

V edícii Západné Karpaty séria geológia doteraz vyšli tieto práce:

In the periodical Západné Karpaty (West Carpathians) so far the following articles have been published in the series Geologie:

Geológia 1:

Marschalko, R. — Mišík, M. — Kamenický, L.: Petrographie der Flysch-Konglomerate und Rekonstruktion ihrer Ursprungszonen (Paläogen der Klippenzone und der angrenzenden tektonischen Einheiten der Ostslowakei) — Petrografia flyšových konglomerátov a rekonštrukcia ich zdrojových zón (paleogén bradlového pásma a príslušných tektonických jednotiek východného Slovenska (117 strán, 13 obr. v texte, 41 fotogr. tab.)

Hanáček, J.: Nové poznatky o triase strážovského a chočského príkrovu v strážovskej hornatine — Information on the Triassic of the Strážov and Choč nappes in the Strážovská hornatina Mts. (26 strán, 1 fareb. príl., 9 fotogr. tab.)

Janáček, J.: Příspěvek k problematice průzkumu živic v předmiocenním podkladu jihozápadní části vídeňské pánve — Beitrag zur Problematik der Erkundung von Bitumina in dem vormiozänen Untergrund des südwestlichen Teiles des Wiener Beckens (30 strán)

Marušiak, I. — Lizoň, I.: Geotermické pole Západných Karpát — Geothermal field of the West Carpathians (25 strán, 5 obr. v texte)

Geológia 2:

Plančár, J. et al.: Geofyzikálna a geologická interpretácia tiažových a magnetických anomálií v Slovenskom rudohorí — Geophysical and geological interpretation of gravity and magnetic anomalies in the Slovenské rudohorie mountain range (138 strán, 53 obr. v texte, 17 fareb. príloh)

Vass, D.: Pribelské vrstvy, ich sedimentárne textúry a genéza — The Pribelce formation, its sedimentary structures and origin (53 strán, 16 obr. v texte, 10 fotogr. tab.)

Geológia 3:

Planderová, E.: Microflorizones in Neogene of Central Paratethys — Mikrofloristické zóny v neogéne centrálnej paratetydy (28 strán, 1 obr. v texte, 18 fotogr. tab.)

Nemčok, J.: Deformácie flyšových sedimentov ako odraz dynamiky podložia — Deformations of Flysch sediments as a reflection of dynamics of the basement (24 strán, 7 obr. v texte)

Jacko, S.: Litologicko-štruktúrna charakteristika centrálnej časti pásma Čiernej Hory — Lithological-structural characteristics of the central part in the Čierna hora region (22 strán, 1 obr. v texte)

Forgáč, J. — Pulec, M.: Vývoj vulkanosedimentárneho a vulkanogénneho komplexu v centrálnej časti stredoslovenských neovulkanitov — History of volcanic-sedimentary and volcanogenic complexes in middle part of Central Slovakian neovolcanic region (20 strán, 5 obr. v texte, 2 fotogr. tab.)

Tematická skupina 03/9  
Cena Kčs 30,—

ZK-Geol 4

1B 107 b

GEOLÓGIA 4 • ZÁPADNÉ KARPATY

1B 107 b

270/79

# ZÁPADNÉ KARPATY

CS ISSN 0036—1372

SÉRIA

*geológia 4*

Michal Maheľ

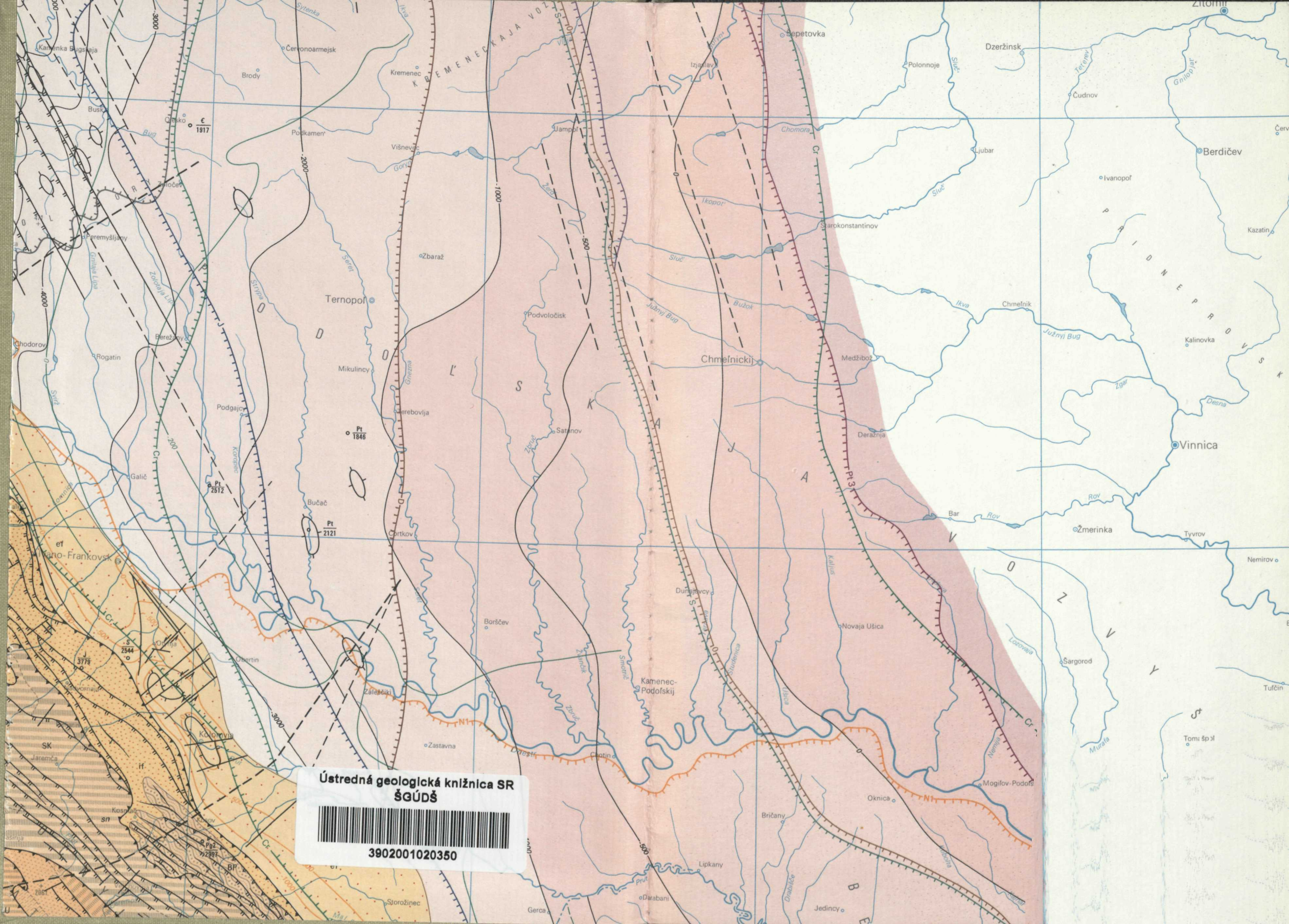
**Geotectonic position of magmatites  
in the Carpathians,  
Balkan and Dinarides**

GEOLOGICKÝ ÚSTAV DIÓNÝZA ŠTÚRA, BRATISLAVA 1978

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**Michal Mahef**

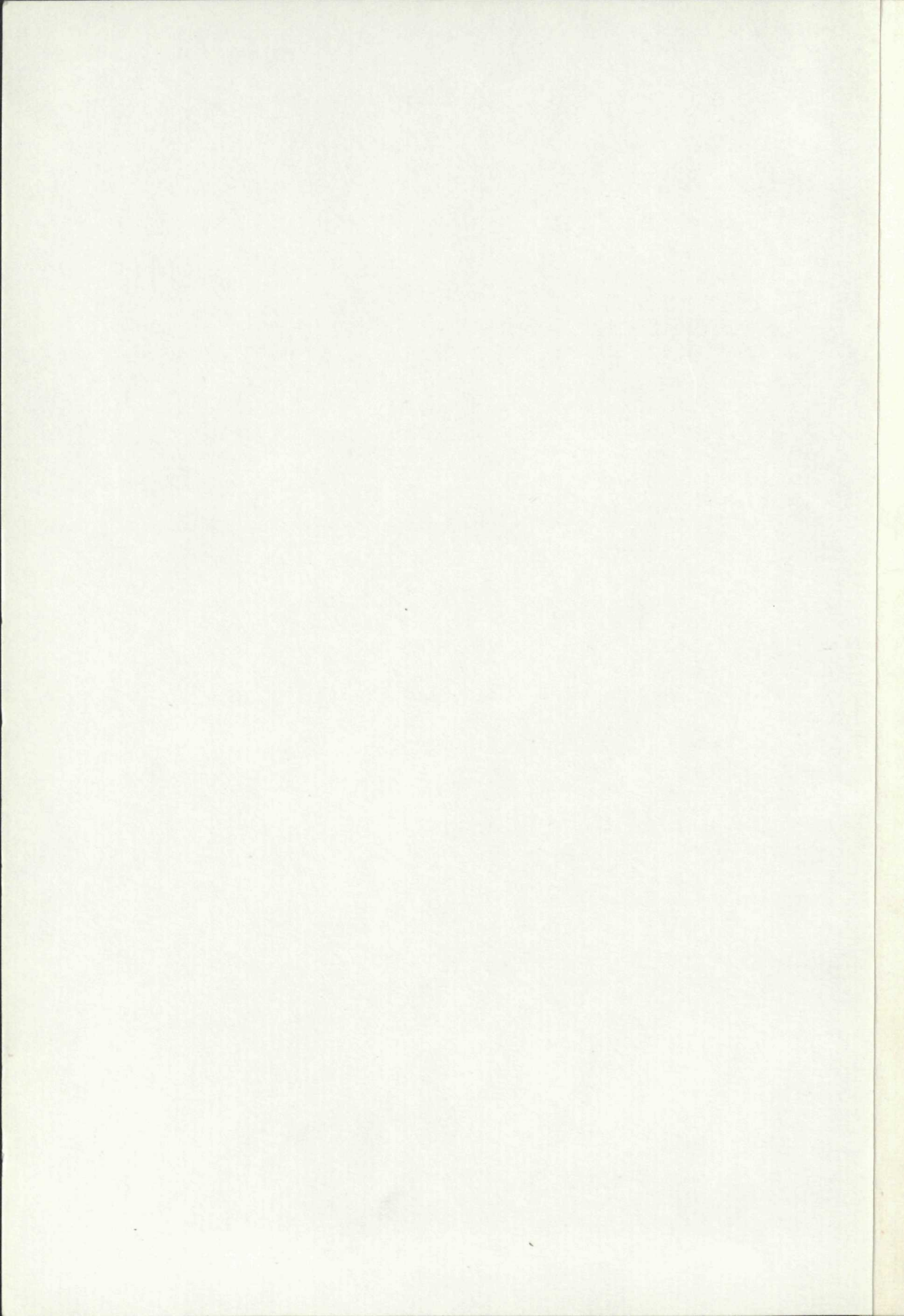
**Geotectonic position of magmatites  
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**SÉRIA**

*Geológia 4*

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# ZÁPADNÉ KARPATY

SÉRIA

*geológia 4*

GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA, BRATISLAVA 1978

# ZÁKLADNÍ KARBY

1978

geologický ústav



**Michal Maheľ**

**Geotectonic position of magmatites  
in the Carpathians, Balkan and Dinarides**

**Geotektonické postavenie magmatitov  
v Karpatoch, na Balkáne a v Dinaridách**

**Геотектоническая позиция магматитов  
в Карпатах, на Балкане и в Динаридах**

1941.10.10

Geotectonic position of magmatic  
in the Carpathians, Balkans and Himalayas

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Geol.	1210/46
KN	30/76
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## I. Geotectonic regional classification of the magmatites

The Tectonic Map of the Carpatho-Balkan regions on a scale of 1:1,000,000 (M. MAHEL' 1973) has revealed that the distribution of magmatites in the Eastern Alps, Carpathians, Balkan and Dinarides shows some characteristic features, such as:

- Abundant, for the most part large bodies of pre-Alpine granitoids and only small, frequently isolated granitoid bodies of Alpine age;
- predominantly acid volcanics (most frequently quartz porphyries), accompanying the oldest Neohercynian molasse (Upper Carboniferous — Permian);
- numerous pre-Carboniferous volcanics (prevalently basites) and a limited distribution of the Alpine volcanics;
- a prominent volcano-plutonic association of Alpine inter-orogenic magmatic rocks associated with the pre-molasse fillings of longitudinal grabens in the Balkan, South Carpathians and Apuseni Mts. (banatite formation);
- extensive occurrences of Tertiary volcanics, predominantly andesites accompanied by rhyolites, confined to young intermontane basins;
- numerous but small bodies of Pliocene-Quaternary basalts scattered particularly in the interior zones of the Alpides.

The editorial work on the Tectonic Map of the Carpatho-Balkan regions and the preparation of synthesizing chapters in its Explanatory text (M. MAHEL' et al. 1974) stimulated the author to study

- a) the role of the magmatites in the separate evolution stages of the Alpides (the timing of magmatism);
- b) the structural position of magmatites and their share in the formation of structures in the Alpine part of south-eastern Europe.

The study of the relationship between magmatites and sedimentary formations and folding processes was aimed at the analysis of the tectonic regime controlling the generation of magmatites and their influence on the tectonic evolution in the individual stages and areas of the Alpides, as well as at the elucidation of the crustal types in the Alpidic area and their changes during the geological history.

The structural lay-out of this paper is presented on the attached maps (Figs. 1,2).

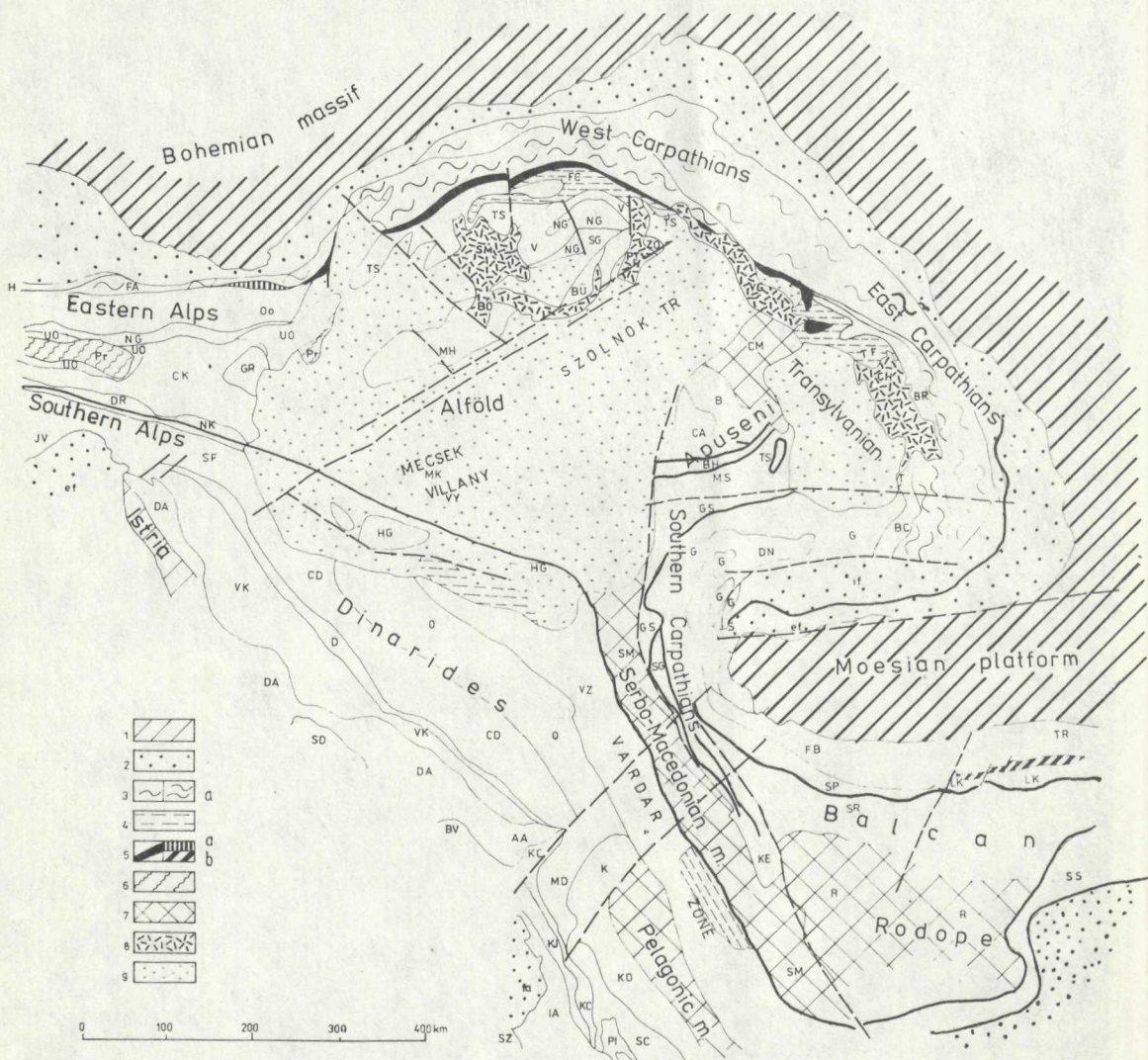


Fig. 1 Scheme showing the principal tectonic units of the Eastern Alps, Carpathians, Balkans and Dinarides

1. Platforms
2. Foredeep
3. Flysch Zone of the Eastern Alps and of the East and West Carpathians, a) inner units
4. Flysch of the inner arc
5. Klippen Belt, a) Gresten Klippen Zone, b) Kotel Zone
6. Penninicum
7. Interior massifs
8. Neovolcanics
9. Large basins

The Tectonic Map of the Carpathian-Balkan Regions and their Foreland was used as a base for their preparation.

The Alpine tectonic complexes were divided according to the date of their individualization into three categories: (a) the Palaeo-alpine units formed during the Austrian-Mediterranean folding phases (not including the more ancient Cimmerian units formed in the Mesozoic); (b) the Meso-alpine units formed during the Laramide-Pyrenean folding phases in the Palaeogene; and (c) the Neo-alpine units individualized by the Savian-Styrian folding phases in the Neogene period. The groups of facies also showing the tectonic regime of sedimentation (denoted as tectonogroups on the Tectonic Map) served us as indicators of the type and dynamic state of the crust and thus of the geotectonic conditions during the formation of magmatites in the individual evolution stages of the Alpide region.



#### Tectonic units

Eastern Alps: Oo — Oberostalpin, NG — Northern Grauwacken Zone, UO — Unterostalpin, GR — Palaeozoic of Graz, CK — East Alpine Crystalline, DR — Drauzug, NK — Northern Karawanken

Inner West Carpathians: TS — Tatrídes and Subtatríc nappes (Križna, Choč, Strážov), V — Veporide units, NG — North Gemeride unit, SG — South Gemeride unit, BÜ — Bükk, FC — Central Carpathian flysch

Inner East Carpathians: ZO — Zemplinicum, BR — Bretilla unit, BC — Bucovina and Subbucovina nappes, T — Transylvanian nappes, FT — Transcarpathian flysch

Apuseni: B — Bihor Zone, Ca — Codru-Arieşani nappe, BH — Biharia, MC — Muncel nappe, MS — Metalliferous Zone, TS — Trascău nappe, CM — Central massif

South Carpathians: DN — Danubicum, G — Geticum, SG — Saska-Gornjak unit, KE — Kraishtides, SG — Suprageticum, S — Severin-(Kraina)

Balkan: TR — Transitional Zone (northern margin) of Fore-Balkan, FB — Fore-Balkan proper, SP — Stara Planina Zone, LK — Luda-Kamchiya Zone, SR — Srednogorie, SS — Strandzha-Sakar Zone, R — Rhodope massif

Dinarides — Hellenides: JV — Venetian, Julian and Savinian Alps, SF — Sava fold Zone, HG — Horst and graben Zone, VZ — Vardar Zone, O — Ophiolitic Zone, CD — Central Dinaric Zone, MD — Mirđita nappe, SC — Subpelagonic nappe, D — Durmitor Zone, VK — Visoki Krš nappe, DA — Zone of Dalmatian Folds, BV — Budva unit, KC — Krasta-Zukali unit, Pl — Pindos unit, KJ — Kruja unit, SD — South Adriatic unit, IA — Ionian unit, SZ — Sazan unit, IS — Istrian platform

Central Hungarian Massif and its projections: MH — Mid-mountains of Hungary, MK — Mecsek, VY — Villány

Neovolcanic mountain ranges: SM — Slovakian Mid-mountains (Slovenské stredohorie), BÖ — Börşöny Mts., CS — Cserhát Mts., MA — Mátra Mts., P-T — Prešov-Tokaj Mts., V-G — Vihorlat Guttin Mts., C-H — Caliman- -Harghita Mts.

On the basis of the above data the magmatites are divided into the following geotectonic types:

- granitoids associated intimately with folding processes and the formation of the fundamental structural Alpine and pre-Alpine patterns;

- Neohercynian late geosynclinal volcanics genetically linked closely to the Hercynian granitoids; they accompany Neohercynian molasse fillings of troughs and depressions;

- pre-Alpine eugeosynclinal volcanics, accompanying pre-Carboniferous series of the Alpidic region;

- Mesozoic ophiolites restricted to particular structural zones;

- Meso-alpine intraorogenic volcano-plutonic magmatites (banatites), accompanying the graben-synclinoria and grabens filled with pre-molasses, and for the greater part with Alpine flysch;

- late geosynclinal young volcanics related genetically with Late Tertiary intermontane basins, which are filled with Cenozoic molasses of several evolution stages, including the closing phase of the development of the large structures forming the recent relief (called the morphostructures);

- final volcanics connected with the postorogenic stage of the Alpides.

The magmatites will be dealt with in the succession given above.



## II. Characteristics of the geotectonic types of magmatites

### 1. Granitoids

In the Alpides of Europe there is a striking difference between the proportion and size of pre-Alpine and Alpine granitoid bodies (Fig. 2).

a) Pre-alpine granitoids most frequently make up large plutons. They are usually divided according to their geotectonic age into pre-Baikalian, Baikalian, Caledonian and Hercynian granitoids. Most of the granitoid bodies occur within the oldest, mostly higher-metamorphosed complexes of the Alpine area, dated as pre-Palaeozoic by many authors. No wonder that their origin was often put into connection with the pre-Baikalian and Baikalian foldings, but recent radiometric dating has shown that the preponderant part of the granitoid bodies was emplaced during the Hercynian orogeny. Other granitoid bodies have been re-dated because the metamorphosed complexes in which they occur have recently been assigned to the Palaeozoic on the basis of biostratigraphic (palynological) study results. Attempt will be made here to outline the present state of knowledge, although the Alpine segments differ widely in the number and extent of granitoid bodies cropping out at the surface, and have not been investigated to the same detail (Fig. 2).

In the Crystalline of the Eastern Alps, the granite-gneisses related genetically to the 'Grobgneis Series' are regarded as pre-Hercynian. This age has been confirmed by radiometric values of  $582 \pm 52$  m.y. obtained for the granite-gneiss at the eastern margin of the Alps in the Sopron Mts. in Hungary (E. SZÁDECKY-KARDOSS et al. 1967). Major massifs of postkinematic granites distributed mainly in the Koralm Crystalline in Austria giving age values of about 270 m.y. are dated as Hercynian. If the Hercynian metamorphism in the Eastern and Southern Alps occurred between 360 and 300 m.y. B.P. (E. JÄGER 1973), these granitoids are contemporaneous with the closing phase of the Hercynian metamorphic processes. The Variscan orogenic plutonism is represented in the Alps to a large extent (granitoids with radiometric ages about 240 m.y.), particularly in the Penninicum and somewhat less in the Crystalline of the Eastern Alps and Northern Karawanken (H. FLÜGEL 1975). Diorites and granites of Northern Karawanken (Eisenkappel), which produced

contact metamorphism of Palaeozoic complexes gave radiometric ages  $227 \pm 7$  m.y. (on biotite) and  $244 \pm 6$  m.y. (on pyroxenes).

The West Carpathians of Slovakia and Poland (Fig. 3) have a particularly high portion of granitoids of different genetic and petrographic types. In the Tatríde Crystalline showing only a slight Alpine reworking, granitoids are the main component of the crystalline 'cores'. The crystalline cores represent tectonic blocks, in which erosion and denudation proceeded to different levels. In some cores only upper parts of postkinematic batholiths or phacoliths with abundant pegmatite and aplite have been laid bare (the Little Carpathians, the Považský Inovec, northern part of the Branisko Mts.), whilst in other cores lower parts of these bodies crop out, containing synkinematic granitoids and broad migmatite belts (the Low Tatra, Sľúbica, southern part of the Branisko). For the most part the rocks are biotite granodiorites, trondhjemites and quartz diorites (the Ďumbier type). The bodies of autometamorphic coarser-grained to porphyritic granite show a more pronounced differentiation (the Prašivá type: J. KAMENICKÝ in M. MAHEL' et al. 1967) distributed in the marginal and upper parts of granitoid bodies.

The more southerly Veporide zone of the West Carpathian Crystalline in Slovakia is distinguished by a prominent zonal arrangement of the granitoid bodies caused by a strong Alpine reworking of the crystalline complex. In its extreme northern (Lubietová) zone, granite-gneisses, migmatites and minor elongated bodies of less stressed granite porphyry and granodiorite porphyry predominate. The latter are thought to be genetically associated with the effusions of Permian quartz porphyry (V. ZOUBEK 1931). The more southerly (Kraklová) zone (with a prevalence of mica-schists and phyllites and rich in basites) encloses minor allochthonous granitoid bodies. These represent nappe outliers translated from the Králová hora zone situated farther to the south. This is built up mainly of granitoids showing a secondary Alpine reworking and containing numerous xenoliths of the metamorphosed rocks of the contact zone (A. KLINEC 1966). The most southerly (Kohút) zone of the Veporides is distinguished by a variety of granitoids (granite-gneisses, migmatites, large granodiorite bodies of several structural types). Leucocratic granites probably of Permian age are fairly frequent on its southern border. Autometamorphic granitoids do not form clearly defined bodies in the Kohút zone. Pneumatolytic processes did not produce a large number of pegmatites and aplites but an extensive migmatization of the country rocks and leucocratic granites containing relics of the primary biotite paragneiss (M. MAHEL' et al. 1967).

A distinctive feature of the Veporide zone is a strong Alpine tectonic reworking of its granitoids, which form broad zones but display only faint metamorphic schistosity, in contrast to the limited number of narrow mylonite zones dissecting the Crystalline of the Tatríde zone. Minor bodies of granitoids without definite structural orientation, located along major tectonic lines of the Veporides, are the products of a later, probably Alpine granitization (A. KLINEC 1966). The most

southerly zone of the West Carpathians with the preserved pre-Alpine basement – the Gemeride zone – is distinguished by lower metamorphism, a subordinate occurrence of minor pre-Alpine granitoid bodies and the presence of Alpine granites (Fig. 10).

The substantial part of the West Carpathian granitoid bodies occur in the metamorphosed gneiss and mica-schist series, which until recently were thought to be pre-Hercynian, and are accompanied by migmatites. However, on the basis of recent palynological data part of the metamorphic, particularly mica-schist series is assigned to the Early Palaeozoic. Since most granitoids have also been radiometrically dated as Palaeozoic, there is a tendency to place the overall granitization in the Hercynian epoch (I. BOJKO 1975). This trend confronts us with a thankless task to decide which of the granitoid bodies are pre-Hercynian. By analogy with other Alpidic segments we consider as the oldest, probably pre-Hercynian, the granite-gneisses, especially in the Lubietová and Kohút zones of the Veporides, and some migmatites in the crystalline complexes of the Tatrides in the High Tatra Mts., the Ďumbier crystalline complex of the Low Tatra Mts. and the Čierna hora Mts.

Radiometric ages of West Carpathian granitoids range mostly from 320 to 280 m.y. (J. KANTOR 1959a, J. BURCHART 1970, A. K. BOJKO 1975), but even the dates of 230–220 m.y. are relatively numerous. Although the radiometric ages should be estimated soberly (with regard to differences caused demonstrably by the differences in sampling and radiometric methods used), they did furnish instructive results in most cases. In the High Tatra Mts., the most frequent dates are 279–308 (317) m.y. for granitoids, 300–250 m.y. for migmatites and 257–254 m.y. for pegmatites. In the Low Tatra Mts., where the granitoid types are unusually varied, radiometric dating gave the following values (J. KANTOR 1959b, A. K. BOJKO et al. 1965): biotite granite (Ďumbier type — on biotite) 305–296 m.y.; autometamorphic granodiorite (Prašivá type — on biotite) 305–288 m.y. and 280 m.y. on feldspars; pegmatites 320–315 m.y. (on muscovite); synkinematic granitoids (Králíčka type) gave higher values — 360 m.y. on muscovite and 320–325 m.y. on zircon — and pegmatites from this granite bodies 330 m.y. Migmatites making up an extensive zone in the crystalline complex of the Low Tatra Mts. also yielded higher dates of 335–395 m.y. (zircon) and 310–315 m.y. (monazite). Also the Muráň gneissose granites show the radiometric values 360 m.y. (zircon) (B. CAMBEL et al. 1977).

The granitoid bodies of the Little Carpathians in Slovakia deserve special attention, as they partly occur amidst the palaeontologically evidenced crystalline complexes. The granitoid of the Bratislava pluton emplaced in the gneiss micaschist-phyllite complex of the Pezinok-Pernek Formation gave ages of  $382 \pm 275$  m.y., migmatites 309–305 m.y. and pegmatites and aplites of this pluton either the same or lower values (227–222 m.y. on muscovite and orthoclase — J. KANTOR 1959b). The northerly Modra pluton which produced contact-metamorphism of the Devonian-Lower Carboniferous Harmónia Group, yielded radiometric ages of 333 to 232

m. y., leucocratic granite 222 m. y. (A. K. BOJKO et al. 1974; G. P. BAGDASARIAN et al. 1977).

Some highly contaminated granitoids of the Veporide type (e.g. of the large Sihla body) gave as high values as 380 m. y. (biotite). Small Veporide granitoid bodies (Hrončok type), which until recently were dated by some authors as latest Permian or Palaeo-alpine, gave the value 370 m. y. (zircon), 260 m. y. but also 113—110 (92) m. y. (feldspars and biotite). The latter date was apparently influenced by Alpine reheating.

Although the radiometric data indicate the Hercynian age even of granite-gneisses and of most of the migmatites, the up to thousands of metres thick quartz porphyries of the Ordovician and partly of Silurian age can hardly be explained without the presence of Late Baikalian and/or Caledonian granitoid bodies. This contention is also suggested by the latest radiometric data on orthogneisses and migmatites from the Veporide Crystalline (Muráň) and the Low Tatra Mts. (Králíčka – Srdiečko) with Pb isochrone of up to 505 m.y., and on the Veporide granitoids (Sihla) with Pb 205/U 238 ratio giving an age of 380 m.y. (personal communication of L. KAMENICKÝ).

In the Mecsek Mts. in Hungary (i.e. in the eastern sector of the Hungarian massif consolidated by pre-Palaeozoic foldings) radiometric values about 500 m. y. but also 305—284 m. y. (Rb/Sr) and 230 m. y. were obtained for migmatites and 360—321 m. y. for granitoids (E. SZÁDECKY-KARDOSS 1973). Microgranite-aplite sills and dykes of the Mecsek Mts. pierce through the Silurian sedimentary complexes.

Appreciably lower values (about 240 m. y.) were yielded by granitoids in the Velence Mts. (at the eastern margin of the western sector of the Hungarian massif, at the contact with the Palaeozoic and Mesozoic of geosynclinal type).

In the Apuseni Mts. (Rumania) (Fig. 4) granite-gneisses and migmatites probably of pre-Hercynian age accompany the pre-Cambrian, predominantly gneiss complexes in the Codru nappe and in the Baia de Arieş Formation of the Biharia nappe system. The minor de Vinta granitoid body was dated radiometrically at 508 m. y. and at about 172 m. y., which value dates obviously the Alpine rejuvenation.

The large granitoid Muntele Mare batholith (up to 35 km long and 10 km broad) in the Bihor autochthon of the Apuseni Mts. produced contact metamorphism not only of the Middle Proterozoic gneiss series but also of the Upper Proterozoic-Cambrian Arada micaschist-phyllite Formation. The opinions on the age of the batholith differ. On the basis of its geological position the granites are thought to be Caledonian (R. DIMITRESCU 1967, 1976), but radiometric values of 552 m. y. indicate the Baikalian age. According to the latest radiometric age values of 232 m. y. (M. SOROIN et al. 1969) the Hercynian age seems more likely. Minor granitoid bodies occur in the nappe units of the more southerly part of the Apuseni Mts. The Codru granitoids (plagiogranites and granites accompanied by migmatites) in the Codru nappe system gave radiometric ages 334—344 m. y. In the Highiş Mts.

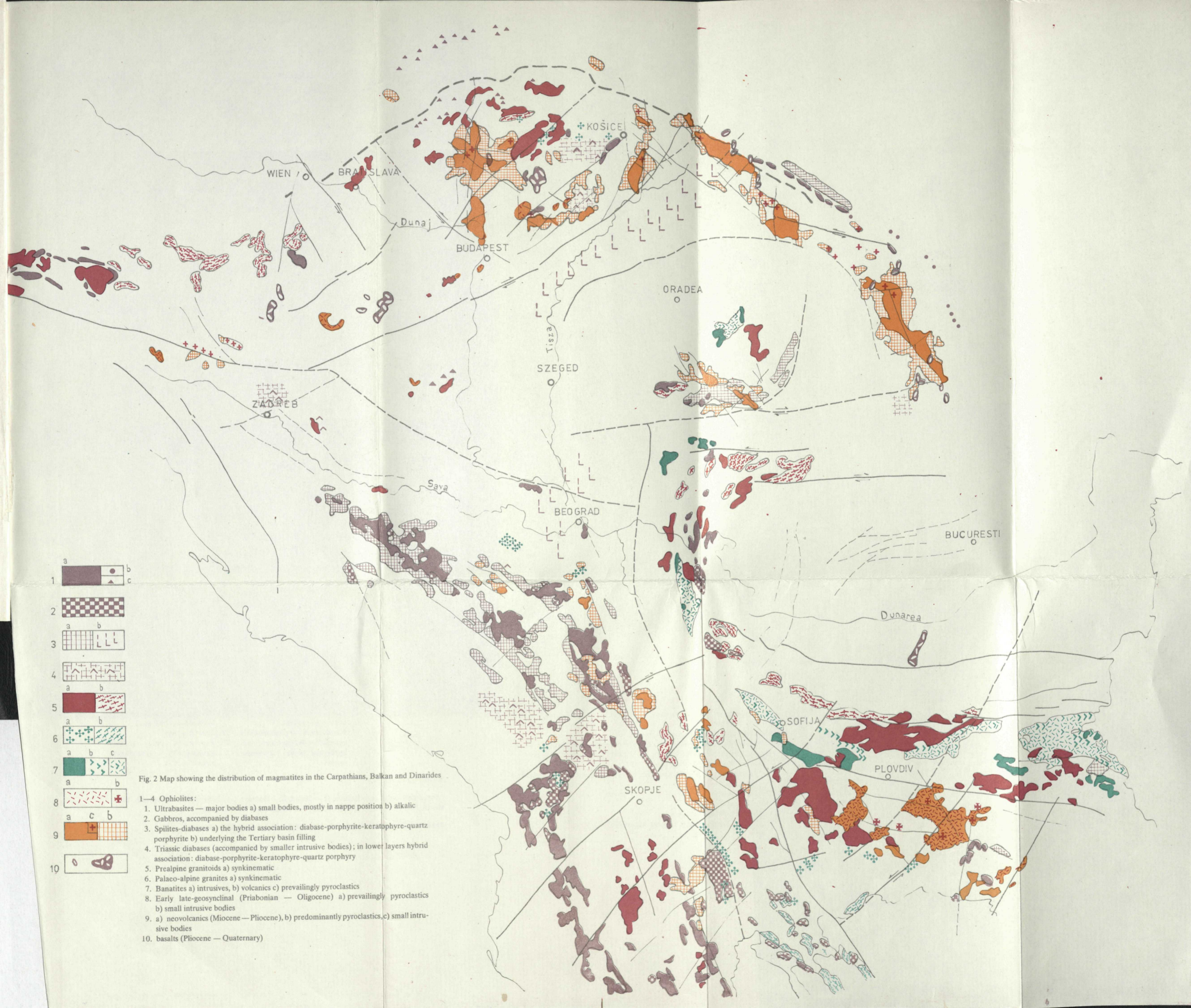
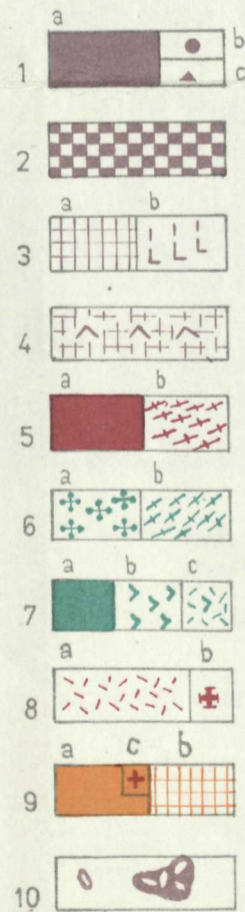


Fig. 2 Map showing the distribution of magmatites in the Carpathians, Balkan and Dinarides



1-4 Ophiolites:

1. Ultrabasites — major bodies a) small bodies, mostly in nappe position b) alkalic
2. Gabbros, accompanied by diabases
3. Spilites-diabases a) the hybrid association: diabase-porphyrine-keratophyre-quartz porphyrite b) underlying the Tertiary basin filling
4. Triassic diabases (accompanied by smaller intrusive bodies); in lower layers hybrid association: diabase-porphyrine-keratophyre-quartz porphyry
5. Prealpine granitoids a) synkinematic
6. Palaeo-alpine granites a) synkinematic
7. Banatites a) intrusives, b) volcanics c) prevailing pyroclastics
8. Early late-geosynclinal (Priabonian — Oligocene) a) prevailing pyroclastics b) small intrusive bodies
9. a) neovolcanics (Miocene — Pliocene), b) predominantly pyroclastics, c) small intrusive bodies
10. basalts (Pliocene — Quaternary)

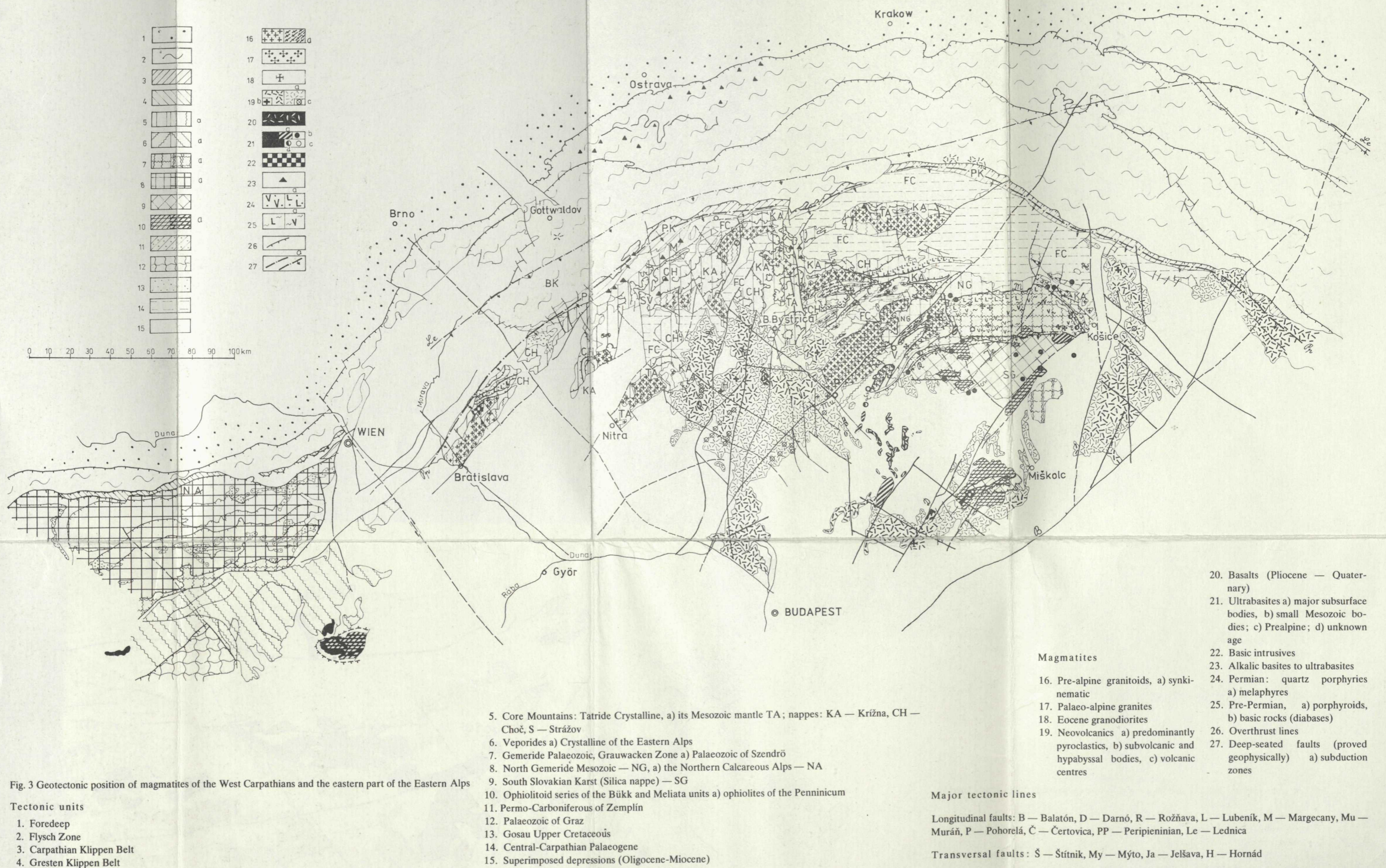


Fig. 3 Geotectonic position of magmatites of the West Carpathians and the eastern part of the Eastern Alps

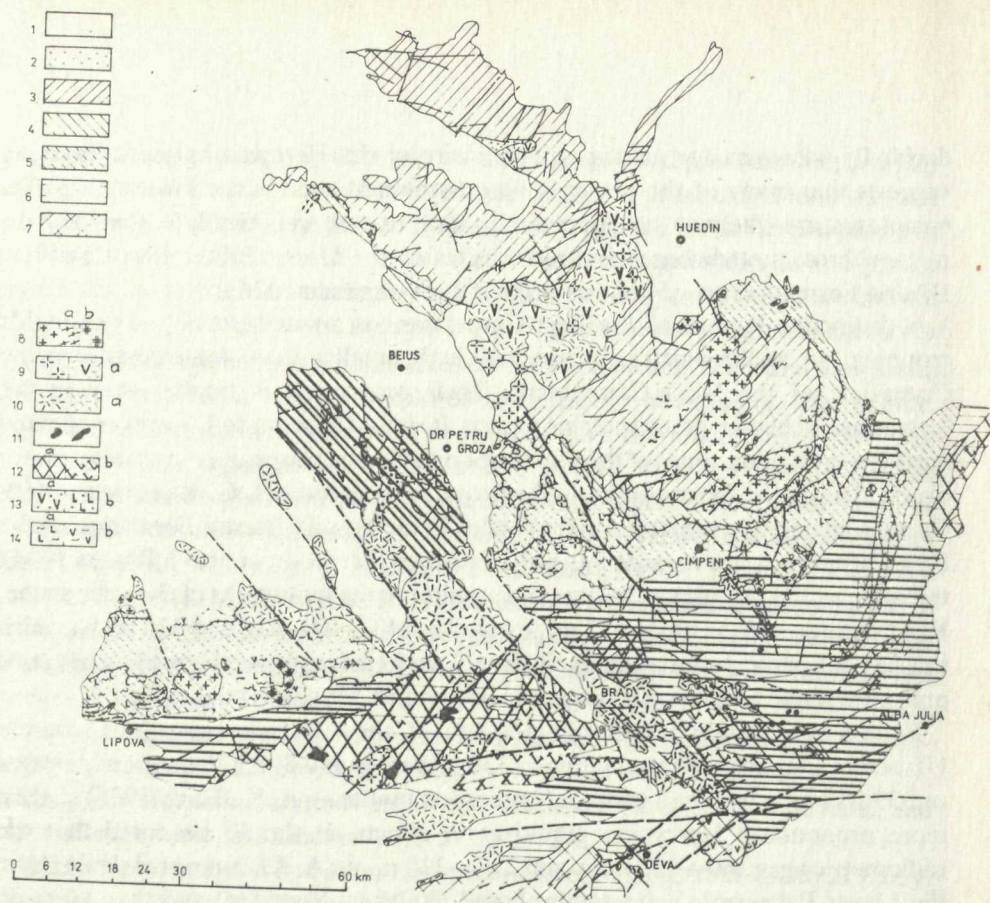


Fig. 4 Geotectonic position of magmatites in the Apuseni Mts. (Compiled on the basis of the CBGA Tectonic Map — edit. M. MaheI, 1974 and the Geological Map of the Apuseni — M. Borcoş et al. 1976)

#### Tectonic units

- |                          |   |
|--------------------------|---|
| 1. Neogene basins        | 5. System of Baia de Aries units (Biharia series) |
| 2. Senonian, Gosau type  | 6. Metalliferous Zone                             |
| 3. Bihor autochthon      | 7. Ultrageticum                                   |
| 4. System of Codru units |   |

#### Magmatites

8. Hercynian and older granitoids, a) synkinematic; b) alkalic,
9. Banatites: granites and granodiorites, a) volcanics (andesite-dacite-rhyolite)
10. Neovolcanics a) pyroclastics
11. Serpentinites (peridotites), partly gabbros
12. a) predominantly basalts, b) spilites-orthophyres-keratophyres-porphyrites
13. Permian volcanics: a) quartz porphyries, b) melaphyres
14. Pre-Upper Carboniferous volcanics: a) metabasalts, amphibolites, b) porphyroids

hypabyssal pegmatitic granites and adamellites are more frequent; they yield late Hercynian K/Ar ages 221—226 m. y. up to 203—172 m. y. In the Codru and Biharia nappe systems some granitoids, granites, syenites and porphyrite dykes penetrate even the sedimentary Devonian-Lower Carboniferous Paiuseni Formation. This

decidedly witnesses for an appreciable share of the Hercynian granitization and suggests that many of the above bodies distributed even in the lower and earlier complexes are likely to be Hercynian. Granites that genetically accompany the metagabbros, metadiabases and ultrabasites of the Upper Proterozoic-Cambrian Biharia Formation are of another origin (R. DIMITRESCU 1976).

A distinctive feature of the East Carpathians is a small portion of granitoids cropping out at the earth's surface, strikingly smaller than, for example, in the Crystalline of the West Carpathians. Their proportion is smaller even in the formations of higher-grade metamorphism. It should be recorded, however, that the northern crystalline zones of the Marmaroš massif, which are richer in granitoids, are overridden by the more southerly Bystrica nappe system (H. G. KRÄUTNER 1969). Granite-gneisses, magmatites and gneisses in the Bretila-Barnar Formation of the East Carpathians are regarded as pre-Hercynian (M. SÂNDULESCU - I. BERCIA 1974), the same as the Hăghimaş diorites and granites in the mountains of the same name. K/Ar dating yielded values 517 m. y. but also 443—409, 328 and 310 m. y., which indicate the participation of the Baikalian, Caledonian and even Hercynian orogenic processes in the formation of granitoid bodies (M. MURESAN et al. 1975).

Radiometric values for granite-gneisses of the Čivčinske gory Mts. (in the Ukrainian Carpathians) amounting to  $330 \pm 35$  m. y. and  $305 \pm 20$  m. y. and the ages of  $339 \pm 12$  m. y. for small granitoid bodies (on biotite — A. K. BOJKO 1975) suggest more pronounced Hercynian granitization effects. It should be noted that the radiometric ages 360—350 m. y. and 330—320 m. y. (A. K. BOJKO et al. 1974) from the Lower Palaeozoic epimetamorphosed (Tulgheş, Repedea) and the older and higher-metamorphosed formations such as the Bely Potok or Bretila-Rarău Fms. also point to a strong Hercynian metamorphism in the East Carpathians. The latter formations, however, also yielded Rb/Sr ages of about 530 m. y. (Baikalian).

The East Carpathians contain some particular magmatite occurrences (Fig. 11). It is chiefly the large dome-like elongated nepheline-syenite massif showing gneissic structure and conformable schistosity to the mantle. Its radiometric age of 396—297 m. y. indicates a genetic association with the Hercynian folding. The massif is pierced by aplite and lamprophyre dykes. Plagiogranite stocks in the Čivčinske gory Mountains are regarded by some authors as members of the gabbro-plagiogranite formation (E. K. LAZARENKO et al. 1973). They are thought to be products of the postkinematic late Hercynian stage on account of their high content of alkalis and oversaturation with respect to  $\text{SiO}_2$  (G. RUDAKOV 1971). The granite-porphry and granodiorite-porphryite stocks accompanying quartz porphyries and quartz keratophyres in the Tulgheş-Delovec epimetamorphosed Form. also belong to specific magmatic types of this part of the East Carpathians (G. RUDAKOV 1971).

The proportion of granitoid rocks in the Crystalline of the Geticum and Danubicum tectonic units of the South Carpathians in Rumania is considerable (Fig. 5). Pre-Palaeozoic age is assigned to synkinematic granites, granodiorites and quartz



diorites distributed mainly in the meso-katamorphosed pre-Baikalian complexes. The best known are the Poneasca and Sichevita granite, granodiorite and quartz-diorite plutons in the Semenic Mts. (Banat) and the Bîrsa Fierului pluton in the Făgăraş Mts., where there are many other granitoid, granodiorite and granite masses; in the Lotru Mts. alkaline syenite and nepheline syenite bodies are known. Major bodies of late kinematic and synkinematic granodiorite and granite also occur in the Baikalian epimetamorphosed series of the Danubicum. The Danubian, Suşita and Tismana plutons are best explored and deserve mentioning. The granite, granodiorite and granite-porphyrries of the last gave radiometric ages 556—520 m. y. and 440 m. y. but prevalently about 296 m. y. (S. MINZATU et al. 1975). The Munte Mic granite pluton in the Danubicum, which is generally considered to be a synkinetic body of Baikalian age (H. SAVU et al. 1975), yielded sporadically higher values of about 411 m. y. (Caledonian) and mostly of ca. 370—290 and even 220—192 m. y. (S. MINZATU et al. 1975; the last date is explained as due to the loss of radiogenic Ar). Granites from the Ogradena, Cherbelezu and Sfîrdinu plutons in the Danubicum of Rumania gave K/Ar dates predominantly of 373—306 and less frequently of about 268 and 203—184 m. y. (S. MINZATU et al. 1975). These granitoid masses are penetrated by granite-pegmatite, granodiorite, granodiorite-porphiry, aplite and lamprophyre dykes.

The pre-Hercynian granitoids are of calc-alkaline type when synkinematic, and more alkaline when late kinematic (H. SAVU 1972).

The above-mentioned and other radiometric dates ranging from 380 to 219 m. y. (N. SEMENKO 1975; M. DIVILJAN—S. DIVILJAN 1967) indicate that the opinions on the pre-Palaeozoic age of the South Carpathian granitoids, particularly post-kinematic (D. GIUSCĂ et al. 1969) should be taken with reservation, the more so as the granitoids in the continuation of the Danubicum tectonic unit in Jugoslavia show decidedly Hercynian age as, for example, the Gorjane pluton which gave Rb/Sr age of 304 m. y. but also 370 and 220 m. y. and the Plavno pluton 208 m. y.

In the Serbo-Macedonian massif, the synkinematic granitoids are regarded as pre-Hercynian, partly affected by Hercynian reworking (M. D. DIMITRIEVIĆ 1967). The Vljajna pluton, for example, yielded 450 m. y., the Bujanovac pluton even 859 m. y. (S. BOYADJIEV 1974). Granite-gneisses and synkinematic granitoids and diorites in the Belasica and Ogražden bodies and in the Upper Proterozoic Osogovo-Lisec and Vlasina metamorphic Formations are dated as pre-Hercynian. Hercynian granitoid plutonic bodies also occur in abundance; more frequent than granitoids linked with the beginning of the Hercynian folding (e.g. the Belasica body — Rb/Sr age of 350 m. y. but also 252 and 114 m. y.) are those indicating the later phases of the dying-out Hercynian folding (e.g. granitoids in the Bujanovac phacolith — 234 m. y., in the bodies of Osogovo —  $220 \pm 50$  m. y. (zircon), Pliučavica — 252 m. y. and Stalač — 250 m. y. — G. DELEON 1966).

Some granitoids of the Pelagonian massif are of pre-Palaeozoic age, as evidenced

by Rb/Sr dating of the Selečko pluton, which gave 1140, 838, 609—577 but also  $135 \pm 5$  and 99—97 m. y. (G. DELEON 1966). This wide range shows a polycyclic character of granitization. Late kinematic granitoids are likely to be predominantly Hercynian (S. KARAMATA 1974).

In the Rhodope massif and in the Srednogorie of Bulgaria, (Fig. 6) granitization and migmatization were most extensive in the oldest complexes which are regarded as Archean or Early Proterozoic as, for example, the synkinematic Lesovo granitoid body (I. BOJANOV—D. KOZUCHAROV 1968). Large, linear batholithic bodies of postkinematic granitoids are abundant in the western Balkan block; the batholith in the Rila anticlinorium pierced by numerous aplite-pegmatite veins and diorite and granite-porphry dykes is one of the largest.

Minor bodies of these 'South Bulgarian' diorites, granodiorites and granites are also in the more easterly Sakar and Strandža blocks and in the Srednogorie Mts. These rocks gave radiometric ages of 360 m. y., more often ca. 305—270 m. y., but also 150—130 and 50—30 m. y. (BOYADJIEV—P. LILOV 1971; S. BOYADJIEV 1974). They are dated by individual authors as Hercynian or Early Caledonian and even Precambrian, because they occur not only in the (Cambrian-Ordovician) phyllite-diabase series but also, and more frequently, amidst the pre-Palaeozoic complexes showing higher-grade metamorphism. Their lower parts are contaminated by assimilates from their mantle, with which they are nearly or fully conformable. Granitoids of the later phases show irregular intrusive contacts. Pegmatite veins, granite-porphry, lamprophyre and dolerite dykes are the youngest (I. ZAGORČEV 1975).

In the Stara Planina Mts. there are fairly frequent bodies of Carboniferous granitoids (granodiorite and granite, which locally grade into gabbrodiorites) of radiometric dates 348—292 m. y. (S. BOYADJIEV 1974). They make up elongated batholiths or stocks in the axial parts of the Hercynian fold structures and predominantly accompany the (Upper Proterozoic? — Ordovician) diabase-phyllitic series. They are in part synkinematic (Klisura and Stakevci plutons), intimately associated with their mantle rocks and accompanied by the zone of thermal metamorphism of their mantle (E. DIMITROVA et al. 1975). The Stara Planina granitoids close the evolution of the Hercynian geosyncline in this area. They are distributed along its interior margin at the flexure connecting this geosyncline (E. BONČEV 1966) with the Hercynian block of the Rhodope-Srednogorie area, which was intruded by the South Balkan granitoid bodies.

Particular granitoid types in the Balkan are, for example, the monzonite and syenite stocks and dykes in the Pirin and Srednogorie Mts., regarded as Permian, and the Struma diorite Formation. This is represented by small bodies of leucocratic granite-gneiss and semiconformable granodiorite and granite bodies, which were emplaced in the diabase-phyllite series (Upper Proterozoic? — Ordovician) in association with gabbros, diorites, gabbrodiorites and small serpentinite bodies.

They occur in the Krajištides, i.e. in the western part of the Balkanides, and also in the marginal part of the Western Rhodope Mts. They are thought to be related genetically with the Caledonian folding (I. HAJDUTOV 1971).

In the interior zones of the Dinarides only minor bodies of Hercynian granitoids, Carboniferous-Early Permian in age, are present, such as Srpsko Crna Gora, Bukulja, Cer, Motajica, Papuk and Moslavočka Gora (Fig. 9).

b) The Alpine granitoids, unlike the pre-Hercynian and Hercynian, usually form only minor bodies. In essentials, three different associations can be distinguished, based on their age and relationship to the evolution of the Alpine orogen.

— Granitoids associated with Alpine folding phases and metamorphism. Most of them are Palaeo-alpine; Meso-alpine granitoids are sporadic (Fig. 2);

— granites accompanying the interorogenic volcanics, called banatites; they are Meso-alpine (Fig. 14b);

— granitoids forming minor plutons and hypabyssal bodies, accompanying late geosynclinal volcanics; they are Neo-alpine (Fig. 15a).

The Palaeo-alpine granitoids are associated mainly with the earlier phases of Alpine folding. Although the Palaeo-alpine folding period is of particular importance in the Alpide system, being the main controlling factor of its structural pattern, the granitoids are of small extent and occur only in the zones showing a higher metamorphic and thermal influence. In the Alps it is primarily the Penninicum, where intensive thermal and metamorphic effects were produced by the Meso-alpine (Pyrenean) orogenic phase, and subordinately some parts of the Austrian Unterostalpin Zone. Isolated bodies of Palaeo-alpine granitoids are also enclosed in the Koralm Crystalline of Austria.

In the West Carpathians of Slovakia the Palaeo-alpine granitoids are represented by granite and granite-porphry bodies in the Cambrian — Silurian Gelnica Group in the Spišsko-gemerské rudohorie Mts. (K/Ar age  $87 \pm 4$  Zlatá Idka,  $94 \pm 3$  Čučma but also  $141 \pm 1$  mil. y. (Čučma) (B. Cambel et al. 1977; J. Kantor 1960). The granitization process occurred in several phases and culminated by minor intrusions (Fig. 10). They are distinguished by abundant volatile components B, F, H<sub>2</sub>, a high background of Sn, W, Be, U and Mo, and low Sr and Ba (J. KAMENICKÝ—L. KAMENICKÝ 1955). The granite bodies are arranged in zones, which like the zones of contact metamorphism, are parallel to the course of the major Palaeo-alpine structures (P. GRECULA 1973). Small granitoid bodies follow the semiarculate weakened zones rimming the Volovec major anticlinorium in the Spišsko-gemerské rudohorie Mts. The northern belt of granite occurrences is restricted to the Hnilec line (L. ROZLOŽNÍK 1972). A relevant problem to be solved in the Gemerides of Slovakia is the relationship of these granite bodies to the overthrust of the Gemeride nappe on the Veporide zone. The well substantiated overthrusts of blocks containing rocks not affected by contact metamorphism on blocks with rocks of the same composition but contact metamorphosed (e.g. along the Jedlovec overthrust) incite

the opinion that the granite bodies together with the genetically related ore deposits are earlier than the nappe structure (P. GRECULA 1973). However, I think that intensive granitization of the country rocks with neoformation of minerals such as blue quartz, feldspars, biotite, and muscovite (found by deep boring — P. GRECULA 1973) rather suggest the 'rootedness' of the granite bodies and their post-deformational character. Of assistance in solving this problem may be minor bodies, especially of biotite granite, emplaced at the contact between the Hercynian granitoids of the Bujanová body and its metamorphosed mantle rocks, strongly reworked by Alpine tectogenesis, in the Čierna hora Mts. north of the Gemerides. They are frequently penetrated by a system of aplite veins and numerous fissures filled with tourmaline, hematite, molybdenite, quartz, pyrite and calcite. Some of these granite bodies pierce the Permian of the Veporides along weakened zones, and on Spálený vrch Hill a diorite body breaks through the overthrust slice of the Gemeride Carboniferous and Permian (J. ŠALÁT 1954). The position of the diorite body and its slight disturbance only by the latest fault system of NE-SW trend indicate that these bodies originated (like the youngest Gemeride granites) in the later phases of the Palaeo-alpine folding processes.

Granitoids with radiometric ages of 107—115 m. y. (J. KANTOR 1960), such as Muránska Zdychova (107 m. y.) or Hrončok intrusive bodies (110—114 m. y. — A. K. BOJKO 1975), which correspond to earlier Palaeo-alpine folding events, are distributed on major tectonic lines in the vicinity of large Hercynian granitoid bodies of the Veporide Crystalline displaying a strong Alpine reworking. Albite granites showing dates of about 86—90 m. y. occur in a major number along the southern margin of the Veporide Crystalline, which suffered a particularly strong Alpine reworking. Small veins and masses of aplite and pegmatite of NE-SW trend that cut across the predominantly mica-schists Hron Formation in the Kraklová zone of the Veporides are of the same age (A. KLINEC 1966), (Fig. 10).

Radiometric ages of 93—140 m. y. determined on granite pebbles in the Palaeogene Proč conglomerates of the Klippen Belt in eastern Slovakia indicate a fairly large extent of Palaeo-alpine granitoids also in the crystalline basement of the northern units of the West Carpathians. The pebbles are probably derived from the 'cordillera' situated either in the central or northern marginal part of the geosyncline of the Klippen Belt.

Palaeo-alpine granitoids are more abundant at the western and partly also eastern margins of the Serbo-Macedonian massif (F. KOCKEL et al. 1971), in the Southern Rhodope Mts., in the eastern block of the Srednogorie Mts. in Bulgaria and in the Pelagonian massif, i.e. in the areas strongly affected by Late Cimmerian and Palaeo-alpine remobilizations (Fig. 9). Granite dykes that penetrate the South Bulgarian granitoids of Hercynian age in the horst of the Srednogorie anticlinorium and the Sakar anticline gave radiometric ages of 130—135 and 140—160 m. y. (S. BOYADJIEV—P. LILOV 1971). These are obviously the dates of Cimmerian regenera-

tion related with metamorphic phenomena observed in the Mesozoic of the Sakar Mts. In Greece there are extensive occurrences even of synkinematic granitoids of this age. Some minor bodies at the eastern margin of the Serbo-Macedonian massif in Ogražden block were dated radiometrically at 124 m. y.

Of Late Jurassic to Early Cretaceous age are the minor granite intrusions with syenite dykes at the western limit of the Serbo-Macedonian massif (Gurnicet, Borkula, Furka — 167—156 m. y., Serta, Štip, Lojane — 120 m. y. after G. DELEON 1966). The same author records the Rb/Sr ages of Palaeo-alpine granitoids from the Pelagonian massif: the bodies of Babuna 148 and 121 m. y., Krušovo 148 m. y., Kajmakčalan 110 m. y., Seleška 131 and 99 m. y. and Prilep 97—170 m. y. The age values about 80—70 m. y. were determined on the granodiorite intrusions Kopaonik and Zelir. This rather wide range of dates suggests that activation of the pre-Alpine basement occurred in several stages in the southern parts of the Alpides.

Granitoids produced by sialic mobilization lack altogether in the Dinarides developed from the Palaeo-alpine eugeosyncline. At several places there are albite granites — differentiates of the gabbro-plagiogranite formation, which accompany and are comagmatic with the Middle Triassic pillow lavas (V. ALEXIĆ—N. KALENIĆ—N. PANTIĆ—E. HADŽI 1974). Quartz diorites, diorites, plagiogranites and granites genetically associated with the Jurassic volcanics of the Mirdita unit in the Albanides are similar (BIÇOKU et al. 1974).

More characteristic for the Meso-alpine evolution stage of the Alpides are the 'banatites' of Upper Cretaceous, Palaeocene to Lower Eocene age. On the basis of geotectonic position and petrographic character two areas can be differentiated: the Timok (including Banat) — Srednogorie area in the south (1) and the Poiana Ruscă — Apuseni area in the north (2) (Fig. 14b).

1. The southern area is distinguished by geosynclinal-graben structures filled with volcano-sedimentary complexes and abundant flysch sequences. The plutons are elongated, and in most cases emplaced in deep boundary faults of grabens or faults running near the margins of grabens and graben-synclinoria. In the Timok graben diorite, syenite-diorite, syenite, monzonite, gabbrodiorite and gabbro bodies are more numerous in its northern part, mainly along the up to 180 km long Ridan — Krepoljin fault line. They are dated as post-Late Cretaceous (70—50 m. y., Laramide to Palaeogene — S. KARAMATA 1974).

Major intrusive masses and hypabyssal bodies occur in Bulgaria (Fig. 6) chiefly between Sofia and Plovdiv along the Marica deep fault at the boundary between the Srednogorie and the Rhodope Mts. Large fissure plutons of Vitoša, Plana and Gucal show a uniform internal structure. They were formed at the epizonal level under low pressure-temperature conditions. In the central part of the Srednogorie Mts. several plutons are situated along the boundary between the Panagiurište synclinorium and the Srednogorie anticlinorium. In the eastern part of the Srednogorie, plutons are found in the Strandža anticlinorium and small bodies at the southern limitation of the

Burgas synclinorium. In these eastern areas, which in Mesozoic-Alpine times were rich in polygenetic volcanoes of central type, the transition from lava and pyroclastic facies to subvolcanic and even abyssal bodies of this age is particularly clear. In the Timok-Srednogorie area granites, granodiorites, syenite-diorites and monzonites, members of the gabbro-monzonite-granite formation (according to E. DIMITROVA et al. 1975) and gabbros form the accompaniment of the rhyolite-andesite-basalt and trachyte-trachyandesite-picrite volcanism. Big plutons, such as Vitoša-Plana, gave radiometric ages 70—90, but also 55 m. y. (Senonian dacites yielded 86—88 m. y., S. BOYADJIEV 1974). This would suggest a Late Cretaceous age of the plutons and their direct genetic association with the Late Cretaceous volcanism. The Upper Cretaceous fillings of grabens together with the marginal plutons were incorporated in the Alpine structural pattern through the effects of the Meso-alpine folding and formed its upper structural level.

2. In the Poiana Rusca (Fig. 5) area in the more northerly part of the Rumanian South Carpathians and in the Apuseni Mts. (Fig. 4), the plutons cut transversely the Palaeo-alpine structures, being associated with the transverse superimposed structures, elongated in the meridional direction. The plutons are late-tectonic and only slightly affected by folding processes. The intrusives of the banatite association belong to diorites, granodiorites and subordinately to granites and aplites; they are generally believed to be post-Laramide in age (M. BLEAHU 1974). They are less varied in composition, more acidic and with less Na than similar rocks in the southern area (1). In the Mureş range in Rumania, especially in the Drocea Mts., small bodies were intruded into the ophiolite complex. They are accompanied by rhyolites, whereas farther north granodiorite and diorite bodies are linked with andesites, dacites and with dykes of granodiorite-porphry and rhyolite.

Granites, granodiorites and diorites also build up the complex Vlădeasa body emplaced in a graben (up to 45 km long and about 30 km broad), which is elongated in the N-S direction and superimposed on the crystalline and Mesozoic basement of the Bihor autochthon of the Apuseni Mts. The accompanying granodiorite porphyrites and granophyres make a connecting link between the intrusives and effusive andesites, dacites and rhyolites.

Granitoid rocks, known as tonalites, accompany the Periadriatic zone. The tonalite bodies of the Karawanken display the radiometric values  $29 \pm 6$  mil. y. or  $28 \pm 4$  mil. y. (on biotite). Interesting is, however, that the accompanying granodiorite porphyry from the Eisenkappel area displays the values  $244 \pm 8$  mil. y. (on pyroxenes) and diorites  $227 \pm 7$  mil. y. (on biotite — E. FANINGER 1976).

The Neo-alpine granitoids form hypabyssal subvolcanic bodies connected with late orogenic geosynclinal rhyolite-andesite volcanism. They occur in three areas and differ accordingly in both the age and geotectonic conditions. The areas of their occurrence are (1) the Rhodope massif, (2) Neotisia including the West and



Fig. 5 Geotectonic position of magmatites in the South Carpathian Mts. (Compiled on the basis of the CBGA Tectonic Map — edit. M. Mahef, 1973, and the Geological Map of Rumania)

**Tectonic units**

1. Moesian platform
2. Danubicum a) interior zone — Stara Planina
3. Geticum
4. Saska-Gornjac tectonic slice — Kraishtides
5. Ultrageticum — Morava—Penkovski nappe
6. Serbo-Macedonian Massif
7. Vardar Zone
8. Foredeep a) inner b) outer
9. Flysch Belt of the Carpathians; Koula Zone

10. Severin nappe; Ceahlău
11. Fore-Balkan
12. Metalliferous Zone (Mureş)
13. Meso-alpine volcano-sedimentary grabens (Timok type)
14. Depressions a) Palaeogene b) Neogene

**Magmatites**

15. Pre-Alpine granitoids a) synkinematic
16. Palaeo-alpine granites a) synkinematic
17. Banatites a) subvolcanic and hypabyssal bodies, b) volcanics, c) pyroclastics
18. Neovolcanics a) volcanics, b) hypabyssal bodies, c) pyroclastics
19. Ultrabasic rocks
20. a) basic intrusive rocks, b) spilite-diabases-keratophyres-porphyroites
21. Pre-Permian basites
22. Permian quartz porphyries
23. Pre-Permian a) spilite-diabases, b) porphyroids

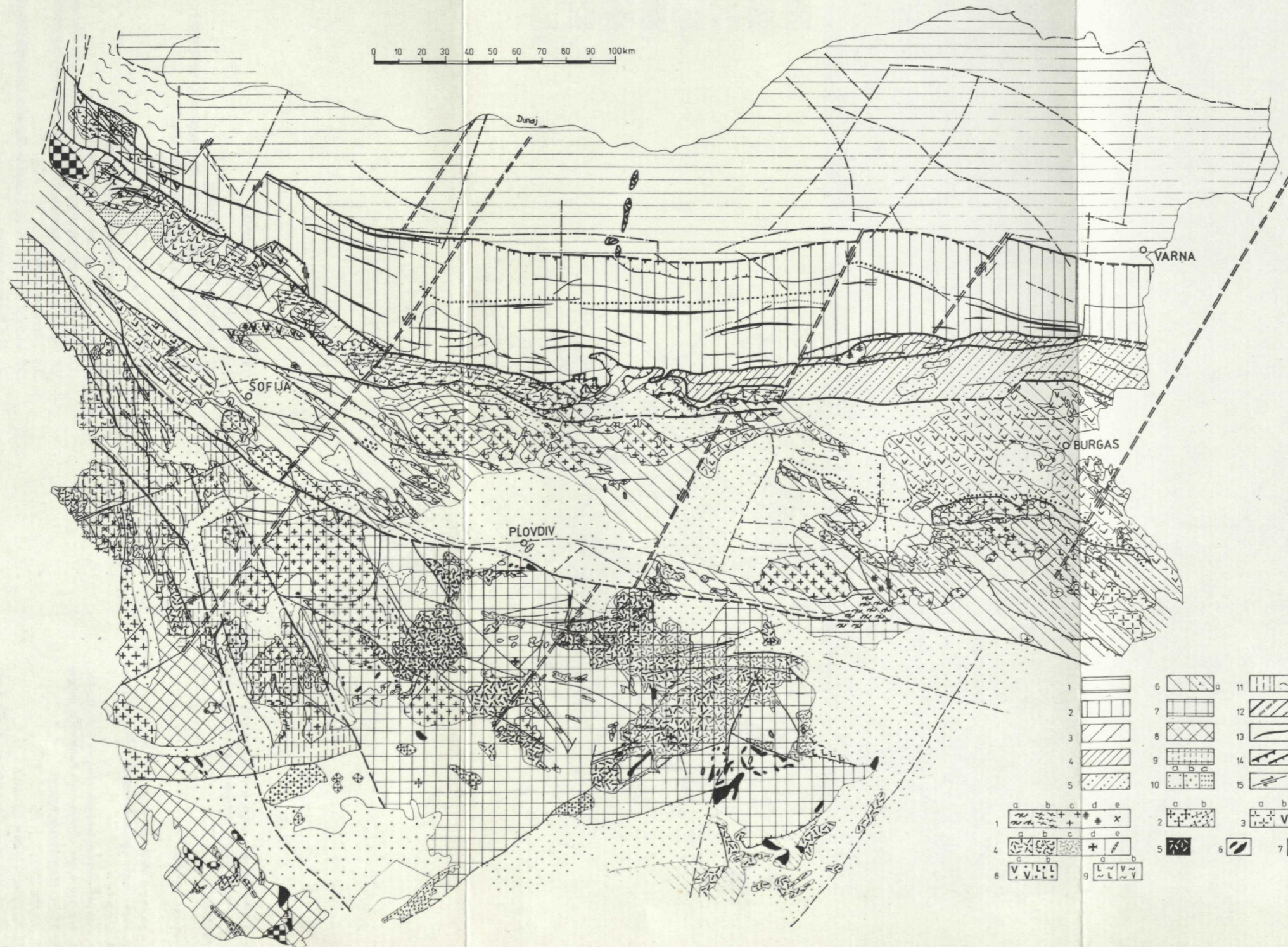


Fig. 6 Geotectonic position of magmatites in the Balkan Mts. (Compiled on the basis of the CBGA Tectonic Map — editor M. Mahef, 1973, and the works of E. Bončev et al.)

Tectonic units

1. Moesian platform
  2. Fore-Balkan
  3. Stara Planina
  4. Kotel Zone
  5. Luda Kamčia Zone
  6. Srednogorie a) Strandja
  7. Rhodope Massif
  8. Serbo-Macedonian Massif
  9. Kraishtides
  10. Depressions a) Neohercynian, b) Neopalpine-Palaeogene, c) Neogene
  11. Kraina unit a) Koula Zone
  12. Crustal deep faults, cryptoruptures, faults
  13. Axes of anticlines
  14. Overthrust lines
  15. Strike-slip faults
- Magmatites

1. Prealpine granitoids a) Prehercynian (synkinematic), b) Stara Planina type (mostly synkinematic), c) Hercynian (?) — South Bulgarian, d) Alkalic intrusives, e) Struma diorite formation
2. Palaeo-alpine granites a) synkinematic
3. Banatites a) intrusives, b) volcanics, c) prevailingly pyroclastics
4. Early late-geosynclinal a) predominantly andesites, b) prevailingly rhyolites, c) pyroclastics, d) subvolcanic bodies, c) dykes
5. Basalts
6. Serpentinities
7. Basic volcanics (predominantly basalts)
8. Neohercynian volcanics a) quartz porphyries, b) melaphyres
9. Pre-Upper-Carboniferous volcanics, a) metabasites, amphibolites, d) porphyroids



East Carpathians, the Apuseni and the Hungarian Mid-Mountains, (3) the marginal parts of the Serbo-Macedonian massif (Fig. 2)

1. In the Rhodope massif the plutonic rocks are found at hypabyssal to subvolcanic levels of the former volcanoes and in the crystalline basement of the Eastern Rhodope Mts. The Madžarovo, Zvezda-Galenica, Jolendere, Lozen and other plutons are of Late Oligocene and sporadically Early Miocene age, as evidenced by radiometric values of 36—45 m. y. (S. BOYADJIEV 1974) (Fig. 6).

2. In Neotisia there are minor, most frequently subvolcanic granitoid bodies, located in the centres of Miocene eruptions. In the West Carpathians of Slovakia (Fig. 3) (around Banská Štiavnica, in the Polana and Javorie Mts.) granodiorite-porphry, granodiorite and diorite laccoliths, radiometrically dated at  $2.5 \pm 0.8$  — 15 m. y. (V. KONEČNÝ—J. SLÁVIK 1974) are present. Minor diorite bodies follow linear faults in the Tisovec area, amidst the Veporide crystalline zone.

In the Transcarpathian Ukraine (U.S.S.R.) diorite-porphyrates, granodiorites and microgranites (14.2—8.2 m. y.) occur chiefly in the Vyškov, Vinograd, Velika Dobroň, Perečin and Slavjansk areas (B. V. MERLIĆ—S. M. SPITKIVSKAJA 1974). In the Rumanian young volcanic areas diorites form major bodies in the Hărghita and Călimăni Mts.; the diorite body in the Călimăni caldera is of unusually large size. At the northern margin of the Hungarian massif, in the Reczk-Lahóca area in the Mátra Mts., occur Upper Eocene hypabyssal and subvolcanic diorites and diorite porphyrites along the Darnó line, which is one of the major deep faults of NNW-SSE trend (T. ZELENKA 1974).

3. Along the margins of the Serbo-Macedonian massif and at the interior margin of the Vardar zone a number of Late Oligocene and Miocene quartz monozite and granodiorite bodies are situated (Fig. 9), being accompanied by andesite-dacite lavas and pyroclastics. These bodies occur in the boundary zone between the crystalline Serbo-Macedonian massif and the Mesozoic eugeosynclinal area rich in ophiolite volcanism. G. DELEON (1968) reported Rb/Sr ages for some major bodies, such as Jastrebac (37 m. y.), Surdulica ( $25 \pm 2$  m. y.), Kremenici, Vrsač (29 m. y.), Motajica 23—17 m. y.), Cer (10.8 m. y.) and Bukulja (23.5—13.1 m. y.). Some of them gave even higher age values (e.g. 36 m. y. — S. BOYADJIEV 1974).

## Discussion

From this overview of granitization in the Alpides several conclusions can be drawn:

- a) Most granitoids of the Alpides are of the Hercynian age;
- b) the granitoids, particularly those of Hercynian age developed during a wide time interval, which points to a polyphasal and polycyclic granitization process;
- c) there is a direct relationship between the position of granitoid bodies and the tectonic style;

d) the granitization is distributed unevenly not only in the Alpidic segments but also in the individual zones of one segment.

a) The Pre-alpine granitoids usually occur amidst higher-metamorphosed crystalline complexes, most of which are dated as pre-Hercynian. Since they are frequently connected with their mantle rocks through migmatites, they were until recently thought to be of pre-Palaeozoic (pre-Hercynian) age. Radiometric ages, however, have shown that most of the granitoid bodies date from the Hercynian period. The palaeontologically evidenced Early Palaeozoic age of the rocks forming the mantle of these bodies also suggests their Hercynian age. As a result, the views have so changed in recent years that, for example, only a smaller part of granitoids of the West Carpathians are thought to be of pre-Palaeozoic age.

The Hercynian granitoids accompany the Hercynian geosynclinal troughs in the central Alpidic (Rhodope, Serbo-Macedonian, Pelagonian and East Hungarian) massifs. They are also abundant in the adjacent Bohemian Massif. The Hercynian granitization along with the formation of the granite-gneiss layer of the earth's crust is of wider, maybe global character. In the Alpidic region of Europe, the Hercynian granitization was connected with the most relevant pre-Alpine palaeogeographic changes in the history of the Alpidics, i.e. the formation of the Hercynian mountain ranges and of longitudinal depressions filled with the first, late Hercynian molasses.

Hercynian granitization and the resulting consolidation of the earth's crust within the present-day Alpidic region were not uniform.

Granite bodies formed in the Alpine orogenic cycle, especially those associated with the Alpidic folding, i.e. with the consolidation of the earth's crust are rather rare and are confined to a limited number of zones. Only granitoids linked closely with andesite and rhyolite Alpidic volcanism are of greater extent.

In the pre-Alpine and Alpine ophiolite zones especially of the Dinaride and Hellenide segments, minor granitoid bodies occur as differentiates of the basic magma.

b) The wide time-range of granitization process is specially noteworthy for the present-day Alpine region. The Hercynian granitoids gave ages from 340 (320—300 most frequently) m. y. to 220 (200) m. y. The granitization process would thus cover the whole Late Palaeozoic, beginning in the Late Devonian at some places. This interval roughly coincides with the duration of the Hercynian folding, from its first (Bretonian) phase towards the end of the Devonian to the Pfalzian phase at the end of the Permian. The period of Hercynian diastrophism of about 100 m. y. is long compared with the preceding Palaeozoic quiescent eugeosynclinal history (the Ordovician or Late Cambrian to the end of the Devonian). At the same time, one cannot omit the signals heralding the Caledonian folding, perhaps also with manifestations of granitization process (E. JÄGER 1973, I. HAJDUTOV 1971).

Also at the eastern margins of the Serbo-Macedonian massif, granitization of the Alpine cycle accompanied by high-pressure metamorphism was active in several

phases within the time range from 170 to 40 m. y. (F. KOCKEL—H. MOLLAT—H. W. WALTHER 1971), i.e. from the beginning of the Jurassic ('late Cimmerian' folding) through Middle Cretaceous to Late Eocene times (J. AUBOUIN 1964). In the Penninicum of the Alps granitization persisted until the beginning of the Oligocene and, surprisingly, together with high-temperature metamorphism. This clearly shows that Alpine granitization also covered a long interval relative to the quiescent Alpine geosynclinal stage in the Triassic and Lower to Middle Jurassic lasting only 80—100 m. y. However, it should be stressed that granite-forming processes of Alpine age were limited only to narrow zones. Extensive areas affected by these processes occur in the southern parts of the Alpides, i.e. in the Pelagonian massif, southern parts of the Serbo-Macedonian and Rhodope massifs and in the Sakar and Strandža blocks. The Hercynian and even earlier granitoid bodies of these areas yielded radiometric ages mainly of 160—97 m. y., which obviously point to the Alpine regeneration and reworking of the crystalline basement. The Alpine reworking also led to the metamorphism of Triassic, Jurassic and Lower Cretaceous series. Some older granitoid bodies of these areas gave even smaller radiometric ages 50 to 30 m. y., which correspond to the Priabonian-Oligocene and in places to the Lower Miocene magmatic activation (S. BOYADJIEV 1974).

The stage evolution, which is so striking in Alpine and Hercynian foldings, is less conspicuous in granite-forming processes. The three major stages in the Alpine cycle (in the sense of M. MAHEL 1974) are the Palaeo-alpine (including Late Cimmerian) stage — Upper Jurassic to Middle Cretaceous; Meso-alpine stage — the latest Cretaceous to the end of the Eocene (or the beginning of the Oligocene), and the Neo-alpine stage (for the most part Neogene). Only two first stages are accompanied by granite generation.

The Hercynian cycle includes the Bretonian-Sudetic, Asturian-Saalian and Pfalzian stages (M. MAHEL 1975). The first is often termed as the early stage with the radiometric date range 350—305 m. y. (S. BOYADJIEV 1974) and the other two as late stages (Dimitrova et al. 1975) with dates 280—230 m. y. The origin of migmatites and synkinematic granitoids in the Alpine region is often placed in the first stage, that of late and post-kinematic granitoids in the second stage and the formation of aplites, pegmatites, albite granites and subalkalic granites in the third stage (A. K. BOJKO 1975). This evolution trend of the Hercynian granitoids and migmatites associated with them is perceptible chiefly in the West Carpathians and the Balkan Mts.; it is in agreement with the proceeding consolidation of the earth's crust during Hercynian diastrophism. However, it is not of general validity, as synkinematic granitoids radiometrically dated as Permian with ages about 270—230 m. y. (e.g. granite-gneisses in the Nerešnički body, South Carpathians of Yugoslavia, or migmatites from the High Tatra in the West Carpathians) and some pegmatites from the Low Tatra with age values of ca. 330 m. y. can hardly be regarded as exceptions to the rule. Within the framework of the above-mentioned evolution, the synkinematic,

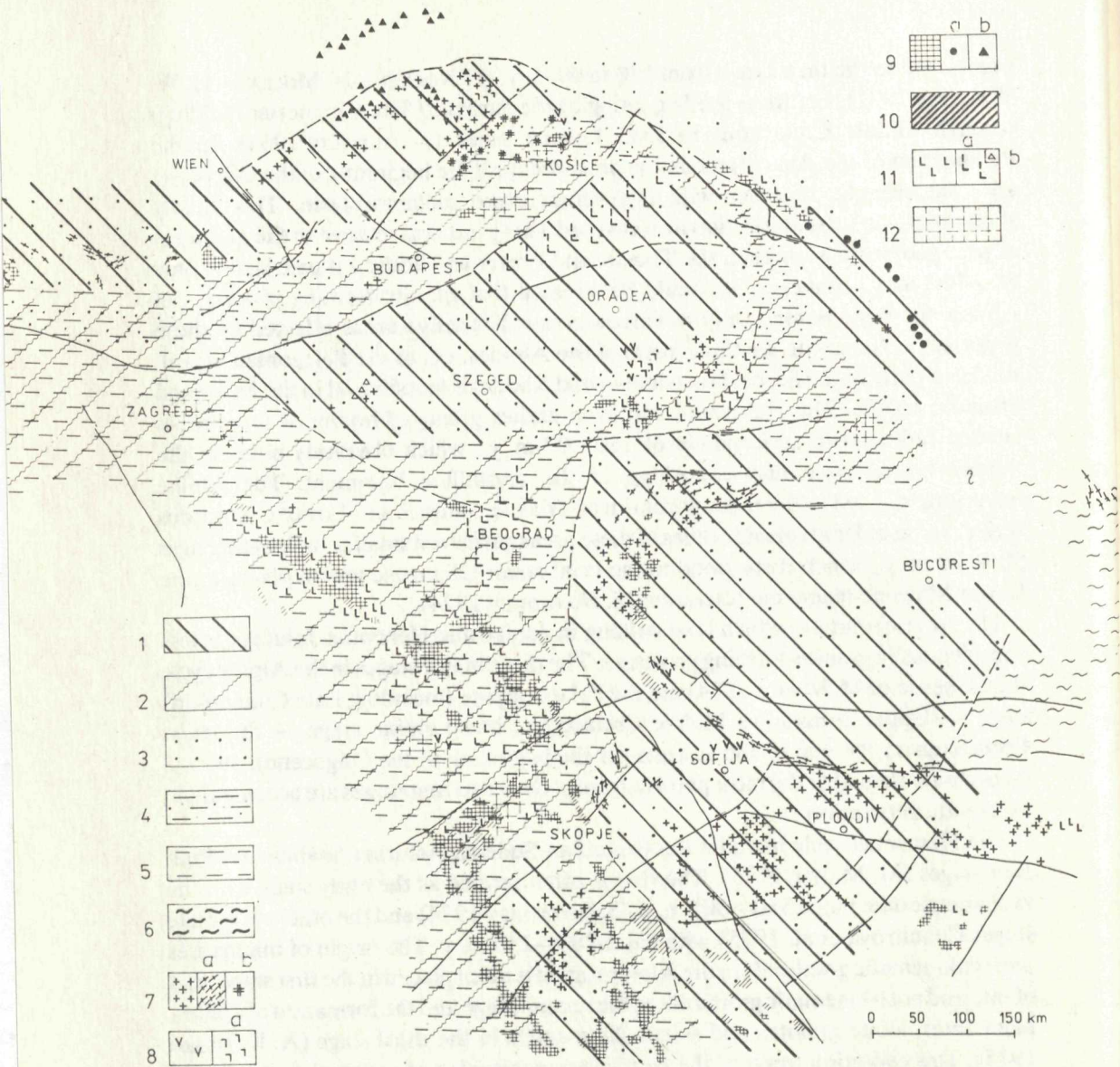


Fig. 7 Sketch showing differences in the Hercynian consolidation in relation to the type of Triassic development and to alpine magmatites

Regions of different Hercynian consolidation: 1. Regions of intensive Hercynian consolidation, 2. Regions of slighter Hercynian consolidation. Types of Triassic development: 3. Quasi-platform type (continental type of crust), 4. Austro-alpine type, mobile shelf with basins, 5. Dinaride-paraliogeosynclinal type, 6. Tauride Flysch type (with flysch tectofacies). Magmatites: 7. Hercynian granitoids, a) synkinematic, b) alkalic, 8. Permian melaphyres (in nappe position), a) quartz porphyries, 9. Ultrabasites a) small bodies — mostly in nappe position, b) alkalic, 10. Gabbros (predominantly accompanied by diabases), 11. a) Spilites-diabases, partly the hybrid association b) diabase-porphyrite-quartz porphyrite, keratophyre, 12. Triassic diabase-porphyrites, keratophyres-quartz porphyrites.

late or post-kinematic types of granitoids depend also on the depth and the pressure and temperature conditions under which they originated. The most intensive granitization of the Alpine orogenic era giving rise to the synkinematic granitoids in the Penninicum took place as late as the end of the Eocene. All these findings suggest that in the general tendency of the evolution of granitoids from synkinematic types to aplite-pegmatite dykes there were marked variations due to evolutionary peculiarities of individual zones, both Hercynian and Alpine. Consequently, the variety in evolution, composition and structure of the zones is a highly characteristic feature of the European Alpides.

Many authors assume not only a polyphasal but also polycyclic formation of granitoid bodies (M. D. DIMITRIEVIĆ 1967, S. KARAMATA 1974, I. ZAGORČEV 1975, E. SZÁDECKY-KARDOSS 1967, S. BOYADJIEV 1967). This concerns especially the large granitoid bodies in the Rhodope and Srednogorie Mts. of South Bulgaria, the granitoids in the Serbo-Macedonian and Pelagonian massifs and partly also those in the South Carpathians, where radiometric dates indicate pre-Hercynian, more often Hercynian and even Alpine ages (e.g. in the Bujanovac pluton in the Serbo-Macedonian massif, in the Belogradčik pluton in the Balkanides, in the Pelagonicum or in the Mecsek Mts.).

c) In the Alpine area there is a striking difference between the Hercynian and the Alpine orogenic cycles in the granitization — folding relationship. Hercynian diastrophism was accompanied by deep-seated processes, i.e. by extensive granitization and metamorphism. The Alpine folding process gave rise to weak granitization and limited extent of metamorphism, in deeper layers accompanied by diaphthoresis. These manifestations are centred mainly to relatively narrow zones, such as the Penninicum of the Alps. On the whole, the Alpine folding is of near-surface character with a dominance of nappe structures. The differences between the Hercynian and Alpine orogenic cycles also concern the structure and type of the earth's crust.

The differences in the degree of metamorphism and granitization in the Alpine segments and even in individual zones of one segment are also reflected in different tectonic styles. To lend support to this contention, three examples of Hercynian structures will be given: In the High Tatra Mts. of Slovakia and Poland (Fig. 3), where occur extensive granitoid bodies, there are broad Hercynian anticlines and synclines. Domal structures are typical of the Alpine areas with huge granitoid batholiths, for example, in the Rhodope Mts. (I. BOJANOV 1973, I. ZAGORČEV 1975) and in the Serbo-Macedonian massif (M. D. DIMITRIEVIĆ 1967). On the other hand in the Kraklová zone of the Veporides and in the Čierna hora Mts. in Slovakia (S. JACKO 1971) tight compressed Hercynian folds were established in the mica-schist zones, where granitoid bodies are small, rare or absent altogether. G. H. KRÄUTNER (1937) reported even Hercynian nappes from the East Carpathian areas containing a negligible amount of Hercynian granitoids (Fig. 11a). The majority of granitoid

domes and cupola-shaped bodies are mostly without any relationship to faults. Although their position is not necessarily conformable with the mantle, most often the mantle is adapted to their contours. The margins of cupolas are steep, often with a narrow zone of migmatite-transition to the mantle. At the margin of granitoid bodies gneissose-granite textures are found. Late tectonic granites form dykes, often of boudinage structure, which occasionally run transversely to the mantle.

d) From the differences in the types and distribution of granitoids in individual segments, and particularly in the proportion of Hercynian and Alpine granitoids in their structure, some features of wider significance can be inferred:

In the West Carpathians (Fig. 3), the abundance of Hercynian synkinematic granitoids (including migmatites) and late kinematic granitoids, aplites and pegmatites in the Tatríde and Veporíde crystalline basement is striking. It is a granodiorite-granite association of relatively low variety in composition. High-temperature conditions and injections of granitoids not only into pre-Palaeozoic but also Palaeozoic sequences have been evidenced. The uneven grade of Hercynian metamorphism (from greenschist facies to amphibolite facies in the zones of higher granitization) is obviously related genetically with the extent and type of Hercynian granitization. Large granitoid bodies are distributed only in two zones, i.e. in the Tatríde zone and the southern part of the Veporíde zone; the north Veporíde is poor in Hercynian granitoids. A Permian molasse graben rich in quartz porphyry is superimposed on it. The southerly parts of the West Carpathians — the Gemerides — were affected by Hercynian granitization to a lesser extent. A substantial proportion of basites and ultrabasites in the Upper Carboniferous and the beginning of the early (Triassic) activation leading to the formation of the Alpine geosyncline might correspond to the slight granitization in the Gemeride zone. In the West Carpathians even the distribution of Hercynian granitoids reflects a distinct zonation (Fig. 16).

In the Apuseni Mts. (Fig. 4) Hercynian granitoids are most extensive in the extreme north, in the Bihor autochthon with the big Muntele Mare body. They are sparsely distributed in the more southern nappe units. Of a different type are the granitoids of the banatite association, which do not show any pronounced geotectonic relationship to the preceding Palaeo-alpine sedimentary trough, but extend transversely to the Palaeo-alpine structure. The East Carpathians contain a small amount of both Hercynian and Alpine granitoids, apart from the extensive alkalic Ditrău body, whose genesis and character are still unclear. The greater distribution of basites in the Upper Carboniferous can be accounted for by the minor formation of the Hercynian granitoids. A small amount of Upper Palaeozoic (including Permian) molasses is probably due to a weak dissection of the Upper Palaeozoic relief, which (according to our experience) indicates lack of granitoid bodies. The Palaeozoic sequences are in most cases affected only by lower-grade Hercynian metamorphism. Some authors, however, believe that metamorphism is of uneven

character also in the East Carpathians and that the upper parts of higher metamorphosed gneiss and mica-schist series, such as the Bretila Rarău and Rebra-Barnar Formations, are of Palaeozoic age, too (A. K. BOJKO 1975). Major granitoid bodies occur in the northern areas partly covered by the nappes translated from the southern crystalline zones. The smaller depth at which the Hercynian folding occurred in the southern zones and the absence of large granitoid bodies is also demonstrated by the presence of Hercynian nappes (H. G. KRÄUTNER 1975), i.e. by the structural forms distinctive of the near-surface folding.

In the South Carpathians the synkinematic granitoids are (Fig. 5) relatively extensive both in the Geticum and Danubicum and the bodies of alkalic character are more numerous, particularly in the Danubicum. Of interest is the difference in the geotectonic type of 'banatite' bodies: in the southern part of the South Carpathians they are interorogenic, near to those of the Srednogorie of Bulgaria, and in the north they are late orogenic associated with the superimposed structures. This difference can be accounted for by more intensive folding subsequent to the Late Cretaceous in the southern parts. The change in the geotectonic type of banatite bodies is accompanied by changes in the petrochemistry of volcanics, i.e. by a decrease in alkalic types in the northern zone.

In the Balkanides (Fig. 6), the granitoids are of a large extent and a great variety of geotectonic types and petrography. The largest bodies are in the Rhodope Mts. North of the South Bulgarian granitoid zone extends the Balkanide type with elongated bodies of the gabbrodiorite-granodiorite association in the Stara Planina Mts., which displays an appreciably more varied petrographic composition. No less varied are the diorite Struma Formation with intrusives ranging from ultrabasites to acid granites, the Permian gabbro-syenite association and the Meso-alpine gabbromonzonite-granite association. This unusual variety is apparently the result of a high palaeotectonic differentiation of the area bearing linear geosynclines (after E. BONČEV 1966) disposed between the zones of a thicker earth's crust. Major plutonic, hypabyssal and subvolcanic bodies of granite, monzonite, syenite and gabbro in the Oligocene and Lower Miocene volcanic areas and the crystalline basement of the Rhodope massif are genetically associated with the regeneration of its ancient basement.

The Dinarides are distinguished by a lack of Hercynian and Alpine granitoids (Fig. 9), except for the bodies in the marginal sectors, which testifies to the presence of the granitic layer. The occurrences of granitic differentiates of basic and ultrabasic magma are more numerous. Outstanding Alpine granitization and metamorphism at the contact with the Pelagonian and the Serbo-Macedonian massifs, the frequency of young granodiorite and monzonite bodies at the western margin of the Serbo-Macedonian massif, i.e. in the boundary zone between it and the Vardar eugeosyncline, are other noteworthy features of the Dinarides.

The peculiar characters of granitoids in the individual segments of the Alpides

have been shown to be connected with the specific history and structure of these segments or their separate zones. The following chapters discuss the influence of the Hercynian granitization on the distribution of late Hercynian geosynclinal volcanics and of the Alpine ophiolite zones in the Mesozoic (Fig. 7).

## 2. Late Hercynian geosynclinal volcanics

Late Hercynian volcanics are a geotectonically particular type, associated genetically with the late Hercynian molasse grabens. They are usually represented by quartz porphyries (palaeo-rhyolites; predominantly pyroclastics) accompanied in places by porphyrites (palaeo-andesites) and by melaphyres (palaeo-basalts). These volcanics are distributed unevenly in the Alpine segments.

In the Eastern Alps, the occurrences of quartz porphyry and its tuff are sparse and those of melaphyre sporadic (near Hallstat). In the Southern Alps the ignimbrites of quartz porphyry are known from the Pohorje Mts. (Yugoslavia). Major quartz porphyry layers in the Carnic Alps separate the lower (conglomerate) horizon from the upper detrital sequence of the Permian Gröden Formation (H. V. FLÜGEL 1975).

In the West Carpathians, the late Hercynian volcanics (Fig. 3, Fig. 8) are widely distributed in several Hercynian molasse grabens. They are less extensive in the northern zones; quartz porphyries occur in the Považský Inovec, Trábeč and Branisko Mts. Quartz-porphyry tuffs and lava flows are more frequent in the marginal grabens of the Veporides, in the northern Lubietová zone and in the contact zone between the Veporides and Gemerides. In the southern part of the North Veporide graben there is metamorphosed Permian with metabasites, melaphyres and porphyrites (A. KLINEC 1966).

Quartz-porphyry is accompanied by porphyrites at the southern margin of the Čierna hora Mts., at the contact with the Gemerides, amidst the Permian molasse (S. JACKO 1971). Dykes of quartz porphyry are known from several places in the Tatride and Veporide Crystalline. The middle part of the North Gemeride Permian contains very thick quartz porphyry layers (predominantly tuffs and tuffites) above the basal Verrucano conglomerates and breccias. Quartz porphyry lava flows predominate in the lower parts of the volcanogenic sequence and pyroclastics are dominant higher up. Porphyrites, quartz porphyrites and diorite-porphyrites are subordinate.

The Choč nappe in the Slovakian West Carpathians differs in the type of its volcanics from other units. A several thousand metres thick Permian molasse which encloses melaphyre bodies, was obviously originally located along a fault zone, several hundred kilometers long. The huge subaqueous melaphyre effusions occurred in two phases (J. VOZÁR 1971); the former (probably Lower Permian) is more acidic (up to 62 %  $\text{SiO}_2$ ) with a negligible proportion of explosive material, the latter is of Upper Permian age, of higher basicity (44—54 %  $\text{SiO}_2$ ) and richer in



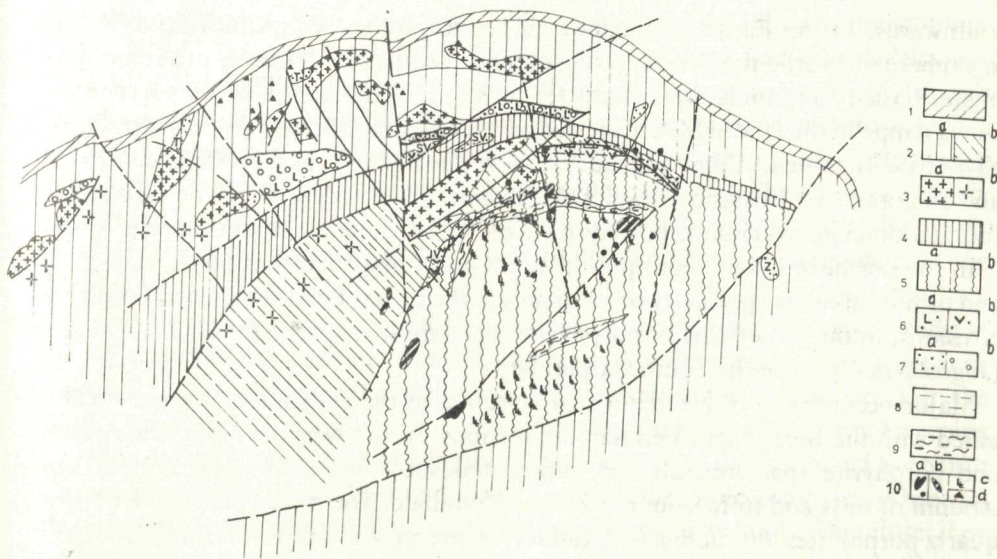


Fig. 8 Zones of different Hercynian stabilization and the distribution of Permian types in the West Carpathians

1. Klippen Belt
2. Zones of intensive granitization and intensive Hercynian folding, a) predominantly autochthonous (Tatrides, Hungarian mass), b) most commonly parautochthonous-allochthonous (South Veporides)
3. Granitoid massifs a) on the surface, b) in the basement of Neogene mantle
4. Zones of intensive Hercynian folding but without large-scale granitization (North Veporides)
5. Zones of minor Hercynian stabilization a) with a small portion of Hercynian granitoids, b) very slightly folded and lacking granitoids
6. Permian a) melaphyres (in nappe position), b) quartz porphyries
7. Permian a) terrestrial type; T = Tatride, SV = North Veporide, JV = South Veporide, Z = Zemplin, b) of the Choč nappe
8. North Gemeride Permian
9. Permian, partly marine, marginal Rožňava-Železník type and Bükk type
10. Ophiolites a) ultrabasites, b) intrusive basites, c) diabases, d) quartz porphyries-keratophyres-diabases

pyroclastics. Melaphyre bodies are accompanied by diorite- and gabbrodiorite-porphyrite dykes, channelways and subvolcanic diorite bodies. The volcanics were probably ascending on the longitudinal faults, which also controlled the formation of Permian grabens. Lower Triassic quartz porphyries are reported from the neighbouring Vernár and Drienok units.

In the Rumanian Apuseni Mts. the Permian volcanics are prevalently components of the nappe systems. Noteworthy is the increase in the basic volcanics (melaphyres)

southwards. In the Finiş-Gîrda nappe quartz porphyries — ignimbrites, tuffs and agglomerates overlie the Permian conglomerate sequence. Volcanics of a somewhat higher basicity and melaphyre character form, for example, the upper parts of the Dieva nappe in the Codru Mts., its lower parts consisting of more acidic quartz-porphry type. In some sections the sequence with quartz porphyry is divided into a lower and an upper part by basic volcanics (M. BLEAHU 1974). Melaphyre and melaphyre tuff predominate over quartz porphyries in the more southerly Moma nappe (Fig. 4).

In the northern (Rumanian) part of the South Carpathians, agglomerates, tuffs and tuffites of quartz porphyry occur mainly in the Lower Permian. Quartz porphyry effusions, in the upper part of porphyrite type, are more frequent in the southern (Jugoslavian) part of the South Carpathians.

Major occurrences of Neohercynian volcanics in the Balkanides are genetically allied with the Stephanian-Permian depressions. In the western block, melaphyre and porphyrite (palaeobasalt and palaeo-andesite) lava flows with a minimum amount of tuffs and tuffaceous breccias predominate. The more acid porphyrites to quartz porphyrites alternating with tuffaceous breccias and tuffs forming part of the Belogradčik and Central-Balkan anticlinoria are somewhat younger, predominantly of Lower Permian and partly probably of Upper Stephanian age. The volcanics of the third phase (Lower Permian), which are present mainly in the eastern block, including the Srednogorie, and less in the western part of the Balkanides, are most extensive. Small basins are filled with tuffaceous breccias, tuffs, tuffites accompanied by quartz porphyry and quartz porphyrite (palaeodacite to palaeorhyolite) lava flows. The appreciable part of Permian magmatites are of volcanic type: subvolcanic domal bodies and microgranite and granite-porphry dykes are also known. The Upper Carboniferous-Permian volcanism of the Balkanides is broadly of linear character, being confined to longitudinal faults that controlled the formation of depressions (D. ČUNEV—P. BONEV 1975).

In the Dinarides, Permian and perhaps also Upper Carboniferous quartz porphyries are sporadic in the Central Bosnian mountain range (S. KARAMATA 1974).

Quartz porphyries of great thickness and abounding in pyroclastics, which are accompanied by porphyrites and assigned to the intra-Permian diastrophism, occur in the eastern block of the Hungarian Mid-Mountains, in the ranges Mecsek and Villány. The aplite dykes that cut across the Upper Carboniferous complex in the Mecsek are probably genetically associated with the quartz porphyries (B. JÁNTSKY 1976). In the western block sporadic quartz porphyries are in the Permian of the Balaton area (E. SZÁDECKY—KARDOSS et al. 1967).

## Discussion

1. Late-geosynclinal late Hercynian volcanics, predominantly of Permian and partly Upper Carboniferous age (Balkanides, South Carpathians) accompany the

complexes of upper, mainly red (lagoon-terrestrial) molasse filling the longitudinal depressions. The formation of the depressions or grabens and of volcanics was controlled by deep faults and connected with the epeirogenic extension of the crust and the genesis of rifts, which completed the Hercynian geosynclinal process.

2. The Neohercynian volcanics are little varied in composition: quartz porphyries with a small amount of pyroclastics predominate and melaphyres are relatively abundant. The transitional intermediate rock types between these two boundary members are strikingly few and are not distributed uniformly in all segments of the Alpides.

In the Balkan region, basic melaphyres give way to the more acid types, whilst in the Apuseni Mts. an opposite trend is observed. A difference also exists between their northern zones (quartz porphyries) and southern zones (melaphyres). In the West Carpathians, melaphyres are characteristic of the Choč graben intervening between the quartz porphyry zones (the Veporide zone in the north and the North Gemeride zone in the south). In all cases, however, the geotectonic type is broadly the same, associated with molasse-filled grabens bounded by faults. Therefore, they can be reasonably assigned to the same basalt-rhyolite formation in both the Balkan (E. DIMITROVA et al. 1975) and the West Carpathians regions (D. HOVORKA 1965). The more basic volcanics are possibly linked with a thinner crust or a greater depth of the faults, on which the magma was ascending.

Melaphyres cannot at all be regarded as analogous volcanics to ophiolites, i.e. as initial volcanics in the sense of Stille (1948), see J. VOZÁR (1972). Melaphyres exert only little influence on the type of sedimentation and are not associated with the early stage of geosyncline development as is the ophiolitic suite, but with the molasse stage at the transition from the Hercynian to the Alpine cycle. However, the opinion that they are the final member of the Hercynian volcanic suite is not convincing either, since they occur chiefly in the southern zones, which were less affected by the Hercynian folding.

The slight alkalic trend of melaphyres in the Choč nappe of the West Carpathians (J. VOZÁR) would suggest a weaker cratonization of the crust. In the West Carpathians and Apuseni Mts., however, the melaphyres are associated with grabens, which indicate the beginning of the Alpine structural-facies zones.

3. A striking feature is the dominantly acid character of the Neohercynian volcanics. The time coincidence of their emplacement with the last phases of granitoid intrusions and dying-out Hercynian metamorphism is also noteworthy. This would suggest a relationship between the acid volcanics and the last Hercynian granitization stages, but no direct evidence for a volcano-plutonic formation is available. Subvolcanic (microgranite) bodies are known especially from the Balkanides. The distribution of the volcanics in marginal grabens, usually adjacent to the zones with granitoid intrusions, rather indicate an indirect relationship. They presumably represent effusions from the marginal parts of major magma reservoirs.

A slight metamorphic alteration of Permian complexes and the abundance of vein quartz suggest an aftermath of hydrothermal activity. The sedimentary formations of molasse type as well as the waning of volcanism and metamorphism point to the terminal stage of the cycle. When the Permian volcanism died out, a quasi-platform regime set in in the prevailing part of the Alpides. This is another reason why the melaphyres can hardly be regarded as the products of initial volcanism.

A quite different situation existed in the southernmost zones of most Alpidic segments and in the greater part of the Dinarides. The Permian of predominantly marine type is there diversified only by interlayers of continental facies; late geosynclinal volcanic manifestations are also isolated. The Hercynian geosyncline was passing into the Alpine geosyncline, undergoing minor changes.

### 3. Pre-alpine eugeosynclinal volcanics

Most of the pre-Carboniferous series of the Alpidic region are volcano-sedimentary sequences including geosynclinal volcanics of basic character. Most of them have been converted into amphibolites and in the oldest series of some segments (in the Rhodope, the Serbo-Macedonian and Pelagonian massifs) into eclogites. They are accompanied by plutonic bodies of gabbro and diorite and quite often also of ultrabasites. Acid volcanics metamorphosed to porphyroids occur to a small or greater extent in some sequences. The study of the stratigraphic position and the regional distribution of the volcanics in the individual segments of the Alpides enabled us to recognize some rules governing their evolution and the peculiarities of individual areas and zones.

Amphibolites, predominantly gabbros, are abundant in the Crystalline of the Eastern Alps. Serpentinites are also present. Basic volcanics of the spilite-diorite association are frequent in the Lower Palaeozoic, mainly Ordovician complexes (Gurtaler Decke, Saualpe, Grazer Paleozoik — P. BECK-MANNAGETTA 1974). Acid and intermediate volcanics, particularly quartz porphyries occur in the Lower Palaeozoic of some zones (Gurtaler Decke, Grauwackenzone, Zieberger Paleozoic), but they are more frequent in the Upper Ordovician to Lower Silurian (H. V. FLÜGEL 1975). Basic volcanics are more abundant in the Lower Devonian, for example, in the northern Grauwackenzone.

In the Karawanken and the Carnic Alps, the earlier Palaeozoic sedimentary (pelitic-psammitic) series enclose fairly frequent basite bodies, even serpentinized pyroxenites. Quartz porphyries, up to 500 m in thickness, occur in the Ordovician. Basic and acid volcanics are also found in the Lower Carboniferous flysch, which replaced the carbonate sedimentation (H. V. FLÜGEL 1975).

The pre-Hercynian and especially Lower Palaeozoic sequences of the West Carpathian Crystalline contain amphibolites derived from diorites, diorite tuffs

and tuffites accompanied by intrusive gabbros and diorites. Metabasites build up conformable, lens-shaped, tabular and even slabby bodies amidst the metasediments (D. HOVORKA 1976). The sequences rich in amphibolites are often affected by anatexis (i.e. penetrated by migmatites), which is related to deep faults on which the magma was ascending; such are, for example, the upper gneiss complex in the High Tatra Mts. and the broad migmatite zone of the Ďumbier, Ľubietová and Kohút crystalline complexes of central Slovakia. D. HOVORKA (1976) reports frequent occurrences of migmatitic amphibolites with an increased content of elements typical of granitoids. Granitization processes strongly affected gabbros near Dobšiná and in the Devonian Rakovec Group in the Gemerides during the Hercynian (L. ROZLOŽNÍK 1974) and Alpine orogenies (P. GREČULA 1973). Metabasites are unusually abundant in the Lower Palaeozoic Pezinok-Pernek Formation of low to medium metamorphic grade in the Little Carpathian Crystalline, in the Kraklová zone of medium-grade metamorphism in the Hron Formation of the Veporide Crystalline (A. KLINEC 1966). Quartz porphyrites altered to porphyroids occur in the upper (Devonian) part of this Formation. Metamorphosed acid volcanics are also found in lower layers of the Hron Formation. According to radiometric dating (zircon) their age value is 370 mil. y. (B. CAMBEL et al. 1977). In the North Veporide Crystalline, particularly in the Kraklová zone, there is a number of meta-ultrabasite bodies showing the character of serpentine-tremolite-talc-chloritic rocks or of peridotites. These members of a gabbro-peridotite suite represent ultrabasic differentiates of a tholeiitic type (D. HOVORKA 1976). The spilite-diabase formation of the Veporides and Tatrides is close to tholeiitic to Al-rich basalts in its chemical composition (D. HOVORKA 1976); probably an oceanic type of crust is concerned. They are less frequent in the Tatríde Crystalline. In the Veľká Lúka massif of the Malá Fatra Mts., hornblende peridotite occurs amidst a hybrid-granitoid mass (D. HOVORKA 1965) and minor serpentinite bodies of lherzolite-harzburgite type are known from the southerly Kohút zone of the Veporide Crystalline. The position of ultrabasites in the mica-schist complexes of the Kohút Crystalline near the deep Lubeník fault suggests that they may be later Alpine protrusions of the upper mantle, but a direct association with the Mesozoic was nowhere proved (Fig. 10).

In the southern zones of the Veporide and the Branisko crystalline complexes are even leptynite gneisses indicating that acid effusives may be present (L. KAMENICKÝ 1973). Metabasites forming sheet-like bodies, often only several cm thick but of considerable strike length, amidst the Muráň granite-gneisses instil the idea that the whole complex is of an effusive-sedimentary origin. Since the granite-gneisses may be a complementary member to basites in the spilite-quartz-keratophyre suite (D. HOVORKA 1976), it can be inferred that the Carpathian Crystalline contains a subordinate spilite-keratophyre or even more acidic association in addition to the spilite-diabase suite.

The presence of acid volcanics is characteristic particularly of the greater part of

the Gemeride Palaeozoic — of the Cambrian-Silurian Gelnica Group (Fig. 10). Radiometric data display the values (zircon) of 420 mil. y. from Dobšiná, 370 mil. y. from Mníšek (B. CAMBEL et al. 1977). Acid low-metamorphosed volcanics form several superposed layers. In addition to quartz porphyries and their abundant tuffs and tuffites, subordinate quartz keratophyres, quartz porphyrites and their pyroclastics, metadiabases and their tuffs and even gabbrodiorite bodies have been identified.

Unlike the Gelnica Group, the younger Devonian Rakovec Group abounds in basic volcanics (diabases, spilites and subordinate quartz porphyries, quartz keratophyres and more acidic porphyrite types), which represent subaqueous effusions of linear type, and occasional pillow lavas in the lower part. Volcanics in the central part of the Rakovec Group are of a more pronounced spilitic character (Š. BAJANIK 1975); they accompany gabbro, gabbrodiorite and quartz diorite intrusions in both the western (near Dobšiná) and eastern parts of the Gemerides. In the Klátov body, a marked transition from gabbros to diabase and diabase-porphyrite facies is seen. The Devonian-Lower Carboniferous Harmónia Formation in the Little Carpathians in south-western Slovakia contains a lower amount of basites.

Peculiar magmatites of the West Carpathians are the basites and ultrabasites in the Namurian (Viséan?)-Westphalian of the Gemerides; they are distinguished by a wide differentiation range, varied modes of occurrence and chiefly by the differing molassoid type of the formation in which they are situated. The formation comprises diabase tuffs, tuffites and diabase and diabase porphyrite bodies; submarine effusions, dykes and hypabyssal bodies of gabbro, gabbro-amphibolite, gabbrodiorite and diorite; bodies of serpentinitized lherzolite, which are even of large size such as Borčok near Brezníčka and Ploske (D. HOVORKA 1965). Minor bodies are more numerous in the area of Železník — Turčok (M. MAHEL 1953).

Early geosynclinal amphibolites occur in the Devonian of the eastern part of the Hungarian Mid-Mountains; in the western part of this block, in the Igal-Bükk geosynclinal trough, diabases and subordinate quartz porphyries are known from the Ordovician and Silurian sequences (G. WEIN 1969).

The following metamorphosed formations of the Apuseni Mts. are rich in acid volcanic rocks: the Upper Proterozoic-Lower Cambrian Arada Formation in the Bihor autochthon has a higher content of metakeratophyres and quartz porphyrites; the synchronous Biharia Formation in the Codru nappe system contains more ortho-amphibolite-metadiorites and metagabbros — and the Muncel Formation bears metamorphosed quartz porphyrites and metadiorites. In the Southern Apuseni area, amphibolites predominate in the Precambrian sequences and a wide range of volcanics from metadiabases to porphyroids, keratophyres with their tuffs and gabbros are present in the Palaeozoic complexes (Paiuseni Formation, M. BLEAHU 1974) (Fig. 4).

In the Eastern Carpathians, amphibolites are frequent in the meso-metamorpho-

sed Proterozoic Bretila-Rarău Formation and the Upper Proterozoic-Lower Cambrian medium- to low-metamorphosed Rebra-Barnar Formation (H. G. KRÄUTNER 1972). In essentials, the formations are of spilite-diabase type with a dyke suite and subvolcanic bodies of basic to intermediate character. The gabbroid magma was slightly alkali-calcic. The Lower Palaeozoic Tulgheş Formation of great thickness contains chiefly acid metatuffs, porphyroids, quartz porphyrites and keratophyres and in the upper part quartz porphyries and quartz keratophyres (A. SÂNDULESCU—I. BERCIĂ—H. G. KRÄUTNER 1974). The volcanics are accompanied by stocks of granite-porphry and granodiorite-porphry (G. RUDAKOV 1971). Basic volcanic rocks and tuffs form only several layers in the middle part of the Tulgheş Formation. They also occur (mainly metadiabases and their tuffs) in the later low-metamorphosed formations; in the Devonian-Lower Carboniferous Repedea Formation they occur together with the metamorphosed sequences of the Lower Carboniferous Rusaia Formation of detrital character. Other basite occurrences have been found amidst the limestones and dolomites of the Lower Carboniferous Tibău Formation (chiefly in its upper part) and in the Upper Carboniferous molassoid complex of the Argestru Formation and in the Radomirova zone.

The gneiss and mica-schist complexes of the oldest South Carpathian formations (Fig. 5) also show a volcano-sedimentary character. Spilite tuffs, tuffites and amphibolites are unusually abundant in the Baikalian complexes of the Geticum and Suprageticum. The gneiss complexes of the Poiana Ruscă Formation contains besides metadiorites and metagabbros also metaperidotite. Magmatic consanguinity of these ultrabasites and basites is apparent (O. MAIER et al. 1975). Serpentinites occur together with gabbros also in the gneiss-mica-schist Ielova Formation and in the Mraconia body of the Danubicum. The presence of acid tuffs has been proved in some gneiss and mica-schist complexes as, for example, in the Lotru and Sebeş Formations of the Geticum. Several layers of acid metatuffs are interlaid in the Crystalline of the Danubicum but the amphibolites predominate.

Basic volcanics, usually with an abundance of tuffs and tuffites are also in the Palaeo-hercynian formations. The Cambrian epimetamorphosed Corbu Formation of the Danubicum unit is rich in metadiabases and acid tuffs, lamprophyres, porphyries, gabbros and gabbrodiorites. The serpentinite bodies are of small dimensions, only the Tisovița body is of more importance (O. MAIER et al. 1975).

Among the varied volcanics of the Devonian epimetamorphosed series, chiefly in the Poiana Ruscă zone, predominate metadiabases and their tuffs over lamprophyres, acid tuffs and granodiorites (O. MAIER 1975). Acid pyroclastics are also found in the Devonian Lescoviț Formation together with the basic tuffogenic rocks (J. BERCIĂ—E. BERCIĂ 1975). Ultrabasic rocks are particularly frequent in the Ordovician to Silurian low-metamorphosed aspidic and flyschoid sequences in the south of the South Carpathians. In the Porečka zone in Jugoslavia, for example, there are diabases and gabbros accompanied by ultramafic rocks. Acid metatuffs, quartz

keratophyres and quartz porphyry dykes are more numerous in the Lower Carboniferous (M. SÂNDULESCU—S. NĂTĂSEANU—H. G. KRÄUTNER 1974).

In the Rhodope massif, amphibolite layers occur in the basal Lower Proterozoic (to Archean) complexes and especially in the Upper Proterozoic crystalline sequences. They are accompanied by serpentized pyroxenites, peridotites, dunites, gabbrodiorites and rare acid volcanics. The age and position of the ultrabasic rocks are difficult to establish. Some of them are closely associated with pre-Palaeozoic complexes but some serpentinite bodies at the northern margin near Plovdiv and at the eastern boundary of the West Rhodopean block follow the tectonic lines. Their contact is pronouncedly tectonic with both the pre-Palaeozoic complexes and the diabase-phyllitic series. The low-grade metamorphism of the serpentinites indicates a younger, possibly even Mesozoic age. The origin and amount of acid volcanics in the Rhodope Crystalline are uncertain; it is difficult to decide whether the orthogneisses are the product of anatexis or were primarily acid volcanics (I. BOJANOV 1974, S. BOYADJIEV 1967). Leptite gneisses would speak in favour of the latter hypothesis.

The crystalline complex of the Serbo-Macedonian massif having extensive basic volcanics and diapiric structures of basic and ultrabasic rocks shows a similar character (M. D. DIMITRIEVIĆ 1967).

Basic volcanic rocks markedly affected the character of younger sequences distributed in the marginal parts of the Rhodope and Serbo-Macedonian massifs, in the Krajištides and Balkanides, i.e. the Upper Proterozoic-Lower Cambrian Vlasina and Osogovo Formations and the Cambrian-Silurian diabase-phyllitic series (Fig. 6). Diabases, spilites and their tuffs are predominant; keratophyres, albitophyres and pyroclastics are subordinate and accompanied by gabbro intrusions and serpentized peridotites and pyroxenites. The pre-Carboniferous spilite-diabases in the Balkans display a distinct sodium tendency (E. DIMITROVA et al. 1975). Quartz porphyries of Riphean-Cambrian age occur at the eastern margin of the Serbo-Macedonian massif (S. KARAMATA 1974). The so-called Struma (diorite) Formation with gabbro, diorite and leucocratic granite bodies is genetically related to the diabase-phyllite series (I. HAJDUTOV 1971).

A particular complex of the Balkanides is the low-metamorphosed flysch Rajanovici Formation, which contains not only diabases but also porphyrites (I. HAJDUTOV 1971).

In the Dinarides (Fig. 9), the proportion of volcanics in the Palaeozoic complexes is larger in the inner units, mainly in the Vardar zone. The volcanic rocks are represented by diabases, partly spilites and gabbros, porphyrites, keratophyres and quartz porphyrites; the accompanying gabbrodiorites, gabbros and serpentized ultrabasites are also frequent. In the Lower Carboniferous diabases and tuffs are of a small extent (S. KARAMATA 1974).

In the Hellenides, chiefly in the Mirdita zone and less frequently in the Korab



zone, the Silurian-Devonian complexes contain diabases and subordinately porphyrites, quartz porphyries and porphyries. They are (genetically) associated with gabbros and diorites. The diabases are pierced by plagiogranite and pegmatite dykes (T. BIÇOKU et al. 1974).

## Discussion

1. The predominant part of pre-Palaeozoic and Early Palaeozoic formations of the area under discussion are of sedimentary-volcanogenic type with a prevalence of basic volcanics. These are represented by the spilite-diabase and spilite-keratophyre associations, frequently accompanied by intrusive bodies of gabbro, diorite, serpentized peridotite, pyroxenite and dunite. Acid effusive rocks, quartz porphyries and their pyroclastics and subordinately keratophyres occasionally occur as separate associations but usually form the accompaniment of other rock types.

Basic volcanics are particularly abundant in the Proterozoic (e.g. the Vlasina and Lisec-Osogovo Formations in the Rhodope and Serbo-Macedonian massifs, the Biharia Formation in the nappe units of the Apuseni Mts., Proterozoic sequences in the South Carpathians). Many Early Palaeozoic formations also contain a large amount of diabases (e.g. the diabase-phyllite formation in the Balkanides, the Devonian phyllite-diabase formation in the Gemeride Palaeozoic of the West Carpathians, the Repedea Formation in the East Carpathians, the Devonian-Lower Carboniferous Paiuseni Formation in the nappe units of the Apuseni).

In most instances the series are of volcanic-plutonic type and they contain even major intrusive bodies, sills or small stocks. In some zones the intrusives are greatly diversified. The presence of the granitoid rocks is in many cases attributed rather to the later granitization (e.g. in the phyllite-diabase series of the West Carpathians), but at many localities the genetic relationship between granitoids and basic rocks can be more intimate.

2. At many places also the ultrabasite bodies are constituent parts of the volcano-plutonic suite. Their magmatic consanguinity with the basic rocks is evident, for example, in the South Carpathian pre-Palaeozoic sequences at Poiana Ruscă, in the Geticum, the Ielova Formation and the Mraconia body in the Danubicum (O. MAIER et al. 1975) as well as in the Palaeozoic sequences as, for example, in the Corbu Formation in the Danubicum. J. JOVČEV et al. (1971) reported major ultrabasic bodies genetically associated with basites from the pre-Palaeozoic complexes of the Rhodope massif. A close relationship between minor serpentized pyroxenite bodies and gabbros and gabbrodiorites is apparent in the Balkanides (I. HAJDUTOV 1971). The conformable ultrabasite bodies in the areas with concentrated amphibolite occurrences as, for example, in the West Carpathian Palaeozoic (Hron Formation in the Kraklová zone of the Veporides) are explained more reasonably as

syngenetic components of the volcano-plutonic association than in terms of subsequent protrusions along the faults. This alternative, however, cannot be fully excluded in some cases, e.g. at the southern margin of the Veporide Crystalline.

3. Of interest is the occurrence of basic and ultrabasic rocks in the Lower Carboniferous (e.g. the Păiușeni Formation in the southern units of the Apuseni Mts., the Tibău Formation in the East Carpathians, the Lower to Middle Carboniferous flysch in the Dinarides) and even in the molassoid Namurian-Westphalian complexes of the Gemeride Palaeozoic (West Carpathians) and the Middle Carboniferous Argeștu Formation (East Carpathians). In all the above-mentioned segments of the Alpides the Hercynian folding, granitization and metamorphism and consequently also the formation of the granite-gneiss layer of the earth's crust, began at the commencement of the Carboniferous or towards the end of the Devonian. However, it can be postulated that the earth's crust remained thinner in the structural-tectonic zones distinguished by Carboniferous basites and ultrabasites. Most of these zones were evidently less consolidated by the Hercynian folding. The presence of basic volcanics in the Carboniferous of some zones can thus be regarded as evidence of marked regional differences in the granite-gneiss layer of the earth's crust existing still during the Hercynian folding.

4. Acid volcanics accompany the basites in a number of formations, being prevalent and characteristic in some of them. Their proportion in the geosynclinal complexes is difficult to determine in the pre-Palaeozoic complexes because of extensive anatexis and higher-grade metamorphism. Leptite gneisses in the Rhodope Mts. (I. BOJANOV—D. KOŽUCHAROV 1968), the Serbo-Macedonian massif (M. D. DIMITRIEVIĆ 1967) and in the Kohút zone of the West Carpathian Crystalline (L. KAMENICKÝ 1973) are often placed in this group. The metakeratophyre and quartz porphyrite Arada Formation of the Upper Proterozoic-Lower Cambrian age in the Apuseni Mts. deserves to be mentioned in this context. Quartz porphyries and their tuffs altered to porphyroids are more frequent in the Lower Palaeozoic epimetamorphosed series of flyschoid to flysch character, as are the Gelnica Formation in the West Carpathian Gemerides and the Tulgheș Formation in the East Carpathians. In the southern nappe units of the Apuseni Mts. the Cambrian Muncel Formation is fairly rich in intermediate and acid volcanics, quartz porphyrites and metamorphosed quartz porphyries. Interlayers of acid volcanics occurring together with basites are reported from the Precambrian complex of the South Carpathians and from the Riphean-Cambrian sequences of the Serbo-Macedonian massif. Layers of quartz porphyry and subordinate quartz keratophyre of great thickness are dated as Ordovician. The quartz porphyries in the Eastern Alps are placed at the Ordovician-Silurian boundary (H. V. FLÜGEL 1975) similarly as those in the Igal-Bükk syncline in north-eastern Hungary (K. BALOGH—L. KÖRÖSSY 1968). This implies that the frequent occurrences of quartz porphyries are often associated with the Caledonian or late Baikalian folding. However, it should be stressed that they accompany clearly

geosynclinal formations, frequently of diastrophic character with a predominance of claystones and sandstones. Numerous layers of basic rocks and both lateral and vertical alternation of acid and basic volcanics lend support to this view.

The larger amount of acid volcanics in the upper layers of the Proterozoic-Cambrian sequences or in the lower part of the Ordovician-Silurian complexes can be explained most logically as the result of the differences in the thickness of the crust, which were caused by the Baikalian, usually late Baikalian or early Caledonian folding. However, no essential change in the geosynclinal character of the complexes took place at that time.

The Devonian-Lower Carboniferous sequences often contain layers of quartz porphyry. In the upper part of the Devonian Hron Formation (West Carpathians), for example, acid and basic effusives alternate. In the East Carpathians, the low-metamorphosed Tibău Formation amidst a complex of limestones, dolomites, phyllites and conglomerates, contains not only basic rocks but also acid metatuffs, similarly as the Devonian-Lower Carboniferous Repedea Formation in the north. In the Apuseni Mts. the Devonian-Lower Carboniferous Păiușeni Formation bears basic rocks, more acid metaquartz porphyrites, keratophyres, quartz porphyries and their tuffs together with phyllites, quartzites and conglomerates. In the Danubicum of the South Carpathians, the tuffs of quartz keratophyre and quartz porphyry are more frequent in the Lower Carboniferous sandstone-shale sequences. The Upper Devonian-Lower Carboniferous sandstone-shale complex of the Balkanides also contains pyroclastics of quartz keratophyre and quartz porphyry.

The above instances show that acid volcanics accompany the basites also in the Upper Devonian and Lower Carboniferous. The differences in the thickness of the crust, which account for their presence, must have been considerable already towards the end of the pre-Hercynian sedimentary cycle and were associated with the first manifestations of the Hercynian orogeny. During this period, the differentiation of the crust in the East and South Carpathian areas into depressions and ridges was also more pronounced. The thicknesses of sequences greatly vary. The limestone and dolomite complexes are very thick as, for example, the Tibău and Repedea Formations in the East Carpathians and the Padeș Formation in the Suprageticum of the South Carpathians (up to 3000 m, H. G. KRÄUTNER 1972).

5. Both the sedimentary and volcanogenic pre-Upper Carboniferous formations are relatively little differentiated. The predominance of diabases with abundant tuffs and accompanied by grauwacken, flysch and aspidic series, as well as the presence of more acid volcanics indicate a specific type of associations, which can hardly be compared to ophiolites; a small proportion of ultrabasites and silicites is an additional contraindication. They obviously represent another crustal type, far less differentiated into structural-facies zones characteristic of the Alpine cycle in the Alps, Carpathians and Dinarides. The differentiation intensified at the end of the Proterozoic or at the beginning of the Early Palaeozoic in result of the Baikalian or

Caledonian folding, but even then the quartz porphyries were closely linked with the spilite-diabase association, which is most characteristic for the pre-Carboniferous period of the Alpine region in south-eastern Europe.

#### 4. Mesozoic ophiolites and basites

In contrast to the pre-Hercynian complexes, the Mesozoic sediments in the most Alpine segments contain a small portion of basic and ultrabasic magmatites (Fig. 2). The interior zones of the Dinarides and Hellenides, and part of the Southern Apuseni Mts. (the Metalliferous zone) regarded as a branch of the Dinaride Vardar zone, are exceptions to this rule, but even there the basites and ultrabasites are confined to a limited number of structural zones. It is usually an association of gabbro, spilite and diabase bodies amidst the abyssal sedimentary rocks, particularly cherts (radiolarites). This association has been long known under the term the 'ophiolite series' or 'ophiolite complex'. The association is often incomplete, isolate occurrences of some of its members are more frequent. In view of the plate tectonics theory, the basic and ultrabasic rocks of the Mesozoic sequences are of particular importance for the recognition of the type of the earth's crust and its changes.

The classical areas of ophiolites are the Dinarides and Hellenides with a Triassic and a Jurassic series (Fig. 9). The 'porphyrite-chert' series of Triassic, mainly Middle Triassic age, is a typical spilite-keratophyre association, which consists of the following members (J. PAMIĆ 1974): a) normal subalkalic volcanics including diabasites (palaeo-basalts), porphyrites (palaeo-andesites), porphyries (palaeo-dacites), quartz porphyries (palaeo-rhyolites) with transitional types and pyroclastics; b) the albite-rich volcanics — spilites, keratophyres (palaeo-trachytes) and quartz keratophyres with transitional types and pyroclastics; c) dykes of quartz albitite, albite granite-porphyry and albite diabase (palaeo-basalt) penetrating the volcanic bodies; d) an intrusive group emplaced amidst the earlier rocks and composed of gabbros, granodiorites, diorites, and albite granites. The volcano-sedimentary character of the spilite-keratophyre association is apparent mainly in the Ladinian. At some places, particularly in Albania, volcanism was active already in the Late Werfenian stage (T. BIÇOKU et al. 1974). The volcanic associations occur in a complex of slates, silicites, cherts, limestones and claystones. Pillow lavas, abundant amygdaloid types, volcanic breccias and pyroclastics alternating with sedimentary rocks provide evidence of submarine volcanism. Intrusive bodies emplaced in the earlier rocks are contemporaneous with the volcanics.

An especially clear picture of the changes in the character of magmatism is evident from detailed studies in the Crna gora (V. KNEŽEVIĆ 1976).

The first magmatic manifestations appeared at the end of the Werfenian in the SE part of the region. Highly hybrid acid magmatites, i.e. quartz keratophyres and keratophyres forming flows and submarine sills date from that time.

More extensive magmatic manifestations are as late as Middle Triassic, following the deposition of Lower Anisian limestones in the Sinjajevina, the Durmitor and the Ljubišnja Mts. (in Bosnia-Herzegovina even of those of Ladinian age). Quartz keratophyres, keratophyres and porphyrites are the main rock types. They are genetically linked to a system of deep-seated, more or less parallel faults of NW-SE direction.

Effusions are more frequent in the Ladinian, being accompanied by breccias. A continuous Budva-Zukali trough-depression was filled with fairly thick basaltoid, dominantly spilite lavas. The volcano-sedimentary series built up the porphyrite chert formation (B. ČIRIĆ 1954). In the marginal parts of the trough (at the margin of the carbonate platform) there are andesites and agglomerates in addition to basalts.

The volcanic manifestations persisted to the Carnian (mainly in the NW part of the Crna gora in the Zukali-Budva Zone — to the end of the Triassic); in the SE part magmatic activity was fading out already towards the end of the Ladinian.

Typical features of Triassic volcanics in the Dinarides are the hybrid character, particularly distinct in the first stages, the increase in basicity and the presence of pillow lavas and volcanic agglomerates in the Ladinian, and the concentration of volcanics in zones that correspond to developing sedimentary troughs, which are separated one from another by broad ridges bearing shallow-water carbonate sediments with a small portion of volcanics. S. Karamata (1974) presumes that this formation was a product of a thermal 'dome', which formed beneath the Dinaric plate at the beginning of the Triassic. This dome was the source of strongly contaminated magma, which extruded along its opening fissures in Middle Triassic time.

A characteristic formation of the Upper Jurassic and Lower Cretaceous is the 'diabase-chert formation' distributed in the interior zones of the Dinarides (Vardar and Ophiolite zones) and in the Mirdita zone of the Hellenides (Subpelagonicum). It includes spilites and diabases (palaeo-basalts), rare keratophyres (palaeo-trachytes), radiolarites, cherty limestones, pelagic limestones and clays as well as gabbros and rare syenites and plagiogranites. Minor or large peridotite and lherzolite bodies are numerous. Some geologists (B. ČIRIĆ 1962, 1974, J. AUBOUIN 1964, J. A. BRUNN 1960) believe it to be an ophiolite sequence with ultrabasites at the base and diabase rocks higher up; intrusive gabbro bodies intervene between them. Radiolarites and pelagic cherty limestones make up the top part.

The composition of the ophiolite association in the Dinarides and Hellenides essentially agrees with the theoretical profile through the crust of present-day oceans. It is therefore not surprising that the Dinarides and Hellenides are in the centre of interest of 'global tectonists' and not only Jugoslavian. Primary attention is paid to the large ultrabasic bodies, their composition and particularly their tectonic position (M. D. DIMITRIEVIĆ 1974, S. KARAMATA—J. PAMIĆ 1972). The ultrabasic

bodies in the Vardar zone are formed dominantly of harzburgites and in the ophiolite zone prevalently of lherzolites; peridotites and dunites form dykes and lenses (S. KARAMATA—J. PAMIĆ 1972). Large serpentinite bodies in the Mirdita zone in Albania (Tropojë, Bulquzi, Krrabë, Gramsh, Vallamarë) are composed of harzburgites and to a minor extent of earlier dunites and pyroxenites. The ultrabasite bodies are pierced by gabbro and gabbrodiorite dykes (T. BIÇOKU et al. 1974). The present position of most major ultrabasic bodies is pronouncedly tectonic, with clear-cut zones (in places up to 100 m thick) of mylonitized serpentinites and a mixture of frequently schistose serpentinites, amphibolites and sedimentary rocks. The major serpentinite bodies themselves are either anticlinal diapirs or blocks uplifted on faults and accompanied by metamorphic phenomena.

The huge bodies of ultrabasites as, for example, those of Brezovica or Zlatibor, are difficult to interpret because of thick amphibolite masses being frequently folded together with serpentinites. This suggests that the rocks of these two types originated jointly at great depths and were brought tectonically into the present position. The great difference in the metamorphic grade of these masses relative to the Jurassic 'diabase-chert formation' showing only occasional low metamorphism, and their predominantly tectonic contact indicate that the major serpentinite bodies with amphibolites, gabbros and gabbrodiorites are possibly older. Some authors (M. ILIĆ 1967) date them as Palaeozoic but with regard to the presence of Palaeozoic metamorphosed limestones in the tectonic mélangé even the pre-Palaeozoic age is admitted. Serpentinites are also regarded as shreds torn from the earth's mantle (J. PAMIĆ 1971), the same as the gabbros and amphibolites belonging to the Precambrian crust of the Jurassic Tethys ocean, which was laid bare owing to the spreading of the granitic layer before the formation of the 'diabase-chert formation' (A. L. KNIPPER 1975). The greater part of the ancient oceanic crust was subducted and partly buried by large nappes such as the Pelagonian and Serbo-Macedonian massifs.

The Ophiolite zone of the Dinarides is interpreted as a zone of ophiolite olistostromes deposited along the oceanic margin subject to subduction. In the cherty matrix are enclosed blocks of basites, ultrabasites, red siltstones, sandstones and limestones of Permian to Jurassic age. Bodies of Triassic and Jurassic limestones accumulated by gravity slumping from the immediate slopes are of a particularly large size (M. D. DIMITRIEVIĆ 1974). The presence of Senonian sediments in the mélangé of the Vardar zone is explained in terms of dextral transcurrent movements at the boundary between the Dinarides and the Serbo-Macedonian massif. The subduction itself, the closing of the 'Zvornik ocean' and the formation of the ophiolite olistostrome is thought to be an older process, initiated in the Late Jurassic (M. D. DIMITRIEVIĆ 1974).

The relationship between serpentinites and gabbros on the one hand and the effusive-sedimentary series on the other is complicated and difficult to recognize. The intricacy of the problem is enhanced by the distribution of the ophiolite series in

two (Ophiolite and Vardar) zones, which are separated by the Central Dinaric zone with the Palaeozoic Drina-Ivanjica belt (M. D. DIMITRIEVIĆ 1974) and with the Mesozoic shallow-water carbonate facies, which are closely connected with the volcano-sedimentary 'diabase-chert formation' (B. ČIRIĆ 1962, J. AUBOUIN 1970, J. RAMPNOUX 1972). This suggests a differentiation of the ocean floor into troughs and ridges and of the corresponding parts of the earth's crust into zones of small and greater thickness.

The ophiolite, effusives and intrusives of gabbro, gabbrodiorite, diorite and even ultrabasic types also occur in the Southern Alps. They are most abundant in the Ladinian, comprising diabases, melaphyres (palaeo-basalts), quartz porphyries (palaeo-rhyolites), kersantites, essexites, nepheline syenites, bostonites, camptonites, diabase-porphyrates, gabbros, pyroxenites and peridotite. In the Ladinian Wengen Formation of the Alps volcanic breccias, tuffites and porphyry, porphyrite and melaphyre lavas are frequent; quartz porphyry tuffs and tuffites (palaeo-rhyolites) occur in the Carnian Buchenstein Formation.

The proportion of Mesozoic ophiolites is smaller in the remaining segments of the Alpides, with the exception of the Bükk Mts. (the southernmost unit of the West Carpathians) and the Mureş zone in the Apuseni Mts. Many authors regard these two ranges as branches of the Dinarides.

In the Eastern Alps volcanic occurrences, rarely of one metre thickness, are known from the Upper Anisian and Ladinian of the Oberostalpin upper nappes. They occur mainly in association with the Reifling Limestone and are represented by porphyrite tuffs and tuffites (palaeo-andesites) and keratophyre tuffs (trachyte-andesites); melaphyres and spilites are sporadic, occurring mainly in the Cordevolian. A small serpentinite body is known from the Werfenian of Hohe Wand (V. J. DIETRICH 1976). Layers up to 120 m in thickness formed prevalently of pyroclastics with layered porphyrites (palaeo-andesites) and melaphyres (palaeo-basalts) are in the Ladinian of the Gailtaler Alpen and in the Lienzer Dolomites. Traces of quartz-porphyry tuffites derived from rather distant areas occur in association with the Jurassic cherts (V. J. DIETRICH 1976). Augite porphyrites, melaphyres and their tuffs are known from many places in the Northern Calcareous Alps, e.g. from the Liassic mottled marlstones and in the Neocomian marly Aptychi limestones and marlstones of the northern nappes (A. TOLLMANN 1961). In the latter sequence lamprophyre dykes and sills are also found. The Albian-Cenomanian 'Randkonglomerat' in the Frankenfels nappe contains diabase, porphyry and quartz porphyry pebbles.

The ophiolite association is characteristic of the Penninicum. The differences in the type of ophiolite associations are in the Alps, and particularly Western Alps, very prominent; they reflect the existence of two oceanic troughs separated by the Briançonnais ridge. In the northern 'Waliser Bünderschiefer Trog' zone, in addition to the schists (Bünder Schiefer) there are only small bodies of basic rocks, such as dolerite and gabbro sills and dykes, pillow lavas, hyaloclasts, serpentinite relics and

rare gabbro occurrences (V. J. DIETRICH 1976). The oceanic crustal type was not very marked in this trough. The southern Penninicum or Piemontian zone is characterized by siliceous slates, cherts, marlstones, shales and fairly rich pillow-lavas and pillow-breccias. In the Groß Glockner area in the Hohe Tauern Mts., amphibolites, eclogites, prasinites and serpentinites are from the Triassic (radiometric age 170 m.y.).

The metamorphism of the ophiolite zones of the Alps (of blue schist and eclogite types) is generally thought to be associated with subduction. From the radiometric ages of 80—60 m. y. (on micas and alkalic amphibolites) it can be inferred that crustal shortening already occurred in the Late Cretaceous. The main subduction phase is placed at the end of the Eocene and connected with the termination of flysch sedimentation.

Of particular interest is the occurrence of ophiolites with serpentinites (probably of Cretaceous age) at the eastern end of the Alps in the neighbourhood of the West Carpathians. They are part of the metamorphosed Rechnitz Formation composed predominantly of calcareous phyllites with greywacke and quartzite-phyllite interlayers (P. BECK—MANNAGETTA 1974), which is considered to be the continuation of the Penninicum (A. TOLLMANN 1967).

Fragments of serpentinite, gabbro, diabase, pillow-basalt and hyaloclasts occur with Jurassic cherts and Aptychi limestone and the Tithonian-Neocomian marlstones in the Klippen Belt near St. Veit. Ophiolite detritus is abundant in the clastic rocks of the Flysch Zone of the Eastern Alps and in the Gosau Cretaceous until the Campanian (V. J. DIETRICH 1976).

In the West Carpathians, the ophiolite rock type is known only in their southernmost units — the Bükk Mts., the Rudabánya area and the Meliata Formation in the Slovakian Karst (Fig. 10).

In the Anisian, Upper Ladinian and Lower Carnian of the Bükk Mts. in Hungary, diabases (palaeobasalts), diabase-porphyrites and gabbro-porphyrites, quartz porphyries and their tuffs occur together with cherty limestones and siliceous shales (E. SZÁDECKY-KARDOSS—G. PANTÓ—K. SZEPESHÁZY 1967). Some magmatites have a higher content of alkalis (spilites, essexites). Massive and layered flows of basic lava alternate with pillow lavas and hyaloclast interlayers. The alternating lava flows and tuffs attain a thickness of several hundred m in the Bükk Mts., in the Rudabánya area thicknesses are smaller (development closest to the Meliata Group). Diabase dykes transect the complex at a greater depth.

The gabbro, harzburgite and peridotite (serpentinite) intrusions, which complement the above association are also believed to be Late Jurassic. The ophiolite formation was metamorphosed under the pumpelyite-prehnite-quartz facies conditions (P. ARKAI 1973). The gabbro-peridotites scattered over a distance of 150 km and the accompanying diabase-spilites follow the Darnó line. They are genetically associated with an ancient fault that bounds a narrow intrageosynclinal ridge (T.



Fig. 10. Tectonic Map of the southern zones of the West Carpathians; magmatite occurrences accentuated

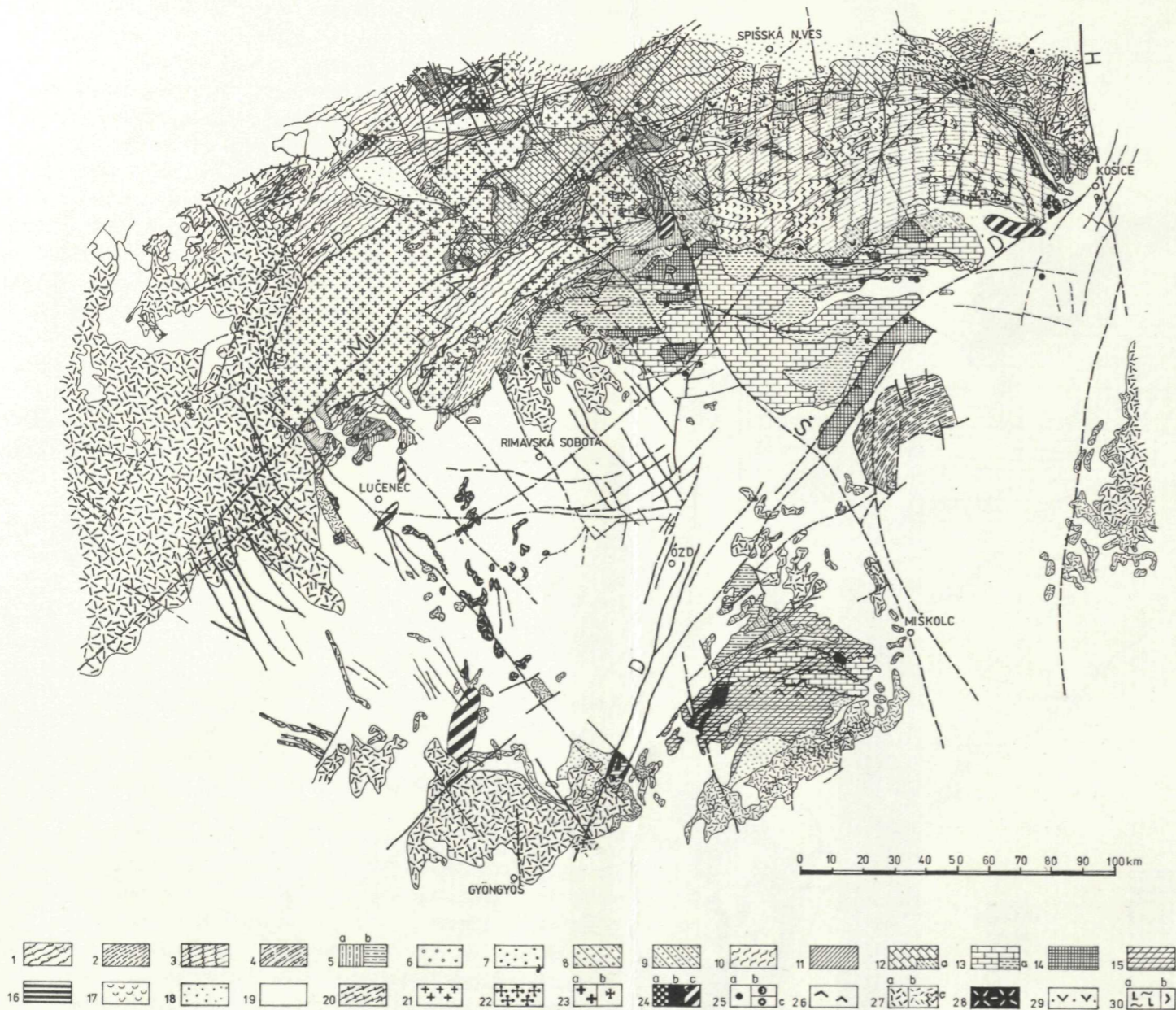
Tectonic units

1. Meso- to katametamorphites of the Veporides, predominantly Early Palaeozoic, partly Proterozoic
2. Hladomorná dolina Group- Early Palaeozoic episerics of the Veporides
3. Early Palaeozoic episerics of the Gemerides
4. Devonian of Szendrő
5. Carboniferous a) of the Bükk Mts., b) North Gemerides
6. Veporide Permian
7. North Gemeride Permian
8. Rožňava-Železník Group — Permian
9. Permian of the Bükk Mts.
10. Krížna unit
11. Stružník unit (Lower Triassic to Jurassic) — envelope of the Veporides
12. North Gemeride Mesozoic a) Lower Triassic
13. Silica nappe a) Lower Triassic
14. Meliata Group (partly ophiolitic) — Triassic
15. Bükk Group
16. Jurassic of the Gemerides
17. Upper Cretaceous
18. Palaeogene depressions
19. Neogene depressions

Magmatites

20. Granite-gneiss of Muráň type
21. Hercynian granites-trondhjemites
22. Palaeo-alpine granites
23. Granodiorites a) Eocene, b) Neo-alpine
24. Gabbros and diorites a) Mesozoic gabbros accompanied by ultrabasic and diabases
25. Minor bodies of Mesozoic ultrabasic a) of uncertain age, b) pre-Alpine, c) ultrabasic at depth
26. Diabases, porphyrites, quartz porphyries (Triassic)
27. Neovolcanics a) predominantly andesites, b) rhyolites, c) pyroclastics
28. Basalts
29. Permian quartz porphyries
30. Palaeozoic a) diabases, b) porphyroids

Principal fault lines: B — Balaton, D — Darnó, R — Rožňava, L — Lubeník, M — Margecany, Mu — Muráň, P — Pohorelá, J — Jelšava deep-seated fault, Š — Štítnik fault, H — Hornád fault



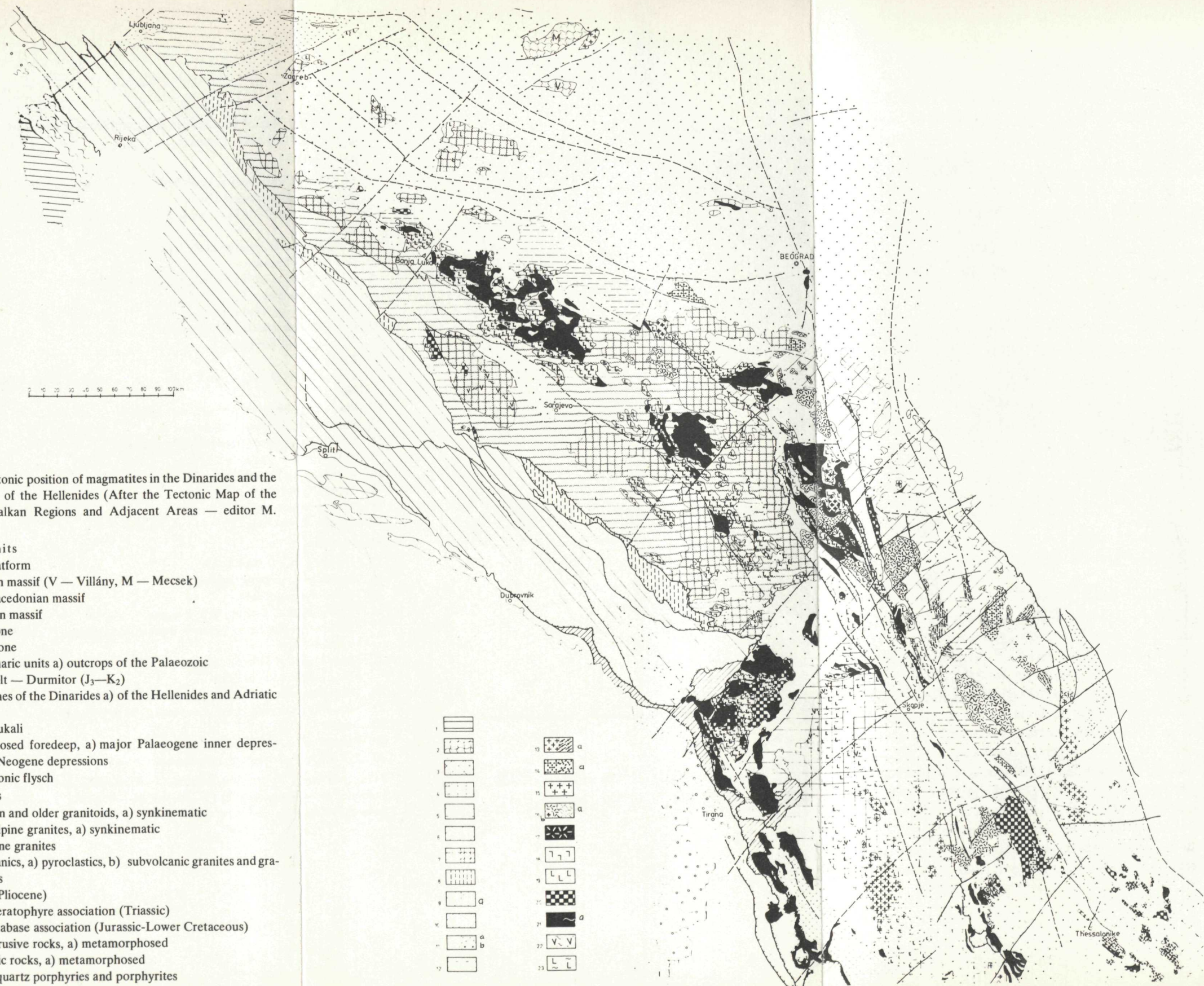


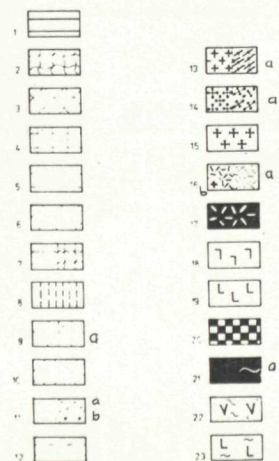
Fig. 9 Geotectonic position of magmatites in the Dinarides and the northern part of the Hellenides (After the Tectonic Map of the Carpathian-Balkan Regions and Adjacent Areas — editor M. Mahef)

**Tectonic units**

1. Istrian platform
2. Hungarian massif (V — Villány, M — Mecsek)
3. Serbo-Macedonian massif
4. Pelagonian massif
5. Korab Zone
6. Vardar Zone
7. Inner Dinaric units a) outcrops of the Palaeozoic
8. Flysch Belt — Durmitor ( $J_3-K_2$ )
9. Outer zones of the Dinarides a) of the Hellenides and Adriatic Zone
10. Budva-Zukali
11. Superimposed foredeep, a) major Palaeogene inner depressions, b) Neogene depressions
12. Late-tectonic flysch

**Magmatites**

13. Hercynian and older granitoids, a) synkinematic
14. Palaeo-alpine granites, a) synkinematic
15. Palaeogene granites
16. Neovolcanics, a) pyroclastics, b) subvolcanic granites and granodiorites
17. Basalts (Pliocene)
18. Spilite-keratophyre association (Triassic)
19. Spilite-diabase association (Jurassic-Lower Cretaceous)
20. Basic intrusive rocks, a) metamorphosed
21. Ultrabasic rocks, a) metamorphosed
22. Permian quartz porphyries and porphyrites
23. Pre-Permian spilite-diabases



ZELENKA 1974). In the south-east they join the south-eastern spur of the Bákony Mts., where the Ladinian and Lower Carnian complexes of cherty limestones and pelites enclose diabase tuffs and tuffites of fairly great thickness.

The basic rocks, diabases (palaeo-basalts) with subordinate basalt tuffs and keratophyres (palaeo-trachytes) occur at the northern margin of the Slovakian Karst in the Meliata Formation, which is composed of dark shales, radiolaritic shales and radiolarites (J. KANTOR 1955). Hematite layers, in places with traces of sulphidic Cu ores are genetically associated with them. The presence of tuffs, spilitic rocks and pillow-lavas indicates a rapid congealing of lavas. The plastically deformed xenoliths of cherty rocks, the volcanogenic material in the Triassic crystalline limestones and layers of glaucophane to sericite quartzites in glaucophanites are evidence that submarine effusions were synchronous with sedimentation (P. REICHWALDER 1970). The diabases are linked up with glaucophanites through transitional types and are related to them in chemistry. J. KAMENICKÝ (1957) explains the formation of glaucophanites by alkaline metasomatism of diabases and P. REICHWALDER (1970) thinks it to be the result of Alpine dynamometamorphism; the radiometric age values about 80 m. y. support the latter opinion (Upper Cretaceous — personal communication of J. Kantor). In the Triassic complexes of the Meliata Formation there are also minor serpentinite bodies (e.g. near Borka, Jelšava, Držkovce), which obviously represent the deeper members of the ophiolite formation (M. MAHEL 1975). The association with the basites of the Meliata Formation should be regarded as being analogous to the Middle Triassic-Carnian volcanics of the Bükk Mts. (H. KOZÚR—R. MOCK 1974).

Several minor serpentinite bodies are located in the proximity of the Rožňava fault-line so that their ascents are placed in this deep fault. Some geologists think the structurally anomalous Rožňava line to be the root zone of the Gemeride nappe (P. GRECULA 1973), and the ultrabasites to be fragments of the simatic basement squeezed into the present position by intense tectonic processes (D. HOVORKA 1965). A large body of serpentinitized peridotites to serpentinites with accompaniment of dunitic to pyroxenitic rocks near Hodkovce is situated at the crossing of the Rožňava and Darnó fault lines.

Minor serpentinite bodies are also found in the Triassic complexes of the North Gemeride synclinal zone, chiefly in the Dobšiná and Jaklovce (J. KAMENICKÝ 1957) areas. Similarly as the South Gemeride serpentinites, they form subhorizontal lenticles and steep tabular tectonic slices or dyke swarms; they belong to the lherzolite-harzburgite association (J. KAMENICKÝ 1956, D. HOVORKA—J. ZLOCHA 1974). In places they are accompanied by diabasic rocks and keratophyres situated amidst the Middle-Upper Triassic sequences. The presence of serpentinites in pebbles of the Upper Cretaceous conglomerates near the Dobšiná ice cave, and contact metamorphism they produced in the Middle to Upper Triassic sediments of this area (M. MAHEL 1957) suggest that at least part of them are of younger, probably

Late Jurassic age. The considerable amount of radiolarite pebbles is puzzling as radiolarites are unknown from the North Gemeric Mesozoic. They occur in the Triassic of the Meliata Formation and in the Malm of the Slovakian Karst. The fragmentary data that are available on the ultrabasites of the West Carpathian southern zones indicate that some bodies (mainly those in the Meliata Formation) might be related genetically to the thin crust of the Meliata trough and its deep-sea sediments (black shales and radiolarites). Most of the ultrabasites of the Slovakian Karst and the North Gemeric synclinal zone seem to be Upper Jurassic. The former are connected with the Rožňava deep fault and the latter with the deep fault bordering the North Gemeric synclinal zone (M. MAHEL 1957). Thin layers of tuffs, tuffites, porphyrite and quartz porphyry are known from the limestone-dolomite complexes of the ranges Stratsenská hornatina, Galmuské pohorie (M. MAHEL 1957, 1967) and of the Slovakian Karst (Silica nappe). They most frequently occur together with the Upper Anisian chert limestones of Steinalm type and Lower Ladinian dark cherty limestones of Reifling type, i.e. the deep-sea facies, which indicate a diversified relief of the Triassic sedimentary area.

The Drienok Formation, which forms the southernmost part of the Strážov nappe, contains (a) small bodies of porphyrites (palaeoandesites), orthofelsites (palaeotrachytes), quartz porphyry (palaeorhyolite) and their tuffs amidst the Campilian marlstones (M. SLAVKAY 1973) and (b) biotite picrites, diabases (palaeobasalts) and keratophyres (alkalic palaeotrachytes — personal communication by D. Hovorka) amidst the dolomites and Reifling Limestone (D. HOVORKA—M. SLAVKAY 1966). These rocks are not 'ophiolites' but volcanics of a hybrid type, which indicate the beginning of fracturing and differentiation of the crust after its Hercynian consolidation.

Small basic bodies, chiefly of augites and their tuffs and tuffites — emplaced in the Neocomian and Lower Albian marlstone complex are particularly frequent in the mountain ranges Strážovská hornatina, Veľká Fatra, Malá Fatra, and in the Low Tatra and Little Carpathians (M. MAHEL 1967) (Fig. 3). With regard to that they follow after a strong deepening of the sedimentary area evidenced by the Dogger-Malm radiolarites, and that they occur together with the pelagic marls, the genesis of the volcanics can be justifiably related to the thinness of the crust in the Krížna—Zliechov sedimentary trough. The basites in the lower part of the pre-flysch Albian in some envelope units (e.g. the Malé Karpaty Formation — M. MAHEL 1967) are of the same type. The frequent presence of basite 'exotic' pebbles of intraformational conglomerates, which accompany pre-flysch complexes suggests that they originated in the period of a more intense differentiation of the earth's crust. In the shallow-water (cordillera) sequences the basic rocks are scarcer. The most important are large bodies of limburgite and limburgite tuffs and tuffites occurring together with neritic Tithon-Neocomian limestones at Osobitá in the High Tatra Mts. In the Trangoška syncline in the Low Tatra, amygdaloidal augites are located as

hypabyssal bodies amidst the Werfenian sedimentary complex and the Triassic limestones and dolomites. They are of Early Cretaceous age and the whole association may be designated (after D. HOVORKA) as "alkalic olivine basalts to basanites". The occurrence of basites in the Mesozoic of cordillera complexes indicates that the earth's crust was disrupted and of diverse thickness during the Late Jurassic and Early Cretaceous.

Palaeo-alpine basic rocks often form minor necks in the Klippen Belt and abundant pebbles of the Upper Cretaceous and Eocene exotic conglomerates. In the Peri-klippen area and in the Klippen Belt there is a wide stratigraphic range of magmatites. In the Peri-klippen area, the Valanginian, Aptian and Albian contain augite tuffs and lavas, and from the Klippen Belt tuffs and tuffites, basites and ultrabasites (picrites), picrite interlayers in the Albian and picrite interlayers in the Cenomanian-Turonian (in Vienna) are known. The magmatites make up minor bodies, necks and dykes. The pebble material of conglomerates, which are mainly Upper Cretaceous (beginning with the Albian) in the west and prevalently Palaeogene in the east, contains magmatites of unusually great variety: limburgites, augite porphyries, spilites, melaphyres, porphyrites (palaeo-andesites), quartz porphyrites (palaeo-dacites) and quartz porphyries (palaeo-rhyolites) (V. ZOUBEK 1931, T. WIESER 1958). Of particular interest are the occurrences of glaucophanites, serpentine fragments and chromite in the Aptian limestones (M. MIŠÍK 1976). The Klippen Belt, including the narrow adjacent Peri-klippen zone, was evidently a specially favourable area for volcanic protrusions during the Early and Late Cretaceous. The variety and contrasts in the material composition — from serpentinites to plagiogranites and from limburgites to quartz porphyries (palaeo-rhyolites)—seem to be allied with the contrasting palaeogeography of the Klippen Belt embracing deep troughs (e.g. the Pieniny-Kysúca trough) and the Czorsztyn ridge and Klape cordillera. The changes in the type of volcanic activity with time should also be considered. The serpentinite protrusions date probably from the Early Cretaceous, but the quartz porphyry and quartz porphyrite tuffs in the Branisko Formation of the Pienides indicate that intermediate volcanism was later, possibly Late Cretaceous, Turonian and even younger. The quartz porphyry (palaeo-rhyolite) and albite porphyry pebbles in the Albian conglomerates are of uncertain age (Permian? — Lower Cretaceous?). The occurrences of glaucophanites are the first signs of metamorphic processes associated with the construction of the basement of the Klippen Belt and subduction. Are not we concerned with oceanic crust subduction of a broad trough in extension of the Penninicum to the east? An especially varied spectrum of magmatites is found in Paleogene conglomerates of the Periklippen area and Klippen Belt in eastern Slovakia: augite porphyries, augitites, limburgites, dacites, kersantites, quartz porphyries, keratophyres, granite porphyries and granites (R. MARSCHALCO—M. MIŠÍK—L. KAMENICKÝ 1976).

From the initial, predominantly pre-flysch geosynclinal stage of the Silesian unit

(in the Flysch Belt of the West Carpathians) date the Lower Cretaceous lava flows, domes, dykes and necks of picrite and teschenite and their tuffs and tuffites (Z. Roth in T. BUDAY et al. 1967), and effusions and hypoabyssal bodies of pronouncedly alkalic rocks (A. MAHMOOD 1973). Picrite lava flows and dykes are also known from the Púchov Marls of the Cenomanian-Turonian of the Biele Karpaty unit in the basement of the Vienna Basin, where they are located along faults that separate the Magura flysch trough from the Klippen Belt cordillera (M. ELIÁŠ 1976). No surface occurrences of ophiolites analogous to those of the Sinaia Beds of the East Carpathians are known in the West Carpathian Flysch, but their existence at depth is probable to the east of the Lednice line. The Zlín, Istebná and Godula Formations (Z. ROTH in T. BUDAY et al. 1967) bear interlayers of Upper Cretaceous to Palaeogene andesite tuff. Similarly as in the East Carpathians, the basic and ultrabasic magmatites are gradually replaced by intermediate volcanics in the Upper Cretaceous of the Flysch Belt.

In the Mecsek Mts. in Hungary, there are Lower Cretaceous submarine effusions of alkalic basaltic diabase ('basalt-trachydolerite') and alkalic diabase ('trachydolerite') and the occurrences of their pyroclastics, hyaloclastites and diabase porphyrite (E. SZÁDECKY—KARDOSS et al. 1967). Volcanic activity began in the pelagic to neritic environment at the beginning of the Berriasian, synchronously with the deposition of marly limestones. Pillow lavas and pillow breccias are frequent. In the course of volcanic activity, the shallowing of the basin, which began along its margin in the Hauterivian, extended to its central parts in the Barremian. Basic volcanism completed the sedimentary cycle in the Mecsek Mts., which was characterized by a prominent subsidence in the Dogger and Malm (G. WEIN 1969). Dykes and sills cut through the Mesozoic and the underlying Permian complex, several thousand metres thick. The ascent of volcanics is placed in the NE-SW fault which was revived during late Cimmerian orogeny (J. BILÍK 1974).

Spilite diabases (palaeo-basalts) with radiolarites were struck by boring also south of the Mecsek Mts., in the basement of the Neogene filling of the Drava valley. They are less frequent in the Villány Mts. On the basis of the boreholes sunk into the basement of the Neogene filling of the Great Hungarian Basin and of geophysical studies, a NE-SW ophiolite zone with trachydolerites and basic diabases (palaeo-basalts) is presumed to run in the area between the Danube and the Tisza, particularly between Nagykörös and Szandaszolos (K. SZEPESHÁZY 1973) (Fig. 13). Jurassic radiolarites and Lower Cretaceous turbidites are its sedimentary components. The zone extends along the western limit of the Upper Cretaceous-Eocene Szolnok flysch zone; in the SW it is linked with Lower Cretaceous volcanics of the Mecsek Mts. and stretches as far as the Transcarpathian region in the NE, where spilites and diabases (palaeo-basalts) were encountered by boring amongst the Jurassic sediments south of the Klippen Belt (C. N. DOLENKO—L. G. DANILOVIČ 1976). In the opinion of some authors the present distribution of basic magmatites in this belt

follows the Cretaceous zone of subduction (A. JUHASZ—G. VASS 1974). Jurassic and Lower Cretaceous dykes and sills, as well as lava flows of basic rocks and their agglomerates and tuffs are also found beneath the flysch of the Szolnok trough, accompanied by claystones, marlstones and radiolarian and pelagic limestones. They mainly consist of diabases (palaeo-basalts, partly alkalic), basaltic porphyrites and porphyrites (palaeo-andesites; K. SZÉPESZHÁZY 1973).

In the Southern Apuseni Mts. (Fig. 4) the ophiolite association is characteristic of the Metalliferous zone (200 km long and 40 km broad). The pre-Oxfordian complex contains basalt pillow lavas and bodies of gabbro and peridotite with radiometric age of max. 180 m. y. (H. SAVU—L. NICOLAE 1974). The Oxfordian-Neocomian ophiolite complexes are of submarine-effusive nature with lava flows and stratovolcanoes. The magmatites are of extreme variety: limburgites, diabases (palaeo-basalts), oligophyres, orthophyres (palaeo-trachytes), porphyrites (palaeo-andesites) and their pyroclastics, quartz porphyrites (palaeo-dacites), dykes and sills of quartz porphyry (palaeo-rhyolite), albite porphyry and oligophyre. The Barremian-Aptian spilitized diabases (palaeo-basalts) are accompanied by gabbro, gabbrodiorite and diorite intrusions (H. SAVU 1967, H. SAVU—G. UDRESCU 1973). Ophiolite volcanics are found together with pelagic facies of jasper limestone, Aptychi limestone (Tithonian-Lower Albian in the Drocea-Criş unit), the Tithonian siliceous and Aptychi limestones in the Mureş unit and its Albian wildflysch, and the Lower and Middle Albian flysch complexes of the Bucium unit. Part of the ophiolites, however, is accompanied by shallow-water limestones in both the Jurassic and Neocomian as, for example, in the south-west of the Metalliferous zone (H. SAVU—I. NICOLAE 1974). The ultrabasic differentiates of the above association represent a tholeiitic magma type. Ophiolite magmatic activity occurred in three stages: Pre-Oxfordian with basic rocks (basalts, gabbros), Oxfordian-Neocomian with neutral and acid rocks (andesites, dacites and rhyolites) and Barremian-Aptian with basic rocks. The volcanics, mainly of the second stage, represent two differentiation lines of initial magma: calc-alkalic (andesite-dacite-rhyolite) and alkalic (oligophyre-albitophyre-trachyandesite-orthophyre). The products are considered to be differentiates of magmas originated in the subduction zone (I. BERBELAC 1975).

In the East Carpathians the basic and ultrabasic rocks are present in several zones, the oldest being in the southernmost zones. The Ladinian basites and ultrabasites in the Transylvanian nappe occur together with the pelagic facies of nodular Hallstatt limestones and cherty and radiolarite limestones. In the Perşani Mts. in Rumania the basic and ultrabasic rocks occur in an allochthonous position in the Barremian-Bedoulian wildflysch zone above the Triassic limestones. According to some authors (G. CIOFLICA et al. 1966), diabase pillow lavas and bodies of gabbrodolerite, gabbro and serpentinite are Middle Triassic in age and porphyrites (palaeo-andesites), oligophyres, bostonitic porphyries and orthophyres (palaeo-trachytes) are somewhat younger (Late Triassic). In the palaeogeographical schemes the ophiolites are

placed in an at least 300 km long zone, extending in the northern part of the Transylvanian basin, which is regarded as the root zone of the Transylvanian nappe (M. SĂNDULESCU 1973).

The Middle Liassic to Dogger volcanics that are known from the Coldea depression, the Braşov area and the southern parts of the Peşani Mts. were derived from the more northerly Bukovina nappe. They comprise granite-porphyrines, orthophyres (palaeo-trachytes), keratophyres (alkalic palaeo-trachytes), porphyrites (palaeo-andesites) and diabases (palaeo-basalts). They were synchronous with the subsidence of the sedimentary area and the deposition of cherty limestones. Radiolarites and radiolarian limestones correspond to the maximum subsidence in the Oxfordian.

The eugeosynclinal zone of the East Carpathians migrated gradually northwards: it occurred in the root area of the Transylvanian nappe group in the Triassic, in the Bukovina area in the Jurassic (M. SĂNDULESCU 1973), and towards the end of the Jurassic and in the Early Cretaceous the northerly flysch eugeosyncline was incorporated in it.

In the Eastern Carpathians the ophiolite formation is characteristic of the Malm and Lower Cretaceous of the boundary zone between the Inner and Outer Carpathians, represented by the 'Black flysch' (Kamenný potok), and the Ceahlău-Rachovo unit (D. RĂDULESCU—M. SĂNDULESCU 1973, M. G. LOMIDZE 1968). In the Black flysch unit, diabases, spilites and their tuffs occur together with silicites, argillites and siltstones and, in its upper members, also with limestones. In the Ceahlău — Rachovo unit the magmatites are found in the carbonate flysch of the Sinaia Formation (Fig. 11a).

The volcanogenic complexes of the boundary zones were studied in greater detail on the Soviet territory. They were divided into the spilite-diabase and the spilite-keratophyre associations. In the former, spilites and porphyritic diabases (palaeobasalts) are accompanied by gabbro-porphyrines, gabbros and serpentinites (harzburgites and lherzolites), and in the latter, diabases (palaeo-basalts), keratophyres (palaeo-trachytes), volcanic breccias and tuffs are found. Major bodies of gabbro and peridotite are thought to occur at depth (M. G. LOMIDZE 1975, E. K. LAZARENKO et al. 1973). The volcanic rocks are accumulated at the northern margin of the Marmaroş massif (M. G. LOMIDZE 1968) in tectonic slices, which consist of fragments of the Oxfordian-Lower Cretaceous, up to 750 m thick volcanogenic-sedimentary sequence. A strong tectonic reworking in the frontal part of the Marmaroş nappe gave rise to a belt of mélangé (M. C. LOMIDZE 1968). The Jurassic to Lower Cretaceous ophiolitic volcanism was related to the disruption of the crust along the Marmaroş suture towards the end of the Jurassic and to the formation of a trough with the oceanic crust. The trough was filled with thick carbonate flysch complexes of the Ceahlău-Rachovo unit (M. G. LOMIDZE 1968, M. SĂNDULESCU 1973). On the Rumanian side, diabases, melaphyres and their agglomerates and tuffs are also



abundant in the Barremian-Aptian wildflysch at the southern margin of the ophiolite zone (M. SÂNDULESCU 1973). Minor basalt effusions are known from the Tithonian-Neocomian of the more southerly Klippen Belt (M. G. LOMIDZE 1975).

Weak volcanic activity took place at the end of the Early Cretaceous and in the Late Cretaceous and Palaeogene in the Flysch Belt of the East Carpathians, even in their exterior units. This volcanism, however, was not of basic type; it was represented by polyphasal andesite-dacite effusions located in several zones amidst the flysch sequences (for details see p. 72—73).

In the South Carpathians the carbonate flysch of the Severin zone is accompanied by diabases, gabbros and serpentinites, particularly in the Tithonian-Lower Neocomian Sinaia Formation (Fig. 5). Minor volcanic occurrences extend to the Danubicum. Volcano-sedimentary complexes of small thickness are known from the Jurassic of the Presacina zone and from the Cenomanian-Turonian wildflysch: diabases, partly spilitized keratophyres, alkalic palaeo-trachytes (H. SAVU 1967). In the Yugoslavian part of the South Carpathians, basic rocks occur in the deep-sea sequences of the Greben zone; the gabbro bodies (near Deli Jovan) bear layers of serpentinitized ultrabasites in the lower part and diabase-porphyrite and diabase dykes higher up. Superjacent are diabases alternating with sedimentary rocks. Some geologists think these magmatites to be relics of the oceanic crust (A. GRUBIĆ 1974) of the 'Danubian ocean', which extended between the Moesian platform (including the greater part of the Danubicum) and the island zone, which is represented by the Geticum and the Serbo-Macedonian massif. The relics of this Tithonian-Lower Cretaceous ocean, closed in the Aptian to Albian, are the dominantly flysch Severin-Kraina unit and the adjacent parts of the Danubicum and Geticum with the deep-sea facies (A. GRUBIĆ 1974).

In the Balkan Mts., the Tithonian-Lower Cretaceous flyschoid complex at Strandža is accompanied by submarine volcanics such as spilites and diabases (palaeo-basalts), keratophyres (ultrabasic palaeotrachytes) and hypabyssal picrite-porphyrites and diabase-porphyrites (S. BOYADJIEV 1967, E. BONČEV 1955). The whole sequence is affected by slight regional metamorphism. The serpentinite bodies along the Marica fault at the northern boundary of the Rhodope massif and the bodies on the transverse faults along the eastern margin of the West Rhodopean crystalline blocks are very likely Alpine in age (E. DIMITROVA et al. 1975) (Fig. 6).

## Discussion

1. Basic volcanics accompanied by gabbro, gabbrodiorite and serpentinite occurrences and termed ophiolites are in the Alpides associated genetically with the following facies:

— the relatively deep-water facies of Triassic pelagic limestones and shales with radiolarite interbeds, e.g. the Budva and Zukali zones, the Bükk Mts., the Meliata

Formation in the Slovakian Karst, the Transylvanian nappe group in the Eastern Carpathians;

— the deep-sea facies of radiolarites, siltstones, Calpionella and Nannocon limestones of Upper Jurassic and Lower Cretaceous age, e.g. the Ophiolite and the Vardar zones in the Dinarides, the Krížna unit in the Central Carpathians, the Bukovina unit in the East Carpathians, lower members in the Mureş zone of the Apuseni Mts., and the ophiolite formation beneath the Neogene of the Great Hungarian Plain west of the Szolnok flysch trough;

— the pre-flysch marlstone-limestone and pelitic complexes, e.g. the 'Black flysch' unit and the Sinaia Formation in the Ceahlău-Rachovo unit;

— in the Mureş zone and partly in the Vardar zone the ophiolite associations characterize both the Jurassic carbonate-radiolarite facies and the Lower Cretaceous flysch complexes, the same as in the Penninicum of the Alps.

In all instances mentioned above, the association of the ophiolites with the stages of maximum subsidence of the geosynclinal zones is clear. It is not only a palaeogeographical relationship but primarily palaeotectonic, which can be reasonably interpreted as due to the relative thinness of the crust in the ophiolite-bearing zones.

The basic rocks, dominantly of alkalic character (augitites, limburgites) often occur together with shallow-water facies of submarine ridges, especially in the marginal faulted sectors as are, for example, Tatrídes, the Peri-klippen area in the West Carpathians and partly the Mecsek Mts. They appeared in the Upper Jurassic and Lower Cretaceous during the most intense basic magmatism in the Alpides.

The Upper Jurassic and Lower Cretaceous basic, frequently alkalic magmatites often accompany the faults bounding cordilleras, ridges or depressions in the flysch geosyncline. Such are, for example, the teschenites in the Silesian unit, basalts and augitites in the Barremian-Albian flysch sequences of the Bukovina and Sub-Bukovina nappes, or diabases (palaeobasalts) in the Jurassic limestones of the Presacina zone (South Carpathians) at the margin of the flysch trough, and augitites in the Carpathian Klippen Belt. The presence of basite pebbles in the wildflysch conglomerates, e.g. in the Albian of the Metalliferous zone, in the Barremian-Aptian at the northern margin of the Marmaros massif, in the Cretaceous conglomerates of the Klippen Belt in the West Carpathians and in the intraformational conglomerates in the Albian pre-flysch envelope units of the West Carpathians, point to an intimate relationship between the genesis of basic rocks and the differentiation of the crust.

As is seen, the distribution of basic magmatites in the Mesozoic sequences clearly depends on the local thickness of the crust, the differentiation of the Alpine geosynclinal belt and on the increased dynamics of the crust during magmatic processes. The basites are predominantly associated with the troughs having thin crust and deep faults that separate the troughs from the ridges of the sea floor or cordilleras.

The basic magmatites reached the maximum extent in the Late Jurassic and Early



Fig. 11a Geotectonic position of magmatites in the East Carpathians (Compiled on the basis of the CBGA Tectonic Map—edit. M. Mahel 1974, and the Geological Map of Rumania)

**Tectonic units**

- 1. Bretila unit — Biely potok
- 2. Sub-Bucovina nappe
- 3. Bucovina nappe
- 4. Group of Transylvanian nappes
- 5. Getic nappe
- 6. Zemplín „island“
- 7. Křížna nappe
- 8. a) Gemerides b) Veporides
- 9. Klippen Belt
- 10. „Black Flysch“ (Kamenný potok) nappe
- 11. Ceahlău (Rachovo) nappe and Baraolt nappe
- 12. Flysch Belt (nappes of middle and external groups)

- 13. Foredeep a) inner, b) outer
- 14. a) Late-tectonic Bucceci paramolasse, b) Late-tectonic Transcarpathian flysch
- 15. Neogene depressions
- 16. Overthrust lines a) Strike-slip faults
- 17. Faults a) inferred (determined by geophysical methods)

**Magmatites**

- 18. Pre-Hercynian and Hercynian granitoids a) synkinematic
- 19. Alkalic plutons (detroits)
- 20. Neovolcanics a) predominance of pyroclastics, b) granites, granodiorites — small intrusions
- 21. Pliocene basalts
- 22. Ophiolites a) basalts, b) spilite-keratophyre association
- 23. Ultrabasites a) small bodies
- 24. Permian quartz porphyries
- 25. Pre-Permian a) metabasalts, amphibolites, b) porphyroids

Cretaceous, in the period of maximum differentiation and tectonic accentuation of the sea-floor relief of the Mesozoic geosynclines and of the most pronounced oceanization in the Alpides.

2. The distribution of ophiolites and their types are related to the differentiation of the Alpine sedimentary area into troughs (trenches) and ridges (cordilleras). The ophiolite zones show particular features in each segment, only some of them pass from one segment into another but never in quite the same character.

In the Eastern Alps, the Triassic facies show a markedly deeper-water character only in the southernmost zones which are connected with the Southern Alps; in the Jurassic and Cretaceous, the Penninic trough and the basic and ophiolite zones are distinctive.

In the Triassic of the West Carpathians, the zones of the Bükk Mts. and the Meliata Formation in the Slovakian Karst are of pelagic character and contain ophiolites. In the Jurassic, the Zliechov trough is more pronounced and specific for the West Carpathians together with the Kysuca trough in the Klippen Belt. In both of them and in the marginal parts of the neighbouring ridges are basic magmatites. The differentiation of the Tatríde ridge is reflected in the frequent Lower Cretaceous necks of alkalic limburgites and augites. The teschenites in the Flysch Belt are also confined to the boundaries between the ridges (or cordilleras) with a thick crust and troughs (or trenches) with a thinner crust.

The distribution of the ophiolite association in the East Carpathians shows an analogous trend. The deeper-water Triassic facies of the southernmost Transylvanian unit are those of the ophiolite zone. In the Jurassic and Early Cretaceous, the most marked subsidence and thinning of the crust was in the northern part of the Central zone and somewhat later, in the Tithonian — Early Cretaceous in the Ceahlău-Rachovo zone. The troughs (trenches) are also there the sites of ophiolites (M. SÂNDULESCU 1973, D. RÂNDULESCU—M. SÂNDULESCU 1973).

In the Apuseni Mts. the ophiolite association occurs in the Metalliferous trough, whose crust was partly thin and of oceanic type during the Jurassic and the Early Cretaceous flysch interval. In the marginal parts of this trough the earth's crust is of continental character.

The Szolnok flysch trough and the neighbouring ophiolite trough, which in the south-west joins the Mecsek zone (in the Jurassic of trough type — E. SZÁDECKY-KARDOSS K. SZEPESHÁZY 1973), are outstanding features of the Hungarian Massif.

In the Mesozoic of the South Carpathians, the Severin and Svinjica Greben zones became the centres of ophiolite or basite occurrences (A. GRUBIĆ 1974, D. RÂNDULESCU—M. SÂNDULESCU 1973).

The Strandža unit with preflysch complexes and ophiolites is one of the peculiar features of the pre-flysch Mesozoic of the Balkan (E. BONČEV 1976).

The Triassic of the Dinarides is of varied composition; it contains deep-water facies, especially in the Budva and Zukali zones, and volcanics that are already in

part basic. The differentiation is stronger in the Jurassic; the troughs originated mainly in the Ophiolite zone, in the Subpelagonicum and the Vardar zone and were filled with the rocks of the diabase-chert ophiolite association. The deep-water sedimentation persisted from the Triassic to the Cretaceous in some areas, just the same as the distribution of ophiolites (M. D. DIMITRIEVIĆ 1974, S. KARAMATA 1974).

From the above it follows that in the Alpides of south-eastern Europe the extent and distribution of ophiolites are related directly to the differentiation of the Mesozoic geosynclinal belt, particularly in the interval of the Middle Triassic-Early Cretaceous.

3. Although the majority of European Alpidic segments is poorer in ophiolites, they are instructive from the point of view of new global-tectonic principles, in several respects: a great palaeotectonic diversity, especially in Late Jurassic and Early Cretaceous times, the demonstrable transitions between shallow- and deep-water sequences, and the existence of several zones with a contrasting crustal thickness. These features reveal that the Jurassic-Lower Cretaceous ocean differed from the type of present-day oceans. In the West Carpathians, for example, at least four zones with a crust of a smaller thickness must be differentiated in the Jurassic sea, which were separated by ridges or cordilleras where the crust was of a greater thickness (M. MAHEL 1975). The sequences consisting of mixed shallow- and deep-water facies are particularly prominent in the Klippen Belt, between the Czorsztyn and Kysuca facies types.

All this indicates that the model of the present Atlantic Ocean cannot be applied to the former deep-sea Alpidic regions. The basins in the Alpine geosynclinal belt did not represent wide, little differentiated oceanic depressions but narrow sea troughs or trenches. The abrupt transition of deep-sea facies into those of shallow-water type is not in keeping with the opinion that such a contact should result primarily from an exceptionally great reduction of space or a tectonic overlapping.

The presence of basic and ultrabasic rocks as early as the Triassic indicates the onset of tensional faulting, i.e. the formation of grabens and troughs, thinning of the crust and, possibly (N. HERZ—H. SAVU 1974), the renting of the granite-gneiss layer and drifting apart of the lithospheric plates.

4. The ophiolite series are believed to represent areas with the oceanic crustal type. The Tethys itself is identified mainly with the ophiolite series in the Dinarides and Hellenides, but even there the microplate of the Drina — Ivanjica — Pelagonian massif divided the ocean into two parts: the Zvornik ocean (Vardar zone) in the east and a western part represented by the Ophiolite zone and the Subpelagonicum (M. DIMITRIEVIĆ 1974). The Triassic ophiolite series in the Bükk Mts. are regarded as a branch of the Tethys joining the Dinarides (E. SZÁDECKY—KARDOSS 1973). The Meliata Formation in the West Carpathians is only the northern rim of this ophiolite series.

The basic rocks struck by boring in the basement of the Alföld Neogene, in basal

sequences of the Szolnok flysch trough (and west of it), which are thought to belong to the ophiolite series, represent another branch of the Tethys extending from the Vardar zone.

In the Alps the ophiolite series of the Penninicum represent the deposits of a more northerly ocean, with a tentative oceanic ridge in their piedmont branch (V. DIETRICH 1976); it was presumably divided by the Briançonnais plate in the Western Alps. The crust in the northern Valais branch is supposed to have been of nearly oceanic (suboceanic) type.

The ophiolite associations in the East Carpathians and the Apuseni Mts. are interpreted as two intrasialic basins having oceanic crust (D. RĂDULESCU—M. SĂNDULESCU 1973). The northern basin is today represented by the ophiolite series in the Ceahlău and 'Black flysch' units and by the basites of the Bukovina nappe. The ophiolite series of the Transylvanian nappe and particularly that of the Metalliferous zone were derived from the southern basin. According to some authors, this ocean situated between the Northern Apuseni and South Carpathians and called the Siret or Mureş Sea, connected the Tethys with the Dobrogea — Crimea — Caucasus ocean (N. HERZ—H. SAVU 1974). To the south-west it joined the Tethys, which is represented by the Vardar zone.

The ophiolites in the Severin zone and basic complexes extending to the Greben zone in the South Carpathians are considered to be representatives of the former 'Danubian' ocean, which spread between the Moesian plate (including the western part of the Danubicum) and the Geticum with the Serbo-Macedonian massif; the latter is thought to have been an island zone between the 'Danubian' ocean and the Tethys (A. GRUBIĆ 1974). The 'Danubian' ocean was part of the Meso-Paratethys, i.e. the Upper Jurassic-Lower Cretaceous ocean, which separated the Eurasian platform from the island arc of the Central Alps, Central Carpathians, Pannonides, Getides, Rhodopides and Anatolides (A. GRUBIĆ 1974).

There is a gap in our knowledge relating to the position of the 'Northern' ocean in the West Carpathian region, where the true ophiolite association does not crop out at the surface. The anomalous structural character of the Klippen Belt, particularly a belt of tectonic breccia along the internal margin of the Flysch Belt closely tied with it, tempts the geologists to regard it as a communication between the Penninic ocean and the northern oceanic basin of the East Carpathians (J. DEWEY et al. 1974, E. SZÁDECKY-KARDOSS 1973, A. GRUBIĆ 1974). However, the Klippen Belt was rather an internal marginal island zone with the Czorsztyn and Klappe partial ridges, separated by a narrow Kysuca-Pieniny trench. The broad oceanic trough itself was obviously the internal zone of the Flysch Belt bounded by two deep, geophysically evidenced fault lines: the Lednice or Pericarpathian, (zone of gravimetric minimum along the tie-line of Hodonín — Valašské Klobouky — Kysucké Nové Mesto — Nový Targ — Domoradz) and the Peripieninian lineament (the zone of the Carpathian block, after W. ŠIKORA 1973); metamorphosed flysch is presumed to

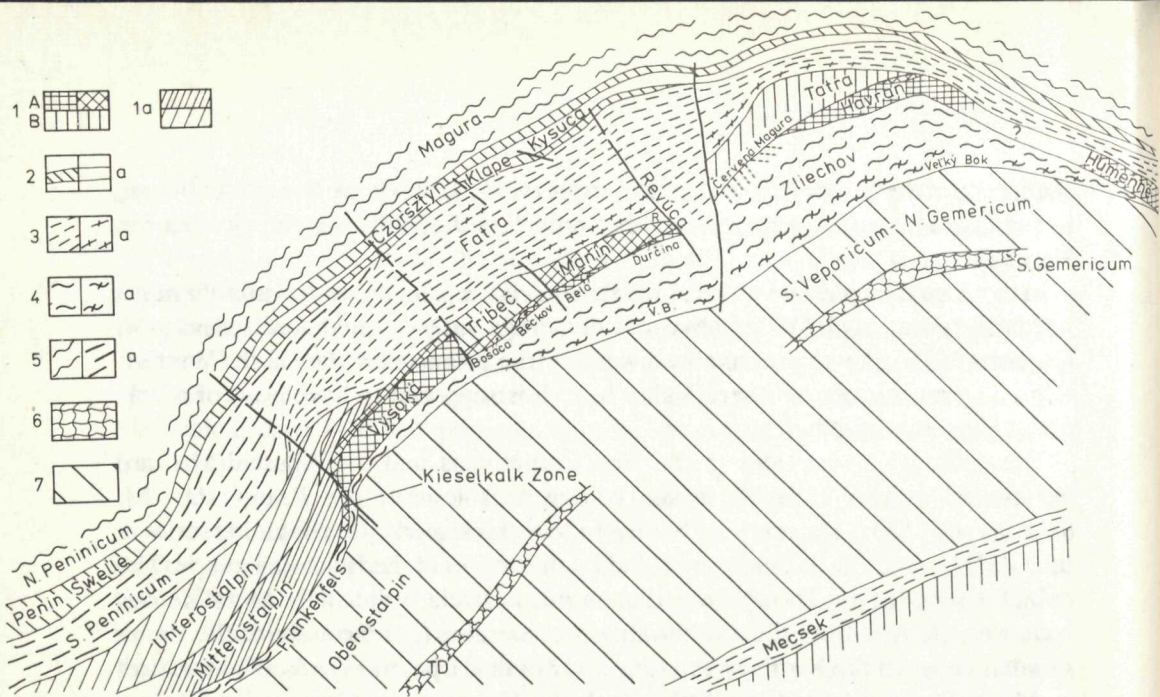


Fig. 11b Palaeogeographical sketch-map of the Inner West Carpathians in the Jurassic and Lower Cretaceous and junction with the Eastern Aops (M. Mahef 1978)

1. West Carpathian ridge subsequently rebuilt tectonically: A. into nappes: a) of Vysoká group, b) Manín nappe; B. into Tatríde units, mostly parautochthonous; 1a. western continuation of the West Carpathian ridge into the Eastern Alps: A. Mittelostalpin, B. Unterostalpin; 2. Cordillera of the Klippen Belt island zone: a) accompanying trench; 3. Trough with thick continental crust: its marginal near-ridge part; 4. Trough with thinner continental (suboceanic) crust: a) its submarginal part; 5. Trough with oceanic crust, a) with thin suboceanic to oceanic crust; 6. Depressions at the intraoceanic margin: SG — South Gemicide, TD — Transdanubian; 7. Intraoceanic margins

occur in the deeper layers of the latter. In the West Carpathians this zone is roughly identical with the area of the Magura unit and is usually considered to be a strongly reduced root zone of the Flysch Belt nappes, but in the west its internal border follows the western margin of the Little Carpathians, at least 30 km south-east of the surface occurrences of the Klippen Belt. In this way, the West Carpathian zone presumably containing the ophiolite association approaches the easternmost surface occurrence of the Penninicum in the Rechnitzer window. Even in this conception, an at least 30 km displacement of the West Carpathians to the north-west must be postulated in the basement of the Vienna Basin, which is the most prominent transverse superimposed structure at the boundary between the Eastern Alps and the West Carpathians (Fig. 11b).

The ophiolite associations representing the 'North Tethydian' sea continue from the Alpine Penninicum through the lower structural layers of the Magura nappe to the Black flysch-Ceahlău zone in the East Carpathians, hence through the Severin unit and the adjacent eastern margin of the Geticum in the Greben syncline to the zone of Trajan Flysch, and further eastwards through or along the Kotel zone.

The southern Tethyian basin is represented by the ophiolite zones of the Dinarides, Hellenides and of Taurides farther east. This scheme presuming the existence of two oceans in Jurassic and Early Cretaceous times is similar to the models developed by the experts acquainted with the ophiolite zones of other continents (J. DEWEY—J. BIRRD 1970, A. L. KNIPPER 1975) and by some students of the Alps (e.g. M. BOCCALETII et al. 1974).

Both parts of the Tethys were subdivided by islands having a thick continental crust as, for example, by the Briançonais island in the Penninicum sea and the Drina-Ivanjica-Pelagonicum island in the southern Tethys. The oceans also encroached upon a broad intervening belt of landmass, dividing it into microplates. Such were the Bükk gulf between the Central Alps and the eastern Hungarian block, the Szolnok gulf between the Northern Apuseni and the Geticum of the South Carpathians. The differentiation of the Tethys can be partly explained by disruption of intraoceanic plates with continental crust into microplates or minor fragments, combined with drifting apart and rotation of the plates, during the development of the Carpathian mountain arc (E. SZÁDECKY-KARDOSS 1973, V. ALEXIĆ et al. 1974). It is, however, hardly possible to regard the Villány and Mecsek Mts. as part of the original Inner Carpathian belt (P. HORVÁTH—L. STEGENA—B. GÉCZY 1974, M. BLEAHU 1974) either from the correlation of the succession of the Mesozoic tectofacies (not isolated facies!) or from the grade of Palaeo-alpine deformation. The Mesozoic complexes of the Villány and Mecsek Mts. lack such markedly Inner Carpathian tectofacies as are the Keuper and chiefly the Albian-Cenomanian flysch. The two mountain ranges were also very little affected by the Palaeo-alpine folding, incomparably less than the West Carpathian units.

In considering the distribution of the basins or troughs with the oceanic crust, also some sequences that cannot be designated as ophiolitic should be taken into account. These are bathyal to abyssal facies with Jurassic-Lower Cretaceous sediments as an essential component; they are analogous to the sediments of the ophiolite sequences but are accompanied with only rare basites and ultrabasites. They are characteristic of the Pieniny — Kysúca unit of the Klippen Belt, the Krížna nappe in the West Carpathians, the Frankenfels nappe and its analogues in the west of the Northern Calcareous Alps, the Bukovina nappe in the East Carpathians, and partly of the Zukali unit and of the Ionic unit in the Hellenides. During the Jurassic and Early Cretaceous, the Alpides were apparently differentiated into zones with the oceanic crust and into deep troughs with continental but very likely thinned crust. In addition, there was a number of zones with a predominance of shallow-water facies, which were not of epicontinental but of geosynclinal type; they were not platform fragments but geosynclinal ridges with a thicker crust (J. AUBOUIN 1964).

It is difficult for a geologist who many years studied the Mesozoic of the West Carpathians and I venture to say that even for those who worked in the Oberostalpinicum or the Southern Alps, to accept the idea that these areas, situated between the



Northern and Southern Tethys, behaved as slightly active plates showing an epicontinental development, especially in Jurassic and Early Cretaceous times. To restrict the activity of the Alpides only to the ophiolite zones seems to me to be far more schematic than was the division into eu- and miogeosynclines, which had to be extended later by a number of other geosynclinal types. The difficulties met with in trying to incorporate the West Carpathian units in these two basic geosynclinal types on the Tectonic Map of Eurasia (N. ŠATSKIJ—A. BOGDANOV 1964) still stick deep in our memory. The first step in establishing the principles for the Tectonic Map of the Carpathians, Balkan and Dinarides (on 1:1,000,000 scale) was to determine a peculiar type of the Alpidic geosynclines; they should not be characterized by the thickness of sediments or by the proportion of magmatites, particularly basic and ultrabasic, but by its differentiation, i.e., an increased dynamics during its history, which was reflected in the number of the structural-facies zones. The increased dynamics was also manifested by an unusual spatial shortening, by the formation of numerous nappes and zones of contraction. We have labelled this geosynclinal type as the Alpine type. Soon afterwards the term 'aristogeosyncline' was coined by A. TOLLMANN. The differentiation of the geosyncline and its variability in the individual stages are shown on the Tectonic Map of the Carpathians, Balkan and Dinarides by tectonogroups, i.e. groups of tectofacies, which define the palaeotectonic character of the area where a given tectonic unit originated (M. MAHEL 1973). For details see the Explanatory Text to the Tectonic Map (M. MAHEL et al. 1974). The areas with outstanding ophiolite and basite occurrences are denoted as specific types of tectonogroups — specific types of troughs. There are several trough types analogously to various types of ridges, which occur between troughs or swells and link the troughs with the platform. The Penninic troughs have recently been divided into the Valaisian, Piemontian and Ligurian types. The Dinaric troughs could also be divided at least into the Vardar type and the Subpelagonian type proper.

The individual ophiolite troughs differ not only in the accompanying type of sediments, abundance of magmatites but also in the type of ophiolites. Characteristic of the Jurassic-Lower Cretaceous troughs of the Dinarides are abundant diabases and large bodies of gabbro and ultrabasites, products of little differentiated tholeiitic magma, occurring besides pelagic carbonates, cherts and pelites. Distinctive of the Mureş trough is the varied succession of magmatites of the ophiolite type: a) in the Jurassic of simatic character, predominantly basalts, less gabbros and ultrabasites, accompanied by silicites of great thickness; b) in the uppermost Jurassic the abundance of differentiated hybrid magmatites from basalts to quartz porphyry (also alkalic keratophyres and orthophyres) in accompaniment of pelitic carbonates (mainly marlstones); c) in the Aptian and Albian spilites accompanied by heterogeneous flysch (varied in facies). More abundant serpentinites are in the unit of the 'Black Flysch' in the East Carpathians with predominantly aleuritic flysch. The rare diabase volcanism and scarce ultrabasites are also in the Ceahlău-Severin-Kraina

trough of carbonate flysch. For the Krížna trough with pelagic carbonates, radiolarites of the Jurassic and Lower Cretaceous and with Albian-Cenomanian flysch scarce augitites are characteristic. In the Mecsek trough with more deep-water carbonate facies of the Dogger and Malm (with shallowing to the end of the Jurassic) Lower Cretaceous alkalic diabbases are abundant.

There are obviously several types of troughs, differing in the stratigraphic range of filling; in the type of sedimentary facies and their changes throughout history; in the amounts and type of ophiolites and thus different also in type of crust (the oceanic, suboceanic, thinned sialic crust).

The different types of troughs were greatly controlled by the differences in their position in the individual segments of the geosynclinal system of the Alpides and by the differences in the dynamics of the crust. But as is evident from the course of the Northern and Southern Tethyian troughs, the nonuniform oceanic troughs are obviously of regional character. The troughs with the oceanic crust are geotectonic elements of a higher order of magnitude connecting several segments, whilst the troughs with a thinner continental crust (e.g. the Krížna trough) are usually characteristic of one Alpe segment only. The global character of these oceanic troughs is also suggested by their opening during the Jurassic synchronously with the formation of the Atlantic Ocean between North America and North Africa (J. DEWEY et al. 1970).

To disregard the differences in the dynamic state of the crust in the individual Alpe zones means to omit the basic character of the development history of the Alpides.

The palaeo-tectonic model of the Alpides during the Alpine cycle was extremely complicated, changing in time and space not only within single sectors but also in single evolution stages. The most prominent and inseparable component of this model from the Middle Triassic to the folding phase were the troughs, some with ophiolites and some with deep-water facies sequences accompanied by basites or ultrabasites.

The zonal and at the same time selective distribution of ophiolite associations and their connection with the troughs result from the differentiation and prominent diversity of the Alpine geosyncline and are associated with the 'oceanization' of the crust in intercontinental conditions. Longitudinal deep faults, which separated ridges and troughs of the Alpine geosyncline, were paths of ascent of basites and ultrabasites from the deeper parts of the crust and were significant factors in the extension of the crust, thinning of its granite-gneiss layer and occasionally in drifting apart of the plates. Another factor that influenced the development of the Alpides, the distribution of ridges and troughs and even the dynamics of the crust, and which should not be forgotten, is the non-uniform Hercynian granitization (Figs. 7, 12a, 12b).

Another characteristic feature of the Alpides is a great, in some zones even

enormous shortening and the formation of often rootless nappes. Particularly the zones with ophiolite associations and those with a thinner continental crust (troughs) have been deformed into extensive nappes. Under these conditions the allochthonous, mostly nappe position of the ophiolite associations and basites accompanied by deep-water sediments is understandable.

5. Two tectonically contradictory events of primary importance in the history of the fold mountains are connected with the ophiolite zones:

— extension of the crust, drifting apart of the plates, i.e. the origin of oceanic basins with ophiolite associations or at least crustal thinning and the origin of troughs with basic rocks;

— shortening of space, closing of troughs, subduction of the oceanic crust and its submerging to the deeper crustal parts, re-melting of the oceanic and even continental crust, which implies the interference with the deep-seated processes.

The extension processes of continental crust represent a new approach to the understanding of the history of fold mountains, and in the Alpine system to a more dynamic interpretation of the development of the geosynclinal systems. They show the differences in the width of the geosynclinal system and in the diversity and type of the crust during its development more distinctly. In the European Alpides it is of particular importance that the geosynclinal development before the Palaeo-alpine folding (often called the carbonate stage) can be, as a result, divided into two stages — the Triassic and Jurassic-Lower Cretaceous.

The shortening of geosynclines was so far associated with the periods of folding. The shortening due to subduction of the oceanic crust, which is regarded as a long-term process, usually preceding folding, opened a new concept. The margins (mainly internal) of the ophiolite zones are thought to be Benioff seismic zones involved in subduction and the basic evolutionary activity of fold mountains is ascribed to them. The frequent metamorphism in the ophiolite series, olistostrome mélanges and olistoliths are often used as evidence of a long-lasting subduction and activity in the contact zones between the ophiolite troughs and the marginal lithospheric plates.

Some of the ophiolite zones show a more or less intensive metamorphism of prehnite-quartz subfacies as, for example, the Meliata Formation, the Bükk Mts., the Strandža, Vardar and Fruška gora zones, and the Penninicum in the Eastern Alps. Strikingly, all of them are interior zones, except for the Penninicum, which occupies an interior position only in the Western Alps. It also shows a higher grade of metamorphism, but in the Alpides the metamorphism of Mesozoic formations is not a characteristic feature. The classics of the nappe theory had long ago interpreted metamorphism as an indicator of the root zones. Metamorphism is also known from the Mesozoic non-ophiolitic associations, more frequently in trough zones (root zones of nappes, e.g. the Veľký Bok Formation, rear part of the Križna nappe in the West Carpathians) than in the zones of strong Alpine compression (e.g. the

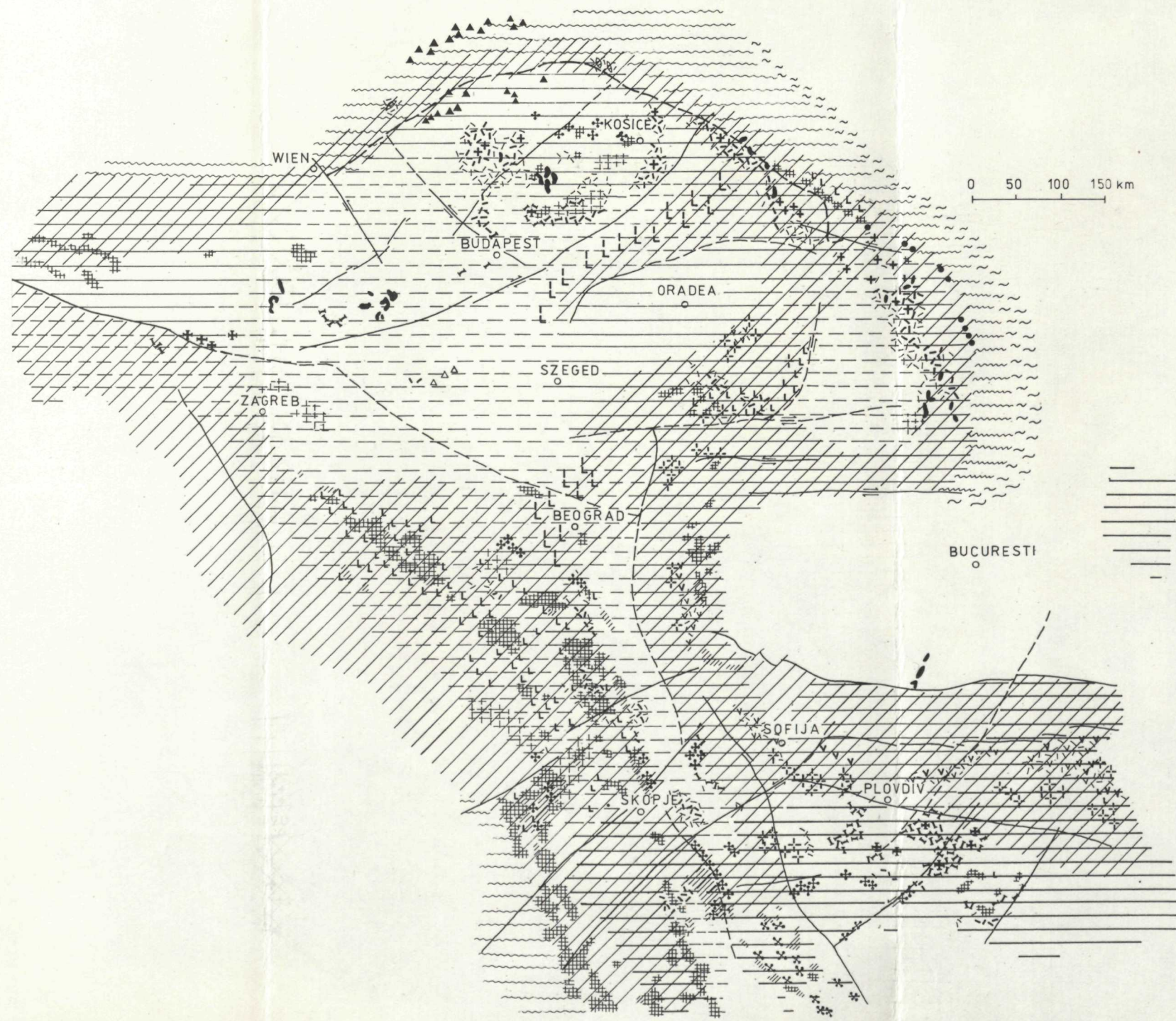
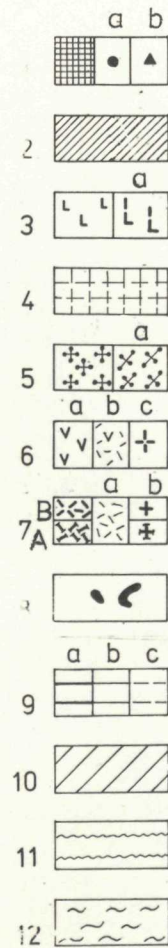


Fig. 12a Sketch-map showing manifestations of Alpine folding periods in relation to magmatites in the Carpathian — Balkan system.

- Magmatites**
1. Ultrabasites a) minor bodies in allochthonous position. b) alkalic ultrabasites
  2. Intrusive basites (mainly gabbros)
  3. Spilites and diabases (Jurassic-Lower-Cretaceous), a) underlying the Tertiary basin filling, b) quartz porphyries, keratophyres, diabases, c) alkalic basites and ultrabasites
  4. Triassic diabases, porphyrites, quartz porphyrites, keratophyres
  5. Palaeoalpine granites a) synkinematic
  6. Mesozoic banatites a) major effusive bodies, b) major pyroclastics, c) subvolcanic and plutonic bodies
  7. Tertiary volcanics A) early late-geosynclinal (Priabone-Oligocene), B) late-geosynclinal (Miocene-Pliocene), a) pyroclastics, b) subvolcanic and intrusive bodies,
  8. Basalts
- Regions**
9. a) Distinctly affected by Kimmerian (mainly Neokimmerian) folding, b) Palaeo-alpine folding (in the Austrian to Subhercynian phase), c) only slightly affected by Palaeo-alpine folding
  10. Folded in the Meso-alpine (Laramide to Pyrenean) phases
  11. Folded in the Neo-alpine (Savian to Styrian) phases
  12. Folded in late Neo-alpine phases (at the end of the Neogene and in the Quaternary)

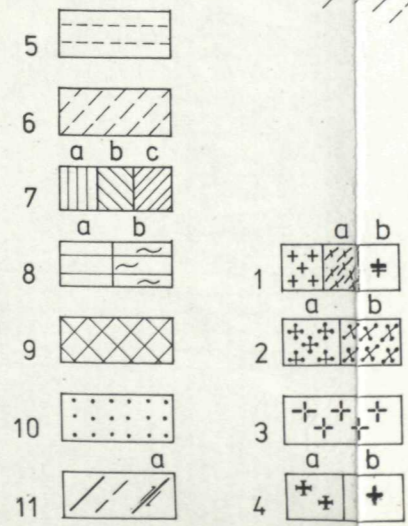
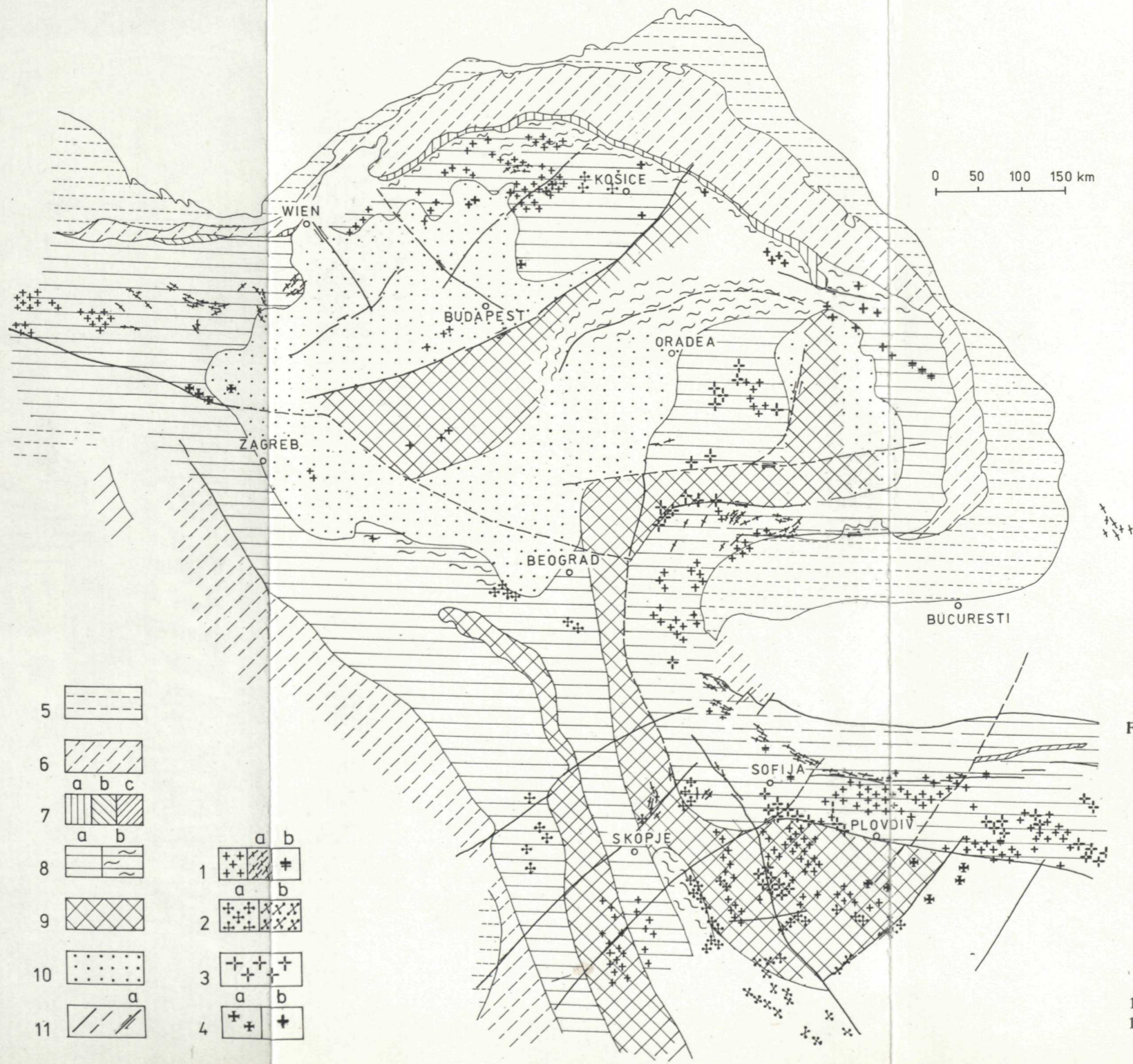


Fig. 12b Geotectonic position of granitoids

1. Prealpine granitoids a) synkinematic, b) alkalic
2. a) Palaeo-alpine granites, b) synkinematic
3. Meso-alpine banatites — interorogenic
4. Subvolcanic and intrusive bodies a) early late-geosynclinal (Priabonian — Oligocene), b) late-geosynclinal (Miocene — Pliocene)
5. Foredeep
6. Flysch Belt of the Carpathians, Eastern Alps, outer zones of the Dinarides-Hellenides and Balkans
7. a) Klippen Belt, b) Gresten Zone, c) Kotel Zone
8. a) inner zones, b) late-tectonic (back-arc) Flysch
9. Intra-geosynclinal massifs
10. Inner depressions
11. Principal tectonic lines a) strike-slip faults

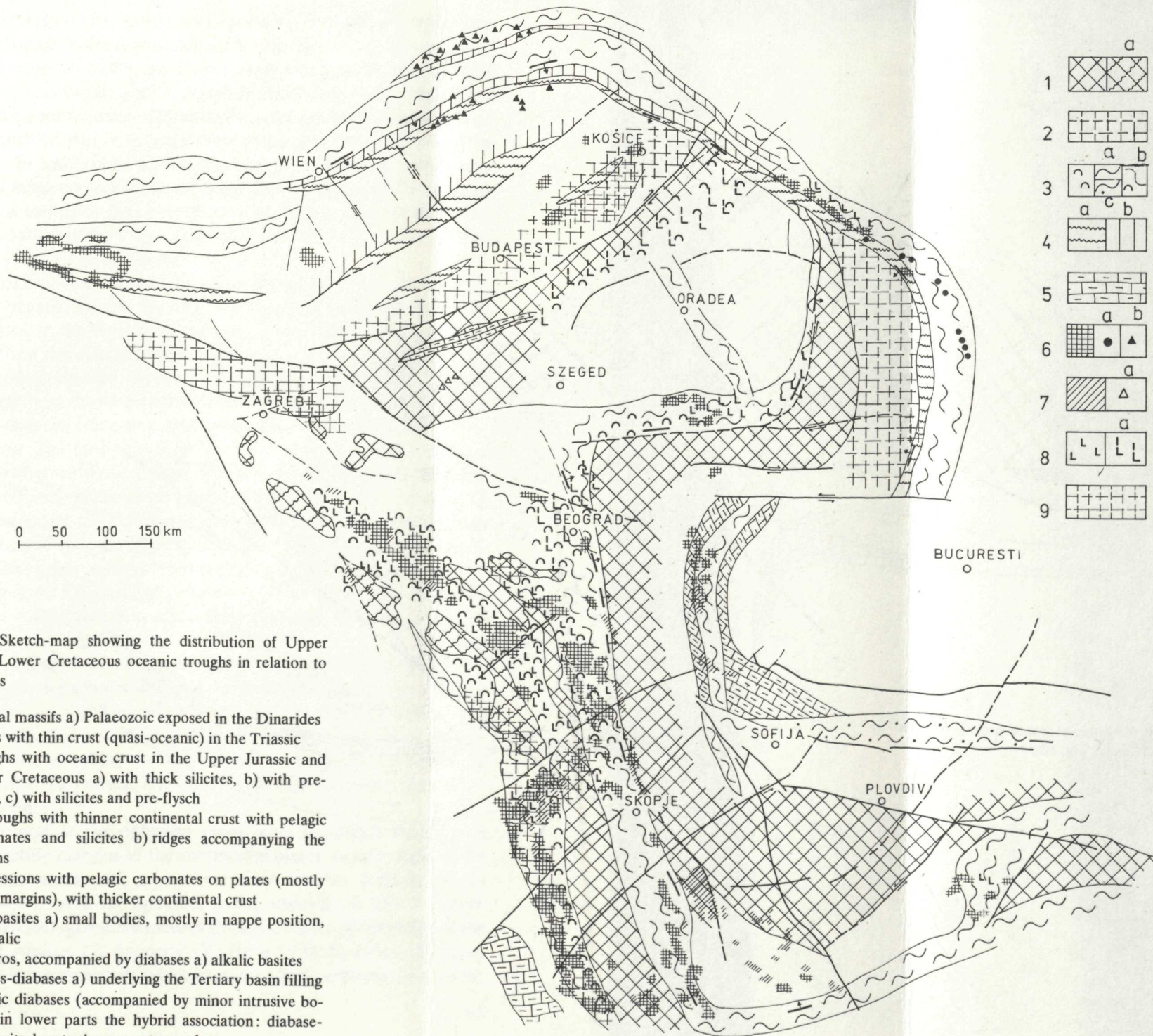


Fig. 13a Sketch-map showing the distribution of Upper Jurassic-Lower Cretaceous oceanic troughs in relation to ophiolites

1. Central massifs a) Palaeozoic exposed in the Dinarides
2. Areas with thin crust (quasi-oceanic) in the Triassic
3. Troughs with oceanic crust in the Upper Jurassic and Lower Cretaceous a) with thick silicites, b) with pre-flysch, c) with silicites and pre-flysch
4. a) Troughs with thinner continental crust with pelagic carbonates and silicites b) ridges accompanying the troughs
5. Depressions with pelagic carbonates on plates (mostly along margins), with thicker continental crust
6. Ultrabasites a) small bodies, mostly in nappe position, b) alkalic
7. Gabbros, accompanied by diabases a) alkalic basites
8. Spilites-diabases a) underlying the Tertiary basin filling
9. Triassic diabases (accompanied by minor intrusive bodies), in lower parts the hybrid association: diabase-porphyrite-keratophyre-quartz porphyry

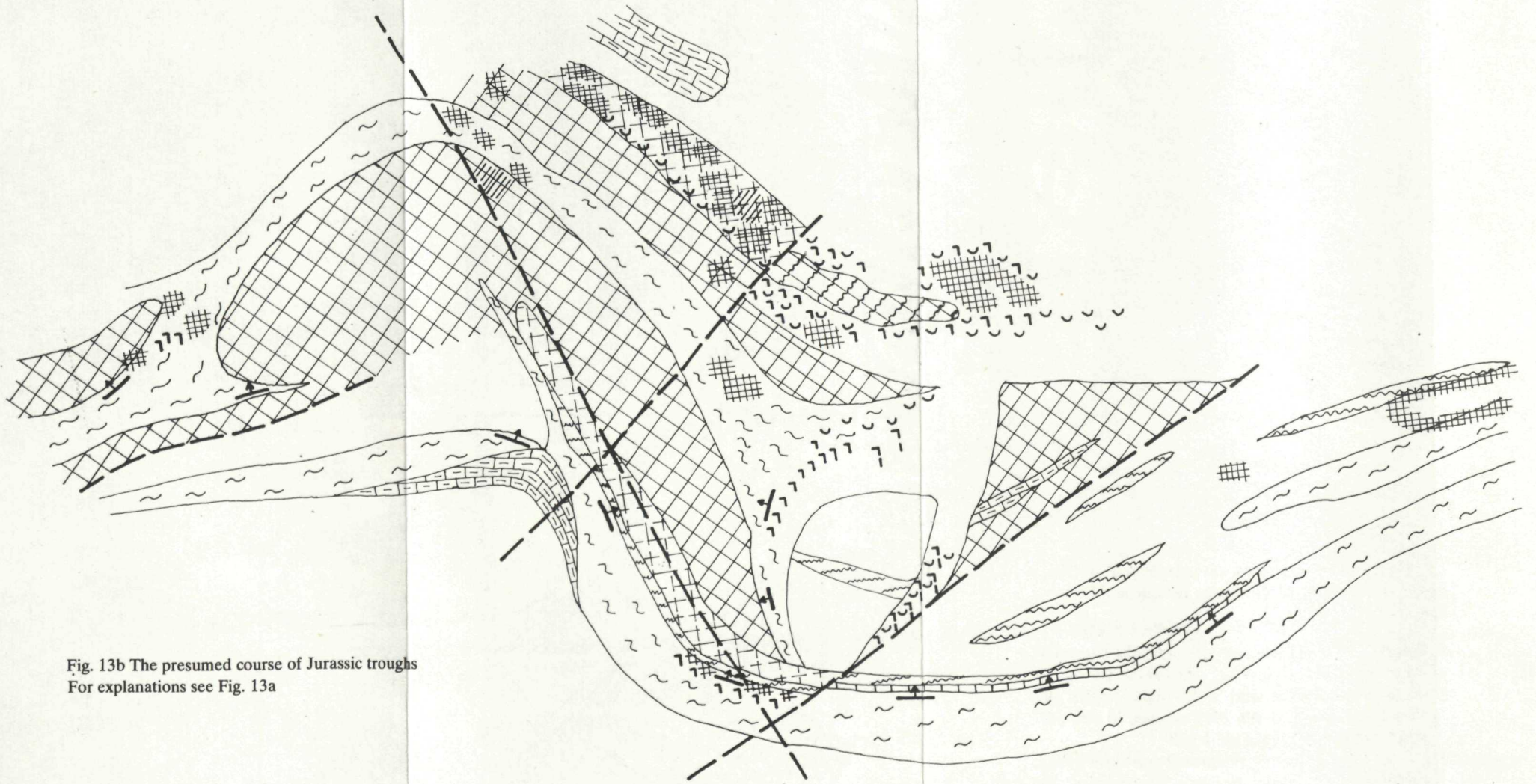


Fig. 13b The presumed course of Jurassic troughs  
For explanations see Fig. 13a

Stružník Formation of ridge type in the South Veporides). It should be emphasized that shortening in the Alpidic realm is not confined only to the ophiolite zones. In the West Carpathians, all zones including the Tatríde zone which is considered to be autochthonous, show at least locally a nappe structure. The nappes in the High Tatra Mts. have been known for some time and thrust faults, overthrusts and minor nappes have recently been identified even in the crystalline basement as, for example, in the Považský Inovec Mts. In addition to a total shortening, some zones, particularly trough zones with a thinner continental crust, were affected by particularly strong shortening. This relates mainly to the root zones of the Krížna nappe in the North Veporides and the zone south of the Veporides, which is connected with the Lubeník suture (the root zone of the Choč and higher nappes).

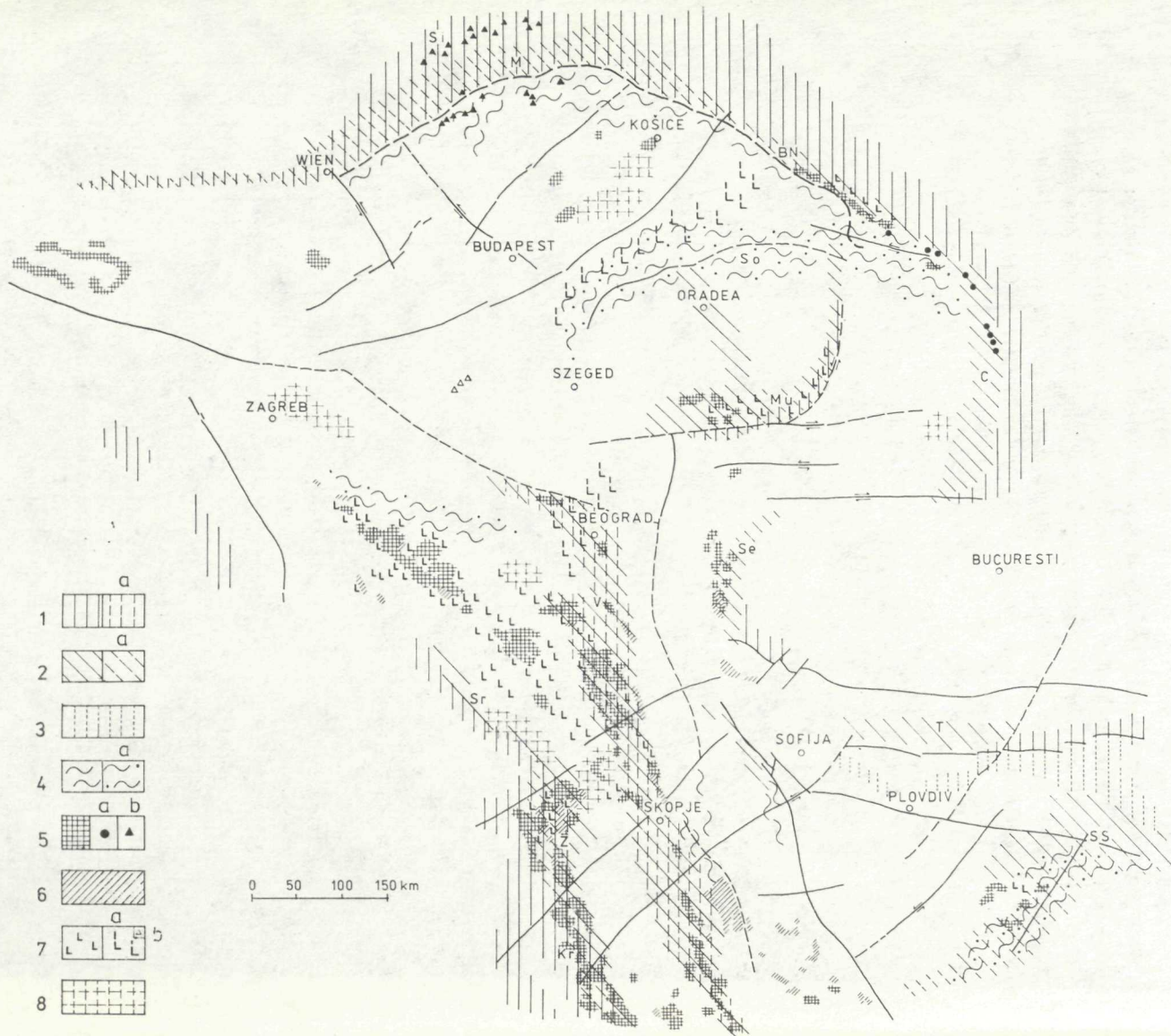
In the Klippen Belt of the West Carpathians, which is rather a marginal zone of the island arc and narrow oceanic trough, the long-term shortening is reflected in the wildflysch conglomerates. In the West Slovakian part of the Klippen Belt polymictic conglomerates occur from the Albian to the Palaeocene and monomictic from the Palaeocene to the Lower Eocene; in its East Slovakian part the polymictic conglomerates are found from the Maastrichtian to the Lutetian (R. MARSCHALKO et al. 1976). The pebble material contains varied exotic magmatites, including basites and even glaucophanites. The most abundant limestone pebbles are derived from several geotectonically contrasting sequences. The changes in the pebble material very likely reflect the changes in the palaeogeographical pattern and in the source areas and suggest the existence of a 'Ultrapieninian exotic ridge' and its subsequent engulfing (D. ANDRUSOV 1968). It should be stressed that the conglomerates occur in the flysch sequence without any marked interruption of sedimentation.

The great shortening of the Klippen Belt, which was (30 to 40 times) greater than shortening of any other Alpidic area, was a long-lasting process. The same also holds for the folding process in the Alpides.

A detailed analysis of folding in the Carpathians (especially West Carpathians), the Balkan Mts. and Dinarides shows that it was active for a long time and in several phases. For this reason, on the Tectonic Map of the Carpathians, Balkan and Dinarides the units are divided according to the period of folding (Palaeo-, Meso- and Neo-alpine) and not according to their formation in shorter phases. The shortening of troughs with the oceanic crust is the only and long-term type of the shortening of a geosyncline.

6. In the neighbourhood of the ophiolite zones were deposited thick flysch complexes along the mobile margins of the continental plates, or in troughs at the boundary between these areas with an oceanic crust and the ophiolite zones (Fig. 14). The distribution of some ophiolite zones adjacent to and in lateral extension of the flysch zones is given as a piece of evidence for the applicability of the global tectonics to the Alpides (E. SZÁDECKY-KARDOSS 1973, N. HERZ—H. SAVU 1974). Since the flysch is extremely abundant in some Alpidic segments, e.g. in the





Alps, East and West Carpathians, Dinarides and Hellenides, the relationship between the flysch and the ophiolite associations is of primary importance.

We have mentioned above that some ophiolite associations grade upwards into flysch associations containing basic and even ultrabasic rocks (the Vardar zone, Metalliferous zone, Penninicum, the 'Black flysch' unit, Strandža). The Preflysch, which is the initial stage of the flysch troughs, contains more basites and ultrabasites. At the same time they alternate chronologically with the ophiolite associations, usually occurring in the neighbourhood (Sarajevo and Vermos along the margin of the Ophiolite and Mirdita zones, Severin at the margin of the Greben, Ceahlău zone at the margin of the 'Black flysch' unit).

The complexes of mature flysch represent another dynamic crustal type than that of the ophiolite zones. It is the type which replaces the period of dilatation, of ocean opening. The Triassic ophiolites (the Bükk Mts., Meliata Formation) lack flysch accompaniment. In the Budva and Zukali units there are flysch complexes of a fairly great thickness but they belong to trough formations characterized by a spilite-keratophyre association.

As is seen, there is a time relation between the ophiolite associations and the flysch complexes in the Alpides. The complexes of mature flysch with predominating flysch s.s. and coarse flysch usually begin in the Middle Cretaceous and lack the ophiolites. The rather rare magmatites are of intermediate to acid character (for details see p. 72—73).

In the structural-facies zones where the carbonate sequences replace the flysch complexes as late as in the Albian, the basites and ultrabasites are more frequent in the transitional (Preflysch) period (the Križna and Tatríde units). This period of the maximum subsidence and the thinnest crust is evidently connected with the occurrence of basites and sporadic ultrabasites in the pebble material of the conglomerate interlayers in the flysch complexes.

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Fig. 14a Sketch-map showing the distribution of ophiolites in relation to flysch tectonogroups

#### Flysch tectonogroups

1. Upper Cretaceous-Palaeogene Flysch Belt a) in places superimposed on pre-flysch zones. Units: M = Magura, Si = Silesian, K = Kruja, J = Ionian
2. Tithonian-Lower Cretaceous flysch (predominantly pre-flysch) a) underlying the later Flysch: So — Szolnok, Bn — 'Black Flysch', C — Ceahlău, Se — Severin, T — Trojan, SS — Strandja, Sr — Sarajevo, V — Vardar Zone, Mu — Mureş, Z — Zukali
3. Upper Cretaceous volcanogenic flysch showing relationship to banatites
4. Flysch to flyschoid of back-arc type a) underlying the molasse filling
5. Ultrabasites a) small bodies, mostly in nappe position, b) alkalic
6. Gabbros (accompanied by diabases)
7. Spilites-diabases, (subordinately the hybrid association a) underlying the Tertiary basin filling, b) alkalic
8. Triassic diabases accompanied by small intrusives; in lower parts the hybrid association: diabase-porphyrite-keratophyre-quartz porphyry

The distribution of flysch complexes and its time range in most of the Alpine segments point to an orogenetic polarity (J. AUBOUIN 1964, 1964, M. MAHEL 1974), which is one of the characteristics of the Alpides in the Alpine cycle. It is this time sequence that shows that not only flysch but also ophiolite complexes fit in the evolution line of the Alpine geosyncline.

7. The ophiolites show the petrochemical character of tholeiitic magma in the zones with oceanic or thin sialic crust and of more alkalic type (augitites, limburgites) in the zones with a thicker crust (e.g. in the West Carpathians — D. HOVORKA 1965, 1976). The character of magmatites of the ophiolite association also changes with time. The Triassic ophiolites contain more acid rocks, even quartz porphyries as, for example, in the Anisian of the Bükk Mts., keratophyres in the Meliata Formation in the Slovakian Karst, quartz keratophyres in the Anisian and Ladinian of the Dinarides and in the Ladinian of the South Carpathians. The petrochemical type of magma corresponds to the first stage of differentiation of the crust leading to a marked crustal thinning. More frequent extrusions are accounted for by a diapir-like ascent of contaminated magma (S. KARAMATA 1974).

In the Jurassic and Lower Cretaceous, which represent the period of most pronounced oceanization, basic rocks predominate and ultrabasic rocks are frequent. In the Middle and Upper Cretaceous, when flysch sedimentation prevailed, the intermediate rock types with more frequent but minor bodies of porphyrites (palaeo-andesites) and quartz porphyrites (palaeo-dacites) became more abundant, e.g. the Flysch Belt of the East Carpathians — M. G. LOMIDZE 1968; the Flysch Belt of the West Carpathians; Mureš zone). The type of the magma obviously corresponds to the changes in the crustal type, as they proceeded from the Hercynian consolidation and granitization through the early Alpine differentiation and crustal thinning to the leptogeosynclinal stage with most extensive oceanization and to the Neo-alpine consolidation.

##### **5. Meso-alpine interorogenic magmatites (the banatite association)**

In the Srednogorie zone in the Balkan Mts., in the Timok and Poiană Rusca zones in the South Carpathians and in the Apuseni Mts. there is a volcano-plutonic formation known from the literature under the label 'banatites'. On the Tectonic Map of the Carpathian-Balkan Region (M. MAHEL 1973) this formation is assigned to the Meso-alpine interorogenic magmatites with regard to its being affected by Meso-alpine folding and to its post-Palaeo-alpine age (Fig. 12a). Each of the areas mentioned above shows, of course, some particularities in geotectonics and the petrological character of magmatites (Fig. 2).

In the Srednogorie (Bulgaria) these volcanics make up a considerable part of the filling of graben-synclinoria, which shows a different character in the eastern part

from that in the west of the Srednogorie (Fig. 6). A flysch filling of a great thickness (Turonian to Maastrichtian) predominates in the east. The magmatites are distributed above the Turonian carbonate flysch and underlie the Upper Senonian flysch, which contains a considerable proportion of marlstones, marly limestones and sandstones with interbeds of andesite tuffs and tuffites (J. JOVČEV et al. 1971). The Turonian-Maastrichtian volcano-sedimentary complex is up to 3000 m thick, with a prevalence of andesite tuffs, effusions of small thickness and subvolcanic andesite bodies. This alkali-calcic andesite-basalt association comprises basalts, andesite-basalts, andesites, dacites, rhyodacites and intrusive bodies of gabbro, gabbrodiorite, diorite, quartz diorite and plagiogranite. In the Upper Senonian, the flysch complex is in places substituted laterally by the alkalic olivine basalt formation consisting of picrites, olivine basalts, basanites, analcime-bearing basalt, limburgites, augites, trachybasalts, trachyandesites, trachytes and intrusive monzonites and syenites (G. STANIŠEVA—VASILJEVA 1973).

In this eastern part of the Balkanides the 'banatite' magmatites occur to a less extent also in the northerly Luda Kamčija flysch zone under the Campanian-Maastrichtian (Emin) flysch. They are more widely distributed south of the Burgas synclinorium in the Strandža area (mainly in the Campanian) and involve tuffs, tuffites, lava flows, small subvolcanic and hypabyssal bodies and numerous dykes of trachyandesite composition. The numerous dykes are most often 2—3 km long, striking along the axes of the main structures of the Burgas synclinorium (100—120°) but also transversally (30—40°) (T. T. VELCHEV et al. 1975).

In the western part of the Srednogorie Mts. a Coniacian-Santonian volcano-sedimentary complex of unusually great thickness is in the Panagurište synclinorium. It overlies a Turonian coal-bearing sandstone-conglomerate sequence. The Upper Santonian variegated marly limestones (Couches rouges facies) are also interbedded with tuffs and tuffites. The overlying thick flysch complex of Maastrichtian age is devoid of volcanics. B. BOGDANOV (1973) reports the following succession of volcanic effusions: a) andesites accompanied by their agglomerates and tuffs, b) dacites and their tuffs and agglomerates (both andesites and dacites are accompanied by diorites and granites), c) dykes and subvolcanic bodies of dacites, d) subvolcanic diorite-porphyrites and e) trachyandesites and basalts filling the joints.

In the Sofia area the stratigraphic range of magmatites is wider. They already appear in the Turonian flysch and in the Senonian sequence. Major plutons constitute subvertical to vertical bodies distributed on faults along the margins of the synclinoria, in places tens of kilometres away from the Upper Cretaceous sedimentary-volcanogenic filling of synclinal basins. Opinions on their age differ; many authors believe them to be younger than the bulk of the volcanic mass associated with the Laramide orogeny. The contact metamorphism of the Upper Cretaceous complexes lends support to this view (e.g. in ancient Vitoša volcano). The above succession of magmatites (G. STANIŠEVA—VASILJEVA 1973, B. BOGDANOV 1973), as

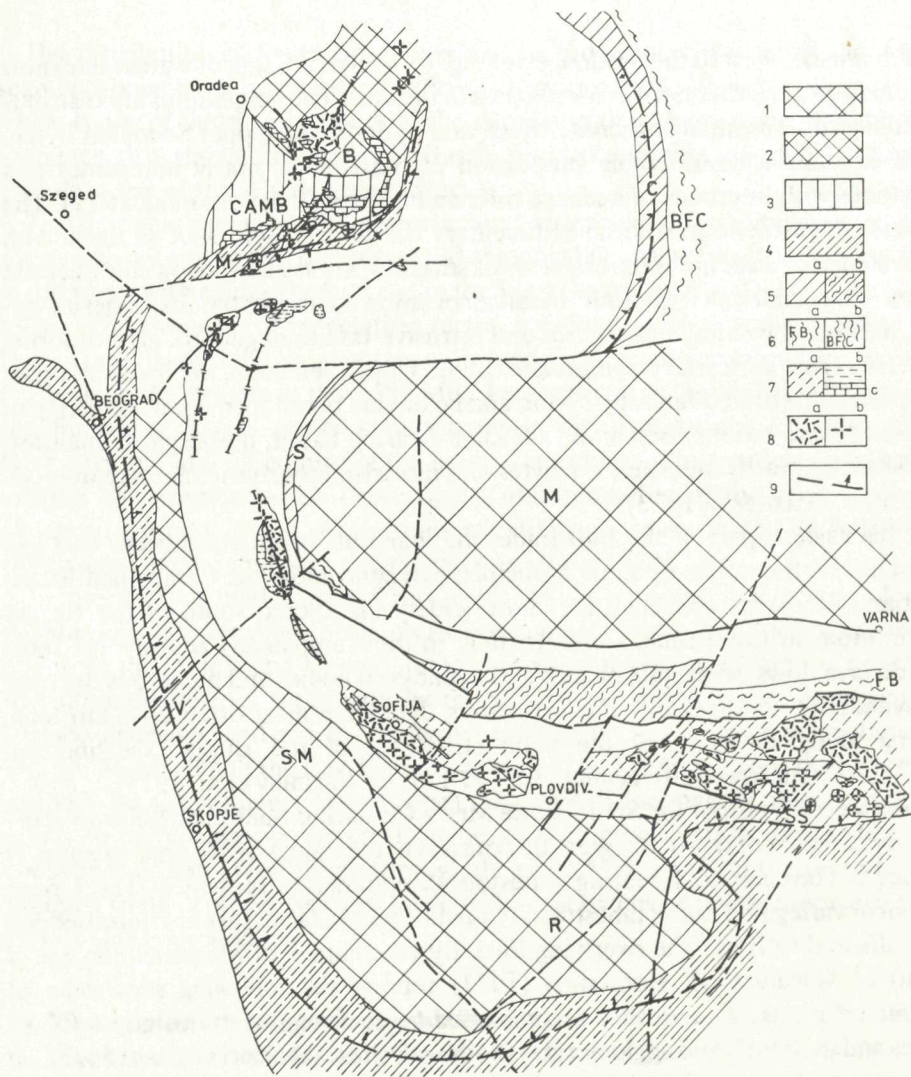


Fig. 14b Sketch showing the geotectonic position of banatites

1. Platforms (M = Moesian) and central massifs (SM = Serbo-Macedonian, R = Rhodope)
2. Bihor autochthon
3. Nappe units of Northern Apuseni Mts.
4. Zones with ophiolites: M = Metalliferous, V = Vardar a) Trojan Flysch zone (Tithonian — Lower Cretaceous)
5. a) Zones of pre-flysch with accompaniment of basites and ultrabasites: C = Ceahlău, S = Severin, SS = Strandja, b) Trojan flysch (Tithonian — Lower Cretaceous)
6. a) Flysch Zone of the Balkanides = FB, b) BFC = the main flysch zone of the East Carpathians
7. Upper Cretaceous complexes accompanying banatites a) flysch, b) various volcanic-sedimentary facies, c) Gosau type, little affected by folding
8. Banatites a) major effusive bodies, b) subvolcanic and plutonic bodies
9. Subduction zones

well as the radiometric ages (70—80 m. y. — S. BOYADJIEV 1974) and occasional contact metamorphic effects (e.g. of the Plana intrusion) only in the lower Upper Cretaceous sedimentary complex suggest that part of the plutonic rocks is Upper Cretaceous in age and syngenetic with the volcanics. The volcano-sedimentary complex of the Srednogorie is intensely folded together with the sedimentary filling of the synclinoria.

In the South Carpathians (Figs. 5, 14b) the volcanics form part of the filling of the Timok synclinorium-graben, having a stratigraphic range of Cenomanian — Eocene. The up to 3000 m thick filling consists of unusually varied rocks: the shallow-water conglomerates, breccias, sandstones (Cenomanian, Maastrichtian, Danian), organogenic limestones (Maastrichtian), deeper-water pelites (Inoceranian marlstones — Turonian and Senonian) and volcanics. Andesites already appear in the lowest Cenomanian complex of conglomerates, breccias and sandstones. Hornblende andesites are more abundant in the Lower Senonian and pyroxene andesites and andesite-basalts in the Upper Senonian. Trachyte-andesites, latites and granite, diorite, quartz diorite, grandiorite, monzonite and syenite intrusions are post-Late Cretaceous (P. STEVANOVIĆ et al. 1967). The products of high-temperature volcanism — andesites and dacites accompanied by quartz diorites and granodiorites — are the oldest; low-temperature latites, quartz latites, granodiorites and quartz monzonites are younger, which indicates a palingenetic character of magma (S. KARAMATA). Frequent bodies of monzonite, diorite, syenite, syenite-diorite, granodiorite and also gabbro, diorite-porphyrite and monzonite-porphyrite occur particularly in the northern part of the Timok graben. The intrusions and alkalic volcanics, mainly K-rich andesites and dacites, and basalts and nepheline basalts are located predominantly along the longitudinal Ridan-Krepoljin deep-fault system. They are mainly attributed to the Laramide to Pyrenean orogenic movements (S. KARAMATA et al. 1967, S. KARAMATA 1974). Subvolcanic bodies, sills and dykes of granite to gabbro composition also accompany the Knjaževac-Pirot tectonic line (in the Majdanpek area of Yugoslavia) to which the andesite and basalt flows and their pyroclastics are related. This line acted as a controlling factor in the evolution of the Timok graben.

In the northern part of the South Carpathians, in the Rusca Montana basin, Maastrichtian andesite pyroclastics, agglomerates and tuffs occur amidst the Campanian to Danian detrital coal-bearing complex (D. GIUSCĂ—G. CIOFLICA—H. SAVU 1966). The dykes of basaltoid andesite trend from west to east. The Eocene sequence is represented by andesites, dacites and rhyolites accompanied by diorites, granodiorites and aplites. At the eastern margin of the Semenic range, quartz diorite, diorite and grandiorite porphyry dykes penetrate the crystalline complex along an up to 25 km long tectonic line.

The banatites in the Apuseni Mts. (Fig. 4) show a specific geotectonic position. Minor diorite and granodiorite massifs accompanied by rhyolite and andesite

occurrences are found especially in the southern Metalliferous zone. They penetrate not only the crystalline schists but also a complex of Mesozoic ophiolites and the Jurassic and Lower Cretaceous sediments (Fig. 14b). They are distributed mainly along two NE-SW fault zones. Both the major Săvârşin and Casanesti bodies in the west and the more eastern linearly arranged minor quartz diorite, diorite, grandiorite and granite-porphry bodies with aplite and pegmatite veins are connected with the basins of Upper Cretaceous beds to the north. This shows their genetic association with the development of the Upper Cretaceous structural pattern (D. GIUSCĂ—G. CIOFLICA—H. SAVU 1966). The intrusions produced intensive contact metamorphism: marblization and skarnization of Jurassic, Neocomian and Senonian limestones and formation of silicites at the contact with crystalline rocks and basalts of the Palaeo-alpine ophiolite association.

In the Northern Apuseni Mts. (Fig. 4) the banatites make up the extensive Vlădeasa mass; it is located in a complex, roughly NE-SW trending graben and for the most part composed of andesite flows cut across by dacite and rhyolite dykes. Agglomerates, tuffs and tuffites, andesites, dacites and rhyolites alternate with sedimentary rocks of the Senonian—Ypresian Gosau facies. The Upper Cretaceous sediments are contact-metamorphosed in places. The lower part of the Vlădeasa mass consists of basalt, andesite, rhyodacite and granite-porphry dykes and diorite, granodiorite and granite bodies (M. BLEAHU 1974). Granodiorite-porphyrines, pegmatites and aplites are also present. At the south-western margin of the mass the granodiorite bodies are penetrated by rhyolite dykes of NW-SE strike (D. GIUSCĂ—G. CIOFLICA—H. SAVU 1966). Volcanism was obviously active to a small degree even after the banatite intrusion, probably in the Palaeocene. The volcanic bodies maintain the basic NE-SW strike also in the north of the Apuseni Mts., in the Padurea Craiului range, both in minor basins of Upper Cretaceous sediments (Borod, Cornăţel) and in the neighbouring crystalline complexes (Cincea area).

In the south-western part of the Apuseni Mts., however, (Biharia massif), numerous dykes, sills and bodies of granodiorite, granodiorite porphyry, andesite, dacite and rhyolite, which pierce the crystalline, Permian, Triassic, Jurassic and occasionally Campanian to Maastrichtian sedimentary complexes are arranged not only in the NE-SW but also in the NW-SE direction.

The volcano-plutonic banatite association in the Apuseni Mts. shows a minor material variety than the banatites of the South Carpathians and the Srednogorie Mts. Alkaline rock types — syenites, monzonites and essexites are very subordinate (G. CIOFLICA 1967, 1973).

The presence of banatites of a similar type to that of the Apuseni (including the Vlădeasa mass) is assumed beneath the thick cover of the Hungarian basin (N. HERZ—H. SAVU 1974). The Upper Eocene diorite and diorite porphyry bodies occurring together with subvolcanic and effusive andesites in the Mátra Mts. (near Reczk, T. ZELENKA 1974) are believed to belong to them (for details see p. 85). L.

ROZLOŽNÍK (1976) postulates analogous banatite bodies to occur also in the Slovenské stredohorie Mts. (e.g. the Hodruša body). This author interprets their appreciably lower radiometric age values (about 15. m. y.) in terms of reheating by the later Badenian-Sarmatian magmatites.

Some geologists extend the occurrence of the banatite association north-westwards into the Klippen and Peri-klippen areas (E. K. LAZARENKO et al. 1973). In support of their view they give the finds of andesite breccias and especially the abundance of alkaline granites with radiometric ages of 88—130 m. y. in the Upper Cretaceous (beginning with the Albian) and Palaeogene conglomerates (S. S. KRUGLOV—A. K. BOJKO—B. B. ZAJDIS 1974, M. KRIVÝ 1969, L. KAMENICKÝ 1973). They include in the banatite association also the stocks and dykes of grandiorite-porphyry and syenite-porphyry in the Rumanian part of the Marmaroš massif and lamprophyres in the Čivčinské gory Mts. in the Ukraine.

Towards the end of the Early Cretaceous and in the Late Cretaceous, volcanism was active in the Flysch Belt of the East Carpathians. Andesite-dacite effusions are concentrated in several zones amidst the flysch sequences. The volcanic activity occurred in several phases and was genetically associated with the diversification of the flysch geosyncline into troughs and cordilleras. Tuffs and bentonites are known from the Barremian of the Silesian unit, from the Barremian-Aptian of the Subsilesian unit, from the Aptian-Albian of the Audia unit, from the Albian-Turonian of the Tarcău unit (Lupchianu Beds), in the Turonian of the Audia unit and in the Čorna hora and the Skibowa tectonic units. The andesite-dacite tuffites in the Late Senonian of the Dukla, Čorna hora, Subsilesian tectonic units and in the Boryslav-Pokuty folds represent the later phase.

Traces of volcanic activity also exist in the Palaeogene sequences. In the Slovakian part of the East Carpathians, for example, the tuffs and tuffites of acid and intermediate rocks date from the Middle Eocene (Zlín Formation), and Late Eocene to Early Oligocene (Malcov Beds and Menilitic Complex; B. LEŠKO et al. 1958). Volcanic breccias and andesites occur in the Palaeogene of the Curbicortale unit (M. SÂNDULESCU et al. 1974). Acid tuffs and tuffites are widely distributed in the lower layers of the Oligocene complexes; they reflect the Pyrenean diastrophism and are connected with the palaeogeographical rebuilding of the Flysch Belt at the Eocene/Oligocene boundary.

## Discussion

1. The interorogenic magmatites (the banatite association) are analogous to the late geosynclinal Late Tertiary volcanics in their petrographic and partly volcanological character. They were generated after the main Palaeo-alpine folding and are tied with the superimposed structures of the new Meso-alpine structural pattern. For this reason they are usually considered to be subsequent or early subsequent (H. STILLE 1953, A. TOLLMANN 1967, D. GIUSCĂ—G. GIOFLICA—H. SAVU 1966, M.



BLEAHU 1974). The principal associations are rhyolite-andesite and trachyte-andesite-latitude, showing an explosivity of up to 90 %, similarly as the late geosynclinal volcanics. In addition to volcanogenic formations of Mediterranean type they contain alkalic rocks such as limburgites, picrites, olivine basalts and trachytes, whose explosivity is often only 40 %. The Cretaceous-Palaeocene volcanics differ from young volcanics not only in age but also in their position within the geotectonic cycle. They are interorogenic, formed after intense Palaeo-alpine folding but affected by Meso-alpine (Laramide-Pyrenean) folding. Except for the banatites of the Apuseni Mts., all magmatites are situated in the graben-synclines, which are generally parallel to the Palaeo-alpine structures. The early molasse and mainly the flysch facies are sedimentary components of the graben-syncline filling. The volcanic complexes are the product of polygenic submarine volcanoes of central type and are characterized by typical volcanics (lavas and pyroclastics) grading to subvolcanic and hypabyssal bodies (G. STANIŠEVA—VASILJEVA 1973). The volcano-plutonic complexes are very abundant and varied, comprising gabbro, gabbrodiorite, diorite, granite, monzonite, syenite, syenite-granite and granodiorite-granosyenite bodies and dykes of porphyrite, diorite porphyrite, granite porphyry and lamprophyre. As is seen, a distinctive feature of the banatite association is the pronounced volcano-plutonic character with a high proportion of subvolcanic bodies: sphenoliths, intrusions of mantle type and abundant cupolas, sills, dykes and necks. Shield volcanoes were formed characteristically by the alkalic olivine diabase and trachyte-andesite formations. The lava flows of shield volcanoes were usually aligned zonally, parallel to the axes of the structures (e.g. in the Burgas synclinorium).

2. Volcanic activity began in the Cenomanian and Turonian to attain its maximum in the Senonian and persisted till the Early Eocene. It is thus typical of the Meso-alpine stage. The Upper Cretaceous is characterized by andesite effusions and extrusions of their pyroclastics, by rhyolites, dacites and their pyroclastics and ignimbrites. These volcanics are part of the volcano-sedimentary fillings of the graben-synclinoria and other early depressions. The intrusives are for the most part regarded as post-Cretaceous, connected with the Laramide paroxysm (S. KARAMATA 1974, M. BLEAHU 1974, D. GIUSCĂ—G. GIOFLICA—H. SAVU 1966, D. RĂDULESCU—M. SĂNDULESCU 1973, N. HERZ—H. SAVU 1974), but the radiometric age values yielded by the Vitoš and Plana plutons in the Srednogorie of Bulgaria (S. Boyadjiev 1974) and contact-metamorphism indicate the Late Cretaceous date. The analysis of the magmatism in the Srednogorie in particular (G. STANIŠEVA—VASILJEVA 1973, B. BOGDANOV 1973) has shown a close chronological connection of the volcanics with plutonic rocks; the latest are probably Palaeocene rhyolite and dacite dykes. The younger, post-Upper Cretaceous age of plutonic rocks in the South Carpathians, and chiefly in the Apuseni Mts. is connected with a different dynamic state of the crust and with the disparate manifestations of the Laramide diastrophism (this would be also testified by radiometric data — 60 m. y).

The banatites of the Apuseni Mts. also differ in a less varied composition and very small amount of alkalic rocks. The greater compositional variety of banatites in the South Carpathians and Srednogorie is explained by the more pronouncedly oceanic type of the subducted crust (D. RĂDULESCU—M. SĂNDULESCU 1973, BOCCALETTI et al. 1974). The different geotectonic character of the structures in which the banatites were formed and their different position in the structural plan are conspicuous. In the South Carpathians and Srednogorie Mts. they are confined to geosynclinal grabens, parallel to the principal features of the structural pattern and controlled by the mobility of the basement. This mobility is expressed not only by the presence of Meso-alpine folding but also by the presence of flysch complexes (Fig. 14). The up to hundreds of kilometres long graben-synclinoria rimmed by deep faults, more or less parallel to the major structures of the Palaeo-alpine plan of the South Carpathians and Srednogorie Mts., a smaller thickness of the crust than in the adjacent zones (about 29 km), the alignment of intrusive bodies, the distribution of earlier sediments in the marginal parts of grabens and the presence itself of extensive magmatic masses — all these facts suggest that the banatites and the Late Cretaceous thinning of the crust are genetically tied with the zone of increased crustal permeability (A. ANTONJEVIĆ—A. GRUBIĆ—M. DJORDJEVIĆ 1974).

In the Apuseni Mts. the magmatites occur only in the late geosynclinal, more or less posttectonic depressions, oriented obliquely to the basic features of the Palaeo-alpine structural plan. Fillings of these depressions probably derive from the early stage of the late-geosynclinal period. The small depth of Upper Cretaceous depressions in the Apuseni Mts. throws doubt on the potential origin of banatites by subsidence palingenesis (as assumed by H. STILLE 1953).

Geotectonically, the banatites of the Apuseni Mts. represent a connecting link between banatites of the Southern Carpathians, Srednogorie and the Carpathian Late Tertiary volcanics. The Eocene volcanics of the Hungarian Mid-Mountains are another member of this genetic series, which stands nearer to the latter.

Spatially the banatites form an arc (R. DIMITRESCU 1966) (Fig. 14b), which is more or less parallel to the course of the Palealpine structures in their principal area, in the Balkanides and South Carpathians. This leads to the opinion of genetic relation of banatites to subduction zones with ophiolites of the Vardar Zone and its eastern continuation (V. BOCCALETTI et al. 1973). The Rhodope and Serbo-Macedonian massifs, in such a conception, formed an island arc in the Upper Cretaceous, and the zone of Srednogorie — Timok a marine basin of back-arc type with strong volcanic activity.

The northern part of the banatite arc (in the northern part of the South Carpathians and in the Apuseni Mts.) — was running diagonally to the Palaeo-alpine structures. Here the banatites display also conspicuous geotectonic and petrochemical differences. Interpretation of the origin of magma from subducted ophiolite zones is not unambiguous, even from the aspects of global tectonics. Some authors

put the genesis of banatites into connection with the subduction zone of the East Carpathians (D. RADULESCU—M. SANDULESCU 1973), others with the subduction of the Metalliferous Zone (N. HERZ—H. SAVU 1974, M. BLEAHU 1974).

Banatites are present in these segments of the Alpides which do not show a distinct polarity-migration of folding. Actually, the renewal of the geosynclinal stage did not take place in juxtaposition but the two main stages of the geosyncline were superimposed, being separated by the molasse period, which preceded magmatic activity. This was evident mainly in the western Srednogorie after the termination of the Palaeo-alpine stage, i.e. after deposition of the Cenomanian-Turonian molasse and before the geosynclinal stage of the younger flysch geosyncline. No banatites are in the West Carpathians, Eastern Alps, Dinarides and Hellenides, where the geosyncline of the 'carbonate' Triassic to Middle Cretaceous stage and the flysch Cretaceous to Eocene (to Oligocene) geosyncline existed broadly side by side, being linked to each other in time and space.

## 6. Late geosynclinal volcanics (neovolcanics)

The late-geosynclinal volcanics are arranged chronologically into two stages: 1. Eocene — Oligocene and 2. Miocene, mainly Badenian — Pliocene. Both groups are linked up with depressions, thinning of crust and disintegration. Their spatial distribution is, however, usually different. Eocene — Oligocene volcanics are characteristic of the inner, intrageosynclinal masses — Rhodope and Hungarian massifs. In time they represent the beginning of the late-geosynclinal stage in the geosynclinal system and rejuvenation and tectonic activation of masses. Appropriate for them is thus the designation early late-geosynclinal. Miocene-Pliocene neovolcanics are mainly restricted to intramontane depressions, in the inner zones of geosynclinal systems. In the time scale they correspond to the late-geosynclinal stage. The two types are linked with each other spatially in contact areas of intrageosynclinal masses with inner zones of the geosynclinal system.

Of particular interest are three geotectonically different areas of their distribution: a) the Hungarian Mid-Mountains with the neighbouring zones of the West and East Carpathians, b) the Rhodope Mts. and c) a narrow zone of activation in the Serbo-Macedonian massif and in the internal Dinarides (Figs. 2, 15a, b).

The classical area of young volcanics is the Hungarian Mid-Mountains with the adjacent zones of the West and East Carpathians and the Apuseni Mts. forming a uniform crustal block called Neotisia (J. SLÁVIK 1971) (Fig. 15), where volcanism was active from the Eocene to Pliocene. During this 50 million years long period, volcanism showed a definite relationship to the type of depressions. In the Neotisia area the depressions evolved in stages and the orogenic and volcanic manifestations migrated from its central part outwards into the peripheral Carpathian areas. As

a result, the individual parts of Neotisia differ in the chronological distribution of volcanic phases and partly also in the types of volcanic rocks.

In the Hungarian massif, young volcanism began as early as in the Eocene. Andesite and their tuffs in the Zala basin reaching a thickness of up to 500 m are for the most part buried under younger sedimentary rocks (K. BALOGH—L. KÖRÖSSY 1968). The Middle and Upper Eocene effusions, cumulo-domes and pyroclastics are known from the Budai Mts., from the northern part of the Velence Mts. and the Gerecze Mts. In the Late Eocene, andesite effusions took place in the Mátra Mts. Diorite and diorite porphyry accompanied by copper deposits near Reczk-Lahóca along the Darnó line of NNW-SSE trend are connected with the stratovolcanic structure. Radiometric age (37 m. y.) and finds of nummulites (*Nummulites fabiani*) in sedimentary interlayers within the andesite complex indicate a Late Eocene age (C. BAKSA—J. CSILLAG—J. FÖLDÖSSY 1974, T. ZELENKA 1974). Eocene effusive rocks are also found in the Mecsek Mts. near Komló.

In the areas adjoining the Hungarian massif, the Eocene volcanism left only rare traces, such as sporadic interlayers of dacite tuffs in the Upper Lutetian of the Central-Carpathian Palaeogene round Banská Štiavnica in the West Carpathians (L. ROZLOŽNÍK—O. SAMUEL—S. JACKO 1974). More frequent layers of dacite and rhyodacite tuffites in various stages of bentonitization are known from Podhale north of the High Tatra Mts. (P. RONIEWICZ—J. WESTWALEWICZ—E. MOGILSKA 1974).

In the Oligocene, volcanic activity was confined to the South Slovakian-North Hungarian basin. Layers of andesite and acid pyroclastics occur mainly amidst the Egerian sediments. Andesites, their tuffs and rhyodacites are also known from the Pilis and Börzsöny Mts. (G. NAGY 1976).

A new volcanic stage (I. KUBOVICS 1974) began in the Early Miocene (Eggenburgian-Ottngian) in the northern parts of the Hungarian massif and the adjacent sectors of the Carpathians. Rhyolites and rhyodacites by far prevail and are of great thickness in the Visegrad and Börzsöny Mts. They also occur in the Mátra Mts. and the Transdanubian region. A series of minor necks, dykes, small domes and cumulo-domes are traceable from Košice to the Cibles area in Rumania. Rhyolite and rhyodacite pyroclastics extend in the neighbourhood of the Klippen Belt from Prešov (eastern Slovakia) to the Transylvanian depression. Pyroclastic deposits cover vast areas.

In the Badenian, thick effusions of andesites (chiefly pyroxene bearing) are at first substituted by dacite tuffs and dacite domes (G. VARGA et al. 1975) in northern Hungary, the adjacent part of the West Carpathian and the Slovenské Stredohorie Mts. Andesite stratovolcanoes and submarine lava flows together with rhyolite effusions build up the predominant part of the Pilis, Börzsöny, Cserhát, Mátra and Bükk mountain ranges. In the central part of the Hungarian massif, especially between Kopovár and Nyiregyház, huge masses of ignimbrite up to 1500 m thick

were emplaced during the Badenian and until the Sarmatian. With the exception of the north-eastern area, the Hungarian block was largely without volcanic activity in the Late Sarmatian and Pliocene.

The alignment of the North Hungarian young volcanic mountain chains along the inner (south-eastern) margin of the Inner Carpathians suggests that the genesis of the bulk of volcanic masses, chiefly of Badenian andesites was associated with the rifts of ENE-WSW trend (G. VARGA et al. 1975).

There is no direct evidence of such a rift. Noteworthy is, however, spatial distribution of Miocene neovolcanics of northern Hungary which overlie the Oligocene basin, obviously with a thinned crust but without more distinct older faults. More important manifestations of volcanism above this thinned crust appeared as late as the Miocene and are related to transversal faults. The effusive centres of lower rhyolite ignimbrites were narrow grabens mostly of NW-SE direction (obviously genetically linked to Lower Miocene Carpathian transversal downwarps — Vienna basin, Jablonica, Dačov lom zones). At the time of deposition of the middle rhyolite tuff (Karpatian) the NE-SW directions of fault lines were already evident (oral communication by G. Hámor).

In the Badenian, volcanic activity in northern Hungary is linked genetically with the ancient NE-SW fault system, which controlled the structural pattern of the Hungarian massif and the Inner West Carpathians. Extensive volcanism of terrestrial type is restricted to horsts bounded by faults of several trends (NW-SE, W-E or W-NW, E-SE, NE-SW and partly also NNW-SSE). In the west, mainly in the Börzsöny Mts., it is the NNE-SSW to N-S Revúca fault system (G. NAGY 1976) affected the transverse downwarp of the Slovenské stredohorie, the structure of the Cserhát and Mátra Mts. (G. VARGA et al. 1975, G. PANTÓ 1970) and the easternmost Tokaj Mts. The NW-SE trending faults had subordinate influence in the more westerly volcanic mountain ranges. The impact of this faulting on the volcanic activity in the Tokaj Mts. was more important.

In the West Carpathians (Fig. 3), neovolcanics are cumulated at the southern margin adjoining the Hungarian massif and particularly at the Central Slovakian transverse downwarp. There they make up the mountain groups Štiavnické hory, Poľana, Javorje, Pohronský Inovec, Vtáčnik and Kremnické hory, included under the name the Slovenské Stredohorie Mts. Volcanism of pyroxene-andesite type began in Badenian time in the Štiavnické pohorie, Javorje and Poľana and to a lesser extent in the Pohronský Inovec and Kremnické pohorie. It persisted until the Pliocene. The time sequences of the individual associations of volcanic rocks are as follows (V. KONEČNÝ—J. SLÁVIK 1974):

— products of andesite volcanism — Lower Badenian (southern periphery-marine environment)

— products of andesite volcanism — Middle Badenian-Sarmatian (major part of stratovolcanic complexes, including intrusive complexes and differentiated rocks)

- products of rhyolite volcanism — Lower Sarmatian (Poľana)
- products of andesite volcanism — Upper Sarmatian-Pannonian (upper parts of the complexes of the Vtáčnik — Kremnické hory Mts.)
- products of rhyolite volcanism — Upper Sarmatian-Pannonian (Štiavnické vrchy, Kremnické pohorie, Vtáčnik, Inovec Mts.)
- basaltoid andesites — Pannonian.

The pyroxene andesite is the principal member of the volcanic suite, dacite and rhyolite are subordinate and basalt andesite is the youngest. The rhyolites are more abundant in the northern part of the Štiavnické pohorie and in the south of the Kremnické hory, being located along the N-S faults. In central Slovakia, the volcanic centres are shifted progressively to the north-west and linked with the NE-SW, N-S and NW-SE fault systems. The central Slovakian transverse downwarp with the inner depressions is genetically associated with the N-S and NW-SE systems of faults. The spectrum of volcanic forms and products is extremely wide: submarine domes, cumulo-domes, central volcanoes, extensive lava fields, big extrusions with coarse-clastic breccias, agglomerates, tuffs and tuffites, but also necks and subvolcanic sills, intrusions of andesite porphyrite and quartz diorite porphyry and hypabyssal intrusions of diorite and granodiorite.

Andesites and basalts form short dykes and occasional laccoliths also in the Outer West Carpathian zones: in the Biele Karpaty unit of the Flysch Belt of Moravia (near Bojkovice and Bánov) and in the Pienides of Poland. According to the radiometric age the andesite from Bojkovice (40 m. y.\*) shows that these volcanics differ from those of central and eastern Slovakia not only in age but also in genesis. They rather resemble volcanic bodies in the East Carpathian Flysch Belt, which appear there from the Upper Cretaceous (see p. 73). Dykes of andesites occur amidst the Paleogene flysch also at the northern margin of the Pieniny Mts, near the contact of the Klippen Belt and Magura nappe. The older of them are more or less parallel (striking W-E or WNW-ESE) with this tectonic line of first order; the younger ones are oriented obliquely to it (K. BIRKENMAJER—A. E. NAIR 1968).

During the Neogene the eastern part of the Hungarian massif was connected with the East Slovakian Plain and the Transcarpathian basin (J. SLÁVIK 1971). Compared with the western part, this area including the Tokaj Mts. and the Vihorlat range was affected by volcanism mainly at the end of the Badenian and in the Sarmatian to Pliocene. In eastern Slovakia the sequence of volcanic products is as follows (V. KONEČNÝ—J. SLÁVIK 1974):

- products of rhyolite volcanism — Badenian
- products of andesite volcanism — Upper Badenian
- products of rhyolite volcanism — Lower Sarmatian
- products of andesite volcanism — Sarmatian (including intrusive and differentiated types)

\* Oral communication.

- products of rhyolite volcanism — Upper Sarmatian-Lower Pannonian
- products of andesite volcanism — Pannonian
- products of basalt-andesite volcanism — Pannonian.

The principal petrological types are andesites, predominantly pyroxene-bearing, and rhyolites and their tuffs, which form several overlying layers. The more acid character of young volcanics is also apparent from the type of subvolcanic bodies, the presence of microgranites, granodiorites and diorite porphyries (B. V. MERLIČ—S. M. SPITKOVSKAJA 1974). Frequent volcanic bodies are in the Guttin Mts. It is noteworthy that volcanic activity extended as far as the Klippen Belt. There is a definite close genetic relationship between volcanism on the one hand and the subsidence of depressions and their small crustal thickness (in places 25 km), on the other. In the whole area, rhyolite and later dacite eruptions prevailed in the first Miocene and Badenian phases. Only in the Sarmatian and Pannonian predominated andesites, which constituted a continuous volcanic Vihorlat — Guttin chain of NW-SE trend, consisting of stratovolcanoes, lava domes, volcano-sedimentary formations with ignimbrites and cupolas. Dykes and subvolcanic bodies are also numerous and cupolas abound in the Bela hora — Mukačevo area. Volcanic activity was governed chiefly by NW-SE trending faults, and in the Prešov—Tokaj Mountains by faults of N-S direction. The major fault system in the Guttin Mts. strikes NE-SW but the N-S faults are also frequent.

In the Apuseni Mts. (Fig. 4) young Tertiary volcanism was active between the Early Badenian and Pliocene. In the Early Badenian, rhyolite and rhyodacite volcanism, partly of ignimbrite type, was dominant and andesite volcanism in the Late Badenian (D. RĂDULESCU—M. BORCOS 1967, M. BORCOS 1974). More basic, especially basaltoid andesites and their pyroclastics, as well as dacites are subordinate. Neovolcanics are distributed in the depressions of the southern parts of the mountain range, particularly in the WNW-ESE to NW-SE, W-E trending fault zone, which cuts across the southern part of the Northern Apuseni Mts. and the Metalliferous zone. In some areas the neovolcanics follow the NE-SW directions (old reactivated faults).

The Călimani-Hărgitha (Fig. 11a) mountain range following the contact of the Transylvanian basin with the interior zone of the East Carpathians is the longest continuous andesite chain (140 km) and the youngest volcanic mountains of the Carpathians. The Badenian volcanism was of minor extent and is represented only by tuffs, predominantly andesitic. The substantial part of the mountain range is built up of volcanics of the Pannonian age among which andesites prevail over basalt-andesites and dacites (M. BORCOS 1974). Some diorite and quartz diorite bodies are known. The craters of stratovolcanoes are well preserved. Compound volcanic structures are aligned in the NW-SE or NNW-SSE, and subordinately E-W directions.

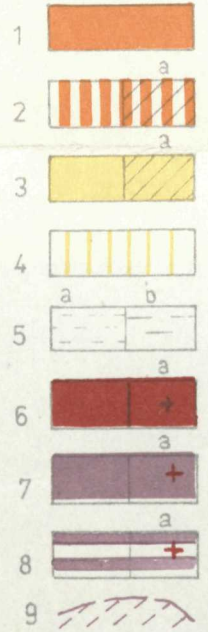
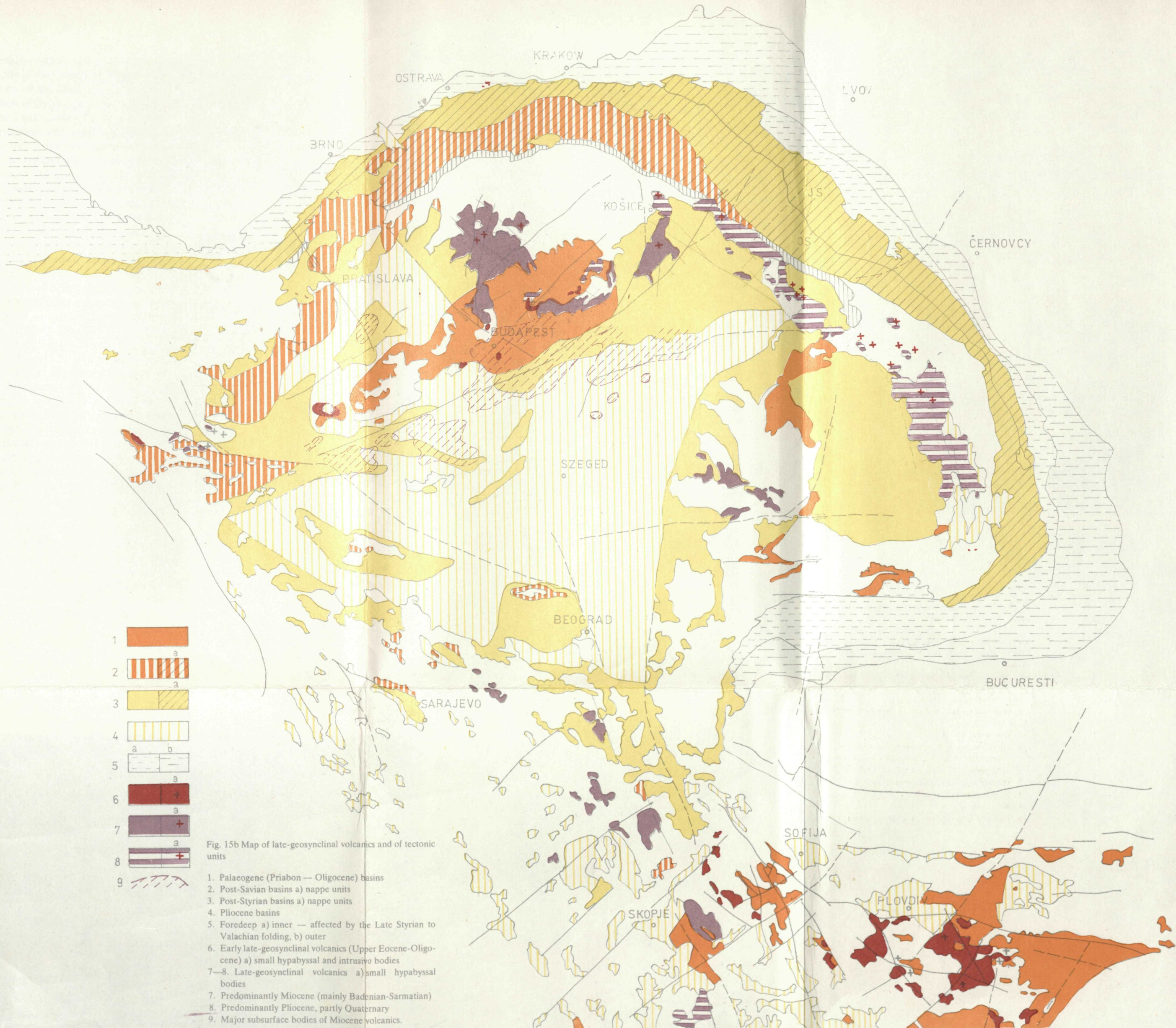


Fig. 15b Map of late-geosynclinal volcanics and of tectonic units

1. Palaeogene (Priabon — Oligocene) basins
2. Post-Savian basins a) nappe units
3. Post-Styrian basins a) nappe units
4. Pliocene basins
5. Foredeep a) inner — affected by the Late Styrian to Valachian folding, b) outer
6. Early late-geosynclinal volcanics (Upper Eocene-Oligocene) a) small hypabyssal and intrusive bodies
- 7-8. Late-geosynclinal volcanics a) small hypabyssal bodies
7. Predominantly Miocene (mainly Badenian-Sarmatian)
8. Predominantly Pliocene, partly Quaternary
9. Major subsurface bodies of Miocene volcanics.



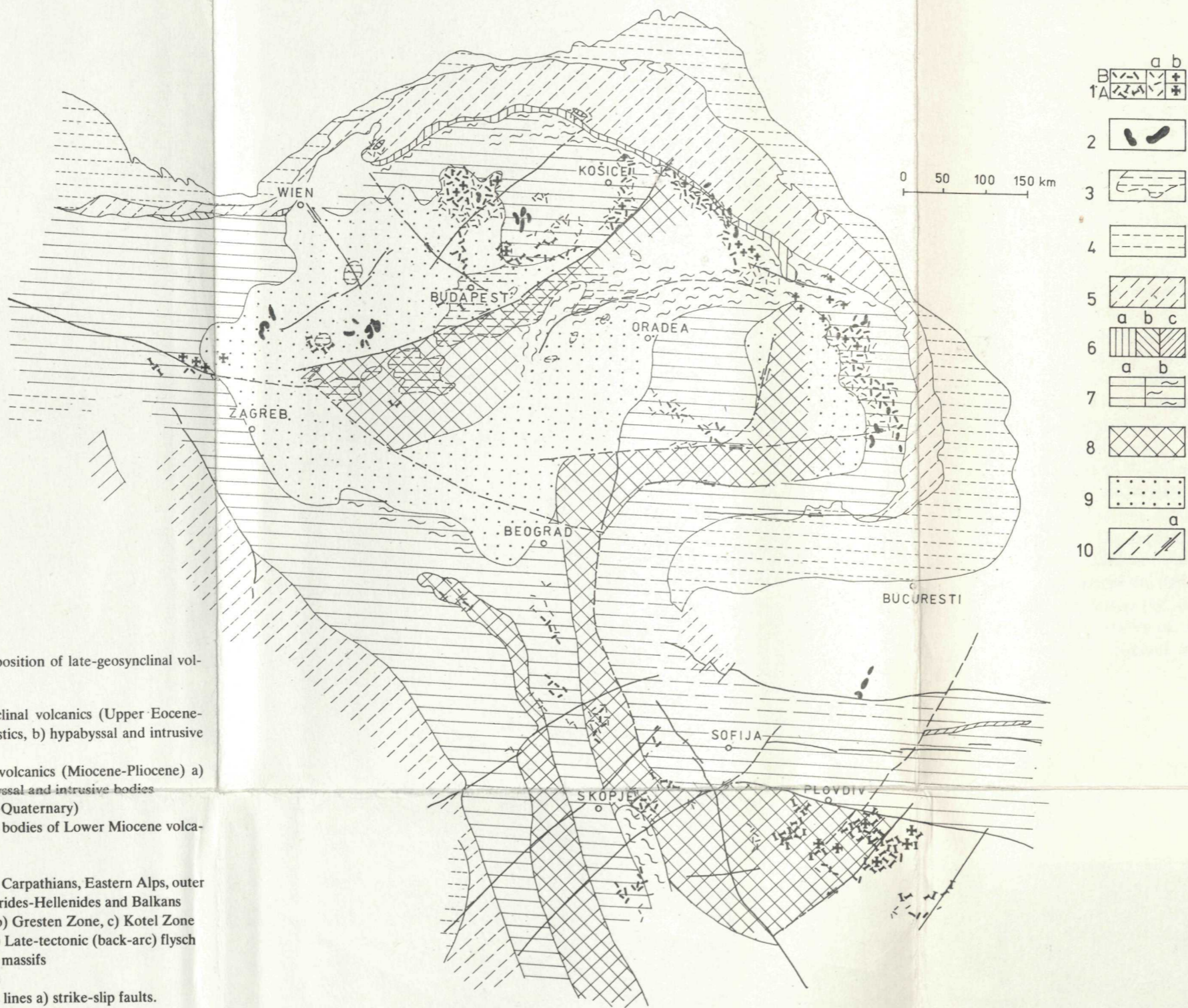


Fig. 15a Geotectonic position of late-geosynclinal volcanics

1A Early late-geosynclinal volcanics (Upper Eocene-Oligocene) a) pyroclastics, b) hypabyssal and intrusive bodies

1B Late-geosynclinal volcanics (Miocene-Pliocene) a) pyroclastics, b) hypabyssal and intrusive bodies

2. Basalts (Pliocene-Quaternary)

3. Major subsurface bodies of Lower Miocene volcanics;

4. Foredeep

5. Flysch Belt of the Carpathians, Eastern Alps, outer zones of the Dinarides-Hellenides and Balkans

6. a) Klippen Belt, b) Gresten Zone, c) Kotel Zone

7. a) Inner zones, b) Late-tectonic (back-arc) flysch

8. Intergeosynclinal massifs

9. Inner depressions

10. Principal tectonic lines a) strike-slip faults.

The neovolcanics occupy particularly extensive areas in the Rhodope Mts. (Fig. 6), especially in the Eastern Rhodope. Over the area of 4000 km<sup>2</sup>, they make up the upper part of the filling of the East Rhodopean Palaeogene basin, whose lower part consists of flysch complexes; submarine and terrestrial effusions of lavas and tuffs alternate with tuffites (R. IVANOV 1965). Volcanic activity was centred to three graben-synclines and exhibited specific features in each of them. Three volcanogenic sectors — the Borovište volcanogenic area, the Momčilgrad volcanogenic area and the Ardino volcanogenic zone — were formed. Complex groups of volcanic occurrences, volcanic centres, dykes and small plutons are aligned along the major fault line, mainly along the Marica fault (90—120°). The Tvardica or Arven fault system (15—35°) played a minor role in the Madžarovo and other areas (I. BOJANOV—D. KOŽUCHAROV 1968). Abundant dykes of andesites and subvolcanic bodies of andesites and diorite porphyrites, 2—3 km long, are mostly of W-E direction. Comagmatic plutonites are of the character of small intrusions (T. T. VELČEV et al. 1976). The oldest slight volcanic activity represented by andesite tuff interlayers is of Middle Priabonian age. More intensive andesite, trachyandesite and trachyte volcanism is of Late Priabonian and Early Oligocene date. The volcanic cycle ends with the acid rhyolite volcanism in the Early Oligocene. Late Priabonian and Early Oligocene are subvolcanic and hypabyssal microgranitoids (B. MAVRUDČIEV 1965). The Kamaran intrusion and diorite porphyrites of the Lozen intrusion are best known. In the Oligocene andesite effusions receded to trachyandesites and rhyolites forming dykes in the gabbro-monzonite-granite formation.

The minor Bracit, Dospat, Persen, Smoljan and Vitin depressions in the western part of the Rhodope massif are located on the faults of the Krajištide NNW-SSE trending system. The volcanic rocks constitute only a small part of the depression filling. Rhyolites (nevadites) with a small portion of pyroclastics dating from the latest Eocene and Early Miocene are predominant. The effusive bodies are connected with the subvolcanic bodies through dykes and sills. Effusions occurred in two phases: the first was characterized by extensive rhyolite effusions and the second by the formation of calderas and additional effusions.

At the end of the Oligocene and the beginning of the Miocene, intrusions of hypabyssal and subvolcanic magmatites occurred in the Rhodope Mts., some of them being situated at deeper levels of volcanic bodies, at the crossings of faults, often in the crystalline basement. Of the norite, gabbro, monzonite, syenite and granite intrusions the best known are the Madžarovo, Zvezdel, Galenica, Pesnopoiska and Malko Gruševo intrusions in the East Rhodopean block and the Zorniška intrusion in the West Rhodope block (B. MAVRUDČIEV 1965).

The Tertiary Rhodope volcanics have relatively little potassium; they are transitional between alkali-calcic and potassium-alkalic types. Latite-andesites, rhyolites and trachytes prevail over normal andesites, basalts and dacites (R. IVANOV 1960,

1965). The composition changes with the geological position of the volcanics; in the upheaved central part of the Rhodope massif occur only rhyolites, whereas a more varied association is present along its periphery.

The genetic association of neovolcanics with the tectonic activation of the Rhodope massif is demonstrated not only by their emplacement along the faults that divide the massif into uplifted and sunken blocks but also by radiometric dating. The ages obtained from the crystalline rocks and Hercynian granitoids of the Pirin horst and the central and eastern blocks of the Rhodope massif — 30—50 m. y. (S. BOYADJIEV—P. LILOV 1971) correspond to the interval Late Eocene — Middle Oligocene. In Greece, where the neovolcanics are later, radiometric ages of 13—27 m. y. (Middle Miocene — Oligocene) indicate the regeneration of the crystalline basement.

At the western margin of the Serbo-Macedonian massif, in the centre and in the zone of the North Dinaride horsts in Jugoslavia as far as the Pohorje, the Savin Alps and Karawanken, extends a belt of scattered major and minor volcanic centres: (Figs. 9, 5) Rožden, Mariovo, Pehčevo-Delčevo, Kratovo-Zletovo, Surdulica, Lece, Novo Brdo, Trebča, Kopaonik, Rogozna, Golija, Kotlenik, Rudnik, Zajaca and Kolnik. They are represented by granodiorite bodies and extrusions of andesite, dacite and rhyolite and their pyroclastics. Quartz latites of Upper Miocene to Quaternary age are often components of the filling of limnic basins. The whole zone was subject to Neo-alpine activation accompanied by differential uplifts and subsidences of crustal blocks amounting up to 2000 m. The presence of abundant mineral springs also indicates a strong tectonic unrest.

Numerous rhyolite, dacite and andesite dykes of Late Cretaceous to Late Oligocene age are also known from the Austro-alpine and South Alpine Crystalline in the south-western Tirol. At the southern border of the Alps, the Maastrichtian-Palaeocene 'scaglia rossa' complex near Trento is overlain by basalt pillow lavas and breccias and hyaloclasts, partly marine and partly subaerial, which extend up to the upper part of the Middle Eocene. Basaltic tuffs are known from the flysch of Belluno with radiometric age of 49—40 m. y. In the South Alps there are numerous occurrences of rhyolite dacite and andesite radiometrically dated at 33—30 m. y., mainly east of Pohorje. The Slovenian-Croatian volcanics of the Karawanken is also of Oligocene age. The intrusive bodies of gabbro, diorite, syenite, granodiorite and tonalite along the Periadriatic lineament are of the same age (45—30 m. y., J. DIETRICH 1976). Their genesis is explained as being related to the subduction of the northern zones of the Alps (G. O. GATTO 1976), mainly the Penninicum.

The mentioned bodies of diorites and diorite porphyries in the Reczk-Lahoca area, resemble tonalites and accompanying volcanics distributed along the Periadriatic lineament not only in their age (around 37 mill. y.) but also in geotectonic position. They are genetically linked with the Darnó fault line, which actually is an offspring of the Periadriatic lineament. Several bodies of granodiorites and granites,

accompanied by andesites and dacites of Upper Eocene age, have been found also in the Zala area in southwestern Hungary, in the zone of the supposed continuation of the Darnó fault line towards the Periadriatic fault zone. Thus in northern Hungary, the fault system should be regarded rather as a prolongation of the Insurbic, Periadriatic zone than of the Banatite zone. . The Rožňava fault zone is also one of the branches of this system and analogous magmatite bodies may be presumed to occur along it and its accompanying faults.

## Discussion

1. The neovolcanics are limited to the system of late geosynclinal post-tectonic depressions, forming part of their filling (Fig. 15a, b). The complex tectono-magmatic structures (volcano-tectonic grabens and horsts, calderas) are associated with faults which controlled the formation of depressions. Young volcanics are thus structurally a component of the Neo-alpine structural pattern, without a direct relationship to the Palaeo- or Meso-Alpine structures. From the above it follows that the late orogenic volcanism differs essentially from the Cretaceous-Palaeocene inter-orogenic volcanism in both the position and structural forms, although the volcanogenic rock associations are analogous; they are specific for the late geosynclinal evolution stage of the Alpides of south-eastern Europe.

The extent of neovolcanic occurrences is not uniform in all segments of the Alpides. They are very scarce in the Alps, virtually absent in the South Carpathians and rather infrequent in the Dinarides, except for the areas adjoining the Hungarian and Serbo-Macedonian massifs. In contrast, young volcanics cover extensive areas in the Hungarian Mid-Mountains, particularly in its northern part and in the adjacent interior zones of the West and East Carpathians. Another such centre is the Transylvanian basin, accompanied by young volcanics at the eastern margin of the Apuseni Mts. and chiefly in the innermost part of the Rumanian East Carpathians. The Rhodope massif, particularly its eastern, central and southern parts and the sectors neighbouring on the Serbo-Macedonian massif were likewise places of extensive late-orogenic volcanism. The neovolcanics are densely distributed in the zones stabilized by pre-Alpine folding (central massifs) and extended into the adjacent areas, which were consolidated by Palaeo-alpine folding (Fig. 12a). After Meso-alpine folding these sectors formed antipolar areas to the zones folded by later Neoalpine foldings. They were the sites of intensive downfaulting, crustal thinning (they show the smallest thickness of the crust) and tectonic regeneration. A classical example is the Hungarian Massif, which underwent activation since the Meso-alpine folding. Together with the neighbouring interior orogenic zones it constituted a specific crustal block — the Neotisia (J. SLÁVIK 1971) with active volcanism during the Eocene — Pliocene interval.

Another area of intensive Neo-alpine activation is the Rhodope massif with an abundance of post-tectonic depressions and neovolcanics. The depression filling

begins with flyschoid to flysch facies, substituted gradually by molasses. It was, however, not a new geosynclinal cycle in the Rhodope Mts. (B. MAVRUDČIEV 1965) but a late geosynclinal stage of the Balkanides extended to the Rhodope massif (the same as in the case of the Carpathians — Hungarian massif). Consequently, not even here can the intrusives be reasonably regarded as syntectonic, but as a component of the volcano-plutonic formation.

2. The analysis of the relationship between neovolcanics and the Inner Molasse sediments provides evidence that the periods of maximum volcanicity coincide with those of block subsidence, and the individual volcanic phases with the periods of the formation of basins (J. SLÁVIK 1971) (Fig. 15a, b). Palaeogeographical and volcanic conditions have revealed that the longitudinal trends, i.e. parallel trends to the margins of the folded geosyncline played an important role. In the west of Hungary and Slovakia it is primarily the NE-SW trend and in eastern Slovakia, north-eastern Hungary, Transcarpathian Ukraine and Rumania the NW-SE trend (J. SLÁVIK 1971, V. KONEČNÝ—J. SLÁVIK 1974) (Fig. 3). The Neogene Inner Molasses sank on faults, particularly longitudinal, even a few thousand metres. The orientation of a number of volcanic mountain ranges (Vihorlat — Guttin and Hărghita— Călimani) is connected with these fault systems. No less important are the transverse faults, for example, in Hungary and in the Rhodope massif, where they controlled the formation of the Neo-alpine basins. In Hungary, the principal volcanic centres in the Börzsöny Mts. follow the NE-SW and NNE-SSW to N-S trends, in the Visegrádi Mts. NW, in the Cerhat-Mátra NW to W-E and in the Tokaj area N-S and NW-SE strikes. The maximum volcanic activity in the Neotisia area falls in the Badenian-Sarmatian, in the period of the most intensive disintegration of the interior zones of the West Carpathians, the Apuseni Mts. and of the Hungarian massif into a system of minor basins, in the formation of which the transverse faults were the primary controlling factors.

Unusually interesting is the relationship between tectonics and volcanism in central Slovakia, where the origin of interior depressions and the distribution of volcanic centres were governed not only by faults of NE (i. e. West Carpathian) trend but also by NW-striking faults and particularly those trending N-S (Kremnické hory, Vtáčnik Mts.). It was essentially the N-S fault system that gave rise to the transverse downwarp, in which the Central Slovakian neovolcanic area developed. Also the N-S trending Hornád fault zone influenced the distribution of volcanic occurrences in the Prešov—Tokaj region. Numerous volcanic bodies are situated at the crossing of fault lines differing in trend, genesis, age and orientation (the Štiavnické hory, Börzsöny Mts.) (Fig. 15a, b).

3. The composition of intermediate magmatites is varied. In Neotisia they are represented by a petrochemically complete differentiation series from rhyolites to basaltoid andesites and basalts. It is a pure alkali-calcic province. The alkalic type of volcanics is represented by alkali basalts produced in a separate later volcanic phase.

This feature suggests a geotectonically homogeneous development of the orogenic region within an interval of about 40 m. y. The spectrum of rock bodies is varied, including all types from intrusions through subvolcanic and hypabyssal bodies and a broad scale of effusive and explosive rocks, from near-crater facies to redeposited rocks of distant volcano-sedimentary facies. The central parts of volcanic edifices in the uplifted areas indicate the presence of intrusions that are linked up with volcanism both in structure and time, but their proportion is smaller than in the banatite association. In the Carpathians this concerns the granodiorite-porphyrites, diorites, granodiorites and granites. In the Rhodope Mts., however, the volcanics represent a transitional type between alkali-calcic to alkalic rocks with a high K-content, the andesite-trachyandesite-rhyolite formation is characteristic of this region. The accompanying plutonites are denoted as the gabbro-monzonite association (E. DIMITROVA et al. 1975).

4. The established presence of major hypabyssal bodies of Palaeogene age mainly in the Mátra Mts. (Figs. 10, 15) (the Reczk body) has promoted the tendency to divide the volcanic activity in the Hungarian Mid-Mountains into the Palaeogene and the Neogene phases (I. KUBOVICZ 1974, T. ZELENKA 1974) by analogy with the Apuseni, where the Upper Cretaceous-Palaeocene early subsequent magmatic phase and the Neogene late subsequent magmatic phase have been differentiated (D. GIUCĂ et al. 1966, M. BLEAHU 1974). The existence of banatite bodies in the West Carpathians is being considered, especially in the Banská Štiavnica area in the Slovenské stredohorie (L. ROZLOŽNÍK 1976). Let us analyse the soundness of these opinions for several Alpine segments:

The Rhodope Mts. (Fig. 6) provide a classical example that shows the independence of Neo-alpine magmatism (beginning with the Late Eocene) of the Late Cretaceous — Meso-alpine (banatite) magmatism and, on the other hand, the connection between the former (Late Eocene — Oligocene) with the Neogene magmatism. On page 73 we have pointed out that the banatites and neovolcanics are associated with sharply different geotectonic structures: the banatites occur in taphrogeosynclines whose flysch filling was affected by Illyrian-Pyrenean and partly Laramide foldings, whereas the neovolcanics are located in transverse late tectonic depressions filled with molasses. In the Rhodope Mts. the two rock associations also differ in spatial distribution: the banatites occur in the Srednogorie zone and the neovolcanics in the Rhodope massif, in the south. However, the time continuity of magmatism in the Balkan cannot be overlooked even if the type and localization change. The post-Pyrenean tectonic and magmatic activation of the Rhodope massif was linked with the Meso-alpine activity in the Srednogorie.

The geotectonic differences between the banatite association and the neovolcanics are not so great in the Apuseni Mts. (p. 75). The banatites are there also developed on transverse structures, deep faults and early depressions, which are filled predominantly with slightly folded Cretaceous and Palaeocene beds of Gosau type.

Essentially analogous depressions, although of small extension, are in the southern zones of the West Carpathians; their Upper Cretaceous sedimentary filling can be denoted as early molasses. The post-Palaeo-alpine model of the Northern Apuseni Mts. obviously resembled more the model of the West Carpathians than that of the Balkan, but the West Carpathians differ from the Northern Apuseni in the type of their hinterland.

The West Carpathians are adjoined in the south by the Hungarian massif and the Apuseni Mts. by their southern part — the Metalliferous zone with a different Palaeo-alpine development and the more mobile crust in the Meso-Alpine stage. The Apuseni became a constituent part of the large Neotisia region as late as in the Neogene; the neovolcanics have many features in common with those of the Inner West Carpathians and in the adjacent part of the Hungarian massif.

The Late Eocene (and in part Oligocene) magmatism developed in the Hungarian massif and in the southernmost part of the West Carpathians, has so far not been proved to exist at a larger extent in the Apuseni Mts. and, in turn, the Upper Cretaceous-Palaeocene magmatism so widely distributed in the Apuseni has not been evidenced either in the Hungarian massif or the West Carpathians, although minor Upper Cretaceous depressions do exist in both areas. There is a question whether the differences in the time intervals of the Upper Cretaceous and Palaeogene magmatism are not caused by the diversity of the hinterland of the Apuseni Mts. and West Carpathians and whether the difference between these two Alpine segments is not due to the origin of Neo-alpine activation. The Hungarian massif as the hinterland of the Inner Carpathians was 'activated' after weak Palaeo-alpine folding at first slowly and only during the Illyrian-Pyreanean tectonic phase the activation increased and extended gradually to the north. The southern part of the Apuseni Mts. — the Metalliferous zone — affected by intensive Palaeo-alpine folding was the source of relatively strong tectonic activity, which called forth intensive Laramide and Illyrian diastrophism also in the adjacent northern part of the Northern Apuseni. The Neo-alpine activation of the Northern Apuseni conditioned by the formation of depressions and Neogene volcanic activity, was linked with the preceding intense diastrophism analogous to that of the Rhodope Mts. but with another vergency. The source of Neo-alpine activation of the Rhodope and Apuseni Mts. is obviously different from that of the Hungarian massif and the Inner West Carpathians. The Recsk-Mátra area (equally as the area of Pohorje with Paleogene tonalites south of the Periadriatic lineament in the Southern Alps), represents in the historical profile a connecting link between early late-geosynclinal Priabonian-Oligocene volcanics of Balkan type and late geosynclinal Carpathian neovolcanics. Genetic relationship of the Recsk area in the Mátra Mts. with the Balkan volcanics is also stressed by the equal type of porphyry copper ore formation.

5. A close genetic association of the neovolcanics with the late geosynclinal interior depressions is evident from the comparison of the age of the basins and of

young volcanics (Fig. 15). In the Rhodope Mts. the Priabon-Oligocene basins predominate over those of Miocene age and the same is true of volcanic bodies. At the same time, the greater extent of Miocene basins in the southern parts of the Rhodope Mts. in Greece is reflected in the greater extent of the Miocene volcanism.

The Hungarian median mass is characterized by extensive Eocene depressions, chiefly in the south-western part. The great extent of post-Styrian (Badenian-Sarmatian) basins in this region and the adjacent Carpathian areas coincides with a great extent of neovolcanics.

In Neotisia, volcanic activity migrated from the centre outwards, and from west to east with time (mainly as the volumes of eruptive masses are concerned), analogously to the advance of orogeny (J. SLÁVIK 1968, 1971, V. KONEČNÝ—J. SLÁVIK 1974). Volcanic activity persisted from the Eocene to the Early Sarmatian in the Hungarian Mid-Mountains, from the Badenian to Sarmatian and Early Pannonian in the Slovenské stredohorie, from the Late Badenian to Pliocene in eastern Slovakia and the Transcarpathian Ukraine, and was more intensive only during the Pliocene in the Călimăni-Hărghita zone. It is possible to assume a causal connection between orogeny and volcanism.

The formation of the neovolcanics of the Carpathians (including the Hungarian median mass) was presumed to be associated with the Neo-alpine subduction in the outer Carpathian zones already by H. STILLE (1953). In the East Carpathians most of the neovolcanics in the Vihorlat-Guttin and Călimăni-Hărghita zones form continuous belts subparallel to the Neo-alpine structures of the Flysch Carpathians, i.e. to the presumed zones of subduction. It is not surprising that the primary genetic relationship between the Neo-alpine subduction and the genesis of volcanics is advocated by the supporters of the new global tectonics (e.g. D. RĂDULESCU—M. SĂNDULESCU 1973, M. BLEAHU et al. 1973, N. HERZ—H. SAVU 1974, J. SLÁVIK 1971, M. BOCCALETTI et al. 1974, E. SZÁDECKY—KARDOSS 1973), who believe the intermediate magmas to be derived from the subducted plates. The distribution of the volcanics in the Apuseni Mts. and the central Slovakian area can hardly be considered subparallel to the assumed zones of subduction. In the period of maximum volcanicity during the Badenian-Sarmatian, the volcanics were distributed in two divergent trends in the North Hungarian massif (with the W-E dominant trend) and in central Slovakia, where the volcanics are aligned in SW-NE and N-S directions. Noteworthy is also the lack of volcanism after the Savian subduction, which is the most characteristic for the West Carpathians.

In the last years, according to similarity in magmatic activity and distribution of heat flow values, crust thickness, thickness and type of sedimentary filling, a character of ensimatic and ensialic "back-arc" basin is ascribed to the Pannonian basin (M. BOCCALETTI et al. 1976; L. STEGENA et al. 1975). The main factors of basin development were spreading and active mantle diapirism, probably evoked by subduction of the lithosphere along the island arc.



In the Carpathians there is a striking non-uniform distribution of rocks: the dominantly acid rocks are in the central part of the basin and the intermediate to basic rocks occur along the margins of the irregular discontinuous volcanic arc. The majority of rhyolitic rocks have lower  $K_2O$  and higher  $SiO_2$  in relation to the extrapolated differentiation trends of intermediate rocks, which is obviously the result of anatexis of the sialic crust.

The youngest volcanics at the margins of the Carpathian arc show higher alkalinity and a lower content of incompatible elements (J. LEXA 1978).

The above findings indicate that the neovolcanics possess very close relations to the structural development of the basins and that the manifestations of volcanism are linked in space and time with block tectonics, in part with subsidence, and with the migration of tectonic mobility of the basins, which is also expressed by the changes in the type of basin filling. Although the neovolcanics are of great material variety, they do not display changes in chemical composition or closer relations to the structure of the presumed subduction zones, in the West Carpathians to the Klippen Belt. Moreover, in the Carpathian arc they are 5—20 m.y. younger than the presumed subduction (J. LEXA).

A direct relationship between the assumed subduction and young volcanism cannot be visualized either for the Rhodope Mts., since the volcanic areas often extend transversely to the contraction, mainly to the subduction zones. However, the relations between the contraction of the outer zones and the initiation of depressions and, consequently, the volcanic activity cannot be omitted. Alternation of compressive movements causing the nappe structure in the Externides in the Flysch Carpathians and relasing movements leading to the formation of interior and intermontane depressions is evident (M. MAHEL 1973), but the interrelations are obviously more complex. There is no such alternation of compression and dilatation in the Rhodope Mts. The solution of this problem can be facilitated by clearing up the formation of the depressions in the Hungarian median mass in terms of crustal thinning caused by subcrustal diapirism (P. HORVÁTH 1974, L. STEGENA—B. GECZY). The formation of the diapir is thought to have been stimulated by dying-out subcrustal horizontal movements, which until the Neogene controlled the evolution of the geosynclinal system and to have been closed with the turning point in the evolution; the vertical movements set in gradually, extended to the central part of the Carpathians and eventually dominated the evolution of the Carpathian system as a whole. It meant not only alternation of compression and dilatation but also of horizontal and vertical movements.

The accompanying volcanism developed only under suitable geotectonic conditions. There is a striking abundance of neovolcanics in the areas with a relatively thin crust as are, for example, the areas of northern Hungaria (with the crustal thickness of about 27—30 km), central Slovakia (ca. 32—34 km) and eastern Slovakia and Transcarpathia (20—24 km). The dependence of the neovolcanics on the crustal

thickness is particularly pronounced in the Rhodope Mts.; in the central block with abundant volcanics the crust is up to 30—40 km thick, whilst it is 40—50 km thick in the western block with a small amount of neovolcanics (B. BOGDANOV 1973). In the Vardar zone, strongly differentiated by young tectonics, and in the Serbo-Macedonian massif the crustal thickness varies around 30 km.

Major interior basins and extensive neovolcanic bodies did not originate in those Alpidic segments in the hinterland of which was not a stabilized faulted block that could disintegrate into areas of uplift and subsidence (e.g. the Eastern Alps).

Volcanism was active in the periods (mainly in the Miocene) and in the areas when and where deep relieving faults were active, causing crustal expansion. The periods of subsidence of the basins independent of the function of deep faults, as it was in Oligocene time in the North-Hungarian-South Slovakian and the Pannonian basins and in the Great Hungarian Plain with the Danube Lowland, were devoid of major volcanic events.

#### **7. Final volcanics — alkalic basalts**

The opinion on the specific geotectonic position of the alkali basalts is accepted almost universally (Figs. 2, 15). They are of Pliocene-Pleistocene age in all Alpidic segments, as in the areas where the andesite and rhyolite volcanism was terminated at the end of the Oligocene or in the Early Miocene (e.g. in the Rhodope Mts.), so in those where it persisted until the Pliocene (e.g. in the East Carpathians). In the latter areas the final basalt volcanism obviously was linked up with the late geosynclinal volcanism. Minor basalt occurrences are present in all Alpidic segments, even where the andesite-rhyolite volcanic phenomena were minimal or absent altogether as, for example, in the South Carpathians and Eastern Alps. In the Hungarian median mass basalt eruptions along the NW, NE and N-S trending faults at the Pliocene-Pleistocene boundary followed after an about 10 m. y. interval of relative volcanic quiescence. They are distributed over an extensive area, particularly north-west of Balaton and in the Salgótarján area. In the West Carpathians they are frequent in the Lučenec-Filákovo area, around Banská Štiavnica and Nova Baňa in Slovakia; in the East Carpathians they are known from the Perşani Mts. and in the South Carpathians from Poiana Ruscă.

Basalt dykes often occur in the Eastern Rhodope Mts. and less frequently in the western Rhodope block and in the Srednogorie. Subvolcanic bodies of lamprophyre-basalt, plagiobasalt and dolerite traverse the Rhodope massif, the Balkanides and the Moesian platform from north to south (Fig. 6).

In the Dinarides, the Boranje basalts and minor basalt bodies in Slovenia are known. In the Eastern Alps, where neovolcanics are extremely rare (Helvetian-Lower Tortonian andesites and dacites in Styria and traces in the Vienna Basin), basalts and their tuffs occur in the Upper Pannonian of the Lavantthal area.

The volcanics are invariably alkalic or subalkalic, mostly olivine basalts and their nepheline-bearing varieties and limburgite-basanites. They are often less than one million years old. Their alkalic character and position generally along the boundaries of the arcuately distributed Carpathian basins are explained as the result of the transition from island-arc volcanism to the inter-arc type of volcanism (M. BLEAHU et al. 1973). Minor bodies of little structural and tectonic importance originated everywhere only with termination of intense sinking (e.g. in the Alföld), in the period when besides sinking movements also uplifts of the individual blocks had begun. The finds of upper mantle spinel peridotites forming xenoliths (up to 25 cm in size) in alkali basalts of Slovakia (near Filákovo) may point to the derivation of magma from the upper mantle; the xenoliths represent its unmelted relicts (D. HOVORKA 1978).

### III. Some general conclusions

#### 1. Basic geotectonic characterization of magmatites

The described magmatite types — granitoids, late Hercynian late geosynclinal volcanics, pre-Alpine early geosynclinal volcanics, Alpine ophiolites, interorogenic Meso-alpine volcanics, Neo-alpine late geosynclinal volcanics and final volcanics differ from one another not only in association and partly also in the type of principal rocks but even in their structural position and share in the structure of the earth's crust. This follows from the differences in the geotectonic conditions of their origin and in the crustal type.

a) The granitoid rocks are usually associated genetically with deep-seated folding processes and the consequent accretion of the crust thickness. However, the maturity of the crust achieved in the individual orogenic cycles, as well as the granitization manifestations were diverse.

The pre-Hercynian granitoids generated in the Baikalian and earlier folding cycles are dominantly synkinematic and closely connected with migmatites, anatexites and intensive regional metamorphism.

The Hercynian granitoids are most numerous and most important in the evolution and structure of the Alpides (Fig. 2). They are distinguished by a great variety in tectonic position, a wide time interval of genesis and a relatively narrow range of material differentiation. Synkinematic granitoids linked up closely with their mantle rocks are in some zones the oldest and occasionally the deepest products of the Hercynian granitization. Major batholiths and phacoliths are most numerous; they are associated genetically not only with folding and thus with the general structural pattern but also with the morpho-tectonic diversity of the area. In many instances, the diapiric ascent up to the erosion level is their characteristic feature. The granodiorite-granites relatively little differentiated are most abundant. Higher differentiated granitoids and of more different associations occur in the Balkanides, i.e. in the part of the Alpidic geosynclinal belt that shows greater diversification, as seen by numerous linear geosynclinal troughs (E. BONČEV 1967). In addition to the granodiorite-granite association, also the gabbro-plagiogranite, gabbrodiorite-granodiorite and alkalic gabbro-syenite associations occur there (E. DIMITROVA et al.

1975). Aplites, pegmatites and minor bodies of leucocratic granites of Permian age are frequent everywhere.

The granites of Alpine age (Fig. 2) played a small part in the formation of the Alpine structural pattern. They are represented dominantly by minor bodies, strongly reworked by Palaeo-alpine to Meso-alpine folding and confined to the zones of the pre-Alpine Crystalline. Major bodies of synkinematic granitoids are chiefly in the more southerly parts of the Serbo-Macedonian massif, in the areas strongly affected by Cimmerian orogeny. The granitoid bodies forming part of volcano-plutonic Meso-alpine formations (banatite association) had a greater share in the structure of the Alpides. The amount of intrusive and hypabyssal bodies and the size of some of them were obviously influenced by the state of the crust, which was governed by recurrent crustal compression (Laramide and Illyrian or Pyrenean). Minor, particularly hypabyssal bodies of granite and granodiorite occurred still in association with neovolcanics. They originated as a result of the late geosynclinal stress relief in the crust and processes of dilatation.

b) The late Hercynian volcanics are genetically linked with the Neohercynian, mainly Permian grabens and depressions filled with molasses. They are a belated superficial manifestation of differentiation during extensive Hercynian granitization of the crust. They are more abundant in the grabens that separate zones of marked Hercynian granitization and at the interior margins of areas consolidated in Hercynian time. In these areas, the deep faults enabled subaqueous effusions of melaphyre, whose sills are accompanied by melaphyre-porphyrity and porphyrite dykes (e.g. the Melaphyre series of the Choč nappe of the West Carpathians and the Permian of the Codru nappe in the Apuseni). The molasse fillings of the Upper Palaeozoic depressions of the more northerly Alpidic zones contain only a small amount of volcanic rocks (quartz porphyry) (Fig. 8).

c) The pre-Alpine eugeosynclinal volcanics are characteristic of the substantial part of geosynclinal series predating the Hercynian granitization. They form a component of the aspidic, partly diastrophic sedimentary-volcanic complexes and belong dominantly to the spilite-diabase and spilite-keratophyre associations. Intrusives of gabbro-leucogranite and peridotite-pyroxenite associations accompany them. Non-uniform distribution of basic magmatites and the greater extent of acid types in some Upper Proterozoic (e.g. Arada Formation) and Lower Palaeozoic, mainly Cambrian-Ordovician and Ordovician units (e.g. the Grauwackenzone of the Eastern Alps, Gelnica Group in the West Carpathians, the Muncel Formation in the Apuseni Mts. and the Tulgheş Formation in the East Carpathians) indicate partial differentiation of the crust during the Baikalian orogeny. It should be stressed, however, that these acid volcanics, mainly quartz porphyries and quartz keratophyres, are components of subaqueous geosynclinal sequences related to the island arcs, i.e. to the areas where the oceanic crust passes into continental crust of greater thickness and which are distributed in the marginal parts of geosynclines.

d) Mesozoic ophiolites occur in the zones which from the presence of deep-sea facies are considered to be trough belts with a thin, occasionally disrupted sialic crust of continental to oceanic type. The ophiolites appear as early as the Triassic, particularly in the Ladinian, often together with shallow-water limestones. At first the intermediate-hybrid type with the predominance of spilites and keratophyres prevails, dating from the interval of crustal breaking and differentiation. As the troughs gradually expanded in the Ladinian and Carnian, being filled with deeper-water facies of carbonates, radiolarites and pelites, the amount of basites also increased and ultrabasites made their appearance. In most segments of the Alpides the Triassic magmatites are confined to the innermost zones of the Alpine geosynclinal system, i.e. to the zones which were less consolidated by the Hercynian granitization. It is the Meliata Formation in the Slovakian Karst of the West Carpathians, an analogous formation in the Bükk Mts., the Vascău Formation in the Apuseni Mts., the Sakar Formation in the Balkan and the Transylvanian nappes in the East Carpathians. In the Dinarides the volcanics of ophiolite type occur in the Triassic in the interior (e.g. Vardar and Ophiolite) zones and in some exterior zones as well (chiefly the Budva-Zukali zone). A lower intensity of Hercynian folding and granitization in the Dinarides, where marine facies predominate even in the Permian, is responsible for this feature.

The basic and ultrabasic volcanics forming part of the ophiolite association are more abundant in the Jurassic and Lower Cretaceous troughs distinguished by marlstone-radiolarite facies. They are mainly distributed in the central zones of the 'Internides' as (Fig. 12a) for example, the Krížna and Kysúca troughs in the West Carpathians, the Bukovina troughs in the East Carpathians, Penninicum troughs in the Alps and the Mureş zone in the Apuseni Mts. In the Dinarides and Hellenides ophiolites occur predominantly in the interior zones (the Vardar and Ophiolite zones in the former and the Mirdita zone in the latter). Ophiolites also frequently accompany thick pre-flysch facies of Tithonian-Early Cretaceous age (e.g. in the Ceahlău-Rachovo tectonic unit in the East Carpathians, the Severin-Krainia tectonic unit in the South Carpathians and the Strandža and Sakar units in the Balkan Mts.). In the Mures and Vardar zones the ophiolites occur in the Upper Jurassic deepsea carbonate-silicite facies and the flysch sequences as well.

e) The Meso-alpine intra-orogenic volcanics and associated plutonic rocks of the banatite association in the Srednogorie Mts. and the Timok zone of the South Carpathians are genetically linked up with the longitudinal graben-synclinoria (Fig. 14b). These are filled with 'pre-molasse' complexes of varied, dominantly marine facies, which are often accompanied by flysch. It is of interest that the occurrence of volcanics increases with the proceeding 'oceanization' of the sedimentary filling. The flysch complex of great thickness is often underlain by volcanics. The grabens having a thinner crust follow the interblock boundaries. The highly differentiated volcanic rocks (rhyolites — dacites, trachytes, andesites, alkali

basalts) form part of the volcano-sedimentary filling of grabens and graben-synclinoria. Abundant subvolcanic and plutonic bodies of equally wide composition range (picrites, gabbros, diorites, monzonites, gabbrosyenites, granosyenites, granites, diorite porphyries and granite-porphyries) mostly rim the synclinoria, demonstrating the increased dynamics of the crust. These volcanic-plutonic formations were at first alkali-calcic and later subalkalic to alkalic. In the Apuseni Mts. the banatites prevalently represent deeper-seated low differentiated bodies (granodiorites and granites with subordinate volcanics). The banatite association shows a different tectonic position, as it is not related to longitudinal graben-synclines but to a transverse deep-fault system. Dating from the beginning of the late geosynclinal stage, the bodies were not affected by major foldings.

f) The late-geosynclinal Tertiary young volcanics (neovolcanics) are linked with young both longitudinal and transverse basins superimposed on the interior zones, which were stabilized in Palaeo-alpine times, and with the hinterland of these zones — the median mass affected subsequently by crustal thinning (Fig. 15). The molasse filling of the basins contains an appreciable amount of pyroclastics. The neovolcanics, predominantly of intermediate composition (predominantly andesites with a large amount of rhyolites) participate in the sedimentary-volcanic formations and also make up volcano-tectonic structures (stratovolcanoes, calderas). Volcanism became an important factor in the history of the basins, particularly in their progressive isolation from one another, as it influenced the morphology of the landscape. The effect of hypabyssal bodies was slighter.

g) The final Pliocene-Pleistocene basalt volcanism of minor extent, mostly subalkaline to alkaline, is representative of the following evolution stage of the geosyncline, when crustal stretching and intense vertical movements played an important role (Fig. 15).

## **2. The evolution of the alpine geosynclinal system and the role of magmatites**

The type and distribution of magmatites are directly related to the evolution of the geosynclinal system and to the crustal type, as are all basic geological processes, i.e. sedimentation, folding, metamorphism and magmatism (Figs. 17—19).

Until the onset of the Hercynian granitization, folding and metamorphism in the Carboniferous period, the facies variety in the Alpides was rather small. The pre-Carboniferous formations are prevalently greywackes, sandstones, siltstones and claystones, partly of diastrophic (flysch) and partly of aspidic nature. This holds broadly for the whole Crystalline of the Alpidic zone (M. MAHEL 1974). Relative to other Alpidic segments, the Crystalline of the West Carpathians is noteworthy for a very small portion of carbonates. In addition to the metamorphic grade, it is the volcanic suite that causes the differentiation of the metamorphosed sedimentary

complexes of the Hercynian Crystalline within the Alpidic zone. Most of the pre-Carboniferous complexes contain abundant magmatites, mainly of the spilite-diabase and subordinately of spilite-keratophyre associations, which were for the most part converted to amphibolites. Gabbros, gabbrodiorites and comagmatic ultrabasites are fairly frequent; porphyroids (quartz porphyries and their tuffs) are found only in some sequences, usually of Ordovician but also of Devonian and Early Carboniferous age. The only major difference in the composition of the Lower Palaeozoic and pre-Palaeozoic complexes is that the orthogneisses and migmatites are more abundant and more intensely metamorphosed in the Proterozoic complexes. This implies that the Caledonian orogeny did not produce any relevant changes either in sedimentary or magmatic conditions. The changes occurred at deeper levels without pronounced morpho-structural manifestations. The more abundant quartz porphyries in the Upper Proterozoic and particularly in Ordovician geosynclinal complexes indicate the differentiation of the crust, formation of the island arcs and possibly of the zones with a greater crustal thickness (M. MAHEL 1974).

The pre-Carboniferous complexes in the Pelagonian and Rhodope massifs differ in the presence of voluminous carbonate masses, often accompanied by basic rocks. They are thought to be of Proterozoic age, because they show a higher grade of metamorphism than the adjacent Lower Palaeozoic diabase-phyllite formation. The distribution of these sequences side by side and the surprising lack of Lower Palaeozoic complexes of platform type in the massif raise the question whether at least part of the carbonate sequence is not a Lower Palaeozoic shallow-water type linked up with the massif.

The presence of quartz porphyries in the Devonian and Lower Carboniferous geosynclinal sequences (the Hron Formation in the West Carpathians, Paiuseni Formation in the Apuseni Mts., some complexes in the South Carpathians), a greater difference in their thicknesses and the occurrences of thicker limestone and dolomite complexes (Repedea, Tibău and Padeş Formations) indicate the effects of the crustal differentiation in the initial stage of the Hercynian orogeny. The sedimentation was not everywhere continuous in the Early Palaeozoic; the unconformities and changes in the palaeogeographical pattern are known from the Middle Cambrian (Serbian phase), at the end of Cambrian (Sardinian phase), at the end of the Ordovician (Taconian) and of the Silurian (Ardenian). In the Balkanides and chiefly in the South Carpathians the separation of the Caledonian cycle from the Hercynian cycle is expressed not only by basal conglomerates and weak unconformities but also by a lower-grade metamorphism (S. NĂSTĂSENAU 1975). However, no marked change in the structural or palaeotectonic pattern, neither in the type of magmatites nor in sedimentary formations are observed either after the Cadomian orogeny or after Caledonian diastrophic manifestations. The frequent radiometric values around 400–300 m. y. rather point to a 'thermal event' (H. J. ZWART 1976). The eugeosynclinal near-to-oceanic crustal type remained unchanged until the Hercy-



nian folding, which was the first to change the character of sedimentation. In Late Palaeozoic times the superimposed basins filled with marine and later with continental molasses appeared for the first time in the Alpine zones.

The formation of huge batholiths and phacoliths and their gradual diapiric rising induced a peculiar tectonic style: large-amplitude folds and major anticlinal domes developed, for example, in some 'Core mountains' of the West Carpathians and brachyanticlines in the Rhodope and Serbo-Macedonian massifs. In areas that were less affected by Hercynian granitization formed narrow constricted folds and thrust faults (e.g. in the Crystalline of the Čierna hora in the West Carpathians, Bystrica nappe group in the East Carpathians, H. G. KRÄUTNER 1972). As is seen, the Hercynian folding together with the formation of granitoid bodies was an important factor in the development of the structural pattern.

Major granitoid bodies showing a rising tendency were also morphostructural agents. The domes became source areas of sediments which filled the adjacent troughs rimming or intervening between the granitoid bodies. The rapid uplift of granitoid bodies in the Late Palaeozoic, their rising often up to the erosion level is conspicuous. It may be a consequence of diapiric squeezing out in result of compression. The cause of the rapid terminal rise may also be a steep geothermal fall of magma.

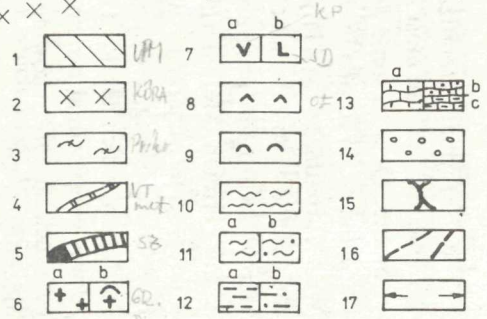
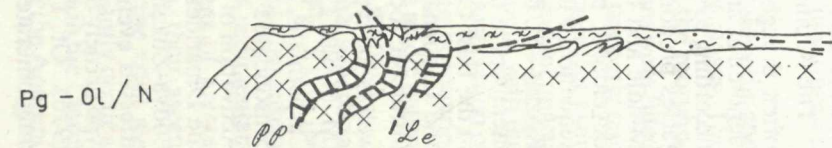
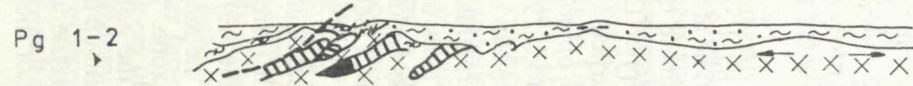
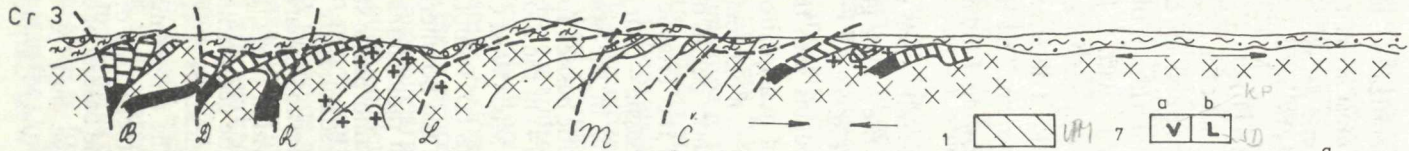
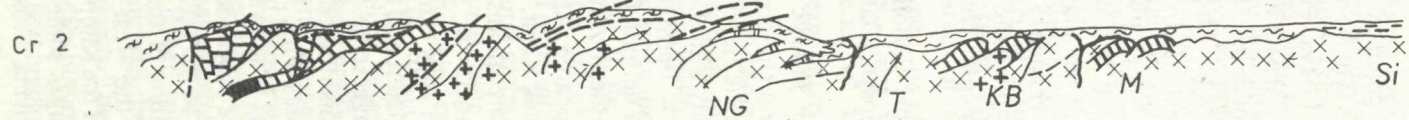
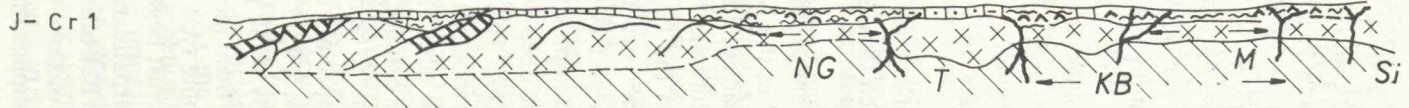
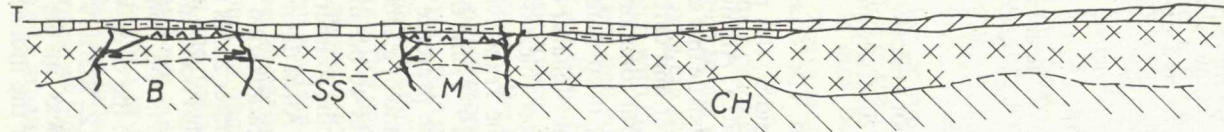
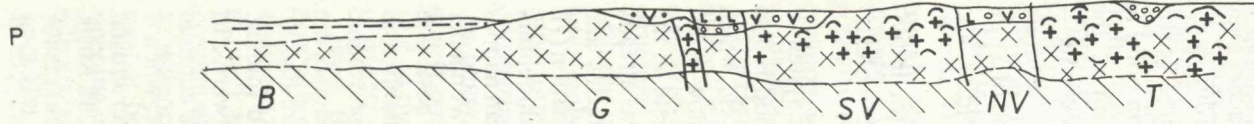
The Hercynian granitization was a long-lasting process. It was active since the end of the Devonian to the end of the Permian, culminating in the folding stages. The structure-forming effects were connected with the earlier stages and the morphostructural effects with the later phases, when the grabens and depressions were filled with the molasse and volcanic deposits. In the less granitized zones, the early and later molasse encloses Permian basites, in the West Carpathian Gemerides even small ultrabasic bodies. In the zones of more intensive granitization, abundant quartz porphyries occur in the later continental molasses, which are synchronous with the last phase of granitization in the adjacent rising areas. A close genetic association of quartz porphyries with granitoids is indicated by the presence of granite-porphyries (e.g. in the North Veporide graben in the West Carpathians) and by the acidic nature of quartz porphyries and Permian aplites, pegmatites and leucocratic granites.

From the above it follows that the Permian quartz porphyries are superficial products of magmatism connected with strongly subsiding grabens (Fig. 8). In the southern zones, melaphyres are found in the Permian upper molasses, but because of a small amount of their pyroclastics they did not markedly affect sedimentary conditions in the basins. As the melaphyre zones do not bear signs of greater subsidence and are not genetically linked with deeper-water facies, their ascent cannot be ascribed to the crustal thinning but to the break-up of the crust and opening of the conduits to a greater depth. The melaphyres predominantly occur as sills and alternate with quartz porphyries in some areas as, for example in the Čierna hora Mts. (West Carpathians), in the Balkanides or in the Codru unit of the Apuseni.

The Hercynian granitization is genetically associated with metamorphism of variable intensity. In most areas it is synkinematic and does not exceed the isograde of biotite. Only in some areas, in Lokva, Poiana Ruscă and Paring in the South Carpathians, in Rodna in the East Carpathians (H. G. KRÄUTNER 1972) and in the West Carpathians (to a larger extent in the Tatrídes and Veporídes) it reaches even the almandite and amphibolite isogrades. The intensity of metamorphism during the Hercynian orogeny progressively decreased, the last weak effects are observable still in the Permian. Metamorphism of recrystallization type and granitization point to a deep-seated character with a high thermal flow. The principal result of the Hercynian folding, granitization and metamorphism was the crustal thickening — sialization, which heralds a decisive change in the evolution of the area of the future Alpine geosynclinal system. This implied the beginning of a new cycle and of a new geosynclinal type showing fundamental changes in sedimentation, folding and metamorphism.

As the Hercynian granitization and consolidation was not uniform in the individual areas and zones, the Alpine geosynclinal belt showed diversity and complexity in its further history (Fig. 7).

The Alpine cycle begins with the Triassic transgression. The Triassic system is often thought to be of quasiplatform character in the Alpídes; this period is also denoted as an interregnum between the Hercynian cycle and the Alpine cycle, which according to some authors begins only with the Jurassic. It is true that shallow-water facies predominate in the Triassic — detrital in the Lower and carbonate in the Middle and Upper Triassic — but a definite facies-structural zonation was already developed, particularly in the southern zones and even deep-water facies and sequences with ophiolitic basites appeared. As examples, the Meliata Formation (H. KOZUR—R. MOCK 1974) in the Slovak Karst, the deep-water volcano-sedimentary members in the Bükk Mts. (West Carpathians) (Fig. 16), the Transylvanian nappes in the East Carpathians (D. RĂDULESCU—M. SĂNDULESCU 1973) and the Wengener and Cassiana Beds in the Southern Alps can be cited. The presence of basic and ultrabasic rocks amidst radiolarites chert and dark shales, and of glaucophanites, as well as the type of metamorphism suggest the crust to be near-to-oceanic or oceanic (M. MAHEL 1975). These zones are distributed especially in the southern parts of the Carpathians, Balkanides and Dinarides (Figs. 3, 4, 6, 9, 11). All of them were slightly consolidated in Hercynian times, as is shown by weak granitization and folding. Activation of the Alpine geosyncline proceeded from the inner zones outwards. In the Dinarides the magmatites near to the ophiolite type also occur in the external Budva-Zukali zone. The differentiation of the crust began in the Triassic, when it affected even the northern area of the West Carpathians; the deep-water crustal type namely built up the basement of the Klíppen Belt area (Z. ROTH 1974, M. MIŠÍK 1976). Of Triassic age may be also some volcanic rocks in the Cretaceous conglomerates of the West Carpathians. This implies that the individual types of

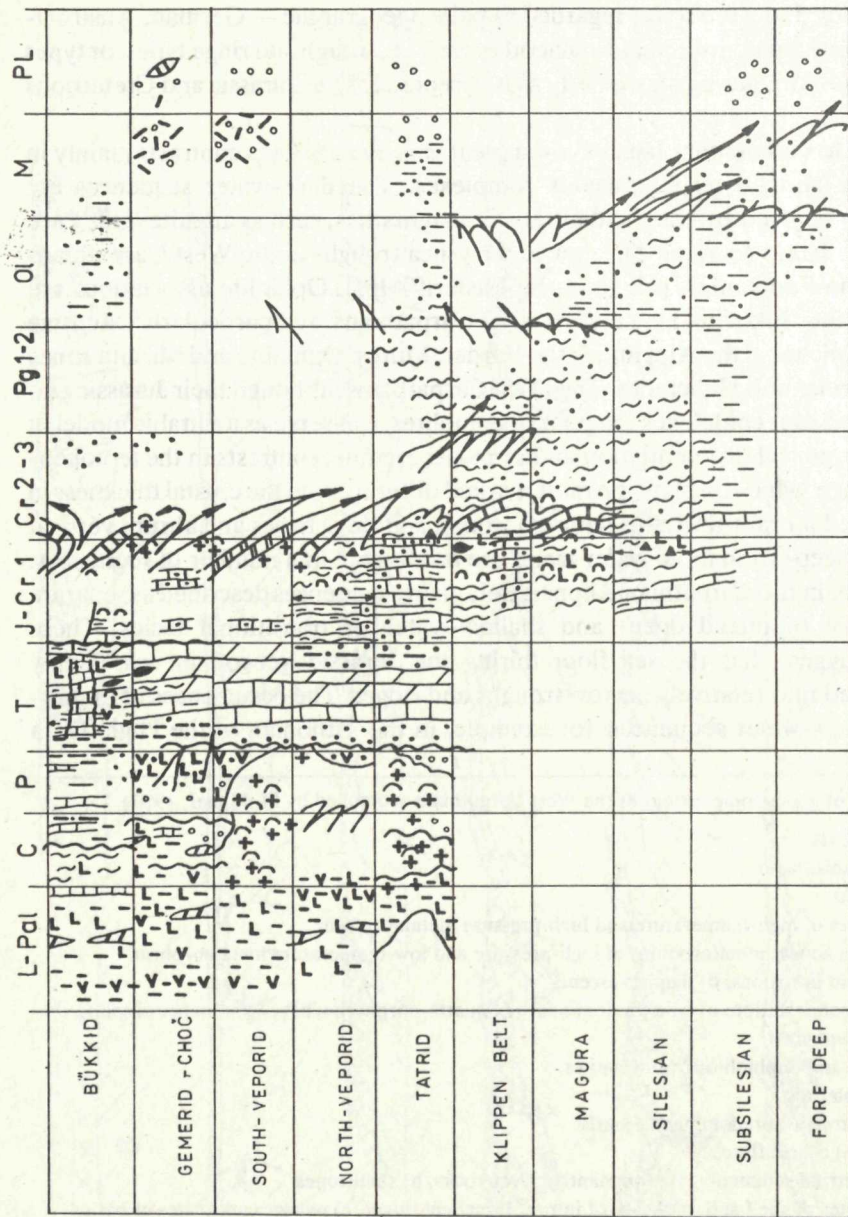


Triassic facies that were so far regarded as palaeogeographic — German, Austroalpine, Dinaride types, are actually palaeotectonic, i.e. trough and ridge types, or types with thick and thinner crust, similarly as is accepted for the Jurassic and Cretaceous facies.

The Alpidic sedimentary facies show a great variety and sharp contrasts mainly in the Jurassic and Lower Cretaceous complexes. The deep-water sequences are usually accompanied by basic and ultrabasic magmatites, such as augitites; they are deep-trough facies (e.g. the Zliechov and Kysuca troughs in the West Carpathians (Fig. 16), the Frankenfels trough in the Eastern Alps). Ophiolite associations are frequent in the Bukovina nappe of the East Carpathians, and particularly extensive in the Penninicum of the Alps and in the Vardar, Mureş, Ophiolite and Mirdita zones in the Dinarides and Hellenides. The West Carpathians, although their Jurassic and Lower Cretaceous complexes are poor in ophiolites, can serve as a suitable model of palaeogeographical differentiation and of palaeotectonic contrasts in the leptogeosynclinal stage, which displays the most marked differences in the crustal thickness in the Alpides. Lateral transitions between trough and ridge facies and abrupt vertical changes of deep- to shallow-water facies are observable in a number of sequences. For example, in the Carpathian Klippen Belt, many sequences described as separate series consist of mixed deep- and shallow-water or transitional facies. These instances suggest that the sea floor during the Jurassic-Neocomian was highly differentiated into relatively narrow troughs and ridges. The occurrences of limburgite in shallow-water sequences, for example, in the Tithonian of the High-Tatra

Fig. 16 Model of geotectonic history of the West Carpathians (compiled by M. Mahef, 1977)

1. Upper Mantle
  2. Granite-gneiss layer
  3. Nappe units
  4. Occurrences of high-temperature and high-pressure metamorphism
  5. Subduction zones; manifestations of high-pressure and low-temperature metamorphism
  6. a) Granitoid intrusions, b) diapiric ascents
  7. Volcanic manifestations a) with predominance of quartz-porphyrines, b) predominance of spilite-dibasites; melaphyres
  8. Ophiolites and "ophiolitoid" magmatites
  9. Thick radiolarites
  10. Facies of troughs, predominantly marls
  11. a) Flysch, b) coarse flysch
  12. a) Shelf detrital sediments, predominantly clayey rocks, b) sandstones
  13. a) Carbonates of shelf and ridges, b) of intrageosynclinal ridges, c) pelagic carbonates of basins
  14. Molasses
  15. Ascents of basic and ultrabasic magmas
  16. Overthrust lines; deep faults
  17. Zones of crust spreading
- Deep faults: B — Balaton line, D — Darnó line, R — Rožňava line, L — Lubeník line, M — Muráň fault, Č — Čertovica line, Po — Peripienian lineament, Le — Lednica line



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tectonic unit in the West Carpathians, evidence the dissection of the Alpidic geosyncline. Not even the ophiolite zones are necessarily indicators of the former vast oceans and, consequently, of large-scale spreading of the ocean floor and drifting apart of continental plates. The distribution of ophiolites in several narrow zones separated by zones with Jurassic and Lower Cretaceous shallow-water facies demonstrates the differentiation and contrasts in the Mesozoic geosyncline, particularly during its leptogeosynclinal stage. Is it not more adequate to consider the zones with ophiolites as extreme types of troughs in the system of geosynclines?

An appreciable part of ophiolites is genetically tied with the initial stage of flysch troughs, all of which are of Tithonian-Early Neocomian age. It is the Silesian trough in the West Carpathians, the Ceahlău trough in the East Carpathians, the Severin trough in the South Carpathians, the Niž-Trojan, and Strandža troughs in the Balkanides, the Sarajevo trough in the Dinarides and the Gramos trough in the Hellenides (M. MAHEL 1973) (Fig. 13). Their filling begins invariably with pre-flysch. The presence of ophiolites or basic and ultrabasic rocks (e.g. in the Ceahlău-Rachovo tectonic unit) indicates that the crust was of oceanic or near to oceanic type (Fig. 11). Some troughs with ophiolites associated with the leptogeosynclinal carbonate stage developed into flysch troughs; such were the Vardar and Mureş zones and the Penninicum. Other ophiolite troughs are linked up even with the flysch troughs (Szolnok, Severin, Sarajevo, Gramos).

The Alpides are distinguished by an abundance of Cretaceous-Palaeogene flysch sequences, which differ from similar sequences of the earlier cycles in the following features:

←  
Fig. 17 Table presenting geotectonic history of the West Carpathians. Compiled by M. MaheI 1977

1. a) Granitoid intrusions, b) diapiric ascents
2. Pre-alpine aspidic to flyschoid and flysch formations
3. a) Quartz porphyries, b) spilite-diabases; melaphyres
4. Bodies of a) of basic intrusives, b) of alkalic basites and ultrabasites
5. Andesites and rhyolites, a) basalts
6. Sandstones, quartzites, a) Carpathian Keuper.
7. a) Shallow-marine detrital sediments, b) with predominance of clayey rocks
8. a) Shelf limestones, b) dolomites
9. Carbonates of intrageosynclinal a) ridges, b) basins, troughs
10. Trough facies a) prevailing marls, b) marly limestones
11. Thick radiolarite layers
12. a) Flysch, b) coarse flysch
13. a) Shelf and riges sediments with predominance of clayey rocks, b) sandstones
14. Molasses
15. a) Manifestations of high-temperature and high-pressure metamorphism, b) subduction zones; manifestations of high-pressure and low-temperature metamorphism
16. Overthrusts
17. Reverse faults; manifestations of folding

- a great facies variety (M. MAHEL 1974), represented by fine-rhythmical flysch s.s., coarse flysch, wildflysch, subflysch, cryptoflysch (most frequently couches rouges), carbonate flysch, flysch with shallow-water limestone beds;
- the presence of numerous structural-facies zones;
- the polarity in the distribution of flysch in the progress of folding and in differences concerning its geotectonic position;
- a wide time interval of flysch formation: from the Tithonian through Albian-Cenomanian (typical flysch) until the Early Miocene in some segments (late flysch grading into molasses — West Carpathians, East Carpathians, Hellenides). Sedimentation of diastrophic type thus predominated from the onset of Palaeo-alpine orogeny to the beginning of the Neo-alpine folding processes. During flysch sedimentation crustal contraction was obviously the main dynamic process;
- a small proportion of magmatites.

As far as the volcanic rocks are present, they are of limited extent and intermediate type (tuffs, tuffites and breccias of andesite and dacite). Their eruptions were synchronous with the differentiation of the crust and with its palaeogeographical and palaeotectonic rebuilding and were located in the peri-cordillera faults.

A peculiar palaeotectonic feature developed after the Palaeo-alpine orogeny is the system of graben-synclinoria in the Timok-Srednogorie zone in the South Carpathians and Balkan (Fig. 14), characterized by intensive intermediate volcanism of rift-type, which followed the deposition of early molasse with coal seams in the Cenomanian-Turonian. During the volcanic activity, especially after the main effusion phases, a thick flysch complex was deposited, containing only thin volcanic layers. In the final stages of flysch sedimentation intrusive bodies were emplaced along the margins of grabens. In the flysch geosyncline itself, represented in the Balkan by the Luda Kamčija zone farther north, volcanic activity was weak.

In the Alpine cycle numerous depressions of several geotectonic types and with different types of magmatites originated (Fig. 15a, b):

- depressions of the foredeep, genetically linked up with the flysch tectonic units in space and time; volcanic phenomena are almost absent;
- intramontane depressions superimposed on the internal Alpidic zones, more or less consolidated by Palaeo-alpine orogeny;
- interior depressions along the inner margin of the geosynclinal system, extending onto the adjacent inner massifs (Hungarian, Transylvanian and Rhodope massifs).

The two last-named types are of multiphasal development and are fairly rich in magmatites. The development phases of these depressions in the West Carpathians and in the Hungarian massif alternated with the formation of nappes in the Carpathian Flysch Belt. The folded Savian, Early Styrian and Late Styrian nappe units correspond to the depressions of the same age. As a consequence of the polarity

of folding the Neo-alpine stage produced crustal shortening and thickening in the outer zones of the geosynclinal system, and in the inner zones crustal stretching, thinning and splitting, accompanied by subsidence and volcanism took place. The stage of inner molasses is designated as late geosynclinal, which term seems therefore appropriate even for the synchronous volcanic rocks.

The abundance of flysch and molasse complexes reflects increased crustal mobility and the changes in their facies and geotectonic types reflect the changes in the Alpine orogenic process. Its stage-evolution connected with the rebuilding of the palaeogeographical and structural patterns as well as its multiphase character in each of three main (Palaeo- Meso- and Neo-alpine) periods were combined with orogenic polarity, which extended the original geosynclinal polarity. Polarity is obviously a hereditary characteristic of the Alpine geosynclinal system, and the substitution of the ophiolite troughs by flysch troughs and of flysch complexes by molasses in vertical and lateral directions is the result of it. The zones with ophiolites as members of the geosyncline systems are involved in the polarity (Figs. 16, 17).

The alternation of compressive movements causing the crustal shortening in the outer Alpine zones and dilatation leading to the origin of depressions is remarkable in the West Carpathians (Fig. 15b).

The synchronicity of subsidence of the basins and of the volcanic activity in the interior zones and the Hungarian hinterland with compression on the outer border of the Carpathians indicates their relationship to the subcrustal processes within the entire width of the geosynclinal system. The changes provoked by subsidence and rooting of the crustal parts during folding very likely induced the ascent of a magmatic diapir in the interior zone. The diapiric intrusion is namely thought to be responsible for crustal thinning and this in turn for subsidence, the formation of basins and the ascent of volcanics. All these processes in the inner zones are controlled predominantly by vertical movements, which also produced all the morphostructures. It was the late geosynclinal effusive volcanics that took part in their formation and no longer the granitoid bodies as they did during the late Hercynian stage.

The Alpine orogeny is characterized by a strong crustal shortening and a variety of tectonic styles (Fig. 16). The latter results from the different thickness and physical properties of the sediments filling the individual structural-facies zones. The most striking manifestations of the crustal shortening are near-surface and subsurface nappes, zones of intense compression and zones of *mélange* indicating subduction. Granitization and metamorphism which had been characteristic of the Hercynian orogeny were confined to narrow zones during the Alpine cycle.

The evolution of the Alpides was completed by an abrupt uplift or subsidence of their individual parts and in result of it the morphostructural pattern initiated before by the formation of basins, became still more expressive. The distinctive alkalic type



of final (basalt) volcanism, which was of a small extent but active in all Alpidic segments, shows that the Pliocene-Quaternary period was a separate evolution stage of the Alpides.

The evolution trend of the Alpides as described above reveals the differences between the geosyncline preceding the Hercynian granitization and the geosynclinal system of the Alpine cycle. Whereas the former was relatively little differentiated, the latter was strongly diversified in vertical and lateral directions and in time as well, and displayed many structural-facies zones. The geosynclinal and orogenic polarity caused the differences in the evolution of individual zones, which are most striking between the inner and outer units. The differences are particularly striking in the Tertiary development. While fold structures and Neo-alpine nappes developed in the outer zones, the late geosynclinal stage with the formation of basins was initiated in the inner zones.

The more dynamic character and diversity of the Alpine geosynclinal system compared to those of the pre-Alpine geosyncline, caused the differences in the distribution and type of the magmatites. Therefore Stille's division of magmatites into initial, orogenic and subsequent seems rather schematic for the geosynclines of the Alpides.

The submitted evolution trend of the Alpine geosynclinal system relates primarily to the Eastern Alp — West Carpathian — East Carpathian region. The Dinarides show some differences in their development caused mainly by the differences in the primary state of the crust, the greater part of which was less consolidated by Hercynian orogenic processes.

### **3. Specific features of magmatism in the individual Alpidic areas and segments**

Although the magmatism of the Alpidic system shows a uniform evolution trend, the quantities of the magmatite types and the times of magmatic phases in the individual segments widely differ. These differences are caused mainly by the unequal crustal consolidation during Hercynian orogeny and unequal portion of Hercynian granitoids. They are reflected in the diversity of the Permian and Triassic facies, in the differentiation of the Mesozoic Alpine geosynclinal system and the dates and extent of ophiolitic and other magmatic types. Each segment of the Alpides shows some specific features in this respect (Fig. 2):

a) The Dinarides (Figs. 9, 12b) are distinguished by a small amount of Hercynian granitoids, which are concentrated to the margins of the Pelagonian and Serbo-Macedonian massifs. The small extent of Hercynian granitization and a minor intensity of Hercynian folding (Fig. 7) resulted in a lower consolidation of the crust and its smaller thickness. This obviously accounts for the predominantly marine development of the Permian, an increased perviousness of the crust in the Triassic and

abundant basic magmatism even in the external zones (e.g. in the Budva-Zukali zone). At first, i.e. in the Anisian, the magmatites were of hybrid character with a greater portion of acid rock types, such as porphyrites, quartz keratophyres, keratophyres and quartz porphyries. Beginning with the Ladinian the basicity of magmatism gradually increased; basaltic pillow lavas became more frequent. Sedimentation of trough type with cherts, siliceous limestones and claystones was more extensive.

None of the Alpidic segments described is so rich in ophiolites and contains so many and so large serpentinite bodies as the Dinarides and Hellenides (Fig. 13a). The ophiolite associations occur in two zones: in the eastern Vardar zone and the western Ophiolite and Mirdita zones (Subpelagonicum). They are separated by the Drina — Ivanjica — Pelagonian massif (with the Korab zone) belt, in which the continental crust is of a greater thickness. Each of these zones shows a different stratigraphic range and content of the ophiolite associations, each of them represents an oceanic trough of another type. The above differences also exist within one trough as, for example, in the Mirdita zone the ophiolite association follows the Triassic volcanogenic series, but in the greater part of the more northerly Ophiolite zone the magmatites are Upper Jurassic. On the other hand, in the north-western extension of the Ophiolite zone in the Southern Alps the ophiolite association is missing and only Triassic hybrid magmatites are there present. Ophiolite melanges of great thickness and transitions of deep-sea ophiolite sequences into shallow-sea sequences are distinctive of the Dinarides and Hellenides, even if they show some peculiarities in each major zone.

The two oceanic ophiolite troughs differ in their relationship to the flysch sequences. In the Vardar zone the flysch complex overlies the ophiolite association and its lower members contain magmatites of the ophiolite type. In the Ophiolite and Mirdita zones the flysch trough of Sarajevo-Vermos developed since the Tithonian along the western margin of the ophiolite trough and after the extinction of the latter it extended westwards as the Durmitor zone.

A characteristic feature of the Dinarides and Hellenides is a strong tectonic activation of their contact zones and of the Serbo-Macedonian and Pelagonian massifs in the east. Syenite dykes and granodiorite bodies (in the Hellenides even synkinematic) became more abundant; their ages have been determined as early Cimmerian (167—156 m. y.), late Cimmerian (148—120 m. y.) and Middle Cretaceous (110—90 m. y.).

The Neo-alpine activation is also responsible for the existence of numerous minor neovolcanic bodies with occasional small volcanic edifices (andesites, dacites, rhyolites, quartz latites) and frequent subvolcanic bodies (granodiorites and quartz monzonites). The volcanoes are generally situated at the crossing of fault systems; where they accompany depressions, these are of small extent. The zone of activation is highly diversified with outstanding uplifts of minor blocks (M. ARZOVSKI 1961).

The period of volcanic activity was relatively long, lasting from the Oligocene to the Sarmatian (37 m. y. — 10.8 m. y. B.P.).

b) The Balkan (Fig. 6) and South Carpathians (Fig. 5) are characterized by extensive Hercynian granitization, which affected virtually their entire areas. In the South Carpathians and the Balkanides, both pre-Palaeozoic and Hercynian granitoids are comparatively abundant. Alkalic granitoids are frequent in the South Carpathians, especially in the Danubicum. The Balkanides are distinguished by an unusual variety of geotectonically differing types of granitoids. In addition to the South Bulgarian granodiorites and granites of the Rhodope massif and the Srednogorie Mts., mention should be made of the Stara Planina gabbrodiorite-granodiorite association, the Struma association ranging from acid granites to ultrabasites and the Permian gabbro-syenite association. The Hercynian granitization and folding and, consequently, the consolidation of the crust were there far more intensive than in any other Alpidic segment described, so that the Triassic of the South Carpathians and Balkanides can be most aptly denoted as being of quasiplatform type (Fig. 7). The heterogeneity of the pre-Alpine development in the Balkanides, reflected in the great variety of granitoid types, also manifested itself by the formation of the first geosynclinal troughs with a thinner crust in the eastern part of the Balkanides, which are represented by the Sakar zone in the south and the Kotel zone in the north. A marked segmentation of the geosyncline began in the Tithonian, simultaneously with the formation of flysch sequences. Ophiolite occurrences linked up with the Tithonian preflysch trough are sparse: Strandža and Sakar areas in the Balkan, serpentinites along the boundary faults of the Rhodope massif, chiefly along the Marica fault and the Severin-Krainia trough in the South Carpathians.

Longitudinal faults such as the Upper Cretaceous Timok-Srednogorie rift, played an important role in the tectonic history of this area. The fault system accompanying the graben-synclinoria and the volcano-plutonic banatite association is of the same kind; it roughly follows the boundary between the Geticum and Danubicum in the South Carpathians and between the Balkanides and the Rhodope massif in the Balkan.

The neovolcanics of the Balkan are linked mainly with the Rhodope massif, which constituted the hinterland to the Balkanides, with interior basins in the rear part of the geosyncline. This linkage is demonstrated by the stratigraphic range of magmatites and of depressions, which in the Bulgarian area is Priabonian to Oligocene and in Greece up to Miocene (Fig. 15). The activation in the rear of the geosyncline in the Rhodope massif had evidently a southern trend. The South Carpathians lack young Tertiary volcanics and major interior basins. The only exception is the Krajištide area in the south, which both genetically and structurally is tied with the Balkanides.

c) In the Apuseni Mts. (Fig. 4) the East (Fig. 11a) and West Carpathians (Fig. 3) and in the Eastern Alps the Hercynian granitization and intense Hercynian folding

mainly affected the northern sectors. The Palaeo-alpine granites are typical of the Gemeride zone of the West Carpathians (Fig. 10).

In all these mountain ranges, thick complexes of quartz porphyry were emplaced at the beginning of the Palaeozoic as a result of the Baikalian crustal differentiation. The Hercynian crustal differentiation (Fig. 7) is reflected in the differences between the Permian facies, diversified especially at the transition to the Moma zone, where melaphyres appear in addition to quartz porphyries. In the extreme south, the Permian is of marine type. Even the Triassic is represented there by varied carbonate sequences and several facies-structural zones expressing a characteristic crustal differentiation of the Alpine geosynclinal system. Besides basic rocks, ophiolite associations occur more frequently in the Transylvanian units of the East Carpathians, in the Slovakian Karst (Meliata Formation) and in the Bükk Mts. of the West Carpathians. The distribution of basites in the southern zones of the Eastern Alps is restricted to sporadic occurrences in the higher nappes of the Calcareous Alps and ophiolite bodies in the Middle Triassic of the Drauzone. This may be the reason why the Bükk Mts. differing from these Alpine zones in the marine type of the Permian and the Triassic rich in volcanics has long been regarded as a branch of the Dinarides linking them with the inner zones of the West Carpathians (K. BALOGH—L. KÖRÖSZY 1968). The establishment of an analogous ophiolite series (Meliata Formation) and other 'southerly' features of the Slovakian Karst provides evidence that the Bükk Mts. is an inseparable part of the West Carpathians. Together with the Slovakian Karst it constitutes the innermost zone called the 'Bukovicum' comprising several tectonic subunits (M. MAHEL 1975). This southern ophiolite-bearing zone is bounded by the Rožňava fault line (at the northern margin of the Slovakian Karst) and the Balatón fault line (at the southern margin of the Bükk Mts.). These two lines along with the Darnó line (at the northern margin of the Bükk Mts.) are probably northern branches of the Insubric line, which does not only bifurcate to the north-east but also changes its geotectonic character. There is no geosynclinal segment to the east of it, as are the Dinarides and the Southern Alps, but the Hungarian massif.

The geosynclinal system of the West and East Carpathians and the Eastern Alps shows a marked differentiation reflected by unusual variety of facies and sequences in the Jurassic and Early Cretaceous leptogeosynclinal stage, particularly in the central and northern Alpine zones. Basic rocks and incomplete ophiolite series are frequent (e.g. in the Zliechov and Kysúca troughs in the West Carpathians, the Bukovina trough in the East Carpathians). Characteristic of the West Carpathians are abundant small augitite and limburgite occurrences in shallow-water facies, which are obviously connected with the faults separating submarine ridges. Their frequency is probably due to the block structure of the Inner West Carpathians. An important ophiolite association of the Alps appears in the Penninicum, including the Rechnitz tectonic window in the east. In the East Carpathians, a well known ophiolite series dating from the initial stage of the flysch trough is in the Ceahlău-Ra-

chovo tectonic unit, at the boundary between the outer and inner Carpathian zones.

In the Flysch Belt of the West Carpathians the ophiolites and basic rocks in general occur in a small amount and particularly along the margins of troughs and cordilleras (for example the teschenites). Minor bodies at the margin of the Klippen cordillera suggest a plausible presence of major ophiolite masses in the root zone of the Magura nappe, with regard to the proximity of the Rechnitz Formation. In contrast to the Eastern Alps, the West Carpathians have a hinterland — the Hungarian Massif, which substantially affected their history in its late geosynclinal stage.

The neovolcanics, which are characteristic of the West and East Carpathians (Fig. 15), are absent in the Eastern Alps as a result of the missing hinterland and a lack of interior basins. The Carpathians were obviously connected with the Hungarian median mass during the Tertiary. The West Carpathians were disrupted by transverse faults and the Central Slovakian transverse graben was the site of extensive volcanism. As the rear depression in the East Carpathians extended up to the Klippen and Flysch Belts at the southern boundary of the Outer Carpathians, the belt of neovolcanics overlaps a large part of the inner zones.

Characteristic of the Hungarian massif south-east of the Balaton line is a thick Permian complex with quartz porphyries. By analogy with the West Carpathians, the differentiation of the crust caused by Hercynian folding, metamorphism and granitization and the formation of a Permian trough along or between Hercynian granitoid zones may be considered. The alkalic character of Lower Cretaceous basites in the Mecsek and Villány Mts., resembling the teschenite association of the outer zones of the West Carpathians, and the 'northern' type of the Mesozoic suggest that the south-eastern block of the Hungarian massif might belong to the northern zone of the Alpides, either as a tectonic window or as a block shifted south-westwards. The drawback to this consideration is the lack of Palaeo-alpine tectonic disturbance in the Mecsek and Villány Mts., and of Lower and Middle Cretaceous flysch sequences, which are so characteristic of the northern West Carpathian zones. It appears that the 'northern' type of the Mesozoic basites and Early Cretaceous alkalic basites are related with the tectonic type of the crust which was relatively thick but broken at that time. It represents most likely a prolongation of the Serbo-Macedonian massif.

From the basites obtained by boring into the basement of the Neogene of the Hungarian Plain (including Transcarpathian region) the presence of ophiolites is assumed west of the Szolnok flysch trough (E. SZÁDECKY—KARDOSS et al. 1967, G. N. DOLENKO—L. G. DANILOVIČ 1976). The ophiolite-filled trough together with the Szolnok trough separate the Apuseni Mts. from the West Carpathians; they extend nearly parallel to the Balaton line and other major structures of the Hungarian massif. The ophiolite trough shows an analogous position to that of the Mureş ophiolite zone; both of them are branches of the Vardar zone penetrating deep

between the blocks of the Alpine mountain system. The Vardar zone itself separates the Serbo-Macedonian massif from the Pelagonian massif.

The northern part of the Apuseni Mts. recalls the Inner West Carpathians in having major bodies of Hercynian granitoids in the north and abundant Permian complexes with thick quartz porphyry and melaphyre layers in the south. The structural-facies diversity of the Triassic increases southwards as well as the number of 'southern' types. Their Neo-alpine volcanics are distributed over an area which forms part of the Neotisia including the inner zones of the West and East Carpathians and the Hungarian Massif. During the Jurassic to Eocene, however, the Apuseni area underwent a development in common with the Mureş ophiolite zone, which is an extension of the Vardar zone structurally linked with the North Apuseni. This mountain range is in turn allied to the South Carpathians by the presence of Upper Cretaceous-Palaeocene banatites. The Apuseni is a median block, which during the Alpine cycle was affected by the volcanism-controlling tectonic features extending there from the neighbouring areas. Their effects overlapped in time at some places. These were the ophiolite zone of Mureş (in the south and east) and the tentative ophiolite zone at the western margin of the Szolnok flysch trough (west of the Apuseni), together with the drawn-out Klippen Belt and its appendage in the Poiana Ruscă to the east. The divergence of these structures and of the Apuseni Mts. indicate a shift and rotation of this block, probably in result of crustal thinning and stretching during the initiation of the ophiolite troughs.

#### 4. Notes on terminology

In describing the magmatite groups in the text above we intentionally avoided the commonly used terms of initial, orogenic and subsequent magmatism. It does not imply, however, that the cyclic evolution is denied. The magmatism displays a cyclicity but within a wider global scope. It directly depends on the type and thickness of the crust and tectonic processes which act in cycles. The Alpine system provides an example how the type, extent and intensity of folding change from one cycle to another (M. MAHEL 1975), the same as the crustal type and magmatism.

The granitoids, which represent magmatites genetically associated with folding, have a different share in the formation of the continental crust and of the structures in individual cycles. In the pre-Palaeozoic cycles the granitization was manifested by anatexis and high metamorphism and probably a long period of 'heat flow', but less marked effects on the structural pattern. The Hercynian cycle was distinguished by the emplacement of vast batholiths and phacoliths, a slighter anatexis, differential metamorphism and extensive crustal consolidation. On account of orogenic multi-stage movements and development of the structural pattern it was an important structural and morpho-structural factor. In the Alpine cycle, synorogenic magma-

tism was represented only by minor bodies of granite within narrow zones and its effect on the formation of the structural pattern was minimum.

The changes in the so-called subsequent magmatism from one cycle to the other are most striking. In all three last cycles the magmatism was of acid or intermediate character, but its geotectonic position and genetic type were different in each cycle. Quartz porphyries are components of geosynclinal sequences subsequent to the Baikalian orogeny; they grade into basic igneous rocks both laterally and vertically. Acid volcanics reflect the differentiation of the crust and the formation of zones of greater crustal thickness — the island arc zones.

The term 'subsequent' does not fit either the late Hercynian volcanics. They followed the main Hercynian granitization and folding, but correspond in age to the latest granites, replacing them laterally. They are confined to grabens with molasse filling, occasionally of the rift type, i.e. the negative morpho-structural forms. Consequently, they are not a consolidating element as are granitoids. Although they complete the geosynclinal cycle, they signify the differentiation of crust before the foundation of the Alpine geosyncline.

The Meso-alpine (Upper Cretaceous-Palaeocene) interorogenic magmatites called 'banatites' (in the literature often denoted as early subsequent — H. STILLE 1953) and neovolcanics (called late subsequent) differ from the Upper Palaeozoic magmatites in a broad differentiation range, and structural and volcano-tectonic forms. In the Srednogorie and Timok areas, which are classical areas of their occurrence, the banatites are folded but do not show a direct relationship to the foregoing Palaeo-alpine orogeny; their volcano-plutonic character is a sign of their close connection with the increased dynamics of the crust. As they formed subsequently only to the first Alpine orogenic stage, they are more conveniently termed as interorogenic or Meso-alpine (on account of their close association with the Meso-alpine stage) than early subsequent.

The designation 'subsequent' is not very apt either for the neovolcanics. Their distribution in the zones that had been consolidated during the Palaeo-alpine or earlier orogenic processes bears witness against their subsequence to folding and granitization. The young volcanics are characteristic products of the late geosynclinal stage and simultaneously of the activation of geological processes in the parts of the geosynclinal system that were stabilized in earlier times or not fully incorporated in the system. The volcanics of the geosynclinal system itself, i.e. of the West and East Carpathians and the Apuseni Mts. can be labelled as late geosynclinal, but those of the massifs (e.g. Rhodope and Pannonian) are better to be called the volcanics of activation zones. Even in this case, however, they represent the late geosynclinal stage of the Alpine geosyncline whose part they form. The young volcanics and the superimposed basins of the Hungarian massif and Rhodope should be regarded as components of the Carpathian or Balkan system, as part of the backdeep just as the foredeep, which is also an activated part of the platform.

The 'initial' type of magmatites predating the Middle Carboniferous is distributed in almost all geosynclinal complexes both at the beginning of the cycle and prior to the folding phase. The large amount of basic magmatites is due to a small thickness and permeability of the crust. A more fitting term would be therefore 'early geosynclinal' or 'immature' or 'pre-Alpine geosynclinal' type of volcanics.

The Mesozoic basites have a higher proportion of ultrabasites and less pyroclastics and more frequently accompany deep-sea sediments, such as pelagic limestones and silicites, in several evolution stages of the geosyncline. They are found in the Triassic, Jurassic and Lower Cretaceous carbonate stages of the geosynclinal history as well as in the pre-flysch stage of the later (Cretaceous) flysch geosyncline. The term 'initial' is not suitable for these magmatites also because they are confined only to some, usually narrow zones of the geosyncline — to the troughs having a thin oceanic or nearly oceanic crust.

Acknowledgement. It is a pleasure to express my sincere thanks to Doc. Dr. Zdeněk Roth, DSc., for many valuable and constructive comments and for suggesting the themes of some chapters. I am also indebted to Professor Dr. Jakub Kamenický for petrographic-terminological advice and to Mrs. Helena Zárubová for careful translation of this paper.

## References

- ABÓNYI, A. 1971: Stratigraficko-tektonický vývoj karbónu gemeríd západne od štítického zlomu. Geol. Práce, Správy 57, Geol. Úst. D. Štúra, Bratislava, 339-348.
- ALEXIĆ, V.—KALENIĆ, M.—PANTIĆ, N.—HADŽI, E. 1974: Historical geology of the evolution of continental transitional-oceanic and oceanic lithosphere in Serbia and surrounding areas. Metallogeny and concepts of the geotectonic development of Yugoslavia, Belgrade, 229-274.
- ANDRUSOV, D. 1968: Grundriss der Tektonik der Nördlichen Karpaten. Vyd. Slov. Akad. Vied, Bratislava, 1-118.
- ANTONJEVIĆ, I.—GRUBIĆ, A.—DJORJEVIĆ, M. 1974: The upper cretaceous paleorift in East Serbia. Metallogeny and concepts of geotectonic development of Yugoslavia, Belgrade, 315—339.
- ARKAY, P. 1973: Pumpelit-prehnit-quarz facies alpine metamorphism in the middle triassic volcanogenic-sedimentary sequence of the Bükk mountains, northeast Hungary. Acta geol. Acad. Sci. hung. 17, 1—3, Budapest, 67—83.
- ARSOVSKI, M. 1961: Bemerkungen zur tektonischen Entwicklung des pelagonischen Horst-Antiklinoriums und seine Stellung im Rahmen der alpidischen Struktur von Macedonien. Bull. Inst. geol. Maced. 8, Skoplje, 65—95.
- AUBOUIN, J. 1964: Geosynclines. Amsterdam, 1—301.
- BAJANIĆ, Š. 1969: Kyslé diferenciáty bázického vulkanizmu v rakoveckej sérii. Geol. Práce, Správy 48, Geol. Úst. D. Štúra, Bratislava, 71—80.
- BAJANIĆ, Š. 1975: Pillow lavas of the Rakovec group in the Spiško-gemerské rudohorie Mts. Geol. Zbor. Geologica carpath. 26,2, Bratislava, 341—348.
- BAKSA, CS.—CSILLAG, J.—FÖLDESSY, J. 1974: Volcanic formations of the NE Matra mountains. Acta Geologica Academiae Hungaricae, 18 (3—4), Budapest, 387—400.



- BALOGH, K.—KÖRÖSSY, L. 1968: Tektonische Karte Ungarns, Masstab 1 : 1 000 000. Acta geol. hung. 12, 1—4, Budapest, 255—264.
- BAGDASARIAN, G, P.—CAMBEL, B.—VESELSKI, J.—GUKSJAN, R. CH. 1977: Kalijargonovije opredelenija vozrasta kristalličeskich kompleksov Zapadnych Karpat i predvaritefnaja interipretacija rezultatov. Geol. zbornik Geologica Carpathica 28, 2, Bratislava.
- BECK—MANNAGETTA, P. 1974: Austrian Eastern Alps. Internal zones. In Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava, 57—74.
- BELOUSOV, V. V. 1975: Osnovy geotektoniki. „Nedra“, Moskva, 1—260.
- BERBELEAC, I. 1975: L'étude pétrographique et métallogénique de la région Valisoara (Porcurec) Monts Metalliferes. Anu. Inst. Géol. St Géophys., 46, Bucuresti, 5—190.
- BERCIA, J.—BERCIA, E. 1975: Formatinnile cristaline din sectoral romanese al Dunarii (Banat-Carpatii Meridionali). Les formations cristallins du secteur roumain du Danube (Banat-Carpates Meridionales). Anu. Inst. Géol. si Geophys., 43, Bucuresti, 5—63.
- BIÇOKU, T.—CILI, P.—DEDE, S.—PAPA, A.—SHEHU, R. 1974: Geological structure of the Albanides. In Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava, 365—389.
- BILIK, J. 1974: Unterkretazische Vulkanite des Mecsek-Gebriges. Acta geol. Acad. Sci. hung. 18, 3—4, Budapest, 315—325.
- BIRKENMAJER, K. 1965: Outlines of geology of the Pieniny Klippen Belt of Poland. Rocznik pol. Tow. geol. 35, Krakow, 401—407.
- BIRKENMAJER, K.—NAIR A. E. M., 1968: Paleomagnetic studies of Polish rocks, III. Neogene igneous rocks of the Pieniny Mountains Carpathians, Ann. Soc. Geol. Pol. 38,4, 475—489.
- BLEAHU, M. 1974: The Apuseni Mountains. In Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava, 221—223.
- BLEAHU, M. 1976: Structural position of the Apuseni mountains in the Alpine System. Rev. Roum. geol. geophys. et géogr. Geologie t. 20, N° 1, Bucuresti, 7—19.
- BLEAHU, M.—BOCCALETTI, M.—MANETTI, P.—PETZ, S. 1973: Neogene Carpathian arc. A continental arc displaying the features of an island arc. Journal of Geophysical Research, 78, 23, 5025—5032.
- BOCCALETTI, M.—MANETTI, P.—PECCERILLO, A. 1974: Balcanides as an instance of back-arc thrust belt. Possible relation with Hellenides. Geol. Soc. Am. Bull. 85, 1077—1084.
- BOGDANOV, B. 1973: Sequence of ore forming processes related to Senonian volcanic activity in Panagyurishte ore district (Bulgaria). International symposium on volcanism and associated metallogenesis, Bucuresti, 29—31.
- BOJANOV, I. 1974: The Rhodope massif: in Tectonics of the Carpathian Balkan regions. Geol. Úst. D. Štúra, Bratislava, 342—345.
- BOJANOV, I.—KOŽUCHAROV, D. 1968: Strojenije i blokovoje rasčlenenije rodopskogo massiva. Bull. of the Geolog. Inst. s. geotectonics, stratigraphy and lithology, 17, Sofia, 199—226.
- BOJKO, A. K. 1975: Voprosy drevnej geologičeskoj istorii Vostočnych i Zapadnych Karpat i radiometričeskoje datirovanije. Naukova dumka, Kijev, 1—41.
- BOJKO, A. K.—KAMENICKÝ, L.—SEMENENKO, N. P.—CAMBEL, B.—ŠČERBAK, N. 1974: Časť rezultatov opredelenija vozrasta gornych porod kristalličeskogo massiva Zapadnych Karpat i sovremenoje sostojanije znani. Geol. Zbor. Geologica carpath. 25, Bratislava, 25—39.
- BONČEV, E. 1955: Geologija na Bulgarija, Sofija.
- BONČEV, E. 1966: Revue générale de la structure géologique de la Bulgarie. Bul. Inst. of Geology, 15, 5—24.
- BONČEV, E. 1976: Lineament-Geosynclinal Zones — Zones of Impermanent Riftogenesis. Geologica balcanica 6, 1, Sofija, 85—100.
- BORKOŠ, M. 1974: Neogene and final volcanism in Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava, 278—283.
- BOYADJIEV, S. 1967: Über die Entwicklung des Magmatismus in Bulgarien. Acta geol. Acad. Sci. hung.

- 11, 1—3, Budapest, 95—114.
- BOYADJIEW, S. 1974: On the results of the radiometric age determinations of the pre Mesozoic basement in parts of the Balkan peninsula. Bulgarian Academy of Sc., geological institute; Mineral genesis, Sofija, 349—363.
- BOYADJIEW, S.—LILOV, P. 1971: On the data from the Pirin horstanticlinorium obtained by K/Ar method. Jahrbuch der Staatlichen Wirtschaft-Vereinigung. Geologische Erkundungen, 20, Sofija, 219—226.
- BOYDJIEW, S.—LILOV, P. 1972: On the data on the age of the Southbulgarian granitoids from the Srednogorian and Sakar-Strandza zones as determined by the K/Ar method. Bull. Geol. Inst. s. Geochem. min, and petrogr. 21, Sofia, 211—220.
- BRUNN, J. A. 1960: Mise en place et differenciation pluto-volcanique du cortege ophiolitique. R v. G ogr. phys. S r. 2, 8, 3.
- BURCHART, J. 1970: The crystalline core of the Tatra mountains, a case of polymetamorfism and polytectonism. Eclogae geol. Helv. 63, Basel, 53—56.
- CAMBEL, B. 1976: Probleme der Metamorphose und der Stratigraphie des Kristallins der Westkarpaten mit Hinsicht auf die Forschungen in dem Bereich der Kleinen Karpathen. Geol. zbor. Slov. Akad. Vied, Geologica carpatica 27, 1 Bratislava, 103—116.
- CIOFLICA, G. 1967: Entwicklung des laramischen Magmatismus in Rum nien. Acta geol. Acad. Sci. hung. 11, Budapest, 153—160.
- CIOFLICA, G.—PATRULIUS, D.—JONESCU, J.—UDUBASA, G. 1966: Les ophiolites triasiques allochtones des mont Persani Carpatas orientales. Revue roum. G ol., G ophys. G ogr. S rie 10, 1, Bucuresti, 75—97.
- CIOFLICA, G.—VLAD, S. 1973: The correlation of laramian metallogenic events belonging to the Carpatho-Balkan area. Revue rouma. G ol. G ophys. G ogr., S r. G ol., 17, 2, Bucuresti, 217—224.
- CIOFLICA, G. et al. 1973: Alpine volcanism metallogenesis in the Apuseni Mountains. Symposium volcanism and metallogenesis. Bucuresti 1973, 13, 5—10.
-  IRIĆ, B. 1962: Le d veloppement des Dinarides Jugoslaves pendant le cycle alpin. Livre jubilaire a la memoire de P. Fallot, 2, Paris, 565—582.
-  UNEV, P.—BONEV, E. 1975: Sur  volution du volcanisme st phanopermien dans le systeme pliss  des Balkanides de Bulgarie. Geologica Balcanica 5,4, Sofija, 3—14.
- DELEON, G. 1966: Pogled na geološku starost granitoidnich stena u Pelagoniji i Zapadnoj Makedoniji na osnovu njihove izotopske starosti (Isotope ages of granitoid rocks in Pelagonia and western Macedonia): Referati VI Savetovanja geologa SFRJ, Ohrid, II (Proceedings of the 6th Conference of Yugoslav Geologists, Ohrid, II): Savez. geol. druŹatva, 59.
- DEWEY, J. F. and BIRD, J. M. 1970: Mountain belts and the new global tectonics. J. Geophys. Res., 75, 14, Washington, 2020—2647.
- DIETRICH, V. J. 1976: Plaketektonik in den Ostalpen. Eine Arbeitshypothese. Geol. Forsch. 50, 1—84, Stuttgart.
- DIMITRESCU, R. 1966: Beitr ge zur Kenntnis der magmatisch-tektonischen Verh ltnisse im Karpatisch-balkanischen Raum. Acta geol. Acad. Sci. hung. 10, Budapest, 357—360.
- DIMITRESCU, R. 1967: Contributti la cunoasterea stratigrafici si structurii partii de NW a masivului cristalin Highiz. D. S. Com G ol. Bul., 53, 1, Bucuresti, 39—50.
- DIMITRESCU, R. 1976: Les terrains m tamorphiques des Mountaignes Apuseni (Roumanie), Formations pr hercyniennes. Geol. Zborn. Geologica carpath. 27, 2, Bratislava, 347—354.
- DIMITRIEVIĆ, M. D. 1967: The geological problems of the Serbomakedonian masiv. CBGA 8. Congr. Belgrade.
- DIMITRIEVIĆ, M. D. 1974: The Dinarides, a model based on the new global tectonics. Metallogeny and Concepts of the geotectonic developement of Jugoslavia, Belgrade, 141—178.
- DIMITRIEVIĆ, M. D. et al. 1967: Metamorphic complexes of the Carpatho-Balcanic arch and adjacent areas.

- Acta geol. Acad. Sci. hung. 11, 1—3, Budapest, 23—34.
- DIMITROVA, E. et al. 1975: Magmatičeskije formaciji Bolgariji. *Geologica balcanica* 5, 1, Sofija, 52—63.
- DIVLJAN, M.—DIVLJAN, S. 1967: Granitoids of the Yugoslav part of the Carpatho-Balkan Arch in the geological Survey of the Carpatho-Balkan Arch. *Carp.—Balk. Geol. Ac.* 8, Congr. Beograde, 27—30.
- DOLENKO, G. N.—DANILOVIČ, L. G. 1976: Novoje učenje o geosinklinalach i jego primenenije v Ukrajinskich Karpatach. *Geol. Zbor. Slov. Akad. Vied, Geologica carpath.* 1, Bratislava, 1—10.
- ELIAŠ, M. 1976: Poznámky k paleogeografickému a paleotektonickému vývoju flyšových Karpát. *Zborník referátov vedeckej konferencie „Čsl. geológia a globálna tektonika“*, Bratislava, 43—53.
- FANINGER, E. 1976. Karavanken tonalit. *Rozprave Geologija, Poročila* 1976, 19, Ljubljana, 192—210.
- FLÜGEL, H. V. 1975: Einige Probleme des Variszikums von Neo-Europa. *Geol. Rdsch.*, 64 1—62.
- FÜLÖP, J. 1968: Formations crétacées de la Hongrie. *Annales Inst. geol. Hungarice* 49, 3, Budapest, 721—738.
- GATTO, G. O. et al. 1976: The "andesitic" magmatism in the South Western Tyrol and its geodynamic significance. *Geol. Rdsch.*, 65, 2, Stuttgart.
- GIUSCA, D.—CIOFLICA, G.—SAVU, H. 1966: Caracterizarea petrologica a provinciei banatice. *Anuarul comitetului de stat al geologiei.* 35, Bucuresti, 13—40.
- GIUSCA, D.—SAVU, H.—BERCIA, I.—KRÄUTNER, H. 1969: Sequence of tectonomagmatic pre-alpine cycles on the territory of Romania. *Acta geol. Acad. Sci. hung.* 13, Budapest, 221—234.
- GIUSCA, D.—GHITULESCU, T.—BORCOS, M.—LUPU, M.—BLEAHU, M.—SAVU, H. 1976: Evolution geologica a Muntilor Metaliferi. *Ed. Academici RSR.*, Bukuresti, 1—741.
- GOČEV, P. 1976: L'évolution géotectonique du mégabloc bulgare pendant le Trias et le Jurassique. *Bull. Soc. géol. France* (7) 18 N°, 209—216. *Coll. intern. CNRS, Paris N° 244*, 13—20.
- GOČEV, P.—KOSTADINOV, V.—SAVOV, S.—ZAGORČEV, S. 1974: Strednogorje. In *Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava*, 322—330.
- GOREK, A. 1959: Übersicht der geologischen und petrographischen Verhältnisse des Kristallins der Hohen Tatra. *Geol. Zborn.* 10 Bratislava, 125—129.
- GRECULA, P. 1973: The homeland of the gemeric and its metallogenesis. *Mineralia slovacae*, 5, 3, Spišská Nová Ves
- GRUBIČ, A. 1974: The Serbian-Macedonian mineralogic province in the light of the neoalpine activation. *Metallogeny and concepts of the geotectonic development of Yugoslavia*, Beograd, 1—311.
- HAJDUTOV, I. 1971: On the structural facies of certain plutons from the Western Balkan (Stará Planina). *Izv. Geol. Inst. r. geotektonika* 20, Sofia, 75—93.
- HAJDUTOV, I. et al. 1974: Stará Planina structural zone; in *Tectonics of the Carpathian Balkan regions. Geol. Úst. D. Štúra, Bratislava* 316—322.
- HERZ, N.—SAVU, H. 1974: Plate tectonics history of Romania. *Geol. Soc. of America. Bulletin* 85, 1429—1440.
- HESSE, R.—BUTT, A. 1976: Paleobathymetry of cretaceous turbidite basins of the East Alps. relative to the calcite compensation level. *Journal of geology* 84, 5, Chicago, 505—533.
- HORVATH, P.—STEGENA, L.—GECZY, B. 1974: Ensialic and ensialic interarc basins. *J. Geophys. res.* (Washington).
- HOVORKA, D. 1965: Ultrabasische Gesteine der Westkarpaten in der Slowakei. *Geol. Zbor. Geologica carpath.*, 1, Bratislava, 129—142.
- HOVORKA, D.—ZLOCHA, J. 1974: Tectonics and origin of Ultrabasic bodies of the Gemeric Mesozoic (West Carpathians). *Sbor. geol. Věd. Geologie* 26, Praha, 185—195.
- HOVORKA, D. 1976: Predterciérne formácie bazitov Západných Karpát. *Mineralia slovacae*, 8, 2, Spišská Nová Ves, 113—132.
- HOVORKA, D.—SLAVKAY, M. 1966: Pikrit od Poník. *Geol. Práce, Správy* 39, *Geol. Úst. D. Štúra, Bratislava*, 41—52.
- ILIČ, M. 1967: Položenije ofiolitov v geotektoničeskom razvitiji Dinarid. *Acta geol. Acad. Sci. hung.* 11,

- Budapest, 77—93.
- IVANOV, R. G. 1960: Magmatizm vostočno-rodopskoj paleogeonovoj vpadiny. Trudove vrchn. geol. na Blgariji s. geochemija i polezny iskopaemie 1, 303—387.
- IVANOV, R. G. 1965: Über Grundgebirgsbau und seinen Einfluss auf Petrochemie und Verteilung des Jungvulkanismus im Zentralmassiv der Rhodopen. Forschungshefte C 190, Geologie, Freiburger, 7—48.
- JACKO, S. 1971: Niektoré osobitosti tektonického vzťahu gemeríd s kryštalinikom Čiernej hory. Geol. Práce, Správy 57, Geol. Úst. D. Štúra, Bratislava, 111—119.
- JÄGER, E. 1973: Die alpine Orogenese im Lichte der radiometrischen Altersbestimmung. *Eclogae geol. Helv.* 66, 1, Basel, 11—21.
- JOVČEV, J. et al. 1971: Tektonski strojež na Blgarija (The explanatory notes to the Tectonic map of Bulgaria sc. 1:200 000 and 1:500 000). Technica, Sofia, 1—558.
- JUHÁSZ, A.—VASS, G. 1974: Mesozoische ophiolite im Beckenuntergrund der grossen Ungarischen Tiefebene. *Acta geol. Acad. Sci. Hung.*, 8, 3—4, 349—358.
- KAHAN, Š. 1969: Eine neue Ansicht über die geologischen Aufbau des Kristallinikums der West Tatra. *Acta geologica et geogr. Univ. Comen.* 18, Bratislava, 19—78.
- KAMENICKÝ, J. 1957: Serpentinity, diabázy a glaukofanické horniny triasu Spišsko-gemerského rudohoria. Geol. Práce, Zošit 45, Geol. Úst. D. Štúra, Bratislava.
- KAMENICKÝ, J.—KAMENICKÝ, L. 1955: Gemeridné granity a zrudnenie Spišsko-gemerského rudohoria. Geol. Práce, Zošit 41, Geol. Úst. D. Štúra, Bratislava.
- KAMENICKÝ, L. 1973: Litologische Studien und strukturelle Rekonstruktion des Kristallinikums der Zentralen Westkarpaten. *Geol. Zbor. Geologica Carpath.* 24, 2, Bratislava, 281—302.
- KANTOR, J. 1955: Diabázy juhoslovenského mezozoika. Geol. Práce, Zoš. 41, Geol. Úst. D. Štúra, Bratislava.
- KANTOR, J. 1956: Serpentinity južnej časti Spišsko-gemerského rudohoria. Geol. Práce, Zoš. 45, Geol. Úst. D. Štúra, Bratislava.
- KANTOR, J. 1957:  $A^{40}/K^{40}$  Methode zur absoluten Alterbestimmungen und das Alter des gemeriden Granites von Betliar, Zips Gömörer Erzgebrige. Geol. Práce, Správy 11, Geol. Úst. D. Štúra, Bratislava, 188—200.
- KANTOR, J. 1959 a: Das Alter gewisser Granitoide und kristalliner Schiefer der Hohen Tatra nach radioaktivem Zerfall von  $K^{40}$ . *Geol. Zbor.* 10, 1, Bratislava, 89—96.
- KANTOR, J. 1959 b: Beitrag zur geochronologie der Granite der N. Tatra. Geol. Práce, Zoš. 55, Geol. Úst. D. Štúra, Bratislava 159—169.
- KANTOR, J. 1960: The Cretaceous orogenetic processes in the light of geochronological research of the crystalline schists of the Kohút-Zone. Geol. Práce 19, Geol. Úst. D. Štúra, Bratislava, 5—26.
- KARAGIULIEVA, J.—KOSTADINOV, V.—CANKOV, C.—GOČEV, P. 1974: Structure of the Panagjuriste strip east of the Topolica siver. *Izv. Geol. Inst., s Geotektonika*, 23, Sofija, 231—301.
- KARAMATA, S. 1974: Evolution of Magmatism in Jugoslavia. In *Tectonics of the Carpathian-Balkan regions*. Geol. Úst. D. Štúra, Bratislava, 354—357.
- KARAMATA, S. et al. 1967: Les roches magmatiques crétacées-tertiaires des Carpatho-Balkanides Jougoslaves. *Acta Geol. Acad. Sci. Hung.* 2, 1—3, Budapest, 115—138.
- KARAMATA, S.—PAMIĆ, J. 1972: Some considerations about the genesis of alpine ultramafites of the Dinarides. 7. Congr. geol. SF.RJ, 2, Zagreb, 139—156.
- KLINEC, A. 1966: Zum Bau und Entstehung der Vepor-Kristallin. *Západné Karpaty* 6, Geol. Úst. D. Štúra, Bratislava, 7—28.
- KNIPPER, A. L. 1975: Okeaničeskaja kora v strukture afpskoj skladčatoj oblasti. *Nauka*, Moskva, 1—208.
- KOCKEL, F.—MOLLAT, H.—WALTHER, H. W. 1971: Geologie des Serbo-Makedonischen Massivs u. seines mesozoischen Rahmes (Nordgriechenland). *Geol. Jb.* 89, 1 Abb., Hannover, 529—551.
- KONEČNÝ, V.—SLÁVIK, J. 1974: Tectonics background of miocene volcanism in the West Carpathians: in

- Tectonics of the Carpathian Balcan regions. Geolog. Ústav D. Štúra, Bratislava, 138—144.
- KOZUR, H.—MOCK, R. 1974: Zum Alter und zur tektonischen Stellung der Meliata-Serie des Slovakischen Karstes. Geol. Zbor. Slov. Akad. Vied, 24, 2, Bratislava, 365—374.
- KRÄUTNER, H. G. 1972: Voralpidische Entwicklung und alpidischer Deckenbau in der Kristallinen Zone der nördlichen Ostkarpaten (Maramurescher Massiv). Rev. roum. Géol., Géophys., Géogr., Série, Géol. 16, 2, Bucuresti, 81—90.
- KRUGLOV, S. S.—BOJKO, A. K.—ZAJDIS, B. B. 1974: O vozmožnom prejaveniji v mezozoje kislého magmatizma vo fliševých Karpatách (po radiologičeských údajoch). Novýje údaje po geológii i neftegazonosnosti USSR (vyp. 9). Ukr. Naučno-Isledovateľskij Geol. rozv. Institut, Lvov, 54—60.
- KRIVÝ, M. 1969: Exotické valúny magmatických hornín západnej časti bradlového pásma. Acta geol. geogr. Univ. Comen. Geologica, 18, Bratislava, 165—197.
- KUBOVICS, I. 1974: Gipo-magmatické procesy sviazanne s tektoničeskimi pojasmami, v neogenových vulkaničeských rajonach severnej Vengriji. Acta geol. Acad. Sci. Hung. 18, 3—4, Budapest, 327—342.
- LAZARENKO, E. K.—LAZARENKO, Z. A.—MOLYGINA, O. A. 1973: Metalogenija Karpat. Materiály komisii mineralogii i geochemiji 2, Izd. Naukova dumka, Kijev.
- LEŠKO, B.—ĎURKOVIČ, T.—ČÍCEL, B. 1958: Spuren des paläogenen Vulkanismus in des Ostslowakei. Geol. Práce, Správy 16, Geol. Úst. D. Štúra, Bratislava.
- LOMIDZE, M. G. 1968: Pozdnejurskij vulkanizm Vostočnych Karpat. Věstnik Mosk. Univ., séria geologija 6.
- MAHEL, M. 1957: Niektoré problémy severogemeridnej synklinály. Geol. Zbor. Slov. Akad. Vied, 4, Bratislava.
- MAHEL, M. 1957: Geológia Stratenskej hornatiny. Geol. Práce, Zošit 48, Geol. Úst. D. Štúra, Bratislava, 1—193.
- MAHEL, M. 1973: Some characteristics of the development und structures of the individual alpine segments. Geol. Zbor. Geologica Carpath. 24, 1, Bratislava, 5—21.
- MAHEL, M. 1974: Introductory articles in Tectonics of the Carpathian-Balkan regions. Explanations to the Tectonic map of the Carpathian-Balkan regions and their foreland. Geol. Úst. D. Štúra, Bratislava, 9—49.
- MAHEL, M. 1974: Some remarks on the European Alpides from the point of view of global tectonics. Geol. Zborník Slov. Akad. Vied — Geologica carpath., 24, 1, (in Prese).
- MAHEL, M. 1975: Postavenie gemerika, Mineralia slov., 73, Spišská Nová Ves, 33—52.
- MAHEL, M. 1975: Predalpínske vrásnenie v Karpatoch, na Balkáne a v Dinarskom pohorí. Geol. Práce, Správy 64, Geol. Úst. D. Štúra, Bratislava, 57—80.
- MAHEL, M. 1976: Vývoj a stavba čl. Karpát vo svetle princípov globálnej tektoniky. Zborník referátov z vedeckej konferencie. Československá geológia a globálna tektonika, Smolenice, 3—22.
- MAHEL, M.—KANTOROVÁ, V. 1960: Alb krížňanskej série na severných svahoch Veľkej Fatry a Nízkych Tatier. Geol. Práce, Správy 19, Geol. Úst. D. Štúra, Bratislava, 59—63.
- MAHEL, M. et al. 1967: Regionální geologie ČSSR 2, Západní Karpaty, 1, Praha, 1—486.
- MAHEL, M. et al. 1973: Tectonic map of the Carpathian-Balkan mountain system and adjacent areas. Geol. Úst. D. Štúra, Kartografia Praha, Bratislava.
- MAHEL, M. et al. 1974: Tectonics of the Carpathian-Balkan system. Explanation of the Tectonic map of Carpathian-Balkan mountain system und adjacent Balkan mountains system and adjacent areas 1:1 000 000, Bratislava, 1—453.
- MAHMOOD, A. 1973: Petrology of the Teschenitic Rocks Series from the Type Area of Cieszyn (Teschen) in Polish Carpathians. Roczn. Pol. Tow. geol. Krakow, 15—212.
- MAIER, O.—SOLOMAN, J.—ZIMMERMANN, P.—ZIMMERMANN, V. 1975: Studium geologic si petrografic cristalulinul din portea sudiča a muntilor Poiana Rusca. Anu. Inst. Géol. si Geophys. 43, Bucuresti, 65—177.
- MARSHALCO, R.—MIŠÍK, M.—KAMENICKÝ, L. 1976: Petrographie der Flysch-Konglomerate und

- Rekonstruktion ihrer Ursprungszonen (Paläogen der Klippenzone und der angrenzenden tektonischen Einheiten der Ostslowakei). *Západné Karpaty — sér. geológia 1*; 7—124 Geol. Úst. D. Štúra, Bratislava 1976.
- MAVRUDČIJEV, B. 1965: Zákonomernosti v rozvitieto na terciernija tektono-magmatičen cikl v rast ot sredinnija Rodopski masiv. *Annuaire de l'Universite de Sofija. Fak. geol. geogr. 1 — geologie*, 58, Sofija.
- MERLIČ, B. V.—SPITKOVSKAJA, S. M. 1974: Problemy tektoniki i magmatizma glubinných razlomov. *Ľvov*, 1—173.
- METZ, K. 1966: New synthetic aspects of the tectonics of the Eastern section of the Austrian central Alps. *Tectonophysics* 3, 2, 1966, Amsterdam, 129—146.
- MINZATU, S. et al. 1975: Données géochronologiques obtenues pour des formations cristallophylliennes et pour des massifs eruptifs de Roumanie. *Dari de seama ale sedintelor 41*, 1973—1974, Bucuresti, 85—111.
- MISÍK, M. 1976: Bradlové pásmo a globálna tektonika. *Zborník referátov z konferencie čsl. geológia a globálna tektonika*. Bratislava, 28—36.
- MURATOV, M. V. 1969: Strojenje skladčatogo osnovanija sredizemnomorskogo pojasa Evropy i zapadnoj Azii i glavnejšie etapy razvitija etogo pojasa. *Geotektonika* 2, 3—21.
- MURESAN, I. 1973: Contributions a l'études des banatites et des pyroclastites de la bordure nord-est des Monts de Giľau. *International symposium on volcanism and associated metallogenesis*, Bucuresti, 133—135.
- MURESAN, M.—JONČICA, M.—TANASESCU, A. 1974: Sur la présence des metamorphides cadédoniennes dans la zone cristallins-mésozoique des Carpatés orientales (serie Damuc). *Daride seama ale sedintelor*. *Tectonics si géologie regionala 60*, Bucuresti, 47—54.
- MURESAN, M. et al. 1975: Synchronisme du métamorphisme régional protérozoique granitoides de Haghimas et du métamorphisme de la série de Bretila — Rarau (Carpatés Orientales). *Dari de seama ale sedintelor 71*, 1973—1974. *Tectonica si géologie regionala*, Bucuresti, 135—149.
- NAGY, G. 1976: Tectonics conditions in the Böršöny Mountains A. R. hung. geolog. Inst. 1973. *Budapest*, 37—40.
- NASTÁSEANU, S. 1975: General outlook on the paleozoic of the Danubian autochthonous (South Carpathians). *Anu Inst. Geol. Geophys.*, 46, Bucuresti, 191—218.
- PAMIĆ, J. 1964: Magmatske i tektonske strukture a ultramorfotima bosanske serpentianske zone. *Poseb. Izd. Geol. Glas.*, 2, Sarajevo, 1108.
- PAMIĆ, J. 1971: Some petrological features of Bosnian peridotite-gabbro complexes in the Dinaride zone of Jugoslavia. *Tchern. Miner. petrogr. Mitteil.* 15, 14—42.
- PAMIĆ, J. 1974: Middle triassic spilite-keratophyre association of the Dinarides and its position in alpine magmatic-tectonics cycle. *From Spilites and Spilitic rocks.*, Berlin-Heidelberg-New York, 161—173.
- PANTÓ, G. 1970: Recent developement in the geological recognition of the Tokaj-Szalanc Mts and Zemplin hills. *Földt. Közl.*, 96, 2, Budapest, 143—154.
- RADULESCU, D.—BORCOȘ, M. 1967: Spätsubsequenter alpinen magmatismus in Rumänien. *Acta geol. Acad. Sci. hung.*, 11, Budapest, 139—152.
- RADULESCU, D.—SANDULESCU, M. 1973: The plate-tectonics concept and the geological structure of the Carpathians. *Tectonophysics*, 16, Amsterdam, 155—161.
- RADULESCU, D.—PELTZ, S.—STANCIU, C. 1973: Neogene volcanism in the East Carpathians (Calimani-Curghiu-Hargita Mts). *Symposium volcanism and metallogenesis Bucuresti. Geol. inst.*, C 12, Bucuresti, 5—69.
- REICHWALDER, P. 1970: Niekoľko poznámok k výskytu glaukofanitických hornín v okolí Hačavy. *Geol. Práce, Geol. Úst. D. Štúra, Správy* 53, Bratislava, 157—165.
- RONIEWICZ, P.—WESTWALEWICZ, J.—MOGILSKA, E. 1974: Tuffites in the Podhole Flysch (Western Carpathians). *Bull. Acad. pol. s. Ses. de la terre*. 12, 1, Varsovie, 37—43.

- ROTH, Z. 1967 in BUDAY, T. et al. 1967: Západní úsek flyšového pásma československých Karpát. Regionální geologie 2, Západní Karpaty, 2, Praha, 169—194.
- ROZLOŽNÍK, L. 1972: Vzťah zrudnenia k tektonike v Spišsko-gemerskom rudohorí. Záverečná správa Archív geol. prieskumu, Spišská Nová Ves.
- ROZLOŽNÍK, L. 1974: The relationship of mineralization to faults of higher order in the Spišsko-gemerské rudohorie Mts. Sbor. geol. vied — geologie 26, Ústí. Úst. geol., Praha.
- ROZLOŽNÍK, L.—1976: Postavenie alpinskej metalogenézy z hľadiska globálnej tektoniky. Čsl. geológia a globálna tektonika (zborník referátov), Bratislava, 151—157.
- ROZLOŽNÍK, L.—SAMUEL, O.—JACKO, S. 1973: Prejavy eocéneho vulkanizmu pri Banskej Štiavnici. Geol. Práce, Správy 61, Geol. Úst. D. Štúra, Bratislava, 143—153.
- RUDAKOV, G. 1971: Stratigrafia domozojskich obrazovaniji Marmaroškogo massiva Vostočnych Karpat. Vestnik Mosk. Univ. Ser. Geol. 2, Moskva.
- SANDULESCU, M. 1973: Essai de reconstitution des éléments préparoxismaux alpins des Dacides (Indernides) orientales. Rev. roum. Géol. Géogr., Geophys. Sér. Géol. 17, 1, Bucuresti.
- SANDULESCU, M.—BERCIA, J. 1974: Carpathian-Crystalline-Mesozoic zone. In Tectonics of the Carpathian-Balkan regions., Bratislava, 240—253.
- SANDULESCU, M.—NASTASEANU, S.—KRÄUTNER, H. G. 1974: Roumanian Carpathians. The South Carpathians. In Tectonics of the Carpathian-Balkan regions. Geol. Úst. D. Štúra, Bratislava, 264—275.
- SAVU, H. 1967: Die mesozoischen Ophiolithe der Rumänischen Karpaten. Acta geol. Acad. Sci. hung. 11, Budapest, 59—70.
- SAVU, H.—UDRESCU, C. 1973: Geotectonic evolution of the Mures zone (Apuseni Mountains) and the distribution of trace elements in its ophiolitic rocks. International symposium on volcanism and associated metallogenesis. Bucuresti, 182—183.
- SAVU, H.—NICOLAE, I. 1974: Evolution of Ophiolitic Volcanism in the Vorta Region and its position in Mures zone Tectogenesis (Apuseni mountains). Dari de seamaale sedintelor 61, 1973—1974. Tectonics si geologie regionala. Bucuresti, 179—196.
- SAVU, H. et al. 1975: Ages K/Ar établis sur les roches du massif granitoique de Muntele Mic. Congr. 10, Asoc. geol. carp.—Balcan, Bratislava.
- SIKORA, W. 1976: On lineaments found in the Carpathians. Annales de la Soc. geol. de Pologne 46, 1, 2, Krakow, 3 — 37.
- SOROIN, M.—POPESCU, GH.—KASPER, N.—DIMITRESCU, R. 1969: Contributions préliminaires a la géochronologie des massifs cristallins des Monts Apuseni. An St. Sect. 2 b, (Geol.), 15, 25—33, (Jazi).
- SOROIN, M. et al. 1970: K/Ar dating by neutron activation of some igneous and metamorphic rocks from the southern branch of the Roumanian Carpathians. Eclogae geol. Helv. 63, 1, Bas.
- SEMENENKO, N. P. 1975: Fázy metamorfizma fundamenta Karpato-balkanskogo regiona po dannym absolutnoj geochronologii. Materialy 10. zjazdu KBGA (v tlači).
- SLÁVIK, J. 1968: Chronology and Tectonics Background of the Neogene Volcanism in Eastern Slovakia. Geol. Práce, Správy 44—45, Geol. Úst. D. Štúra, Bratislava, 199—214.
- SLÁVIK, J. 1971: Subsekventný neovulkanizmus a jeho vzťah k tektonickému vývoju; in Geologická stavba podložia zakrytých oblastí južnej časti vnútorných Západných Karpát. Západné Karpaty 15, Geol. Úst. D. Štúra, Bratislava, 92—104.
- SLAVKAY, M. 1971: Ložiská polymetalických rúd pri Ponikách. Mineralia slov., 3, 11, Bratislava, 181—213.
- STANISCHEVA,—VASSILEVA, G. 1973: Critères pétrochimiques des formations volcaniques (Ca-alkaline et K-alkaline) du Crétacé supérieur de la zone de Srednogorje orientale — Bulgarie. International symposium on volcanism and associated metallogenesis, Bucuresti, 195—197.
- STEVANOVIČ, P. et al. 1967: The structural — facial zone and the stratigraphy of the East Serbian Carpatho-Balkanids. Carpatho-Balkan geolog. Association, 8, Kongres, Beograd, 1—47.

- STILLE, H. 1953: Der geotektonische Werdegang der Karpatenbogens. Geol. Jahrb. Beihefte 8, Hannover.
- SZADECKY—KARDOSS, E. 1973: Gesteinmetamorphose und Tektonik im karpatisch-balkanisch-dinarischen Gebiet. Bulletin of the 9. th. Congress of the Carpatho-Balkan geological Association 4, Akademia Kiadó, Budapest, 445—464.
- SZADECKY—KARDOSS, E. et al. 1967: Metamorphose in Ungarn. Acta geol. Acad. Sci. Hung., 11, 1—3, Budapest, 49—58.
- SZADECKY—KARDOSS, E.—PANTÓ, G.—SZEKY, FUX, V.—KISS, J.—KUBOVICS, I. 1967: Die Vulkanite Ungarns. Acta geol. Acad. Sci. Hung., 11, Budapest, 161—180.
- SZEPESHÁZY, K. 1973: A Tiszántúl északnyugati részének felsőkréta és paleogén korú képződményei. Akadémiai Kiadó, Budapest, 1—96.
- ŠALÁT, J. 1954: Petrochémia žúl Čiernej hory, oblasť medzi Margecanmi a Košicami, Archív Geol. Úst. D. Štúra, Bratislava.
- ŠATSKIJ, N.—BOGDANOV, A.—MURATOV, M. 1964: Tectonique de l'Europe. Notice explicative pour la Carte Tectonique internationale de l'Europe au 1:2 500 000, Moskva.
- ŠČERBAK, N. P.—CAMEL, B.—BARTNICKIJ, E. N.—KAMENICKI, L. 1977: Rezultaty izotopno-geochemičeskich i mineralogičeskich issledovanij kristalličeskich porod Karpatskogo rajona vypolennyh v IGFMA AN USSR v 1973—75 godach. Geol. zborník Geologica Carpathica, 28, 2, Bratislava.
- TOLLIMANN, A. 1967: Tektonische Karte der Nördlichen Kalkalpen, 1. Teil. Der Ostabschnitt. Mitt. Geol. Geo. Wien 59, 238—253.
- VARGA, I. 1971: Prejavy hercýnských orogénnych fáz vo vývoji permu gemeríd. Geol. Práce, Správy 57, Geol. Úst. D. Štúra, Bratislava, 349—360.
- VARGA, G. et al. 1975: Geology of the Matra mountains. Annales instituti geologici publici Hungarici 57, 1, Budapest, 1—480.
- VASS, D. 1976: Molassové panvy a globálno-tektonický model Karpát. Zborník ref. z konf. Čsl. geológia a globálna tektonika, Bratislava, 111—117.
- WEIN, G. 1969: Tectonic review of the Neogene-covered areas of Hungary. Acta geol. Acad. Sci. Hung., 13, 399—436.
- WELCHEV, T. T. et al. 1976: On the deep structure and the young igneous activity in the southeast Bulgaria. Annuaire de l'Univ. de Sofia, Fac. géol. et geogr., vol. 67, t. 1, Géologie.
- WIESER, T. 1958: Magmowe i metamorficzne skaly egzotyczne z kredy i paleogenu pieninskiego pasa skalkowego. Z Badán geol. v Karpatach. T 3, Inst. geol. Buletyn 13, 97—136.
- VOZÁR, J. 1971: Der mehrphasige Charakter des Perm-Vulkanismus der Choč-Einheit in der Niederen Tatra. Geol. Práce, Správy 55, Geol. Úst. D. Štúra, Bratislava, 131—137.
- ZAGORČEV, I. 1975: Phénomènes tectoniques paléozoïques superposés dans les socle précambrien de certaines parties de la région Thraco-Anatoliene. Geologica Balcanica 5, 4, Sofia, 15—28.
- ZELEŇKA, T. 1974: Istorija megatektoničeskogo i magmatogeologičeskogo rozvitija severo-vostočnoj Matry. Acta geol. Sci. hung. T 18, 3—4, Budapest, 377—385.
- ZOUBEK, V. 1931: Caractéristique de quelques roches cristallophylliennes et éruptives des galets exotiques des conglomérats sénéoniens et paléogènes des Carpathes Occidentales. Knih. Stát. geol. Úst., 13, A.
- ZOUBEK, V. 1935: La tectonique de la vallée supérieure du Hron et sa relation avec la distribution des sources minérales. Věst. Stát. geol. Úst. ČSR 2, Praha, 85—115.
- ZOUBEK, V. et al. 1961: Tectonic development of Czechoslovakia. Praha. 1—224.
- ZWART, H. J. 1976: Regional metamorphism in the Variscan orogeny of Europe. Franz Kossmat Symposium. Nova Acta Leopoldina, 224, 45, Halle, 361—367.



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## Geotektonické postavenie magmatitov v Karpatoch, na Balkáne a v dinaridách

Resumé anglického textu

### Charakteristika geotektonických typov magmatitov

Sledovanie štruktúrneho, a hlavne geotektonického postavenia magmatitov v jednotlivých vývinových cykloch alpíd nás viedlo k ich rozdeleniu do siedmich skupín (obr. 1,2):

- granitoidy — úzko späté s procesmi vrásnenia a s utváraním štruktúrnych plánov,
- neohercýnske neskorogeosynklinálne vulkanity — geneticky priamo nadväzujúce na hercýnske granitoidy, sprevádzajúce molasové výplne brázd a prepadlín,
- predalpínske eugeosynklinálne vulkanity — v asociácii s predkarbónskymi sériami alpíd,
  - mezozoické ofiolity a bázické horniny viazané na úzke, osobitné, štruktúrne pásma,
  - mezoalpínske magmatity intraorogénneho typu vulkano-plutonickej formácie banatitov v sprievode tektonických prepadlín vyplnených zväčša flyšom, čiastočne i včasnými molasami,
  - neskorogeosynklinálne neovulkanity — geneticky späté s treťohornými kotlinami, ktoré sú vyplnené molasami,
  - finálne vulkanity — späté s poorogénnym štádiom dotvárania morfoštruktúry alpíd.

#### 1. Granitoidy

Väčšina granitoidných telies vystupuje uprostred najstarších, zväčša vyššie metamorfovaných komplexov, zaradovaných medzi predpaleozoické útvary. Preto prevažovala mienka, že sa vznik väčšiny granitoidov spája predovšetkým so starším predbajkalským a bajkalským vrásnením. V posledných rokoch sa rádiometrickým štúdiom získali hodnoty ukazujúce, že staré granitoidy sú tu síce prítomné, ale ich podstatná časť geneticky súvisí s hercýnskym vrásnením. Väčšina granitoidov sa vekove radí do karbónu aj preto, že sa metamorfované komplexy, v rámci ktorých vystupujú, zaradili vďaka biostratigrafickým metódam (najmä palinologickým) do staršieho paleozoika.

Predhercýnske granitoidy, viazané geneticky na bajkalské a staršie vrásnenia, sú zväčša synkinematické, úzko súvisia s migmatizáciou, anatexiou a s výraznou regionálnou metamorfózou (M. MAHEĽ 1974). Sú to, napr. séria Grobgneiss vo Východných Alpách, žuloruly a migmatity v Iubietovskom a kohútskom pásme veporíd, niektoré migmatity v kryštaliniku

Nízkyh Tatier a Čiernej hory, migmatity a anatektity v pohorí Moragy v južnom Maďarsku (rádiometrické údaje okolo 505 mil. r.); ďalej žuloruly a migmatity v príkrove Codru a v príkrovovom systéme Biharia v Apusenách a v Južných Apusenách sprevádzajúce sériu Baia de Aries (s rádiometrickými hodnotami okolo 508 mil. r.); žuloruly a okaté ruly v sérii Bretila Barnau vo Východných Karpatoch. Početnejšie synkinematické granitoidy sú v Južných Karpatoch, kde tvoria aj väčšie plutóny (napr. Poneasca a Sichevita v getiku, plutóny Súsita a Tismana v danubiku — rádiometrické hodnoty 556—520 mil. r. — obr. 5). Väčšie telesá predhercýnskych synkinematických granitoidov sú aj v Srbsko-macedónskom masíve (rádiometrické údaje okolo 450 mil. r.), sú však zväčša hercýnsky remobilizované; ďalej sú v Pelagónskom masíve (dokonca s hodnotami 1140, 838, 609 — 577 mil. r.; G. Gedeon 1968). Pre rodopský masív sú zvlášť charakteristické synkinematické lessovské granitoidy vystupujúce uprostred ultrametamorfítov, považovaných za archaické (I. BOJANOV—D. KOŽUCHAROV 1966; obr. 6).

Najpočetnejšie a čo do vývoja a stavby alpíd najvýznamnejšie sú hercýnske granitoidy. Príznačná je pre ne štruktúrno-tektonická pestrosť, široké časové rozpätie a nevelká rôznorodosť obsahu.

Synkinematické hercýnske granitoidy, úzko späté s plášťom, predstavujú obvykle najhlbšie pásma hercýnskej granitizácie (napr. južné časti Ľumbierskeho kryštalinika Západných Karpát (obr. 3). Charakteristické sú i pre niektoré zóny s osobitým tektonickým postavením, napr. klisúrsky a stakevský masív v Starej Planine na Balkáne — súčasť zóny granitizácie pozdĺž rozsiahleho prehybu severne od rodopského masívu (včítane Srednogorja; E. BONČEV 1975). Hercýnske granitoidy tvoria veľké telesá batolitov a fakolitov, a to nielen vo vlastných alpídach, ale i v priľahlých masívoch (srbsko-macedónskom, rodopskom), ktoré štruktúrne spájajú s vyvrásneným systémom.

Rozsiahlejšie telesá batolitov podmieňujú vytváranie širokých antiklinál a synklinál, napr. vo Vysokých Tatrách (A. GOREK 1967, Š. KAHAN 1968), alebo aj formovanie dómovitých štruktúr, typických najmä pre rodopský masív (I. BOJANOV 1973, I. ZAGORČEV 1975), ale i pre srbsko-macedónsky masív (M. D. DIMITRIEVIČ 1967).

Najmladšie sú zväčša permské intrúzie malých granitoidov, ktoré majú prevažne charakter pňov (štokov). Častejšie sa vyskytujú na Balkáne a v rodopskom masíve, ale nájdu sa aj v iných oblastiach, napr. v Iubietovskej zóne Západných Karpát nadväzujúcej geneticky tesne na neohercýnske vulkanity (kremenné porfýry; V. ZOUBEK 1931).

Pozoruhodné je široké časové rozpätie granitotvorných procesov; hercýnske granitoidy vykazujú hodnoty od 340, najčastejšie 320 až 220 (200) mil. r. (J. BURCHARD 1970, J. KANTOR, 1959, B. CAMBEL 1976, M. DIVILJAN—S. DIVILJAN 1967, M. D. DIMITRIEVIČ 1967, S. BOYADJIEV 1974, 1967). Tieto procesy prebiehajú približne v tom istom období ako účinky hercýnskeho vrásnenia počnúc jeho prvými prejavmi koncom devónu (bretónskymi) až po koniec permu (falckými). Toto obdobie hercýnskeho diastrofizmu, celkove asi 120 mil. r., je veľmi dlhé v porovnaní s etapou prevažne pokojnejšej paleozoickej eugeosynklinálnej sedimentácie (ordovik resp. vrchné kambrium — koniec devónu). Pritom nemožno opomenúť náznaky kaledónskeho diastrofizmu, azda so sprievodnou granitizáciou (E. JÄGER 1973, I. HAJDUTOV 1971) — časté rádiometrické údaje okolo 400 mil. r. značia najskôr „termálnu udalosť“ (H. ZWART 1976).

Hercýnske vrásnenie i súčasná granitizácia majú 3 výrazné štádiá: bretónsko-sudetské, astúrsko-saalske a falcké štádium (M. MAHEL 1975). Prvé štádium sa často označuje ako

rané (rádiometrické údaje o granitoidoch 340—305 mil. r. — S. BOYADJIEV 1974), ďalšie dve sú označované ako neskoré štádiá (E. DIMITROVA et al. 1975 — rádiometrické údaje o granitoidoch okolo 280—250 mil. r.). Migmatity a synkinematické granitoidy sa zvyknú spájať s prvým štádiom, neskorokinematické granitoidy s druhým a vznik aplítov, pegmatitov, aplitických granitov a subalkalických granitov sa obvykle spája s tretím štádiom (A. K. BOJKO 1975). Naznačená tendencia vývoja hercýnskych granitoidov je bežná najmä v Západných Karpatoch a na Balkáne; je v zhode s postupnou stabilizáciou kôry počas hercýnskeho diastrofizmu. Nemožno však hovoriť o všeobecne platnom pravidle. Niektoré synkinematické granitoidy s rádiometrickými hodnotami permu okolo 270—230 mil. r. sotva možno považovať len za výnimky v rámci pravidla (napr. žuloruly v Južných Karpatoch v masíve Nerešnički, alebo migmatity, napr. z Vysokých Tatier s hodnotami 250 mil. r., a naopak niektoré pegmatity z Nízkych Tatier v Západných Karpatoch s hodnotami 330 mil. r.). I keď zaznamenávame iba všeobecnú tendenciu vývoja granitoidov, nezávisí ich typ len od vývojového štádia, ale aj od hĺbky ich vzniku. Všeobecná tendencia vývoja granitoidov od synkinematických k apliticko-pegmatitickým žilám má výrazné odklony vo vývoji jednotlivých areálov a zón. Je známe, že rozmanitosť vývoja, stavby i obsahu jednotlivých zón je príznačná najmä pre alpidy.

Častejšie sa objavujú názory nielen o polyfázovosti, ale aj o polycykličnosti vývoja granitoidných telies (M. D. DIMITRIEVIĆ 1967, S. KARAMATA 1974, I. ZAGORČEV 1975, I. DIMITRESCU 1976, E. SZÁDECKY—KARDOSS 1967, S. BOYADJIEV 1967). Ide hlavne o veľké telesá južnobulharských granitoidov (v Rodopách a Srednogorji), granitoidov v srbsko-macedónskom a v pelagónskom masíve, sčasti aj v Južných Karpatoch; rádiometrické údaje tu poukazujú na predhercýnsky, častejšie na hercýnsky, ale aj na alpínsky vek (napr. v plutóne Bujanovac v Srbsko-macedónskom masíve, či v belogradčickom plutóne v balkanidách, v Pelagónskom masíve, v magmatitoch pohoria Mecsek atď.).

Hercýnske granitoidy predstavuje predovšetkým formácia granodioritovo-granitová. Bohatší látkový obsah a viac formácií granitoidov majú balkanidy, t. j. oblasť geosynklinálne členitejšia s lineamentnými geosynklinálami (E. BONČEV 1966). Tu je popri granodioritovo-granitovej formácii vyvinutá aj formácia gabroplagiogranitová, gabrodioritová-granodioritová a alkalická gabrosyenitová formácia (E. DIMITROVA et al. 1975). Všade sú hojné aplity a pegmatity, ale i menšie telesá najmä permských leukokrátnych granitov.

Rozdiely v stupni metamorfózy a granitizácie v jednotlivých úsekoch alpíd, ale aj v jednotlivých zónach toho istého segmentu, sa prejavujú i odlišnosťou tektonických štýlov. Uvedieme príklady z hercýnskych štruktúr. Vo Vysokých Tatrách, kde sú rozsiahle telesá granitoidov, boli preukázané hercýnske štruktúry — antiklinály a synklinály širokého rozpätia (A. GOREK 1959, Š. KAHAN 1969). Dómovité štruktúry sú zvlášť typické pre oblasti s veľkými batolitmi granitoidov, napr. Rodopský masív (I. BOJANOV 1974, I. ZAGORČEV 1975) a Srbsko-macedónsky masív (M. D. DIMITRIEVIĆ 1967). V krakovskom pásme veporíd a v Čiernej hore na Slovensku (S. JACKO 1971) boli v svorových zónach s malým podielom granitoidov, resp. bez granitoidov preukázané úzke stlačené hercýnske vrásy. Z Východných Karpát sa uvádzajú aj hercýnske príkrovy s nepatrným podielom hercýnskych granitoidov (H. G. KRÄUTNER 1972).

Podiel granitoidov v alpínskom cykle je nevelký (obr. 2). Týka sa to hlavne granitoidov súvisiacich s vrásnením, a teda aj so stabilizáciou kôry. Sú zastúpené len menšími telesami, ktoré sa viažu na zóny predalpínskeho kryštalinika, sú silne prepracované paleoalpínskymi,

niektoré až mezoalpínskymi vrásneniami. Telesá alpínskych synkinematických granitoidov sú rozšírené hlavne v južnejších častiach Srbsko-macedónskeho masívu, v oblastiach už silne postihnutých kimerským vrásnením. Pritom sa v alpínskom cykle, pri východných okrajoch Srbsko-macedónskeho masívu, granitizácia spolu s vysokotlakovým metamorfizmom prejavila viacfázovo, v časovom rozpätí 170—40 mil. r. (E. KOCKEL—H. MOLLAT—H. J. WALTHER 1972), t. j. od začiatku strednojurského „mladokimerského“ cez strednokriedové až po vrchnoecénne vrásnenia (J. AUBOUIN 1964, obr. 9). V penniniku Álp siaha granitizácia až do začiatkov oligocénu a i tu sa spája s vysokotermálnou metamorfózou. Je teda zrejme, že i alpínske granitotvorné procesy majú široké časové rozpätie, ak ich porovnáme s pokojným geosynklinálnym obdobím triasu a spodnej až strednej jury (len 80—100 mil. r.). Pri alpínskej granitizácii treba zdôrazniť, že sa prejavuje skromne a viaže sa iba na úzke zóny. Rozsiahlejší areál postihnutý týmito procesmi predstavujú južné oblasti alpíd, hlavne pelagónskeho masívu; južné oblasti srbsko-macedónskeho masívu, rodopského masívu a bloky sakarský a strandžanský. V týchto oblastiach sú u starších granitoidov časté rádiometrické hodnoty okolo 160—97 mil. r. Ide zrejme o alpínsku regeneráciu kryštallického podkladu. Pravda, v niektorých granitoidoch sa zistili aj hodnoty 50—30 mil. r., čo je späté s magmatickou priabónsko-oligocénovou, miestami až spodnomiocénovou aktivizáciou (S. BOYADJIEV 1974).

V Západných Karpatoch (obr. 10) sú výrazným predstaviteľom paleoalpínskych granitoidov telesá granitov a granitporfýrov uprostred kambrosilúrskych gelnickej episérie v Spišsko-gemerskom rudohorí (rádiometrický vek (K/Ar) 93—103 mil. r. — J. KANTOR 1960). Granitotvorný proces bol tu azda viacfázový a vyvrcholil malými intrúziami. Výstup žulových telies i priebeh zón kontaktnej metamorfózy je pásmový, súdežný s priebehom hlavných paleoalpínskych štruktúr (P. GRECULA 1973). Malé telesá sa viažu na oslabené poloblúkované zóny pri okrajoch megaantiklinória Volovca (L. ROZLOŽNÍK 1972). V gemeridách je závažnou otázkou vzťah granitových telies k príkrovu prvého rádu, k násunu gemeríd na veporidy. Skutočnosť, že sa preukázali šmyky s násunom kryh bez kontaktnej metamorfózy na kryhy rovnakého zloženia, ale kontaktne metamorfované (napr. v prešmyku Jedlovec), vedie k názoru o príkrovej pozícii granitových telies i geneticky s nimi zviazaných rudných ložísk (P. GRECULA 1973). Výraznejšie prejavy granitizácie plášťa s novovzniknutými minerálmi, ako je „dymový“ kremeň, živce, biotit, muskovit, ktoré boli zistené hlbokými vrtmi (P. GRECULA 1973), naznačujú podľa môjho názoru skôr „zakorenenosť“ žulových telies. Pri riešení tohto problému poslúžia menšie masívy hlavne biotitických granitov, ktoré vystupujú na styku hercýnskych granitoidov masívu Bujanovej a cez metamorfný, alpínsky silne poprerašaný plášť v Čiernej hore, severne od gemeríd. Tieto masívky granitoidov sú často poprerašané systémom aplitových žíl a hojných trhlín s výplňou turmalínu, hematitu, molybdénu, kremeňa, pyritu a kalcitu. Niektoré zo spomínaných telies prerážajú cez oslabené zóny permu veporidného obalu. Na Spálenom vrchu teleso dioritu preráža i presunutú kryhu gemeridného karbónu a permu (J. ŠALÁT 1954). Pozícia tohto telesa a len slabý tektonický zásah do najmladšieho systému dislokácií sv.—jz. smeru (S. JACKO 1971) nasvedčujú, rovnako ako u gemeridných granitov, že ich intrúzia časove spadá i do neskorších procesov paleoalpínskeho vrásnenia.

Popri veľkých hercýnskych telesách granitoidov v kryštaliniku veporíd, ktoré je silne alpínsky zvrásnené, vystupujú na význačných tektonických líniách granitoidy s rádiometrickými hodnotami 107—115 mil. r. (J. KANTOR 1960), napr. Muránska Zdychava 107 mil. r.,

Hrončok 110—115 mil. r. (A. K. BOJKO 1975), ktoré patria k starším prejavom paleoalpínskeho vrásnenia. Sú tam však i telesá albitických granitov s hodnotami okolo 86—90 mil. r. Alpínske granity sú hojnejšie pri južnom okraji veporidného kryštalinika v zónach zvlášť intenzívne zasiahnutých alpínskym vrásnením. Taký vek vykazujú i malé žily a masívky aplítov a pegmatitov sv.—jz. smeru, ktoré prerážajú prevažne svorovú hronskú sériu v krakovskom pásme veporíd (A. KLINEC 1966, obr. 10).

Rádiometrické hodnoty 93 až 140 mil. r. získané z granitov valúnového materiálu v paleogénnych pročských zlepencoch bradlového pásma východného Slovenska (R. MARS-CHALCO et al. 1976) naznačujú prítomnosť paleoalpínskych granitoidov i v kryštalickej podklade severnejších jednotiek Západných Karpát. Valúny spomínaných granitov pochádzajú pravdepodobne z kordiliery uprostred geosynklinály bradlového pásma, alebo pri jej severnom okraji.

Granitoidné horniny, známe pod názvom tonality, sprevádzajú periadriatickú zlomovú zónu. Rádiometrické údaje z telesa tonalitov z Karavaniak vykazujú hodnoty 29 +6 mil. r., resp. 28 +4 mil. r. (z biotitu). Zaujímavé je však, že sprievodné granodioritové porfýry z oblasti Eizenkappel sú zrejme hercýnske s hodnotami  $24 \pm 8$  mil. r. (z pyroxénov) a diority s hodnotami  $227 \pm 7$  mil. r. (z biotitu) (E. FANINGER 1976).

Dôležitejšiu úlohu v stavbe alpíd majú telesá granitoidov, tvoriace súčasť vulkanoplutonických formácií mezoalpínskeho banatitového typu (bližšie viď na str. 139). Ich vznik súvisí so stavom kôry, striedavo ovplyvňovaným kompresnými pohybmi laramskými a ilýrskymi, resp. pyrenejskými (obr. 6).

Vo vzťahoch medzi vrásnením a granitizáciou je nápadný rozdiel medzi hercýnskym a alpínskym tektonomagmatogénnym a metamorfným cyklom. Zatiaľ čo je pre hercýnsky diafektorizmus príznačná hlbinná metamorfóza a rozsiahla granitizácia, charakteristickým znakom alpínskeho vrásnenia je obmedzený rozsah metamorfózy, ktorá nastala v plytších zónach a prejavila sa ako progresívna metamorfóza mladšieho paleozoika a mezozoika a ako lokálna diafektoréza kryštalinika. Alpínske vrásnenie sprevádzajú aj drobné intrúzie granitov. Celkove má však pripovrchový charakter s prevahou príkrovov (M. MAHEL 1974).

## 2. Mladohercýnske neskorogeosynklinálne vulkanity

Neskorogeosynklinálne mladohercýnske vulkanity, prevažne permské, sčasti už i vrchnokarbónske (balkanidy, Južné Karpaty) sprevádzajú komplexy vrchnej, lagunárne suchozemskej molasy, uloženej v pozdĺžnych žlaboch. Vznik žlabov, tektonických prepadlín i vulkanitov je dôsledkom hlbokých zlomov, sprevádzajúcich epiorogénne rozťahnutie, niekde i riftogénu završujúcu geosynklinálny proces.

Neohercýnske vulkanity majú pomerne jednotvárny obsah, najčastejšie sú kremenné porfýry; vertikálne i laterálne ich miestami zastupujú melafýry. Medzi obidvoma krajnými typmi týchto vulkanitov sú síce prechody, no i tak je podiel intermediárnych vulkanitov, porfýritov a kremenných porfýritov nápadne malý. Vzťahy medzi melafýrmi a kremennými porfýrmi sú v niektorých segmentoch rozdielne. Na Balkáne možno pozorovať prechod od bázičejších melafýrov ku kyslejšim (P. ČUNEV—P. BONEV 1975); v Apusenách je to skôr naopak (M. BLEACHU 1974). V severnejších zónach výrazne prevažujú kremenné porfýry, v južnejších zónach výrazne melafýry. V Západných Karpatoch sú melafýry charakteristické

pre chočský žľab, rozložený medzi zónami s kremennými porfýrmi, severnejšou veporidnou a južnejšou severogemeridnou. Geotektonické postavenie melafýrov a kremenných porfýrov je v podstate rovnaké. Preto je logické i ich zaradenie do jednej bazaltovo-ryolitovej formácie na Balkáne (E. DIMITROVA et al, 1975) ako i v Západných Karpatoch (D. HOVORKA 1965). Lineárny charakter erupcií bázického vulkanizmu v niektorých zónach spôsobuje tenšia kôra, ale i hlbšie siahajúce zlomy, ktoré slúžili ako prívodné kanály výlevov. V nijakom prípade nemožno melafýry považovať za vulkanity, analogické s ofiolitmi, teda za iniciálne vulkanity v ponímaní H. STILLEHO (1953)—(J. VOZÁR 1971). Melafýry totiž len málo ovplyvňujú sedimentáciu a nesúvisia s prehĺbením geosynklinály ako ofiolity, resp. ofiolitoidy, ale s molasovým štádiom prechodu od hercýnskej geosynklinály k alpínskej. Slabý alkalický trend melafýrov v chočskom príkrove Západných Karpát (J. VOZÁR 1972) síce dáva podnet k úvahám o slabšej kratonizácii kôry, melafýry v Západných Karpatoch i v Apusenách sú však viazané na grabeny, ktoré znamenajú začiatky alpínskych štruktúrno-faciálnych zón.

Lenže ani zaradenie melafýrov ako finálnych článkov hercýnskeho vulkanického sledu (D. HOVORKA 1976) nie je presvedčivé, a to preto, že melafýry sa spájajú najmä s južnejšími zónami alpíd, menej stabilizovanými hercýnskym vrásnením.

Prevažne kyslý charakter neohercýnských vulkanitov a ich časová zhoda s poslednými štádiami granitoidných intruzií, ako aj prejavy doznievajúceho hercýnskeho metamorfizmu, naznačujú úzky vzťah neskorogeosynklinálnych vulkanitov ku granitotvorným procesom. Niet však priamych dôkazov, že by išlo o vulkanicko-plutonickú formáciu, i keď subvulkanické telesá povahy mikrogranitov sú známe napr. z balkaníd. Vulkanity sú rozložené v okrajových tektonických prepadielinách spravidla vedľa zón s granitoidnými intruzivami; to naznačuje skôr nepriamu väzbu na granitizáciu. Ide azda o výlevy magmy v okrajových častiach väčších tumorov a možno ich považovať za oneskorené povrchové prejavy rozsiahlej hercýnskej sializácie kôry. Slabšie metamorfovaný perm a hojný žilný kremeň v tomto útvare znamenajú asi dozvuky tepelných prejavov. Molasové sedimentárne formácie, doznievajúci vulkanizmus a metamorfizmus svedčia o završovaní hercýnskeho cyklu.

Mladohercýnske neskorogeosynklinálne vulkanity sú hojnejšie v brázdach, oddeľujúcich pásma výraznejšej hercýnskej granitizácie, a pri vnútorných okrajoch hercýnsky stabilizovaných areálov (obr. 8). V týchto priestoroch došlo vďaka hlbokosiahajúcim zlomom aj k subakválnym výlevom melafýrov, zväčša sillov, v sprievode dajok melafýr-porfyritov a porfyritov (napr. melafýrová séria chočského príkrova Západných Karpát a perm v jednotke Codru v Apusenách). Vrchnopaleozoické depresie severnejších oblastí alpíd majú molasové výplne, s malým podielom vulkanitov, kremenných porfýrov, ich tufov a tufitov.

### 3. Predalpínske eugeosynklinálne vulkanity

Väčšina predkarbónskych sérií v alpídach predstavuje série vulkanosedimentárne s geosynklinálnymi prevažne bázickými vulkanitmi. Tieto sú zväčša metamorfované na amfibolity, v najstarších sériách niektorých oblastí (Rodopy, Srbsko-macedónsky, Pelagónsky masív) až na eklogity. Prevládajú bázické vulkanity spilitovo-diabázovej a spilitovo-keratofýrovej formácie. Často ich sprevádzajú intrúzie gabier a dioritov, ale aj serpentinizovaných peridotitov, pyroxenitov, dunitov. Kyslé efuzíva sú obyčajne len sprievodných javom, prípadne charakterizujú jednotlivé formácie. Ide zväčša o kremenné porfýry a ich pyroklastická, menej o kremenné keratofýry.

Bázické vulkanity sú zvlášť hojné v sériách proterozoika, napr. vo vlasínskej a lisecko-oso-govskej sérii na okraji rodopského masívu a v srbsko-macedónskom masíve (M. DIMITRIEVIČ 1967), v sérii Biharia v prikrivových jednotkách Apusén (R. DIMITRESCU 1966), v proterozoických sériách Južných Karpát, v sérii Poiana Rusca, Ielová, Lotru, Sebes v Južných Karpatoch (O. MAYER et al. 1975 — obr. 4, 5).

Aj celý rad staropaleozoických sérií má vysoký podiel diabázov, napr. diabázovo-fylitová séria balkaníd, devónska fylitovo-diabázová séria v gemeridnom paleozoiku Západných Karpát, séria Repedea vo Východných Karpatoch, devónsko-spodnokarbónska séria Paiuse-ni v prikrivových jednotkách Apusén, série Corbu, Leșcovița v Južných Karpatoch (M. SANDULESCU—S. NATASEANU—H. G. KRÄUTNER 1974). Vo väčšine sérií ide o vulkanicko-plutonickú formáciu, v niektorých prípadoch dokonca s väčšími telesami intruzív. Tieto vytvárajú najčastejšie žily, prípadne menšie pne. Rôznorodosť intruzívnych telies je v niektorých oblastiach zvlášť veľká. Týka sa to predovšetkým strumskej dioritovej formácie na Balkáne, ktorú tvoria gabrá, diority, ale aj plagiogranity (J. JOVČEV et al. 1971). Pravda, prítomnosť granotoidných hornín sa vo viacerých prípadoch pripisuje skôr neskorším granitoidným procesom (napr. vo fylitovo-diabázovej sérii Západných Karpát). V mnohých prípadoch sú však genetické vzťahy granitoidov k bázickým horninám nejasné.

Súčasťou vulkanicko-plutonickkej formácie sú vo väčšine prípadov i ultrabázické telesá. Ich magmatická príbuznosť s bázickými horninami je zjavná napr. v Južných Karpatoch v predpaleozoických sériách v Poiana Rusca, v getiku v sérii Jelova a v masíve Mrakonia v danubiku (O. MAYER et al. 1975), ale aj v paleozoických sériách napr. v sérii Corbu v danubiku. Genetický vzťah medzi väčšími ultrabázickými a bázickými telesami sa zistil v predpaleozoických komplexoch Rodop (J. JOVČEV et al. 1971). Spätosť menších telies serpentinizovaných pyroxenitov s gabrami a gabrodioritmi je zjavná v balkanidách (I. HAJDUTOV 1971). V priestoroch nahustenia amfibolitov (napr. v západokarpatskom paleozoiku hronskej série, v kraklovskej zóne veporíd) je logickejšie vysvetľovať prítomnosť ložných ultrabázických telies ako syngenetických. Sú súčasťou vulkanicko-plutonickkej formácie a ich výstupy možno spájať s protrúziami pozdĺž zlomov. Spilitovo-diabázová formácia veporíd i tatrid je chemickým zložením blízka toleitickým, na hliník bohatým bazaltom (D. HOVORKA 1976); ide pravdepodobne o oceanický typ kôry. V niektorých prípadoch, napr. pri južnom okraji veporidného kryštalinika, rozloženie ultrabázických telies nevyklučuje v kohútskej zóne ani túto druhú možnosť (D. HOVORKA 1976; obr. 10).

Zaujímavý je výskyt bázických, ba v niektorých prípadoch dokonca aj ultrabázických telies v spodnom karbóne (napr. séria Paiuseni v južnejších jednotkách Apusén, séria Tibau so značným podielom vo Východných Karpatoch, spodnokarbónsky až strednokarbónsky flyš dinaríd), ba i v molasoidných sériách namur-westfalu gemeridného paleozoika Západných Karpát, v strednokarbónskej sérii Argestu vo Východných Karpatoch (M. SANDULESCU et al. 1974). Vo všetkých spomenutých oblastiach alpíd sa začína hercýnske vrásnenie, granitizácia, metamorfóza a s tým spojené vytváranie rulovo-žulového sloja už začiatkom karbónu, ba miestami už koncom devónu. Zrejme však v tých štruktúrno-tektonických zónach, v ktorých sú karbónske bázické prípadne i ultrabázické telesá, bola kôra tenšia. Ide zväčša o zóny slabšie zasiahnuté hercýnskym vrásnením. Prítomnosť bázických vulkanitov v karbóne možno považovať za dôkaz výraznejšej diferenciacie pri utváraní žulovo-rulového sloja počas hercýnskeho vrásnenia.

Bázické vulkanity v celom rade predkarbónskych sérií asociujú aj s kyslými vulkanitmi.



Tieto v niektorých sériách prevládajú a sú pre ne charakteristické. V predpaleozoických komplexoch sa pre rozsiahlejšiu anatexiu a vyššiu metamorfózu ťažko stanovuje ich podiel na tvorbe geosynklinálnych sérií. Za také sa často považujú leptitové ruly, napr. v Rodopách (I. BOJANOV—D. KOŽUCHAROV 1968), v Srbsko-macedónskom masíve (M. DIMITRIEVIČ 1967), v kohútskej zóne západokarpatského kryštalinika (L. KAMENICKÝ 1974). Zaujímavá je v tejto súvislosti metakeratofýrová a kremito-porfýrová episéria Arada, vrchnoproterozoického — spodnokambrického veku. Častejšie sú kremenné porfýry a ich tufy, zmenené na porfýroidy v staropaleozoických flyšoidných až flyšových episériách, ako je gelnická séria v gemeridách Západných Karpát a séria Tulges vo Východných Karpatoch. V južnejších príkrovových jednotkách Apusén má séria Muncel (kambrium) častejšie intermediárne a kyslé vulkanity, kremenné porfýry a kremenné porfýry. I v proterozoických komplexoch Južných Karpát sa zistili kyslé vulkanity poväčšine ako polohy v sprievode bázických vulkanitov; sú známe aj v rifejsko-kambrijských sériách Srbsko-macedónskeho masívu. Hrubé polohy kremenných porfýrov, menej kremenných keratofýrov, zodpovedajú hlavne ordoviku. Vo Východných Alpách sa kremenné porfýry kladú na rozhranie ordoviku — silúru (H. V. FLÜGEL 1975), obdobne ako v igalsko-bükkskej synklinále v Maďarsku. Častejšie sa teda spája hojnejší výskyt kremenných porfýrov s účinkami kaledónskeho vrásnenia. Treba však zdôrazniť, že sa združujú s geosynklinálnymi formáciami s prevahou ílovcov a pieskovcov, často diastrofického charakteru, teda s formáciami výrazne geosynklinálnymi. Zvýrazňujú to i častejšie polohy bázických telies, prípadne laterálne a vertikálne striedanie kyslých a bázických vulkanitov.

Väčší podiel kyslých vulkanitov vo vrchnejších polohách proterozoicko-kambrijských sérií, príp. v spodnejších polohách ordovicko-silúrskeho vrásnenia, možno logicky vysvetliť ako dôsledok rôznej hrúbky kôry, čo spôsobilo bajkalské, najčastejšie neskorobajkalské, príp. ranokaledónske vrásnenie. Tu treba zdôrazniť, že hoci má kôra rozličnú hrúbku, obsah sérií ani ich geosynklinálny charakter sa podstatne nemenia. Kyslé vulkanity geneticky súvisia s ostrovnými oblúkmi, resp. s oblasťami hrubšej prechodnej subkontinentálnej kôry.

Kremenné porfýry často asociujú aj s devónsko-spodnokarbónskymi sériami. Napr. vo vrchných polohách hronskej série (devón) sa kyslé efuzívne telesá striedajú s bázickými (A. KLINEC 1966). Vo Východných Karpatoch má slabometamorfovaná séria Tibau uprostred série vápencov a dolomitov, fylitov a zlepencov nielen bázické telesá, ale i kyslé metatufy; v severnejších zónach je to devónsko-spodnokarbónska séria Repedeá (H. G. KRÄUTNER 1972). V Apusenách devónsko-spodnokarbónska séria Paiuseni má okrem bázických telies i kyslejšie kremenné dioritové porfýry, keratofýry, ale i kremenné porfýry a ich tufy; aj v tejto sérii sa popri fylitoch a kremencoch vyskytujú aj zlepence (R. DIMITRESCU 1966). V Južných Karpatoch, v danubiku, sú tufy kremenných keratofýrov a kremenných porfýrov častejšie v spodnokarbónskych pieskovcovo-bridličnatých sériách. Rovnako i v balkanidách obsahuje vrchnodevónsko-spodnokarbónsky pieskovcovo-bridličnatý komplex pyroklastické kremenné keratofýry a kremenné porfýry. Uvedené príklady naznačujú, že i vo vrchnom devóne a spodnom karbóne sprevádzajú kyslé vulkanity bázické vulkanity. Dá sa to logicky vysvetliť ako dôsledok rozličnej hrúbky kôry. Rozdiely v hrúbke kôry boli zrejme veľmi výrazné i koncom staropaleozoického sedimentačného cyklu a súvisia s prvými prejavmi nastupujúceho hercýnskeho vrásnenia.

Rovnako ako predvrchnokarbónske sedimentárne komplexy aj vulkanogénne formácie sú pomerne málo diferencované a najčastejšie ich sprevádzajú intruzívne telesá. Sú to zrejme

produkty nevzretej kôry, včasných geosynklinál. Náznaky výraznejšej diferenciácie koncom proterozoika, prípadne začiatkom staršieho paleozoika sú dôsledkom bajkalského, prípadne kaledónskeho vrásnenia.

#### 4. Mezozoické ofiolity a bázické horniny

Na rozdiel od predhercýnskych sérií je podiel bázických a ultrabázických telies v mezozoických komplexoch vo väčšine segmentov alpíd malý. Výnimku tvoria vnútorné zóny dinaríd a heleníd a čiastočne i Južné Apuseny, považované za odnož dinaríd (vardarskej zóny (obr. 2, 13) ale i v týchto areáloch sa viažu iba na niektoré štruktúrne zóny. Spravidla ide o asociáciu ultrabázických telies, gabier, spilítov, diabázov v sprievode hlbokomorských sedimentárnych hornín, hlavne rádiolaritov. Už dávnejšie sa pre túto asociáciu používa označenie „ofiolitová séria“ či „ofiolitový komplex“. Táto asociácia býva neúplná a hojnejší je ojedinelý výskyt niektorých jej článkov. Jednako nadobúdajú bázické a ultrabázické telesá v mezozoických sériách vzhľadom na tektonickú teóriu litosférických platní osobitný význam najmä pri určovaní typov a zmien kôry.

Bázické vulkanity, častejšie spolu s gabrami a serpentinitmi, inde s keratofýrmi a porfýritmi až kremennými porfýrmi, sa v mezozoických sériách alpíd viažu na: a) hlbokovodné fácie stredného triasu (hlavne ladinu), sčasti vrchného triasu (hlavne karnu); b) hlbokovodné fácie vrchnej jury a spodnej kriedy; c) súvrstvie preflyšu, hlavne titónsko-spodnokriedového; d) kompletnejší sled pelagických vápencov, rádiolaritov (ale i flyšu), s časovým rozpätím od strednej jury po strednú kriedu.

a) V triase sa vulkanity nachádzajú spravidla s pelagickými vápencami a bridlicami s polohami rádiolaritov.

— V jednotke Budva — Zukali v dinaridách a v helenidách: diabázy, kremenné porfýrity, kremenné porfýry, keratofýry, kremenné keratofýry a uvedené pyroklastické výlevné horniny, ďalej granitporfýry, gabrá, gabrodiority, diority, granity (J. PAMIĆ 1964; obr. 9).

— V pohorí Bükk sú kremenné porfýry, diabázy a ich pyroklastické produkty, diabáz-porfýrity, gabroporfýrity, gabrá, harzburgity a peridotity (obr. 10).

— V meliatskej sérii v Slovenskom kráse sú diabázy a ich tufy, podradne keratofýry, glaukofanity, serpentinity.

— V transylvánskych príkrovoch vo Východných Karpatoch sú diabázy, gabrodolerity, gabrá (v ladine), porfýrity, oligofýry, ortofýry (vo vrchnom triase; G. CIOFLIKA et al. 1966).

— V Južných Karpatoch sú diabázy, kremenné porfýry a ich pyroklastické produkty, diabáz-porfýrity, gabrá, nefelinické syenity, pyroxenity, peridotity.

— Vo vyšších príkrovoch oberostalpinika v severných vápencových Alpách sú menšie výskyty tufov, tufitických porfýritov, keratofýrov (hlavne vo vrchnom anise a v ladine), melafýrov (hlavne v kordevole). V drauzóne sú porfýrity a melafýry — hlavne ich pyroklastické produkty (A. TOLLMAN 1967).

b) V sprievode hlbokovodných facií rádiolaritov, slieňovcov a pelagických kalpionelových a nanokonových vápencov vrchnej jury a spodnej kriedy sú známejšie výskyty bázických telies, prípadne i ofiolitových asociácií.

— Vo vardarskej zóne a v ofiolitovej zóne dinaríd i v zóne Mirdita v helenidách sú diabázy, zriedka keratofýry, gabrá, gabrodiority, syenity a plagiogranity; peridotity, lherzolity, dunity (B. ČIRIĆ 1962 J. AUBOIN 1968, J. PAMIĆ 1964; obr. 9).

— V bukovinských príkrovoch Východných Karpát sú diabázy, porfyryty, ortofýry, keratofýry).

— V Maďarskom masíve v podloží neogénu Velkej maďarskej nížiny západne od flyšového solnockého trógu a v jeho vrchnojursko-spodnokriedovom súvrství sú diabázy, bazaltické porfyryty a porfýry (G. N. DOLENKO—L. G. DANILOVIČ 1976, E. SZÁDECKY—KARDOSS et al. 1967, K. SZEPEŠHÁZY 1973).

— V krížňanskom príkrove Západných Karpát sú augity a ich pyroklastické produkty (D. HOVORKA 1976; obr. 3).

— Vo frankenfelskom a v ďalších severnejších príkrovoch Severných vápencových Álp sú augit-porfyryty, melafýry a ich tufy.

c) V slienovcovo-vápencových a pelitických súvrstviach preflyšu sú známe bázické horniny, niekde aj úplná ofiolitová asociácia:

— V jednotke Ceahlău-Rachov vo Východných Karpatoch a v jednotke čierneho flyšu — Kamenný potok sú diabázy, gabroporfyryty, gabrá, serpentinity (harcburgity a lherzolity), ale i keratofýry; často tufobrekcie a tufy (M. G. LOMIDZE 1968, D. RADULESCU—M. SANDULESCU 1973; obr. 11a).

— V severínskom príkrove Južných Karpát sú diabázy, gabrá, serpentinity, menej keratofýry, diabáz-porfyryty (H. SAVU 1967).

— V jednotke Strandža na Balkáne sú diabázy, keratofýry pikritporfyryty, diabáz-porfyryty (E. DIMITROVA et al. 1975; obr. 6).

— V penniniku, vrátane série Rechnitzer sú diabázy, gabrá, serpentinity.

— V tatridných celkoch i v krížňanskej jednotke a v pribradlovej oblasti Západných Karpát vo vrchnoneokómskom až albskom preflyši sú diabázy, augity a pikryty.

d) V zóne Mureš a čiastočne i vo vardarskej zóne charakterizuje ofiolitová asociácia, podobne ako v penniniku tak jurské karbonátovo-rádiolaritové fácie, ako i mladšie flyšové komplexy spodnej kriedy (obr. 4). Pestrejšie zoskupenie magmatitov je hlavne v Mureši. V jure sú diabázy, gabrá a peridotity; vyššie vo vrchnej jure a v spodnom neokóme sú limburgity, ale i ortofýry, porfyryty, kremenné porfyryty, oligofýry; v apte znova prevládajú bázickejšie magmatity; diabázy, gabrodiority až diority a gabrá (H. SAVU—C. UDRESCU 1973, H. SAVU 1967).

Vo všetkých uvedených prípadoch je zreteľná väzba ofiolitov na štádiá najväčšieho prehĺbenia jednotlivých geosynklinálnych zón. Nejde však len o väzbu paleogeografickú, ale aj o prejav aktivity, teda o väzbu predovšetkým paleotektonickú, ktorá logicky súvisí s tenšou kôrou ofiolitových zón a zón s bázickými magmatitmi a hlbokovodnejšími faciami.

Bázické až ultrabázické telesá, prevažne alkalického charakteru (augitivity, limburgity) sú častejšie i v sprievode plytkovodných facií prahového typu, hlavne v okrajových, zlomami porušených častiach, napr. v tatridách (Vysoké Tatry) a v pribradlovej oblasti Západných Karpát, a to vo vrchnej jure a spodnej kriede. V tomto období sa najväčmi uplatnila činnosť bázického magmatizmu v alpidách.

Podrobnejšie prejavy vrchnojurského a spodnokriedového bázického až ultrabázického magmatizmu často alkalického charakteru, sa viažu na zlomy obmedzujúce kordiliery, prípadne prahy a prehĺbeniny. Známe sú z flyšovej geosynklinály, napr. tešinity v sliezskej jednotke, diabázy a augitivity vo flyšových barémsko-albských súvrstviach bukovinského a podbukovinského príkrovu, prípadne pri okrajoch flyšových trógov, napr. diabázy v jur-

ských vápencoch v zóne Presečina v Južných Karpatoch, ale i augitivity v bradlovom pásme Karpát.

Alkalické sú i spodnokriedové podmorské výlevy v pohorí Mecsek v Maďarsku, alkalické diabázy („bazalt-trachydolerity“) v sprievode pyroklastických produktov a diabáz-porfyrítov (E. SZÁDECKY—KARDOSS et al. 1967). Bázickým vulkanizmom, ktorý sa začína v morskom pelagickom až neritickom prostredí, sa začína splytčovanie, až sa napokon sedimentačný cyklus skončil (J. FÜLÖP 1968).

V zlepencoch wildflyšu, napr. v albe v zóne Metaliferes, či v baréme — apte pri severnom okraji marmaroškého masívu, v kriedových zlepencoch bradlového pásma Západných Karpát, ale i v intraformačných zlepencoch uprostred albského preflyšu obalových jednotiek Západných Karpát sa vyskytujú valúny bázických hornín, čo ukazuje na ich genetický vzťah k diferenciácii kôry.

Táto závislosť je zrejmä i z rozšírenia bázických telies v mezozoických sériách, ale tu hrá úlohu i členitosť geosynklinály a zvýšená dynamika pohybov kôry v období magmatických prejavov. Bázické telesá sú sprievodným javom hlavne prehĺbenia tenkej kôry a hlbokosiahajúcich zlomov oddeľujúcich prehĺbeniny a prahy, prípadne kordiliery.

Najväčší rozsah bázických a ultrabázických magmatitov je vo vrchnej jure a v spodnej kriede, t. j. v období najväčšej členitosti a najväčšej paleotektonickej kontrastnosti mezozoických geosynklinál, ale i najvýraznejšej oceanizácie v alpidách.

Rozloženie ofiolitov a ich typ závisia od členenia sedimentačnej oblasti na trógy (prehĺbeniny) a prahy (príp. kordiliery). Tak ako má členitosť každého segmentu svoje osobitosti, majú ich aj ofiolitové zóny. Len niektoré z nich prechádzajú z jedného segmentu do druhého, aj to s určitými odlišnosťami.

Vo Východných Alpách majú fácie hlbnejší charakter v triase len v najjužnejších zónach nadväzujúcich na Južné Alpy; v jure a v kriede je výrazný severnejší penninský tróg.

Rovnako je to i s rozložením zón s bázickými horninami a ofiolitových zón. Ofiolitová asociácia v Alpách je práve charakteristická pre penninikum. Rozdiely v type ofiolitových asociácií sú v Alpách zvlášť výrazné, a to hlavne v Západných Alpách, kde sú dve vetvy oceánskych trógov, ktoré oddeľuje briansonský prah. V severnej zóne, označovanej ako „Waliser Bünderschiefer Trog“ v sprievode bridlíc (Bünder Schiefer) sú len menšie výskyty bázických hornín, a to doleritické a gabrové žily, poduškové (pillow) lávy, hyaloklasty, relikty serpentinitov, zriedkavé gabrá (V. DIETRICH 1976). Ide o tróg s málo výrazným oceanickým charakterom kôry, skôr o typ suboceanický. Južnú penninskú, či piemontskú zónu charakterizujú: kremité bridlice, rádiolarity, slieňovce, bridlice, ktoré sprevádzajú výrazne poduškové lávy a poduškové bazaltové brekcie.

V oblasti Glockner vo Vysokých Tatrách amfibolity, eklogity, prasinity a serpentinity vykazujú rádiometrický vek 170 mil. rokov.

Úlomky serpentinitov, gabier, diabázov, poduškových bazaltov a hyaloklastov sa nájdu aj v spoločenstve hlbokomorských jurských rádiolaritov a aptychových vápencov i slieňovcov titónsko-neokómskeho veku v bradlovom pásme pri St. Veit (V. DIETRICH 1976). Hojný ofiolitový detrit je pozoruhodný v klastikách flyšového pásma Východných Álp, ba i v gosauskej kriede až po kampán (V. DIETRICH 1976, R. HESSE—A. BUTT 1976). Tieto severnejšie výskyty bázických a ultrabázických telies súvisia zrejme so severnejšou vetvou ofiolitového trógu, geneticky zviazanou s flyšovým pásmom. Výstupy ofiolitov v tektonickom okne

Rechnitzer predstavujú s najväčšou pravdepodobnosťou južnú vetvu penninika (A. TOLLMANN 1967).

V Západných Karpatoch v triase majú hlbokomorský charakter najjužnejšie zóny Bük, rudabanské pásmo a meliatska séria v Slovenskom krase; len tieto zóny majú triasové báziká a ultrabáziká (obr. 10). V sprievode rohovcových vápencov, silicitických bridlic, rádiolaritov i tmavých ílovcov, vystupujú vo vrchnom anise až spodnom karne diabázy, porfýry a ich tufy. Nájdu sa aj spility a esexity. V Bükku dosahujú komplexy lávových prúdov miestami hrúbku niekoľko sto metrov (P. ARKAY 1973). Sprievodné gabrá, harcburgity a peridotity sú podľa niektorých autorov vrchnojurské (C. BAKSA—J. CZILAG—J. FÖLDESZI 1974); pravdepodobnejšia je však ich genetická väzba s triasovými vulkanitmi.

V jure je pre Západné Karpaty výraznejší špecifický zliechovský a kysucký tróg. V ich okrajových častiach na styku so susednými prahmi, napr. kordilierami sú bázické telesá. Členitosť tatrídneho pruhu sa odráža v častejších alkalických spodnokriedových prienkoch limburgitov a augititov. I tešiny sú vo flyšovom pásme späté s rozhraniami hrubšej kôry prahov a tenšej kôry prehĺbenín (M. ELIÁŠ 1976).

Ofiolitovú zónu možno predpokladať v spodnej štruktúrnej etáži magurského príkrovu v Západných Karpatoch (obr. 3, 13). Predstavuje pokračovanie flyšového pásma Východných Álp (Z. ROTH 1964) i s nimi geneticky zviazaného penninika (s ofiolitmi) a na východe nadväzuje svojou račiansko-bystrickou časťou na jednotku Rachovo-Ceahlau s ofiolitmi a na sprievodnú ofiolitovú sériu „čierneho flyšu“ (rozloženú pri vnútornej strane flyšovej zóny a v podloží presunutého marmaroškého masívu).

Svedectvom existencie článkov ofiolitovej série sú šupiny pri styku magurskej jednotky s bradlovým pásmom, miestami blízke typom známym z Východných Álp i z Východných Karpát. Takými sú predovšetkým posidoniové vrstvy (M. ELIÁŠ 1976) i členy série Grajčarka pri severnom okraji (posidoniové vrstvy, rádiolarity — malm, vápence typu biancone — titón — spodný neokóm, čierny flyš — spodná — stredná krieda). Azda z tejto „skrytej“ ofiolitovej série pochádzajú početné bázické a ojedinelé ultrabázické útvary: melafýry, diabázové porfýry, augitové porfýry, augitivity, limburgity, gabrodioritovo-diabázové porfýry, ale ojedinele i dacity, kersantity, kremenné porfýry, keratofýry, granitové porfýry a granity s rádiometrickými hodnotami (stanovenia J. Kantora) 140—98,3 mil. r. Ide o valúnový materiál z kriedových zlepenčov bradlového pásma v jeho západnom úseku. Zvlášť pestrá paleta magmatitov je v paleogénnych zlepencoch pribradlovej oblasti s bradlového pásma na východnom Slovensku: augitové porfýry, augitivity, limburgity, melafýry, diabázové porfýry, gabrodioritovo-diabázové porfýry, dacity, kersantity, kremité porfýry, keratofýry, granitové porfýry a granity (R. MARSCHALCO—M. MIŠÍK—L. KAMENICKÝ 1976).

Polarita v rozložení ofiolitovej asociácie sa vyskytuje aj vo Východných Karpatoch. Hlbokovodné fácie triasu s hojnými, hlavne bázickými i ultrabázickými magmatitmi sú aj v príkrovoch transylvánskej skupiny, pôvodom najjužnejších. Najvýraznejšie prehĺbenie a stenčenie kôry jurských a spodnokriedových členov sa javí u bukovinských príkrovov (zo severných častí centrálného pásma) a o niečo neskôr v titóne a v spodnej kriede jednotiek „čierny flyš“ — Kamenný potok a Ceahlau. Aj tu sú prehĺbeniny zónami ofiolitov (M. SANDULESCU 1973, D. RADULESCU—M. SANDULESCU 1973; obr. 11).

Pre Apusény je charakteristická ofiolitová asociácia v zóne Metaliferes — typu trógu sčasti s tenkou, azda oceanickou kôrou s mocnými rádiolaritmi a so sedimentáciou pestrých (heterogénnych) typov flyšu sčasti v jure i počas obdobia spodnej kriedy; obr. 4). V okrajo-

vých častiach má kôra kontinentálny charakter so sedimentáciou plytkovodných vápencov.

Osobitosťou maďarského masívu je predovšetkým solnocký flyšový tróg so sprievodnými bazickými telesami; tento tróg smerom k Z nadväzuje na pásmo pohoria Mecsek (E. SZÁDECKY—KARDOSS 1973). Vrchnojurské až spodnokriedové diabázy, alkalické diabázy a porfyrity vystupujú v peliticko-karbonátovom súvrství so sprievodom silicítov v podloží flyšu aj západnejšie od flyšového trógu. Jurské a spodnokriedové dajky, silly a lávové prúdy diabázov (sčasti alkalických), bazaltických porfyrítov a porfýrov sa nachádzajú v sprievode ílovcov, slieňovcov, rádilarítov a pelagických vápencov (K. SZEPEŠHÁZY 1973; obr. 13).

Osobitosťou Južných Karpát v mezozoiku je tróg s ofiolitmi v jednotke Severín a v prebiehajúcej vnútornej zóne danubika, v zóne Svinjica — Greben (D. RADULESCU—M. SANDULESCU 1973; obr. 5).

K osobitostiam Balkánu v období predflyšového štádia mezozoika patrí jednotka Strandža s preflyšom obsahujúcim ultrabázické a bazické horniny (E. BONČEV 1976; obr. 6).

Dinaridy majú pestrý trias s hlbokovodnými fáciami hlavne v zóne Budva (s vulkanitmi sčasti bazickými už v triase; obr. 9). V jure je členitosť ešte výraznejšia s prehĺbeninami hlavne v jednotkách ofiolitovej zóny — subpelagoniku a vo vardarskej zóne, s charakteristickou diabáz-silicítovou formáciou. V niektorých priestoroch pokračuje hlbokomorský typ sedimentácie od triasu až do kriedy tak ako zastúpenie ofiolitov (M. DIMITRIEVIĆ 1974, S. KARAMATA 1974). Ofiolitová zóna dinaríd sa interpretuje ako ofiolitová melanž, zóna olistostrómov, ktoré sedimentovali pri okraji oceánu s prepádavajúcou sa kôrou. Matrix tvoria silicity; bloky predstavujú bazické, ultrabázické horniny, červené aleurolity, pieskovce a vápence od permských po jurské. Zvlášť veľké sú telesá triasových a jurských vápencov, gravitačne sklíznutých z bezprostredného susedstva (M. D. DIMITRIEVIĆ — 1967). Účasť senónskych sedimentov v melanži vardarskej zóny sa vysvetľuje ako dôsledok mladších dextrálnych transkurentných pohybov na rozhraní dinaríd a srbsko-macedónskeho masívu. Vlastná subdukcia, pohltie „oceánu Zvornik“ so vznikom ofiolitového olistostrómu sa považuje za starší proces, počínajúci vrchnou jurou (M. D. DIMITRIEVIĆ 1974).

Ako vidíme z uvedených skutočností, môžeme v alpidách hovoriť o priamej závislosti rozsahu i rozloženia ofiolitov od členitosti mezozoickej geosynklinály, hlavne v časovom rozpätí stredný trias — spodná krieda.

I keď je vo väčšine segmentov európskych alpíd menej ofiolitov, pre aplikáciu princípov novej globálnej tektoniky má význam najmä ich veľká paleotektonická členitosť, hlavne vo vrchnej jure a spodnej kriede a ich preukázateľné prechody z hlbokomorských sekvencií do plytkovodných. Majú viaceré zóny s tenkou kôrou a zóny s hrubšou kôrou. Jursko-spodnokriedový tetýdny oceán sa líšil od dnešných oceánov. Napr. v jure Západných Karpát treba rozlíšiť najmenej 4 zóny s tenšou kôrou, ktoré sú oddelené prahmi (príp. kordilierami) s hrubšou kôrou (M. MAHEL 1975; obr. 13, 11b). Vrstevné sledy so zmiešanými plytkovodnými a hlbokovodnými fáciami sú zvlášť výrazné v bradlovom pásme, v pásme azda najvýraznejšej kontrastnosti medzi czorsztyńským a kysuckým typom. Všetko poukazuje na to, že model dnešného Atlantického oceánu nemožno aplikovať na alpidy. Jednotlivé prehĺbeniny nikdy nedosahovali takú šírku ako Atlantik a neboli tak málo členené. Boli to oceánske priehlbne, úzke trógy. Náhly stratigrafický a bočný prechod hlbokomorských fácií do plytkovodných nie je v súlade s názorom, podľa ktorého styk antagonistických hlbokomorských sekvencií s plytkovodnými by mal byť dôsledkom zvlášť veľkého tektonického skrátenia priestoru, prípadne tektonického prekrytia.

Prítomnosť bázických a ultrabázických hornín naznačuje začiatok rozťahovacej zlomovej aktivity, vznik tektonických prepادلín, stenčenie kôry už v triase a podľa niektorých autorov (N. HERZ—H. SAVU 1974) je dôkazom toho, že sa roztrhla žulovo-rulová kôrová vrstva a že sa rozostúpili litosférické platne.

Ofiolitové série jury a spodnej kriedy sa považujú za oblasti s oceánickým typom kôry. Vlastná tetýda sa stotožňuje hlavne s ofiolitovými sériami v dinaridách a helenidách. I tu však litosférická mikroplatňa Drina — Ivanjica a Pelagónsky masív členia „oceán“ na dve vetvy: východnú — oceán Zvornik (vardarská zóna) a západnú, reprezentovanú ofiolitovou zónou a subpelagonikum (M. D. DIMITRIEVIĆ 1974). Triasové ofiolitové série v pohorí Bükk sa považujú za odnož tetýdy, nadväzujúcu na dinaridy (E. SZÁDECKY—KARDOSS 1973). Meliatska séria v Západných Karpatoch je len severným okrajom tejto „dinaridnej“ odnože. Bázické telesá navrátné v podloží neogénu Alföldu západne od solnockého flyšového trógu a jeho spodnej štruktúrnej etáže, považované za článok ofiolitovej asociácie, tvoria ďalšiu vetvu členitej tetýdy, rozloženej medzi východo-maďarským blokom a Apusenami (A. JUHASZ—G. VASZ 1974). Ofiolitové série na Balkáne v Sakare a v Strandži sa považujú za severný výbežok tetýdy, pokračujúcej z vardarskej zóny do tauríd (P. GOČEV 1976).

V Alpách ofiolitové série penninika predstavujú „severnú tetýdu“, v piemontskej časti dokonca s predpokladaným oceánickým chrptom (V. DIETRICH 1976), v Západných Alpách rozčlenenú briansonskou platňou (Briançonnais). Pritom u severnej Valaiskej vetvy (Bünderschiefer) sa predpokladá suboceánický typ kôry.

Ofiolitové asociácie vo Východných Karpatoch spolu s Apusenami sú interpretované ako dva intrasialické bazény s oceánickou kôrou (D. RADULESCU—M. SANDULESCU 1973). Severný bazén predstavujú dnes ofiolitové série v jednotke Ceahlău a v jednotke „čierneho flyšu“, ale i bázické telesá bukovinského príkrovu; južný bazén zastupuje ofiolitová séria transylvánskeho príkrovu a hlavne ofiolity pásma Murešu (Metaliferes). Podľa niektorých autorov tento „oceán“, nazývaný Siretský, prepája tetýdu s oceánom Dobruđa — Krym — Kaukaz (N. HERZ—H. SAVU 1974). K juhozápadu sa tento „oceán“, rozložený medzi Severnými Apusenami a Južnými Karpatmi — nazývaný tiež murešský — spája s tetýdnym „oceánom“, ktorý reprezentuje vardarská zóna.

Ofiolity v zóne Severín a bázické telesá, zasahujúce i do zóny Svinjica — Greben v Južných Karpatoch sa ponímajú ako reprezentanti bývalého „danubijského“ oceánu rozloženého medzi Moezijskou platňou (vrátane západnej časti danubika) a zónou getika spolu so srbsko-macedónskym masívom, považovaným za ostrovnú zónu medzi „danubickým oceánom“ a tetýdou.

Zostáva medzera v rozložení „severného oceánu“, a to v Západných karpatoch, kde je pravá ofiolitová asociácia na povrchu neznáma. Pre svoj anomálny štruktúrny charakter, bradlové pásmo (pásmo tektonickej brekcie) rozložené pri vnútornom okraji s ním užšie zviazaného flyšového pásma — považujú niektorí autori za spojku penninského oceánu a severného oceánického bazénu Východných Karpát (J. DEWEY et al. 1978, E. SZÁDECKY—KARDOSS 1973). Je to však skôr vnútorná okrajová zóna s kordilierovým ostrovným pásmom czorsztyńským a klapským, oddelenými úzkou kysucko-pieninskou prehĺbeninou. Vlastný „oceánický“ tróg zrejme predstavovala vnútorná zóna flyšového pásma, zovretá medzi dvoma hlbinnými, geofyzikálne preukázanými líniami: lednickou či perikarpatskou peripieninskou, čiže zónou karpatského bloku v zmysle W. ŠIKORU (1976) s predpokladaným metamorfovaným flyšom v hlbších polohách. V Západných Karpatoch sa taká zóna, ktorú

možno považovať i za domovskú, silne redukovanú oblasť príkrovov flyšového pásma, zhruba stotožňuje s rozsahom magurskej jednotky. Ale smerom na západ prebieha jej vnútorný okraj po západnom okraji Malých Karpát — teda aspoň 30 km jv. od povrchových výstupov bradlového pásma. Tým sa toto pásmo s predpokladanou ofiolitovou asociáciou v Západných Karpatoch približuje ku krajnému východnému výstupu penninika v okne Rechnitzer (obr. 3, 13). Z toho vyplýva predpoklad, že na rozhraní Východných Álp a Západných Karpát, v podloží viedenskej panvy — najvýznamnejšej priečnej naloženej štruktúry medzi dvoma segmentmi — sa Západné Karpaty posunuli najmenej o 30 km k SZ.

Ofiolitové asociácie — predstaviť „oceánu“ severná tetýda pokračujú tak z penninika Álp cez spodné štruktúrne etáže magurského príkrovu do zóny „čierny flyš“ — Ceahlău (obr. 13) vo Východných Karpatoch, cez jednotku Severín (a príľahlý západný okraj danubika Svinjica — Greben) do zóny trojanského flyšu a cez zónu Kotel, resp. pri jej okraji ďalej na východ. Južný bazén tetýdy predstavujú ofiolitové zóny asociácie dinaríd a heleníd a ďalej na východ tauridy. Taká schéma s dvoma oceánmi v jure a v spodnej kriede je blízka vzorovým modelom znalcov ofiolitových zón z iných kontinentov (J. DEWEY—J. BIRD 1970, A. KNIPPER 1975), ale i niektorých alpských geológov (M. BOCCALETTI et al. 1973 a i.).

Samotné „oceány“ a to tak severná, ako aj južná tetýda boli rozčlenené ostrovmi s hrubou kontinentálnou kôrou; napr. brianský ostrov v oceáne penninika, Drina — Ivanjica—pelagonikum v južnej tetýde. Výbežky „oceánov“ však zasahujú i do širokého pásma rozloženého medzi nimi a členia ho na malé mikroplatne: bükkský „záliv“ medzi centrálnymi Alpami a východným panónskym blokom, solnocký medzi týmto blokom a severnými Apusenami, murešský medzi severnými Apusenami a getikom Južných Karpát. Členitosť tetýdy možno síce sčasti vysvetliť rozlámaním intraoceánskych platní s kontinentálnou kôrou na mikroplatne, či menšie fragmenty kontinentálnej kôry, ktoré sa navzájom vzdalujú a otáčajú hlavne pri vytváraní karpatského oblúka (E. SZÁDECKY—KARDOSS 1973, V. ALEXIČ et al. 1974). Sotva však možno považovať Villany i Mecsek za súčasť pôvodného vnútrokarpatského pásma (P. HORVÁTH—L. STEGENA—B. GÉCZY 1974, M. BLEAHU 1976), a to ani na základe porovnania sledu mezozoických tektofácií (nie ojedinelých fácií!) a už vonkoncom nie na základe stupňa paleoalpínskej deformácie. V mezozoiku v pohorí Villány a Mecsek chýbajú také výrazné vnútrokarpatské tektofácie ako keuper a hlavne albsko-cenomanský flyš; zásah paleoalpínskym vrásnením je u oboch malý, neporovnateľný so západokarpatskými jednotkami.

Pri uvažovaní o rozložení bazénov a trógov s oceánickou kôrou nemožno opomenúť ani sekvencie, ktoré nemožno označiť za ofiolitové. Ich podstatnou zložkou sú jursko-spodnokriedové sedimenty, batyálne až abysálne fácie analogické so sedimentami ofiolitových sekvencií, ale sprevádzané iba skromnými výskytmi bázických alebo ultrabázických telies. Charakteristické sú pre pieninsko-kysuckú jednotku bradlového pásma, krížňanský príkrov v Západných Karpatoch, frankenfeldský príkrov a západnejšie analogické príkrovy v Severných vápencových Alpách, bukovinský príkrov vo Východných Karpatoch, čiastočne i pre jednotku Budva — Zukali a jursko-spodnokriedové časti jónskej jednotky v helenidách. Alpidná geosynklinála hlavne v období jury a spodnej kriedy bola zrejme rozčlenená na prehĺbeniny s oceánickou kôrou a na hlbokovodné trógy s kontinentálnou kôrou, ale pravdepodobne stenčenou. Medzi tým je niekoľko zón s prevahou plytkovodných fácií geosynklinálneho typu. Nejde o platne, útržky platformy, ale o zóny s hrubšou kôrou, o prahy geosynklinály (J. AUBOUIN 1964).



Pre geológa, ktorý roky spracúval mezozoikum vnútorných Západných Karpát a azda aj pre toho, kto pracoval v oberostalpiniku, prípadne v Južných Alpách, je ťažké zmieriť sa s názorom, že oblasti, rozložené medzi „severnou“ a „južnou“ tetýdou, predstavovali hlavne v priebehu jury a spodnej kriedy málo aktívne platne s epikontinentálnym vývojom. Zužovať aktivitu alpíd len na ofiolitové zóny je schematizácia dokonca väčšia, ako bolo členenie na eugeosynklinály a miogeosynklinály, ktoré bolo treba rozšíriť o celý rad ďalších typov geosynklinál. V živej pamäti máme ešte ťažkosti, s akými sme sa stretávali pri úlohe začleniť do rámca týchto dvoch základných geosynklinálnych typov jednotky Západných Karpát na tektonickej mape Euroázie 1:2 500 000 (N. ŠATSKIJ—A. BOGDANOV 1964). Prvým krokom pri rozpracovaní princípov pre tektonickú mapu Karpát, Balkánu a Dinárskeho pohoria v mierke 1:1 000 000 bolo vyčlenenie geosynklinály alpíd ako osobitného typu, pre ktorý nie je charakteristická hrúbka sedimentov, ani väčší či menší podiel magmatitov, hlavne bázických a ultrabázických telies, ale členitosť meniac sa v priebehu vývinu, teda zvýšená dynamika počas vývoja charakterizovaného početnosťou štruktúrno-faciálnych zón. Zvýšená dynamika sa prejavila i v neobyčajnom skrátaní priestoru, v tvorbe početných príkrovov zón kontrakcie. Tento typ geosynklinály sme nazvali alpínskym typom. Skoro nato sa objavilo zvučnejšie pomenovanie — aristogeosynklinála (A. TOLLMANN 1967). Členitosť geosynklinály a jej menlivosť v jednotlivých štádiách je znázornená na Tektonickej mape Karpát, Balkánu, Dinárskeho pohoria a priľahlých oblastí tektonogrupami; grupy tektofácií bližšie dokumentujú paleotektonický charakter priestoru, v ktorom vznikla daná tektonická jednotka či oblasť (M. MAHEL 1973). Bližšie je to rozvedené vo Vysvetlivkách k tektonickej mape (M. MAHEL et al. 1974). Oblasti s výraznejšími prejavmi ofiolitov a bázických telies sú vyznačené ako osobitné typy tektonogrup — osobitné typy trógov. Jestvuje niekoľko typov trógov a prahov rozložených medzi nimi a taktiež aj šelfov. Osobitný typ predstavujú trógy penninika. V poslednom období sa rozlišujú 3 typy: valaiský, piemontský, ligurský. Aj v Dinarskom pohorí sa vyčleňuje niekoľko typov: vardarský a subpelagonský. Pásmo Ceahláu a Severín s preflyšom predstavuje iný typ ofiolitových trógov než typ murešský obsahovo heterogénny. Ani typy trógov s tenšou sialickou kôrou nie sú jednotné stratigrafickým rozsahom, ani obsahom vo vývoji. Napr. krížňanský typ s hlbokovodnými faciami od liasu po cenoman, s augititmi v neokóme a v spodnom albe sa výrazne líši od typu Mecsek. Od vrchného liasu po spodnú kriedu sa v ňom striedajú sedimenty plytkomorské a hlbokomorské, spodnokriedové magmatity pohoria Mecsek majú alkalický charakter (alkalické bazalt-diabázy a alkalické diabázy).

Rozdiely v typoch trógov vyplývajú do značnej miery z odlišnosti postavenia v jednotlivých segmentoch geosynklinálneho systému alpíd, z rozdielov v členitosti, ale i v dynamike kôry. Prítom však členité trógy s oceanickou kôrou majú zrejme regionálny charakter, ako to vyplýva z priebehu „severnej tetýdy“: penninikum — spodná etáž magurského príkrovu — Ceahláu — Severín — pásmo Kraina trojanského flyšu — Kotel, ale i priebeh južnej tetýdy: vnútorné dinaridy — vnútorné helenidy — tauridy. Trógy s oceanickým typom kôry sú najskôr geotektonickým prvkom vyššieho rádu — spojkami viacerých segmentov; trógy s tenšou kontinentálnou kôrou, ako je napr. krížňanský typ, sú charakteristické len pre niektoré segmenty alpíd. Paleotektonický model alpíd bol počas alpínskeho cyklu zrejme zložité a menil sa v jednotlivých úsekoch i v jednotlivých vývinových štádiách v čase i priestore. Neoddeliteľnou a najvýraznejšou súčasťou tohto modelu, a to od stredného triasu až po

vyvrásnenie, boli trógy s ofiolitmi, prípadne s hlbokovodnými sekvenciami facií v sprievode bázických, prípadne ultrabázických hornín.

Členitosť a výrazná paleotektonická protikladnosť alpskej geosynklinály — najvýraznejšia vo vrchnej jure a v spodnej kriede — je jedným z charakteristických znakov alpskej geosynklinály. Dokazuje to zonárne a pritom selektívne rozmiestnenie ofiolitových asociácií a ich väzba na trógy. Súvisí to zrejme s „oceanizáciou“ kôry v interkontinentálnych podmienkach. Výraznú úlohu zohrali aj pozdĺžne hlbinné zlomy, ktoré oddeľovali v alpskej geosynklinále prahy a prehĺbeniny, resp. trógy. Po nich dochádzalo k výstupu hlbinných bázických i ultrabázických magmatitov, ktoré zohrali význačnú úlohu pri rozťahovaní kôry, pri stenčovaní jej žulovo-rulového sloja a miestami aj pri rozostupovaní platní. Nemožno však zabúdať, že vo vývoji alpíd i v rozložení trógov a prahov, ako aj v dynamike kôry zohrala významnú úlohu nerovnomernosť hercýnskej granitizácie (obr. 7).

Ďalším charakteristickým znakom alpíd je ich veľké, v niektorých zónach enormné skrátenie, často s „nezakorenými“ príkrovmi. Zvlášť zóny s ofiolitovými asociáciami a zóny s tenšou kontinentálnou kôrou — trógy — vytvárajú rozsiahle príkrovy. Za takých okolností je logická štruktúrna pozícia ofiolitových asociácií a bázických telies so sprievodnými hlbokovodnými sedimentmi zväčša v alochtónnej príkrovovej pozícii.

S ofiolitovými zónami sa v histórii pásmových pohorí spájajú dva tektonicky protichodné procesy prvoradého významu a to:

— rozťahnutie kôry, rozostúpenie platní, čiže vznik oceánskych bazénov s ofiolitovými asociáciami, alebo aspoň stenčenie kôry, vznik trógov s prítomnosťou bázických telies

— skracovanie priestoru, zanikanie trógov, pohlcovanie oceánskej kôry, jej ponáranie pod kontinentálne platne, čiže subdukcia, pokles častí kôry do hĺbky, pretavenie oceánskej i kontinentálnej kôry, čo znamenalo zásah do hlbinných procesov.

Procesy rozťahovania kontinentálnej kôry sú novým prvkom v chápaní vývoja pásmového pohoria a v alpskom systéme znamenajú dynamickejší prístup k poňatiu vývoja geosynklinálnych systémov. Zvýrazňujú rozdiely v šírke geosynklinálneho systému, rozdiely v členitosti a type kôry v jeho jednotlivých vývinových štádiách. Pritom je v európskych alpídach osobitne cenným prvkom zvýraznenie rozdielov medzi triasom a jurou, čiže rozčleňovanie obdobia geosynklinálneho vývoja pred paleoalpínskym vrásnením, často známeho pod pojmom karbonátové štádium, na dve štádiá: triasové a jursko-spodnokriedové (M. MAHEL 1973).

Proces skracovania geosynklinálneho priestoru sa doteraz spájal s obdobiami vrásnenia. Novým prvkom je skracovanie priestoru subdukciou oceánskej kôry, pričom ho považujeme za dlhodobý proces obvykle predchádzajúceho vrásnenia. Súčasne okraje ofiolitových zón, hlavne vnútorné, s ktorými sa spája subdukcia, považujeme za seizmicky aktívne zóny, ako zóny Benioffove. S nimi sa často spája základná aktivita vývoja pásmového pohoria.

Podrobnejšia analýza prejavov vrásnenia v Karpatoch, na Balkáne, Dinarskom pohorí a v Západných Karpatoch (M. MAHEL 1973) ukazuje, že tento proces, s ktorým sa spája predovšetkým skrátenie geosynklinálneho priestoru, mal viac štádií, a teda bol dlhodobý. Preto sme na tektonickej mape Karpát, Balkánu a Dinárskeho pohoria vyčlenili jednotky sformované v periódach vrásnenia (paleoalpínske, mezoalpínske, neoalpínske), a nie v krátkodobých fázach. Skrátenie trógov s oceánskou kôrou je iba jedným typom dlhšie prebiehajúceho skrátenia geosynklinály. Treba však zdôrazniť, že skrátenie priestoru v alpídach nie je charakteristické len pre ofiolitové zóny. V Západných Karpatoch majú všetky zóny, dokonca

i zóna tatríd považovaná za autochtón, menšie príkrovy. Z nich sú dávnejšie známe príkrovy vo Vysokých Tatrách a v posledných rokoch boli zistené prešmyky, presuny a menšie príkrovy v kryštalickom podloží, napr. v Považskom Inovci. Popri všeobecnom skrátaní boli najviac postihnuté niektoré zóny s tenšou kontinentálnou kôrou, najmä trógy. Týka sa to predovšetkým severných veporíd — koreňovej zóny krížňanského príkrovu i zóny južne od veporíd, spájanej s lubeníckou jazvou (koreňovej zóny chočského príkrovu a vyšších príkrovov — obr. 15).

V bradlovom pásme sa dlhodobé skracovanie odráža v hojnosti wildflyšu, a to v jeho západnej časti s polymiktnými zlepenkami od albu po paleocén, monomiktnými od paleocénu do spodného eocénu; vo východoslovenskej časti sú polymiktné zlepenky veku maastricht až lutét (R. MARSCHALCO et al. 1976). Pritom chýba akákoľvek výraznejšia diskordancia i prerušenie sedimentácie. Je zrejmé, že 30- až 40-násobné skrátanie bradlového pásma (D. ANDRUSOV 1968) bolo dlhodobým procesom.

Rozloženie niektorých ofiolitových zón v susedstve a v laterálnej nadväznosti na flyšové zóny, slúži ako jeden z dôkazov platnosti zásad novej globálnej tektoniky v alpidách (E. SZÁDECKY—KARDOSS 1974, N. HERZ—H. SAVU 1974). Keďže v niektorých segmentoch európskych alpíd je hypertrofia flyšu, posúdenie vzťahu flyšu k ofiolitovým asociáciám má širší význam (obr. 14a).

Už predtým sme poukázali na to, že v niektorých zónach prerastajú ofiolitové asociácie do flyšových, v ktorých sú bázické a ultrabázické telesá (vardarská zóna, zóna Metaliferes, penninikum, jednotka „čierneho“ flyšu). Bázické a ultrabázické telesá často vystupujú v preflyši. V iných prípadoch vystupujú flyšové bazény v susedstve ofiolitových zón, napr. sarajevský flyšový tróg a Gramos pri západnom okraji ofiolitového pásma dinaríd a pásma Merdita-subpelagonikum. Severínsky flyšový tróg vytvára okraj zóny Greben, flyšové pásma Ceahlău severný okraj jednotky „čierny“ flyš. Ako vidieť, medzi ofiolitovými asociáciami a flyšom sú v alpidách popri laterálnych väzbách aj väzby vertikálne.

V zónach, kde karbonátové sekvencie zamieňa flyš až v albe, sú bázické a ultrabázické telesá častejšie práve v období striedania týchto facií, v období reprezentovaného preflyšom (krížňanská jednotka a tatrídne jednotky Západných Karpát).

Ofiolity v triase buď nemajú flyšový sprievod (Bükk, meliatska séria), alebo flyš u nich vystupuje ako sprievodný článok, napr. v jednotke Zukali.

Rozloženie flyšu a jeho časový rozsah je spravidla ukazovateľom orogenetickej polarít (J. AUBOUIN 1964, M. MAHEL 1974), jedného z najcharakteristickejších znakov alpíd. A tak vzájomná časová a priestorová spätosť ofiolitových asociácií a flyšu ukazujú, že obidve formácie sú súčasťou vývojového rádu alpínskej geosynklinály.

Flyšové komplexy „zrelého“ štádia s prevahou flyšu s.s. a hrubého flyšu sú spravidla chudobné na magmatity, častejšie sa v nich vyskytujú vulkanity intermediárneho až kyslého charakteru. Tieto flyšové komplexy vznikli zrejme za iného dynamického stavu kôry ako ofiolity (obr. 3, 11b).

Popri zmenách petrochemického charakteru ofiolitov, pôvodne s magmatitmi toleitového typu v zónach oceánskej, prípadne tenkej sialickej kôry a s alkalickými typmi (augitity, limburgity) v zónach s hrubšou kôrou, je zaujímavá zmena charakteru magmatitov ofiolitovej formácie v časovom slede. Triasové ofiolity častejšie obsahujú kyslé horniny, ba i kremenné porfýry (napr. v anise Bukových hôr, keratofýry v gemeridách Západných Karpát i v meliatskej sérii, kremenné keratofýry v anise, v ladine dinaríd a v ladine Južných Karpát).

Petrochemický typ ich magmy zodpovedá začínajúcej diferenciácii kôry a výraznejšiemu stenčeniu v niektorých zónach. Častejšie sa výstup tejto magmy spája s diapirom bohatým na kontaminované magmy (S. KARAMATA 1974). V jure a v spodnej kriede, t. j. v čase najvýraznejšej oceanizácie, prevládajú hlavne bázické a časté sú ultrabázické telesá toleitového typu magmy. V strednej a vrchnej kriede, t. j. v období prevahy flyšovej sedimentácie, je výrazný prechod k intermediárnym typom magmatitov s častejšími, rozsahom veľmi obmedzenými porfyritmi a kremennými porfyritmi (napr. flyšové pásmo Východných Karpát; M. G. LOMIDZE 1968; flyšové pásmo Západných Karpát; Metaliferes). Typ magmy zrejme zodpovedá zmenám v type kôry v priebehu vývinu alpínskej geosynklinály od hercýnskej granitizácie a konsolidácie cez relatívnu diferenciáciu kôry a jej stenčenie až po leptogeosynklinálne štádium s najrozsiahlejšou oceanizáciou pôvodnej sialickej kôry až k neopalpskej konsolidácii.

## 5. Mezoalpínske interorogénne vulkanity, „banatitový typ“

V zóne Srednogorje na Balkáne, v timockej zóne, v Poiana Rusca v Južných Karpatoch a v Apusenách vystupuje vulkanoplutonická formácia, známa v literatúre pod názvom banatity (obr. 2).

Svojím petrografickým, sčasti vulkanologickým charakterom sú analogické s neskororoogénnymi trefohornými vulkanitmi, preto sú spravidla ponímané ako subsekventné, resp. včasne subsekventné (H. STILLE 1953, A. TOLLMANN 1969). Ich základnými formáciami sú ryolitovo-andezitová, trachyandezitovo-latitová formácia s explozivitou až 90 %. Popri týchto vulkanogénnych formáciách mediterárneho typu obsahujú banatity aj alkalické formácie: (limburgity, pikrity, olivinické bazalty a trachyty) s menšou explozivitou, často len asi 40 %. Kriedovo-paleocénne vulkanity sa líšia od neovulkanitov nielen vekom, ale aj postavením v geotektonickom cykle — sú interorogénne. Vznikli po silnom paleoalpínskom vrásnení, ale aj sami boli zasiahnuté mezoalpínskym vrásnením (M. MAHEL 1973 — obr. 12). Charakteristická je aj ich tektonická väzba s prepادلínovými synklinálami, zväčša paralelnými s pozdĺžnymi paleoalpínskymi štruktúrami. Vulkanické komplexy banatitov sú podmorské, objavujú sa s postupnou oceanizáciou kôry. Na ich báze sú obvykle vyvinuté včasné molasy, nad nimi leží hrubý flyš.

Pre banatity sú charakteristické predovšetkým polygénne vulkány centrálneho typu s prechodom od typických vulkanických facií (lávovej a pyroklastickej) k subvulkanickým formám až hypoabysálnym telesám (G. STANISCHEVA—VASSILEVA 1973). Súvisí to s početnosťou a pestrosťou vulkanoplutonických komplexov banatitov, s hojnosťou telies gabier, gabrodioritov, dioritov i granitov, ale i monzonitov, syenitov (syenit-granitov a granodiorit-granosyenitov) i s dajkami porfyritov, dioritových porfyritov a granit-porfýrov a lamprofýrov. Jedným z typických znakov banatitov je ich výrazný vulkanoplutonický charakter s veľkým podielom hlbinných telies. V Južných Karpatoch (obr. 5) a v bulharskom Srednogorji (obr. 6) sa banatity viažu na lineamentné geosynklinálne prepadliny, súbežné s paleoalpínskou štruktúrou. Mobilitu podložia vyjadruje nielen mezoalpínske vrásnenie, ale i flyšové komplexy. Priebeh týchto prepادلínových synklinórií lemovaných hlbokosiahajúcimi zlomami v dĺžke niekoľko sto km (hrúbka kôry tenšia než u susedných zón — okolo 29 km), lineárny priebeh intruzívnych telies, rozloženie starších sedimentov najmä v okrajových častiach

predliny a prítomnosť rozsiahlych vulkanických más vnútri — to všetko naznačuje genetickú väzbu banatitov so zónou vrchnokriedového stenčenia kôry a zvýšenej magmatickej priepustnosti (A. ANATONJEVIČ—M. GRUBIČ—M. DJORDJEVIČ 1974).

Banatitové vulkanoplutonické formácie sú v prvých štádiách alkalicko-vápenaté, neskôr subbalkalické až alkalické. Pre alkalické vulkanity (olivíndiabázovej a trachyt-andezitovej formácie) sú charakteristické tzv. štítové vulkány. Lávové prúdy sú usporiadané obvykle zonálne, v smere osi štruktúr (napr. v burgasskom synklinóriu). Osi tektonických štruktúr sledujú i početné kanály, malé kupoly, silly, necky a dajky.

Vulkanizmus sa začína prejavovať už v cenomane a v turóne, vrcholí v senóne a siaha až do eocénu. Je teda typický pre mezoalpínske štádium. Intruzívne články sa často považujú za mladšie ako vulkanity — za pokriedové (S. KARAMATA 1974, M. BLEAHU 1974, D. RADULESCU—M. SANDULESCU 1973, N. HERZ—H. SAVU 1974). Rádiometrické údaje uvádzané z plutónov Vitoš a Plana v Srednogorji (S. BOYADJIEV 1974) však ukazujú na ich vrchnokriedový vek.

V Apusenách (obr. 4) zastupujú banatity prevažne hlbinné subvulkanické telesá menšej diferenciačnej škály: granodiority a granity s podradnejším zastúpením vulkanitov. Sú sprievodným javom plytkých neskorogeosynklinálnych depresií, diagonálne orientovaných k paleoalpínskej štruktúre s väzbou na hlbinný priečny zlomový systém. Nezasiahlo ich výraznejšie vrásnenie. Neskorotektonické výplne týchto depresií, a teda i magmatity považujeme za včasné neskorogeosynklinálne. Z geotektonického hľadiska tvoria medzičlánok interorogénnych typov banatitov Južných Karpát a balkánskeho Srednogorja a neovulkanitov Karpát. Eocénne vulkanity maďarského medzihoria sú ďalším článkom tejto genetickej refaze (bližším k neovulkanitom).

Priestorove vytvárajú banality oblúk (I. DIMITRESCU 1966, obr. 14b), ktorý je viac-menej súbežný s paleoalpínskymi štruktúrami v hlavnom areáli ich rozšírenia na Balkáne a v Južných Karpatoch. To zväzda k názoru, že je tu genetický vzťah medzi banatitmi a subdukciou zón s ofiolitmi. Ide o vardarskú zónu a jej východné pokračovanie (V. BOCCALETTI et al. 1973). Rodopský a Srbsko-macedónsky masív v takom ponímaní predstavovali vo vrchnej kriede ostrovný oblúk a pásmo Srednogorje — Timok zasa morský bazén s bohatou vulkanickou činnosťou typu back-arc.

Severná časť banatitového oblúku (v severnej časti Južných Karpát a v Apusenách) prebieha diagonálne k paleoalpínskym štruktúram. Tu vykazujú banatity i výrazne geotektonické ale i petrochemické odlišnosti. Pôvod magmy nemožno jednoznačne vysvetľovať zo subdukovaných ofiolitových zón, a to ani z hľadísk globálnej tektoniky. Jedni spájajú genézu banatitov so subdukciou Východných Karpát (D. RADULESCU—M. SANDULESCU 1973), iní so subdukciou pásma Metaliferes (N. HERZ—H. SAVU 1974, M. BLEAHU 1976). Pravda, ani staršie vysvetlenie vzniku banatitov subsidenčnou palingenézou (H. STILLE 1953) nie je presvedčivé vzhľadom na pomerne malú hrúbku vrchnokriedových depresií v Apusenách. Je zaujímavé, že banatity vystupujú v tých segmentoch alpíd, ktoré nevykazujú výraznejšiu polaritu a mladšia prepadlinová synklinála či tatrogeosynklinála je naložená na paleoalpínske štruktúry.

## 6. Neskorogeosynklinálne vulkanity

Neovulkanity sú geneticky späté s neskorogeosynklinálnymi neoalpínskymi depresiami (obr. 15) a to ako súčasť ich výplne, prípadne ako súčasť zložitých vulkanotektonických prepadlín a hrastí; ako kaldery a stratovulkány viazané na zlomy, ktoré určovali vývoj depresíí. V štruktúrnom zmysle sú teda neovulkanity súčasťou neoalpínskeho štruktúrneho, ale i neskoršieho morfoštruktúrneho plánu a nemajú bezprostredný vzťah k štruktúram paleoalpínskymi či mezoalpínskym (obr. 12a). Štruktúrnymi formami i postavením sa teda neskoroorogénny vulkanizmus zásadne líši od kriedovo-paleocénneho interorogénneho banatitového vulkanizmu, i keď vulkanogénne formácie sú v oboch prípadoch analogické; neovulkanity zodpovedajú osobitnému vývinovému štádiu alpíd — neskorogeosynklinálnemu.

Neskorogeosynklinálne vulkanity sú zoradené v časovom rozmere do dvoch štádií: 1. eocén-oligocén a 2. miocén, hlavne bádén-pliocén. Obidve skupiny sú viazané na depresie, stenčenie kôry a dezintegráciu. Ich priestorové rozloženie je však spravidla odlišné. Eocénno-oligocénne vulkanity sú charakteristické pre vnútorné intrageosynklinálne masívy — rodopský a maďarský masív. Časove predstavujú začiatok neskorogeosynklinálneho štádia v geosynklinálnom systéme a omladenie, či tektonické oživenie masívov. Je teda priliehavé označiť ich ako včasne neskorogeosynklinálne. Miocénno-pliocénne neovulkanity sa viažu hlavne na vnútrohorské depresie vo vnútorných zónach geosynklinálnych systémov. V časovom meradle zodpovedajú pokročilejšiemu neskorogeosynklinálnemu štádiu. Obidva typy sa priestorove prekrývajú v stykových oblastiach intrageosynklinálnych masívov s vnútornými zónami geosynklinálneho systému.

Tri areály ich rozšírenia sú geotektonicky odlišné, a to:

1. Maďarské medzihorie s priľahlými vnútornými zónami Západných a Východných Karpát, tzv. neotisia (obr. 2, 15);
2. Rodopy (obr. 6);
3. úzka zóna aktivizácie v Srbsko-macedónskom masíve a vo vnútorných dinaridách (obr. 5, 9).

Analýza vzťahu neovulkanitov k sedimentom vnútorných molás nasvedčuje, že maximum vulkanickej aktivity splyva s obdobiami blokovej subsidencie a jednotlivé vulkanické periódy s periódami formovania panví (J. SLÁVIK 1971; obr. 15). Úzka genetická spätosť neovulkanitov s neskorogeosynklinálnymi vnútornými depresiami je očividná z porovnania veku panví a neovulkanitov. Názorným príkladom toho je Rodopský masív s priabónsko-oligocénnymi panvami; miocénne panvy sú podradnejšie, to isté platí i o veku rodopských neovulkanitov. Väčší význam miocénnych panví v južných častiach Rodop (na gréckom území) sa prejavuje väčším rozsahom miocénneho vulkanizmu. U neovulkanitov a u depresíí je priama väzba na pozdĺžne a priečne zlomy. Pri pozdĺžnych zlomoch zohrali zvlášť významnú úlohu smery paralelné s okrajom vrásnenej geosynklinály. V západnej časti Maďarska a na strednom Slovensku je to hlavne smer JZ-SZ, v severovýchodnom Maďarsku, v Zakarpatskej Ukrajine a v Rumunsku sú to smery SZ-JV (J. SLÁVIK 1971, V. KONEČNÝ—J. SLÁVIK 1974; obr. 15). Pozdĺž týchto okrajových zlomov zaklesávali neogénne vnútorné depresie miestami i o niekoľko tisícov metrov. S týmito systémami súvisí orientácia celého radu vulkanických pohorí, najvýraznejšie v reťaziach; Vihorlat — Gutin a Hargita — Calimani. Výraznú úlohu zohrali i zlomy priečne, napr. v Maďarsku hlavné vulkanické centrá sledujú línie: v pohorí Börzsöny

SV-JZ a SSV-JJZ až SJ; vo Višegrade smer SZ; v pohorí Cserhát — Mátra SZ až ZV a v tokajskej oblasti SJ. Najhojnejšia vulkanická činnosť v neotisii je v bádene až sarmate, t. j. v období najväčšieho rozpadu vnútorných zón Západných Karpát, Apusen a maďarského masívu na systém menších panví, pri formovaní ktorých zohrali najvýznačnejšiu úlohu priečne zlomy. Významne sa podieľali aj na formovaní nealpínskych panví v Rodopskom masíve.

V stredoslovenskej oblasti zohrali osobitnú úlohu pri vzniku vnútorných depresí i pri rozložení vulkanických centier popri tzv. západokarpatských zlomoch sv.—jz. smeru i zlomy sz.—jv. a sj. smerov, najmä zlomy sj. smeru (Kremnické hory, Vtáčnik). Práve posledne uvedený systém zlomov vytvoril priečnu zlomovú zónu, na ktorú sa viaže i stredoslovenská neovulkanická oblasť. Pri rozložení vulkanitov v oblasti Slanských vrchov a Tokaja zohral dôležitú úlohu systém zlomov sj. smeru, tzv. hornádske zlomové pásmo. Mnohé vulkanity vystúpili v tektonických uzloch, kde dochádzalo ku kríženiu disjunktívnych línií rôznych smerov, genézy a veku (Štiavnické hory, Börzsöny; obr. 15).

Látkové zloženie neovulkanitov je pestré. V neotisii ich predstavuje úplný diferenciálny rad od ryolitov po bazaltoidné andezity až bazalty. Je to alkalicko-vápenatá provincia. Alkalický typ vulkanitov predstavujú až bazalty viazané na mladšie samostatné štádium vulkanizmu. Svedčí to o geotektonickej homogenite vývoja orogénnej oblasti v rozpätí asi 50 mil. rokov. Pestrá je paleta typov magmatických telies, od intrúzií cez subvulkanické telesá, prikráterové efúzíva a explozíva až po distálne vulkanosedimentárne fácie. Denudované centrálné časti vulkanických aparátov obnažujú vo vyzdvihnutých oblastiach intrúzie, štruktúrne i časovo späté s vulkanizmom. Podiel intrúzií je však menší ako u banatitov. V Karpatoch sú to granodioritové porfyry, diority, granodiority, granity. V rodopskom masíve predstavujú vulkanity prechodný typ od vápenato-alkalických k alkalickým, s vysokým podielom kálie (R. IVANOV 1960, 1965); charakteristická je tu andezit-trachandezit-ryolitová formácia. Sprievodné plutonity sú zaradované k gabro-monzonitovej formácii (E. DIMITROVA et al. 1975).

Zistenie, že sa, hlavne v Mátre, nachádzajú väčšie hypoabysálne telesá, a to dokonca paleogénneho veku (teleso Reczk), vyvolalo tendenciu rozčleniť vulkanickú činnosť v maďarskom medzihorí na dve etapy — na etapu paleogénnu a neogénnu (I. KUBOVICS 1974; T. ZELENKA 1974). Hľadajú sa paralely s Apusenami, kde sa odlišuje vrchnokriedovo-paleocénna etapa včasne subsekventného magmatizmu a neogénna etapa neskorosubsekventného magmatizmu (D. GIUSCA et al. 1966, M. BLEAHU 1973). O prítomnosti mezoalpínskych banatitov, teda magmatických telies staršej etapy, sa uvažuje i v Západných Karpatoch, hlavne v banskoštiavnickej oblasti slovenského stredohoria (L. ROZLOŽNÍK 1976). Pokúsme sa posúdiť opodstatnenosť takých názorov.

Rodopy (obr. 6) sú klasickým príkladom nezávislosti nealpínskeho magmatizmu počnúc vrchným eocénom na vrchnokriedovom — mezoalpínskome — banatitovom magmatizme a sú naopak príkladom súvislosti medzi vrchnoeocénno-oligocénnym a neogénnym magmatizmom. Už na str. 139 sme poukázali na výrazné rozdiely väzby banatitov a neovulkanitov na geotektonické štruktúry. U „banatitov“ pozorujeme väzbu na pozdĺžne prepadliny či tafrogeosynklinály prevažne s flyšovou výplňou zasiahnutou ilýrsko-pyrenejským, sčasti laramským vrásnením; u neovulkanitov sa banatity viažu na priečne neskorotektonické depresie, vyplnené molasami. Na Balkáne je rozdielne aj priestorové postavenie oboch typov: banatity sú v zóne Srednogorja, neovulkanity v južnejšom rodopskom masíve.

Popyrenejská tektonická a magmatická aktivizácia Rodopského masívu nadväzuje na mezoalpínsku aktivitu Srednogorja.

V Apusenách nie sú geotektonické rozdiely medzi banatitovou asociáciou a neovulkanitami také veľké (str. 140; obr. 4) ako na Balkáne. I banatity v Apusenách sa viažu na priečne orientované depresie (vyplnené hlavne gosauským typom kriedy a paleocénu, ktoré sú len slabo zvrásnené). V podstate rovnaké geotektonické pomery sú aj vo vnútorných Západných Karpatoch; vrchnokriedové sedimenty predstavujú výplň pásma menších depresíí, ktorú možno označiť ako včasné molasy. Popalealpínsky model severných Apusén je zrejme bližší modelu Západných Karpát než popalealpínsky model Balkánu. Rozdiely medzi Západnými Karpátmi a severnými Apusenami sú v type ich zázemia. K vnútorným Západným Karpatom sa primyká z juhu maďarský masív, k severným Apusenám ich južná časť — zóna Métaliferes s odlišným, palealpínskym vývojom, s dynamickejším stavom kôry v mezoalpínskej etape. Až počas neogénu sa Apuseny stali súčasťou širokého areálu neotisie a vývoj neovulkanitov v nich nadobudol spoločné znaky s vývojom vo vnútorných Západných Karpatoch i v príľahlej časti maďarského masívu.

Vrchnoeocénny a sčasti oligocénny magmatizmus rozšírený v maďarskom masíve nie je zatiaľ preukázaný v širšom rozsahu ani v Západných Karpatoch, ani v Apusenách. Naopak, vrchnokriedovo-paleocénne magmatické prejavy, rozšírené v Apusenách, nie sú zatiaľ dokázané ani v maďarskom masíve, ani v Západných Karpatoch, i keď v oboch týchto areáloch sú menšie vrchnokriedové depresie. Nesúvisia rozdiely v časových prejavoch vrchnokriedového i paleogénneho magmatizmu práve so spomínanými odlišnosťami zázemia Apusén a Západných Karpát? Nie je rozdiel medzi týmito dvoma celkami alpíd v zdroji nealpínskej aktivizácie? Maďarský masív ako zázemie vnútorných Karpát sa „aktivizuje“ po slabom palealpínskom vrásnení pomaly, výraznejšie až počas pyrenejských pohybov. Vlna aktivity sa v ňom postupne šíri smerom na sever. Južný príviesok Apusén — zóna Métaliferes, intenzívne zvrásnená palealpínskym vrásnením, zostala pohyblivejšia a vyvolala výraznejší mezoalpínsky (laramský a ilýrsky) diastrofizmus i v príľahlej severnej časti Apusén. Nealpínska tektonická aktivizácia severných Apusén, ktorej prejavom je vznik depresíí a neovulkanická činnosť, nadväzuje na intenzívny mezoalpínsky diastrofizmus, analogický ako v Rodopách, ale s inou vergenciou. Zdroj nealpínskej aktivizácie Rodop a Apusén je zrejme iný ako v maďarskom masíve a vo vnútorných Západných Karpatoch.

V maďarskom medzihorí je nápadná geotektonická nadväznosť eocénneho vulkanizmu na zónu zlomov, ktoré možno považovať za vetvy periadriatického zlomového pásma. Týka sa to predovšetkým línie Darnó, na ktorej leží i známe hypoabysálne teleso dioritov v oblasti Reck.

Rozsah neovulkanitov v jednotlivých segmentoch je nerovnomerný a ich typy nie sú všade rovnaké. Ich podiel v Alpách je rudimentárny, v Južných Karpatoch prakticky chýbajú, nevelký rozsah majú i v Dinárskom pohorí, s výnimkou oblastí príľahlých k maďarskému medzihoríu a v srbsko-macedónskom masíve. Mocné komplexy rozložené na veľkých priestranstvách, budujú neovulkanity v maďarskom medzihorí, hlavne v jeho severnej časti a v príľahlých zónach Západných i Východných Karpát. Analogickú „centrálnu“ pozíciu ako maďarské medzihorie má i transylvánska panva sprevádzaná neovulkanitmi pri východnom okraji Apusén a hlavne pri jej severnom okraji v najvnútornejšej časti rumuských Východných Karpát. I rodopský masív, hlavne jeho východná, stredná a južná časť, i časti príľahlé k srbsko-macedónskemu masívu, sú oblasťami rozsiahlej činnosti neskoroorogénneho vulka-



nizmu. Rozsiahle neovulkanity sú — ako vidieť — viazané na zóny stabilizované už predalpínskym vrásnením (centrálne masívy), rozšírené o priľahlé oblasti stabilizované paleoalpínskym vrásnením. Tieto oblasti po mezoalpínskom vrásnení vytvorili antipólové areály zónam, vyvrásneným mladšími neoalpínskymi vrásneniami. V týchto zónach dochádzalo k rozsiahlym poklesom, stenčeniu kôry (vykazujú najnižšie hodnoty hrúbky kôry) a nastalo tektonické oživenie. Klasické postavenie má v tom smere maďarský masív, ktorý sa aktivizoval už od mezoalpínskeho vrásnenia. Spolu s priľahlými vnútornými zónami orogénov vytvára osobitný krustálny blok — neotisiu (J. SLÁVIK 1971) s časovým rozpätím vulkanizmu od eocénu po pliocén.

Výraznou oblasťou neoalpínskej aktivizácie je i rodopský masív s hojnosťou potektonických depresií s neovulkanitmi. Výplň depresií sa začína flyšoidnými až flyšovými faciami, ktoré postupne vystriedali molasy. Nejde tu však o nový geosynklinálny cyklus (B. MAVRUDČIEV 1965), ale o neskorogeosynklinálne štádium balkaníd (rovnako ako v prípade Karpát a maďarského masívu). Preto i tu sotva možno považovať intruzívne telesá za synorogénne, ale za súčasť vulkanoplutonickéj formácie.

V neotisii sú zjavné prejavy migrácie vulkanizmu z jej centrálnej časti k okrajom a od západu na východ (hlavne pokiaľ ide o objemy eruptívnych hmôt), t. j. analogické s postupom orogénou (J. SLÁVIK 1968, V. KONEČNÝ—J. SLÁVIK 1974). V maďarskom medzihorí je totiž vulkanizmus od eocénu po spodný sarmat (E. SZÁDECKY—KARDOSS et al. 1967), v Slovenskom stredohorí od bádenu po vrchný sarmat až spodný panón, vo východoslovenskej oblasti a v Zakarpatsku od vrchného bádenu po pliocén (B. V. MERLIČ—S. M. SPITKOVSKAJA 1974) a v Calimani-Hargita je hlavný vulkanizmus až pliocénny (M. BORKOS 1974). Medzi orogénou a vulkanizmom je zrejme určitá súvislosť.

Výrazná polarita neovulkanitov v maďarskom medzihorí a flyšových príkrovoch Karpát už dávnejšie vedie k vysvetľovaniu vzniku neovulkanitov neoalpínskou subdukciou litosféry vo vonkajších pásmach Karpát (H. STILLE 1953). K takému vysvetľovaniu nabádajú i súvislé pásma vulkanitov vo Východných Karpatoch, kde pásma Vihorlat — Gutin, ale hlavne Calimani-Hargita sú subparalelné s neoalpínskymi štruktúrami flyšových Karpát, teda i s predpokladanými zónami subdukcie. Nečudo, že zástancami priamej genetickej súvislosti medzi neoalpínskou subdukciou a vznikom neovulkanitov sú pod vplyvom globálnej tektoniky hlavne rumunskí geológovia (D. RADULESCU—M. SANDULESCU 1974, M. BLEAHU et al. 1974, N. HERZ—H. SAVU 1975, ale i ďalší — J. SLÁVIK 1971, Z. ROTH 1957, V. BOCCALETTI et al. 1973, E. SZÁDECKY—KARDOSS 1973), ktorí odvodzujú vznik intermediárnych magiem od pohltých častí kôry. Lenže rozloženie vulkanitov Apusén i stredoslovenskej oblasti sotva možno považovať za subparalelné s predpokladanými subdukčnými zónami. Pritom v období vrcholiacej vulkanickej činnosti v bádene až sarmate sa smery rozloženia vulkanitov v susedných areáloch výrazne rozchádzajú; v severomaďarskom areáli s dominantným smerom Z—V a v stredoslovenskom areáli s dominantným smerom JZ—SV a S—J. Zdôrazniť treba i nedostatok vulkanizmu po sávskej subdukcii, azda najvýraznejšej pre Západné Karpaty. Priamy vzťah medzi subdukciou a neovulkanitmi nie je ani v rodopskom masíve, kde sú vulkanické areály často rozložené prične na pyrenejské štruktúry a sú zväčša mladšie. Tým nevylučujeme vzťahy medzi orogénou vonkajších zón na jednej strane a založením depresií a s nimi spätého vulkanizmu na strane druhej. Evidentná je alternácia kompresných pohybov so vznikom príkrovov vo flyšových Karpatoch a uvoľňovacích pohybov spojených so vznikom vnútorných a medzihorských depresií (M. MAHEE 1973). Vulkanizmus sa uplatnil len

v obdobiach činnosti hlbokých zlomov, ktoré rozťahovali kôru. V obdobiach, keď sa formovali prehybové panvy bez výraznejšej činnosti hlbších zlomov, (napr. v oligocéne severomaďarsko-južnoslovenská panva a v panóne Veľká maďarská nížina s Podunajskou nížinou) neprebíhala výraznejšia vulkanická činnosť (D. VASS 1966). K vyjasneniu vzťahov medzi vulkanizmom a tektonikou sa azda dostaneme, keď objasníme vznik depresii v Maďarskom medzihorí, ich subsidencie stenčením kôry spôsobeným subkrustálnou eróziou vyvolanou podkôrovým diapírom (P. HORVÁTH—L. STEGENA—B. GÉCZY 1974). Vznik diapíru mohli stimulovať doznievajúce horizontálne podkôrové a kôrové pohyby, ktoré až do neogénu usmerňovali vývoj geosynklinálneho systému. Vznik diapíru je viazaný so zvratom vo vývoji, keď vertikálna zložka pohybov sa rozšírila i na centrálné časti Karpát a usmerňovala formovanie celého systému.

Z uvedeného teda vyplýva, že neovulkanity vykazujú veľmi úzke vzťahy k štruktúrnemu vývoju bazénov a prejavy vulkanizmu sú viazané priestorovo i časovo na blokovú tektoniku, sčasti subsidenciu a na migráciu tektonickej mobility bazénov, vyjadrené i zmenami typu panvovej výplne. I pri svojej látkovej pestrosti neovulkanity nevykazujú zmeny chemického zloženia, prípadne užšieho vzťahu k štruktúre predpokladaných subdukčných zón, v Západných Karpatoch k bradlovému pásmu. Pritom sú v karpatskom oblúku 5—20 mil. rokov mladšie ako predpokladané subdukcie (J. LEXA).

Rozloženie neovulkanitov v alpídach naznačuje, že staré masívy uprostred alpínskej geosynklinály — panónsky, transylvánsky, srbsko-macedónsky, rodopský — poskytujú vhodné geotektonické podmienky pre väčší rozsah neovulkanitov. Všetky tieto masívy sú na rozhraní alpínskych vetví odlišnej vergencie. Zrejme hrali dôležitú úlohu pri usmerňovaní hlbinných procesov počas vývinu geosynklinály, hlavne v jej neskorom štádiu. Vtedy sa vertikálne podkôrové pohyby menili na horizontálne a práve pod starými masívmi sa dvíhal podkôrový diapír a v dôsledku toho sa omladili a užšie včlenili do alpínskeho systému. V tých častiach alpíd, ktoré nemali v zázemí ani stabilizovaný blok, rozčlenený zlomami, schopný rozpadu na kryhy, ani diferenciacie na časti vystupujúce a klesajúce (napr. Východné Alpy) nevznikli ani väčšie vnútorné panvy, ani rozsiahlejšie neovulkanity.

## 7. Finálne vulkanity — alkalické bazalty

Všeobecne sa prijíma názor o zvláštnom geotektonickom postavení alkalických bazaltov (obr. 2, 15). Vo všetkých častiach alpíd majú zhruba rovnaký vek; pliocén-pleistocén, často vek mladší ako 1 mil. rokov. Platí to tak pre areály, kde vulkanická činnosť andezitov a ryolitov skončila už koncom oligocénu, resp. v spodnom miocéne — napr. v Rodopách, ako aj pre oblasti, kde siaha až do pliocénu napr. vo Východných Karpatoch. V posledne uvedených prípadoch nadväzuje finálny bazaltový vulkanizmus na vulkanizmus neskororoogénny. Menšie výskyty bazaltov sa nájdu vo všetkých segmentoch alpíd, a to i v takých, kde boli prejavy andezitovo-ryolitového vulkanizmu minimálne alebo žiadne, napr. v Južných Karpatoch, vo Východných Alpách. V maďarskom medzihorí boli erupcie bazaltov až vo vrchnom panóne a na hranici pliocénu a pleistocénu, pozdĺž zlomov sz.—jv., sv.—jz. a sj. smerov. Predchádza im 8 až 10 mil. rokov dlhé obdobie bez výraznejšej vulkanickej činnosti. Bazalty sú rozšírené na rozsiahlej ploche najmä sz. od Balatónu a v šalgotarjánskej oblasti, sú v Západných Karpatoch hojné v lučensko-fifakovskej oblasti, v okolí Banskej Štiavnice,

Novej Bane, vo Východných Karpatoch v pohorí Persani a v Južných Karpatoch v Poiana Rusca.

Hojné žily bazaltov sú vo východných Rodopách (E. DIMITROVA et al. 1975), nájdu sa však i v západnom bloku a v Srednogorji. Subvulkanické telesá lamprobazaltov, plagiobazaltov i doleritov presekávajú naprieč v meridionálnom smere nielen rodopský masív a balkanidy, ale i moesijskú platformu (obr. 6).

V dinaridách sú známejšie bazalty Boranje a malé telesá v Slovenii (S. KARAMATA 1974). Ba i vo Východných Alpách, kde sú neovulkanity zastúpené len rudimentárne (helvet — spodnotortónske andezity a dacity v Štajersku a stopy vo viedenskej panve) sa vyskytujú bazalty a ich tufy vo vrchnom panóne v oblasti Lavanthal (P. BECK—MANNAGETTA 1974).

Všade ide o alkalické, resp. subalkalické vulkanity, najčastejšie o olivinické bazalty a ich variety (obyčajne s nefelínom) a o limburgitové bazanity. Alkalická povaha, ale i poloha zväčša na hraniciach veľkých karpatských panví zvädza k tomu, aby sa ich genéza vysvetľovala ako dôsledok prechodu od vulkanizmu typu ostrovného oblúka k vulkanizmu medzioblúkového typu (M. BLEAHU et al. 1974). Všade ide o menšie telesá štruktúrno-tektonicky málo významné, ktoré vznikli až pri ukončení intenzívneho klesania (napr. v Alfölde), v období, keď sa začali uplatňovať popri poklesových pohyboch i zdvihové pohyby jednotlivých krýh.

Nálezy vrchnoplášťových spinelových peridotitov ako xenolitov (veľkosti do 25 cm) v slovenských alkalických bazaltoch poukazujú azda na to, že sa alkalické bazalty vytavili z vrchného plášťa, zatiaľ čo xenolity predstavujú jeho nevytavený relikt (D. Hovorka v tlači).

## Niektoré zovšeobecnenia

### 1. Základná geotektonická charakteristika magmatitov

Opísané typy magmatitov — granitoidy, mladohercýnske neskorogeosynklinálne vulkanity, predalpínske včasne geosynklinálne vulkanity, alpínske ofiolity a interorogénne mezoalpínske vulkanity, neoalpínske neskorogeosynklinálne vulkanity a finálne vulkanity sa od seba líšia nielen súborom a sčasti i typom hlavných hornín, ale aj ich štruktúrnym postavením a úlohou v stavbe kôry. Vyplýva to z rozdielnych geotektonických podmienok vzniku i z odlišného typu kôry (obr. 2).

a) Granitoidné horniny sú geneticky späté zväčša s hlbinnym vrásnením a s procesom hrubnutia kôry. Prejavy granitizácie sú v jednotlivých cykloch odlišné.

Predhercýnske granitoidy, geneticky viazané na bajkalské a staršie vrásnenia, sú zväčša synkinematické, úzko späté s migmatizáciou s anatexiou i s vysokým stupňom regionálnej metamorfózy.

Najpočetnejšie a v stavbe alpíd najdôležitejšie sú hercýnske granitoidy. Príznačná je pre ne štruktúrna pestrosť, široké časové rozpätie medzi ich výstupmi a nevelká diferenciálna škála. Synkinematické granitoidy (úzko späté s plášťom) predstavujú spravidla najstaršie a najhlbšie prejavy hercýnskej granitizácie. Najhojnejšie sú batolity a fakolity. Ich vznik úzko súvisí s vrásnením a hercýnskym štruktúrnym plánom, ale i s mladšou morfofotektonickou členitosťou, ktorú do značnej miery spôsobili. Charakteristický je pre ne diapirický výstup, a to až po eróznou úroveň.

Najhojnejšie sú granodiority a granity. Petrograficky i vekom rozmanitejšie sú granitoidy v balkanidách v oblasti geosynklinálne členitejšej, s lineamentnými geosynklinálami (E. BONČEV 1966). Popri granodioritovo-granitovej asociácii je v balkanidách asociácia gabroplagiogranitová, gabrodioritovo-granodioritová a alkalická gabrosyenitová (E. DIMITROVA et al. 1975). Hojně aplity a pegmatity, ale i menšie telesá leukokrátnych granitov sú často permského veku.

Účasť alpských granitov pri formovaní alpskej štruktúry je malá (obr. 2). Sú to prevažne len menšie telesá, obmedzené na zóny predalpínskeho kryštalinika, silne paleoalpínsky až mezoalpínsky prepracované. Väčšie telesá, a to i synkinematických alpských granitoidov sa nachádzajú v južnejších častiach srbsko-macedónskeho masívu, už silne zasiahnutých kimerským vrásnením. Význačnejšiu úlohu v stavbe alpíd majú telesá mezoalpínskych granitoidov, ktoré tvoria súčasť vulkanoplutonickéj banatitovej formácie. Podiel intruzív a ich veľkosť súvisia so stavom kôry, zasiahnutej striedajúcimi sa kompresnými pohybmi (Iaramskými a ilýrskymi, resp. pyrenejskými). Menšie telesá granitov a granodioritov, hlavne hypoabysálne, sprevádzajú aj neovulkanity. Ich výstup je viazaný na uvoľnenie napätia v kôre (obr. 1).

b) Neskorogeosynklinálne mladohercýnske vulkanity sa geneticky viažu na neohercýnske (zväčša permské) žľaby a depresie vyplnené molasami. Sú oneskoreným povrchovým prejavom rozsiahlej hercýnskej sializácie kôry. Hojnejšie sú v žľaboch, ktoré oddeľujú pásma výraznejšej hercýnskej granitizácie a pri vnútorných okrajoch hercýnsky stabilizovaných areálov. Tam (vďaka hlbokosiahajúcim zlomom) došlo aj k subakválnym výlevom melafýrov zväčša typu sillov, sprevádzaných dajkami melafýrov-porfýritov a porfýritov (napr. melafýrová séria chočského príkrovu Západných Karpát a perm v príkrove Codru v Apusenách). Vrchnopaleozoické depresie severnejších zón alpíd majú molasové výplne len s malým podielom vulkanitov (kremenných porfýrov); obr. 8.

c) Predalpínske eugeosynklinálne vulkanity sú charakteristické pre podstatnú časť sérií geosynklinálnych štádií starších ako hercýnska granitizácia. Predstavujú súčasť sedimentárno-vulkanických aspidných komplexov, sčasti diastrofických. Patria prevažne spilitovo-diabázovej, spilitovo-keratofýrovej formácii a sprevádzajú ich aj intruzíva gabrodioritovo-leukogranitovej a peridotitovo-pyroxenitovej formácie. Nerovnomerné priestorové rozloženie bazických vulkanitov a väčší rozsah kyslých vulkanitov v niektorých vrchnoproterozoických sériách (napr. séria Arada), predovšetkým staropaleozoických, hlavne kambrium-ordovických, resp. ordovických (napr. zóna Grauwacken vo Východných Alpách, gelnická séria v Západných Karpatoch, séria Muncel v Apusenách. séria Tulges vo Východných Karpatoch) naznačujú čiastočnú diferenciáciu kôry počas bajkalského vrásnenia. Treba však zdôrazniť, že tieto kyslé vulkanity, prevažne kremenné porfýry a kremenné keratofýry, sú súčasťou geosynklinálnych sérií. Geneticky súvisia s ostrovnými oblúkmi, resp. s oblasťami hrubšej prechodnej subkontinentálnej kôry, rozloženými v okrajových častiach geosynklinál.

d) Mezozoické ofiolity vystupujú v pásmach, ktoré zaraďujeme na základe prítomnosti hlbokovodných facií ako trógové, s tenkou, prípadne roztrhanou sialickou subkontinentálnou resp. paraoceanickou kôrou. Objavujú sa už v triase (hlavne v ladine), často v sprievode plytkovodných vápencov. Spočiatku (v anise) predstavujú hybridný, prevažne intermediárny typ s prevahou spilitov a keratofýrov, geneticky spätých s lámaním a diferenciáciou kôry. S postupným narastaním trógov v ladine, a hlavne v karne sú pre ne typické hlbokomorské karbonáty, rádiolarity a pelity. Rastie aj podiel bazickejších hornín a objavujú sa ultrabázické

telesá. U väčšiny segmentov alpíd sa triasové vulkanity viažu na najvnútornejšie zóny alpínskeho geosynklinálneho systému, t. j. na zóny slabšie konsolidované hercýnskou granitizáciou. V Západných Karpatoch je to meliatska séria v Slovenskom krase, analogická séria v pohorí Bükk, séria Vascau v Apusenách, na Balkáne Sakar, vo Východných Karpatoch transylvánske príkrovy. V dinaridách sa už v triase vulkanity ofiolitového typu viažu nielen na vnútorné zóny (napr. vardarská a ofiolitová), ale sčasti i na vonkajšie zóny (hlavne Budva—Zukali). Súvisí to s menšou intenzitou hercýnskeho vrásnenia i granitizácie v dinaridách, kde má dokonca i perm prevažne morské fácie.

Hojnejší výskyt bázických a ultrabázických vulkanitov je ako súčasť ofiolitovej asociácie viazaný na jurské a spodnokriedové trógy, charakterizované slieňovcovo-rádiolaritovými fáciami. Sú rozložené hlavne v stredných pásmach interníd (obr. 13), napr. križňanský a kysucký tróg Západných Karpát, bukovinský tróg Východných Karpát, trógy penninika Álp, v Apusenách zóna Mureš. V dinaridách a helenidách sú ofiolity viazané hlavne na vnútorné zóny (vardarskú, ofiolitovú, Mirdita — subpelagonikum). Častejšie sprevádzajú ofiolity i mocné preflyšové fácie (titón — spodná krieda, napr. v jednotkách Rachov — Ceahlău vo Východných Karpatoch, Severín — Kraina v Južných Karpatoch, Strandža na Balkáne). V Mureši a vo vardarskej zóne siaha ofiolitová séria od hlbokovodných vrchnojurských karbonatiko-silicitických facií do flyšových.

e) Mezoalpínske intraorogénne vulkanity a sprievodné plutonity banatitového typu sú v balkánskom Strednogorí a v timockej zóne Južných karpát geneticky späté s pozdĺžnymi prepaddinovými synklinóriami (obr. 14b), vyplnenými predmolosovými, prevažne morskými sériami, so značným podielom flyšu. Pritom je príznačné, že sa vulkanity objavujú s postupnou oceanizáciou typu sedimentov. Pod mocnými vulkanickými komplexmi sú včasné molasy, nad nimi hrubý flyš. Vulkanity širokej diferenciácie škály (ryolity — dacity, trachyty, andezity, alkalické bazalty) sa podieľajú na vulkanosedimentárnej výplni priekopových prepaddin a synklinórií. Popri nich vystupujú hojne, najmä pri okrajoch synklinórií, subvulkanické i hlbinné telesá rovnako širokého petrochemického rozpätia (pikrity, gabrá, diority, monzonity, gabrosyenity, granosyenity, granity, dioritové porfyrity a granit-porfýry). Ide o vulkano-plutonické formácie, v prvých štádiách alkalicko-vápenaté, neskôr subalkalické až alkalické. V Apusenách zastupujú banatitovú formáciu prevažne hlbinné telesá (granodiority a granity s podradnejším zastúpením vulkanitov). Neviažu sa na výraznú pozdĺžnu tektonicko-prepaddinovú synklinálu, ale na hlbinný priečny zlomový systém. Nezasiahlo ich ani výraznejšie vrásnenie (obr. 4).

f) Neskorogeosynklinálne trefohorné neovulkanity sa viažu na mladé (pozdĺžne i priečne) depresie, naložené na vnútorné, paleoalpínsky stabilizované zóny a hlavne na ich zázemie (medzihorie), neskôr tektonicky oživené pri súčasnom stenčení kôry (obr. 15). Vysokým podielom neovulkanických pyroklastík je výrazne ovplyvnená molosová výplň depresii. Charakteristickou črtou neovulkanitov (s prevahou intermediárnych hornín, hlavne andezitov a so značným podielom ryolitov) sú vulkanotektonické formy, hlavne stratovulkány, kaldery. Vulkanizmus sa uplatňuje v postupnej izolácii panví a výrazne sa prejavuje geomorfologicky. Podiel sprievodných hlbinných telies je malý.

g) Finálne pliocénno-pleistocénne bazalty menšieho rozsahu, zväčša subalkalické až alkalické, sú produktom osobitného štádia vo vývine geosynklinály, keď výrazne pôsobilo rozťahovanie kôry pri zvýšenom účinku vertikálnych pohybov (obr. 2).

## 2. Vývinový trend alpinskej geosynklinály a pozícia magmatitov

Typy a rozloženie magmatitov sú v priamom vzťahu s vývinom alpinskej geosynklinály a kôry, čo sa odráža v geologických procesoch (v sedimentácii, vrásnení, metamorfizme i magmatizme; obr. 16, 17).

Až do karbónu, t. j. do nástupu hercýnskej granitizácie a metamorfózy sa formácie vyznačujú menšou faciálnou pestrosťou. Predkarbónske sedimentárne série mali prevahu grauwackov, pieskocov, aleurolitov a ílovcov, sčasti diastrofického (flyšového), častejšie však aspidného charakteru. Platí to v podstate pre celé kryštalinikum alpíd (M. MAHEL 1974). Pre Západné Karpaty je príznačný malý podiel karbonátov v kryštaliniku. Hlavným činiteľom spôsobujúcim pestrosť monotónnych sedimentárnych komplexov sú popri metamorfóze vulkanity. U väčšiny predkarbónskych sérií sú hojne zastúpené vulkanické horniny hlavne spilitovo-diabázovej, podradne spilitovo-keratofýrovej formácie, zmenených zväčša na amfibolity. Častejšie sú gabrá, gabrodiority, ale i komagmatické ultrabázické telesá. Len v niektorých sériách sú hojnejšie porfýroidy (kremenné porfýry a ich tufy), najčastejšie v ordoviku, zriedkavejšie i v devóne. Pritom niet výraznejších rozdielov v zložení medzi staropaleozoickým komplexom a staršími komplexmi, azda s tou výnimkou, že v proterozoických a v starších komplexoch sú častejšie ortoruly a migmatity a spravidla vyššia metamorfóza. V podstate to znamená, že kadomské, resp. bajkalské vrásnenie nespôsobilo výraznejšie zmeny ani v podmienkach sedimentácie, ani v magmatizme. Toto vrásnenie vyvolalo skôr zmeny hlbinného charakteru bez výraznejších morfoštruktúrnych prejavov. Hojnejšie kremenné porfýry v geosynklinálnych sériách koncom proterozoika, hlavne však v ordoviku, sú azda prejavom diferenciacie kôry (napr. vytvorenia ostrovných zón, prípadne zón s hrubšou kôrou (M. MAHEL 1974).

Odlíšny charakter majú predkarbónske komplexy v pelagómskom a hlavne v rodopskom masíve i prítomnosťou mocných máz karbonátov častejšie s bázickými telesami. Považujú sa za proterozoické, a to vďaka vyššej metamorfóze, hlavne v porovnaní so spodnopaleozoickou diabázovo-fylitoidnou sériou. Rozloženie týchto sérií prevažne vedľa seba a nepochopiteľný nedostatok staropaleozoických sekvencií patfomného typu v uvedených masívoch vnucuje otázku, či karbonátový komplex nepredstavuje aspoň sčasti typ plytkovodných staropaleozoických sledov.

Prítomnosť kremenných porfýrov (spravidla v sprievode bázických telies) v geosynklinálnych sériách devónu až spodného karbónu (hronská séria vo veporidách Západných Karpát, niektoré série v Južných Karpatoch) poukazuje na účinky diferenciacie kôry, ktorá prebiehala súčasne s nástupom hercýnskeho vrásnenia. Sedimentácia v starom paleozoiku nebola všade plynulá. Sú známe diskordancie i paleogeografické zmeny i nielen uprostred vrchného proterozoika (kadomská fáza), ale i uprostred kambria (srbská fáza), koncom kambria (sardinská fáza), koncom ordoviku (takonská fáza) i koncom silúru (ardenská fáza). Oddelenosť kaledónskeho a hercýnskeho cyklu na Balkáne a v Južných Karpatoch sa prejavuje nielen bazálnymi zlepenkami, diskordanciami, ale oproti starším sériám i slabšou metamorfózou (S. NASTASEANU 1975). Lenže ani po kadomskom vrásnení, ani po prejavoch kaledónskeho diastrofizmu sa neprejavuje výraznejšia zmena v type kôry. Eugeosynklinálny typ, blízky oceanickému, pretrváva až do hercýnskeho vrásnenia. Až toto vrásnenie, sprevádzané rozsiahlym výstupom granitoidných telies, mení charakter sedimentácie (v

mladšom paleozoiku sa prvýkrát objavujú v alpidách naložené depresie vyplnené molasami, spočiatku morskými, včasnými, neskôr kontinentálnymi) a kôry.

Následkom tvorby veľkých batolitov a fakolitov i ich pozvoľného diapirického vystupovania, vzniká osobitný tektonický štýl. Vytvárajú sa široké vrásy, megaantiklinálne klenby (napr. v tatridách v niektorých jadrových pohoriach Západných Karpát, brachyantiklinály rodopského a srbsko-macedónskeho masívu). Oblasti menej zasiahnuté granitizáciou sa vyznačujú zónami úzko stlačených vrás i prešmykov (napr. v kryštaliniku Čiernej hory v Západných Karpatoch, v bystrickej skupine príkrovov vo Východných Karpatoch; H. G. KRÄUTNER 1972). Hercýnske vrásnenie sa spolu s tvorbou granitoidných telies výrazne uplatňovalo pri formovaní celkového štruktúrneho plánu.

Väčšie granitoidné telesá s tendenciou k zdvihui sú dlhodobým morfoštruktúrnym činiteľom. Klenby sa stávajú znosovými oblasťami. Klasktické horniny z nich vyplňujú príahlé žľaby sformované pri okraji, resp. medzi zónami granitoidných telies.

Hercýnska granitizácia má rozpätie od konca devónu do konca permu; je však sústredená hlavne v jednotlivých orogenetických štádiách hercýnskeho vrásnenia. Štruktúrnoverné prejavy súvisia so strašími štádiami, morfoštruktúrne prejavy zasa s mladšími štádiami, t. j. s tvorbou molasových žľabov a depresí s vulkanitmi. V sprievode spodnej molasy, v pásmach slabšie zasiahnutých hercýnskou granitizáciou, sú v karbóne časté bázické telesá, v gemeridách dokonca i drobné telesá ultrabázické. V oblastiach intenzívnejšie zasiahnutých granitizáciou sú kremenné porfýry v sprievode mladšej (suchozemskej) molasy. Časove sa kryjú s najmladšími prejavmi granitizácie v príahlých zdvihových oblastiach. Úzku genetickú nadväznosť kremenných porfýrov na granitoidy naznačuje prítomnosť granitporfýrov (napr. v severoveporidnom žľabe Západných Karpát), ale i kyslejší charakter magmy (kremenných porfýrov, permských aplitov, pegmatitov a leukokrátnych granitov, napr. v Rodopskom masíve a v balkanidách — E. DIMITROVA et al. 1975).

Permské kremenné porfýry sú teda povrchovým prejavom magmatizmu, viazaným na tektonické prepadliny s výraznou tendenciou klesania (obr. 8). V južnejších zónach v perme vystupujú v sprievode vrchných molás i melafýry. Následkom slabej produkcie pyroklastík sa melafýry podielajú na sedimentačnej výplni panví oveľa menšou mierou. Zóny s melafýrmi nevykazujú známky výraznejšieho prehĺbenia. Nejde o stenčenie kôry pri ich výstupe, ale o jej roztrhnutie, otvorenie prírodných kanálov do hlbších častí. Nasvedčuje tomu i charakter telies typu sillov, i striedanie melafýrov a kremenných porfýrov v niektorých oblastiach, napr. v Čiernej hore v Západných Karpatoch, v balkanidách, hlavne však v Apusenách (v jednotke Codru).

Hercýnska granitizácia geneticky úzko súvisí s metamorfózou nerovnomernej intenzity. Napr. v Západných Karpatoch, v zónach výraznejšej granitizácie, je staršie paleozoikum metamorfované do fácie amfibolitov, zóny menej postihnuté granitizáciou sú metamorfované len do fácie zelených bridlíc. Intenzita metamorfózy od štádia k štádiu počas hercýnskeho vrásnenia klesá, jej slabšie prejavy sú badateľné ešte i v perme. Rekryštalizačná metamorfóza a granitizácia svedčia o hlbinnom charaktere hercýnskeho vrásnenia. Pritom sa však vytvorila zložitá viacetážová štruktúra so zjavnými diskordanciami. Hlavným dôsledkom hercýnskeho vrásnenia, granitizácie a metamorfizmu je však zhrubnutie kôry — jej sializácia, a to znamená zásadný zvrät vo vývine geosynklinálneho systému. Pritom treba zdôrazniť nerovnomernosť hercýnskej granitizácie i stabilizácie v jednotlivých areáloch i zónach, čo sa stáva základom odlišnosti a do značnej miery i členitosti neskoršej alpínskej geosynklinály (obr. 7).

Začiatok alpínskeho cyklu sa spája s transgresiou triasu, útvaru, ktorému sa často pripisuje v alpidách kvázi platformný charakter (V. CHAIN 1972), príp. sa nazýva štádiom interregna medzi hercýnskou a alpínskou etapou, ktorá sa podľa niektorých autorov začína až jurou. V triase prevládajú fácie plytkovodné — v spodnom triase detritické, v strednom a vrchnom triase hlavne karbonatické. V južných zónach je už výrazná štruktúrno-faciálna zonálnosť a objavujú sa i fácie a celé sledy s prevahou hlbokovodných facií s bázickými vulkanitmi ofiolitového typu, napr. meliatska séria (H. KOZUR—R. MOCK 1974), hlbokomorské vulkanosedimentárne členy v pohorí Bükk (obr. 16) a v transylvánskych príkrovoch Východných Karpát (D. RADULESCU—M. SANDULESCU 1973). Prítomnosť bázických a ultrabázických telies v sprievode rádiolaritov a tmavých bridlíc, glaukofanity a prejavy metamorfózy, to všetko naznačuje oceanický resp. paraoceanický typ kôry (M. MAHEL 1975). Pozoruhodné je rozloženie týchto zón v južných oblastiach Karpát i Balkánu, v dinaridách (obr. 3, 4, 6, 9, 11a) sú i v niektorých vonkajších jednotkách — Budva—Zukali. V každom prípade ide o zóny bez výraznejšej hercýnskej granitizácie a výraznejšieho hercýnskeho vrásnenia, teda o zóny hercýnsky menej konsolidované. Z tých sa šíri aktivizácia geosynklinály a postupuje smerom von. Už v triase teda pozorujeme diferenciaciu v type kôry.

Hlbokomorské sledy jury sú spravidla sprevádzané početnými prejavmi bázického a ultrabázického vulkanizmu, napr. augititmi. Často sa totiž viažu na hlbokovodné fácie trógov, ako je napr. zliechovský a kysucký tróg v Západných Zarpatoch, frankenfeldský tróg vo Východných Alpách (obr. 16). Častejšie sú bázické vulkanity v bukovinskom príkrove Východných Karpát, hlavne však v dinaridách a helenidách. Západné Karpaty môžu poslúžiť ako model paleogeografickej členitosti, ale i paleotektonickej kontrastnosti s najvýraznejšími rozdielmi v hrúbke kôry v jure. Preukázateľné vzájomné laterálne prechody prehĺbeninových a prahových sledov a rýchle vertikálne zmeny hlbokovodných a plytkovodných facií sú napr. v bradlovom pásme Karpát. Je tam celý rad sledov opisovaných ako osobitné výviny, so zmiešanými hlbokomorskými i s prechodnými faciami. Obdobná situácia je v krížňanskom príkrove, kde je vyvinutý hlbokovodný zliechovský i plytkovodný vysoký typ sledov. Išlo teda o silne členité more s pomerne úzkymi žlabmi a prahmi. Ofiolitové zóny nemusia predstavovať široké oceány a nie sú samy osebe svedectvom ďalekosiahlych zblžiení kryh kontinentálnej kôry, sú však svedectvom členitosti mezozoickej geosynklinály, hlavne v jej leptogeosynklinálnom štádiu.

Značná časť ofiolitov je geneticky spätá so začiatočným štádiom flyšových trógov. Zrejme je dosť všeobecným javom, že sú mnohé titónsko-spodnokriedového veku: ceahläu — rachovský tróg vo Východných Karpatoch, severínsky tróg v Južných Karpatoch, nižsko-trojanský a strandžanský na Balkáne, sarajevský tróg v dinaridách, Gramos v helenidách (M. MAHEL 1973; obr. 13). Všetky sa začínajú preflyšom. Prítomnosť ofiolitov, prípadne bázických a ultrabázických telies (napr. v jednotke Ceahläu-Rachov) svedčí o tom, že ide o oceanický typ kôry, prípadne jemu blízky (obr. 14a). Už sme uviedli, že niektoré trógy s ofiolitmi leptogeosynklinálneho štádia prerastajú vo flyšové. Napokon takým je i peninulum Alp popri vardarskej zóne dinarid a zóne Mureš v Apusenách. Diastrofický typ sedimentácie (flyš s.s. a hrubý flyš) prevláda v alpidách od nástupu paleoalpínskych procesov vrásnenia až po začiatočné neoalpínske pohyby (M. MAHEL 1973). Sprevádza ho len malý podiel magmatitov. Počas sedimentácie flyšu dynamicky pôsobila v kôre hlavne kontrakcia. Pokiaľ sú prítomné vulkanity, ide o slabšie prejavy intermediárneho vulkanizmu; tufy, tufity a brekcie andezitov i dacitov. Ich erupcie sú viazané hlavne na prikordilierové zlomy.



Osobitným paleotektonickým typom vytvoreným po paleoalpínskom vrásnení je sústava prepادلínových synklinórií: Timok-Srednogorje v Južných Karpatoch a na Balkáne (obr. 14b). Tento typ je sprevádzaný hrubým mezoalpínskym, látkovo pestrým intermediárnym magmatizmom. Treba však zdôrazniť, že ide o osobitný typ vulkanitov viazaný na paleoalpínsky intenzívne zvrásnené podložie po uložení včasnej molasy (t. j. po cenomane — turóne so slojmi uhlia). V priebehu vulkanickej činnosti, hlavne po výlevoch lávy, nastupuje hrubý flyš s menšími polohami vulkanitov pri okrajoch tektonických prepادلín s intruzívnymi telesami. Vo vlastnej flyšovej geosynklinále, ktorú predstavovala na Balkáne susedná severnejšia ludokamčijská zóna, sú však prejavy tohto vulkanizmu len slabé.

Alpínsky cyklus alpíd sa vyznačuje i hojnosťou depresíí viacerých geotektonických typov:

— depresie pásma čelnej predhlbne, ktoré nadväzujú časove, priestorove a geneticky na flyšové jednotky: sú bez výraznejších prejavov vulkanizmu;

— depresie vnútrohorské, naložené hlavne v priečných prelomoch vnútorných zón alpíd, konsolidovaných prevažne už paleoalpínsky;

— depresie vnútorné (pri vnútornom okraji geosynklinálneho systému), presahujúce na príslahlé vnútorné masívy v zázemí (maďarský, transylvánsky, rodopský).

Obidva posledne uvedené typy depresíí sa vyznačujú častými vulkanitmi a štadiálnosťou vývoja. Tá v Západných Karpatoch a maďarskom masíve časovo alternuje so vznikom príkrovových jednotiek vo flyšových Karpatoch. Tak, ako vyvrásnené jednotky sávske, staroštajerské, mladoštajerské, i depresie sú rovnakého veku. Neoalpínsku etapu v geosynklinálnom systéme charakterizuje vo vonkajších pásmach skrakovanie a hrubnutie kôry, vo vnútorných oblastiach rozťahnutie, lámanie a stenčenie kôry, subsidencia a vulkanická činnosť. Vznik vnútorných molás sa považuje za neskorogeosynklinálny, preto je azda priliehavý i názov neskorogeosynklinálne vulkanity (obr. 15a, 15b).

Hojnosť útvarov, ktoré vznikli pri zvýšenej pohyblivosti kôry, ich faciálna i geotektonická pestrosť, sú odrazom štadiálnosti alpínskeho vrásnenia, spojenej s prestavbou paleogeografického i štruktúrneho plánu a s viacaktnosťou paleoalpínskych, mezoalpínskych a neoalpínskych prejavov. Táto štadiálnosť alpínskeho vrásnenia je kombinovaná s orogenetickou polaritou, ktorá je zrejme genetickým znakom alpínskej geosynklinály. Jej dôsledkom je postupné vertikálne i laterálne vystriedanie flyšu molasami (obr. 17).

Súčasnosť poklesov v depresíách, ale i súhra vulkanickej činnosti vo vnútorných zónach a v zázemí s kompresnými procesmi na vonkajšej strane Karpát, signalizuje určité vzťahy podkôrových procesov v celej šírke geosynklinálneho systému. Zmeny spôsobené zahĺbením, zakotvením kôrových častí pri vrásnení vyvolávali zrejme výstup magmatického diapiru na vnútornej strane (P. HORVÁTH—L. STEGENA—B. GÉCZY 1974). S diapírom sa totiž spája stenčenie kôry podkôrovou eróziou, v dôsledku toho poklesy a vznik depresíí i výstup vulkanitov. Všetky tieto pohyby vo vnútorných pásmach sú ovládané prevažne vertikálnymi pohybmi, ktoré utvárajú i celkovú morfoštruktúru. Neskorogeosynklinálny vulkanizmus sa svojimi vulkanotektonickými i morfológickými formami podieľa i na tomto procese. Teda nie granitoidy — ako v mladohercýnskej etape — ale efuzívne neovulkanity sa z alpínskych magmatitov najviac podieľajú na formovaní morfoštruktúrneho plánu.

Alpínske vrásnenie sa vyznačuje rozsiahlym skrátением kôry, pestrosťou tektonických štýlov (dôsledok členitosti a rozdielnosti hrúbky a mechanických vlastností látkovej výplne jednotlivých sedimentačných zón) (obr. 16). Najvýraznejším prejavom tohto skrátения sú pripovrchové i hlbinné príkrovy, intenzívne stlačené zóny, zóny štruktúrnych anomalí, tektonických

brekcií a melanží, naznačujúce subdukciu. Granitizácia a metamorfóza, také charakteristické prejavy hercýnskeho vrásnenia, sa obmedzujú v alpínskom cykle len na úzke zóny. Prejavy alpínskeho vrásnenia majú zrejme výraznejší pripovrchový charakter.

Uvedený vývinový trend geosynklinály sa vzťahuje predovšetkým na oblasť Východných Álp, Západných Karpát a Východných Karpát. V dinaridách má určité odlišnosti, spôsobené hlavne iným východiskovým stavom kôry, v areáli Južných Karpát — Balkánu sú tieto odlišnosti ešte výraznejšie.

### 3. Osobitosti magmatizmu v jednotlivých oblastiach a segmentoch

Napriek jednotnému vývinovému trendu magmatizmu sú medzi jednotlivými segmentmi alpid značné rozdiely v kvantite jednotlivých typov magmatitov i v jeho časových prejavoch. Príčinou toho je hlavne nerovnomerná hercýnska konsolidácia kôry (vrátane nerovnakého podielu hercýnskych granitoidov). Odráža sa to v odlišných typoch permu a triasu, a hlavne v členitosti geosynklinály v mezozoiku, v čase a množstve prejavov ofiolitového vulkanizmu i ostatných typov magmatitov. Ako osobitné typy alpínskych geosynklinál, odlišné zastúpením a rozložením jednotlivých skupín magmatitov možno vyčleniť tri areály (obr. 2).

Dinaridy. Vyznačujú sa malým podielom hercýnskych granitoidov, ktoré sú sústredené pri okrajoch v pelagónskom a srbsko-macedónskom masíve (obr. 9).

Malý rozsah hercýnskych granitoidných telies a menšia intenzita hercýnskeho vrásnenia mali za následok nižšiu konsolidáciu kôry a zrejme i jej menšiu hrúbku. S tým zrejme súvisí aj prevažne morský vývin permu a zvýšená priedušnosť kôry v triase, hojnejšie prejavy bázického magmatizmu, a to i vo vonkajších zónach, v Budva—Zukali. Spočiatku (anis) majú magmatity hybridný charakter s väčším podielom kyslejších typov: porfyrity, keratofýry, kremenné keratofýry, ba i kremenné porfýry. Postupne však (ladin) dochádza k zvyšovaniu bázicity. Častejšie sú bazaltové poduškové lávy typu „pillow“; výraznejšie sa prejavuje i trógový typ sedimentácie so silicitmi, kremíťmi vápencami a ílovcami.

Nijaký z opisovaných segmentov alpid nevykazuje také rozsiahle zastúpenie ofiolitovej asociácie, a hlavne toľko a takých veľkých telies serpentinitov ako dinaridy a helenidy (obr. 13). Pritom sú rozložené v dvoch pásmach: východnom vo vardarskej zóne a v západnom v ofiolitovej zóne a v zóne Mirdita (subpelagonikum), oddelených pásmom hrubšej kontinentálnej kôry: Drina — Ivanjica — pelagónsky masív so zónou Korab. Každé z týchto pásiem má osobitosti v stratigrafickom rozpätí ofiolitovej asociácie i v obsahu; každé predstavuje v podstate oceanický tróg odlišného typu. Uvedené odlišnosti sa však menia i v jednotlivých úsekoch toho istého trógu, napr. v zóne Mirdita nadväzuje ofiolitová asociácia na vulkanogénnu sériu triasu, kým v podstatnej časti severnejšej ofiolitovej zóny sú magmatity až vrchnojurské. naopak, v severozápadnom pokračovaní ofiolitovej zóny v Južných Alpách ofiolitová asociácia chýba a zastúpené sú len hybridné triasové magmatity. Mocné ofiolitové melanže (M. DIMITRIEVIĆ 1974), ale i vzájomné prechody hlbokomorských ofiolitových sledov do sledov plytkomorských patria k osobitostiam dinaríd a heleníd, opäť s odlišnosťami v dvoch hlavných zónach.

Dve opisované ofiolitové oceanické trógy dinaríd majú odlišnosti vo vzťahu k flyšovým sekvenciám (obr. 14a). Vo vardarskej zóne nadväzuje flyš vertikálne na ofiolitovú asociáciu. Pritom jeho spodné členy obsahujú magmatity ofiolitového typu. V ofiolitovej zóne a v zóne

Mirdita sa flyšový tróg Sarajevo — Gramos vyvíja od titónu pri západnom okraji ofiolitového koryta — laterálne naň nadväzuje a v neskoršom období sa po zániku ofiolitového trógu rozširuje na západ ako zóna Durmitor.

Charakteristická pre dinaridy a helenidy je výrazná alpínska tektonická aktivizácia ich stykovej zóny a východného Srbsko-macedónskeho a pelagónskeho masívu. Častejšie sú syenitové dajky a granodioritové telesá, a to i synkinematické (v helenidách, starokimerské 167—156 mil. r., mladokimerské 148—120 mil. r., ale i strednokriedové 110—90 mil. r.).

Vnútorňý okraj dinaríd, vrátane srbsko-macedónskeho masívu, hlavne jeho západného okraja je i zónou rozloženia početných menších telies neovulkanitov s menšími vulkanickými aparátmi (andezity — dacity, ryolity, kremenné latity) zato s početnými subvulkanickými telesami (granodiority a kremenné monzonity) sú viazané na zónu výraznej nealpínskej aktivizácie. Jednotlivé vulkány sú častejšie na križovatkách viacerých zlomových systémov. Miestami sprevádzajú menšie depresie. Ide o zónu aktivizácie, veľmi členitú, s výraznými zdvihmi menších kryh (M. ARSOVSKI 1961). Pritom interval vulkanickej činnosti je pomerne dlhý (37 mil. r. — 10,9 mil. r., t. j. oligocén — sarmat).

Balkán a Južné Karpaty. Pre túto oblasť je prakticky v celom rozsahu príznačná hercýnska granitizácia, ktorá ich spája s rodopským a srbsko-macedónskym masívom (obr. 5,6). Tak v Južných Karpatoch, ako i na Balkáne sú pomerne značne zastúpené synkinematické granitoidy, a to predpaleozoické i hercýnske. V Južných Karpatoch, hlavne v danubiku, sú časté i granitoidy alkalického charakteru. Pre Balkán je zvlášť charakteristická pestrosť formačných, geotektonicky odlišných typov granitoidov. Popri južnobulharských granodioritoch a granitoch, hojných v rodopskom masíve a v bulharskom Stredohorí, sú to: staroplanínska gabrodioritovo-granodioritová asociácia, strumská asociácia s petrografickým rozpätím od kyslých granitov až po ultrabázické telesá, permská gabro-syeintová asociácia. To poukazuje na väčšiu geotektonickú členitosť Balkánu v priebehu paleozoika.

Hercýnska granitizácia i vrásnenie, čiže konsolidácia kôry bola intenzívnejšia než vo všetkých opisovaných segmentoch alpíd, takže práve pre Južné Karpaty a balkanidy je najpriliehavejšie hovoriť o kvázi platformnom type triasu (obr. 7). Heterogénnosť predalpínskeho vývoja na Balkáne, vyjadrená pestrosťou typov granitoidov, sa však prejavila v triase vznikom prvých geosynklinálnych trógov — oblastí s tenšou kôrou vo východnej časti Balkánu, a to na juhu v zóne Sakar, severnejšie v zóne Kotel.

Pre alpínsky vývojový cyklus Balkánu a Južných Karpát je príznačná členitosť geosynklinály v titóne pri nástupe flyšových formácií. S nimi súvisí prítomnosť serpentinitov pri okrajoch rodopského masívu a v Strandži (hlavne pozdĺž marického zlomu) a v trógu Severín-Kraina (v Južných Karpatoch).

V magmatickom vývoji tohto areálu osobitnú úlohu zohrali pozdĺžne zlomy (vrchnokriedový rift Timok — Srednogorje) i sprievodný systém prepادلínových synklinórií s banatitovou vulkanicko-plutonickou formáciou, ktorý sleduje rozhranie dvoch základných blokov v Južných Karpatoch: (getika a danubika), na Balkáne zlomové rozhranie balkaníd a rodopského masívu.

Pre Balkán sú príznačné neovulkanity spojené s rodopským masívom (vytváral balkanidám zázemie) a s vnútornými kotlinami (v priľahlej tylovej časti geosynklinály). Túto zviazanosť dokumentuje stratigrafický rozsah vulkanitov i výplne depresii (v bulharskej časti hlavne priabón — oligocén, na gréckom území i miocén (obr. 15). Tylová aktivizácia geosynklinály sa v rodopskom masíve zrejme šírila na juh. V Južných Karpatoch je nedostatok neovulkanitov

i významnejších vnútorných kotlín. Výnimkou je južná časť Východných Karpát — Krajštidy, ktoré zdieľali osud balkaníd, resp. rodopského masívu.

Západné Karpaty (obr. 3), Východné Alpy, Východné Karpaty (obr. 11a) a Apuseny (obr. 4). Hercýnska granitizácia a intenzívnejšie hercýnske vrásnenie postihli hlavne severné zóny tohto areálu. Pre Západné Karpaty (gemeridy) sú typické paleoalpínske granity (obr. 10). Areál vykazuje určitú jednotnosť čo do zastúpenia kremenných porfýrov zo začiatku paleozoika (dôsledok bajkalskej diferenciácie kôry v Západných Karpatoch, Východných Alpách, Východných Karpatoch i v Apusenách). Rôzny stupeň hercýnskej konsolidácie sa odráža v značnej rozmanitosti facií permu, ktorý obsahuje nielen kremenné porfýry, ale i melaťfýry (obr. 7). Najjužnejšie oblasti vykazujú morský typ permu a faciálne pestrý karbonatický trias. Možno to pokladať za prejav diferenciácie kôry. V týchto južných zónach sú častejšie bázické telesá, ba i ofiolitové série už v období triasu, a to v transylvánskej skupine príkrovov Východných Karpát, v Slovenskom krase (meliatska séria), v pohorí Bükk v Západných Karpatoch. Azda i preto sa pohorie Bükk (na rozdiel od južnejších zón Severných vápencových Álp s morským typom permu a triasu a hojnými vulkanitmi) považuje za odnož dinaríd, ktorá nadväzuje na vnútorné zóny Západných Karpát (K. BALOGH—L. KÖRÖSSY 1968). Po preukázaní analogického ofiolitového sledu v Slovenskom krase (meliatska séria) a celého radu ďalších „južnejších“ znakov Slovenského krasu, je však zrejmé, že pohorie Bükk je neoddeliteľnou časťou Západných Karpát a spolu so Slovenským krasom predstavuje ich najvnútornejšiu zónu, zv. bukovikum, s niekoľkými čiastkovými jednotkami (M. MAHEL 1975). Rožňavská línia (pri severnom okraji Slovenského krasu) a balatónska línia (pri južnom okraji pohoria Bükk) sú hraničnými líniami tejto južnej zóny s ofiolitmi. Obe spolu s líniou Darnó (pri severnom okraji pohoria Bükk) predstavujú severné odnože insubricko- -drávskej línie. Táto línia sa však smerom na SV nielen rozvetvuje, ale mení aj svoje postavenie a význam. Južne (jv.) od nej nie je ďalšia geosynklinálna vetva, ako sú v západnej časti dinaridy a Južné Alpy, ale maďarský masív.

Vo Východných Alpách je rozsah bázických telies v južných zónach obmedzený na ojedinelé výskyty vo vyšších príkrovoch vápencových Álp a na polohy v strednom triase v „Drauzóne“.

Zvlášť výrazná diferenciácia kôry vnútri geosynklinálneho systému Západných Karpát, Východných Álp a Východných Karpát bola v jure a v spodnej kriede, t. j. v leptogeosynklinálnom štádiu hlavne v stredných až severnejších zónach alpíd. V trógoch sú početné bázické telesá, ale i nekompletné ofiolitové sledy (ofiolitoidy): (zliechovský a kysucký tróg v Západných Karpatoch, bukovinský tróg vo Východných Karpatoch). Pre Západné Karpaty sú charakteristické hojné drobné výskyty augititov a limburgitov v plytkovodných faciách (vysokotatranská jednotka v Osobitej, výskyty v obale Nízkych Tatier), zrejme zviazaných so zlomami oddeľujúcimi prahy. Súvisia azda s blokovitou stavbou vnútorných Západných Karpát. V Alpách je známa výrazná ofiolitová formácia v penniniku (vo východnej časti Álp v sérii Rechnitz). Vo Východných Karpatoch sú ofiolity charakteristické pre preflyšové štádium jednotiek rozložených severne od Marmarošského masívu, t. j. pre jednotky „čierny“ flyš a Ceahlău — Rachovo.

Podiel ofiolitov a vôbec bázických telies vo flyšovom pásme Západných Karpát je obmedzený, viazaný hlavne na okraje trógov s kordilierami (tešinity). Drobné výskyty ultrabázických telies sú v bradlovom pásme a pravdepodobne sú výskyty ofiolitov v koreňovej časti magurského príkrovu.

Pre Západné a Východné Karpaty sú zvlášť charakteristické neovulkanity, vo Východných Alpách, zrejme v dôsledku nedostatku zázemia i nedostatku vnútorných tylových depresí chýbajú (obr. 15). Úzke spojenie Karpát s maďarským masívom (medzihorím) v trefohorách je evidentné. Priečne zlomy Západných Karpát sa prejavujú rozsiahlym vulkanizmom viazaným hlavne na stredoslovenský prelom. Vo Východných Karpatoch následkom zásahu tylovej depresie až k južnej hranici vonkajších Karpát (k bradlovému a flyšovému pásmu) prekrývajú neovulkanity podstatnú časť vnútorných zón.

K osobitostiam maďarského masívu jv. od balatónskej línie rátame mocný vývoj permu s kremennými porfýrmi. Zo skúseností v Západných Karpatoch možno uvažovať o diferencovanosti kôry už pri hercýnskom vrásnení, metamorfóze a granitizácii, o vzniku permských žľabov pri okrajoch, resp. medzi blokmi s výraznejšími zónami granitoidov. Alkalický charakter spodnokriedových bazických telies (pripomínajúcich tešiny vonkajších zón Západných Karpát) v pohoriach Mecsek a Villány, rovnako ako „severný“ typ mezozoika dáva podnet k úvahám o príslušnosti jv. bloku maďarského masívu k severným zónam alpíd, a to buď v postavení tektonického okna (P. HORVÁTH—L. STEGENA—B. GÉCZY 1974), alebo v postavení bloku posunutého k juhozápadu (M. BLEAHU 1976). Slabinou týchto úvah je nedostatok nielen výraznejšieho tektonického paleoalpínskeho zásahu v pohorí Mecsek a Villány v Maďarsku, ale i nedostatok už spodnokriedových, prípadne strednokriedových flyšových sledov — znakov charakteristických pre severné zóny Západných Karpát. Zrejme tak „severný“ typ triasu, ako i alkalické bazické telesá spodnej kriedy súvisia s paleotektonickým typom kôry maďarského masívu, ktorá bola v triase až v spodnej kriede pomerne hrubá.

Na základe nálezov bazických telies (z vrtných jadier z podlažia neogénnej výplne Maďarskej nížiny vrátane Zakarpatska) predpokladá sa pásmo ofiolitov (K. SZEPESHÁZY 1973, E. SZÁDECKY—KARDOSS et al. 1967, G. N. DOLENKO—L. G. DANILOVIČ 1976) západne od solnockého flyšového koryta i v jeho spodných predflyšových horizontoch. Obidva oddeľujú Apuseny od Západných Karpát lemovaných klinom maďarského masívu. Pásmo s bazickými útvarmi rovnako ako solnocké flyšové pásmo je zhruba súbežné s balatónskou líniou. Má pozíciu analogickú s murešskou ofiolitovou flyšovou zónou. Obidve tieto ofiolitové korytá predstavujú odnože vardarskej zóny vnikajúce hlboko na sever medzi bloky karpatského systému. Takúto medziblokovú pozíciu má i vardarská zóna, ktorá oddeľuje Srbsko-macedónsky a pelagónsky masív.

Severné Apuseny pripomínajú vnútorné Západné Karpaty hlavne rozložením väčších telies hercýnskych granitoidov, hojnosťou permu s mocnými kremennými porfýrmi a s melafýrmi v južnejších zónach. Ich nealpínske vulkanity sú súčasťou okrajovej časti neotisie, ktorá zahrňuje vnútorné zóny Západných Karpát i Východných Karpát a maďarský masív. V priebehu jury až eocénu prekonali Apuseny vývoj, ktorý ich zviazal s dinaridami, a to vďaka vzniku ofiolitovej zóny Mureš. Tá je totiž paleogeograficky predĺžením vardarskej zóny, ale štruktúrne súčasťou Apusén (M. LUPU 1975). Prítomnosť vrchnokriedovo-paleocénnych banatitov zväzuje Apuseny s Južnými Karpatmi. Apuseny teda predstavujú medziblok, na ktorom sa počas alpínskeho cyklu uplatňovali a miestami časove prekrývali vplyvy zo susedných areálov, ktoré usmerňovali magmatickú činnosť. Pásmo ofiolitov v Mureši, pásmo (predpokladaných) ofiolitov pri západnom okraji solnockého flyšového trógu v. od Apusén, spolu s prívskom (apendixom) bradlového pásma v oblasti Poiana Rusca a divergencia štruktúr Apusén naznačujú pohyb k V a rotáciu tohto bloku v smere hodinových ručičiek,

azda i v súvislosti so stenčovaním a rozťahovaním kôry v období zakladania ofiolitových trógov.

### Vysvetlivky k obrázkom

Obr. 1 Schéma hlavných tektonických jednotiek Východných Álp, Karpát, Balkánu a dinaríd

1. platformy, 2. čelná predhlbeň, 3. flyšové pásmo Východných Álp a Západných a Východných Karpát, a) vnútorné jednotky, 4. neskorotektonický flyš, 5. bradlové pásmo Karpát a) grestenská zóna Východných Álp, b) zóna Kotel, 6. peninikum, 7. vnútorné masívy, 8. neovulkanity, 9. väčšie kotliny

tektonické jednotky

Východné Alpy: Oo — vrchné východoalpinikum, NG — severná Grauwacken zóna, UO — spodné východoalpinikum, GR — gracké paleozoikum, CK — východoalpské kryštalinikum, DR — zóna Drávy, NK — severné Karavanky

Vnútorné Západné Karpaty: TS — tatridné jednotky a subatranské príkrovy (križňanský, chočský, strážovský), V — veporidné jednotky, NG — severogemeridná jednotka, SG — juhogemeridná jednotka, BŮ — Bukové hory, FC — vnútrokarpatský flyš

Vnútorné Východné Karpaty: ZO — zemplinkum, BR — Bretilla jednotka, BC — bukovinské a subbukovinské príkrovy, T — transylvánske príkrovy, FT — transkarpatský flyš

Apuseny: B — zóna Bihor, CA — príkrov Codru-Arieșani, BH — Biharia, MC — príkrov Muncel, MS — zóna Metaliferes, TS — príkrov Traskau, CM — centrálny masív

Južné Karpaty: DN — danubikum, G — getikum, SG — jednotka Saska-Gornjak, KE — Krajštidy, SG — supragetikum, S — príkrov Severin (Kraina)

Balkan: TR — prechodná zóna (severný okraj) Predbalkánu, FB — vlastný Predbalkán, SP — Stara planina, LK — Luda Kamčija, SR — Srednogorje, SS — Strandža — Sakar, R — rodopský masív

Dinaridy — helenidy: JV — Venetské, Julské a Savinské Alpy, SF — vrásy oblasti Sávy, HG — zóna hrastí a priekopových prepadlín, VZ — vardarská zóna, O — ofiolitová zóna, CD — centrálna-dinárska zóna, MD — Mirdita príkrov, SC — subpelagónsky príkrov, D — zóna Durmitor, VK príkrov Vysoký Krš, DA — zóna dalmatínskych vrás, BV — jednotka Budva, KC — Krasta — jednotka Zukali, PI — jednotka Pindos, KJ — jednotka Kruja, SD — južnoadriatická jednotka, IA — jónska jednotka, SZ — sazanská jednotka, IS — istrijská platforma

Centrálny maďarský masív a jeho výbežky: MH — maďarské stredohorie, MK — Mecsek, VY — Villány

Neovulkanické pohoria: SM — Slovenské stredohorie, BÖ — Börzsöny, CS — Cserhát, MA — Mátra, P-T — Prešovsko-tokajské vrchy, V-G — Vihorlatsko-gutinské hory, C-H — Caliman-Hargita

Obr. 2 Mapa magmatitov Karpát, Balkánu a dinaríd

1—4. ofiolity, 1. ultrabázika — väčšie telesá a) malé telesá zväčša v príkrovej pozícii, b) alkalické, 2. gabrá v sprievode diabázov, 3. spility — diabázy a) miestami hybridná asociácia: diabáz-porfýrit-kerato-

fýr-kremenný porfýrit, b) v podloží trefohornej panvovej výplne, 4. triasové diabázy (v sprievode menších intruzívnych telies); v spodnejších polohách hybridná asociácia: diabáz-porfýrit-keratofýr-kremenný porfýr, 5. predalpínske granitoidy a) synkinematické, 6. paleoalpínske granity a) synkinematické, 7. banatity a) intruzíva, b) vulkanity, c) prevažne pyroklastiká, 8. včasne neskorogeosynklinálne (priabón-oligocén) a) prevažne pyroklastiká, b) malé intruzíva, 9. a) neovulkanity (miocén-pliocén), b) prevažne pyroklastiká, c) malé intruzíva, 10. bazalty (pliocén-kvartér)

Obr. 3 Geotektonické postavenie magmatitov Západných Karpát a východnej časti Východných Alp  
tektonické jednotky

1. čelná predhlbeň, 2. flyšové pásmo, 3. bradlové pásmo Karpát, 4. grestenské bradlové pásmo, 5. jadrové pohoria: tatridné kryštalinikum, a) jeho mezozoický obal TA; príkrovy: KA — krížňanský, CH — chočský, S — strážovský, 6. veporidy a) východoalpínske kryštalinikum, 7. gemeridné paleozoikum, Grauwacken zóna a) paleozoikum Szendrő, 8. severogemeridné mezozoikum — NG a) Severné vápen-cové Alpy — NA, 9. juhoslovenský kras (silický príkrov) — SG, 10. ofiolitoidná séria Bükk a) meliatskej série, a) ofiolity peninika, 11. zemplínsky permokarbón, 12. gracké paleozoikum, 13. gosavská vrchná krieda, 14. centrálnokarpatský paleogén, 15. naložené depresie (oligocén — miocén)

magmatity

16. predalpínske granitoidy a) synkinematické, 17. paleoalpínske granity, 18. eocénne granodiority, 19. neovulkanity a) prevažne pyroklastiká, b) subulkanické a) hypoabysálne telesá, c) vulkanické centrá, 20. čadiče (pliocén — kvartér), 21. ultrabáziká a) väčšie „skryté“ telesá, b) malé mezozoické telesá, c) predalpínske, d) neznámeho veku, 22. bázické intruzíva, 23. alkalické báziká až ultrabáziká, 24. permské: kremenné porfýry a) melafýry, 25. predpermské báziká (diabázy) a) porfýroidy, 26. presu-nové línie, 27. hlbinné zlomy (geofyzikálne určené) a) zóny subdukcií

význačnejšie tektonické línie

Pozdĺžne: B — balatónska, D — Darnó, R — rožňavská, L — lubenická, M — margeciánska, Mu — muránska, P — pohorelská, Č — čertovická, Pp — peripieninský lineament, Le — lednická

Priečne zlomy: Š — štítický zlom, My — mytniansky zlom, Ja — jelšavský zlom, H — hornádsky zlom

Obr. 4 Geotektonické postavenie magmatitov Apusén (zostavené na základe Tektonickej mapy KBO — edit. M. Mahef 1974 a geologickej mapy Apusén — edit. M. Borcoş et al. 1976)

tektonické jednotky

1. neogénne panvy, 2. senón-gosauský typ, 3. bihorský autochtón, 4. systém jednotiek Codru, 5. systém jednotiek Baia de Arieş (Biharia), 6. zóna Metaliferes, 7. ultragetikum

magmatity

8. granitoidy hercýnske a) staršie a) synkinematické, b) alkalické, 9. banatity: granity a) granodiority a) vulkanity (andezit-dacit-ryolit), 10. neovulkanity a) pyroklastiká, 11. serpentinity (peridotity) sčasti gabrá, 12. a) prevažne bazalty, b) spility-ortofýry-keratofýry-porfýry, 13. permské vulkanity a) kremenné porfýry, b) melafýry, 14. predvrchnokarbónske vulkanity a) metabazalty, amfibolity, b) porfýroidy

Obr. 5. Geotektonická pozícia magmatitov Južných Karpát (zostavené na základe Tektonickej mapy KBO — editor M. Mahef 1973 a geologickej mapy Rumunska)

#### tektonické jednotky

1. moezijská platforma, 2. danubikum a) vnútorná zóna — Stara Planina, 3. getikum, 4. tektonická šupina Saska-Gornjak-Krajštidy, 5. ultragetikum-moravsko-penkovský príkrov, 6. srbsko-macedónsky masív, 7. vardarská zóna, 8. predhľbeň a) vnútorná, b) vonkajšia, 9. flyšové pásmo Karpát, zóna Koula, 10. severínsky príkrov, Ceahlău, 11. Predbalkán, 12. zóna Metaliferi (Mureš), 13. mezoalpínske vulkanicko-sedimentárne priekopové prepadliny (typu Timok), 14. depresie a) paleogénne, b) neogénne

#### magmatity

15. predalpínske granitoidy a) synkinematické, 16. paleoalpínske žuly a) synkinematické, 17. banatity a) subvulkanické a hypoabysálne telesá, b) vulkanity, c) pyroklastiká, 18. neovulkanity a) vulkanity, b) hypoabysálne telesá, c) pyroklastiká, 19. ultrabázické horniny, 20. a) bázické intruzívne horniny, b) spilitdiabázy-keratofýry-porfýry, 21. predpermské báziká, 22. permské kremité porfýry, 23. predpermské a) spilit-diabázy, b) porfýroidy

Obr. 6 Geotektonická pozícia magmatitov na Balkáne (zostavené na základe Tektonickej mapy KBO — edit. M. Mahef 1973 a prác E. Bončeva a kol.)

#### tektonické jednotky

1. moezijská platforma, 2. predbalkán, 3. Stara Planina, 4. zona Kotel, 5. ludakamčijská zóna, 6. Srednogorje a) Strandža, 7. rodopský masív, 8. srbsko-macedónsky masív, 9. Krajštidy, 10. depresie a) neohercýnske, b) neoalpínske-paleogénne, c) neogénne, 11. jednotka Kraina a) zóna Koula, 12. hlbinné kôrové zlomy, kryptoruptúry, zlomy, 13. osi antiklinál, 14. presunové línie, 15. posuny

#### magmatity

1. predalpínske granitoidy a) predhercýnske (synkinematické), b) typ Stara Planina (zväčša synkinematické), c) hercýnske (?) — južnobulharské, d) alkalické intruzíva, e) strunská dioritová formácia, 2. paleoalpínske granity a) synkinematické, 3. banatity a) intruzíva, b) vulkanity, c) prevažne pyroklastiká, 4. včasne neskorogeosynklinálne a) prevažne andezity, b) prevažne ryolity, c) pyroklastiká, d) subvulkanické telesá, e) dajky, 5. čadiče, 6. serpentinity, 7. bázické vulkanity (prevažne bazalty), 8. neohercýnske vulkanity a) kremenné porfýry, b) melafýry, 9. predvrchnokarbónske vulkanity a) metabázity, amfibolity, b) porfýroidy

Obr. 7 Skica rozdielov hercýnskej konsolidácie vo vzťahu k typu triasu a k alpínskym magmatitom

#### regióny rozdielnej hercýnskej konsolidácie

1. regióny intenzívneho hercýnskeho vrásnenia, 2. regióny slabšej hercýnskej konsolidácie

#### typy triasu

3. kvaziplatformný typ (kontinentálny typ kôry), 4. austroalpínsky typ triasu (členitý šelf), 5. dinaridný typ — paraliogeosynklinálny typ, 6. tauridný typ (s flyšovou tektofáciou)



## magmatity

7. hercýnske granitoidy a) synkinematické, b) alkalické, 8. permské melafýry (v príkrovovej pozícii), a) kremenné porfýry, 9. ultrabáziká a) malé telesá — zväčša v príkrovovej pozícii, b) alkalické, 10. gabrá (prevažne v sprievode diabázov), 11. a) spility — diabázy; sčasti asociácie; b) diabáz-porfýrit-kremenný porfýrit, keratofýr, 12. triasové diabáz-porfýry, keratofýry — kremenné porfýry

Obr. 8 Skica zón rozdielnej hercýnskej stabilizácie a rozloženie jednotlivých typov permu v Západných Karpatoch

1. bradlové pásmo, 2. zóny intenzívnej granitizácie a intenzívneho hercýnskeho vrásnenia a) prevažne autochtónne (tatrity a maďarský masív), b) v pozícii zväčša paraautochtónnej až alochtónnej (južné veporidy), 3. masívy granitoidov a) na povrchu, b) v podloží neogénneho pokryvu, 4. zóny intenzívnejšieho hercýnskeho vrásnenia, ale bez rozsiahlejšej granitizácie (severné veporidy), 5. zóny menšej hercýnskej stabilizácie a) s malým podielom hercýnskych granitoidov, b) veľmi slabo postihnuté vrásnením a bez granitoidov, 6. perm a) melafýry, (v príkrovovej pozícii), b) kremenné porfýry, 7. perm a) suchozemského typu, T = tatrídny; SV — severoveporidny; JV — južnoveporidny; Z — zemplínsky, b) chošského príkrovu, 8. severogemeridny perm, 9. perm sčasti morský okrajový typ rožňavsko-železnický, bükkský typ, 10. ofiolity a) ultrabáziká, b) bázické intruzíva, c) diabázy, d) kremité porfýry-keratofýry-diabázy

Obr. 9 Geotektonické postavenie magmatitov dinaríd a severnej časti heleníd (podľa tektonickej mapy karpatskobalkánskych regiónov a príslušných oblastí — edit. M. Maheľ)

## tektonické jednotky

1. istrijská platforma, 2. maďarský masív (V — Villany, M — Mecsek), 3. srbsko-macedónsky masív, 4. pelagónsky masív, 5. zóna Korab, 6. vardarská zóna, 7. vnútrodinárske jednotky a) výstupy paleozoika, 8. flyšové pásmo — Durmitor ( $J_3$ — $K_2$ ), 9. vonkajšie zóny dinaríd a) heleníd a adriatická zóna, 10. Budva-Zukali, 11. naložená čelná predhlbeň a) väčšie vnútorné depresie paleogénne, b) neogénne, 12. flyš neskorotektonický

## magmatity

13. hercýnske a staršie granitoidy a) synkinematické, 14. paleoalpínske granity a) synkinematické, 15. paleogénne granity, 16. neovulkanity, a) pyroklastiká, b) subvulkanické granity a granodiority, 17. čadiče (pliocén), 18. spilit-keratofýrová asociácia (trias), 19. spilit-diabázová asociácia (jura-spodná krieda), 20. intruzíva bázik a) metamorfované, 21. ultrabáziká a) metamorfované, 22. permské kremenné porfýry a porfýry, 23. predpermské spilit-diabázy

Obr. 10 Tektonická mapa južných zón Západných Karpát so zvýraznením magmatitov

## tektonické jednotky

1. mezometamorfity až katametamorfity veporíd, prevažne staropaleozoické, sčasti proterozoické, 2. séria Hladomornej doliny — staropaleozoická episéria veporíd, 3. staropaleozoická episéria gemeríd, 4. devón Szendrő, 5. karbón a) bükkský, b) severogemeridny, 6. veporidny perm, 7. severogemeridny perm, 8. rožňavsko-železnická séria — perm, 9. perm Bükku, 10. krížňanská jednotka, 11. stružnícka jednotka (spodny trias až jura) — obal veporíd, 12. severogemeridné mezozikum a) spodny trias, 13. silický príkrov a) spodny trias, 14. meliatska séria (sčasti ofiolitová) — trias, 15. séria Bükku, 16. jura gemeríd, 17. vrchná krieda, 18. paleogénne depresie, 19. neogénne depresie

## magmatity

20. žulo-ruly muránskeho typu, 21. hercýnske žuly — trondhjemity, 22. paleoalpínske žuly, 23. granodiority a) eocénne, b) nealpínske, 24. gabrá a diority a) mezozoické gabrá v sprievode ultrabázik a diabázov, 25. menšie telesá mezozoických ultrabázik a) neistého veku, b) predalpínske, c) ultrabáziká v hĺbke, 26. diabázy, porfyryty, kremenné porfýry (trias), 27. neovulkanity a) prevažne andezity, b) ryolity, c) pyroklastiká, 28. čadiče, 29. permské kremenné porfýry, 30. paleozoické diabázy a) porfyroidy

## hlavné zlomové línie

B — balatónska, D — Darnó, R — rožňavská, L — lubenícka, M — margeciánska, Mu — muránska, P — pohorelská, J — hlbinný zlom jelšavský, Š — štítnický zlom, H — hornádsky zlom

Obr. 11a Geotektonické postavenie magmatitov vo Východných Karpatoch (na podklade Tektonickej mapy KBO — edit. M. Maheľ 1974 a geologickej mapy Rumunska)

## tektonické jednotky

1. jednotka Bretila — Biely potok, 2. subbukovinský príkrov, 3. bukovinský príkrov, 4. skupina transylvánských príkrovov, 5. getický príkrov, 6. zemplínsky ostrov, 7. križňanský príkrov, 8. a) gemeridy, b) veporidy, 9. bradlové pásmo, 10. príkrov „Čierny flyš“ (Kamenný potok), 11. príkrov Ceahlău (Rachovo) spolu s príkrovom Barault, 12. flyšové pásmo (príkrovy strednej a vonkajšej skupiny), 13. čelná predhľbeň a) vnútorná, b) vonkajšia, 14. a) neskorotektonická paramolasa Buccoci, b) neskorotektonický transkarpatský flyš, 15. neogénne depresie, 16. presunové línie a) posuny, 17. zlomové línie a) predpokladané (určené geofyzikálnymi metódami)

## magmatity

18. predhercýnske a hercýnske granitoidy a) synkinematické, 19. alkalické plutóny (detroitidy), 20. neovulkanity a) prevažne pyroklastiká, b) žuly, granodiority — malé intrúzie, 21. pliocénne čadiče, 22. ofiolity a) bazalty, b) spilit-keratofýrová asociácia, 23. ultrabázické horniny, a) malé telesá, 24. permské kremenné porfýry, 25. predpermské a) metabazalty, amfibolity, b) porfyroidy

Obr. 11b Paleogeografická skica vnútorných Západných Karpát počas jury a spodnej kriedy a návaznosť na Východné Alpy M. MAHEĽ, 1978)

1. Západokarpatský intraoceanický prah, neskoršie tektonicky preformovaný v A — príkrovy: a) vysokej skupiny, b) manínsky, B — tatrídne jednotky zväčša paraautochtónne; 1a. západné pokračovanie západokarpatského intraoceanického prahu do Východných Álp: A — Mittelostalpin, B — Unterostalpin; 2. Kordiliery ostrovnej zóny bradlového pásma; a) sprievodný žľab (trenč); 3. tróg s hrubou kontinentálnou kôrou; a) jeho okrajová prahová časť; 4. tróg s tenšou kontinentálnou (suboceanickou kôrou; a) jeho príprahová časť; 5. tróg s oceanickou kôrou; a) s paraoceanickou až oceanickou kôrou; 6. krátkodobé žľaby na širokom oceanickom prahu: SG — juhogemeridný, TD — transdanubijský; 7. široký vnútrooceanický prah

Obr. 12a Skica vzťahu prejavov alpínskych vrásnivých periód k magmatitom v karpatsko-balkánskom systéme

## magmatity

1. ultrabáziká a) malé alochtónne telesá, b) alkalické, 2. intruzívne baziká (hlavné gabrá), 3. spility a diabázy (jura — spodná krieda) a) v podloží trefohorných sedimentov, b) kremenné porfýry, keratofýry,

diabázy, c) alkalické báziká a ultrabáziká, 4. triasové diabázy, porfyry, kremenné porfýry, keratofýry, 5. paleoalpínske granitoidy a) synkinematické, 6. mezoalpínske banatity a) vulkanity, b) prevažne pyroklastiká, c) subvulkanické a plutonické telesá, 7. trefohorné vulkanity, A) včasne neskorogeosynklinálne (priabón-oligocénne), B) neskorogeosynklinálne (miocén-pliocén), a) pyroklastiká, b) subvulkanické a intruzívne telesá, 8. bazalty

#### regióny

9. a) výraznejšie postihnuté kimerským (hlavne neokimerským) vrásnením, b) paleoalpínsky zvrásnené (za austrijskej až subhercýnskej fázy), c) len slabo paleoalpínsky postihnuté, 10. mezoalpínsky zvrásnené (za laramskej až pyrenejskej fázy), 11. neoalpínsky zvrásnené (za sávskej až štajerskej fázy), 12. neskoralpínsky (koncom neogénu a vo štvrtohorách) postihnuté vrásnením

#### Obr. 12b Geotektonická pozícia granitoidov

1. predalpínske granitoidy a) synkinematické, b) alkalické, 2. a) paleoalpínske granity, b) synkinematické, 3. mezoalpínske banatity — interorogénne, 4. subvulkanické a intruzívne telesá a) včasne neskorogeosynklinálne (priabón-oligocén), b) neskorogeosynklinálne (miocén-pliocén), 5. čelná predhľbeň, 6. flyšové pásmo Karpát, Východných Álp, vonkajšie zóny dinarid, helenid a Balkánu, 7. a) bradlové pásmo, b) grestenská zóna, c) zóna Kotel, 8. a) vnútorné zóny, b) neskorotektonický (back-arc) flyš, 9. intrageosynklinálne masívy, 10. vnútorné depresie, 11. hlavné tektonické línie a) posuny

#### Obr. 13a Skica rozloženia vrchnojursko-spodnokriedových oceanických žľabov vo vzťahu k ofiolitom

1. stredné masívy a) výstupy paleozoika v dinaridách, 2. oblasti s tenkou kôrou (kvazioceanickou) v triase, 3. žľaby s oceanickou kôrou vo vrchnej jure a v spodnej kriede, a) s mocnými silicitmi, b) s preflyšom, c) so silicitmi a s preflyšom, 4. a) žľaby s tenšou kontinentálnou kôrou s pelagickými karbonátmi a so silicitmi, b) sprievodné prahy, 5. priehľbeniny s pelagickými karbonátmi na platniach (zväčša na okrajoch) s hrubšou kontinentálnou kôrou, 6. ultrabáziká, a) malé telesá zväčša v príkrovovej pozícii, b) alkalické, 7. gabrá v sprievode diabázov a) alkalické báziká, 8. spility — diabázy a) v podloží terciérnej výplne bazénov, 9. triasové diabázy (v sprievode menších intruzívnych telies); v spodnejších polohách hybridná asociácia: diabáz-porfýrit-keratofýr-kremenný porfýr

#### Obr. 13b Predpokladaný priebeh trógov. Vysvetlivky pozri pri obr. 13a.

#### Obr. 14a Skica rozloženia ofiolitov vo vzťahu k flyšovým tektonogrupám

##### flyšové tektonogrupy

1. vrchnokriedovo-paleogénne flyšové pásmo a) miestami naložené v pásmach preflyšu; jednotky: M = magurská, Si = sliezská, K = Kruja, J = jónska, 2. titón-spodnokriedový flyš (prevažne preflyš) a) v podloží mladšieho flyšu: So — solnocký, BN — „čierny flyš“, C — Ceahlău, Se — Severin, T — Trojan, SS — Strandža, Sr — Sarajevo, V — vardarská zóna, Mu — Mureš, Z — Zukali (Gassau), 3. vrchnokriedový vulkanogénny flyš so vzťahmi k banatitom, 4. flyš až flyšoid typu back-arc a) v podloží molasovej výplne, 5. ultrabáziká a) malé telesá, zväčša v príkrovovej pozícii, b) alkalické, 6. gabrá (v sprievode diabázov), 7. spility — diabázy; v menšej miere hybridná asociácia a) v podloží trefohornej výplne bazénov, b) alkalická, 8. triasové diabázy v sprievode menších intruzív; v spodnejších polohách hybridná asociácia diabáz-porfýrit-keratofýr-kremenný porfýr

Obr. 14b Skica geotektonického postavenia banatitov

1. platformy (M = moesijské) a stredné masívy (SM = srbsko-macedónsky; R — rodopský), 2. bihorský autochtón, 3. príkrovové jednotky severných Apusén, 4. ofiolitové zóny: M = Metaliferes; V = vardarská, a) trojanská flyšová zóna (titón — spodná krieda), 5. a) zóny preflyšu so sprievodom bázik a ultrabázik: C = Ceahlău, S = Severin, SS = Strandža, b) trojanský flyš (titón — spodná krieda), 6. a) flyšové pásmo balkaníd = FB, b) BFC = hlavné flyšové pásmo východných Karpát, 7. vrchnokriedové formácie sprevádzajúce banatity, a) flyš, b) pestré vulkanosedimentárne fácie, c) gosauský typ málo postihnutý vrásnením, 8. banatity a) väčšie telesá efuzív, b) subvulkanické a plutonické telesá, 9. zóny subdukcie

Obr. 15a Geotektonická pozícia neskorogeosynklinálnych vulkanitov

1A — včasne neskorogeosynklinálne vulkanity (vrchný eocén — oligocén) a) pyroklastiká, b) hypoabyzálske a intruzívne telesá 1B — neskorogeosynklinálne vulkanity (miocén — pliocén) a) pyroklastiká, b) hypoabyzálske a intruzívne telesá, 2. bazalty (pliocén — kvartér), 3. väčšie podpovrchové telesá spodnomiocénnych vulkanitov, 4. čelná predhľbeň, 5. flyšové pásmo Karpát, Východných Álp, vonkajšie zóny dinaríd-helenid a Balkánu, 6. a) bradlové pásmo, b) grestenská zóna, c) zóna Kotel, 7. a) vnútorné zóny, b) neskorotektonický (back-arc) flyš, 8. intrageosynklinálne masívy, 9. vnútorné depresie, 10. hlavné tektonické línie a) posuny

Obr. 15b Mapa neskorogeosynklinálnych vulkanitov a tektonických jednotiek

1. paleogénne (priabón — oligocén) panvy, 2. posávske panvy a) príkrovové jednotky, 3. poštýrske panvy a) príkrovové jednotky, 4. pliocénne panvy, 5. čelná predhľbeň a) vnútorná — zasiahnutá neskorostýrskym až valašským vrásnením, b) vonkajšia, 6. včasne neskorogeosynklinálne vulkanity (vrchný eocén — oligocénne) a) malé hypoabyzálske a intruzívne telesá, 7—8. neskorogeosynklinálne vulkanity a) hypoabyzálske telesá, 7. prevažne miocénne (hlavne báden-sarmat), 8. prevažne pliocénne, sčasti až kvartérne, 9. väčšie podpovrchové telesá miocénnych vulkanitov

Obr. 16 Vývojový model Západných Karpát — zostavil M. Maheľ 1977

1. plášť, 2. ruložulový sloj, 3. príkrovové komplexy, 4. prejavy vysokotermálnej a vysokotlakovej metamorfózy, 5. zóny subdukcie vysokotlakovej a nízkotermálnej metamorfózy, 6. a) preniky granitoidov, b) diapirické výstupy, 7. prejavy vulkanizmu s prevahou kremenných porfýrov s prevahou spilitov, diabázov; melafýrov, 8. ofiolity a ofiolitoidné magmatity, 9. mocnejšie radiolarity, 10. trógové fácie, prevažne slieňovce, 11. a) flyš, b) piesčité, 12. šelfové detritiká s prevahou pelítov a) pieskovecov, 13. a) šelfové a prahové karbonáty, b) intrageosynklinálnych prahov, c) pelagické karbonáty bazénov, 14. molasy, 15. výstupy bázičkej až ultrabázičkej magmy, 16. presunové línie; hlbinné zlomy, 17. zóny roztiahnutia kôry

hlbinné zlomy

B — balatónska línia, D — Darnó línia, R — rožňavská línia, L — lubenícka línia, M — muránsky zlom, Č — čertovická línia, Pp — peripieninský lineament, Le — lednická línia

Obr. 17 Tabuľka ukazujúca vývoj Západných Karpát — zostavil M. Maheľ 1977

1. a) preniky granitoidov, b) diapirické výstupy, 2. aspidná až flyšoidná, sčasti flyšová formácia,
3. a) kremenné porfýry, b) spilit-diabázy; melafýry, 4. telesá ultrabázik a) bazických intruzív, b) alkalických bázik a ultrabázik, 5. andezity a ryolity a) čadiče, 6. pieskovce, kremence a) karpatský keuper,
7. a) plytkomorské detritické sedimenty, b) s prevahou pelitov, 8. a) šelfové vápence, b) dolomity,
9. karbonáty intrageosynklinálnych a) prahov, b) bazénov až trógov, 10. trógové fácie a) prevažne slieňovce, b) slienité vápence, 11. mocnejšie rádiolarity, 12. a) flyš, b) piesčité, 13. šelfové a prahové sedimenty s prevahou pelitov, a) pieskovcov, 14. molasy, 15. a) prejavy vysokotermálnej a vysokotlakovej metamorfózy, b) zóny subdukcie; vysokotlakovej a nízkotermálnej metamorfózy, 16. presuny, 17. prešmyky; prejavy vrásnenia

Михал Магель

## Геотектоническая позиция магматитов в Карпатах, на Балкане и в Динаридах

Резюме

Введение

Тектоническая карта Карпато-балканских регионов 1:1000000 (М. Магель 1973) показывает некоторые характерные черты позиции магматитов в Восточных Альпах, Карпатах, на Балкане и в Динаридах. Такими чертами являются:

- изобилие часто крупных тел гранитоидов доальпийских и только небольшие, часто спорадические тела альпийских гранитоидов;
- сопровождение древнейших неогерцинских моласс (верхний карбон – пермь) вулканитами преимущественно кислыми, особенно кварцевыми порфирами;
- многочисленные вулканиты (преимущественно основные) в докарбонских комплексах и их органичное распространение в альпийских комплексах;
- яркая вулcano-плутоническая формация межорогенического типа, генетически связанная с домолассовыми выполнениями продольных грабен на Балкане, в Южных Карпатах и Апусенах (банатитовая формация);
- обширные третичные вулканиты с преимуществом андезитов, сопровождаемые риолитами, связанными с молодыми депрессиями;
- многочисленные, но небольшие тела плиоценово-четвертичных базальтов, рассеянные особенно в интернидных зонах.

С точки зрения геотектоники, опираясь на вышеприведенные данные, магматиты выделяются в следующие группы:

- гранитоиды, тесно связанные с процессами складчатообразования и с формированием структурных планов;
  - неогерцинские позднегеосинклинальные вулканиты, тесно генетически связанные с гранитоидами и сопровождающие молассовые выполнения желобов и депрессий;
  - доальпийские эвгеосинклинальные вулканиты – сопровождение докарбонских свит альпид;
  - мезоальпийские офиолиты и базальтоиды, связанные с отдельными структурными поясами;
  - мезоальпийские межорогенические магматиты вулcano-плутонические (банатиты) в сопровождении грабен-синклиналиев и грабен, выполненных отчасти альпийскими молассами, в большинстве случаев флишем;
  - позднегеосинклинальные неовулканиты, генетически связанные с третичными депрессиями, выполненными молассами нескольких стадий развития, стадий окончания морфоструктурного плана;
  - финальные вулканиты, связанные с послеорогенической стадией альпид.
- В этом порядке мы и будем магматитами заниматься.

а) Гранитоидные породы в большинстве случаев генетически связаны с процессами складкообразования более глубинного характера и с ними вызванным процессом растения толщины коры. Однако, состояние зрелости коры в отдельных циклах развития различно. Различны также проявления гранитизации в отдельных циклах.

Догерцинские гранитоиды, связанные генетически с байкальской и более древними складчатостями, в большинстве случаев являются синкинематическими, тесно связанными с мигматитизацией, анатексисом и ярким региональным метаморфизмом.

Самыми частыми и также по своему значению в развитии и строении альпид самыми важными являются герцинские гранитоиды. Для них характерно широкое структурное тектоническое разнообразие, широкий диапазон времени, сравнительно небольшая шкала дифференциации содержания. Синкинематические гранитоиды, тесно связанные с мантией в некоторых зонах, представляют собой обыкновенно самые древние и отчасти самые глубинные проявления герцинской гранитизации. Чаще всего встречаются большие тела батолитов и факолитов, оказывающие тесную генетическую связь не только с процессами складкообразования но, следовательно, также со структурным планом. В большом количестве случаев для них характерно диапировое поднятие, достигающее уровня эрозии.

Благодаря образованиям больших батолитов и факолитов и их постепенному поднятию, диапировому подъему, возникает особый тектонический стиль. Образуются складки широкого распространения, мегаантиклинальные своды, напр. в некоторых керновых горных массивах Западных Карпат в татридах, брахиантиклинали Родопского массива и Сербийско-македонского массива. Регионы, менее затронутые гранитизацией, выявляют зоны узко сжатых складок и надвигов (напр. в кристаллическом основании Черной горы в Западных Карпатах; система бистрицкой группы покровов в Восточных Карпатах) (КРЕЙТНЕР 1972). Герцинская складчатость рядом с образованием гранитоидных тел, по-видимому, проявляется очень ярко при формировании структурного плана.

Более крупные гранитоидные тела с тенденцией к поднятию являются также морфоструктурным деятелем. Своды становятся материалом сноса, выполняющим прилегающие желобы, сформированные вдоль окраин или же между зонами гранитоидных тел.

Процесс герцинской гранитизации проходил в широком промежутке времени, с конца девона до конца перми и он сосредоточился особенно на отдельных стадиях герцинской складчатости. Структуроформирующие проявления связаны с более древними, морфоструктурные с более молодыми стадиями с образованием молассовых желобов и депрессий, сопровождаемых вулканистами. В сопровождении более древней или нижней молассы в поясах менее постигнутых герцинской гранитизацией в перми встречаются чаще основные породы, в гемеридах даже в сопровождении небольших ультраосновных тел. В зонах более интенсивно затронутых гранитизацией часто встречаются кварцевые порфиры в сопровождении более молодой сухопутной молассы. В то же время они синхронны с самыми молодыми проявлениями гранитизации в примыкающих регионах поднятий. О тесной генетической связи кварцевых порфиров и гранитоидов свидетельствует присутствие гранитпорфиров напр. в северовепоридном желобе Западных Карпат, но также и более кислый

характер магмы кварцевых порфиров и пермских аплитов, пегматитов и лейкократовых гранитов.

Кварцевые порфиры перми являются, следовательно, поверхностным проявлением магматизма, связанным с грабенами с яркой тенденцией опускания. В более южных зонах в перми в сопровождении верхних моласс встречаются также мелафиры. Но благодаря небольшому присутствию пирокластического материала они в далеко меньшей степени участвуют во влиянии на осадконакопление. Но эти зоны с мелафирами не оказывают признаков более яркого проглубления. Речь идет не об уточнении коры во время их поднятия, а о ее прерывании, о раскрытии подводных каналов в более глубокие части. Об этом свидетельствует также характер тел типа силлов и также чередование мелафиров и кварцевых порфиров в некоторых регионах, напр. в Черной горе в Западных Карпатах, в Балканидах, но особенно в Апусенах в единице Кодру.

Герцинская гранитизация генетически тесно связана с метаморфизмом неравномерной интенсивности. Например в Западных Карпатах в зонах более яркой гранитизации более древний палеозой подвергнут метаморфизму вплоть до фации амфиболитов, зоны менее постигнутые гранитизацией метаморфизованы только до фации зеленых сланцев. Интенсивность метаморфизма во время герцинской складчатости на каждой степени понижается, более слабые ее проявления можно наблюдать даже в перми. Метаморфизм перекристаллизационного характера и гранитизация свидетельствуют о глубинном характере герцинской складчатости. Но при этом образовался сложный структурный план более этажный, с явными несогласиями. Но главным результатом герцинской складчатости, гранитизации и метаморфизма является утолщение коры — ее сиализация, что значит существенный переворот в развитии альпийской геосинклинальной системы, начало нового цикла, но одновременно и нового типа геосинклинали с существенными изменениями в главных процессах в седиментации, складчатости и в метаморфизме. Но необходимо подчеркнуть неравномерность герцинской гранитизации и стабилизации в отдельных пространствах и зонах, обусловившую различие и в значительной степени также расчлененность альпийской геосинклинали во время ее дальнейшего развития.

Чаще всего встречаются гранодиорит-граниты с более узкой шкалой дифференциации. Более широкое вещественное содержание и большее количество формаций гранитоидов встречается в балканидах и в области геосинклинально более расчлененной (Е. Бончев 1967), где рядом с гранодиорит-гранитной ассоциацией встречаются тоже в балканидах габбро-плагиогранитная, габбродиорит-гранодиоритовая и щелочная габбро-сиенитовая ассоциации (Е. Димитрова и др. 1975). Везде часто встречаются аплиты и пегматиты, но также и меньшие тела лейкократовых гранитов, часто относящихся к пермскому возрасту.

Участие альпийских гранитов в формировании альпийского структурного плана небольшое. Это главным образом небольшие тела, ограниченные зонами доальпийского кристаллического основания и сильно обработанные палео- вплоть до мезоальпийской складчатостями. Более крупные тела также синкинематических гранитоидов встречаются, главным образом, в более южных частях Сербийско-македонского массива в областях уже значительно постигнутых киммерийской складчатостью. Альпийская складчатость отличается обширным сокращением коры и разнообразием тектонических стилей



(результат расчлененности и различия толщины и механических качеств вещественного выполнения отдельных зон седиментации). Самым ярким проявлением этого сокращения являются покровы приповерхностные и более глубинные, зоны интенсивного сжатия и зоны меланжей, свидетельствующие о субдукции. Такие характерные проявления, как гранитизация и метаморфизм, связанные с герцинской складчатостью, в альпийской складчатости наверно имеют более значительный приповерхностный характер. Геосинклиналь альпийского типа отличается более значительной динамикой и, главным образом, расчлененностью, чем геосинклиналь доальпийская, и это приводит также к основным различиям в распространении и типе магматитов. Поэтому распространенное расчленение магматитов по Штилле на начальные, орогенические и субсеквентные с более широкой глобальной силой является слишком схематическим для геосинклинали альпид.

Более важную роль в строении альпид играют тела гранитоидов, образующие составную часть вулканоплутонических формаций мезоальпийских (банатитовый тип). Доля интрузий и гипабиссальных тел и величина некоторых наверно имеют связь с состоянием коры, обусловленным чередующимися сжимающими движениями (ларамийскими и илирийскими или же пиренейскими). Меньшие тела гранодиоритов, особенно гипабиссальные, сопровождаются также неовулканистами. Их возникновение связано с ослаблением напряжения в коре.

б) Молодогерцинские позднегеосинклинальные вулканисты генетически связаны с неогерцинскими, главным образом пермскими желобами и депрессиями, выполненными молассаами. Они являются запоздалым поверхностным проявлением обширной герцинской сиализации коры. Чаще всего они встречаются: а) в желобах, отделяющих поясы более значительной герцинской гранитизации, б) вдоль внутренних окраин герцински стабилизированных пространств. Благодаря глубоко затрагивающим разломам там произошли также подводные извержения мелафиров, в большинстве случаев типа силлов, сопровождаемых дайками мелафир-порфиритов и порфиритов (напр. мелафировая свита хочского покрова Западных Карпат и пермь в покрове Кодру и Апусенах). Верхнепалеозойские депрессии более северных зон альпид имеют молассовые выполнения только с небольшой долей вулканистов, а именно кварцевых порфиритов.

в) Доальпийские эвгеосинклинальные вулканисты характерны для существенной части геосинклинальных свит более древних стадий, чем герцинская гранитизация. Они представляют составную часть осадочно-вулканических комплексов аспидных, отчасти диастрофических. Они относятся преимущественно к спилит-диабазовой, спилит-кератофировой формациям и сопровождаются интрузиями габбродиорит-лейкогранитной и перидотит-пироксенитовой формаций. Неравномерное пространственное распространение основных вулканистов и большее распространение кислых вулканистов в некоторых верхнепротерозойских свитах (напр. Арада свита), прежде всего древнепалеозойских, особенно кембрий-ордовикских или ордовикских (напр. зона Грауваккен Восточных Альп, гельницкая свита в Западных Карпатах, свита Мунцель в Апусенах, свита Тулгеш в Восточных Карпатах) свидетельствуют о частичной дифференциации коры, связанной с позднебайкальской складчатостью. Но следует подчеркнуть, что эти кислые вулканисты (преимущественно кварцевые порфиры, кварцевые кератофиры) являются составной частью

подводных геосинклинальных свит. Генетически они связаны с островными дугами, или же с регионами более мощной переходной субконтинентальной коры, расположенными в краевой части геосинклиналей. Присутствие кварцевых порфиров в геосинклинальных свитах девона (гронская свита в вепоридах Западных Карпат, некоторые свиты в Южных Карпатах) свидетельствует о последствиях дифференциации коры, связанной с началом герцинской складчатости. Правда, седиментация в древнем палеозое не была всегда беспрерывной.

г) Мезозойские офиолиты и базальтоиды встречаются в поясах, приуроченных на основе присутствия глубокоководных фаций к троговым с тонкой или же прерванной сиалической корой субконтинентальной или же субокеанической. Они появляются уже во время триаса, главным образом ладина, часто в сопровождении более мелкоководных известняков. Сначала они представляют преимущественно интермедиаальный — гибридный тип с преобладанием спилитов и кератофилов, генетически связанных с утончением, дроблением и дифференциацией коры. С постепенным нарастанием трогов во время ладина и особенно карна — для которых характерны более глубококоморские фации карбонатов, но также и радиоляритов и пеллитов — нарастает также доля более основных пород и появляются тоже ультраосновные породы. В большинстве сегментов альпид вулканы триаса связаны с самыми внутренними зонами альпийской геосинклинальной системы, т.е. с зонами, слабо стабилизированными герцинской гранитизацией. В Западных Карпатах такими являются мелиатская свита в Словацком карсте и аналогичная свита в Буковых горах, единица Башкау в Апусенах, а на Балкане Сакар и в Восточных Карпатах трансильванские покровы. В динаридах уже во время триаса вулканы офиолитового типа связаны не только с внутренними зонами (например вардарская и офиолитовая), но отчасти также с зонами внешними (особенно Будва-Зукали). Это является результатом меньшей интенсивности герцинской складчатости и гранитизации в динаридах главным образом в древнегерцинских трогах, где даже в перми встречаются преимущественно морские фации.

Более часто встречаемые основные и ультраосновные породы в качестве составной части офиолитовой ассоциации связаны с трогами юры и нижнего мела, для которых характерны мергелисто-радиоляритовые фации; например крижянский и кисуцкий трог Западных Карпат, буковинский трог Восточных Карпат, трог пениникума Альп, зона Муреш в Апусенах, в динаридах офиолиты связаны особенно с внутренними зонами (вардарской и офиолитовой), а в геленидах с зоной Мирдита (субпелагоникум). Чаще всего офиолиты сопровождаются также мощными дофлишевыми фациями (титон — нижний мел — напр. в единицах Рахов — Чеахлау в Восточных Карпатах, Северин — Краина в Южных Карпатах, Странджа на Балкане). В Муреше и вардарской зоне офиолитовая свита располагается от глубокоководных верхнеюрских карбонатно-силицитовых фаций вплоть до флишевых.

Во всех приведенных случаях видима связь офиолитов со стадией самого большого углубления отдельных геосинклинальных зон. Но это не только связь палеогеографическая, но также и проявление активности и следовательно также связь прежде всего палеотектоническая, логически связанная с более тонкой корой зон офиолитов и зон с присутствием базальтоидов и более глубокоководных фаций.

Основные породы, преимущественно щелочного характера (авгититы, лимбургиты), чаще всего встречаются в сопровождении мелководных фаций типа порогов, особенно в краевых, нарушенных разломами частях, например в татридах (Высокие Татры) и в приутесовом регионе Западных Карпат, а именно во верхней юре и нижнем мелу.

Более численные проявления верхнеюрского и нижнемелового основного магматизма, часто щелочного характера, связаны с разломами, ограничивающими кордильеры, или же пороги и прогибы. Известны они из флишевой геосинклинали, напр. тешиниты в силезской единице, диабазы и авгититы во флишевых баррем-альбских свитах буковинского и суббуковинского покровов, или же на окраинах флишевых трогов (например диабазы) в юрских известняках в зоне Пресачина в Южных Карпатах.

Щелочной характер имеют также нижнемеловые подводные излияния в горном массиве Мечек в Венгрии; щелочные диабазы («базальт-трахидолериты») в сопровождении пирокластических пород и диабаз-порфиритов (Э. Садецки-Кардош и др. 1967).

Видима зависимость распространения основных пород в мезозойских свитах от толщины коры, но также от расчлененности геосинклинали и от повышенной динамики коры во время магматических проявлений. Базальтоиды являются сопровождающим явлением, главным образом, в прогибах с тонкой корой и в глубоко затрагивающих разломах, отделяющих друг от друга прогибы и пороги или же кордильеры. Самое большое распространение базальтоидов наблюдается в верхней юре и в нижнем мелу, т.е. во время самой большой расчлененности и самого большого палеотектонического противоречия мезозойских геосинклиналей, но также и самой значительной океанизации в альпидах.

Расположение офиолитов и базальтоидов и их тип зависят от расчленения региона осадконакопления на трого (прогибы) и пороги (или же кордильеры). Подобно, как расчлененность у каждого сегмента имеет свои особые черты, этому и у зон офиолитов. Только некоторые из них переходят из одного сегмента в другой, и это тоже с определенными отличиями.

Хотя в большинстве сегментов европейских альпид встречается менее офиолитов, для применения принципов новой глобальной тектоники значение имеет особенно их большая палеотектоническая расчлененность, главным образом во время верхней юры и нижнего мела и их доказываемые переходы от глубоководных к мелководным секвенциям. В них наблюдаются несколько зон с тонкой корой и зоны с более мощной корой. Например в Западных Карпатах во время юры следует отличить по крайней мере четыре зоны с более тонкой корой, отделенные друг от друга порогами (или же кордильерами) с более мощной корой (М. Магель 1975). Секвенции с перемешанными фациями мелко- и глубоководными выявляются особенно ярко в поясе утесов, проявляющем наверно самое большое противоречие между чорштынским и кисуцким типами. Все свидетельствует о том, что модель настоящего Атлантического океана нельзя применить в альпидах. Отдельные прогибы никогда не достигли такой широты, как Атлантический океан, а проявляли большую расчлененность. Это были океанические прогибы, узкие трого. Внезапный стратиграфический и боковой переход глубоководных фаций в мелководные не соответствует представлению, согласно которому антагонистический контакт глубоководных с мелководными секвенций должен быть

результатом особенно большого тектонического сокращения пространства, или же тектонического перекрытия.

Рассуждая о распространении бассейнов или трогов с океанической корой, нельзя упустить из виду секвенций, которые невозможно назвать офиолитовыми. Их существенной составной частью являются юрско-нижнемеловые осадки, батиальные вплоть до абиссальной фации подобные осадкам офиолитовых секвенций, но сопровождаемые только небольшими проявлениями основных или же ультраосновных пород.

Существуют трюги большего количества типов, также как и пороги нескольких типов (расположенные между трюгами) и шельфы. Особый тип представляют собой трюги пениникума, у которых выделяются три типа: валезский, пьемонтский, лигурский (П. Цуфарди и др. 1976). В динарском типе также надо выделить несколько типов: тип вардарский и собственный тип субпелагонский. Поясы Чахлау и Северин с префлишем представляют другой тип офиолитовых трюгов, чем тип Муреш, с точки зрения содержания гетерогенный. Типы трюгов с тонкой сиалической корой, хотя без мощных базальтоидов, также не являются унифицированными ни по своему стратиграфическому распространению, ни по содержанию и развитию. Например крижнянскому трюг с глубоководными фациями от лейаса до сеномана, с авгититами в неоконе и нижнем альбе, ярко отличается от типа Мечек. В нем с верхнего лейаса до нижнего мела чередуются условия осадконакопления трюга с порога и нижнемеловые магматиты имеют щелочной и особый петрологический характер (трахидодериты, фонолиты и щелочные диабазы).

Различия в типах трюгов обусловлены в значительной мере отличиями позиции в отдельных сегментах геосинклиальной системы альпид, отличиями в расчлененности, но также и динамике коры. При этом трюги с океанической корой наверно имеют региональный характер, как это выявляет трюг «Северный Тетис»: пениникум – нижний этаж магурского покрова – Чахлау – Северин – пояс Краина – Троянский флиш – Котель и простираение Южный Тетис: внутренние динариды–гелениды–тауриды. Трюги с океаническим типом коры вероятнее всего являются геотектоническим элементом высшего порядка – соединениями нескольких сегментов; трюги с более тонкой континентальной корой, как например крижнянский трюг, характерны только для некоторых сегментов альпид. Палеотектоническая модель во время альпийского цикла была наверно сложной и изменялась во времени и пространстве на отдельных этапах и стадиях развития. Неотделимой и самой значительной составной частью этой модели, а именно от среднего триаса до окончания складчатости, были трюги с офиолитами или с глубоководными секвенциями фаций, сопровождаемые базальтоидами или же ультраосновными породами.

д) Мезоальпийские межорогенические вулканиты и сопровождающие их плутониты банатитового типа в Среднегорье и в тимоцкой зоне Южных Карпат генетически связаны с продольными грабенами и синклиниями, выполненными свитами разнообразной фациальной палитры преимущественно морских фаций, со значительной долей флиша. При том характерно, что вулканиты появляются с постепенной океанизацией осадков. Под мощными флишевыми комплексами залегают ранние молассы, над ними мощный флиш. Более тонкая кора грабенов, расположенных в большинстве случаев на пределах блоков, напоминает рифтовые зоны. Рядом с вулканитами широкой шкалы дифференциации (риолиты – дациты, трахиты, андезиты, щелочные ба-

зальты), участвующими в вулcano-осадочном выполнении грабенов и грабен-синклиналиев, проявляются особенно часто на окраинах синклиналиев субвулканические и глубинные тела одинаково широкого вещественного диапазона (пикриты, габбро, диориты, монзониты, габбро-сиениты, граносиениты, граниты, диоритовые порфириды и гранит-порфиры) в качестве проявления повышенной динамики коры. Речь идет о вулcano-плутонических формациях на первых стадиях щелочно-известковых, позже субщелочного вплоть до щелочного типа. В Апусенах банатиты представляют преимущественно более глубинные тела с меньшей шкалой дифференциации (гранодиориты и граниты с незначительным проявлением вулкаников). От среднегорских они отличаются геотектонической позицией; связаны не со значительной долготной грабен-синклиналей, но с системой глубинных поперечных разломов. Эти ранне-позднегеосинклинальные тела не подверглись значительной складчатости.

Пространственно банатиты образуют дугу (Р. Димитреску 1965; рис. 146), более или менее параллельную с протеканием палеоальпийских структур в главной области их распространения, на Балкане и в Южных Карпатах. На основе этого легко можно склониться к выводу о генетической связи банатитов к субдукции зон с офиолитами, а именно Вардарской зоны и ее восточного продолжения (В. Бокалетти и др. 1973). Родопский и Сербийско-македонский массивы в таком понятии во время верхнего мела представляли собой островную дугу и пояс Среднегорье – Тимок – морской бассейн с большой вулканической деятельностью типа бацк-арк.

Северная часть дуги банатитов (в северной части Южных Карпат и в Апусенах) протекает диагонально к палеоальпийским структурам. Банатиты здесь проявляют значительные отличия геотектонические, но также и петрохимические. Одни соединяют генезис банатитов с субдукцией Восточных Карпат, (Д. Радулеску–М. Сандулеску 1973), другие связывают ее с субдукцией пояса Металиферес (Н. Герц–Н. Саву 1974, М. Блеаху 1976). Однако, ни предыдущие объяснения возникновения банатитов субиденционным палингенезисом (Г. Штиле 1953) из-за сравнительно малой мощности верхнемеловых депрессий в Апусенах не убедительны. Любопытно, что банатиты встречаются в тех сегментах альпид, которые не выявляют значительной полярности и более молодая грабен-синклиналь или же тафрогеосинклиналь наложена на палеоальпийские структуры.

е) Позднегеосинклинальные третичные вулканиды связаны с молодыми продольными и поперечными депрессиями, наложенными на палеоальпийские стабилизированные зоны и особенно на их тыловую зону – межгорье, позже тектонически активизированное из-за утончения коры. Высокая доля пирокластического материала значительно влияет на молассовое выполнение депрессий. Кроме участия в образовании осадочно-вулканических формаций значительной чертой позднегеосинклинальных вулкаников (с преобладанием интермедиальных пород, особенно андезитов, и с большой долей риолитов), являются собственные вулкан-тектонические формы, главным образом стратовулканы и кальдеры. Вулканизм проявляется в развитии бассейнов, в их постепенной изоляции и стабилизации, и также в геоморфологии. Доля сопровождающих более глубинных тел небольшая.

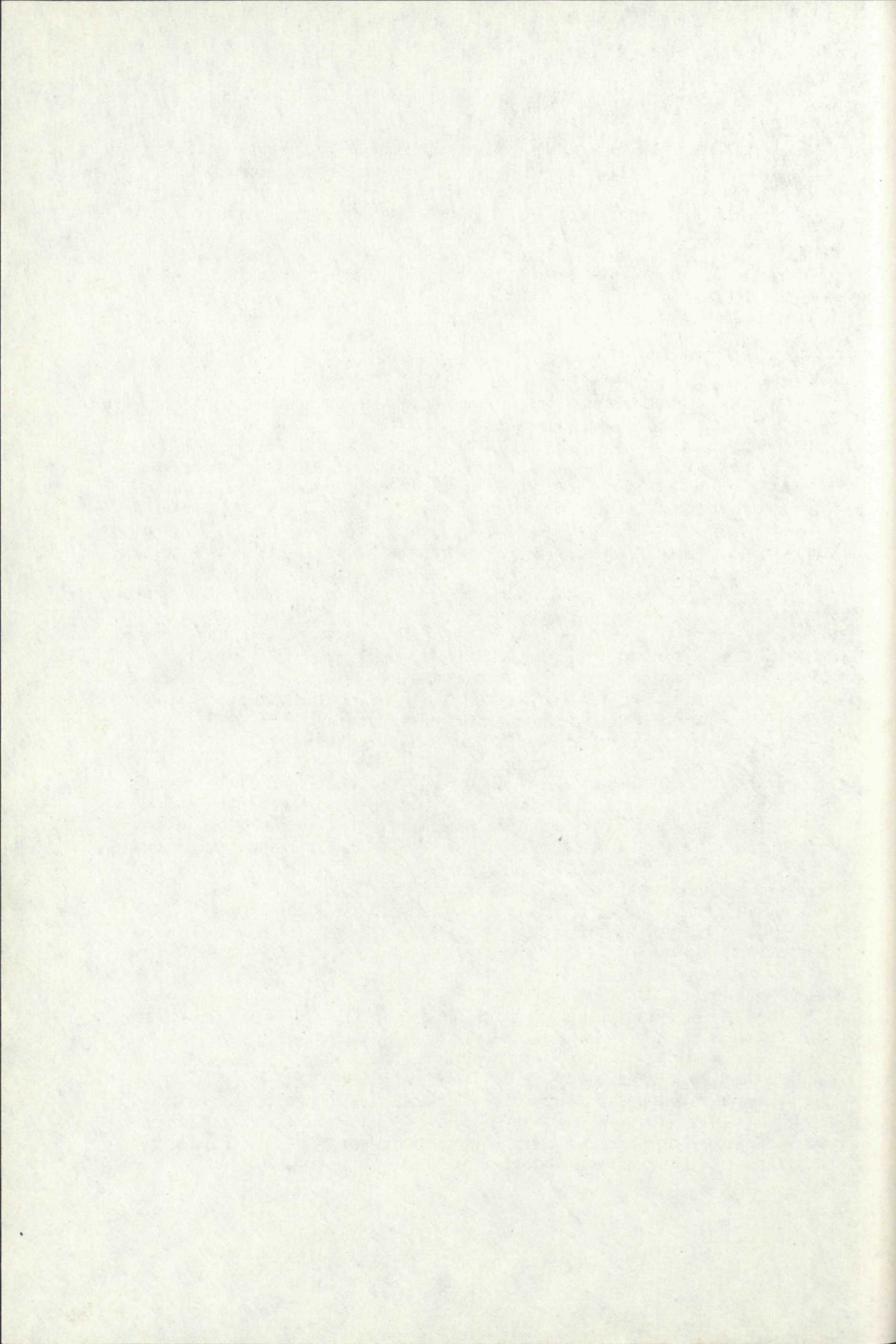
Распространение позднегеосинклинальных вулкаников в отдельных сегментах неравномерно и их типы не везде одинаковы. Их доля в Альпах рудиментарна, в Южных Карпатах они почти совсем отсутствуют, небольшое распрос-

транение у них в динаридах, за исключением областей, примыкающих к Венгерскому межгорью, и в Сербийско-македонском массиве. Мощные комплексы, расположенные на больших пространствах, строят вулканыты в Венгерском межгорье, особенно в его северной части и в примыкающих зонах Западных и Восточных Карпат. Аналогична «центральная» позиция как у Венгерского межгорья, так у трансильванского бассейна, сопровождаемого вулканытами на восточной окраине Апусен и особенно вдоль их северной окраины в самой внутренней части румынских Восточных Карпат. Тоже Родопский массив, особенно его восточная, средняя и южная части, и участки примыкающие к Сербийско-македонскому массиву, являются областями обширной деятельности позднеорогенического вулканизма. Большое количество вулканытов, повидимому, связано с зонами, стабилизированными уже доальпийской складчатостью и расширенными примыкающими областями, стабилизированными палеоальпийской складчатостью.

Позднегеосинклинальные вулканыты отнесены к двум стадиям: 1. эоцен-олигоценовой и 2. баден-плиоценовой. Обе группы связаны с депрессиями, утончением коры и дезинтеграцией. Но их пространственное расположение, как правило, различно. Эоцен-олигеновые вулканыты характерны для внутренних интрагеосинклинальных массивов Родопского и Венгерского. С точки зрения времени они представляют начало позднегеосинклинальной стадии и омоложение, или тектоническую активизацию массивов. Точное название для них — ранне-позднегеосинклинальные. Миоцен-плиоценовые неовулканыты связаны главным образом с внутригорными, интрамонтанными депрессиями и во внутренних зонах геосинклинальных систем они отвечают позднегеосинклинальной стадии. Оба типа пространственно перекрываются в регионах контакта интрагеосинклинальных массивов с внутренними зонами геосинклинальной системы, напр. в массиве Матра в Венгрии. В порядке времен оба типа связывают нижнемиоценовые риолиты, главным образом, типа игнимбритов, распространенные особенно в Венгерском массиве.

Распространение позднегеосинклинальных вулканытов в альпидах свидетельствует о том, что древние массивы в середине альпийской геосинклинали: венгерский, трансильванский, сербийско-македонский и родопский предоставляют подходящие геотектонические условия для большего распространения вулканытов. Все эти массивы находятся на рубеже альпийских ветвей различной вергенции. Они наверно сыграли важную роль в направлении глубинных процессов в течение развития геосинклинали, особенно на ее позднегеосинклинальной стадии. Тогда при замещении горизонтальных подкорных движений вертикальными произошло также поднятие подкорного диапира именно под древними массивами и следовательно и их омоложение, которое привело к их приурочению к альпийской системе. В тех сегментах альпид, у которых в тыловой области отсутствовал стабилизированный блок, расчлененный разломами и способствующий распаду на блоки или дифференциации на поднимающиеся и опускающиеся участки (напр. Восточные Альпы), не возникли ни большие внутренние бассейны, ни более распространенные позднегеосинклинальные вулканыты.

3) Финальные плиоцен-плейстоценовые базальты небольшого распространения, главным образом субщелочные вплоть до щелочных, являются представителями новой стадии в развитии геосинклинали с ярким проявлением растяжения коры при возвышенном действии вертикальных движений.



## ZÁPADNÉ KARPATY

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SÉRIA

*geológia 4*

Vydal Geologický ústav Dionýza Štúra vo vydavateľskom oprávnení Vedy, vydavateľstva Slovenskej akadémie vied v Bratislave roku 1978.

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Zodpovedná redaktorka: Irena Bročková

Technický redaktor: Jozef Bielik

Vytlačila Kníhtlačiareň Svornosť v roku 1978. 1. vyd., náklad 1200 kusov, tem. skup. 03/9. Povol. SÚKK 1197/I-1977. Rozsah AH 17,34, VH 17,59. Cena Kčs 30,—



