Geology of the pre-Cenozoic basement of Hungary

Explanatory notes for
"Pre-Cenozoic geological map of Hungary"
(1:500 000)

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Introduction

The Carpathian Basin is a specific geo-region; it is a young basin system filled with sedimentary and volcanic rocks, formed within the Alpine Mountain System. Geophysical and geological research over the last few decades has revealed that the basement of the basin system is built up of lithosphere fragments of different origin. These reached their current position in the Early Neogene, right before the development of the basin system. In order to see the geological relations between the basins and the surrounding mountain range clearly, as well as to understand the geological history of the region, the basement must be surveyed by mapping; this means synthesizing the geological and geophysical data regarding the basement. Up until the 1960s there were mostly only assumptions concerning the build-up of the basement; different authors published maps reflecting different concepts on the structural geology (Vadász 1960, Schmidt 1961, Szalai 1961, Horusitzky 1961). The Geological Institute of Hungary published the first basement map to a 1:500 000-scale, in 1967 (Csalagovits et al. 1967). Two decades later, in 1987, this was followed by the publication of a basement map for which much more data was available (Fülöp et al. 1987). The tectonic version of this map was completed in 1990 (Dank et al. 1990), and a few years later the explanatory book was also published for both maps (Haas ed. 1996).

Following the 1990s, the geological research which aimed to survey the basement of the basin system proceeded. Geophysical surveys applying increasingly modern methods were conducted on large areas and the rapid evolution of computer methods made it possible to process old and new measurements thus producing results of a superior quality. New geological mapping programs refined previous knowledge; several new national geological maps were compiled; the research significantly modified the stratigraphic classification of the geological formations and the determination of the type and age of metamorphism; the knowledge about structural geological interpretation has also been changed. Earlier, maps were compiled in the traditional way since effective use of computer tools was not yet an option. However, the informatics background that also enabled mapping techniques to be used was already available at the beginning of the 21st century, providing more options for a wider utilization and presentation of map information.

Simultaneously, the completion of theoretical and practical research tasks raised an urgent need to compile a new basement map that would reflect the synthesis of current knowledge, and at the same time, would be an essential element of the nationwide geological spatial model. This requirement produced the initiative for a project that started in 2005. After a long period of theoretical preparation and preliminary organizing work, and with the collaboration of experts from many institutes, a new basement map (normal and raised relief) was published in 2010 (Haas et al. 2010a). The preparation work was conducted at the Eötvös Loránd University, but the conditions of the actual mapping were created through an agreement in principle between the Geological Institute of Hungary (MÁFI) and the MOL Plc. (the Hungarian Oil and Gas Company).

The map depicts the rocks making up the basement of the young basins formed in the Cenozoic, the tectonic elements that border the basins and which determine the relations and structures of rocks, as well as the contour representation of the basement relief of the basins. Given the fact that the rock masses building up the basement of the Pannonian Basin moved closer to each other only in the Cenozoic, units of significantly different origin must be separated. (Their respective positions are shown on separate inset maps). The symbols in the legend show genetically related geological formations within each unit. The map also depicts the most significant Mesozoic and Cenozoic tectonic elements determining the structure and spatial position of the rock bodies of the basement, and the depth of the basement relative to the sea level (with a contour interval of 500m).

The basement map may provide a starting point for future works on geodynamics and structural developments. It could be the first stage for a map depicting the whole basement of the Carpathian Basin System, although this would only be possible with international cooperation. The practical importance of the map is prominent in terms of further prospecting for hydrocarbon materials in Hungary, as well as with respect to groundwater explorations. It is also an important tool in the hands of users when the prognosis of geothermal energy opportunities and the designation of adequate exploration areas are concerned.

However, in order to use the map to gain adequate and effective help for basic and applied geological research, potential users must be familiar with (i) the principles of map compilation, (ii) the geological meaning of the geological units and rock bodies depicted on the map, (iii) the logic and reasons behind the establishment of rock units, (iv) the data and concepts related to the most important tectonic elements, (v) as well as the degree of the geological knowledge and the reliability of data. That is the purpose of this volume: it will provide a comprehensive overview of the geology of the pre-Cenozoic basement of Hungary, and at the same time, it is also the explanatory notes of the 1:500 000-scale geological map depicting the basement (HAAS et al. 2010a). It briefly describes the principles and processes of map compilation, as well as the utilization possibilities provided by the informatics background. This is followed by the presentation of the structural position of the area within the mega-units, and the short definition-like description of the key features of the tectonic elements bordering the structural units making up the area. The bulk of the book comprises the description of the pre-Cenozoic formations depicted on the map, which build up the structural units. They are discussed in the same order as shown in the legend. The description is illustrated by conceptual stratigraphic columns that, in several cases, show the formations that are grouped on the map in a more detailed manner. The last chapter of the volume gives a summary of the main events of the tectonic evolution. The undivided stratigraphic units and formations belonging to them which are depicted on the map and mentioned in the explanatory notes are listed in the Index at the end of the volume.

The transverse regional cross-sections — that, in addition to the geological build-up of the structural units, also show the position of these units relative to each other and the type of displacement — are annexes of this volume and the map.

Compiling the map

Works of map compilation were brought together by a compilation committee led by János Haas, with the participation of Tamás Budai, László Csontos, László Fodor and Gyula Konrád. The compilation was made using a scale of 1:250 000, by research groups responsible for different regions. Each research group was coordinated by a member of the committee. Regions were designated partly on the basis of structural, partly on the basis of practical aspects. Compilers sought to take into account available maps and geological data of neighbouring countries with special regard to the zones along the border. The geoinformation background of the map compilation was provided by the Geological Institute of Hungary (MÁFI), under the leadership of Tibor Tullner and Gábor Turczi.

The regional geological maps and the data of the deep borehole national database served as the basis for the map compilation. The compilers assessed all the used boreholes and classified the formations into the units of the map representation. They took into account the previous pre-Cenozoic basement and structural geological maps of the Geological Institute of Hungary (Fülöp et al. 1987, Dank et al. 1990) and used (i) the 'prognosis maps' compiled by the MOL Plc. in the 2000s, (ii) the materials of the Mecsekérc Ltd. recently compiled in the frame of the BAF (Boda Claystone Formation) project, (iii) the basement maps of the MÁFI compiled for the MOL Plc. around the Millennium, as well as (iv) the published geological and geophysical data available in public repositories.

The formations older than Cenozoic are depicted on the map thus the geological build-up of the pre-Cenozoic surface is also presented. The only exception is the Upper Cretaceous–Palaeogene flysch found in the Mecsek Unit of the Tisza Mega-unit that is not subdivided on the map.

Within each structural unit, the map depicts the genetically-related geological formations. These are mainly groups (for example: Upper Triassic and Lower Jurassic platform limestones, Variscan granitoids) and complexes. In order to indicate the age of the sequences dominantly made up of sedimentary rocks, as well the age of very low-grade and low-grade metamorphic rock sequences, the internationally recognised colour chart was applied. Igneous rocks are indicated by reddish colours, the medium-grade metamorphic rocks by pink shades. Areas where the build-up of the basement is barely known or unknown are coloured light grey. The outcrops of the pre-Cenozoic formations are marked with white contour lines and dotting. In the case of tectonic elements, colours distinguish Cenozoic elements from those older than Cenozoic. Line thicknesses refer to the importance of the structural elements (1st, 2nd, 3rd order) while the line style refers to the nature of displacement. Basement relief is depicted by contour lines with an interval of 500m above sea level.

The map shows the most important deep boreholes (ca. 1200) that reached the basement. Due to the limitations of the graphical representation, there was a strict selection of boreholes. In exceptional cases boreholes which did not reach the basement were added to the map, given their importance with regard to the determination of the depth of the basement.

The geoinformation background of the map compilation and the database created during the compilation allow such presentations and computer operations that would help a great deal with the utilization. Every aspect of the map content can be presented separately: i.e. formations, tectonic elements, and contour lines. It is possible to select and separately depict certain groups of formations (such as carbonate rocks, granitoid rocks), even by depth criteria. The geoinformation background allows the correction or expansion of the map database, as well as the refinement of certain parts.

Structural position of Hungary

The Pannonian Basin is a unique geological structure, surrounded by the Alps, the Carpathian Mountains and the Dinarides. The basin is interspersed with "inselbergs", though its overwhelming part is filled with several kilometres-thick young, loose sediments. The latter formed over the last 19 million years when plate tectonic processes resulted in lithosphere attenuation and intensive subsidence. Under the sediment mass which fills, the young basins, the basement is extremely diverse and shows a complex structure. This kind of diversity is attributed to the fact that the region was situated between two large tectonic plates (African and Eurasian) over a long (tens of millions of years) period. The formation and disappearance of oceanic basins, as well as continental collisions, resulted in the disintegration of the plate margins and the formation of the so-called terranes. As such, the basement of the Pannonian Basin consists of such structural elements (terranes) that were formed far from each other, under different conditions. They drifted towards each other in the early stage of the Cenozoic (Kovács et al. 2000).

The region belongs to the Alpine Orogenic Belt: more precisely, to the Mediterranean Mountain System that stretches from the Atlas Mountains in North Africa — through Europe — to South-east Asia. During the development of the western part of the Eurasian Plate, the successive plate tectonic cycles always ended with an orogenic phase. As a result of this, over the course of the Palaeozoic, continuously younger mountain systems joined the Precambrian core of the continent: namely, the Caledonides to the western margin of the plate in the Cambrian-Silurian, and the Variscides (Hercynides) to the southern margin in the Late Devonian – Carboniferous (Stille 1924, Peive et al. 1981, Fülöp 1989, McCann 2008).

The opening of the ocean within the frame of the Alpine tectonic cycle started in the Permian ca. 270 million years ago. During the formation of oceanic basins belonging to the Tethys system, microcontinent-sized fragments broke off both from the edge of the African and from that of the Eurasian Plate. At the time of the closure of the oceanic basins the fragments changed their position and were affected by deformation and metamorphism due to their collision with each other and with the large tectonic plates. At the end of the process the terrane-mosaic making up the basement of the Pannonian Basin assembled. Only then could the formation of the huge basin-system begin (Kázmér, Kovács 1985, Fodor et al. 1999, Tari et al. 1999, Kovács et al. 2000, Csontos, Vörös 2004, Horváth et al. 2006, Schmid et al. 2008, Haas et el. 2012).

The structural position of the pre-Miocene basement of the Pannonian Basin System is the result of the above-described complex structural evolution. The main features of the current structure of the basin are determined by the significant displacement of the early (pre-19 Ma) Cenozoic (Balla 1984, 1986; Csontos et al. 1992; Csontos, Nagymarosy 1998; Fodor et al. 1998; Haas, Kovács 2001; Schmid et al. 2008). They are shown on the pre-Cenozoic basement map, as well as on the inset maps depicting the main structural units (Haas et al. 2010a). The most important shear zone is the Mid-Hungarian Mega-unit (MH MU). This is a relatively narrow shear zone between the Mid-Hungarian Line and the Balaton Line, the latter representing the continuation of the Periadriatic Line (Figure 1). The Mid-Hungarian Mega-unit comprises the Mid-Transdanubian Unit that consists of elements also of South Alpine and Dinaric origin, as well as the Szarvaskő–Darnó, the Bükk and the Szendrő–Uppony units that moved a considerable distance from their place of origin as a result of strike-slip tectonics (Kovács, Haas 2010). The Late Palaeozoic – Mesozoic succession of the Mid-Transdanubian (Sava) Unit shows similarities with the formations occurring in the eastern part of the Southern Alps (Carnic Alps, Southern Karawanks, Julian Alps, Savinja Alps). From the boreholes reaching the basement of the Zala Basin, such formations have been described that can be related to the Jurassic ophiolite melange complex and the low-grade Alpine metamorphic rocks; these are known to occur in the Medvednica and in the Zagorje area (Pamić, Tomljenović 1998, Halamić et al. 1999, Haas et al. 2000, Haas, Kovács 2001, Pamić 2003, Judik et al. 2006). The Jurassic melange complex of the Szarvaskő–Darnó Unit can be related to that of the Dinaric Ophiolite Belt, while the Palaeozoic–Triassic succession of the Bükk Unit shows similarities with the Jadar and Sana–Una units of the Inner Dinarides (Pešić et al. 1986, Filipović et al. 2003, Dimitrijević et al. 2003, Haas, Kovács 2010).

South-east of the Mid-Hungarian shear zone the basement is represented by the rocks of the Tisza Mega-unit.This was broken off from the Variscan Orogenic Zone of the Eurasian Plate and deformed at the time of the Alpine orogeny. The area

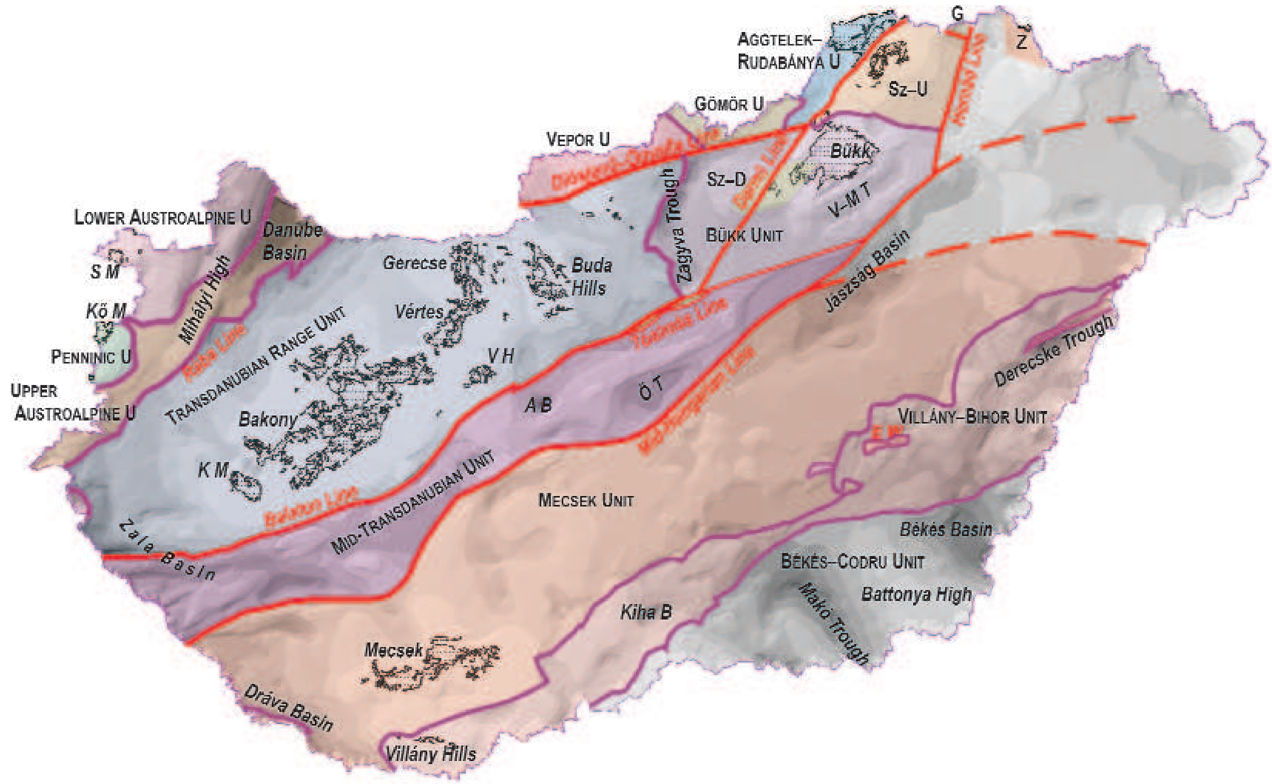


Figure 1. Basement map of Hungary depicting the tectonic units, the main structural elements separating the mega-units, as well as the main topographic elements of the basement

Lilac lines mark the boundaries of the Mesozoic nappes, the red lines those of Cenozoic structures. Dotted areas indicate the superficial occurrences of the basement. Abbreviations: A B: Adony Basin; E W: Endrőd Window; G: Gemer Unit; K M: Keszthely Mountains; Kiha B: Kiskunhalas Basin; Kő M: Kőszeg Mountains; Ö T: Örkény Trough; S M: Sopron Mountains; Sz–D: Szarvaskő–Darnó Unit; Sz–U: Szendrő–Uppony Unit; V H: Velence Hills; V–M T: Vatta–Maklár Trough; Z: Zemplén Unit

of the Tisza Mega-unit extends beyond the border of the country, encompassing the Apuseni Mountains and a part of the Slavonian Mountains as, well (Bleahu et al. 1996, Kovács et al. 2000, Haas, Péró 2004). Researches carried out in the last decade, however, pointed out that some of the metamorphic complexes (Moslavacka Gora) that formerly had been considered to belong to the Tisza Mega-unit, are in fact Mesozoic formations affected by Alpine metamorphism; as such, they are assigned to the Sava Unit along with the Cretaceous–Palaeogene flysch formation that at one time had been considered part of the Western Vardar Zone (Schmid et al. 2008, Ustaszewski et al. 2009). Accordingly, south-westward, the extension of the Tisza Mega-unit decreased.

The Alcapa Mega-unit originating from the African Plate consists of the Austroalpine and the Western Carpathian units. The Sopron Mountains and the basement of the Danube Basin are the direct continuation of the Austroalpine Nappe Systems. From beneath the latter, in the area of the Kőszeg Mountains, the Penninicum crops out in the form of a nappe window. In a structural sense, above them the Transdanubian Range Unit is situated; this is considered to be an Upper Austroalpine nappe which has slid from its original position (Tari 1994, Fodor et al. 2003). With regard to its facies characteristics, it shows a South Alpine affinity (Haas et al. 1995). In North Hungary, the Veporic Unit, the Zemplém Unit (probably an equivalent of the Veporic Unit), as well as the Gemer Unit and the Aggtelek–Rudabánya Unit (consisting of certain elements of the Inner Western Carpathian Nappe System) can be assigned to the Alcapa Mega-unit (Plašienka et al 1997, Vozárova, Vozár 1996, Kovač, Plašienka 2002, Kövér et al. 2009).

The boundarys of the mega-units

MID-HUNGARIAN LINE

It is a major structural line that dissects the basement of the Pannonian Basin in W–SW – E–SE direction (Figure 1). It is the boundary of the Tisza Mega-units, separating it from the transpression shear zone of the Mid-Hungarian Mega-unit. Its track line can be clearly determined from around Zagreb till the southern foreland of the Bükk. This assertion is primarily on the basis of the information on the rocks making up the basement. However, its track is debatable on the Nyírség area of the Tiszántúl where the basement is covered with a thick succession of Miocene volcanic rocks.

Scheffer (1959) was the first to refer to the existence of this important structural line, on the basis of geophysical anomalies. It is depicted as a major displacement zone on a basement structure map published by Kőrössy (1963). The lineament has a similar track line on the regional tectonic maps of Tollmann (1969) and Sandulescu (1975) (referred to as the boundary of the Tisia Unit), as well as on that of Flügel (1975) (referred to as "Zagreb Lineament"). The name "Mid-Hungarian Line" was first used by Szepesházy (1975). Wein (1978) referred to it on his structural map as the "Zagreb–Kulcs–Zemplén Lineament" and described it as the boundary of two mega-units that had drifted towards each other. Kovács (1982) depicted it on his structural map as the northern boundary of the "Tiszia" under the name "Mid-Hungarian Lineament". Brezsnyánszky, Haas (1986) highlighted the fact that the Mid-Hungarian Lineament is actually the margin of the Tisza microplate and that it determined the track line of the Cenozoic displacement.

In subsequent geodynamic interpretations the lineament appears as the southern boundary fault of the "Mid-Hungarian Belt" — i.e. that facing the Tisza Mega-unit (Balla 1984, Balla, Dudko 1989, Csontos 1995, Csontos, Nagymarosy 1998, Fodor et al. 1998, Csontos, Vörös 2004, Schmid et al. 2008, Kovács, Haas 2010). The most significant displacement along the lineament probably occurred during the Palaeogene – Early Miocene.

Post-Neogene activity can be inferred along several sections (Kapos Line, Tamási Line) on the basis of the morphostructural appearance around the Mid-Hungarian Line (Síkhegyi 2002).

RÁBA LINE

The Austroalpine units making up the basement of the Danube Basin are structurally separated from the Transdanubian Range Unit from the south-east by the NE–SW-trending Rába Line (Figure 1). In contrast to the Austroalpine and the Penninic units, the Transdanubian Range Unit was not affected by Alpine metamorphism. The Rába Line is covered with a thick succession of Neogene deposits and thus it can be studied only in detail on seismic profiles. Earlier, the Rába Line had been interpreted as representing a primary structural boundary. The concept, that used the Permo-Triassic facies as palaeogeographic indicators and correlated them along the line (Majoros 1980, Kovács 1982) assumed a significant, several hundreds of kilometres strike-slip movement (Kázmér, Kovács 1985). This process would have resulted in the "escape" of the Bakony Unit (as a crustal fragment) from the Alpine compression, and it would have arrived in its current position during the Late Palaeogene – Early Miocene. This concept interpreted the Rába Line as a deep-rooted strike-slip zone that might penetrate the entire crust and which separates crust fragments of significantly different geology.

New interpretations treat the Rába Line as Cretaceous nappe boundary (Horváth F. 1993, Fodor, Koroknai 2000, Haas et al. 2010a). Consequently, the Transdanubian Range Unit, in accordance with its non-metamorphic nature, represents the highest-positioned Austroalpine nappe within the Alpine system (Tari 1994, Fodor et al. 2003, Tari, Horváth F. 2010).

However, during the opening (extension) of the Pannonian Basin, the original Cretaceous structure was significantly modified; the northwestern margin of the Transdanubian Range Unit was affected and reactivated by an intensive extensional deformation (or more precisely an oblique-slip deformation). The structural pattern as seen today is attributed to this deformation. On the other hand, the geometry of the Miocene deformation, as well as the intensity of the reactivation, shows significant change along the strike of the Rába Line. It is also due to the alongstrike changing Miocene fault kinematics that the track line of the Miocene fault and that of the Cretaceous nappe boundary is slightly different. This can be best seen at the eastern margin of the Mihályi High where the normal fault is NNE–SSW-striking, while the Cretaceous nappe boundary keeps the NE–SW direction (TARI 1994). In the vicinity of Ikervár and Nemeskolta the Cretaceous nappe boundary runs in the footwall of the Miocene fault, while north-east of it, it runs in the hanging wall (TARI 1994, FODOR et al. 2013). As a consequence, a tectonic zone of complex inner structure developed; this was made up of Early Cretaceous, possibly Late Cretaceous and Miocene elements, showing significant, location-dependent differences. Further investigations are needed to clarify what kind of structural element (i.e. of what geometry and of what age) the Rába Line should be considered as being.

DIÓSJENŐ–ÓGYALLA LINE

The major structural units of the Inner Western Carpathians (the Vepor and the Gemer units) are separated from the Transdanubian Range Unit and the Bükk Unit by the ca. E–W-striking Diósjenő–Ógyalla (Diósjenő–Hurbanovo) Line (Figure 1). The Rába Line (that has an approximately NE–SW strike on the Danube Basin) joins the E–W-striking Diósjenő–Ógyalla Line when reaching the southern part of Slovakia. The Diósjenő–Ógyalla Line reaches the Hungarian border once more at the western margin of the Börzsöny Mountains, with the same E–W strike line. The track line can be traced on the basis of boreholes that have reached different kinds of basement units (those of the Transdanubian Range and the Inner Western Carpathians) (BALLA 1989). The structural element is reflected also on the magnetic anomaly map (KISS J., GULYÁS 2006). There is no direct information on the eastern termination of the line; most probably it runs into the Darnó Zone that might displace it to a significant degree. In this case the displaced continuation of the line would border (from the south) the presumed small block of the Gemer Unit at Hidasnémeti. According to an alternative concept, it might be related to Late Cretaceous low-angle detachment planes and it continues as such a structural element in an eastward direction (under Salgótarján) (FODOR, KOROKNAI 2000).

There is little information on the kinematics of the Diósjenő–Ógyalla Line due to Cenozoic cover rocks. The primary structural character is quite uncertain. By the projection of the surficial fault geometry at Southern Slovakia, FODOR, KOROKNAI (2000) assumed an oblique sinistral-normal kinematics along the line in the Late Cretaceous that was overprinted by right-lateral strike-slip kinematics in the Early Cenozoic (BALLA 1989, BADA et al. 1996). In the Oligocene, basement rooted reverse faulting occurred along the line, characterised by large amplitude and southern vergency (thick-skinned backthrust, TARI et al. 1993). A movement of sinistral character is assumed from the late Middle Miocene.

BALATON LINE

The Balaton Line is a structural element running parallel with Lake Balaton at a distance of about 10km to the south (Figure 1). It does not occur on the surface; its position is assumed on the basis of geophysical and borehole data. Most scientists agree that south-westward it forms a unique structure with the Periadriatic Line (PREMRU 1981, KÁZMÉR, KOVÁCS 1985, CSONTOS et al. 1992, CSONTOS, NAGYMAROSY 1988, FODOR et al. 1998, JÓSVAI et al. 2005). The line can be easily traced till the Velence Hills in the north-east (BALLA 1999b). According to BALLA et al. (1987), the line continues north from the Velence Lake until it eventually terminates. On our map it continues in the Tóalmás Line east to the Danube, similarly as shown in the work of BREZSNYÁNSZKY et al. (1986).

The Balaton Line is a primary structural boundary between the Transdanubian Range Unit and the Mid-Hungarian Mega-unit, as shown also by early works, albeit using different terminology (SZENTES 1961; KŐRÖSSY 1963; WEIN 1969, 1978; BREZSNÁNYSZKY et al. 1986). The Transdanubian Range Unit is only slightly deformed north-west of the line, while the Mid-Hungarian Mega-unit is made up of highly-sheared elements. In other words, the Balaton Zone is the south-eastern boundary of the rigid Alcapa Mega-unit and at the same time the northernmost major structure of the Mid-Hungarian Shear Zone (CSONTOS et al. 1992, FODOR et al. 1999). Looking at the details of the Balaton Line, there is the possibility that it behaves as a complex shear zone, but this can mostly be inferred from Slovenian analogues on the surface. According to this, the zone is made up of strike-slip duplexes with anastomosing fault boundaries. The duplexes are consisted of blocks sheared from the two main bounding units or even from exotic units. Such duplexes were encountered by boreholes originating from the Velence Granite, as well as the rock lenses sheared from the Palaeogene basins (CSONTOS et al. 1992, TARI 1994, FODOR et al. 1998). West of Balaton, low- or medium-grade metamorphites probably also form duplexes. The extent of the zone in a south-eastward direction is debateable; it depends on whether the next strike-slip zones are considered individual zones or parts of the Balaton Zone. If the former concept is applied (BALLA 1999a), south of the zone the Buzsák Line (Zone) follows.

As for the kinematics of the zone, with regard to the most important, Late Oligocene – Early Miocene age of the activity, a right-lateral strike-slip movement is generally accepted (KÁZMÉR, KOVÁCS 1985, BALLA 1988, DUDKO 1988, BALLA, DUDKO 1989, CSONTOS et al. 1992, FODOR et al. 1998). The amount of movement can be determined by the correlation of the Permian–Mesozoic facies zones, the Velence and the Eisenkappel granites, as well as the Palaeogene basin remnants. Estimations calculate with a distance of 280–350km for the total movement observed today (BÁLDI 1983, KÁZMÉR, KOVÁCS 1985, TARI 1994, CSONTOS et al. 1992), though a part of it is attributed to subsequent (syn-rift) stretching. The right-laterial strike-slip fault is covered with Karpathian sediments. In the Middle Miocene certain segments of the zone were possibly reactivated (either as reverse faults or as transpressive structures), as indicated by borehole and seismic data (BALLA et al. 1987, CSONTOS et al. 2005). Sinistral kinematics is assumed from the Middle Miocene (FODOR et al. 1999). Sinistral movement probably continued also on some of the elements of the zone in the neotectonic phase (MAGYARI et al. 2005, BADA et al. 2010).

DARNÓ LINE (DARNÓ ZONE)

The Darnó Line is a structural element situated in North Hungary, in the northern foreland of the Bükk and the Mátra Mountains (Figure 1). According to the original definition the Darnó Fault lies between the Mesozoic of the Darnó Hill and the Cenozoic sediments to the west (TELEGDI ROTH 1937, SCHRÉTER 1942). Its south-western continuation under the Mátra till the Tóalmás Line can only be assumed on the basis of geophysical data. North-eastward it can be linked to many other structural lines and can be followed till the south-eastern side of the Rudabány Hills (PANTÓ 1956). Along with the nearby, parallel structural elements the Darnó Zone might include the whole Uppony and Rudabánya Hills (HERNYÁK 1977, LESS et al. 1988, LESS, MELLO eds 2004, FODOR et al. 2005c). In an even broader sense the zone between Ózd and the Bükk can be considered as a Darnó Deformation Belt (VASS 2002, FODOR et al. 2005). The structural zone is well-recognisable on the Bouguer anomaly map and can be identified on seismic reflection profiles (SZALAY, ZELENKA 1977, BRAUN et al. 1989). On the basis of these data and borehole information the zone borders the deep sub-basin of the North Hungarian Palaeogene basin from the east (BÁLDI 1986); moving eastwards from the zone, a Palaeogene succession of significant thickness is known only south and north-east of the Bükk.

There are different views on the structural role of the line and the zone, as well as on the amount of displacement. Boreholes along the close vicinity of the Darnó Line justified thrusting (TELEGDI ROTH 1951) since they penetrated Kiscellian sediments under the Permo-Triassic formations. The age of the thrusting might be post-Kiscellian–pre-Ottnangian (~30-19 Ma). The Palaeozoic was thrust over the Mesozoic and also over the Oligocene in the foreland of the Uppony Hills (SCHRÉTER 1952, PANTÓ 1954). This kind of kinematics is justified by field measurements (FODOR et al. 2005) and seismic profile interpretations (SZTANÓ, TARI 1993). However, it should be pointed out that JASKÓ (1946), later ZELENKA et al. (1983), GRILL et al. (1984), LESS et al. (1988) and SZENTPÉTERY (1997) had already identified a strike-slip kinematics of the line and postulated 20–30km sinistral displacement. According to FODOR et al. (2005c), this sinistral movement could have been active in the Ottnangian – Early Badenian (~18–15 Ma). Starting from the Late Badenian, normal kinematics can be assumed on the basis of borehole information and field structures (TELEGDI ROTH 1951, RADÓCZ 1966, FODOR et al. 2005c).

Most of the Palaeozoic–Mesozoic rocks of the Darnó Zone underwent plastic deformation and metamorphism during the Cretaceous. Along the zone, the Cretaceous deformation along the proto-Darnó Zone can be indicated by ductile structural elements; this assumption is confirmed by the foliations, folds and lineations parallel with the possibly left-lateral strike-slip line (KOROKNAI 2004). Almost synchronously, the Bükk and Szendrő rocks and older structures were bended in a ductile manner toward the proto-Darnó Line (ZELENKA et al. 1983, CSONTOS 1999).

There are divergent views concerning the role of the Darnó Line in the distribution of the pre-Cenozoic formations. WEIN (1969) and SZALAY, ZELENKA (1979) considered it to be an important Palaeo-Mesozoic palaeogeographic boundary. SCHMID et al. (2008) thought it could be one of the most important structures separating mega-units — as the continuation of the Palaeogene – Early Miocene Periadriatic Line — in the form of a right-lateral strike-slip fault. However, according to this map and the interpretation of ZELENKA et al. (1983) and HAAS, KOVÁCS (2001), with regard to the Mesozoic formations, no significant differences are shown on the two sides of the zone; only different nappes of the same nappe series occur, depending on the variable pre-Cenozoic denudation and deformation. As such, on this map the line holds no primary importance and does not represent a boundary between tectonic mega-units.

Geological build-up of the mega-units of the basement

The pre-Cenozoic basement of Hungary consists of three mega-units (Figure 1): the Tisza Mega-unit that derives from the European Plate; the complex Alcapa Mega-unit deriving from the African Plate; and the shear zone between these two, known as the Mid-Hungarian Mega-unit. In this chapter a description is given of the geological build-up of the structural mega-units and the structural units within them. Formations are described from older to younger usually according to the order of the legend of the map. The colouring of the formations on the stratigraphic columns attached to the descriptions is the same as that used on the map. The explanatory table of symbols marking the lithology and the depositional environment is placed in the back cover of the book.

TISZA MEGA-UNIT

The pre-Cenozoic basement of the Tisza Mega-unit consists of three large units that were formed during the Cretaceous: the Mecsek, the Villány–Bihor and the Békés–Codru Nappe Systems (Figure 2). Superficial outcrops of the basement occur only on the area of the Mecsek Mountains in the Mecsek Unit and on the area of the Villány Hills in the Villány–Bihor Unit. On the remaining parts of these two units and in the Békés–Codru Unit the basement is known solely from the successions exposed by boreholes. The Cretaceous nappes of the Tisza Mega-unit have a NNW vergency, similarly to most of the compression structures within the structural units (such as the slivers of the Villány Hills). In the basement of the Great Plain the lower-positioned nappes crop out from beneath the higher-positioned nappes, on the surface of the basement, in the form of tectonic windows. Examples include the outcrop of the Mecsek Unit from below the Villány–Bihor Unit on the area of the Endrőd Window (Figure 1). According to the seismic profiles and to the integrated interpretation of low-temperature geochronological data (Tari et al. 1999), in several areas the exhumation of the deeper-positioned units is unmistakably a neo-Alpine process in connection with the extension of the Pannonian Basin (Figure 2).

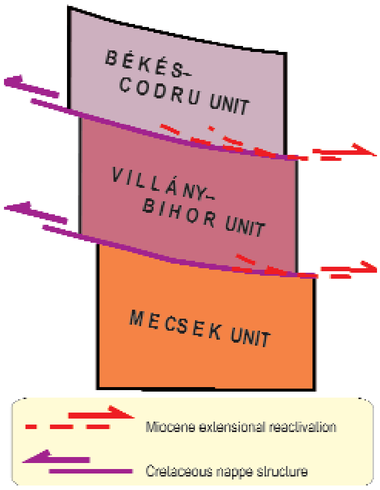


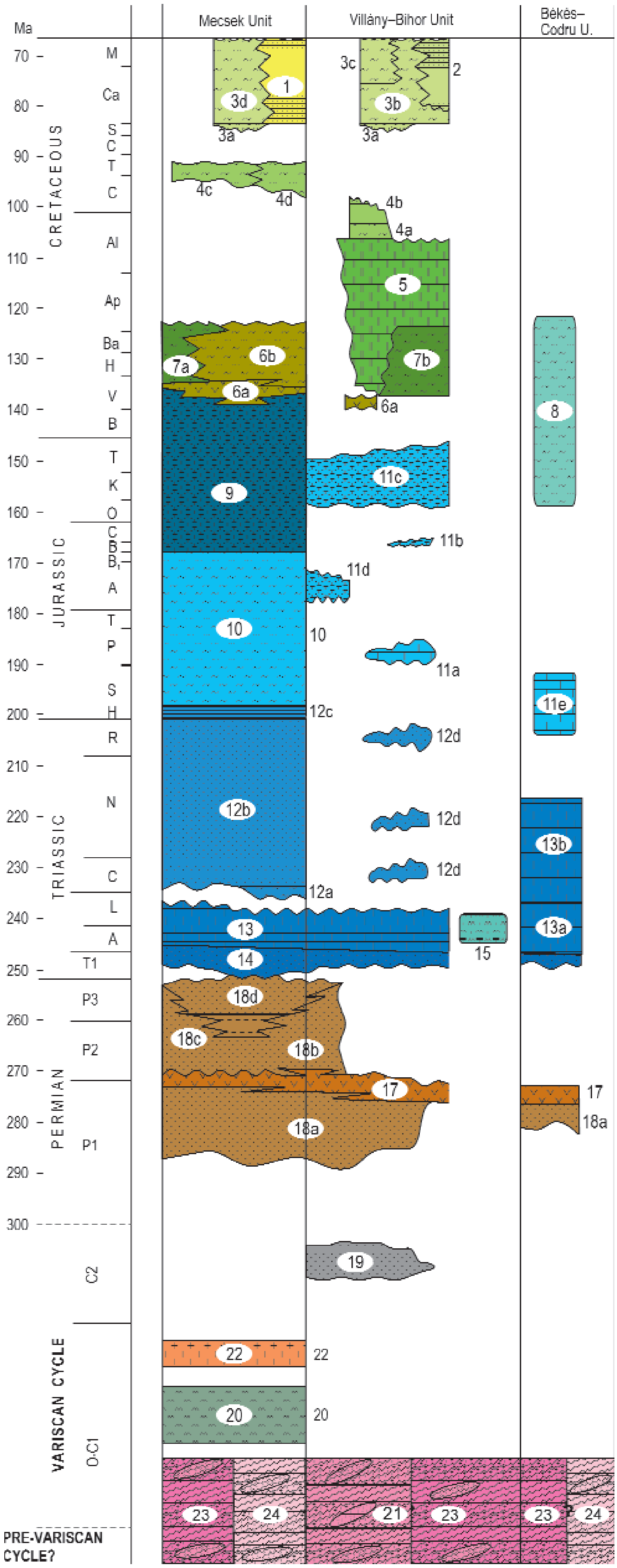
Figure 2. Schematic tectonic model of the pre-Cenozoic structural units of the Tisza Mega-unit

Variscan tectonic cycle

Pre-Alpine basement (20–24)

The oldest formations of the pre-Cenozoic basement of the Tisza Mega-unit are assigned to many crystalline basement complexes (Szederkényi 1996, 1998). The subdivision was mostly based on the differences of the mineralogical/lithological facies and the type of metamorphism. The distinguished complexes, according to the basic features, can be sorted into three major classes (Figure 3): (i) The overwhelming proportion includes medium-grade Variscan metamorphites (gneiss-micaschist-amphibolite, locally with marble, quartzite, eclogite and serpentinite lenses) (21, 23, 24). Beyond them there are also units dominantly consisting of (ii) granitoids (22) and (iii) low-grade metamorphites. Their occurrence is limited to a much smaller area.

The subdivided and unsubdivided **Variscan medium-grade metamorphites (21, 23, 24)** of the first group making up the bulk of the pre-Alpine basement are known solely from deep boreholes. Complexes in this group include the earlier identified Babócsa, Baksa, Körös, Kelebia, Tisza, Battonya and Sarkadkeresztúr Complexes.



Their lithology is quite similar: it consists of gneiss, micaschist, as well as amphibolite intercalations. The protolith was probably a carbonate-free (low-carbonate) greywacke– psammite–pelite formation, into which basic lava and tuff rocks were deposited. In addition to the above-mentioned rock types, in the Baksa Complex (21), carbonate formations (marble, dolo-marble, calcareous silicate rocks) occur as well. Locally, smaller eclogite (e.g. Görcsöny, Jánoshalma, Szarvas, Mezősas, Körösladány) and serpentinite bodies (Gyód) have been deposited in the host mica schist-gneiss succession.

There is little information on the inner structure of the crystalline basement. However, with regard to the mega-structural features it is an important aspect that the crystalline formations of the basement in SW Hungary (Babócsa and Baksa Sub-units) are mainly characterised by a (W)NW–E(SE) strike (as indicated by the observed south-western dips in the their continuation in the Slavonian mountains [Papuk, Psunj]). In the meantime, in the other parts of the Tisza Mega-unit (namely, the crystalline basement of the Mecsek, the Villány–Bihor and the Békés–Codru units) an overall NE–SW strike is probable. The intensive inner deformation of the formations is indicated by the frequent highly-folded character of the formations and the appearance of mylonitic/ cataclastic shear zones (mostly associated with intense retrograde effects).

The rocks of this group have the same (or at least very similar) pre-Alpine metamorphic evolutionary history, as shown with these major stages:

1) Early high-pressure metamorphism of eclogite facies (p_{min}=15–16 kbar, T=600–700 °C), characterised by a mineral assemblage of garnet (rich in almandine–pyrope components), omphacite, kyanite, phengite, zoisite and rutile. The relicts of the metamorphism have been preserved in only a few cases (e.g. Ravasz-Baranyai 1969).

2) Medium-grade Barrow-type metamorphism characterised by garnet–staurolite–kyanite± sillimanite index minerals, determining the petrography of the crystalline rocks of the Tisza Mega-unit (p=6–9 kbar, T=550–650 °C; Árkai et al. 1985, Török K. 1990).

3) Low-pressure, medium-grade metamorphism with high thermal gradient, with andalusite index mineral (p=2–3 kbar, T=550–600 °C; Lelkes-Felvári et al. 1989).

Figure 3. Stratigraphic column of the pre-Cenozoic formations of the Tisza Mega-unit (Barabás, Barabásné 1998, Bérczi-Makk 1998, Császár ed. 1996, Főzy ed. 2012, Bérczi-Makk et al. 2004, Szederkényi 1998)

On the basis of geochronological data (K-Ar- and Ar-Ar-dating on mica minerals) the cooling following the amphibolite facies metamorphism occurred ca. 320–290 million years ago (Balogh Kad. et al. 1983, Lelkes-Felvári et al. 2003, Lelkes-Felvári, Frank 2006).

In the Papuk Mountains of Croatia Balen et al. (2006) found signs of a medium-pressure metamorphism of amphibolite facies (p=8–11 kbar, T=600–650 °C). This is similar in every respect to phase 2, only it is of a much older age (428–445 Ma). This indicates that the above-described pre-Alpine evolutionary model is not applicable to the whole area of the Tisza Megaunit and new research might come up with significant changes to the existing model.

The Permian (~275 Ma, Sm-Nd data obtained from garnet) high-temperature/low-pressure metamorphism described in the south-eastern part of the Great Plain (Algyő High) also indicates the recently determined event of the metamorphic evolution (Lelkes-Felvári et al. 2003). Published geochronological data indicate this Permo-Triassic event also affected the eastern part of the Great Plain (e.g. Szeghalom) (Lelkes-Felvári et al. 2003, Balogh Kad. et al. 2009). During the Alpine orogenic phase ca. 90 million years ago the crystalline basement complexes were affected by a low-grade retrograde metamorphism of green schist facies; this is evident at many places on the Great Plain (pl. Sáránd; Árkai et al. 1998). On the other hand, the eo-Alpine metamorphism (95–82 Ma, Ar-Ar muscovite) on the area of the Algyő-Ferencszállás basement-high also reached even the amphibolite facies (with disthene, staurolite and garnet index minerals) (Horváth, P., Árkai 2002, Lelkes-Felvári et al. 2003). The Permo-Mesozoic formations overlying the crystalline complexes were affected by eo-Alpine very low-grade/low-grade prograde metamorphism over a large area in the vicinity of the Cretaceous nappe fronts of the Great Plain, as well in the deeper-situated units cropping out in tectonic windows (Árkai et al. 2000). There are reasonable grounds to assume the occurrence of Alpine retrograde metamorphism of the underlying crystalline rocks in these areas.

Variscan granitoid rocks (22)

These rocks form the second group of the formations making up the pre-Alpine basement. With regard to the areal extension, they are of high importance. Examples include the Mórágy Complex that was formed as a result of the mixing of magmas of different composition. It mostly consists of monzogranite, monzonite, diverse contaminated rock types and late-magmatic leucocratic dykes. The Lower Carboniferous (~340 Ma) intrusive body was affected by a regional metamorphism of green schist facies over the course of the Variscan orogeny, during the cooling following the intrusion. The intensity of the metamorphism was highly diverse in space (Balla, Gyalog eds 2009). The original contact of the granite body is unknown; it has a tectonic contact in the north with the metamorphites exposed in the Mecsekalja Belt (Ófalu Complex, see below). A smaller outcrop is known to be present in the western part of the Mecsek, while the granite body can also be followed in the basement north-east from the superficial formations of the East Mecsek (Mórágy Block), reaching Kecskemét, and Szolnok as well (Bátaszék–Miske– Soltvadkert–Kecel). In the northern foreland of the Mecsek Mountains (Mágocs) it is known only from boreholes.

There is also a significant Variscan granite body in the southern part of the Battonya Complex.

Early Palaeozoic low-grade metamorphic complex (20)

This complex represents the third group of formations making up the pre-Alpine basement. The formations in this group have the smallest areal extension. They include the Ófalu, Szalatnak, Horváthertelend, Tázlár and Álmosd Complexes, as well as the Gyód and Helesfa Serpentinites.

The ca. 1km-wide tectonic zone of the Mecsekalja Belt stretching out from the southern foreland of the Mecsek dips north-westward at 50° and is made up of Early Palaeozoic (Silurian?–Devonian) folded, mylonitised rock types (Balla, Gyalog eds 2009). The bulk of the Ófalu Complex comprises gneiss and quartz phyllite, with less crystalline limestone and amphibolite. A small serpentinite body is tectonically wedged into the phyllite. The main metamorphic event in the Ófalu Complex is of greenschist facies, though in certain gneiss types there are also traces of an earlier metamorphism of amphibolite facies (Lelkes-Felvári et al. 2000). The age of the mylonitic deformation of green schist facies — related to the strike-slip movement along the zone (Szederkényi 1977) — is ca. 300 million years (Lelkes-Felvári et al. 2000).

In the foreland of the Mecsek Mountain dark grey, folded Lower Silurian slate–metasiltstone–metasandstone series and associated metavolcanites are known to occur (Szalatnak Slate) in a drilled thickness of 600m. A totally similar formation occurs in the western foreland of the Mecsek as well, at Horváthertelend. Both have undergone transitional very low / low-grade low-pressure metamorphism (~350 °C; Árkai et al. 1995b). Szederkényi (1996, 1998) interpreted the formations at Szalatnak and Horváthertelend as Variscan nappe remnants of unknown vergency, similarly to the Ófalu Phyllite. However, the latter, despite being partly similar to them with regard to the composition, was affected by higher-temperature metamorphism of greenschist facies. The pre-Alpine age of the metamorphism is justified by the non-metamorphic Permian – Lower Triassic formations at Szalatnak, which overlie the metamorphites.

A near-vertical serpentinite body is known in the Mórágy Complex, in the axis zone of the anticline of the West Mecsek

(Helesfa) as well as in the Baksa Complex, in the area of the Görcsöny High (Gyód). They are situated with a tectonic contact.

At Tázlár, in the Danube–Tisza Interfluve (in the vicinity of the contact of the Mecsek and the Villány–Bihor nappes) folded, very low-grade metamorphic, often graphite- and pyrite-rich carbonate phyllite, quartz phyllite and sericite phyllite are known above the polymetamorphic basement in small tectonic remnants; they are tectonically pinched in the gneiss making up the basement. The thickness of the formation is several tens of metres. The age is unknown. FÜLÖP (1994) assumes an Early Palaeozoic, possibly Lower Carboniferous sandy marl / marl protolith, rich in organic material content.

Several boreholes in the eastern part of the Great Plain in the vicinity of Álmosd have exposed chlorite schist, biotite-muscovite schist, graphitic biotite schist and tremolite-actinolite schist — all affected by greenschist facies metamorphism. SZEDERKÉNYI (1996) interpreted these rocks as NW-vergent Upper Cretaceous nappe remnants of small areal extent above the polymetamorphic basement of the Körös Complex. With respect to the assignment of the pre-Cenozoic formations of the nearby Bam–1 and –2 (Bagamér), as well as the Nyáb–1 (Nyírábrány) boreholes to the Mesozoic of the Mecsek Mountains (HAAS et al. 2010a), it seems more probable that these formations also belong to the very low-grade/low-grade metamorphic Mesozoic of the Mecsek Mountains. They crop out in tectonic windows from beneath the medium-grade metamorphites of the Körös Complex.

Upper Carboniferous continental siliciclastic complex (19)

The Variscan evolutionary cycle of the Tisza Mega-unit ended in the Late Carboniferous with the molasse formation following the orogenesis. The terrestrial Téseny Sandstone made up of grey, cyclic conglomerate–sandstone–siltstone beds is known to occur in the "Görcsöny High" area (Bogádmindszent, Siklósbodony, Téseny), as well as in the area of the Dráva Basin (Csokonyavisonta, Darány, Kálmáncsa Szulok). The maximal drilled thickness is 1100m (in borehole Bogádmindszent Bg–1). Carboniferous molasse formations also occur in smaller areas in the Danube–Tisza Interfluve (Soltvadkert, Nagykőrös, Törtel), as well as in the northern foreland of the Villány Hills (Túrony) (JÁMBOR 1998).

ALPINE TECTONIC CYCLE

Permian – Middle Triassic evolutionary cycle

The Alpine respective evolutionary histories of the three units of the Tisza Mega-unit were uniform until the beginning of the Late Triassic. In this period the microplate belonged to the European shelf of the Tethys. On the basis of the facies image of the structural units, the Mecsek Unit was probably the closest to the continent; the Villány–Bihor Unit was situated in the middle part of the shelf and the Békés–Codru Unit in that side of the shelf that faced the pelagic region (BLEAHU et al. 1994).

Permian continental siliciclastic formations (18)

The denudation period following the Variscan orogeny is represented by a 3–4km-thick Permian terrestrial molasse that is deposited either on the Carboniferous granite or on the metamorphic Variscan basement. The mostly red fluvial succession of clastic sediments is made up of fining-upward cycles in the Lower Permian (Korpád Sandstone, 18a). In the Middle Permian (Cserd Formation, 18b) it interfingers with fine-grained lacustrine formations (Boda Claystone, 18c). The cyclic fluvial sedimentation (to which uranium ore genesis was associated in the Late Permian) continued even in the beginning of the Early Triassic (Kővágószőlős Sandstone, 18d). The clastic series is known from the whole area of the Tisza Mega-unit in variable thicknesses (Mecsek Unit [Mecsek, Vajta V–1], Villány–Bihor Unit [Máriagyüd Mgy–1], Bihor parautochton, Békés–Codru Unit [Palics, Tótkomlós T–1]). Within the series, the Korpád Sandstone has the largest areal extension. It is several hundred metres thick and underlies the rhyolite (17). The occurrence of the Boda Claystone and the Kővágószőlős Sandstone is restricted to the Mecsek. The thickness of the former is 1000m, while that of the latter is 150–1200m.

Permian rhyolite (17)

The early continental rifting of the Alpine cycle was accompanied by extensive acidic volcanism. The Gyűrűfű Rhyolite — deposited in the Lower Permian terrestrial molasse succession — is mostly made up of greyish lilac extrusive and sub-volcanic lava rocks, frequently with intercalating tuff, agglomerate and ignimbrite (R. VARGA 2009). The maximal thickness of the formation exceeds 800m (for example, south of the Mecsek Mountains, Egerág E–7). Although it is subordinately present in the Mecsek Unit (Gyűrűfű, Jákó), it is more significant in the Villány–Bihor Unit (Somberek, Vokány, Bisse), especially in the deep boreholes of the Danube–Tisza Interfluve (Kiskunmajsa, Kömpöc, Csólyospálos).

Its extent is significant in the basement of the Békés–Codru Unit on the area of the Battonya High (Battonya, Mezőkovácsháza, Pitvaros).

Mesozoic rocks in general (16)

Formations of uncertain stratigraphic position — probably of Mesozoic age — are indicated in those areas where sporadic borehole data made it impossible to depict the geological build-up. Examples include a part of the Battonya High north of Orosháza, where Lower Permian rhyolite (Nagyszénás Nsz–2), Lower Triassic sandstone (Nsz–3) as well as Middle Triassic dolomite (Orosháza Oros–2) equally occur. Boreholes drilled above a high of the basement at the north-western boundary of the Mecsek Unit (Szenta Szta–1, Szta–2) have exposed andesite of uncertain position, Mesozoic age of which cannot be excluded.

Low-grade metamorphic Mesozoic formations (15)

In the basement of the Tiszántúl region very low-grade / low-grade metamorphic formations (ÁRKAI et al. 1998, 2000) crop out in a tectonic window (in the vicinity of Sáránd and Endrőd) from beneath the Variscan crystalline formations of the Villány–Bihor Unit and in the vicinity of the nappe front. These can be assigned to the Mesozoic of the Mecsek Unit. The upper section of the series exposed by the Sáránd S–I borehole is made up of very low-grade metamorphic Anisian dolomite, dolomitic limestone and shale, under which low-grade metamorphic marble, slate and basic volcanic rocks are found (PAP 1990, ÁRKAI et al. 1998). The original sediment of the siliceous metalimestone exposed by the Bagamér Bam–1 borehole is considered to be open-marine (radiolarian?) limestone in which the dark green chlorite schist was originally siltstone or neutral-basic volcanic rock. Under these formations — conditionally assigned to the Middle Triassic — a Lower Triassic succession of metamarl, metasandstone, calcareous schist and metadolomite is deposited.

The very low-grade metamorphic dark grey marl and Calpionella-bearing limestone exposed by the boreholes in the area of the Endrőd Window can be assigned to the Upper Jurassic, while the underlying sandstone and coal-bearing siltstone can be placed in the Lower Jurassic (SZILI GY.-NÉ 1985, BÉRCZI-MAKK 1998, KŐRÖSSY 2005).

Mesozoic very low-grade metamorphic formations crop out on the surface of the basement also in the area of the Dráva Basin. The succession of the boreholes marked as Barcs-Ny is made up of acidic metavolcanite (metaryholite), metasandstone, lamellar folded dolomitic chlorite schist, metadolomite with sericite- and chlorite layers, dolomitic anhydrite and crystalline dolomite (KÓKAI et al. 1987, ÁRKAI 1990). This succession probably correlates with the low-grade Mesozoic formations occurring in the pre-Cenozoic basement in North Croatia, where they have been exposed by many boreholes (HAAS et al. 2000b).

Lower Triassic siliciclastic formation of fluvial and delta facies (14)

The coarse conglomerate formed at the beginning of the Early Triassic indicates the uplift of the denuding background and the start of a new sedimentary cycle. The red, fluvial succession of the Jakabhegy Sandstone has a fining-upward character. Over a large area the formation is directly deposited on the metamorphic Variscan basement (BARABÁS, BARABÁS-STUHL 2005).

On the surface the formation is known from the Mecsek, but it also occurs in the Apuseni Mountains. It has been exposed by boreholes in the Dráva Basin (Komlósd Kom–1, Cún Cun–1), in the forelands of the Mecsek Mountain (Szalatnak Sza–3, Somberek Sb–3, Himesháza Hh–1) and in the basement of the Great Plain (Kecskemét–Nagykőrös, Kisköre, Zsana, Battonya). Its typical thickness is 250m (BÉRCZI-MAKK et al. 2004).

Middle Triassic shallow marine siliciclastic and carbonate formations (13)

As a result of the Anisian transgression over the Early Triassic subaerial region, a shallow marine ramp began to evolve, characterised by the deposition of fine-grained terrigenous clastic sediments (Patacs Siltstone). During the rest of the Early Anisian a succession of alternating dolomite, dolomitic marl, claystone, siltstone, anhydrite and gypsum was deposited (Magyarürög Evaporite, Hetvehely Dolomite).

In the Mecsek and in the Villány–Bihor units sedimentation took place on a carbonate ramp during the Middle Anisian and in the Ladinian. The succession is represented by dark grey limestones (Víganvár Limestone, Lapis Limestone, Zuhánya Limestone) and dolomitised carbonate rocks (Rókahegy Dolomite, Csukma Dolomite, Templomhegy Dolomite). The Anisian–Norian succession of the Békés–Codru Unit is overwhelmingly made up of dolomitised rocks formed on a carbonate platform (Szeged Dolomite, 13a, Csanádapáca Dolomite, 13b) (BLEAHU et al. 1994, BÉRCZI-MAKK 1998, BÉRCZI-MAKK et al. 2004). The thickness of the carbonate succession exceeds 600m.

Late Triassic – Early Cretaceous evolutionary cycle

The rifting of the Penninic Ocean started in the Late Triassic. The uniform evolution of the shallow marine shelf came to a halt at the beginning of the Late Triassic and the extensional tectonics resulted in the segmentation of the platform. After that, during the Mesozoic, the evolutionary history of the three structural units branched and became differentiated.

In the "half-graben" sedimentary basin of the Mecsek Unit, a large amount of siliciclastic sediments was deposited in fluvial, delta and later, coastal swamp environments over the course of Late Triassic and in the early phase of the Early Jurassic (NAGY E. 1969). In the Early Jurassic, increasingly deeper marine depositional environmental evolved, characterised by the deposition of pelagic carbonate mud and fine terrigenous clastic sediments. In the middle of the Middle Jurassic, as a result of the breaking off of the Tisza microplate from the European Plate, the terrigenous influx significantly decreased and in the deep marine depositional environments pelagic calcareous mud and siliceous mud started to be deposited. Both the sedimentary rocks and the fauna assemblages became similar to those of the Mediterranean-type Jurassic formations (GÉCZY 1973, VÖRÖS 1993). During the Early Cretaceous, intensive alkali basaltic volcanism occurred (HARANGI et al 1996).

The Villány–Bihor Unit between the deep basins of the Mecsek and the Békés–Codru units was in a relatively emerged position during the Jurassic and the Early Cretaceous.

Upper Triassic to Lower Jurassic coal bearing siliciclastic formation (12)

The shallow carbonate ramp which existed in the Middle Triassic became subaerially exposed during the Ladinian in the central part of the Mecsek. Subsequent transgression started with the deposition of the Kisrét Limestone, formed in brackish water. In a lagoonal environment of gradually decreasing salinity, bituminous limestone and calcareous marl were deposited (Kantavár Formation, 12a). The thickness of the overlying Upper Triassic (Karolinavölgy Sandstone, 12b) and Lower Jurassic fluvial, as well as delta sandstone-siltstone-claystone successions (Mecsek Coal, 12c) increases from NE to SW, from 400m to 1800m. The formation is significantly thinner in the basement of the Great Plain, as indicated by the boreholes around Nagykőrös, Szank and Tázlár. The clastic Upper Triassic is known only in the form of a several metres-thick terrestrial – shallow marine facies in the Villány–Bihor Unit (Mészhegy Sandstone, 12d; BÉRCZI-MAKK et al. 2004).

Jurassic shallow marine and condensed pelagic limestone formations (11)

In the slivers of the Villány Hills, the Jurassic pelagic limestone succession is made up of several metres-thick Lower Jurassic pebbly limestone (Somssichhegy Limestone, 11a), Middle Jurassic pelagic condensed limestone (Villány Limestone, 11b), as well as Upper Jurassic thick-bedded, bedded pelagic limestone (Szársomlyó Limestone, 11c); the latter reaches 300m in thickness. The succession contains several hiatus levels (VÖRÖS 2009).

The Middle Jurassic formations (11d) of the Villány–Bihor Unit occur on the surface as well, in the Máriakéménd–Somberek range. They are represented by a 100–300m thick succession of alternating thinly-bedded marl and crinoidal limestone (Somberek Limestone), and a 50m-thick cherty limestone succession (Máriakéménd Limestone, 11d) (CSÁSZÁR 2012, RAUCSIK 2012).

The oldest member of the incomplete Jurassic succession exposed in the area of the Békés–Codru Unit (Tótkomlós T–I) is represented by reddish-brown crionidal limestone ("menyháza limestone" 11e; BÉRCZI-MAKK 1998).

Lower and Middle Jurassic pelagic, fine-grained siliciclastic formations (10)

In the area of the East Mecsek, the Lower Jurassic ("spotted marl") succession of pelagic basin facies is made up of sandstone, siltstone and marl (Vasas and Hosszúhetény Marl, Mecseknádasd Sandstone, Óbánya Siltstone, Komló Calcareous Marl). Its thickness increases southwestward where it may attain 2000m (NÉMEDI VARGA 1988). The spotted marl is known in significant thicknesses north-east from the Mecsek, in the vicinity of Tolnanémedi in the Transdanubian region, as well as in the basement of the Danube–Tisza Interfluve around Kecskemét–Nagykőrös (BÉRCZI-MAKK 1998).

Middle Jurassic to Lower Cretaceous pelagic limestone, cherty limestone (9)

In the East Mecsek the pelagic Middle Jurassic is represented by thin-bedded limestone (Óbánya Limestone), the Upper Jurassic by siliceous limestone and radiolarite (Fonyászó Limestone) as well as thin-bedded limestone (Kisújbánya and Márévár Limestone). These formations are known from the Mecsek Unit: in the Kecel–Kiskőrös–Soltvadkert–Páhi region in the Danube–Tisza Interfluve, and in the vicinity of Tiszagyenda and Hajdúszoboszló in the Tiszántúl region (BÉRCZI-MAKK 1998).

Jurassic to Lower Cretaceous pelagic limestones, marls (8)

In the area of the Békés–Codru Unit the Upper Jurassic – Lower Cretaceous pelagic Calpionella and Saccocoma-bearing, radiolarian Pusztaszőlős Marl has a large areal extent (Pusztaszőlős, Tótkomlós T-K–2).

Lower Cretaceous pelagic marls, limestones (7)

The Lower Cretaceous succession of the Mecsek Unit (7a) consists of clay marl and bentonitic basalt tuff (Hidasivölgy Marl), as well as bedded crinoidal limestone (Apátvarasd Limestone). Its largest extent is known in the basement in the region of Kerekegyháza and Lajosmizse, in the Danube–Tisza Interfluve (Kőrössy 1992). The oldest section of the Lower Cretaceous succession in the Villány–Bihor Unit is made up of calcareous marl and ooidic limestone (Biharugra Formation, 7b) which overlies a succession of conglomerate, sandstone and siltstone (Bérczi-Makk 1996).

Lower Cretaceous basic volcanites and their reworked marine deposits (6)

The Lower Cretaceous volcanites (6a) are represented by alkali basalt, trachybasalt, tephrite and phonolite rocks (Mecsekjános Basalt). Their appearance (pillow lava, hyaloclastite) and characteristic alteration patterns indicate submarine effusion. The thickness of the debris (Magyaregregy Conglomerate, 6b) deriving from the erosion of the volcanoes and the reefs (evolved on the volcanoes) is several hundred metres (Császár 1998b, 2002). There is a significant occurrence in the north-eastern part of the Mecsek Unit, primarily in the vicinity of Martfű, Tiszagyenda and Nagykőrös–Kecskemét.

Lower Cretaceous platform limestone (5)

In the slivers of the Villány Hills thick-bedded, cyclic, shallow marine, biogene limestone of “Urgonian facies” (Nagyharsány Limestone) overlies the Jurassic limestone with erosional unconformity (Császár 1996, 1998b, 2002). Its thickness attains 600m in the southern foreland of the Villány Hills. It occurs in the southern part of the Danube–Tisza Interfluve in the vicinity of Öttömös, as well as in the Tiszántúl region in the Doboz–Sarkad–Biharugra Range.

Albian–Late Cretaceous evolutionary cycle

The appearance of coarse siliciclastic turbidite formations in the Villány–Bihor Unit suggests significant changes in the Alpine evolution of the Tisza Mega-unit at the end of the Albian. The turbidite formations were probably formed in the foreland basin of the nappe fronts. Turonian pelagic basins evolved in relation with the advance of the nappe fronts. In the Apuseni Mountains it is clearly shown that the major Alpine nappe-stacking phase was in the Coniacian. According to many observations, this also can be applied to the whole Tisza Mega-unit. During the orogenic phase a significant uplift occurred, as a result of which the overwhelming part of the Albian–Turonian pelagic formations became denuded. In the Santonian, in the foreland of the nappe systems, basins evolved; namely, the Villány–Bihor Basin in the foreland of the Békés–Codru Nappe System and the Szolnok Basin in the foreland of the Villány–Bihor Nappe System (in the Mecsek Unit). In these basins the sedimentation began in the Santonian–Campanian and continued in the Szolnok Basin even in the Palaeogene.

Albian and Turonian basinal marls and clastic slope deposits (4)

The grey silty basinal marl of the Late Aptian – Early Cenomanian (Bisse Marl, 4a) is deposited on Lower Cretaceous (Aptian) platform limestone in the Villány Hills and on Upper Jurassic limestone in the Bóly Basin north-east from it. Infrequently, the marl also contains siliciclastic turbidite intercalations (Császár 1998b). A formation similar in age and facies has been exposed in the southern part of the Danube–Tisza Interfluve as well (Pusztamérges–ÉK–1). In the Bóly Basin, the marl gradually passes upward into a turbidite succession of alternating feldspathic sandstone, pebbly sandstone and conglomerate beds, with a thickness of around 230m (Bóly Sandstone, 4b). Its age is Middle Albian – Middle Cenomanian (Császár 1998b). A similar Cenomanian formation has been exposed in the Danube–Tisza Interfluve (Üllés-ÉNy–3, Alpár–I). Red bathyal basinal marl assigned to the Turonian is known from the northern slivers of the Mecsek Mountains (Vékény Marl, 4c). Marl similar in age and facies has been exposed in the Danube–Tisza Interfluve (Kerekegyháza–5). Turonian basinal grey silty marl and clay marl with conglomerate and breccia intercalations have been exposed in the eastern part of the Danube–Tisza Interfluve (Gátér Marl, 4d; Gátér–2) (Haas 1998b).

Senonian continental, shallow and deep marine formations (3)

The formations of the Senonian sedimentary cycle are deposited with erosional and angular unconformity mostly on older Mesozoic rocks (Triassic, Jurassic, Lower Cretaceous) or occasionally, on Variscan crystalline rocks. The Senonian succession usually starts with coarse-grained formations (Szank Formation, 3a) consisting of alternating breccia, conglomerate, pebbly sandstone and greyish white sandstone. The thickness of the formation is 20–180m. It is terrestrial sediment; the poorly sorted breccia (fanglomerate) is toe-of-slope sediment, while the conglomerate and the sandstone were probably deposited following a short fluvial sediment transport. According to sporomorphs, the age of the formation is Late Santonian – Early Campanian (HAAS 1987).

In the Villány–Bihor Unit, in the southern part of the Danube–Tisza Interfluve, a 60–120m thick, Lower Campanian succession of deep-marine grey clay marl, marl and silty marl overlies the terrestrial sediments (locally, directly the basement) (Csikéria Marl, 3b) (SZENTGYÖRGYI 1983, 1989; HAAS 1987). With respect to the number of individual species, it has a rich planktonic foraminifera and nannoplankton assemblage. Its sporomorph assemblage indicates the closeness of the shore. The dominantly marly succession is overlain by a lithologically heterogeneous formation comprising a siliciclastic-carbonate rock mix (Bácsalmás Formation, 3c), reaching a thickness of 100–300m. The dominant rock type in the lower part of the formation is marl; in the middle part, limestone; and in the upper part, sandstone (SZENTGYÖRGYI 1983, 1989; HAAS 1987). Gravel and breccia can be observed everywhere in the formation, either being present sporadically or as intercalations. Their respective amounts, however, decrease upward. The depositional environment was possibly a submarine slope.

In the Mecsek Unit, in the middle part of the Danube–Tisza Interfluve Campanian–Maastrichtian deep-marine marl and calcareous marl, clayey limestone is known to occur (Izsák Marl, 3d). The lower part of the formation is red marl that gradually passes upward into grey silty marl. The amount of silt increases, while the carbonate content decreases north-eastward. Its thickness is around 400m in the Danube–Tisza Interfluve (borehole Izsák–1) and 60–300m in the Tiszántúl region (Kunmadaras, Kisújszállás, Nádudvar) (SZENTGYÖRGYI 1989).

Senonian flysch (2)

The Campanian–Maastrichtian sedimentary cycle is deposited on the crystalline basement and also on the older Mesozoic formations with angular or erosional unconformity. It is a dark grey formation (Körös Fm), consisting of alternating sandstone and siltstone, as well as silty clay marl with sandstone interbeddings. It contains conglomerate intercalations. At its base, basal breccia is to be found. It is known from the Villány–Bihor Zone in the eastern part of the Danube–Tisza Interfluve (Mélykút, Kisszállás, Zsana-N, Kiskunmajsa, Gátér), where it is exposed with a range of thickness of 100–550m. It also occurs in the Tiszántúl region, in the area of the Körös Rivers, where two facies have been distinguished: a thinner one with a thickness of 100–150m, dominantly made up of silty clay marl, and a thicker one of alternating sandstone and siltstone, reaching a thickness of 1000m (SZENTGYÖRGYI 1989).

Senonian–Palaeogene pelagic marls, flysch (1)

According to earlier assumptions, the sedimentation of flysch sediments was continuous from the Late Cretaceous till the Oligocene in the north-eastern part of the Mecsek Unit (“Szolnok Flysch”). However, detailed investigation of nannoplankton assemblages has indicated that the sedimentation was not continuous, since in most parts of the area there are significant hiatus levels in the successions, probably related to submarine erosion (BÁLDI-BEKE et al. 1981, BÁLDI-BEKE, NAGYMAROSY 1993, NAGYMAROSY 1998).

In the Mecsek Unit, in the north-eastern part of the Tiszántúl region (in the vicinity of Nádudvar and Debrecen) a Campanian-Maastrichtian formation of alternating sandstone and siltstone beds has been exposed by boreholes, with a maximum thickness of 500m (Debrecen Sandstone). Although the rocks are in most cases heavily pressed and deformed, the turbidite features are easily recognisable.

On the basis of nannoplankton, the red, variegated and greenish grey marl, claystone and sandstone turbidite formations are assigned to the Palaeocene and the Lower Eocene. They have been exposed in only a few boreholes. They occur sporadically, in patches (BÁLDI-BEKE, NAGYMAROSY 1993, NAGYMAROSY 1998). The Middle and Upper Eocene formations have a much larger areal extent. Characteristic rock types include sandstone turbidite and silty marl with grey and variegated thin sandstone beds, as well as poorly sorted sandstone, pebbly sandstone, conglomerate and breccia (Nádudvar Formation). Sandy marl and limestone are also exposed and they contain redeposited shallow marine fossils (Hajdúszoboszló) (NAGYMAROSY in HAAS et al. 2012). The extent of the Oligocene formations is restricted to the north-eastern part of the Tiszántúl section of the Mecsek Unit. They are dominantly composed of grey clay marl with sandstone intercalations (NAGYMAROSY in HAAS et al. 2012).

The pre-Cenozoic geological build-up of the north-western part of Hungary is determined by the successions known in the neighbouring Eastern Alps. The mega-units of the Eastern Alps in Hungary are represented by the Penninicum (structurally in the deepest position), as well as (above it) the elements of the Austroalpine (Eastern Alpine) nappe system. These are characterised by a complex inner structure and their diverse rock assemblages can be studied in some isolated superficial outcrops around the Austrian–Hungarian border. These outcrops at the same time represent the easternmost extension of the Eastern Alps. Eastward, in the north-western part of the Transdanubian region, many (mainly hydrocarbon exploratory) boreholes have exposed these units in the basement.

Both structural units are characterised by an inner nappe structure (and by related Alpine metamorphism) developed during the eo-Alpine phase in the Cretaceous in the Lower and the Upper Austroalpine units, and during the Meso-Alpine phase in the Early Cenozoic in the Penninicum. The original nappe structure underwent significant alteration at the beginning of the formation of the Pannonian Basin, during the Miocene extensional deformation (neo-Alpine phase) (TARI 1994, 1996).

Recently, the small Ikervár Unit has been distinguished at the boundary of the Austroalpine and the Transdanubian Range units (HAAS et al. 2010a). However, given the unverified stratigraphic and the insufficient metamorphic lithological and tectonic data, it is still an element of uncertain affinity.

The Transdanubian Range Unit is the uppermost, non-metamorphic member of the Austroalpine Nappe System (HORVÁTH F. 1993, TARI, HORVÁTH F. 2010).

The schematic tectonic model of the above-mentioned structural units is shown on Figure 4. Detailed descriptions of the respective units are provided below.

PENNINIC UNIT

Low-grade metamorphic Jurassic – Lower Cretaceous formations (25)

The Penninic Unit comprises the magmatic basic-ultrabasic formations of the crust of the Penninic Ocean (which opened in the Middle Mesozoic and then vanished in the Palaeogene), as well as the diverse rock association of clastic and carbonate sediments, overlying the magmatic rocks (Figure 5). In the Meso-Alpine phase these formations underwent metamorphism of blueschist and greenschist facies.

The deeper-positioned lithological units (Kőszeg Quartz Phyllite, 25a, Velem Calcareous Phyllite, 25b and Cák Conglomerate, 25c) can be assigned to the Jurassic, or perhaps to the Early Cretaceous. The quartzite and the

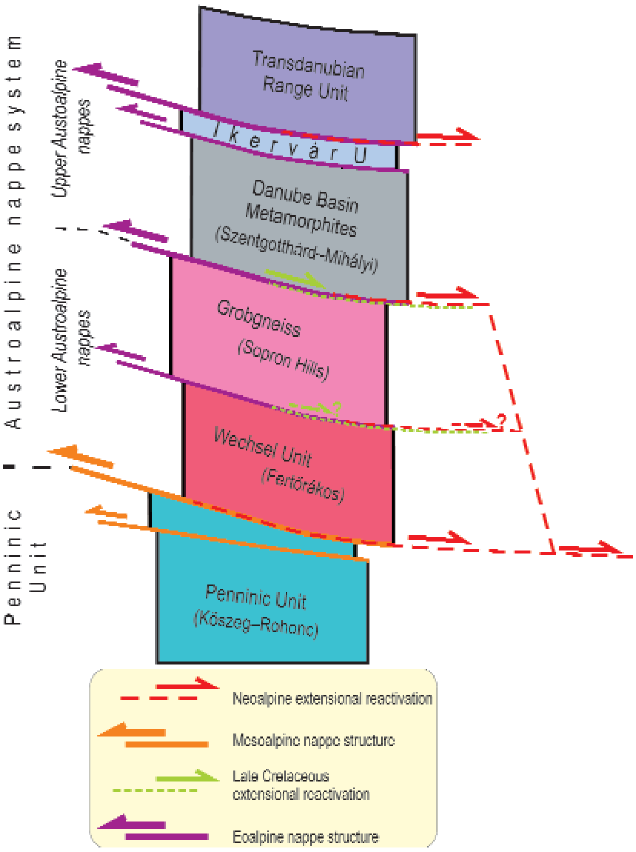


Figure 4. Schematic tectonic model of the Pre-Cenozoic structural units of North Transdanubia

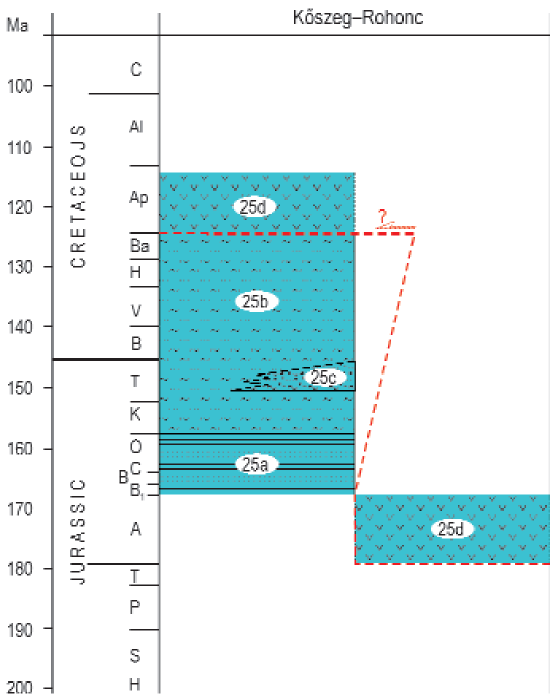


Figure 5. Schematic stratigraphic column of the Penninic Unit. Due to the almost complete absence of biostratigraphic data the age of the depicted formations is uncertain

quartz phyllite were formed by the alteration of sandy and clayey sediments. Locally, the calcareous phyllite contains pure marble intercalations and these represent the metamorphic rocks of the deep marine basin still existing in the Cretaceous; they are characterised by calcareous-clayey sedimentation (Schönlaub 1973). The conglomerate intercalating in the calcareous phyllite in several metres-thick lenses dominantly consists of the fragments of re-crystallised Triassic dolomite and dolomitic limestone. The pebbles of the conglomerate derive from the shallow marine region, from where they were transported to the deeper basin by gravitational movement (redeposition). The (stratigraphically? structurally?) highest-positioned (Figure 5) greenschist (Felcsőtár Greenschist, 25d) was formed as a result of the metamorphism of basaltic lavas and pyroclastic rocks. Serpentinite and talc schist occur between the greenschist and the calcareous phyllite in the Vashegy area, along with metagabbro. These rocks were formed through the metamorphism of the deeper-positioned members (harzburgite, ferrogabbro/diorite) of the ophiolite series which once made up the oceanic crust. In Hungary, these rocks crop out in the Kőszeg Mountains and in the Vas Hill.

The Penninicum in the discussed area crops out from beneath the units of the Austroalpine Nappe System in tectonic windows (Kőszeg-Rohonc and Vashegy Windows). Its inner structure is characterised by nappes and intensive, multiphase folding (Pahr 1984, Ratschbacher et al. 1990, Dudko, Toufik 1990).

The earliest traces of metamorphic alteration have been preserved in the gabbroic rocks of the unit: namely, during the ocean-floor metamorphism (T=450–750 °C, P=1–2 kbar) that occurred as a result of the hydrothermal solution activity affecting the crust high-temperature, amphiboles were formed. This was followed by a subduction-related metamorphic event of blueschist facies (T=330–370 °C, P=6–8 kbar) the traces of which have also been preserved in the ophiolite series. The age of metamorphism is presumably Late Cretaceous or Palaeogene. The greenschist metamorphism (T=350–430 °C, P~3 kbar) determining the petrographic image of the rocks occurred in the Oligocene and in the Early Miocene (muscovite K-Ar: 31–28 és 23–19 Ma) (Lelkes-Felvári 1982). The degree of the greenschist metamorphism increases southward (Vas Hill) as indicated by the appearance of the biotite in the greenschists. The final exhumation and the related cooling of the Penninicum occurred during the Miocene (zircon FT: 22–15 million years, apatite FT: 10–6 million years) (Balogh Kad. et al. 1983, Dunkl, Demény 1997).

Lower Austroalpine Unit

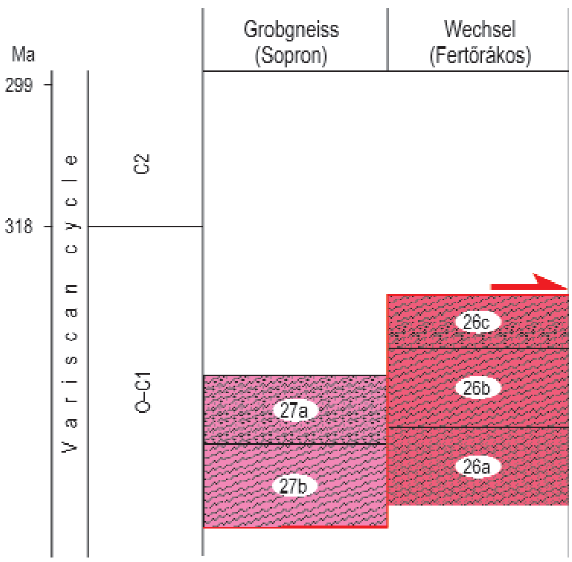


Figure 6. Schematic stratigraphic column of the Lower Austroalpine nappes

The Austroalpine Unit is a nappe system comprising complicated, polimetamorphic basement complexes related to the Alpine tectogenesis and metamorphism which occurred in the Cretaceous. It can be further divided into two sub-units (Lower and Upper Austroalpine Nappe Systems) (Figure 4), the boundary of which in Hungary lies in the basement of the Danube Basin. There is no detailed knowledge available on this particular subject. Further inner nappes can be distinguished within the sub-units, made up of protoliths of quite different type and age. As a consequence, the Austroalpine nappes could not have derived from one single, geologically uniform "mother unit" (i.e. from the so-called "Altkristalline"); rather, they represent an association of many units of different origin. These units were formed in different geodynamic environments and were joined together in the Cretaceous.

In Hungary, the Lower Austroalpine Unit is traditionally accepted as being further divided into two nappes: the structurally deeper-positioned Wechsel and the overlying Semmering (Grobgneis) units (Figures 4 and 6; cf. Pahr 1991). There is no direct information on the Createous nappe contact of these two.

Medium-grade polymetamorphic formations* (26)

The tectonically deeper-positioned Wechsel Unit is made up of medium-grade polimetamorphic formations. The lower part of the succession dominantly consists of amphibolite, amphibole (actinolite) and biotite schist (26a). Above them feldspathic and graphitic, and, locally, phyllonitic mica schist follows (26b). In the latter, frequently, apatite-rich rock lenses are situated. Upward in the succession marble–dolomarble also appears. In the uppermost part there are muscovite–albite–microcline gneiss (26c) and mica schist with phyllonitic zones. These formations crop out on the surface in the area of the

* Incorrectly, in the legend of the pre-Cenozoic map, high-grade metamorphic formations are listed too.

small Fertőrákos "schist island", situated in the vicinity of Lake Fertő. However, knowledge about these formations largely comes from borehole data.

The inner structure of the unit is its least known aspect. The phyllonitic zones mark ductile Alpine shear zones. The gneiss in the uppermost structural position belongs to the higher-positioned Grobgneiss Nappe, which is known from the Sopron Mountains.

In a petrological sense, two metamorphic events can be distinguished within the series: namely, greenschist facies metamorphism overprints the older, amphibolite facies event. In addition to some Variscan (Rb-Sr: 339±12 Ma) and Late Cretaceous ages, geochronological investigations indicated ages mostly in the 170–220 Ma interval (Ar-Ar muscovite and biotite). The latter has been explained by the entry of additional Ar (FRANK et al. 1996). Taking into account also the results from the Sopron area, the following metamorphic evolutionary history can be given: (1) the early amphibolite facies metamorphism is probably of Variscan age (~330 Ma); (2) ages dominantly in the 170–220 Ma interval might indicate the effect of the Permo-Triassic high-temperature and low-pressure event with the intensive re-crystallisation or neo-crystallisation of micas; (3) the Late Cretaceous ages indicate the last, eo-Alpine metamorphism of greenschist facies.

Medium-grade polymetamorphicformations (27)

The tectonically higher-positioned Semmering (Grobgneisz) Unit is made up of medium-grade polimetamorphic formations. At its lower part, various orthogneiss rocks are characteristic (27a); the most common type is the greyish white, well-foliated, medium-grained muscovite-gneiss (Sopronbánfalva Gneiss). Upward in the succession mica schist is mainly the dominant rock type (27b), with a highly diverse mineral composition (Óbrennberg Mica Schist, Vöröshíd Mica Schist). In the mica schists disthene-quartzite lenses intercalate. A rare, but a characteristic rock type is the excellently foliated, white leucophyllite which represents a tectonite formed from gneiss and mica schist. Albeit very subordinately, amphibolite occurs as well. All these rocks crop out in the Sopron Mountains.

The inner structure of the Grobgneiss Unit is generally characterised by a flat, southward dip. Mylonitic rock types formed in low-angle shear zones include the leucophyllite and the Rókaháza Gneiss; the latter is characterised by a well-developed stretching lineation. These are partly related to the Cretaceous nappe stacking, partly to the exhumation of the unit (Figure 4). DRAGANITS (1998) assigned the orthogneiss and mica schist rocks (Vöröshíd Mica Schist — that show pronounced Alpine overprints) to the Lower Austroalpine Unit, while the mica schists (Óbrennberg – Kaltes Bründl series), that are characterised by well-preserved pre-Alpine mineral paragenesis, correlate with the higher-positioned Strallegg complex. SCHMID et al. (2004, 2008) assigned both the Grobgneiss and the Stralleg Complex to the Upper Austroalpine Unit (Koralpe–Pohorje–Wölz Nappe; see details later). As such, in this classification only the Fertőrákos Schist Island, currently assigned to the Wechsel Nappe, is considered to be part of the Lower Austroalpine Unit.

The succession is characterised by a complex polimetamorphic evolutionary history (KISHÁZI, IVANCSICS 1985; LELKES-FELVÁRI et al. 1984; TÖRÖK 1996, 1998, 1999). As a consequence, the early (pre-Alpine) events can be reconstructed only with uncertainty. The earliest, amphibolite facies metamorphism is reflected by mica schists (T=600–700 °C, P=2–4 kbar?). On the basis of data of various methods (monazite U-Pb-Th: 310±34 Ma, biotite K-Ar: ~320±12 Ma), the age of the event is evidently Variscan (NAGY G. et al. 2002, BALOGH KAD., DUNKL 2005). The pronounced andalusite-sillimanite-biotite paragenesis, that locally contains staurolite relicts, was formed during a low-pressure metamorphism of amphibolite facies. This event can be correlated with the large-scale Permo-Triassic HT/LP event of the Eastern Alps; this assertion is based on geochronology (biotite K-Ar and muscovite Ar-Ar: 275–236 Ma; BALOGH KAD., DUNKL 2005). The typical index mineral assemblage of the third event indicates high-pressure metamorphism (T=450–550 °C, P=11–14 kbar; TÖRÖK K. 1996, 1998). Its age is undoubtedly Late Cretaceous (70–80 Ma; biotite and phengite muscovite K-Ar and Ar-Ar, monazite U-Pb-Th; BALOGH KAD., DUNKL 2005, NAGY G. et al. 2002). Fission track data indicates the rapid tectonic exhumation of the formation in the Late Cretaceous (Figure 4.) and its persistent shallow position during the Early Palaeogene (zircon FT: ~70 Ma, apatite FT: 60–40 Ma; BALOGH KAD., DUNKL 2005).

UPPER AUSTROALPINE UNIT

The Upper Austroalpine Unit occurs in the basement of the Danube Basin, east and south of the Penninic and Lower Austroalpine units, extending till the Rába Line (previously known as Rába Metamorphic Complex, FÜLÖP 1990). It is made up of low-grade metamorphic rocks (Figure 7.).

Variscan, low-grade metamorphic Lower Palaeozoic formations (28) and Devonian marble, calcareous schist (29)

The lower part of the complex dominantly consists of phyillites (Mihályi Phyillite, 28a), with intercalating neutral-basic volcanic rocks (Sótony Metavolcanite, 28b). In the absence of biostratigraphic data these are conditionally assigned to

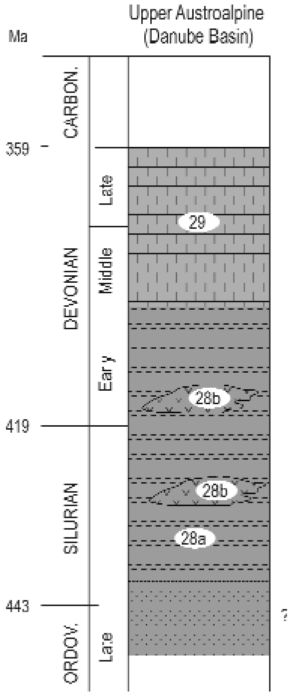


Figure 7. Schematic stratigraphic column of the Upper Austroalpine Unit (Danube Basin). In the absence of biostratigraphic data the age and relations of the depicted formations are debatable

Silurian–Devonian. The Devonian age of the carbonates (Bük Dolomite, 29) is also, for similar reasons, suppositious. It is uncertain whether the low-grade slate penetrated at Szentgotthárd can be correlated with the Mihályi Phyllite. FÜLÖP (1990) interpreted the very low-grade metasandstone exposed by boreholes Nemeskolta Kol–2 and –3 and Ikervár Ike–4 and –7 as the basal formation of the succession. However, HAAS et al. (2010a) assigned it to the Early Palaeozoic metamorphic basement (54) of the Transdanubian Range Unit.

The Upper Austroalpine Unit structurally overlies the Penninic and the Lower Austroalpine units (Figure 4.) and can be traced in deep boreholes from Szentgotthárd, through the Ölbő area, till the north-northeastern edge of the Mihályi High. With regard to the lithological and structural features, it can be correlated with the Graz Palaeozoic (BALÁZS 1975, FÜLÖP 1990). However, BALLA (1993), in his detailed work considers this analogy to be highly debatable. The transverse foliation typical of phyllites justifies the intensively folded inner structure of the complex.

The complex was affected mostly by low-grade metamorphism (T=350–400 °C) (ÁRKAI et al. 1987), with K-Ar ages of 180–116 Ma (ÁRKAI, BALOGH KAD.1989). These ages (in contrast with the K-Ar ages (310–330 Ma) of the lithologically similar metamorphic rocks of the Transdanubian Range Unit) can be interpreted as mixed ages; younger ages are due to eo-Alpine (Cretaceous) metamorphic effects.

The **Variscan** (medium-grade) **metamorphic formations** (**30**), exposed by the hydrocarbon exploratory borehole Baján–M–1 near to the Slovenian border, are also assigned to the Upper Austroalpine Nappe System. These formations are different from the above-mentioned formations both concerning their distribution and their geological features. The borehole has exposed mylonitic mica schist and gneiss, as well as amphibolite of symplectite texture at a depth of 3800–4100m; these lie, under the Miocene sediments. These crystalline basement formations crop out from beneath the pre-Cenozoic formations of the Transdanubian Range Unit in a tectonic window (FODOR et al. 2003, HAAS et al. 2010a). The penetrated ductile shear zone, which is of extensional origin and low-angle dip, is related to the Late Cretaceous – Early Cenozoic extension of the Austroalpine units (LELKES-FELVÁRI et al. 2002, FODOR et al. 2003).

The unit is characterised by a complex, multiphase metamorphic evolution (LELKES-FELVÁRI et al. 2002): the early, high-pressure metamorphism of eclogite facies (preserved only in symplectite relicts) was followed by medium-grade eo-Alpine metamorphism (T=570 °C, P>8 kbar). It was later overprinted by Early Palaeogene (Mu, Ar-Ar 67–60 Ma), high-temperature mylonitisation of greenschist facies (~450 °C). These rocks, both with regard to their appearance and metamorphic evolution, show conspicuous similarity with the characteristic rock types making up the bulk of the Pohorje Mountains of North-east Slovenia. As such, they can be correlated with the crystalline rock series of the Koralpe–Pohorje–Wölz Nappe System (SCHMID et al. 2004, 2008).

TRANSDANUBIAN RANGE UNIT

The Transdanubian Range Unit is bordered by the Balaton Line south-eastward and by the Rába Line north-westward (Figure 1). It is also tectonically bordered (i) in the south-west, towards the Pohorje Unit of the Austroalpine Nappe System, (ii) in the north, at the Diósjenő–Ógyalla Line, (iii) and in the north-east, towards the Bükk Unit. According to the current concepts regarding the related tectonics, the Transdanubian Range Unit is the uppermost member of the Austroalpine Nappe System (Figure 4). Its pre-Cenozoic structure is determined by compressional tectonic elements of SW-NE strike. The syncline structure of the unit was formed during the Austrian and pre-Gosau phases in the mid-Cretaceous (BALLA, DUDKO 1989). In the axis zone, Jurassic and Lower Cretaceous formations are to be found, while the formations progressively get older towards the limbs, ranging in age from the Upper Triassic to the Early Palaeozoic. Both sides of the syncline are accompanied by antiforms, in the axes of which Early Palaeozoic formations are found. Such an anticline stretches from the Zala Basin through the basement of Lake Balaton, till the southern foreland of the Buda Hills, as well as in the basement of the Danube Basin, in the vicinity of Tét, Vaszar and Takácsi. On the limbs of the syncline most exposed to the effects of compression, overturned and overthrust folds have been formed, among which the most important one is the Litér thrust. This stretches over a length of 100km through the Balaton Highland and through the East Bakony (BÖCKH 1872 1873; TELEKI 1939; DUDKO in BUDAI et al. 1999). The inner structure of the synform is complicated by the Bakonybél Thrust (TARI 1994).

Between the low-grade metamorphic formations (metamorphic events which occurred during the Variscan orogeny) of the Upper Austroalpine Unit (assigned to the Palaeozoic and the formations of the Transdanubian Range Unit with similar age and metamorphic grade) there is a narrow zone probably comprising several tectonic slivers. This zone is in the vicinity of Sótony and Ikervár, where several boreholes have exposed very low-grade metamorphic slate, metamarl, sandstone, conglomerate, and volcanic tuff. From one single borehole some Jurassic – Lower Cretaceous microfossils have also been described (JUHÁSZ Á., KŐHÁTI 1966). These formations are depicted on the map as **"Mesozoic very low-grade metamorphic formations" (48)** and have been assigned to the conditional Ikervár structure.

Variscan tectonic cycle

The Early Palaeozoic succession of the Transdanubian Range was formed during the Variscan orogenic cycle (Figure 8). The open-marine, fine sedimentation during the Ordovician and the Devonian was occasionally interrupted by basic and acidic volcanic activity. A shallow marine carbonate platform evolved during the Devonian, connected by slope to the open sea. Early Palaeozoic formations underwent very low-grade and low-grade metamorphism during the Variscan orogeny in the Early Carboniferous. Over the course of the Late Carboniferous, molasse sediments accumulated in the shallow marine and continental basins, the material of which originated from the denudation of the Variscan Mountain System.

Variscan, Early Palaeozoic low-grade metamorphic formations (54)

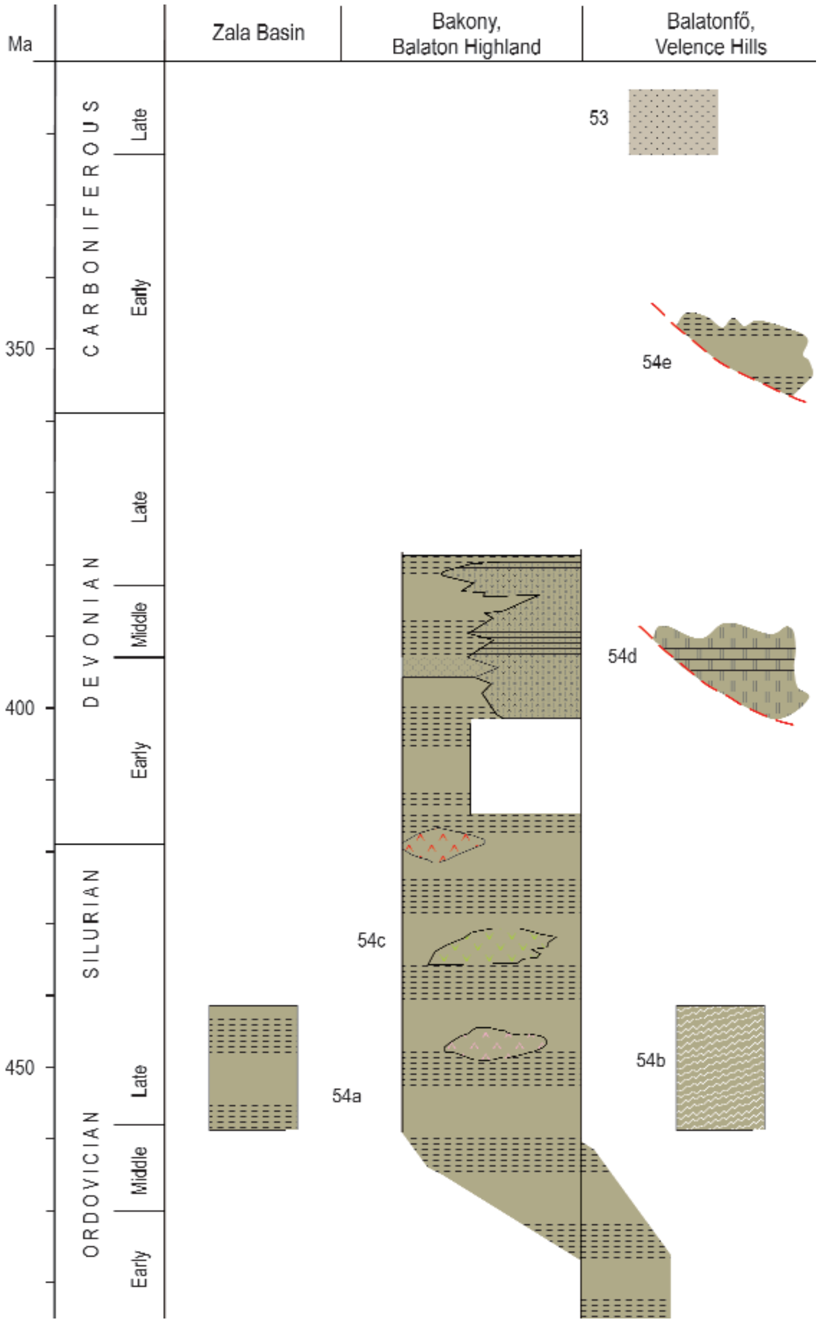


Figure 8. Variscan formations of the Transdanubian Range Unit (after FÜLÖP 1990, CSÁSZÁR in BUDAI et al. 1999)

The oldest part of the succession is made up of Ordovician and Devonian metasiltstone and slate (54a). These also contain siliceous shale and sandstone intercalations (Lovas Slate). Within the succession, the quartz phyillite at Balatonfő (Balatonfőkajár Quartz Phyillite, 54b) underwent higher-grade, greenschist metamorphism. The formation is known between the Balatonfő Line and the Balaton Line (FÜLÖP 1990, LELKESNÉ-FELVÁRI 1998). In this area, from several boreholes (Sávoly–7, Balatonhídvég Hi–1, –2, Garabonc–1), medium-grade metamorphites are also known (garnet mica schist, garnet-staurolite mica schist, andalusite-biotite-muscovite schist) (TÖRÖK K. 1992). Due to the limited amount of borehole data, neither the distribution of the rocks in the pre-Cenozoic basement, nor their relationship to the adjacent low-grade metamorphites are known; what is more, it is also doubtful, whether a Variscan or an Alpine metamorphism was involved. In the Ordovician–Devonian slates there are metavolcanic bodies of an acidic (Alsóörs Metarhyolite) and basic (Litér Metabasalt) character (54c). The Devonian limestone bodies of different facies (54d) are characteristic between the Balatonfő and the Velence Hills (GYALOG et al. 2004). The largest mass amongst these limestone bodies is the shallow marine platform carbonate (Polgárdi Limestone) that is connected to the open-basin facies (Kékkút Limestone) by a slope facies (Úrhida Limestone). There is only one occurrence of the clay shale (Szabadbattyán Clay Shale, 54e) that is tectonically in contact with the Polgárdi Limestone and which contains limestone intercalations rich in shallow marine fossils.

Upper Carboniferous continental siliciclastic formation (53)

Molassic, fluvial formations accumulated in the basins that had formed at the end of the Carboniferous, following the Variscan orogeny. These formations occur only on the Balatonfő area between Lake Balaton and Lake Velence (Majoros, in Budai et al. 1999). The succession, which reaches a thickness of 600m (Füle Conglomerate), is made up of the cyclic alternation of fanglomerate, conglomerate, sandstone, siltstone and claystone. The material is made up of pebble-sized grains and is derived mainly from the Early Palaeozoic succession (quartzite, quartz phyllite, phyllite, and rarely, metarhyolite). According to macroflora-sporomorph assemblages, the age of the succession is Late Carboniferous (Westphalian–Stephanian).

Alpine tectonic cycle

Permian – Early Cretaceous evolutionary cycle

In the Transdanubian Range Unit the Alpine tectonic cycle began in the Permian with a continental rifting, during which the uneven subsidence resulted in the formation of basins and highs. On the overwhelming part of the unit the basement, which had been metamorphised and deformed during the Variscan orogeny, was denuded, and in the Middle Permian terrestrial-fluvial sediments started to accumulate on the dissected surface (Figure 9). According to borehole data, this was locally preceded by dacite volcanism (Majoros 1998).

In the Late Permian, only the north-eastern part of the unit (north-east of Lake Velence) was part of the wide western margin of the Tethys, but by the very beginning of the Triassic period the whole unit became part this margin, represented by a shallow sea. The rifting related to the opening of the Neotethys reached the region in the Middle Triassic, resulting in the differentiation of the shelf; as a result of this process shallow platforms and deeper basins evolved (Budai, Vörös 1992, 2006). In the early stage of the Late Triassic most of the basins were filled-up and extensive carbonate platforms dominated the overwhelming part of the unit (Haas, Budai 1995). In the Early Jurassic the Neotethys was still — and the Alpine Tethys (Ligurian–Penninic oceanic basin) was already — in the phase of the formation of oceanic crust. In the Transdanubian Range Unit situated between these two oceanic realms, the extension resulted in the formation of deeper basins and, between the basins, highs. From the end of the Early Jurassic till the later stage of the Early Cretaceous the unit was a deep marine basin with an uneven basement, mostly characterised by carbonate sediment accumulation (which occurred during the course of the Jurassic). The obduction of the oceanic basement during the Cretaceous caused a significant change in the nature of sedimentation in the north-eastern part of the Transdanubian Range Unit. Fold deformation was related to the compression stage at the end of Early Cretaceous, when the synform structure of the Transdanubian Range Unit also developed. This structure significantly determined subsequent denudation and sediment accumulation. Most parts of the unit were subjected to subaerial exposure at the beginning of the Albian.

Upper Carboniferous – Lower Permian granitoid plutons (52)

Post-collisional granitoid intrusions are known along the Balaton Line. They are of crustal origin with alkaline, granodiorite features (Buda 1985). The intrusion outcropping in the Velence Hills was solidified at a hypabyssal depth and is made up of biotitic orthoclase granite. In the latter smaller pegmatite bodies, as well as microgranite and granite porphyry dykes, are also to be found (Velence Granite). Similar rocks are also known to occur beneath the surface in the southern foreland of Lake Velence and that of Lake Balaton. Their radiometric age ranges from 260 to 320 Ma, though more recent measurements have indicated an Early Permian age (274–282 Ma) (Gyalog et al. 2004).

In the same zone, deep boreholes have also exposed quartz diorite (Gárdony Quartz Diorite). According to K-Ar dating, its age is 272 ±11 Ma (Balogh Kad. et al. 1983).

Middle – Upper Permian continental siliciclastic formation (51)

This typically red, terrestrial, dominantly fluvial succession is deposited unconformably on the uneven Variscan basement (Balatonfelvidék Sandstone). The lower part of the unit comprises alternating polymict conglomerate and sandstone. The unit passes upward into a succession of sandstone and siltstone (Majoros 1983, Fülöp 1990, Majoros in Budai et al. 1999). Its thickness is 200–800m on the Balaton Highland, and only 50–150m in the north-eastern part of the Transdanubian Range Unit. According to sporomorphs, its age is Middle(?) – Late Permian.

Upper Permian shallow marine carbonates and evaporites (50)

The north-eastern part of the Transdanubian Range Unit was flooded by sea during the Late Permian, as a result of which peritidal and shallow lagoonal environments evolved (Majoros 1980, Fülöp 1990). In the peritidal zone, under warm, arid

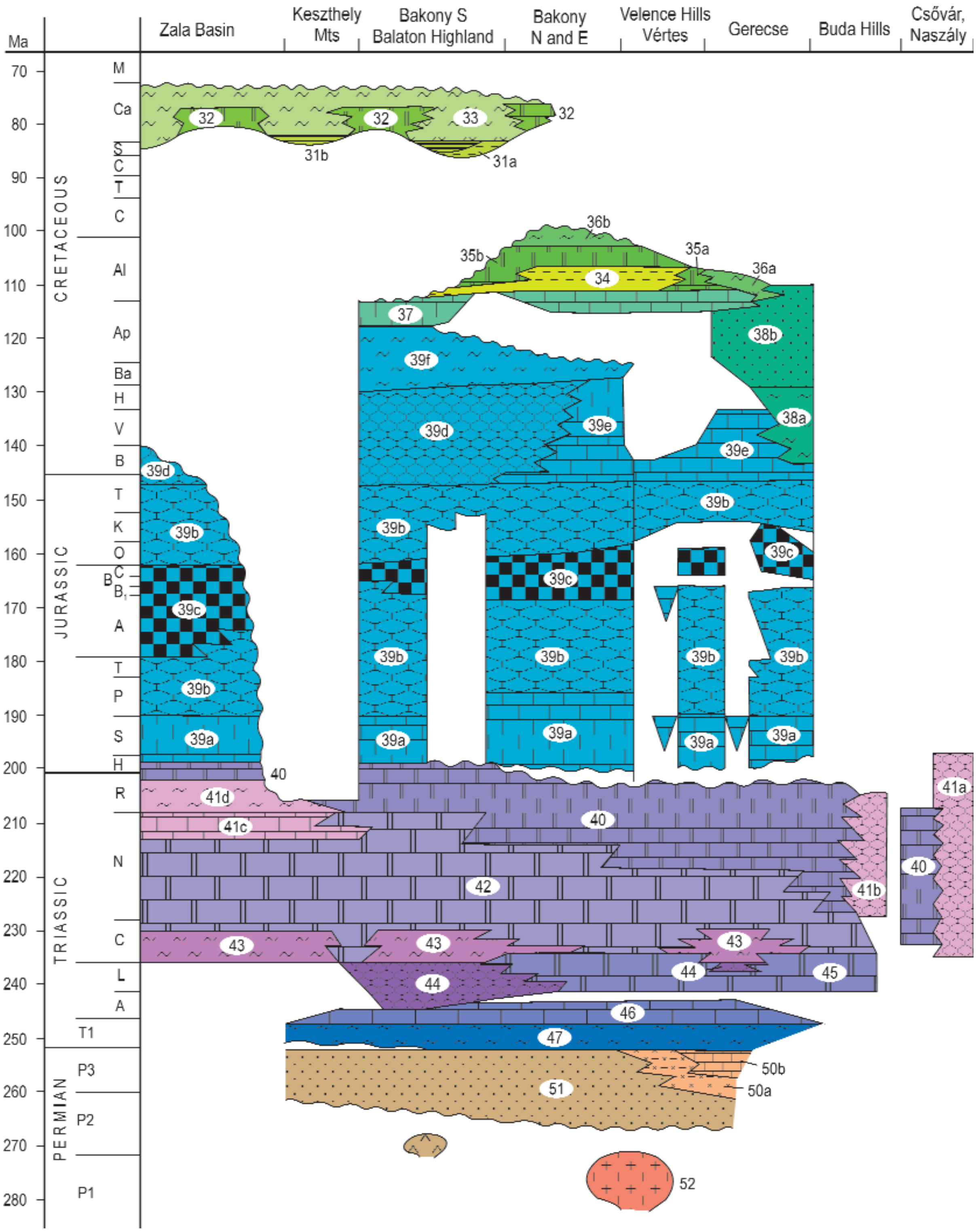


Figure 9. Stratigraphic column of the pre-Cenozoic Alpine formations of the Transdanubian Range Unit (after HAAS, BUDAI 1995, 1999, 2004; VÖRÖS, GALÁCZ 1998; CSÁSZÁR ed. 1996)

climatic conditions a succession of alternating siltstone, marl, dolomite and anhydrite was formed, representing sabkha facies (Tabajd Anhydrite, 50a). In the more inner part of the lagoon a succession of cyclic alternating marine fossil-bearing dolomite and anhydrite beds was deposited (Dinnyés Dolomite, 50b). None of these formations appear on the surface, but they have been exposed by many boreholes. The thickness of the former is 200–250m, which of the latter is 450–500m. Their Late Permian age was verified by sporomorphs, as well as by calcareous algae and foraminiferans (HAAS et al. 1986, GÓCZÁN et al. 1987).

In some parts of the Transdanubian Range Unit the basement has been penetrated by only a few boreholes in which, occasionally, even the stratigraphical classification of the pre-Cenozoic formations is doubtful. For the representation of such basement segments the **"Upper Palaeozoic and Mesozoic formations in general" (49)** label and mark is used.

Lower Triassic shallow marine fine siliciclastic and carbonate formations (47)

Because of the eustatic sea level rise at the beginning of the Triassic, an extensive shallow marine ramp evolved. On the ramp siliciclastic and carbonate sedimentation took place. The inflow of the debris was intensified in the middle of the Early Triassic, followed by a sea level drop and climate change (increasing aridity). An open shelf-sea developed in the last phase of the Early Triassic as a result of the transgression that, at the end of the Early Triassic, was followed by a new regression event (HAAS, BUDAI 1999).

In the north-eastern part of the Transdanubian Range (in the southern foreland of the Vértes and in the vicinity of Lake Velence) the Lower Triassic limestone and calcareous marl succession (Alcsútdoboz Limestone) overlies the Upper Permian formations of marine facies (HAAS et al. 1988b). The thickness of the formation is 150–200 m. South-westward, this formation interfingers with a succession of marl, calcareous marl and limestone (Arács Marl); this is characteristic for the north-eastern part of the Veszprém Plateau and the Balaton Highland. The thickness of the marly succession is 80–120m. In the south-western part of the Balaton Highland and in the foreland of the North Bakony, coastal lagoonal sandy dolomite and sandstone are deposited with erosional unconformity on the Upper Permian red sandstone (Köveskál Dolomite), with a thickness of around 100m.

The lower section of the Olenekian succession overlying the Induan formations comprises 70m-thick shallow marine red siltstone (Zánka Sandstone) and 40m-thick lagoonal dolomite (Hidegkút Dolomite). The uppermost part of the Lower Triassic succession is a 150–200m-thick sequence consisting of marl and siltstone of basin facies, with intercalating limestone and sandstone (Csopak Formation) (HAAS, BUDAI 2004).

Anisian shallow marine limestones and dolomites (46)

On the carbonate ramp which evolved at the beginning of the Middle Triassic, the sedimentation took place first on arid tidal flat, later in a lagoon of a poorly-oxygenated basement. Eventually, a shallow marine carbonate shelf developed (BUDAI et al. 1993, 1999).

An approximately 250m thick sequence of thinly bedded-lamellar cellular dolomite is lying on the Lower Triassic clastic succession (Aszófő Dolomite). The lower part of the overlying, 250–300m-thick bituminous limestone is lamellar, while the upper part is bedded and has been affected by bioturbation (Iszkahegy Limestone). The thick-bedded dolomite overlying the limestone is the last member of the Lower Anisian succession, ranging in thickness from 20m to 200m (Megyehegy Dolomite). In the middle part of the Balaton Highland and in the north-eastern part of the Transdanubian Range the middle section of the Anisian stage is made up of cyclic shallow marine platform carbonates with a thickness of 80–150m (Tagyon Fm).

Ladinian–Carnian platform dolomites (45)

Cyclic, peritidal-shallow subtidal dolomite overlies the Anisian platform carbonates (Budaörs Dolomite). The maximum thickness of the Late Anisian – Early Carnian dolomite in the Vértes and in the Buda Hills attains 800m (BUDAI et al. 2008). South-westward it interfingers with Anisian–Ladinian basinal facies (HAAS, BUDAI 1999, 2004).

Anisian–Ladinian basinal limestones and cherty limestones with tuff intercalations (44)

As a result of extensional tectonics activated in the middle of the Anisian, the shallow marine carbonate ramp had been disintegrated along normal faults. On the down-faulted areas half-graben basins evolved, while on areas that had remained, shallow marine carbonate platforms emerged (BUDAI, VÖRÖS 1992, 2006). Due to the sea level rise the Anisian basins continued to deepen. In them, the deposition of pelagic carbonates was accompanied by that of volcanics till the end of the Ladinian. In the meantime, on the emerged platforms, shallow marine carbonate sedimentation took place until the beginning of the Carnian.

On the Balaton Highland and in the Eastern Bakony, the Middle – Upper Anisian basin facies is represented by bedded, cherty, as well as thinly-bedded–lamellar bituminous limestone (Felsőörs Limestone) (BUDAI et al. 1999, 2001). Its thickness ranges from 10m to 150m. The overlying, uppermost Anisian succession comprises tuff and cherty limestone to a maximum thickness of 20–30m (Vászoly Formation). The Ladinian stage is made up of 30–50m-thick, bedded, cherty limestone and radiolarite (Buchenstein Limestone). This succession passes upward into a 20–50m-thick limestone that contains marl intercalations (Füred Limestone) at the top of the Ladinian .

Carnian basinal marls and limestones (43)

Possibly as a result of the climate change at the beginning of the Carnian, the deposition of pelagic carbonate sediments came to a halt, giving way to the deposition of grey clay marl, marl and silty marl in the basins that had evolved during the Middle Triassic (Veszprém Marl). Due to the change of the relative sea level, during the transgression stages the marl sedimentation extended to most of the neighbouring platforms, while the slowdown of the progress of the sea level rise resulted in the extension of the platforms (Haas, Budai 1999, 2004). At the foot of the slopes breccia and carbonate turbidite appeared, while in the inner parts of the basins, due to the inflow of a large amount of carbonate mud, limestone intercalations came into existence. The thickness of the succession in the depocentre is up to 1000m, while in the marginal zones it is only about 10m. By the Late Carnian the basins had already been filled-up and in shallow, isolated basins bituminous limestone, as well as shallow marine dolomite and limestone were deposited (Sándorhegy Formation). However, the marl intercalations in the succession indicate an increased transport of fine terrigenous material into the basins in more humid periods (Budai et al. 1999). The maximum thickness of the formation is 120m.

Carnian–Norian platform dolomites (42)

In the humid period of the Carnian the area of the once-extensive platforms decreased, but on certain areas significant isolated platforms survived. In the south-western part of the Transdanubian Range Unit (in the Keszthely Mountains and its vicinity), reef and lagoonal limestone (Ederics Limestone), as well as it is partly or totally dolomitised type is known to occur (Budai et al. 1999, Haas et al. 2014). At the same time, in the inner part of larger platforms, cyclic successions were deposited with alternating peritidal and subtidal facies. They have been fully dolomitised (Gémhegy Dolomite). Such successions are known on the Veszprém Plateau in the Bakony, as well as in the Vértes and in the Buda Hills (Haas, Budai 1995, 1999, 2014). In the later phase of the Late Carnian a large carbonate platform evolved, extending over the area of the basins that had already been filled-up. The bulk of the Transdanubian Range Unit became an inner platform where a 1–1.5km-thick cyclic succession of thick peritidal and shallow lagoonal beds were formed up until the Middle – Late Norian (Main Dolomite). The whole succession essentially underwent a syngenetic – early diagenetic dolomitisation (Haas, Budai 2004).

Norian–Rhaetian and lowermost Jurassic basinal cherty limestones, dolomites and marl (41)

Over the course of the Carnian and the Norian, extensional basins evolved in the north-eastern part of the Transdanubian Range Unit (namely, in that part of the Neotethys that is situated closer to the shelf margin) as a result of the intensive subsidence of a part of the platforms along fault lines. The cherty dolomite and limestone of slope and basin facies formed in the basin of the Carnian are known east of the Danube, in the vicinity of Csővár (Csővár Formation, 41a). Their formation lasted right up until the Early Jurassic. The thickness of the formation is 600–700m (Haas et al. 1997). It can be assumed that the formation of the cherty, frequently laminitic dolomite and limestone in the Buda Hills also started in the Carnian (Mátyáshegy Formation, 41b), in isolated intraplatform basins (Haas et al. 2000a). In the basin which evolved in the Pilis over the course of the Norian, thinly-bedded, lamellar limestone was deposited (Feketehegy Formation), the thickness of which is around 300m (Haas et al. 2010b).

During the Late Norian, extensional basins began to evolve also in the south-western part of the Transdanubian Range Unit, in that part of the future Alpine Tethys that is situated closer to the margin (Haas 2002). As a consequence, the deposition of the Main Dolomite was replaced by that of a thinly-bedded–lamellar bituminous dolomite, the formation of which was restricted to an isolated basin (Rezi Dolomite, 41c). Its thickness is 100–300m (Budai et al. 1999). At the very end of the Norian the dolomite formation was replaced by one of organic-rich clay marl, marl and silty marl (Kössen Formation, 41d). This can probably be attributed to a more humid climate. The deposition of the clayey sediments continued during the Rhaetian, resulting in the deposition of a succession in the depocentres, reaching a total thickness of ca. 400m. On the margins, a succession of alternating marl and limestone was deposited. This reflects the periodic change of the sea level, which was accompanied by significant platform progradation in the later stage of the Rhaetian (Haas 2002, Haas, Budai 2004).

Upper Triassic – lowermost Jurassic platform limestones (40)

From the Late Carnian till the end of the Triassic, the overwhelming part of the Transdanubian Range Unit was a carbonate platform (Haas, Budai 1995, 1999). Until the Middle Norian, in the inner part of the platform, syngenetic and early diagenetic dolomite was formed under semiarid conditions (Haas 2004). However, in the outer zone of the platform (which rarely got close to the peritidal zone) no dolomite was formed; here only the oncoid facies of the Dachstein Limestone is present, representing a heteropic facies of the Main Dolomite. This formation was formed during the Late Carnian – Norian in the Csővár area, and during the Norian – Early Rhaetian in the Buda Hills (Haas, Budai 2014). In these areas the formation is known to occur to a thickness of several 100m. Presumably due to the more humid climatic conditions, from the Middle Norian the dolomitisation of the sediments of the inner platform gradually ceased and on the overwhelming part of the Transdanubian Range Unit a ca. 1000m-thick limestone succession was developed. This comprises alternating peritidal and subtidal beds. The formation of this limestone lasted till the end of the Rhaetian (Lofer cyclic Dachstein Limestone) (HAAS 2004).

In the north-eastern part of the Transdanubian Range Unit the carbonate platform was flooded at the Triassic–Jurassic boundary, but in the south-west the development of the platform proceeded even in the Early Jurassic (Hettangian). Furthermore, in a shallow subtidal environment of ca. 150m-thick limestone (Kardosrét Limestone) accumulated until the platform was finally flooded in the Early Sinemurian (VÖRÖS, GALÁCZ 1998).

Jurassic and Lower Cretaceous pelagic limestones, cherty limestones and radiolarite (39)

The extensive carbonate platform formed during the Late Triassic gradually disintegrated at the beginning of the Jurassic period. On emerged areas, submarine ridges (in tectonically opened trenches) and deep marine basins evolved (GALÁCZ 1988, VÖRÖS 1998, VÖRÖS, GALÁCZ 1998). In the basins continuous but slow sedimentation took place, while on the submarine ridges the deposition of the sediments was periodic. In the south-western part of the Transdanubian Range the formation of pelagic carbonates continued even over the course of the Early Cretaceous. During the Hauterivian–Barremian, however, inflow and accumulation of terrigenous debris began.

The Lower Jurassic succession overlying the Upper Triassic – Lower Jurassic platform carbonates, comprises thick-bedded (Pisznice Limestone, 39a) and cherty limestone (Isztimér Limestone, 39a). The upper section of the Lower Jurassic, as well as the bulk of the Middle Jurassic and the lower section of the Upper Jurassic, is represented by clayey, nodular, ‘ammonitico rosso’ limestone (Tűskövesárok Limestone, Törökbükk Limestone, Kisgerecse Marl, Tölgyhát Limestone, Pálihálás Limestone, 39b) (VÖRÖS 1998, VÖRÖS, GALÁCZ 1998, CSÁSZÁR et al. 1998). The pelagic basin facies of the Middle Jurassic is represented by the thinly bedded-lamellar ‘bositra limestone’ (Eplény Limestone) and lamellar-siliceous marl and cherty limestone (Lókút Radiolarite, 39c). The total thickness of the Jurassic succession exceeds 300m in the area of the Bakony basins, while it barely reaches 20–50m in the area of the ridges and in the Gerecse.

The thinly-bedded, cherty limestone (‘maiolica’) facies of the Upper Jurassic – Lower Cretaceous is characteristic in the south-western part of the Transdanubian Range (Mogyorósdomb Limestone, 39d). Its maximum thickness attains 150m (HAAS et al. 1984). Its heteropic facies are significantly thinner in the northeastern region (Szentivánhegy Limestone and Borzavár Limestone, 39e). The upper section of the Lower Cretaceous is made up of pelagic marl in the Bakony, the thickness of which reaches 250–300m in the southwest (Sümeg Marl, 39f).

Lower Cretaceous flyschoid formations (marl, sandstone, conglomerate) (38)

There was a significant change in the nature of sedimentation in the Early Cretaceous in the area of the Gerecse Basin; thick, flysch-type, deep basinal clastic sediments were accumulated until the beginning of the Albian (SZTANÓ 1990, CSÁSZÁR, ÁRGYELÁN 1994, FOGARASI 1995).

In the basement of the Gerecse, the Tatabánya Basin and the Dorog Basin, the lower section of the Lower Cretaceous is made up of silty clay marl and calcareous marl of turbidite facies, with limestone and graded sandstone intercalations (Bersek Marl, 38a). Above it, thick-bedded, coarse-grained, graded sandstone is deposited, representing abyssal fan facies, with characteristic channel-filling coarse debris at its upper section (Lábatlan Sandstone, 38b). The maximum thickness of the Lower Cretaceous sequence can reach 600m, within which that of the sandstone can be up to 500m.

Aptian–Albian shallow marine limestone (37)

Crinoidal limestone (Tata Limestone) occurs in the axis zone of the syncline of the Transdanubian Range Unit from the western foreland of the Gerecse till the South Bakony; it was formed under shallow marine conditions in the Late Aptian – Early Albian (LELKES 1990). Its thickness exceeds 200m. It frequently contains older Cretaceous and Jurassic rock fragments of sand-size and locally, also siliciclasts. In the south-western part of the Transdanubian Range Unit it continuously develops from the pelagic Sümeg Marl (HAAS et al. 1984). At other places it is deposited on the older Cretaceous and Jurassic formations with erosional disconformity and with significant hiatus.

Albian–Cenomanian evolutionary cycle

Following the formation of the Tata Limestone, at the beginning of the Early Albian tectonic processes in the compression field led to the development of the syncline structure of the Transdanubian Range Unit and the fold deformation of the formerly deposited sequences. As a result of the tectonic processes, the overwhelming part of the Transdanubian Range Unit was subjected to subaerial exposure and consequently, denudation. However, in the north-eastern part of the Transdanubian Range Unit (in the Gerecse) the marine sedimentation was continuous. The transgression which started here again resulted in the flooding of the overwhelming part of the unit in the middle of the Albian, so that by the Late Albian it had become a pelagic basin. However, new tectonic movements in the Turonian–Coniacian led to the subaerial exposure of the area once more, resulting in the denudation of the bulk of the deposited sediments.

Albian continental, lacustrine and lagoonal formations (34)

The western part of the northern foreland of the Vértes, as well as the Bakony area, was subjected to subaerial exposure and denudation during the Early Albian. Locally, the Triassic and Lower Jurassic platform limestone was exposed subaerially and karstified. In the karstic depressions bauxite accumulated (Alsópere Bauxite). Onto the eroded surface — and locally, onto the bauxite — fluvial, lacustrine, swamp and marine grey and variegated clay, clay marl, marl and limestone, and (locally) sandstone were deposited, mostly in brackish water (Tés Clay Marl). The thickness occasionally exceeded 200m (CSÁSZÁR 1986, BUDAI et al. 2008).

Albian platform limestones (35)

The eastern part of the northern foreland of the Vértes was the margin of the Albian basin in which calcarenitic shallow marine limestone was deposited, followed by coral and stromatopora-bioherm–bearing rudistid limestone (Környe Limestone, 35a), to a thickness of 200m (CSÁSZÁR 1986, 2002; BUDAI et al. 2008). During the transgression process the sedimentation of the succession of lagoonal facies gave way to that of the shallow marine, typically rudistid limestone, from the northern foreland of the Vértes up to the Bakony (Zirc Limestone, 35b). The thickness of the successions representing the different shallow marine facies varies between 25 and 250m.

Albian–Cenomanian basinal marls (36)

In the western part of the Gerecse and in its western foreland the Aptian–Albian crinoidal limestone is continuously overlain (locally along with the Lábatlan Sandstone) by dark grey siltstone, clay marl and marl (Vértessomló Siltstone, 36a); this formed in a shallow, bathyal environment. The thickness of the formation can exceed 200m (CSÁSZÁR 1986, BUDAI et al. 2008). In the Bakony and in the forelands of the Vértes, situated on the Albian shallow marine carbonate rocks, Late Albian – Cenomanian grey silty marl is deposited, representing deep, pelagic facies. In the upper part of the succession siltstone and sandstone occur (Pénzeskút Marl, 36b). The thickness of the formation exceeds 450m.

The map depicts the pelagic siltstone and marl formations — which are lithologically similar, but separated in space and time — with a similar symbol.

Senonian evolutionary cycle

The structural development and subtropical denudation in the uplift stage preceding the Senonian sediment accumulation cycle resulted in a highly dissected morphology (HAAS 1983, KAISER 1997). Roughly parallel with the strike of the Bakony, structural highs and basins were formed. During the denudation, on the karstified surface of the basement exposed at the surface, bauxite was deposited (MINDSZENTY et al. 2000); while on the area of the basins fluvial, lacustrine and swamp sediments were deposited. Later, the slopes between the basins and the highs, as well as the highs themselves were flooded by the sea. On the highs, platform carbonates were formed, while in the basins pelagic carbonates and clayey rocks developed. Later, the whole area became a pelagic sedimentary basin.

The Senonian cycle is closed by a tectonically-determined uplift and as a consequence, by a significant denudation. The formations that had formed during the cycle were preserved in the south-western part of the Transdanubian Range Unit, as well as in the western part of the North Bakony, in the South Bakony, in the southern marginal part of the Danube Basin, and in the North Zala Basin.

Senonian continental siliciclastic and swamp formations (31)

Formations in this unit include (i) those formed in the early stage of the Senonian cycle in the depressions of the karstic basement and in the terrestrial, fluvial and lacustrine sedimentary basin and (ii) later, those deposited on the delta plain, as well as on the area of coastal swamps (HAAS 1998a).

Bauxite sediments (Nagytárkány Bauxite and Halimba Bauxite Formations) fill karstic depressions (deep dolines and shallower depressions) or occasionally, they occur over a larger area in the stratiform bodies. Their thickness usually varies between 10 and 30m, though in certain dolines the thickness can attain 100m.

On the pre-Senonian basement and the bauxite, as well as on the coal-bearing beds in certain basins, a succession of alternating variegated clay, clay marl, siltstone and sandstone is deposited, locally containing gravel and conglomerate beds (Csehbánya Formation, 31a). The succession represents fluvial and delta plain facies (JOCHA-EDELÉNYI 1988). Its thickness is 100–200m; in the western part of the North Bakony it is thicker and more coarse-grained, while south-westward, its thickness, as well as its grain size decreases. Locally, it is rich in vertebrate fauna (MAKÁDI et al. 2006, ŐSI, RABI 2006). According to palynological data, its age is Santonian.

On the pre-Senonian basement, bauxite or terrestrial sediments in certain sub-basins are overlain by alternating brown coal, coal-bearing clay, clay, clay marl, calcareous marl, limestone, siltstone and sandstone (Ajka Coal, 31b). The succession usually starts with a fresh-water section, followed by sediments formed in a mixed-water lagoon environment and in a related area of coastal swamps. The thickness of the Santonian formation occurring in the South Bakony and in its north-western foreland varies between 30 and 100m.

Senonian platform limestone (32)

Highs that were formed due to the tectonic disintegration and erosion preceding the Santonian transgression were flooded by the sea only during the Campanian. On these areas the dominant feature was the formation of shallow marine limestone, consisting of rudist bivalves and their fragments (Ugod Limestone). The thickness of the formation can attain 400m (HAAS 1979).

Senonian basinal limestones and marls (33)

During the transgression in the Early Campanian, on the fluvial-lacustrine, delta plain and swamp facies, shallow marine (first, in brackish water, later in normal saline water) grey marl was deposited (Jákó Marl); its thickness can reach 100m (HAAS 1983, 1998a). Due to the continuing sea level rise deep marine calcareous marl, marl and silty marl were deposited; initially, this occurred only in the sub-basins that were separated by the highs while later, following the drowning of the carbonate platforms (that had evolved on the highs) in the late stage of the Late Campanian, it took place over the whole area of the basin (Polány Marl). The maximum thickness of this formation is 800m, but probably its extent and thickness were significantly higher than today, since after its formation the area was uplifted and during a long subaerial period it was partly or totally eroded. Its oldest overlying formation is Middle Eocene.

MID-HUNGARIAN MEGA-UNIT

The Mid-Hungarian Mega-unit is a narrow zone between the Alcapa and the Tisza Mega-units. Its heterogeneous neo-Palaeozoic and Mesozoic formations can be related to the facies units of the Carnic Alps, the South Karawanks, the Julian Alps, the Zagorje region, as well as to some of the facies units of the Dinarides. The mega-unit consists of the Mid-Transdanubian Unit, the Bükk Unit, the Szarvaskő–Darnó Unit and the Szendrő–Uppony Unit. According to their facies features, these units could have originated from the province of the Tethys margin situated between the South Alpine and the Dinaric segments. In the Cenozoic the Alcapa and the Tisza Mega-units drifted next to each other. As a consequence, some parts of the mega-units were torn from their place of origin and by dislocations along shear planes moved close to each other before the formation of the Pannonian Basin began (BALLA et al. 1987, CSONTOS, NAGYMAROSY 1998, FODOR et al. 1998).

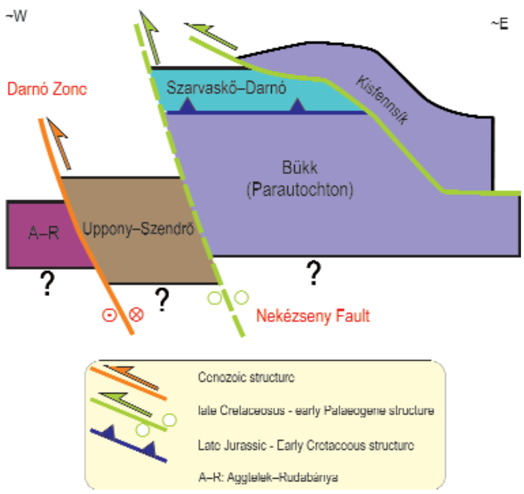


Figure 10. Structure of the northern part of the Mid-Hungarian Mega-unit

SZENDRŐ–UPPONY UNIT

The north-western boundary of the unit towards the Aggtelek–Rudabánya Unit is represented by the north-western and south-eastern marginal faults of the Darnó Zone (Figure 10). Its contact with the Bükk Unit is known only at the southern margin of the Uppony Hills. This is where the Permo-Triassic formations of the Bükk were thrust over the Senonian Gosau conglomerate (overlying the Palaeozoic Upponyian complex) at the very end of the Cretaceous or in the Palaeogene. The eastern boundary of the structural unit towards the Zemplén Unit is unknown.

The Szendrő–Uppony Unit is dominantly made up of Middle Devonian – Middle Carboniferous sedimentary formations belonging to the Variscan cycle (Figure 11). They were affected by metamorphism of greenschist facies during the Alpine orogeny. These formations (which have also been exposed by many deep boreholes in the broader vicinity of the Uppony and the Szendrő Hills) have similar characteristics in the area of the two major superficial occurrences, and thus they will be discussed together.

The structural relationship of the Szendrő–Uppony Unit with the neighbouring units is uncertain; on the basis of the stratigraphical features, the Szendrő–Uppony Unit is considered to be the original basement of the Bükk. Its inner structure is determined by the eo-Alpine NW-vergent multi-phase folding and the associated small-scale imbrication (Kovács, Péró 1983, Koroknai 2004).

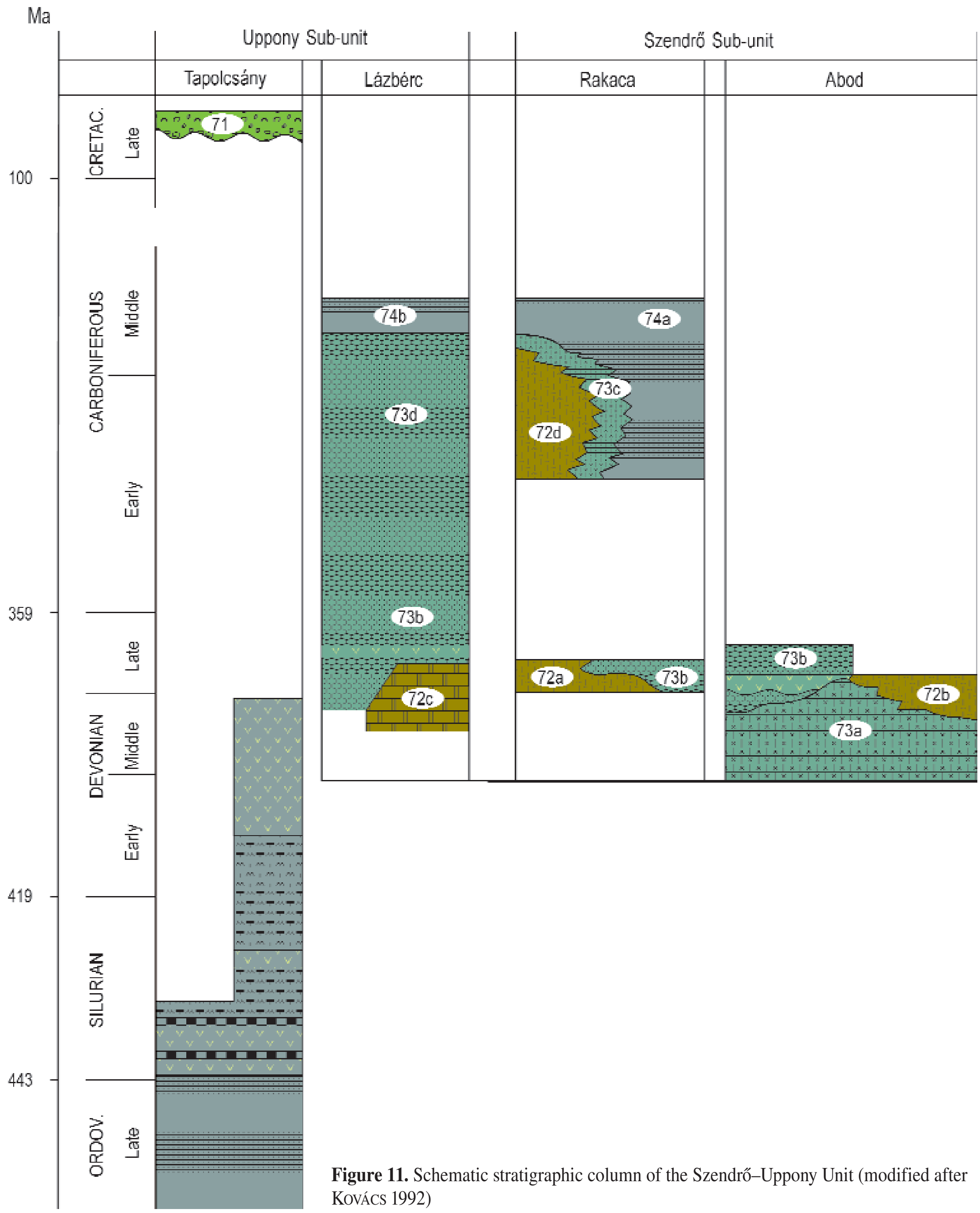


Figure 11. Schematic stratigraphic column of the Szendrő–Uppony Unit (modified after Kovács 1992)

The low-grade metamorphism (300–450 °C and 2–3 kbar; ÁRKAI 1983) which affected the unit occurred during the eo-Alpine phase (K-Ar muscovite: 110–120 Ma; ÁRKAI et al. 1995a). The grade of the metamorphism is smaller in the area of the Palaeozoic of the Uppony Sub-unit (namely, the boundary of the very low-grade and low-grade metamorphism) than in the Palaeozoic of the Szendrő Sub-unit (low-grade, locally with neomorphic biotite). There are no traces of the pre-Alpine metamorphic event in the unit. Fission track ages (zircon: ~97–99 Ma) can be interpreted as the date of the cooling following the temperature peak of the metamorphism (ÁRKAI et al. 1995a). The unit was subjected to subaerial exposure in the Late Cretaceous, as suggested by the unconformably deposited Senonian marine conglomerate.

The metasandstone–slate series that occurs at the southern part of the Uppony Hills and in the tectonic zone at the south-eastern margin of the Rudabánya Hills (where it forms smaller lenses) is assigned to the lower part of the Palaeozoic (Ordovician–Silurian?) (Tapolcsány Fm; KOVÁCS 1992, FÜLÖP 1994). The series also contains basic metavolcanite intercalations. As such, these formations do not belong to the Lower – Middle Carboniferous siliciclastic series (that will be discussed later), though they are not depicted on the map with a separate symbol. The same applies to the Middle – Upper Devonian(?) formation group of diverse rock associations that overlies the Tapolcsány Formation and is characterised by olistostrome features (Strázsahegy Fm; KOVÁCS 1992, FÜLÖP 1994).

Low-grade metamorphic Devonian–(Carboniferous?) platform carbonates (72)

The oldest elements of this succession are made up of shallow marine limestones (Rakacaszend Marble, 72a and Rakaca Marble, 72d, Bükhegy Marble, 72b, as well as Uppony Limestone, 72c). They were mostly formed in the Middle Devonian. There is no direct data to justify the occurrence of platform development in the Carboniferous (KOVÁCS 1992).

Low-grade metamorphic Devonian–Carboniferous basinal carbonates (73)

The platform limestones are overlain by Late Devonian – Early Carboniferous basinal limestones. They are coeval with the platform facies to a lesser extent (Szendrőlád Limestone, 73a), but dominantly represent a condensed successon formed after the drowning of the platform (Abod Limestone, 73b, Verebeshegy Limestone and Kopaszhegy Limestone, 73c, as well as Dedevár Limestone and Lázbérc Limestone, 73d). Among these limestones there is a characteristic rock type occurring in the area: the Upper Devonian chlorite–sericite–meshed limestone (“cipollino” facies of the Abod Limestone, 73b); this formed due to the metamorphic alteration of the carbonatic sediment and the intercalating basic tuffitic material. The basic volcanism accompanied the disintegration and the drowning of the platform.

Low-grade metamorphic Carboniferous basinal siliciclastic formations (74)

The condensed carbonate succession gradually passes into basinal siliciclastic formations (Szendrő Phyillite, 74a, as well as the clayey-sandy succession of the Lázbérc Formation, 74b). These formations represent already the Middle Carboniferous flysch stage (KOVÁCS 1992).

Senonian marine conglomerate (71)

The Late Mesozoic, post-orogenic overlying beds of the Palaeozoic rocks consist of Senonian (Campanian) conglomerate with sandstone and marl intercalations and rudistid limestone blocks. The conglomerate was deposited mostly as a result of debris flow (Nekézseny Formation). In the clastic material of the formation — known from the southern margin of the Uppony Hills — there are rock fragments which originated from the Palaeozoic of the Uppony Sub-unit and from the Aggtelek–Rudabánya Unit. However, there is no evidence for clast transport from the neighbouring Bükk Unit (BREZSNYÁNSZKY, HAAS 1984).

BÜKK UNIT

The Bükk Unit represents the parautochthonous, low-grade metamorphic Palaeozoic–Mesozoic succession of the Bükk Mountains, and is situated in a lower structural position (Figure 12.). The bulk of the carbonate – clay shale succession of the unit was formed along a continental margin. The rocks have been affected solely by Alpine metamorphism (ÁRKAI 1973, 1983); Variscan deformation and alteration have not affected any of the rocks, not even the Palaeozoic rocks. According to these indicators and the characteristic features of the succession, the Palaeozoic–Mesozoic formations of the Bükk show similarities with those of the South Alpine – Dinaric units (SCHRÉTER 1943; BALOGH K. 1964; CSONTOS 1988, 2000; FILIPOVIĆ et al. 2003).

The rocks of the Bükk Unit can be studied on the surface in the Bükk Mountains in North Hungary; borehole information is provided by the boreholes of Verpelét, Mezőkövesd, Mezőkeresztes, Emőd and Hejőpapi, located south of the Bükk.

Towards west and north-west, low-grade metamorphic rock successions have been exposed by the Nagybátony Nb–324 (Kozur 1984) and the Nagykökényes Nks–I boreholes and these undoubtedly belong to this unit.

The northern boundary of the unit towards the Szendrő–Uppony Unit is represented by the Nekézseny Thrust, along which the Bükk Unit has been thrust over the Uppony Sub-unit (Figure 10.). The age of this structural element is uncertain; it could have been formed anytime over the course of the Late Cretaceous and the Palaeogene; before its development the Bükk and the Szendrő–Uppony units had probably formed one single unit. North to Nagybátony the boundary of the unit is represented by the Diósjenő–Ógyalla (Hurbanovo) Line. It can be assumed that at the western margin of the Zagyva Trough (west from the Nagykökényes Nks–I borehole and east from the boreholes of Tura) the Transdanubian Range Unit is thrust over the Bükk Unit. In the south and in the south-west the Bükk Unit is bordered by the Transdanubian Unit with which (more precisely, with certain elements of which) the Bükk Unit shows close relationship. This is indicated by the fact that the boundary of the two units is a young Cenozoic strike-slip–reverse fault zone. The boundary of the two units south of the Bükk Mountains is the Tóalmás Line and its continuation, the southern marginal fault of the Vatta–Maklár Trough. Eastward, the unit stretches as far as the Hernád Line (Figure 1). The basement is unknown in this area; deep boreholes have not exposed identifiable pre-Miocene rocks.

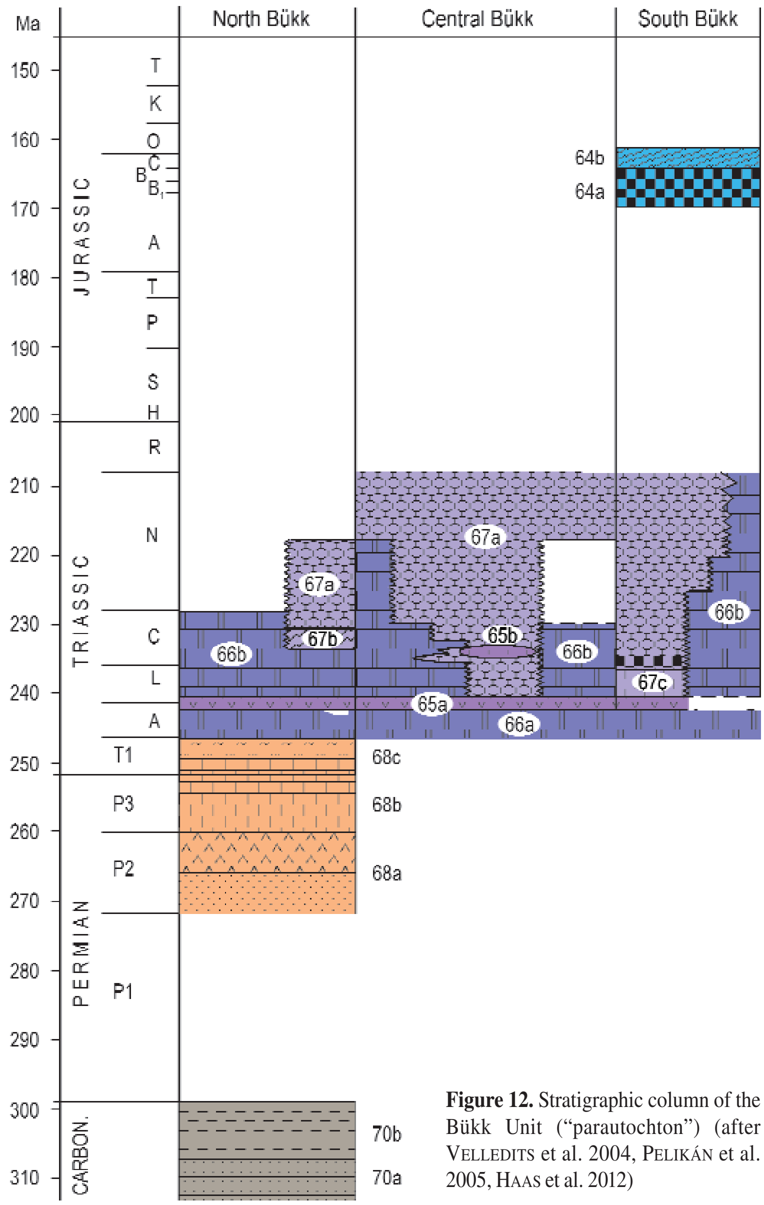


Figure 12. Stratigraphic column of the Bükk Unit ("parautochton") (after Velledits et al. 2004, Pelikán et al. 2005, Haas et al. 2012)

There are two nappe units on the Bükk Unit (Figure 10.). One of them is the Szarvaskő–Darnó (Melliata) Unit; this comprises pelagic sediments which were deposited on the ocean floor and magmatites that are differentiated from the lithosphere. The other nappe unit is the Kisfennsík ("Little Plateau") Nappe, the platform carbonate of which is very similar to that of the "parautochton"; as such, it is discussed along with the description of the Bükk Unit.

The nappe units probably juxtaposed at the time when the intra-oceanic subduction zone of the Neotethys (Szarvaskő–Darnó Unit) was thrust over onto the thinned crust (Bükk Unit). This occurred at some time during the Late Jurassic and the Early Cretaceous (Csontos 1999). The juxtaposed nappes underwent common folding and metamorphism. The age of the low-grade metamorphism (Árkai 1973, 1983, Csontos 1988) has been successfully determined at some places; the ages are in the 130–120 Ma interval (Barremian) (Árva-Soós et al. 1987). Metamorphism occurred at a depth of ca. 6–9km, under relatively cold (250–300 °C) conditions (Árkai 1983).

The original structures (whose age is the same as that of the metamorphism) are tight, foliated folds with E–W strike. On the overwhelming part of the Bükk area they have a southern vergency (Csontos 1988, 1999). Mostly, the thick platform carbonates (e.g. on the Nagyfennsík — "Great Plateau" —, in the vicinity of Répáshuta) are folded into large anticlines. The Palaeozoic – Early Mesozoic formations of the succession also form a large anticline in the Northern Bükk (Balogh 1964). On the basin areas of the Triassic there are many smaller folds. The Jurassic shales and the Szarvaskő–Darnó Unit are preserved in large synforms, mostly in the western parts (Balla 1983, Csontos 1988). Following the formation of the original folds many secondary box folds and slivers evolved, still with E–W strike. After that the Bükk Unit (along with the Szendrő–Uppony Unit) was dissected by a large-scale, partly ductile strike-slip fault system that has given a listric character to the original structures. The sinistral dislocations along the ancient Darnó Zone and the dextral dislocations along the

Bükkszentkereszt Zone (CSONTOS 1999) dragged the original structures and shaped the characteristic structure of the mountains, which resembles a downward-pointing moustache. Along the Bükkszentkereszt Zone the age of the shearing can be placed at the very end of the Late Cretaceous, and the very beginning of the Cenozoic (younger than 76 Ma) (KOROKNAI et al. 2008).

The pre-Cenozoic rock complex became subaerially exposed over the course of the Eocene. Even after that many deformations (e.g. normal and reverse fault, strike-slip fault) affected the area. Among them the most important ones are: (i) the presumably Palaeogene right-lateral strike-slip fault along the Darnó Zone (SCHMID et al. 2008); (ii) the reverse and strike-slip faults along the Darnó Zone and the Nekézseny Thrust (TELEGDI ROTH 1951); and (iii) the Late Miocene normal fault along the Darnó Zone (FODOR et al. 2005) and the Hernád Zone.

Large areas of the unit in the deep zones of the Zagyva Trough and Vatta–Maklár Trough, as well as in the basement of the Mátra, are made up of Bükk-type rocks — that is to say, rocks that have undergone low-grade Alpine metamorphism. Their classification is uncertain. These areas are depicted on the map as "**Very low-grade metamorphic Upper Palaeozoic and Mesozoic formations in general (69)**".

Variscan tectonic cycle

There is no trace of Variscan deformation in the Bükk Unit, but there are Palaeozoic rocks representing the Variscan tectonic cycle.

Very low-grade metamorphic Upper Palaeozoic marine formations (70)

The oldest elements of the unit are Carboniferous slates. The successions that show similarities with the Southern Alpine facies were deposited first in a deep marine environment (Szilvásvárad Fm = Hochwipfel flysch), and later in a gradually shallowing environment (Mályinka Fm) (PELIKÁN et al. 2005). The pelagic formation is a turbidite succession (70a) made up of siltstone and claystone, and it also contains sandstone intercalations. In some levels, in the shallow marine slate–metasilt–fine-grained metasandstone succession (70b), there are calcareous–dolomitic lenses, rich in fossils (corals, phylloid algae and larger foraminifera). Their individual thickness can reach 50m. The carbonate lenses were formed on shallow marine ramps. Upward in the succession coarse conglomerate interbeddings (Tarófő Conglomerate) frequently occur. These are polymict, well-rounded beds, dominantly consisting of quartz and quartzite pebbles (FÜLÖP 1994). The age of the formation is Late Carboniferous (Bashkirian–Moscovian–Gzhelian).

Alpine tectonic cycle

Permian–Jurassic evolutionary cycle

The cycle started with the development of a shallow marine ramp that later, following the rifting related to the opening of the Neotethys, disintegrated (VELLEDITS 2000, VELLEDITS et al. 2004). The resulting platform-basin system existed until the end of the Triassic, while in the Jurassic the whole area further subsided. This resulted in the formation of a deep marine, redeposited slope and toe-of-slope sediments (olistostromes), as well as condensed pelagic sediments.

Very low-grade metamorphic Upper Permian – Lower Triassic shallow marine limestones, sandstones, marls (68)

The Carboniferous slates are overlain — with erosional contact and hiatus — by clastic–evaporitic Permian beds (Szentlélek Fm = Val Gardena Sandstone), onto which carbonatic beds (Nagyvisnyói Limestone = Bellerophon Limestone) are deposited (FÜLÖP 1994, PELIKÁN et al. 2005). The sandstone is medium-grained, red or faded yellow; upward it alternates with anhydritic chloritic green slate (68a). In the limestone intercalations of the evaporitic member, Middle Permian ostracods have been found (KOZUR 1990).

The evaporitic member is overlain by well-bedded, dark grey-black bituminous limestone with claystone intercalations (Nagyvisnyó Limestone, 68b). The succession was formed on a shallow marine ramp and it is rich in Late Permian fossils, mainly green algae and foraminifera.

The Permian-Triassic boundary within the continuous shallow marine succession can be precisely drawn by bio- and chemostratigraphic methods (HAAS et al. 2007). Above the boundary several metres-thick stromatolitic succession occurs (HIPS, HAAS 2006). The shallow marine environment lasted throughout the Early Triassic (68c). The lower part of the Lower Triassic is made up of ooidic limestone (Gerennavár Limestone) that passes upward into a succession of carbonates and clay shale (Ablakoskővölgy Fm) (PELIKÁN et al. 2005).

Low-grade metamorphic Middle and Upper Triassic platform carbonates (66)

The ca. 400m thick dolomite making up the lower part of the Middle Triassic (Hámor Dolomite, 66a) occurs practically everywhere in the Bükk Mountains. The grey, thick-bedded rock contains foraminifera and green algae, and there are also stromatolitic beds within it (Pelikán et al. 2005). Its general distribution indicates that at the beginning of the Middle Triassic an extensive, uniform carbonate platform came into existence.

During the Middle and Late Triassic a thick limestone succession was deposited on the carbonate platforms that had evolved following the Middle Triassic rifting (66b). The rock is white, light grey. Its original bedded structure is hardly recognisable due to the deformation and rock alteration; locally re-crystallised remnants can be observed, somewhat resembling green algae or coral sections (Fehérkő Limestone, Bükkfennsík Limestone). The thickness of the massive limestones can attain several hundred metres. In the Southern Bükk the thick-bedded platform limestones are least metamorphosed and contain relatively well-preserved Ladinian-Carnian fossils that, locally, indicate a reef environment (Balogh K. 1964, Velledits 2000, Velledits et al. 2004).

The Kisfennsík Limestone occurs north of the Garadna Valley, in a tectonic situation above the formations of the Szarvaskő–Darnó and the Bükk units (Forián Szabó, Csontos 2002). The less metamorphosed thick-bedded, light grey limestone contains stromatolitic beds, locally Megalodus sections and Carnian foraminiferans (Bérczi-Makk in Pelikán et al. 2005).

Middle – Upper Triassic metavolcanites (65)

In the Anisian and in the Early Ladinian, a thick rhyolite–andesite series (Szentistvánhegy Meta-andesite, 65a) was deposited locally on the carbonate platform. The calc-alkalic series — consisting of lava flows, ignimbrites and tuffs (Szoldán 1990, Harangi et al. 1996) — is characteristic for the whole Bükk Unit and it shows pronounced similarities with the South Dinaric and South Alpine region. The volcanism was related to a rifting event, during which the uniform carbonate platform disintegrated and as a consequence, smaller remnant-platforms evolved (Csontos 2000, Velledits 2000, Velledits et al. 2004).

In the northern part of the Bükk and in the vicinity of Odorvár, basic volcanic rocks can be found in the Carnian sedimentary rocks (65b). The composition of these metabasalts (Létrás and Szinva Metabasalt) is alkalic-tholeiitic (Szoldán 1990). It is can be assumed that they are related to the rifting event. The formation is made up of microcrystalline lava rocks, as well as siltstone-rich tuffs.

Very low-grade metamorphic Middle – Upper Triassic cherty limestones of toe-of- slope and basin facies (67)

The rifting event in the Middle Triassic resulted in the formation of large intra-platform basins in which pelagic sedimentation occurred from the Ladinian. The characteristic formation of the succession is the well-bedded cherty limestone (Felsőtárkány Limestone) which contains micritic black or dark brown chert bands and layers (67a). The Ladinian – Early Rhaetian age of the rock was determined on the basis of conodonts and foraminiferans (Velledits 2000, Velledits et al. 2004). The cherty limestone is highly foliated and folded.

There are slight differences concerning the evolution of the basin areas of the Bükk. In the southeastern Bükk the deposition of the cherty limestone above the Ladinian limestone and radiolarite (Várhegy Fm, 67c) lasted till the end of the Late Triassic. In the Northern Bükk a carbonate platform had evolved that later, due to the rifting event, subsided. Initially clay shale (Vesszős Fm, 67b) was formed, but later, in the Norian, basinal cherty limestone occurred (Kozur, Mock 1977, Csontos 2000).

Very low-grade metamorphic Middle – Upper Jurassic pelagic formation (radiolarite, slate) (64)

Over the course of the Jurassic the sedimentary basin of the Bükk further deepened, resulting in the drowning of the former platform areas and the development of a starving pelagic environment.

The Upper Triassic carbonates are overlain by thin basinal succession of diverse lithology (Répáshuta Limestone). The martix of the rock is yellowish, pink, lilac, thinly-bedded, foliated pelagic limestone that locally contains echinodermata fragments and red chert layers. In it dominantly Upper Triassic (Norian), shallow marine-origin limestone clasts of variable size occur. The olistostrome is probably Lower Jurassic (Riedel et al. 1988), or perhaps uppermost Triassic.

This is overlain by red radiolarite (Bányahegy Radiolarite) which occurs in the whole Southern Bükk. It also contains redeposited limestone beds and lenses (64a). The thinly-bedded radiolarite is highly foliated; its age is Bajocian–Oxfordian (Kozur 1984, Csontos et al. 1991a).

The radiolarite is overlain by clay shale with intercalating, fine-grained sandstone layers. The shale comprises graded, distal turbidite cycles (Lökvölgy Shale, 64b) and locally, it also contains redeposited conglomerate. It can attain several hundred metres thickness; however, because of the high-grade deformation, the original thickness can only be assumed. Due to the slight metamorphism, it is platy-foliated. The assumed age is Middle – Late Jurassic; the assignation is only approximate due to the low number of uncertain biostratigraphic data (PELIKÁN et al. 2005, HAAS et al. 2013).

SZARVASKŐ–DARNÓ UNIT

The highly tectonised succession of the Szarvaskő–Darnó Unit probably derives from the accretionary prism (subduction complex) from the front of the oceanic island arc of the Neotethys (CSONTOS 1999, CSONTOS 2000). On the basis of the features of the magmatites within the sedimentary sequence, the rocks of the nappe were formed on an oceanic lithosphere (BALLA et al. 1983, BALLA 1983, HARANGI et al. 1996).

The largest superficial, continuous occurrence of the unit is in the western Bükk (Szarvaskő synform; BALLA 1983); while in the south-eastern Bükk it appears above the Bükk Unit in the form of several smaller nappe remnants (CSONTOS 1988, 1999). Westward the unit occurs on the Darnó Hill and in the boreholes of the region (Recsk deep-level Rm). Its western boundary under the Mátra Mountains is unknown. In the Mid-Transdanubian Unit low-grade metamorphic Jurassic slates with basic magmatites have been exposed at Tóalmás. Similar rock-association is known in the vicinity of Inke in the Zala Basin.

Middle Jurassic olistostrome melange (63)

This diverse Jurassic succession is made up of sandy clay shale, clay shale, black radiolarite and basinal ooilitic limestone (the latter being redeposited as carbonate turbidite).

The Lower – Middle Jurassic is made up of pelagic sandstone (Vaskapu Fm), as well as dark grey clay shale and grey cherty limestone (Oldalvölgy Fm); olistostrome intercalations are also present (PELIKÁN et al. 2005, HAAS et al 2013). The pelagic limestone is well-bedded, though this feature is hardly recognisable due to the high-grade deformation. Locally, it contains black chert layers and bands, Jurassic ooilitic limestone blocks and clasts of volcanic rocks (BÉRCZI-MAKK, PELIKÁN 1984; CSONTOS et al. 1991a, b; HAAS et al. 2013) and Upper Triassic pelagic carbonate clasts also occur (KOZUR, MOCK 1977).

In the Nagy-Rézoldal of Recsk (in the south-eastern Bükk) and in the boreholes signed with Rm–, the clay shale contains Triassic olistoliths of complex lithology and intercalating magmatic bodies. Between the green, amygdaloidal, microcrystalline pillow lava rocks of the Nagy-Rézoldal, reddish volcanic glass and locally, Middle Triassic red radiolarites are to be found (DE WEWEER 1984). The geochemical composition of the basalt indicates a mid-oceanic ridge depositional environment (MORB) (RÉTI 1985, HARANGI et al. 1996). It can be interpreted as a block which was ripped off from the Triassic oceanic basement.

Locally, lenticular black radiolarite can be observed in the clay shale; this corresponds to the red Bányahegy Radiolarite with regard both to the facies and the age (Bajocian–Callovian) (KOZUR 1984, CSONTOS et al. 1991a).

The uppermost part of the reconstructed succession is made up of Middle – Upper Jurassic carbonates with clay shale intercalations and ooilitic limestone beds (Bükkzsérc Limestone; BÉRCZI-MAKK 1999, CSONTOS 1991b, HAAS et al. 2006). Within some of the beds the ooids are graded and, locally, black chert also occurs.

Jurassic basic magmatites (62)

Basalt pillow lava is deposited on the lenticular clay shale, as well in the sedimentary rocks in the vicinity of Szarvaskő. The black-dark green hyaloclastic volcanite is at least 500m thick. Basaltic rocks also occur east of the volcanic centre (Kós Valley) and west from it (Recsk, Darnó Hill).

Downward from the extrusion level, gabbro sills are to be found of increasing crystal size and thickness. The green rock contains many lath-shaped feldspar crystals, pyroxenes, a few amphiboles and locally, also pyrite. In certain cases the plutonic bodies created metamorphic aureoles within the sedimentary rocks. According to the dating of the contact-metamorphic minerals of one of these zones (ÁRVÁNÉ SÓS et al. 1987), the age of the whole magmatic complex is Middle Jurassic (165 Ma).

According to geochemical investigations, the Jurassic palaeo-volcano (BALLA et al. 1983, BALLA 1983, HARANGI et al 1996) is a magmatic formation of either oceanic affinity (NMORB) or supra-subduction zone origin (SSZ).

MID-TRANSDANUBIAN UNIT

The Mid-Transdanubian Unit is a narrow, highly tectonised zone between the Mid-Hungarian Line and the Balaton Line, the latter representing the continuation of the Periadriatic Line (Figure 1). It comprises the south-western part of the Mid-Hungarian Mega-unit. Its pre-Alpine basement does not occur on the surface of the area of Hungary; direct information on its Late Palaeozoic and Mesozoic succession is provided only by boreholes.

For most parts of the zone knowledge on the Cenozoic basement is significantly limited; the map depicts these areas as unknown basements (**88**). There are areas where, according to geophysical data and data of limited number of boreholes, it is can be assumed that the basement is made up of rocks assigned to the Mid-Transdanubian Unit. However, the rocks could not be assigned to lithological units, nor could the distribution of the units be depicted. These areas are marked **Upper Palaeozoic and Mesozoic formations in general (60)**.

Within the structural unit such sub-units can be drawn that are characterised by Late Palaeozoic – Mesozoic successions of different facies and different metamorphic grades (Figure 13). These represent blocks that were formed in different palaeo-environments and which were later pushed next to each other by structural movements. According to outcrops and borehole information from Slovenia and Croatia, it is clear that the unit continues westward. As a result of the comparative analysis of the joint cooperation of researchers of the countries involved, the following sub-units have been distinguished (HAAS et al. 2000b):

— South Karawanks – Julian Sub-unit, dominantly consisting of Upper Palaeozoic shallow marine, as well as Triassic shallow marine, dominantly carbonate rocks. The rocks of the unit were not affected by Alpine metamorphism.

— South Zala Sub-unit, made up of highly diagenised, very low-grade (Alpine) metamorphic Upper Palaeozoic and characteristically deeper marine Mesozoic formations. Palaeozoic and Mesozoic formations of similar metamorphic grade and age are known in and around the Medvednica Mountains (Medvednica Unit).

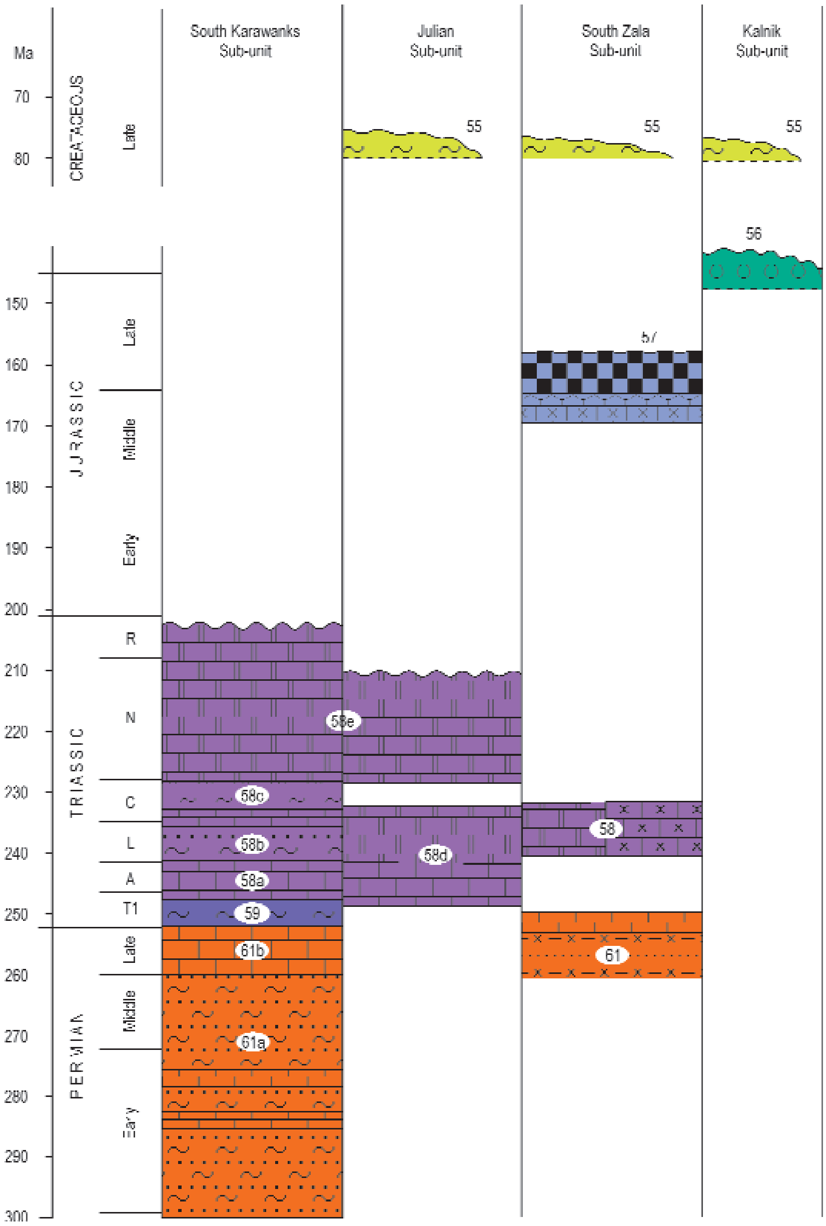


Figure 13. Stratigraphic column of the pre-Cenozoic formations of the Mid-Transdanubian Unit (after FÜLÖP 1990, HAAS, RÁLISCH 2004)

— The ophiolite melange of the Kalnik Sub-unit, which is the continuation of the Dinaric Ophiolite Belt. The unit also comprises the heterogeneous succession of the boreholes of Inke, where Triassic formations have also been penetrated.

Permian shallow marine siliciclastic and carbonate formations (61)

The Újfalu U–I borehole in the southern foreland of the Balaton Line, close to the Slovenian border, has exposed a 700m-thick Permian marine succession. The lower part of the succession is made up of fine siliciclastic rocks containing limestone intercalations (Trogkofel Fm, 61a). The facies of the formation is similar to that of the Košna Formation of the Velebit Mountains, while the foraminifera fauna can be related to that of the Lower Permian formations of the South Karawanks and the Carnic Alps (BÉRCZI-MAKK, KOCHANSKY-DEVIDÉ 1981). A similar Lower Permian succession is exposed by some boreholes in the 5–10km-wide zone south of the Balaton Line (BÉRCZI-MAKK 1988, BÉRCZI-MAKK et al. 1993). According to lithological analogies, the anhydritic dolomite, sandstone and silty marl (metamarl) exposed in the western part of the South Zala Basin (Semjénháza) are all considered to be Permian or Early Triassic.

Upper Permian shallow marine limestone and dolomite (Tab Dolomite, 61b) are known from several boreholes east of the Danube (in the vicinity of Sári and Bugyi) (BÉRCZI-MAKK 1978). It is likely that this area also belongs to the Mid-Transdanubian Unit, though the Permian formations do not provide conclusive evidence in this matter because the Upper Permian might be of similar facies in the north-eastern part of the Transdanubian Unit and in the Bükk Unit. A Lower Permian marine succession has recently been observed in the eastern continuation of the strike of the Mid-Transdanubian Unit, in the vicinity of Jászberény (CSEREPES et al. 2008).

Lower Triassic shallow marine claystones, marls, limestones (59)

In the northern zone of the South Karawanks – Julian Sub-unit, a large number of boreholes have exposed a succession of Lower Triassic shallow marine formations that overlie the Permian marine formations (Buzsák Group) (BÉRCZI-MAKK et al. 1993). The lower section of the succession is characterised by the dominance of variegated marl of lilac shade. Moving upwards, limestone, ooilitic limestone, dolomitic limestone, calcareous marl and marl rock types can be found (with sandstone and anhydritic dolomite – marl intercalations). In the upper part of the succession dark grey dolomite is present.

Middle – Upper Triassic formations of platform and basin facies (58)

In the Anisian stage, in the northern zone of the South Karawanks – Julian Sub-unit, a foraminifera-rich succession of grey dolomite, dolomitic limestone, dolomitic marl and limestone is to be found, formed in a shallow marine setting (Táska Fm, 58a). The Ladinian stage is represented by basinal dark grey, siliceous, cherty limestone, clayey limestone and marl; in these, intercalating volcanic tuff and radiolarian tuffite are also evident (Sávoly Limestone, 58b). In several boreholes Ladinian tuff-sandstone has been exposed, comprising sand-sized rhyolite fragments. The sandstone also contains radiolarian claystone intercalations (Murakeresztúr Sandstone, 58b) (RÁLISCH-FELGENHAUER 1998). The Carnian stage is represented by basinal and slope dark grey marl, siltstone, sandstone and limestone (Újudvar Marl, 58c).

In the southern part of the South Karawanks – Julian Sub-unit there are successions in which the whole Middle Triassic and also the lower part of the Carnian are represented by platform dolomite and limestone (Som Limestone, 58d). The sedimentological features and biofacies of these show similarities with those features of the Wetterstein-type platforms (HAAS et al. 1988). Volcanic tuff has been exposed under it by the borehole Seregélyes Sg–1.

Over the whole area of the unit the uppermost part of the Triassic succession is represented by light grey platform dolomite and Norian–Rhaetian brownish grey, light grey limestone (Igal Fm, 58e). According to the lithological characteristics and the biofacies, the setting of the deposition might have been the inner part of a Dachstein-type platform (HAAS et al. 1988a).

Very-low grade metamorphic Triassic–Jurassic formations of slope and basin facies (57)

The pre-Cenozoic basement of the South Zala Sub-unit is made up of diagenetically highly-altered shallow marine carbonate formations (that occasionally show signs of very low-grade or low-grade metamorphism), and pelitic, silicified rocks and magmatites. Their contact is usually tectonic. On the basis of K-Ar dating the age of the very low-grade and low-grade metamorphism is Cretaceous (illite-muscovite: 96.7 Ma, acidic metavolcanoclastite: 93–97 Ma; ÁRKAI et al. 1991).

The most reliable information on the Triassic formations is given by the core Iharosberény Ib–1 borehole. This borehole exposed Ladinian–Carnian platform limestone and platform foreslope brecciated limestone with, respectively, considerable thicknesses. The brecciated limestone comprises clasts of reef origin (58, RÁLISCHNÉ FELGENHAUER 1998). The succession is overlain — with a tectonic boundary — by radiolarite, metasiltstone and sericite schist. From the radiolarites Middle and Upper Jurassic radiolarians have been described (DOSZTÁLY 1994). Many other boreholes have exposed Triassic and Jurassic formations of similar lithologies in the basement of the South Zala Basin, but their age could not be determined due to the lack of adequate biostratigraphic data (RÁLISCH-FELGENHAUER 1998).

Jurassic–Cretaceous melange (56)

In borehole Inke–I, acidic and intermediate metavolcanites and serpentinite, as well as shallow marine carbonate formations have been exposed. These are underlain by dark grey radiolarite that contains Ladinian (DOSZTÁLY 1994) and Carnian radiolarian fauna. Similar rock types have been described from other boreholes around Inke. This mixed rock association indicates that the exposed formations probably belonged to a sedimentary-tectonic melange complex that can be interpreted as the equivalent of the ophiolite melange known from the Medvednica, Ivanščica and Kalnik Mountains (HAAS et al. 2000b). On the basis of data from Croatia, the clay shale-radiolarite matrix of the melange complex is of Middle Jurassic age (HALAMIĆ et al. 1999). It contains serpentinite, basalt, Middle and Upper Triassic radiolarite, sandstone, siltstone blocks and pebble-sized grains. During the Late Cretaceous — and locally, also in the Oligocene — the melange was deformed and redeposited (PAMIĆ, TOMLJENOVIĆ 1998).

Senonian pelagic marl (55)

In the western part of the Mid-Transdanubian Unit, Senonian formations are known from several boreholes. In the vicinity of Sávoly (Sáv–2) sandy siltstone (with limestone fragments) is deposited on the Triassic platform carbonates. The siltstone also contains freshwater (Chara) and marine (benthic and planktic foraminifera) microfossils (RÁLISCH-FELGENHAUER 1998). In the vicinity of Bagolasánc (Bag–1) and Gyékényes (Gyék–1) pelagic marl has been exposed, with microfossils indicating Senonian age.

ALCAPA MEGA-UNIT II — WESTERN CARPATHIAN UNITS

The pre-Cenozoic geological structure of the northern–north-eastern part of Hungary is determined by the mega-units known from the Inner Western Carpathians (Figure 1), situated in the northern vicinity of the Diósjenő–Ógyalla Line (as well as of its eastern continuation). From west to east (structurally, from bottom to top) three units can be distinguished: namely, the Vepor Unit, the Gemer Unit and the Aggtelek–Rudabánya Unit (the latter is in fact a complex unit in itself). The structural position of the isolated Zemplén Unit (situated much farther eastward) within the Carpathian system is debatable, even in the Slovakian scientific literature (Veporicum?, Gemericum? Tatricum?). Among the listed units only the Aggtelek–Rudabánya Unit can be studied on the surface; the respective build-ups of the others are known from mineral exploratory boreholes.

The nappe structure of the above-mentioned mega-units (and the relating Alpine metamorphism) evolved during the Mesozoic (Jurassic-Cretaceous). However, due to the Cenozoic cover the contacts between the units are unknown in the area of Hungary. In the western part of the area the Vepor and the Gemer units are separated structurally in the south from the Transdanubian Range Unit and the Bükk Unit by the E-W–striking Diósjenő–Ógyalla Line (BALLA 1988). The Aggtelek–Rudabánya Unit is separated from the Szendrő–Uppony and the Bükk units in the east by the NE–SW–striking Darnó Zone. The schematic tectonic model of the above-described structural units is shown on Figure 14.

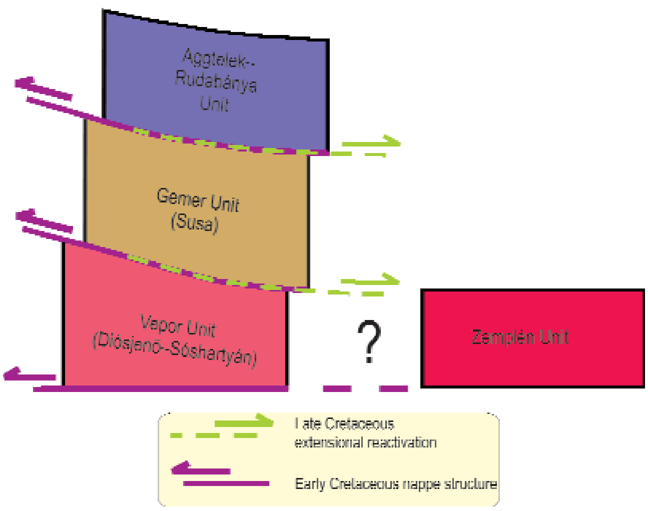


Figure 14. Schematic tectonic model of the Western Carpathian pre-Cenozoic structural units

VEPOR UNIT

The Vepor Unit in Hungary is known only from boreholes. It is made up of a **medium-grade polimetamorphic complex (75)** (formerly: Ipoly Crystalline Complex, FÜLÖP 1990), that underwent amphibolite metamorphism during the eo-Alpine phase.

Two main lithological groups are present in the crystalline basement: (1) The garnet-biotite-muscovite-feldspar-quartz gneisses are accompanied by (locally graphitic) mica schist and, in smaller quantities, quartzites. Rarely, phyllonites also occur. (2) Amphibolites (formerly considered as greenschists) occur as smaller or larger intercalations in the previous complex. The above-described rocks have been exposed by a large number of boreholes in the vicinity of Perőcsény, Hont, Diósjenő, Balassagyarmat, Szécsény and Sóshartyán.

The Vepor Unit is the deepest-positioned Carpathian mega-unit in the area. There is no information with regard to its inner structure due to the overlying Cenozoic cover. Microstructural investigations of the complex indicate folding in at least two phases, as well as plastic shear zones of amphibolite and greenschist facies.

The unit is characterised by polimetamorphic evolution (KOROKNAI et al. 2001): the earlier, probably Variscan amphibolite(?) facies event is indicated by almandine-pyrope-rich cores. The next amphibolite facies event — which was once considered to be Variscan (RAVASZ-BARANYAI, VICZIÁN 1976) (T=550 °C, P=9 kbar) — occurred in fact in the Cretaceous eo-Alpine phase (87–95 Ma, K-Ar and Ar-Ar data obtained from muscovite, biotite and amphibole KOROKNAI et al. 2001). This assertion is based on radiometric dating. In the later phase of the Alpine metamorphism the former mineral assemblages (mainly in the amphibolite group) underwent lower or higher-grade greenschist facies retrogression. The unit was affected by rapid exhumation in the Late Cretaceous (zircon FT: ~75 Ma, apatite FT: ~50 Ma) in connection with the east-directed sliding of the units from above the Veporicum (PLAŠIENKA et al. 1999, Figure 14.).

GEMER UNIT

The rocks of the unit (**Palaeozoic and Mesozoic formations in general, 76**) were exposed in Hungary only by one single borehole (Susa, S–1), in the north-western vicinity of Ózd. Conditionally, on the basis of data from Slovakia, a small area around Hidasnémeti has also been assigned to this unit. The borehole penetrated mostly poorly metamorphised clastic sediments (lithoclastic metasandstone, slate, metasiltstone) and near to the bottom, highly-altered, carbonatic, vesicular metavolcanite. The debris material of the metasandstone is poorly sorted; most of the grains are angular or slightly rounded. It mainly consists of quartz and to a lesser degree, feldspar and mica. Polycrystalline quartzite occurs in the form of lithoclasts, derived from low-grade metamorphic rocks. It is characterised by its poorly-oriented texture. The fine-grained metapelites underlying the metasandstone (downwardly alternating with it) have a transverse foliated structure, indicating the folded inner structure of the unit. Metabasites at the lowermost position are foliated to various degrees; the original mineral composition cannot be determined because of the alteration. The age of the formation cannot be determined due to lack of fossils.

On the basis of the rock material of the boreholes in the vicinity of Rimavská Sobota, the above-described formations are assigned to the Palaeozoic of the Gemericum (FÜLÖP 1994). At the same time, the lithological character and grade of metamorphism make it possible to suggest a correlation with the Meliata Unit.

AGGTELEK–RUDABÁNYA UNIT

The Aggtelek–Rudabánya Unit comprises several nappe units and nappes, made up of Permian-Jurassic sedimentary complexes (BALOGH K. 1948, GRILL et al. 1984, LESS et al. 1988, GRILL 1989, SZENTPÉTERY, LESS 2006). The number, composition and the boundaries of the nappes raise questions even today (GRILL et al. 1984, LESS et al. 1988, LESS 2000, FODOR, KOROKNAI 2000, KÖVÉR et al. 2008). However, the existence of the main nappe units is basically accepted. To these nappe units belong the non-metamorphic Szilice Nappe, the metamorphic Torna Sub-unit and the Tornakápolna Sub-unit; the latter is made up of oceanic crustal fragments embedded in the evaporitic sole of the base of the Szilice Nappe (Figure 15). In this volume the Bódva Sub-unit (which comprises mostly pelagic sediments) and the Jurassic Telekesoldal Sub-unit are described separately.

The sedimentation in the Aggtelek–Rudabánya Unit can be traced from the end of the Permian. In the Late Permian and at the beginning of the Triassic, the sedimentary environment of the Szilice, the Bódva and possibly the Torna sub-units was the wide shelf of the Tethys. The clastic-carbonatic sedimentation of the shallow marine ramps (HIPS 1996, 2007) was replaced at the end of the Early Triassic by shallow marine carbonatic sedimentation. In the middle part of the Triassic, in most of the subsequent nappe units, rifting began, accompanied by crustal-thinning (GRILL et al. 1984). The starting date of rifting was different for each of the units. Rifting in the Ladinian led to the break-off the continental margins, resulting in the formation of oceanic crust, as indicated by the pelagic sediments accompanying the basaltic pillow lavas (RÉTI 1985, KOZUR, RÉTI 1986). Pelagic sedimentation took place on the deep marine slope and in the basins (Szőlősardó, Bódva, Torna sub-units) that evolved above the thinned continental crust, as well as above the carbonate platforms that gradually subsided and drowned while moving away from the margin (Aggtelek facies unit). This was interrupted in the Carnian by siliciclastic sedimentation (KOVÁCS et al. 1988) that later, at the very end of the Triassic, became once again the dominant form of sedimentation. Around the same time when the rifting took place (Middle and Late Triassic) acidic and locally neutral

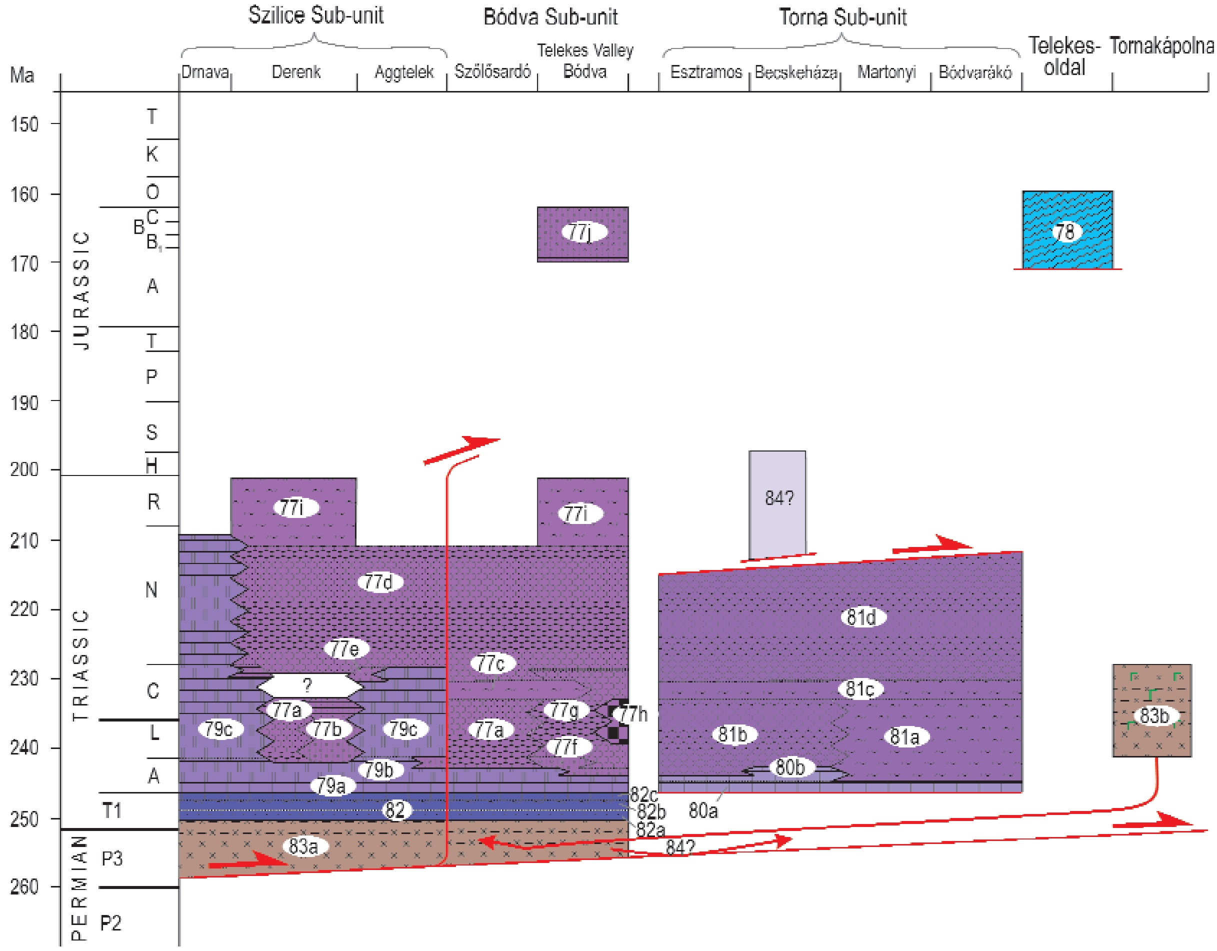


Figure 15. Stratigraphic column of the pre-Cenozoic structural sub-units of the Aggtelek–Rudabánya Unit (after Kovács et al. 2004)

magmatism occurred. The traces of this process have been preserved in the form of thin tuff intercalations, as well as olistolithes in the Jurassic formations. The geodynamic relation between the magmatism and the rifting is still unclear (Deák-Kövér 2012).

In the Vardar–Meliata branch of the Neotethys the intraoceanic subduction of the oceanic crust possibly began at the end of the Early Jurassic (Bortolotti et al. 2012, Schmid et al. 2008). At the same time it is also likely that the deposition of the clastic and siliceous rocks started on the continental slope (Bódva–Telekesvölgy facies). However, the Early Jurassic sediments are not documented by fossils rather the Middle Jurassic age has been undoubtedly proved (Grill, Kozur 1986). The Telekesoldal Nappe consists of gravity mass flow sediments deposited in a deep marine trench or on its slope. Olistostromes in the nappe could (indirectly) indicate an obduction event that accompanied the subduction. The process resulted in the deformation and denudation of the slopes of the trench. As a consequence of the subduction and, from the end of the Middle Jurassic, that of the obduction, some sub-units (Telekesoldal, Torna) were detached from their basement and buried tectonically. The highest-pressure metamorphism (blueschist facies) affected the sedimentary-magmatic rocks which occur around Bôrka and Hačava (Faryad 1995). The very low-grade and low-grade metamorphism which affected both the Telekesoldal and the Torna sub-units was accompanied by multi-phase ductile deformation (Árkai, Kovács 1986; Kövér et al. 2009a). On the basis of metamorphic petrological data and the character of deformation, the maximum grade of the metamorphism was probably 300–350 °C and 3–5 kbar (Árkai, Kovács 1986, Kövér et al. 2009a). Some K-Ar ages fix this event as Early Cretaceous (Kövér et al. 2009a). At the same time, a large-scale strike-slip movement occurred, affecting the northern part of the Neotethys. The movements changed the position of the Szilice and the Bódva sub-units and their situation in the nappe stacking (Stüwe, Schuster 2010, Deák-Kövér 2012): according to the strike-slip model, the units were probably displaced from the northern (lower) margin to an upper position.

In the mid-Cretaceous ca. 105–100 million years ago, the metamorphosed, and later exhumed and cooled Torna and Telekesoldal sub-units were thrust over the non-metamorphic Bódva Sub-unit (Kövér et al. 2009a). Around 90 million years ago new nappe movements started and these rearranged the order of the nappes, locally resulting in a young-on-older order contacts between formations or nappe slices. At the bottom of the uppermost Szilice Nappe, Neotethian oceanic crustal fragments (Bódvavölgy Ophiolite) were incorporated into the basal evaporite due to the last two nappe emplacement events (Grill et al. 1984). In addition to the evaporite, the overthrust was also facilitated by the rocks of the relatively high-temperature (~300 °C) and high-pressure (2–3 kbar) rauhwacke lenses forming at the base of nappes (Deák-Kövér 2012). These two nappe emplacement had probably south-eastern and/or southern vergency looking at present position With regard to the earlier nappe movements there is no direct data. A Senonian sediment remnant found in a karstic doline in the Szilice Sub-unit (Slovakia) indicate that most of the nappe movements occurred before the Late Cretaceous. This is also suggested by the fission track cooling ages (Árkai et al. 1995a).

Permian–Mesozoic formations in general (84)

The classification of the formations in this group is unclear. One example is the "Hidvégardó Formation", and its part, "Tornaszentjakab beds" which occur on the surface and which were exposed by boreholes in the northern part of the Rudabánya Hills (Hidvégardó Ha–3, Tornaszentjakab Tj–1) (Szentpétery, Less 2006). The complex is probably made up of slate, thin-bedded limestone, marl and silty calcareous marl. However, due to the intense nappe stacking it is unclear which rocks form one succession, or even more, one structural unit. The "Permian – Lower Triassic(?)" assignment of Szentpétery, Less (2006) is based on the Permian fossils found in Slovakia, as well as on the presence of evaporite intercalations. At the same time, it is possible that the monotonous black shales are of Jurassic age, whereas the thin-bedded limestone and the silty calcareous limestone are of Late Triassic – Early Jurassic age (Deák-Kövér 2012). The low-grade metamorphic series underwent ductile deformation (foliation, kink folds). As such, it can be separated from the non-metamorphic evaporitic complex and the related Permian laminated claystone.

Upper Permian – Lower Triassic anhydrite (83)

The evaporite complex (Perkupa Anhidrite, 83a) that was sheared from the basement, is basically made up of anhydrite, gypsum, as well as intercalating dolomite, shale and siltstone (Mészáros M. 1961). Due to subsequent nappe stacking, it is undoubtedly present only in the Aggtelek facies unit, whereas in the Bódva Sub-unit its occurrence can only be assumed. On the other hand, the complex probably emplaced between or beneath the Torna and the Telekesoldal sub-units by salt tectonics. Frequently grey, white bands can be observed in the evaporite; these either represent the original sedimentary bedding or a ductile deformation element (foliation, Zelenka et al. 2005). The formation assemblage was formed in a hypersaline lagoonal setting. Its original thickness in unknown, but according to Szentpétery, Less (2006) a thickness of 250m is probable. The age of the formation assemblage cannot be determined due to lack of data, but on the basis of the age of the direct overlying rocks the evaporite formation probably continued also in the Early Triassic (namely, until the end of the Early Induan) (Hips 1996).

The ultrabasic-basic blocks which are derived from the oceanic crust of the Neotethys are depicted as part of the evaporite complex into which the blocks have been tectonically emplaced (83b, Havas 1984, Réti 1985). They are known to occur on the surface only at one location, but they have been exposed by several boreholes (Réti 1985; Horváth P. 1997, 2000). The rocks in Hungary belong to the Tornakápolna sequence and in fact were formed in the same ocean as the basic and ultrabasic rock known in redeposited olistholites in the Meliata sequence. According to the pelagic sediments within the pillow basalt, their age is Ladinian (Kozur, Réti 1986).

Lower Triassic shallow marine sandstones, marls, limestones (82)

The Early Triassic was characterised mainly by clastic sedimentation, which occurred on a shallow marine ramp (Hips 1996, 2001). The evaporite series is overlain by lilac red or greenish grey sandstone, siltstone and claystone (Bódvaszilas Fm, 82a). The clastic complex was formed on a microtidal inner ramp, in lagoons isolated by sand banks (Hips 1996) at the beginning of the Early Triassic. Its thickness is 200–250m. The overlying Szin Marl (82b) is of diverse lithology; it is made up of marl, fine-crystalline limestone, crinoidal limestone, ooidic limestone and clay marl. The ooid-mounds were formed on the inner ramp; their material redeposited by storms was formed in lagoons and on the middle ramp, while the fine-grained sediments were formed on outer ramp below the storm base (Hips 1996). The 350m-thick succession was formed in the later phase of the Early Triassic (Hips 1996). The overlying dark grey, vermicular Szintpetri Limestone (82c) was deposited at the end of the Early Triassic, in a periodically isolated, oxygene-poor lagoon (Hips 1996).

Middle Triassic – Carnian shallow marine carbonates (79)

The Lower Triassic clastic-carbonate sequence is overlain by Lower Anisian shallow marine carbonates in every tectonic unit. These formations consist of dark or light grey, bituminous or bioclastic, thinly-bedded, bedded or massive dolomite and limestone (Gutensteini Fm, 79a). The thickness of the formation is 250–400m (Szentpétery, Less 2006, Hips

2007). During the carbonate deposition the outer ramp, that was characterised by oxygen-poor water, was gradually replaced by microbial or sponge-dominated mud-mounds of the middle-inner ramp and later, by a peritidal environment (HIPS 2007). The Steinalm Formation (79b), which represents the younger section of the Anisian complex, consists of the cyclic alternation of limestone beds of different microfacies (algae, bioclastic, oncoidic microfacies). In addition to the lagoonal-sediments, reef bodies are also known dominantly made up of calcareous sponges (VELLEDITS et al. 2011). In most of the structural units the formation of the shallow marine carbonates came to an end in the Anisian (KOVÁCS et al. 1989). However, in the Aggtelek facies unit the lagoonal and reef carbonate (Wetterstein Limestone, 79c) continued to be deposited in the Ladinian and the Carnian (KOVÁCS 1979). The thickness of the formation can exceed 1000m. On the basis of foraminiferans, algae and sponges its age is Early Ladinian – Late Carnian (BÉRCZI-MAKK 1996a, PIROS 2002, SENOWBARI-DARYAN et al. 2011). In the vicinity of Drnava (Slovakia) the formaion of platform carbonates continued in the Norian, as well.

Low-grade metamorphic Middle Triassic and carnian shallow marine carbonates (80)

The oldest known formation of the metamorphic Torna Sub-unit is the dark grey, massive, bedded, locally thinly-bedded dolomite (Gutenstein Fm, 80a). In addition to the occurrences of the formation at Bódvarákó, Esztramos and Becskeháza, FODOR, KOROKNAI (2000) assigned to this formation the Gutenstein dolomite of the Martonyi Nappe. The latter, on the basis of its colour and higher pyrite content, was probably formed on the more isolated part of the shallow marine ramp. The largest exposed thickness is 125m, though the underlying formation is not known. It represents the Early Anisian stage of the carbonate ramp evolution.

In the Middle Anisian, the evolution of the ramp continued with the deposition of light grey crystalline shallow marine limestone and dolomite; however, the original relationship of the Steinalm Limestone (80b) with the overlying formation was erased by subsequent structural deformations. The algae and crinoids in the limestone indicate a shallow marine ramp environment (KOVÁCS et al. 1988) that, during the Middle Anisian, was followed by a pelagic basin environment (Esztramos, KOVÁCS 1986). In certain sub-units (e.g. Becskeháza facies), however, this occurred only after the Middle Anisian. At Bódvarákó and in the succession of the Martonyi Nappe the Gutenstein Formation is overlain by basinal formations (FODOR, KOROKNAI 2000). The shallow marine carbonate formations usually underwent very low-grade metamorphism (in case of the Esztramos, low-grade metamorphism) (ÁRKAI, KOVÁCS 1986).

Middle and Upper Triassic and Jurassic toe-of-slope and basin facies (77)

Over the course of the Anisian, on the ramp which subsided as a result of crustal stretching, diverse carbonate sedimentary environments evolved. The oldest basinal formations in the Aggtelek facies unit occur in the foreland of the platform framed by reefs (VELLEDITS et al. 2011). In the Derenk facies unit the disintegration of the ramp is indicated by the syndiagenetically brecciated drusy limestone (Derenk Limestone, 77a). At the same time, on the developing slope, micritic and variously filamented, radiolarian, crinoidal limestone was formed (Nádaska Limestone). Locally, grey cherty, bedded limestone occurs (Reifling Limestone, 77b); this was interpreted by SZENTPÉTERY, LESS (2006) as a near-platform basinal sediment. In the mid-Carnian these formations were partly replaced by the Szőlősardó Marl (77c). These diverse formations are usually overlain by the pelagic, mostly reddish coloured Hallstatt Limestone (77d) or the grey, cherty Pötschen Limestone. Their thickness is 50–200m and their age ranges between the end of the Anisian up to the mid-Carnian; in the case of the Hallstatt and the Pötschen Formations, the deposition of the sediments lasted till the Middle Norian or possibly, till the middle of the Late Norian (KOVÁCS et al. 1989, SZENTPÉTERY, LESS 2006). At the end of the Carnian, above the platform formations, crinoidal, brachiopodic, ammonitic limestone was deposited in the northern part of the Aggtelek facies unit (Szádvárborsa Limestone 77e) (SZENTPÉTERY, LESS 2006).

The transition of the shallow marine formations to the pelagic formations in the Bódva Sub-unit is represented by the light brown, pink radiolarian “filament” limestone (Dunnatető Limestone, 77f). It is overlain by red, lilac, usually thinly-bedded, “filament”-radiolarian pelagic limestone (Bódvalenke Limestone, 77g) (KOVÁCS et al. 2011). Locally, radiolarite also appears with the absence of the limestone beds (Szárhegy Radiolarite, 77h). The grading and the slump folds indicate gravitational redeposition. The thickness of the described pelagic formations ranges between 10m and 80m; their formation took place during different phases of the Anisian till the mid-Carnian. After that, till the middle part of the Norian, the pelagic limestone of the Hallstatt Formation followed. The pelagic limestone both in the Szilice and the Bódva sub-units is overlain by marl, clay marl and calcareous marl (Zlambach Marl, 77i; NÁDOR 1990). Its age is Late Norian – Rhaetian (SZENTPÉTERY, LESS 2006, KÖVÉR et al. 2009b). It is probably overlain with a hiatus by a Jurassic succession of grey limestone, marl, calcareous sandstone, claystone, siliceous marl and conglomerate (Telekesvölgy Group, 77j). A significant part of these sediments was deposited by gravity mass flows on slopes or in deeper basins (KÖVÉR et al. 2009b). From the Jurassic succession Bajocian–Bathonian radiolarians and foraminiferans are known (DOSZTÁLY 1994, GRILL 1988, KÖVÉR et al. 2009b).

Very low-grade and low-grade metamorphic Middle and Upper Triassic formations (81)

In the metamorphic (Torna) subunit the carbonates of the shallow marine ramp pass into pelagic formations in different levels of the Anisian. The pelagic formations are represented partly by dark grey, black limestone with chert layers, calcareous dolomite, marl, dolomarl and siltstone (Bódvarákó Fm, 81a), and partly by grey, drab, pink limestone that also contains — albeit rarely — chert lenses (Szentjánoshegy Limestone, 81b; Kovács 1986). In the Martonyi Nappe the succession is dominantly made up of metamarlstone and siltstone, containing thin limestone intercalations (Rednekvölgy beds, 81a; Szentpétery, Less 2006); its age is Middle Anisian – Early Carnian (Kovács 1986).

The clastic Tornaszentandrás Slate (81c) is made up of slate and metasiltstone and is related to the regional climate change of the mid-Carnian occurs also in the metamorphic successions (Less 2000). On the basis of the conodont fauna the 30-150m-thick formation was deposited in the mid-Carnian (Julian) (Kovács 1986). The deposition of the basinal siliciclasts,was gradually followed by pelagic carbonates from the Late Carnian till the Middle Norian. The thickness of the Pötschen Limestone (81d) is 20–150m and it contains chert layers, chert lenses and marl intercalations. Its sedimentary cover is unknown, but probably a part of the Hidvégardó Formation (84) also belongs here, and it comprises calcareous marl and metamarlstone. It underwent a deformation similar to that of the Triassic part of the Torna Sub-unit (Fodor, Koroknai 2003). The formations underwent very low-grade or low-grade metamorphism (Árkai, Kovács 1986) and ductile (plastic) deformation (Kövér et al. 2008, 2009a).

Low-grade metamorphic Jurassic slope and basin facies (78)

At many locations in the Rudabánya Hills and at the eastern edge of the Aggtelek Hills (at Szögliget) Jurassic metamorphic rocks are known. In a broader sense, it is the Telekesoldal Formation that forms an individual nappe (Kövér et al. 2008, 2009a). It is made up of slate, sandstone, siliceous or carbonaceous slate, and olistostrome levels (Grill 1988, Kövér et al. 2009b). The bulk of the rocks were deposited as a result of gravity mass flows. The material of the clasts of the olistostromes is dominantly pelagic limestone and (rarely) radiolarite or basalt. The material of the largest clasts is rhyolite; these can be as large as 100m. Previous studies considered them Jurassic subvolcanic bodies (Szakmány et al. 1989), but more recent U-Pb dating indicated Triassic ages (~204–220 Ma, Deák-Kövér 2012).

The formations were probably deposited in a deep marine basin or on the slope of a deep oceanic trench. Clasts probably originated from the thinned continental margin (Deák-Kövér 2012). On the basis of radiolarians and dinoflagellate cysts, the age of the formation is Bajocian–Callovian (Grill, Kozur 1986, Kövér et al. 2009b). The complex was sheared from the basement at the very end of the Jurassic or at the beginning of the Cretaceous (between 120 Ma and 140 Ma) and was subducted, metamorphosed and suffered multi-phase ductile (plastic) deformation (Kövér et al. 2009a). According to the grade of metamorphism and its facies, the formation is quite similar to the Meliata sequence (Mock et al. 1998) and in a broader sense it can be assigned to it.

Zemplén Unit

The Zemplén Unit is made up of Variscan medium-grade metamorphic formations (which have undergone amphibolite metamorphism), the Upper Carboniferous–Permian continental clastic and rhyolite complex (unconformably deposited on the former), and Triassic formations. These formations together were affected by a new tectogenesis during the Cretaceous. The inner structure of the unit is characterised by SW-vergent slivers/nappes(?) and folds, mostly related to the Alpine orogeny.

Variscan medium-grade metamorphic complex (87)

The pre-Alpine basement complex is dominantly built up of mica schists and gneiss, in which there are a few thin (garnetiferous) amphibolite intercalations. Protoliths of these rocks are pelitic-psammitic sediments and smaller intercalating basic (neutral) magmatic bodies. As a result of subsequent deformations, mylonitic and cataclastic zones frequently occur in the metamorphites of amphibolite facies. In these zones the rock material underwent greenschist retrograde metamorphism. These formations occur on the surface in the north-eastern corner of Hungary, in the vicinity of Vilyvitány and Felsőregmec.

The upper-amphibolite metamorphism of the crystalline rocks (T=640–700 °C, P=5–8 kbar; Vozárová 1991) is related to the Variscan orogeny (monazite U-Pb-Th and amphibole K-Ar: ~338 Ma; Finger, Faryad 1999, Faryad, Balogh Kad. 2002). Beyond geochronological data, the Early Carboniferous metamorphism is also verified by the fact that the pebbles of the above-described rocks appear in the post-orogenic sedimentary overlyer of the crystalline complex as well — namely, in the Upper Carboniferous molasse sediments. Cretaceous ages (126–106 Ma) obtained from the fine-grained muscovite of retrograde origin, as well as from the feldspar, indicate a weak Alpine effect in the unit. The greenschist retrogression accompanying the intensive deformation was probably, at the least partly, related to this phase.

Carboniferous–Permian continental siliciclastic formations and rhyolite (86)

In the northern part of the Tokaj Hills the polimetamorphic crystalline basement is overlain by a ca. 800m-thick Permo-Carboniferous molasse (Fülöp 1994). The Felsőregmec Fr–3 borehole has exposed an Upper Carboniferous fluvial-marshy succession of alternating sandstone and conglomerate. In the Lower Permian fluvial sandstone, rhyolite tuff (ignimbrite) and tuffite intercalations are frequent. The Upper Permian ends with a fluvial complex of alternating red and variegated sandstone and conglomerate (Széphalom Szh–2, Sátoraljaújhely Suh–8).

Middle and Upper Triassic platform carbonates (85)

In the area of the Zemplém Unit, in the vicinity of Sátoraljaújhely (Suh–8) and Komoró (Komoró–I), dark grey dolomitic limestone, dolomite, clay marl and claystone are known to occur; these are equivalents of the Middle Triassic shallow marine ramp facies of Slovakia (Ladmóc) (Haas, Pentelényi 2004). According to several boreholes and to the inclusions of the Miocene volcanic complex, the Triassic facies of the unit in Hungary significantly differs from that known from Slovakia. From inclusions Middle Triassic Wetterstein-type limestone is also evident. The Sárospatak S–7 borehole has exposed a ca. 400m-thick succession of Upper Triassic carbonate platform facies, made up of Dachstein-type thick-bedded stromatolite limestone and dolomite (Pentelényi et al. 2003). These Triassic formations probably belong to an upper-positioned nappe.

Structural evolution

Hungary is situated in the Pannonian Basin which was formed in the Neogene within the Alpine (Mediterranean) orogenic zone stretching from North Africa, through South Europe and Asia Minor and going on until South-east Asia. The basement of the Pannonian Basin is complex; Palaeozoic rocks are of high importance and a part of the basement is represented by lithosphere fragments broken off from the Variscan Orogenic Belt.

PRE-VARISCAN AND VARISCAN EVOLUTIONARY STAGES

The structure of the overwhelming part of the European crust was formed by the orogeny following the Late Cambrian – Early Devonian Caledonian tectonic cycle, the Late Palaeozoic Variscan orogeny and the Mesozoic–Cenozoic Alpine orogeny. However, the East European Platform and the Baltic Shield, made up of several billion years old rocks, were not affected by post-Mesoproterozoic orogenic events.

The history of the evolution of the Variscan mountain range starts in the Early Palaeozoic, at the time of the Pan-African orogeny. The event resulted in the suture of the super continent Gondwana, comprising South America, India, Australia and the Antarctic. As a result of the closure of oceanic basins during the plate tectonic processes lasting from the Ordovician till the Devonian, different continental lithosphere blocks collided and were sutured to the continent of Laurussia. A large ocean (called Palaeotethys) had formed between Laurussia and Gondwana by the Early Carboniferous. The plate tectonic cylce ended ca. 320 million years ago, with the collision of Gondwana and Laurussia and the formation of the Variscan Orogenic Belt (its North American equivalent is the Appalachian Orogenic Belt). As a result of this event one single super-continent was formed on the Earth, called Pangea.

In Europe, in the axis zone of the Variscan orogeny, highly intensive metamorphism took place and as a result of the melting of the colliding blocks a large amount of granite was formed. This central zone of the orogeny (also known as Moldanubian Zone) can be followed from the Massif Central of France, through the Vosges and the Black Forest till the southern part of the Bohemian Massif. The evolution of the basins was determined primarily by the position of the basins relative to the deformation front of the orogeny (McCann et al. 2008). In the direct vicinity of the deformation front and in its northern foreland, flexural and foreland basins evolved. In the inner part of the Variscan Orogenic Belt, in the early stage of the orogenic phase, extension basins were formed, while in the last stage compression and transpression tectonics had a greater role. The uplifted mountain ranges were instantly affected by intensive denudation; the resulting gravel, sand and clay were accumulated in the intermontane basins and primarily, in the foreland basins.

The southern parts of the ranges formed during the Variscan orogeny were once again deformed and altered during the Alpine orogeny, and some of their elements were placed into the ranges of the Alps and the Western Carpathians (Ebner et al 2010). The southern part of the Pannonian Basin consists of the Tisza microplate (Tisza Mega-unit); this broke off from the Moldanubian Zone during the Jurassic, and this is also reflected by the similarities of the granitoid rocks with those of the Bohemian Massif, which are of the same age (Buda et al. 2004, Kovács et al. 2010). The Upper Carboniferous fluvial complex of Southern Transdanubia is an equivalent of the Variscan molasse basins.

In the Early Carboniferous, south from the later-formed Variscan Mountain Range between Laurussia and Gondwana, in the shallow marginal sea basins of the Palaeotethys Ocean, limestone and marl were deposited. The Lower Carboniferous formations known from the Balatonfő area probably belonged to this zone (Ebner et al. 1991). In the deeper inner basins which have direct contact with the open sea, flysch-type formations were deposited, comprising alternating sandstone and marl beds. In the Late Carboniferous, however, shallow marine sediments were deposited here as well. These successions, deposited far from the collision zone, underwent only weak alteration during the Variscan orogeny. It is very likely that the Carboniferous successions of the Bükk Unit were deposited in this palaeogeographic unit (Ebner et al. 1991, 2010).

During the Variscan orogenic phase, a uniform continent existed only west from the East Mediterranean zone, and this is still present today. However, an oceanic area remained open between Eurasia and Gondwana, east of this region. Significant parts of the Palaeotethys subsisted in the form of the large bay of the Panthalassic Ocean protruding deep into the Pangea super continent.

ALPINE EVOLUTIONARY STAGES

Palaeotethys margin — Neotethys rifting (Permian – Middle Triassic)

In this region, the break-up of the Pangea started in the Permian. The continental rifting was related to the acidic volcanism and terrestrial sedimentation, the latter occurring in rapidly subsiding grabens.

In the area of Hungary the continental rifting stage of the Alpine cycle was recorded most strikingly in the Tisza Mega-unit which belonged to the Moldanubian Orogenic Zone of the Variscan Mountain Range streching along the edge the Eurasian Plate. In the graben system known from South Transdanubia, a several kilometres-thick fluvial, lacustrine sediment mass had been deposited from the Early Permian till the Early Triassic, accompanied by rhyolite volcanism at the boundary of the Early and Middle Permian (Barabás, Barabás-Stuhl 1998).

In the Late Permian, in the southern foreland of the Variscan Mountain Range, a wide zone was formed which was in a position to collect the debris deriving from the intensive denudation. This zone included the Tisza Mega-unit and, west of it, the bulk of the area of the Central Western Carpathians and the Austroalpine Unit. Southward, on the shores of the bay of the Palaeotethys and in its isolated lagoons, evaporite was formed under arid climate. This evaporite series is of a significant extent and is known in the Aggtelek area in Hungary; it had an important role in the subsequent structural evolution.

In the Permian, the area of the Transdanubian Range was situated between the sedimentary basins of the Northern Limestone Alps and the Southern Alps, at the western end of the Palaeotethys. During the Late Carboniferous – Early Permian a series of pull-apart basins evolved in the area of the Southern Alps, in relation with the slip motion of Laurasia and Gondwana (Ziegler 1988). The basins were filled with terrestrial clastic sediments and volcanic rocks of significant thickness. This stage of the rifting in the Transdanubian Range is represented by the intrusion of the granitoid pluton of Velence. In the Late Permian transpressional basins were formed in the area of the Southern Alps; these were also later filled up with terrestrial sediments and acidic volcanic rocks (Cassinis et al. 1988). The Transdanubian Range Unit was also a landmass, but there is little few evidence of the one-time volcanic activity. The Late Permian palaeogeographical situation and evolutionary history of the two areas are very similar. Sedimentation on both areas started with the deposition of fluvial sediments, while in a later stage of the Late Permian the ocean-ward part of the units was flooded by shallow sea, edged by a wide tidal flat (Majoros 1983, Cassinis, Neri 1992). At the beginning of the Triassic the former landmasses were also inundated by the sea. On the area of the Bükk Unit, similarly to the Sana–Una and Jadar units of the Inner Dinarides, the shallow marine Late Carboniferous sedimentation was followed by subaerial exposure; in the Middle Permian terrestrial, later shallow marine sedimentation took place that uninterruptedly continued even in the Early Triassic (Filipović et al. 2003).

At the beginning of the Early Triassic the overwhelming part of the Tisza Mega-unit was still dry land and here the deposition of fluvial sediments continued. The bulk of the region, however, already belonged to the shallow, marginal zone of the Palaeotethys Ocean. On the ramp connecting the shore to the deeper shelf quite similar successions were formed, made up of alternating sandstone, marl, limestone and dolomite. In the early stage of the Anisian dominantly carbonate rocks were deposited. At the beginning of the Middle Triassic the continent-ward parts of the Tisza Mega-unit (Mecsek Zone and Villány–Bihor Zone) were also flooded by the sea and a carbonate ramp evolved in these zones as well (Török Á. 1993).

During the Middle Triassic, at the northern margin of the one-time Gondwana continent, a new ocean — known as Neotethys — began to form simultaneously with the subduction of the Palaeotethys beneath the Eurasian Plate (Dercourt et al. 1986). The rifting related to the opening of the ocean expanded from east to west and reached the region in the later stage of the Anisian. Locally, it started with upwarping and typically led to the tectonic disintegration of the shallow ramps along the ocean margin. Grabens and submarine highs were formed in extensional stress regime (Budai, Vörös 1992, 2006; Velledits 2006). In the grabens pelagic limestone and siliceous limestone started to be deposited. Basaltic volcanic activity began in the axis of the subsequent mid-ocean ridge, in grabens located on thinned continental lithosphere. The lava flows poured into the pelagic carbonate sediments, resulting in the characteristic peperite-facies of the basalts, the remnants of which (in the form of olistolithes) are known from the Jurassic accretionary complex of the Dinaric Ophiolite Belt. They also occur in North Hungary, in the Szarvaskő–Darnó Nappe (Kiss G. et al. 2008, Kovács et al. 2008). The stretch of the lithosphere starting from the Ladinian led to the formation of oceanic crust; this now appears as blocks kneaded into the evaporitic sole of the Szilice Nappe (Grill et al. 1984). It is likely that the Triassic ocean crustal fragments are also present in the Jurassic melanges of Szarvaskő and the Darnó Hill (Kovács et al. 2008, Haas et al. 2011).

Beyond the axial zone, the seafloor was tectonically disintegrated in the passive margin. In certain zones intensive volcanic activity took place over the course of the Ladinian and Carnian. Volcanic centres evolved in the Dolomites, the volcanic material of which was spread widely, so that it reached even the Transdanubian Range. The Slovenian Graben was also formed as a result of the rifting. The process was accompanied by intensive magmatism. As a consequence, the block of the Julian Alps was separated from the Adriatic Carbonate Platform. The separation of the Sana–Una and the Jadar units, as well as the Bükk Unit can be interpreted similarly. In the Bükk, significant magmatic activity was related to the disintegration of the seafloor (SZOLDÁN 1990, HARANGI et al. 1996, VELLEDITS 2006).

OPENING OF THE NEOTETHYS AND PASSIVE MARGIN EVOLUTION (LADINIAN – EARLY JURASSIC)

Following the onset of the spreading of the ocean floor in a late Middle Triassic (Ladinian), the characteristic feature of the passive margins was the general subsidence. The Variscan Mountains stretched along the European margin and had by this time been highly denuded. Drainage system running from the Variscan Mountains (especially in humid intervals) transported large amounts of gravel and sand to the continental basins formed at the foot the mountain range, such as into the intensively subsiding half-graben-like basins of the Mecsek Belt (NAGY E. 1969).

The deep Middle Triassic basins were filled up with terrigenous sediments in the humid stage of the Carnian. This process can be traced in the Transdanubian Range Unit (HAAS et al. 2012). On the even surface several kilometres-thick platform limestone and dolomite had been accumulating from the Late Carnian till the end of the Triassic. At the time of the opening of the ocean the development of extensional grabens continued on the edge of the shelf facing the ocean (for example: at the margin of the Transdanubian Range Unit); this shelf is characterised by a thinned crust. Certain marginal blocks were faulted down. Here, the carbonate platform sedimentation was replaced by pelagic sedimentation. In the Aggtelek and Rudabánya subunits, in the blocks subsided during the Anisian rifting, pelagic carbonate sedimentation usually took place from the Ladinian till the Middle Norian (KOVÁCS et al. 1989). Over the entire period platform carbonate sedimentation was characteristic only in few blocks in the northern part of the Aggtelek facies unit. Few intraplatform grabens were marked by pelagic sedimentation and were filled up and were characterised by shallow marine sedimentation from the Carnian.

The smaller carbonate platforms of the Bükk Unit which had broken off from the Adriatic Plate drowned at the end of the Triassic. After that, over a long time, no sediments were preserved. Pelagic sedimentation started only in the Middle Jurassic.

PENNINIC RIFTING (END OF THE TRIASSIC – EARLY JURASSIC)

Rifting leading to the opening of the Atlantic Ocean had started already in the latest Triassic. As part of this process, the rifting event of the Penninic oceanic branch (“Alpine Tethys”, SCHMID et al. 2004) also started from the Mid-Atlanticum and headed west to east. In the western part of the Southern Alps intensively subsiding extensional basins evolved in the Late Norian (BERTOTTI 1993). This process can be related to the initial evolution of the new ocean-branch. A similar process can also be traced in the western side of the Transdanubian Range Unit (in the Keszthely Mountains and in the basement of the Zala Basin) as well as on the area of the Northern Calcareous Alps. The process led to the formation of the Kössen Basins (KOVÁCS et al. 2011).

Around the same time, extensional grabens were also formed in the Tisza Mega-unit. From the Variscan Ranges at the margin of the European Plate a large amount of terrigenous material was transported to the sedimentary basins situated close to the continent (Mecsek Unit). The climate became more humid and in coastal swamps a siliciclastic succession was deposited with coal being formed from the dense vegetation. In the later phase of the Early Jurassic the coastal swamps were inundated by the sea nearly everywhere and shallow marine, later deep marine siliciclastic sedimentation became prevailing.

In zones situated farther away from the rifting margin of the continent the extensive carbonate platforms that had formed in the Late Triassic broke along faults in the Early Jurassic, and the so-formed unevenly subsiding blocks gave place to pelagic carbonate sedimentation. This event can be traced in those marginal zones of the Neotethys that later broke off from the European Plate at the time of the opening of the Ligurian–Penninic ocean branch: from the Southern Alpine segment, through the Transdanubian Range, the Austroalpine and the Central Western Carpathian units, up to the Tisza Mega-unit (HAAS et al. 2011).

However, on the carbonate platform (Adriatic–Dinaric Carbonate Platform) which had evolved on the northern and eastern margin of the Adriatic microcontinent the shallow marine sedimentation of the Triassic continued also in the Jurassic. At the same time, in the basins surrounding the platform (Belluno Graben, Slovenian Basin, Bosnian Basin), limestone bodies intercalated in the pelagic sediments which were made up of redeposited platform-derived grains (ROŽIČ, POPIT 2006).

NEOTETHYS SUBDUCTION AND THE OPENING OF THE PENNINIC OCEAN BRANCH (MIDDLE JURASSIC – EARLY CRETACEOUS)

The opening of the Penninic oceanic branch and the starting of the spreading of the oceanic lithosphere occurred at the end of the Early Jurassic. The remnants of the oceanic basement are known from beneath the Austroalpine Nappe System, in tectonic windows (in Hungary, in the Kőszeg–Rechnitz Window) (KOLLER, PAHR 1980).

The opening of the Penninic oceanic branch from west to east led to the breaking of the Tisza Unit off from the European Plate in the Middle Jurassic. This event is reflected by the carbonate-siliceous pelagic sedimentation of the Mecsek Zone that replaced the terrigenous siliciclastic sedimentation. In the Neotethys-ward side of the area between the Penninic oceanic branch and the already closing Neotethys, pelagic carbonate-siliceous sediments were formed on the uneven basement during the Middle and the Late Jurassic.

In the region, the subduction of the Neotethys commenced in the latest Early Jurassic – earliest Middle Jurassic interval. On the basis of their petrological and geochemical features supra-subduction type ophiolites occur in the Dinarides, the Albanides and the Hellenides can be related to the subduction; they are of a supra-subduction type. Their metamorphic sole was probably formed by the subduction of the cold oceanic plate beneath the hot, obducting ophiolite. As such, the age of the metamorphic sole is close to the age of the ophiolite formation. According to these ages (147–174 Ma) of the Dinaric ophiolites these processes probably occurred in the Middle to Late Jurassic (LIATI et al. 2004, BORTOLOTTI et al. 2013). The olistostrome melange known in the Dinaric Ophiolite Belt and in the Vardar Zone can be interpreted as an accretionary complex formed during the subduction processes. Closure of the Neotethys Ocean probably started with intra-plate subduction, while the ophiolite formed above the subduction zone was later thrust over the continental margins surrounding the oceanic basin: namely, over the Adriatic microcontinent and over the lithosphere blocks broken off from the European Plate (Austroalpine, Central Western Carpathian, Tisza Mega-units). In the Middle – Late Jurassic (Callovian–Kimmeridgian), in the Northern Calcareous Alps belonging to the Upper Austroalpine Unit, a series of flexural basins evolved in front of the nappe front during the nappe stacking of the Austroalpine margin. The nappe thrusting began in the external zone (Hallstatt facies belt) and progressed towards the internal zones (Hauptdolomite/Dachstein facies belt) (GAWLICK, FRISCH 2003, MISSONI, GAWLICK 2010).

The Middle Jurassic olistostrome melange known in the Szarvaskő–Darnó Unit derives from the accretionary complex of the Dinaric Ophiolite Belt. After it had been sheared, it arrived at its current position as a result of large-scale dextral movements along the Mid-Hungarian Shear Zone (CSONTOS et al. 1992). Clasts of deep sea sediments make up the bulk of the olistostromes in the basement of the Mátra Mountains. In the vicinity of the Darnó Hill the upper part of the accretionary complex is made up of clasts of Triassic basalt and Jurassic basalt and gabbro. The disintegrating rocks got to the thrust front partly due to the imbrication of the marginal zones of the continental plate, and partly as a result of shearing from the oceanic basins. The Szarvaskő Nappe moved above the Midde Jurassic (Bathonian–Callovian) polymict olistostrome probably at the same time, breaking off from its original place.

The melange formations of the Bükk, Szarvaskő–Darnó and Rudabánya units which formed in the foreland of the Dinaric ophiolite nappes, broke off from their basement and were thrust under other nappes probably already during the Late Jurassic. In the meantime, they underwent low-grade metamorphism (CSONTOS 2000, KÖVÉR et al. 2009a). As a result of nappe movements, strong folding occurred in the melange nappes and underthrust units (for example in the Bükk "parautochton") (CSONTOS 2000). Nappe movements almost certainly continued at the earliest Cretaceous, since the cooling of the nappe-related metamorphic units stopped only ca. 120 million years ago (ÁRKAI et al. 1995a).

In the Transdanubian Range Unit the Dinaric subduction/obduction events can be studied only indirectly. In the northern part of the unit, in the Gerecse Mountains, the debris of the ophiolite or the melange occurs already in Late Jurassic sediments (CSÁSZÁR, ÁRGYELÁN 1994). However, the area was probably situated on the subducted plate, far from the subduction front (FODOR et al. 2013a). At the beginning of the Cretaceous this distance gradually decreased, as a result of which the carbonate sedimentation was replaced first by fine-grained, and later — starting from the Barremian (130 Ma) — coarser-grained clastic sedimentation. Due to remote nappe stackings a flexural basin was formed (TARI 1994). As a result of the approaching deformation front gentle folds formed in the north-east (SASVÁRI 2008). On the other hand one can assume that the thin Early-Cretaceous succession, uplift and possible denudation of the middle part of the Transdanubian Range Unit can also be related to this process (MINDSZENTY et al. 2000).

THE SUBDUCTION OF THE ALPINE TETHYS (PENNINIC OCEAN) AND RELATING NAPPE STACKING (END OF EARLY CRETACEOUS – LATE CRETACEOUS)

At the end of the Early Cretaceous, the NW–SE-trending shortening related to the nappe stackings of the Eastern Alpines started in the Transdanubian Range Unit. At the end of the process the Transdanubian Range Unit was thrust over other Alpine units, the traces of which can be observed in the basement of the Danube Basin (TARI 1994, HORVÁTH F. 1993). The deformation accompanying the nappe stacking came to a halt or slowed down in the Middle Albian, ca. 110–105 million years ago, when continental denudation took place, locally accompanied by bauxite formation.

During the next, Middle Albian – Cenomanian sedimentary cycle the initial terrestrial or shallow marine clastic sediments were replaced by carbonate or marl sedimentation; these occurred on a shallow marine platform or in a deeper basin. The structural character of this sedimentary cycle that lasted for ca. 10 million years is unknown.

In the middle part of the Cretaceous, nappe stacking continued in the Aggtelek–Rudabánya Hills and as a result these structural movements the former order of the nappes had changed. The age of this deformation is indicated by thermochronological data ranging between 120 and 100 million years, marking the cooling age of the deformed units (ÁRKAI et al. 1995, KÖVÉR et al. 2009b).

Nappe stacking probably started around the Albian–Cenomanian boundary in the Tisza Mega-unit (CSÁSZÁR 1992). The coarse-grained turbidite successions known in the Villány–Bihor Unit, in the Transdanubian region (Bóly Basin) and in the basement of the Great Plain accumulated in foreland basins of the nappes. In the Mecsek Unit, during the Late Cenomanian and the Turonian (90–95 Ma) pelagic sediments were deposited.

In the Transdanubian Range Unit, the deposition of the sediments of the Middle Albian – Cenomanian cycle was followed once again by a contractional deformation which occurred between 95 and 85 million years ago (Turonian–Coniacian). The folding stage which started in the Aptian or the Albian and led to creation of the folded structure of the Transdanubian Range and the Litér Thrust, terminated at this time. Simultaneously, continental denudation occurred at the beginning of the Late Cretaceous and an extensive, even "denudation surface" evolved, characteristic for subtropical areas (KAISER 1997). In the karstic depressions of the surface bauxite accumulated over a large area. In the deeper parts of the developing Late Cretaceous (Senonian) basin (probably compressional; TARI 1994) fluvial and coal-bearing swamp sediments were deposited in the Santonian. I was followed by the deposition of a shallow to deep marine marl succession during the Campanian. At the same time, shallow marine platforms were formed on the inundated highs, but later these regions also became pelagic sedimentary basins (HAAS 1988, 1999). At the end of the Late Cretaceous sedimentation, a slight deformation can be assumed, although its character is rather ambiguous.

In the Tisza Mega-unit the nappe stackings and the main stage of the deformation occurred in the Coniacian (IANOVIĆI et al. 1976). At that time, the area of the former basins uplifted and eroded, while in the Santonian, deep basins were once again formed in the foreland of the nappes. In the western part of the Villány–Bihor Unit (Danube–Tisza Interfluve) the Santonian terrestrial sediments are overlain by Campanian pelagic marl; this was followed by slope sediments containing conglomerate levels and redeposited shallow marine carbonates (HAAS 1987). In the meantime, in the eastern part of the zone (Tiszántúl region) flysch-type siliciclastic formations represent this stratigraphic interval (SZENTGYÖRGYI 1989). In the Campanian and Maastrichtian, deep marine limestone and marl were deposited in the Mecsek Unit (in the Danube–Tisza Interfluve), while in the Tiszántúl Region flysch-type sandstone was formed.

PALAEOGENE DEFORMATIONS AND BASIN EVOLUTION (49–23 MA)

The subduction of the rocks of the Penninic Unit (known in the Kőszeg Window) occurred in the Palaeogene; the unit subducted beneath the older Austroalpine Nappe stack. There is no direct data in Hungary on the age of the deformation, but rocks of the similar unit (Tauern Window) unequivocally underwent Palaeogene subduction and metamorphism (KURZ et al. 1998).

In the region north of the Mid-Hungarian Shear Zone the sedimentation restarted at the beginning of the Middle Eocene. The basins had a compressional; possibly transpressional character and they were probably situated in back-arc position behind the forming Alpine–Carpathian orogenic arc (TARI et al. 1993, FODOR et al. 1994). Accordingly, the succession of the basins was characterised by a rapid subsidence that followed the deposition of the thin continental or shallow marine units. The age of the subsidence was increasingly younger from SW to NE, along the strike of the basins (BÁLDI 1986, BÁLDI, BÁLDI-BEKE 1985, KÁZMÉR et al. 2003). This character and the geodynamic nature of the basins can be clearly related to the subduction and the nappe stacking occurring on the Eastern Alpine – Western Carpathian outer front. Since the collision is believed to have occurred in the Eocene in the Alps, the evolution of the Palaeogene basins in Hungary probably started with a post-collision character.

South-east of the Mid-Hungarian Shear Zone, deep marine sedimentation took place over a long period in the Szolnok Flysch Belt, often in relation with turbidity currents; palaeontological data indicate Late Cretaceous, Palaeocene and Eocene-Oligocene ages, though the continuity of the successions is unclear (BÁLDI-BEKE, NAGYMAROSY 1993). Although there is insufficient amount of specific data (LŐRINCZ-DETZKY, SZABÓ 1993), according to the Alpine–Carpathian frame a compressional basin character can be assumed here, as well (TARI 1996). Beyond the Szolnok Flysch Belt, in the Hungarian part of the Tisza Mega-unit, Palaeogene sediments are known to occur only south-west of the Mecsek (in the vicinity of Szigetvár), in the form of terrestrial sediment remnants (WÉBER 1985); as such, here the reconstruction of the structural evolution of this period is difficult.

Significant magmatism also accompanied the Palaeogene sedimentation north of the Mid-Hungarian Shear Zone. The dominantly neutral volcanism started in the Middle Eocene. It is believed that in the Alps it is related to the slab break-off of the subducting plate (VON BLANCKENBURG et al. 1995). The main phase of the magmatism probably occurred in the Early Oligocene, as indicated by Transdanubian K-Ar data (BENEDEK 2002) and the radiometric age data of the magmatic body at Recsk, as well as the palaeontological data justifying its stratigraphic position (LESS et al. 2008). These magmatic bodies are unequivocally related to the tonalites along the Periadriatic Line; and they form their continuation (BENEDEK et al. 2004). Intrusions indicate the "mid-Oligocene" birth or reactivation of the Periadriatic Line and those of its continuation in Hungary — namely, the Balaton Line. After the major magmatic phase (31–29 Ma) the structural activity during the Late Oligocene stage is not known for certain.

Early Miocene large-scale strike-slip movement (23–19 Ma)

The Palaeogene basin evolution in the northern region ended in the Eggenburgian in such a way that the compressional (flexural) geodynamic nature of the basin did not change much compared to the conditions of the Palaeogene (Sztanó, Tari 1993). However, palaeogeographical–sedimentological data show that southward the basin had already had a structural boundary along the Balaton–Tóalmás Fault System (Sztanó 1994). It indicates also that the Palaeogene – Early Miocene basin evolution and strike-slip deformations overlap in time. As such, one of the most significant structural movement of the Carpathian Basin occurred in the Early Miocene, it was a large-scale strike-slip deformation along the Mid-Hungarian Shear Zone and along its several shear zones. Although the determination of the exact geometry of the zone needs further research, its existence and its kinematics is unequivocally interpreted in the works of most authors (Balla 1984, 1988; Balla, Dudko 1989; Kázmér, Kovács 1985; Csontos et al. 1992; Tari 1994; Fodor et al. 1998). The right-lateral strike-slip fault could have combined with thrusting at many places (Csontos, Nagymarosy 1998, Palotai, Csontos 2010). The Darnó Zone branching from the Balaton–Tóalmás Line probably operated as a reverse fault at this time (Fodor et al. 2005c). The consequence of the large dextral displacement was that the North Hungarian Palaeogene Basin was separated from its Slovenian continuation and moved eastward. Along the shear zone many strike-slip duplexes were formed and these are also shown on the map to which this text refers. The blocks that were dragged into the strike-slip zone were broken off partly from the Palaeogene basins, partly from the Velence Granite, and partly from exotic Palaeozoic–Mesozoic rocks. The right-lateral strike-slip faulting is the continuation of the similar movement of the Periadriatic Fault. In the Alps, north from the Periadriatic Fault normal detachment faulting was clearly documented along the eastern margin of the Tauern Window, (Scharf et al. 2013); while further to the north, left-lateral strike-slip faults occurred (Linzer et al. 1995). As such, in the Early Miocene the Alcapa block certainly existed and moved eastward: this is the “escape” or the “extrusion” phase (Kázmér, Kovács 1985, Ratschbacher et al. 1987).

The other consequence of the eastward movement of the Alcapa is that along its eastern edge it is thrust over the Tisza–Dacia block (Györfi, Csontos 1999, Tischler et al. 2007). The process also affected different flsych units as well at the eastern boundary of the block. It is possible that the thrusts affecting the Szolnok Flysch Belt also belong here (Pápa 1993).

The right-lateral strike-slip movements had stopped by the end of the Early Miocene, probably by the beginning of the Ottnangian (ca. 19 Ma). In the Pannonian Basin the Karpatian sediments unconformably overlie the structures of the zone, as a rule.

The development and evolution of the Pannonian Basin (19–5 Ma)

The beginning of the development of the Pannonian Basin can be defined from the onset of the lithospheric extension. According to the traditional classification, two phases can be distinguished: the syn-rift phase (coeval with the normal faults) and the following post-rift phase (Horváth F., Royden 1981). The exact age of initiation of the crust-stretching is unknown, but many authors consider it to have happened 19 million years ago (Tari 1994, Fodor et al. 1999). At this time a new sedimentary cycle began following the denudation phase that terminated the Palaeogene–Eggenburgian basin (Hámor 1985), and on the denuded Palaeogene surface in the Transdanubian Range (Kókay 1991). Extensional deformation was basically accommodated by normal faults, tilted blocks and half-grabens. In the latter asymmetric sediment wedges were deposited. These wedges were consisted of shallow marine sediments along the faults and, in the top parts of the footwall blocks and by pelagic sediments in the deeper parts of the grabens. The map depicts the largest syn-rift normal faults — namely, the Rába Fault along the Mihályi High, the marginal faults of the Zagyva, and the Hódmezővásárhely, the Békés, and the Derecske troughs. In several cases strike-slip transfer faults occurred between the normal faults (Tari et al. 1993). This led to the formation of small, but deep sub-basins, like those at Kiskunhalas.

According to the research of the last 20 years, it is clear that the extensional deformation was locally so significant that even metamorphic core complexes could have occurred in the Pannonian Basin. Above them low-angle detachment faults were formed, while the metamorphic rocks themselves were exhumed near to the surface, sometimes even reaching the surface. The exhumation was shown by thermochronological methods. Areas that had undergone such a significant extension include the tectonic window of the Kőszeg Mountains that exposed the Penninic nappes, the Muraszombat High (Mura–Zala Basin), and probably also the Algyő High (Tari et al. 1993, 1999; Tari 1996; Dunkl, Demény 1997; Fodor et al. 2003; Tari et al. 1999). A part of the deep basins of Hungary is situated above these important detachment faults (i.e. the Danube Basin, Zala Basin, Makó [Hódmezővásárhely] Trough).

Recent studies have clarified that during the syn-rift phase the degree of deformation changed spatially within the basin system. While at the western side of the Pannonian Basin the main phase of crustal stretching occurred in the Ottnangian – Early Badenian (19–15 Ma); at the eastern side the extension culminated probably at the end of the Badenian and in the Sarmatian (14–11.5 Ma) while only modest extension occurred at the same time in the western basin segment. This duality was interpreted by Fodor et al. (1999) and Fodor (2010) as two phases: an early and a late syn-rift phase.

Another feature of the syn-rift deformation is the vertical-axis rotation of the rocks . This is clearly shown in the northern Alcapa block where in two phases (between 18.5 and 17.5 Ma, as well as between 16 and 14.5 Ma), counterclockwise rotation occurred at 75–80° (MÁRTON E., MÁRTON P. 1996, MÁRTON E., FODOR 1995, MÁRTON E., PÉCSKAY 1998). Clockwise rotation of the Tisza-Dacia block may have taken place at the same time, though it could have started earlier and it certainly lasted longer in the Transylvanian region (PATRASCU et al. 1994, MÁRTON E. et al. 1999, CSONTOS et al. 2002).

The syn-rift deformations were replaced in the Transdanubian Range by strike-slip deformation. As a consequence WNW–ESE–trending, right-lateral strike-slip faults were formed, and these are well-known in the mountains: for example the Telegdi Roth, Herend and Padrag faults (MÉSZÁROS J. 1983, KÓKAY 1996). Also the Csesznek transpression belt can be assigned here (KISS A., FODOR 2007). Dextral movements were accommodated at the eastern fault terminations in the form of steep reverse faults and low-angle thrusts (TARI 1991). An element of this could be the Nagymező Sliver of the Balaton Highland (FODOR et al. 2005b). Along the Balaton Zone the combination of strike-slip faults, steep reverse faults and low-angle thrusts can also be observed (CSONTOS et al. 2005, TÖRŐ et al. 2012). As such, at the end of the Sarmatian or possibly at up to the earliest Pannonian, the bulk of the Transdanubian region was characterised by transpression deformation. Its duration differs region by region; in a narrower sense it lasted for 12–11 Ma, in a broader sense for 14–10 Ma.

The transpression deformation probably extended to the whole Pannonian Basin in the Late Sarmatian and possibly at the beginning of the Pannonian. According to several models it resulted in the absence (denudation?) of the Sarmatian sediments. This is the first inversion phase of the Pannonian Basin (HORVÁTH F. 1995).

In the Late Miocene the Pannonian Basin underwent a general subsidence; this is the classical post-rift phase (ROYDEN, HORVÁTH F. 1988). As a result of the subsidence and the accompanying general rise in the water level, the water of the Pannonian Lake reached its largest extent around 9.5–9 million years ago: at that time most of the mountain ranges were covered by water. Moreover, the salinity of the lake was clearly different from the normal marine salinity; due to the salinity and water balance of the lake, the salinity approached a fresh-water state (UHRIN 2011). Progressive isolation of the lake from the sea resulted in the formation of endemic flora and fauna (MAGYAR 2010).

The general crustal subsidence was locally increased by reactivation of fault movements. In the so-formed lake, diverse sedimentary environments evolved. From the north-western, northern and north-eastern margins of the lake large amounts of sediments were transported byrivers into the lake. Since the intensity of the subsidence was not as intense as that of the sediment transport, the lake was gradually filled up and its shoreline moved southward. Along the shoreline delta environments occurred, forming several tens of metres-thick sediment cycles (SZTANÓ et al. 2010). The deltas continued in shallow shelves, connected by a slope to the deep basins that were mostly characterised by clayey sedimentation. On the slope (partly in canyons) gravity mass movements transported sediments which were deposited at the toe of slopes, and on the bottom of the deep basin. Accordingly the deep lacustrine marls are usually overlain by redeposited sandy sediments. This sedimentation pattern has been analysed and specified since the 1980s by many studies (BÉRCZI, PHILIPS 1985; JUHÁSZ GY. 1988; JUHÁSZ GY. et al. 2007; UHRIN et al. 2009; MAGYAR et al. 2013; SZTANÓ et al. 2013a, b).

The direction of the slope progradation was basically determined by the faults that were active during the sedimentation. As such, active deformation occurred also in this phase, primarily through the rejuvenation of some former syn-rift faults (PALOTAI, CSONTOS 2010, TÖRŐ et al. 2012, FODOR et al. 2013). Furthermore, the intensity of the deformation after 8 million years seems to decrease.

THE INVERSION OF THE PANNONIAN BASIN (5–0 MA)

The post-rift subsidence of the Pannonian Basin and the parallel fault reactivation was replaced by the inversion of the basin. The cause of this is clear: along the Carpathian arc the roll-back of the subduction was no longer possible, yet the Adriatic plate continued to excerpt a push from the south (BADA et al. 2007). The inversion of the basin started from the south and gradually expanded northwards (TARI 1994, FODOR et al. 2005a). Accordingly, in the southernmost areas the first folded structures were developed ca. 8–7.5 million years ago in the vicinity of Budafa (UHRIN et al. 2009). As a result of the inversion many syn-rift normal faults were reactivated as reverse faults or blind reverse faults (DANK 1962, HORVÁTH F. 1995). Many strike-slip faults were reactivated, for example the Balaton Line (MAGYARI et al. 2005). An important consequence of the inversion was the uplift of the mountains and hilly areas and the continuing subsidence of the plains (RÓNAI 1985, HORVÁTH F., CLOETHING 1996, RUSZKICZAY-RÜDIGER et al. 2005). Many studies dealt with the earthquake hazards in relation to the active faults (e.g. HORVÁTH F. et al. 2005), but further investigations are still required.

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