bedset'. *Hindeodus parvus* (Kozur et Pjatakova), which is regarded as the index fossil of the basal Triassic (Orchard 2001), is not reported near the lower boundary of the Gerennavár Limestone Formation. The 'basal' and the 'transitional bedsets' contain Permian and longer-life faunal elements.

Above a conodont-free interval the first appearance of the conodont *Hindeodus parvus* (Kozur et Pjatakova) with an advanced form was recorded from ca. 15 m (in the core drilling Mályinka-8, Pelikán 1985b), and 20 m (Kozur 1988) above the lower boundary of the formation. Still higher, approximately from the middle part of the formation, the conodonts *Isarcicella isarcica* (Huckriede), *Hindeodus parvus* (Kozur et Pjatakova), and *Ellisonia aequabilis* Staesche, and the ostracod *Hollinella tingi* (Patte) were encountered (Kozur in Pelikán 1985a; Kozur 1985, 1988). Kozur determined *Callicythere postiangulata* Wei, *Liuzhinia parva* Wei, *Liuzhinia* sp., *Bairdia* sp., *Polycope* sp. from the lower half of the formation (in Pelikán 1985b). From the upper half of the formation, *Cyclogyra? mahajeri* Brönnimann, Zaninetti et Bozorgnia, *Spirorbis phlyctaena* Brönnimann et Zaninetti were also reported (Oravecz-Scheffer in Pelikán 1985a). The bivalve *Claraia clarai* (Emmrich) was mentioned by Schréter (1953) from the Bálvány section, but only one dubious specimen is present in the collection of the Hungarian Geological Institute. Moreover *Claraia aurita* (Hauer) and *C.* gr. *aurita*

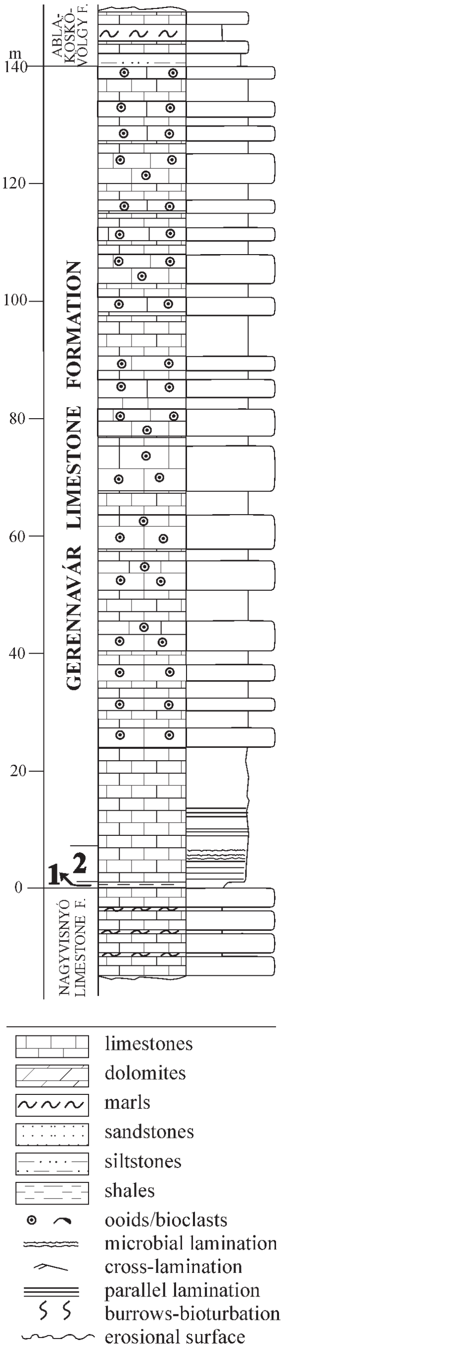

Fig. 3. Generalized log of the Gerennavár Limestone Formation in the Gerennavár section (after Péró 1983). **1** — 'Basal bedset'; **2** — 'Transitional bedset'.

(Fig. 5) occur in the upper part of the formation and in the transitional beds towards the overlying unit (Schréter 1935, 1953, 1954).

Biostratigraphic correlation and chronostratigraphy. Conodonts from the 'transitional bedset' refer to the H. latidentatus–C. meishanensis Zone (see Zhang et al. 1996). Between the 'transitional bedset' and the first occurrence of the advanced form of *Hindeodus parvus*, which does not occur before the Isarcica Zone (Kozur pers. commun.), there is no marker in the Bükk Mountains. On the basis of the conodont data, at least a ca. 55–60 m thick mudstone–oolite succession within the lower half of the formation, upwards from the appearance of *Hindeodus parvus* could be correlated to the Isarcica Zone. The lower boundary of the C. wangi-griesbachi Subzone, which is a part of the Claraia Zone and defined in the Dolomites (Broglio Loriga et al. 1983), coincides roughly with the appearance of the *Isarcicella isarcica* (Kozur 1985; Broglio Loriga et al. 1990). The uppermost part of the formation yields abundant *Claraia* gr. *aurita* allowing us to correlate these beds with the lower part of the C. aurita Subzone.

On the basis of the biostratigraphic results, the lowermost part ('basal' and 'transitional bedsets') of the formation represents the uppermost Permian (Changhsingian). The Permian/Triassic boundary could not be marked at present, further detailed studies are necessary for delineation of the boundary. The appearance of the advanced form of *Hindeodus parvus* and *Isarcicella isarcica* marks the beginning of the Upper Griesbachian, and that of the *Claraia aurita* assigns the beginning of the Dienerian (Nakazawa 1977; Broglio Loriga et al. 1990). There are several hints, which suggest that the formation comprises the entire Griesbachian and reaches up to the lower part of Dienerian (lower part of the Nammalian). These indirect pieces of evidence are 1) the sedimentology, that is continuous transitions are recorded between the different lithologies and facies, 2) the biostratigraphy, that is the ambiguous specimen of ?*Claraia clarai* (Emmrich) might record the C. clarai Subzone. Moreover, the Griesbachian/Nammalian boundary is detectable in the uppermost part of the formation, with the appearance of *Claraia aurita*.

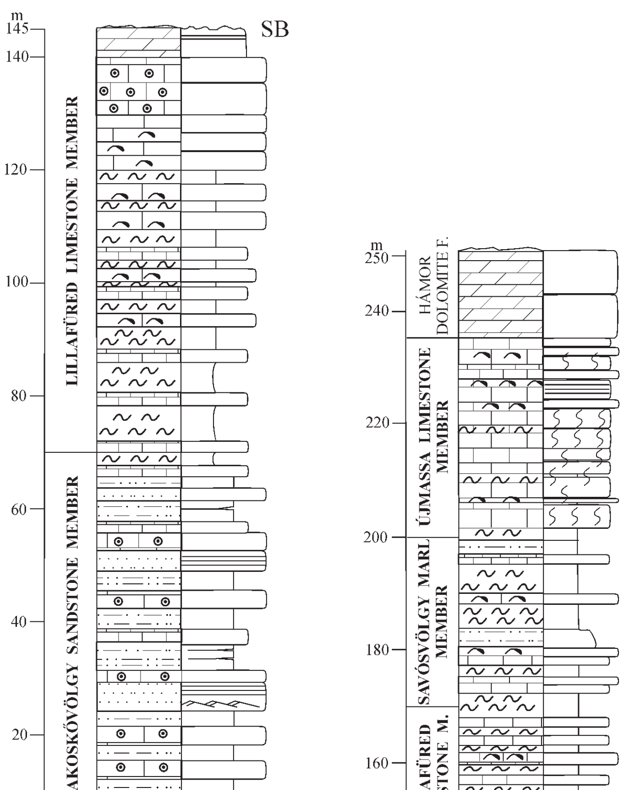

Fig. 4. Generalized log of the Ablakoskővölgy Formation in the Lillafüred section. Legend on Fig. 3. SB: a supposed sequence boundary.

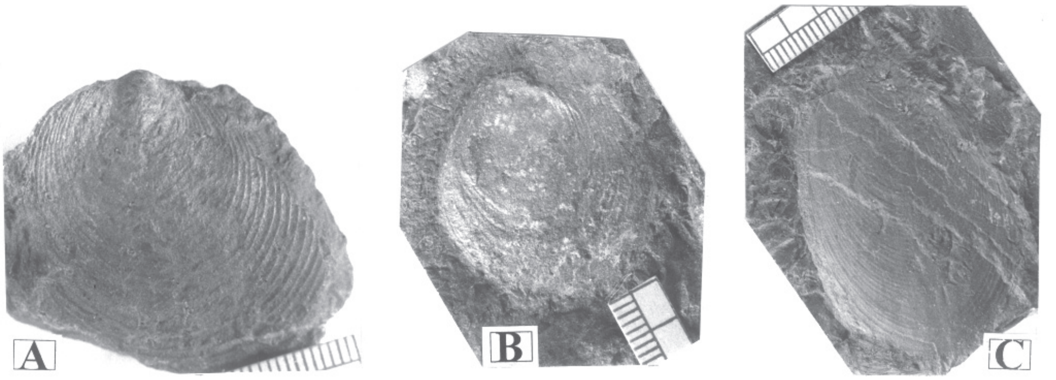

Fig. 5. Specimens of *Claraia* gr. *aurita* from the upper part of the Gerennavár Limestone Formation; left valves, **A)** and **B)** from Nekézseny, Bikkfolyás Valley (51 and 45 in the collection of the Hungarian Geological Institute (cHGI further on)), **C)** near Bánkút (cHGI 204). Scale bars have mm subdivision.

Ablakoskővölgy Formation

This formation is ca. 300 m thick, and composed by alternation of limestones, dolomites, and sandstones–shales. It can be divided into four members (Figs. 2 and 4), each displaying a typical lithological and facies development.

The *Ablakoskővölgy Sandstone Member* is defined by alternation of red and green sandstones, siltstones, and shales 40–100 m in thickness, but thin or thick intercalated limestone beds generally occur. There are gradual lithological changes both from the underlying formation, and towards the overlying member. The lowermost part of the member yields rich *Claraia* gr. *aurita* (Hauer) assemblages. *Eumorphotis* gr. *multiformis* (Bittner), which is incorporated within marly fine-crystalline limestones, and *Costatoria subrotunda* (Bittner) (Fig. 6), *Eumorphotis* cf. *hinnitidea* (Bittner), and *'Pseudomonotis' lóczyi* Bittner (Fig. 7), which is preserved within brownish sandstones–siltstones, are reviewed from the collections of Schréter (1935) and Balogh (1964). The lower part of this member corresponds to the uppermost Claraia Zone, that is the upper C. aurita Subzone on the basis of occurrence of the index fossil. Otherwise, this member contains index fossils of the lower part of the Eumorphotis Zone, that is the E. multiformis and E. hinnitidea Subzones (see Broglio Loriga & Posenato 1986; Broglio Loriga et al. 1990). According to the biostratigraphic data this member is Nammalian in age ranging from the lower Dienerian to the top of the Smithian (see Broglio Loriga et al. 1990). The C. aurita Subzone is attributed to the Dienerian, and the E. multiformis–E. hinnitidea Subzones together have been indirectly referred to the Nammalian, because they are located below the Tirolites cassianus Beds (Broglio Loriga et al. 1990).

The *Lillafüred Limestone Member* is up to 150 m in thickness and composed of grey limestones dissected by greenish, yellowish grey marl intercalations. Schréter (1935, 1954) and Balogh (1964) mentioned *Natiria costata* (Münster) (Fig. 8A) from many outcrops from the area of the Northern anticline. According to the revision of the fauna, the occurrence of *Eumorphotis kittli* (Bittner) shows that (at least the lower part of) the member belongs to the E. kittli Subzone (see Broglio Loriga et al. 1990), which suggests Spathian age for this member. Schréter (1935) collected *Tirolites* specimens from the Bogdány-tető location. He described the fossil assemblages partly from brownish limestones (identified as Lillafüred Limestone) and partly from marls with thin limestone beds (identified as Savósvölgy Marl), both with '*Tirolites cassianus* (Quenstedt)'. But specimens in the collection are not differentiated according to two lithologies (to two lithostratigraphic units), since he recognized and mapped only one 'Campil' Member. Thus, after many years these two *Tirolites* groups could not be separated. It is supposed that a pile of fragments, and a couple of whole specimens, *Tirolites illyricus* Mojsisovics (Fig. 9A) were collected from the Lillafüred Limestone Member. None of them could be identified as *Tirolites cas-*

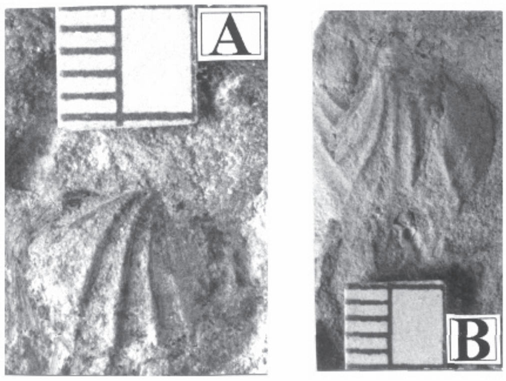

Fig. 6. *Costatoria subrotunda* (Bittner) from Ablakoskővölgy Sandstone Member. **A** — right valve, Mályinka, southwest of Bogdány-tető 424 (cHGI 74); **B** — left valve, Mályinka, from the topmost bed south of Bogdány-tető 424 (cHGI 75). Scale bars have mm subdivision.

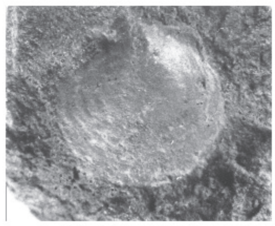

Fig. 7. *'Pseudomonotis' lóczyi* Bittner, left valve, from Ablakoskővölgy Sandstone Member, Mályinka, east of Bogdány-tető (cHGI 68). M: 1.5×.

sianus (Quenstedt) in accordance with the taxonomical study by Posenato (1992).

The *Savósvölgy Marl Member* consists predominantly of grey, greenish grey shales, clayey marls, and marls, 40–150 m in thickness. Thin micritic limestone intercalations are very characteristic for the entire member. Sand-rich deposits are recorded from the western part of the Northern anticline. A relatively rich gastropod fauna is reported by Schréter (1935) and Balogh (1964) including *Natiria costata* (Münster), *Naticella subtilistriata* (Frech), and '*Turbo*' *rectecostatus* Hauer. *Costatoria costata* (Zenker) was found from the thin-bedded limestones–marls (Schréter 1935; Balogh 1964). In the collection of the Hungarian Geological Institute, *Tirolites seminudus* Mojsisovics (Fig. 9B), and *Dinarites* sp.? are supposed to have been collected from this member from the Bogdány-tető location (see Schréter 1935). According to the study by Posenato (1992) these above-mentioned species represent the early and middle stages of the tirolitids phylogenetic trend belonging to the T. cassianus Zone proposed by Krystyn (1974). On the basis of the occurrence of *Tirolites* and *Costatoria costata*, this member also belongs to the middle part of the Spathian.

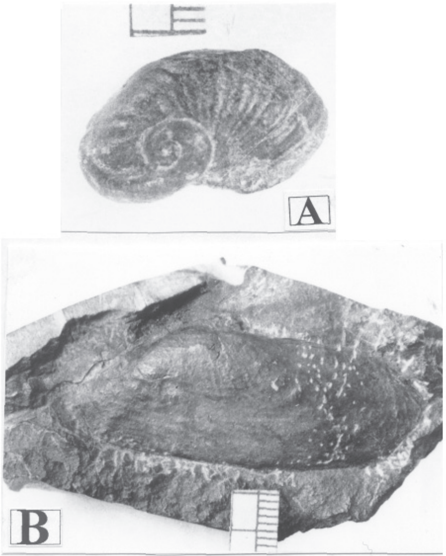

Fig. 8. A — *Natiria costata* (Münster) from the Lillafüred section (cHGI 152); **B** — *Unionites canalensis* (Cat.) left valve, from the Lillafüred section (cHGI 160). Scale bars have mm subdivision.

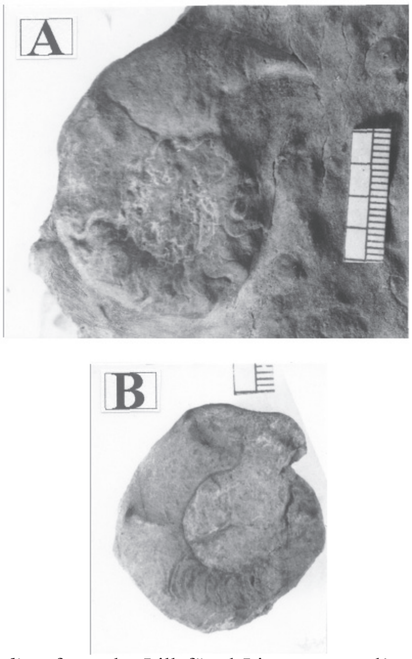

Fig. 9. *Tirolites* from the Lillafüred Limestone and/or Savósvölgy Marl Members; **A** — *Tirolites illyricus* Mojsisovics, Mályinka, east of Bogdány-tető 424 (cHGI 93); **B** — *Tirolites seminudus* Mojsisovics, Mályinka, west of Bogdány-tető 424 (cHGI 82). Scale bars have mm subdivision.

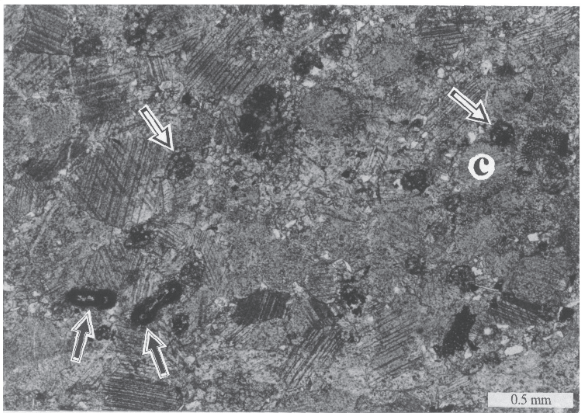

Fig. 10. Microfacies of crinoidal tempestite layer in the Újmassa Limestone Member, Lillafüred section — packstones with crinoids (C), and co-occurrence of *Meandrospira pusilla* (Ho) (white arrows) and *Cyclogyra? mahajeri* Brönnimann, Zaninetti et Bozorgnia (black arrows). Sample: 84.

The *Újmassa Limestone Member* is composed of dark grey platy and nodular, strongly bioturbated limestones with marl and shale intercalations or flasers. In its upper half, the bioclastic limestones often alternate with laminated mudstones. Dolomite lenses, or layers also occur. Its maximum thickness is ca. 60 m. *Rectocornuspira kalhori* Brönnimann, Zaninetti et Bozorgnia and *Cyclogyra? mahajeri* Brönnimann, Zaninetti et Bozorgnia were recently recognized in the Bükk Mountains (Fig. 10). On the basis of the occurrence of *Costatoria costata* (Zenker), *Meandrospira pusilla* (Ho), this member corresponds to the upper part of the C. costata Zone (Broglio Loriga et al. 1990), and its age is upper Spathian (uppermost Olenekian). There is a gradual lithological and facies transition towards the overlying Hámor Dolomite Formation, which is regarded to be Anisian in age. However, similarly to many other Western Tethyan sections, there is no evidence for the Scythian/Anisian boundary in the Bükk Mountains.

Glomospira sinensis Ho, *Glomospirella shengi* Ho, and *Meandrospira pusilla* (Ho) are the most frequent foraminifers in the carbonate part of the Ablakoskővölgy Formation (Pelikán 1995). They refer to Spathian age. Schréter (1935, 1954) and Balogh (1964) collected many Alpine faunal elements from outcrops hardly identifiable at present. Most of them have no chronostratigraphic value within the Lower Triassic, like *Unionites canalensis* (Catullo) (Fig. 8B), *U. fassaensis* (Wissmann), *Bakevellia* sp., '*Pecten*' sp., *Neoschizodus laevigatus* (Ziethen).

Sedimentology

Gerennavár Limestone Formation

The lowermost part of the formation (ca. 7–8 m) is different from the bulk of the formation, especially the lowermost 1 m (Fig. 3). Clayey marls with fine siliciclastic limestone layers and sandstones of the 'basal bedset' were deposited with a sharp lithological change from the underlying carbonate succession. The 'transitional bedset' is composed of thin-bedded dark grey limestones (ca. 6–7 m) characterized by alternations of microbial lamination and fine even lamination, with silt and sand-sized biodetritus sometimes normally graded (Figs. 11 and 12).

The predominant part of the formation consists of two types of limestone: 1) thin- to thick-bedded, dark grey laminated or massive finely crystalline limestones (mudstones, occasionally wackestones); 2) thick-bedded well-sorted oolites (Fig. 13), or pure bioclastic grainstones, which are mainly crinoidal grainstones (Figs. 14–15). The wackestones are composed of recrystallized microsparitic matrix with fragments of ostracods and bivalves. The thick-bedded oolites are massive, and there is no visible structure inside the beds. The lack of cross-bedding could be explained by well-sorting of the ooid sand, and absence of micrite.

The lower part of the succession is composed of mudstones–wackestones, whereas oolite grainstones first appear ca. 20 m above the top of the 'transitional bedset'. The oolite beds become predominant upsection. There is a short, gradual transition with increasing siliciclastic interlayers towards the overlying formation.

The oolite beds contain ooids, which display different degrees of diagenetic alterations. The following ooid-types can be recognized on the basis of their microfabrics. The first type contains typical coated grains with multiple tangential concentric laminae (Fig. 13.1). This type represents the initial depositional form. However, the nuclei of these ooids are generally already neomorphosed. The second type of rounded grains are fully micritized, thus, they are regarded as peloids (Fig. 13.2). Grains of the third type, which are recrystallized into microspars, have rounded surfaces, but no internal structures (Fig. 13.3). They preserve remnants of micrite envelopes, or micritized patches. Dolomitized grains, the fourth type, display only the hollows of the ooids (Fig. 16). In some layers and lenses sand-sized grains have a characteristic yellowish color. The coloration could be due to the weathering of dolomitized ooids and bioclasts originally most probably composed of aragonite and/or high Mg-calcite. Otherwise, the other types of ooid (types of 1–3) are typically grey.

The above-described types represent alteration stages of the diagenesis. In the first phase micritization took place, and resulted in rounded peloids. The micritization is regarded as a very early, syndepositional diagenetic processes accomplished by micro-organisms (Bathurst 1971; Reid & Macintyre 2000). In the second phase most probably the micritized grains and the nuclei of the ooids were recrystallized. The triggering processes are interpreted as aggrading neomorphism which took place during burial diagenesis (see Bathurst 1971).

Dolomitization occurred in the burial diagenetic realm and represents the third phase of the diagenetic alterations. The burial diagenesis is indicated by the petrography — the coarse crystalline baroque dolomite (Fig. 16) was formed by fabric selective and fabric destructive replacive dolomitization.

Additionally, the pseudo-oolite type (cortoid) of the grainstones is very common consisting of rounded bioclasts, mostly echinoderm and mollusc fragments (0.2–10 mm), whereas each grain is coated by a thin micritic envelope (Fig. 15). These are not strictly coated grains. These micritic crusts represent alteration of the grain surfaces by boring micro-organisms. In the crinoidal limestones degrading neomorphism, as a result of burial diagenesis, is recorded (Fig. 15).

Facies interpretation. The lithological change at the base of the formation most probably also reflects the fundamental en-

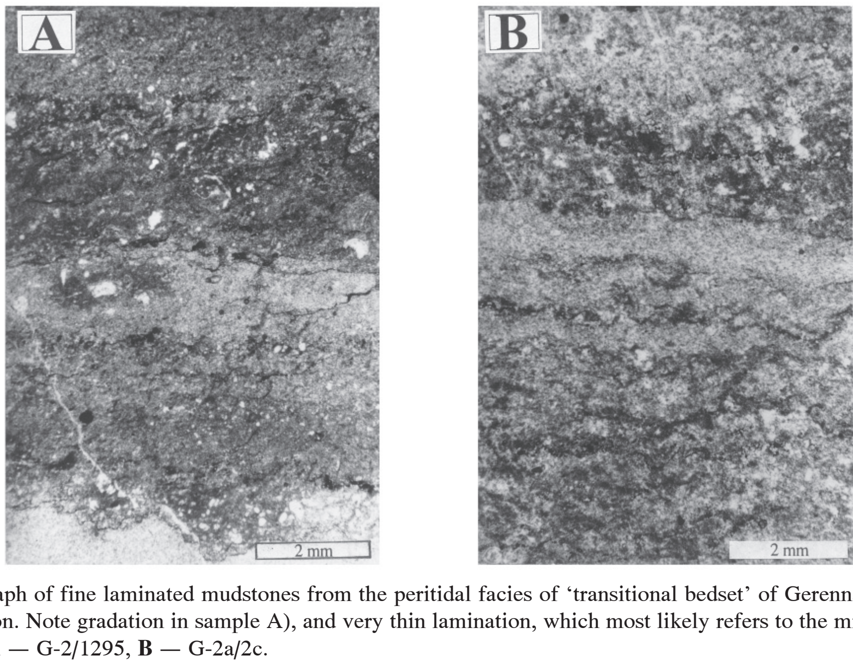

Fig. 11. Photomicrograph of fine laminated mudstones from the peritidal facies of 'transitional bedset' of Gerennavár Limestone Formation, Gerennavár section. Note gradation in sample A), and very thin lamination, which most likely refers to the microbial micritization in sample B). Samples: **A** — G-2/1295, **B** — G-2a/2c.

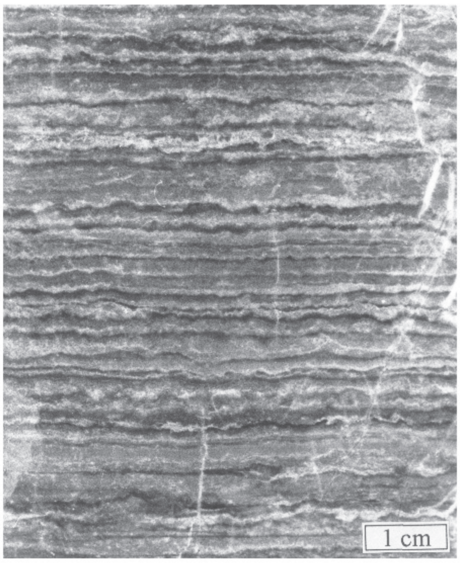

Fig. 12. Thin, irregular lamination in microbial mat, peritidal facies in the upper part of the 'transitional bedset' of Gerennavár Limestone Formation, Bálvány section.

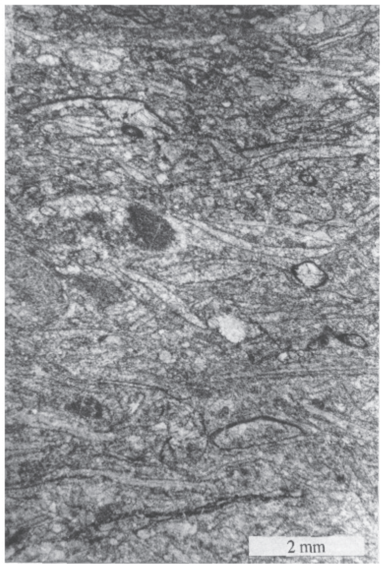

Fig. 14. Photomicrograph of bioclastic grainstones with mollusc shell fragments, Gerennavár Limestone Formation, Gerennavár section, sample G-2a/10a.

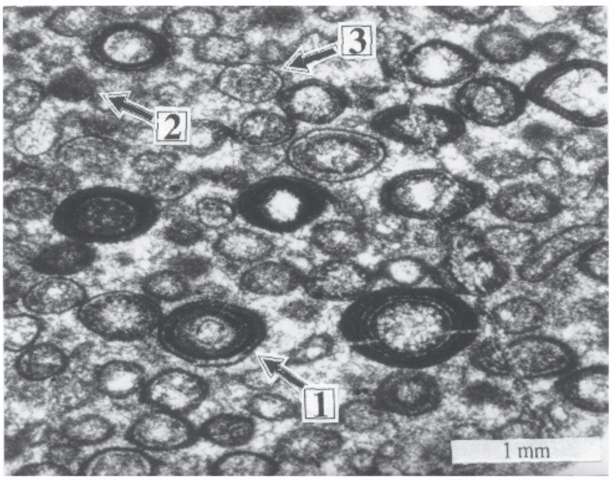

Fig. 13. Typical microfacies of oolite grainstones from the Gerennavár Limestone Formation with the types of ooids: **1** — ooid with recrystallized nucleus and a number of concentric, thin micrite crusts, **2** — micritized ooid, **3** — entirely recrystallized ooid with only one micrite envelope. Note the elongation of ooids according to the shear stress. Sample: G-1/19.

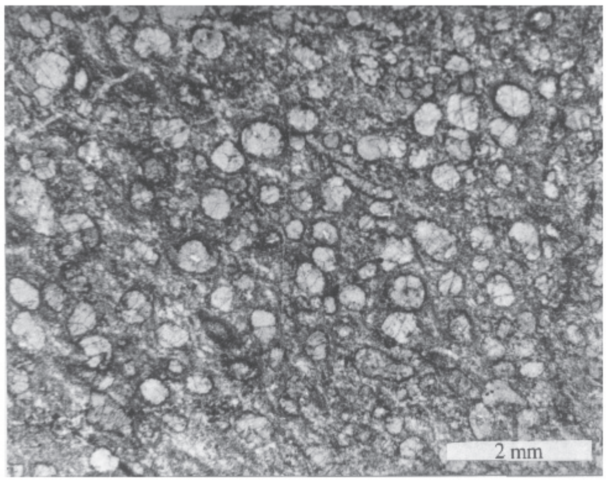

Fig. 15. Photomicrograph of packstones with recrystallized, rounded crinoid fragments which preserved their thin micrite envelopes, Gerennavár Limestone Formation, Gerennavár section, sample G-1/15. Note the result of degrading neomorphism (burial diagenesis). This means that the crinoid fragments are composed of not only one but a few spars.

vironmental change during the latest Permian. Thick laminites of the 'transitional bedset' were probably deposited in a restricted, peritidal environment. Crinkled, discontinuous, thin stromatolite lamination refers to microbial colonization. Graded planar laminae indicate deposition from waning-flow, or storm currents. Weak bioturbation is most likely due to the changeable environmental conditions. Summarizing, the 'transitional bedset' suggests a highly stressed environment.

Formation of ooids took place in a shallow subtidal zone, where the energy of water was sufficiently high to keep the sand grains in permanent movement. Studies of modern environments (summary e.g. in Halley et al. 1983; Tucker & Wright 1990), have shown that oolite sand bodies are associated with bank, platform or shelf margins, which are either tide-dominated or windward open, storm- and tide-dominated. The recent analogies suggest that the Lower Triassic oolite shoals were initiated on topographic highs whereas skeletal sand bodies were deposited at first, since bioclastic grainstone layers underlie the oolite succession.

Thick successions of oolite beds suggest amalgamation of the shoal lobes in the major part of the Gerennavár Limestone Formation. In sand belts, grains in the shield zone are commonly reworked to wide fans, which migrate towards the 'platform interior' during storms (e.g. Lilly Bank, see Ball

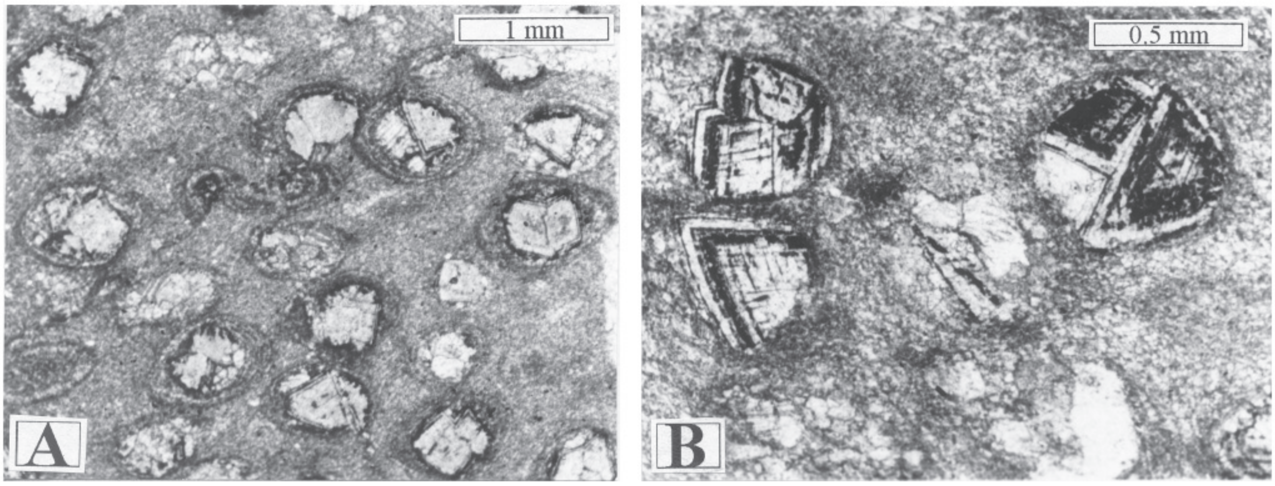

Fig. 16. Photomicrograph of dolomitized ooids in microspar matrix. Late diagenetic baroque dolomite partly, or fully replaced the ooids. Note zonation and curved crystal surface of the baroque dolomite (e.g. B — lower left), and the preserved round surface of the host grains (B — two uppers). Gerennavár Limestone Formation, Gerennavár section, samples: **A** — G-1/5, **B** — G-1/22b.

1967). Another example of oolite formation is the Joulter's Cays, where the ooids were deposited in lobe-shape fans developed at the ends of tidal channels, through which the ooids were transported from the shoal, and at bank spillovers (Harris 1979). The oolite deposits of the Gerennavár Limestone Formation could be compared to the above mentioned oolite sand bodies. Thin-bedded mudstone interlayers within the thick-bedded oolite grainstones are interpreted as deposits of major storms similar to those described from the Bahamas by Shinn et al. (1993), and by Major et al. (1996). The cortoid containing beds may have formed shallower flanks of the oolite shoal, where microbes altered the grain surfaces during calm periods.

The interlayering mudstones–wackestones in the lower and upper part of the formation most probably represent a low-energy subtidal facies. The facies of the beds, where scattered ooids embedded within micrite matrix in the upper part of the formation, could be interpreted as shallow subtidal stabilized sand flat bankward to the shoal. Its mixed muddy peloidal–ooidic sand deposits and facies could be compared to the modern Joulter's area, in the Bahamas (see Harris 1979).

The poor fossil record of this formation could be mainly the result of the depopulation effect of the Permian/Triassic extinction, and partly due to the disadvantageous conditions for the population in the niche of a high-energy mobile sand belt. Otherwise, the preservation potential must also have been low in this environment because of the disintegration and fragmentation of skeletons. This could explain the missing marker of the Claraia clarai Subzone.

In summary, the following vertical facies changes are recorded in the succession of the formation: a restricted very shallow marine environment at the base of the formation (peritidal), which evolves upwards into a low-energy shallow subtidal (lagoon), and then into a high-energy intertidal–shallow subtidal (oolite shoal) environment. At the upper part, the mixing of the muddy and peloidal–oolitic sand refers to decreasing water energy (stabilized sand flat).

Ablakoskővölgy Formation

Ablakoskővölgy Sandstone Member

Fine siliciclastics prevail with intercalations of carbonate layers (Fig. 4). The following litho-types are observed in the siliciclastic bedsets (not in stratigraphic order): 1) reddish brown shales; 2) alternating greyish green, rarely brown, parallel laminated clayey or carbonate siltstones and silty marls; 3) alternating greyish green sand-streaked siltstones, and parallel, or sometimes cross-laminated thin-bedded, mica-rich sandstones. The mature sandstones contain predominantly detrital quartz grains and additionally detrital feldspar up to 600 μm in size. Clay minerals are illite, and rarely kaolinite. Because of low-grade metamorphism and strong schistosity, the original sedimentary structures of this member are hardly recognizable.

The thick-bedded limestones (wackestones–packstones) contain red ooids or fragments of ooids and detrital quartz grains. This lithofacies is characterized by reddish staining due to the high Fe-content. The thin-bedded grey or beige limestones are typically mudstones, generally recrystallized microspar with much less siliciclastics, and often with juvenile pelecypod shells. The laminated or thin-bedded mudstones typically alternate with greenish grey calcareous siltstones.

An upward thickening and coarsening stacking pattern characterizes the major part of this member. However, an upward fining trend and darkening of colors becomes prevalent in the upper part and the transitional interval towards the overlying carbonate member. With upwardly increasing carbonate content this member continuously evolves into the overlying carbonate member.

Facies interpretation. Sediment deposition most likely took place mainly in the coastal, shoreface and transitional zones of a shallow shelf. In the siliciclastic succession, the oolite grainstones are intercalated as distinct thick beds, which refers to the reworking of ooids most probably by storms. They were either fed by temporal, patchily distributed shoals, or by thin tidal delta lobes. However, the *in situ* cross-bedded deposits of the oolite shoal or tidal delta were not recognized.

Lillafüred Limestone Member

This member can be subdivided into three parts (Fig. 4). The lower part shows a gradual transition from the underlying member with upwardly increasing carbonate content. The bulk of the lower part is composed of dark grey, thin-bedded, partly thick-bedded finely crystalline limestones (mudstones–wackestones) and marly bedsets. The latter are either dark grey calcareous marls or greenish brown clayey siltstones occasionally with thin platy mudstones. Lenses, or thin- to thick-bedded bioclastic limestone beds, generally crinoidal wackestones–packstones, are more abundant upwards. A thickening-upward trend up to the middle part of the member is recognized.

The middle part of the member consists of upward thickening and coarsening bedsets. In the basal part dark grey laminated or thin-bedded mudstones alternate with thick-bedded crinoidal limestones separated by marl layers. Thick-bedded, well rounded and well sorted crinoidal–oolite limestones (grainstones–packstones) occur upsection (Fig. 17). The sequence is terminated by a dolomite cap consisting of massive, finely crystalline beds with occasionally preserved thin irregular lamination (dolocrete crusts). Its top is truncated by an erosional surface.

Above the erosional surface, dark grey crinoidal limestones with intraclasts and peloids occur which form the base of the upper part of the member (Fig. 18). Upsection, upward fining and thinning bioclastic, crinoidal limestones (packstones), and mudstones with an upwardly increasing number of marl intercalations are characteristic. Crinoidal limestones (packstones) occasionally exhibit ripples on the bedding surfaces (Fig. 19).

Facies interpretation. The lower part of the member was deposited in the deeper subtidal zone, mostly below the storm wave-base indicated by the fine deposits and the dominant thin bedding. The thinning- and fining-upward stacking pattern of the basal transitional and lower parts may refer to a relative sea-level rise. The crinoidal wackestones–packstones represent storm events between fairweather and storm wave-base which are generally rare in the lower part and become more frequent upwards. Thin lenses composed of washed crinoidal fragments suggest patchily distributed thin storm veneers.

Shoals consisting of washed and well-rounded crinoidal and ooid sand represent deposition around the fairweather wave-base. Their position in the succession reflects highstand progradation. Mudstones above the oolites were probably deposited in a shallow peritidal zone and subjected to dolomitization afterwards. Thin uneven, laminated dolocrete crusts inside the dolomite body most probably indicate subaerial exposures. Early diagenetic near surface dolomitization by relative sea-level drop is indicated by pervasive finely crystalline stratiform dolomites and the associated erosional surface on top of the succession. In the upper part of the member, a

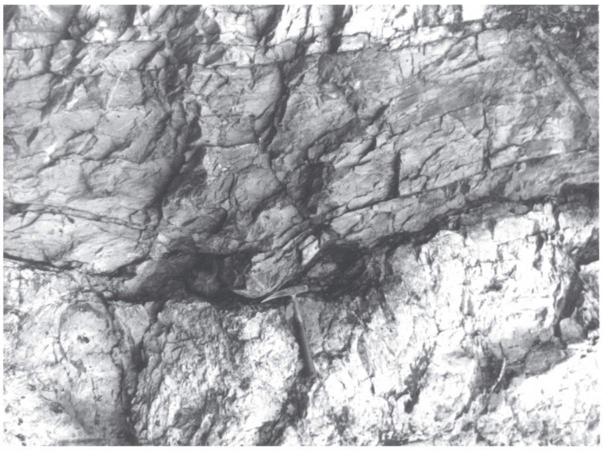

Fig. 18. Sharp surface (at the head of hammer) truncating the massive, light grey dolomite succession, peritidal facies, is overlain by dark grey intraclastic crinoidal limestones, shoal facies, in about the mid-part of the Lillafüred Limestone Member, Lillafüred section (overturned section but the photograph is rotated into the original depositional position). Hammer is 33 cm long.

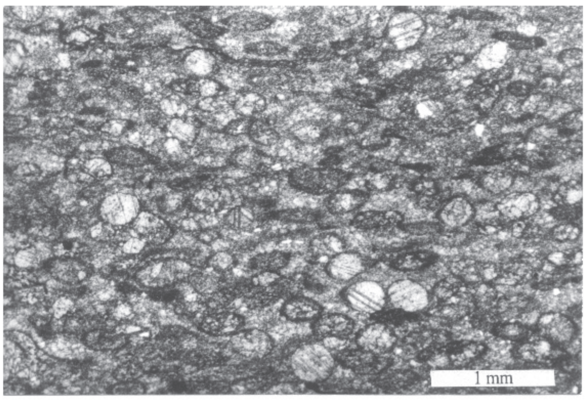

Fig. 17. Photomicrograph of crinoidal–oolite limestones, shoal facies, from the mid part of Lillafüred Limestone Member, Lillafüred section. Sample: 55.

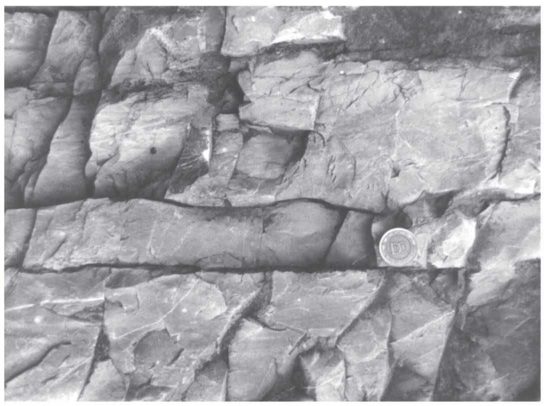

Fig. 19. Ripples in crinoidal limestones, shoals facies, in upper part of the Lillafüred Limestone Member, Lillafüred section. Coin is 2.5 cm in diameter.

thinning- and fining-upward succession of crinoidal limestones and mudstones–marls recorded an upwardly deepening trend from shallow subtidal below the storm wave-base. This is most likely due to a relative sea-level rise.

In summary, the vertical stacking pattern of the facies in this member exhibit a deepening-, followed by a shallowing-, and then a deepening-upward trend. Considering the thickness of the deepening–shallowing-upward succession (ca. 75 m) in the lower part of the member, this trend most likely suggests third-order relative sea-level changes. A sequence boundary is recognized on top of the shallowing-upward cycle, that is on top of the finely crystalline massive dolomites, approximately in the middle part of this member (Fig. 4).

Savósvölgy Marl Member

This member is characterized by monotonous grey, greenish grey shales, clayey marls, and marls (containing gastropods and ammonites) with frequent, thin platy mudstone intercalations (Fig. 4). Sandier development was reported in the western part of the Northern anticline.

Facies interpretation. Lack of wave- and storm-induced structures, and the relatively rich ammonite fauna in the succession suggest that the deposition took place in an open shelf environment, most probably below the storm-wave base. It is an open question what was the control of terrestrial siliciclastic input.

Újmassa Limestone Member

This member (Fig. 4) is characterized by platy, thin-bedded (1–4 cm), dark grey nodular limestones, mudstones–wackestones–packstones (Fig. 10). Their bedding surfaces are often covered by clayey marl flasers (Fig. 20). The intensity of bioturbation varies from minor individual burrows to strong bioturbation, and related nodular structure. The burrow-mottled pattern resembles *Thalassinoides*-type burrows. Beds with only minor bioturbation preserved their original depositional structures, for example normal gradation in thin skeletal lenses or layers, or faint planar lamination becoming more common upwards. Dolomite lenses, or layers appear at the transition zone to the overlying formation.

Facies interpretation. The dark colour of the rocks and faint planar lamination refer to partial restriction near the bottom. Lack of wave structures and common graded bedding indicates that this succession was composed of a series of distal tempestites deposited within a muddy sequence below the storm wave-base.

Facies model and correlation of sedimentary units

The Alpine sedimentary history of the Dinarides–Alps–Carpathians started in the middle Permian with terrigenous clastics of alluvial plains. First marine ingressions reached the southeastern domains (according to the recent co-ordinates) of the Western Tethys, and cyclic sabkha, and then subtidal open-marine ramp carbonates of ‘Bellerophon-type’ formations were deposited in the late Permian (e.g. Nagyvisnyó Limestone Formation in the Bükk Mountains). A renewed long-term sea-level rise shifted the coastline further west–northwest, and the coastal zone of the shallow sea reached the western part of the Northern Calcareous Alps in the late Scythian (Gwinner 1971; Tollmann 1976; Mostler & Rossner 1984; Broglio Loriga et al. 1990; Krainer 1993; Rüffer & Zühlke 1995).

Fig. 20. Dark grey platy, nodular limestones, restricted outer ramp facies of the Újmassa Limestone Member, Lillafüred section (overturned section but the photograph is rotated into the original depositional position). Coin is 2.5 cm in diameter.

The distribution of carbonate and siliciclastic sediments temporally and laterally changed on the extensive shallow seafloor of the Western Tethys. On a large-scale a general lateral trend can be recognized in the decreasing amount and grain size of siliciclastics from the coastal areas towards the shelf interior. However, the temporal distribution exhibited different trends in different facies domains, but generally the siliciclastic influx intensified during the Dienerian and Smithian in many depositional areas.

According to the Paleozoic-Mesozoic paleogeographic reconstruction the depositional area of the Bükk Mountains was in the northwestern neighbourhood of the Inner Dinarides, next to that of the Jadar Block (Yugoslavia), and the Sana–Una Unit (Bosnia-Herzegovina) (Protić et al. 2000). During the Early Triassic very similar sequences were deposited in all three units, which is reflected in very uniform lithological and facies development of time equivalent formations. These three units together composed a uniform facies domain during the Early Triassic.

In the depositional area of the Bükk Mountains, siliciclastic sedimentation prevailed during the Nammalian, otherwise carbonates deposited. The disposition of sedimentary environments was controlled by the position of the fairweather and storm wave-base. On the basis of the vertical hierarchy of lithofacies, three characteristic parts can be distinguished. They are represented by the Gerennavár Limestone, and the siliciclastic and carbonate parts of the Ablakoskővölgy Formation. On the basis of grain types such as ooids, and early diagenetic processes (dolomitization) the sediments may have been generally deposited under arid–semi-arid climate. Controls of the periodic siliciclastic input might have been either the climate (temporal humid periods) or tectonic uplifting of the hinterland.

The global absence of reef-builders is a general characteristic of the Lower Triassic sequences (Heckel 1974; James 1984; Flügel 1982). The faunal assemblage is generally characterized by cosmopolitan, simple, opportunistic forms, which belong to the heterozoan association (a term of James 1997), and reflects a mass extinction aftermath (Schubert & Bottjer 1995). They indicate that stressful environmental conditions persisted during Scythian time.

Gerennavár Limestone Formation

Peritidal microbial and fine evenly laminated mudstones, mudstones–wackestones of a shallow subtidal lagoon environment, and bioclastic-oolite limestones of a high-energy subtidal zone characterize the Gerennavár Limestone. The ooids were formed in the high-energy, permanently agitated surge-zone and transported into shallower (stabilized sand flat) environments. The ooid-dominated ramps in general probably reflect low rates of biological carbonate production in shallow-water environments, which were widespread in periods when shallow-water framework reefs were globally absent or scarce as during the Mississippian and Jurassic (Wright & Faulkner 1990; Burchette & Wright 1992).

The depositional interval of the Gerennavár Limestone Formation can be correlated to that of the Tesero Horizon, Mazzin Member, Andraz Horizon, and about the lower half of the Siusi Member in the Dolomites (see Broglio Loriga et al. 1983). The differences are more obvious than the similarities in comparison of the two successions. The uniform development of mudstones and ooidic grainstones is a characteristic feature of the succession of the Gerennavár Limestone in the Bükk Mountains, whereas alternations of variable limestones, early diagenetic dolomites, marls, and fine siliciclastics, make up the succession in the Dolomites. The lithology of the latter one is more marly and rich in detrital clay and silt.

Similarities can hardly be found in the lower units of the two areas, which overlie the Bellerophon-type upper Permian carbonates, that is siliciclastics of the 'basal bedset' and fine evenly and microbially laminated mudstones (Bükk Mountains) and oolite grainstones–packstones of the Tesero Horizon, or offshore mudstones in the Cadore, Comelico, and Southwest Carnia areas (Dolomites).

Mostly muddy lithology characterizes the time equivalent succeeding parts of the sequences, that is in the Gerennavár Limestone above the 'transitional bedset', up to the first appearance of ooids (ca. 20 m in thickness)(see Fig. 3), which can be correlated with the lower part of the Mazzin Member. However, the sedimentary structures and their facies are different, since deposition occurred in low-energy environments, in a lagoon in the area of the Bükk Mountains, but on a deeper subtidal shelf in the area of Dolomites. The upward succeeding time equivalent part of oolite limestones of the Gerennavár Limestone in the Isarcia Zone (ca. 45 m in thickness) and the upper part of the Mazzin Member are also rather different.

The facies developments of the overlying part in the sequences display great differences, as well. High-energy oolite shoal and stabilized sand flat is interpreted for the major part of the Gerennavár Limestone. Because of the highly monotonous development, trends were not recognized in the oolite sequence. This is partly due to the fact, that the system had a potential to accrete vertically and therefore the shallow marine environment could have persisted for a relatively longer time. There is no analogy for the Andraz Horizon for correlation in the sequence of the Bükk Mountains. However, the lower unit of the Siusi Member consists partly of oolites, but their colour and bedding is rather different. Lithology and facies development of the lower half of the Siusi Member is also quite different from the equivalent part of the Gerennavár Limestone Formation, except for the Trento and Valsugana area. In these paleohighs, the oolitic-bioclastic bodies several metres in thickness occur at various levels within the Siusi Member.

A 3rd order depositional sequence has been recognized within the uppermost Bellerophon Formation, the Tesero Horizon and the Mazzin Member, and the self margin wedge and the transgressive systems tract of another one in the Andraz Horizon and the lower part of the Siusi Member in the Dolomites (Broglio Loriga et al. 1983; Neri 1991; De Zanche et al. 1993). Facies changes within the lower and upper part of the Gerennavár Limestone Formation mark a shift between low-energy and high-energy environments, with no significant change in depositional depth. As a consequence the individual depositional sequences of the Dolomites cannot be correlated, which is most likely due to local controls dominated on sedimentation in the depositional area of the Bükk Mountains.

The south-easterly zones of the Northern Calcareous Alps and Inner Western Carpathians were reached by the sea during the latest Permian–earliest Scythian. The deposition on the shallow shelf, for example in the tidal flat, coastal zone and shallow shoreface zone, is controlled by the strong detrital siliciclastic input. Evaporite facies near the Permian/Triassic boundary, that is the Haselgebirge Formation in the Northern Calcareous Alps, Perkupa Evaporite Formation in the Silica Nappe, Inner Western Carpathians, and the succeeding siliciclastics, that is the Werfen Schist, and Bódvaszilas Sandstone Formations are rather different from the time equivalent Gerennavár Limestone Formations in the Bükk Mountains (see Tollmann 1976; Mostler & Rossner 1984; Kovács 1992; Hips 1996).

Ablakoskővölgy Sandstone Member

During the deposition of the Ablakoskővölgy Formation the terrigenous clastic input increased in two periods, in the Nammalian and mid-Spathian. Sediment supply was drastically changed during the Dienerian (in the Claraia aurita Zone) in the depositional area of the Bükk Mountains. Various marine sediments, clastics and carbonates, characterized many Western Tethyan depositional areas during the Nammalian, as in the upper part of the Siusi Member, Gastropod Oolite Member, and Campil Member in the Dolomites (Broglio Loriga et al. 1983; Broglio Loriga et al. 1990). The terrigenous input achieved a peak during Smithian defined as the 'Campil event' in the Dolomites (Italy) and in the Transdanubian Mid-Mountains (Hungary) (Broglio Loriga et al. 1990). In the depositional area of the Bükk Mountains the temporal distribution of siliciclastic supply was rather steady. Siliciclastic deposits unambiguously predominate over the carbonates, although limestone intercalations are recorded in the entire succession as well as in the Smithian (Ablakoskővölgy Sandstone Member). In the Outer Dinarides (see Aljinović 1995), Northern Calcareous Alps (see Mostler & Rossner 1984) and Inner Western Carpathians (see Bystrický 1964, 1973; Salaj et al. 1983; Hips 1996, 1998) the clastic input was continuously significant and inhibited carbonate production in the course of the early Scythian time.

The similarity all over the Western Tethyan depositional area is that the intense siliciclastic input seems to be terminated at the beginning of the Spathian (see Bystrický 1964, 1973; Salaj et al. 1983; Herak et al. 1983; Sćavničar & Šušnjara 1983; Mostler & Rossner 1984; Broglio Loriga et al. 1990; Michalík 1994; Aljinović 1995; Hips 1996, 1998). By late Scythian–earliest Anisian time the fine terrigenous components almost completely disappeared and consequently carbonates and evaporites were deposited as in the Hámor Dolomite Formation of the Bükk Mountains.

The time equivalent sedimentary units of the Ablakoskővölgy Sandstone Member are the upper part of the Siusi Member, Gastropod Oolite Member, and Campil Member in the Dolomites. They have generally similar lithological and facies developments with the consideration that the red oolite beds are distributed throughout the entire Ablakoskővölgy Sandstone Member, and they do not compose a distinct member as in the Dolomites (Gastropod Oolite). Depositional sequences recognized in the Dolomites, that is the highstand systems tract in the upper Siusi Member and lower Gastropod Oolite, and a transgressive systems tract and a highstand systems tract, a part of the succeeding sequence in the majority of Gastropod Oolite and Campil Members were not recorded in the Bükk Mountains.

From Lillafüred Limestone to Újmassa Limestone Member

The carbonate stage of the Ablakoskővölgy Formation from the Spathian clearly indicates predominantly lower energy depositional environments, whereas finely crystalline limestones and marls prevail in the deeper ramp setting. However, while distal storm layers are relatively abundant in intervals, deposits of the high-energy shallower environment as bioclastic beds and thicker beds of partly dolomitized oolite–crinoidal 'shoals–amalgamated storms sheets' appear only in a relative short interval in the middle part of the sequence. The lithological and facies characters of the Lillafüred Limestone correlate well to the Val Badia Member except for its basal peritidal horizon and uppermost cross-bedded sandstone bodies.

In the lower part of the Lillafüred Limestone, the vertical facies arrangement may refer to a relative sea-level rise. The striking transgression at the beginning of the Spathian with the appearance of *Tirolites* sp. was recorded not only in the whole Western Tethys, that is in the Southern Alps (Broglio Loriga et al. 1990), Balaton Highland (Budai & Haas 1997), Drau Range (Krainer 1987, 1993), Northern Calcareous Alps (Mostler & Rossner 1984), Carpathians (Patrulius et al. 1971; Hips 1998), Dinarides (Herak et al. 1983; Aljinović 1995), but also from the Peri-Tethyan areas (Bleahu et al. 1994), and in the western USA (Schubert & Bottjer 1995) and Western Canada Sedimentary Basin (Davies 1997).

The stacking pattern of the Lillafüred Limestone Formation, except for its lower part, displays a regressive–transgressive trend with a sequence boundary at the top of the finely crystalline dolomites approximately in the middle part of this member. Because of the lack of detailed biostratigraphy in the sequence of the Bükk Mountains, reliable correlation to the depositional sequences of the Dolomites (De Zanche et al. 1993; Rüffer & Zühlke 1995) is not possible.

The time equivalent formations in the uppermost part of the Scythian sequences exhibit major differences in terms of lithology and sedimentology between the Bükk Mountains and the Dolomites. Only a few similar features are recognized. Small-scale sequences consist of the Cencenighe Member in the Dolomites, with open shelf muddy basal part and tide-controlled oolitic-bioclastic calcarenites, and siltstones–marls of supratidal mud flat (Broglio Loriga et al. 1983, 1990). Whereas no small-scale sequences are recognized within the deep ramp mudstones–marls of the Savósvölgy Marl Member in the Bükk Mountains. On the one hand, the open shelf muddy basal parts of the small-scale sequences of the Cencenighe Member display a similar lithological character to the mudstones–marls of Savósvölgy Marl Member. On the other hand, the rest of the small-scale sequences of the Cencenighe Member are markedly different in every respect from the succession of the Savósvölgy Marl. Moreover, no correspondence between the peritidal varicoloured fine siliciclastics–dolomites of San Lucano Member (Dolomites) and the deep ramp distal tempestites–mudstones of the Újmassa Limestone Member (Bükk Mountains). Nevertheless, the Újmassa Limestone closely resemble the time equivalent Szinpetri Limestone Formation in the Silica Nappe of the Inner Western Carpathians (see Hips 1998; Koša & Janočko 1999) since both have the same bio- and lithofacies. According to Michalík (1994) the strong restriction of sea-water circulation between the extremely shallow seas in the depositional area of the Western Carpathians caused hyper-salinity and low oxygen regime during the Spathian/Anisian boundary time interval. A similar situation most likely occurred in the area of the Bükk Mountains.

Epeiric shelf and ramp models

On the basis of the comparison of time equivalent facies of the Dinaridic–Alpine–Carpathian depositional areas of Western Tethys (the above-mentioned examples) two similar facies models can be obtained for two intervals of the Early Triassic. They are on the one hand an epeiric shelf model (see Pratt & James 1986; Wright & Burchette 1996) for the Griesbachian–Nammalian, and on the other hand a carbonate ramp model (see Ahr 1973; Read 1985; Burchette & Wright 1992) for the Spathian.

The Western Tethyan depositional area was initiated as a ramp during the middle-late Permian, but evolved into an epeiric shelf during the Early Scythian, when the sea flooded more territories on the wide continental shelf (see Mostler & Rossner 1984; Krainer 1993; Rüffer & Zühlke 1995). As a result, most probably very shallow sea-water covered the extensive shelf area in the course of the Griesbachian and Nammalian. Storms and winds strongly influenced the circulation and deposition (see e.g. Aljinović 1995), but tidal activity was also recognized locally (e.g. in the formation of longer life oolite shoals, Bükk Mountains).

Although there are lithological variations in the characters of the sequences of the different depositional areas, which are partly carbonate, or pure siliciclastic in periods, but mainly mixed carbonate–siliciclastic, the successions are generally represented by either dominantly shallow subtidal facies, or shallowing-upward cycles from shallow subtidal to intertidal, and supratidal deposits. These cycles are well documented for example in the Outer Dinarides (Sćavničar & Šušnjara 1983; Aljinović 1995), and in the Dolomites (Broglio Loriga et al. 1983). These cycles appear to represent deposition on and around scattered, ephemeral, low-relief shoals whose shoreline prograded into the surrounding shallow subtidal areas to generate areally restricted shallowing-upward small-scale cycles. This type of epeiric platform model was described from the Ordovician by Pratt & James (1986).

Subsidence over such a large area was most probably not uniform. The different subsidence rate could be an explanation for the difference of time equivalent formations, for example between the Mazzin Member (Dolomites) and the oolites of Gerennavár Limestone. Additionally, the different local physical regime was most likely the other control factor, for example during mid-Griesbachian the low-energy environment in the facies domain of the Dolomites and high-energy, tide- and wave-dominated environment in the facies domain of the Bükk Mountains.

A prominent feature of the sequences is a characteristic deepening facies trend as a result of a relative sea-level rise at the beginning of the Spathian. As a consequence a deeper subtidal environment mostly below or around the storm wave-base was the site of deposition in the marine settings and further zones were reached by the sea. Generally this event was coupled with increase of carbonate content in the sediments and decrease of the grain size of detrital clastics. However, the northwestern margin of the basin is characterized by stronger terrigenous influx (Mostler & Rossner 1984). Different degrees of storm redeposition are widely recognized in the sequences, thus a major part of the shelf was storm-controlled (e.g. Broglio Loriga et al. 1990; Aljinović 1995; Hips 1998).

Generally in the Western Tethyan sequences, appearance of oolite bedsets above a deeper shelf facies reflects a shallowing-upward trend in the successions. These oolite shoals framed a marine basin extending to the northwest with normal salinity (model by Mostler & Rossner 1984). During the late Scythian, the sedimentation continued in the inner ramp zone with carbonates and partly evaporites in the depositional areas of the Dolomites, Northern Calcareous Alps, and Transdanubian Mid-Mountains of Hungary, whereas thin-bedded carbonates deposited in the deep ramp zone around or below the storm wave-base in the depositional areas of Bükk Mountains, Silica Nappe in Hungary and Slovakia, and part of the Outer Dinarides. Most probably the facies differentiation was controlled by the different subsidence. Appearance of the basinal red nodular limestones in the Hellenides (Chios section, Gaetani et al. 1992) indicated the early stage of a rifting (Mostler & Rossner 1984). As a consequence of the accelerated subsidence in some part of the Western Tethyan depositional area a gentle slope was generated.

Conclusions

Biostratigraphic correlation of the sequence in the Bükk Mountains, as in other Western Tethyan sections, is mostly based on molluscs, bivalves and partly on conodonts and foraminifers, and from the Spathian on ammonites. Several index fossils have been identified from the Bükk Mountains: *Hindeodus parvus* (Kozur et Pjatakova), *Isarcicella isarcica* (Huckriede), *Claraia aurita* (Hauer), *Eumorphotis* gr. *multiformis* (Bittner), *Eumorphotis* cf. *hinnitidea* (Bittner), *Costatoria subrotunda* (Bittner), *Eumorphotis kittli* (Bittner), *Tirolites illyricus* Mojsisovics, *Tirolites seminudus* Mojsisovics, *Costatoria costata* (Zenker), *Rectocornuspira kalhori* Brönnimann, Zaninetti et Bozorgnia, *Cyclogyra? mahajeri* Brönnimann, Zaninetti et Bozorgnia, and *Meandrospira pusilla* (Ho).

On the basis of the revised fossil-collections of Schréter (1935) and Balogh (1964), and conodont studies by Kozur (1985, 1989, 1996) the Lower Triassic sequence of the Bükk Mountains was subdivided on the basis of the biostratigraphic zonation proposed in the Dolomites by Broglio Loriga et al. (1983, 1990), and Yin et al. (1996), Zhang et al. (1996). The generally scanty occurrence of index fossils allowed only a basic comparison to the biozonation proposed in the Western Tethyan areas. Using this zonation a chronostratigraphic subdivision of the formations and members is assigned. The Gerennavár Limestone Formation contains the Isarcica conodont Zone (with advanced form of *Hindeodus parvus*) and C. aurita Subzone, (C. clarai Subzone is ambiguous because of scarce specimens), thus this formation is Griesbachian and lower Dienerian in age. In the Ablakoskővölgy Formation the Ablakoskővölgy Sandstone Member contains fossils of the C. aurita Subzone, and Eumorphotis gr. multiformis–E. hinnitidea Subzones (with *Costatoria subrotunda*), thus its age is Nammalian. In the Lillafüred Limestone and Savósvölgy Marl Members, ammonites of the Tirolites cassianus Zone are recognized, thus they are Spathian in age. The Újmassa Limestone Member contains *Costatoria costata*, *Rectocornuspira kalhori–Cyclogyra? mahajeri* and *Meandrospira pusilla*, which refer to the upper Spathian age.

The Lower Triassic succession can be subdivided into three stages. The first stage is represented by the Gerennavár Limestone Formation and reflected a) partly, in the lowermost part in low-energy peritidal and shallow subtidal lagoon facies, and b) predominantly in a typical high-energy, tide- and wave-dominated sand belt and shallow subtidal sand flat. The second stage includes the siliciclastic, lower part of the Ablakoskővölgy Formation. Since the terrigenous clastic supply increased in this period this is a mixed, but predominantly siliciclastic shallow shelf with sedimentation in the coastal, shoreface and transitional zones. The third stage is represented by the carbonate, upper part of the Ablakoskővölgy Formation, and reflected in a storm-controlled sedimentation in the whole spectrum of environments from peritidal to deeper subtidal below the storm wave-base.

Although, the sedimentation of the 'Werfen Formation' in the Western Tethyan area occurred generally on a gently sloping shallow, epeiric shelf, and partly on a ramp at the end of the Scythian, the deposits and the facies can be very different in the time equivalent sequences. It suggests the importance of local control factors in sedimentation, that is terrigenous siliciclastic input, different subsidence rate, antecedent topography, and dominance of local physical regime. The Lower Triassic succession in the Bükk Mountains, in term of sedimentology, displays many lithological and facies differences relative to that of the Dolomites. Nevertheless, close similarities can only be documented in short intervals between sequences of different paleogeographical position.

Acknowledgments: The first author thanks Gy. Less for his guidance in the field. We appreciate the encouragement from J. Haas and S. Kovács by means of which the first Hungarian version of this review was written. The paper benefited from critical comments of the GC reviewers, R. Posenato, J. Michalík and an anonymous reviewer. We are grateful to M. Pellérdyné and Cs. Péró for the photographs. The fieldwork was partly sponsored by the Hungarian Scientific Research Fund (OTKA) No. T 037966 and T 037595. We are indebted to the Bükk National Park for supporting our studies in the mountains.

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