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GEODYNAMIC EVOLUTION OF THE CARPATHIANS

(5 Figs.)

Ewolucja geodynamiczna Karpat

(5 fig.)

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Abstract: The Carpathian arc is heterogenous. The western and the eastern parts of the Carpathians differ in style of tectonic structure and history of deformation. The linkage of the paraoceanic crust zone along the Alps, the Carpathians and the Dinarides is proposed for the Jurassic-Cretaceous phase of structural evolution of these Alpine ranges. The Neogene phase of structural evolution of the Carpathians was controlled by relative movements of the western and the eastern blocks of the inner structural units of the Carpathians. The Pieniny Klippen Belt is interpreted as the trace of a major sinistral transcurrent fault. The Neogene volcanic rocks of the Carpathians form two belts associated with the Western and the Eastern Carpathians. The curvature of the Carpathians arc is controlled by the boundaries of the rigid units of the Foreland.

Key words: Carpathians, tectogenesis, palinspastic reconstruction, paleogeography.

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Treść: W heterogenicznym łuku karpackim część zachodnia i wschodnia różnią się stylem i historią deformacji. Przedstawiono hipotetyczny przebieg strefy skorupy paraoceanicznej w Alpach, Karpatach i Dynarydach dla jurajsko-wczesnokredowego

etapu ich rozwoju. Neogeńska faza ewolucji strukturalnej Karpat podporządkowana była względnym ruchom zachodniego i wschodniego bloku jednostek skonsolidowanych we wcześniejszych, kredowych fazach tektogenicznych. Bloki te rozdzielał basen fliszu Szolnok w podłożu Wielkiej Niziny Węgierskiej. Pieniński Pas Skałkowy interpretowany jest jako ślad lewego uskoku przesuwczego. Neogeńskie skały wulkaniczne Karpat tworzą dwa niezależne łuki związane z Zachodnimi i z Wschodnimi Karpatami. Krzywizna łuku karpackiego jest wymuszona przez granice sztywnych bloków przedgórza.

THE STRUCTURAL UNITS OF THE CARPATIANS

The Carpathian Mts. are forming a nearly perfect arc, and their apparent regular structural pattern was a misleading factor in some models of structural evolution of this range proposed during the past fifteen years. In reality, the structure of the Carpathian arc is markedly heterogenous, polyphase and heterochronic. Great differences of structure exist between the western and the eastern part of the Carpathians.

The following major structural units are distinguished in the Carpathian arc and its interior (Fig. 1, Fig. 2).

— the Foreland

— the Foredeep, with in inner zone folded and thrust over the outer zone,

— the Outer Flysch Belt,

These units are extending along both the western and the eastern part of the Carpathians.

Inside the Outer Flysch Belt, the Western Carpathians comprise the following units:

— the Pieniny Klippen Belt

— the Central West Carpathians

— the Inner West Carpathians and the Hungarian Mts.

In the Eastern Carpathians the following units are situated inside the Outer Flysch Belt:

— the Inner East Carpathians,

— the Northern Apuseni Mts.,

— the Southern Apuseni Mts.,

— the Southern Carpathians.

The eastern extremity of the Pieniny Klippen Belt is situated south of the western extremity of the Inner East Carpathians. This is one of the important points determining the relations between the western and the eastern part of the Carpathians. The other important point is the presence of the Szolnok Flysch Belt in the basement of the Great Pannonian Basin. The folded Szolnok Flysch Belt separates the inner structural units of the Western and the Eastern Carpathians.

Fig. 1. Principal structural units of the Carpathians. 1 — Neogene volcanic rocks exposed: MS — Central Slovakia, M — Matra, PT — Prešov-Tokay, VG — Vihorlat-Gutin-Gutii, CH — Calimani-Hargitha. 2 — Neogene volcanic rocks in subsurface

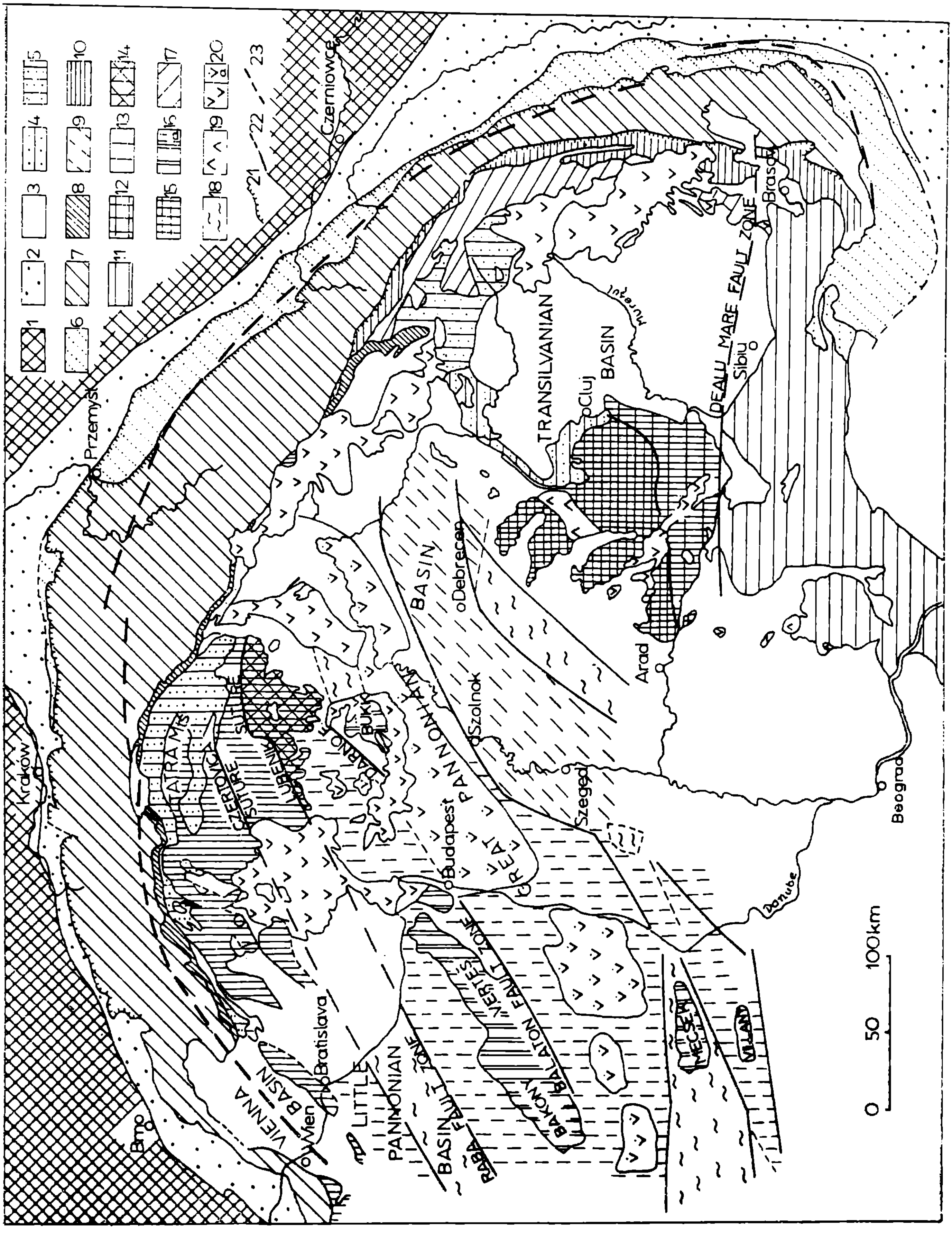
Fig. 1. Główne jednostki strukturalne Karpat. 1 — Neogeńskie skały wulkaniczne na powierzchni, symbole literowe jak wyżej. 2 — Neogeńskie skały wulkaniczne w głębi, w obrębie osadowego wypełnienia basenu panońskiego

Two Palaeogene basins, superposed on folded structural units of the Inner Carpathians and filled with flysch sediments, only slightly deformed, are adjoining the northern extremity of the Szolnok Flysch Belt. These are: the Central Carpathian Flysch in the Western Carpathians and the Maramureș Flysch in the Eastern Carpathians.

Young Neogene basins, superposed on various structural units are: the Vienna Basin, the Pannonian Basin, and the Transilvanian Basin.

Volcanic belts of Neogene-Quarternary age are completing the list of the major structural units of the Carpathians.

The following brief description of the structural units listed above is extending an earlier outline (Unrug, 1979).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VIENNA BASIN
LITTLE PANNONIAN BASIN
RABA FAULT ZONE
BAKONY
BALATON FAULT ZONE
VERTES ZONE
MECSEK
VILLANY
TATRA Mts
SUTURE
Budapest
GREAT PANNONIAN BASIN
Szolnok
Debrecen
Szeged
Arad
Beograd
Danube
Cluj
TRANSILVANIAN BASIN
Muresul
DEALU MARE FAULT ZONE
Sibiu
Brasov
0 50 100km

The Foreland

The Western extremity of the Foreland of the Carpathians is formed by the Bohemian massif. To the east the following units form the Foreland: the Variscan fold belt of the Eastern Sudetes Mts., the epi-Assynthian platform of the Upper Silesia Massif (Ślączka, 1976), the epi-Caledonian--Variscan platform, the heterogenous Dobrudja block, and the Moesian platform.

Fig. 2. Tectonic map of the Carpathians. 1 — Foreland, 2 — Foredeep. Post tectogenetic cover: 3 — Neogene (faulted) and Quarternary in intramontane basins, not folded, (basement of neogene shown in the Panonian Basin, symbols 9, 16, 18). 4 — Paleogene in epicontinental facies, not folded, 5 — Paleogene in flysch facies, weakly folded in intramontane basins. Units formed during Neogene tectogenesis: 6 — inner zone of the Foredeep folded and thrust-faulted, 7 — Outer Flysch Belt, 8 — Pieniny Klippen Belt, 9 — Szolnok Flysch Belt. Units formed in Late Senonian tectogenesis: 10 — Ceahlau nappe, 11 — Southern Carpathians, 12 — Southern Apuseni. Structural units formed in pre-Senonian tectogenesis: 13 — Central West Carpathians, 14 — Inner West Carpathians, 15 — Northern Apuseni, 16 — Hungarian Mts., and Late Paleozoic-Mesozoic rocks in the basement of the Pannonian Basin (a). Structural units formed in pre--Cenomanian tectogenesis: 17 — Inner East Carpathians. Other units: 18 — Crystalline rocks in the basement of the Pannonian Basin, 19 — Late Cretaceous — Paleogene volcanic rocks, 20 — Neogene volcanic rocks, a — in subsurface, 21 — Northern limit of overthrust of the Inner Foredeep and the Outer Flysch Belt, 22 — Major rejuvenated in Neogene time fault zones, 23 — Axis of negative gravimetric anomaly

Fig. 2. Mapa tektoniczna Karpat. 1 — Przedgórze, 2 — Rów przedgórski. Pokrywa post--tektoniczna: 3 — Neogen i czwartorzęd w basenach śródgórskich, niesfałdowany (podłoże neogenu przedstawione dla basenu panońskiego, symbole 9, 16, 18), 4 — Paleogen w facji epikontynentalnej, 5 — Paleogen w facji fliszowej, słabo sfałdowany (flisz Podhala i Zachodnich Karpat Centralnych, flisz marmaroski). Jednostki utworzone podczas tektogenezy neogeńskiej: 6 — wewnętrzna część rowu przedgórskiego, sfałdowana i nasunięta płaszczowinowo (płaszczowina stebnicka, fałdy pokucko-borysławskie), 7 — Flisz zewnętrzny, 8 — Pieniński Pas Skałkowy, 9 — Flisz Szolnok. Jednostki utworzone podczas tektogenezy późnosenońskiej: 10 — Płaszczowina Ceahlau, 11 — Karpaty Południowe, 12 — Apuseni Południowe. Jednostki utworzone podczas tektogenezy przedsenońskiej: 13 — Zachodnie Karpaty Centralne, 14 — Zachodnie Karpaty Wewnętrzne, 15 — Północne Apuseni, 16 — Góry Węgierskie i późnopaleozoiczne i mezozoiczne utwory podłoża Basenu Panońskiego (a). Jednostki strukturalne utworzone podczas tektogenezy przedcenomańskiej: 17 — Wschodnie Karpaty Wewnętrzne. Inne jednostki: 18 — Skały krystaliczne w podłożu Basenu Panońskiego, 19 — późnokredowe i paleogeńskie skały wulkaniczne i banatyty, 20 — Neogeńskie skały wulkaniczne, (a) występowanie wgłębne, 21 — Północna granica nasunięcia Fliszu Zewnętrznego i wewnętrznej części rowu przedgórskiego, 22 — Główne strefy uskokowe odmłodzone w neogenie, 23 — Oś ujemnej anomalii grawimetrycznej

The Foredeep

The Foredeep forms a continuous zone along the whole Carpathians. It is filled with sediments of Miocene-Pleistocene age, overlying Proterozoic, Paleozoic and Mesozoic rocks of the Foreland. The sedimentary fill of the foredeep, progressively younger and thicker towards the east and south, consists of predominantly detrital material, ranging in grain size from coarse conglomerates to marly clays with two evaporitic horizons. In the Western part of the Foredeep the sedimentary fill is 600 m thick and ends with Upper Badenian deposits, while in the south-eastern part it is c. 10 000 m thick and ends with Pleistocene sediments.

The autochtonous sedimentary fill of the outer zone of the Foredeep is faulted and overriden by the inner zone of the Foredeep, and by the Outer Flysch Belt. The inner zone of the Foredeep consists of a belt of folded sediments of Miocene age in front of the nappes of the Outer Flysch Belt in the Eastern part of the Carpathians. In the innermost part of this zone the Lower Miocene sediments are overlying concordantly the Cretaceous-Paleogene flysch, and are involved in the nappe structure. In the Western Carpathians the zone of folded Miocene rocks at the front of the Outer Flysch Belt is discontinuous and poorly developed. Late folding, posterior to the formation of nappes, affected the Pleistocene sediments in the south-eastern part of the Foredeep.

The Outer Flysch Belt

The Outer Flysch Belt forms a continuous zone extending along the western and eastern part the Carpathians. The age of the Flysch sequence ranges from Tithonian to late Oligocene — Early Miocene. Facial diversity is conspicuous, especially during the late Cretaceous and Eocene, while Early Cretaceous and Oligocene facies are more uniform. The major part of the flysch sequence consists of detrital rocks whose material was supplied from marginal and intrabasinal tectonic lands (cordilleras) intermittently uplifted and eroded.

The tectonic structure of the Outer Flysch Belt was formed during two tectogenetic phases in Miocene time. Earlier, Late Senonian deformation affected the innermost part of the Outer Flysch Belt in the Eastern Carpathians. The folded flysch rocks form several *décollement nappes* thrust outwards, i.e. to the north — west, north, north — east and east. The style of thrusting and folding differs markedly in the western and the eastern part of the Outer Flysch Belt. This problem is discussed below.

The Pieniny Klippen Belt

The Pieniny Klippen Belt is flanking from the south the western part of the Outer Flysch Belt and the north-western extremity of the Inner East Carpathians. This very narrow and elongated unit (a few km to maximum 20 km wide, more then 600 km long) has an extremely complex tectonic structure. No crystalline basement rocks are involved in the structure of the Pieniny Klippen Belt. The marine sedimentary sequence comprised Jurassic and Early Cretaceous limestones and marls with radiolarite cherts in the Callovian and Oxfordian, followed by late Cretaceous marls and Senonian (locally Albian) — Paleogene flysch. Triassic rocks are presented only locally.

The main tectogenetic phase is post-Paleogene, while Early Cretaceous and Late Cretaceous deformations are recognized, partly by indirect evidence (Birkenmajer 1970, Mišik 1977, Marschalko 1979).

The Central West Carpathians

The Central West Carpathians are defined here as consisting of the Tatrides and Veporides crystalline basement units with their own sedimentary cover, and the pile of rootless nappes thrust from the south, and with the exception of the Križna nappe, displaying Austroalpine facial affinities. The full sequence of the sedimentary cover comprise continental — volcanogenic Permian — marine Triassic — Jurassic — Cretaceous, up to Cenomanian inclusive. The crystalline basement consists of Variscan granitoids. The tectogenesis is pre-Senonian. The post-tectogenetic cover consists of Senonian and Paleogene rocks, the latter developed mostly in nummulitic limestone and flysch facies.

The Inner West Carpathians

Recent stratigraphic and facial studies summarized by Mock (1978) led to the definition of the Inner West Carpathians with a continuous marine sedimentary sequence comprising Late Paleozoic-Triassic rocks facially related to the southern Alps, and of the Silitsa nappe belonging to the Austroalpine facial domain, but thrust southward. The tectogenesis is pre--Senonian, but younger, probably post-Paleogene movements affected markedly this area (Mock 1978). The pre-Alpine basement consists of pre-

-Cambrian and Early Paleozoic parametamorphic rocks. Small granitoid plutons of Late Cretaceous age and coeval epizonal metamorphizm is present in the north-eastern part of this unit (Varček 1973). The Lubenik suture is a major tectonic line separating the Central West Carpathians from the Inner West Carpathians (Andrusov 1975, Mock 1978).

The Hungarian Mts. and the basement of the Pannonian Basin

The Pannonian Basin, situated in the centre of the Carpathian composite arc has a key position for the understanding of the structural evolution of the Carpathians. Recent progress in the investigations of the structure of the basement of the Pannonian Basin (Balogh and Korossy 1968, 1974, Wein 1969, 1975, Szadeczky-Kardoss 1974, Juhasz and Vass 1974, Zelenka 1973, Geczy 1973) can be summarized as follows: — the basement of the Pannonian Basin is composed of pre-Cambrian and Early Paleozoic cristalline rocks, covered by Late Paleozoic and Mesozoic marine sedimentary rocks. Marine sedimentation persisted since the Late Paleozoic into Mesozoic time with local breaks in continuity. The area was subject to pre-Senonian folding. The Senonian is post-tectogenetic with predominant marl and limestone facies. Paleogene and Neogene faulting strongly affected the area. Horsts formed during Neogene faulting, are constituting the Bakony-Vertes, Mecsek and Villany ranges of the Hungarian Mts., in which Mesozoic and Paleozoic rocks are exposed. North of the Mecsek-Villany area most of the fault-bound structural units are narrow and elongated in the ENE-WSW direction.

The Szolnok Flysch Belt

The Szolnok Flysch Belt extending the Debrecen-Szolnok-Seged area in the basement of the Pannonian Basin is a structural unit of major importance, separating the inner parts of the Western and the Eastern Carpathians. This elongated basin is filled with a folded sequence of Senonian-Oligocene flyschoid rocks. An unconformity exist between the Cretaceous and the Paleogene. The Senonian is underlain by basic volcanic rocks intercalating with radiolarites of Jurassic and Early Cretaceous age.

The Inner East Carpathians

The Inner East Carpathians are continuous to the inner margin of the eastern part of the Outer Flysch Belt. They consists of several nappes thrust eastward, composed of continental-volcanogenic Permian — marine Triassic — Jurassic — Early Cretaceous sedimentary cover, and pre--Alpine crystalline basement present in some of the nappes. The lower group of Bucovinian nappes is overriden by the higher Transilvanian nappes which contain ophiolites, present also in the uppermost Bucovinian nappes (Sandulescu 1975). The tectogenesis is pre-Cenomanian. The post-tectogenetic cover of Cenomanian-Oligocene age begins with conglomerates, followed by Late Cretaceous red marls and limestones, and Paleogene flysch.

The Black Flysch and Ceahlau flysch nappes present along the Eastern Carpathians were subject to Late Cretaceous deformations and usually are considered as a part of the Inner East Carpathians. However from the point of view of basin evolution they are closely related with the Outer Flysch Belt.

The Northern Apuseni Mts.

The Northern Apuseni Mts. comprise the Bihor Massif composed of a pre-Alpine crystalline basement and a sedimentary cover including continental-volcanogenic Permian — marine — Triassic — Jurassic — Cretaceous (up to Upper Turonian), with epizodic continental sedimentation is Early Jurassic and Early Cretaceous. The Bihor Massif is overriden by a pile of nappes thrust northward, consisting of pre-Alpine crystalline basement and a sedimentary cover covering a similar stratigraphic range as in the Bihor. The emplacement of nappes is pre-Senonian. The post-tectogenetic cover begins with Lower Senonian.

The Southern Apuseni Mts.

The Southern Apuseni Mts. are composed of Jurassic ophiolites, and a Tithonian — Lower Cretaceous sedimentary cover. Sialic metamorphic basement is present locally. The tectogenesis is polyphase, with pre--Cenomanian and Late Senonian deformations. The post tectogenic cover of Late Cretaceous age is developed partly in flysch facies.

The Southern Carpathians

The Southern Carpathians consists of nappes thrust southward on the autochtonous Danubian domain. The sequence of continental-volcanogenic Permian and marine Triassic — Jurassic — Lower Cretaceous (with a major lacune in Late Triassic and continental Lower Jurassic), and large slices of crystalline and sedimentary pre-Alpine basement are forming the nappes emplaced during the Late Senonian tectogenetic phase. Earlier deformations attributed to pre-Cenomanian movements ars also present.

The Neogene basins and the volcanic belts of the Carpathians will be described later.

THE PROBLEMS OF TECTONIC DEFORMATIONS IN CONTINENT-CONTINENT COLLISION

As indicated by McKenzie (1977) in the case of continent — continent collision the zone of deformations is vide and has diffused boundaries in comparison with oceanic plate boundaries. Tapponier (1977a, 1977b) proposed an approach based on deformation of a continuous medium in the analysis of continent — continent collision. This approach resulted in the recognition of the importance of horizontal translations along strike-slip faults, and of formation of secondary subduction zones in unconstrained directions of a blocked system of colliding continental plates.

This approach is bringing useful results in the analysis of the structural evolution in the Carpathians.

THE GENERAL PALINSPASTIC RECONSTRUCTION OF THE CARPATHIANS BEFORE THE NEOGENE TECTOGENESIS

The structural and paleogeographic analysis of the Carpathian units deformed in the post-Paleogene tectogenesis (Unrug 1979) provided a general palinspastic reconstruction for Paleogene time, i.e. for the period between the major tectogenetic phases of the Cretaceous which deformed the inner zones of the Carpathians, and the Neogene tectogenetic phases which deformed the outer zones of the Carpathians (Fig. 3).

This reconstruction is based upon the following premises: — the margin of the Foreland platform — and the European plate — is indicated (Ślączka 1975, Airinei 1977) by the axis of the negative gravimetric anomaly extending along the Outer Flysch Belt and the Foredeep of the Carpathians (Fig. 2);

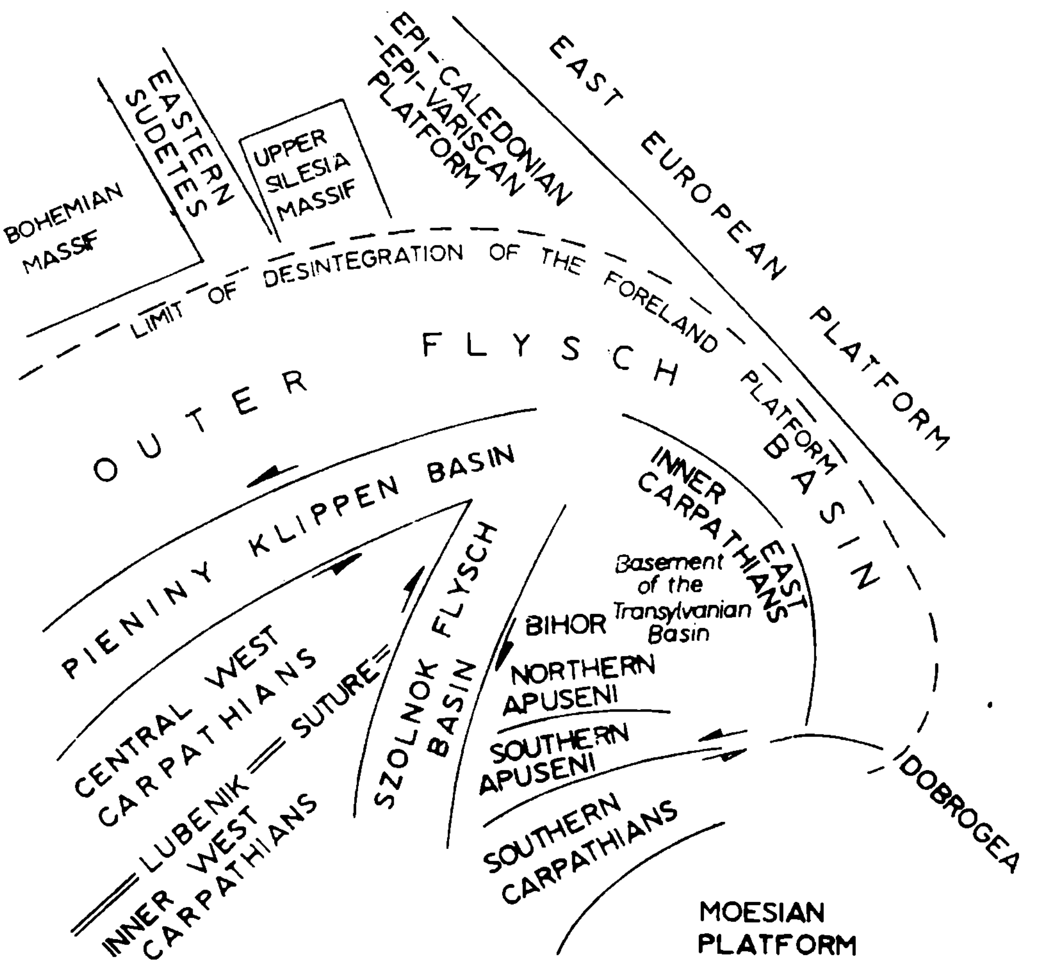

Fig. 3. Palinspastic sketch-map of the Carpathians before the Neogene tectogenesis

Fig. 3. Szkic palinspastyczny Karpat przed tektogenezą neogeńską

— the arcuate shape of the Carpathians is caused by the curvature of the margin of the Foreland platform controlled by the edges of the rigid East European Platform and of the Bohemian massif;
— in the Western Carpathians the Outer Flysch basin and the Pieniny Klippen basin were situated between the margin of the Foreland platform and the inner zone deformed in Cretaceous tectogenetic Phases;
— in the Eastern Carpathians only the Outer Flysch basin separated the margin of the Foreland platform and the inner zones deformed in Cretaceous tectogenetic phases;
— dextral strike-slip displacement along fault zones bounding the Szolnok Flysch Belt, and sinistral strike-slip displacements along fault zones bounding the Pieniny Klippen Belt occurred during the Neogene tectogenesis.

The palinspastic reconstruction for the Carpathians before the Neogene tectogenesis forms a reference frame for the analysis of the Cretaceous tectogenesis.

LATE JURASSIC — EARLY CRETACEOUS PALEOGEOGRAPHY OF THE CARPATHIANS

The major translations during the Neogene tectogenesis mentioned above, reduce the accuracy of the paleogeographic reconstruction and kinematic analysis for the Cretaceous tectogenesis of the Carpathians. However, recent results of facial and tectonic studies (Andrusov 1975, Mišik 1977, 1978, Mock 1978, Bleahu 1976a, 1976b, Lupu 1976a, 1976b, Patrulius 1976, Savu 1976, Juhasz and Vass 1974, Książkiewicz 1977) create the necessity of revision of the older opinions.

The palinspastic and paleogeographic situation of the Carpathians must be analysed in connection with adjoining segments of the Alpine fold of Europe. The following discussion is based upon the generally accepted paleogeographic scheme for Alps (Trümpy 1975, Lemoine 1978) and on the model of Dimitrijević (1974) for the Dinarides area.

The proposed paleogeographic model for the Carpathians in the Late Jurassic — Early Cretaceous is based upon two premises:
— the continuity of the paraoceanic tract from the Pennine belt of the Alps along the Carpathians into the Dinarides area (the term para-oceanic is used here in the meaning of Trümpy (1975), for an area with partly oceanic, partly continental crust with reduced thickness);
— the major facial changes and discontinuity of isopic zones and structural units along the Carpathians.

The Western Carpathians

In the model presented in Fig. 4 it is proposed that the Pennine zone of the Alps is prolonged into the Križna basin and the Vepor massif of the West Carpathians, on account of the metamorphism of the Mesozoic cover of the Vepor zone (a unique situation in the Central West Carpathians) and the presence of deep water Jurassic sediments in the Križna succession. The Tatride region, situated between the deep water zones of the Pieniny Klippen Basin in the north and the Križna basin in the south, had a similar paleogeographic position as the Briançonnais zone. The similitudes of facial development of the Mesozoic succession in the Briançonnais and the Tatric zones were recognized long ago (Kotański 1959, Debelmas 1960); but the conclusions resulting from these similitudes were not formulated hitherto.

The Austroalpine zone continues from the Eastern Alps into the Central West Carpathians, while the Southern Alps facial zones are prolonged into the Inner West Carpathians and the Hungarian Mts. (Wein 1969).

In this model the Lubenik suture of the Carpathians (Andrusov 1965) in considered as the prolongation of the Gail line of the Alps (Kozur and

Mock 1973, Mock 1978). The similitudes of the Austroalpine — South Alpine boundary to the Central West Carpathians — Inner West Carpathians boundary comprise the following features:

-- the Austroalpine facial development of the Carpathian nappes thrust northward from the Lubenik suture;

-- the presence of Alpine metamorphism and of plutonic intrusions (Varček 1973). Their position is generally similar to that of the plutons associated with the Periadriatic line, although the age of the Inner West Carpathian pluton may be older than that of the plutons of the Periadriatic line. High — pressure low-temperature metamorphism is indicated by the presence of glaucophanites (Kamenicky 1957);

— the southern vergency of the structural units situated south of the Lubenik suture, including the Silitsa nappe thrust southward (Kozur and Mock 1973, Andrusov 1975, Mock 1978), and the folds of the Bükk — Uppony region.

The northward thrust of the Inner West Carpathians on the Vepor units along the Lubenik suture and the formation of north-vergent structures in the Meliata group and the Silitsa nappe are considered by Mock (1978) as formed during late Alpine, possibly Miocene movements. This idea fits well the tectonic mobility of the inner parts of the West Carpathians during the Neogene tectogenesis discovered by Hungarian geologists (Wein 1969, 1975, Zelenka 1973).

The problem of the Pieniny Klippen Basin will be considered now. A great amount of evidence on the evolution of this basin was obtained recently by Slovakian geologists, by petrographic and microfacial studies of pebbles of exotic rocks in conglomerates of the Cretaceous Flysch of the Pieniny Klippen Belt, derived from source areas situated north of the Inner West Carpathians and south of the Pieniny Klippen Belt.

This region subsided rapidly already in the Triassic, since a sequence of Triassic carbonate rocks deposited in deeper water than the coeval rocks of the Tatric block of the Central West Carpathians was determinated by microfacial studies of exotic carbonate pebbles by Mišik (1977, 1978).

In the Pieniny Klippen Belt deep water pelagic facies predominate in the Jurassic and Lower Cretaceous sequence. The region situated south of the Pieniny Klippen Belt and north of the Central West Carpathians was the site of collision of lithospheric plates active in Late Jurassic and Early Cretaceous time, as indicated by the association of radiometrically dated plutonic and volcanic rocks (Marschalko 1979) extending far to the east (Lomize 1968). Pre-Albian deformation of this region resulted in uplift of source areas for the Cretaceous Flysch sediments of the Pieniny Klippen Belt. These source areas were subsequently overriden by the Inner West Carpathians during the Late Cretaceous and Paleocene (Marschalko 1979)

It is generally accepted that the Pieniny Klippen of the Carpathians are corresponding to the St. Veit Klippen of the Eastern Alps, which are situated south of the Helvetic-Ultrahelvetic facial zone. In the Alps the Pennine flysch is strongly developed south of the St. Veit Klippen, and suggestions were expressed (Tollman 1965, Lemoine 1978) on the obliquity of the Alpino-Carpathian flysch basin, which had a more internal position in the Alps-south of the St. Veit Klippen, and a more external position in the Carpathians-north of the Pieniny Klippen. Yet, the Pieniny Klippen Belt has its own flysch succession, locally beginning already in the Albian (Marschalko 1973). Also the Tatric region has a Cenomanian — Turonian flysch, mostly eroded and preserved only locally. Certainly the Carpathians, similary as the Alps, have several different flysch basins. The hypothesis on the reasons of the exceptionally strong development of the Outer Carpathian Flysch with comparison of the adjoining Alpine ranges will be presented below.

The Eastern Carpathians

Turning now to the Eastern Carpathians, it is proposed that east of the Central West Carpathians the para-oceanic tract was shifted northward by a transform fault, and continued in the sedimentary basin of the Transilvanian nappes of the Inner East Carpathians (Fig. 4). This conclusion is based upon the occurrence of ophiolites in the Transilvanian nappes, in the olistostromes of the Bucovinian nappe and in the Maramureș Klippen (Kruglow 1974) which form part of the Bucovinian nappe (Sandulescu 1975). This eastern segment of the para-oceanic tract was probably divided by another transform fault into the northern part of the Inner East Carpathians and the southern part represented by the ophiolites of the Southern Apuseni Mts.

The transform fault between the West Carpathian and the East Carpathians segment of the para-oceanic tract was situated probably along the future Szolnok Flysch basin. It may be indicated by the presence of the basic rocks in the basement of the Szolnok Flysch Basin (Juhasz and Vass 1974). The importance of intrusions of mafic material along transform faults and other deep-seated fractures was indicated by Thomson and Melson (1972) in the oceanic areas. The existence of such intrusions along para-oceanic or ensialic transform faults is also probable.

The Southern prolongation of the South Apuseni ophiolite zone is situated probably rather in the Ophiolite Belt of the Dinarides than in the Vardar zone as proposed by Bleahu (1976a). According to the model of Dimitrijević (1974) the Vardar zone was formed by dextral transcurrent faulting during the Cretaceous. Moreover the opening of the oceanic

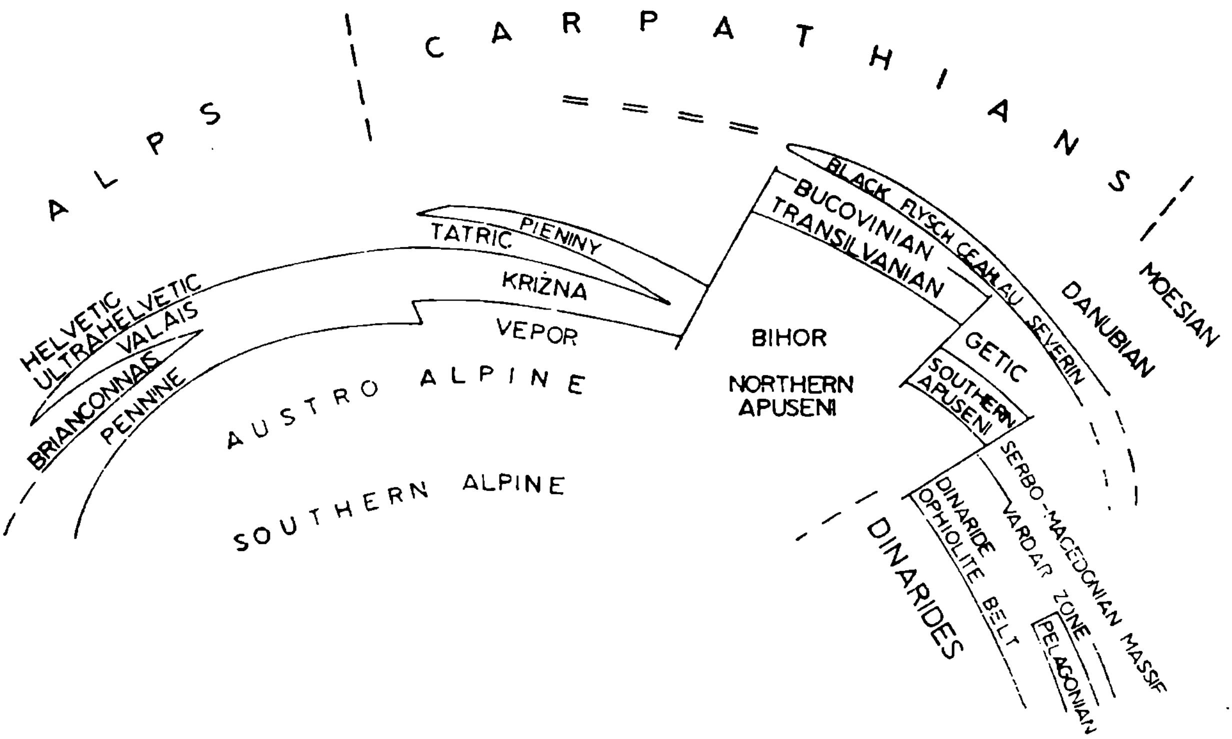

Fig. 4. Paleogeographic sketch-map of the Alpine-Carpathian-Dinaride zone in Late Jurassic/Early Cretaceous time

Fig. 4. Szkic paleogeograficzny obszaru alpejsko-karpacko-dynarskiego na przełomie jury i kredy

tract along the ophiolite zone of the Dinarides in the Ladinian (Dimitrijević 1974) corresponds to the age of submarine basic effusions in the Transilvanian nappes of the East Carpathian segment.

THE EVOLUTION OF THE OUTER FLYSCH BASIN

The innermost nappes of the Outer Flysch Belt in the Eastern Carpathians (the Black Flysch nappe and the Ceahlau nappe) contain Tithonian basic effusive rocks in the black flyschoid Sinaia Beds. This zone continues in the Southern Carpathians in the Severin nappe[1].

It is speculated, that the eastern part of the Outer Flysch Basin developed as a marginal basin related with an east dipping subduction plane in the Inner East Carpathians. Spreading was short — lived in this basin and restricted to its inner part. In the Western part of the Outer Flysch Basin there are no indications of development of a spreading zone. The mobility of the floor of the Outer Flysch Basin witnessed by the development of the rapidly subsiding troughs filled with turbidite sediments

[1] The tectonic position of the Deli Jovan ophiolite massif of the Southern Carpathians with its typical section of oceanic crust (Grubić 1974) does not permit a definite correlation with the sedimentary basin of the Sinaia beds of the Severin nappe.

and the rising intrabasinal tectonic lands could be related with the incipient development of a marginal basin without the formation of a definite spreading axis. The high alcali subvolcanic intrusions and submarine effusions of teschenitic lavas in the western part of the Outer Flysch Basin may also be regarded as an indication of the evolution of the basin into a marginal basin type.

The tectonic lands (cordilleras) of the Outer Flysch Basin (Książkiewicz 1965) were composed of continental crust. The morphology of the basin floor reconstructed for the period of Flysch sedimentation reminds the Basin and Range province of North America. It may suggest a plunging of the incipient rifting and spreading zone of the eastern segment of the Outer Flysch Basin under the continental crust of the western part of the basin, causing its desintegration and exceptional mobility — by analogy with the Bay of California and the Basin and Range Province.

The limit of the desintegration of the Foreland platform and of the newly formed Outer Flysch Basin is evidently structurally controlled. The north-eastern margin of the Outer Flysch Basin is markedly parallel to the Teisseyre-Tornquist line forming the south-western border of the East European Platform. The north-western margin of the Outer Flysch Basin is parallel to the south-eastern margin of the Bohemian Massif. Between these two rigid areas the northern boundary of the Outher Flysch Belt cut across various structural units, generally belonging to the Epi-Hercynian and Epi-Caledonian platforms.

Petrographic and microfacial studies of exotic blocks and pebbles present in conglomeratic membres of the Outer Flysch (Książkiewicz 1954, Malik 1978, Wieser 1949) provided ample evidence of the evolution of the Outer Flysch Basin. The pre-Alpine basement of the basin consisted of Precambrian crystalline rocks, covered by Paleozoic sedimentary rocks, probably deformed by Caledonian and Variscan orogenic movements. During Permian, Triassic and Early Jurassic time the area was emerged. The marine transgression came in the Callovian and a shallow-water carbonate sedimentation was installed. The carbonate platform extending into the Foreland persisted till the end of the Jurassic. The desintegration of the carbonate platform with continuity of sedimentation in the Outer Flysch Basin was accompanied by a regression in the Foreland area. The desintegration of the carbonate platform is evidenced by the appearance of a late Tithonian calcareous flysch (Książkiewicz 1962) in the incipient basin, and of Berriasian detrital limestones on the carbonate platform (Malik 1979). Rapid subsidence of the basin began, accompanied by the uplift of marginal and intrabasinal tectonic lands which supplied detrital material to the lithoclastic flysch sandstones and conglomerates.

Till the Cenomanian the sedimentary conditions were relatively uniform over the major part of the Outer Flysch Basin with black-coloured shales predominating (Książkiewicz 1962, Danysz et al. 1963, Sandulescu

1975). The Late Cretaceous and Paleogene were a period of intense vertical movements of blocks of the basin floor and of tectonic lands, resulting in a major facial diversity. Only the terminal period of flysch sedimentation (Late Oligocene — Earliest Miocene) was marked by a more uniform facial development.

THE CRETACEOUS TECTOGENESIS IN THE CARPATHIANS

The Cretaceous tectogenesis in the Carpathians was polyphase, with major pre-Albian, pre-Cenomanian, pre-Senonian and Late Senonian deformations. The kinematics of these deformations involved both north-south and east-west crustal shortening. In the Western Carpathians the deformations are pre-Albian, pre-Senonian and Late Senonian, while in the Eastern Carpathians the deformations are pre-Cenomanian, pre-Senonian and Late Senonian.

In the Western Carpathians the pre-Albian deformation is evidenced indirectly by the uplift of the source area for the coarse clastic of the Albian Flysch in the Pieniny Klippen Belt, yielding a rich petrographic assemblage of more than one hundred rock types (Mišik 1978). This deformation affected the outermost part of the Central West Carpathians Basin an the continuous innermost part of the Pieniny Klippen Basin. In the Tatric crustal block the pre-Albian tectogenetic phase active further north in registered by vertical movements and penaccordance of the Albian and the Lower Cretaceous (Lafeld 1968, Bac-Moszaszwili et al. 1979).

The pre-Senonian phase is the most important in the Western Carpathians, since it produced the fold deformations and large scale thrusting of nappes in the Central West Carpathians, the Inner West Carpathians, in the Hungarian Mts. and parts of the basement of the Pannonian Basin.

The Late Senonian phase resulted in the overriding of the source areas for the Cretaceous flysch of the Pieniny Klippen Belt (Marschalko 1979) by the Central West Carpathians behaving as a rigid block. Vertical movements and possibly some deformation in the Pieniny Klippen Belt (Birkenmajer 1970, Książkiewicz 1972) were associated with this phase.

The pre-Senonian (pre-Gosau) deformation in the Eastern Alps and the Western Carpathians, although broadly synchronous, display important differences in these two regions. The nappes formed in the Austroalpine facial domain of the Western Carpathians (the Choč and Stražov nappes) were transported far to the north already in the pre-Senonian phase, and there were no Paleogene northward thrusting which is so important in the Eeastern Alps. This may indicate the reality of the transcurrent fault zone between the Eastern Alps and the Western Carpathians suggested by Chorowicz and Geyssant (1976).

The structure of the Central West Carpathians and the Inner West Carpathians with the "Lubenik fan" (Andrusov 1975) may be interpreted as a rootless rotation zone in the meaning of Roeder (1973) over a subduction zone dipping north, and then flipped to dip south. Such a interpretation would account for the active processes in the boundary zone of the Central West Carpathians and the Pieniny Klippen Belt in the Early Cretaceous, as deformation antithetic to the subduction zone.

In the Eastern Carpathians, the earliest, pre-Cenomanian deformation affecting the Inner East Carpathians is only slightly younger than the first deformation in the Western Carpathians. The Bucovinian nappes of the Inner East Carpathians are all basement nappes and this may suggest that they were sheared along relatively steeply dipping planes antithetic to a subduction zone (Roeder 1973) situated in the sedimentary basin of the Transilvanian nappes and dipping under the basin of the Bucovinian nappes.

During the pre-Cenomanian deformation the Southern Apuseni segment of the para-oceanic tract began to rotate, possibly between a pair of transcurrent faults, breaking the continuity with the Inner East Carpathians segment.

The pre-Senonian deformation phase resulted in the Eastern Carpathians in the formation of the pile of nappes of the Northern Apuseni Mts. The interpretation of this thrusting is difficult, but it may be related with a clockwise rotation of the southern Apuseni and of the contiguous Getic domain.

The differences in structural evolution of the Western and the Eastern Carpathians during the pre-Senonian phase of deformation resulted in distension and formation of the Szolnok Flysch Basin in the boundary zone between these two regions.

The Late Senonian tectogenetic phase produced major thrusting in the Southern Carpathians, where the Danubian domain has been overriden by the Getic basement nappe. The Inner East Carpathians were pushed outwards on the Outer Flysch Basin forming the innermost flysch nappes (the Black Flysch nappe and the Ceahlau nappe) in front of them.

THE TECTONIC EVOLUTION OF THE CARPATHIANS DURING THE TERTIARY

During the Tertiary the tectonic evolution of the Carpathians continued along different lineas in the inner zones subject to Cretaceous tectogenesis and in the Outer zones deformed by the Neogene tectogenesis.

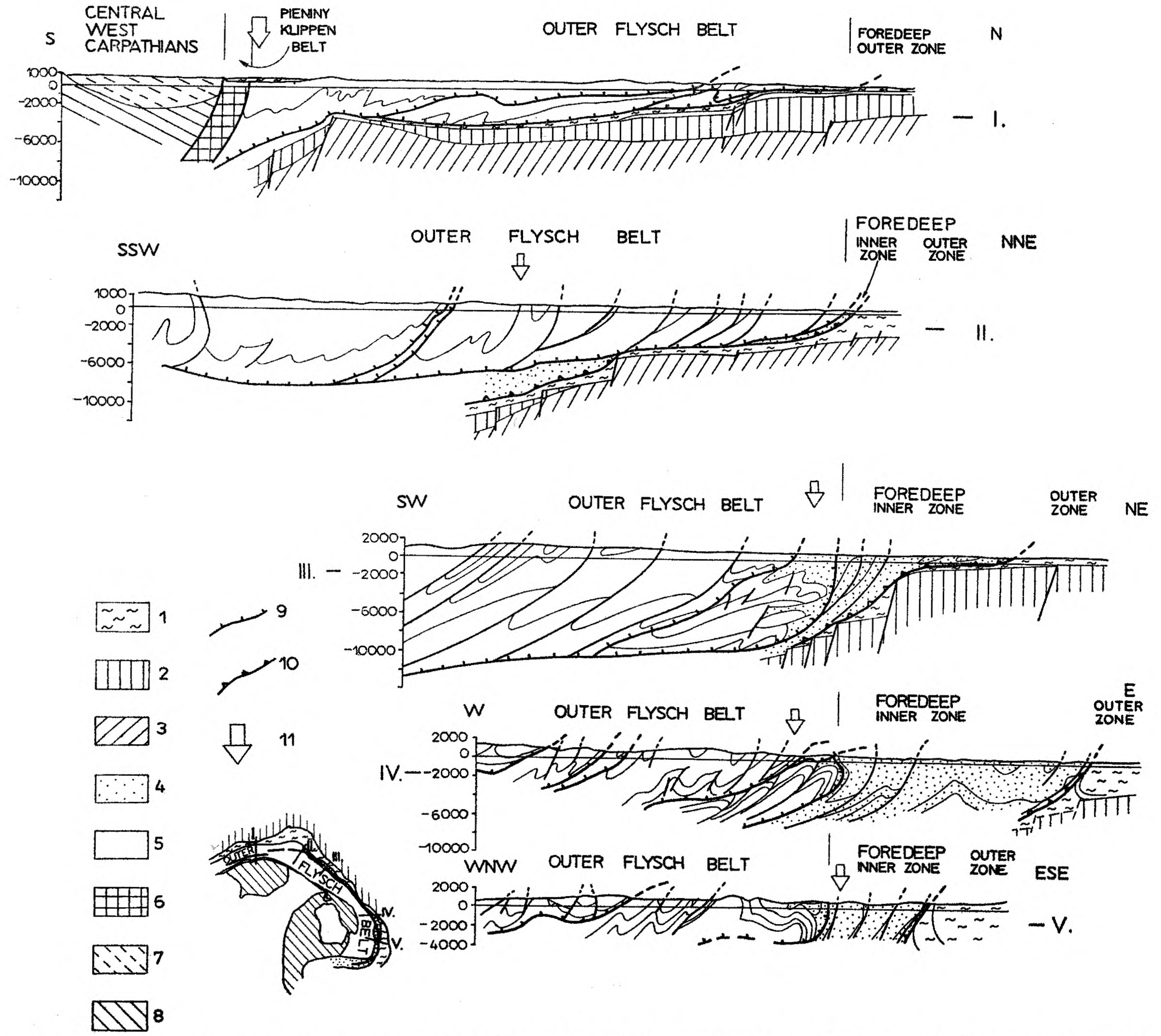

Fig. 5. Cross-sections showing the relation of the Outer Flysch Belt and the Inner zone in the Foredeep to the Foreland Platform. 1 — Neogene of the Outer zone of the Foredeep, 2 — Foreland Platform, sedimentary rocks, 3 — Foreland Platform, metamorphic basement, 4 — Inner Zone of the Foredeep, folded and thrust-faulted, 5 — Outer Flysch Belt, 6 — Pieniny Klippen Belt, 7 — Paleogene Flysch weakly folded, 8 — Central West Carpathians, 9 — Major overthrust planes in the Outer Flysch Belt, 10 — Overthrust plane of the Inner Zone of the Foredeep, 11 — Axis of the negative gravimetric anomaly. Inset: Location of the cross section along the Outer Flysch Belt. Cross-section I after Ślączka (1975), cros-section II after Wdowiarz (1976)

Fig. 5. Przekroje przedstawiające stosunek Fliszu Zewnętrznego i wewnętrznej strefy rowu przedgórskiego do platformowego przedgórza. 1 — Neogen rowu przedgórskiego, 2 — Platforma przedgórska, skały osadowe, 3 — Platforma przedgórska, podłoże metamorficzne, 4 — Wewnętrzna strefa rowu przedgórskiego, sfałdowana i nasunięta, 5 — Flisz Zewnętrzny, 6 — Pieniński Pas Skałkowy, 7 — Flisz Podhalański, 8 — Centralne Karpaty Zachodnie, 9 — Główne nasunięcia w obrębie Fliszu Zewnętrznego, 10 — Nasunięcie wewnętrznej strefy rowu przedgórskiego, 11 — oś ujemnej anomalii grawimetrycznej. Lokalizacja przekrojów podana na mapce. Przekrój I według Ślączki (1975), przekrój II według Wdowiarza (1976)

In the Paleogene the Hungarian Mts. and the Inner West Carpathians were subject to important thrust faulting (continuing from Latest Cretaceous time), accompanied by formation of normal and transcurrent faults. The majority of the faults are oriented NE-SW, and the thrusts have south-eastern vergency (Wein 1969, Zelenka 1973).

Sedimentation continued in Paleogene time in the Szolnok Flysch Basin and the adjacent Central West Carpathians and Maramureș Flysch basins, in the Pieniny Klippen Basin, and in the Outer Flysch Basin. In the latter, facial changes and thickness distribution of various lithostratigraphic units indicate a great mobility of the basin floor, and a general eastward and outward migration of the zones of subsidence (Książkiewicz 1956, Ślączka 1969, Sandulescu 1975).

The Neogene tectogenesis of the Carpathians deformed chiefly the external structural unite — the inner zone of the Foredeep and the Outer Flysch Belt in the Eastern Carpathians, and the Outer Flysch Belt and the Pieniny Klippen Belt in the Western Carpathians (Książkiewicz 1972). In the spite of the continuity of the Outer Flysch Belt the kinematics and age of deformation is different in the Western and the Eastern Carpathians. The large scale deformation in the outer zones of the Carpathians were accompanied by faulting and volcanism in the inner zones of the Carpathians.

In the Western Carpathians the deformation of the Outer Flysch Belt is characterized by the presence of several piled thin thrust sheets overriding flatly the Foredeep and the Foreland Platform (Fig. 5). The relation of this pile of thrust sheets to the edge of the Foreland Platform implies a clockwise rotation of the Outer Flysch Belt segment west of Przemyśl (Unrug 1979). Such a rotation explains kinematically the sharp Z-bend of the front of the Outer Flysch Belt — the so-called Przemyśl sigmoid. The deformation is polyphase, the last phase being of intra-Badenian age in the Western Carpathians. The Inner zones of the Carpathians situated inside the Outer Flysch Basin and the Pieniny Klippen Belt were also subject to clockwise rotation, as indicated by paleomagnetic data (Kotasek and Krs 1965) and confirmed their varying distance from the southern edge of the Foreland Platform. The consequences of this rotation are indicated by the distribution of the volcanic rocks in the inner regions of the Western Carpathians.

Contrary to the current opinion, the Neogene volcanic rocks are forming not one, but two arcs within the inner regions of the Carpathians. One of these arcs is related with the Western Carpathians and comprise the Prešov — Tokay — Matra — Börzsony belt and the elongated zone of subsurface occurrences of volcanic rocks in the Great Pannonian Basin extending parallel to the Balaton Fault zone. The Central Slovakia volcanic massif is lying in the centre of this volcanic arc. The prolongation of the West Carpathians volcanic arc is intersecting the front of the

Outer Flysch Belt in the region of the Przemyśl sigmoid. It is proposed here, that distensional fractures related with the clockwise rotation of the Central West Carpathians and Inner West Carpathians determined the location of the volcanic effusions and eruptions, which according to Lexa and Koneczny (1974) range in age from the Eggenburgien to the Pliocene.

The second volcanic arc, situated in the East Carpathians and comprising the Vihorlat — Gutin — Gutii — Calimani — Hargitha ranges is younger, representing the Badenian — Pliocene interval (Lexa and Koneczny 1974). The Neogene volcanics of the Southern Apuseni Mts. are situated in the centre of this arc.

It is worth noting that important ore mineralisations are related with the volcanic bodies situated in the centres of the two Carpathians volcanic arcs; the Banska Stiavnica and Matra Mts. areas in the Western Carpathians arc, and the Southern Apuseni area in the Eastern Carpathians arc. This problem merits a separate discussion and is not considered here in detail.

In the Eastern Carpathians the relation of the Outer Flysch Belt to the edge of the Foreland Platform and the style of deformation is completely different from that in the Western Carpathians. The Outer Flysch Belt is not thrust over the Foreland Platform but is butting against it. Only the folded inner zone of the Foredeep is partly thrust over the steep step-faulted edge of the Foreland Platform. Imbricated folds with steeply dipping axial planes are characteristic for the style of deformations of the Outer Flysch Belt in the Eastern Carpathians (Fig. 5).

The last phase of deformation associated with thrusting is intra-Sarmatian, while the folding of Pliocene and Pleistocene sediments of the Foredeep is at least partly related to halokinetic movements.

These differences is structure and style of deformation of the western and the eastern part of the Outer Flysch Belt and of the Foredeep provide evidence of different kinematics associated with the Neogene tectogenesis in the discussed two parts of the Carpathians. In the Western Carpathians a major transcurrent displacement of the inner zones of the Carpathians obliquely to the edge of the Foreland Platform along the Pieniny Klippen Belt (Unrug 1979) resulted in shearing along the Pieniny Klippen Belt responsible for the peculiarites of the tectonic structure of this unit, clockwise rotation of both the inner and the outer zones of the Carpathians (Unrug 1980) and thrusting of the sedimentary fill of the Outer Flysch basin as thin thrust sheets on the Foreland Platform. Instead, in the Eastern Carpathians there was no strike-slip displacement between the inner and the outer Carpathians and no rotation. In consequence the Outer Flysch Belt did not override the Foreland Platform.

Lexa and Koneczny (1974) provided convincing evidence disproving the relation between a subduction in the outer zones of the Carpathians and the generation of the andesitic magmas of the volcanic ranges of the Carpathians. According to these authors the location of the volcanic ranges is related to subsidence of the Neogene basins along fracture lines caused by a diapiric uprise of the mantle (Stegna 1974) an the generation of magmas is related to the diapiric mantle uprise itself. In the western Carpathians the location of the fractures controlling the position of the eruptive centres was probably associated with a general clockwise rotation of the whole western Carpathians in relation to the Foreland Platform.

CONCLUDING REMARKS

The preceding discussion leads to the conclusion, that subduction of para-oceanic crust and accretion at active plate margins was associated with the Cretaceous tectogenesis in the Carpathians. The deformation of the Outer Flysch Belt during the Neogene tectogenesis was caused by the overriding of the marginal basin of the Outer Flysch by the inner zones of the Carpathians. The sedimentary fill of the Outer Flysch Basin was pushed in front of the overriding (and in the Western Carpathians also rotating) blocks of the internal parts of the Carpathians, while the basement of the Outer Flysch Basin, possibly consisting of para-oceanic crust and relatively heavy, was partly engulfed beneath the overriding blocks. Gravity data (Airinei 1977) suggest that the active subduction zone in Vrancea area is a local structural feature, which can be possibly related with an unconstrained direction in a blocked system of colliding lithospheric plates, but should not be extrapolated along the Outer Flysch Belt.

The difference in structure of the western and the eastern part of the Outer Flysch Belt and the Neogene deformation on of the Szolnok Flysch Belt indicate the importance of the Kinematic activity of the southern plate and its accretion zone formed during Cretaceous tectogenesis in producing the deformation of the outer zones of the Carpathians during the Neogene collision with the northern European plate.

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STRESZCZENIE

Łuk karpacki wykazuje zróżnicowanie stylu tektonicznego w swej części zachodniej i wschodniej. Jednostkami ciągnącymi się na całej długości łuku karpackiego są: przedgórze heterogeniczne i składające się z jednostek różnego wieku, rów przedgórski z wewnętrzną strefą sfałdowaną i nasuniętą na część zewnętrzną oraz flisz Karpat Zewnętrznych. W Karpatach Zachodnich w bardziej wewnętrznej pozycji występują: Pieniński Pas Skałkowy, Centralne Karpaty Zachodnie, Wewnętrzne Karpaty Zachodnie i Góry Węgierskie. W Karpatach wschodnich występują w pozycji wewnętrznej następujące jednostki: Wewnętrzne Karpaty Wschodnie, Północne Apuseni, Południowe Apuseni i Karpaty Południowe.

Centralne Karpaty Zachodnie obejmują jednostki tatrzańskie i veporskie znajdujące się na północ od linii tektonicznej Lubenika. Wewnętrzne Karpaty Zachodnie obejmują obszar na południe od tej linii tektonicznej, to jest jednostkę gemerską wraz z płaszczowiną silicką o wergencji południowej. Góry Węgierskie są zrębowymi wypiętrzeniami podłoża Basenu Pannońskiego.

Pomiędzy jednostkami zachodniokarpackimi i wschodniokarpackimi występuje w podłożu basenu panońskiego flisz Szolnok, sfałdowany w neogenie. Do jego części północnej przylegają baseny fliszu Centralnych Karpat Zachodnich i fliszu Marmaroskiego, nieznacznie tylko zdeformowane w wyniku zróżnicowanych ruchów pionowych ich podłoża skonsolidowanego podczas kredowych faz tektogenetycznych.

Neogeńskie skały wulkaniczne Karpat tworzą dwa łuki. Łuk związany z Karpatami Zachodnimi obejmuje pasma: Prešov — Tokaj, Matra, oraz wgłębne wystąpienia skał wulkanicznych na południe od gór Bükk i Bakony-Vertes. W środku tego łuku znajduje się masyw wulkaniczny Słowacji środkowej. Łuk związany z Karpatami Wschodnimi obejmuje pasma Vihorlat, Gutin — Gutii i Calimani — Hargitha. W środku tego łuku znajdują się wystąpienia neogeńskich skał wulkanicznych w Południowym Apuseni (fig. 1, fig. 2).

Północną granicę zasięgu mobilnego podłoża basenów sedymentacyjnych zewnętrznych jednostek karpackich stanowi krawędź platformowego przedgórza, wyznaczona przez oś ujemnej anomalii grawimetrycznej. Krzywizna tej granicy wymuszona jest przez krawędzie sztywnych jednostek platformowego przedgórza: Masywu Czeskiego, Bloku Górnośląskiego i Platformy Wschodnioeuropejskiej (fig. 2, fig. 3).

Stosunek jednostek wewnętrznych, zdeformowanych i skonsolidowanych w kredowych fazach tektogenetycznych, do platformowego przedgórza był podczas paleogenu różny w zachodniej i we wschodniej części łuku karpackiego. Pomiędzy jednostkami wewnętrznymi Karpat Zachodnich a krawędzią platformowego przedgórza mieściły się baseny sedymentacyjne Fliszu Zewnętrznego i Pienińskiego Pasa Skałkowego, natomiast w Karpatach Wschodnich między krawędzią platformowego przedgórza a jednostkami wewnętrznymi rozciągał się tylko basen Fliszu Zewnętrznego. Podczas tektogenezy neogeńskiej przesunięcie mas wewnętrznokarpackich ku krawędzi platformowego przedgórza było zatem większe w Karpatach Zachodnich niż w Karpatach Wschodnich. W wyniku prawego przemieszczenia przesuwczego jednostek wewnętrznych Karpat Zachodnich względem jednostek wewnętrznych Karpat Wschodnich sfałdowany został flisz Szolnok.

Blok zachodniokarpackich jednostek wewnętrznych znajdował się przed tektogenezą neogeńską na zachód od swego obecnego położenia, w odległości stanowiącej sumę składowej wschodniej przemieszczenia bloku wschodniokarpackich jednostek wewnętrznych i zmniejszenia pierwotnej szerokości basenu fliszu Szolnok podczas tektogenezy neogeńskiej. Przemieszczenie bloku zachodniokarpackich jednostek wewnętrznych względem położonego na zewnątrz basenu Pienińskiego Pasa Skałkowego miało składową przesuwczą lewą. Struktura Pienińskiego Pasa Skałkowego jest związana z ruchem przesuwczym na uskoku ograniczającym od północy zachodniokarpacki blok jednostek wewnętrznych.

Przedstawiona tu rekonstrukcja palinspastyczna Karpat przed tektogenezą neogeńską (fig. 3) wyznacza ramy dla analizy paleogeografii i tektogenezy kredowej.

Model paleogeograficzny Karpat dla przełomu jury i kredy (fig. 4) oparty jest na założeniu ciągłości strefy paraoceanicznej skorupy od strefy pennińskiej Alp do obszaru Dynarydów. Model uwzględnia nieciągłość izopowych stref facjalnych w wewnętrznych jednostkach strukturalnych Karpat.

W modelu przyjęto, że przedłużeniem strefy pennińskiej Alp jest basen płaszczowiny kriżniańskiej i strefa veporska Centralnych Karpat Zachodnich, na podstawie słabego metamorfizmu pokrywy mezozoicznej w strefie veporskiej i obecności głębokowodnych facji w profilu płaszczowiny kriżniańskiej. Płytkowodna strefa tatrzańska zajmuje pozycję pomiędzy dwiema strefami głębokowodnymi — pienińską na północy i kriżniańską na południu, analogicznie do strefy briansońskiej Alp zachodnich.

Wschodnie przedłużenie austroalpejskiej strefy facjalnej obejmuje baseny sedymentacyjne wyższych jednostek tektonicznych Zachodnich Karpat Centralnych, oraz basen płaszczowiny silickiej, a wschodnie przedłużenie południowoalpejskiej strefy facjalnej obejmuje baseny sedymentacyjne Zachodnich Karpat Wewnętrznych i Gór Węgierskich, z ciągłością sedymentacji od górnego paleozoiku do mezozoiku. Linia tektoniczna Lubenika odpowiada linii peryadriatyckiej Alp. Alpy Wschodnie oddziela od Karpat strefa uskokowa Split — Karlovač — Wiedeń, wzdłuż której jednostki karpackie zostały przemieszczone ku północy dalej i w innym czasie, niż odpowiadające im jednostki alpejskie.

W modelu przyjęto, że wschodnie przedłużenie strefy paraoceanicznej skorupy przemieszczone było na północ na uskoku transformującym biegnącym na osi basenu fliszu Szolnok, gdzie pod sekwencją fliszową występują skały zasadowe oraz jurajsko-kredowe radiolaryty, w strefę basenu sedymentacyjnego płaszczowin transylwańskich Wschodnich Karpat Wewnętrznych, skąd poprzez Południowe Apuseni łączyło się ze strefą ofiolitową Dynarydów.

Basen Pienińskiego Pasa Skałkowego i jego południowe obrzeżenie będące miejscem wczesnokredowej aktywności wulkanicznej i plutonicznej, a zdeformowane w przedalbskiej fazie tektogenetycznej, interpretowane są jako basen marginalny. Również basen fliszu zewnętrznego, zawierający w późnojurajskich i wczesnokredowych osadach skały zasadowe interpretowany jest jako basen marginalny o podłożu rozbitym na ruchliwe bloki. Strefy subdukcji w pasie występowania skorupy paraoceanicznej zapadały ku północy i oba wymienione wyżej baseny marginalne rozwijają się na północ od stref subdukcji paraoceanicznej skorupy na karpackim odcinku Alpidów.

Podczas tektogenezy neogeńskiej zdeformowane zostały: Pieniński Pas Skałkowy, Flisz Zewnętrzny, oraz wewnętrzna część rowu przedgórskiego. Na obszarze Karpat Zachodnich strefy szwów tektonicznych rozdzielających jednostki wewnętrzne zostały odmłodzone już w paleogenie i przekształcone w strefy uskokowe z przemieszczeniami zrzutowymi, nasuwczymi i przesuwczymi, trwającymi również w neogenie. Nowo powstałe neogeńskie strefy uskokowe przecinają wewnętrzne jednostki strukturalne Karpat Zachodnich.

Zróżnicowanym przesunięciom bloków uskokowych w wewnętrznej strefie Karpat Zachodnich towarzyszyła rotacja zachodniej części Karpat Zachodnich zgodna z ruchem wskazówek zegara i nasunięcie płaszczowin fliszowych na platformowe przedgórze w obszarze na zachód od Przemyśla. Wygięcie czoła nasunięcia fliszowego w sigmoidzie przemyskiej interpretowane jest jako wynik tej rotacji. W obszarze na wschód od Przemyśla nasunięte masy fliszowe są przyparte do krawędzi platformowego przedgórza i tylko płaszczowina stebnicka, reprezentująca wewnętrzną, sfałdowaną strefę rowu przedgórskiego nasunięta jest częściowo na platformę przedgórską (fig. 5).

W Karpatach subdukcja paraoceanicznej skorupy związana była z wczesnymi fazami tektogenezy kredowej. Późne fazy tektogenezy kredowej i tektogeneza neogeńska miały raczej charakter kolizji kier kontynentalnej skorupy ze znacznymi translacjami na uskokach przesuwczych i z pochłonięciem podłoża marginalnych basenów Pienińskiego Pasa Skałkowego i Fliszu Zewnętrznego. Badania geologów węgierskich wykazały, że zróżnicowanie geochemiczne wulkanicznych skał karpackich wieku neogeńskiego nie daje się pogodzić z modelem subdukcji skorupy w podłożu basenu fliszowego pod strefy wewnętrznokarpackie. Geneza magm andezytowych i riolitowych wiązana jest z aktywnym diapirem płaszcza na obszarze basenu panońskiego, a lokalizacja neogeńskiego wulkanizmu wiązana jest z pęknięciem na granicach objętych subsydencją obszarów basenu panońskiego i basenu transylwańskiego. Zachodni łuk wulkaniczny związany jest też zapewne z rotacją zachodnich Karpat względem krawędzi platformowego przedgórza. Przedłużenie tego łuku ku północy przecina się z krawędzią platformowego przedgórza w rejonie sigmoidy przemyskiej, sugerując położenie osi tej rotacji.