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The Tisza Mega-unit (Tisia Terrane) forms the basement of the Pannonian Basin south of the Mid-Hungarian Lineament (see Fig. 1 in “Introduction”). In the territory of Hungary the pre-Neogene basement crops out only in two relatively small, isolated areas in South Transdanubia – the Mecsek Mountains and Villány Hills (Figs. 1, 2 in “Introduction”). However, more than 3,000 wells, oil and uranium prospecting boreholes provide information on the geologic features of the basement and the younger overburden of the basins. Significant parts of the Tisza Mega-unit extend into Croatian (Slavonian), Serbian (North Vojvodina) and Romanian (West Transylvanian) territories (Szederkényi 1974, 1984; Kovács 1982; Fülöp 1994; Kovács et al. 2000). In Romania and Croatia large mountains and uplands (Apuseni Mts., Papuk Mts., Psunj Mts.) provide an excellent opportunity to study its structural setting and stratigraphy.

The Tisza Mega-unit forms a more than 100,000 km²-large lithosphere fragment broken off of the southern margin of Variscan Europe during the Jurassic (Bathonian), and after complicated drifting and rotational processes it occupied its present-day setting in the Pannonian Basin during the Early Miocene (Balla 1986; Csontos et al. 1992; Horváth 1993). The crystalline basement and the overlying Upper Palaeozoic and Mesozoic overstep sequences show heterogeneous lithology and lithostratigraphy, indicating various phases of geologic evolution.

On the basis of these features both pre-Mesozoic and Mesozoic sequences are classified into numerous units and subunits. The pre-Mesozoic (Variscan) basement of the Tisza Mega-unit is actually a composite terrane which was accreted during the Variscan Orogeny. During the Alpine cycle true terrane dispersion did not occur, but facies zones were differentiated in the Jurassic and nappe-systems were formed in the Cretaceous.

2.1 Pre-Variscan to Variscan Evolution

The Tisza Mega-unit is a Variscan orogenic collage which was accreted during the Carboniferous–Permian, becoming part of the European continent. However, at present the boundaries of the Tisza Mega-unit are determined by the Alpine structural evolution (Fig. 1 in “Introduction”, 1.1). Its north-western boundary is the Mid-Hungarian Lineament (Szepesházy 1975; in the earlier literature: Zagreb–Kulcs–Hernád Lineament: Wein 1969; Zagreb–Zemplin Lineament: Grecula and Varga 1979). It is overthrust southwestward onto the ophiolite-bearing Sava Zone (in sense of Schmid et al. 2008; Ustaszewski et al. 2008). Overthrusts of the Vardar and Mures (Transylvanian) ophiolite belts mark the southern and southeastern boundary of the Tisza Mega-unit (Săndulescu et al. 1981; Csontos and Vörös 2004; Schmid et al. 2008).

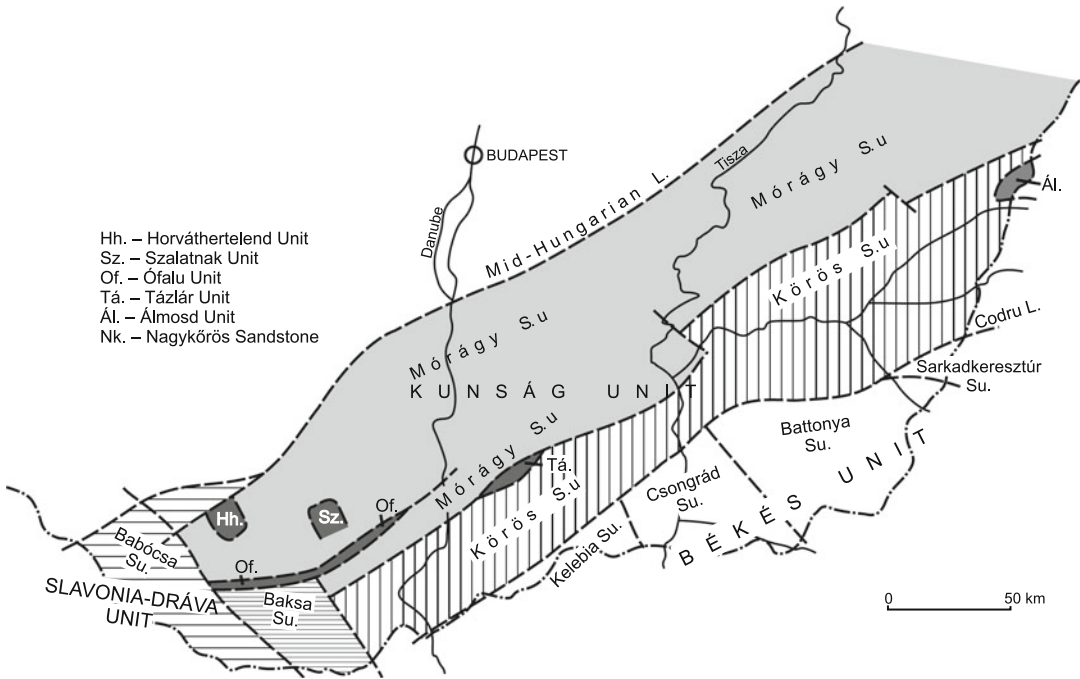


Fig. 2.1 Pre-Alpine structural units of the Tisza Mega-unit (After Szederkényi 1997).

The northeastern border is provided by the North Transylvanian or Somes Line.

The name “Tisza Mega-unit” is derived from the term Tisia, of the so-called “median-mass concept” which arose at the beginning of the twentieth century in the Hungarian geology (Prinz 1914, 1923, 1926; Lóczy 1918; Kober 1921). A common feature of these hypotheses was the assumption of an old (Palaeozoic or older) and rigid crystalline central massif which – much as a rigid boot-stretcher – strained the Carpathians during their uplifting. The name of this hypothetical ancient core-massif was Tisia, named after the Tisza River by Prinz (1914). Today it is plausible that the basement of the Pannonian Basin is not a rigid crystalline massif. However, the old name has been preserved, although in a sense significantly different from the original one.

2.1.1 Crystalline Complexes

Within the crystalline basement of the Tisza Mega-unit, three units (terrane) have been distinguished, separated from each other by major

fracture zones. The units can be subdivided into subunits bounded by fracture zones of secondary importance. Thus, all units and subunits have tectonically determined extensions and boundaries and show characteristic lithostratigraphic columns and evolution.

The pre-Alpine units (terrane) and the subunits constituting them are as follows (Kovács et al. 2000; Fig. 2.1):

Slavonia–Drava Unit (Terrane)

Babócsa Subunit

Baksa Subunit

Kunság Unit (Terrane)

Mórággy Subunit

Kőrös Subunit

Békés Unit (Terrane)

Kelebia Subunit

Csongrád Subunit

Battonya Subunit

Sarkadkeresztúr Subunit

In addition to the large units (terrane) listed above, several small units (“outliers” – nappe remnants, tectonic wedges) are also found in the Tisza Mega-unit which show entirely different lithological and metamorphic features than

those of large terranes. They are: Horváthertelek Unit, Szalatnak Unit, Ófalu Unit, Tázlár Unit, Álmosd Unit.

2.1.2 Lithostratigraphy of the Tectono-stratigraphic Units and Tectono-metamorphic Evolution

Conventional stratigraphic methods cannot be applied in the metamorphic complexes because the original features of the rocks are largely or totally destroyed, and boundaries of metamorphic units generally cut across those of the pre-metamorphic lithological units. The basement of the Tisza Mega-unit is affected, in the overwhelming majority of cases, by medium to high-grade metamorphism, as a result of which the former units have been amalgamated. Consequently, new lithological units were developed, with new borders and a new metamorphic age reflecting the age of last heating.

The lithostratigraphic chart (Fig. 2.2) displays both metamorphic features and the tectono-metamorphic history of the rock columns. The columns indicate the time-range of the protolith accumulation. Top of the columns shows the timing of the last progressive metamorphic event. The outliers (nappe remnants and tectonic wedges) either “hang in space” in the lithostratigraphic table (because their parent complexes are unknown), or within the rock column of the host unit.

Subunits of the Slavonia–Drava and Kunság Unit are relatively autochthonous (“parautochthonous”) compared to those of Békés Unit, representing the crystalline basement of an Alpine nappe-system.

The lateral relationships of metamorphic lithostratigraphic units are clearly expressed in the general similarity of protoliths and metamorphic character as well as in time of culmination of metamorphic phases, which are accompanied by migmatization and palingenesis. Apart from the outliers there are some essential differences between the two main groups of lithostratigraphic charts. They are manifested in the following points:

- Pre-Variscan deformation has only been detected in the “parautochthonous” units;
- Variscan late kinematic heating (with andalusite mineralisation) occurs primarily in the nappe units;
- Late Cretaceous contact metamorphism belonging to the banatitic intrusions in the Hungarian part of the Tisza Mega-unit only occurs in the nappe units.

2.1.2.1 Slavonia–Drava Unit

The Slavonia–Drava Unit is located in southeastern Transdanubia, extending southward into eastern Croatia. It is bordered by the Mid–Hungarian Lineament to the northwest and by the Mecsek–Kajka Fracture Zone to the east. No outcrops of this terrane are found in Hungarian territory, but there are some in the Papuk and Psunj Mts. in Croatia. A general NW–SE-striking of formations is characteristic all over the Slavonia–Drava Terrane, underscoring its relationship to the East Croatian crystalline basement.

Babócsa Complex

This complex makes up the northwestern part of the Slavonia–Drava Unit. It consists mostly of medium-grade gneiss with subordinate micaschist and amphibolite intercalations. Apart from an uncertain Caledonian measurement (Jantsky 1979) a double Variscan metamorphism has been established. The first phase is represented by a Barrow-type deformation with 6–9 kbar pressure and 17–27°C/km thermal gradients; the second is an andalusitic, higher-temperature phase (34°C/km thermal gradients – Árkai 1984; Török 1989). On the SE part of the complex the crystalline rocks are overlain by Upper Carboniferous molasse.

Baksa Complex

This unit makes up the crystalline basement of Villány Hills and its northern foreland up to the Mecsek Mts. Its southwestern border is a fracture zone (transcurrent fault) running between the Kunság and Slavonia–Drava Units (Kassai 1977). Petrographically this complex consists of weakly-folded migmatite, gneiss, micaschist,

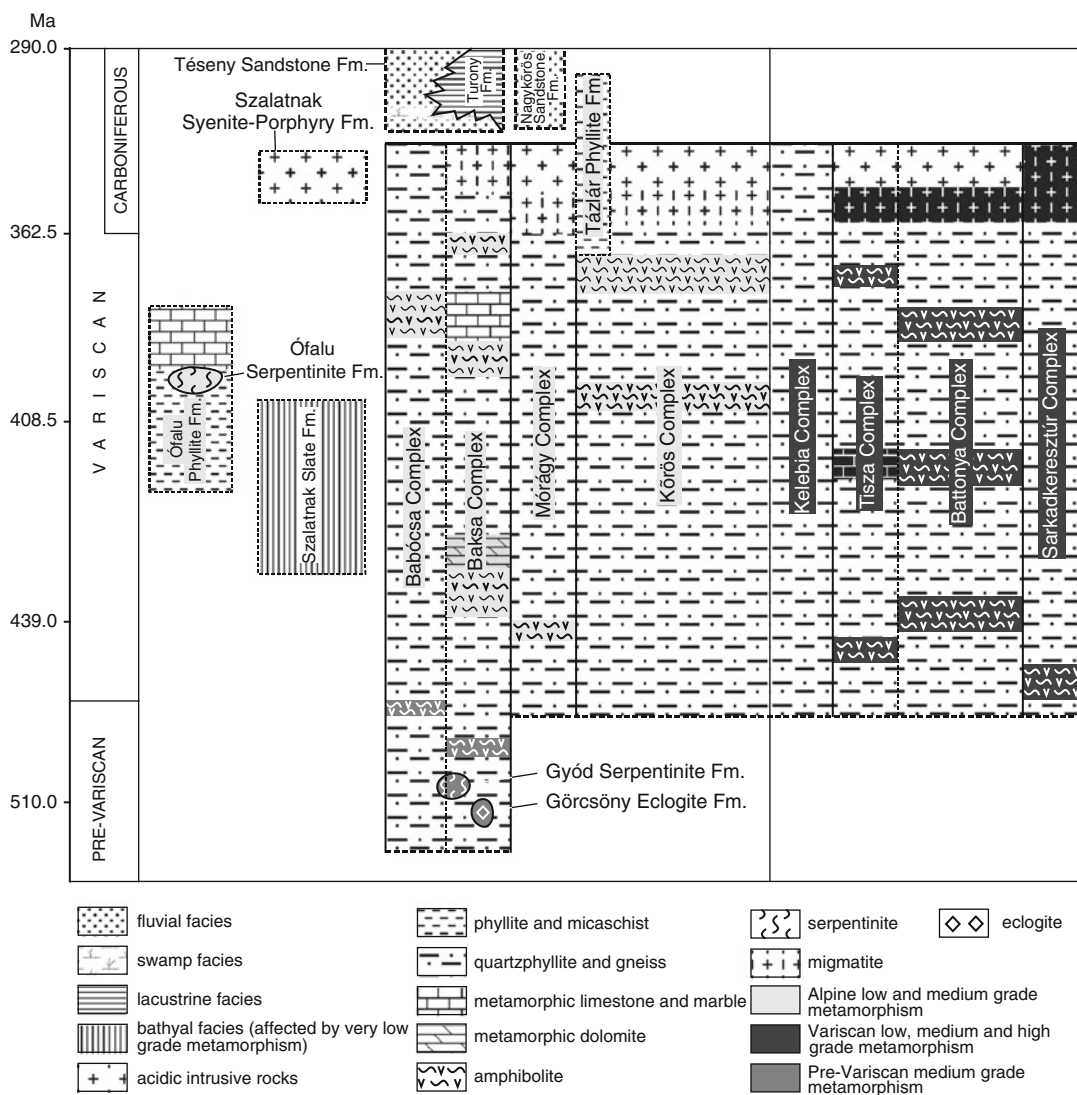


Fig. 2.2 Pre-Alpine metamorphic complexes of the Tisza Mega-unit and their sedimentary cover (After Császár (ed.) 1997)

marble, dolomitic marble, and calc-silicate gneiss (Fig. 2.3) characterised by an isograd system with sillimanite to chlorite zones and isograds showing a southwest progressive trend (Szederkényi 1976).

The thickness of this complex exceeds 10 km. Two marble and dolomitic marble members (250 m and 25 m-thick, respectively) occur in the sillimanite zone accompanied by fairly thick (23–30 m) amphibolite beds. At the northern margin of the complex high-temperature over-

printing with andalusite was encountered (Lelkes-Felvári and Sassi 1983). Altogether, at least three phases of polymetamorphism could be recognised.

2.1.2.2 Kunság Unit

The Kunság Unit extends over the area located between the Middle Hungarian Lineament and the Mecsekalja Fracture Zone as well as the northern front of the Békés Unit (South Hungarian Nappe Belt). An eastward continuation towards

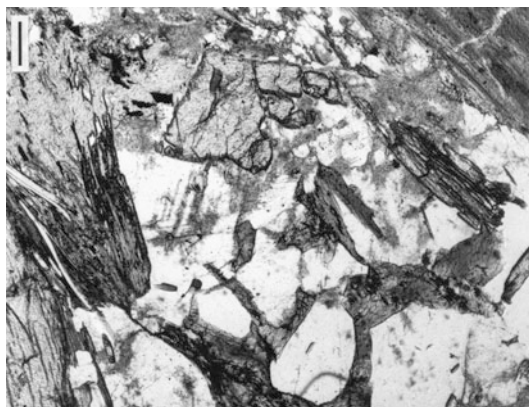


Fig. 2.3 Sillimanite–staurolite gneiss from the Baksa Complex, core Baksa-2, 950.4 m. Scale bar: 0.2 mm (Photo: T. Szederkényi)

the Apuseni Mts. of Romania can be postulated, but a true correlation between them is lacking so far (c.f. [Kräutner 1996–1997](#)). In Hungary, crystalline rocks of this terrane only crop out in the Mecsek Mts. (Mórág Hills).

Mórág Complex

It constitutes the Mórág-Kecskemét granitoid range and the accompanying migmatite-gneiss-micaschist flanks on both sides. The most characteristic part of the complex is the granitoid range itself. Forming an axial belt of an ENE-WSW-striking synclinal zone this body is about 200 km long and 25–30 km wide, forming a continuous zone from Szigetvár (South Transdanubia) to Szolnok (central part of the Great Hungarian Plain), where it disappears beneath the Upper Cretaceous – Paleogene flysch complex ([Jantsky 1979](#)).

It is a granite-granodiorite-diorite rock association; 340–354 Ma old (U/Pb dating; [Klötzli et al. 2004](#)); the 307–312 Ma age data can be regarded as cooling time ([Lelkes-Felvári and Frank 2006](#)). It contains biotite and/or amphibole rich xenolites (Fig. 2.4) which was dated at 440–400 Ma (Rb/Sr ages; [Svingor and Kovách 1981](#); [Kovách et al. 1985](#)), suggesting a pre-Variscan (Caledonian?) metamorphic event. Detailed investigations carried out in connection with the establishment of a radioactive waste

repository in the Mórág Hills (SE Mecsek), resulted in the discovery of a huge monzodiorite body of ENE – WSW strike in the axial part of a granitoid range. All rocks in this body are of monzonitic character. In contrast to the former hypothesis about the migmatitic origin of granitoids ([Szádeczky-Kardoss 1959](#)), the major characteristics of the rock association indicates rather plutonic than migmatitic origin ([Király 2009](#)). The granitoids are syn-collisional, S-type, mixed meta-, and peraluminous ([Buda 1981, 1985, 1995](#)). They are accompanied by crystalline schists showing typical polymetamorphism which flank the syncline. In the first phase of Variscan deformation a Barrow-type event took place at 6–8 kbar pressure and 14–26°C/km thermal gradients ([Szederkényi et al. 1991](#)). In the second phase a low-pressure/high temperature retrogression occurred along the Mecsekalja Fracture Zone and in the eastern continuation of Mecsek Mts. ([Lelkes-Felvári et al. 1989](#)) with late kinematic (322 Ma) andalusites. The granitic rocks were affected by multistage deformation subsequently ([Maros et al. 2010](#)).

Körös Complex

It constitutes a more than 250 km-long narrow, discontinuous granite range embedded in a 15–20 km-wide migmatite belt (Fig. 2.1). Within the range five lens-shaped granite bodies, 5–10 km wide and 15–25 km long, can be found. They are made up of S- and I-type porphyroblastic biotite–granite/granodiorite rocks ([Buda 1985, 1995](#)), and were formed in the axis belt of a syncline. The granite-migmatite range is accompanied on both sides by medium-grade ortho- and paragneiss – micaschist – amphibolite associations as flanks of the ENE – WSW –striking syncline. Based on revision of the isotope ages of the Körös Complex two metamorphic events were determined ([Lelkes-Felvári et al. 2003](#)). The first was dated at 310 Ma; it is characterised by a general Barrow-type metamorphism. It was followed by a local overprint in the NE part of the Körös Complex at 202–299 Ma. From the same rock body 330 Ma age was measured on the zircon and 310–295 from biotite of orthogneiss ([Balogh et al. 2009](#)). Apart from the heated area, the phase

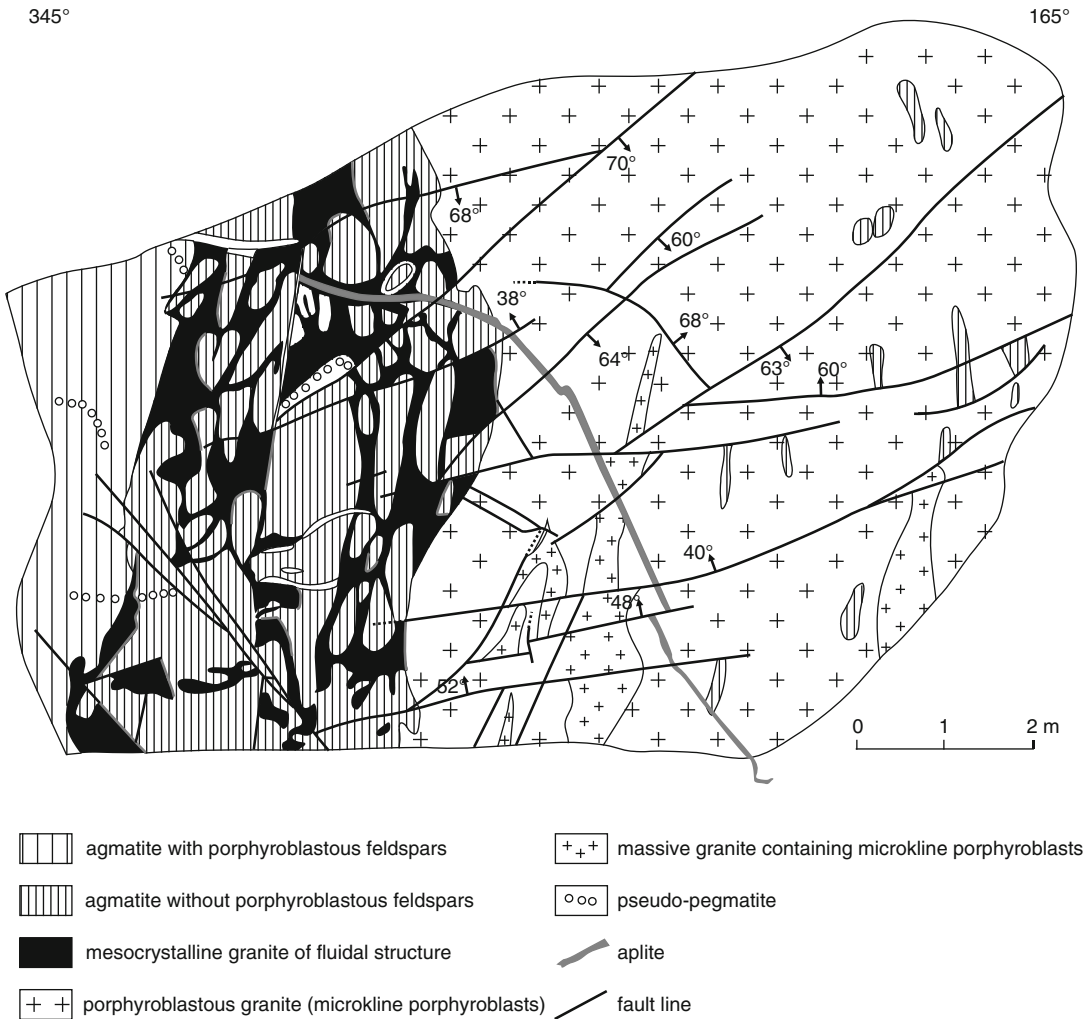


Fig. 2.4 Exposure of the Mórág Granite, Pince Hill, Mórág (After Szederkényi 1987)

typified by andalusite is absent from the metamorphic history of the Körös Complex. In its axial zone several small eklogite bodies (Göröcsöny Eklogite) were encountered.

of this terrane is uncertain due to the lack of relevant tectonic and lithologic evidence. However, based on sporadic data it may extend south of the Mecsek Mts. (Baksa Unit).

2.1.2.3 Békés Unit

This unit extends over the area of the southern part of the Great Plain, corresponding to the western continuation of the Romanian Codru and Bihor Nappe System into Hungary and Northern Serbia. Its northern border coincides with the northern front of the Upper Cretaceous nappe system (Békés–Codru Zone; South Hungarian Nappe System). The western continuation

Kelebia Complex

This complex is located in the westernmost part of the Békés Unit. It is limited by a nappe boundary to the west and north, and by the Ásotthalom–Bordány Depression to the east. Southward it extends into Serbian territory. Strongly folded two-mica-schist (and locally chlorite-schist) forms this low and medium-grade metamorphic rock complex of unknown thickness. A Barrow-

type Variscan metamorphic phase and several small, Upper Cretaceous quartz diorite intrusions and dikes with definite contact zones characterise the progressive metamorphism of the complex.

Csongrád Complex

This part of the Békés Unit is limited by a nappe boundary to the north, by the Ásotthalom–Bordány Depression to the west, and the Makó Trough to the east. Its southern border is in the Serbian Bačka. A characteristic peculiarity of the Csongrád Complex is a 200 m-thick marble/dolomitic marble member which was encountered near Szeged; it has the distinction of being the only carbonatic rock association in the crystalline basement of the Great Plain. Besides this marble a small, deep-plutonic granite occurrence and related migmatites, as well as medium-grade, slightly folded gneiss–micaschist, are also typical. The main metamorphic events were a first Variscan (Barrow-type) phase with 6–8 kbar pressure and 500–570°C temperature at 350–330 Ma, a second Variscan phase with blastomylonitisation at 330–320 Ma, a third, late kinematic, high-temperature and low-pressure retrogression ($P = 3\text{--}4$ kbar pressure and $T = 580\text{--}600^\circ\text{C}$ temperature) at 320–315 Ma, and finally Late Cretaceous quartz diorite magmatism and related contact metamorphism. The latter is comprised of small, elongated intrusions and accompanied by relatively broad (400–600 m), tourmaline-rich muscovite schist aureoles, with a ENE–WSW strike (Szederkényi 1984; Szederkényi et al. 1991).

Battonya Complex

It is known in a 15–25 km-long and 10–15 km-wide body consisting mainly of granite and a few associated migmatite and crystalline schist occurrences. In Hungary, the boundaries are the Makó Trough to the west, the Békés Basin to the east and the nappe boundary to the north. Porphyroblastic orthoclase–biotite granite and associated enclaves make up the predominant portion of this deep-plutonic body. This pluton forms a more than 150 km-long and not very wide continuous range, stretching from the Ser-

bian Bačka to the Apuseni Mts. of Romania. The deep-plutonic granite magma, after in-situ melting, moved upward a little as an intrusion during the Variscan late kinematic period (Szepesházy 1969; Szederkényi 1984; Kovách et al. 1985). All deformational and age data are the same as those of the Csongrád Complex.

Sarkadkeresztúr Complex

This is an isolated, 15 km-long and 5 km-wide crystalline ridge on the eastern side of the Békés Basin which consists of light grey gneiss–granite. It is accompanied on both sides of the range by a high and medium-grade gneiss–micaschist–amphibolite association showing the same deformational characters and age as those of the Csongrád Complex (Szederkényi 1984).

2.1.2.4 Outliers

Lithostratigraphic units encountered in small nappe remnants, or in tectonic wedges are characterised below. Their location is shown in Fig. 2.1.

Ófalu Phyllite

Meta-graywacke (Fig. 2.5), phyllite, crystalline limestone and interbedded meta-basalt, actinolite–schist, porphyrite and porphyroid form a low-grade metamorphic sequence which is stuck as a wedge within the “Mecsekalja Tectonic Belt” in a length of 40 km and a width of more than 2 km. The weakly folded and tilted (locally vertical) rock slabs are strongly sheared in most cases except for a few siliceous shale and crystalline limestone intercalations. The silicification of these exceptions is attributed to synsedimentary submarine volcanic activity. Some plant remnants and conodont fragments have been preserved. Since the carbonised plant remnants show supporting tissue the fossils must have been derived from botanically fairly advanced plants (Kedves and Szederkényi 1997), suggesting that the age of the protoliths is not older than Late Silurian. The strongest shearing took place at the northern margin of the formation and due to considerable friction-generated heat a weak melting event also developed within it.

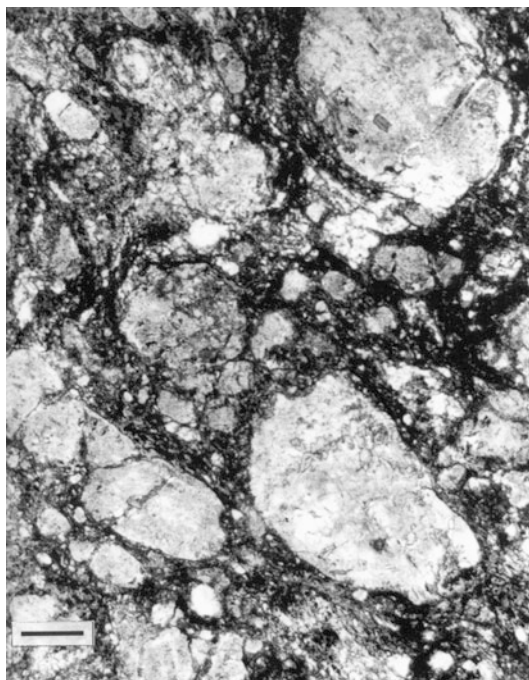


Fig. 2.5 Meta-graywacke from the Ófalu Formation. Scale bar: 5.5 mm (Photo: T. Szederkényi)

Ófalu Serpentine

It is near a small (12 m wide and about 100 m long), nearly vertical serpentinite body; a tectonic wedge within the Ófalu Phyllite. These rocks of Iherzolitic origin (Ghoneim and Szederkényi 1979) are interpreted as an obducted lower lithosphere remnant (Balla 1983b).

Gyód Serpentine

It consists of two occurrences about 5–6 km long and 600–700 m wide (Fig. 2.1). Serpentine and talc-schist at Helesfa form a nearly vertical lenticular body, wedged into Variscan granites along broad shearing zones. The complex consists of sheared and perfectly serpentinised harzburgite showing a diapiric structure (Szederkényi 1974, 1977). The other occurrence at Gyód is located at the northern margin of the Baksa Sub-unit. Its host rocks are medium-grade crystalline schist belonging to the Baksa Complex. No traces of shearing are observable, so the process of serpentinisation could not have been completed. In a central narrow slab less serpentinised

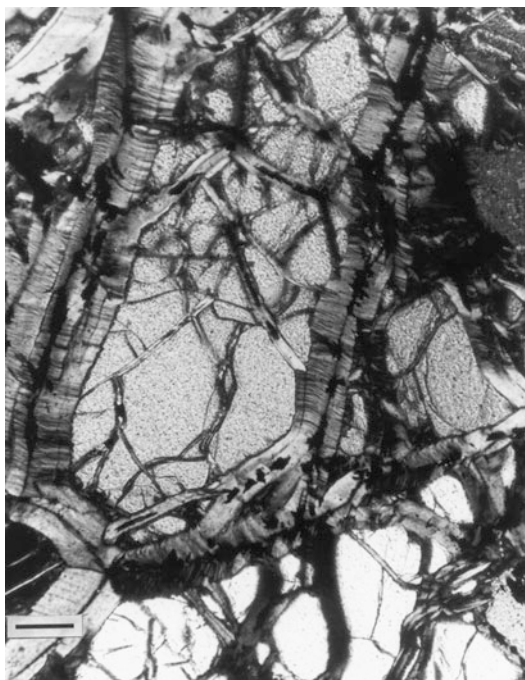


Fig. 2.6 Serpentinised harzburgite from the Gyód Serpentine. Core Gyód-2, 84.8 m. Scale bar: 0.05 mm (Photo: T. Szederkényi)

harzburgites were preserved (Fig. 2.6). According to Balla (1983b) both members of the Gyód Serpentine can be regarded as remnants of obducted oceanic lithosphere.

Görcsöny Eclogite

In the Görcsöny area, south to the Mecsek Mts. and in the southern part of the Great Hungarian Plain small eklogite bodies were encountered in the metamorphic basement in several drillings. These bodies represent partly tectonic wedges in paragneiss (Ravasz-Baranyai 1969, partly xenoliths in orthogneiss (Zachar 2008). The eklogites are generally strongly altered and they consist mostly of simplectitic rocks. However, in one of the occurrences at Jánoshalma, the xenoliths which are enclaved in orthogneiss largely retained their original characteristics. The typical minerals are as follows: garnet, klinopiroxene, kyanite, K-feldspar and secondary amphibole, plagioclase, klinozoizite and phengite. 710°C temperature and 26 kbar pressure were calculated by

Zachar and M. Tóth (2009). The occurrences of the Göröcsöny Eklogite as well as Gyód, Helesfa, and Ófalu serpentinites are interpreted as parts of a supposed 5–10 km broad Variscan suture zone (Szederkényi 1977, 1984; Tóth and Zachar 2002; Zachar 2008).

Szalatnak Shale

It is comprised of strongly folded, dark grey shale which was encountered at two localities. At Szalatnak, in the Eastern Mecsek Mts., basalt agglomerate (80 m-thick) occurs within the shale. The more than 1,500 m-thick sequence is tectonically underlain by the Mórágý Complex. Thin, siliceous stripes are characteristic for the shale. Thin anthracite intercalations also occur, mainly in the lower member, containing graptolite fragments (Oravecz 1964) and a characteristic Llandoveryan conodont and Muellerisphaeridae fauna (Kozur 1984). The sequence was affected by very low-grade metamorphism (prehnite-quartz facies; Szederkényi 1974) which turns into a low-grade one (Árkai et al. 1996) in the lower part of the formation. The Szalatnak Shale extends over an area of 200 km² covered by Permian and/or Lower Triassic sandstone. Tectonically, it is a Late Variscan nappe remnant of unknown vergency. During the Carboniferous (before the nappe movements) a small (about 1 km-large) granodiorite body was intruded into the lower member. Its Rb/Sr ages (Svingor and Kovách 1981) indicate a Variscan late kinematic origin (328–310 Ma). Its geochemical characteristics differ from those of the Mórágý Granite. Beneath Karpatian–Badenian terrestrial sediments similar shale was encountered in the western foreland of the Mecsek Mts. (Horváthertelend – see Fig. 2.1)

Tázlár Phyllite

This unit makes up an approximately 15 km-long and 300 m-wide double body wedged into the gneiss of the Mórágý Subunit along a NE-SW-striking fault zone in the central area of the Danube–Tisza Interfluve (Fig. 2.1). The lithology of these bodies is defined as greenish-grey carbonate-phyllite with black graphitic phyllite.

Their age is uncertain; according to Fülöp (1994) it may be Early Palaeozoic or Early Carboniferous.

Álmosd Formation

This is a low-grade chlorite-schist/two-micas-chist and graphite-bearing biotite-schist association which forms an Upper Cretaceous nappe outlier (over an area of about 20 km²) thrust over the metamorphics of the Kőrös Complex at the Romanian–Hungarian border. It shows NW vergency and is genetically the same as the low-grade metamorphics of the Békés Unit (South Hungarian Nappe Belt).

2.1.3 Protoliths and Polymetamorphic Deformations

In the pre-Alpine basement of the Tisza Mega-unit the prevailing rock association consists of gneiss and micaschist as well as related anatectic granitoids which were derived from graywacke/argillite-type sedimentary sequences (Szederkényi 1984), with mafic lava and tuff intercalations several metres thick. The latter generally show a tholeiitic basalt and tuff character (Szederkényi 1983). Based on geochemical data and discrimination analyses these volcanics represent back-arc basin tholeiite (T-MORB; Tóth 1995). In the rock sequences of South Transdanubia and the southern part of the Great Plain some acidic tuff intercalations also occur, indicating a presumed continental margin volcanic effect. In the Baksa and Csongrád Complexes carbonatic (marl, limestone, dolomitic marl, dolomite) interlayers several metres thick occur in a psammitic–pelitic sedimentary sequence. Carbonatic layers or lenses are completely absent from the other units.

Apart from several “outliers” (eroded fragments of nappes and tectonic wedges) the metamorphic evolution comprised one or more progressive and several retrograde phases. According to basic metamorphic conditions three characteristic fields can be separated (Fig. 2.7). (1) High-pressure, relatively low-temperature metamorphism (25 kbar pressure and 12°C/km thermal gradient) was encountered

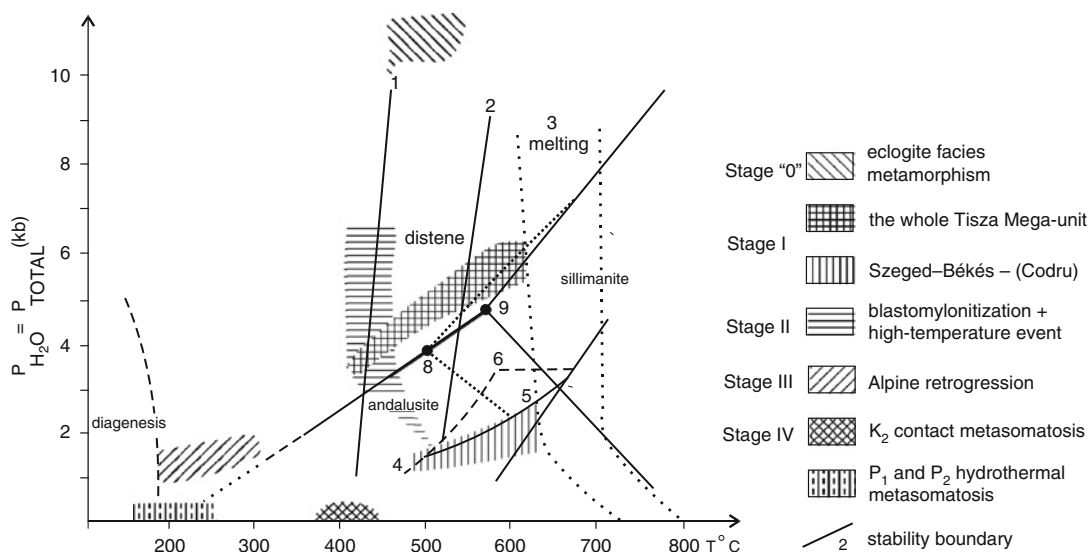


Fig. 2.7 P-T diagram of the metamorphic events in the Tisza Mega-unit (After Szederkényi 1997)

in a few smaller, covered occurrences extending along the axis of Kunság Unit. (2) Metamorphism characterized by medium-pressure and temperature (Barrow-type) deformation with 4–6.5 kbar pressure and 24–27°C/km thermal gradient. This type predominates in the Kunság Unit but was detected in the entire area of the Tisza Mega-unit. (3) Low-pressure and high temperature metamorphism characterised by 2–3 kbar pressure and 70°C/km thermal gradient overprinting the Barrow-type metamorphic complex mainly in the southern and north-eastern sections of Tisza Mega-unit.

The complete succession of polymetamorphic deformations can be interpreted as follows:

1. The first phase corresponds to the previously described deformation type 1.
2. A medium pressure and temperature Barrow-type progressive metamorphic event corresponding to deformation type 2. It is probably the very first manifestation of Variscan metamorphism and the most powerful deformation in the metamorphic history of the Tisza Mega-unit.
3. Blastomylonitisation and a subsequent low-pressure and high-temperature event occurred in the metamorphics of southern part of the Tisza Mega-unit belonging to the deformation type 3.
4. Tensional and compressive phase of Tethyan development during the Mesozoic ended with retrograde metamorphism in the Late Cretaceous nappe movements.
5. Thermal (contact) metamorphism related to Late Cretaceous-Eocene quartz diorite intrusions which were encountered in the Békés Unit.
6. Hydrothermal metasomatism linked to the subsequent volcanic events (Lower Permian, Lower Cretaceous, and Miocene); effect of thermal waters in the fractured basement domes.

2.1.4 Tectono-metamorphic Events

Based on the determined ages of the succession of metamorphic events and related deformation characters, as well as on P-T conditions and the presence of specific “indicator rocks” (eclogite, blueschist, ultramafics), the following pre-Alpine tectono-metamorphic evolution can be established for the crystalline rocks of the Hungarian part of the Tisza Mega-unit:

1. A Variscan orogenic event (its exact age is unknown). Its remnants are preserved in a narrow (5–10 km wide), poorly explored belt in the axis on the Kunság and Slavonia-Drava

units. High-pressure and low-temperature eclogites occur in some places (Kőrösvidék, Jánoshalma, Görcsöny), as well as obducted, serpentinised ultramafic bodies in South-east Transdanubia have been interpreted as Variscan in general.

2. A Variscan collisional event that can be regarded as culmination of Variscan orogeny. When accretion of Variscan Europe occurred the formation of crystalline rock associations was accompanied by mega, macro and micro-folding, shearing and blastomylonitisation. Palingenetic (?) granite belts were formed in the axial zones of synclines during the same period, although the Late Variscan low-pressure and high-temperature regime (late orogenic heating in the 330–270 Ma period) undoubtedly contributed to the granitisation as well.
3. Following granite genesis but prior to the Late Carboniferous basin formation several important tectonic events took place. They are as follows:
 - Nappe formation, producing nappes of unknown vergency. Their remnants were encountered at Horváthertelend and Szalatnak,
 - NW-SE-striking transcurrent faulting bordering the Slavonia–Drava Unit to the east, and
 - Strike-slip faults with an ENE-WSW strike (the oldest manifestation of the so-called “Mecsekajka Fracture Zone” as well as the Baja–Tázlár–Túrkeve–Nyírábrány Fracture Zone) which surround the tectonic wedges of the Ófalu and Tázlár units.

varying thickness. Permian formations cropping out in the Mecsek Mts. (Fig. 2.8) have been known for a long time, but the covered Upper Carboniferous–Permian ones only for 30–40 years, as a result of uranium prospecting.

The oldest such overstep formation is of Late Carboniferous age. It overlaps both subunits of the Slavonia–Drava Unit. The next overstep stage is represented by the Lower Permian Korpád Sandstone and/or Gyűrűfű Rhyolite, which appear in every unit but did not cover them entirely. The third stage occurred in the Early Triassic as manifested by the widespread extension of the Jakabhegy Sandstone.

2.2.2 Late Carboniferous–Permian Cover of the Slavonia–Drava Unit

Variscan post-orogenic sedimentation began earlier in the Villány area than in the Mecsek one and produced a Late Carboniferous and/or Early Permian, molasse-type overstep sequence which was draped over the eroded surface of the crystalline basement (Jámbor 1969; Hetényi and Ravasz-Baranyai 1976; Kassai 1976; Barabás-Stuhl 1988). It covers the basement between the Villány Hills and Mecsek Mts. as well as in the Drava Basin, and also occurs in erosional patches or wedges within tectonic zones in the area of the Great Plain. The lithostratigraphic classification of this sequence is as follows (Fig. 2.9).

The Téseny Sandstone is found partly in the northern foreland of the Villány Hills and also in the Drava Basin. It is underlain by the Baksa and Babócsa metamorphic complexes and overlain by the Turony Formation in the Villány Hills and/or Miocene–Pliocene sediments of the Drava Basin. This 1,500 m-thick formation consists of a cyclic alternation of grey and dark grey conglomerate, sandstone, siltstone, shale and thin coal (anthracite) seams with signs of very low-grade (burial) metamorphism. The dark grey fine-grained sandstone is full of remnants of a rich fern flora permitting this sequence to be assigned a Westphalian D as well as Stephanian age.

2.2 Post-Variscan Evolution

2.2.1 Late Carboniferous–Permian Continental Formations

Non-metamorphic (locally anchimetamorphic) molasse-type overstep sequences cover the crystalline basement of the Tisza Mega-unit, locally. They show varying stratigraphy and strongly

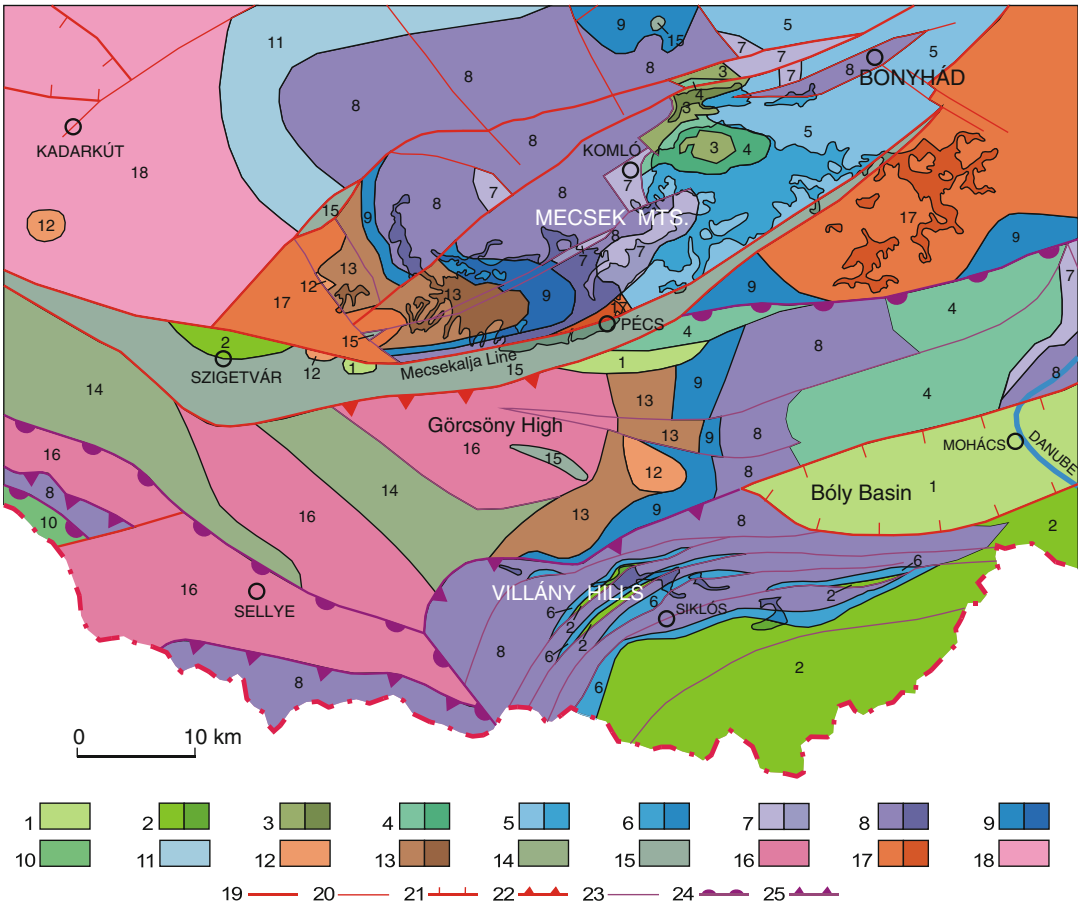


Fig. 2.8 Pre-Cenozoic geological map of South Transdanubia (after Haas et al. 2010). Legend: 1 Senonian continental, shallow and deep marine formations, 2 Lower Cretaceous platform limestone, 3 Lower Cretaceous basic volcanites and their reworked deposits, 4 Middle Jurassic to Lower Cretaceous pelagic limestones, cherty limestones, 5 Lower and Middle Jurassic pelagic fine siliciclastic formations, 6 Jurassic shallow marine and condensed pelagic limestone formations, 7 Upper Triassic to Lower Jurassic coal bearing siliciclastic formation, 8 Middle Triassic shallow marine siliciclastic and carbonate formations, 9 Lower Triassic siliciclastic

formation of fluvial and delta facies, 10 Low-grade metamorphic Mesozoic formations, 11 Mesozoic rocks in general, 12 Permian rhyolite, 13 Permian continental siliciclastic formations, 14 Upper Carboniferous continental siliciclastic formation, 15 Low-grade metamorphic Lower Paleozoic formations, 16 Variscan medium-grade metamorphic complex (gneiss, mica schists, marble), 17 Variscan granitoid rocks, 18 Variscan metamorphic rocks (gneiss, mica schists, amphibolite), 19 regional Cenozoic tectonic line, 20 local Cenozoic tectonic line, 21 Cenozoic fault, 22 Cenozoic overthrust, 23 Mesozoic tectonic line, 24 Mesozoic nappe boundary, 25 Mesozoic overthrust

The Tázlár Carbonate–phyllite consists of an approximately 100 m-thick, 1–2 km²-large, narrow tectonic wedge that might be the northeastern continuation of the Mecsekajka Fracture Zone. It is composed of sheared, dark grey carbonate–phyllite with quartz, sericite, and graphitic phyllite intercalations, which underwent Alpine low-grade metamorphism (Árkai et al.

1985). Based on lithological analogy its age is conditionally assumed to be Late Carboniferous.

The Nagykőrös Sandstone is composed of grey, sometimes organic matter-rich, fossil-free, non-metamorphic, medium and fine-grained molasse-type sandstone wedged into the NE continuation of the Kapos Line(?) at Nagykőrös. Based on analogues its age is tentatively

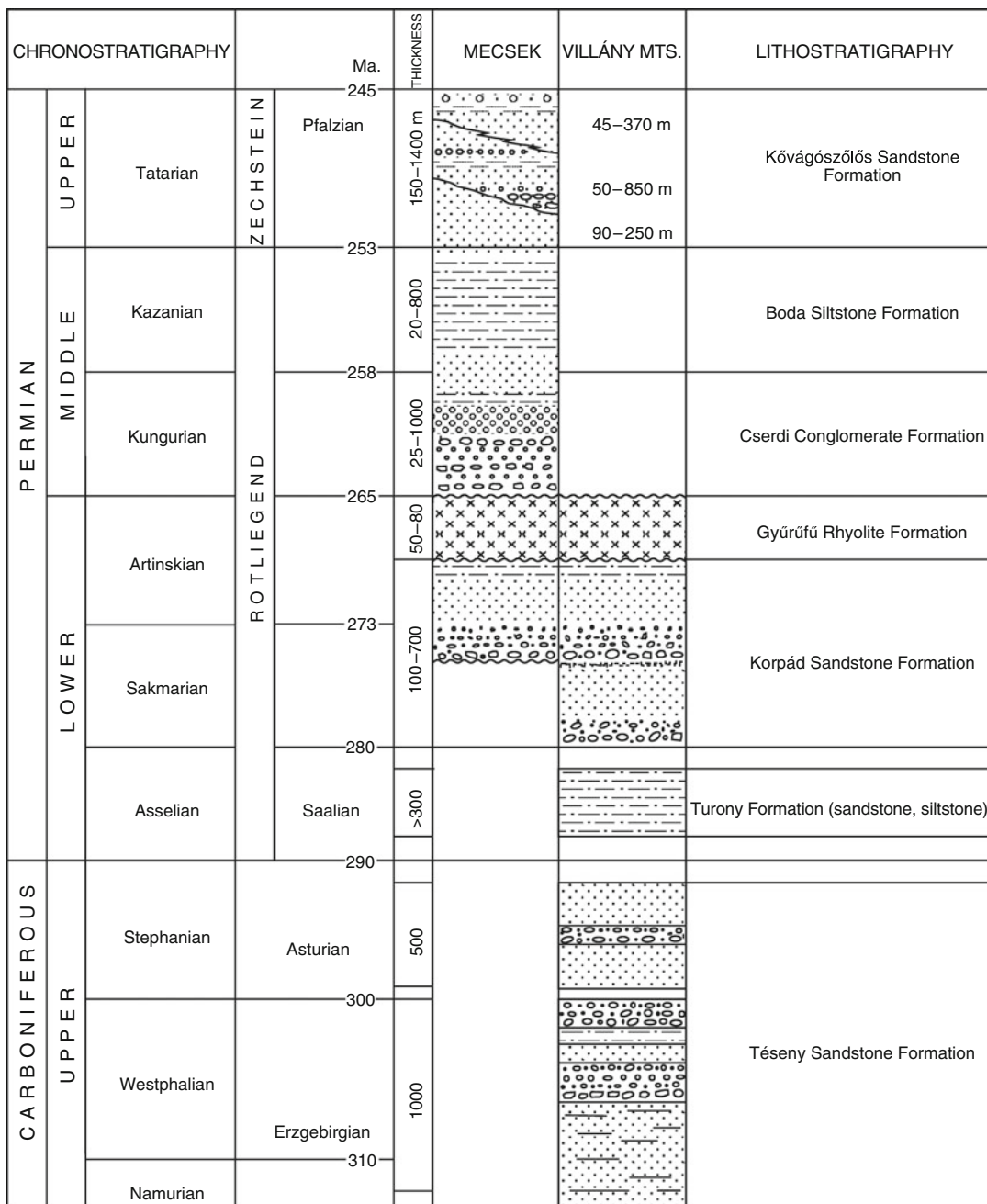


Fig. 2.9 Stratigraphic chart of the Upper Palaeozoic Formations of South Transdanubia (After Barabás-Stuhl in Fülöp 1994)

regarded as Late Carboniferous. The relationship with the tectonically similar Lower Permian sandstone wedge (belonging to the Korpád Sandstone Formation), which was also encountered

near Nagykörs, is unknown (Szederkényi 1984).

The 300 m-thick Turony Formation is made up of an alternation of violet-brown siltstone and

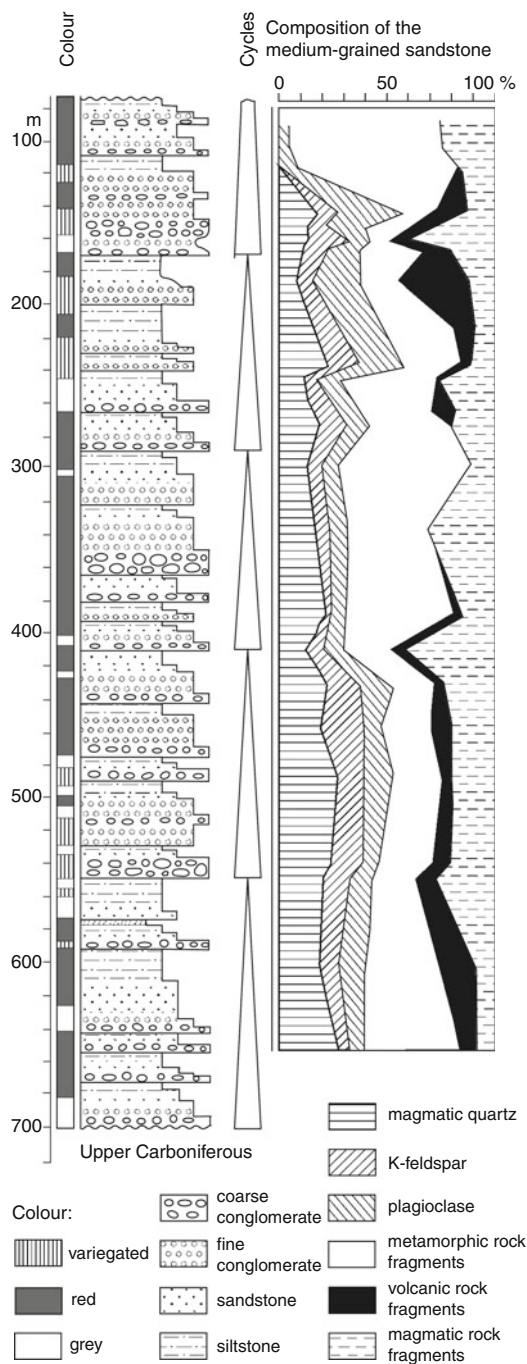


Fig. 2.10 Key-section of the Korpád Sandstone Formation in core Siklósbodony Sb-1 (After Barabás-Stuhl in Fülöp 1994)

fine-grained sandstone. Several thin, rhyolite tuff and dolomitic marl intercalations also occur.

Amphibian footprints of *Anthichnium* (*Saurichnites*) *salamandroides* and *Platytherium psamobates* were also found, suggesting a Stephanian age (Barabás-Stuhl 1975). Kassai (1976) regards this formation as an equivalent of the Boda Siltstone of the Mecsek Subunit; Fülöp (1994) places it at the Carboniferous–Permian boundary.

Following the deposition of the Turony Formation in the Permian, continental sedimentation extended over the accreted terranes of the Tisza Mega-unit, resulting in the formation of uniform sequences above previously different units. The characteristics of the Korpád Sandstone Formation are the same as those of its Mecsek equivalent. Its thickness varies from 100 to 700 m. The extension of the formation is continuous beneath the Villány Hills and in its northern foreland. It is generally covered by the Gyűrűfű Rhyolite and/or the Lower Triassic Jakabhegy Sandstone Formations. It occurs not only in the southeastern Transdanubian area but also in the basement of the Danube–Tisza Interfluve, in the form of a 300–330 m-thick, not entirely well-defined tectonic wedge near the town of Nagykovács, settled upon the granite basement (Szepesházy 1962). Furthermore, as a small denudation remnant above the crystalline schist and below the rhyolite lava sheet, this formation was also encountered at Tótkomlós in the Tiszántúl (Trans-Tisza area) in a thickness of 150 m, with the same lithologic characteristics as in the Villány Hills.

The Gyűrűfű Rhyolite Formation covers either the before-mentioned formation (in the northern foreland of the Villány Hills) or the crystalline basement. Petrographically and chemically it is the same as its Mecsek equivalent, but its thickness and extension are the largest among the Hungarian rhyolite occurrences. Its thickness in the northern foreland of the Villány Hills reaches 450 m, decreasing eastward, but in the neighbourhood of the village of Egerág – as a vent facies – it exceeds 800 m (Fazekas et al. 1987). All rhyolite and related rocks form a single strato-volcano characterised by lava-ignimbrite and tuff alternation, with the lava rocks predominating. Another but much smaller (several m-thick), rhyolite lava sheet was found at Mélykút (Danube–Tisza Interfluve), lying on

crystalline schist and covered by Upper Cretaceous sediments.

The Cserdi Conglomerate Formation was encountered in a few boreholes (Nagykozár, Már-
iakéménd, and Bába) in very variable thickness
(from 25 to 350 m). It overlies the Gyűrűfű Rhyo-
lite Formation and is covered by the Boda Silt-
stone or the Lower Triassic Jakabhegy Sandstone.
In each case the properties are the same as those of
the equivalent formations in the Mecsek Mts.

The Boda Siltstone Formation occurs in a very
thin (max. 20 m-thick) development, and only in
the northern foreland of the Villány Hills beneath
Lower Triassic redbeds (Barabás-Stuhl 1988).
Both the Cserdi and Boda Formations are miss-
ing in the basement of the Great Plain.

2.2.3 Permian Cover of the Kunság Unit

Covering various types of crystalline rock
(granite, crystalline schist, serpentinite) a
2,500–3,200 m-thick Permian sequence occurs
in the entire area of the Mecsek Mts. In the
axial zone of the Western Mecsek brachianticline
they are at the surface, but in other parts of the
mountains they are covered by Lower Triassic
red sandstone (Jakabhegy Sandstone Formation)
and younger Mesozoic rocks (Barabás 1979;
Barabás-Stuhl 1981). The Permian succession
of the Mecsek Mts. is made up of the following
formations:

The Korpád Sandstone consists of
300–320 m-thick, variegated (red, grey and
green) but predominantly red, coarse-grained
sandstone and conglomerate (Fig. 2.8) consisting
of polymict rock fragments and pebbles (Jámbor
1964; Barabás-Stuhl 1988). The frequently inter-
bedding reddish-brown, fine-grained sandstone
and siltstone exhibit a characteristic slurry struc-
ture. The sequence displays a definite cyclicity.
As a basal formation the Korpád Sandstone over-
lies the granite of the Mórág Complex in the
western sector of the Mecsek Mts. in the form of
a basal conglomerate. Its upper boundary is an
erosional surface covered by rhyolite lava. Based

on sporomorphs and megaflores the age of this
formation is Early Permian (Barabás-Stuhl
1981).

The Gyűrűfű Rhyolite consists of a reddish-
brown or reddish-lilac volcanic body of
50–130 m thickness. In several places the rather
monotonous lava masses are punctuated by thin
rhyolite ignimbrite layers. The upper boundary
of the formation is a typical erosional surface
(Fazekas et al. 1987). The whole rock Rb/Sr
age is 277 ± 45 Ma (Balogh and Kovács 1973).

The Cserdi Conglomerate, as a transgressive
sequence, overlies the eroded surface of the
Gyűrűfű Rhyolite. It gradually passes upward
into the overlying Boda Siltstone. The thickness
of the formation varies between 250 and 1,000 m.
Forming a typical fluviatile, cyclic, redbed
sequence, it consists of a regular (sometimes
irregular) alternation of conglomerate, sandstone
of various grain-sizes and siltstone (Jámbor
1964; Barabás and Barabás-Stuhl 1998).

The Boda Siltstone develops from the under-
lying Cserdi Conglomerate with a 100 m-thick
transitional interval, and a similar transitional
zone occurs at the top of the formation. The
formation is made up of a 900 m-thick, monoto-
nous, reddish-brown siltstone with a few fine-
grained sandstone and dolomite-rich interlayers.
At the basin margins the thickness of the forma-
tion decreases dramatically (10–50 m). Lamina-
tion and ripple marks are common in the entire
succession, indicating a lacustrine sedimentary
environment in an arid climate. Phyllopods indi-
cate an Early Permian age (Fülöp 1994) but
according to sporomorph studies the formation
is early Late Permian (Barabás-Stuhl 1981; Bar-
abás and Barabás-Stuhl 1998).

The Kővágószőlős Sandstone is the youngest
lithostratigraphic unit of the Mecsek Permian
sequence. As a result of copper and lead ore
traces in its lowermost member and uranium
ores in its upper part, detailed prospecting was
carried out, yielding detailed data on the charac-
teristics of the unit which were summarised by
Barabás (1979), Barabás-Stuhl (1988), Fülöp
(1994), and Barabás and Barabás-Stuhl (1998).
The thickness of the formation varies between
150 and 1,400 m from west to east. Four

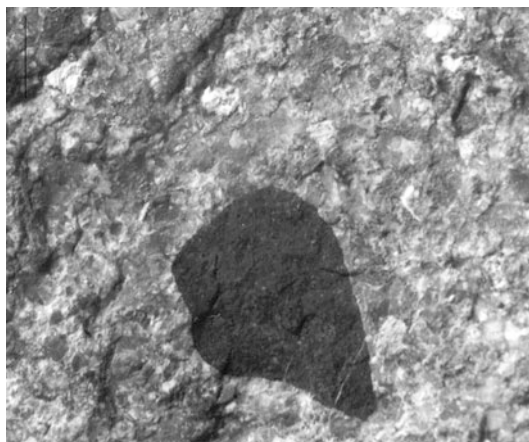


Fig. 2.11 Coarse-grained green arkosic sandstone with brown felsite clasts from the U-ore-bearing horizon of the Kővágószőlős Sandstone Formations, Bakonya, core No. 2061. Scale bar: 0.2 mm (Photo: T. Szederkényi)

members were distinguished (the uppermost one being questionable). The Bakonya Sandstone Member consists of variegated (grey, green and red) sandstone with disseminated chalcopyrite and galenite enrichments; the Kővágótöttös Sandstone Member contains grey sandstone and siltstone with characteristic (Upper Permian) plant remnants and a greenish uranium ore-bearing level in its uppermost segment (Fig. 2.11); the Cserkút Sandstone Member is composed of red sandstone beds, and finally the Tótvár Sandstone Member consists of a violet gravel-rich sandstone. Based on sporomorphs the uppermost part of the Kővágószőlős Formation can be assigned to the Triassic, i.e. the Permian/Triassic boundary can be drawn within the formation (Barabás and Barabás-Stuhl 1998).

The Kővágószőlős Sandstone consists of well-bedded (locally cross-bedded), fluvatile, coarse, medium, and fine-grained sandstone and lacustrine–paludal siltstone. Numerous grey-coloured beds contain coalified macroflora represented by *Ullmannites*, *Voltzites* and *Baiera* species (Heer 1877).

Uranium ores and enrichments were formed at the contact of the Kővágószőlős and Cserkút Sandstone Members (Virágh and Vincze 1967; Vincze and Somogyi 1984). The uranium ions originated from the weathering of adjacent

granite, rhyolite, and crystalline masses and were precipitated and enriched by physico-chemical processes along fracture zones as U-oxides, phosphates, etc., together with chrome and vanadium-bearing silicates in the transitional interfingering interval of the before-mentioned members. After 40 years of exploitation the uranium mining operation went bankrupt and ended in the late 90s.

2.2.4 Permian Cover of the Békés–Codru Unit

Remnants of Upper Palaeozoic lithostratigraphic units are rare above the crystalline rocks of the Békés Unit, with the exception of the Gyűrűfű Rhyolite Formation. The Korpád Sandstone was encountered in a single hydrocarbon exploration well near Tótkomlós. It is made up of 130 m of red conglomerate and coarse- to fine-grained sandstone.

The Gyűrűfű Rhyolite is the only widespread Permian formation in the area of the Békés–Codru Unit. It forms numerous isolated lava sheets which lie on the erosional surface of the crystalline basement or the Korpád Sandstone, and is covered by Lower Triassic redbeds or Miocene as well as Pliocene sediments. At least three effusion centres have been identified, represented by their feeder facies (Fazekas et al. 1987.) The lava sheets are rarely intercalated by ignimbrite or tuff layers, with the exception of Kiskunmajsa where ignimbrites and crystal tuff form a fairly high percentage of the volcanics. The thickness of the occurrences varies between 20 m and more than 200 m; their chemical character shows an alkaline nature. Generally they are made up of rhyolite with a few tuff or ignimbrite intercalations, but subordinately rhyodacite also occurs. The age of the volcanism, according to Rb/Sr measurements on whole rock samples from Battonya, is 240 ± 45 Ma (Balogh and Kovách 1973).

2.3 Alpine Evolution

In the early stages of the Alpine evolutionary cycle the Tisza Mega-unit (microplate) was

located at the southern margin of the European continental plate, being a segment of the northern shelf of the Tethys (Fig. 1.24). This palaeogeographic reconstruction is also constrained by characteristics and distribution of the Triassic to Middle Jurassic facies (Bleahu et al. 1996; Haas and Péro 2004).

During the Permian to Middle Jurassic interval, within the Tisza Mega-unit the Mecsek Zone was subjected to the strongest terrigenous influence. Consequently, this unit must have been located in the external part of the shelf, relatively close to continental source areas. The Villány–Bihor Zone may have belonged to the middle shelf and the Békés–Codru Zone to the outer shelf.

Following the filling up of the Permian continental rift basins a fairly uniform ramp came into being by the Middle Triassic. In the Mecsek Zone segmentation of this ramp (formation of half-grabens) already began in the Late Triassic and became even more pronounced in the Early Jurassic.

In the Middle to Late Jurassic the opening of the Ligurian–Penninic Oceanic Branch led to the breaking off of the Tisza Microplate (Tisia Terrane) from the European Plate and the formation of deep pelagic basins in the Mecsek Zone. In connection with the rifting, basic volcanic activity commenced in the Mecsek Zone; however, it only reached its paroxysm in the Early Cretaceous (Harangi et al. 1996).

The Villány–Bihor Zone was a relatively elevated threshold between the deep basins of the Mecsek and Békés–Codru Zones. In the latter unit deep marine siliceous–carbonate sedimentation predominated in the Late Jurassic, and then shifted gradually toward flysch-type deposition in the Early Cretaceous.

Compressional zones appear to have shifted from the internal belts toward the external ones during the Cretaceous. This is reflected by the appearance of turbiditic and related pelagic basin formations in the Villány–Bihor Zone in the Albian and in the Mecsek Zone in the Cenomanian–Turonian (Császár 2002). However, the Late Turonian–Coniacian period was the major time of thrusting (nappe formation) in the Apuseni Mts. and most probably also in the basement of the

Great Plain. Turbiditic siliciclastic sequences indicate the evolution of flexural basins both in the Mecsek and Villány–Bihor Zones.

Flysch sediments characterise the Palaeogene sequences in the Mecsek Zone (Szolnok Sub-zone). The strongly deformed and imbricated structure of these rocks may be related to the collision of the ALCAPA and Tisza Mega-units at the time they were emplaced into their juxtaposed position in the Late Palaeogene–Early Neogene.

2.3.1 Fluvial Sedimentation in the Early Triassic

The continental rift troughs of the Tisza Mega-unit had been filled up by terrestrial sediments and volcanic rocks by the end of the Permian. In the course of the Early Triassic the marine transgression only reached the innermost zones of the mega-unit. Outliers of the Codru Nappes in the Apuseni Mts. (Romania), containing marine Lower Triassic deposits, indicate this. In a predominant part of the mega-unit continental siliciclastic sedimentation predominated (Alpine Buntsandstein facies). On the basis of spore-morph studies the Permian/Triassic boundary can be drawn within the upper part of the fluvialite Kővágószőlő Sandstone (Barabás and Barabás-Stuhl 2005) which represents the final stage of filling of the Late Palaeozoic rift troughs (Fig. 2.12).

In the Early Triassic a new sedimentary cycle began with the deposition of coarse conglomerate and red sandstone. These sequences extended over a large area, far beyond the Permian rift troughs, even onto the eroded surface of the Variscan metamorphic complex.

At the base of the red siliciclastic formation (Jakabhegy Sandstone) coarse conglomerate occurs in 1–10 m thickness (Fig. 2.13). The size of the components may attain 20 cm and quartz, rhyolite, ignimbrite, and granite are the most common rock types encountered. Above the conglomerate unit the 150–400 m-thick formation is made up of cross-bedded sandstone (Fig. 2.14). The sandstone succession is punctuated by

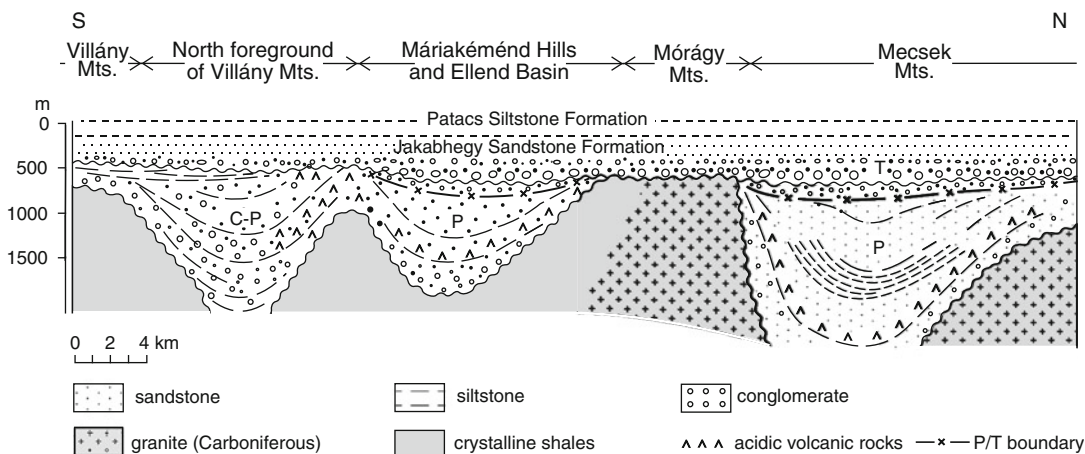


Fig. 2.12 Late Palaeozoic continental basin in the area of South Transdanubia (After Haas et al. 1986)



Fig. 2.13 Pebbly sandstone and conglomerate beds at the basal part of the Lower Triassic Jakabhegy Sandstone, Kővágószőlős (Photo: Cs. Péró)

pebbly horizons and siltstone interlayers which show a definite cyclicity. The material of the sand is predominantly quartz; however, the amount of feldspar is also significant (20–30%).

Based on sedimentological features of the rock sequences a fluvial depositional environment can be assumed for the lower part of the formation, and for its upper part a tide-dominated delta facies has been proposed (Csicsák and Szakmány 1998). Statistical analysis of cross-bedding directions in the fluvial facies suggests transportation from N to S (Nagy 1968). No marine fossils were found in the formation; however, sporomorphs found in the upper part of the unit point to the

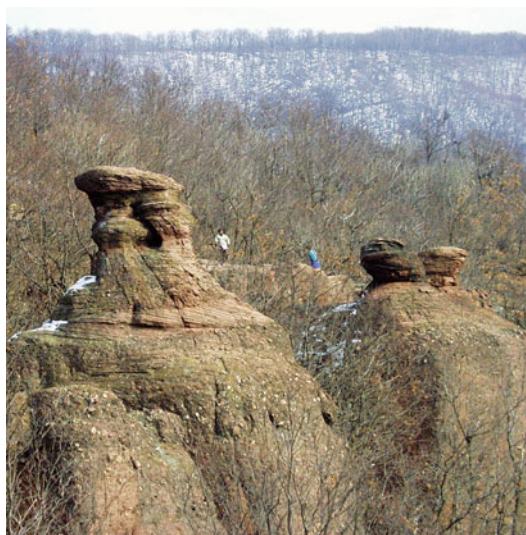


Fig. 2.14 Cross-bedded sandstone in the Lower Triassic Jakabhegy Formation, Kővágószőlős, Mecsek Mts. (Photo: Cs. Péró)

upper part of the Lower Triassic (Barabás-Stuhl 1981).

2.3.2 Transgression in the Anisian – Siliciclastic Ramp Sedimentation

In the tectonically calm interval of the Early Anisian a eustatic sea level rise may have led to

age \ area	MECSEK ZONE		VILLÁNY-BIHAR ZONE			BÉKÉS BASIN
	Mecsek Mts.		Villány Mts.	Bácska	Körös	
RHAETIAN	Mecsek Coal Fm.					
NORIAN	Karolinavölgy Sandstone Fm.		?		?	
CARNIAN			Mészhegy Sandstone Fm.		Mészhegy Sandstone Fm.	
LADINIAN	Kantavár Fm.					Csanádapáca Dolomite Fm.
			Csukma Fm.			
ANISIAN			Zuhánya Limestone Fm.			Szeged Dolomite Fm.
			Lapis Limestone Fm.		Rókahegy Dolomite Fm.	
			Hetvehely Dolomite Fm.			
SCYTHIAN	Patacs Siltstone Fm.					
	Kővágószőlős Sandstone Fm.		Jakabhegy Sandstone Fm.			

Fig. 2.15 Triassic formations of the Tisza Mega-unit (After Császár et al. 1997)

general transgression in the Tisza Mega-unit area. The inundation of the topographically levelled area resulted in the formation of a widely extended ramp. On the ramp a very wide tidal flat and a relatively narrow subtidal zone came into being. Coevally the influx of fine-grained terrigenous material continued. In this environment red and green fine-grained sandstone and siltstone and green siltstone were deposited (Patacs Siltstone).

In the outcrops of the Mecsek Mts. the red sandstone layers of the Patacs Formation (Fig. 2.15) show parallel and cross-lamination, and ripple marks are common. In the green siltstone-claystone layers a rich phyllopod fauna was found; brachiopods (*Lingula tenuissima*) and bivalves (*Costatoria costata*) also occur. The sporomorph assemblage indicates an Early Anisian age (Barabás-Stuhl 1981). The thickness of

the formation is about 200 m in the Mecsek Mts. According to core data, however, it is not more than 15 m in the area of the Villány Hills and between the two areas. The formation was also encountered in exploration wells in the basement of the Great Plain.

In the Mecsek Mts. the siltstone layers of the Patacs Formation of shallow-marine to tidal flat facies are gradually substituted by dolomitic marl and dolomite layers with evaporitic nodules and laminae of sabkha facies (Freyt et Cross 1984). Thereafter grey marl layers with thick evaporitic dolomite interbeds become predominant. The thickness of the formation, consisting of an alternation of fine terrigenous, dolomitic and evaporitic layers (Hetvehely Formation) may attain 200 m in the Mecsek Mts., and only 70 m in the Villány Hills.

This evaporitic facies shows a definite similarity to the “Röt” facies of the Germanic facies realm. Periodically a restricted, shallow inner ramp/lagoon may have been the site of deposition. During sea level highstands the gradually fining terrigenous material may have been deposited in the subtidal zone of the open ramp. However, in the lowstand intervals the inner parts of the ramp may have become restricted and superhaline conditions have come into being. This was followed by sabkha-type evaporite formation and dolomitization during the subaerial exposure intervals.

2.3.3 Shallow Carbonate Ramp Evolution in the Middle Triassic

At the end of the Early Anisian terrigenous influx decreased, resulting in a predominance of carbonate deposition in the shallow ramp and platform areas which extended over most of the area of the Tisza Mega-unit. Carbonate sequences were formed on a wide, back-platform ramp, showing a close genetic relationship with the “Muschelkalk” facies of the Germanic Basin and European margin of the Tethys (Nagy 1968; Török 1993, 1998).

Within the Tisza Mega-unit the Mecsek Unit represents the deeper zone of the ramp and the Villány–Bihor Unit the shallower one, where the ramp passed over into a platform. In the Codru nappes even the offshore shelf margin facies appear. Although sequences of the Mecsek and Villány–Bihor zones differ from each other the facies transitions between them are plausible and differences in their lithology are caused partly by differences in their burial diagenetic history, resulting in variable grades of dolomitization.

The evaporite formation is overlain by dolomite (Rókahegy Dolomite). In the Villány Hills the thickness of the dolomite unit may attain 100 m. It is laminitic and often contains peloidal or ooidic interlayers. In the Mecsek Mts. the thickness of the formation is less than 20 m; it may have formed on a shallow ramp. Dolomitization

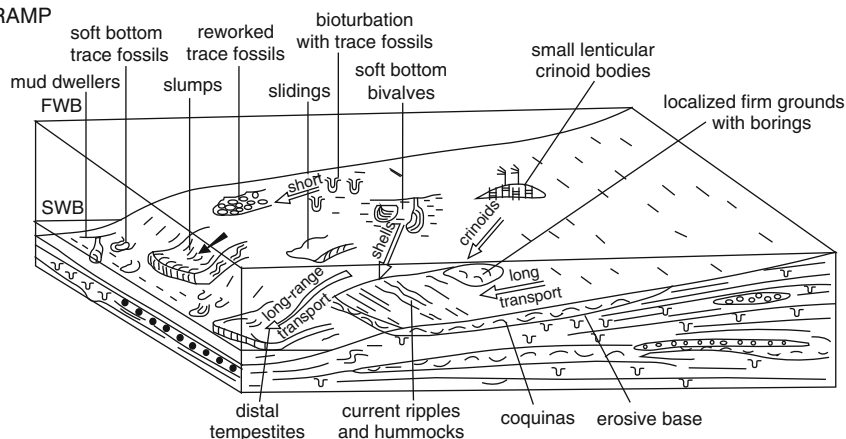
may have taken place under subaerial conditions during sea level lowstands.

The next stages of ramp evolution are represented by sequences consisting of an alternation of thin-bedded and marly, nodular limestone and marl with bioturbated beds and crinoid and mollusc coquina interlayers. The colour of the rocks becomes lighter and the thickness of the beds increases upward; bioturbation is common and oncoidal-oolitic intercalations also occur. The features of this 300 m-thick formation (Lapis Limestone) are similar to those of the “Wellenkalk” facies of the Germanic Triassic (Török 1993; Figs. 2.16, 2.17). The site of deposition of the lower part of the formation may have been the deeper ramp, near to and just beneath the storm wave base. The bivalve-gastropod coquina layers and graded crinoidal beds were deposited by storm-generated currents. The upper part of the unit was formed in a shallower, occasionally strongly agitated environment above the wave base.

Dark grey, nodular, intraclastic clayey limestone with calcareous marl intercalations represents the upper part of the Anisian (Zuhány Limestone). Brachiopod coquinas are common and typical (Fig. 2.18). *Coenothyris vulgaris* (Schlotheim), *Tetractinella trigonella* (Schlotheim), and *Punctospirella fragilis* (Schlotheim) are the most characteristic species of the brachiopod assemblage (Török 1993). *Encrinurus radiates* (Schaueroth) and *Holocrinurus* sp. are typical and age-diagnostic representatives of the crinoids (Hagdorn et al. 1997). These layers may have been deposited on a deeper open ramp. This is also indicated by the appearance of conodonts and a few ammonites (Kovács and Papsová 1986). The characteristic nodular structure was probably caused by intense bioturbation and, in connection with it, by early diagenetic deformations of the semi-consolidated sediments. An early diagenetic origin of the nodules is also supported by the occurrence of redeposited nodules in the storm coquinas.

The upper member of the Zuhány Limestone consists of dark grey, thin-bedded limestone with lilac, reddish, or yellowish patches. Bedding

MIDDLE ANISIAN MID-RAMP



MIDDLE-LATE ANISIAN OUTER RAMP

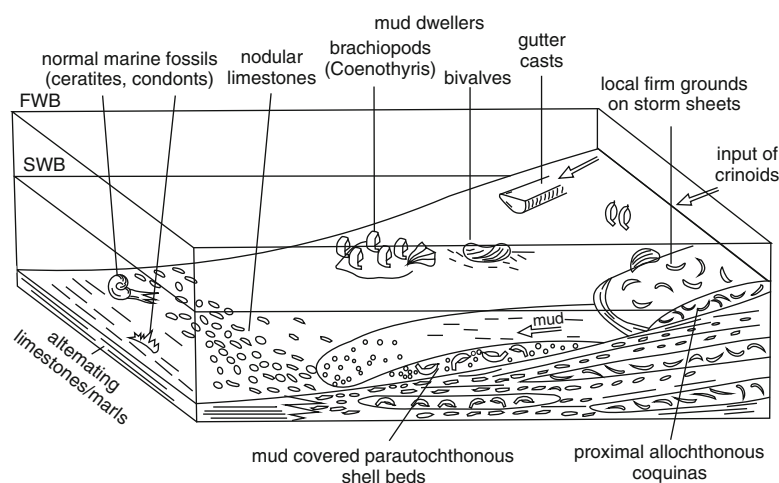


Fig. 2.16 Characteristic features of the Anisian carbonate formations in the Mecsek Mts. and the interpreted depositional environments (After Török 1993)

planes are wavy and clayey; bituminous interlayers are common. These features indicate restriction of the basin which led to oxygen depletion in the near-bottom zone. Non-stratiform, patchy dolomitization, common in the Zuhány Formation, is the result of deep burial diagenesis.

In the Békés–Codru Zone, in the basement of the southeastern part of the Great Plain, above the Lower Anisian shallow marine siliciclastic and evaporite units, dark grey massive dolomite with a poor foraminiferal, mollusc and ostracode fauna was encountered in many boreholes. This

formation (Szeged Dolomite), which probably includes a large part of the Anisian, may have been deposited in an oxygen-depleted, relatively deep marine basin (Bérczi-Makk 1986).

A general shallowing of the “Muschelkalk” basin began in the latest Anisian to Early Ladinian (Török 1998). In the Mecsek Mts. this shallowing-upward trend is reflected by the appearance of thick-bedded limestone with ooidal and crinoidal interlayers. In certain parts of the Mecsek Mts. and Villány Hills, however, this stratigraphic interval is represented by brownish-grey, yellowish-grey dolomite with



Fig. 2.17 Thin-bedded, laminated limestone with soft sediment deformation structures in the Anisian Lapis Limestone. Lapis road cut, Mecsek Mts. (Photo: Cs. Péro)



Fig. 2.18 Brachiopod coquina in the Anisian Zuhánya Limestone, Pécs, Mecsek Mts. (Photo: Gy. Konrád)

dolomitic marl intercalations in the upper part of the formation.

In the Békés–Codru Zone light grey dolomite was encountered in cores (Csanádapáca Dolomite). Characteristic calcareous algae (*Gyroporella ampleforata* Gümbel and *Diploporella annulata* Schafhäütl) indicate a Ladinian age and a protected inner platform depositional environment (Bérczi-Makk 1986).

2.3.4 Differentiation of the Facies Zones of the Tisza Mega-Unit

During the Middle Triassic more or less uniform, shallow marine carbonate sedimentation charac-

terised the area of the Tisza Mega-unit. A definite differentiation of the facies zones was initiated in the Late Triassic, when extensional half-grabens began to be formed in the Mecsek Zone (Fig. 2.19). Subsequently subsidence accelerated in the grabens in connection with the opening of the Penninic Ocean Branch in the Jurassic (Haas and Péro 2004) and intensive rift volcanism commenced in the Early Cretaceous.

The facies of the Villány–Bihor Unit also show significantly distinctive features from the Late Triassic on; in the Jurassic–Early Cretaceous interval the threshold nature of this zone became quite definite, manifesting itself in shallow marine facies punctuated by gaps in the Jurassic and establishment of carbonate platforms in the Cretaceous.

In the Békés–Codru Zone carbonate platform facies are known from the Upper Triassic (in the Codru Mts.). In the Jurassic increasingly deeper marine facies appear which may be related to the evolution of the Neotethys. The separate evolution of the zones also continued during the Cretaceous–Palaeogene convergent regime.

Due to their definitely divergent evolution the further history of the differentiated zones (units) of the Tisza Mega-unit will be discussed separately.

2.3.5 Mecsek Facies Unit

2.3.5.1 Intensification of Continental Input in the Late Triassic

Shallowing at the end of the Middle Triassic, accompanied by intensification of terrigenous input, led to fundamental changes in the sedimentary pattern all over the Tisza Mega-unit. In the Mecsek Unit, above the carbonate ramp facies, there is a black limestone horizon characterised by pebble-sized oncoids and bivalve (*Trigonodus*) coquina. In the central part of the Mecsek Mts. it is overlain by black, thin-bedded argillaceous limestone with black marl interlayers, coal stripes, and sandstone layers appears (Kantavár Formation). The faunal assemblage is poor in species, but the number of specimens of the monospecific ostracode fauna (*Darvinula liassica*) is extremely large (Monostori 1996).

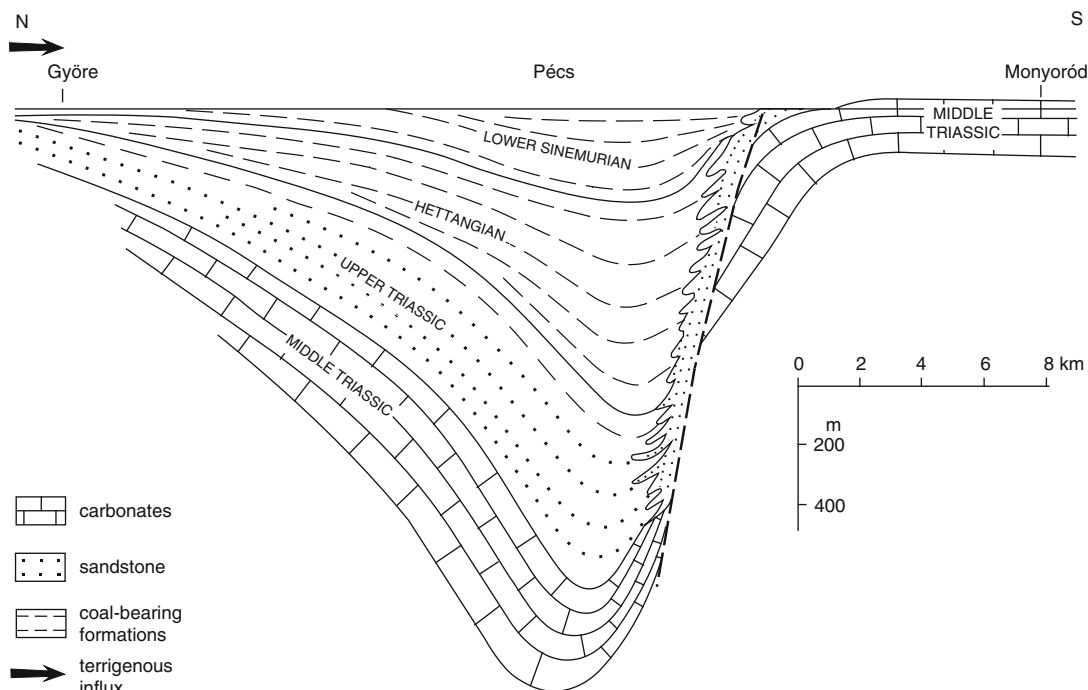


Fig. 2.19 Conceptual cross-section of the Mecsek half-graben showing the depositional pattern until the Early Jurassic (After Nagy 1969)

Small gastropods, charophytes, and carbonised plant remnants (e.g. *Equisetites* and *Anatopteris*) are also common. Changes of features of this sequence reflect a transition from brackish water to freshwater environment. Based on sporomorphs the age of the formation is latest Ladinian to earliest Carnian (Bóna 1995).

The Kantavár Formation is overlain by grey arkosic sandstone and siltstone and grey or greenish/reddish shale (Karolinavölgy Sandstone). The thickness of the formation in the Mecsek Mts. is about 500 m, but significantly thinner in the Danube–Tisza Interfluvial segment of the Mecsek Zone.

The lower part of the formation contains thin coal interlayers. In addition to ostracodes and phyllopod marine bivalves and gastropods also occur in a restricted number. The rich sporomorph assemblage suggests a Carnian age (Bóna 1995). Depending on the sea level, sedimentation took place in lagoonal, lacustrine, or deltaic depositional environments. A significant increase in terrigenous input may have corresponded to a climatic change (pluvial event)

which resulted in similar trends both in the Neotethys basins and the Germanic Basin during the Carnian. The middle part of the formation is made up mainly of sandstone and siltstone of lacustrine or lacustrine-deltaic facies; lagoonal facies are subordinate. The basal part of the upper member of the formation is characterised by coarse-grained, cross-bedded fluvial sandstone (Nagy 1968). Greenish-grey claystone of lacustrine facies becomes predominant upsection. Further upward there is a gradual transition into the overlying coal-bearing succession. Based on sporomorphs the upper member can be assigned to the Rhaetian (Bóna 1995).

2.3.5.2 Coastal Swamp and Shallow Marine Siliciclastic Ramp in the Early Liassic

Thin coal interlayers already appear in the fluvial succession in the latest Rhaetian. At the beginning of the Liassic fluvial–lacustrine–palustrine sedimentation continued but paralic coal-swamp deposits became predominant in the sedimentary record (Mecsek Coal). The thickness of the coal-

bearing series is usually 150–300 m; in the southern part of the Mecsek Mts., however, it may attain 1,200 m. This asymmetric thickening, already encountered in the Upper Triassic Karolinvölgy Sandstone, may be explained by the formation of an extensional half-graben (Nagy 1969). In the Alpine-Carpathian region Lower Liassic, coal-bearing, siliciclastic sequences, showing features similar to those in the Mecsek Zone, are classified as “Gresten Facies”, which is considered to be a characteristic facies of the European shelf of the Tethys.

The basal uppermost Triassic part of the Mecsek Coal was formed predominantly in lacustrine as well as lacustrine/deltaic facies (Fig. 2.20). No mass extinction event was encountered in the micro- and macroflora in the Triassic/Jurassic boundary interval (Ruckwied et al. 2008). The formation is made up of a cyclic alternation of arkosic sandstone, siltstone, claystone and coal layers; the periodical environmental changes reflect also in the diverse palynomorph assemblages (Ruckwied et al. 2008). In some horizons of the succession well-preserved prints of plants were found (*Equisites*, *Thaumatoporphites*, *Nilssonina*, etc.). The numerous and diverse ferns, large amount of horstails, and dominance of *Komlopteris* indicates wet and warm climatic conditions (Barbacka 1994; Barbacka and Bodor 2008). In the marine sublittoral layers euryhaline molluscs occur, locally in large amounts (*Cardinia*, *Gervilleia*, *Gryphaea*, *Anomia*, etc.). The Hettangian middle member of the formation is mainly fluvial with channel, flood plain, and swamp facies; however, passing upward, coquinas of brackish-water molluscs appear in increasing frequency. Thin (0.5–1.5 m) rhyolitic tuffite interlayers occur in this member (Némedi-Varga 1983). The Lower Sinemurian upper member of the Mecsek Coal may have been deposited in a tidal flat marsh environment. However in some layers remnants of crinoids also appear, indicating a temporary establishment of normal salinity conditions.

Steeply dipping, strongly deformed, 0.4–6 m-thick coal seams of the Mecsek Coal was exploited during more than 200 years until 2002 (Fig. 2.21). This is the only coking coal deposit in

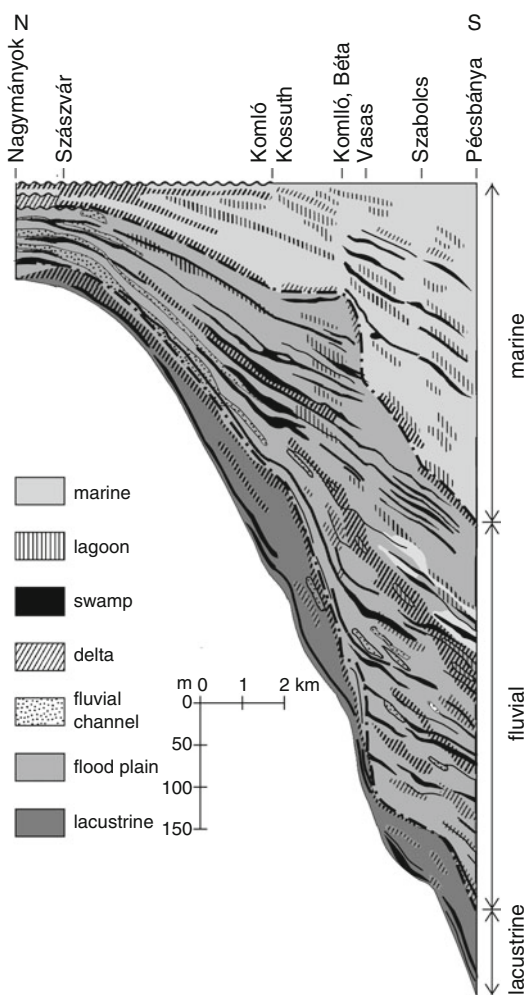


Fig. 2.20 Changes in the facies characteristics and thickness of the Mecsek Coal Formation in the Mecsek Mts. (After Nagy 1969)

Hungary. The mining region was located in the environs of Komló and Pécs extending over an area of 350–400 km². Number of coal seams of commercial value was 15 in Pécs area with 28 m total thickness and 9 in Komló area with 24 m total thickness.

The coal formation is overlain by fine-grained sandstone and dark grey marl with crinoidal limestone interlayers, Late Sinemurian in age (Vasas Marl). The thickness of the formation in the Mecsek Mts. is 250–650 m. In the lower part of the succession *Liogryphaea* beds occur which may have formed in a very shallow marine



Fig. 2.21 Alternation of sandstone and coal beds in the Mecsek Coal Formation. Open-pit mine at Pécs–Vasár (Photo J. Haas)

environment. Other bivalves and crinoid ossicles also occur locally in rock-forming quantity. Poorly-preserved plant remnants are common. The upper part of the formation is rich in molluscs, ammonites, belemnites, echinoderms, brachiopods, and foraminifera, indicating a deeper ramp environment of normal salinity (Császár et al. 2007). The entire sequence suggests gradual deepening and transgression. The sea level rise may also have caused a decrease in terrigenous input, and no deltas appear to have been located close to the Mecsek Basin.

2.3.5.3 Pelagic Marl Facies in the Middle Liassic to Early Dogger Interval

During the later part of the Sinemurian, most probably as a combined result of eustatic sea level rise and accelerated subsidence, water depth continued to increase. Coevally the continental source area, which still provided a large amount of terrigenous material, moved even

farther away from the site of deposition. In accordance with this palaeogeographic setting an open marine deep basin had been the site of deposition until the middle part of the Jurassic. In this basin fine-grained terrigenous material and pelagic biogenic ooze were deposited together; however, their ratio continuously changed. This heavily bioturbated marl sequence, the so-called “Fleckenmergel” or “Allgäu” Facies is also characteristic of the European Tethys margin. In the rapidly subsiding southern zone of the Mecsek half-graben its thickness may attain 2,000 m, whereas in the northern part of this structural unit, as well as in the subsurface parts of the Mecsek Zone (i.e. in the basement of the Transdanubian area and the Great Plain), it is generally only 150–300 m.

In the upper part of the Sinemurian the carbonate content increases and grey, slightly bioturbated marl, silty marl and calcareous marl with crinoidal limestone intercalations become characteristic (Hosszúhetény Marl; Fig. 2.22). Ammonites, belemnites, brachiopods, bivalves (Vadász 1935) and sponge spicules and foraminifera are common (Raucsik and Merényi 2000). This 50–350 m thick formation of Late Sinemurian to Early Pliensbachian age was deposited under open marine conditions in the deeper zone of the open shelf.

The higher part of the Pliensbachian is characterised by rhythmic alternation of hemipelagic spotted marl, calcareous marl, redeposited crinoidal limestone and mixed carbonate–siliciclastic turbidite (Mecseknádasd Sandstone) (Raucsik and Merényi 2000; Raucsik and Varga 2008). In the rapidly subsiding southern zone of the Mecsek Mts. area the thickness of the formation may attain 1,000 m, and only 70 m in its northern zone. The depositional environment may have been a shallow bathyal basin. The predominance of the sand-sized terrigenous material may reflect a lowering sea level and/or a climatic change.

Upsection, silty marl becomes predominant again and the share of sandstone decreases (Obánya Shale). This approximately 150 m-thick formation represents the Lower and Middle Toarcian. In the Lower Toarcian a 10 m-thick,

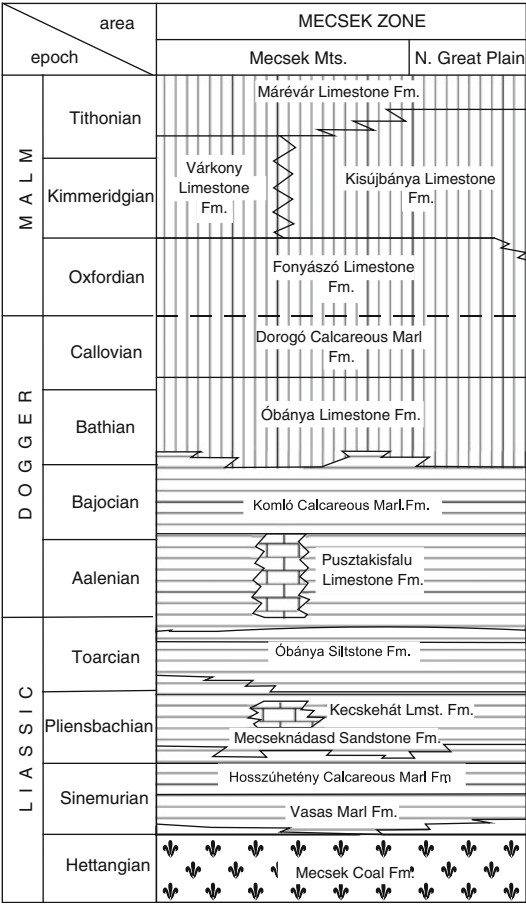


Fig. 2.22 Jurassic formations of the Mecsek Zone (After Császár ed. 1997)

very peculiar black shale intercalation with thin sandstone and crinoidal limestone interlayers as well as ammonites, belemnites, thin-shelled pelagic bivalves and fish remnants can be found (Dulai et al. 1992). The black shale facies indicates that anoxic conditions prevailed near the sea bottom. Based on ammonites this anoxic layer can be exactly correlated with the Early Toarcian global anoxic event (Jenkyns and Clayton 1986; Jenkyns 1988).

Above the black shale intercalation the typical “Fleckenmergel” facies resumes. The Upper Toarcian–Bajocian is characterised by rhythmic alternation of spotted marl, calcareous marl and clayey limestone, 200–500 m in thickness (Komló Calcareous Marl) (Raucsik and Varga 2008). It contains predominantly pelagic fossil

elements: *Bositra* shell fragments, radiolarians, sponge spicules, echinoderm fragments, belemnite rostra, and ammonoids (Vadász 1935). The pelagic fossil assemblage suggests a relatively deep marine depositional environment supplied with large amount of fine-grained terrigenous material from a distant source area.

In the southern and northern margins of the “Mecsek Basin” grey and red crinoidal–brachiopodal limestone, coeval with the Komló Calcareous Marl has been encountered. This facies may characterise uplifted, shallower marginal blocks of the “Mecsek Basin”.

2.3.5.4 Siliceous and Carbonate Deep-Sea Facies in the Late Dogger to Malm Interval

At the end of the Bajocian the sedimentation character fundamentally changed: the amount of terrigenous material and consequently the sedimentation rate significantly decreased; continuous and probably accelerated subsidence led to increased water depth. These changes in the sedimentary pattern can be related to the separation of the Tisza Mega-unit from the European plate. Due to oceanic opening between the European Plate and the Tisza Block the Mecsek Zone was cut off from its previous continental source area. Consequently, condensed pelagic carbonates and siliceous sediments, similar to coeval sequences in the Transdanubian Range or the Alpine region, were formed. Changes in the fossil assemblage (e.g. ammonites, brachiopods), that is the appearance and then prevalence of Mediterranean elements, may also be attributed to this process (Géczy 1973; Vörös 1993).

Reflecting the changes in the depositional regime the spotty marl (“Fleckenmergel”) facies is overlain by greenish-yellowish-reddish marl, then by red calcareous marl and finally by nodular, argillaceous limestone rich in poorly-preserved ammonoids and pelagic microfossils (*Bositra* shell fragments, *Protoglobigerina*, radiolarians). This 10–20 m-thick formation (Óbánya Limestone) of Bathonian age (Galács 1995) was deposited in a deep, pelagic, starved basin, above the aragonite compensation depth.

The next unit consists of brownish- and greenish-grey, thin-bedded, siliceous calcareous marl (Dorogó Marl) with a few poorly preserved ammonites, *Bositra* fragments, and radiolarians. It also contains altered pyroclastics but only in a small quantity. The thickness of this formation of Callovian age does not exceed 10–20 m. The calcareous marl passes upward into siliceous limestone (Fonyászó Limestone). In the basal part of the formation brownish–greenish, highly silicified radiolarite occurs. Above it, the 30–120 m-thick formation is made up of thin-bedded, yellowish-grey, reddish, and greenish cherty limestone. The rocks are poor in megafossils; the microfossils suggest an Oxfordian age.

The Kimmeridgian to Lower Tithonian interval is represented by red, nodular, locally cherty limestone with ammonoids and aptychi (Kisújbánya Limestone) with features of the Mediterranean “Ammonitico rosso” facies. The limestone consists predominantly of *Saccocoma* ossicles. The thickness of the formation is 10 to 50 m.

The red nodular limestone passes upward into greyish- or yellowish-white, thin-bedded limestone and argillaceous limestone, locally with intraclasts and chert nodules (Márévár Limestone), similar to the Mediterranean Maiolica Facies. Its thickness may attain 100 m. The layers are poor in megafossils but they contain a rich calpionellid microfauna which emplaces the formation into the Upper Tithonian–Berriasian. The site of deposition may have been a deep pelagic basin. Intrabreccia intercalations in the pelagic sequences and the chronostratigraphically mixed microfossil assemblage (Nagy 1986) indicate a significant gravity mass flow activity resulting in the redeposition of the unconsolidated and semiconsolidated sediments.

In the upper part of the formation, in addition to the redeposited carbonate grains, fine pyroclastics and volcanic bombs appear in the layers, indicating the intensification of volcanic activity in the Berriasian.

2.3.5.5 Basaltic Magmatism in the Early Cretaceous

The very intensive and areally extensive Early Cretaceous alkaline basalt magmatism is one of

the most characteristic features of the Mecsek Unit. It was connected to rifting which had already been initiated in the Late Triassic and led to crustal attenuation and formation of extensional basins. Products of the magmatic activity crop out in the Mecsek Mts. and can also be traced in other parts of the Mecsek Zone, in the basement of the Tertiary basins, in Transdanubia and the Great Plain (Haas and Péró 2004).

Traces of volcanic activity can already be observed in the Jurassic formations; however, the culmination of the volcanism occurred in the Early Cretaceous, mainly in the Valanginian, although it extended into the Hauterivian (Mecsekjános Basalt). In the western Mecsek Mts. basaltic and trachytic small intrusions and dykes are typical, generally with alkaline metasomatism, whereas in the eastern part of the range rocks of basalt–tephrite–phonolite series and kalithrachte occur in equal amounts (Kubovics et al. 1990). Pillow lava, lava breccias, and hyaloclastite indicate submarine volcanism. Dikes, sills, and subvolcanic bodies are common. In the basement of the Great Plain feldspar-rich basalt showing spilitisation, Mg-metasomatic alteration, and carbonatisation was also encountered. The geochemical and mineralogical–petrographic features suggest continental rift-type volcanism (Kubovics and Billik 1984; Harangi et al. 1996).

In the Mecsek Mts. basalt is overlain by conglomerate and sandstone, and marl. Locally the clastic layers are again covered by basaltic lava rocks. In the clastic beds (Magyaregregy Conglomerate) a significant part of the clasts were derived from the volcanic build-up. It contains a rich shallow-marine fauna (rudists, gastropods, and corals) was found together with cephalopods and other pelagic fossils. It is obvious that the top of the volcanic centres reached sea level, allowing colonisation by shallow marine biota around the volcanoes; atoll-like carbonate build-ups were formed (“Mecsek-type atolls” – Császár 2002). As a consequence of rapid erosion of the volcanoes together with the encroaching atolls lithoclasts and bioclasts accumulated in the deeper basins between the volcanic highs while volcanic activity may have been continuing (Fig. 2.23). Rhythmic alternation of limestone or

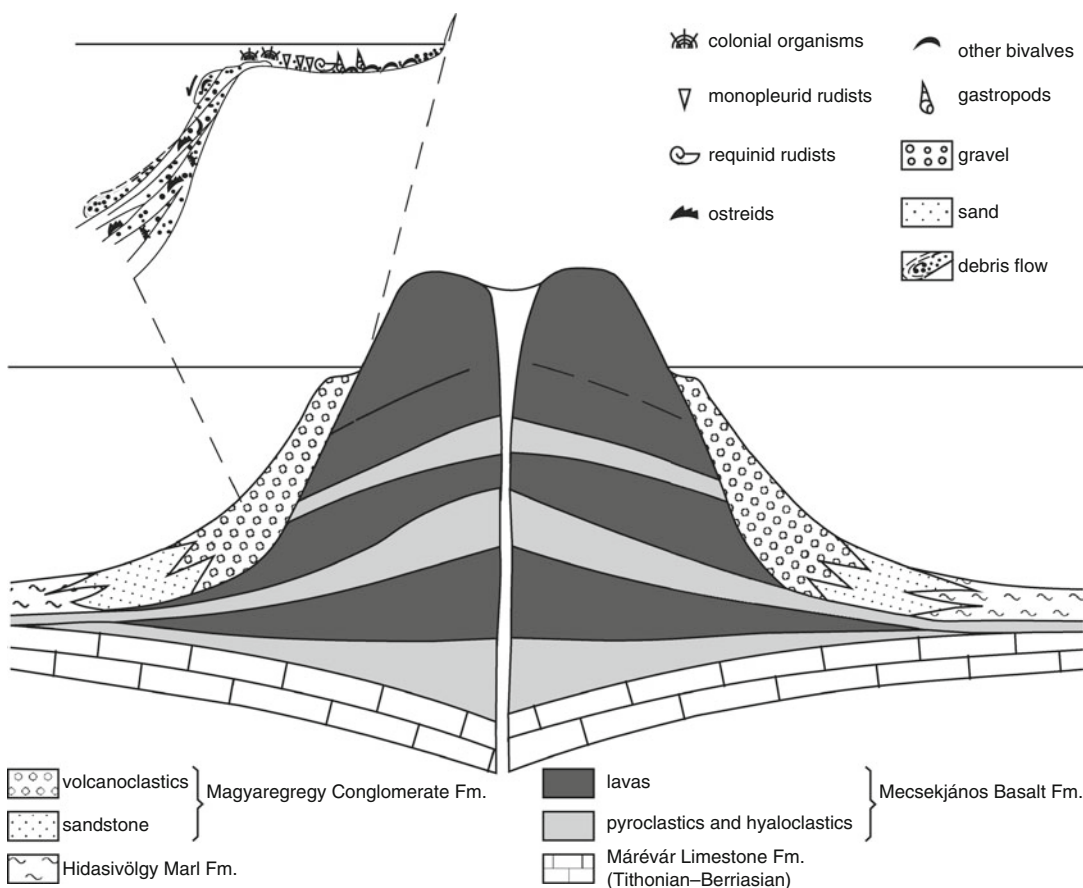


Fig. 2.23 Conceptual cross-section of the Lower Cretaceous volcanic and volcano-sedimentary complex in the Mecsek Mts. (After Császár and Turnšek 1997)

calcareous marl and siltstone or silty marl (Hidasivölgy Marl) characterises the successions of the central part of the basins located relatively far from the volcanoes (Császár 2002).

In the southern part of the Mecsek Mts. the thickness of the volcanic complex is significantly reduced, and volcanites are overlain by a 400–500 m-thick Valanginian–Hauterivian sequence consisting predominantly of crinoidal limestone.

2.3.5.6 Tectogenic Episodes and Flexural Basins in the Late Cretaceous

In the Mecsek Mts. Cretaceous formations younger than Barremian do not occur, except for a single locality where Turonian pelagic marl was

found. Similarly there is no firm evidence for deposition between the Barremian and the Turonian in any other part of the Mecsek Zone. Based on observations in the Villány–Bihar Zone it seems probable that in the Tisza Mega-unit the first phase of Alpine orogeny occurred in the Late Albian–Cenomanian, and may have resulted in the erosion of a significant part of the previously deposited Cretaceous formations.

Coevally with and/or subsequent to the assumed orogenic movements deep basins came into being in the foreland of the thrust belts. There is only little evidence for the existence of this kind of basin in the Turonian. In the Mecsek Mts. red marl and calcareous marl rich in planktonic foraminifera were encountered (Vékény

Marl; Balla and Bodrogi 1993; Császár 2002). Similar rocks and fossil assemblages were found in a well in the Great Plain area; however, in some boreholes south of the previously-mentioned well grey, foraminifera-bearing, pelagic marl representing the same stratigraphic interval was encountered.

In the Romanian Apuseni Mts., in the eastern part of the Tisza Mega-unit, the most intensive nappe formation occurred in the Coniacian (Ianovici et al. 1976); this is probably true for the entire mega-unit (Haas and Péró 2004). It follows that pelagic sedimentation was probably interrupted, uplifting and subaerial denudation may have taken place, and subsequently new basins came into existence. Senonian (Campanian to Maastrichtian) pelagic sequences encountered in many wells in the Great Plain area may have been deposited in these basins.

In the Trans-Tisza area, in the northeastern part of the Mecsek Unit, Campanian-Maastrichtian sequences made up of an alternation of sandstone and siltstone layers and subordinately shale and conglomerate interlayers were also found (Debrecen Sandstone; Szepesházy 1973; Szentgyörgyi 1989). The thickness of the formation is not known exactly but certainly exceeds 500 m. The composition of the sand-sized grains indicates a predominantly metamorphic source area. The lithologic features of the formation indicate turbiditic (flysch-type) sedimentation (Szolnok Flysch Complex, hereafter SFC); however, the fact that the rocks are generally strongly compressed, and also that relatively few core data are available, make it difficult to recognise typical flysch characteristics.

Southwest of the area of extent of the Debrecen Formation the Campanian-Maastrichtian interval is represented by marl, calcareous marl, silty marl, and argillaceous limestone (Izsák Marl). In the lower part of the formation rusty brown, and reddish colours predominate and grey becomes dominant upsection. In the westernmost occurrence of the formation in the Danube-Tisza Interfluvium area red colour prevails in the entire sequence and the carbonate content is the highest (Szentgyörgyi 1989). This red marl shows features very similar to those of

the Puhov Marl, a characteristic Upper Cretaceous formation of the Pieniny Klippen Belt in the Carpathians. The formation is rich in pelagic microfossils (calcisphaerulids and foraminifera). The Izsák Marl is a typical bathyal basin facies, deposited far from the continental source areas.

2.3.5.7 Palaeogene Flysch Deposition in the “Szolnok Flysch Trough”

The “Szolnok Flysch Zone” (hereafter SFZ) is located in the basement of the Great Plain (Figs. 1.86a, 1.88), extending along SW–NE strike in a length of 130 km from the town of Szolnok toward the Nyírség and Maramureş area in Romania and Sub-Carpathian Ukraine (Dudich 1982). The “Szolnok Flysch” has been drilled all along this belt in Hungary and also in some hydrocarbon exploration wells in East Romania as well (Paraschiv 1979).

The SFZ is located at the northeastern edge of the Tisza Mega-unit. The width of this belt is 15–20 km, but locally it may attain 30–40 km. The belt is parallel to a relatively elevated crystalline ridge to its south. North of the ridge the crystalline basement drops to a considerable depth and is overlain by the flysch sequence (Kőrössi 1959). The contact may be a tectonic one.

Hydrocarbon exploration wells encountered thick sequences of Senonian to Oligocene age, locally showing turbiditic sedimentation features. It has often been argued in international geologic literature whether the “Szolnok Flysch” is a true flysch or not. Based on the detailed study of the available cores Szepesházy (1973) claimed several times that it is a flysch sequence, although not all members of the succession show clear turbiditic features. Therefore the flysch character of the entire sedimentary succession of the Szolnok Zone has been generally accepted by the Hungarian geologic community and is cited overwhelmingly as “Szolnok Flysch”. The term “Szolnok Flysch Zone” has been used by several authors as a name of a tectonic unit as well. However, if we define a “true flysch” not only based upon its turbiditic character, but also to its geodynamic setting (e. g. submarine trench deposit, subduction

related sequence, a foreland basin of prograding nappe piles, etc.), the Szolnok Flysch cannot be considered as “true flysch”. This is why several authors distinct Inner Carpathian flysches (e.g. Podhale Flysch, Maramures Flysch, Szolnok Flysch) and Outer Carpathian “true” flysches in the region.

Beneath a 2,000 to 3,000 m-thick Quaternary and Neogene sedimentary and volcanic cover the over 1,000 m-thick clastic sedimentary series of the SFZ has been encountered in several wells in the last 40 years. The total thickness of the “Szolnok Flysch” is unknown since no well has ever reached its base. One of the exploration wells stopped within the complex after 1,400 m of penetration.

Previously the “Szolnok Flysch” was thought to have been deposited continuously from the Cretaceous to the Oligocene (Szepesházy 1973). Detailed studies of the available core samples (Báldi-Beke et al. 1981; Báldi-Beke and Nagymarosy 1993; Nagymarosy and Báldi-Beke 1993; Nagymarosy 1998) have shown that only a few Cretaceous and Palaeogene nannoplankton zones can be proved, and that others are completely missing. According to the most recent information the “Szolnok Flysch” sequence is non-continuous and it can be subdivided into several discrete units. The hiatuses must be interpreted as submarine unconformities or submarine erosional events, since it is difficult to suppose several very quick basin inversions and subaeric erosions.

Although more than 100 wells reached the flysch sequence its lithological composition is only incompletely known, because of insufficient coring and the lack of any continuously cored section.

The age of the oldest beds is Campanian and the cores can be assigned either to the Izsák (Puchóv) Marl or to the Debrecen Sandstone Formations. The top part of the Cretaceous (Maastrichtian) and practically the entire Palaeocene is missing in the core material. Nannoplankton assemblages of a few isolated cores have been interpreted by Báldi-Beke and Nagymarosy (1993) as transitional beds between the Palaeocene and Lower Eocene (NP 9–10 nannoplankton zones). These cores consist of red, variegated,

green or greenish-grey, rarely dark grey marl, locally non-calcareous shale and also finely rhythmic, turbiditic sandstone. Their distribution is restricted to a few drilling sites in the southwestern segment of the flysch belt (Fig. 2.24).

The Middle to Upper Eocene part of the flysch series was encountered much more frequently than the Cretaceous and Palaeocene one. Grey and variegated shale with finely rhythmic sandstone, poorly sorted sandstone, polymict conglomeratic sandstone, conglomerate and breccia are the characteristic rock types. Limestone and sandy marl, rarely *Nummulites* and *Lithothamnium*-bearing and very probably redeposited, were observed in the surroundings of Hajdúszoboszló. The Middle and Late Eocene deposits were reported in the entire area of the Szolnok Zone (Fig. 2.25).

Most cores were classified into the Upper Lutetian and Bartonian. Lower Lutetian is entirely missing and Priabonian biozones in the cores are rare.

The distribution of the Oligocene deposits is restricted to the northeastern part of the Szolnok Zone (Figs. 1.94, 2.26). They are practically absent in the Tisza Valley. Flysch characters are absent from the Oligocene part of the sequence. The most typical Oligocene lithofacies consists of clayey marl (very similar to the Kiscell Clay) with sandstone intercalations (rarely cross-bedded). *Lepidocyclina*-bearing conglomerate was encountered at one site.

The presence of the Lower Oligocene is uncertain in the Szolnok Zone. Upper Kiscellian and Egerian (NP 24–25 nannoplankton zones) deposits were cored quite frequently.

The actual preservation and recent distribution of the SFZ is strongly controlled by Early Miocene compressional tectonics and subsequent denudation until the Middle Miocene. Tectonic imbrication and erosion can be suspected as well, mainly for the deeper part of the sequence. This may also explain the lack or rarity of the Palaeocene and Cretaceous parts of the complex. In some Palaeocene cores red and variegated, non-calcareous and non-fossiliferous shale may suggest deposition below or about at the CCD.

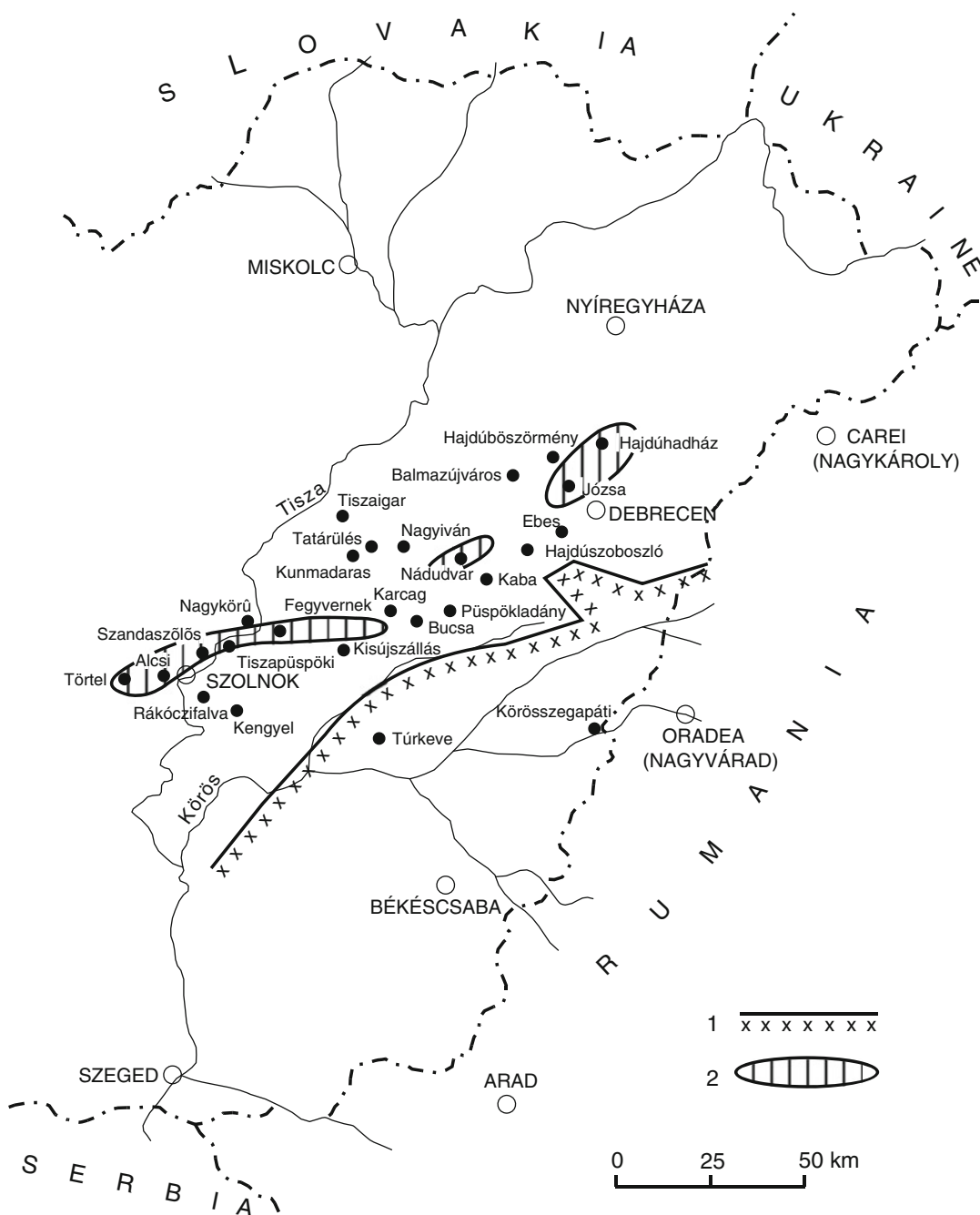


Fig. 2.24 A distribution of the Palaeocene to Early Eocene (NP 9–10 nannoplankton zones) beds in the Szolnok Flysch trough. 1 the northern boundary on the crystalline basement; 2 extension of the Palaeocene to Early Eocene beds

For the Eocene the calcareous nannoplankton assemblages of the “Szolnok Flysch” sequence indicate pelagic conditions. The composition of nannoplankton assemblages differs strongly

from that of the coeval near-shore assemblages of the Hungarian Palaeogene Basin (see Báldi-Beke 1984; Báldi-Beke and Nagymarosy 1993), underlining the possibility that no *direct*

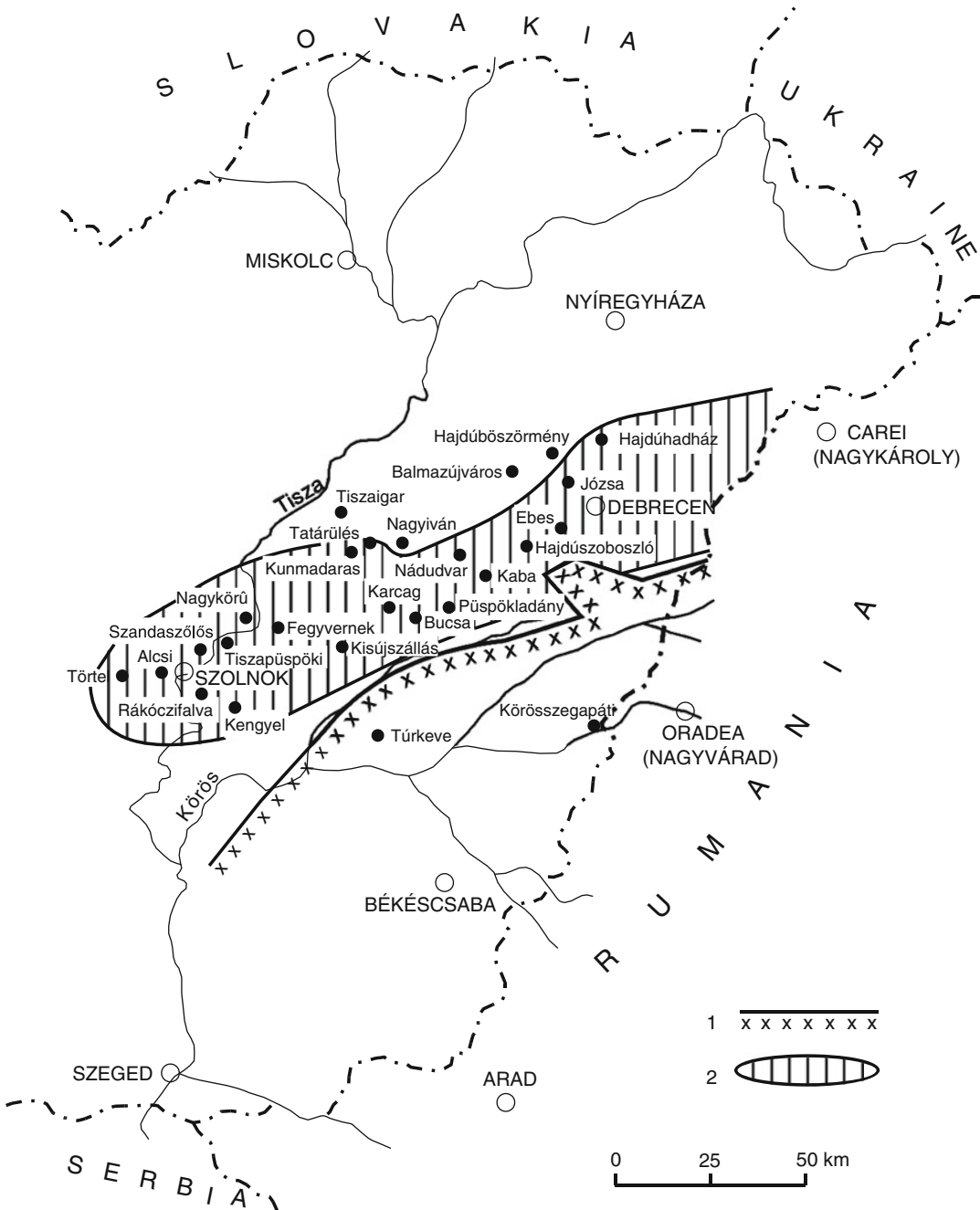


Fig. 2.25 Distribution of the Middle to Late Eocene (NP 16–19 nannoplankton zones) beds in the Szolnok Flysch trough. 1 the northern boundary on the crystalline basement; 2 extension of the Middle to Late Eocene beds

palaeogeographic connection between the two basins existed. The Oligocene nannoplankton assemblages of the “Szolnok Flysch” show less pelagic and more near-shore features than the

Eocene ones (Báldi-Beke and Nagymarosy 1993). Taking into consideration all the petrographic and palaeontological features of the “Szolnok Flysch” a gradual change in

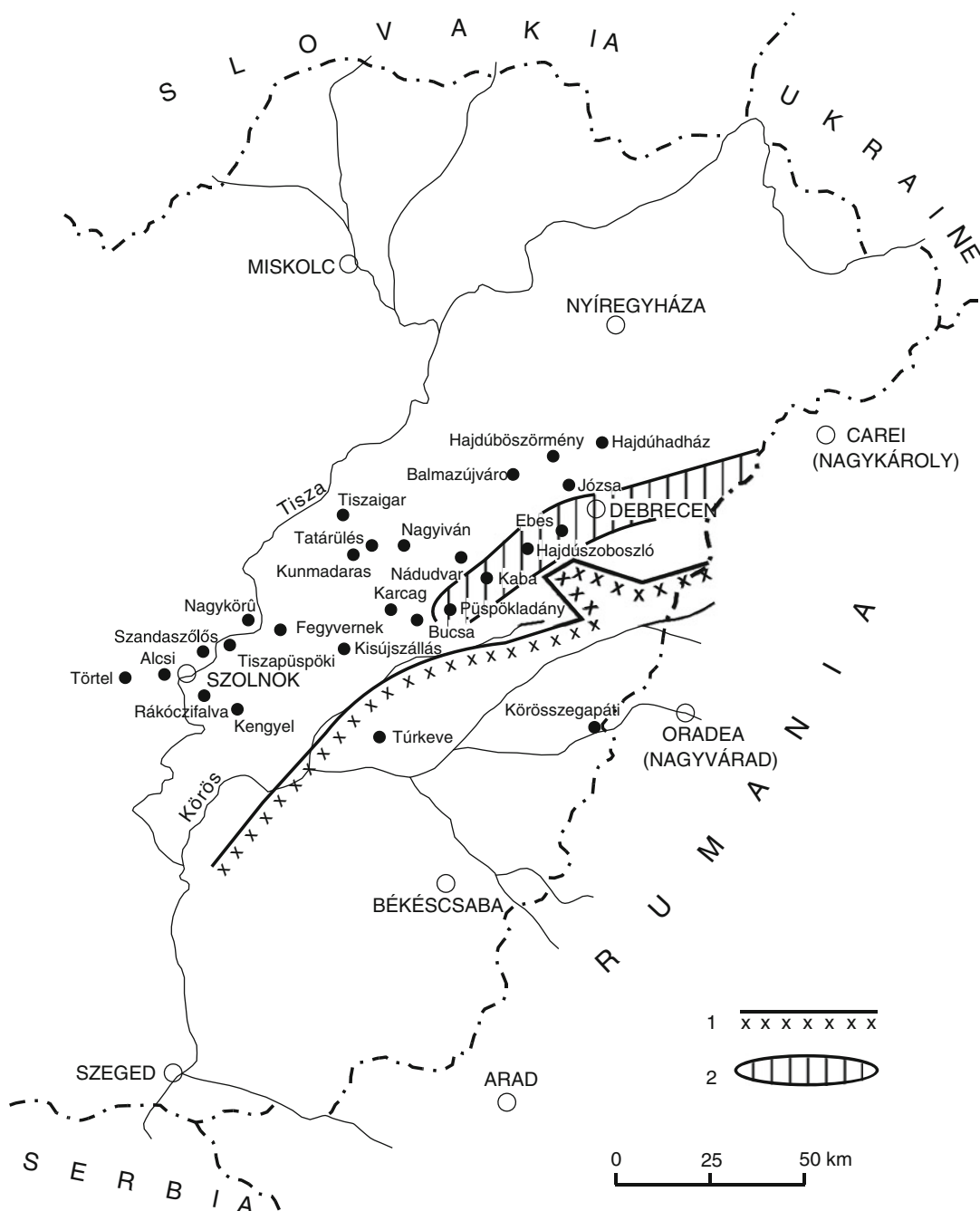


Fig. 2.26 Distribution of the Oligocene (NP 24–25 nannoplankton zones) beds in the Szolnok Flysch trough. 1 the northern boundary on the crystalline basement; 2 extension of the Oligocene beds

depositional conditions, from deep water/pelagic to shallower/nearshore, can be assumed.

Kőrössy (1959, 1977), Juhász (1966) and Szepesházy (1973) emphasised the strongly tec-

tonised character of the “Szolnok Flysch”. Dips between 70° and 90° as well as sheared and compressed sections (e.g. cores with vertical dip and folded structures) were frequently

reported. However, strongly compressed and imbricated structures are not only confined to the Szolnok Zone in this region. Pap (1990) described a number of imbrication structures from the basement of the eastern part of the Great Plain, also outside of the Szolnok Zone. He mentioned a borehole section located in the southern part of the flysch belt where even Tertiary (i.e. Upper Eocene) rocks were involved in the compressional structures. In two cases nanoplankton studies revealed that older, Cretaceous deposits have been thrust over younger, Tertiary ones in the vicinity of Debrecen and Nádudvar (Nagymarosy and Báldi-Beke 1993).

The examples presented here confirm that the “Szolnok Flysch” has been subjected to compressional tectonics and significant displacement following deposition. The youngest formation involved with certainly in these imbricated structures is of Late Eocene age; however, the heavily tectonised character of some Oligocene beds (sheared rocks with steep dips) suggests that compression must have taken place after the deposition of the Oligocene layers, in the Early Miocene before deposition of the overlying, non-compressed Middle Miocene (Badenian) rocks. The significant differences in thickness of the sequences may also be attributed to tectonic erosion.

The “Szolnok Flysch” is in an unusual position inside the Carpathian arc. No direct connection with any Carpathian or Dinaridic flysch units is evident. The SFZ can be traced eastward to Carei and Satu Mare (Romania) and then disappears beneath young volcanic masses of the Gutin Mts. However, the stratigraphic content of the “Szolnok Flysch” sequence provides some help in correlating it with the other flysch units.

It could be possible, on the one hand, that the “Szolnok Flysch” might be a displaced continuation of one of the Outer Carpathian flysch belts which disappears near the northeastern termination of the Mid-Hungarian Lineament. The northeastern termination of the “Szolnok Flysch”

points toward the Maramureş area. In this sector, the Pieniny Klippen Belt (Botiza Klippen; see Săndulescu 1980; Săndulescu et al. 1981) forms a bend and strikes southwestward in the direction of the “Szolnok Flysch”, as does the Outer Carpathian Magura Nappe that also ends in this region. The Magura Nappe and the Pieniny Klippen Belt (Botiza Klippen) pinch out toward the west. There is no known prolongation of the Pieniny Klippen Belt and the Magura Nappe further to the SE or SW. Consequently, the “Szolnok Flysch” (or a part of it) may be the continuation of the above-mentioned tectonic units.

According to another hypothesis, the “Szolnok Flysch” would be the continuation of the Inner Carpathian Flysch Belt, i.e. of the “Transcarpathian Flysch” in Maramureş. Thus, the “Szolnok Flysch” may be the subsurface prolongation of any of these three units (or a tectonically “mixed” structure of all three units). However, the autochthonous Central Inner Transcarpathian Flysch also pinches out toward the SW at the surface.

The comparison of the stratigraphic patterns of the individual units permits a choice among these solutions (Györfi et al. 1999). The several hiatuses in the “Szolnok Flysch” sequence suggest its marginal position during the Senonian and Palaeogene. Therefore its correlation with the continuous, basinal Outer Carpathian sequences seems improbable and can be excluded.

Similar Palaeocene and Eocene gaps have been observed, both in the Pieniny Klippen Belt and the Transcarpathian (Maramureş) Flysch Basin, where the diagnostic Puchov Marl also occurs (Bombița 1972; Dicea et al. 1980; Szász 1975).

The SFZ may have been deposited under conditions similar to those of the “Transcarpathian or Maramures Flysch” on the northern slope of a continental microplate (Tisza Mega-unit, Fig. 2.27). It can therefore be assumed that the “Szolnok Flysch” and “Transcarpathian Flysch” units are strongly related in their genesis and

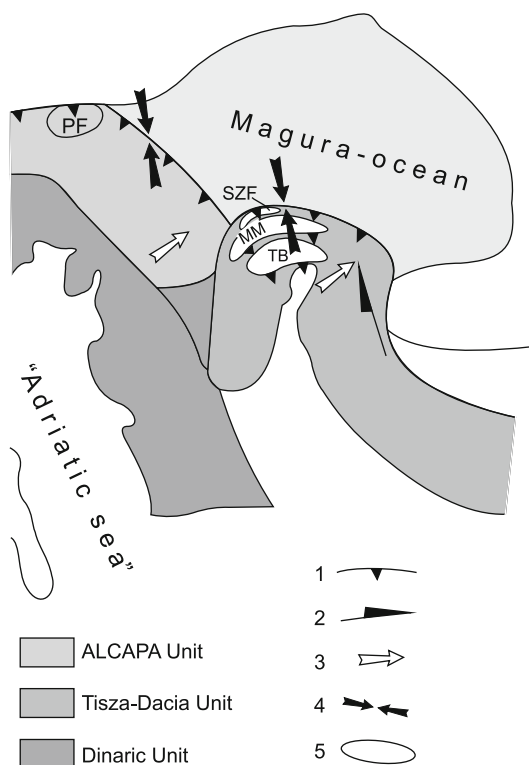


Fig. 2.27 Palinspastic reconstruction of the Carpathian-Pannonian region in the mid-Tertiary and the position of the so-called “Inner Carpathian flysch belts”. PF – Podhale Flysch, SZF – Szolnok Flysch, MM – Maramures Flysch, TB – Transylvanian basin, 1 upthrust zone, 2 transcurrent fault, 3 direction of the microplate-drift, 4 compressional field, 5 compressional basin (Csontos and Nagymarosy)

might form a continuous subsurface belt. This can be confirmed by a strong similarity between their stratigraphic columns.

The present-day distribution of the “Szolnok Flysch” shows considerable variation, which is not due to changes in the depositional systems but only to the effects of post-depositional erosion. In the southwestern segment of the unit only Cretaceous–Palaeocene–Eocene deposits occur, without younger layers. In the middle part of the Szolnok Zone Upper Cretaceous to Oligocene deposits occur. In the northeastern continuation of the belt, i.e. in the Maramureş area (Romania), Upper Cretaceous to Lower Miocene successions are known.

An explanation for this distribution pattern, i.e. the shift in the ending of sedimentation in the Szolnok–Maramureş Belt, could be a result of Late Eocene–Oligocene palaeogeographic–palaeotectonic dynamics (Fig. 2.27; Csontos et al. 1992).

In the northeastern prolongation of the Szolnok Zone the Botiza and Lapuş Nappes are thrust over Early Miocene deposits. This post-Oligocene compression shows a strong similarity to that in the Szolnok sector (Săndulescu et al. 1981), indicating a pre-Badenian tectogenesis for this belt.

2.3.5.8 Continental Palaeogene Basin in the Mecsek

South of the Mecsek Mts. (see Nagymarosy in Császár et al. 1990; Wéber 1982, 1985) more than 150 m of Palaeogene continental clastic deposits (maximum thickness 400 m) were encountered in the Szigetvár area. Palynologic studies have identified the lower part of this sequence as Eocene; the higher part was assigned to the Late Oligocene. According to Varga et al. (2004) the age of the whole sequence is Late Eocene.

Diagnostic lithofacies are variegated shales, clays, marls, sandstones, clast-supported conglomerates and breccias. Thin beds of coal and paleosols also occur. The Palaeogene rocks of this South Mecsek graben were most probably formed in an isolated, continental depositional basin.

2.3.6 Villány–Bihar Facies Unit

2.3.6.1 Coastal–Terrestrial Sedimentation in the Late Triassic

In contrast to the thick siliciclastic sequences of the Mecsek Unit the Villány–Bihar Unit is characterised by a thin, coastal–continental Upper Triassic succession akin to the “Carpathian Keuper” facies of the European shelf of the Tethys. In the Villány Hills, Ladinian dolomite is conformably overlain by a formation made up of an alternation of yellowish-grey dolomitic marl and

dolomite, brownish- or greenish-grey sandy siltstone, and greyish-white quartzarenite (Mészhegy Formation; Fig. 2.26). In the upper part of the 15–40 m-thick formation the dolomite layers disappear and greenish–reddish variegated siltstone becomes predominant. Marine fossils are completely absent from these layers. In addition to plant remnants only bones of reptiles have been found so far, which do not allow an exact age determination. Above the Middle Triassic carbonates a few wells also encountered similar sequences in the basement of the Great Plain.

2.3.6.2 Discontinuous Shallow Marine Deposition in the Jurassic

The Jurassic sequence of the Villány–Bihor Zone is known mainly from the Villány Hills. As far as the basement of the Great Plain is concerned core data is only available for the Malm formations.

In the Villány Hills, unconformably overlying the Upper Triassic rocks, the Jurassic series begins with a quartzarenite bed, grading upward into shallow marine, sandy, crinoidal limestone with conglomerate interlayers. In the conglomerate quartzite and dolomite components are recognised, indicating the proximity of a continental hinterland. The next bed contains large pebbles and conglomerate and limestone boulders. These beds are overlain by yellowish-grey limestone with ammonoids, belemnoids, and brachiopods, followed by grey, strongly bioturbated, thick-bedded, cherty, crinoidal limestone (Fig. 2.29). Based on ammonites the only 6–8 m-thick limestone formation (Somsichhegy Limestone) can be assigned to the Pliensbachian (Vörös 2009). The appearance of a condensed marine series following a long-lasting subaerial hiatus can be explained by tectonically-controlled transgression which was followed by another gap of about 20 million years duration.

Above the hiatus a thin, yellow, sandy limestone bed, rich in Bathonian ammonites, occurs. It is overlain by an extremely condensed limestone layer, very rich in ammonites (Fig. 2.28). In the 30–40 cm-thick layer more than 150 species were found (Géczy 1982b). The fossil

assemblage indicates a pelagic environment. It is assumed that a pelagic plateau came into being in the Middle Jurassic which was strongly affected by currents, leading to permanent removal of the sediments. Thin layers of the Villány Limestone may have been preserved due to microbial encrustation. Most of the ammonites are characteristic of the European province; a smaller part of the assemblage, however, is Mediterranean, indicating the incipient break-off and separation of the “Tisza Block” from the European Plate during the Dogger (Géczy 1973).

The ammonite-bearing limestone is overlain by 300 m of thick-bedded, grey, brownish- or yellowish-grey limestone (Szársomlyó Limestone; Fig. 2.28). It is characterised by a peloidal, oolitic–oncolidal, micritic texture. Megafossils are very scarce. In the lower part of the formation pelagic elements prevail in the microfossil assemblage: protoglobigerinids and *Saccocoma* fragments. In the upper part, however, a typical shallow marine biofacies appears and the pelagic elements disappear. Based on the microfossils the age of the formation is Oxfordian–Tithonian and its upper member may extend into the Berriasian (Bodrogi et al. 1993). In the Jurassic/Cretaceous boundary interval a significant part of the Villány–Bihor Unit became subaerially exposed and the Upper Jurassic carbonates were affected by karstification.

2.3.6.3 Carbonate Platform Development in the Early–Middle Cretaceous

At the beginning of the Cretaceous bauxite was accumulated in the karstic depressions of the Upper Jurassic limestone. Small deposits have been found in the Villány Hills (Dudich and Mindszenty 1984); larger deposits of commercial value are known from the Apuseni Mts. in Romania. Pyroclastics connected with Early Cretaceous volcanic activity in the Mecsek Zone may have been the source for the bauxite.

In the more intensively subsiding parts of the unit (i.e. the area of the southern tectonic slices in the Villány Hills) transgression probably began as early as the Berriasian or Valanginian.

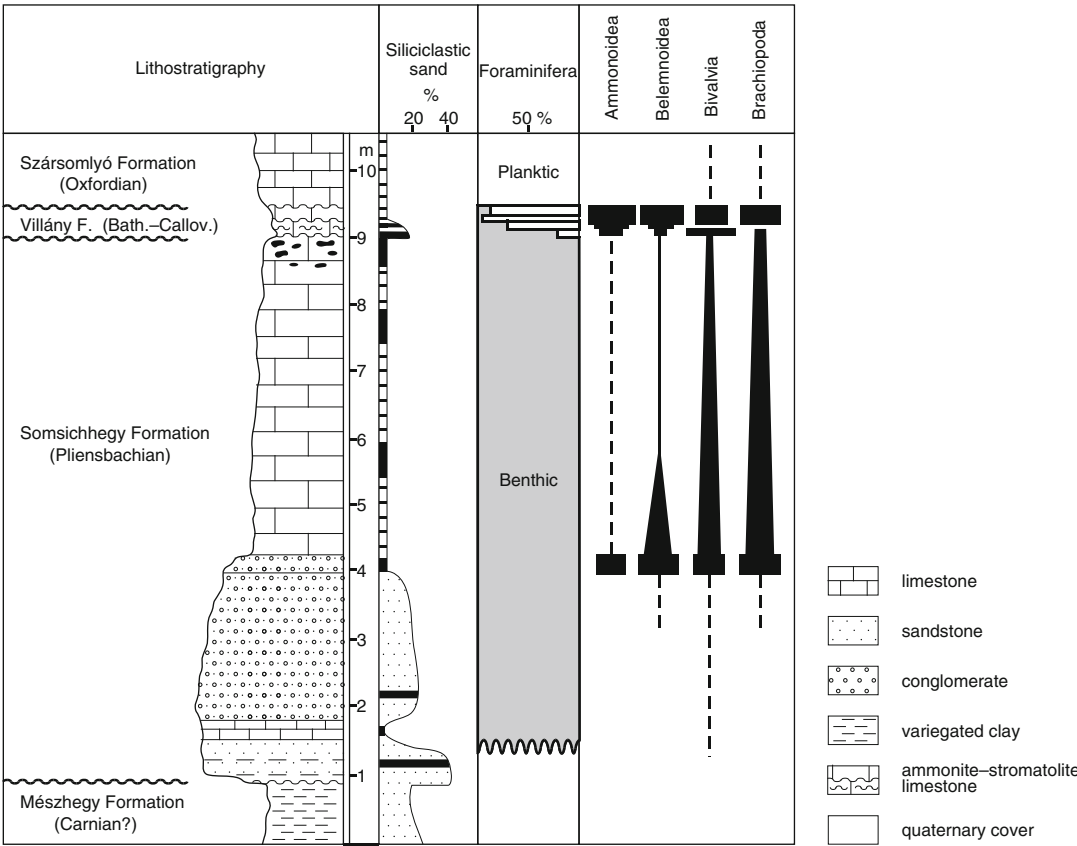


Fig. 2.28 Upper Triassic and Jurassic formations on the Templom Hill, Villány, Villány Hills. (After Vörös 1990)



Fig. 2.29 Ammonite-bearing bedding plane (hard-ground) in the Bathonian-Callovian Villány Formation (Photo: J. Haas)

A carbonate platform came into existence which maintained itself until the Albian. In the slowly

subsiding areas shallow marine carbonate accumulation began later and some areas only became inundated in the Albian (Fig. 2.30).

On the carbonate platform 400–500 m-thick, light grey, thick-bedded limestone with Urgon facies characters was formed (Nagyharsány Limestone). Outside of the outcrops in the Villány Hills the formation was also encountered in wells in the Danube–Tisza Interfluvium and Trans-Tisza regions.

In the Villány Hills, in the lower part of the sequence intraclastic and stromatolite layers alternate with thick grey limestone beds (Császár 2002). In the basal layers fresh or brackish water *Chara* biofacies are characteristic, whereas normal marine fossil elements prevail gradually upsection. This part of the formation was deposited in a tidal flat and in a

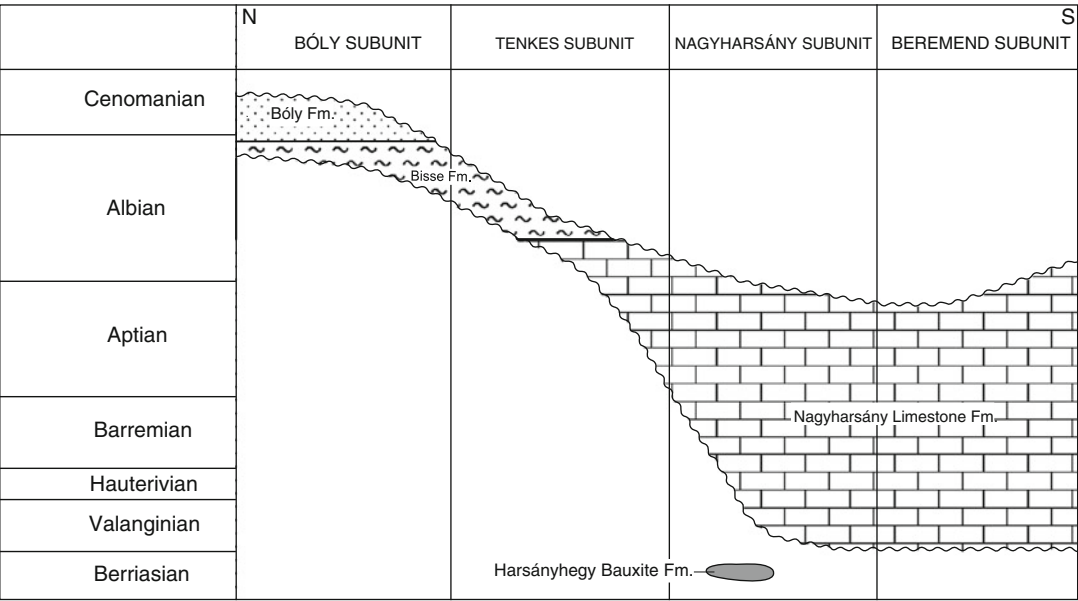


Fig. 2.30 Cretaceous formations of the Villány Hills (After Császár 1992)

shallow lagoon. Metre-scale cyclicity is a result of short-term sea level oscillation. The next interval of the succession is characterised by a prevalence of rudists (*Requena*, *Toucasia*) and the appearance of *Chondrodonta*, dasycladacean algae and orbitolinids. Subsequently hermatypic corals appear in rock-forming quantity. The former bio-associations may have occupied the inner part of the carbonate platform, whereas the corals may have colonised the outer platform. The uppermost part of the succession is characterised by massive limestone with orbitolinids (*Orbitolina texana*). This facies represents the upper foreslope of the carbonate platform (Császár 2002).

In the southern part of the Danube–Tisza Interfluvial area, in addition to the typical “Urgonian” biota (i.e. shallow marine fossils) planktonic foraminifera also appear, indicating a foreslope depositional environment (Bérczi-Makk 1986). In the Trans-Tisza area the Lower Cretaceous sequence begins with an upward-fining, clastic, continental-coastal sequence: conglomerate, sandstone, and siltstone. It is overlain by dark grey, oolitic limestone and calcareous marl (Bérczi-Makk 1986).

2.3.6.4 Pelagic Basin Formation at the End of the Mid-Cretaceous

In the Albian, probably as a result of eustatic sea level rise and increased influx of fine-grained terrigenous material, the carbonate platforms were drowned and new basins began forming, probably in connection with the initiation of compressional structural evolution. In the northern part of the Villány Hills, above the strongly thinned Nagyharsány Limestone, grey marl, containing rich pelagic foraminifera and ammonites faunas, has been encountered (Bisse Marl). The age of the marl layers is Late Albian (Fülöp 1966). Directly overlying the Malm limestone (Szár-somlyó Limestone), marl with *Rotalipora* was encountered in a cored well in the northern foreland of the Villány Hills (Császár 1998; Bodrogi 1998).

In the same well section the Bisse Marl is overlain by a succession made up of an alternation of conglomerate, sandstone and marl (Bóly Formation) which was assigned to the Lower Cenomanian (Császár 1998). In the conglomerate layers components originating from the crystalline basement and the older Mesozoic

formations were found. They may have been transported into the pelagic basin by gravity mass flow. In the sandstone layers gradation and slump structures were observed.

Upper Albian–Cenomanian sequences consisting of coarse clastics and siltstone and marl layers were also encountered in the basement of the Great Plain in the Villány–Bihor Zone (Szederkényi 1984). The features of the Bóly Formation indicate a basin formed in the foreland of a thrust zone. It suggests the initiation of nappe tectonics in the Villány–Bihor Zone in the Cenomanian.

2.3.6.5 Senonian Basin Evolution

In the Apuseni Mts., Romania, i.e. in the eastern parts of the Tisza Mega-unit, the main phase of nappe formation occurred in the Coniacian (Ianovici et al. 1976), and this is probably true for the entire mega-unit. In the foreland of the newly-formed thrust belts basins began forming in the Late Santonian–Campanian (Haas and Péro 2004).

Within the Hungarian part of the Villány–Bihor Zone that is in the basement of the Great Plain there are significant differences in the characteristics of the Senonian sequences. In the western part of the Danube–Tisza Interfluvium area light grey, arkosic sandstone, conglomerate and breccia (Szank Conglomerate) occur at the base of the Senonian series (Fig. 2.31). The coarse components originated from granitoids, micaschists, and Mesozoic basement rocks. No marine fossils and only a few sporomorphs were found in the formation which may have been accumulated in a terrestrial basin during the early Senonian (Haas 1987; Szentgyörgyi 1983).

Above an uneven erosion surface, but without any significant angular unconformity, the Szank Conglomerate is overlain by dark grey shale (silty marl – Csikéria Marl; Fig. 2.31). The 50–120 m-thick shale unit also contains in some horizons pebbles which are predominantly of granitoid origin. A large part of the formation is intensely bioturbated. Fragments of *Inoceramus* bivalves are common. The marl layers are rich in planktonic foraminifera (*Globotruncana*, *Hed-*

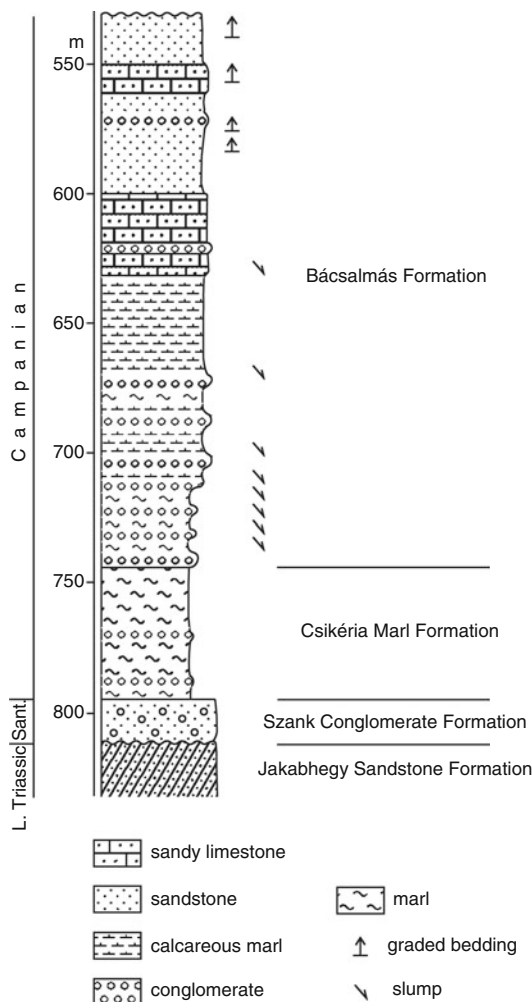


Fig. 2.31 Senonian formations in the southern part of the Danube–Tisza Interfluvium area, core Bácsalmás Bá-1 (After Haas 1987)

bergella, *Heterohelix*), nannoplankton, and sporomorphs. The fossil assemblage indicates a pelagic depositional environment; the coarse terrigenous clastics and the significant amount of continental plant remnants, however, suggest uplifted ranges (fronts of the thrust belts) in proximity of the basin.

The shale unit passes upward into a 100–200 m-thick, lithologically extremely variable, formation (Bácsalmás Formation; Fig. 2.31). It is made up of clay, fine siliciclastics, and biocalcarene. Although argillaceous, siliciclastic

and carbonate layers may occur everywhere in the sequence, marl is predominant in the lower, limestone in the middle, and siliciclastics in the upper part of the formation. In the lower part of the sequence terrigenous conglomerate and intraclastic intercalations, together with slump structures, are common (Haas 1987). The marl layers are rich in planktonic calcisphaerulids. The carbonate layers are characterised by calcarenitic packstone texture with fragments of molluscs, red algae, and benthonic foraminifera (*Pseudosiderolites*). Based on nannoplankton and foraminifera the age of the formation is Campanian (Siegl-Farkas 1999).

In the eastern part of the Villány–Bihar Unit, in the basement of the Great Plain, siliciclastics became predominant in the Senonian sequences (Körös Formation). Here, above basal breccia, the successions consist of an alternation of dark grey sandstone and siltstone layers, locally with conglomerate interlayers. Lamination and convolute bedding are common. In the eastern part of the Trans-Tisza area the thickness of this formation may attain 1,000 m, whereas in other places it is only a few hundred metres thick (Szentgyörgyi 1983). The thicker sequences show features similar to that of typical flysch, whereas the thinner sequences of finer grain size may have been deposited farther from the thrust belt in the inner part of the foreland basin.

2.3.7 Békés–Codru Facies Unit

In the basement of the Békés Basin Lower to Middle Jurassic, red limestone was encountered

in a few wells; it was identified as the Moneasa Limestone of the Finis Nappe, part of the Codru Nappe System in Romania (Haas and Péro 2004). It was deposited in a well-oxygenated basin as well as on slope areas and has yielded a “Germanic-type” brachiopod, *Gryphaea* and belemnite fauna in the Codru Mts.

A number of wells in the Békés Basin encountered Upper Jurassic to Lower Cretaceous formations. The several hundred m-thick series consists of grey and red clayey marl, marl, calcareous marl and limestone layers with sandstone intercalations in the upper part of the formation. The lower part of the series contains poor pelagic microfauna with radiolarians, *Saccocoma* fragments, and calpionellids, indicating a Late Jurassic and Early Cretaceous age (Bérczi-Makk 1986). A non-fossiliferous, dark grey limestone and marl sequence which was also encountered in the Békés Basin was assigned to the Lower Cretaceous (Grow et al. 1994).

Nappe formation and the most intensive deformations probably took place in the Late Cretaceous. Upper Cretaceous and Palaeogene formations are absent from the Hungarian sector of the Békés–Codru Zone.

2.4 Regional Geological Cross-sections

Geological setting and relationships of the basement rocks of the Pannonian Basin and basic characteristics of the Cenozoic basin filling are displayed on a series of regional cross-sections (Figs. 2.33–2.37). The position of the sections is shown on Fig. 2.32

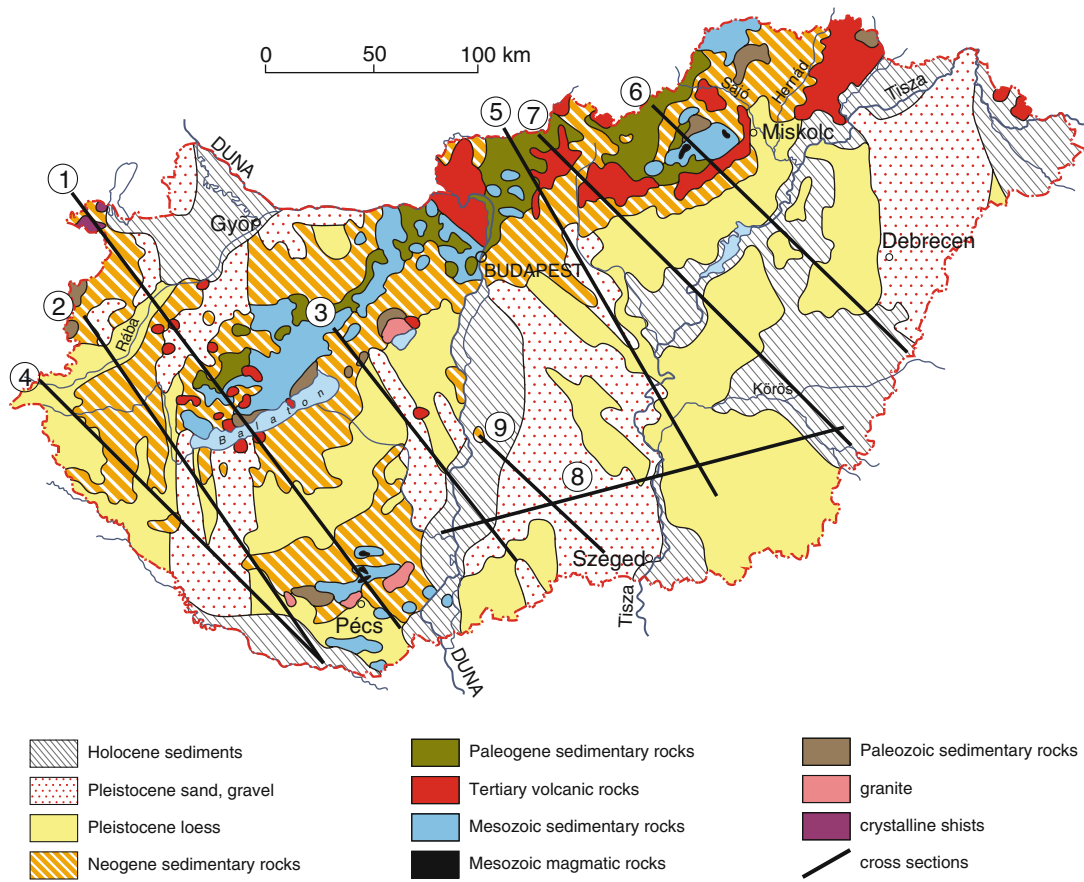


Fig. 2.32 Simplified geological map of Hungary displaying position of geologic cross-sections Figs. 2.33–2.37

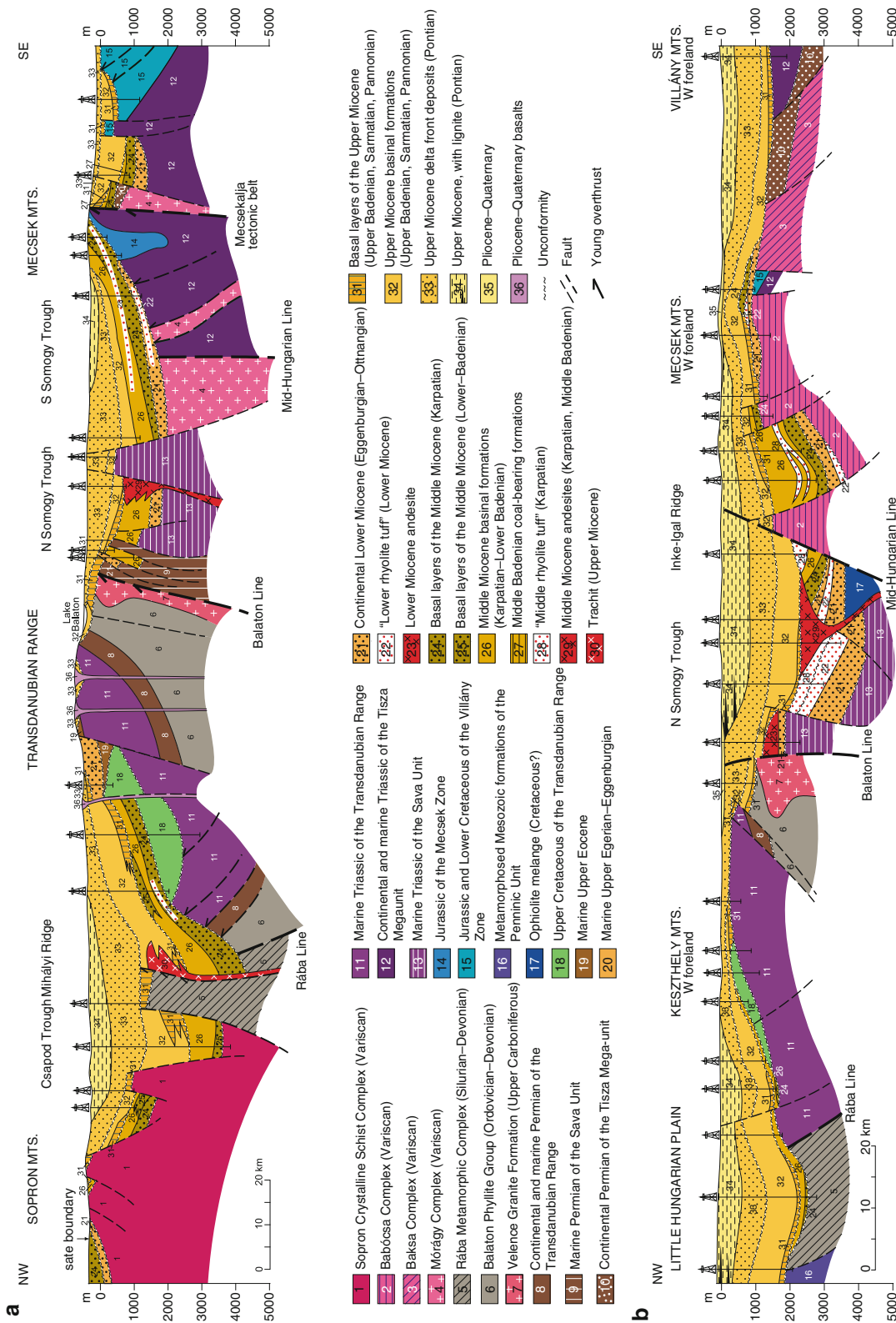
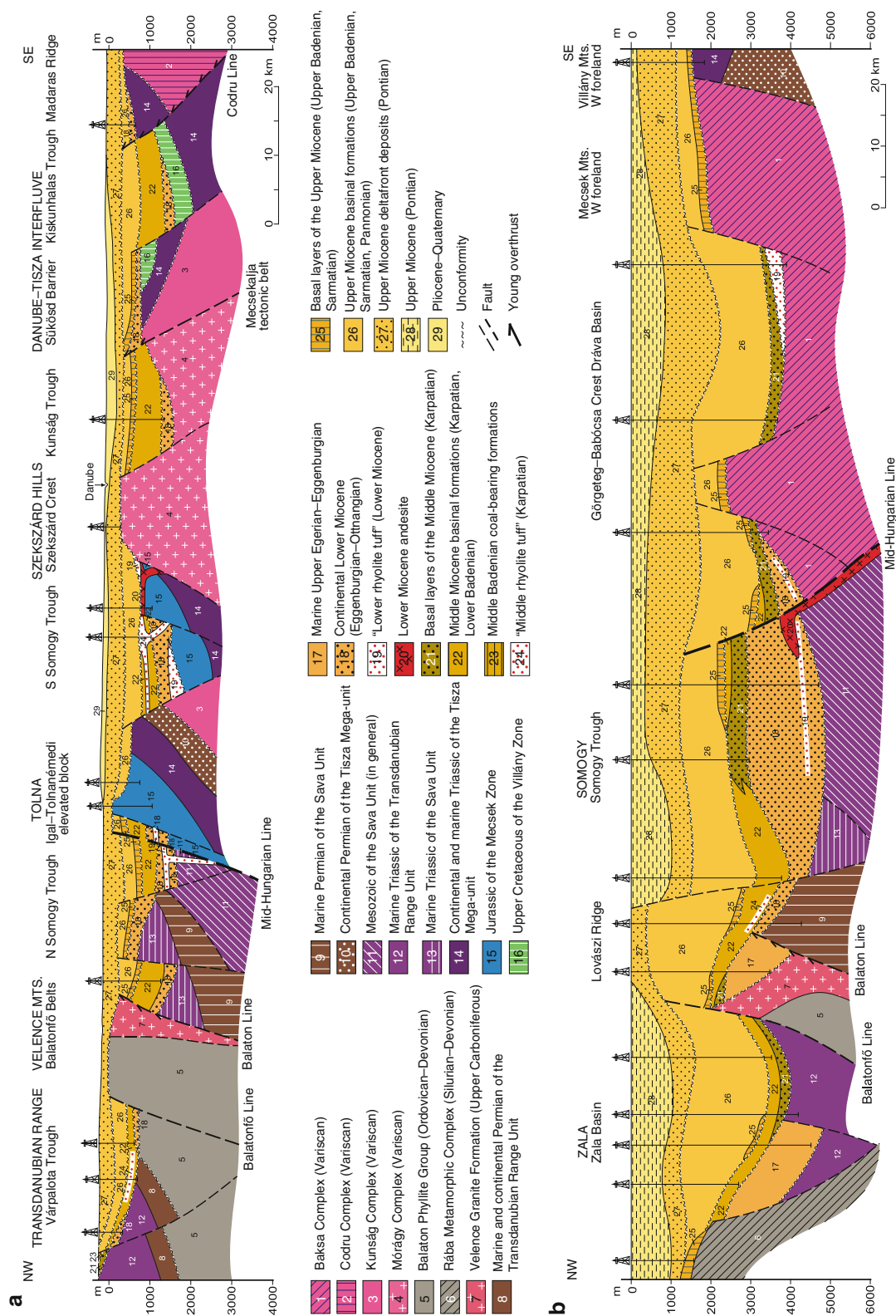


Fig. 2.33 Geologic cross-section through the Sopron Mts – Little Plain – Transdanubian Range – Somogy Trough – Mecsek Mts. (a) (No. 1 on Fig. 2.32). Geologic cross-section through the south-western part of the Little Plain – Transdanubian Range – Somogy Trough – Kadarkút Trough – SW Mecsek Mts. – western foreland of the Villány Hills (b) (No. 2 on Fig. 2.32)



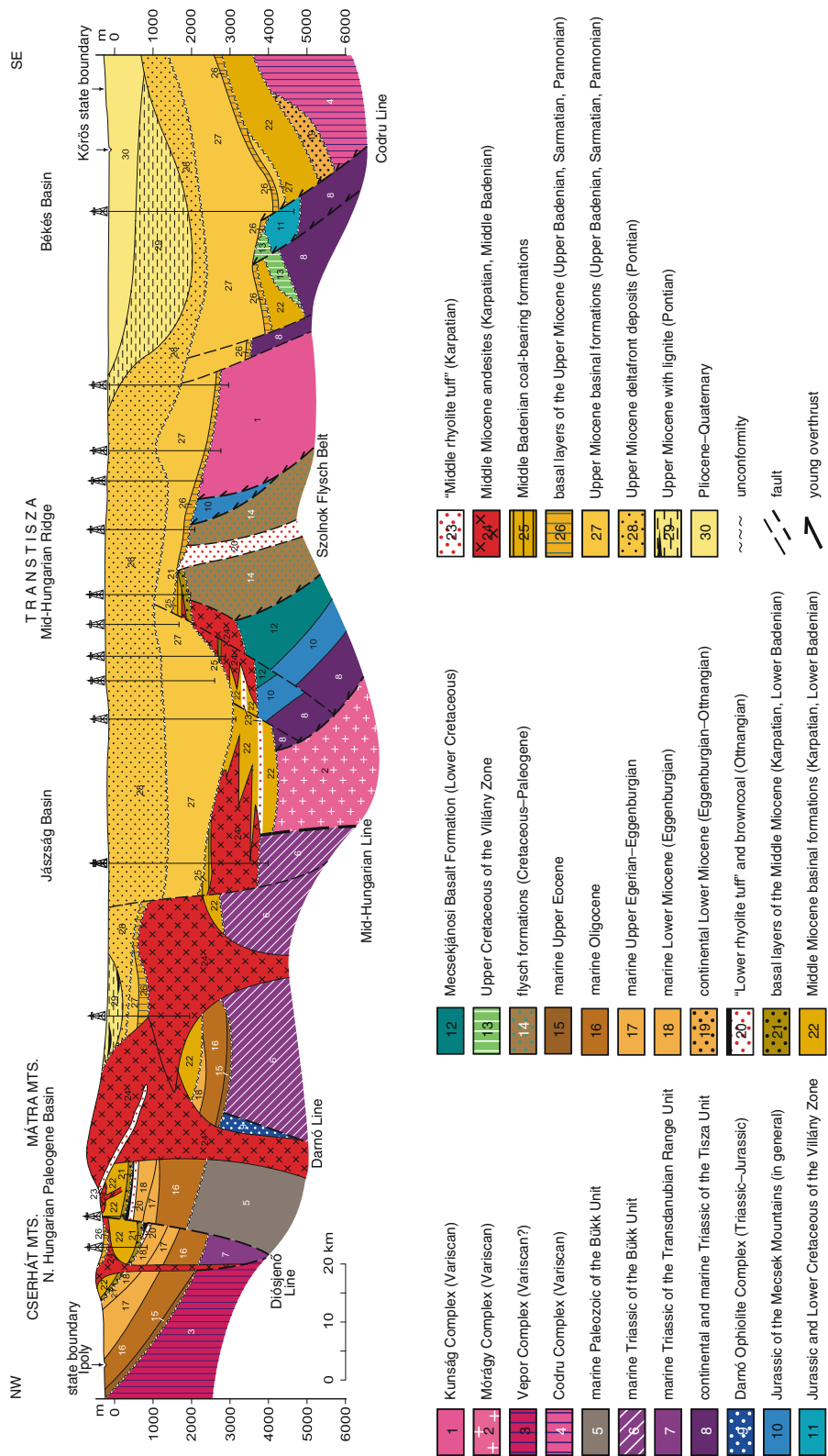


Fig. 2.35 Geologic cross-section between the Cserhát Mts, North Hungary and the south-eastern part of the Trans-Tisza area (No. 7 on Fig. 2.32)

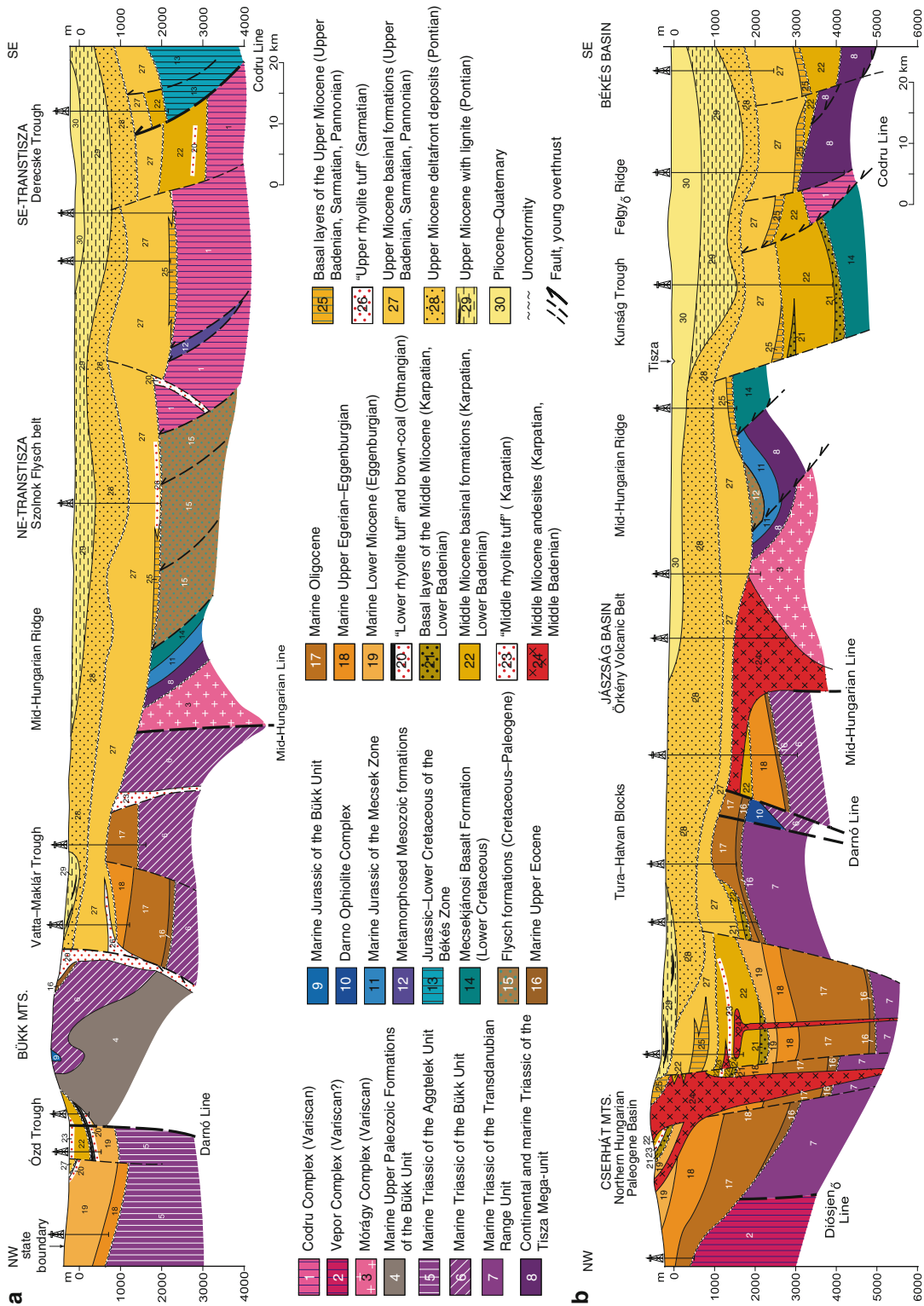


Fig. 2.36 Geologic cross-section between the Bükk Mts., NE Hungary and the Dercske Trough in the Trans-Tisza area (a) (No. 6 on Fig. 2.32. Geologic cross-section between the Cserhát Mts., North Hungary and the Békés Basin in the southern part of the Trans-Tisza area (b) (No. 5 on Fig. 2.32)

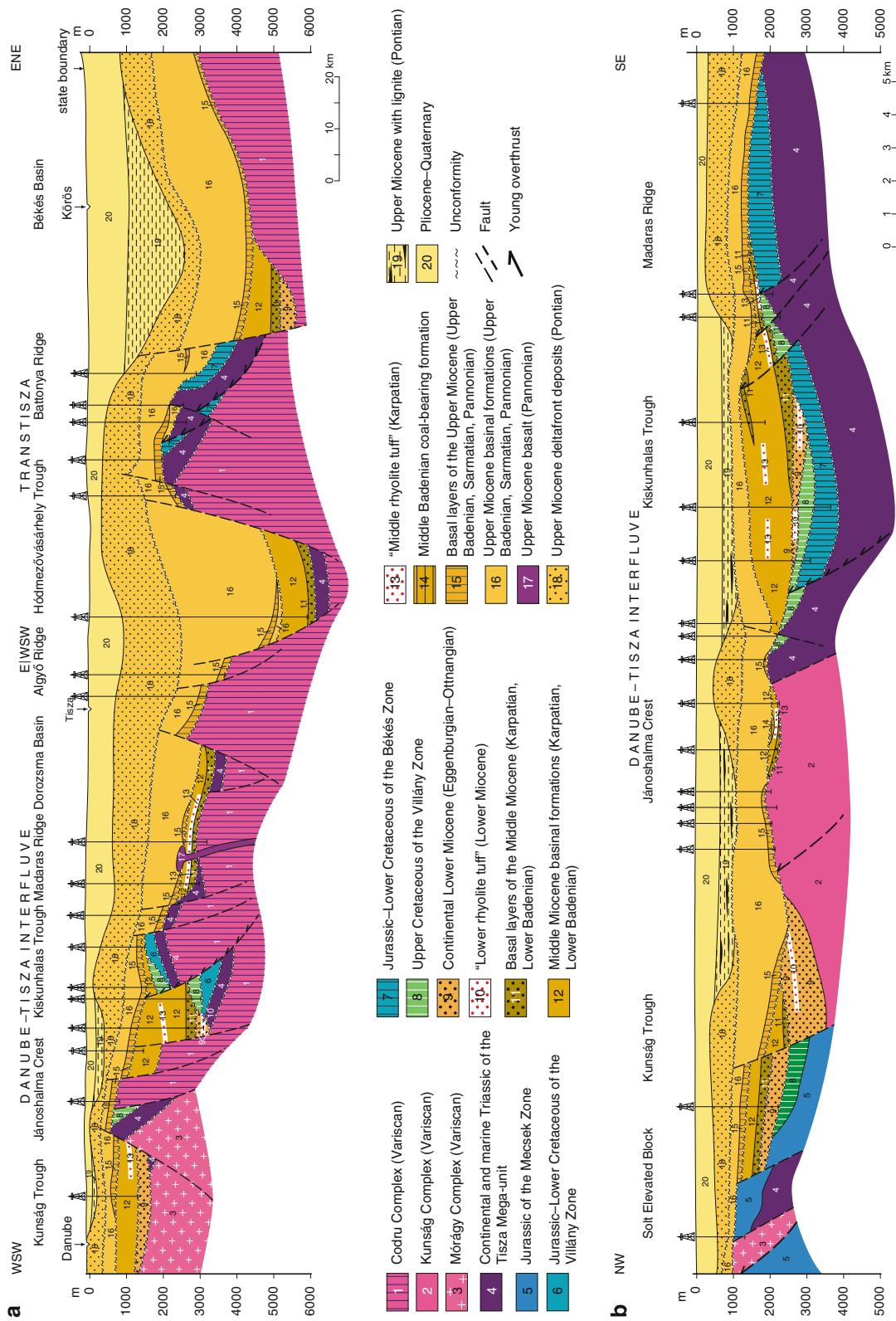


Fig. 2.37 Geologic cross-section between the Kunság Trough and Békés Basin (a) (No. 8 on Fig. 2.32. Geologic cross-section trough the southern part of the Danube-Tisza Interfluvium (b) (No. 9 on Fig. 2.32)



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