# ZÁPADNÉ KARPATY

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*geológia 12*

František Čech

## Dynamics of Neogene Carpathian Basins in Relation to Deep Structure, Crustal Type and Fuel Deposits

## Dynamika neogénych karpatských panví a ich vzťah k hlbinnej stavbe, typom kôry a ložiskám palív

GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA, BRATISLAVA 1988

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## Contents

## Introduction

Sedimentary basins forming on continents or on their margins are economically significant with respect to deposits of coal, oil and gas. Some basins also comprise geothermal energy resources. In the sedimentary filling of these basins are frequent deposits of non-metalliferous raw materials. Basins adjacent to orogens and volcanic zones are more significant owing to ore deposits occurring near energetic resources. Extensive geological and geophysical explorations inspired by the economical significance of basins revealed the deeper structure of the basins. The geological and geophysical data facilitate the study of the origin, history and geologic character of the basins.

These phenomena enable a rational management of further explorations and the determination of their strategy. Since exploration of the basins is very expensive, its strategy is economically important and influences the risks of the explorations.

Direct information on the structure and composition of deeper parts of the Earth's crust can only be obtained from deep boreholes realized in basins mostly for the purpose of prospection for oil and gas. Also the second deepest borehole in the Soviet Union — Saatly — is situated in a basinal structure.

In spite of many deep boreholes we still must rely upon indirect — geophysical information when studying the deep parts of the Earth. This is why we still do not know the causes of deep processes; nor their dynamics. With respect to this, geology is only dealing with models and hypotheses, frequently contradictory. The main contradiction is in the study by means of fixistic and mobilistic tectonic models.

With respect to human life and a short development of geology as science, we have only partial information on deep processes, based on their short manifestations. Studying the processes we are aware of the seemingly paradox Heisenberg's idea about "the amount of unknown phenomena increasing owing to the knowing process". This is why in the early stage of geology only few hypotheses about the origin of geologic units were presented. At present there are tens of hypotheses and models of only the origin of basins. And many hypotheses are contradictory. Since we do not know the substance of endogenous dynamics we cannot formulate a categorical global hypothesis or theory.

In this book I present a survey of the most important data and opinions about the origin of sedimentary basins, and discuss actual models of their origin, particularly of basins in the Carpathian system. Many information and opinions were presented at the 27th International Geological Congress in Moscow, in August 1984.

Some Carpathian basins are close to Mediterranean basins and have some features in common with them. So I also discuss the basins with respect to their characters correlable to the Carpathian basins. As it is impossible to accept uncritically new hypotheses, the reader of this book is asked for a critical approach to hypotheses formulated or advocated here.

It is difficult to reconstruct the history of sedimentary basins, so the concepts are and will be discussed by authors of various opinions. I am mainly considering plate-tectonic aspects with the purpose of a critical evaluation of models — sometimes extremely dynamic, without the necessary data known.

The main purpose of this book is the solution of the problem concerning the relation of the coal-bearing and oil- and gas-bearing Carpathian basins to the deep structure. My interpretation of the deep structure and the solution of the problem concerning the foundation of basins is based upon data on the block structure and deep faults, and on information about crustal types in the surroundings of the basins. I have tried to reconstruct the crustal type in the basement of the basins on the basis of their history and according to scarce information about the composition of the basin basement rocks. The complex approach to deep structure facilitated the solution of the problem on a new qualitative level, and affected the determination of other prognostic criteria and areas.

The book consists of four parts:

1. The general part concerning endogenous processes and problems of the foundation of basins and their classification.
2. The regional part concerning basins in the Alpine-mobile Europe including the Mediterranean area.
3. The part dealing with the deep structure of Neogene basins in the West Carpathians.
4. The part concerning productive basins in the Carpathian-Balkan region.

## 1. Processes stimulating the generation of continental basins

Continental basins s. l. comprise all sedimentary basins — marine and non-marine, forming on the continent and on continental crust or at the margins of the continent. These are non-oceanic basins.

The history of basins represents another problem. At the beginning of their development or during their development the continental basins formed on the non-continental crust, perhaps on oceanic. Generally we can base our interpretation on a simple evolution model: oceanic — intermediary — continental crust. The types of basins changed respectively. Continental basins formed in areas, segments or belts of new or ancient crustal sialization but I think that the basins may represent areas or zones of unevolved or stopped, slowed or very weak sialization as a progressive process. The second trend is a regressive process in crustal evolution: basification, oceanization of continental crust. Generally, in the terms of plate tectonics the process may be denoted as a new formation of the oceanic crust — a process with various mechanisms and dynamics.

The origin of basins is associated with endogenous primary or secondary thinning of the Earth's crust. The thinning and weakening of continental crust also comprises the fault fragmentation of the crust besides basification.

At present, geologists associate the origin of basins with endogenous regimes with their source in the Earth's mantle, and perhaps even in the outer core. Older ideas of the exogenous origin are being abandoned at present (F. L. SCHWAB 1976).

a) The exogenous origin was associated with the denudation — erosion processes, and relief depressions were presumed to form on the former mountain range, levelled by denudation. Their sedimentary filling caused the crustal bending (J. HALL 1859). E. G. HAUG (1900) also infered primary deep-sea basins, gradually filled with sediments and altered to fresh-water basins.

b) Endogenous origin of basins was studied by J. Dana who explained the bend (a geosyncline in Dana's sense) by a lateral pressure. The first hypotheses about the origin of basins concerned geosynclines, and many continental basins were regarded as parts or remains of geosynclines. The role of faults in the genesis and subsidence of basins started to be emphasized as late as in the thirties of the century, owing to investigations and experiments by H. CLOCS, followed by R. A. SONDER (1953) who know the primary role of lineaments. H. STILLE (1924) presumed that the faults arose in the later stage of the development of geosynclines under the influence of the mass of sediments.

A geosynclinal model comprising a couple system of geosyncline-geanticline, or several couples of intrageosynclines-intrageanticlines (V. V. BELOUSOV 1954, J. AUBOIN 1965 a. o.), undation movements (J. W. UMBGROVE, R. W. VAN BEMMELEN), inversion movements associated with the origin of foredeeps, intramontane and intermontane depressions, movements of median massifs serve as a basis for various models presented by Soviet geotectonists. They study deep faults and their role in the dissection of geosynclines, in the borders of basins and in inversion movements. The geosynclinal model is still accepted and plate tectonists discuss its importance and reasoning. The concepts of the origin of continental basins are based on geotectonic hypotheses. One of the oldest — but still actual in renewed variants — is the hypothesis associating the origin of megastructures with intrusive magmatism. These ideas resulted in models of convection, subcrustal erosion, basification, a. o.

It is difficult to classify single hypotheses because the substance and dynamics of processes are either similar or have the same deep sources. So the classification of the problems of models is concentrated to the character of processes and some specific features — as far as the models are not based upon quite different principles. So the problems concerning analogous themes also comprise different models if aimed at the explanation of geotectonic processes other than the origin of continental basins, and based upon the same endogenous processes like magmatism. Present models applied on the Carpathian basins will be discussed in detail in the chapter concerning the basins.

Foundation and subsidence of basins have endogenous causes, according to A. G. FISCHER (1978) resulting from:

1. Changes in isostatic regime, due to: a) movements of masses because of changes in the mobility of material (its rheologic properties), and mainly to magmatism; b) volume changes (thermal differences, phase changes); c) changes

in crustal thickness and density; d) disequilibrate states in the asthenosphere;

2. changes in tectonic regime, caused by mass migration and changes in material composition and physical state of masses.

The last factor actually comprises the preceding processes and the changes are due to folding, fault movements or movements of lithospheric plates. Processes presented under a, b, and d, are associated with diapirism, and processes under d should be complemented with the phenomenon of the crustal type.

### 1.1 Magmatism — the principal factor in mass migration and in density changes in the lithosphere

The influence of magmatism upon the stretching and fault segmentation of the Earth's crust in an elevated dome above the magma pillow intrusion was for the first time mentioned by L. v. Buch as early as the first half of the 19th century. It is the principle of the formation of basins above mantle diapirs. The principle is actual at present.

Most hypotheses on dynamics of the crust or of the lithosphere are based upon the influence of magmatism.

Crustal types differ in the composition of magmatic rocks. If sialization is a process of enrichment of the crust with light lithophilous elements, then basification is a reverse process.

Both processes form:

a) density stratification in the crust, i. e. distribution of rocks with their density increasing from the Earth's surface to the depth,

b) density inversion — when rocks with a higher density are underlain by rocks of a lower density.

The density inversions in crust cause isostatic instability and higher mobility of higher or hot masses, leading to the formation of diapirs (H. RAMBERG 1967). Melting of rocks causes changes in their rheologic properties, and their differentiation results in a separation of lower density rocks which try to migrate toward the level of a lower geostatic (lithostatic) pressure where rocks of lower density are stable. Diapirism is a frequent process in the Earth's crust (salt diapirs, granitoid diapirs) and the upper mantle. G. N. BUGAEVSKYJ (1977) basing on seismologic data interpreted diapirism from the Earth's outer core and in the lower mantle. Due to volume changes diapirism causes instability in the crust (mantle) and may be a source of further endogenous processes like thermal differentiation, phase changes, fluid and mass movements, the formation of secondary magma chambers, a. o.

The density inversion owing to granitization (e. g. in paragneiss complexes with a higher density) is associated with the formation of granitoid diapirs. Around diapirs rim synclines form with rests of denser rocks, into which the heavier masses, mainly basic rocks, can migrate and concentrate. They mostly fringe granitoid elevations and also occur in depressions among elevations. W. KREBS and H. WACHENDORF (1973) applied the principle of granitoid diapirism associated with subsidence of adjacent basins, upon the evolution of Hercynian Europe. Diapirism will also be discussed in a separate chapter (5.1).

Phenomena of tension above a dome caused by magmatism, of rifts and graben-like basins were applied in the hypothesis of geo-tumors of E. HAARMANN (1930), in the undation hypothesis elaborated in detail by R. W. VAN BEMMELEN (e. g. 1972) — concept of asthenoliths in the mantle —, in the hypothesis of convection currents, actualized in the plate tectonic model (mantle plumes), or in the concept of advection of hot masses, treated by Soviet tectonists, mainly V. V. BELOUSOV. The advection model is mostly based on vertical movements: uprise of lower-density masses and descent of higher-density masses in crust and in mantle without any more extensive horizontal movements — as presumed in convection models. In the case of continental basins advection is reasoned. Advection of basic rocks differenciated from the mantle, may be explained by basification.

## 1.2 Basification

Basification is a process of partial or total replacement of continental crust by oceanic. This is why the process, is often denoted as oceanization. It also comprises volume changes in the lithosphere and in crustal thickness, but it is treated in this part of the book because of its prevailingly magmatic nature. Basification proceeds also on continents without the formation of an ocean. In the model the primary heterogeneity of the crust is denied and the global continental crust is preferred.

From the isostatic view, basification disturbs the equilibrium in the crust by increasing its density. It is derived from some changes in the nature (e. g. plateau basalts or the so-called sandwich structures in metamorphites). In the global relation the approach to the mechanism is speculative. After a certain hesitation also defendants of mobilism accepted basification as a fixistic hypothesis. For example Z. ROTH (1980) associated the high Alpine foldability of the North--European Platform and the formation of some Mediterranean oceanic basins to basification.

Basification is rejected by many authors, mainly with respect to the origin of oceans. There are many doubts about the origin of basins as well. D. J. J. KINSMAN (1978) doubts about the effects of basic intrusions — unless their amounts are greater than masses known. There is a little difference in density when compared to the lower crust density. So only mantle intrusions (diapirs) can display a gravity effect. For example E. V. ARTYUSHKOV (1979) does not agree with Belousov's model, and although he admits basification, his model is different.

V. V. BELOUSOV is the greatest propagator of global basification. According to him (1982) the basification model was introduced by A. D. ARKHANGELSKY who wrote in 1941 that the crust in the area of the Atlantic Ocean was the same like beneath adjacent continents, only the rocks in the lower parts of the crust changed. The change might have been caused by huge intrusions of heavy magmatites whose effusions in the form of plateau basalts are known on adjacent continents.

The idea of basic metasomatosis as basification mechanism was presented by V. V. TIKHOMIROV (1958) and S. V. MOSKALEVA (1956) in V. V. BELOUSOV (1982). Mediterranean basification was ascribed a specific metasomatic character by R. W. VAN BEMMELEN (1972). According to him the suprising branches of convection cells destroy the crust (subcrustal erosion — see Chap. 3.1) and break off blocks dragged by currents into the mantle. This results in disequilibrate states and phase changes in the mantle, and crustal blocks get oversaturated with femic components and their density increases. Crustal material "dissolves" in the mantle. It is a process similar to subduction — consumption of the lithosphere.

The problems in metasomatic concept are removal and new accumulation of sialic masses. It is similar in models of deep erosion a. o. (see Chap. 3.1). In the surrounding of basified crustal segments equivalent volumes of altered sialic crust miss acid magmatites and volcanics younger than or coeval with basification. H. RAMBERG (1967) explained basification by granite crust being covered with thick lava flows of basalts, by subsidence and lateral flows of granite rocks. Examples of such a process resulting from inversion of densities in the crust are known from Precambrian complexes (the Kola peninsula) but they do not enable us to explain sialic orogenic rims around basins, neither the circumoceanic orogenic sialic crust. V. ŠKVOR and J. ZEMAN (1976) denote the process of saturation of the crust with basalt intrusions and covering of the crust with basalt effusions as basaltization, i. e. a kind of basification. Basalt traps represent one phenomenom of basaltization. E. G. ZHURAVLIOV (1984) revealed a 1.222 m thick layer of 11 Mesozoic basalt lava flows of basalts, covered with a 3.274 m thick layer of Meso-Cenozoic sediments in the borehole Ombinskaya 2 in Western Siberia. The traps are very frequent and cause increasing crustal density and its subsidence.

V. V. BELOUSOV (1982) presented a new variant of global basification. Its mechanism is in principle similar to R. W. VAN BEMMELEN'S (1972) model. The cause of basification is penetration of low-viscous basics and ultrabasics from the mantle to the continental crust. According to van Bemmelen, density of more viscous lithosphere increases owing to basification and surpasses density of the asthenosphere. The problem of density is a weak point of the model, often criticized by physicists. Blocks of the lithosphere sink into the mantle to dissolve there. The continental crust should be replaced by new oceanic crust.

A deep cause is in the anomalous heat flow resulting in the melting of peridotites or pyrolite, and differentiation of basalt, and decreasing viscosity and average mantle density. This should lead to elevation of the asthenosphere whose density in its upper part is hypothetically lower than density of the lithosphere and so it sinks into the asthenosphere. Belousov's concept is based on geophysical data about rift asthenoliths with velocities of P-waves by 0.3—0.6 km/s lower than in the surrounding mantle, i. e. also in the lower part of the lithosphere.

The calculated density difference is by 0.03—0.04 . $10^3$ kg . $m^{-3}$ smaller than in the normal mantle.

The volume of basalt melt is in average estimated to 15 percent of volume of the asthenosphere. The melt with the density by 0.1 . $10^{-3}$ kg . $m^{-3}$ lower than the density of ultrabasics leads to density disequilibrium and to the ascent of basalts (by diapirism or along deep faults) into the lithosphere. A system of basaltoid intrusions forms then in the lithosphere.

In the elevation of the asthenosphere less viscous masses flow downside the limbs of the elevation. The mass movements are associated with tearing away of the lower lithosphere blocks. The blocks (with basic intrusions) are heavier, have a greater density and sink into the anomalous lighter hot mantle. The outpressed mantle extends higher up into the crust. This is the first stage of basification. (The mechanism is close to the model of subcrustal erosion.)

The invasion of the crust by the mantle commences another stage of basification favoured by thermal activation of the mantle and fault fragmentation of the crust, supporting dense basic intrusions. Statified crust is favourable for the formation of sills. Most basics are in the sills and in plateau basalt sheets. So it is not inevitable to presume crustal spreading or the Earth's expansion. Density and weight of the crust increase with its saturation with basalt and gabbro. The density and weight of the crust also increase at overheating of rocks, causing decrease in porosity, dehydration and higher-grade metamorphism. It should be noticed that zones with basics (simatic crust) are not associated with higher--grade metamorphism. V. V. Belousov assumes that basification is facilitated by high-grade metamorphism, for example granulite facies.

Waters released from overheated rocks and from magma (average $H_2O$ content in plateau basalts is 1.82 percent) should concentrate in basins (oceans). Their volume decreases (by about 10 percent) at the increasing crust density, the formation of rifts and faults proceeds without any crustal spreading.

V. V. Belousov (1982) explains intrusions of hot ultrabasic rocks into the crust and thus increasing density of the lower parts of the crust, by anomalous overheating of the mantle. Melting of ultrabasics (at partial presence of water t 1.300 °C) may cause decrease in the mantle density. Blocks of basified crust break off into the anomalously low-viscous mantle and are replaced by basic (± ultrabasic) intrusions. Further basic extrusions extend on the Earth's surface.

Belousov presents other factors in favour of basification, partly included in other models (see lower):

1. Extensive surface erosion and deep denudation thinning the granite crust (basification mostly affects platforms and median massifs).

2. Low viscosity of basalts, facilitating their mobility and extension.

3. Petrochemistry of oceanic basalts is indicative of their origin from the mantle, partly devoid of volatile elements, and similar to the mantle below continents.

4. Contents of radioactive elements in mantle below continents and oceans are approximately equal and should so prove the common mantle of pancontinental character.

5. Increased number of basic intrusions in some places of continental rims, and plateau basalts on the continental crust close to oceanic. The paragraph on basification is somewhat larger bacause some phenomena and ideas must be considered in the evaluation of models of the genesis of continental basins. Since there are different concepts of basification, it is better to use the general term, namely basification s.l. It is a process of increase in density and weight of the crust (without respect to crustal types) resulting in its higher mobility, fault fragmentation and subsidence.

## 1.3 Volume changes in lithosphere

Volume changes may be caused by thermal differences, mainly by increasing temperatures. They also cause phase changes, control the fluid content and pressure changes. Volume changes caused by other dynamics of the lithosphere will be discussed below.

### 1.3.1 Thermal differences

Changes in density and volume of rocks in the crust may be caused by increasing heat flows in active volcanic areas and in areas of accomplished neovolcanism and postvolcanic phenomena. Thermal differentiation is also due to hot spots, ascents of asthenoliths and magma. It is most frequently associated with high tectonic mobility and subsurface activity — as proved by positive correlation of elevations of the asthenosphere to high heat flow. In such areas are active basins with volcanism (e. g. Tyrrhenian Sea, Aegean Sea). In the Pannonian Basin, where volcanism at the margins of the basin ended in the Quaternary, it is not clear whether the endogenous activity is ceasing or renewed. Then the basin would be in a stage preceding a new intense subsidence.

According to V. V. BELOUSOV (1978a) basalt melt originates in the depth of 100 km at 1400 °C. A hot mantle needs not be overlain by a hot crust and vice versa. A hot crust with a cold mantle should characterize the post-inversion stage of geosynclinal development, i. e. the formation of continental basins. This idea is, however, denied by data on the areas of the Mediterranean Sea, Black Sea and the Pannonian Basin. Subsidence in basins may proceed under the conditions of normal and anomalously high heat flow originating from the mantle and not localized only in the crust.

D. P. MC KENZIE (1978) regards increasing temperatures and consequent gradual cooling as very important for subsidence. Rapidly increasing temperatures cause overheating and stretching of the lithosphere, i. e. a reverse process, different from models in which overheating is associated with increasing volume and crustal elevation. The model of the stretching lithosphere will be discussed in the Chapter 4.2.

E. V. ARTYUSHKOV and M. A. BEER (1983) regard thermoelastic contraction of crust and mantle at the cooling of the lithosphere as one of possible causes of crustal subsidence following the overheating and elevation of the crust. This process might only have resulted in small intramontane basins, to A. E. SVYATLOVSKY (1984) relates the volcano-tectonic dome collapse, associated with the end of the main volcanic activity in oceans to regional regressions of seas on continental margins, and to the formation of oceanic troughs. Marginal basins on the inner side of volcanic island area are denoted as collpase-subsiding basins by SUZUKI YASUMOTO (1984).

All the models quoted relate the origin of basins and the collapse to movements in mantle diapirs.

Thermal differences caused the phase- and material changes in mantle and crust, and changes in their thickness. The origin of basins and subsidence ranging to 10 km are besides basification explained by dehydration of ultrabasic rocks or by eclogitization of gabbros.

### 1.3.2 Serpentinization — deserpentinization and metamorphic differentiation

H. H. HESS'S (1962) model of serpentinization of ultrabasic rocks was modified and extended mainly by I. A. REZANOV (1977). In a crust predestined to oceanization owing to its material composition, ultrabasic rocks in its lower part and in the mantle should be hydrated at a low heat flow and metamorphosed into serpentinite with the density of about $2.6 \,.\, 10^3$ kg . m$^{-3}$. The decrease of density causes isostatic uprise. The Mohorovičić discontinuity, regarded by I. A. REZANOV (1977) as a phase boundary, i. e. the boundary of (de-) serpentinization migrates deeper in the elevating units. Geophysical records show a great crustal thickness. The increase of temperature to 500 °C in predestined crust (in my concept in the simatic crust) is followed by deserpentinization. Light serpentinites in the lower part of the crust, loosing water, change into heavy ultrabasic rocks with the density of $3.3 \,.\, 10^3$ kg . m$^{-3}$. The lower part of the crust and the mantle have the same composition and physical properties. The Mohorovičić discontinuity rises up and the crust gets thin. This is followed by subsidence, by the formation of basins and oceans. I. A. REZANOV (1980) formulated another model, namely the model of metamorphic differentiation into acid and basic complexes in the crust. He assumed that owing to the differentiation the deep part of the crust had a stripped migmatite structure, and light aplite layers alternated with dark amphibolites. Rezanov's concept is based upon sections of deep parts of the crust in shields. This concept is partly supported by data from a superdeep borehole SG-3 on the Kola peninsula (Fig. 1). There in depths surpassing 10 km are sialic metamorphosed rocks of granitoid nature (mostly granitized gneisses), laminated with amphibolite bands — the sandwich structure type. According to REZANOV by the overheating of the crust the light sialic components are segregated due to anatexis or hydrothermal solutions and accumulate in the upper part of the crust. Dark femic components remained in their place or sagged down and formed the present lower "basalt" part of the crust. In its composition this part of the crust may even be related to the ultrabasic mantle. Acid components enter both the crust and the sea water. The lower basic part of the crust is affected by (de)serpentinization processes.

According to V. V. BELOUSOV (1982) the process of serpentinization in a global scale is doubtful since ultrabasic rocks and xenoliths in volcanics do not display any indications of present or fossil serpentinization. V. V. BELOUSOV doubts about crustal serpentinization as a dominant process and admits only its local extent in the zone of water circulation, perhaps not even deeper than

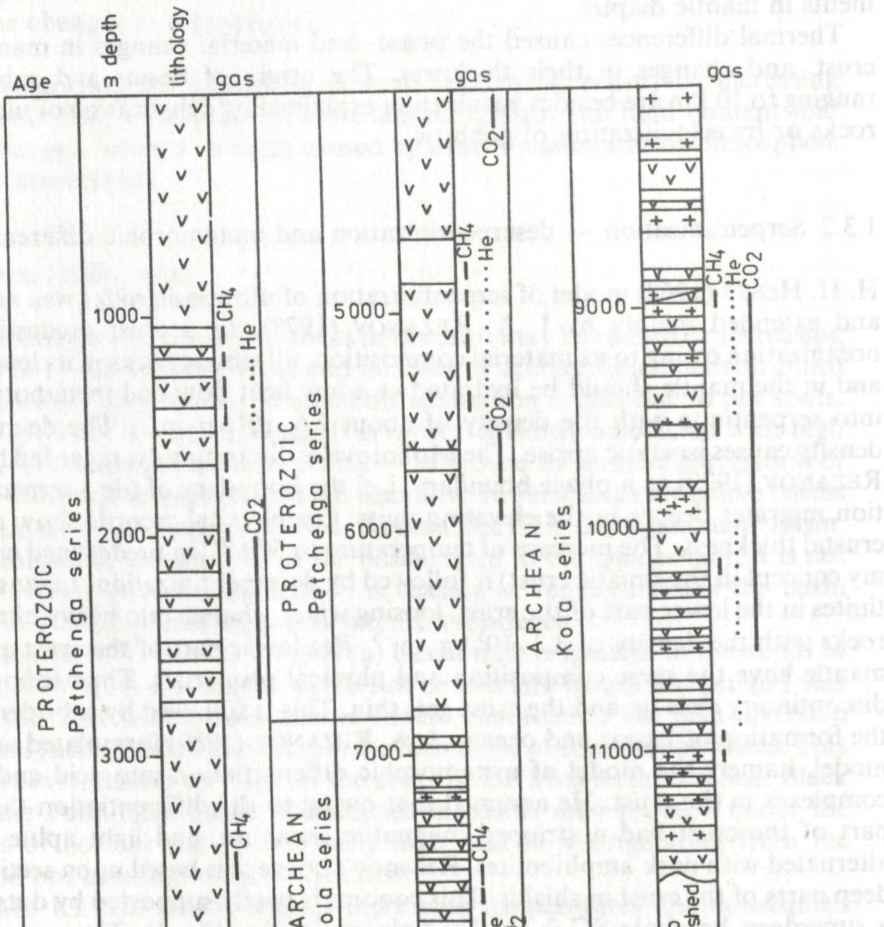



Fig. 1 Scheme of the geological profile of the superdeep — hole Kola SG-3 in the U.S.S.R. (After E. A. KOZLOVSKY et al. 1984, strongly simplified and modified).
Gas content is related to 1 $m^3$ of the rock: Only intervals of anomalous gas contents are plotted on the left side: He 4—15 . $10^{-3}$ $m^3/m^3$, $CH_4$ 0.5—2.5 . $10^{-1}$ $m^3/m^3$, $CO_2$ 1—3 . $10^3$ $m^3/m^3$. Profile illustrates the composition of the simatic-type crust.

20 km. According to V. V. BELOUSOV (1982) the model of the crust as a megamigmatite would only be reasoned with respect to temperatures about 700 °C necessary for anatexis of sialic components, and only in depths exceeding 5 km. Further problems concern release of $SiO_2$ into sea water which should be then extremely rich in $SiO_2$. No high $SiO_2$ content in sea water is known.

Both Rezanov's models are suggestive for non-oceanic crustal units and should be taken in consideration in cases where fossil circulation of water may be presumed as well as a favourable type of the crust. We do not exactly know the deep distribution of waters but B. G. LUTC'S (1975) data show that dissociated H + OH, acid character of fluids and leaching ability of acids may also

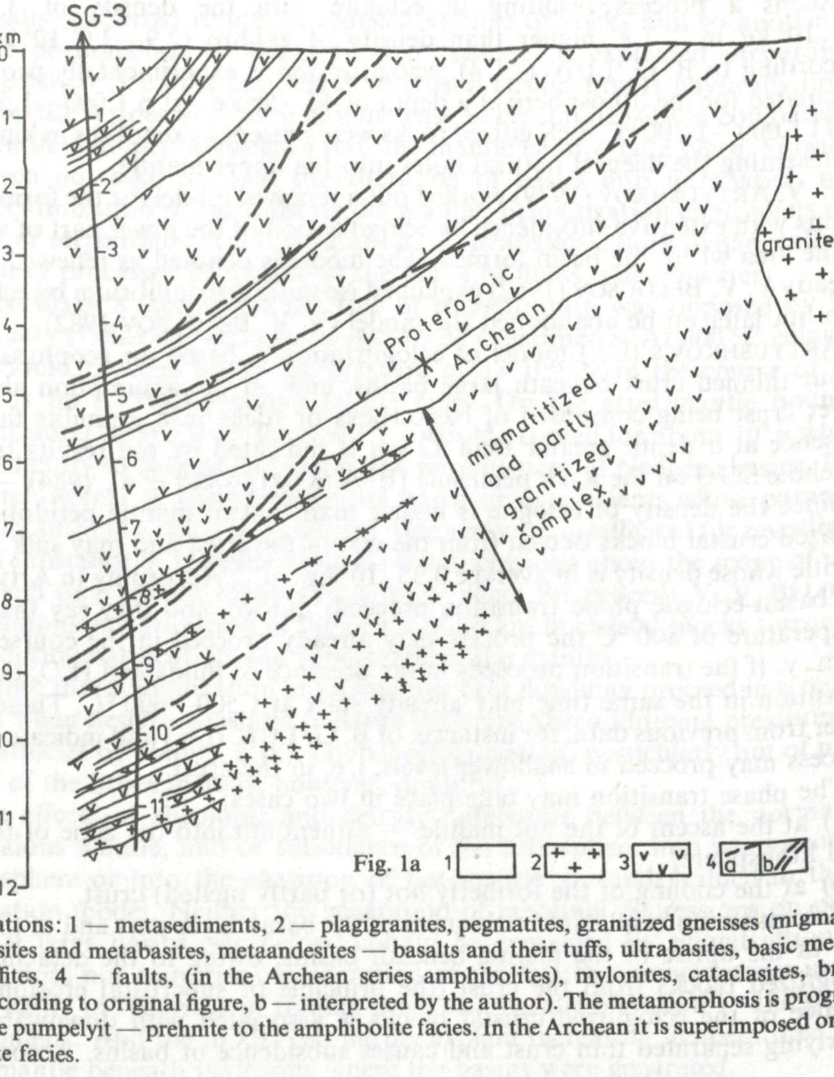



Fig. 1a

Explanations: 1 — metasediments, 2 — plagigranites, pegmatites, granitized gneisses (migmatites), 3 — basites and metabasites, metaandesites — basalts and their tuffs, ultrabasites, basic metatuffs and tuffites, 4 — faults, (in the Archean series amphibolites), mylonites, cataclasites, breccias (a — according to original figure, b — interpreted by the author). The metamorphosis is progressive from the pumpelyit — prehnite to the amphibolite facies. In the Archean it is superimposed on older granulite facies.

exist in the mantle. So far we know only little about fossil temperatures in the crust, necessary for metamorphic differentiation. In complexes, metamorphosed in amphibolite facies, differentiation results in accumulation of leucosome in migmatites in anticlines and in decrease of leucosome in synclines. Rezanov's model of deep crustal differentiation is not proper for depths smaller than 5 km because there processes characteristic of shallow crust prevailed.

### 1.3.3 Eclogitization

Eclogitization of gabbroid rocks, presumed to exist in the lower part of the crust, is a process resulting in eclogite with the density of $3.5 . 10^3$—$3.6 . 10^3$ kg . $m^{-3}$, i. e. higher than density of gabbro (2.9—$3.0 . 10^3$ kg . $m^{-3}$). According to B. G. LUTC (1974), eclogitization is experimentally proved and calculated for the lithosphere at a depth of 40—50 km, at p 1.000—1.500 MPa and t 1.000—1.200 °C. Eclogitized rocks were present as xenoliths in kimberlites representing the deepest natural holes into the upper mantle.

E. V. ARTYUSHKOV (1979) worked out a renewed model for the formation of basins with extensive subsidence by eclogitization of the lower part of the crust in the area where the basin formed. The model is denoted as renewed because already V. V. BELOUSOV (1962) explained isostatic disequilibrium by eclogitization but later on he abandoned the model (V. V. BELOUSOV 1982).

ARTYUSHKOV'S (l. c.) model of eclogitization is based on geophysical data about thinned crust beneath large basins, and on the assumption about the lower crust being composed of basic rocks or rocks in a granulite facies. Its presence at a depth greater than 12 km is indicated by the results from the borehole SG-3 on the Kola peninsula (E. A. KOZLOVSKY et al. 1984) — Fig. 1.

Since the density of eclogite is higher than that of mantle peridotites, eclogitized crustal blocks detach from the rest of the crust and may sink into the mantle whose density is in average $3.35 . 10^3$ kg . $m^{-3}$. According to Artyushkov the basalt-eclogite phase transition proceeds quickly and they say that at the temperature of 800 °C the process may already proceed in the course of 1—10 m . y. If the transition proceeds in the presence of fluids and $H_2O$, the phase transition in the same time may already start at t 500—600 °C. These values differ from previous data, for instance, of B. G. LUTC (l. c.) and indicate that the process may proceed in shallower levels, i. e. in the crust.

The phase transition may take place in two cases:

a) at the ascent of the hot mantle — asthenolith into the zone of formerly cold lithosphere,

b) at the cooling of the formerly hot (or partly melted) crust.

Eclogitization results in the thinning of the basalt layer part and of the crust, and in the uprise of the mantle beneath basins, owing to the detachment of eclogitized blocks from the crust (the principle of subcrustal erosion). Subsidence of the eclogitized crustal blocks is associated with downwarping of overlying separated thin crust and causes subsidence of basins. According to

Artyushkov (l. c.) the extent of subsidence is controlled by the original thickness of the basalt layer prior to eclogitization, and by the extent of eclogitization. Additional subsidence (about 1—2 km) may be associated with the cooling of the low velocity mantle beneath the crust. The influence of the weight of sediments is negligible.

Eclogitization at the cooling of the crust may proceed beneath orogens with volcanism and the mantle temperature about 1200 °C.

If basalts at the cooling of the crust or mantle get to depths with temperature 700—800 °C, rapid eclogitization may take place.

Artyushkov also explains quick subsidence in basins and rapid transition from shallow-water to deep-sea sedimentations by eclogitization. Eclogitization of basalts (gabbros) results in reduced volume of rocks and so another subsidence phenomenon — filling of free "spaces" is explained. According to Artyushkov (l. c.) the hypothesis of crustal basification is physically impossible because it does not cause any increase in crustal density in comparison to the mantle and any subsidence into the mantle cannot take place. Subsidence can even not take place at the alteration of basics into ultrabasics, nor at deserpentinization of the crust or the mantle. Eclogitization contradicts to the model of serpentinization (hydratation of ultrabasics). Both dynamically contradictory processes take place at the influx of fluids, mainly water.

Artyushkov's concept of the role of eclogitization was opposed by V. V. Belousov (1982) on the basis of petrologic experiments. At 800 °C eclogitization should proceed under the pressure of 2 × 109 Pa in the course of $10^6$ y. There is no so high pressure in the crust. On the crust/mantle boundary, eclogitization (at p $1 \times 10^9$ Pa, t 500 °C) would proceed for about $10^9$ y which is contradictory to data on the origin and development of Neogene basins. Rejection and defense of eclogitization are based on experiments whose parameters were obtained in a short (laboratory) time and were insufficient for revealing the actual substance of eclogitization. So the discussions about the scope of values of pt and time cannot reliably confirm or reject the process. V. V. Belousov (l. c.) admits eclogitization at the depth of 60 km in crustal blocks torn off the crust during basification and submerged in the mantle.

I think that eclogitization at anomalous heat flow may proceed in a regional extent when besides geostatic pressure there are also additional pressures (tectonic stresses by diapirs, a. o.). Global eclogitization, particularly out of mobile zones of the lithosphere is, however, doubtful.

The effects of rheologic and density differences between the normal and anomalous mantle, and of subsidence of the lithosphere into less viscous asthenosphere or into the elevation of hot mantle are undervalued in the eclogitization model. Neither the generation of marginal depressions on peripheries of large basins (see below) cannot be explained by eclogitization. The model does not comprise consideration of the fact that temperatures increase more easily in segments of thinner crust and mantle elevation — the primary phenomena. And the model can hardly explain increasing temperatures in the cold mantle beneath platforms where the basins were generated.

## 1.4 Changes in crustal thickness and in the lithosphere

Changes in crustal thickness as a cause of the generation of large basins became a basis for hypotheses after geophysics had revealed that the crust was thinner beneath basins (and oceans) than in the surroundings of basins. Besides the above mentioned hypotheses the causes were also searched in the infracrustal erosion and lithosphere spreading.

### 1.4.1 Crustal erosion

A hypothesis on erosion of the basal part of the crust was introduced by J. GILLULY (1955). He presumed that roots of orogens, made of lighter masses than the surrounding mantle, are "eroded off" by subcrustal processes. A contemporaneous denudation of mountains will result in crustal thinning and the thin crust will become the basement for the formation of a basin. Erosion should be caused by convection currents. The emplacement of eroded-off masses is problematic. If they would accumulate in adjacent crustal elevations in the form of magma and volcanics, the volume of these rocks should be equal to the volume of eroded rocks beneath new-formed basins. But, for example in the West Carpathians the area of Tertiary magmatites is by more than 50 percent smaller than the area of basins near the periphery of the Pannonian Basin. The volume of volcanics will be still smaller than the volume of sediments in basins. The same concerns the Pannonian Basin where the crustal thinning is explained by deep erosion by L. STEGENA et al. (1975). A particular modification of subcrustal erosion is presented in a model by R. W. VAN BEMMELEN (1972), V. V. BELOUSOV (1982) and in E. V. ARTYUSHKOV'S (1979) eclogitization model — chapters 2.1 and 3.3. Evidence about the relation of mantle elevations to basins in the Mediterranean area inspired an idea of the origin of a supra-mantle crustal elevation with a granite layer removed by surface erosion and denudation. The remaining thinned crust with heavier basalt layer was presumed to sink into the mantle elevation (K. J. HSÜ 1968, R. D. SCHUILING 1973).

According to R. BRINKMAN (1974) the Black Sea was generated by a similar mechanism. There, however, is a problem that around the basin accumulated clastic sediments as products of extensive and deep denudation of to 20 km thick granite crust are absent. The generation of basins (often of molasse type) should be preceded by immense accumulation typical of molasse basins in Alpine (high-mountain) areas. The model of surface crustal erosion as a cause of the generation of thin to oceanic type crust (i. e. when the composition of the continental basalt layer should be identic with oceanic crust) beneath basins is unacceptable. Accumulation of masses removed from the lower parts of the crust also represents an unsolvable problem in this hypothesis as well as in others, like in the hypothesis of basification, convection currents a. o. Sometimes the term "deep erosion" is generally used for crustal thinning without any explanation of the mechanism of crustal thinning or the term is used instead of "basification".

### 1.4.2 Stretching of the lithosphere

The generation of basins was already explained by crustal stretching in Wegener's concept of the continental drift mechanism. The stretching associated with the generation of faults was studied on rifts and related to uplifts. The ideas of viscous-elastic stretching of the crust or lithosphere as a whole were corrected by geophysical information on the relation of rifts to deep mantle elevations and asthenoliths of basic rocks with P-wavers velocities higher than in the crust and lower than in the normal mantle. In the model the rigid (brittle) upper part of the crust and the ductile lower level are presumed. As far as there is no evidence of a new-formed oceanic crust in the basement of basins (and rifts), we cannot assume crustal stretching neither distancing of crustal blocks — by divergent movements of blocks or plates. D. P. MC KENZIE (1978) presumes stretching of the lithosphere owing to its quick overheating. He also presumes that the upper part of the lithosphere is rigid and with deformations in block segmentation. The lower part of the lithosphere is ductile and can get thinner by stretching due to horizontal movements. D. P. MC KENZIE (l. c.) presumes even double stretching of the lithosphere, in the rigid level associated with the generation of listric faults and with subsidence of blocks accompanied by the generation of basins. Rheologic properties of the ductile lithosphere result from its overheating and its thinning is associated with passive ascent of anomalous mantle. The subsidence continues after the cooling of the lithosphere, and is associated with its thermoelastic shrinking and downwarping of blocks or microplates. D. P. MC KENZIE (l. c.) based his model on geophysical data on the Aegean Sea with a hot mantle diapir beneath the sea. Since the primary active ascent of diapir should be associated with the arching of a dome several km high, then the origin of a basin could not be associated with a dome (upwelling) about which there is no geologic evidence in the pre-basinal crustal history. This is why Mc Kenzie came to the above conclusion about the stretching of the lithosphere and the secondary passive upwelling of the mantle diapir.

According to L. ROYDEN and J. G. SCLATER (1981) the rapidly increasing temperatures caused thermal expansion of the lithosphere and this is associated with the so-called initial subsidence. The second phase is characterized by a long cooling of lithosphere, and its extension phase is followed by thermal contraction and the so-called linear thermal subsidence. It may be correlated to the collapse of R. W. VAN BEMMELEN'S mantle diapir.

The problem of the model is in the presumable rapid overheating and consequent decreasing viscosity of the lithosphere. The ductile lithosphere should migrate and expand to the area of lower viscosity. There is no evidence of continuous subduction- or obduction zones under/over margins of developing basins. The model cannot explain the generation of basins in areas out of the increasing heat flow. The only explanation might be in accumulation of the lithospheric masses in the rim syncline around the mantle diapir where the roots of the orogenic lithosphere could be created. This idea is however, unacceptable, because marginal depressions of most basins above diapirs rest on unthickened crust. It is more logic to associate the generation of marginal depressions with the rim synclines.

According to E. V. ARTYUSHKOV and M. A. BEER (1983) the movement of the ductile lithosphere should either be associated with unfolded of rigid blocks in the upper lithosphere and with immediate generation of abyssal-marine basins, or with the generation of listric faults and antithetic dipping of faulted blocks (Fig. 2). Owing to this pattern, the subsidence compensated with sedimentation should be associated with angular unconformities (up to 270 — according to calculations) in sedimentary sequences. All these phenomena are absent in non-rift basins and so the authors reject D. P. MC KENZIE'S (l. c.) model of the generation of non-rift and especially of non-linear basins. They only admit this mechanism in rifts with evidenced listric faults.

At present the idea of the relation of basins to the stretching of the crust-lithosphere is widely spread. The double-layered rigid and ductile — rheologic model of the lithosphere is proved by geophysical data on the depth of fault disturbances, by experimental and petrostructural data. Some geologists, partly influenced by information from the superdeep borehole SG-3 on the Kola peninsula, place the lower border of the rigid crust to the depth of 8—12 km. With the increasing depth viscosity should decrease and rifts may disappear.

According to S. N. IVANOV'S (1984) calculations, viscosity of rocks in the depth of 30 km is 10.000 times lower than in the rigid zone. The model of the stretching of the lithosphere is accepted by those who reject basification.

The concept of the stretching of the lithosphere is also supported by indications of layering and great lateral and vertical heterogeneity of the lithosphere, derived from geophysical measurements: the latest data are presented by A. L. YANSHIN and ZH. S. ERZHANOV (1984), C. FROIDEVAUX and L. FLEITOUT (1984). B. V. ERMAKOV et al. (1984) assume that layering results from accretion of the continental lithosphere material differentiation in the course of the development of the lithosphere, and some even regard it is a result of dynamics of the lithosphere. According to S. V. RUZHENTSEV and S. D. SOKOLOV (1984) the tectonic layering is a result of gliding movements among "microplates", inside the lithosphere, of movements controlling stretching and compression of lithosphere. Nappes and other types of detachment of sedimentary sequences from the basement, the level structure of ophiolite suites (A. V. PEÌVE et al. 1984), subhorizontal discontinuites associated with tectonic breccias revealed by drilling (e. g. borehole SG-3 in the Kola Peninsula at the depth of 8.700—8.800 m and 9.900—10.000 m; E. A. KOZLOVSKY et al. 1984 — Fig. 1) are geologic evidences of subhorizontal and/or horizontal layering in the lithosphere.

Geologists and geophysicists prefer the model of stretching of the lithosphere owing to measured increasing heat flow. Some presume the primary stretching (e. g. due to expansion of the Earth — N. P. IVANKIN 1984) was caused by the uprise of the hot mantle, indicated geophysically.

From the mechanical view there are two types of stretching: thermoplastic and mechanical. The second type is associated with — for instance — the origin of back arc basins or marginal basins on the convergent continental margins owing to the extension and downwarping of the overriding plate on the subduction zone (the type of convergent-extensional active margin after J. AUBOIN et al. 1984). Admittance of extension (stretching) in the zone of convergence and active overthrust of the continental plate (A. YA. SHARASKIN 1984) causes changes in opinions about dynamics of plates in the subduction zone. Some modern opinions are getting in accordance with the original Benioff's concept.

The extension of the back-arc and inter-arc basins i. e. a process that cannot be associated with the mechanism of convergent plate margins was revealed during deep investigation of back-arc basins and deep-sea trenches (J. AUBOIN et al. 1984). The results of the investigations proved the rift character of most depressions. It is difficult to explain it only from the plate tectonic aspects so far some geologists admit the activity of mantle diapirs — i. e. the phenomenon abandoned in convergence models and phenomena preferred in fixistic models. ZHANG YONGXIN (1984) and SUZUKI YASUMOTO (1984) associate the back-arc and marginal basins with the expansion of the lithosphere with partial mantle diapirs above the zone of convergence. V. D. TCHEKHOVITCH (1984) considers expansion and diapirism, and associates the origin of small basins with rapid centrifugal crustal stretching above the subduction zone. He assumes that the basins may form on both the convergent and divergent plate margins.

In my opinion, changes in crustal thickness result from crustal (lithospheric), physical-chemical and mechanical processes and are significant for the generation of basins. I am partly following those who defend the physical-chemical

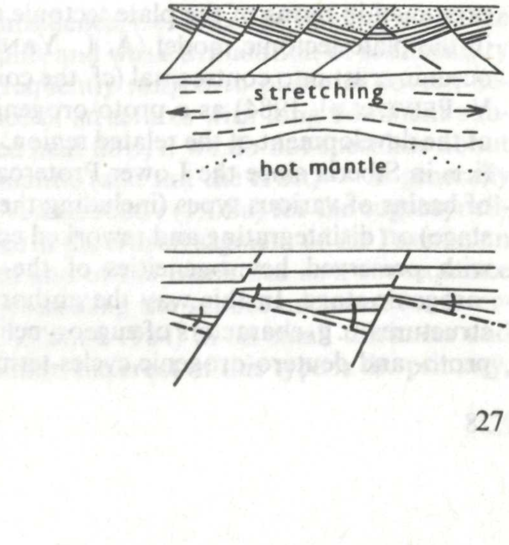






Fig. 2 The origin of the basin by the stretching of the lithosphere. Explanations: 1 — shallow-water sediments, 2 — deep-water sediments, 3 — angular unconformity.

character of endogenous processes and also phase explanation of geophysical discontinuities and indications of crustal thickness variability. N. I. PAVLENKOVA (1984) assumes that beneath basins with thinned crust the P-wave velocities increase and there is a general equilibrium in the upper 40—60 km thick layer of the lithosphere. Amplitude of Moho-elevations and volume of rock intrusions from the mantle into the lower part of the crust are proportional to thickness of sedimentary filling and to the grade of their diagenesis. In "mature" basins (e. g. the Donec—Dniepr Basin) with metamorphosed sediments the Moho elevation disappears in geophysical records. It is a basis for assumption that the deep seismic boundaries are not mechanical, so they did not result from pure tectónic discontinuities (cf. S. V. RUZHENTSEV and S. D. SOKOLOV 1984) but more-or-less from physical-chemical rock changes in the crust. It may be in accordance with the opinions that some rapidly subsiding basins (e. g. the South-Caspian depression) have yet not reached the level and responding phase changes in crust and so the Moho beneath them is recorded as a subhorizontal zone without elevations.

The problem of stretching of the crust — lithosphere is unsolved. According to I. B. RAMBERG and P. MORGAN (1984), the MC KENZIE'S simple model often used in speculations concerning the extent of the stretching and subsidence rate cannot be applied when the crust structure and composition were anomalous before the stretching. Simatic crust or deep faults may be included.

### 1.4.3 Primary changes in crustal thickness

Dynamic models are based on presumable permanent changes in the development and types of the crust. The oceanic crust is regarded as a new-formed crust in the place of the opening of the continental crust. Only shields are regarded as rather conservative element. Platforms may be mobilized (H. STILLE 1924) and changed into the basement of geosynclines and orogens (e. g. V. E. KHAIN 1984).

Existence of the continental crust prior to the generation of basins by riftogenesis and by the creation of new oceanic crust among divergent plates is presumed in terms of the plate tectonic model. Those who propose corrections in the plate tectonic model (A. L. YANSHIN et al. 1984) regard the change of oceanic crust into continental (cf. the concept of A. V. PEIVE et al. 1972, and A. V. PEIVE et al. 1984) as a proto-orogenic cycle existing only in the early stage of the development of the related region. The consequent deutero-orogenic cycle (i. e. in Siberia since the Lower Proterozoic) is characterized by the generation of basins of various types (including the development from rift to geosynclinal stage) on disintegrating and reworked continental (formerly consolidated) crust with preserved heterogeneities of the structure pertaining since the proto-orogenic stage. In this way the authors also explain heterogeneity of crustal structures, e. g. characters of eugeosynclines in intracontinental basins. Both the proto- and deutero-orogenic cycles terminate with molasses formed in various stages of crustal development and metamorphism. So according to stages of crustal development, intermontane basins of two types may form. Both cycles may exist next to each other, for example geologic units with proto-orogenic molasses side by side with deutero-orogenic volcanic sedimentary zones and basins. This model rejects the concept of gradual progressive accretion of the continental crust (e. g. A. V. PEIVE et al. 1984, M. M. LEBEDEV et al. 1984), and if it is admitted, then the accretion is very slight in comparison to the process of disintegration of the sialic crust into blocks and megablocks. In the course of further development the blocks are affected by spreading and convergence, and this dynamics changes the structural pattern of the respective continents.

In no one of the models the primary heterogeneity of crustal types, the existence of thick sialic crust and thinner restite simatic crust resulting from the generation of continental crust in one evolutionary cycle are considered. Geologists often doubt the existence of remnants of the crust of oceanic character inside the continental crust.

There, however, are evidences about the existence of heavy mafic crust in continents. The crust is very frequent and forms the basement of basins (the so-called deprogens; N. YA. KUNIN 1984). Basing on the occurrence of ophiolite arcs, S. S. KARAPETOV (1984) presumes the existence of intracontinental basins with oceanic basement but he associates the presence of oceanic crust mechanically with spreading. Basite crust is regarded as a product of the original spreading or unaccomplished oceanization which also can be a result of basification. In this case the crust is mostly regarded as the secondary, altered platform crust.

In plate models the oceanic crust enclosed in continental plates is explained in three ways:

a) As a rest of non-subducted crust at the closing of the ocean ( J. F. DEWEY et al. 1973, A. TOLLMANN 1978, a. o.);

b) by the trapping of oceanic crust on the margin of a continental plate due to subduction or collision;

c) by obduction, mostly on fossil collision margins of a continent (e. g. A. V. PEIVE et al. 1972).

Theoretically only in the first two cases a marine basin or a trough with deep-sea sedimentation may form.

The existence of basins with long subsidence, i. e. since the Paleozoic or Late Proterozoic, with shorter phases of uplift and weak denudation of sedimentary complexes without metamorphism (frequently ranged to platform cover), devoid of granites, indicates the existence of structures with long dominant subsidence trend, without recent increased heat flow, if we do not speculate about the high heat flow only affecting the mantle (and not the crust) — as generally assumed by MC KENZIE (1978) or V. V. BELOUSOV (1978a) for the eugeosynclinal regime. I regard the long subsidence in the crustal segment of the Pannonian Basin, the so-called Moesian Platform and of the Black Sea as a consequence of a primarily different crustal type, resembling the suboceanic crust (or intermediary type of crust F. ČECH — J. ZEMAN 1984) in its small thickness and often reduced granite layer. In my opinion the crust of this type is the primary, slightly sialized block or crustal belt tending to greater mobility and repeated subsidence. Mobile units formed owing to the existence of simatic, perhaps originally Precambrian oceanic crust with decising influence upon the creation of simatic segments of mobile platforms, i.e. the simatic internal massifs (F. ČECH — J. ZEMAN 1982) or zones inside the continents and on their margins. This problem will also be discussed in Ch. 7.

## 1.5 Disequilibrium in asthenosphere

In the present models of the generation of basins the sources of dynamics are placed in the mantle, and mainly in the asthenosphere whose depth is estimated on the basis of the records of decreasing velocities of P- and S-waves. So far the nature of low velocities is unknown and interpretations are different (F. BIRCH 1969, D. H. GREEN and A. E. RINGWOOD 1969, V. V. BELOUSOV 1973, a. o.). For example, R. J. O'CONNELL and J. H. JACKSON (in T. H. JORDAN and W. S. FYFE 1976) do not regard the seismically indicated asthenosphere as identic with the zone of partial melting, and the amount of the melt is small according to them. The opinion that it is a consequence of decreasing viscosity or partial melting of mantle rocks in several depth levels, due to stepped gravity differentiation (P. M. SYCHEV 1984) is, however, dominant. This interpretation is supported by elevations of the asthenosphere beneath volcanic-active basins with a high heat flow (e. g. the Tyrrhenian Sea). Considerations of the possibility that subsidence of the colder and consistent lithosphere is enabled by the astenosphere or by the anomalous (hot) mantle with the P-waves velocities by even 10 percent lower than in the normal mantle are less frequent. The asthenosphere has variable thickness and depth, and perhaps a different character beneath shields, orogens and basins. For example the low-velocity layer underlying the Basin and Range province is approximately 20—40 km thick (E. V. ARTYUSHKOV 1981).

The model concepts of the role of the asthenosphere are based on mostly unsystematic geophysical data on the existence of low velocity elevation zones at the base of the crust in oceanic rifts (P. R. VOGT et al. 1969 a. o.). There is an opinion about the mixture of the crustal and mantle rocks, or about uprises of mantle diapirs with high heat flows (J. H. ILLIES 1970). Density- and volume changes cause the uprise of granitoid diapirs, mainly tonalites, if lighter masses result from deep differentiation of basic rocks.

It is presumed (e. g. R. W. VAN BEMMELEN 1972, H. RAMBERG 1967) that heavier masses from the sides of overlier of unstable masses migrated into areas vacant after the diapiric transport of lighter masses. Volume changes are also associated with phase changes, hydration or dehydration of rocks.

Seismic investigations revealed layered structure, indicated by variable seismic waves velocities (for details see N. A. BELJAJEVSKIJ 1974), both in the upper mantle and in the asthenosphere, as well as megablock structure of the mantle, with horizontal and vertical low velocity channels (G. N. BUGAEVSKIJ 1977). Subduction is also presumed to cause changes in the asthenosphere (A. G.

FISHER 1978 a. o.). Melted basalt and sialic masses migrate back to a higher level in the subducted lithosphere. Such magmas are presumed to pass not only to the young orogenic zone but also beneath adjacent stable platforms. The resulting heterogeneities in the asthenosphere change dynamics of the lithosphere which is then divided into elevations and depressions due to the differentiated ductility of the lithosphere. A slight revival of tectonic movements on platforms, resulting from the activated asthenosphere is also presumed. In this way the generation of inner and marginal platform depressions could be explained, as well as partial elevations and depressions in basins, or the platform regeneration as a result of the activity of the adjacent orogenic zone; for example the West Carpathians and the epivariscan platform in the Cretaceous up to the Tertiary.

### 1.5.1. Mantle diapirs

In the areas of the increased heat flow (± volcanism) the mantle elevations beneath thin crust are associated with the asthenosphere elevations. It is repeatedly proved by complex geophysical researches. The uprise of the mantle facilitates — owing to decreased geostatic pressure — the melting of rocks and the generation of magma chambers in higher crustal levels (N. A. FLORENSOV 1977). Problems, concerning the mechanism of the uprise of hot mantle diapirs, are solved by various models (J. NIKOLSKIJ 1982): convection, advection, radioactive decay, differentiation of the mantle rocks into zones or segments of different densities, physical-chemical changes of rocks, expansion or pulsation of the Earth (e. g. E. E. MILANOVSKY 1977).

#### 1.5.1.1 The origin of diapirs

According to J. NIKOLSKY (1982) hot diapirs form owing to the heating in the depth of 800—900 km near the contact between the upper and lower mantles. Nikolsky basing on seismologic records, assumes that at this depth is a 200 —300 km thick layer of rocks partly melted by the deep heat. The layer condensates thermal energy released by melting. In the condensator layer the convection causes increasing thermal conductivity of the overlying mantle in the places of advection. The accumulating thermal energy supports further melting of rocks in smaller depths. Ascending hot mass in the form of vertical lens with density lower than that of the surrounding mantle causes the density inversion and disequilibrium in the mantle. The uprise of the vertical hot lens to colder shallower levels is associated with crystallization. The crystallization rate increases and slows or stops the ascent of the hot melt lens. In a shallower zone with a lower pressure the heat from the stabilized lens may give rise to the secondary melting focuses, and these may be a cause of the generation of new partial mantle diapirs. J. NIKOLSKY (l. c.) calculates periodicity of a deep cycle to 140—500 Ma on the basis of the development of a Mediterranean belt, presumably founded as early as the Late Proterozoic. In this deep diapir belt the mantle diapirs got nearest to the surface and caused crust oceanization. The existence of shallower diapirs is proved by prominent positive gravity anomalies. Nikolsky assumes that hot lenses form in deep mantle along the continuation of the diapir belt to Central Asia. Lower density of the lenses causes lower gravity effects and the lenses should be indicated by negative gravity anomalies. Lenses in the depth cause enlarging volume and so the Central Asiatic segment of the diapir zone (extending to the Pacific Ocean) is orogenically uplifted. Uplifted depressions should prove the existence of a shallow mantle diapirism more than 140 Ma ago. A new deep cycle appeared beneath Asiatic high mountains. The author of this model predicts a new epoch of basification and oceanization in the Asiatic orogenic belt in the geologic future.

The generation of the mantle diapir should cause crustal elevation and tensions (V. V. BELOUSOV 1969, R. W. VAN BEMMELEN 1972). In the asthenosphere the lower-viscous masses may flow from the top down the flanks (V. V. BELOUSOV 1982). Owing to tensions in the lithosphere, shear- and listric faults form. Blocks of the lithosphere sag down and glide along the faults. The mechanism is similar to that of P. D. MC KENZIE'S (1978) model. This hypothesis may explain the origin of basins in the case when the asthenosphere is elevated in zones without evidence of fossil mantle elevation and without crustal elevation prior to the generation of a basin. The model of the lithosphere gliding over the elevations of the asthenosphere may be criticized in the same way as the model of the spreading of the lithosphere (Ch. 4.2.).

### 1.5.1.2 Does the global diapirism exist?

J. NIKOLSKY'S (1982) model is interesting although highly hypothetic. The idea of diapirism as the main dynamic phenomenon in the evolution of the crust (also see R. W. VAN BEMMELEN 1972, H. RAMBERG 1967) could be globally reasoned for the lunar pregeologic stage of evolution of the Earth. Also results of the remote sensing of the Earth (satellite images), the existence of circular and ring structures and megastructures on the Earth's surface could indicate the planenary extent of the mantle diapirism. Although the circular morphostructures of planets are still mostly explained by meteorite impacts. Their volcanic origin can however, not be denied. Lunar mares collapse stages of ascending deep diapirs would be in accordance with their tectonic control and link with linear structures.

The origin of magmatic rocks, mostly granitoids, and diapiric uprise of rocks with a lower density like light granites, orthogneisses a. o. may support the idea of a considerable frequency of diapirism in the crust. The present trend in geology indicates a return to the ideas of R. W. VAN BEMMELEN and his predecessors. For example G. B. UDINTSEV et al. (1984) regard diapirism as the main factor of the formation of oceanic rifts. They reject the divergent horizontal movements of plates.

### 1.5.1.3 Setting of diapirs

There is the same problem with the emplacement of mantle diapirs in the mantle and in the crust like with the emplacement of granitoids, gabbros, a. o. The idea of the fracturing and/or ductile stretching of the lithosphere under the influence of expansion of the Earth (S. W. CAREY 1976) is simplest. There is no problem in the mechanism, and data on the global riftogenesis and taphrogenesis support the expansion hypothesis.

The formation of a "free" space by the stretching of the ductile lithosphere is possible — unless there are serious arguments against the model (cf. Ch. 4.2 ).

F. HORVÁTH et al. (1981) tried to explain the area by a lateral splitting of the convergent plate margin and by the detachment of the upper rigid lithosphere from the lower ductile lithosphere. Hot mantle masses entered the resulting traps and formed diapirs. J. F. DEWEY'S et al. (1973), K. BIRKENMAJER'S (1976) and A. TOLLMANN'S (1978) models also consider the splitting of collision plate margins and possible penetration of diapirs into cracks and areas resulting from the break-off of blocks. V. V. BELOUSOV (1982) and E. V. ARTYUSHKOV (1979) consider the break-off of basified blocks of the lithosphere or of eclogitized crustal blocks in relation to the formation of an area for ascent of diapirs.

The emplacement problem must also be considered in concepts of the origin of diapirs in two convergently subducting plates (L. STEGENA et al. 1975), the stretching of the lithosphere to the bending and dipping subducting zone (L. H. ROYDEN et al. 1982), or in other variants of models concerning subduction, spreading or opening of the crust in rifts (J. M. LORT 1977, B. BIJU — DUVAL et al. 1974 a. o.).

Data on velocity rates of P-waves changes of seismic waves in the mantle beneath the Baikalian rift and its surroundings in southern Siberia (G. N. BUGAEVSKY 1977) show that low viscosity affected a large area of the mantle extending to hundreds or thousand km.

Hot spots are likely to form in zones of gradual vertical heating to melting point. They prefer orientation to fossil mantle elevations, below the thin primary simatic crust or to deep fault zones. A temperature increase by 200 °C is followed by volume expansion 0,04 percent and arching with uplifts in the mantle by 480 m, in the crust by 160 m (calculations made by K. D. SCHUILING 1973).

Arcogenesis of mantle and crust causes decreasing tension in lithosphere. It is very likely that the diapir uprise is accompanied by the ascent of a high-thermal front, causing heating and partly melting of the rocks. The front forms a path for the moving magmatic melts. Their migration from the surroundings of diapirs proceeds most probably in the asthenosphere whose thickness is controlled by the volume of the low-viscous mantle. In the place of decreasing mass volume the lithosphere sinks into a circumdiapir rim syncline. The extreme heat and the ascending melts can sooner and easier invade the base of the crust in places with primary thin simatic crust underlain by a fossil mantle elevation. In this elevation also vertical and partly horizontal expansions of diapirs are easier. Lateral expansion may in a small scale be controlled by divergent or centrifugal mantle currents under the lithosphere causing tension in the lithosphere above the diapir of the approximate calculated value 10—$10^2$ MPa (LIM YUAN-LONG et al. 1984).

In the dome of the lithosphere, magmatic rocks (differentiated and non--differentiated) move laterally by means of the mechanism of hydrorupture (P. M. SYCHEV 1984), exploiting heterogeneity and layering of the lithosphere. So it is unnecessary to presume the existence of areas open prior to the penetration of the hot mantle masses.

The mantle expansion causes vertical and horizontal shear and compression tensions in the crust. Decompression in the upper lithosphere is gradually manifesting in the deep lithosphere or beneath the asthenosphere. There decompression may occur at the melting of rocks and after the flow-away of melted rocks. So deep earthquake epicentres form a semiarcuate zone in colder rigid mantle close to the contact with active hot diapirs (Tyrrhenian Sea, Aegean Sea; V. V. BELOUSOV 1982, F. ČECH — J. ZEMAN 1984). The arcuate epicentres are regarded as partial Benioff's subduction zones (e. g. M. BOCCALETTI et al. 1976).

The effects of decompression in the crust and in the upper mantle upon the generation of magma are at present emphasized by A. A. KADIK and M. JA. FRENKEL (1982). According to them not only magma but also fluids form in the zones of decompression, e. g. on deep faults and in asthenoliths. Solutions then penetrate into rocks, cause decrease in temperature necessary for melting, and support, for instance the generation of andesites in smaller depths.

### 1.5.2 Fault pattern above mantle diapirs

The active mantle diapir provokes a tension field in the overlying crust. Shear tensions dominate above the diapir margin. M. V. GZOVSKY (1960) presented the model of their origin. It is proved that shear tensions occur above the margin of the ascending body (in models of the piston) whereas above the top are decompressed and tensional zones with blocks tending to sag down. Modelling on a photoelasticimeter showed spatial tension differences and facilitated the construction of tension trajectories. Maximal tangential tensions (associated with shear deformations) are diagonal to the margins of the ascending body, and incline towards the body. Maximal normal tensions, causing tension deformations, occur above the top of the elevation and above adjacent "depressions" — non-ascending parts of the compressed mass (Fig. 3). The tension pattern is the same above the diapir.

Faults on margins of basins genetically related to diapirs, prove concentrations of shear tensions above the margin of the mantle diapir.

Explorations of hydrocarbon deposits in the Red Sea rift structure revealed the development and structure of a basin forming on basified continental crust owing to the diapir of asthenosphere (J. D. LOWELL et al. 1978). The elevation of the crust was associated with bending. The length of elliptic elevation in the area of the Red Sea was 2000 km, its width was 500 km. Although it is a rift, data on the course of faults are also valuable for depressions forming on the elevation of mantle — asthenosphere.

The data on the history of the Red Sea show that the area started developing as a shallow depression by "downsagging" of the crust. Drilling revealed alternation of lavas with sediments, i. e. basaltization with the total thickness of basalts 3 km. No oceanic tholeiites were found in this part, because the rift formed on the continental crust.

So the initial stage corresponds to the generation of lakes and of the oceanic basin in depressions bordered with faults. The initial volcanism had a character of continental basalts.

During the accretion of the mantle diapir the width of the rift structure increased with the increasing frequency of faults. The faults are antithetic to the elevation limbs and their density is the greatest above the limbs — the elevation slopes (Fig. 4). Some geologists were inspired by the dip of faults to explain the fact by subduction zones compensating the stretching of the lithosphere. The orientation of faults does not support this model.

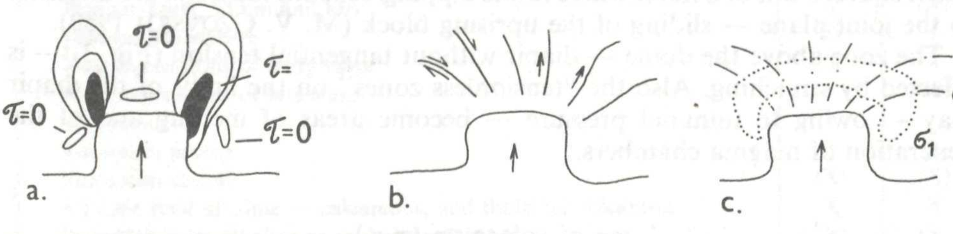



Fig. 3 Distribution of stress above the diapir body — modelled after M. V. Gzovsky (1960). a) — origin of tangential stress (τ), b) trajectory scheme of tangential stress c) trajectory scheme of normal stress (δ).

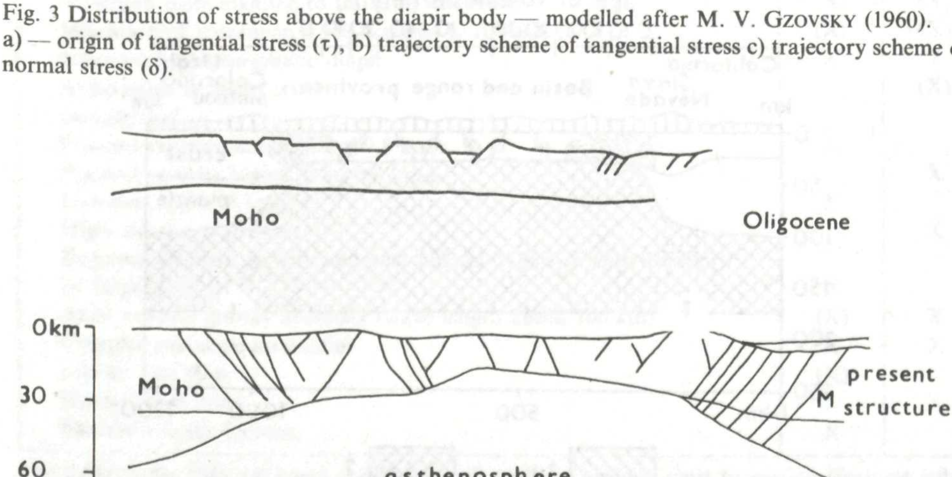



Fig. 4 Evolution of faults in the Red Sea rift (simplified according to J. D. Lowell et al. 1978).

Seismoactive faults in the Great Basin show the same arrangement (CH. H. SCHOLZ et al. 1971) — Fig. 5. Two explored basins display a striking accord not only in the setting of marginal shear zones, but also in the internal segmentation into horsts and grabens — Figures 4, 5. There is a substantial accord in the distribution of the main faults, mainly deep reaching, also in the Pannonian Basin. Naturally, a fault pattern generally shows different fault strikes in the elliptical, circular and linear basins. In the last case, i. e. in the case of a rift, parallel faults prevail, whereas in oval basins the faults are oriented variably. Dominant is tangential orientation to the margins of mantle diapir.

There is a good correlation between models and large structures above geophysically indicated lithosphere diapirs. The models support the idea of the penetration of the mantle diapir through lithosphere, causing tension in the overlying crust. Orientation of maximal concentration of faults, and shear tensions of seismoactive faults above margins of the diapir indicate its participation in the fault structure of basins and in the activity or even in the generation of deep faults at the margin of the diapir. There is a striking accord between the dip of marginal faults (to the diapir) and the dip of trajectories in models. Faults, resulting from the plastic bent of a plate, show the same orientation. They were evidenced on folds whose core diapirically penetrates the layers, or at the squeeze-out of a horst where faults dipping beneath the horst are diagonal to the joint plane — sliding of the uprising block (M. V. GZOVSKIJ 1960).

The zone above the dome — diapir without tangential tension (Fig. 3) — is affected by stretching. Also the "tensionless zones" on the limbs of the diapir may — owing to minimal pressure — become areas of melting and of the generation of magma chambers.

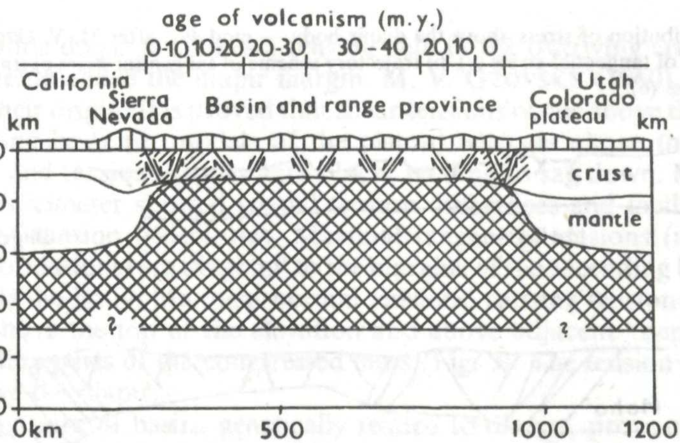



Fig. 5 An east-west section across central Nevada. (After CH. H. SCHOLZ et al. 1971, modified). Explanations: 1 — seismic zone, 2 — low-velocity, partially melted zone.

### 1.5.3 Continental rifts

Principal geological and geophysical characters of rifts are presented in Table 1. Terminology, concerning rifts is in Table 2. The tables show a great variability of continental rift types. Owing to new plate tectonics and data on oceanic floor obtained in the past twenty years, the genesis of continental rifts has become an interesting object of geological and geophysical investigations. With respect to endogenous activity significant are data on the relation of the Baikalian rift to

Main features of continental rifts

Table 1

| Features | Rifts present | Rifts failed |
|---|---|---|
| 1. Structural and geomorphological: | | |
| Width (generally) 35—60 km | X | X |
| Length (order) 1000 km | X | X |
| Existence of rift valley | X | (X) |
| Normal faults, basic veins — spreading feature | X | X |
| Frequent cases: asymmetric profile | X | X |
| inner horsts and grabens | X | X |
| Wide domes | (X) | ? |
| Thinned crust, 35 km and less | X | (X) |
| Thinned lithosphere | X | ? |
| Wide depressions of early stage | X | X |
| Wide depressions of late stage | | X |
| 2. Magmatic | | |
| Volcanism present | X | X |
| Volcanism absent | (X) | (X) |
| Alkaline (and alkaline — calcareous, and tholeiite) volcanism | X | X |
| Evolution from alkaline to tholeiite volcanism | (X) | (X) |
| Migration of volcanism from margins to rift axis | (X) | (X) |
| Subcrustal asthenospheric diapir | X | ? |
| Asthenolith in lower crust | | (X) |
| 3. Geophysical | | |
| P-waves velocity smaller than 7,8 km/s | X | |
| P-waves velocity higher than 8,0 km/s | | X |
| Low-rate zones in crust | X | |
| High-rate zones in crust | | X |
| Regional negative gravity anomaly (effects of low-density mantle) on large area | X | |
| Axial positive gravity anomaly (wave length about 100 km) | (X) | X |
| Complex magnetic anomalies | X | X |
| Intense heat flow | (X) | |
| Normal heat flow | | X |
| Shallow seismic focuses | X | |

The table shows that not every continental rift tends to oceanic crust formation (features sub 2). Modified and simplified after I. B. Ramberg — P. Morgan (1984) Symbols: X — typical phenomenon, (X) — occasional occurrence, ? — unknown.

Terminology of graben structures (modified after I. B. RAMBERG — P. MORGAN, 1984) Table 2

| Term | Characteristics |
|---|---|
| active riftogenesis | riftogenesis as a result of isotherms (asthenosphere) uprise |
| passive riftogenesis | riftogenesis controlled by regional stress field |
| present rift | rift with young tectonics and volcanism |
| failed rift — paleorift | rift active in geologic past, neotectonically nonactive (sleeping rift) |
| failed arm | triplet branch which does not open farther and does not change into ocean |
| aulacogen | paleorift on ancient platform, often compression-deformed |

geophysically indicated heterogeneity of the mantle up to the outer core (G. N. BUGAEVSKY 1977). The relation between the origin of the rift and the ascending asthenolith of hot differentiates from the mantle, or the mantle diapirism is generally accepted (N. A. FLORENSOV 1977, M. H. P. BOTT 1981, E. V. ARTJUSHKOV 1981, J. H. ILLIES 1981, a. o.).

Crust doming (antiforms, arching, upwelling, plateau uplifts) preceded rifting-according to most authors (J. H. ILLIES 1981, S. BHATTACHARJI 1984 a. o.). According to M. H. P. BOTT (1981) doming is associated with tension. About 200 MPa are required for the generation of troughs with 5 km thickness of sediments. I. B. RAMBERG and P. MORGAN (1984) doubt about the necessity of the doming — rift relation. According to the authors quoted, some others assume that rifting was preceded by downwarping of the crust. The problem is unsolved, since the crust elevation is not always preserved. Rifting is also controlled by pre-existing lineaments (J. H. ILLIES 1981, P. J. BUREK 1978, N. A. BOZHKO 1984, A. V. RAZVALIAEV 1984). The influence of disturbed lithosphere upon the upwarping of mantle masses intruding into the crust must also be considered in the study of the genesis of non-rift basins. At present the rift origin is also ascribed to island arc basins, marginal seas and geosynclines (A. V. PEIVE et al. 1984, V. E. KHAIN 1984 a. o.).

At rifting the lithosphere is overheated. So it may get stretched with formation of listric faults (MC KENZIE'S model). Another explanation of rift genesis is basification by asthenolith products, resulting in increasing lithosphere density. Heavier lithosphere (perhaps due to eclogitization of lower parts) may sink into the asthenosphere. Opening mechanism of continental rifts is a widely discussed, yet unsolved problem.

M. H. P. BOTH (1981) assumes that a rift forms only in the upper, 10—20 km thick brittle crust whereas the lower ductile crust is deformed by visco-elastic creep. This process (cf. MC KENZIE'S model) may result in a tension, and it is unnecessary to presume any extensive crustal doming. According to M. E. ARTEMYEV and E. V. ARTYUSHKOV'S (1971) calculations (in P. A. ZIEGLER 1982), crustal extension only represents 200 m at 200 km width and 3—4 km height of the dome. In BOTT'S model of tensions also the formation of dyke intrusions is considered if magma is present. According to this model, rifts formed above hot mantle diapir (Fig. 6).

E. V. ARTYUSHKOV (1981) presumes overheating of the lithosphere by heat flow from the asthenosphere, followed by decreasing viscosity of the lower crust and decreasing density of the upper crust as well as by its high uparching (min. 2 km). A sideward shift of crustal blocks may result in tensions (200—300 MPa) required for the formation of a rift, for example the Baikalian rift.

Fig. 7 shows crustal elevation due to the heating of the mantle at the depth 100 km, and to the formation of asthenolith (diapir). Active rifts show indica-

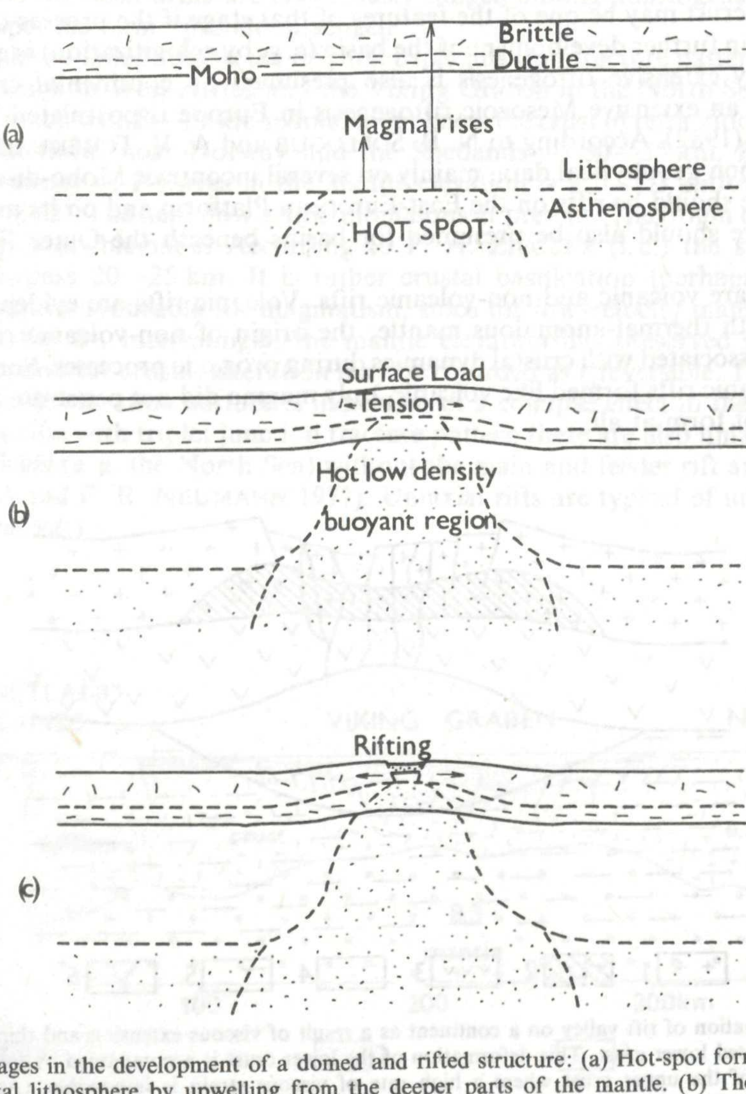



Fig. 6 Stages in the development of a domed and rifted structure: (a) Hot-spot forms below the continental lithosphere by upwelling from the deeper parts of the mantle. (b) The continental lithosphere becomes heated and thinned, with consequent isostatic uplift and development of tensile stress system in the upper crust. (c) Graben formation starts when the tensile stresses become sufficiently large. (After M. H. P. BOTT 1981, modified).

tions of dyke intrusions and basic effusions regarded as commencement of the generation of oceanic crust. This process is illustrated by the example of the Red Sea and California Bay, and basalt magma is regarded as a dynamic factor in divergent plate movements (L. E. LEVIN 1984). Dead rifts lack indications of basification affecting the entire continental crust.

The rift stage is inferred or evidenced as the initial phase of the development of basins which lost their rift characters in their further development. Basins in their early stage may represent an undeveloped ocean (J. R. CURRAY 1978). Thinned crust may be one of the features of that stage if the process of crustal thinning in further development of the basin (e. g. by eclogitization) is excluded. Extremely extensive riftogenesis is also presumed on continental crust. For example, an extensive Mesozoic riftogenesis in Europe is postulated by P. A. ZIEGLER (1982). According to N. B. SOLLOGUB and A. V. TCHEKUNOV (1983) basing upon geophysical data, mainly on several incontrast Moho-discontinuities, there should be rifts on the East-European Platform and on its margin. A rift nature should also be presumed for basins beneath the Outer East Carpathians.

There are volcanic and non-volcanic rifts. Volcanic rifts are evidently associated with thermal-anomalous mantle; the origin of non-volcanic rifts (grabens) is associated with crustal dynamics during orogenic processes. Some failed non-volcanic rifts formed like volcanic, only magma did not penetrate the crust or did not form at all.

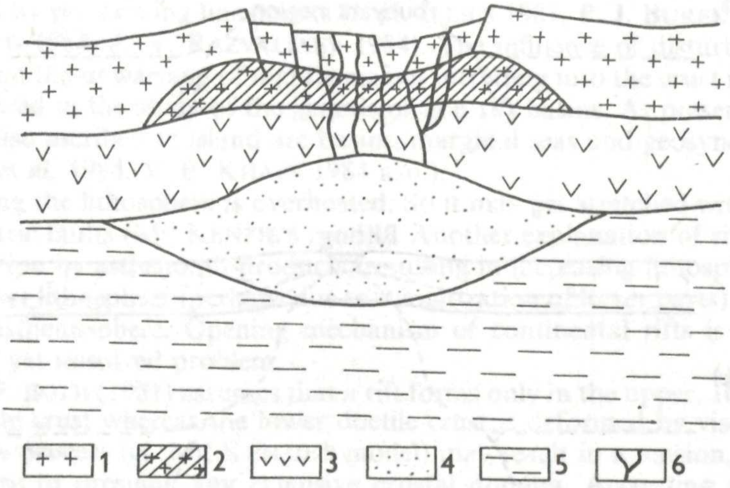



Fig. 7 Formation of rift valley on a continent as a result of viscous extension and thinning of a strongly heated lower crust. This deformation of the lower crust is accompanied by faulting and subsidence of the upper crust where a high rate of viscous strain is impossible. (After E. V. ARTYUSHKOV 1981, modified). Explanations: 1 — part of granitic layer with brittle deformations, 2 — part of granitic layer with viscous deformations, 3 — basaltic layer, 4 — low-velocity mantle, 5 — normal mantle, 6 — faults in the crust along which magma can ascend.

Rifts frequently display a triple junction fracture pattern which is the proof of the formation of a thermal-controlled dome. K. C. A. BURKE and J. F. DEWEY (1973) assume opening of all the three arms, and divergent movements of three plates, or one arm oriented towards the continent, and two opening arms in which the new oceanic crust formed. The continental arm represents a transport way of sediments accumulating in a delta on the continental slope. This arm was not open and turned into a failed rift arm. According to D. M. CURTIS (1980), such arms are erroneously ranged among aulacogens. CURTIS only accepts the term "pseudoaulacogen".

Examples of failed rifts in the original triple junction fracture pattern are the Benue Trough in West Africa and the Viking Graben in the North Sea. In the Viking Graben with 8—10 km sediment thickness the crust in refraction sections was 20 km thick, near Norway and the Shetlands — 30—35 km. Recorded normal velocity of P-waves in the mantle elevation is 8.1—8.3 km/sec. (P. A. ZIEGLER 1982). The heat flow is not anomalous at present. The origin of crustal stretching is problematic. According to P. A. ZIEGLER (l. c.) the stretching cannot surpass 20—25 km. It is rather crustal basification (perhaps by eclogitization), infavourable for magmatism, since the low-velocity mantle is not recorded (Fig. 8). Interestingly, the mantle elevation has preserved since the Mesozoic, and the crustal alteration may be regarded as irreversible. The shape of the mantle elevation surface is indicative of a collapse effect in the diapir.

Besides rifts with triplet junction fracture pattern there are also uniaxial rifts and rift fields (e. g. the North Sea) without the main and feeder rift axes (I. B. RAMBERG and E. R. NEUMANN 1977). Uniaxial rifts are typical of unopening continental rifts.

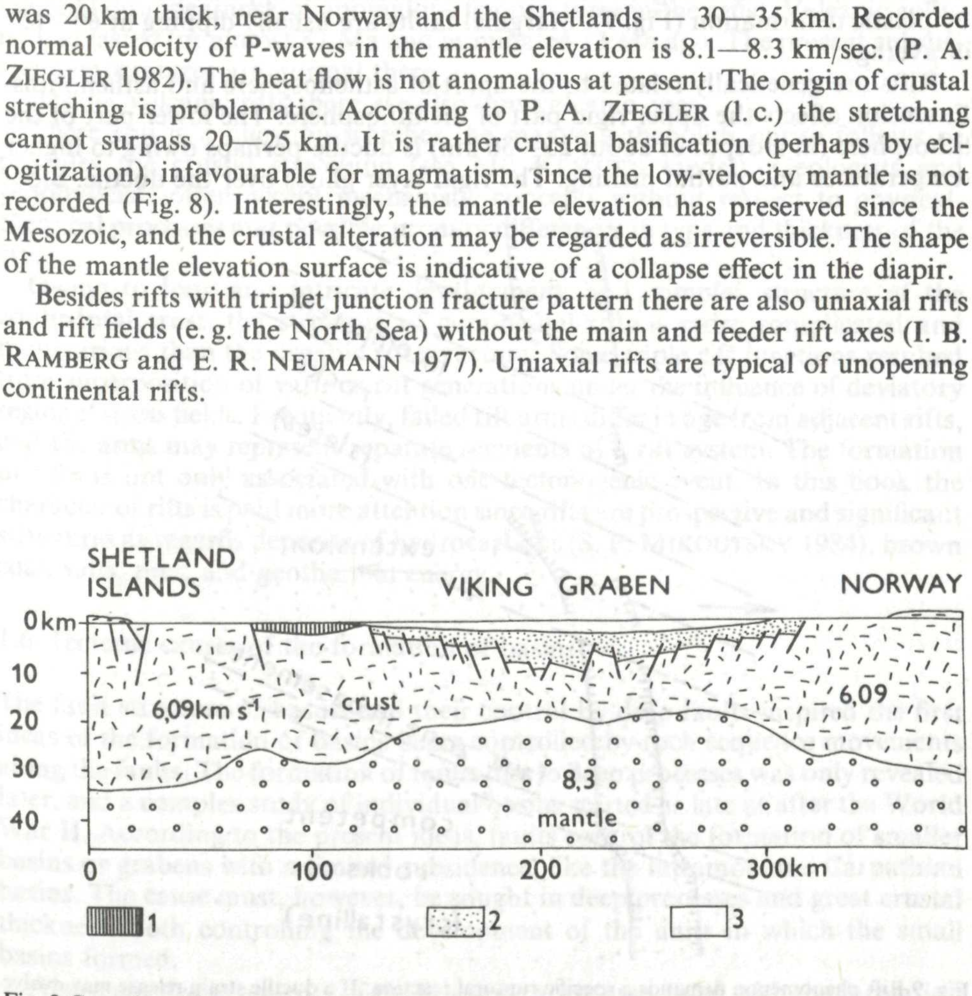



Fig. 8 Crustal profile across Viking Graben. (After P. A. ZIEGLER 1982, modified). Explanations: 1 — pre-rift sediments, 2 — rift fill, 3 — postrift sediments.

Most information on rifts resulted from extensive prospection for oil-, gas- and ore deposits. Basing on plentiful geological and geophysical data, J. H. ILLIES (1981) presented both the general and specific characteristics of rifts. Rifts are linear trenches (mostly with volcanic activity) controlled by and associated with mantle — asthenospheric processes. Rifts form by lateral extension normal to the rift axis, or by extensional shear governing the rift propagation. Rifts form inside continents (intraplate rifts), in forelands of orogens or on the contact between plates.

Most rifts follow pre-existing, often Precambrian deep faults (lineaments). Transversal deep faults control orientation and strike deviations of rifts. Complexes of incompetent rocks (e. g. schist massifs) represent the place of rift virgation or fault splitting — i. e. brittle deformation disappears and is replaced by plastic deformation (Fig. 9). Marginal faults are usually dipping at 60—65° in average.

Rifts are genetically related to the uplift of asthenosphere and asthenoliths. Extension affects the upper rigid part of the lithosphere. The lower part of the lithosphere (at a depth of about 25—30 km) is ductile-perhaps owing to higher temperatures and partial melting. The rigid layer glides over the ductile. Side-

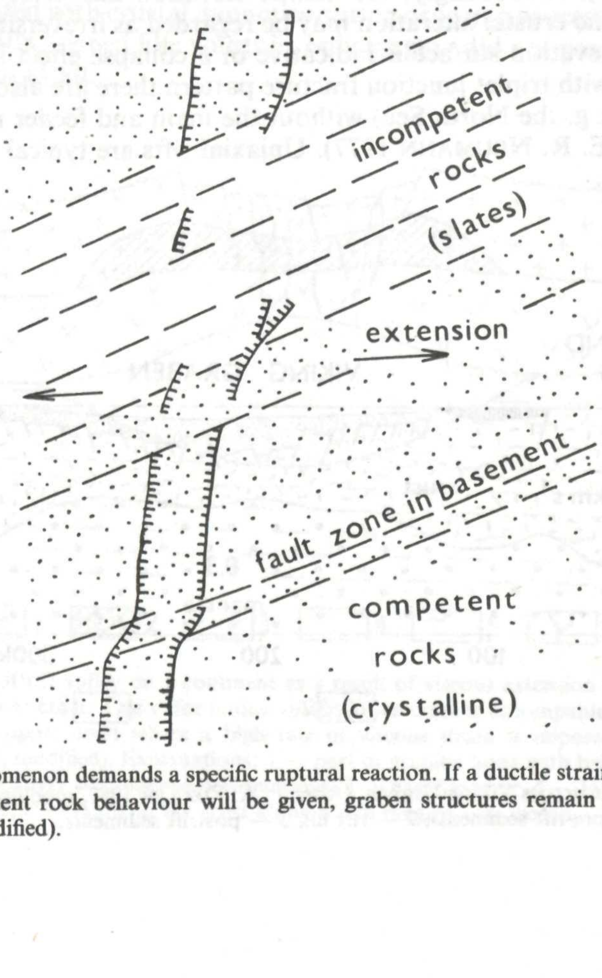



**Fig. 9 Rift phenomenon demands a specific ruptural reaction. If a ductile strain release may evolve i. e. an incompetent rock behaviour will be given, graben structures remain absent. (After J. H. ILLIES 1981, modified).**

ward movements are extremely slow, most likely governed by elastic properties of the lithosphere out of the rift. Opening of the upper lithosphere correlates with the thickness of basalt dykes, channels and subvolcanic bodies (cf. also calculations by I. I. ARTEMYEV and E. V. ARTYUSHKOV 1971). The sideward sliding of the lithosphere is facilitated by mantle elevation. If the sliding blocks are stopped by — for example — the adjacent more rigid tectonic unit, the uprising asthenolith starts uplifting the sunken blocks in the rift and a regional uplift follows.

The formation of a rift is not only associated with but also preceded by alcalic volcanism (e. g. 50 Ma prior to deposition of oldest sediments of the Rhinegraben rift). Volcanism is associated with a high heat flow. The Kaiserstuhl volcano in the southern part of the Rhine rift (Rhinegraben) indicates a possible time of regional disappearance of anomalous temperatures in the crust. Volcanic activity had lasted for almost 87 Ma and terminated 13 Ma ago. The present subsurface temperatures are normal there.

Besides volcanic rifts there are also non-volcanic ones.

There still is a question whether the mantle asthenolith uprise follows or precedes the crustal stretching (see MC KENZIE'S model). Geologists and geophysicists often accept mechanistic concepts without respect to physical-chemical processes and possible primary differences in type and thickness of the crust.

Owing to long and intricate development, and complex structure of the continental crust, the structure of continental rifts is more complicated and multivarious than the oceanic rift structure. Some triple rift junctions resulted from superposition of various rift generations under the influence of deviatory regional stress fields. Frequently, failed rift arms differ in age from adjacent rifts, and the arms may represent separate segments of a rift system. The formation of rifts is not only associated with one tectonogenic event. In this book the character of rifts is paid more attention since rifts are prospective and significant structures as regards deposits of hydrocarbons (S. P. MIKOUTSKY 1984), brown coal, salts, ores, and geothermal energy.

## 1.6 Tectonic causes of the formation of basins

The fault structure of basins and their control by deep faults inspired the first ideas of the formation of basins being controlled by rock sequence movements along the faults. The formation of faults due to deep processes was only revealed later, and a complex study of individual basins started as late as after the World War II. According to the present ideas, faults control the formation of smaller basins or grabens with a limited subsidence, like the intramontane Carpathian basins. The cause must, however, be sought in deep processes and great crustal thickness, both controlling the development of the units in which the small basins formed.

Also tectonic movements of the lithosphere are associated with the formation of basins.

### 1.6.1 Subduction and the origin of basins

Although the subduction zone is a compression belt, some geologists explain the formation of basins by this dynamics. I shall only discuss Carpathian basins. L. ROYDEN et al. (1982) explained the formation of the Pannonian Basin by westward subduction beneath the East Carpathians (Fig. 10). The stretching of the lithosphere might be caused by gradual bending of the plate, at enlarging angle between the Earth's surface and the subducing plate surface. This mechanism cannot be applied on the Transylvanian Basin, so a curious process is presumed: the bending is associated with a kind of funnel-like "suction" of the lithosphere, resulting in its downwarping. L. STEGENA et al. (1975) (Fig. 11) explained the formation of the mantle diapir and of the Pannonian Basin by subduction of two opposite submerging plates. R. JIŘÍČEK (1981) presumed the formation of the Vienna Basin by subduction of the Bohemian Massif plate and of an autonomous block beneath the West-Carpathian plate.

Because of the lack of geological and geophysical evidence of such subductions they cannot be associated with the origin of basins (F. ČECH 1984, 1985).

### 1.6.2 Strike-slip faults and the origin of basins

The formation of basins by strike-slip faults was explained by A. R. GREEN (1977). He denoted them as transverse shear basins (Fig. 12). Formerly they were defined for transcurrent and transform faults on the oceanic crusts and on continental margins.

The development of a basin of this type commences with strike-slip faults and basalt extrusions. The mantle rocks intrude into the broken thinning crust. Horsts and grabens formed beneath the Earth's surface. Further strike-slip faults might cause deepening of basins and increasing metamorphism of sediments. The formation of oceanic crust with an enlarging volume beneath the basin is followed by the spreading of the oceanic basin. Melting of the crust on peripheries of transverse shear fault basins results in rhyolite volcanism. Mantle elevation and basification must be considered in this model as well.

J. C. CROWELL (1974) explained the origin of partial basins along the San Andreas Fault by strike-slip movements. He described the basins resulting from strike-slip fault movements along discontinuous fault-plane as "pull-apart basins" (Fig. 13). The basins are associated with the Atlantic type of divergent plate margins, and form by the plate detachment along the future primary rift. The basins remain on the divergent continental margin as parts of the young passive margin (D. M. CURTIS 1980). J. C. CROWELL (l. c.) in his model also presented further variants of the formation of basins, for example on arcuate or virgating faults (Fig. 14).

Basins of the pull-apart type generally form during strike-slip movements along faults also out of divergent zones, on fault borders of blocks, microplates and plates — as far as the borders are transform or transcurrent faults.

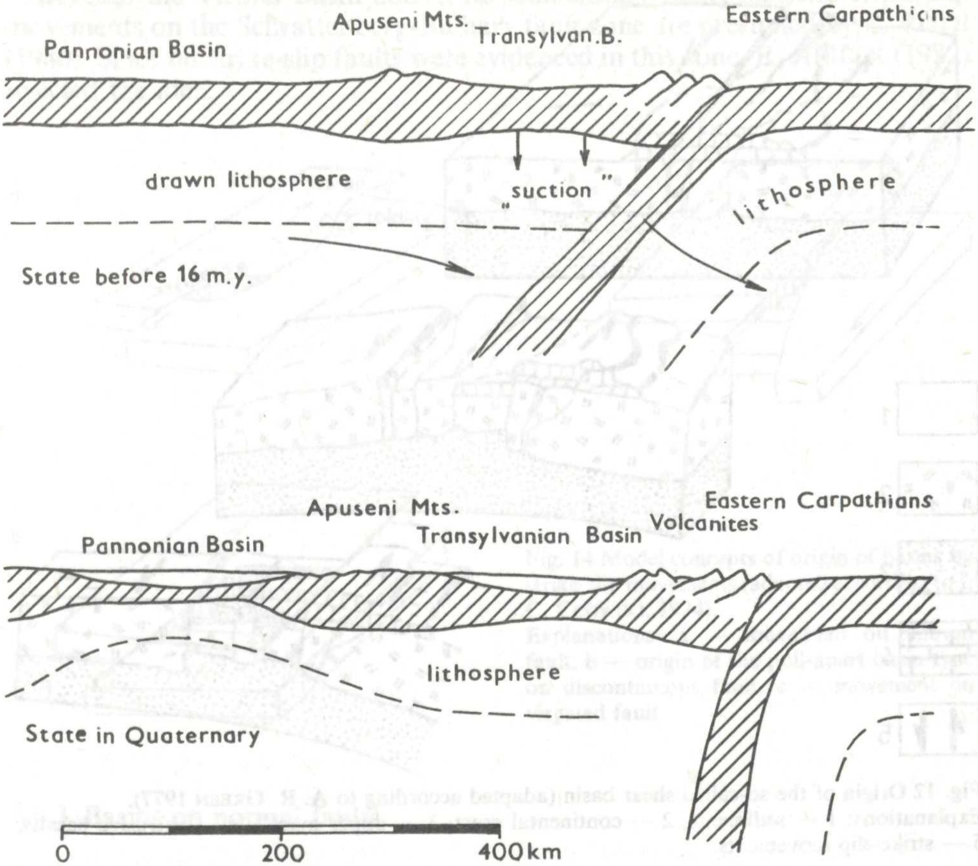



Fig. 10 Model of formation of Inner-Carpathian basins by subduction of lithospheric plates (according to L. ROYDEN et al. 1982).

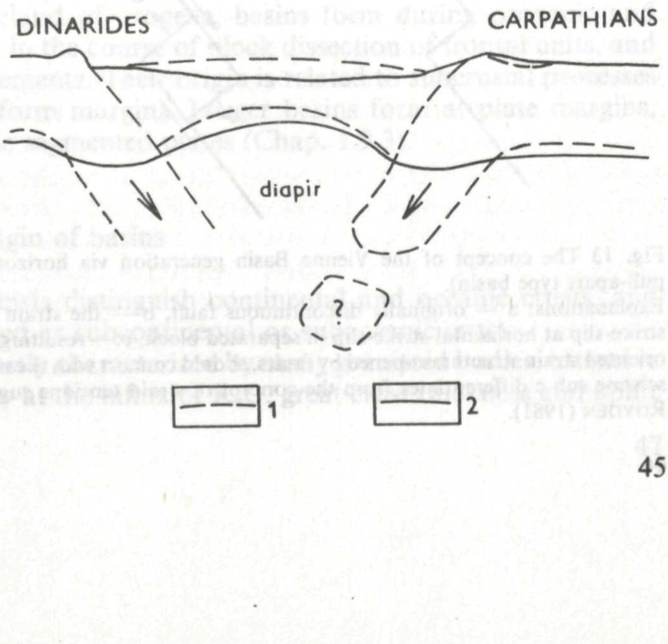



Fig. 11 Model of formation of Pannonian Basin by subduction of lithospheric plates (according to L. STEGENA et al. 1975).

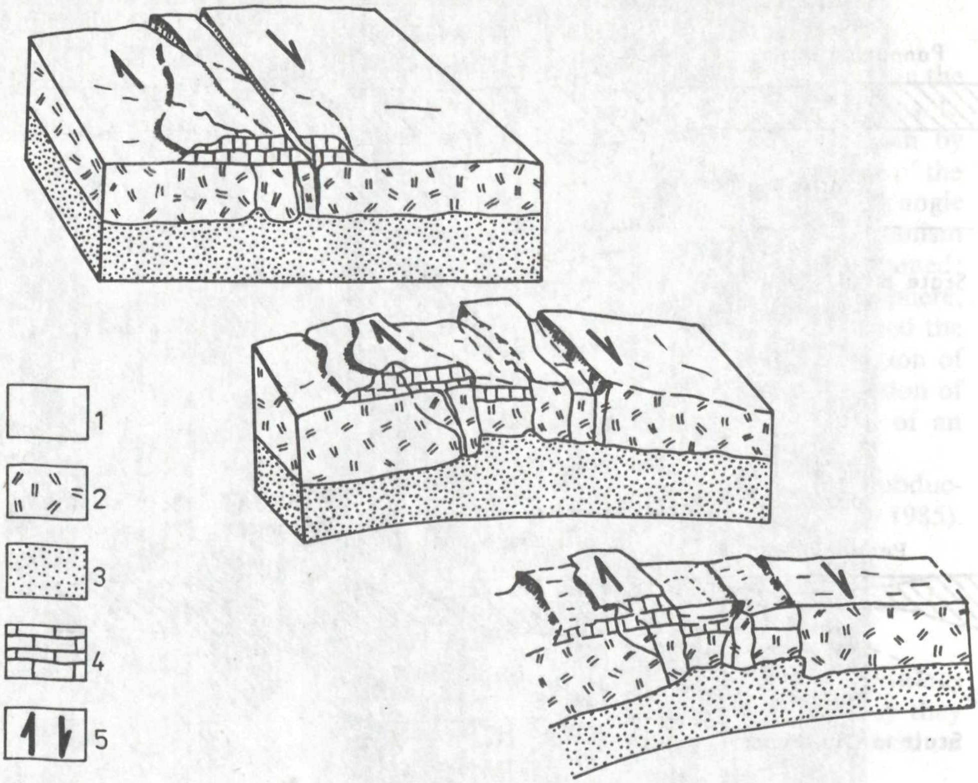



Fig. 12 Origin of the so-called shear basin (adapted according to A. R. GREEN 1977).
Explanations: 1 — sediments, 2 — continental crust, 3 — upper mantle, 4 — extrusive basalts, 5 — strike-slip movements.

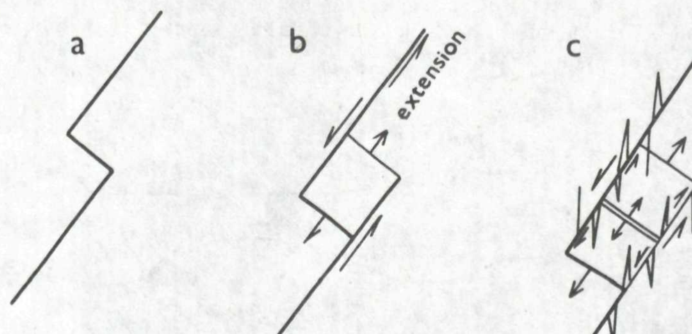



Fig. 13 The concept of the Vienna Basin generation via horizontal strike-slip movement (the pull-apart type basin).
Explanations: a — originally discontinuous fault, b — the strain is oriented in the direction of strike-slip at horizontal strike-slip in separated block, c — resulting fault picture in the basin (N-S oriented strain structures opened by faults, should connect with the main horizontal strike-slip). The scheme sub c differentiates from the concept of strain tensions suggested by F. HORVÁTH and L. ROYDEN (1981).

Beneath the Vienna Basin and in its sedimentary filling sinistral strike-slip movements on the Schrattenberg-Bulhary fault zone are presumed by Z. ROTH (1980). Since no strike-slip faults were evidenced in this zone, R. JIŘÍČEK (1982) rejected the idea.

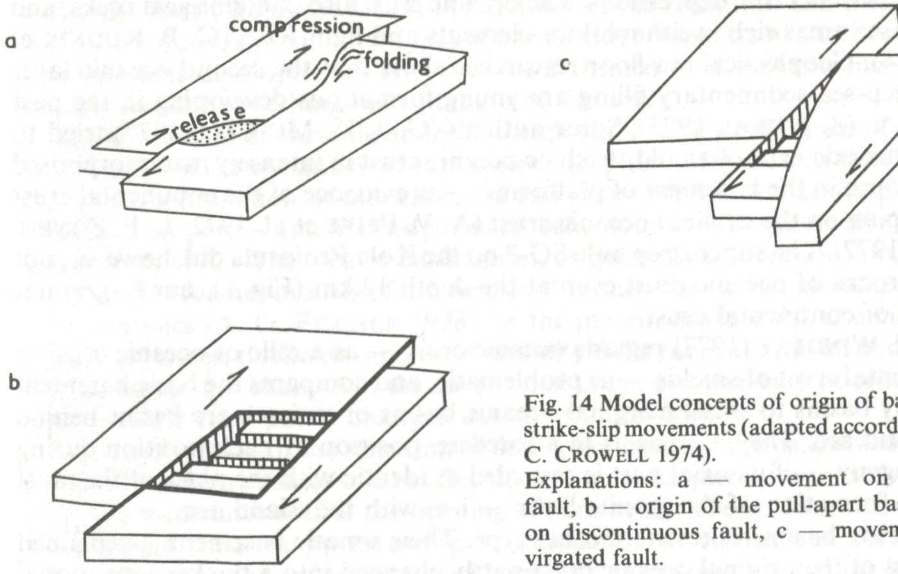



Fig. 14 Model concepts of origin of basins by strike-slip movements (adapted according to I. C. CROWELL 1974).
Explanations: a — movement on uneven fault, b — origin of the pull-apart basin type on discontinuous fault, c — movement on virgated fault.

### 1.6.3 Basins on normal faults

Basins form most frequently on normal- and deep faults. Subsidence is also associated with shear diagonal and horizontal movements. Along faults the basins are smaller and form as longitudinal basins in the hinterland of folded zones, and as transverse basins oriented towards orogen, on transverse faults. On platforms in the foreland of orogens, basins form during orogenic and postorogenic movements, in the course of block dissection of frontal units, and during crustal block movements. Their origin is related to subcrustal processes like re-activation of platform margins. Larger basins form at plate margins, between blocks and inside segmented plates (Chap. 1.9.3).

## 1.7 Crustal types and origin of basins

Geologists and geophysicists distinguish continental and oceanic crusts, and intermediary types denoted as subcontinental or suboceanic crust.

Continental crust is mostly characterized by many granitoid bodies, extensive migmatitization — mainly at the influx of $K_2O$, great crust thickness and uplift tendency. The two types of the crust differ in the style of tectonic deformations (J. ZEMAN 1980).

Oceanic crust is characterized by small thickness, dominant basic and ultrabasic rocks of a varied composition — according to the latest data (B. P. ZOLOTAREV 1984, G. B. UDINTSEV et al. 1984). The oceanic crust is regarded as typical of oceanic depressions. The oceanic crust also contains acid rocks, and comprises areas rich in lithophilous elements and light REE (G. B. RUDNIK et al. 1984). Geophysical sea floor researches show that the second oceanic layer and deep-sea sedimentary filling are young formations developing in the past 150 m.y. (Z. KUKAL 1973). Some authors (G. J. H. MC CALL 1977) tried to prove the existence of an old, Archaic oceanic crust in intensely metamorphosed shields and in the basement of platforms — as evidence of the continental crust developing on the original oceanic crust (A. V. PEIVE et al. 1972, L. P. ZONENSHAIN 1972). The super-deep hole SG-3 on the Kola Peninsula did, however, not reveal rocks of oceanic crust even at the depth 12 km (Fig. 1), nor K-granites typical of continental crust.

B. F. WINDLEY (1977) regards oceanic crust — as a relic of oceanic crust in continental crust of shields — as problematic, and compares the basic basement of many basins to small marginal oceanic basins or to back-arc basins behind the island arc. They are basins in a foredeep position or back position during the orogeny — if its axial part is regarded as identic with the place of the most intense formation of the granite layer — i.e. with the island arc.

Inter-arc basins represent another type. Their simatic basement is explained as a rest of the original oceanic crust partly changed into a thicker transitional crust (suboceanic or subcontinental). Both the back-arc basins and the inter-arc basins are characterized by great mobility, great thickness of sediments, fault dissection and by volcanism, mostly of the basalt-rhyolite or basalt-andesite type. Gradual increase of crustal thickness, and the change of the crust type into continental are asociated with the change of basins into shallow-water and shelf basins to become part of a system of continental structures.

Basins on the continental margin with mafic (simatic) composition are denoted as deprogens by N. YA. KUNIN (1984). They are parts of platforms, like the Pericaspian depression in the northern extension of the Caspian Sea. According to the latest data (A. L. YANSHIN et al. 1984) the depression had already since the Late Proterozoic and during the Early Paleozoic been a sea, often deep, in which shallow-water carbonate sedimentation had proceeded at the margins since the Devonian time. The depression represents a long subsidence, with mobility controlled by the simatic composition of the crust. It is regarded as the southern margin of the East-European Platform. The depression is situated at the southern margin of the Uralian ensimatic orogen. If the intense post-Paleozoic subsidence was associated with crustal basification (V. V. BELOUSOV 1982), then basification proceeded in old simatic continental crust. Ensimatic inter-arc basins resulting — according to CH. H. SCHOLZ et. al. (1971) — from crustal basification by mantle diapir squeezed out by the subducing lithosphere, represent a specific case.

In the new plate tectonic model two types of crust and lithosphere, and their significance for tectonic mobility and types of tectonic units are emphasized.

Recent researches in oil- and gas-bearing areas offered plentiful data on the variable structure of the crust in the basement of basins. Mainly seismic investigations and two super-deep drill holes on the Kola Peninsula (SG-3) and near Baku (the Saatly hole) show that the division of the crust into continental and oceanic is insufficient, and that types of basins differ from one to another in their structure and development. They were not formed in a uniform mechanical manner like the stretching and thinning of the crust.

The knowledge of variable material composition, and significance of density differences for the dynamics of tectogenesis inspired a revision of mechanistic ideas, and compilation of models based rather on information about physical--chemical properties and hydrodynamical behaviour of lithosphere (R. W. VAN BEMMELEN 1972). The idea of basification, or generally of oceanization, served as a basis for further models of the genesis of basins, also accepted in the present plate tectonics (A. G. FISCHER 1978). In the present models also chemical and physical changes, thickness- and crust mobility differences are considered. There may be secondary changes in the continental crust, or there may be relics of an older oceanic crust partly changed due to granitization.

In an application of the structure of lithospheric plates, the basins with clastic sediments and carbonates of a smaller thickness are frequently regarded as relics of shelf sedimentation after the opening of continents. The new-formed oceanic crust should then disappear owing to subduction or collision of continental plates. The question is important with respect to prospection for oil and gas: eventual erroneous determination of a sedimentary unit as a shelf complex, more prospective for hydrocarbon deposits — may cause performation of superfluous explorations. So the genesis and history of sedimentary non-oceanic basins must be paid a greater attention. Reconstruction of the history of continental margins is very significant as well.

Authors of some tectonic models presume oceanic crust preserved as a restite of incompletely subduced oceanic crust in the place of present basins (S. S. KARAPETOV 1984). In a link with this presumption M. Mahel' (1978b) presented a model of crustal types and their development of the crust in the West Carpathians and adjacent regions. M. MAHEL' (l. c.) introduced a geologic-historical approach to the problem. Basing upon variable distribution of granitoids, geophysical data on thickness of the crust, on the relation between the so-called granite and basalt layers, upon various degres of Hercynian consolidation (consolidation of the crust is according to M. MAHEL' controlled by the intensity and extent of granitization). M. Mahel' divided the Carpathian crust into continental, intermediary and paraoceanic to oceanic (in the geologic past). In belts of Precambrian and Phanerozoic oceanic or paraoceanic crust the Upper Paleozoic, Mesozoic and Cainozoic troughs formed. In Mahel's models the denser oceanic crust was subduced beneath the lighter continental granite crust and the process resulted in reduction of partial troughs and in the folding of their sedimentary filling (Fig. 15). Subduction in this sense is not associated only with convergent plate margins and mobile troughs are not always parts of a former ocean. In most tectonic models it is difficult to solve the problem of closing of the Paleotethys ocean, because spatial reduction of large areas of oceanic crust at once is then inevitable. And there are no complexes of formerly extensive areas of oceanic (deep-sea) sediments in the folded sediments. So the crust of the oceanic type need necessarily be associated with the existence of the ocean floor. If every type of the crust should represent a rest of an ocean, we had to presume generation of extremely numerous closed oceans in areas with the presumable oceanic stage of their development denied by geologic data. For example, in the basement of the Pannonian Basin the existence of closed oceans (L. STEGENA et al. 1975, I. VARGA 1978) is not geologically evidenced (cf. L. TRUNKÖ 1977, V. G. SVIRIDENKO 1976, F. ČECH and J. ZEMAN 1982). Deep faults cannot be regarded as subduction sutures (cf. M. MAHEĽ 1978b), because they represent vertical boundaries of blocks (F. ČECH — J. ZEMAN 1980).

In the second model (F. ČECH 1982, F. ČECH — J. ZEMAN 1984) the rests — belts or blocks — of slightly or very slightly consolidated and granitized crust in the present continental crust are presumed. The crust preserved its smaller thickness than the granitized consolidated crust, and also a greater mobility. It was affected by repeated subsidence. J. ZEMAN (1978) denoted blocks with the crust in the Bohemian Massif as simatic, and the crust with dominant mafic rocks, with a small portion of granitoids, migmatization and often with a slight metamorphosis — as simatic or suboceanic. On this crust the repeated generation of basins proceeded — with some breaks — since the Upper Proterozoic to the Cretaceous.

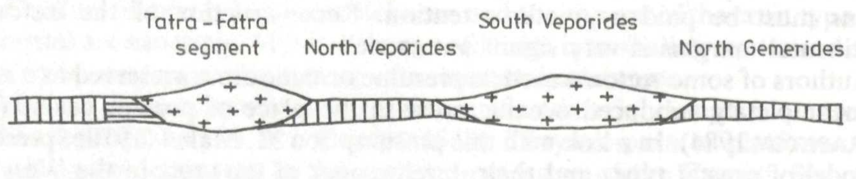



Fig. 15 Simplified model of the evolution of Paleozoic crust in the Western Carpathians (adapted according to M. MAHEĽ 1978).
Explanations: 1 — sediments, 2 — continental crust, 3 — oceanic crust, 4 — para-oceanic crust.

Terminology concerning the crust of a transitional type is not unambiguous. R. TRÜMPY (1975) introduced the term "paraoceanic crust" for belts with partly oceanic and partly continental crust of a reduced thickness. Since the actual composition of the crust is not known, it cannot be defined and classified. We only have geologic information on surficial geologic phenomena, like petrographic and petrochemic composition of rocks, the history of sedimentation, the thickness and degree of deformation of sediments. All those features reflect higher or lower crustal mobility.

It, however, should be noticed that to postulate the existence of fossil oceanic crust on the basis of indirect evidence — like facies or magmatic development, is difficult and may result in subjective interpretations. Petrographic criteria applied on the genesis of mafic rocks (e. g. A. MIYASHIRO 1973, J. A. PEARCE 1975, R. G. COLEMAN 1977), by some of our petrographers (F. FIALA 1977, D. HOVORKA 1978) are not generally accepted and their interpretation is often criticized. The existence of basic crust in the basement of basins should be based upon a complex of data, mainly of geophysical and historical-geological character. A purely petrochemical or other (e. g. geophysical) criterion is insufficient for the definition of a crustal type.

The first exact data on the composition of the upper part of the crust beneath a basin — the Middle Kura intermontane depression in the Transcaucasian Median Massif resulted from the super-deep hole Saatly 1, planned to reach the depth 15 km. It reached to the depth 8.500 m (Y. A. KOZLOVSKY and A. L. YANSHIN 1984). According to geophysical data a basalt layer should be at a depth of 8—9 km. In the basement of Pliocene to Quaternary molasses ranging up to 2.830 m in thickness, is a 700 m thick layer of Cretaceous and Late Jurassic carbonates, underlain by a more than 4.900 m thick volcanodetrital Jurassic sequence: lavas, tuffs, subvolcanic bodies of basalt-andesite-dacite formation. Neither the Saatly borehole nor the borehole SG-3 in the Kola Peninsula (in the second half of 1984, 12.065 m deep) confirmed the geophysical interpretation of the existence of the basalt layer. The Saatly borehole was the first in the world to reveal the composition of the upper part of the simatic crust, without surficial geophysical and geological indications of the presence of the granite layer.

At the present state of knowledge I regard the terms simatic, mafic, suboceanic, paraoceanic and quasi-oceanic crust as synonymous terms. Some Soviet geophysicists (like N. A. BELYAEVSKY 1974) used the term "suboceanic crust" for the interpretation of seismic sections through the crust without indications of the granite layer and with the crust thickness greater than that of the oceanic crust and smaller than the thickness of the continental crust. Geologists use the term in a wider sense for the crust under larger basins where in geophysical records the oceanic crustal type may be calculated. This crustal type is mostly supposed for inter-arc and back-arc basins, a. o. So I range the term "suboceanic crust" among the above mentioned synonyms.

The mode of preservation of simatic crust in continental crust is controlled by variable degree of sialization and consolidation of the crust, formerly perhaps of oceanic origin. Uneven distribution of granitoids — also in shields — indicates that the areal extent of granitization was neither great nor planetary, and that granitization was restricted to continental nuclei or mobile orogenic zones with partly sialized or non-sialized crust preserved among them. The problem was solved by J. ZEMAN (1980), and by F. ČECH — J. ZEMAN (1984) with respect to older crystalline units in Alpine mobile Europe and in foreland of Alpine Europe (J. ZEMAN 1979).

I tend to accept the hypothesis of accretion for crustal development, i. e. from oceanic to continental crust (e. g. A. V. PEIVE et al. 1972) because of evidenced zonal arrangement from ovoidal sialized nuclei in shields (T. N. SPIZHARSKY 1984) and in continents, for instance through island arcs (A. V. PEIVE et al. 1972, a. o.) or riftoid structures (J. ZEMAN 1979, 1984, N. A. BOZHKO 1984), or intracontinental ensialic folded belts evolving from original continental rifts of higher permeability. Accretion zones of the Andean type, a. o. are also accepted. This concept is opposed by H. STILLE (1924) who explained all phanerozoic geologic processes by reactivation, dissection and mobilization of the Precambrian Platform. H. STILLE'S concept is still accepted by some Czechoslovak (e. g. Z. ROTH 1980, V. ŠKVOR 1981) and foreign geologists (V. V. BELOUSOV 1954, 1962, geologists of GDR, a. o.). In accepting this concept of the origin of basins it is necessary to consider preferably global processes of basification s. l. or spreading mechanisms in plate tectonic terms. So my approach to the genesis of basins is based on the accretion model. I do not deny mobilization (by the process of basification s. l.) of plate margins or ancient sialization nuclei — microcontinents — median sialic massifs.

A type of the crust is a phenomenon, most frequently omitted in models of the genesis of basins, or it is more-or-less simplified assuming the continental crust of the platform. Mantle elevations beneath basins with accomplished subsidence indicate the resistent character (some kind of permanence) of thinned crust and also indicate that basification if associated with the crustal thinning — is an irreversible process. It is likely that the original simatic relic crust, primarily thinner than the sialic one is a conservative element preserving its mobile character in areas of denser endogenous activity. So basification s. l. is a leading process permanently tending to crustal mobility, subsidence and to intense fault deformations. Rocks in mantle elevations are permanently affected by lower pressure (geostatic load) than beneath the surrounding thick sialic crust. So elevations may represent natural draining paths of the thermal flux, ascent of astenoliths and secondary diapirs of rocks differentiated in upper mantle or lower crust.

The concept of restite simatic crust, partly sialized or generally non-sialized in the course of the evolution of continental crust has fixistic features and signs of petrochemical permanence of crust composition. This idea is rejected also by some fixists, like N. V. BELOUSOV. Interestingly, elements of the hypothesis of crust permanence also occur in some plate tectonic concepts concerning the existence of the ancient ocean, like Panthalassa inherited by the Pacific Ocean or the "Permanent Tethys" presumably joined with the ancient Pacific Ocean (J.

AUBOIN 1984). According to Auboin the principle of the ocean opening as a global process cannot be accepted. There are further authors postulating the permanence of oceans or crustal types in their concepts based on presumed basinal heredity: Mesotethys as a precursor of the Mediterranean Sea and of South-Asiatic basins (I. V. ARKHIPOV 1984, a. o.), or Mesotethys as a system of depressions having been formed since the Triassic time (M. A. BAER 1984), predisposition of rifts as hereditary mobile structures on ancient weakened zones — lineaments (H. J. ILLIES 1981, N. A. BOZHKO 1984). In these concepts the pre-riftogenic activity is presumed to be controlled by the linear mantle activity followed by the generation of riftogenic linear mantle diapirs (A. V. RAZVALYAEV 1984).

This is the basis of my approach to the history of linear or semilinear basins associated with a hot, less viscous mantle. Dynamic effects of mantle diapirs upon the origin and development of sedimentary basins may be so intense as to remind of a rapid spreading of the upper lithosphere. This process may also be favourable for the generation of smaller dispersed non-linear basins according to V. D. TCHEKHOVITCH (1984). There also is geophysical evidence of the link with mantle diapirism, in marginal seas in the West Pacific Ocean (ZHANG YONGXIA 1984). Data concerning riftogenic features in the early stages of the genesis of back-arc or inter-arc basins (A. YA. SHARASKIN 1984) are increasing at present. The data are in accordance with the further explanation of the genesis of rifts and superimposed basins by deep diapirism or by lithosphere stretching due to the overthrust of the continental lithosphere margin over a subducting oceanic plate (J. AUBOIN et al. 1984, A. YA. SHARASKIN 1984 a. o.). The development and mobility of linear rift basins or aulacogens are mostly controlled by deep-seated faults whose endogenous activity is due to the hot mantle material uprise. Their function is stressed by linear draining control and orientation of endogenous processes.

In contrast to sialic crust the denser simatic crust tends to subside also without basification and additional phenomena usually increasing the density of crustal segments. If such processes (like eclogitization) proceed, then their activity and efficiency in the simatic crust (and their action radius) are higher than in the thicker sialic crust. The history of crystalline elevations shows that the sialic crust was hardly affected by basification — except its margins where the crust was thinner on the transition to the simatic crust.

The explanation of increasing crustal density is still hypothetic but most geologists and geophysicists treating non-oceanic basins agree in the effects of mantle diapirs basifying the crust and uprising from asthenosphere to lithosphere (asthenolith of R. W. VAN BEMMELEN 1972, "mantle plume" in terms of plate tectonics — B. F. WINDLEY 1977). The same principle is applied on geosynclines lying on the thin oceanic crust adjacent to continental crust (J. R. CURRAY 1978).

All data and their interpretation indicate that no larger basins formed on the thick continental crust without changes in its thickness and density, i. e. without its alteration into a crust close to oceanic crust or without its oceanization.

Oceanic crust is a prototype of mobile crust with uncompensated subsidence. This crust is in gravity disequilibrium owing to variable heating or cooling. It is not thick and is easily broken by large faults. Continental crust shows different gravity properties. On the contact between the two types of the crust an inequilibrate state rises to cause mobility.

With respect to the genesis of tectogens and associated basins, the contact between the two types of crust is a significant dynamic phenomenon (A. V. PEIVE et al. 1972, a. o.).

Geophysical results (N. A. BELYAEVSKIJ 1974, M. P. H. BOTT et al. 1971, S. V. CAREY 1976 a. o.) and petrochemical data (D. H. GREEN — A. E. RINGWOOD 1969, B. G. LUTC 1975 a. o.) indicate a different composition of upper mantle beneath oceans and continents. Quantitative data are missing so far. On the grounds of the existing information the lithosphere is divided into oceanic and continental.

So there are two principal types of lithosphere:

a) oceanic, 60—80 km thick, with a highly mobile subsidence, represented by oceans,

b) continental, 100—150 km thick, with elevation mobility represented by shields.

Both lithospheric (and crustal) types are divided into subsiding and relatively uplifting units or vice versa. This division proves the autonomous development and internal heterogeneity of lithospheric structure and dynamics. Basing upon variable mobility, the two types mentioned may further be divided to distinguish intermediary types. Following is W. D. Mac Donald's (1972) classification:

1. normal oceanic crust,

2. oceanic crust with a thicker basalt layer — the so-called platillo crust, a type of abyssal plains,

3. crust of intermediary type — around island arcs and in small oceanic basins (partly corresponds to suboceanic crust in my concept),

4. young tectonic crust — continental crust of a smaller thickness,

5. continental crust — of the shield type with thick granite — metamorphic complexes.

The type sub 4 needs not necessarily correspond only to younger crust but also to older undeveloped continental (or suboceanic) crust. The origin of primary oceanic basins is associated with type 1. Secondary basins — non-oceanic — are associated with the crust of types 3 and 4. It is actually only the type 4 — but in places where type 3 developed (some inter-arc and back-arc marine basins). So far it has nowhere been proved that oceanic crust changed into continental without the evolutionary stage of island arc with adjacent seas — oceanic basins of restite character. V. V. BELOUSOV (1982) has a different opinion but without any direct evidence. So type 3 becomes part of type 4, and in a more advanced stage — also of type 5. On the other hand, platforms plates partly regenerate by the formation of rifts and by basification along platform margins with the process of a change from type 5 to type 4 (incompletely), later on to type 2 and perhaps even 1. The stage of the island arc cannot appear but type 3 may be represented in a qualitatively new form — as non-basified continental crust. The Black Sea is a possible example; the Pannonian Basin is in an unaccomplished (or initial? — cf. V. V. BELOUSOV 1982) stage of crustal change. It is difficult to distinguish rests of oceanic crust and their incorporation into continental crust. The genesis may only be judged according to indirect information on the geologic history of the region. The oceanic crust may be a rest of crust of both the types between 1—4 (5) and between 5—4 (3) — 1. The second trend is dominant in rift structures, on margins of an ancient unstable plate — as far as I regard it as heterogeneous rather in the sense of M. MAHEĽ (1978b) than of Z. ROTH (1962, 1974) supposing the existence of a homogeneous plate, corresponding to the so-called Triassic carbonate platform in models of Alpine geologists (like J. DEBELMAS 1980).

Because of the lack of historical data, geologists differ in their opinions about the question whether the primary crust was oceanic or continental (V. ŠKVOR — J. ZEMAN 1976). The plate tectonic model points out to the younger, secondary emplacement of oceanic crust. Most geologists try to prove that the primary oceanic crust has been forming since the Archaean time (G. J. H. MC CALL 1977). It will, however, be necessary to correct the tendency to regard every Precambrian basin with basic volcanism as a relic of oceanic crust (B. F. WINDLEY 1977). Relics of oceanic crust are in the crust of shields and ancient platforms in the place of longtermed mobile depressions. The crust was altered by later sialization resulting in consolidation of respective parts of lithospheric plates. The primary oceanic character may hardly be proved — it can only be inferred according to indirect geological information which is not a unambiguous criterion.

The character and petrochemical indications of oceanic crust were repeatedly discussed from various views (A. MIYASHIRO 1973, B. G. LUTC 1975, R. G. COLEMAN 1977 a. o.). From the geological view most reliable are

a) tholeiites of oceanic character,

b) ophiolites, mainly ultramafites,

c) areas of long subsidence with volcanic-sedimentary filling,

d) hereditary subsidence mobility and anti-consolidation trend of development,

e) absence of larger amounts of granite masses and migmatization, mainly of $K_2O$ migmatization,

f) petrochemical prevalence of rocks with Na > K,

g) sedimentary lithofacies, mainly fauna with deep-sea indications, pelites, radiolarites, a. o.

Among indicative geophysical data on oceanic crust are positive gravity and magnetic anomalies, thinned crust, and a thick zone of high velocities (basalt layer).

Geophysical indications show the present geophysical pattern and may reflect the youngest, basification processes. The problem of the existence of oceanic crust and its relics is usually a matter of geologists because they know best about the history of the geologic units under study.

Lithofacies development of sediments is a complementary recognoscation factor but actuogeologic data are not absolutely unambiguous. Deep-sea sediments may also form in depressions on the ocean-continent contact on continental crust and continental sediments may accumulate on the deepest part of the continental slope. The sediments may be regarded as a proof of the existence of continental crust (arcoses, polymict sandstones and conglomerates).

Investigations in the North Atlantic Ocean between Greenland and Norway revealed Paleozoic metamorphosed rocks typical of the continental crust (S. W. CAREY 1976). Even on the ocean floor are submerged blocks (or microcontinents) with reduced granite layer (the Indian Ocean, Falkland Plateau, Bermuda Plateau — R. MARTIN 1972).

On the other hand, the existence of intermediate or alcalic lavas and sediments of continental and shelf character need not exclude the existence of the originally oceanic crust in the basement of basins.

Zones of oceanic crust, mostly ophiolite belts, are regarded as subducted relics or as obducted oceanic crust. It is, however, impossible to explain all occurrences and especially nonlinear relics in this way. J. R. CURRAY (1978) distinguishes geosynclines:

a) in intraplate position without signs of subduction, ranged to an embryonic ocean,

b) on plate margins with their dynamics controlled by subduction.

The crust (we do not know if including lithosphere) of continents is heterogeneous also in shields and mainly in the basement of plates. It is evidenced by different lithologic and magmatic development, especially by concentrations of mafic rocks in smaller zones or areas of more-or-less isometric shape. Classical areas are represented by the African and the Australian shields (G. J. H. MC CALL 1977). Sometimes the basic crust is partly altered by sialization, or preserved in mafic relics — xenoliths in granitoid plutons, i.e. completely replaced by the granite layer. Tonalites — which may result from melting oceanic crust — are a significant phenomenon in the process.

Continental crust extending around sialization nuclei encloses gradually the relics of oceanic crust and stops their subsidence mobility. If reactivation — mobilization of the continental crust takes place, then in areas with relics of the primary oceanic crust or of suboceanic crust later partly altered by sialization. M. MAHEĽ (1978b) presented evidence of this process in the Carpathian-Balkan region. This theoretical interpretation is not in accordance with the model of lithospheric plate because of the lack of geologic evidence of the collision of plates and of the ocean closing.

Relics of oceanic crust are most frequent near the continent-ocean contact. The influence of the type of crust upon the subsidence rate of basins is illustrated by a series of basins along the coast of the Mexican Gulf and Atlantic Ocean (A. G. FISCHER 1978). There riftogenesis commenced in the Jurassic time. The area adjacent to continental crust shows a smaller subsidence than the central part (the south of Florida, Cape Hatteras). The subsidence is greatest in the area closer to the Atlantic coast. A. G. FISCHER (1978) presumed there a new-formed geophysically indicated oceanic crust.

Basins with oceanic crust on deep floor belong among uncompensated and are compared to small oceanic depressions (P. R. VOGT et al. 1969).

The oceanic crust — young and ancient, slightly or intensely altered — has a dominant or even decising influence upon the foundation of basins and the rate of subsidence.

So the existence of the heavier crust of the oceanic type may be regarded as significant in forming conditions favourable for the foundation and development of basins. Relics of this crust may also play a decising role in the generation of basins in continental crust, consolidated earlier but remobilized by new tectogenesis associated with deep processes. Relics of crust of the oceanic type — either primary or basified in the past owing to variable dynamics of crustal development (prior to the foundation of the basin) are economically significant with respect to raw-material deposits. So the deep research of the crust should be paid a great attention. It is not right to derive dynamics only from crust types and to deduce speculative paleogeography. Dynamic problems cannot be neglected in relation to the study of the origin of raw-material deposits.

The Carpathian-Balkan region served as a basis for the model of the development of basins on continental crust (S. I. SUBBOTIN et al. 1972). The thinning of crust as part of the destructive process in the development of crust was regarded as secondary. The Black Sea is presented as an extreme example of oceanization in accordance with the opinion of other Soviet geologists (V. E. KHAIN — V. I. SLAVIN 1972). The genesis of the Black Sea is not reliably explained, although S. I. SUBBOTIN et al. (1972) and later on P. GOTCHEV (1976) admitted the primary riftogenic stage. But the concept of the non-rift mantle diapir cannot be excluded as well (F. ČECH — J. ZEMAN 1985).

So the basins may also form in the progressive trend of the development of crust from the oceanic to the continental (cf. S. I. SUBBOTIN et al. 1972), namely in the postgeosynclinal stage on the rests of slightly sialized or in places non-sialized crust enclosed in continental crust within which they were most frequently delimited tectonically. The seismically active Odesa tectonic zone on the Black Sea coast is a typical example of the contact between the oceanic and continental crust. Analogues of this type in other basins with a similar crust type will be discussed later.

Differences in density and thickness of the crust cause differentiated isostatic movements resulting in tectonic detachment of the single blocks or zones of basite crust from the continental. Faults, particularly those in tangential contact with the margins of mantle elevations (F. ČECH — J. ZEMAN 1984), facilitate crust subsidence and formation of a depression. It is likely that the subsidence is stimulated by deep activity with its source in the asthenosphere, owing to the disturbed gravity equilibrium. Increasing density of crust in a segment to become subsidence-mobile, is significant for dynamics. The process is characterized by:

a) progressive trend — if the restite (sub)oceanic crust is further basified — basaltized,

b) regressive trend — if basaltization affects continental crust.

The existence of faults and fault zones facilitating the ascent of magma lighter than the mantle, oversaturating the ancient crust, is structurally predisposed. It is likely that the process mentioned under a) causes rapid subsidence of the basin and accumulation of sediments, and the process mentioned under b) causes a slowlier subsidence and formation of basins with sediments of a smaller thickness.

Subsidence associated with sedimentary filling (compensated basins) causes increasing load acting on the underlying crust and phase transitions in rocks (eclogitization?). These result in greater density differences in crust and in acceleration of subsidence. The subsidence is likely to stop in the time of renewal of the equilibrate state in the mantle, when phase-altered rocks in the crust beneath a basin acquire physical properties of rocks of the mantle. According to thickness of sediments in compensated basins the subsidence would stop at the depth of 10—15 km. In uncompensated basins-like oceans — the subsidence will stop at the depth of 5—6 km which corresponds to the average depth of the ocean floor.

## 1.8 Influence of sediment mass upon subsidence

In the present models the influence of sediment mass in the basin is evaluated really, at most as an additional phenomenon (I. A. REZANOV 1972, V. E. ARTYUSHKOV 1981). Sedimentary filling cannot cause or accelerate subsidence unless it overcomes elastic properties of crust. It may have an additional influence in asymmetrical basins if its mass affects a narrow crust belt by concentrated geostatic pressure. In symmetrical basins, particularly in those with a small sediment thickness, the influence of density upon subsidence and physical conditions equals zero.

Subsidence, uncompensated by sedimentation, requires isostatic disequilibrium of the actively sinking part of lithosphere. In such case, sedimentary filling may cause an additional pressure due to sediment mass affecting the physical conditions in deeper lithosphere beneath a basin. Such phenomena are represented by phase change of rocks, for example eclogitization of basic magmatic rocks and amphibolites. Increasing density causes further subsidence. In this way I. A. REZANOV (1977) explains a geophysically proved fact, namely that Mohorovičić discontinuity beneath larger basins according to seismologic data rises up and forms elevations. Then the nature of Moho may be explained as a boundary of a combined chemical and physical character. This results in basification or ultrabasification of crust with the features of oceanic crust. This process may proceed in both the oceanic and continental crust.

The sediment density has a less significant influence upon compression of basement rocks, decrease of their porosity, recrystallization and increase of their density. It is also supported by overheating in the area of higher temperatures and escape of water from rocks. So far we, however, do not know whether the increasing density of basement rocks exceeds the deficit in sediment density or not.

In basins with overthrust (for example in foredeeps) the overloading is rapidly increasing, porosity decreases, rocks get compressed and an uncompensated subsidence may take place.

To test the above mentioned ideas of the subsidiary influence of sediment density upon subsidence, I have performed the following consideration and calculation.

The principal factor in differentiation of density of the lithosphere is its layered structure with the increasing density of rock complexes with the increasing depth. The increase of density is due to chemical changes, i. e. the change of petrographic rock types in the series granite-gabbro-peridotite, and physical changes, i. e. in porosity and density of rocks owing to compression by increasing geostatic and tectonic pressures. It is a problem to determine the mean density of lithosphere. At the chosen value $3.15 \cdot 10^3$ kg . m$^{-3}$ we can at the base of lithosphere, i. e. on the contact with asthenosphere at the depth of 100 km, calculate with geostatic pressure $3.15 \cdot 10^3$ MPa.

If the basinal filling consists of sediments, according to the grade of diagenesis ranging from $2.2 \cdot 10^3$ to $2.7 \cdot 10^3$ kg . m$^{-3}$ (about $2.5 \cdot 10^3$ kg . m$^{-3}$ in average), then the pressure upon the basement beneath the basin, for example at the depth of 5 km, will be $12.5 \cdot 10^{-1}$ MPa.

The pressure performed by an equal granite column will be greater (density of granites ranges to $2.65 \cdot 10^3$ kg . m$^{-3}$). In the case of metamorphosed rocks with density as high as $2.7 \cdot 10^3$ kg . m$^{-3}$. The pressure will be still higher.

So when the deeper part of the crust is not downwarping under the load of granites or metamorphites with an overthrust sedimentary cover, the influence of the basinal sedimentary filling cannot be decising for subsidence of the deeper part of the crust or even lithosphere. The causes of basinal subsidence cannot be found in simple filling — compensation of basins with sediments, but — as already mentioned — in deeper processes.

Systematic deep seismic investigations of the crust in Eruopean socialist countries and in the Asian part of the U.S.S.R. revealed changes in geologic structure and in physical properties beneath depressions and elevations (S. I. SUBBOTIN et al. 1972, N. A. BELYAEVSKIJ 1974). If we do not consider rapidly subsiding depressions with a high heat flow (rifts, Tyrrhenian Sea, a. o.), then beneath the crust of other basins a denser upper mantle dominates. Higher P-waves velocities in the mantle correspond to a greater rock density. The fact has so far not been explained unambiguously.

When we consider the density of a rock column out of the basin and the density of a rock column with basinal filling, then the initial difference in density will get levelled with the increasing depth, if rocks in the lower crust will have the same composition. At the depth of 50 or 100 km the density difference $0.1 \cdot 10^3$—$0.2 \cdot 10^3$ kg . m$^{-3}$ for the first 5 km will be levelled or negligible.

The pressure on the top of asthenosphere at the depth of 100 km out of the basin will be $3.15 \cdot 10^3$ MPa.

If we subtract the effect of lighter sediments with the thickness of 5 km, the resulting pressure beneath the basin will be $(3.15 \cdot 10^3)$ — $(0.1 \text{ to } 0.2) \cdot 10^1$ MPa.

If the thickness of basinal sedimentary filling would increase and the thickness of crust would be the same as in the surroundings of the basin, then the geostatic pressure in the area of the basin can only decrease. It follows that we cannot explain the increase in density (?) of the mantle by overloading with basinal sediments. It can be explained as follows:

a) The mantle density was primarily greater owing to preceding differentiation when for example, a heavier ultramafic restite remained in the mantle after the separation of lighter basaltic masses. This is why the ultramafic column has a greater density than its surroundings and tends to subsidence.

b) Owing to its basic composition the density of crust beneath the basin is greater and so the mantle also gets more compressed.

c) The lower part of lithosphere is ductile, thinning by stretching, and the upper rigid brittle lithosphere is splitting and downsagging into the zone of thinned lower lithosphere.

d) The mantle in the zone of asthenosphere is more ductile and the crust is downwarping into the mantle.

A comparison to oceans with the more dense mantle beneath them than beneath continents — as proved by seismic data (N. A. BELYAEVSKIJ 1974) supports the first explanation (under a.).

Generally, subsidence is explained by the existence of "larger masses" beneath the depression, and they may be geologically explained as follows:

a) by the presence of denser oceanic crust which — in contrast to its less dense surroundings (including the respective mantle segment) — will tend to subsidence;

b) by decrease in density of the surroundings of continental crust, due — for example — to granitization, to emplacement of granitoids beneath the orogenic surroundings of the basin, a. o. A zone of non-granitized crust will get heavier and tend to subsidence in relation to its uprising granitized rim. This explanation might be applied on some intermontane or intramontane depressions.

## 1.9 Classification of basins

Basins may be classified according to sedimentary filling, types of crust, their position in relation to tectonic units, time interval in the evolutionary cycle (geosynclinal model) or tectonic character.

### 1.9.1 According to the amount of sedimentary filling

Classification of basins according to the amount of sedimentary filling is simplest. Distinguished are basins fully compensated with sediments and non-compensated with sediments $\pm$ volcanics (J. E. ADAMS et al. 1951). Non-compensated basins resemble oceans like the Tyrrhenian Sea, the Black Sea a. o. Compensated basins are similar to or identic with intermontane, mainly molasse depressions. The South-Caspian Depression may also be ranged to them (F. ČECH — J. ZEMAN 1985). Mobility and subsidence rate may be derived from thickness, facies and lithologic character of sediments.

### 1.9.2 According to crust types

According to crust types two kinds of basins may be distinguished: ensialic and ensimatic. These terms were introduced for geosynclines at first (F. G. WELLS 1949). Ensialic geosynclines are geosynclines in classical sense, with sedimentary filling derived from continental crust — mainly polymict sandstones, arcoses, kaolinite shales a. o. Ensimatic geosynclines (type of eugeosynclines) are filled with mafic, intermediary, dacite lavas, pyroclastics, shales, graywackes including volcanic material and generally rocks with a lower $SiO_2$ content. Sedimentary filling contains intercalations of quartzose effusive rocks. According to F. G. WELLS (l. c.) the sialic crust comprises increments of ensimatic geosynclines. On his concept the model of mafic-simatic restites in continental crust is based. The model is preferred in this book.

Ensialic basins form on continental crust with indications of the sialic crust type. Most frequently all basins on continents are regarded as ensialic. This type comprises intracratonic and intramontane basins, some intermontane basins, foredeeps, perhaps most molasse basins, some marginal depressions of platforms and of sialic inner massifs; continental rifts and all aulacogens. The ensialic basins may also form on the boundary between sialic and simatic crusts. Ensialic basins are usually superimposed on the ancient fault and fold structure.

Ensimatic basins form on fossil oceanic or simatic crust which will become part of continental crust. The basins may belong to complexes of ensimatic geosynclines (e. g. Uralian basins). Ensimatic basins comprise marginal seas, basins at island arcs (inter-arc and back-arc basins), some marginal depressions of platforms and inner massifs. Most basins of the pre-platform evolutionary stage of the crust belong among ensimatic basins. Subsidence in some basins persists during the tectogenic cycle or several cycles and in the platform regime. In block structure the basins are on simatic blocks (J. ZEMAN 1978).

The heterogenic type of basins is represented by basins generated in the early stage as ensimatic and later on changed into ensialic (by crust sialization). Such are some basins of the Andean type near the Pacific Coast (J. JAROŠ 1975) or marginal geosynclines like the Caledonian British-Norwegian geosyncline (F. W. DUNNING 1977). Paleorifts in which the oceanic crust changes into continental, represent a specific case (J. ZEMAN 1979, 1984).

### 1.9.3 According to position in tectonic units

Classifications of basins were based on the geosynclinal model. New classifications are derived from the classification of plate tectonic units.

### 1.9.3.1 Classifications in terms of geosynclinal model

The latest classification of basins, based on various endogenous regimes, was done by V. V. BELOUSOV (1978b). From this classification I have selected basins with possible coal deposit occurrences and hydrocarbon deposits.

1. In the geosynclinal regime: flysch troughs of eu-, mio- and parageosynclines, basins on internal massifs mobilized during this regime (inner basins or intermontane basins).

2. Basins on the boundary between geosynclines and platform regime: marginal depressions of mixed development, more mobile on the side adjacent to the geosynclinal zone. During the migration of geosynclines the more stable zones will be mobilized, and older sediments will be covered by the younger. At the beginning the process is favourable, later on it may become unfavourable for the quality and preservation of hydrocarbon deposits.

3. In the platform regime: syneclises and aulacogens (their molasse stage). This group may also comprise depressions of the Atlantic type i. e. those which are not on the contact with the folded geosyncline and lack the alcalic-calcareous volcanism.

4. In the orogenic epigeosynclinal or epiplatform regime (orogenic activation of plates, like Tian-Shan): foredeeps, intramontane and intermontane depressions.

5. In the rift regime — continental rifts.

The position of riftogenesis in the geosynclinal model is not determined unambiguously. It is most frequently associated with tectonic or orogenic activation of platforms — the platform orogenic regime — and then it is more proper to denote depressions as tectonic grabens. Rifts may also form in the geosynclinal-orogenic regime on plate margins or in the initial stage of the generation of geosynclines. This is illustrated by the example of riftogeosynclines (W. R. CHURCH — R. K. STEVENS 1971) — a frequent type of developing mobile zones. E. BONTCHEV (1976) denoted similar structures with space-fixed long subsidence as lineament-geosynclinal zones. Z. P. ZONENSHAIN (1972) regards geosynclines as ensimatic and presumes that oceanic rift was the initial stage of geosyncline between two continents.

The type of crust beneath geosynclines represents another problem. According to A. N. PEIVE et al. (1972) eugeosynclines were always ensimatic or oceanic. Mobile troughs (intrageosynclines) are very likely to have ensimatic character. The extending geosynclinal regime also occupies sialic crust blocks and massifs, particularly at margins of ancient continents (i. e. also around sialic crust nuclei and/or microcontinents). We do not know whether mobilization of continental margins affects mostly belts or blocks with simatic crust (for example in the foreland of the West Carpathians). H. STILLE (1953) presumed the existence of a linear marginal depression in a green schist belt beneath the outer part of the East Carpathian belt running along the SW margin of the East-European Platform. (We would rather range the depression among ensimatic basins.)

In classification of basins in terms of the geosynclinal model some problems occurred concerning the time and space range of units, incompleteness of the geosynclinal cycle and many deviations from the geosynclinal cycle. The model was based on the presumption of the foundation of geosynclines on continental crust. Many geologists (e. g. V. V. BELOUSOV 1954, 1962, a. o.) assumed global extension of the continental crust. In geosynclinal models the historically variable crust development was omitted, or the development was ascribed the only trend: either sialization or basification (cf. V. ŠKVOR — J. ZEMAN 1976). So the geosynclinal model was criticized with proposals to reject it (P. J. CONEY 1970, A. D. MIALL 1983, B. F. WINDLEY 1984, a. o.) or revise it thoroughly. Discussion on this problem is still not accomplished (e. g. J. CHÁB et al. 1984).

### 1.9.3.2 Classification in terms of plate tectonic model

In new classifications of basins based on plate tectonics, crust types, their development and different dynamics are considered in addition to the above mentioned problems. Marginal depressions and their position in relation to passive (see the Atlantic type in the geosynclinal model) and active plate margins are thoroughly studied. For the classification applied on the Czechoslovak territory without units of plate size I prefer the terms block, group of blocks or megablock, basins on block margins, on megablock margins, a. o. (F. ČECH 1982).

Marginal basins are associated with both the continental and the transitional crust. The present passive margins occupy 40 percent or 200.7 mil. $km^3$ of sediments if we accept L. E. LEVIN'S (1984) presumption about all basins on the earth (without miogeosynclines) occupying 510 mil. $km^3$. The transitional zone to the fossil margins — the so-called Mesogea occupies 6.5 percent or 33 mil. $km^3$, the transitional zone between passive and recent active margins occupies 7.5 percent or 38 mil. $km^3$. Recent active margins including the Columbian margin (Columbian-type transitional zone after V. V. BELOUSOV 1982) only occupy 4.8 percent or 24.3 mil. $km^3$. Basins associated with the Atlantic--type continent margins (divergent type in terms of plate tectonics) occupy a larger area because there were most favourable dynamic conditions for the formation of basins of this type. With the Atlantic-type continental margins (divergent type in the sense of plate tectonics) are most frequent owing to favourable dynamic conditions of their origin and development.

#### 1.9.3.2.1 General classification of sedimentary basins

The general basin classification (W. R. DICKINSON 1971) is based on their tectonic and paleotectonic setting mainly on their distance from plate margins, on the character of plate margins and crust types differing in their dynamic regime.

The application of classification must be based on knowledge of deep structure, history of basins, and especially of tectonic development — if the basin type changed. The knowledge is significant in practice for the study of changes in hydrocarbon production potential of basins. This is why I also present classifications of basins with hydrocarbon deposits.

Relations between tectonic units and sedimentary basins were evaluated by H. G. READING (1978). He relates sedimentation to three principal types of movements, preferred in plate-tectonic models: divergent, convergent and strike-slip movements along faults. H. G. READING (l. c.) emphasizes that in subduction zones are basins whose origin is associated also with tension movements and vice versa — that rifts as a tension phenomenon might have formed in the foreland of compressional movements. Also basins forming on strike-slip faults may form in divergent and convergent zones. H. G. READING'S (l. c.) relativistic concept represents one of starting points of sedimentary basin classification.

#### 1.9.3.2.1.1 Basin classification after A. W. BALLY and S. SNELSON

A. W. BALLY and S. SNELSON (1980) presented general classification of sedimentary basins with respect to their depositional potential. In their classification the authors considered different types of ancient and recent margins of plates. They emphasize the necessity to study the history of basins and their basement, since the causes of subsidence might or might not have been in the basement, for example in paleorifts or paleosutures (fold zones or lineaments) as relics of former plate convergence. A. W. BALLY and S. SNELSON (l. c.) distinguish three principal categories of basins and 21 basin types within the categories. Their position in relation to plate margins and to the character (rigidity) of lithosphere is the main classification criterion. In mobile zones the authors distinguish perisutural basins on zone margins, and episutural basins in the zone. Like in other classifications, the authors distinguish convergent contact — suture associated with A-subduction, and B-subduction with several basin types. Subduction extending along the steeply dipping Benioff zone to the depth of 600 km is denoted as B-subduction. Not deep intracrustal nappe subduction, or subduction inside lithosphere along a flat zone is denoted as A-subduction (Alpine subduction) according to O. AMPFERER who was the first to interpret extensive continental crust shifts. Basin classification after A. V. BALLY and S. SNELSON (1980):

1. Basins Located on the Rigid Lithosphere. Not Associated with Formation of Megasutures

Related to formation of oceanic crust

Rifts

Oceanic transform fault associated basins

Oceanic abyssal plains

Atlantic type passive margins (shelf, slope rise) which straddle continental and oceanic crust

Overlying earlier rift systems
Overlying earlier transform systems
Overlying earlier backarc basins type

Located on pre-Mesozoic continental lithosphere

Cratonic basins
Located on earlier rifted grabens
Located on former backarc basins

2. Perisutural Basins on Rigid Lithosphere Associated with Formation of Compressional Megasuture

Deep sea trench or moat on oceanic crust adjacent to B-subduction margin

Foredeep and underlying platform sediments, or moat on continental crust adjacent to A-subduction margin
Ramp with buried grabens, but with little or no blockfaulting
Dominated by blockfaulting
Chinese-type basins associated with distal blockfaulting related to compressional or megasuture and without associated A-subduction margin

3. Episutural Basins Located and Mostly Contained in Compressional Megasuture

Associated with B-subduction zone

Forearc basins
Circum Pacific backarc basins
Backarc basins floored by oceanic crust and associated with B-subduction (marginal sea sensu stricto)
Backarc basins floored by continental or intermediate crust, associated with B-subduction

Backarc basins, associated with continental collision and on concave side of A-subduction arc

On continental crust or Pannonian-type basins
On transitional and oceanic crust or W. Mediterranean-type basins

Basins related to episutural megashear systems

Great basins-type basins
California-type basins

1.9.3.2.1.2 Basin classification after A. D. Miall

A. D. Miall (1984) accepted basical principles of basin classification by A. W. Bally and S. Snelson into his modified classification. He emphasizes dynamics of transform and transcurrent faults associated with origin of several basin types. His classification is based upon the Wilson cycle model, i. e. on the interval of the development of lithosphere from the opening to the closing of the ocean. Position of basins in relation to plate margins as a classification criterion is of secondary importance in contrast to A. W. Bally and S. Snelson's (l. c.) classification. The plate margin behaviour is the most important criterion.

Basins are divided into five categories with many geologically similar types formed in different stages of the Wilson cycle. According to A. D. MIALL (l. c.), classification criteria of W. R. DICKINSON (1971, 1974), i. e. crust type, position of a basin in relation to the plate margin, type of plate interaction and distance of the interaction from sedimentary basin, changed in time. For example, basins formed on divergent plate margins get into the convergent, for instance foredeep position during collision at the end of the Wilson cycle.

Basin classification after A. D. MIALL (1984).

1. Divergent margin basins

Rift basins
Rifted arch basins
Rim basins
Ocean margin basins
Red Sea type ("youthful")
Atlantic type ("mature")
Aulacogens and failed rifts

2. Convergent margin basins

Trenches and subduction complexes
Forearc basins
Interarc and backarc basins
Retroarc (foreland) basins

3. Transform and transcurrent fault basins

Basin setting:
Plate boundary transform fault
Divergent margin transform fault
Convergent margin transcurrent fault
Suture zone transcurrent fault
Basin type:
Basins in braided fault systems
Fault termination basins
Pull-apart basins in en echelon fault systems

4. Basins developed during continental collision and suturing

Foreland basins
Peripheral (foredeep) basins
Intra-suture embayment basins (remnant ocean basins)
Associated transcurrent fault basins[1]

5. Cratonic basins

In the individual basin types A. D. MIALL (l. c.) evaluates four phenomena:
1. plate tectonic process generating the basin,
2. mechanism of crustal subsidence,
3. structure of the basin,
4. typical evolutionary development of depositional systems.

[1] this type of basin may be classified under two headings.

In the study of subsidence mechanism the thermal, mechanistic and other basinal subsidence models were considered. In classification the petrologic characters of sedimentary basinal filling were emphasized (cf. D. R. KINGSTON'S et al. classification 1983a).

A. D. MIALL (1984) emphasized the mobilistic dynamic aspect of his classification in contrast to the quasi-fixistic geosynclinal model. MIALL rejects to use terms of this model like miogeosyncline, intermontane basin, a. o. He does even not regard cratonic basins as a fixistic phenomenon because they only seem to be out of the plate dynamics. The basins might have formed on marginal paleolineaments or paleosutures, and might have been connected with mobility outside the craton — a mobility controlled by the transmission of dynamic effects of distant plate interaction along transcurrent faults. Some theoretical problems and phenomena treated by A. D. MIALL (1984), presented in the following parts of this book, complement the chapters on plate tectonic models treated in the first part of this book.

1. Divergent margin basins

Rifts are the main representatives of this basin category. Many authors associate their origin with ancient lineaments — perhaps also paleosutures of ancient convergent plate margins. Some riftogenic basins formed in the island arc zones owing to extension, and embayments with the character of marginal arch basins of active plate margins formed along mature passive divergent margins, i. e. margins with persistent spreading (Atlantic type). But the rift stage may also be identified in some segments of island arcs.

On divergent margins also troughs perpendicular to plate margins form particularly on the crossing of the margins with transform and transcurrent faults. Marinal basins and grabens on the continental slope are also controlled by fault crossings. These structures may be covered with sedimentary prisms or deltas. According to A. D. MIALL (l. c.) the basin variableness is greater on mature margins than on the youthful where rifts dominate (the Red Sea type).

2. Convergent margin basins

In contrast to subduction zones of the A-type, the B-type subduction zones are associated with extremely variable basin types. (A. D. MIALL (l. c.) doubts about the subduction character of the A-subduction zone.) In the second case the arch, or the outer rise as an upwarp in the oceanic crust, trench, and often also the volcanic arc are missing. Fore-arc and back-arc sedimentation regions represent the only character in common.

Basins in back-arc regions of the A- and B-subduction zones differ in the history of subsidence, facies complexes and crust types. In the case of B-subduction the back-arc regions may have oceanic or intermediate crust in the basement of basins, whereas there is only continental crust in the A-subduction. The Pannonian Basin is presented as an example (cf. A. W. BALLY — S. SNELSON 1980). The shape and/or the type of basin changes according to subduction rate, the shape of the subducing plate and its variable position. Migration of the subducting plate towards the ocean results in extension in the back-arc region of the overriding plate (cf. the convergent-extensional margins of AUBOUIN et al. 1984). Back-arc basins then develop under shelf-like conditions. The extension may also be associated with formation of riftogenic basins and basins associated with transform and transcurrent faults. The process may be accompanied with the stretching of continental lithosphere and uprise of mantle diapirs.

Dimensions of basins in the fore-arc region vary in accordance with coast morphology, distance and shape of a ridge between the ocean and the island arc.

The development and shape of the retro-arc (foreland) basins in regions of A-subduction are in the front of thrusting nappes affected by olistostromes and by the extension of melange. Tectonism changes facies and their extent. Retro-arc (foreland) basins also form in compression zones of island arcs where mountain ranges form, and along sutures resulting from plate collisions.

3. Transform margins and strike-slip fault basins

The transform boundary represents the third type of dynamic plate contact. Distinguished are oceanic/continental crust boundaries (San Andreas Fault) and intraoceanic boundaries (like fractures Vema, Mendocino, a. o.). Besides transform faults in the zone of divergent margins there are also convergent margins with transcurrent faults oriented obliquely or parallel to the margins. Many types of fault-bordered basins (shear basins, pull-apart basins), are related to transcurrent faults with strike-slip movements. According to J. C. CROWELL (1974) a volcanic complex and a spreading centre may form in the middle of pull-apart basins. The Salton Trough at the head of the Gulf of California is quoted as an example. Frequent are basins on the crossing of transform faults in the ocean with divergent continent margin. According to A. D. MIALL (1984) 42 basins (Santos Basin, Campos Basin, Reconcavo Basin a. o. in South America) formed in this way.

4. Basins associated with suture zones

In plate tectonic models sutures are regarded as a typical character of a former convergent plate margin. The plate margin was segmented into microplates along the suture. Basins associated with sutures may be divided into four types:

a) Back-arc or retro-arc (foreland) basins, transformed into grabens or segmented by movements of the overriding plate along transform faults.

b) Peripheral or foredeep basins, often at least partly folded.

c) Intra-suture embayment basins (remnant ocean basins), mostly undeformed, with preserved outlier of oceanic crust detached from the subducing plate.

d) Associated graben and transcurrent fault basins on continental crust.

Their formation was controlled by fault invasion from the margin of the subducing plate into continent. The basin may also have a rift character. Associated suture basins were denoted as flysch and molasse basins in the geosynclinal model. Conditions of their formation were more complicated than presumed in the geosynclinal model and the basins might also have formed without orogenic movements. Associated ancient suture basins are part of accretion zones of continents. In these basins subsidence may be renewed in the cratonic stage of crust development.

5. Cratonic basins
It is not possible to explain the origin of cratonic basins by plate tectonic models. They are first of all basin structures in platform cover. They may reflect plate dynamics out of cratons.

The origin of Precambrian, mostly Archaean sedimentary basins represents another problem. It still is not known exactly whether plate tectonics existed in the Precambrian time and whether the early stage of crust formation was affected by other dynamics. According to A. D. MIALL (l. c.) it is necessary to analyze in detail Precambrian sediments and the structure of ancient basins and their basement. Then the problem of the origin of Precambrian basins may be solved more reliably and they may also be classified.

### 1.9.3.2.2. New classifications of basins with hydrocarbon deposits

Classifications of sedimentary basins with hydrocarbon deposits are derived from general classification models. They are variably simplified or detailed. The first classification model aimed at evaluation of basin prospectiveness — i. e. for economic geology — was presented by D. M. CURTIS (1980).

#### 1.9.3.2.2.1. Basin classification after D. M. CURTIS

D. M. CURTIS (1980) distinguishes following basins according to their tectonic position:

1. Cratonic basins in intraplates or on passive continental margins.
2. Riftogenic basins.
3. Marginal basins on active margins, associated with A- and B-subductions.
4. Basins along strike-slip faults.

The position of basins controlled the heat flow, the character of surface and ground waters in the time of deposition, and salt content.

1. Intraplate basins inside plates show a small thickness of sediments and an extremely low heat flow. These factors are not always favourable for the formation of hydrocarbons. Geomorphologic dissection of the surface might have influenced climatic conditions in favour of accumulation and preservation of organic matter. The large areal extent of basins is a favourable factor — as shown by the example of West-Siberian oil-bearing basins.

2. Rifts are inside plates and on their margins, in zones of transition of continental crust to intermediate or oceanic crust. Continental rifts and tectonic grabens are usually in an intraplate position.

A rift opening is often associated with pull-apart basins rimming divergent margins. The pull-apart graben depressions remain at the continental margin. They are detached from the active rift axes and become part of the young passive continental margin (of Atlantic type). The basins are denoted as "pull--apart basins" (Fig. 16). In these basins water circulation is limited, sometimes also a high heat flow is present. The basins are mostly favourable for preservation and ripening of organic matter. Naturally, the amount of organic matter depends upon communication of a basin with the sea-ocean.

Rifts pass over several evolutionary stages:

a) Rifts which do not open and remain continental rifts without the formation of oceanic crust (e. g. the Rhine rift, the Ohře rift in ČSR, a. o.);

b) Rifts with branched arms in triplet junction. One unopening failed arm is oriented towards the continent and two opening arms towards the ocean. In the opening arms oceanic crust may form. In the continental arm subsidence may occur. Then the valley is crossed by a river transporting clastic material to the continent margin. The clastic material usually accumulates in a delta (Fig. 17). The oceanic crust forming in the opening arms is gradually spreading. In the next stage clastic material from the continent is transported by pericoastal currents away from the delta and sediments form on continental margin or slope. They are the so-called miogeoclinal prisms. In literature a continental failed rift was erroneously denoted as aulacogen, i. e. rift entering the sea. In this sense F. Pícha (1979) regarded the Nesvačilka and Vranovice grabens in the Carpathian Foredeep as aulacogens. According to D. M. Curtis (1980) it is a "pseudoaulacogen", i. e. the failed arm of a rift triplet. Examples of such

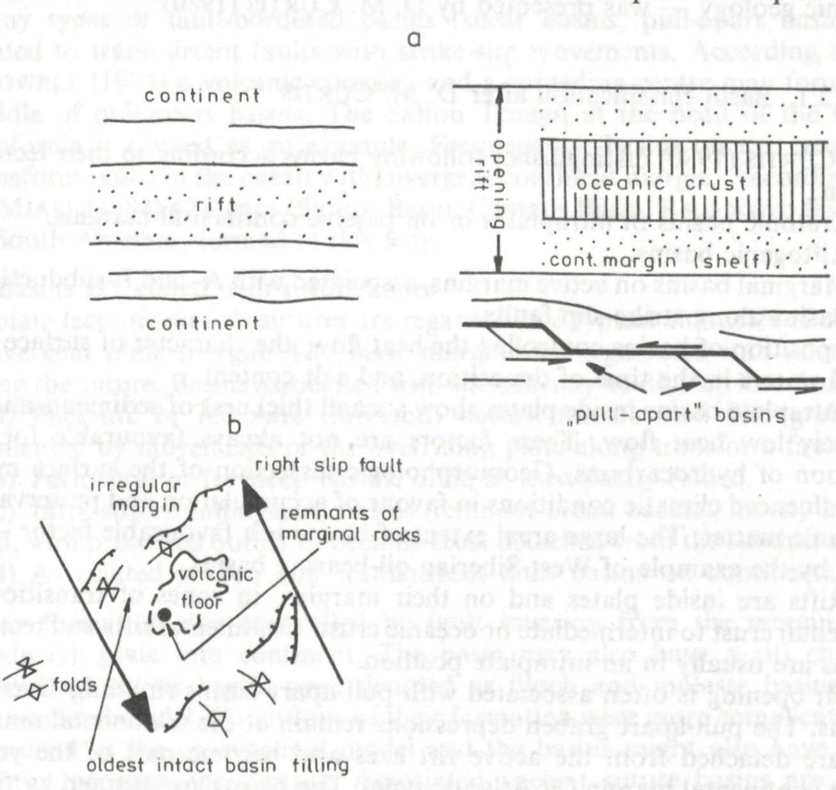



Fig. 16 Origin of the pull-apart basin type on transform border. Explanations: a — at divergent margin, b — general model (simplified and adapted according to I. C. Crowell 1974).

structures on the Atlantic continental margin are the Benue Trough in West Africa and the Viking Graben in the North Sea.

On the continental margin the sediments of failed rift arms, cañyons, and pull-apart basins may be covered with miogeoclinal sediments at the subsidence of continental shelf. Thus ancient sediments may start ripening and become source of hydrocarbons. Examples are the Suez and the Aden Bays.

When organic matter accumulated in some beds inside a miogeocline, its maturity was controlled by the depth of downwarping and by subsidence rate of a continental margin. Shelf sediments on the submerged continental margin covered with the miogeocline may represent a significant source of hydrocarbons. Deltas covering the ancient shelf may be equally significant for shelf sediments. The deltaic sediments are enriched with $C_{org}$ transported by rivers. The delta of Niger is one of the examples. Other deltas on the Atlantic coast are also prospective.

3. Basins on active continents and plate margins. We have already discussed basins on divergent margins, rifts and "pull-apart basins". Their marginal position is secondary. Basins on convergent margins, in the compression zone are associated with collision zones of the type continent-continent (A-subduction) or ocean-continent (B-subduction). There are also basins associated with transform faults.

Basins in the zone of A-subduction of two continental plates are represented by aulacogens, foredeeps, and intermontane depressions.

a) Aulacogen is — in accordance with the original N. S. ŠATSKIJ'S (1955) definition — a graben, either partly overthrust orogen or extending into a

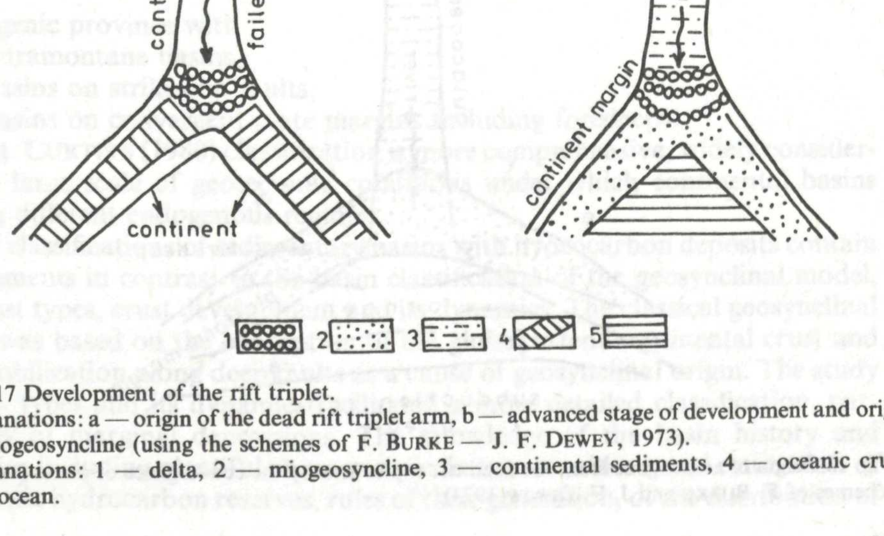



Fig. 17 Development of the rift triplet.
Explanations: a — origin of the dead rift triplet arm, b — advanced stage of development and origin of miogeosyncline (using the schemes of F. BURKE — J. F. DEWEY, 1973).
Explanations: 1 — delta, 2 — miogeosyncline, 3 — continental sediments, 4 — oceanic crust, 5 — ocean.

foredeep (Fig. 18). Thus it differs from the "pseudoaulacogen". Its junction with orogen related to plate tectonic convergent margin is shown in Fig. 18. According to K. C. A. BURKE and J. F. DEWEY (1973) aulacogen represents a miscarried attempt for the rift opening. It forms as a basin mainly at the termination of the collision stage or in the orogenic stage (uplift of mountain range). In this concept, based on BURKE and DEWEY'S (1973) definition, aulacogens cannot be expected to occur on the divergent (Atlantic) continental margin. Sediments of the early riftogenic stage ripen in the orogenic stage and become a reservoir of hydrocarbons generated in the foredeep development stage. In this stage the material is transported from orogen, i. e. in a reversed direction when compared to pseudoaulacogen. So the parts of aulacogen covered with the foredeep and continental margins covered with the foredeep or with nappes, are most prospective. This model may evidently be applied on the zone of the contact between the West Carpathians and the Bohemian Massif (cf. for example F. PÍCHA, l. c.).

b) Foredeeps may in the initial stage represent miogeosynclines with a shelf basement. If the passive continental margin of the Atlantic type preserves, no folding takes place. If the subduction zone migrates from the ocean to the continent or if the continent-continent collision takes place, the folding follows. From the dynamic view the second case is less favourable for the preservation of hydrocarbon deposits. It is, however, favourable in relation to B-subduction with respect to a lower heat flow.

c) Intermontane basins represent the final stage of foredeep basins and their disintegration into postorogenic depressions. The basins of this type are particularly prospective for hydrocarbon occurrences because of their low heat flow, rapid subsidence and closed ground water circulation. The Vienna Basin may be ranged among these basins.

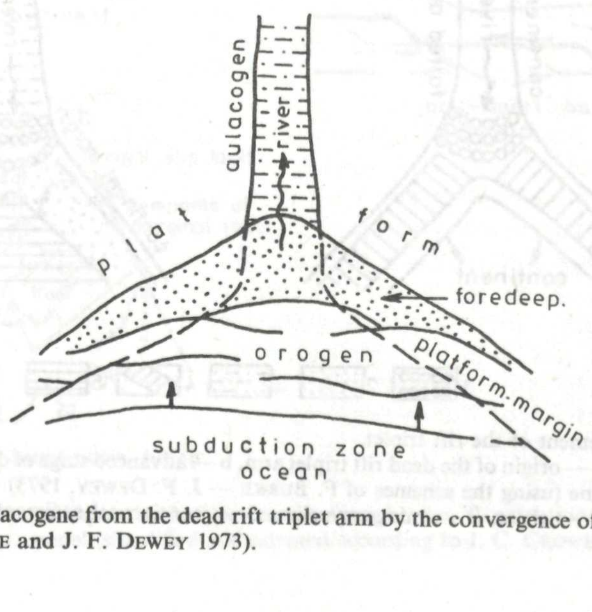



Fig. 18 Origin of aulacogene from the dead rift triplet arm by the convergence of plates (using the schemes of F. BURKE and J. F. DEWEY 1973).

Basins associated with B-subduction are mostly unfavourable for hydrocarbon deposits and have no analogues in our country. A particular type of basins is represented by basins at the Pacific continental margin. They were ensimatic in the early stage and then became ensialic.

4. Basins on strike-slips along convergent margins (in case of divergent margins it is the pull-apart type) display a character similar to those in B-subduction zones. Their origin is also associated with B-subduction. Examples: partial basins Los Angeles, Ventura, Maracaibo, a. o.

According to D. M. CURTIS (1980) riftoid grabens, aulacogens, pseudoaulacogens are optimal for hydrocarbon deposits because of the limited water circulation and a lower heat flow. Intensive subsidence prevents destruction of sediments. Intermontane basins, miogeoclines on passive plate margins, pull-apart basins and some foredeeps are also favourable.

1.9.3.2.2.2. Basin classification after A. Perrodon

A. PERRODON (1980) applied D. M. CURTIS'S (1980) classification in a simplified modification. He distinguishes the following basins according to their position in three principal geostructural provinces:

1. Platform province with
   a) stable-platform basins,
   b) unstable-platform basins,
   c) basins of carbonate complexes including reefs.
2. Graben province with
   a) rift basins,
   b) passive continental (plate) margin basins,
   c) deltas on continental margins,
3. Orogenic province with
   a) intramontane basins,
   b) basins on strike-slip faults,
   c) basins on convergent plate margins including foredeeps.

D. M. CURTIS'S (1980) classification is more comprehensive, widely considering the large scale of geotectonic conditions under which continental basins form in different endogenous regimes.

New classifications of sedimentary basins with hydrocarbon deposits contain new elements in contrast to the basin classification of the geosynclinal model, like crust types, crust development and its dynamics. The classical geosynclinal model was based on the conception of the global extent continental crust and its remobilization along deep faults as a cause of geosynclinal origin. The study of crust types and its dynamics facilitated a more detailed classification, particularly of marginal depressions. The knowledge of the basin history and dynamics including plentiful characteristic features enables a real estimation of prognostic hydrocarbon reserves, rules of their generation, of the distribution of traps, and orientation of the strategy of deep basin exploration. The seemingly theoretical problems of classification are actually extremely significant for economic geology.

1.9.3.2.2.3. Geologic-economical basin classification after H. D. KLEMME

H. D. KLEMME (1980) presented a classification including principal economic-geologic parameters for prospection and exploration of hydrocarbon deposits. The classification is based on the position of basins on continents and on the character of sedimentary filling decising for the hydrocarbon potential.

H. D. KLEMME (1980) distinguished eight types of basins in intra- and extracontinental positions.

a) Intracontinental basins: simple basins (type 1), complex basins (type 2) with hiatuses or unconformabilities. Subtype 2 A is particularly intricate when

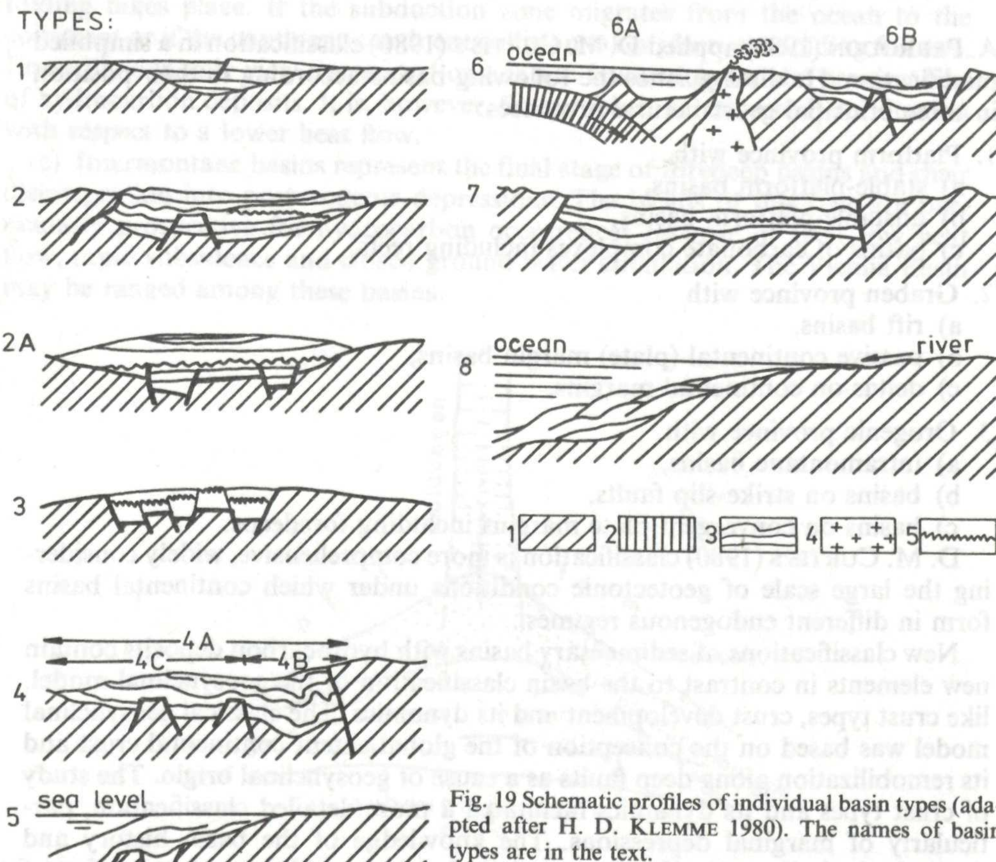



Fig. 19 Schematic profiles of individual basin types (adapted after H. D. KLEMME 1980). The names of basin types are in the text.
Explanations: 1 — continental crust, 2 — oceanic crust, 3 — sediments, 4 — volcanic island arc, 5 — discordance.

the lower stage has a graben structure (more grabens or riftoid structures) and the upper stage shows a simple — platform structure (Fig. 19).

Rifts — type 3 are analogous to subtype 2 A but the width of intracontinental rifts is four times smaller than with basins 2 A.

b) Extracontinental basins, marginal basins are most frequent. One of the types is type 4 represented by downwarp basins. These are on transition from continent to marginal seas. They are closed without direct connection to the ocean (subtype 4 A) or resemble a trough (4 B), or they are open towards the ocean.

Pull-apart basins (type 5) are parallel or transversal to the passive continental margin. The early stage of their development may have the character of a satellite rift.

Associated subduction — and island arc basins belong to type 6: including subtype 6 A — forearc basins and 6 B — backarc basins.

Inner basins (type 7) form in the course of orogenesis after the end of subduction. They are actually intermontane basins.

Deltaic sediments deposited on a continental slope on a margin of the Atlantic type. Types 3, 5 and 8 are associated with divergent plate margins, types 6 and 7 — with convergent margins.

In the series from type 1 to 8 the age of basins (time of their origin) and the structural influence of the basement upon the sedimentary filling decrease, the amount of clastic sediments increases and the content of carbonate rocks decreases (except in type 8 — if the delta is on a shelf). The size of basins decreases, the extent of offshore plain increases, and the amount of abyssal sediments decreases.

Basins of types 1 and 2 comprise large traps with immense hydrocarbon accumulations. Basins of other types are little prospective as for larger deposits. Parent rocks in types 1 — 4 B show a higher maturity, those in types 4 C — 8 show a low maturity. In basins with persistent subsidence like types 3, 6, 7 also diagenesis continues and results in changes of hydrocarbon migration and water circulation. Most important data about basin hydrocarbon potential factors are in Table 3.

Representative basin types (after H. D. KLEMM 1980):

Type 1: midcontinents of North and South America, Aquitanian Basin in France.

Type 2: Western and Eastern Siberia, western part of U.S.A. and of South America, Australia.

Type 3: Northern Ice Sea, coast of Norway, Nigeria, Dnepr—Don aulacogen.

Type 4: Gulf of Mexico, Northern Alaska, Middle East, Caspian Sea, West India, Indonesia, Carpathian basins (type 4 A).

Type 5: coast of West Australia, coast of West Africa, eastern coasts of South and North America.

Type 6: smaller basins on Pacific coasts of U.S.A. and of Indonesia.

Type 7: New Zealand, Indonesia, Alaska, Northern part of South America, Mediterranean region, Iran, Pannonian Basin.

Type 8: Japanese Islands, northern coast of East Siberia, West-African coast, Central-American basins, Mediterranean Sea.

Some data on deposits in basins of individual types by H. D. KLEMME (1980) Table 3

| Basin type | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
|---|---|---|---|---|---|---|---|---|
| % present world reserves | 1,5 | 25 | 10 | 48 | 0,5 | 7,5 | 2,5 | 5 |
| % world basin areal/% productive | 18/3,5 | 27/20 | 5,5/2,4 | 18/12 | 18/2 | 7,2 | 3,5/1 | 2,5/1 |
| % reservoir clastic/carbonate | 60/40 | 75/25 | 60/40 | 4B: 50/50<br>35/65 | 70/30 | 90/10 | 90/10 | 100/? |
| large traps<br>large basins | | +<br>+ | + | +<br>+ | | + | + | |
| depth of deposits | shallow | shallow<br>medium | shallow<br>medium | average<br>medium | deep<br>medium | deep<br>(70 %) | deep<br>(70 %) | shallow |
| secondary traps:<br>cf-compression folds<br>l-lithostratigraphic<br>c-combination | | cf, l, c | c, l | cf, c, l | c | c(l) | c | l |
| genesis:<br>c-compression<br>e-extension<br>w-wrench | e | l, c | e, w | c(4B) | e | c, w | c, w, e | gravit. |
| crust type | | continental | | | transitional | | oceanic | continental |

Gigantic deposits (type 2) represent approximately 1 percent of world deposits. With respect to newly discovered deposits most prospective are types 4 (A, C), 3, and partly some basins of type 6.

Classification in terms of plate tectonics is mostly comparable to types of basins in the geosynclinal model (see Table 4). The comparison shows for example that the more detailed division of marginal depressions was included in classifications in terms of plate tectonics.

### 1.9.3.2.4. Basin classification after D. R. KINGSTON, C. P. DISHROOM and P. A. WILLIAMS

In contrast to the above mentioned simpler classification for the purpose of economic geology, D. R. KINGSTON et al. (1983a) elaborated a global basin classification respecting general and specific features of mutual basin correlation. The classification (like that of A. D. MIALL 1984) also respects the history of basins and variableness of types.

Comparison of basin types to hydrocarbon deposits according to geosynclinal and plate tectonic models

Table 4

| Model | Geosynclinal | Plate tectonic | | | | |
|---|---|---|---|---|---|---|
| regime | division by V. V. BELOUSOV | classification by D. M. CURTIS | classification H. D. KLEMME | D. R. KINGSTON et al. | | A. D. MIALL |
| | | basin | basin | type | type of cycles/basins | basin |
| geosynclinal | flysch intrageosyncline | interarc | forearc | 6 A | TA | forearc, interarc basins |
| | flysch miogeosyncline | backarc | backarc | 6 B | TA—LL (OS, OSLL) | backarc basins, trenches |
| | marginal depression | on strike — slip faults | downwarp | 4 A | LL, IF (OSLL) | ocean margin basins (rim basins), transform. + transcurr. fault basins |
| | inner depression | intermontane | inner | 7 | IS, IF | intra + suture basins suture zone transcurr. fault basins |
| platform | syneclise<br>parageosyncline<br>aulacogen | intracratonic<br>pseudoaulacogen<br>aulacogen | intracontinent.<br>complex basin | 1,2<br>2 A | IS<br>ISIF | cratonic basin<br>aulacogens |
| | marginal depression (Atlantic type) | "pull-apart"<br>miogeocline | "pull-apart" | 5 | LL | pull-apart |
| | | | deltaic type | 8 | IFMS | ocean margin basins |
| rift | rift | continental rift<br>rift with failed arm | rift | 3 | IF | rifts<br>failed rifts |
| | foredeep | foredeep | | 6 A<br>(4 B) | LL | foreland and peripheral basins |
| | intermontane depression | intermontane | inner | 7 | IS (LL) | intrasuture basins |

Remark: Comparison of some types is only approximate, only some features are correlated. Basin types are explained in text.

Reconstruction of basin history is based on three principal parameters:
a) depositional sequences,
b) basin-forming tectonics,
c) basin-modifying tectonics.

Basins are simple — with one or two tectonic — sedimentary cycles, and complex basins with a complicated development with several different cycles and events — like the so-called polyhistorical basins.

In contrast to D. M. CURTIS'S (l. c.) and H. D. KLEMME'S (l. c.) classifications, D. R. KINGSTON et al. (l. c.) emphasizes the analysis of sedimentary cycles registered in the depositional sequence and sedimentary filling of overy basin. The classification is based upon the study of depositional sequences of about 600 world basins with different hydrocarbon productivity. Kingston distinguished eight principal types of cycles (basin types) applicable in basin classifications.

a) Depositional sequences

Determination of depositional cycles and their stages is the first principal element of D. R. KINGSTON'S et al. (1983a) classification. Each cycle consists of three stages: the initial stage with continental sedimentation passing into sedimentation of the intermediary type on the coast. As far as marine intercalations in the sequence do not exceed 50 percent, the sequence may be ranged to the first stage. The further second stage is a marine stage with typical marine and abyssal oceanic facies. Non-marine final stage with more than 50 percent of non-marine sediments is the third stage. The cycles are separated by unconformities. Unconformities frequently border individual stages within a cycle.

Determination of basin-forming tectonics and of wiping of paleogeographic contact between a continent and an ocean is based on delineation of cycles.

b) Basin-forming tectonics

Three identification criteria must be determined for the analysis of the foundation of a basin, namely:

1. Type of the crust beneath the basin — only two principal types — continental and oceanic — were accepted for classification.

2. Type of past plate movements. Only divergent and convergent types are distinguished. In contrast to A. D. MIALL (l. c.), the transform type is not considered. This type only represents one form of both principal plate dynamics. Convergent movements along faults may invade intraplate regions.

3. Position of sedimentary — tectonic cycles of a basin on a plate (interior, marginal), character of primary movements during the formation of a basin — sagging, normal or wrench faults.

Combination of the three criteria is the basis of classification. With respect to prospection for hydrocarbons economically most significant are basins on continental crust, inside and on margins of continents, controlled by faults.

b 1. Divergent cycle types

b 1a) Interior sag cycles (basins/IS) may also be on ancient margins — now in intraplate position. Basin axes are oriented diagonally to recent margins. In basin structure dominant is non-fault pattern i. e. a megasynclinal (syneclise) downwarp. Basins are characterized by oval to circular shape, and small sediment thickness dominates.

b 1b) Interior fracture cycles (basins/IF) may be joined with paleosutures and/or lineaments of ancient divergent margins — now structures in intraplate position. In their structure horsts and grabens dominate. They also form by tension at the beginning of divergent movements. The IF type may change into MS type.

b 1c) Margin sag cycles (basins/MS). Basin axes are parallel to the Atlantic type continental margins. Sediments may pass from continental to oceanic crust. Polyhistoric type basins may for instance have the lower stage of the IF type and the upper stage covers the faulted lower stage in the form of a sedimentary sequence on a continental slope or as a delta.

Cycles may consist of the 1$^{st}$ stage of continental sedimentation or of the 2$^{nd}$ stage of marine sedimentation (shelf type, evaporites, a. o.) at the subsiding continental margin. During the sea regression sediments of the 3$^{rd}$ stage of the cycle consisting of continental sediments develop.

On some faults, intruding basalts may indicate the commencement of spreading.

On oceanic crust, basins form under conditions of marine sedimentation in the form of oceanic fracture basins (OF) or as oceanic sags (OS). Oceanic wrench basins (OSLL) have a marginal position. They are less significant with respect to hydrocarbon production.

b 2. Convergent cycle types

This category comprises basins situated on margins or close to margins of two convergent plates. Most basins show tensional features.

b 2a) Wrench or shear cycles basins (LL)

The basins result from strike-slip faults and are denoted as wrench strike-slip or shear basins. Those on continental crust are prospective for hydrocarbons. They result from reverse movement of two blocks or microplates over a strike--slip fault. Their development is controlled by strike-slip rate and by folding along the basin margin if the basin is in a foredeep position. In the initial stage of lateral movements of blocks the basins show the character of tension grabens with continental sedimentation. Their character is the same as that of pull-apart basins. Oil accumulations are along block margins inside basins. Continuous subsidence on the side oriented towards the sea is associated with marine transgression and marine sedimentation. In the part close to the forming fold zone, oil accumulates on embryonal folds above the wrench zone. Formation of the fold zone destroys sediments and causes hydrocarbon migration out of the folding zone.

b 2b) Trenches and trench associated basins (T and TA). Basins of TA type are associated with the zone between a trench and a volcanic island arc. TA basins are only little prospective for hydrocarbons. Trenches (T) are on oceanic crust or above converging plates. Ancient trenches were folded and faded out. Trenches are not prospective for hydrocarbons (small thickness of only deep-sea sediments).

b 2c) Oceanic sags — wrench couplet cycles (basins/OSLL). The group comprises small oceanic basins and marginal seas on oceanic or transitional crust, in the zone of plate convergence. Tension on an overriding plate causes crustal stretching. Many geologists associate the origin of these basins with strike-slip displacement on wrench couplet.

c) Basin-modifying tectonics

Basins forming under some of above mentioned dynamic conditions are further modified by structural events. Authors of the classification model distinguish three types of basin-modifying tectonics:

1. Episodic wrenches (L),
2. adjacent fold belts (FB),
3. folded basins (FB 3).

Folded basins (FB 3) are incorporated in the fold belt and are practically unprospective for hydrocarbons.

The formulae placed at basin types for simple expression of basin history and for evaluation of their economic-geological prospectiveness with the use of a computer. Single formulae may further be divided, for example according to various effects of basin-modifying tectonics. Wrenches along a basin L may only affect the sourroundings of the basin without influencing its sedimentary filling (La, Lb — very weak and weak influence) or they affect the basin and divide it into blocks with variable portions of hydrocarbons (Lc — stronger influence, Lf — very strong influence). Folding (FB) only has marginal effects without deforming the sedimentary filling (FBa) or may strongly affect the basin and change its structure completely (FBf, FB3).

Basing on deep drilling and geophysical investigations several complete (with stages 1—3) and incomplete (stage 1, or 1—2, 2—3) cycles may be identified in the basin development, as well as various basin types (IS, MS or OSLL — LL — MS — IS). For instance the formula MS-Lb-Le-Lf-FBc-FBf (FB3) denotes the development from the marginal basin through its entering the folding up to complete folding preceded by many strike-slip fault movements.

### 1.9.3.3 Hydrocarbon potential of basin types

Evaluation of basin types also offered information on productivity of individual cycle (basin) types (D. R. KINGSTON et al. 1983b).

Basin types:

IS — mostly Paleozoic basins lacking structural traps. Owing to no tectonic disturbance, hydrocarbons accumulated in extensive stratigraphic traps restricted to large basins, i.e. on extensive platforms. IS cycles in sections of polyhistorical basins are significant for economic geology. Forty-nine percent of investigated IS basins are interesting economically.

IF — most productive basins like Sirte in Libya, West Siberia or Gulf of Suez. According to the authors quoted also smaller basins like the Vienna Basin are significant. Sediments of IS basins with a less intense subsidence are not mature enough for oil production. Thirty-five percent of world basins studied contain economically significant hydrocarbon reserves.

MS — owing to intense subsidence sediments of the basins are extremely thick and hardly reachable in the depth of 9000 m. The authors' analysis shows that deltaic sediments, regarded by some geologists as prospective, are unprospective. Among sixty studied large recent deltaic accumulations only four are on the limit of economical rentableness of hydrocarbon exploitation. They are the deltas of Mississippi, Niger, Mahakam and Mackenzie rivers. Ancient deltas, later tectonic-structured and covered with younger cycles, are more prospective. Only twenty percent of MS cycles are productive and exploration means a risk in most MS basins.

Margin-sag — interior-sag cycles (basins/MSIS) inside plates are highly prospective. The basins were on ancient margins and most sediments produced hydrocarbons. Shelf carbonates contained large hydrocarbon deposits, for example in the Gulf of Persia, in Texas, West Canada. Approximately 50 percent of MSIS cycles produce commercial hydrocarbons.

The LL-type had a quick geologic development. Younger basins are immature, the older ones are overstructured. Their exploration is a risk. The lack of reservoir layers and a high heat flow are unfavourable factors. The basins are small. Folding resulted in numerous structural traps. About 47 percent of studied LL cycles-basins contain commercial hydrocarbon reserves.

Other basin types are less significant as for hydrocarbon accumulations. T and TA types cycles represent limited productive levels in smaller deposits like the Gulf of Alaska, the Abukuma Basin in Japan and Mentawe in Indonesia. In OS basins the deltaic sediments extending to oceanic crust are partly significant. The Niger delta is a prospective basin on oceanic crust.

Basins with the maximal hydrocarbon potential belong among polyhistorical basins where structuring was followed by renewed subsidence and by dynamic regime changes in structures. Polyhistorical basins terminated with the IS cycle are most favourable for hydrocarbon preservation. Owing to sediment load the final interior sag accelerates ripening of sediments and regional hydrocarbon migration.

#### 1.9.3.4 Role of basin classification in prospection for hydrocarbons and coal

A. W. Bally and S. Snelson (1980) do not consider heterogeneity of crust types inside continental crust and exclude so the main cause of long, repeated subsidence, namely heredity of mobility caused by simatic crust. They only associate the heredity principle with ancient or recent active plate margins, and the intraplate mobility is less considered. The intraplate mobility need not only be a consequence of faults, sutures, megasutures but also of differences in crust types. The above authors do not consider oceanic or intermediate crust restites owing to unaccomplished (embryonic) spreading or incomplete ocean closing (e. g. oceanic crust detached or obducted due to the splitting of lithosphere). A historical approach to the development of a lithospheric segment with a basin is a positive feature of this and other classifications. This approach facilitates interpretation of conditions and dynamics of the genesis and migration of hydrocarbons in productive basins.

The above comments may also concern A. D. MIALL'S (1984) basin classification. Also he neglegted physical-chemical crust changes, and paid little attention to the lithosphere/asthenosphere interaction or regarded it as a consequence of plate mobility — like in plate models. Rejection of terms "intermontane" and "intramontane" basin may cause errors in paleogeographic reconstructions, in evaluation of economic-geological potential of basins and in determination of prospective areas. Overestimation of strike-slips and application of extreme dynamics may cause erroneous evaluation of conditions of the genesis and migration of hydrocarbons, and even exclusion of possible existence of hydrocarbon-bearing layers. It should also be noticed that discussions and frequent speculative interpretations of the character of ancient plate margins, continental margins and of the existence of hypothetic closed oceans are very difficult.

Basins included in D. R. KINGSTON'S et al. (1983a) classification contain hydrocarbon deposits, coal-, salt deposits and deposits of other nonmetalliferous raw materials. Association of individual raw materials to cycle stages is evident. Moreover, sediments of single cycle stages either have productive properties or in the course of the basin development they become lithologic and structural traps. Historical analysis of basin development enables evaluation of conditions of the genesis, migration, and accumulation of hydrocarbons in traps. So it is important to study the relations of the present basin type to its geologic past.

D. R. KINGSTON'S et al. (1983a) classification is simpler than the general basin classification of A. W. BALLY and S. SNELSON (1980), and of A. D. MIALL (1984). D. R. KINGSTON et al. (l. c.) have a more exact approach to the history of basins with a complicated development. This classification enables incorporation of all world basins into one simple scheme in case we generalize their specific features and exclude transitional crust types. It is not right to omit rifts from basin types because they represent a significant basin category with specific characters of deep structure and lithospheric dynamics. Little attention is also paid to intermontane basins and to Mediterranean basin types where mantle diapirs cannot be associated with strike-slip faults. With respect to dynamics it is not right to omit normal faults among tectonic modifiers. In fact, normal faults are better evidenced than wrenches, especially than those covered with younger sediments. So the formulae written for basin-modifying tectonics are nor complete neither objective. In addition to symbol L (lateral for wrench) also symbol V (vertical) should be applied (for normal faults and block-fault movements).

## 1.10. Summary of model ideas

Individual factors of basin origin and development alone are not sufficient for a complete explanation of the genesis of basins. No factor is active isolatedly.

It is generally evident that the origin of basins has a deeper subcrustal cause in asthenosphere, with possible exception of small basins with subsidence ranging to several hundreds of metres.

We do not know the substance of deep physical-chemical changes. We may suppose — basing on changes in heat flow in the basin areas (H. D. KLEMME 1978) and their surroundings — that deep temperature differences and rock volume changes in asthenosphere cause further changes in lithosphere due to tectonic processes.

So the basin origin is first of all stimulated by a deep endogenous cause, namely increasing mantle temperature causing decrease in viscosity and melting of rocks. Sedimentary filling gradually presses the basement and the formation of a basin is also supported by increasing geostatic pressure. Submergence of crust into mantle is most likely followed by phase changes in deep crustal rocks, by increasing rock density, and further subsidence with possible acceleration. If subsidence rate increases in time, phase changes may be admitted — as far as subsidence is not associated with volcanism which might cause downsagging of crust into emptied magmatic chambers.

Volcanism and volcanic intrusions cause the same crust density increase. In accordance with other authors, I regard deep changes in isostatic equilibrium and a deeper density differentiation with progressively increasing rock density in belts or areas as decising for basin formation. Now I want to say that treatment of the problem concerning kinematic basin history with respect to their economic-geological productiveness is only at its beginning. The problem is a wide one including for example, causes of larger units subsidence (median massifs, platforms), but it is not possible to treat it so thoroughly. I shall discuss some general phenomena possibly affecting the basin development in relation to a broader area.

### 1.10.1. Subsidence process

Stages of extensive transgressions and regressions, associated with emergence of continents, thalassogenous and cratogenous stages (R. A. SONDER 1956) are well known in geology, and inspired a concept of cyclical alternation of these stages (S. VON BUBNOFF 1954). Most interpretations were based on fluctuation of the world ocean level (E. SUESS in his "Antlitz der Erde"). Data on variable oceanic and continental crust endogenous regime and new plate tectonic information are indicative of other causes of subsidence and uplift of lithosphere.

"Ocean closing and opening" facilitate interpretation of changes in water mass migration, their deficiency or excess.

P. A. RONA (1972 in A. G. FISHER 1978) presumes variable oceanic crust thickening associated with different rift opening rates and changes in the extent of rift doming. In the time of rapid crust thickening extensive rift elevations expelled water masses out upon continents subsiding together with the surrounding oceanic crust. The solution of this problem will require further stratigraphic and paleogeographic correlation studies — now in the initial stage (e. g. E. JAEGER 1977 et al.). A. G. FISCHER (1978) studied the relation between the history of the Atlantic Ocean and adjacent basins and revealed a considerable part of differentiated basin development controlled by their tectonic position. Platform units were least mobile (less than 1 km subsidence per 300 m. y.; orogenic units subsided 5—9 km per 200 m. y.). Fischer ascribes about 2/3 of subsidence value to the load of sediments and water upon the basement after deposition of 2—4 km of sediments. Then the subsidence rate increased owing to the secondary covering by additional loading. Commencing subsidence was, however, controlled by other factors. Analogous characters were revealed in the Michigan Basin on the mobile part of the platform. Most geologists and geophysicists regard the thermal flow uprise as the cause of basin foundation. Significant are also historical causes of changes in subsidence rate, the relationship between subsidence acceleration and folding phases in the adjacent (source) area (for example a couple tectogen-foredeep) has been known since long. On the other hand, orogenic uplifts also affect depositional areas and either interrupt or accelerate deposition, there. The relationship was studied by S. VON BUBNOFF (1954), V. V. BELOUSOV (1954, 1962) a. o.

Continental rifts in the area of the U.S.A. show a higher subsidence rate in the time of uplift of the surroundings — at the elevation of the rift-bearing dome (A. G. FISCHER 1978). Examples of other basins of different types (foredeeps, inner molasse basins) show a differentiated development slightly affected by regional orogeny. Intense subsidence of many basins (not of eugeosynclinal character) preceded tectogenesis.

Analysis of sediment thickness in basins shows (F. L. SCHWAB 1976) that most mobile basins are on plate — or continent margins and the slowest subsidence is inside the continents. Further examples are in Chapter 9.3.2. Marginal basins are most prospective as for economic geology, although most frequent giant hydrocarbon deposits are in intracontinental basins.

### 1.10.2. Formation of basins above mantle diapir — modified model

A detailed model of basin formation above mantle diapir was elaborated by R. W. VAN BEMMELEN (1972). His concept was further treated by geologists studying genesis of rifts, Mediterranean basins, the Great Basin a. o. (references are partly quoted in single chapters of this volume).

The following model represents a modification of the existing models and a trial of a general model for non-rift basins and/or for basins with an early rift stage, late altered to isometric basins. The model is based upon these phenomena:

a) crust type,
b) mechanism of formation and re-formation of diapir,
c) presumable subsidence of lithosphere into less viscous asthenosphere.

In the zone of active hot diapir also basification s. l. and collapses during the diapir cooling are admitted.

The diapir activity is supported by primarily thin simatic crust preserved in sialic crust. The latter is tectogenic-orogenically active during the diapir uplift. Heterogeneous variably thick crust causes differentiation in the mantle, variable restite ratio after segregation of magma, and different tension in the mantle.

Elliptical, isometric or triangular basins of intermontane types are rimmed with orogenetic crust, usually twice so thick as simatic crust above diapir. Beneath the thin crust is permanent mantle elevation causing disequilibrate state in pressure and tension in the mantle. It is intensified by deep faults forming on the contact between thin and thick crust. During the formation of hot diapir some faults cause its uprise and divide the cold rigid mantle and deep asthenosphere from hot mantle and asthenosphere elevations. This is illustrated by the example of the arched "Benioff zones" in the Tyrrhenian and Aegean Seas (F. ČECH — J. ZEMAN 1985, — Fig. 20).

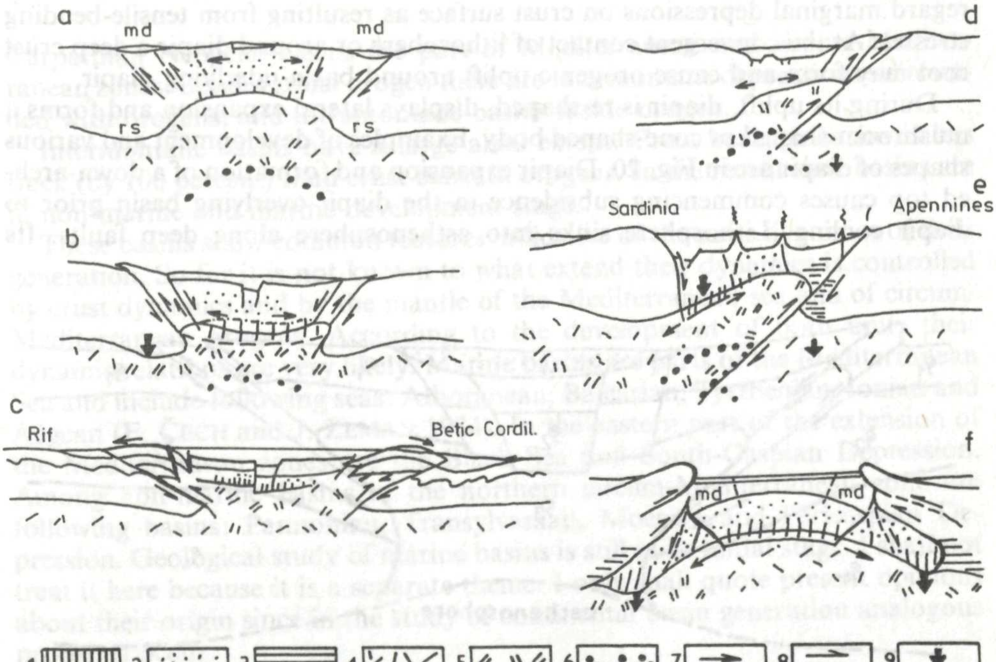



Fig. 20 Models of mantle diapirs and their dynamics.
a — stress in symmetric diapir (md-marginal depression, rs-rim syncline), b — spreading of diapir peak into fan-like shape, c — dynamic model of lithosphere of the Alboranean sea (modified according to M. LEMOINE 1978), d — stress in asymmetric diapir, e — dynamic model of lithosphere of Tyrrhenian sea, f — stress and dynamics in linear diapir (Moesian-Pontic depression).
Explanations: 1 — basification (closer relations between physical properties and mantle), 2 — simatic crust, 3 — anatexis, 4 — cold mantle, 5 — hot mantle, 6 — surface of asthenosphere, 7 — tension stress, 8 — shear stress, 9 — subsidence.

Density disequilibrium arises in lithosphere and asthenosphere (Fig. 21). Hot diapir and basic differentiates lighter than mantle migrate from the density inversion to the higher lithosphere level by the mechanism of melting of the mantle elevation. Melts uprise and non-melted restites sink into asthenosphere to accumulate at its base. In lithosphere diapir stabilizes in the lower crust in a density-equivalent environment.

Formation of diapir is associated with generation of rim synclines (H. RAMBERG 1967). They result from the runoff of masses into diapir. Above the synclines the crust is downsagging and marginal depressions form on its surface. In the depression, sedimentation commences earlier than in the basin above diapir (F. ČECH and J. ZEMAN 1985; see Table 5).

The diapir uprise may be associated with creep of ductile lithosphere or of overheated lower crust down the diapir slopes into rim synclines. Creeping may also occur in asthenosphere. Subsidence in a mantle rim syncline causes bending and downwarping of lithosphere margins around diapir. Thus local subductions modified by marginal faults form. I denote the subductions as passive and regard marginal depressions on crust surface as resulting from tensile-bending stresses. At the convergent contact of lithosphere or around diapir a deep crust root may form and cause orogenic uplift around basin-uderlying diapir.

During its uplift, diapir is re-shaped, displays lateral expansion and forms a mushroom-shaped or cone-shaped body. Examples of development and various shapes of diapir are in Fig. 20. Diapir expansion and formation of a down-arched top causes commencing subsidence in the diapir-overlying basin prior to diapir cooling. Lithosphere sinks into asthenosphere along deep faults. Its

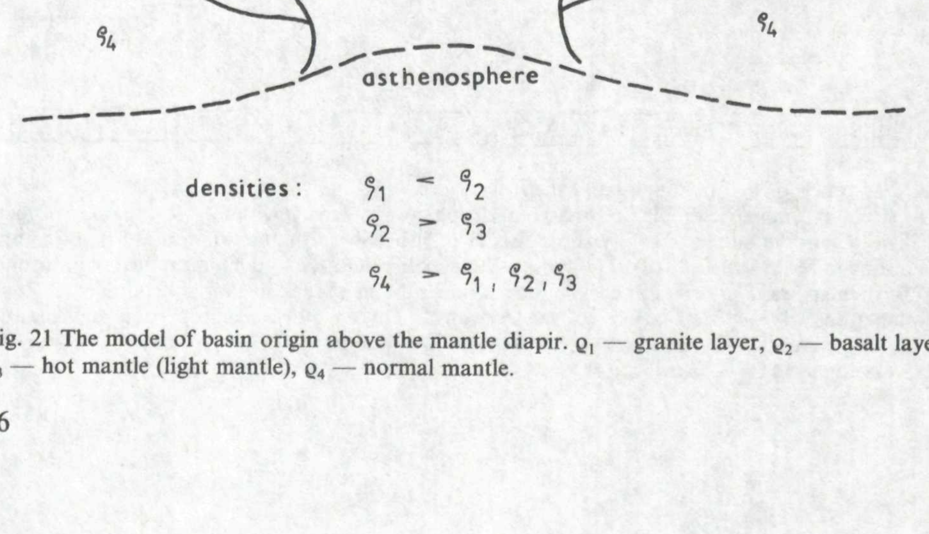



Fig. 21 The model of basin origin above the mantle diapir. $\varrho_1$ — granite layer, $\varrho_2$ — basalt layer, $\varrho_3$ — hot mantle (light mantle), $\varrho_4$ — normal mantle.

subsidence also supports emptying of magma chambers in volcanic-active diapirs. Shear stresses above diapir support formation of shear faults or rejuvenation of shear movements on pre-diapir faults. Tension stress above diapir supports formation of normal faults. Diapir expansion may be associated with formation of listric faults, i. e. under conditions different from those considered in D. P. Mc Kenzie's (1978).

Diapir cooling and collapse may cause acceleration of subsidence. When no mushroom-shaped diapir is formed, then domed diapirs can possibly be produced by diapir expansion and "diffusion" although it formerly was cone--shaped (diapirs beneath recent volcanic-active basins). Domed or mushroom--shaped diapir (Alborane Basin) represents a final irreversible mantle structure. It is also preserved after heat flow decrease (Viking depression, Black Sea).

## 2. Neogene Basins in Alpine mobile Europe

Carpathian Neogene basins are part of Mediterranean and circum-Mediterranean zones. Besides zonal orogen there are intermontane depressions surrounded with orogens, and intramontane basins inside orogen.

Intermontane basins have a large areal extent. Crust beneath basins is less thick (by 100 percent) than crust beneath orogens. In the recent time basins are in non-marine and marine development stage.

These basins show common features indicative of the same dynamics of their generation. So far it is not known to what extend their dynamics is controlled by crust dynamics and by the mantle of the Mediterranean sea and of circum-Mediterranean orogens. According to the development of both units their dynamic relations are very likely. Marine basins are parts of the Mediterranean Sea and include following seas: Alboranean, Balearian, Tyrrhenian, Ionian and Aegean (F. Čech and J. Zeman 1984). In the eastern part of the extension of the Mediterranean zones are the Black Sea and South-Caspian Depression. Among non-marine basins in the northern circum-Mediterranean zone are following basins: Pannonian, Transylvanian, Moesian and Pericaspian Depression. Geological study of marine basins is still in its initial stage. I shall not treat it here because it is a separate theme. I only shall quote present opinions about their origin since in the study of continental basin generation analogous problems occur.

### 2.1. Origin of Mediterranean Sea

Origin of the Mediterranean Depression is a problem that has since long been solved by relic- or new-formed oceanic crust models:

1. Depression-underlying crust is a relic of the original ocean (Paleotethys). In its development the crust did not reach the stage of geosynclinal alteration

Phanerozoic subsidence rate in main basins and in their marginal depressions (bed thickness in m)

| Formations | Pannonian Basin | Marginal depressions | | | | | |
|---|---|---|---|---|---|---|---|
| | | Danube | Transcarpathian | Dráva | Sáva | Mur | Styrian |
| Quaternary Pliocene | 2000–4000 | <500 | 0–500 | 300–2700 | 0<200 | 1000–2000 | 500–1000 |
| Miocene | 100–1000 | 2000–4500 | 1000–3000 | 2000–4000 | 3000–5000 | 2000–4000 | 1000–2000 |
| Paleogene | 0–2000 | 1000 | 1500 | ?300–1000* | 100–1000?* | | |
| Cretaceous | 400–2100 | | 100–150 | | | | |
| Jurassic | 350–4000 | | 300 | ? | 0 | see Pannonian | |
| Triassic | 850–2700 | | 100 | | | | |
| Permian | 800–3000 | >100** | >500** | >300* | 300* | | |
| Carboniferous | 0–1500 | >100?** | >500** | ~500* | ~500* | see Pannonian | |
| Devonian | >800 | >200** | ? | ? | ? | | |
| Silurian-Cambrian | 1000–3500 (to Ordovician) | ? | | | | | |
| Late Proterozoic | >1000 | | | | | | |

Explanations: * — estimation according to Dinaride region; ** — estimation; + — maxima from marginal depressions.

Data: on Inner–Carpathian basins, Moesian Platform — by V. V. SEMENOVICH — JU. G. NAMESTNIKOV

into continental crust (V. A. PEIVE 1969). The model of relic oceanic crust was in several variations already presented by E. SUESS and other authors.

2. The relic model also comprises L. GLANGEAUD'S (1970) idea of the basalt crust in the sea basement being a rest of distorted continental (granite) crust, its drift and underthrust beneath young orogens. A. FAURE and G. CHOUBERT'S (1975) opinion is similar: they regard oceanic crust as identic with exposed basalt layer from which the sialic layer was detached and transported beneath orogens.

3. Formation of new oceanic crust is presumed in plate tectonic models of D. P. MC KENZIE (1972), J. F. DEWEY et al. (1973) a. o. They presume partial spreading from the Mediterranean Ridge with variable plate movement azimuth and splitting of divergent plate margins. J. F. DEWEY et al. (1973), K. BIRKENMAYER (1976), A. TOLLMANN (1978) presume (without geologic evidence) even fourfold opening and closing of Tethys and relics of new-formed oceanic crust among continental crust fragments. A. TOLLMANN (l. c.) imagined complicated splitting of collision plates and defined specific Mediterranean plate convergence. Mesozoic new-formed oceanic crust is defined by ophiolites, for example according to A. T. ASLANYAN and M. A. SATIAN (1984).

Table 5

| Transylvanian Basin | Moesian depression | Marginal depressions | | | | | |
|---|---|---|---|---|---|---|---|
| | | Varna | Getic | Beilesht | Lom | Dolná Kamchiya | Forebalkan |
| 150–800 | 1300–1600 | > 30 | 0–500 | 50–300 | 0 > 20 | 0 > 10 | |
| 3200–6400+ | 500–1000 | 350 | 2000–3000 | 500 | 900 | 0–370 | uplifts |
| 700–1470 | 1700–4000 | 30–675 | 3600–3750 | 100 | 1000 | 2100–3000 | denudation |
| 1700–2400+ | 1500–2500 | 320–1600 | v. 2000 | v. 3800 | 4300 | 1600–2500 | s. 2800 |
| 300–500 | 1000–2000 | 70–380 | 250–300 | ? | 250–800 | 1500 | 4100 |
| 600–700 | 1800–2000 | 2500 | ? | | 1600 | 3000 | 3800 |
| > 100 | 500–1000 | | | | ? | Permian-Triassic | 2000 |
| ? | 1000 | 2000–2600 | | | | 1000? | |
| | 1500–2000 | | | | | 4000? | |
| | | 4000 | | | | | |
| | ~ 5000 | ~ 5000 | | | | ~ 5000? | |

(1981) higher values of Miocene–Pliocene sediments thickness in Pannonian Basin — by D. VASS (1979) on Transylvanian Basin — by J. G. CIUPAGEA et al. (1970), on Moesian Depression — by T. V. STRELCOVA — J. A. SAPUNDZIEVA (1970) in V. V. SEMENOVICH — JU. G. NAMESTNIKOV (1981).

4. Formation of basins in West- and East Mediterranean areas during collision of European and African plates when the Adriatic prominence left empty areas W and E of the promontory (V. D. TCHEKHOVITCH 1984).

5. Oceanization of continental crust by a specific Mediterranean type of process (of metasomatic character) and under the influence of mantle asthenolith and diapirs was for the first time used by R. W. VAN BEMMELEN (1972) to explain the origin of the Mediterranean Depression.

6. V. V. BELOUSOV (1982) explains the origin of the Mediterranean depression by crust basification associated with intrusions and effusions of basic and ultrabasic rocks when in some regions the continental crust fully changed into oceanic. Belousov also explains the existence of partial Benioff zones by differences in temperature and viscosity of mantle.

7. M. V. MURATOV (1975) explains the origin of marine depressions by increasing upper mantle density and by metamorphosis of original continental crust to a more dense crust. E. V. ARTYUSHKOV and M. A. BAER (1983) take in consideration basification based on lower-crust eclogitization, and gradual alteration of upper crust into mantle rocks.

8. Specific development and structure of the Mediterranean Sea and its surroundings, stressed by R. W. VAN BEMMELEN (1972) are also emphasized by others. According to I. V. ARKHIPOV (1984) there was a heterogeneous system of partial depressions with oceanic crust in the Mesozoic (Mesotethys). Basins were separated by continental crust belts and blocks and by shallow-water sedimentation. According to Arkhipov it is not possible to compare the Mediterranean Sea to oceans and interpret its origin in the same way. The "oceanization" process represented destruction of continental crust blocks by differentiated basification. M. A. BAER (1984) presumes existence of basins on continental and oceanic crust, and that rapid subsidence was due to crust eclogitization when divergence of continental crust in places of spreading was followed by active mantle uprise in which eclogitization proceeded.

9. F. ČECH and J. ZEMAN'S (1985) opinion about predisposition of basification processes and hot mantle diapirs uprise in regions with original suboceanic (simatic) thinner crust is closer to I. V. ARKHIPOV'S (l. c.) concept. The Mediterranean area is the zone of the weakest crust granitization proceeding discontinuously on a limited area (e. g. islands in the Mediterranean Sea, a. o.). Mantle diapirism and development of depressions comprise various stages from the cool post-collapse stage through fading out thermal activity to culminating activity and volcanism. Subsidence of crust thinned by basification is caused by hot less viscous mantle (asthenosphere elevation) into which heavy crustal blocks submerge and assimilate there. Owing to rests of basalt or andesite differentiation heavy rock blocks form in asthenosphere and support circumdiapir subsidence and underthrust of lithosphere beneath diapir. This model accepts partial centrifugal or centripetal (convergent and divergent) subductions along the margin of uprising mantle (partial Benioff zone) into mantle circumdiapir rim synclines. In contrast to other oceanization models it stresses heterogeneity of crust types, pre-existence of simatic crust, its nonconsolidability, i. e. rejects the concept of ancient platform. It does not consider ancient oceanic crust or ancient ocean which is not directly evidenced (e. g. YA.P. MALOVITSKY et al. 1982 — Fig. 20).

Mantle density differences owing to crust eclogitization or to heavy restites of crustal differentiation are manifested by high values of gravity anomalies. M. E. ARTEMYEV (1975) also presumes heavy masses below asthenosphere. They might have got to that depth level by circumdiapir subduction.

Literature concernign the Mediterranean region contains frequent discussions about the existence of original restite or unsubducted oceanic crust. Plate tectonic models also comprise concepts of various origin of the Mediterranean: K. J. HSÜ (1977) prefers continent riftogenesis and ocean opening for the West-Mediterranean segment, and the existence of restite oceanic crust with opening in the Jurassic time and plate convergence in the Paleogene — for the East-Mediterranean segment. J. M. LORT (1977) also presumes riftogenesis and formation of new oceanic crust and intermediary crust beneath basins for the western segment whereas for the eastern part he presumes basification of heterogeneous continental crust. Authors, basing on paleotectonic and paleogeographic reconstructions, reject the idea of the existence of former oceanic crust. J. I. Nikolskii (1982) does not accept the existence of suboceanic crust and presumes extensive basification owing to mantle diapirs, basing on geophysical data in thin lithosphere (in the West Mediterranean 35—100 km, in

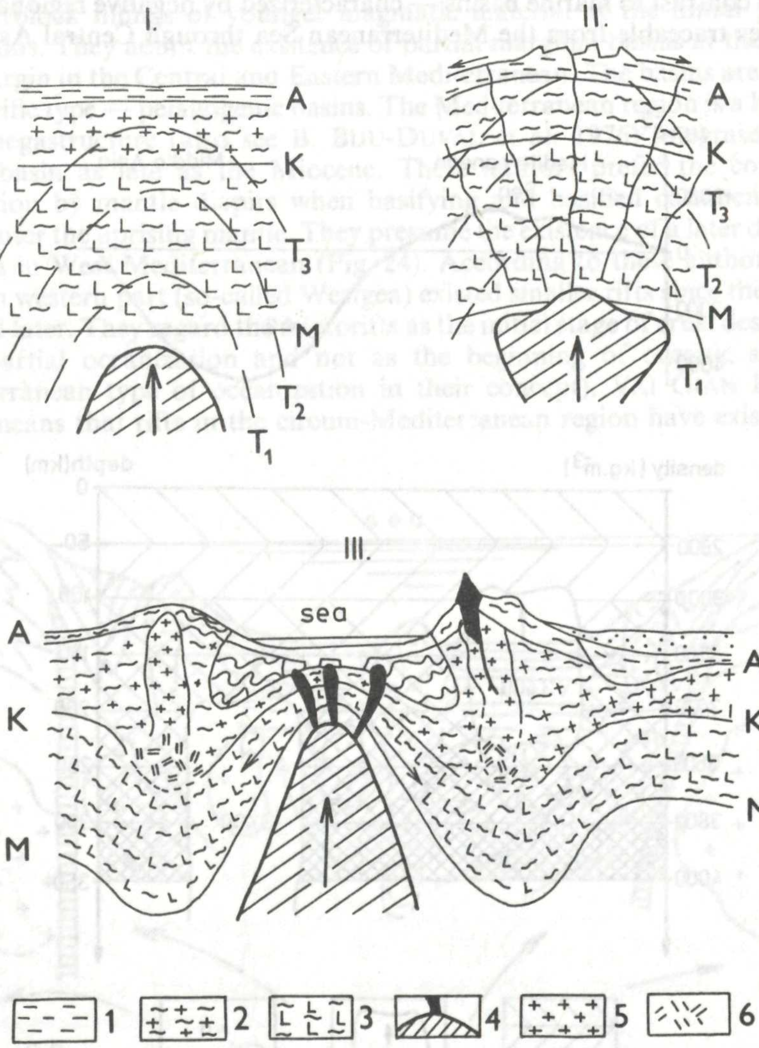



Fig. 22 The scheme of intrusion of mantle diapir (according to Ju. I. Nikolsky 1982).
Explanations: 1 — sedimentary and metamorphic complexes, 2 — "granite layer", 3 — "basalt layer", 4 — mantle diapir with intrusions of basic magma, 5 — granite-gneiss dome, 6 — magmatic chambers with acid magma. Discontinuities: A — surface of the crystalline basement, K — Conrad discontinuity, M — Moho-discontinuity.
Isotherms: $T_1$, $T_2$, $T_3$. Development stages of diapir: I, II, III. Arrows designate movements of the upper part of the lithosphere and those of deep-seated substances.

the East — 60—100 km). Active diapirism is associated with a high heat flow (Fig. 22). He presumes a 200—300 km thick zone of partially melted mantle rocks at a depth of 800—900 km. The zone should be a condenser of thermal energy uprising by convection. In shallow mantle the secondary melting centres form as a source of mantle diapirs. Diapirs which did not reach the crust base, are — in contrast to marine basins — characterized by negative regional gravity anomalies traceable from the Mediterranean Sea through Central Asia to the

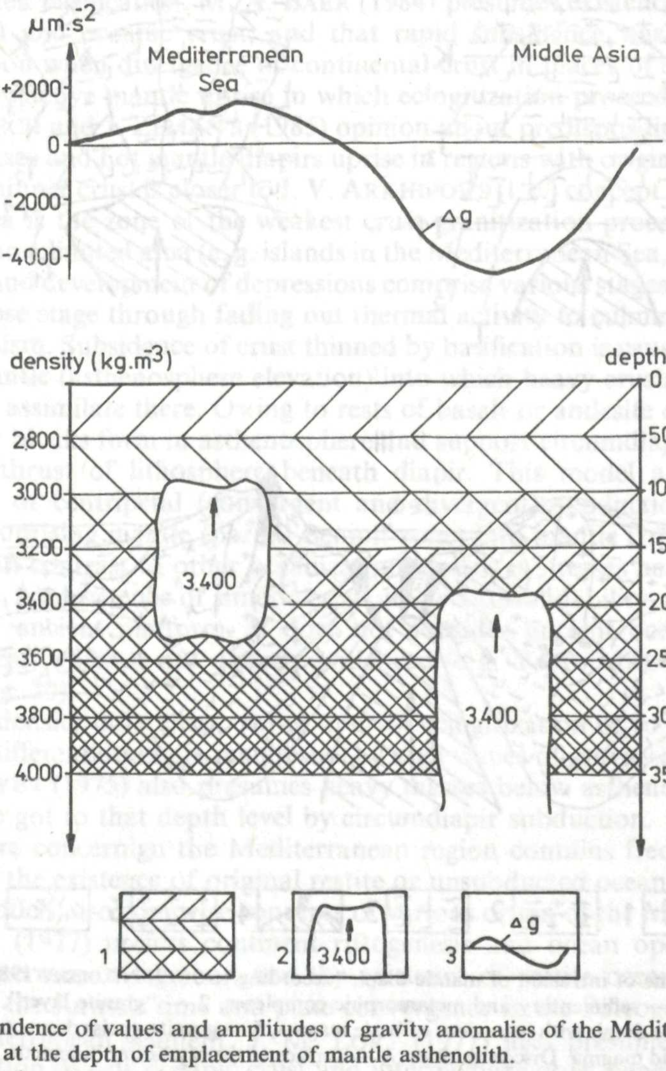



Fig. 23 Dependence of values and amplitudes of gravity anomalies of the Mediterranean sea and Central Asia at the depth of emplacement of mantle asthenolith.
Explanations: 1 — the crust with the designation of increasing density from 2,8—4,0 . $10^3$ kg . m$^{-3}$ — on the left side of the scheme, 2 — asthenolith with the density 3,4 . $10^3$ kg . m$^{-3}$, 3 — the curve of gravity field anomaly.

Pacific Ocean (Fig. 23). New-forming mantle asthenoliths display increasing depth in this direction, and they cause regional uplifts, like that of basins in Asia to alpine levels.

YA. P. MALOVITSKY et al. (1982) deny the existence of the Tethys Ocean and prefer the existence of heterogeneous mosaic of ancient continental fragments and interblock fillings of younger magmatic material at the initial phase of depressions. They admit the existence of partial marginal basins at the African plate margin in the Central and Eastern Mediterranean. The basins are denoted as a specific type — pelagogenic basins. The Mediterranean region is a heterogeneous megastructure (also see B. BIJU-DUVAL et al. 1976) integrated into a marine basin as late as the Miocene. These authors prefer the concept of basification by mantle diapirs when basifying and basified continental crust blocks enter the uprising mantle. They presume the existence of a later destroyed platform in West Mediterranean (Fig. 24). According to these authors in this platform western part (so-called Westgea) existed smaller rifts since the Triassic time and later. They regard the microrifts as the initial stage of crust destruction, of its partial oceanization and not as the beginning of oceanic spreading (Mediterranean type of oceanization in their concept). VAI GIAN BATTISTA (1984) means that rifts in the circum-Mediterranean region have existed since

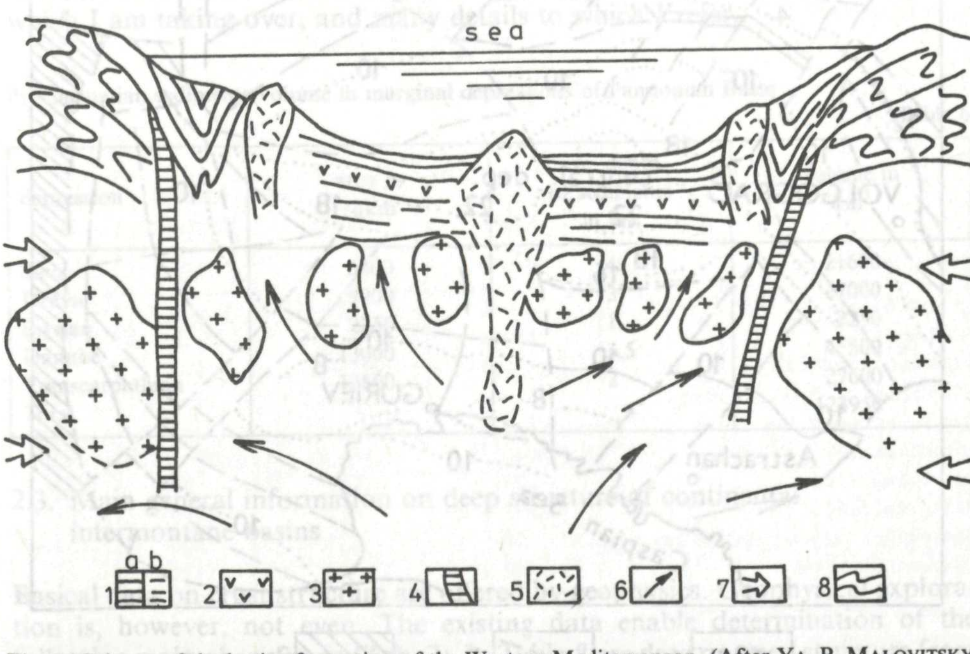



Fig. 24 Scheme of the basins formation of the Western Mediterranean. (After YA. P. MALOVITSKY et al. 1982, modified and simplified by the author).
Explanations: 1 — sediments: a — Late Miocene-Quaternary, b — older, 2 — evaporites (Messinian), 3 — blocks of basement, 4 — deep-seated faults, 5 — volcanoes and magma chambers, 6 — intrusions of mantle asthenolith, 7 — horizontal stresses, 8 — folded sequences and overthrusted.

the Cambrian and their number increased during the Paleozoic and Mesozoic. There are still more opinions rejecting the idea of the formation of the Mediterranean by dynamics of plate tectonics applied on the origin of oceans. There also are still more evidences of specific character of development, for the first time stressed by R. W. VAN BEMMELEN at the end of the sixties.

Considering that the concept of extensive diapirism and polyriftogenesis is also applied on the origin of oceans (e. g. G. B. UDINTSEV et al. 1984a; A. E. SVYATLOVSKY 1984), we can regard the present opinions about the origin of the Mediterranean as a new stage in geologic study of oceanic basins.

## 2.2. Characteristic features of continental intermontane basins

Intermontane basins are often elliptical, less frequently triangular and in isolated cases also circular. V. S. ZHURAVLEV (1972) ranges the Pericaspian and the Pannonian Basins among the so-called hexagonal basins according to the course of marginal faults (Fig. 25). The basins are underlain by thin continental, simatic and oceanic crust. All continental basins and some marine basins passed

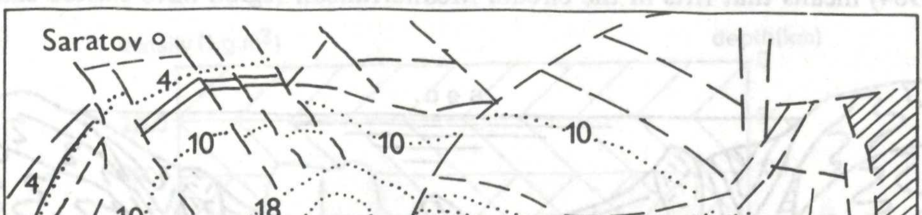



Fig. 25 Structure of the Pericaspian depression basement (After V. V. BELOUSOV 1982, simplified). Explanations: 1 — isohypse of the basis of sedimentary complex in km of depth (according to P-waves velocities), 2 — faults (terraces in the basement), 3 — South Uralides in outcrops.

over a long subsidence development and their basement often shows heterogeneous zonal or block structure with heterogeneous crust type. V. V. BELOUSOV (1982) denotes the basement of the basins as inner massifs whose shape was inherited by intermontane basins.

On peripheries of most basins are marginal depressions conformable with the main basinal axis, or their longer axis is diagonal. Subsidence intensity in marginal depressions was higher when subsidence in the intermontane basin was less intense, and vice versa. Table 5 shows the inverse relation between subsidence intensity in marginal depressions and in the main depression. Volume of sedimentary filling of marginal depressions (Table 6) is indicative of a significant peripheral subsidence. With respect to time dynamics of subsidence the Vienna Basin may also be ranged among peripheral depressions. The thickness of the Pliocene and the Pannonian is 500—1000 m, of the remaining Miocene — 3000—5000 m. It, however, differs from other depressions in its position (behind the Peripieninian lineament in the Outer Carpathians) and in its deep structure. Data presented in text and in tables are from available publications, mainly: B. BIJU-DUVAL et al. (1974), N. E. NAIRN et al. (1977), M. LEMOINE (1978). Most publications concern narrower areas: L. CONSTANTINESCU et al. (1973), N. A. BELYAEVSKY (1974), YU. D. BULANZHE et al. (1975), M. BOCCALETTI et al. (1976), F. HORVATH et al. (1981), F. HORVATH and L. ROYDEN (1981), V. V. SEMENOVICH — YU. G. NAMESTNIKOV (1981). They published basical data which I am taking over, and many details to which I refer.

Pre-Sarmatian sediments volume in marginal depressions of Pannonian Basin

Table 6

| depression | area in sqkm | ave. depth of sediments in km | volume in km$^3$ |
|---|---|---|---|
| Sáva | 5400 | 4 | 21600 |
| Dráva | 9000 | 3 | 27000 |
| Styrian | 2250 | 1 | 2250 |
| Danube | 13000 | 3,5 | 45500 |
| Transcarpathian | 13800 | 2 | 27600 |
| Total | | | 123950 |

## 2.3. Main general information on deep structure of continental intermontane basins

Basical data on crust structure are offered by geophysics. Geophysical exploration is, however, not even. The existing data enable determination of the following main phenomena (Tab. 7). In Table 8 are data on deep structure from the main Carpathian non-marine basins. Further data on other basins and basinal structures were published by F. ČECH and J. ZEMAN (1984). Deep seismic focuses are indicative of partial arcuate deep discontinuity which is by some geophysicists identified with the Benioff zone and subduction (e. g. B. BIJU-DUVAL et al. 1974, M. BOCCALETTI et al. 1976 — Fig. 26, 27).

The crust-mantle boundary is evidenced as a sharp 5 km thick zone, in places characterized by two Moho-discontinuities. Basins with a high heat flow(Tab. 8) have an anomalous upper mantle with Vp velocities below 8.0—8.2 km/s. The anomalous zone either has a crust-mantle mixed character or is formed by lower-density mantle. The anomalous mantle is aseismic. The colder rigid

Characteristic features of dynamics of intermontane basins

Table 7

| | intermontane basins | orogenic rim |
|---|---|---|
| gravitational field | positive | negative |
| magnetic field | mostly positive, heterogeneous | mostly negative, frequently stripped |
| crust thickness | small | great |
| asthenosphere depth | shallow | deep |
| seismoactivity | near margins, mostly aseismic inside | frequent and intense |
| recent movements | negative | positive |

Data on deep structure and Neogene subsidence in recent non-marine sedimentary basins in Alpine-mobile Europe

Table 8

| | Pannonian Basin | Transylvanian Basin | Moesian depression | Pericaspian depression |
|---|---|---|---|---|
| crust thickness in km | 24—32 | 24—30 | 28—35 | 26—30 |
| of granite layer | 10—19 | 8—10 | 5—10 | 0—15 |
| of sediments | 11—14 | 7—8 | 7—15 | 10—22 |
| of Miocene sediments | 1—5* | 3—6 | 0,5—1 | 3? |
| of Pliocene-Quaternary sediments | 2—4 | 0,1—0,8 | 1,3—1,6 | 2 |
| age of volcanism in Ma | 1,3—20 | 2—6 | — | — |
| maximum heat flow in $mWm^{-2}$ | 100—130 | 50—60 | 60—70 | 50 |
| (mean value) | (95) | (40) | (50) | (40) |
| gravitational field in m | +100 až +150 | ±0 | ±100 | +100 |
| asthenosphere surface depth in km | 50 | 115 | 150 | 140 |
| lithosphere thickness | 50 | 115 | 150 | |

*Maxima in marginal depressions

Geological and geophysical data from U.S.S.R by N. A. BELYAEVSKI (1974), JU. D. BULANZHE et al. 1975, V. B. SOLLOGUB et al. 1984; from Pannonian Basin by L. STEGENA et al. (1975), from Transylvanian Basin and Moesian depression by L. CONSTANTINESCU et al. (1973), C. DEMETRESCU et al. (1983). Heat flow accord to V. ČERMÁK — L. RYBACH (1979), lithosphere thickness calculated from heat flow by V. ČERMÁK (1980), in Pannonian and Transylvanian Basin — according to V. B. SOLLOGUB et al. (1984).

mantle beneath the orogen accumulates more energy and is seismically very active. The highest seismicity is on the contact between thick and thin crusts.

The crust underlying the entire Mediterranean Sea (Fig. 28) is thin or

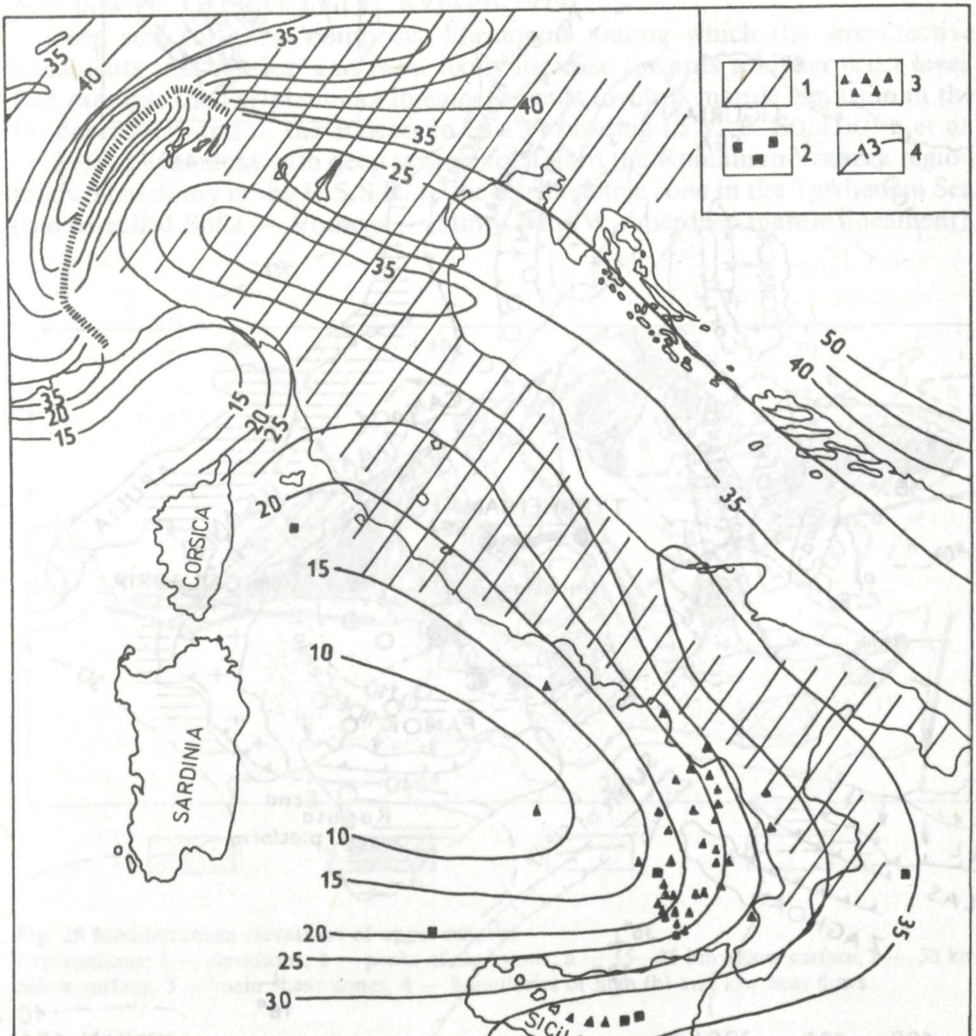
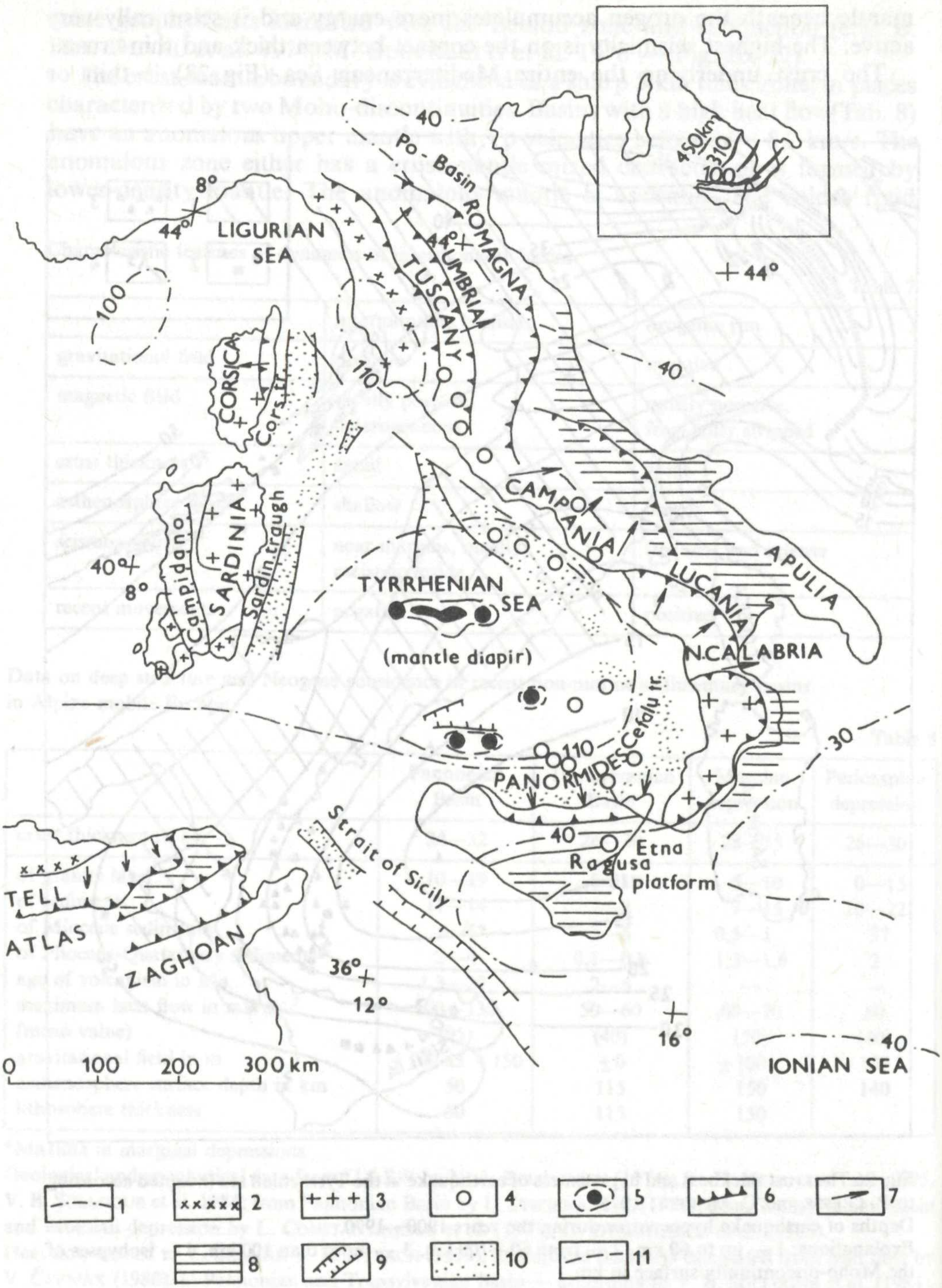



Fig. 26 The crust thickness and hypocentres of earthquake in the Tyrrhenian sea (adapted according to P. Giesse and C. Morelli 1975).
Depths of earthquake hypocentres during the years 1900—1970.
Explanations: 1 — up to 60 km, 2 — from 60—100 km, 3 — more than 100 km, 4 — isohypses of the Moho-discontinuity surface in km.



LIGURIAN SEA
Po - Basin
ROMAGNA
UMBRIA
TUSCANY
CORSICA
Cors. tr.
SARDINIA
Campidano
Sardin. traugh
TYRRHENIAN SEA
(mantle diapir)
CAMPANIA
LUCANIA
APULIA
N. CALABRIA
PANORMIDES
Cefalu tr.
Etna
Ragusa platform
Strait of Sicily
TELL ATLAS
ZAGHOAN
IONIAN SEA
450 km
310
100
44°
8°
40°
36°
12°
16°
40
100
110
30
0 100 200 300 km
1
2
3
4
5
6
7
8
9
10
11

thinned. The asthenosphere forms a dissected elevation. Some basins did not show any Vp typical of the granite layer which is missing (in the Black Sea, the South-Caspian Depression, Ligurian Sea).

Basins with Miocene and mainly recent volcanism have anomalously high heat flow (V. ČERMÁK and L. RYBACH 1979).

Deep structure also comprises lineaments among which the seismoactive lineaments still develop and most likely increase towards a higher crust level. For example the Peripieninian lineament tends towards mantle elevation in the Po depression and in the Balearean Sea. According to V. B. SOLLOGUB et al. (1984) the lineament with deep seismic focuses in the Rumanian Vrancea region tends from Sumy in the U.S.S.R. to the seismoactive zone in the Tyrrhenian Sea (the so-called Sofia — Vrancea — Sumy NE-SW superdeep mantle lineament).

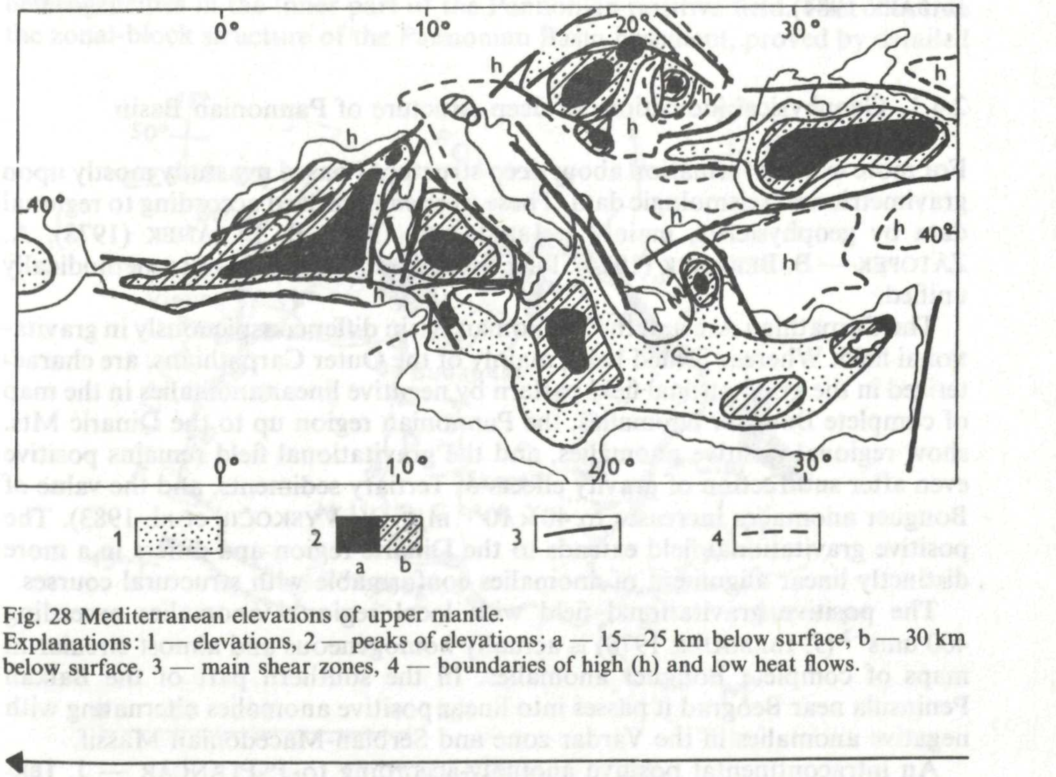



Fig. 28 Mediterranean elevations of upper mantle.
Explanations: 1 — elevations, 2 — peaks of elevations; a — 15—25 km below surface, b — 30 km below surface, 3 — main shear zones, 4 — boundaries of high (h) and low heat flows.

Fig. 27 Geologic structure of Tyrrhenian sea.
Explanations: 1 — heat flow isolines in $mWm^{-2}$, 2 — Kabylides and Toscanian ridge, 3 — granitoides, 4 — alkaline-calc volcanics, 5 — oceanic tholeiites, 6 — nappes and overthrusts, 7 — tectonic transport direction, 8 — Neogene Foredeep, 9 — grabens, 10 — marginal depressions of sea, 11 — supposed deep-seated faults. In the upper right corner: approximate shape of Benioff zone. Modified according to A. CAIRE (1973), M. BOCCALETTI et al. (1976) and F. HORVÁTH et al. (1981).

## 2.4. Deep structure of Neogene basins in Carpathian and Dinaric regions

The Alpine—Carpathian orogen and Dinarides join or bound the Pannonian Basin separated by the Apuseni Mts. from the Transylvanian Basin. Both basins are almost circular and show many characters in common with Mediterranean and circum-Mediterranean basins including the so-called Moesian Depression. The Moesian Depression is bounded by marginal depressions (Tab. 5) with indications of inversion in subsidence in relation to the Moesian Depression, similar to inversions in the Pannonian and Transylvanian Basins.

Some authors compare the Pannonian Basin to the Great Basin in the U.S.A. (L. STEGENA et al. 1975). But the Pannonian Basin shows few features in common with the Great Basin and many with Asiatic intermontane basins, like the Tadjic-Afghanian Depression (F. ČECH — J. ZEMAN 1980). Crust thickness in this depression is about 31—35 km, in peripheral orogens 55—70 km (A. M. BABAEV 1984).

### 2.4.1. Geophysical indications of deep structure of Pannonian Basin

For more exact information about deep structure I based my study mostly upon gravimetric and seismologic data. These were summarized according to regional data by geophysicists, mainly J. IBRMAJER (1978), B. BERÁNEK (1978), A. ZÁTOPEK — B. BERÁNEK (1974). I use these data because they are methodically unified.

The Carpathian Arc and the Pannonian Basin differ conspicuously in gravitational field. Whereas folded belts, mainly of the Outer Carpathians, are characterized in the gravitational field pattern by negative linear anomalies in the map of complete Bouguer anomalies, the Pannonian region up to the Dinaric Mts. show regional positive anomalies, and the gravitational field remains positive even after subtraction of gravity effects of Tertiary sediments, and the value of Bouguer anomalies increases to $40 \times 10^{-5}$ m/s$^2$ (V. VYSKOČIL et al. 1983). The positive gravitational field extends to the Dinaric region and differs in a more distinctly linear alignment of anomalies conformable with structural courses.

The positive gravitational field with local regional anomalies exceeding 400 ums$^{-2}$ (J. IBRMAJER 1978) is actually homogeneous and almost circular in maps of complete Bouguer anomalies. In the southern part of the Balkan Peninsula near Beograd it passes into linear positive anomalies alternating with negative anomalies in the Vardar zone and Serbian-Macedonian Massif.

An intracontinental positive anomaly-according to J. PLANČÁR — J. IBRMAJER (in O. FUSÁN et al. 1971) corresponds in its areal extent to anomalies of the highest category with their source in the lower crust and in upper mantle.

The positive field (I denoted it as the Pannonian positive gravitational field) only extends to margins of Alpine-Carpathian orogen: partly to the East Alps, more to the Inner Carpathians and completely to the Serbian-Macedonian Massif, the Moesian Depression and the Apuseni Mts. in Romania (Fig. 29, 30).

For orientation and subsequent correlation of data I have compiled three schemes generalizing considerably the reality, yet enabling a comparison among the principal manifestations of crustal structures (Fig. 29, 30, 31).

Comparing the Pannonian Basin to the North-American Great Basin (L. STEGENA et al. 1975), I want to mention diametral differences in their gravitational field characters. Gravitational field of the Great Basin is negative. It is explained by a lesser density of basin-underlying mantle (CH. H. SCHOLZ et al. 1971) and its crust is thicker (30 and more km — Fig. 5) than that of the Pannonian Basin. I think that the source of gravity differences are in material composition of crust of both basins.

The positive Pannonian field extends to the southern part of the West Carpathians and farther northwards in East Slovakia — deepest in the Carpathian Arc.

It is important that the deep gravity source mass excess covers all structural heterogeneities in the inner part of the Pannonian positive field. This concerns the zonal-block structure of the Pannonian Basin basement, proved by detailed

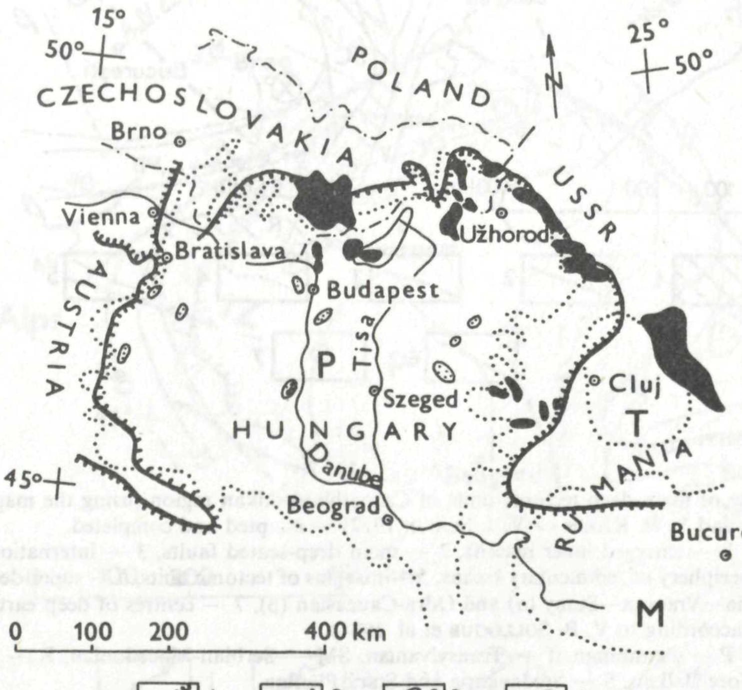



Fig. 29 Extent of Carpathian-Balkan region disturbed by positive gravity in relation to Inner-Carpathian Neogene basins.
Explanations: 1 — zero isoanomaly μms$^{-2}$ confines positive plane to negative isoanomaly (according to J. IBRMAJER, 1978), 2 — periphery of Neogene-Quaternary basins, 3 — neovolcanics, 4 — anomalies above +300 μms$^{-2}$; P — Pannonian Basin, T — Transylvanian Basin, M — Moesian Platform.

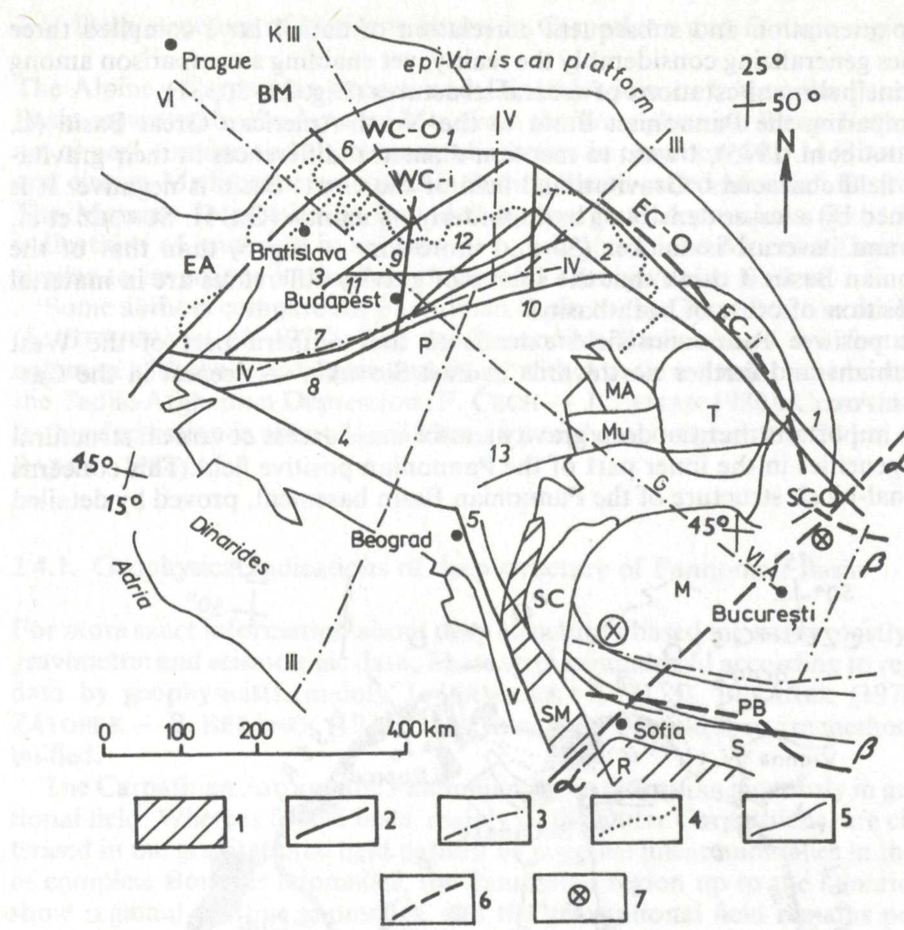



Fig. 30 Scheme of main deep tectonic units of Carpathian-Balkan region (using the maps of M. MAHEL 1978b and V. E. KHAIN — V. I. SLAVIN 1972) — adapted and completed.
Explanations: 1 — emerged inner massifs, 2 — main deep-seated faults, 3 — international DSS profiles, 4 — periphery of sedimentary basins, 5 — margins of tectonic units, 6 — super-deep matle fault zone Sofia—Vrancea—Sumy (α) and Odra-Caucasian (β), 7 — centres of deep earthquakes Vrancea (6, 7 according to V. B. SOLLOGUB et al. 1984).
Inner massifs: P — Pannonian, T — Transylvanian, SM — Serbian-Macedonian, R — Rhodop Mts., PB — Fore-Balkan, S — Srednegorie and Stará Planina.
Forelands: BM — Bohemian massif, V — Vardar zone, WC — Western Carpathians (O — outer, I — inner). EC — Eastern Carpathians (O — outer, I — inner), SC — Southern Carpathians, G — getic units, EA — Eastern Alps, MA — Muntii Apuseni Mts., Mu — Muresh trough, M — Moesian megablock.
Deep-seated faults: 1 — Peripieninian lineament, 2 — Számos line, 3 — Rába line, 4 — Insubric line, 5 — Vardar-Kraishtide zone, 6 — Labe lineament, 7 — Balaton line, 8 — Zagreb-Kulcs line, 9 — N-S Danube line, 10 — N-S Hornád line, 11 — Hurbanovo fault, 12 — Darnó line, 13 — Muresh fault.

geologic and geophysical investigations and by many boreholes (G. WEIN 1969). In marginal parts of gravitational field the deep effect is weaker — as proved by detailed gravimetric interpretations by J. IBRMAJER and J. PLANČÁR (in O. FUSÁN et al. 1971).

A complicated heterogeneous block structure was manifested at the use of transformed fields for the study of deep structure of Czechoslovakia (B. BERÁNEK 1979). In the study of crust structure the method enabled diminution of disturbing bodies anomalies at depth more than 10—15 km, the weakening of subcrustal effects, and excluding of the effects of shallower bodies to depths of 5—8 km. Gravimetric interpretations also included the northern part of the Pannonian Basin and proved unambiguously the alternation of zonal positive and negative fields with prominent NE-SW gravity gradients. With respect to the use of maps of complete Bouguer anomalies, regional and residual anoma-

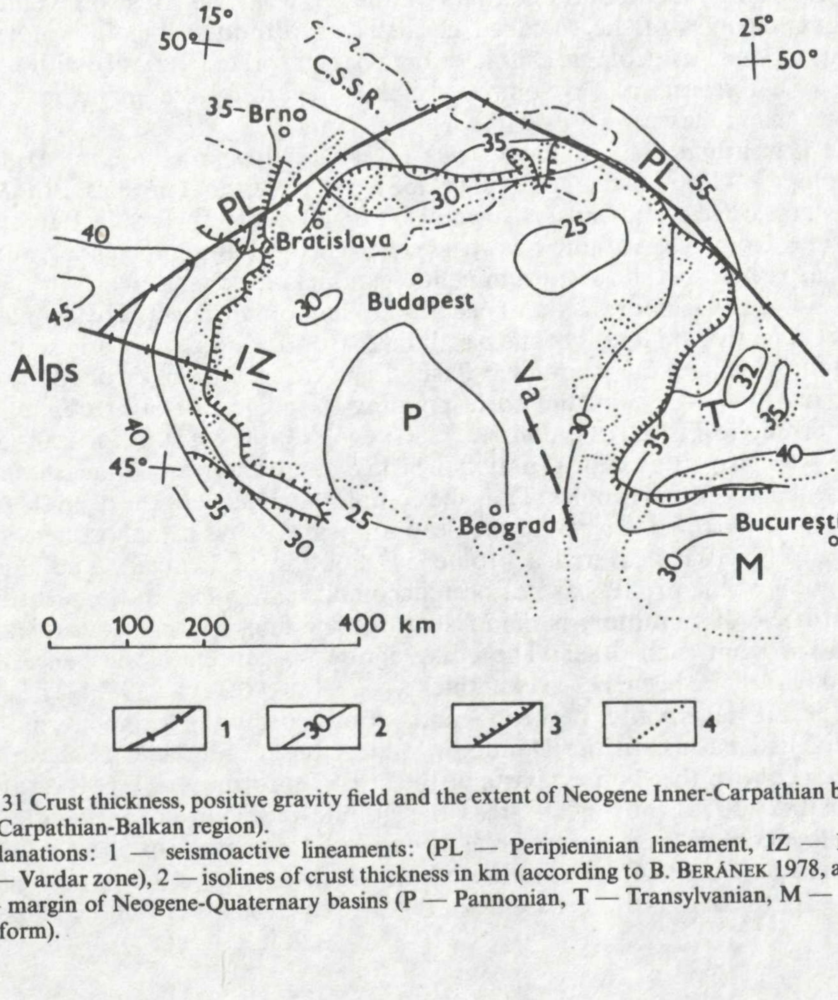



Fig. 31 Crust thickness, positive gravity field and the extent of Neogene Inner-Carpathian basins (in the Carpathian-Balkan region).
Explanations: 1 — seismoactive lineaments: (PL — Peripieninian lineament, IZ — Insubric, Va — Vardar zone), 2 — isolines of crust thickness in km (according to B. BERÁNEK 1978, adapted), 3 — margin of Neogene-Quaternary basins (P — Pannonian, T — Transylvanian, M — Moesian Platform).

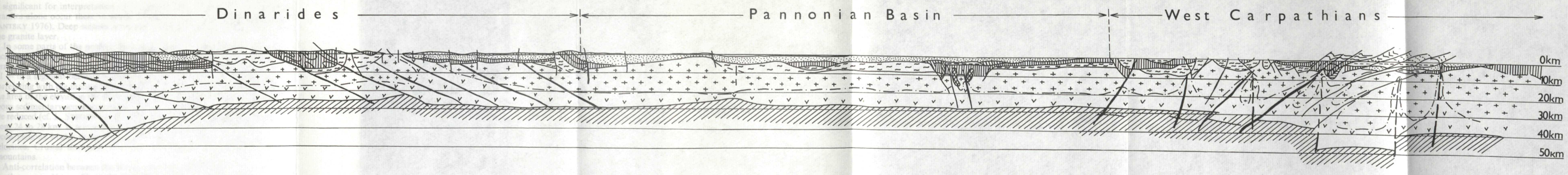
lies for r = 85 km, it is very valuable for the study of deeper basin structure that maps of second gravity derivations prove the pattern resulting from the gravitational field analysis from the above mentioned maps.

Identic positive and negative gravity anomalies along the Slovak — Hungarian frontiers are indicative of a continuous extension of geologic units including Neogene basins and elevations, separating the basins on the Hungarian territory. I shall return to the results of geologic correlation. Here I only want to stress significance of the indication of the Darnó line (G. WEIN 1969). B. BERÁNEK (1979) denoted the line erroneously as the Balaton line. Prominent gradient and identic course of zero isoanomals of second derivations of N-S gravity in deep equidirectional faults extending from Slovakia to Hungary. Their existence is also proved by ultrabasic rock dykes in the N-S zone (I. HORVÁTH — L. ÓDOR 1984). This course corresponds to the Central-Carpathian lineament of J. ŠTOHL (1976).

For the study of deep structure most significant was explosive seismology. It is so far the only method to offer data on the physical crust division, mainly on thickness changes of the so-called basalt layer, although it is interpreted by various models. Its geologic character may be interpreted variably with respect to historical-geological development, mainly regarding old mafic rocks and younger mafic masses.

The Carpathian-Balkan region was investigated by methods of explosive seismology — deep seismic sounding (DSS) in international profiles. Profiles III and V are longest and most instructive (Fig. 30 and 32). Profile III cuts the Dinarides from the Adriatic Sea, passes across the Pannonian block and the East Carpathians. It shows a prominent crust thickness change from 45—30 km beneath the Dinarides to 24 km beneath the Pannonian Basin. The profile also shows decreasing thickness of the basalt layer from 20—25 km in adjacent units to 5—8 km beneath the Pannonian Basin. In a vertical section through the crust the zone of P-waves velocities corresponding to the granite layer dominates.

According to E. MITUCH and K. POSGAY (1972) the crust in the Pannonian region was also measured in national profiles and gravimetric measurements were aimed at seismic profiles. Thus the complex approach to the deep structure researches is enabled. To reveal the deep lithosphere and asthenosphere structure geophysicists measured a profile by shooting in Hungary and Poland. Evaluation of the profile has not been accomplished so far. In the records the Mohorovičić discontinuity is manifested by twinning, i. e. horizons distanced 1—1.5 km from each other: The crust-mantle transitional zone beneath the eastern part of the basin is 2—4 km thick (V. LAZARESCU et al. 1983). The depth of the granite layer varies between 4 and 9 km according to sediment thickness in individual basins. In the Danube r. valley the granite layer is elevated in accordance with the thinnest parts of the crust. According to E. MITUCH and K. POSGAY (1972), not in all areas the granite layer surface is identic with crystalline complexes. In most areas it corresponds to the surface of the pre--Austrian basement of depositional basins. In fact, seismic data are in accordance with the depths of the basement, revealed by deep holes. This explanation



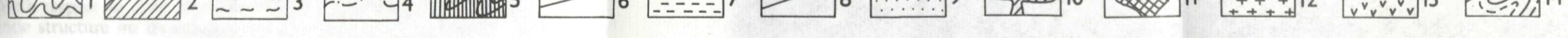




Fig. 32 Geological-seismic cross section through Dinarides-Pannonia-Western Carpathians. Cross section III-V (Dubrovnik—Szeged—Varszawa) SSW—NNE. According to J. ILAVSKÝ et al. (1979), adapted.
Explanations: 1 — Precambrian, 2 — Mohorovicic discontinuity, 3 — Paleozoic, 4 — Conrad discontinuity, 5 — Mesozoic, 6 — deep transcrustal faults, 7 — Paleogene (Flysch), 8 — regional crustal faults, 9 — Neogene to Quaternary, 10 — neovolcanics (andesites — basalts), 11 — ultrabasic igneous rocks, 12 — granitoides, 13 — basalt layer, 14 — vergency of fold structures and nappes.

M = 1 : 1 000 000

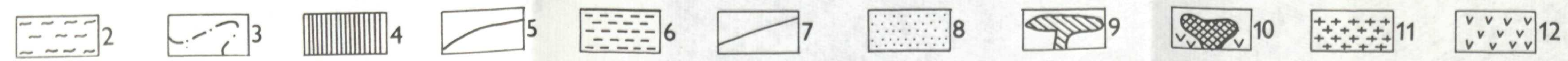




Fig. 33 Longitudinal geological-seismic profile VI and VI A (Brno—Budapest—Bucurest). According to J. ILAVSKÝ et al. (1979). adapted.
Explanations: 1 — Mohorovicic discontinuity, 2 — Paleozoic and crystalline complex, 3 — Conrad discontinuity, 4 — Mesozoic, 5 — deep transcrustal faults, 6 — Paleogene (Flysch), 7 — regional crustal faults, 8 — Neogene to Quaternary, 9 — neovolcanites (andesites — basalts), 10 — ultrabasic rocks, 11 — granitoides, 12 — basalt layer.

M = 1 : 1 000 000

is significant for interpretation of the granite layer character. It is likely that granites alone occur there in smaller amounts — as proved by deep holes (B. JANTSKY 1976). Deep seismic sounding profiles show 10—15 km thickness of the granite layer.

In some parts of the profiles the granite layer has in its upper part zones of increased P-waves velocities (6.5—6.85 km/sec.) possibly corresponding to intermediary rocks intrusions. A granite layer elevation was recorded in the profile VI — in a segment between Kaposvár and Nagykorös (Fig. 33). The elevation correlating with the mantle elevation ranges to 3—5 km along the Danube r. In the area of the elevation the thicknesses of the Upper Paleozoic—Lower Cretaceous sedimentary sequence and of the Upper Cretaceous — Neogene sequence are reduced (cf. L. STEGENA et al. 1975). The Moho discontinuity depth ranges from 24 to 28 km. It was proved by the Hungarian national profile NP-2. The inconspicuous Moho discontinuity consists of 4 levels of increased P-waves velocities. Crustal thickness 30—32 km is only in the Transdanubian Mid--mountains.

Anti-correlation between the Moho relief and basin basement was also found in other areas. No profile in the Pannonian region shows indications of deep--seated faults. The only deep profile III on the contact with the East Carpathians corresponds to the Számos line. The fault is manifested in the crust — not in the mantle, like for example, deep-seated faults beneath the East Carpathians.

With respect to deep structure it is very important to know the stratified mantle structure on its contact with the crust, elevation of the high electric conductivity zone (a depth of 40—60 km), shallow position of the low P-waves velocity channel (depth of asthenosphere surface 75 km — according to L. STEGENA et al. 1975; 50 km — according to V. B. SOLLOGUB et al. 1984). It is more likely that the depth of the asthenosphere surface is smaller and corresponds to results of magnetotelluric measurements. Beneath the East Carpathians the asthenosphere surface is sinking to 80 km and beneath the Ukrainian Shield even to the depth of 200 km. According to Hungarian geophysicists four Moho levels represent the scope of phase transitions on the contact of crust with mantle, and thus the existence of a discontinuity is excluded.

Upper mantle in the Pannonian region is less dense than beneath the West- and East Carpathians. P-waves velocities are by 0.5 km/sec. lower (L. P. VINNIK et al. 1975) beneath the Pannonian Basin. Velocities in the mantle beneath the Carpathians are identic with velocities beneath the East-European platform in the Carpathian foreland. The mantle beneath the Pannonian Basin is seismically inactive. The mantle forms a dome with its peaks situated S and NE of Budapest. West of the dome is the deepest (30 km) depression in mantle. The depression is dipping centrifugally to the depth of 30—55 km beneath the internal West Carpathian units. In external units the crustal thickness is generally increasing. Its thickness is greatest (55—57 km; B. BERÁNEK 1978) beneath the outer East-Carpathian Flysch zone and beneath the Central Taurus in the East Alps (63 km; K. ARIC — R. GUTDEUTSCH 1975). Thinned crust (27—30 km) extends to South-Slovak and East Slovak Neogene basins (24 km).

The mantle elevation corresponds to the asthenosphere elevation at the depth of 50 km.

Eastwards to Romania the mantle is downsagging beneath the South Carpathians 35—40 km deep, and beneath the East Carpathians 50—55 km deep (V. LAZARESCU et al. 1983). There is a deep-seated fault beneath North Dobrodgea, separating it from the Moesian Platform (Profile II). The fault downwarps the North Dobrodgea unit by 10—12 km deeper and is most likely a continuation of deep-seated faults from NE part of the East Carpathians, corresponding to the Számos line according to H. STILLE (1953). New data prove that it is not the Számos line but parallel structures belonging to the Peripieninian lineament. In the area of Vrancea with an intense deep seismic activity crustal thickness ranges up to 54 km (L. CONSTANTINESCU et al. 1975). Thickness of the crust-mantle transition zone exceeds 10 km. Crust thickness beneath the Moesian Platform in international profile II ranges from 20 to 30 km. With the increasing sediment thickness in the overlier of the platform the thickness of the granite layer decreases to 8 km, that of the basalt layer to 18 km in contrast to blocks with thick crust (25—28 km).

The Pannonian Basin is a region with a high heat flow (T. BOLDIZSÁR 1964). In the depth of 1 km temperatures range up to 75 °C, mainly in places of thermal water ascents. The average increase of temperature is 50 °C per 1 km of the depth. According to data presented by L. STEGENA (1972) the areas of maximal temperatures are represented by elliptical centres with the character of thermal domes:

— E of the Tisa river (the area of the East-Pannonian block — see below);

— W of the Danube r., between the Danube r. and Sáva r. (SW of Budapest).

Another high temperature zone is between the Danube and Tisa rivers NW of Beograd.

By its temperature the Pannonian area belongs to tectonically active crust areas similar to those with active volcanism. Heat flow in the Pannonian Basin decreases towards the Carpathians. On the contact with the West Carpathians it decreases in a leap by 41.9 $mWm^{-2}$ (T. BOLDIZSÁR 1964) which is quite a high value.

Towards orogenic zones of the Pannonian Basin the temperature anomaly drops to 45—35 °C in the depth of 1 km; to 25 °C on the adjacent East-European Platform which is still more than inside the Platform.

Paleotemperature evidence is missing. During the shortly increased temperature no thermic metamorphosis took place but the high-degree diagenesis of sediments also in pre-Miocene elevations may be indicative of a higher temperature as early as the Paleogene. Coalification and the brown-coal stage of Eocene coal metamorphosis above bauxite deposits in the Transdanubian Midmountains are sensitive indicators. They also indicate higher heating in coal seams of the northwestern basin where they do not lie beneath any thick overlier and are distant from andesite effusions. A high heat flow in NW Hungary was evidently associated with Lower Miocene volcanism as well. Causes of the present heat production are not exactly known: either the Miocene — Pliocene higher heat flow is fading out or thermal activity associated with the new uprise of mantle diapir and future volcanism, is renewed. All geophysical indications, mainly the circular shape of regional gravitational field, thin crust, the domal shape of mantle and asthenosphere, and a high heat flow prove the existence of mantle diapir.

Seismoactive faults are in tangential contact with the Pannonian Basin and other faults run across the Basin. Contact of the Pannonian megablock (in its size corresponding to the Pannonian median massif — Fig. 30) with West-Carpathian internal units (O. FUSÁN et al. 1971), the contact with the Eastern Alps along the Peripieninian lineament geophysically defined as the Pericarpathian lineament for the deep crust (A. ZÁTOPEK — B. BERÁNEK 1974, B. BERÁNEK 1978), seismoactive in the line Malé Karpaty Mts. — Semmering — Verona, are typical examples. The eastern part of the Insubric line (Fig. 30, 31) is in a special, perhaps radial relation to the positive Pannonian gravity anomaly.

The Carpathian-Balkan region belongs among areas of increased to high seismoactivity — as proved by the macroseismic intensity map of Central Europe (V. KÁRNÍK 1975; Fig. 34). Increased seismoactivity is along the contact of the Pannonian megablock with the central Carpathians, at the crossing of deep-seated faults, mainly at the Rába fault crossing with the Central-Carpathian lineament. The area between the maximally and minimally thinned Pannonian block crust around the Rába and Darnó faults is most active. The north-eastern mantle elevation (NE of Budapest) and a crustal belt about 25—30 km thick, oriented towards Beograd, are inactive. The contacts with thicker crust at the northern periphery of the Pannonian Basin and with the Dinarides are active. An elliptical area W and NW of Beograd, spatially identic with the mantle elevation margin in the southern part of the Pannonian Basin and with the partial high temperature centre shows a higher seismoactivity. Seismic focuses in the Pannonian Basin are shallow (5—16 km), controlled by balancing tension in the basement of depositional basins (L. STEGENA et al. 1975).

Most active is the contact of the Pannonian Basin with the continuation of the Peripieninian (Pericarpathian) lineament, with the Semmering — Verona line (A. ZÁTOPEK — B. BERÁNEK 1974). The line is particularly mobile on the contact between two units differing in density and isostatic tendency. Isolated activity manifestations are on the Hornád fault — denied by P. GRECULA et al. (1977) as a deep-seated fault — and along the eastern part of the Peripieninian lineament. Seismoactive zones run along marginal parts of the basin, mainly along its contact with the Peripieninian lineaments or along faults parallel or oblique to the lineament. It is in accordance with the tectonophysical model (main shear tension zones); analogous to the Red Sea rift and identic with the Great Basin where — according to CH. H. SCHOLZ et al. (1971) — seismoactive zones rest above the mantle diapir margin — the limit of crustal thickness changes (Fig. 5).

Another seismoactive deep-seated fault is the Vardar zone oriented radially to the basin centre. In the NW continuation of the zone are more intense seismic focuses tending towards Budapest (V. KÁRNÍK 1975). Faults on margins of mantle diapir indicate that their Tertiary and present activity is controlled by the diapir dynamic evolution.

### 2.4.2. Geophysical indications of deep structure of Transylvanian Basin

The shape of the Transylvanian Basin shows analogous features with the Pannonian Basin: greater thicknesses of sedimentary filling, orogenic ring, partly also the position of marginal faults with an overthrust character.

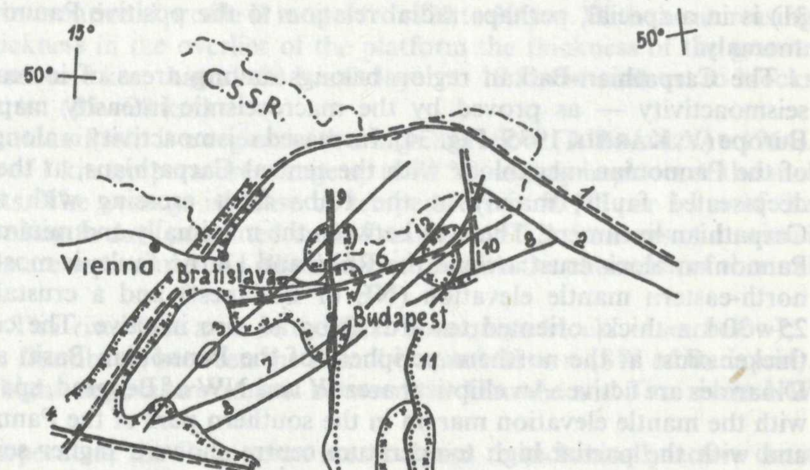



Fig. 34 The main deep-seated faults and recent seismicity in Pannonian Basin.
Explanations: 1 — deep-seated faults, 2 — deep-seated faults with neotectonic activity, 3 — upper crustal faults, 4 — areas of higher seismic activity (according to V. KÁRNÍK 1975), simplified.
Deep-seated faults and lineaments: 1 — Peripieninian, 2 — Számos, 3 — Rába, 4 — Periadriatic (Insubric), 5 — Vardar, 6 — Hurbanovo, 7 — Balaton, 8 — Zágreb-Kulcs, 9 — Danube, 10 — Hornád, 11— Tisa, 12 — Darnó.

The faults are in a tangential contact with the basin margins. Some faults are parallel to the basin margins. There also is accordance in smaller crustal thickness.

According to point seismic data the crustal thickness beneath the basin is about 24—30 km — in contrast to 35 km thickness of the Mt. Apuseni block. Southwards to the Getic crystalline block of the South Carpathians the thickness ranges up to 35—40 km beneath the Getic elevation, and to 45—50 km (L. CONSTANTINESCU et al. 1975) or to 50—55 km eastwards beneath the East Carpathians. The average thickness of the granite layer beneath the Transylvanian Basin is 8 km. The crust-mantle transitional zone is 2—3 km thick beneath the basin and beneath the Apuseni Mts., more than 10 km thick beneath the East Carpathians, and 4—6 km beneath the South Carpathians (V. LĂZĂRESCU et al. 1983). Depth of the asthenosphere surface is indicated at 50 km (V. B. SOLLOGUB et al. 1984). There is an anomalous positive magnetic field up to 400 nT in the basin (S. AIRINEI et al. 1983). The basin crustal block is sharply detached by deep faults from the thick crust belts. The Muresh fault (Fig. 35) shows the most intense seismic activity. In the present dynamic pattern the discontinuity joins the Kraishtide lineament and separates seismotectonically the Getic unit (block) from the Apuseni Mts. and from the Pannonian Basin. Most intense and most frequent quakes in the Vrancea region (V. KÁRNÍK 1979) occur on the bend of the East Carpathians, also caused by faults (the Odra—Caucasus lineament in the sense of V. B. SOLLOGUB et al. 1984). The seismoactive line is regarded as the northern microplate margin (L. CONSTANTINESCU et al. 1975). In contrast to this, V. B. SOLLOGUB et al. (1984) explain the Vrancea centre as a crossing of the Ondra—Carpathian (NW—SE) and the Sofia—Vrancea—Sumy (NE-SW) lineaments originating from the mantle. Asthenosphere forms a narrow diapir on the Sofia—Vrancea—Sumy lineament and uprises from the depth of 200 km beneath the East-European Platform to the depth of 80—70 km. In the northern part the Transylvanian Basin is deep-bordered by the Számos fault in relation to the East-Carpathian crust, ranging to 50 km in thickness. The deep crustal block of the basin is triangular-shaped and its peak is oriented northwards.

As for geophysics, the Transylvanian Basin is similar to the Pannonian Basin but it differs in some deep features typical of the Pannonian Basin, applied on the interpretation of its genesis by mantle diapir. The difference is in a little thicker crust in the depth of asthenosphere beneath the basin and a very low heat flow ranging in average to 60 $mWm^{-2}$ (V. ČERMÁK 1979). According to C. DEMETRESCU et al. (1983) the heat flow in the basin centre is 29—30 $mWm^{-2}$ and on its margins 50—60 $mWm^{-2}$.

Geophysical data on deep structure of the basin basement show that the crust is thinner here than in the Inner West Carpathians or Bohemian Massif, and may be compared to simatic crust of median massifs. With respect to this the Transylvanian Basin may be ranged among depressions with heterogeneous thinning crust. The basin rests upon primarily subsiding massifs like the Moesian Platform. Beneath the Transylvanian Basin the crust thickness is about

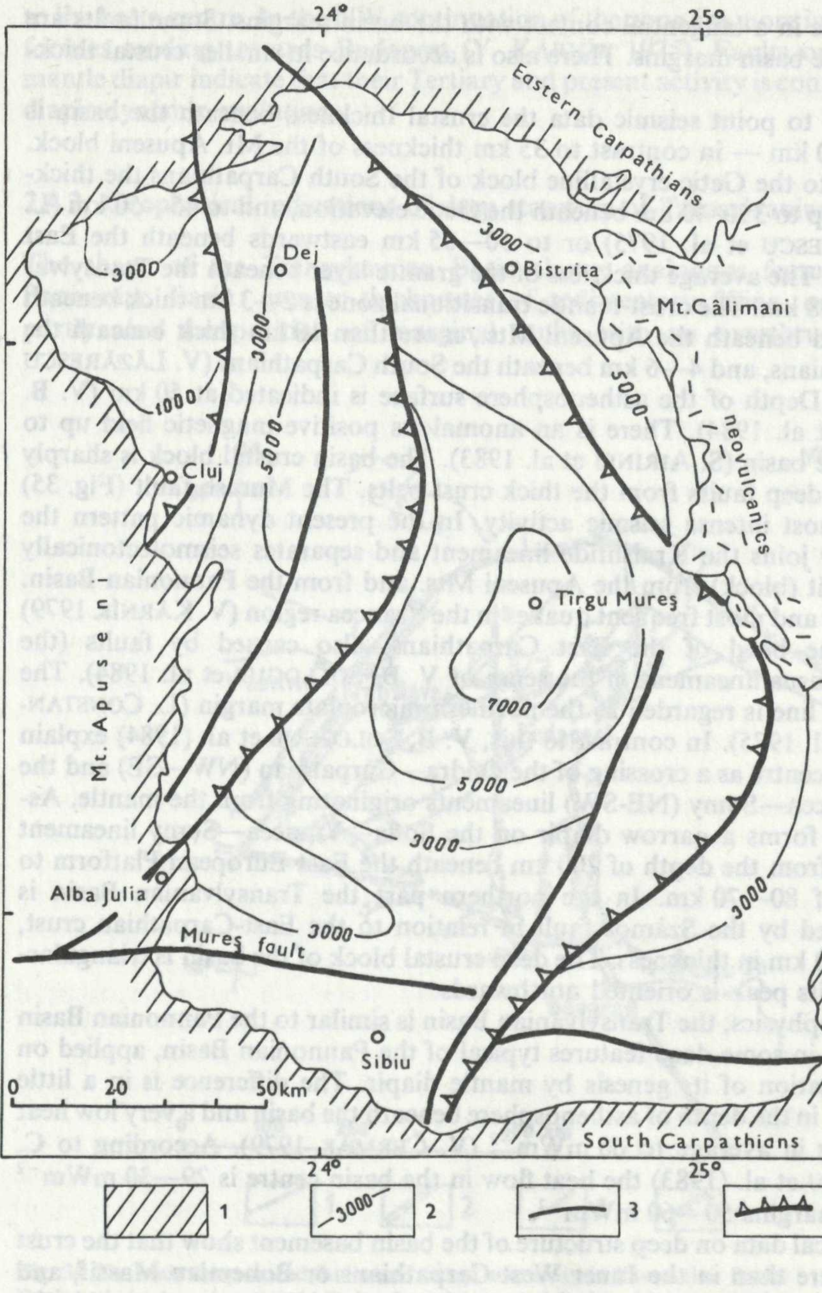



Fig. 35 Tectonic structure of Transylvanian Basin. [Adapted according to Tectonic map of Romania (1970) and V. V. SEMENOVICH — JU. G. NAMESTNIKOV (1981)].
Explanations: 1 — pre-Neogene complexes, 2 — basement surface isohypses in m, 3 — faults, 4 overthrusts.

24 km, in the southern part 32 km. If 3—7 km of sediments are subtracted, then the pre-Tertiary crust was only 20—29 km thick, in the central part of the basin in the Muresh depression — 17—20 km. These are thicknesses of continental crust of mobile median massifs. So the diapir influenced crustal reduction in a minimal extent. Its effects were most likely restricted to the old suboceanic crust. The present low heat flow does not exclude the past diapir effects evidenced by the different basin tectonics.

### 2.4.3. Geophysical indications of deep structure of the Moesian Platform-Depression

The elliptic Moesian depression is surrounded with South-Carpathian orogen and Balkanides (Pre-Balkan) — Fig. 36. On the east it passes into the Black Sea depression. In the Tertiary and Quaternary times the Moesian Depression and the Black Sea formed the Moesian-Pontic Depression as a less mobile part. The depression shows more features in common with intermontane depressions than with a typical platform. Crust thickness in the depression is about 30 km, in the surrounding orogenic belts-up to 50 km. The granite layer is only 10 km thick, the crust-mantle transional zone is 1—2 km (V. LĂZĂRESCU et al. 1983). Gravitational field is mostly positive, with an eastward values increase. Magnetic field is homogeneous, positive and negative, with low values — except in marginal fault zones with numerous positive anomalies up to 400 nT (S. AIRINEI et al. 1983). The depression centre is almost aseismic, the margins are associated with seismoactive faults. In relation to the Balkanides (Pre-Balkan) the depression is bordered with the South-Moesian fault zone (E. BONTCHEV — D. YOSIFOV 1984).

Geophysical data show that the structure of the Lom Basin at the southern margin of the Moesian Depression (Fig. 36, Tab. 5) differs from the structure of the latter.

In the Lom Depression the basement of the Tertiary-Paleozoic sediments is at the depth of 8—10 km (K. DAČEV et al. 1972). The Conrad discontinuity is inconspicuous, undulated. It is about 20 km beneath the depression surface. Thickness of the granite layer E of the Lom Depression is only 7—8 km. Indicated basalt layer elevations in the crust are in places in the depth of 8—9 km. They may be regarded as ancient partial diapirs of basic and ultrabasic rocks.

Seismic focuses with M 5 are distributed in two NE-SW zones. They run in the basement of the Moesian Depression and border the NE-SW linear blocks: the West Bulgarian, Central-Bulgarian and the East-Bulgarian (E. BONTCHEV and D. YOSIFOV 1984). According to these authors the single blocks differ physically in the degree of sialization increasing towards the East-Bulgarian block, i. e. towards the Black Sea. In the eastern part of the Moesian Depression basement the simatic crust prevails. So it is in the East Rhodopes devoid of granitoids and with dominant metapelites and metabasites in crystalline complexes. It is not exactly known whether the primary simatic crust controls the structure, mainly in the Black Sea region or the secondary changes i.e. crust basification — control oceanization in the Black Sea region.

In my opinion, if basification proceeded at all, then in the area of ancient simatic crust controlling the long subsidence tendency in the Moesian Depression and in the Black Sea Depression. The basification may also be evidenced by a heterogeneous geothermal field characterized by the values 70—80 $mWm^{-2}$. According to C. DEMETRESCU et al. (1983) it may be caused by subrecent basalt intrusions (sill-like) in upper mantle. In the places of the more simatic crust are calculated temperatures up to 700 °C.

Geophysical data indicate the thickness of the Quaternary up to 2 km and the Paleogene base depth to 8 km. Subsidence proceeded in the Mesozoic time (D. A. TUGOLESOV et al. 1984). According to these authors the Black Sea sedimentary filling is segmented into transversal rises and depressions. This is, however, not in accordance with the idea of a continental rift genesis of the depression, parallel to the sea axis, nor with the concept of the rift genesis of the Moesian

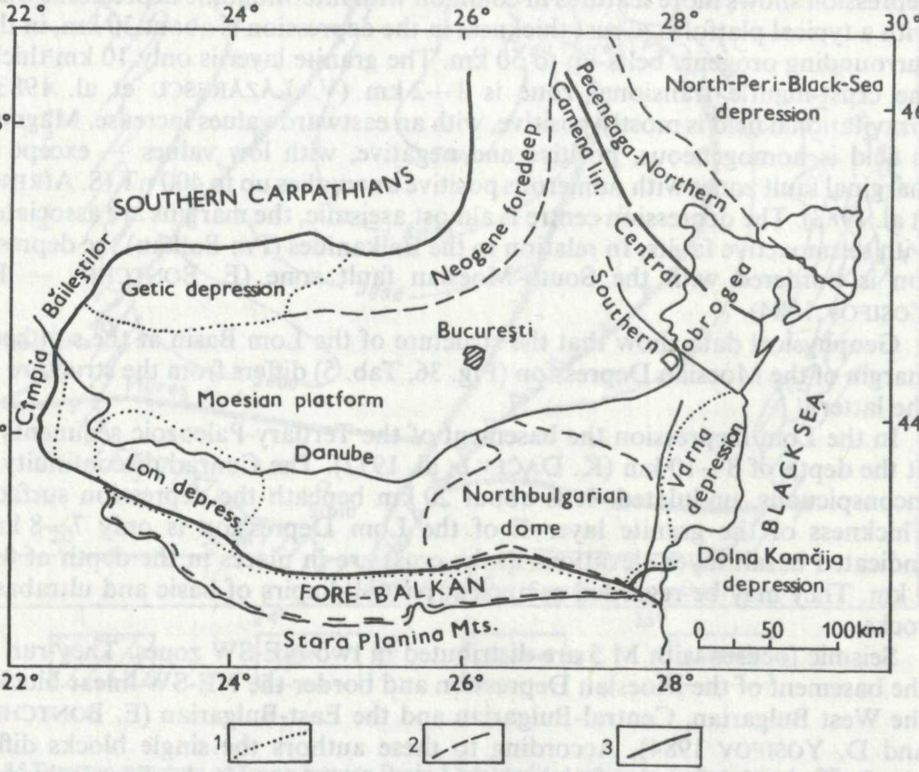



Fig. 36 Moesian Platform — depression and its division.
Explanations: 1 — marginal depressions, 2 — significant tectonic units, 3 — margin of the Moesian plate.

Depression — as presumed by CH. I. DATCHEV (1984). The rift structure is not indicated geophysically. But we can agree with CH. I. DATCHEV'S (l. c.) opinion about the Moesian and Black Sea Depressions having been the inner sea with a long subsidence since the Late Paleozoic and especially since the Mesozoic. The Moesian Depression is likely to have had the character of an ensimatic back arc basin in the Paleozoic time (F. ČECH — J. ZEMAN 1985).

The seismologic pattern of the Carpathian-Balkan region are three elliptical to circular centres (the Pannonian, Transylvanian Basins, the Moesian Depression) of crust reduced to 32—55 km (Fig. 31). Zones of a thicker crust correspond to folded units, to active orogenic, morphologically manifesting mountain- and alpine reliefs in contrast to basins on median massifs.

## 2.5. Geologic characteristics of Neogene intermontane basins

Neogene intermontane basins. Sedimentary filling of depressions depends — with respect to variable thickness of sediments — upon the structure (± block) of the basin basement. Post-Miocene sediments — except the Pannonian Basin — are poorly differentiated and non-differentiated according to thickness. At the end of the Miocene and during the Pliocene, re-structuring and partial homogenization proceeded in depressions. The basement subsided (mostly as a complex) and acquired the character of a homogeneous megablock — median massif. In the Pannonian Basin a higher degree of subsidence differentiation preserved up to the Quaternary time.

### 2.5.1. Pannonian Basin

The deeper basin structure is well explored by holes: there are 85 m per 1 sqkm (V. V. SEMENOVICH — JU. G. NAMESTNIKOV 1981).

The Pannonian Basin is in the position of an intermontane depression. Thickness of basin sediments is markedly variable in the central depression and in marginal depressions bordering the basin on the contact with orogenic margins (Fig. 37).

Miocene sediments range to the greatest thickness (above 1500 m) in NE Hungary. The partial depression formed in NW and SW Hungary (L. STEGENA et al. 1975). The maximal subsidence is in correlation to atypical continental-, formerly suboceanic crust (Fig. 38). Pliocene sediments are indicative of inver--sion. Maximal thicknesses are concentrated in areas with a reduced thickness in the Miocene: NW and W margins of Hungary, SE Hungary.

During the Quaternary time, in the western part of the Pannonian Basin (W of the Danube r.) the thickness of sediments was less than 100 m, and more than 500 m in SE Hungary. Main sedimentary basin axes in NW Hungary migrated from NW to SE — like along the Balaton line. In contrast to Miocene, the Quaternary main mobility axis migrated to NW.

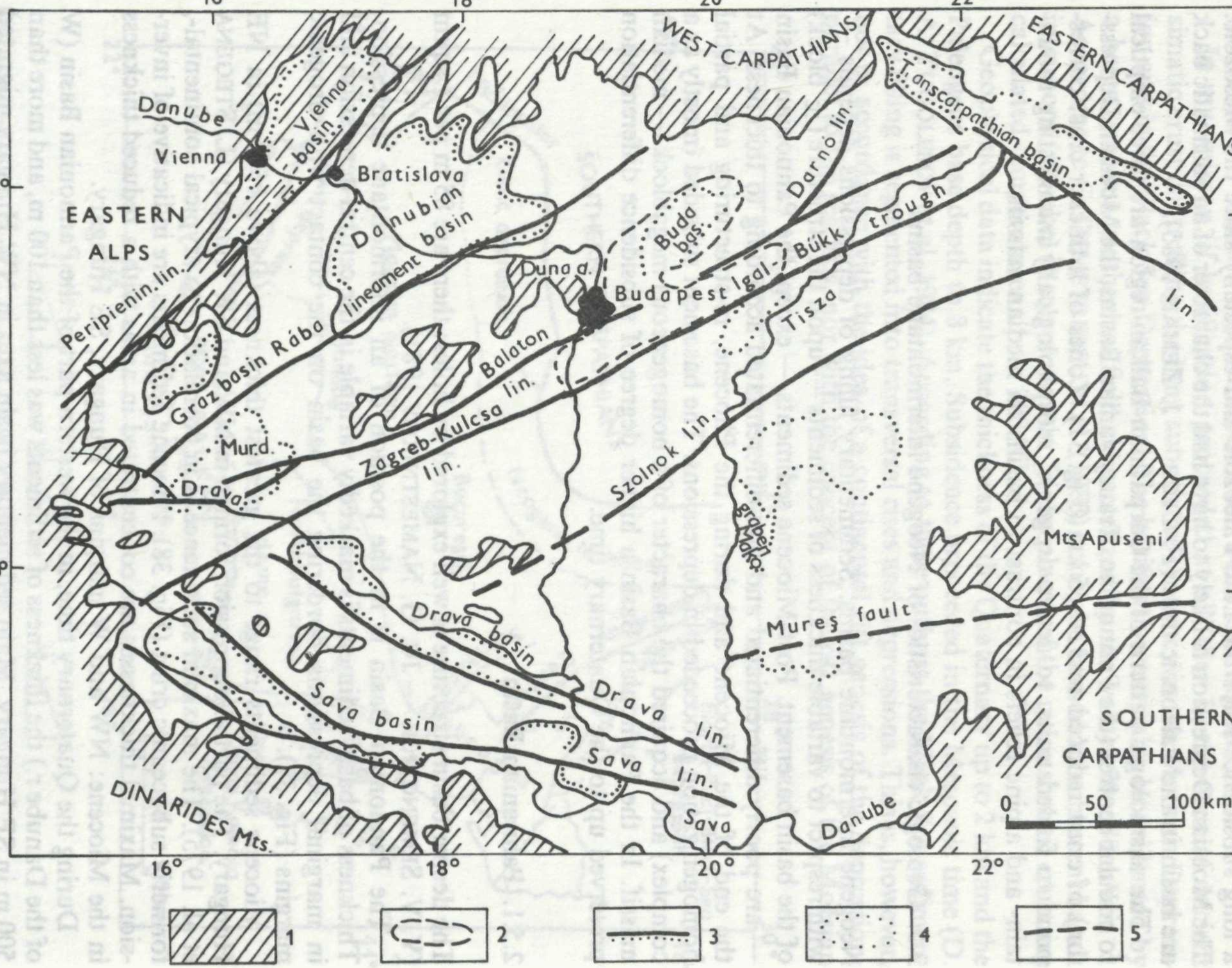

16°
18°
20°
22°
48°
46°
WEST CARPATHIANS
EASTERN CARPATHIANS
Danube
Vienna
Vienna basin
Bratislava
Danubian basin
EASTERN ALPS
Peripienin. lin.
Rába lineament
Graz basin
Duna d.
Buda bas.
Darnó lin.
Budapest
Igal-Bükk trough
Tisza
Transcarpathian basin
lin.
Balaton lin.
Zagreb-Kulcsa lin.
Mur.d.
Drava
Szolnok lin.
Makó graben
Mts.Apuseni
Mureş fault
Drava basin
Drava lin.
Sava basin
Sava lin.
Sava
Danube
SOUTHERN CARPATHIANS
DINARIDES Mts.
0
50
100km
1
2
3
4
5

Fig. 37 Main lineaments and marginal depressions of Pannonian Basin. [Modified according to M. BOCCALETTI et al. (1976), D. VASS (1979), F. ČECH and J. ZEMAN (1982)].
Explanations: 1 — elevations with pre-Neogene complexes, 2 — Paleogene Buda basin, 3 — marginal depressions, 4 — highly mobile segments inside the basin with Neogene-Quaternary sediments more than 2 km thick, 5 — deep-seated faults.

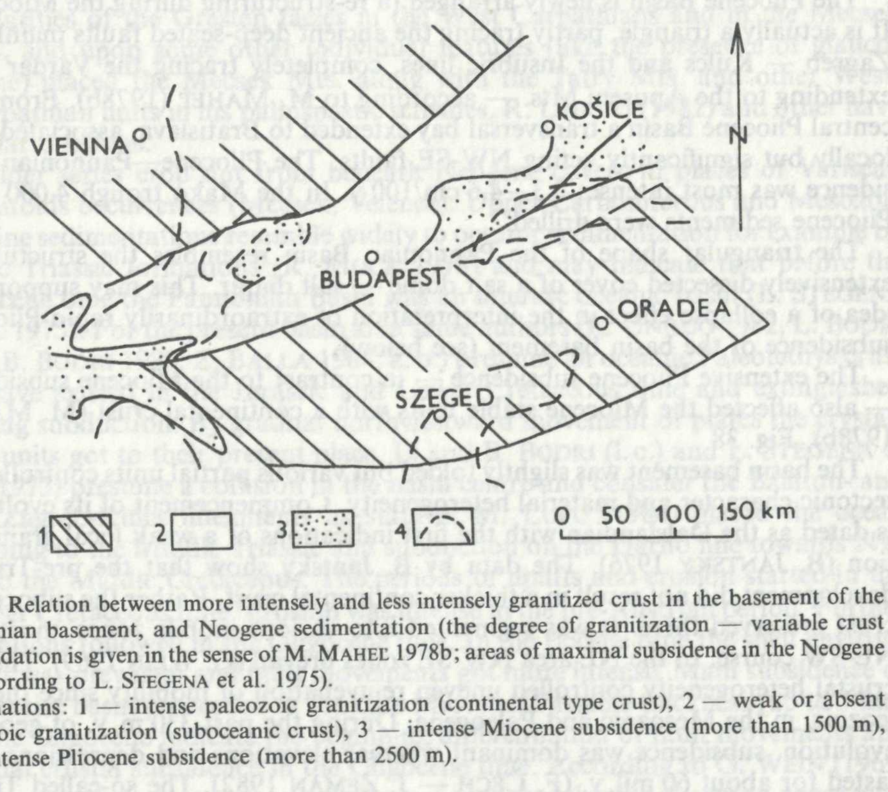



Fig. 38 Relation between more intensely and less intensely granitized crust in the basement of the Pannonian basement, and Neogene sedimentation (the degree of granitization — variable crust consolidation is given in the sense of M. MAHEĽ 1978b; areas of maximal subsidence in the Neogene — according to L. STEGENA et al. 1975).
Explanations: 1 — intense paleozoic granitization (continental type crust), 2 — weak or absent Paleozoic granitization (suboceanic crust), 3 — intense Miocene subsidence (more than 1500 m), 4 — intense Pliocene subsidence (more than 2500 m).

In the course of the Pannonian Basin evolution four main depositional stages proceeded: the Priabonian—Oligocene, post-Savian, post-Styrian and Pliocene.

The Paleogene stage of the basin development is typical of this area and proceeded only along the northern, eastern and south-western margins of the present Pannonian Basin. The post-Savian deepening only extended to the basin on the west — along the Peripieninian lineament and in a short bay along the Insubric line. This stage — in contrast to the preceding one — was not in an areal but in a linear association with the activity of largest lineaments. In the Ipeľská kotlina basin and in the Bzovík depression (part of the Buda depression) the deposition rate in the Upper Cretaceous — Eocene times was 0.14 cm per 100 years (D. VASS — F. ČECH 1983).

The evolution of post-Styrian basins is characterized by depressions extending inside the Carpathians and inside the Pannonian Basin. The depositional area expands southwards along the Vardar zone. The subsidence met an immense response in the Transylvanian Basin and in the Moesian Depression (Tab. 5). Whereas in marginal depressions the Miocene subsidence was very intense (5.6—10.9 cm/100 years, 2.1—7.8 cm/100 y in the Danube Basin — D. VASS and F. ČECH 1983) subsidence in the Pannonian Basin was slow (Tab. 5).

The Pliocene Basin is newly arranged (a re-structuring during the Miocene). It is actually a triangle, partly tracing the ancient deep-seated faults mainly the Zagreb — Kulcs and the Insubric lines, completely tracing the Vardar zone extending to the Apuseni Mts. — according to M. MAHEĽ (1978b). From this central Pliocene Basin a transversal bay extended to Bratislava, associated with locally but significantly acting NW-SE faults. The Pliocene—Pannonian subsidence was most intense: 2.3—4.6 cm/100 y. In the Mako trough 4.000 m of Pliocene sediments were drilled.

The triangular shape of the Pannonian Basin resembles the structure of extensively dissected cover of a salt dome — salt diapir. This may support the idea of a collapse effect in the interpretation of extraordinarily rapid Pliocene subsidence of the basin basement (see below).

The extensive Pliocene subsidence — in contrast to the Miocene subsidence — also affected the Miocene stable units with a continental crust (M. MAHEĽ 1978b), Fig. 38.

The basin basement was slightly folded but various partial units controlled its tectonic character and material heterogeneity. Commencement of its evolution is dated as the Dalslandian with the first indications of a weak local granitization (B. JANTSKY 1976). The data by B. Jantsky show that the pre-Triassic development did not result in a thicker continental crust. Rather the suboceanic crust preserved there. It was segmented into various mobile zones, mostly of NE-SW course. In the NE area NW-SE zones prevail (G. WEIN 1969). The old crustal heterogeneity controlled uneven rejuvenation of mobility since the Paleozoic, in the Mesozoic and Paleogene. During the past 330 m. y. of geologic evolution, subsidence was dominant whereas elevations and depressions only lasted for about 60 mil. y. (F. ČECH — J. ZEMAN 1982). The so-called Transdanubian Midmountains geosyncline or the Igal-Bükk riftogeosyncline resting on an Early—Paleozoic metamorphosed basement in the Permian time and evolving in the Triassic, are good examples. The Transdanubian Midmountains geosyncline is placed between the Rába line and the Balaton—Velence granite zone. Coal-bearing sediments occurred in the Upper Cretaceous (Coniacian) and lasted up to the Oligocene time. The Szolnok Flysch trough, mainly E of the Tisza r. where the trough estended from the NW part of the Transylvanian Basin, shows the first indications of volcanic activity since the Upper Cretaceous, associated with intense subsidence.

The long evolution of sedimentary troughs is in accordance with a presumably discontinuous thinner granite layer with a very small portion of granitoid rocks. This is why the Hungarian territory except continental sediments is characterized in its NE part by marine sediments also in the Carboniferous. There also is other evidence of unaccomplished Hercynian tectogenesis (M. MAHEĽ 1978b).

The basement of the Pannonian Basin is slightly affected by Cretaceous folding, partly rebuilt and occurs in the Transdanubian Midmountains, in the Mecsek and Villányi Mts. These two mountain ranges may represent a slightly folded part — a promontory — of the South Carpathians or of the Serbian-Macedonian Massif (M. MAHEĽ 1978a). I. VARGA (1978) basing upon facies similarities of the Gresten facies in the West Carpathians and in the Mecsek Mts., and upon some other individual features (like the presence of glaucophane) placed the Mecsek Mts. along with the Tatry Mts. and other West--Carpathian units in his palinspastic schemes. R. UNRUG (1982) and other have similar opinions.

Older series crop out from beneath Neogene covers in places of Variscan granitoids occurrences (Mecsek, Velence). Upper Carboniferous and Mesozoic marine sedimentations resemble widely to oceanic sedimentation for example by some Triassic formations (R. MOCK 1978) and may indicate that before the Miocene time the Pannonian Basin was an interarc oceanic basin (L. STEGENA et al. 1975). For the present basin area some authors (R. UNRUG 1982, L. BODRI and B. BODRI 1980, Z. BALLA 1984, a. o.) presume an oceanic Paleotethys crust to have existed in the Jurassic and Lower Cretaceous time and extinguished during subduction. By gradual northwestward movement of plates the crystalline units got to their present place. L. and B. BODRI (l. c.) and L. STEGENA et al. (1975) presume a collision in the basin centre and consider the Balaton- and the Zagreb-Kulcs lineaments as sutures. M. LUPU (1984) placed the ocean opening to the Middle Triassic and subduction on the Darnó line towards NW — to the Middle Cretaceous. The periods of uplifts and erosion started in the Upper Cretaceous time. Erosion was intense in the pre-Austrian period. Further elevations followed in the Oligocene time. In the Eocene time the shelf environment was prevalent and fault movements got more intense. Main subsidence of the entire basin proceeded in the Late Miocene. Variable thickness (up to 3 km) of the Oligocene indicate the beginning differentiation of fault movements and partial crustal subsidence in the Oligocene time. According to G. WEIN (1969) the movements started mostly in former troughs whereas elevation zones with granitoids resisted the subsidence. Activation of the Pannonian region was mostly controlled by the intensity of tectonic disturbation and crust character. The crust of suboceanic character was more liable to mobility than the crust with a thicker granite layer. Andesite volcanic activity ended in the Late Miocene and was followed by gradually accelerating subsidence.

Folding was unintense in most cases, even in the Paleozoic. The next folding proceeded as Late as the Cretaceous time. Pre-late Cretaceous deformations only proceeded along margins of variably mobile units. There was no further folding in the interval between the Late Cretaceous and the end of the Tertiary — except the Alföld Depression affected by unintense pre-Miocene folding. The sediment beds are mostly subhorizontal, only deformed by subsidence.

The structure of the southeastern margin of the Pannonian Basin is affected by NNW-SSE lines. The Kraishtide zone is tectonically deformed by faults and belongs among strongly compressed areas. The Serbian-Macedonian Massif horst with its Late-Cadomian folded basement represents another linear unit including the Struma zone. The massif represents there a scar-like median massif between the South-Carpathian east-vergent folds and the west-vergent Dinarides. The Serbian-Macedonian Massif is markedly compressed by neotectonic movements as well. The Vardar zone with elevations and subsidence faults also belongs among the linear zones. Along its elevations and subsidence faults (tectonic) graben of ensialic nature formed. The zone had a character of oceanic crust with many Jurassic ophiolite bodies up to the Eocene time when granitoid plutons appeared.

The Pannonian Depression is in tectonic contact with the Dinarides. It is indicated by considerably variable thickness of the Neogene. In the Outer Dinarides extensive carbonate sedimentation prevailed in the time from the Triassic up to the Eocene and since the Late Senonian it has been replaced by flysch sedimentation. In the Inner Dinarides the flysch sedimentation commenced in the Jurassic and terminated in the Cenomanian. Folding and elevations commenced in the Senonian time in this period. The Mesozoic sediments are 2—3 km thick, sediments in Cretaceous troughs a 3—8 km thick. Multiple overthrusts and imbrications caused increased thickness of sediments and basement subsidence to the depth of 6—10 km.

Basins along the Pannonian block margin differ from one another in individual tectonic units (M. MAHEĽ 1971b). Ancient basins mostly formed along longitudinal basement structures. Internal basins, closer to the Pannonian block resemble superimposed basins (T. BUDAY 1961). A comparison of data on the thickness of sediments and on the basin basement depth show that basins inheriting ancient tectonic lines are not more mobile than superimposed basins. On the contrary: tectonic lines alone, even if deep-seated, are not decising for subsidence intensity. They only control differentiated movements inside a basin or a block.

### 2.5.1.1. Deep-seated faults and block structure

The first systematic survey of deep-seated faults related to the Inner-Carpathian basins was presented by T. BUDAY (1961). In the past fifteen years further, mainly geophysical data on the deep-seated faults were collected: G. WEIN (1969), B. BERÁNEK et al. (1972), E. BONTCHEV (1971, 1978), B. SIKOŠEK (1976), M. MAHEĽ (1971a, 1971b, 1978b), O. FUSÁN et al. (1971, 1979), A. ZÁTOPEK — B. BERÁNEK (1974), I. VARGA (1978), P. GRECULA — I. VARGA (1979), J. ZEMAN (1978), unpublished data by B. BERÁNEK (1978), J. IBRMAJER (1978).

We may approach to the deep-seated fault concept from geophysical, geological or a complex viewpoint. Geophysical data frequently indicate deep-seated faults without geologic evidence of their deep character. For example B. BERÁNEK (1978) proves that the Peripieninian lineament in its SW branch is not identic with the Klippen Belt as defined by M. MÁŠKA and V. ZOUBEK (1961 in T. BUDAY et al. 1960), but according to seismic indications it is shifted to SE — to the western margins of the Malé Karpaty Mts. The lineament is by B. BERÁNEK (l. c.) extended to Verona, although there is no reliable geologic evidence of the surface course of the lineament in this extent. It is likely that the lineament only forms there in deeper zones and has so far not reached the surface (D. PROCHÁZKOVÁ and J. ZEMAN 1982). Because of different courses of the Klippen Belt and seismoactive deep zone, B. BERÁNEK (1978) introduced the term Pericarpathian lineament, and O. FUSÁN (1979) the term Záhorie — Humenné deep fault for the seismoactive deep fault.

The recently inactive Moravian-Silesian lineament extended in the Paleozoic time to the abrupt eastern termination of the Eastern Alps, according to J. ZEMAN (1978). But there is no geophysical evidence about this post-Paleozoic failed deep structure. It, however, has some structural influence upon orientation of some lines in the Vienna Basin (perhaps in the pre-Neogene basement) and in seismologic indications (D. PROCHÁZKOVÁ — J. ZEMAN 1982).

Many deep-seated faults were recorded in international DSS sections. Eleven deep faults were distinguished in Profile III between the Adriatic Sea and the East Carpathians (platform contact). In the Pannonian Basin basement no active deep faults are seismologically recorded, although their past existence is reliably proved by geologic evidence. There are no conspicuous physical indications of neotectonically (i. e. Miocene—Pliocene—Quaternary) inactive faults, or of faults whose activity extinguished after the end of volcanism. My opinion is based upon the fact that most faults recorded by DSS are in active areas. The only exception are pre-Miocene neotectonically inactive faults separating crust of various types and thicknesses, preserved up to now including the preserved Moho displacement. The deep-seated Száva fault and faults, rimming flysch troughs, are good examples (F. ČECH — J. ZEMAN 1980).

The fact that thinned crust beneath the Pannonian Basin extends to marginal folded units, is significant for the deep faults evolution. The positive gravitational field also extends to the marginal folded units and the most mobile Neogene basins formed there. The thinner/thicker crust boundary is associated with seismoactive deep faults bordering geologically (Neogene) and physically different basement units.

Among significant tectonic lines on the Pannonian Basin margin is an arcuate crustal fault — the Hurbanovo fault (O. FUSÁN et al. 1971), separating the Slovak megablock from the Pannonian megablock in the Neogene basement. The Hurbanovo fault connecting marginal lineaments inspired the concept of the Peripannonian lineament (V. G. SVIRIDENKO 1976, 1978).

The Pannonian Basin margin is also bordered by the Számos line parallel to the eastern branch of the Peripieninian (Záhorie — Humenné in the sense of O. FUSÁN et al. 1979) lineament. The Számos line extends to the Romanian Carpathians. The seismoactive Rába line is also in marginal position to the

Pannonian Basin. The Rába line either branches in its NW part or is associated with parallel faults. Some authors also join it with neovolcanic active faults of local importance (L. ZBOŘIL et al. 1971) which need not be deep-seated faults.

Soviet geologists regard also the Rožňava fault as part of the Peripannonian lineament (V. G. SVIRIDENKO 1976). The Rožňava fault conformable to the Pannonian Basin margin might have resulted from the connection of ancient tangential faults in the Neogene time. The fault border of the Pannonian Basin shows some features of polygonal or hexagonal basin shape, according to V. S. ZHURAVLEV (1972).

The basin basement is most disturbed by deep faults in the northern and western parts. The southern part is missing the peripheral fault features. One of most significant faults — the Vardar fault — extends to the basin centre and has morphotectonic influence on the Danube r. bent near Beograd and on the lower course of the Tisza r. E. BONTCHEV (1971) joined this deep fault with the Kraishtide zone into the broad Kraishtide-Vardar lineament and regarded it as a continuation of the Labe lineament. This is in accordance with P. BANKWITZ'S et al. (1977) opinion. In contrast to this, B. SIKOŠEK (1976) means that up to the Variscan tectogenesis no Vardar zone existed and the W-E dissection prevailed in the area. The lineament is neoid, does not extend to Szeged and is not joined to the Labe lineament. This SIKOŠEK'S opinion agrees with V. KÁRNÍK'S (1975) view based on recent seismic activity. Kárník admits possible gradual displacement of activity towards Budapest in the past. The "discontinuity" also follows from the determination of the Labe lineament by O. FUSÁN et al. (1979). (They denoted its equivalent in the West Carpathians basement as the Přerov — Štiavnica fault ceasing to extend southeastwards of Banská Štiavnica).

V. N. GERGELCHEV et al. (1977) regard the Vardar zone, the Struma zone (the so-called Struma-Vardar rift zone) and the Prilep-Euboya zone as rifts connecting the Balaton rift with East-African rifts. I have already mentioned that there are no geophysical, nor geological evidences on the Hungarian territory. There is a connection among structures of different course, tectonic nature and genesis. The rift concept, including the Balaton Lake, contradicts to data on thicknesses of Pliocene and Quaternary sediments (L. STEGENA et al. 1975).

Another N-S deep fault (I denoted it as the Danube fault) begins in the basement of the Central Slovakian neovolcanic region and is parallel to the Danube r. in Northern Hungary and ends as a deep fault on the Balaton line. The fault arose most likely before the Miocene time but its deep activity in the area of Tertiary volcanism is Miocene and lasts up to the present (seismic indications). J. ŠTOHL (1976) denoted it as the Central-Carpathian lineament. The fault — or the surface fault zone — has other names, like the Zázrivá — Revúca — Budapest fault (D. KUBÍNY 1962). I regard my term (the Danube fault) as a preliminary name for the Late Tertiary volcanic-active structure. On the surface it continues southwards parallel to the Danube r. course. Their past deep connection in this segment may be indicated by N-S basic and ultrabasic bodies (I. HORVÁTH and L. ODOR 1984).

The Danube fault is associated with the parallel Tisza fault controlling probably the Tisza river (Fig. 34). The fault is not evidenced as a deep-seated one, it is only traceable in the morphologic pattern. In my opinion the two faults delimitate a young graben. I regard it as tectonically young because it cuts the Central-Hungarian upland which is recently not too active and because it was bordering a horst until the Pliocene (thinned Pliocene sediments). The graben is a morphologically prominent young structure as proved by satellite images as well (L. TRUNKO 1977).

Miocene-Quaternary isopachyte maps presented by L. STEGENA et al. (1975) show the role of deep faults in crustal restructure (geodynamic inversions) proceeding in the Pannonian Basin. The maximal Miocene subsidence proceeded in the eastern part of the Balaton block, mainly between the Balaton and Darnó lines. A partial depression formed NW of the Rába line, a linear depression also along the Insubric line. The zone joining on NW the Zagreb—Kulcs and Balaton lines was most mobile.

Mobilization of the Rába proceeded in the Pliocene time (more than 2500 m thick sedimentary formation in the Danube Basin). The maximum subsidence displaced to the east of Danube (East-Pannonian block; M. MAHEĽ 1978b). The Pannonian Basin disintegrated in two stable elliptical basins separated from the mobile depression by an elevation horst-like structure bordered by the Danube- and the Tisza faults. In the Quaternary time only the East-Pannonian block was mobile.

In the Pannonian Basin basement are almost inactive deep faults Darnó and those parallel to the Számos line. The NE part of the Balaton line was partly active in the Quaternary time. The Hornád fault is rejuvenated. It is not sure whether it is a deep fault in its entire course (I. VARGA 1978). Its deep character on the Slovak territory is proved by O. FUSÁN et al. (1971).

Among active larger lineaments the Insubric line extends to the Pannonian Basin and causes faults mostly in the basin basement. Neotectonic activity on the contact with the Vardar zone, is indicated by the changing course of Danube, conformable to both discontinuities (Fig. 34). M. MAHEĽ (1978b) regards the Darnó line as a NE-SW branch of the Insubric line. This may indicate the pre-Neogene structure of the deep line, now without seismoactivity which is likely to have passed to the main trunk of the subequatorial structure.

I shall mention some neotectonically slightly active lines — strongly active in pre-Mesozoic time —, which controlled structural division of the Pannonian Basin crust. They are the Balaton line — gravimetrically prominent as well, and the Zagreb—Kulcs line. SE of the line is the Lóczy ridge which underwent neotectonic subsidence.

All the lines are NE-SW striking. According to G. WEIN (1969) the Balaton- and the Zagreb—Kulcs lines bordered the so-called Igal-Bükk eugeosyncline which is rather a rift geosyncline or a lineament — geosyncline in the sense of E. BONTCHEV (1976).

L. KÖRÖSY (1983) introduced new terms for some deep faults, e. g. Zagreb—Kulcs fault is denoted as Inke-felsötisza line (High Tisza lineament). He ascribed deep faults an older age and dynamics associated with lithospheric microplates movement. L. KÖRÖSY (l. c.) regards the Balaton line linking to the Darnó line as the extension of the Periadriatic lineament (cf. G. WEIN 1978) and emphasizes different structure and mobility of individual units (blocks in my opinion).

Division of the pre-Miocene basin basement, and the division into troughs and ridges remind of subparallel faults on the sea floor. M. MAHEĽ (1978b) also mentions deep NW-SE faults running out of the Bohemian Massif. It is, however, difficult to prove their existence in the covered Pannonian Basin basement.

Ancient and young deep faults dissect the crust in the northern and partly the western parts of the Pannonian Basin into polygonal blocks (F. ČECH — J. ZEMAN 1980). Faults with variable tectonic and seismic activity — except the Paleozoic-Mesozoic active faults — form irregular block mosaic controlled by variable courses of deep faults. The rather dense segmentation indicates heterogeneity of the Pannonian Basin crust and its marked deformation in the western part. The deformation is less intense in the eastern part but elevations at the basin margin (Fig. 30 and 34) are more intensely segmented by faults.

Variable deformations may also indicate different structure of the eastern block (the eastern Pannonian block, M. MAHEĽ 1978b). The NE-SW course of main faults and structural units changes eastwards into ENE-WSW course. Faults of the northern part even change their course to W-E (the Bükk Mts., the eastern part of the Szendró Mts., M. MAHEĽ 1978b).

The changed structural courses inspired V. DANK (in V. DANK — J. BODZAY 1971) to the concept of the secondary Alpine arcuate structure of the Pannonian Basin basement, namely in the continuation of the Eastern Alps structure following the Carpathian arc. This is a hypothesis because eventual W-E division should be pre-Paleozoic (J. ZEMAN 1981), with a dividing line on the Danube river. B. SIKOŠEK (1976) described similar pre-Variscan segmentation of the Vardar zone and adjacent areas. In the eastern block a Late Mesozoic flysch basin formed. In the basin the extent (width) of tectonic-paleogeographic units (e. g. the Lóczy's ridge) changes with the changing course of the basin. In SE Hungary arcuate structural elements appear (e. g. Battony; the bent of the SE crystalline ridge — G. WEIN 1969). The block has been mobile since the Pliocene inversion, G. WEIN (1978) in his last tectonic model presumed Alpine movements and nappe structure.

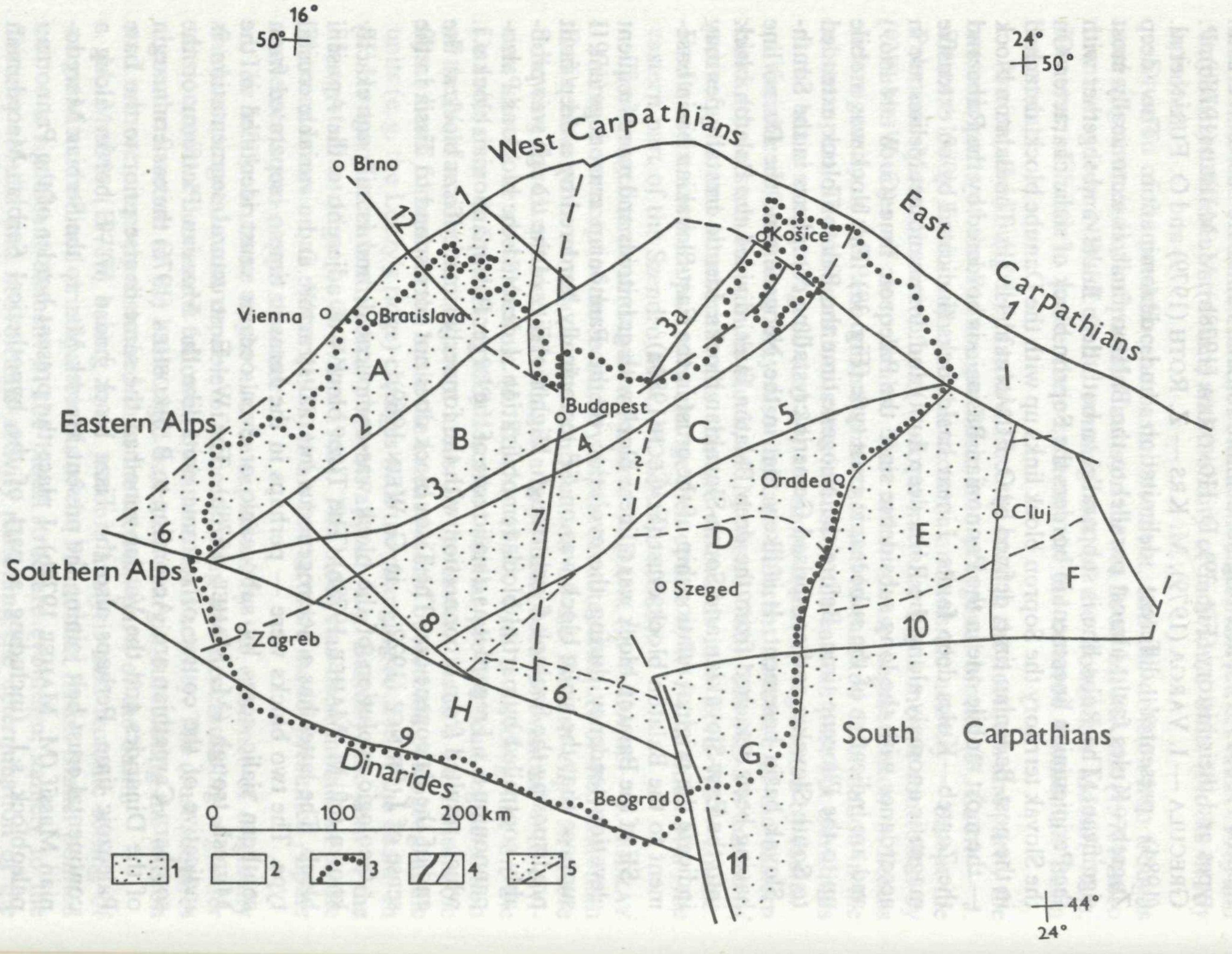
The block structure might also be explained by the existence of originally thick continental crust capable of concentrating and accumulating a greater tension, whose decrease is followed by intense fault activity. The crust acquires there the function of the basin rigid basement margin cumulating the tension and is more intensely disturbed than the internal parts of the basin.

M. MAHEĽ (1978b) basing upon tectonic analysis and tectonic setting of magmatites, outlined principal blocks of the Pannonian Basin. Maheľ pointed out to a significant element in the block structure evolution, namely to the Tertiary rebuild and neovolcanic activity reasoning the determination of the large crustal block Neotissia in the sense of J. SLÁVIK (1974).

Pre-neoid NE-SW deep faults are distinguished in the basin basement (J. Zeman 1980, 1981). Their original delimitation and names are based on crustal types and their history (Fig. 39). D. Hovorka (1978b), M. Maheľ (1978b), P. Grecula — I. Varga (1979), M. Krs — Z. Roth (1979) and O. Fusán et al. (1984) presented different delimitation and denomination. The deep Zagreb—Kulcs fault, almost parallel to the Balaton fault, is tectonically most significant. The Rába line is subparallel to both deep faults, and together with the Peripieninian lineament in borders the Sopron block of sialic character. On the Slovak territory the Sopron block links up with the Danube block, detached in the post-Badenian time, defined by O. Fusán et al. (1971). The Balaton block — the most mobile one in the Pannonian Basin — is bordered by the Rába and the Zagreb—Kulcs deep faults. Linear basic rocks (indicated by an extensive magnetic anomaly along the Rába deep fault) and dominant metapelites are in accordance with the long subsidence since the Paleozoic time (G. Wein 1969) and are indicative of the suboceanic crust type (Fig. 39). The block was mobile up to the Pliocene time. Before the Neogene time the Balaton block extended to South Slovakia and comprised Gemeric crystalline complexes in the South--Slovak basin basement. It is likely that in the Neogene time the Darnó line having been detached from the deep Balaton fault, divided the Balaton block into the East-Slovakian and South-Slovakian blocks. Identic crustal types may indicate the original relationship of the crust in the East-Slovakian basin basement to the Balaton block crust (F. Čech 1980a).

SE of the Balaton block was a linear block with granitoids and more frequent elevation tendency during the evolution of the Pannonian crust segment. I suppose that the sialic block was on SE tectonically bordered by a deep fault predisposing the Szolnok flysch trough. High mobility of the trough was probably controlled by partial block remobilization. I denoted the block as Paleopannonian s. s. I regard it as a sialic core of the larger Paleopannonian block s. l. which resulted from a connection with the formerly simatic Tisza block at the end of the Paleogene era. The Tisza block does not correspond to Tissia in the sense of J. Princz (1922 — in G. Wein 1969).

Geologic evolution of the block was complicated, and is still not exactly known (cf. M. Maheľ 1978b). The Tisza block was adjacent to the Apuseni block. The latter has a heterogeneous internal structure and a variable crustal type. The two blocks were — perhaps in the Jurassic time — separated from southern sialic cores by suboceanic or even oceanic crust identified in the Muresh trough (M. Maheľ 1978b). The W—E structural segmentation is indicative of the course of tectonic units like the Moesian Platform or the southern Carpathian arc. According to B. Sikoshek (1976) the eastern margin of the Dinarides and the Vardar zone had the same course prior to the Late Paleozoic time. Perhaps also the Tisza block had a W—E border along a continental crust belt joining the present Mecsek Mts. to the Serbian Macedonian Massif (M. Maheľ 1978b). I place the present border of the Pannonian paleoblock s. l. (including a part of the hypothetical Serbian-Macedonian block) on the Insubric line or on the deep Dráva fault which is likely to have


16°
50°
24°
50°
West Carpathians
East Carpathians
Brno
Košice
Vienna
Bratislava
Eastern Alps
Southern Alps
Budapest
Oradea
Cluj
Szeged
Zagreb
Beograd
South Carpathians
Dinarides
A
B
C
D
E
F
G
H
I
1
2
3
3a
4
5
6
7
8
9
10
11
12
0
100
200 km
44°
24°

detached in the Neogene time as a separated structure predisposing the graben like Dráva depression.

The blocks, including the Apuseni and the Transylvanian blocks, were on the NE bordered by the Számos line — formerly a significant lineament, evaluated by H. STILLE (1953).

The block structure was rebuilt approximately in the Paleogene time when the Pannonian Basin megablock started to form. The megablock is a new deep tectonic unit including the southern part of the Inner West Carpathians and SW margin of the East Carpathians — Neotissia of J. SLÁVIK (1974). The new structure comprises various units with different crust types. So the older term "Tissia", based on the idea of a homogeneous block, is problematic.

My considerations about the origin and evolution of Neogene basins are based on the following assumptions:

a) Since the Tertiary period, crustal segmentation into variably mobile blocks or zones controlled by variable extent and intensity of Paleozoic granitization, has resulted in different crustal consolidation. In this region the consolidation process has not reached the platform stage;

b) repeating mobility of slightly Hercynian-consolidated units;

c) Tertiary, mainly Neogene restructuring with preferred subsidence in mobile units. Local formation of overthrusts and nappes in compressed depressions is not excluded.

With respect to dynamics the following facts are significant:

a) Faults (including deep ones) mostly concentrate along basin margins.

b) Two structural-fault plans denoted by myself as the Paleozoic-Mesozoic and the Tertiary plans (cf. M. MAHEĽ 1971a — paleoalpine and neoalpine fault groups).

c) Changes in tectonic regime by the end of the Paleogene — Lower Miocene times, later since the Pliocene time, inversion in subsidence dynamics.

d) Seismoactive segments indicative of recent deep movements on faults and their continuous evolution.

e) Concentration of deep neotectonic activity in the northern part of the basin, and of the upper-crustal activity in the eastern part of the basin in accordance with intense fault activity of the Central Carpathians.

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Fig. 39 Block structure of the Pannonian Basin.
Deep-seated faults and lineaments: 1 — Peripieninian, 2 — Rába-Vepor, 3 — Balaton (3a — Darnó), 4 — Zagreb-Kulcs, 5 — Szolnok, 6 — Periadriatic (Insubric), 7 — Danube (Central-Carpathian), 8 — Dravian, 9 — Savian (Peridinaric), 10 — Muresh, 11 — Vardar, 12 — Nesvačilka —Trnava, 13 — Számos.

Blocks: A — Sopron, Danube, B — Balaton, C — Paleopannonian s. s., D — Tisa (C + D Paleopannonian s. l.), 4 — Apuseni, F — Transylvanian, G — Serbian-Macedonian, H — Foredinaric, I — East-Carpathian.
Explanations: 1 — sialic blocks, zones of Variscan consolidation and sialization, 2 — simatic blocks consolidated at the end of Cretaceous time and in the Tertiary, 3 — boundaries of the Pannonian Basin, 4 — main deep-seated faults and lineaments.

With respect to tectonic development of the study area I paid most attention to the two structural-fault plans (b) and to the activity concentrations (e).

The study of structural and paleotectonic effects of deep faults and their courses shows that pre-Neogene NE-SW faults inside the Pannonian Basin lost their deep activity at the end of the Miocene time. V. G. SVIRIDENKO (1976) regards faults of this course as structurally live also after the Pannonian period and delineates there the transversal Pannonian-Volhynian depression in the area of Uzhgorod — Rachov in the East Carpathians, the Miskolc — Debrecen depression up to the western margins of the Apuseni Mts. in the Pannonian Basin. The transversal depression shows, however, seismologic features (D. PROCHÁZKOVÁ — J. ZEMAN 1982). Among NE-SW faults only those near the basin margins were active in the Neogene time. They were the Rába line, the Darnó line (inactive after the Ottnangian — L. TRUNKÓ 1977), and faults associated with the lines (cf. L. ZBOŘIL et al. 1971). I have already mentioned active faults in tangential relations to the Pannonian Basin.

Since the Late Miocene time the new N-S faults had been active even up to the Quaternary and are still active in recent movements (J. KVITKOVIČ — J. PLANČÁR 1979). A correlation of data on recent seismic activity and tectonic fault activity (Fig. 4) shows that the N-E Danube and Hornád lines ended as deep structures on the ancient Zagreb — Kulcs discontinuity (in accordance with G. WEIN 1969 and in contradiction to T. BUDAY 1961). They only continue southwards in the upper crust in the form of increasing upper-crustal discontinuity owing to tensions resulting from uneven downwarping of the Pannonian block since the Pliocene time. The discontinuities are still living in upper crustal levels — as proved by shallow seismic focuses. This opinion of mine is also based upon the fact that no of the faults S of the Zagreb — Kulcs line served as an ascending way for Neogene volcanism and is seismically inactive (Fig. 34). The area of deep Neogene activity was in the northern part of the Pannonian Basin in the linear Balaton block and in the Central Carpathians where many faults formed or ancient lines got revived and the largest Štiavnica — Kremnica volcanic centre formed. In contrast to other units, the West Carpathians have most faults of variable courses, as already mentioned by M. MAHEĽ (1971b). The central part of the basin was in the sphere of a decreased tension.

### 2.5.1.2. Genesis of Pannonian Basin

The origin of the Pannonian Basin was paid attention by geologists and geophysicists in the past decade. There are several new models of the basin genesis.

1. The existence of normal faults always inspired the basin genesis explanation by tension. L. STEGENA et al. (1975) — Fig. 11 explained the basin genesis or better say the mantle diapir genesis — by convergent subduction of two hypothetic plates owing to compression in lithosphere. Dynamics of this model was discussed in Chapter 1.5.1. Z. BALLA (1982) presented a new variant: subduction first from the south, in the Eocene also from the north, of presum-

able North-, South- and Central-Pannonian microplates. According to Z. BALLA (1982) the basin basement is a lithospheric block mosaic resulting from Neogene convergent microplates movement and from movements on transcurrent faults. Relics of oceanic sediments should be in the flysch Szolnok trough. The Pannonian Basin was regarded as an interarc basin (L. STEGENA et al. 1975) or a back arc basin (J. MÉSZÁROS, personal information 1982).

2. Tertiary volcanism inspires a question about the relation between subsidence and empty magma chambers in asthenosphere. If the effused lava volume in the Pannonian basin was 200.000 km$^3$ (including non-ascending lava — F. HORVÁTH et al. 1981), then the sediment volume exceeded the lava volume several times. (Only pre-Sarmatian sediments of marginal depressions of the Pannonian Basins have the volume 124.000 km$^3$). The relation between volcanism and subsidence cannot be considered.

3. V. V. BELOUSOV (1982) presumes thin crust basification. According to him the crust is in the initial basification stage and the Pannonian Basin should change into a sea. E. V. ARTYUSHKOV — A. M. BAER (1983) explain subsidence by eclogitization.

4. The W-E lithosphere stretching is assumed in accordance with P. D. MC KENZIE'S (1978) model. Strike-slip faults formed in the upper brittle lithosphere along W-E and NE-SW faults and box-formed depressions of the pull-apart basin type formed on these faults (F. HORVÁTH — L. ROYDEN 1981), Fig. 40. According to this model also the Pannonian Basin marginal depressions should represent the pull-apart basins. But the authors only accept movements on faults oriented appropriately for their model. They do not consider movements on all NW-SE, NE-SW faults (e. g. the Darnó line) and on faults of different courses, like on the Vardar lineament. The authors do not deal with the origin of partial horst structures between some depressions. The eastward strike-slip faults cause the lithosphere stretching by at least 50 per cent.

The lithosphere subduction from E beneath the East Carpathians should attract the overriding lithospheric plate beneath the Pannonian Basin. L. ROYDEN et al. (1982), Fig. 10, presume that the Transylvanian Basin resulted from gradual bending and decreasing dip of the subducted plate and from suction of the lithosphere in the foreland of the bending subducted plate. The model is curious, however, without any geologic evidence. It was discussed by F. ČECH — J. ZEMAN (1984).

In subduction models the origin of the Pannonian mantle diapir is explained by passively upwelling mantle rocks owing to the stretching and thinning of lithosphere or laterally bulged elevation by convergent subducting plates.

### 2.5.2. Transylvanian Basin

The basin with gas-bearing structures was explored by 1.100 boreholes with the maximal depth of 4.533 m (Filitelnik hole) in the central part of the basin.

The Transylvanian Basin is between the internal East-Carpathian units (the

Danube autochthonous unit with Getic overthrusts) and the Apuseni Mts. elevation. Some Hungarian geologists join the elliptical basin with the Pannonian Basin into the so-called Carpathian Basin (e. g. G. WEIN 1969). I regard it as an independently evolving basin, as indicated by different tectonic structure and age of sedimentary fillings. The basins were related to each other in their deep dynamic evolution. This opinion of mine is also based on differences in crustal thicknesses of the Apuseni Mts. and of the basin. The Transylvanian Basin basement is not exactly known. According to drilling data the basin basement consists of Triassic limestones (N. ONCESCU 1959). A more intense subsidence in the basin commenced at the end of the Cretaceous time (about 640 m in the central part, about 1350 m thick sediments at margins). The structure of adjacent units, mostly the Apuseni Mts. indicate that in the base-

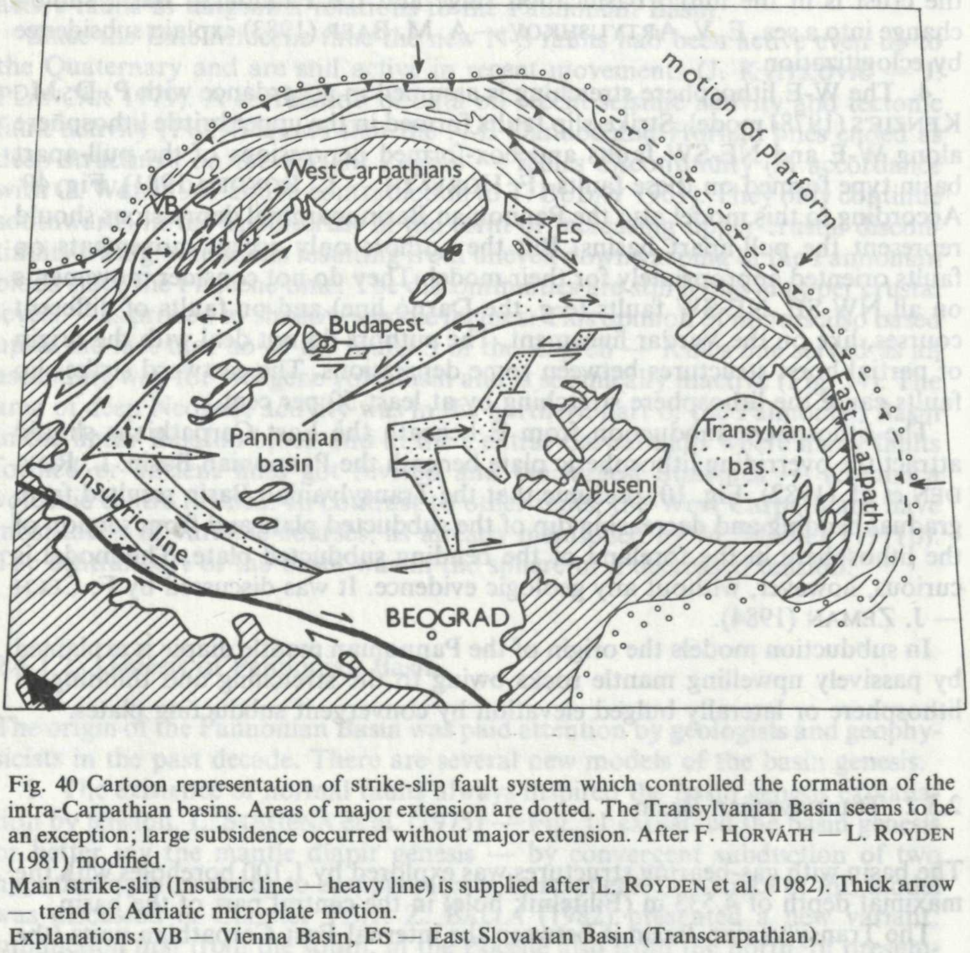



Fig. 40 Cartoon representation of strike-slip fault system which controlled the formation of the intra-Carpathian basins. Areas of major extension are dotted. The Transylvanian Basin seems to be an exception; large subsidence occurred without major extension. After F. HORVÁTH — L. ROYDEN (1981) modified.
Main strike-slip (Insubric line — heavy line) is supplied after L. ROYDEN et al. (1982). Thick arrow — trend of Adriatic microplate motion.
Explanations: VB — Vienna Basin, ES — East Slovakian Basin (Transcarpathian).

ment of the northern part of the basin is the Bihor crystalline complex (Precambrian) joining the Rodnai block in the East Carpathians. The two crystalline zones are in the area of Baia Mare separated from each other by the deep Számos fault. It is likely that the basement of the basin centre consists of the Muresh trough ophiolite belt also indicated by the positive gravity anomalies. N. HERZ and H. SAVU (1974) presume the existence of the Mesozoic oceanic crust beneath the basin. The rest of the oceanic crust might have been trapped by the Carpathian subduction and preserved in the basin. In the southern part with the thickest crust — 35 km the submerged Precambrian crystalline complex of the Getic unit may be presumed. The complex might have formed a separate block, divided from the Serbian-Macedonian block by the Kraishtide lineament. The orogenic crust in the NE and S rims of the basin is 38—55 km thick.

The sedimentary filling of the Transylvanian Basin consists of Eocene—Pliocene sequence ranging up to 500—1500 m in thickness at the basin margins, and up to 6.500 m in the central part and along the Tirnava Mica river. Thickness of Tertiary sediments is indicative of gradual subsidence, increasing from basin margins to its centre. In the Paleogene time the maximum thickness of Eocene sediments was 1000 m, of Oligocene sediments — 1300 m (G. N. DOLENKO 1962). The Eocene formation was reached by a hole in the structure Saros below a 3900 m thick Upper Miocene (beneath the Dej tuff). This area was uplifted in the Oligocene — Ottnangian time. The Miocene is 3.200—6.400 m thick. Actually, the basement mobility in the Neogene and the Paleogene times had an oscillation character. In the areas of great Sarmatian and Badenian subsidence in the basin's southern part by the maximum subsidence of the basin's northern part in the Lower Miocene time preceded. The most intensive Paleogene subsidence proceeded in the northern part of the basin. The maximum subsidence axis migrated from N to S during the basin evolution. It was associated with synsedimentary formation of the dome, and the Muresh fault controlled localization of the southern mobile depression in the Pliocene time. According to geophysical data the basin basement in its central part is at the depth of 8.000 m (V. V. SEMENOVICH and YU. G. NAMESTNIKOV 1981); Fig. 35. Pre-Badenian sediments in marginal depressions range up to 3.000 m in thickness; Lower-Badenian sediments are 1.800—2.000 m thick whereas in the central part of the basin they only range to 900 m in thickness. At the end of the Miocene the northern and the eastern parts of the basin started to uprise — evidently during volcanic activity along the basin margins. Pliocene sedimentation did already show freshwater lacustrine character (N. ONCESCU 1959). Variable thickness of sedimentary filling shows that the subsidence regime was like in the Pannonian Basin and its marginal depressions. The sialized crust affected slow subsidence in the northern and southern parts of the basin, whereas the simatic crust controlled the maximum subsidence in the basin centre. The axis of the maximum depression follows the basic crust course. Although the basin basement since the Tertiary time had been behaving as one block (Transylvanian), crust heterogeneity controlled differential movements and variable subsidence. Different thicknesses and the course of Tertiary is-

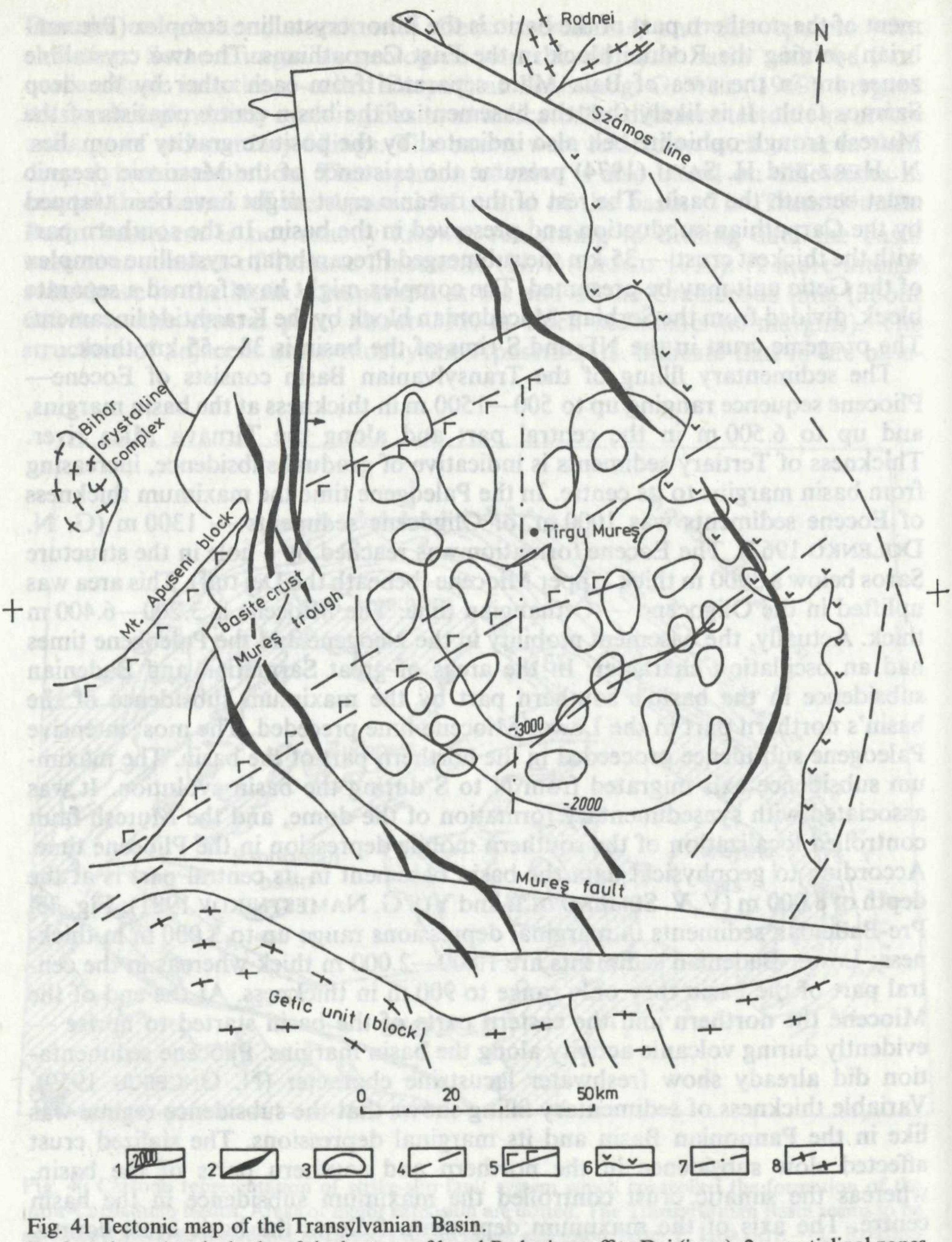



Fig. 41 Tectonic map of the Transylvanian Basin.
Explanations: 1 — isobaths of the bottom of basal Badenian tuffite Dej (in m), 2 — anticlinal zones of diapir folds, 3 — domed structures with natural gas, 4 — assumed concealed faults, 5 — margin of basite crust, 6 — margin of neovolcanites, 7 — symmetry axis of distribution of gas deposits — axis of basin curve, 8 — crystalline complexes; based on tectonic map of Romania (1970).

opachs enable reconstruction of faults inside the basin. Their course is indicative of variable crust types (Fig. 41). Fault tectonics in the basin has so far not been valuated exactly (cf. I. DUMITRESCU — M. SANDULESCU 1970). Boreholes and geophysical researches revealed pre-Badenian thrusts of crystalline complexes over Mesozoic formation (Fig. 35).

The tectonic structure is characterized by diapir structures. Besides the marginal zone dipping at 5—15° to the basin centre there also is a diapir fold zone bordering the central area of domal structures (G. N. DOLENKO 1962). The diapir fold zone forms two arcuate longitudinal belts along the periphery of the basin. In the southern part the folds follow faults extending there from the underlying Getic block. The folds do not follow the basin margin course. In some fold axes are salt diapirs, mainly in lines of the presumable Transylvanian block margins.

The central part of domal structures consists of 67 cryptostructures, testified by geophysical and drilling explorations. The presumable N-S axis divides the structures into the western and the eastern areas forming limbs of the N-S elongated elliptical depression. The axis of symmetric domes pattern is on the line between the towns Tirgu Muresh and Dumbraveni. Limbs of domal structures are flat, dipping at angles up to 10°. The division into two areas is more conspicuous and tectonically more significant than the usual division into three regional structural groups (cf. G. N. DOLENKO 1962; some Romanian geologists). Hidden low-amplitude domes also belong among salt structures with the initial diapirism stage.

The diapir movements are associated with pre-Pliocene and Pliocene movements and with general sea regression changed by Pliocene lacustrine sedimentation. Diapir uprise and folding are associated with post-Pliocene activity — the Rhodanian folding phase. From this period also earth gas accumulations in anticlinal structures originate.

Besides models discussed in Chap. 1.4.1.1, the Transylvanian Basin origin is usually associated with orogenic movements in the Carpathians. In contrast to the Pannonian Basin more data for the mantle diapir interpretation are missing. The existence of thin crust unconsolidated by the Hercynian event is presumed from historical viewpoint. If we admit the existence of ancient partial diapir, then its influence must have been weak, it must have originated in the Paleogene, and its main uprise proceeded in the Lower Miocene time. Neither diapir, nor crustal basification got to higher levels, so they did not affect processes in the basin by a higher heat flow. Since heat transporting media, like water and hot gases were missing, the higher temperature could not pass from 30 km depth to the upper crust and/or to the sedimentary basin filling by the normal heat conduct. According to thermodynamic calculations (VERHOOGEN et al. in V. ŠKVOR — J. ZEMAN 1976) a million of years is necessary for the heat transport by radiation for 1 km of the rock plate thickness. The time since the Miocene is rather short for high temperatures to get to the basement/sediment boundary in the basin.

The basin evolution shows that in the Pliocene the diapir was inactive and if no, then its effects ended by volcanism. Its collapse stage in the central part commenced in the Miocene. The diapir spatial extension was limited by a thicker sialic crust and by the size of the Transylvanian block enclosed by ancient peripheral massifs on a smaller area. The existence of autonomous mantle diapir is indicated by different subsidence intensities in the marginal and central parts of the basin. In areas bordered by more rigid units of the deeper structure the spatial mantle diapirism evolution was retarded and the activity is likely to have ended prior to the culmination of the Pannonian diapir activity. Owing to a greater crustal heterogeneity and its more intense segmentation into blocks, the Pannonian diapir was dynamically and thermally more active, and its evolution was longer than in the Transylvanian Basin.

The origin of salt diapirism may be controlled by the overlying sediments load, by fault-disturbation of layers, and mainly by the presence of the salt layer, absent in the Pannonian Basin. The uprise of the basin orogenic ring resulted in compression and overthrusts of the basin filling, and triggered the diapir mechanism. Uplift of the basin margins was also supported by the continental crust thickening in young volcanic zones. The greatest geostatic pressure concentrated in the basin centre where the most extensive basement movements proceeded in the Miocene time. This was followed by light- and plastic salts escape into the decompressed zone near the surface. The mechanism might have caused gas migrations in Sarmatian horizons and in the Oligocene formations affected by the most intense pressure.

On the basin margins, in the place of the Transylvanian block marginal faults the layers were most compressed and a diapir fold zone with overthrusts formed there. The fold zone is indicative of areal limitation of horizontal pressure effects. Deformations also prevented pressure effects in smaller depths, so the horizontal pressure was less intense inside the basin. In the basin centre the pressure was intensified owing to the sediment load.

The Apuseni Mts. are neotectonically separated from the Transylvanian Basin. According to M. MAHEL (1978a) since the Late Cretaceous they are a single unit which cannot be associated with the original continuous continental microplate together with the West Carpathians and the Mecsek Mts. in Hungary. The Apuseni Mts. have an E- and S-vergent nappe structure. They are separated by the Muresh ophiolite zone from the South Carpathians. M. MAHEL (1978b) compares the zone to a W-E trough on oceanic crust. The trough is eastwards downwarping beneath the Transylvanian Depression Neogene and indicates thus crustal heterogeneity in the depression basement as well as the presence of highly mobile suboceanic crust (Fig. 41).

We (F. ČECH and J. ZEMAN 1980) presume the same crustal type in the Pannonian Basin basement. The Muresh trough plunges westwards beneath the Neogene filling of the Pannonian Basin.

### 2.5.2.1. Relationship between Transylvanian Basin and Pannonian Basin

The two basins have similar tectonic position and form. They are bordered with orogenic belts and tangential faults. In comparison to the orogenic rim their heterogeneous crust is thinned and they differ in the following features.

1. The Transylvanian Basin is smaller and its basement is likely to consist of one pre-Tertiary block.
2. Its tectonic structure is more complex.
3. Salt diapir structures are the typical feature of the Transylvanian Basin.
4. Subsidence in the Transylvanian Basin was more symmetrical. The sedimentation evolution and subsidence intensity were similar to those in intramontane basins (the Vienna Basin and the East-Slovakian Basin), comparable to marginal basins on the Pannonian diapir periphery. Subsidence culminated in the Miocene time and was reduced or extinguished in the Pliocene. In the Pliocene the intense subsidence in the Pannonian Basin preceeded.
5. The Transylvanian Basin is characterized by a low heat flow.
6. Volcanic rocks do not extend into the basin and border the northern and eastern margins of the Transylvanian Basin.
7. Subsidence was followed by compression of the basin, whereas in the Pannonian Basin were tensions. In the time of the greatest subsidence rate elevations and salt diapir uprise. The present surface in the Transylvanian Basin in relation to other basins is uplifted above more than 600 m.

### 2.5.3. Moesian Depression

The Moesian megastructure is explored by holes. Thirty holes reached deeper than 4.000 m. The deepest hole in Romania — Gerjasa — in this depression, reached to the depth of 6.204 m (V. V. SEMENOVICH — YU. G. NAMESTNIKOV 1981).

From the geotectonic view it is regarded as a platform (H. STILLE 1953) in the Carpathian foreland, as a median massif (YU. D. BULANNBE et al. 1975) or as a fossil back arc basin altered into an ensimatic median massif in the Alpine tectogenesis (F. ČECH — J. ZEMAN 1984). The Paleozoic sediment basement consists of a folded metapelite and metabasite series of the green-schist facies, outcropping in the Dobrodgea elevation. The series is 634—711 Ma old (K-Ar method D. GIUSCA et al. 1969). The basement shows features of simatic crust with subsidence proceeding more than 700 Ma, with hiatuses at the end of the Triassic, in the Jurassic, Cretaceous and Neogene times.

In the Neogene the depression was structurally rebuilt. The term depression (Moesian — Pontic depression) is used (as mentioned above) because the megastructure shows features in common with other Neogene intermontane depressions: the Carpathian — Balkanid (Forebalkan) orogenic rim, and the Hercynian, neotectonically activated North-Dobrodgean orogen. The (central) depression is bordered with marginal depressions with an inverted subsidence regime (F. ČECH — J. ZEMAN 1985) — Fig. 36 and Table 5. Other feature in common are long subsidence, rapid Pliocene subsidence, structural rebuild at the end of the Miocene, and thin crust.

The sediment thickness in the central part of the Moesian Depression is 7—15 km (D. PARASCHIV in V. V. SEMENOVICH and YU. G. NAMESTNIKOV 1981). The depression is comparable to the Black Sea in its sediment thickness and in other characters. The maximum mobility axis oscillated: in the Paleozoic — Triassic time the axis was in the area of the present Danube river, since the Jurassic it had been in the southern half of the depression. In the Miocene the axis migrated N of the Danube r. towards the Carpathians.

The South-Carpathian arc bounding the depression, was already formed in the Lower Paleozoic or Precambrian time (H. G. KRÄUTNER — H. SAVU 1978). This is why the existence of an ancient back-arc depression with unfolded sedimentary filling is presumed. Sedimentation was frequently pelagic; Cretaceous sediments in the Lom marginal depression show abyssal-marine facies. Total sediment thickness in the depression is estimated to 7—8 km, most likely without the Paleozoic whose thickness is estimated to 2—3 km (D. PARASCHIV l. c.).

The origin of the Moesian Depression in the Mesozoic and Tertiary times was identic with the Black Sea origin. There basification due to mantle diapir (YU. D. BULANZHE et al. 1975, V. V. BELOUSOV 1982, a. o.) is preferred. Low heat flows in both units are indicative of recently inactive diapir or a new activity (cf. J. I. NIKOLSKY 1982) commencing in deeper, less viscous mantle (aseismic mantle and crust in the central part) when high heat flow has not yet reached the upper crust. Mantle diapir beneath the Moesian Depression might already had existed in the Precambrian time and reactivated (draining function of endogenous processes) in the Meso-Cainozoic.

### 2.5.4. Crustal types and evolution beneath intermontane Neogene basins

A complex survey of crustal types and its evolution in the Carpathian-Balkan region from the tectonic view was presented by M. MAHEĽ (1978b). His analysis resulted in the following information about the Pannonian Basin and its peripheral areas.

Precambrian and Paleozoic granitoid rocks were nuclei of thicker continental crust belts surrounding the basin. Their thickness is in accordance with geophysical data on crustal thickness. The granitoid rocks form a NE-SW elongated block adjacent to the East-Carpathian zonal unit. Another small nucleus is in the Apuseni Mts. tectonically separated from the arcuate South--Carpathian and Balkan belts.

The basin proper is segmented by two reduced granitoid belts of the Transcarpathian Midmountains, the Mecsek and Villány Mts. Now the basin extends on thinned crust. I think that prior to the Tertiary time it was on thin continental crust (in contrast to M. MAHEĽ'S concept); Fig. 38 and 39. Another markedly reduced ancient granitoid belt is in the Dinarides with their long-preserved simatic crust. The crust of the Dinarides is linked to the Pannonian simatic crust whose evolution was controlled by a different dynamics.

Hercynian granitization was polyphasic. According to M. Maheľ (1978b) it proceeded in three stages, resulted in gradual crustal consolidation which in the Carpathian and Mediterranean areas did not reach the platform stage. The portion of Alpine granitoids is small, and the Alpine crustal thickening was limited to the Serbian-Macedonian Massif, to the western part of the Rhodopes Mts. and partly to the West Carpathians. Banatites are in the South Carpathians, Apuseni Mts. and Balkan. Basing on the high age of sialic rocks, J. ZEMAN (1980) presumes nuclei of sialization and continental crust growth in these regions. The sialization corresponds to the origin of the South Carpathian paleoarc (Chapters 2.5.3) and sialic median massifs.

Continental crust thickening did not reach the Pannonian region. As late as the Paleogene, a zonal crustal accretion proceeded along the Zagreb — Kulcs deep fault, among ancient granitoid belts. The Paleozoic and earlier granitization had an abyssal character in contrast to the upper-crustal Meso-Cenozoic granitization (M. MAHEĽ 1978b). The size of granitoid bodies is also different.

Late-Hercynian volcanics do not display fading-out sialization in all areas. In some places (e. g. the Apuseni Mts.) they show initiating oceanization in less stabilized or unconsolidated zones. The Mesozoic mobility in the Pannonian region is linked with the preceeding mobility. Eugeosynclinal evolution at the end of the Early Proterozoic, and rather ensimatic evolution in the Paleozoic (B. JANTSKY 1976) were proved by deep drilling. Analogous evolution in geologic units adjacent to the Pannonian region is indicative of a common evolution of the entire crustal segment, and of a reduced continental crust up to the Late Paleozoic. According to M. MAHEĽ (1978b) there was oceanic crust in the region, including a part of the West Carpathians. At the end of the Paleozoic the crust was irregularly sialized, mainly in the Pannonian region. Crustal segmentation into elevations with granitoids and into troughs with basic rocks is indicative of variable crustal thickness. Suboceanic crust in the Bükk Mts. Triassic is indicated by basic and ultrabasic rocks. This crust extended to the South—Slovakian Basin (Chap. 3.4.3; Fig. 42).

Another suboceanic crust belt is W of the Szolnok flysch trough in the Alföld Neogene basement (G. WEIN 1969). Basic rocks in the Late Jurassic — Early Cretaceous filling are indicative of a continuation of this crust beneath trough joining the Mecsek Mts. basic rocks. The Muresh zone W-E elongated ophiolite belt extends beneath the Transylvanian and the Pannonian Basins — like the basic rock belts from the Hungarian zone.

Divergent crustal types orientation suggests the contact of three troughs most likely of the rift type, in the Danube region, S of Szeged: NE-SW, W-E, NNW-SSE. They are crustal thickness reduced zones of a higher mobility preventing the formation of a consolidated megablock. Alcaline basalt volcanism is indicative of faults extending deep in the mantle, and of a thicker crust — besides troughs with tholeiites.

In contrast to N. HERZ — H. SAVA'S (1974) opinion about the distortion of a granite continental plate, I regard suboceanic crustal belts as remnants of slightly granitized or non-granitized (closer to oceanic) crust or of the basified crust belt. My opinion is based on the above mentioned data, mainly on poor Hercynian consolidation evidenced by tectonic information published by M. MAHEĽ (1978a, b).

Basic rock occurrences associated with troughs are also in sialic crust belts but their size is smaller and segmentation less intensive than in the Pannonian region or the Dinarides. Numerous troughs prove intense segmentation of the Tethys (M. MAHEĽ 1978a).

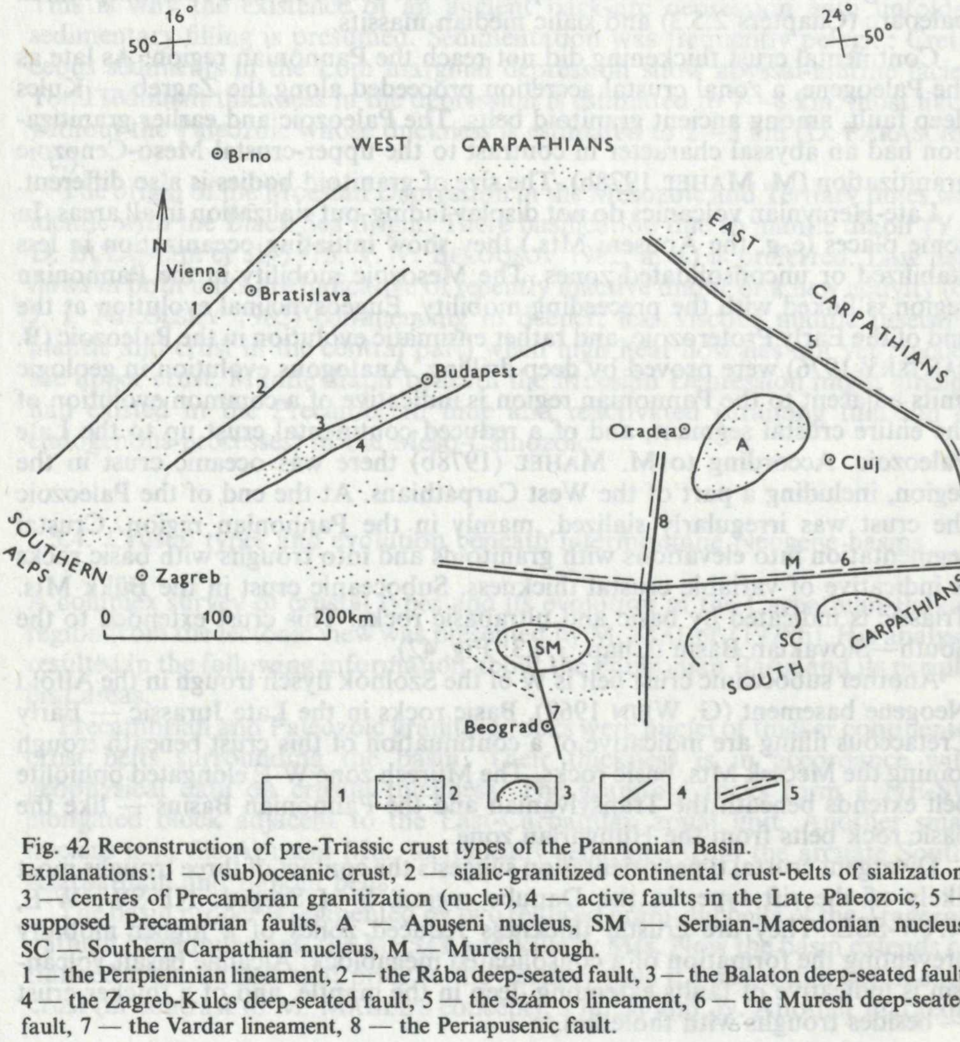



Fig. 42 Reconstruction of pre-Triassic crust types of the Pannonian Basin.
Explanations: 1 — (sub)oceanic crust, 2 — sialic-granitized continental crust-belts of sialization, 3 — centres of Precambrian granitization (nuclei), 4 — active faults in the Late Paleozoic, 5 — supposed Precambrian faults, A — Apuşeni nucleus, SM — Serbian-Macedonian nucleus, SC — Southern Carpathian nucleus, M — Muresh trough.
1 — the Peripieninian lineament, 2 — the Rába deep-seated fault, 3 — the Balaton deep-seated fault, 4 — the Zagreb-Kulcs deep-seated fault, 5 — the Számos lineament, 6 — the Muresh deep-seated fault, 7 — the Vardar lineament, 8 — the Periapusenic fault.

So in the Mesozoic time the Pannonian crust was segmented into mobile suboceanic belts: the Bükk Mts. belt separated by the more stable eastern Pannonian block according to M. MAHEĽ (1978a), the Szolnok belt on the SE, separated by the Apuseni Mts. from the Muresh zone. M. MAHEĽ (1978a) regards elevations as intrageosynclinal massifs among which the East-Pannonian massif mostly affected further basin evolution. Early Miocene rhyolite volcanism proceeded along the Zagreb—Kulcs fault. As a surface product, the volcanism did not affect the origin of continental crust. The massif was involved in Late-Miocene, mainly Pliocene subsidence. So the temporary elevations were still in the area of increased crustal mobility. Since the Pannonian Basin has not been intensely folded although being mobile and missing the platform character (suboceanic crust), it need not be ranged among common types of median massifs. It rather displays some features characteristic of simatic median massifs (F. ČECH — J. ZEMAN 1982). Poor local folding may be indicative of dominant tension (basins of the rift character, graben faults) in crustal evolution and inspires the opinion that the mantle elevation predated the Tertiary time (L. TRUNKÓ 1977). It may also be evidenced by oceanization-basification elements, in troughs associated with crustal thinning. Areas of Mesozoic mobilization agree with areas of poor Hercynian consolidation of local temporary duration (marine Carboniferous and Permian). The beginning of volcanic activity might be represented by the banatite volcanic-plutonic formation. It is also evidenced by its association with trench-fault structures superimposed on paleoalpine structures i. e. with structural rebuild.

Following are further effects of the mantle elevation:

a) Eocene-Oligocene activity with the main areal effects on the central part of the Pannonian Basin and side effects on its marginal parts (the Central-Slovakian area, Mátra Mts.). They might be associated with hypothetic partial diapirs or with thermal columns uprise.

b) Miocene, mainly Badenian-Pliocene activity with the maximum intensity in the N and NW parts of the Pannonian Basin upon poorly Hercynian--consolidated structures (Fig. 38). Volcanic belts were on the thick crust/thin crust boundary. They show the character of continental crust accretion around suboceanic massifs. So the suboceanic crustal type is emphasized. It is partly reflected in the block rebuild.

The origin of andesite lavas was associated with crust melting (H. STILLE 1953) owing to its subduction into high-temperature zones. H. STILLE'S concept was renewed in subduction models (L. STEGENA et al. 1975, M. BOCCALETTI et al. 1976, N. HERZ and H. SAVA 1974 a. o.). J. Mészáros (unpublished) regards Paleogene volcanism in NW Hungary as a product of the so-called internid subduction (in the back of orogen). But subduction did not produce volcanism. It is a result of the backars basin formation. The role of subsidence in the origin of volcanism is not clear, and even doubtful (J. LEXA — V. KONEČNÝ 1974).

The alkaline basalt (final) volcanism may reflect crustal consolidation or deepening of feeding (faults) in a mantle elevation. (Generally the alkalic basalt source is put deeper in the mantle — e. g. B. G. LUTC 1975). It is also indicated by finds of spinel peridotite inclusions in basanite near Mašková (D. HOVORKA 1978a). The faults deepening is likely to have been associated with the cooling of the top part of the mantle and with its collapse. This is in accordance with M. MAHEĽ'S critical comment on hypothetic partial subductions in the Panonian region and on unreasoned presumption of closing microoceans (L. STEGENA et al. 1975).

M. MAHEĽ (1978b) is right in rejecting the idea of a relationship between Carpathian neovolcanics and presumable subduction zones. In the period of volcanism the subduction zone could not be between the Outer and Inner West Carpathians like — for example — in the East Carpathians. The subduction zone should be between the Pannonian and the West Carpathian units which would be reasoned with respect to the suboceanic-continental crust boundary. Only the respective folding and other phenomena are missing. Either it is a specific case of subduction in the West Carpathians — which is an unreasoned presumption — or it is an interplate (i. e. between megablocks) discontinuity due to the mantle diapir contacting overlying continental crust. The plate tectonic model of the origin of the Carpathians and of the Pannonian Basin was also criticized and rejected by L. TRUNKÓ (1977), and by V. G. SVIRIDENKO (1978).

Petrologic experiments prove the possible origin of andesites from mantle peridotites in the presence of water and alkali dotation (B. O. MYSEN — A. L. BOETTCHER 1975). Andesites may originate from shallower depths beneath the crust, and from fractioned basalts altered into amphibolites undergoing further melting on the crustal base at the presence of $H_2O$ (V. V. BELOUSOV 1982). So the origin of intermediary magma may be associated with differentiation processes in the mantle. The same process may also proceed in mantle diapir. Its inner dynamics and expansion controlled the S-N volcanism migration, and persistent effusions in Central-Slovakian stratovolcanoes (J. LEXA — V. KONEČNÝ 1979).

Following is my concept: The present Pannonian Basin underlying crust did not result from interaction of plates or microplates but as autochthonous crust it was poorly differentiated in the Precambrian and Early Paleozoic times and generally resembled the original oceanic crust. Late Paleozoic differentiation under local alkalies influx resulted in granite layer zones and nuclei; so the crust was heterogeneous and mostly consisted of sialized small massifs (nuclei) and belts: Balaton-Velence, Apuseni, a part of the Serbian-Macedonian massif. The belts were surrounded by enclosed rests of suboceanic crust, slightly Hercyniansialized. The massifs tended to higher Alpine mobility. The crust between the massifs was partly Hercynian-sialized, perhaps also non-sialized and might have had suboceanic character (Fig. 39). The mobile crust in the Mesozoic time became the basement of troughs, partly of a rift nature. The trough system was controlled by faults separating more sialized massif belts, forming elevations. Ultrabasic rocks may indicate initiating crust mantle elevations interactions. In the Mesozoic time, oceanization of the Mediterranean type (R. V. BEMMELEN 1972) and gradual alteration of thin continental crust segments with suboceanic crust relics into thinned crust "oceanized" in the Tertiary by expanding mantle diapirs (YA. P. MALOVITSKY et al. 1982) began. In the Tertiary time the crust with a dynamics close to oceanic basins dynamics, i. e. with areal subsidence and rift structures, formed. The dynamics of oscillation processes and volcanism were most likely affected by uneven mantle elevation, its expansion and collapse processes. Mantle diapirism and the maximum Neogene subsidence developed mostly under partly Hercynian-granitized suboceanic crust (Fig. 38). As late as the Pliocene, also originally stable crust got partly regenerated, mainly in the Pannonian Basin basement.

The Pannonian Basin geologic future: According to L. BODRI'S (1981) calculations, a high heat flow enables partial rock melting beneath the basin — in the crust base, since temperatures are by 400—500 °C higher than in surrounding units. According to L. BODRI (l. c.) the mantle surface temperature is 700—900 °C. V. ČERMÁK (1981) calculated 800—1000 °C temperatures on the crust/mantle boundary — which is temperature suitable for the generation of andesites and acid lava. Cenozoic evolution ended with the diapir collapse. Other basins with rapid subsidence (Black Sea, South-Caspian Depression) have a low heat flow. There is a question whether it is a postvolcanic fading out heat flow or the beginning of a new thermal event. V. V. BELOUSOV (1982) assumes that the basin-underlying crust is in the initial basification stage. The basin is, however, an exception (cf. Chap. 2.5.5) among basins with mantle diapir in the series of active diapir + volcanism — inactive diapir with subsidence without anomalous heat flow. So the Pannonian Basin might even represent a stage predating new volcanism and more intense subsidence. After all, the present subsidence values in the Danube lowland, ranging to 2 mm/year (P. MARČÁK 1978), surpass the Miocene and Pliocene subsidence rates. If we shall not consider the existing pulsating character of movements (P. MARČÁK 1978), then in 1.000 years subsidence will range up to 2 m in average, and the present basin surface will in 45.000—50.000 years get beneath the present sea level. The future existence of an intracontinental lake like the Caspian Sea or an oceanic depression would not be excluded.

### 2.5.4.1. Basement type and age

My study of relations between crustal mobility and age of endogenous continental crust-forming processes is based on complex geochronologic data presented in a map published by the CBGA Subcommission on absolute age (M. P. SEMENENKO 1977). I do not apply age — except high ages — as an absolute criterion, since a measured lower age may be due to younger endogenous processes.

The 1200—1700 Ma old Moesian Depression basement (D. GIUSCA et al. 1969) is the oldest unit known. In this area no higher crustal metamorphism took place. Crustal mobility corresponds there to young platforms. The highest known age in the Carpathian region indicates that crustal thinning began in the Early Tertiary time when the unit got mobile and acquired the character of an intermontane depression on a simatic (?) median massif.

The age of the Transylvanian Basin basement is not exactly known, it is estimated to pre-Cadomian age about 1.000—1.200 Ma, with Hercynian and Cimeric magmatic regeneration. The data are extrapolated from the Pannonian Basin basement where the above values have been measured. The Transylvanian Basin basement was only locally affected by Paleozoic continental crust-forming processes, mainly around granitoids of the Mecsek and Villány Mts., and along the Balaton-Velence belt. In other places the original processes preserved their age.

The Zagreb—Kulcs line is a boundary between units of variable ages. It separates the Proterozoic Sopron sialic crust from the Early Paleozoic crust of the Transdanubian Midmountains. The age of the Sopron block granitoids is pre-Cadomian. Around Szeged is the centre of pre-Laramide rejuvenation. The area is in the place of the thinnest crust and a younger age might speculatively be associated with overheating during asthenosphere advective processes.

In the Carpathian lowland hinterland the Tertiary mobility affected the crust of variable maturity as far as the maturity is measured with the age of the last endogenous process. Although repeated endogenous processes — mostly Hercynian — affecting the evolving crust, are geochronologically indicated, the crust age is not deciding for the basin mobility (see the Moesian Depression). No shield crust associable with the Tissia block was generated. For more than one milliard years the Pannonian crust was tending to repeated subsidence (F. ČECH — J. ZEMAN 1982). In the course of its evolution no typical platform continental crust with a continuous granite layer was generated because no one of crust-forming tectonic-magmatic cycles had any considerable continental accretion effects. This is why I do not agree with L. STEGENA et al. (1975) in their assumption about the original continental crust having the thickness of platform crust (up to 38 km) and being thinned by at least 8 km owing to lower-crust erosion by mantle diapir. L. Stegena et al. (l. c.) explain the Neogene subsidence by this crust thinning.

### 2.5.5. Pannonian mantle diapir dynamics

R. W. VAN BEMMELEN (1972) considered geotectonic significance of mantle diapir activity. According to this concept the crustal stretching resulted from spreading low-viscous overheated differentiated rocks from the mantle elevation top (and/or from the top of asthenospheric elevation), followed by crustal downsagging at diapir collapse (L. STEGENA et al. 1975, D. VASS 1979). The relation between subsidence in the Pannonian Basin and on its margins is best explained by active dome-formed mantle diapir. This explanation is in accordance with the geophysical and geological data about the basin structure and evolution. The diapir might also have caused crust basification s. l. Also neovolcanism is associated with diapirism. Diapirsm (see Chap. 1.10.2) was also associated with the generation of marginal depressions and with movements on marginal shear faults. Diapirism is very frequent in the Mediterranean and circum-Mediterranean areas.

The diapir centre was beneath the Pannonian Basin in places of original thin heterogeneous crust of the non-platform type. A. R. CRAWFORD'S (1977) assumption about the diapir centre beneath the Transylvanian Basin on a hypothetic lineaments contact is unreasoned. There are no geological neither geophysical indications (e. g. prominent crust thinning) beneath the Transylvanian Basin. L. POSPÍŠIL and D. VASS (1983) presume the diapir centre in the area of Oradea—Arad.

Because of lithosphere density it is difficult to explain mechanically F. HORVÁTH'S et al. (1981) idea about the diapir penetrating into a free area, horizontally open due to splitting of plate front during collision; causing there rigid lithosphere updoming and its detachment from ductile lithosphere. Neither the collision plate margin in the place of diapir can geologically be proved because any evidence of intense sediment folding above a hypothetical suture is missing.

The time of mantle diapir or other diapirs inside the extensive mantle dome is problematic. L. TRUNKÓ (1977) dated the diapir as Late Cretaceous. In the Apuseni Mts. and in Banát volcanic activity commenced in the Senonian (K. KAROLUS 1978). In the northern part of the basin further magma was generated in the Eocene and Oligocene times. Cretaceous age of the diapir is acceptable. L. BODRI (1981) presumes increased heat flow beneath the Pannonian Basin since the Triassic. It is likely that the largest diapir expansion in the area of thinner crust in the southern part of the Inner West Carpathians and along deep faults in the East Carpathians proceeded in the Miocene. In the East Alps diapir probably extended to the deep compression zone of the Alps bordered by the Pericarpathian lineament. The lineament also represents the spatial limit of thermal activity. Therefore neovolcanism is absent in the Alps. Magmatic activity affected ancient suboceanic dehydrated crust and also in my opinion caused poor alkali and water contents in the volcanic association in the East-Carpathian volcanic series (J. LEXA — V. KONEČNÝ 1979). In the Central-Slovak region, chemical composition of lava was affected by sialic crustal blocks, by a high heat flow which also controlled dehydration and enabled formation of rhyolite magma.

Migration of volcanism outwards from the Pannonian Depression (with local deviations) is indicative of anatectic influence of the expanding diapir. It is also indicated by tension relaxation migrating towards the basin centre and associated with subsidence, formation of faults and partial troughs. In the NW part the subsidence mobility extended to the depression margins and was associated with mobilization of Central-Carpathian units. Diapir activized deep faults and altered some, mostly N-S lines into deep active faults.

In my opinion the uprise of alkalic basalts indicates deep faults penetrating deeper in the mantle in the Pliocene-Quaternary time in connexion with the cooling of its upper part, resulting in spatially limited or stopped diapirism. It is likely that diapir cooling and reduction resulted — besides emptying of primary magma chambers — in intense horizontal tensions and bending. Rigid mantle behaviour caused intensified subsidence of increased rate. But collapses might also have resulted from diapir topical downsagging. The syncline-dish form of the Pliocene Pannonian Basin and its triangular border may be the result.

Changed tectonic regime and reorientation of tension in the Late Miocene and Pliocene time, resulting in preferred tectonic course independent of Late Cretaceous lines may also be explained by mantle activity. The independence was also revealed in the Outer Flysch Carpathians (J. LEXA — V. KONEČNÝ 1979).

The role of heterogeneous crust in the Pannonian Basin evolution has so far not been evaluated properly. The heterogeneous crust could not be found by common geophysical methods. Deep gravity sources affected the quasi-homogeneous gravity pattern which inspired the idea of a uniform platform block. The transformed gravitational fields method offered new data but still more information resulted from drilling explorations (G. WEIN 1969, B. JANTSKY 1976).

Evidently, basification has a selective nature, and the mantle diapir occurred in poorly consolidated, thinner segment of partly sialized crust (in contrast to L. STEGENA'S et al. 1975, V. V. BELOUSOV'S 1982, a. o. opinions). The emplacement of master diapir and of satellite diapirs was decisingly controlled by isostatic differences between this and Paleozoic (or Precambrian) granitized crust of adjacent units.

Migration of volcanism and maximal subsidence in the Pannonian Basin and on its periphery from inner zones to outer and from W to E along the West--Carpathian arc (T. BUDAY 1961) shows locally an opposite trend in migration, namely from periphery to centre. This migration course is in accordance with spatial distribution of mantle diapirs and satellite diapirs — with advancing basification front and decompression in the master diapir centre. In outer units the volcanism migration is not oriented towards the Pannonian Basin centre (J. LEXA — V. KONEČNÝ 1979, M. MAHEĽ 1978b). The relation to hypothetical subduction is controlled by dominant tension and no compression in the Inner -Carpathian crust, since the Paleogene. The genesis of basins as well as of volcanism seems to be rather an autonomous process, whose character is also indicated by the asthenosphere elevation.

According to geophysical data the Inner Carpathians and the Apuseni Mts. are part of the Pannonian Basin (L. STEGENA et al. 1975), and it is a reflection of the young process. They are geologically different units with variable original crustal types (M. MAHEĽ 1978b). I think that plentiful acid lavas in the volcanic arc and their smaller amount in the Pannonian Basin Paleogene also prove the existence of the original geochemically different crust beneath the basin and on its periphery.

From the dynamic view I — in contrast to L. STEGENA et al. (1975) — do not consider diapir as a consequence of subduction (for the reasons see above) but as an autonomous process with riftogenic features, due to density disequilibrium in mantle and crust, and to different dynamics of continental and (sub) oceanic crust movements (R. W. VAN BEMMELEN 1972). I think that according to geologic history of the Pannonian Basin basement the process already had a partly riftogenic character in the Mesozoic time. It is rather a riftogenesis with oceanic features. I also explain by mantle diapirs the origin of the Transylvanian Basin and the Moesian Depression subsidence with the weakest diapir effects. In accordance with some Soviet geologists (S. I. SUBBOTIN et al. 1972) I regard the Black Sea, as the result of stretching of the crust due to the initial stage of riftogenesis. In contrast to S. I. SUBBOTIN et al. 1972, I suppose that the primary crust was not a typical platform one but rather a relic suboceanic, partly sialized crust (P. GOTCHEV 1976 also presented some ideas like this).

Now I shall discuss correlation between the Pannonian Basin and intermontane oceanic basins.

The Pannonian Basin with its deep structure resembles the Tyrrhenian Sea where the hot diapir surface (lower-velocity channel) ranges to the depth of 10—6 km (M. BOCCALETTI et al. 1976). Close diapir/crust interaction is manifested in the present acid-intermediate and basalt volcanism. The Tyrrhenian Basin is surrounded by continental crust. It is likely that the continental crust ring was destroyed by oceanization to a considerable extent.

Relics of Variscan poorly consolidated crust are on islands like SECCHIA, CASSINIS, CORNEGLIA FLAVIO GIOIA, ISSEL a. o.

The alkaline-calcite lavas are closer to basin margins whereas the central part of the sea consists of submarine volcanic islands, most likely with tholeiites.

There also are alkaline basalt effusions.

The Aeolian volcanic arc formed in two phases: the andesite arc in the Early and Middle Pleistocene;; the younger andesite — shoshonite is a recent one. The inner tholeiite basic volcanism and Na-rhyolites have Pliocene age ranging from 5 (Na-rhyolites) to 0.7 Ma. Alkaline basalts are younger than 1 Ma. The volcanic evolution shows that mantle derivates in places of intermediary stage of volcanism are oldest — perhaps due to crustal contamination (basification), and the youngest — final mantle invasion stage is alkaline. The evolution is similar to that of the West-Carpathian neovolcanism with freqvent rhyolites. The lack of rhyolites in the area of the Tyrrhenian Sea may be indicative of the original absence of the granite layer in the place of mantle/crust interaction.

The Tyrrhenian Sea shows further features in common with the Pannonian Basin: thin crust, positive gravitational field (more prominent than in the Pannonian Basin) and a high heat flow. Seismoactivity is restricted to basin margins — to the areas of active volcanoes. So there is a partial correlation with peripheral faults of diapir on the contact between positively and negatively with peripheral faults of diapir on the contact between positively and negatively gravity-disturbed crusts. Interpretation of the Benioff zone (M. BOCCALETTI et al. 1976) is violent, without seismologic evidence of its link with the surrounding continent, and the area of recorded seismic focuses is only associated with the eastern part of the Tyrrhenian Sea. Seismic focuses do not exceed the depth of 100 km. The zone between 100—230 km is dispersed, most likely because of mantle melting. Only two focuses were recorded 450 km deep, beneath the diapir centre. The existence of the arcuate Benioff zone is explained by the contact of the hot ductile mantle with rigid cool mantle during the hot diapir ascent (F. ČECH — J. ZEMAN 1985) — Fig. 20.

In contrast to the Pannonian Basin the Tyrrhenian Sea shows a greater recent dynamic activity. It is shown by differences in the depth of lower-velocity channels: 50 km beneath the Pannonian Basin, 30 km beneath the Tyrrhenian Sea. The asthenosphere elevation sinks rapidly beneath thicker continental crust — it is recorded 50 km below the western margin of Sardinia.

The Aegean Sea surrounded with a ring of the thicker Hellenic arc continental crust, has a very young diapir structure. A mushroom-shaped diapir of a less dense hot mantle (according to gravimetric data) has a high heat flow in upper crust. According to geophysical calculations, of J. MAKRIS (1977 in J. LEXA — V. KONEČNÝ 1979) the top of mantle (asthenosphere) diapir reaches to the depth of 25—23 km where calculated temperatures should be 700—900 °C. The "asthenospheric" diapir height is 100 km in the diapir axis. Explosive volcanism in the Aegean Sea is a proof of mantle activity.

The Black Sea is most likely to represent a diapir relic — the probably cool mantle elevation is in the depth of 20 km, the asthenosphere surface — in a depth of about 150 km. Low heat flow density (20—30 $mWm^{-2}$; V. ČERMÁK 1979) is indicative of extinct activity and post-collapse stage associated with rapid crust subsidence with sediments ranging to 10—12 km in thickness.

Geophysical and geological indications of different endogenous activity characterize the evolution sequence and various stages of basin evolution:
— Tyrrhenian and Aegean Seas — culminating hot diapir activity,
— Pannonian Basin — extinct or renewing hot mantle diapir activity,
— Black Sea and South-Caspian Depression — extinct diapir activity.

### 2.5.6. Conclusions concerning genesis and deep structure of basins, based on correlation of geological and geophysical data

Following are most important facts, opinions, models and relations:

1. The Miocene, and mostly Pliocene sedimentation areas in the Pannonian Basin, in adjacent intramontane basins, in the Transylvanian and Moesian depressions are associated with primarily thinned simatic crust. The secondary thinning associated with mantle elevation affected the basalt layer, and the present crustal thickness is in average by 1/3 smaller than in adjacent orogenic belts. The thinner crust segment is positively gravity-disturbed, frequently associated with increased heat flow. The crust (mantle boundary is not sharp, it shows a character of a transitional zone. P-waves velocities in upper mantle are lower than in normal platform-underlying mantle, and the zone with velocities 7 km/s and 8 km/s forms elevations in lithosphere. They are associated with asthenosphere elevations. There is a positive correlation between the high heat flow density and asthenosphere dome.

2. The granite layer is thinner than in thicker crust zones in orogens or on the platform, which is indicative of its primary thinning.

3. Basins are bounded with deep faults. The internal fault segmentation is controlled by the basement block structure, mostly in the Pannonian Basin. Basins rest on simatic median massifs.

4. In the West Carpathians and in the depression area are two fault-deformational planes, corresponding to two segmentation periods. In the neotectonic period the basin subsidence was affected by both ancient fault activation and by the formation of new active faults.

5. Specific features of pre-basin evolution are due to the thinner granite layer and perhaps also to primarily thinned crust because of poor Hercynian folding and incomplete tectorogenic cycle. The cycle resulted in a locally poor consolidation, since the crust — irrespective of age — stayed mobile in simatic units, and as a whole it rather had a suboceanic character. Its evolution was mostly associated with subsidence without any prominent orogenic inversion. Pre--Tertiary depressions often have a character of rifts and rift geosynclines.

6. Faults of the greatest structural significance are on median massifs margins where deep faults with a shear fault character are in a tangential relation to the mantle elevation margin. These faults are most seismoactive. The mantle elevation has a slightly undulated surface. Partial elevations in their position do not correlate with the internal deep fault segmentation of the basin basement into blocks.

7. The upwarping mantle affected tectonically mobilization of all ancient deep faults. In the Pannonian Basin, melted mantle rocks or crust rocks only got in contact with faults in the northern part of the elevation. In the Transylvanian Basin, volcanic rocks form the NE margin of the basin. In the area of the Moesian Depression the active volcanic belt of the Srednogorije is out of the depression. Seismically recorded lower velocity elevation zones with an intense heat flow in the Pannonian Basin are indicative of unaccomplished crust anatexis. (Low velocities may also be explained by material change, for example dioritization of original basite crust, possibly representing a relic of andesite volcanism.

8. Relics of perhaps formerly oceanic crust are on the Pannonian (Transylvanian depressions contact — the Muresh trough.

9. Neovolcanic rocks occurrences are most frequent in the Carpathians (M. Maheľ 1978a). They are associated with the thick/thinned crust boundary in the area of the Pannonian positive gravitational field. In Romania, Cretaceous and Paleogene volcanic rocks are in a similar position in relation to the Transylvanian Basin. Neovolcanics are concentrated along the mantle elevation margins in the place of most significant deep faults, in the Balaton block and along the internal periphery of the East Carpathians.

10. Character of the basement of active volcanic lines is so far not known in the whole extent. Detailed research revealed volcanically active parallel faults in the zone of the Rába line in the Pannonian Basin near the Czechoslovak frontier. There the basic volcanism was already active prior to the Neogene time, perhaps since the end of the Proterozoic (L. Zbořil et al. 1971). In this segment, like in the Balaton block (R. Mock 1978) the basement rocks are indicative of suboceanic crust with long repeated basic activity. The belt is in derived regional gravity maps characterized by a positive anomaly. Data about the Moesian Platform and on other parts of the Pannonian Basin show that positive gravity

anomalies are not only indicative of basement elevation but also of basic rocks in the basement. As far as the basement consists of older sedimentary complexes of a nappe character (the Transylvanian Basin, the Vienna Basin), the nappe subsidence to a great depth might have been caused by the basite substratum tending to subsidence. Nappes with denser limestones overburdened the crust and thus caused a more intense subsidence.

## 3. Deep structure of Neogene basins in Inner Carpathians and in Vienna Basin

Neogene basins are longitudinal intramontane and superimposed intermontane basins in the West Carpathians, comprising coal-, oil and gas deposits. I shall deal with foredeep basins only with respect to some correlation problems.

D. Vass (1979) studies the genesis and distribution of the West Carpathian intramontane and intermontane basins in relation to mantle diapir. He distinguished the following basins:

a) basins on the active diapir periphery (intramontane basins),
b) basins in the central part of the active diapir (episodic basins),
c) basins above the central part in the diapir final stage (superimposed intermontane basins; T. Buday 1961).

With respect to deep structure I divide Neogene basins in the Inner West Carpathians into two groups, partly in accordance with D. Vass (1979):

a) Intramontane basins: the Turčianska, Trenčianska, Ilavská, Hornonitrianska, Handlovská, Žiarska, Zvolensko-Slatinská, Banskobystrická, Breznianska and Rožňavská panva basins;
b) Intermontane basins: the East-Slovakian Basin — Transcarpathian Basin, the Vienna Basin having a special position since it also extends on the outer side of the West Carpathians, the Danube- and the South-Slovakian Basins. The latter two are genetically related to the Pannonian Basin.

### 3.1. Geophysical indications of structure

Geophysical basin explorations were performed with respect to hydrocarbon deposits prospection. Explorations of basins prospective for new deposit structures still continue.

#### 3.1.1. Gravimetry

The area of Neogene basins was gravimetrically explored in detail within regional research. Individual basins, mostly economically prospective, were then examined in detail by prospection geophysics.

My study of deep structure is based upon gravity map of complete Bouguer anomalies and maps derived according to B. Beránek (1979), J. Plančár — J. Ibrmajer (in O. Fusán et al. 1971) and J. Ibrmajer (1978).

Maps depicting the entire area of the West Carpathians with the Bohemian Massif for the purpose of deep structure interpretation are most suitable for the study of the problems. J. PLANČÁR — J. IBRMAJER'S detailed maps were methodically aimed at the study of the basin basement. Methods and gravity anomalies interpretation had the same purpose.

I compared the geological map with marked deep structure units defined by O. FUSÁN et al. (1979) to the maps of regional field and residual gravity anomalies (O. FUSÁN et al. 1979) and their interpretation. I correlated gravimetric data to geological results of deep drilling (O. FUSÁN et al. 1971, A. BIELA 1978).

The West-Carpathian positive gravitational field occupies the area of all large Neogene basins except the Vienna Basin and the Košice part of the East-Slovakian Basin. In derived gravity maps the southern part of the Vienna Basin on the Czechoslovak territory (Č. TOMEK — L. BUDÍK 1981) and the Košická kotlina depression are manifested as regional positive anomalies. They are tectonically disturbed basins with deeper and deep faults. Some of them are seismoactive. Another zone of negative anomalies is the eastern part of the Danube Basin in the area of the Vepor deep fault (O. FUSÁN et al. 1979). As for the negative gravitational field — the situation is more complicated in the East-

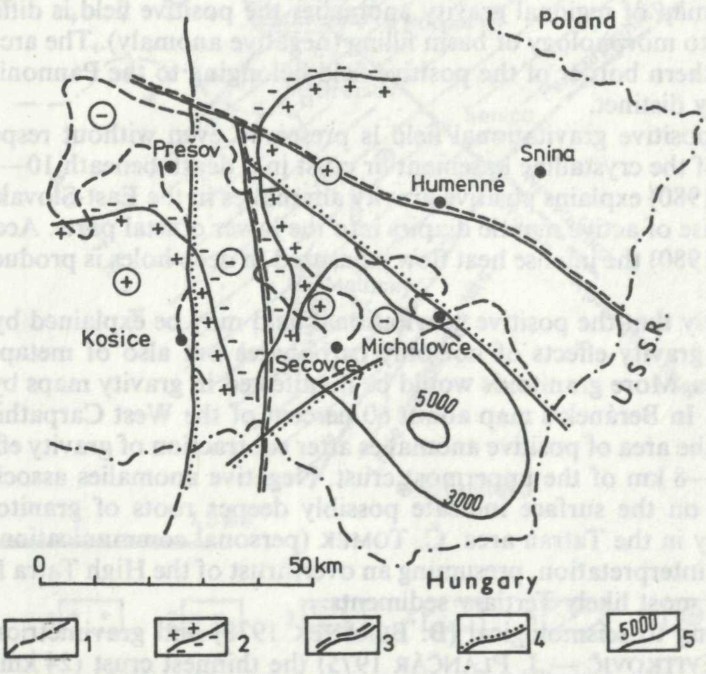



Fig. 43 Deep structure of East-Slovakian Basin.
Explanations: 1 — confinement of negative residual anomaly at the depth of 12 km, 2 — regional anomaly at the depth of 8 km (boundary of positive and negative anomalies), 3 — deep-seated faults (according to O. FUSÁN et al. 1979), 4 — seismoactive faults (according to O. FUSÁN et al. 1979), 5 — isohypses of basin basement in m.

-Slovakian Basin (Fig. 43). The field has the form of a three-tipped star with its peak on the contact of the Slánske vrchy (hills) fault with the Humenné fault. These faults are doubled by a parallel seismoactive fault. The anomaly extends along both seismoactive fault courses. All negative anomalies in the positive-gravity disturbed Carpathian field are prominent in transformed maps of the second gravity derivation (B. BERÁNEK 1979) and in maps of residual anomalies.

Upper-crustal inhomogeneities disappear in regional gravitational field maps according to Griffin formula with excentricity radius $5\sqrt{5}$ km, and in maps comprising greater depths (cf. O. FUSÁN et al. 1971). Positive gravity anomalies blend into a continuous belt linking up with the Pannonian gravitational field. The prominent Transdanubian Mid-mountains anomaly with basite crust in its basement joins the southern part of the Spišsko-gemerské rudohorie ore mountains. Regional positive gravity anomalies in western Slovakia beneath the Danube Basin terminate on a NW-SE line corresponding to the Skýcov fault zone. The westernmost part of the regional positive anomaly is near Prešov. From Prešov the anomaly bounded with prominent gravity gradient of the Peripieninian lineament is running southeastwards beneath the East Carpathians.

On the map of regional gravity anomalies the positive field is differentiated according to morphology of basin filling (negative anomaly). The arcuate form of the northern border of the positive field belonging to the Pannonian field is particularly distinct.

So the positive gravitational field is preserved even without respect to the influence of the crystalline basement or crust in a depth beneath 10—15 km. L. POSPÍŠIL (1980) explains positive gravity anomalies in the East-Slovakian Basin by the uprise of active mantle diapirs into the lower crustal parts. According to POSPÍŠIL (1980) the intense heat flow measured in deep holes is produced by the diapirs.

It is likely that the positive gravitational field may be explained by a denser crust, i. e. gravity effects of not only carbonates but also of metapelites and metabasites. More granitoids would be manifested in gravity maps by negative anomalies. In Beránek's map about 60 percent of the West Carpathian region remain in the area of positive anomalies after subtraction of gravity effects from the first 5—8 km of the uppermost crust. Negative anomalies associated with granitoids on the surface indicate possibly deeper roots of granitoid bodies particularly in the Tatran area. Č. TOMEK (personal communication) presents a different interpretation, presuming an overthrust of the High Tatra Mts. upon less dense, most likely Tertiary sediments.

According to seismological (B. BERÁNEK 1978) and gravimetrical calculations (J. KVITKOVIČ — J. PLANČÁR 1975) the thinnest crust (24 km) is in the southern part of the East-Slovakian basin and joins the thin crust of the Pannonian Basin. It is indicative of possible partial diapirs. The basins area is divided from elevations with up to 40 km thick crust by an isoline of 30 km crustal thickness.

The Vienna Basin is in a regional negative gravitational field. After subtraction of Neogene sediments gravity effects, positive anomalies appear in the southern part (Č. TOMEK — L. BUDÍK 1981; R. JIŘÍČEK — Č. TOMEK 1981). This area is separated from the northern part with the negative field by the Nesvačilka — Trnava deep fault separating variably mobile blocks with different hydrocarbon accumulation concentrations (F. ČECH 1982) — Fig. 44 and 45. In my opinion the positive gravitational field is due to the denser mobile simatic crust which is part of a submerged margin of the Brunovistulicum block in A. DUDEK'S (1980) sense. My opinion about the simatic crust character is based upon geophysical indication of basic rock bodies presumed in the area of

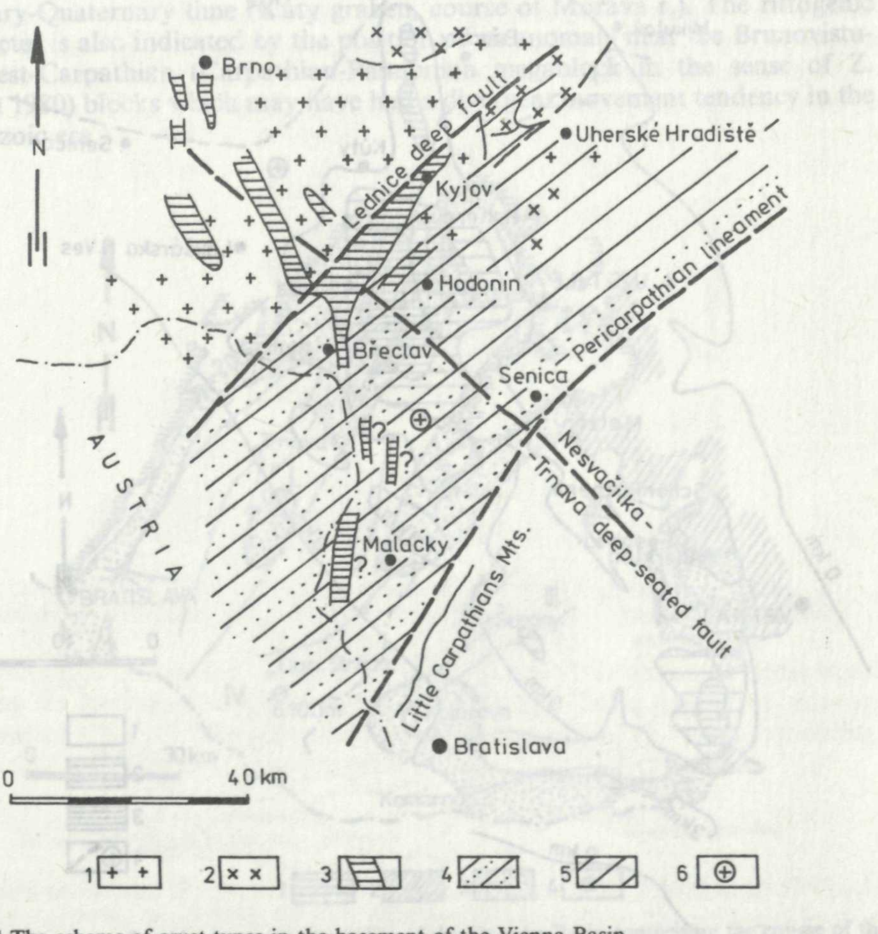



Fig. 44 The scheme of crust types in the basement of the Vienna Basin.
Explanations: 1 — granitoides, 2 — tonalites; supposed complexes: 3 — basics (metabasites) and ultrabasics; 4 — paragneisses with metabasites, 5 — paragneisses, in places weakly migmatized, 6 — positive gravity field in exposed gravity map (Č. TOMEK — L. BUDÍK 1981). (Granitoid rocks adapted according to A. DUDEK 1981).

Břeclav (Č. TOMEK — L. BUDÍK 1981). Surrounding rocks are likely to have a non-granite character (cf. J. KADLEČÍK et al. 1980).

Orientation of N-S gravity isoanomals is anomalous in relation to the general NE-SW basin course. Isoanomals orientation is correlated with the course of isoseists of seismic energy propagation from the East-Alpine earthquakes (D. PROCHÁZKOVÁ — J. ZEMAN 1982). A. DUDEK (1980) found pre-Hercynian structural orientation in the Brunovistulicum, E. CLAR (1965) in Paleozoic

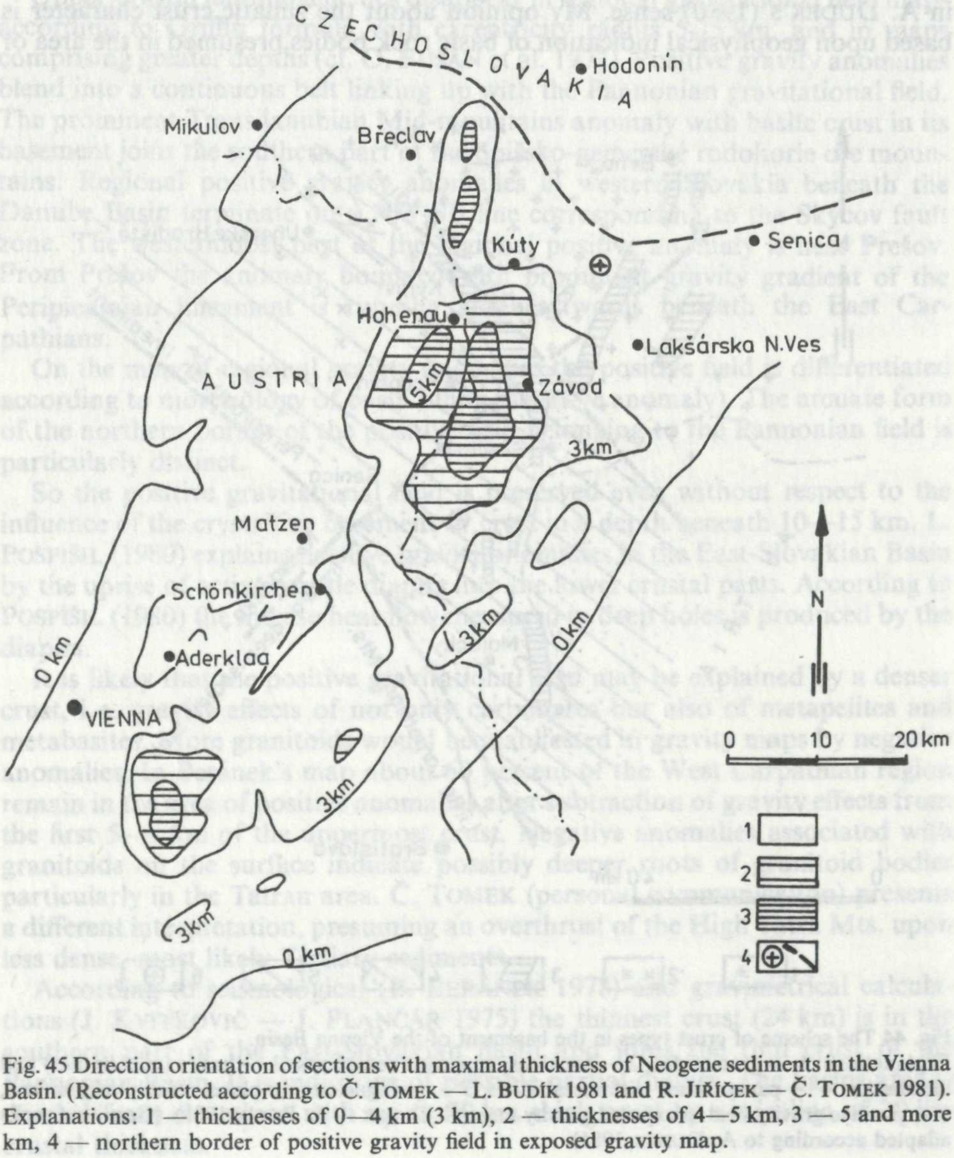



Fig. 45 Direction orientation of sections with maximal thickness of Neogene sediments in the Vienna Basin. (Reconstructed according to Č. TOMEK — L. BUDÍK 1981 and R. JIŘÍČEK — Č. TOMEK 1981). Explanations: 1 — thicknesses of 0—4 km (3 km), 2 — thicknesses of 4—5 km, 3 — 5 and more km, 4 — northern border of positive gravity field in exposed gravity map.

crystalline complexes of the Eastern Alps. I prefer association of the N-S gravity isoanomals pattern with hidden structures in the pre-Neogene substratum of the Vienna Basin nappes, for example with Mesozoic-Early Tertiary structures rift depressions beneath the nappe sequences or with structural pattern in simatic (oceanic) crust beneath the hypothetical rifts. Z. ROTH (1980) correlates the gravity minimum axis to the Outer/Inner West Carpathians boundary, and interprets a geophysical anomaly as a secondary tectonic boundary. My interpretation is based on changing negative anomaly axis course from NE-SW to N-S. So I think its source is in the basement of Alpine and West-Carpathian overthrusts. I presume rejuvenation of ancient discontinuity in the Neogene time, in pre-disposition of the origin of maximally mobile zones inherited in the Tertiary-Quaternary time (Kúty graben, course of Morava r.). The riftogenic character is also indicated by the position of an anomaly near the Brunovistulic/West-Carpathian (Carpathian-Pannonian megablock in the sense of Z. ROTH 1980) blocks which may have had a divergent movement tendency in the Mesozoic era.

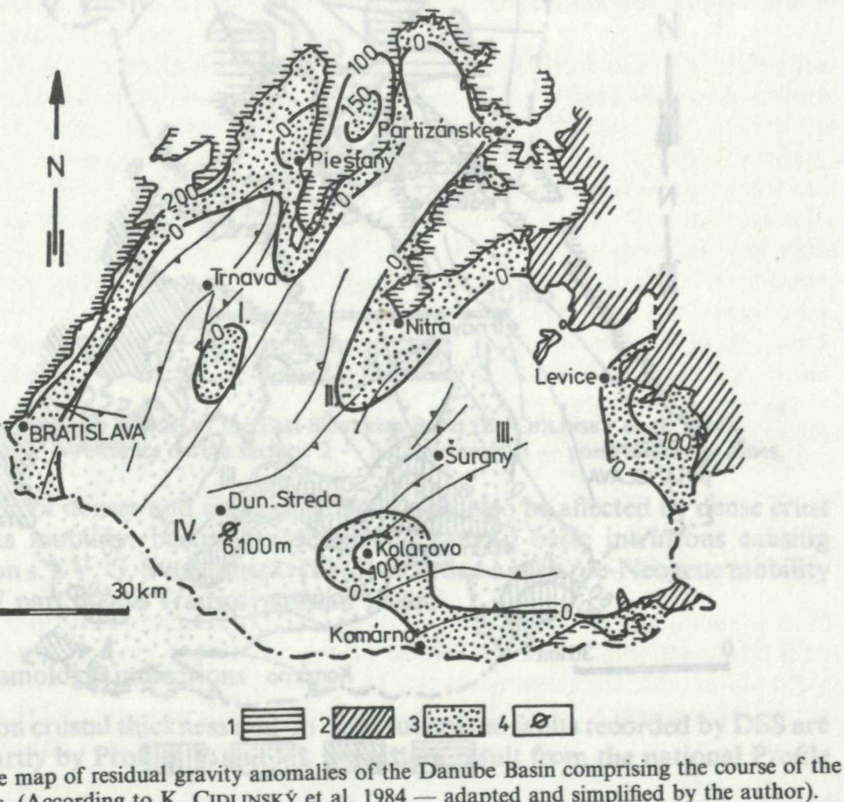



Fig. 46 The map of residual gravity anomalies of the Danube Basin comprising the course of the main faults. (According to K. CIDLINSKÝ et al. 1984 — adapted and simplified by the author). Explanations: 1 — margin of the basin, 2 — neovolcanites, 3 — positive residual gravity anomalies ($\mu m \cdot s^{-2}$), 4 — maximal thickness of Neogene sediments in central Pliocene depression. I — Inovec horst, II — Nitra horst, III — Levice horst, IV — central depression.

In the Danube Basin the gravitational field is usually explained by the relief of the pre-Tertiary basin basement (K. CIDLINSKÝ et al. 1984) — Fig. 46, 47. This concept is reasoned as regards marginal parts of the basin. In deeper parts of the basin the gravitational field may be explained by crustal composition and by gravity effects of mantle elevation. The positive Kolárovo anomaly may originate from a basic body — perhaps of a mantle origin—, from a body located in crust at a depth of about 10 km (K. CIDLINSKÝ et al. 1984). Positive vertical magnetic intensity anomalies are explained by neovolcanic bodies beneath Neogene sediments. The bodies are likely to form an arcuate belt around a satellite deep mantle diapir genetically related to the Pannonian diapir.

The East-Slovakian Basin has a segmented gravitational field also in derived anomalies. Sources of positive anomalies (Fig. 48) are not only in Neogene basement elevations, in hidden neovolcanic bodies but perhaps also in basite vari-aged crustal bodies (cf. magnetic anomalies — Fig. 49). L. POSPÍŠIL'S (1980) opinion about the presence of partial mantle diapirs is right. Great areal

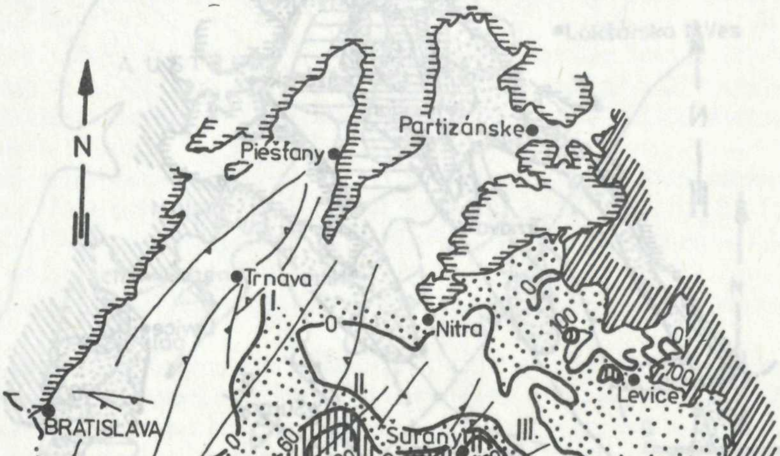



Fig. 47 The map of vertical magnetic intensity. (According to K. CIDLINSKÝ et al. 1984 — adapted and simplified by the author).
Explanations: 1 — margin of the basin, 2 — neovolcanites, 3 — 0—60 nT, 4—60 and more nT.

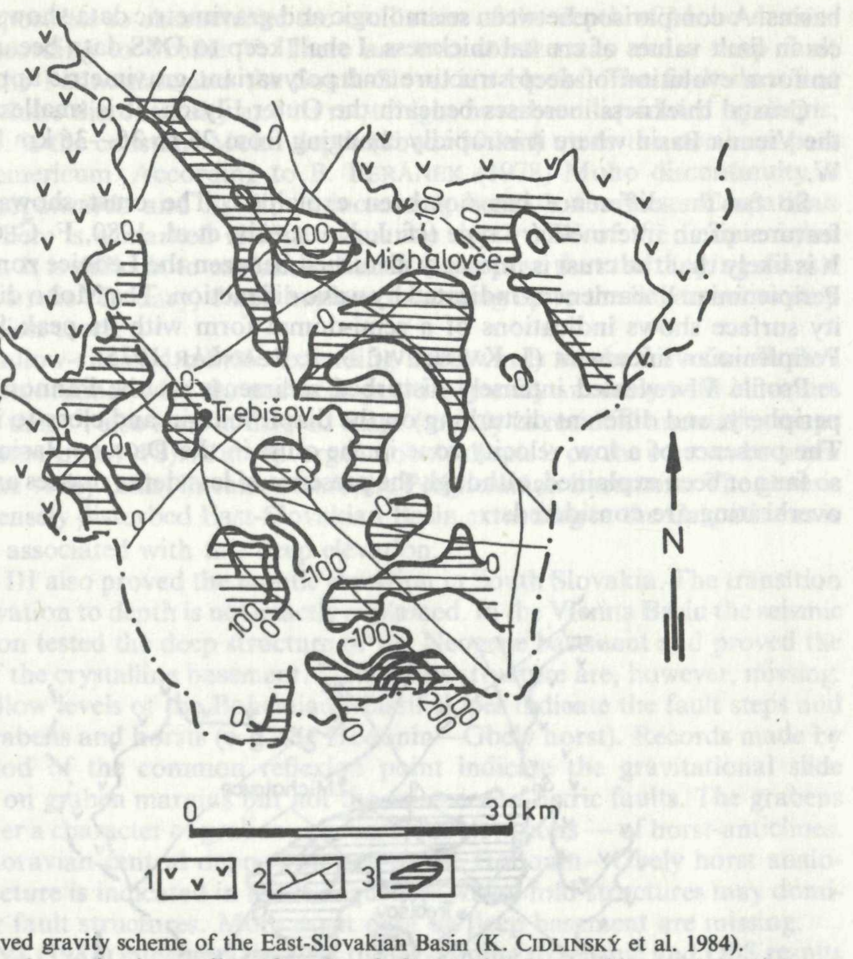



Fig. 48 Derived gravity scheme of the East-Slovakian Basin (K. CIDLINSKÝ et al. 1984).
Explanations: 1 — volcanics on the surface, 2 — fault tectonics, 3 — positive gravity zones.

distribution of denser and magnetic masses may also be affected by dense crust causing its mobility, basin subsidence and (ultra-) basic intrusions causing basification s. l. V. G. SVIRIDENKO (1976) recorded a high pre-Neogene mobility in the SW part of the Transcarpathian Basin.

### 3.1.2. Seismologic indications

The data on crustal thickness and on deep subcrustal faults recorded by DSS are offered partly by Profiles V and VI, more data result from the national Profile III.

In the West Carpathians the crustal thickness increases from the Pannonian block towards the High Tatra Mts. There it ranges up to 42 km. There is a 10 km difference in crustal thicknesses of the High Tatra Mts. and of southern Neogene basins. A comparison between seismologic and gravimetric data shows differences in limit values of crustal thickness. I shall keep to DSS data because of the uniform evolution of deep structure and polyvariant gravimetric approach.

Crustal thickness increases beneath the Outer Flysch. It is smallest beneath the Vienna Basin where it is rapidly changing from 28 to 35—36 km from E to W.

So far the difference has not been explained. The crust shows physical features of an intermediary type (cf. J. KADLEČÍK et al. 1980, F. ČECH 1984). It is likely that the crust is intensely disturbed between the Lednice zone and the Peripieninian lineament as indicated by wave diffraction. The Moho discontinuity surface shows indications of a semi-domal form with its peak SE of the Peripieninian lineament (J. KVITKOVIČ — J. PLANČÁR 1975).

Profile VI revealed intensely disturbed sediments on the Pannonian Basin periphery, and different disturbing on the diapir margin and close to its centre. The presence of a low velocity zone in the crust in the Danube Basin area has so far not been explained, although the presence of less dense masses and crustal overheating are considered.

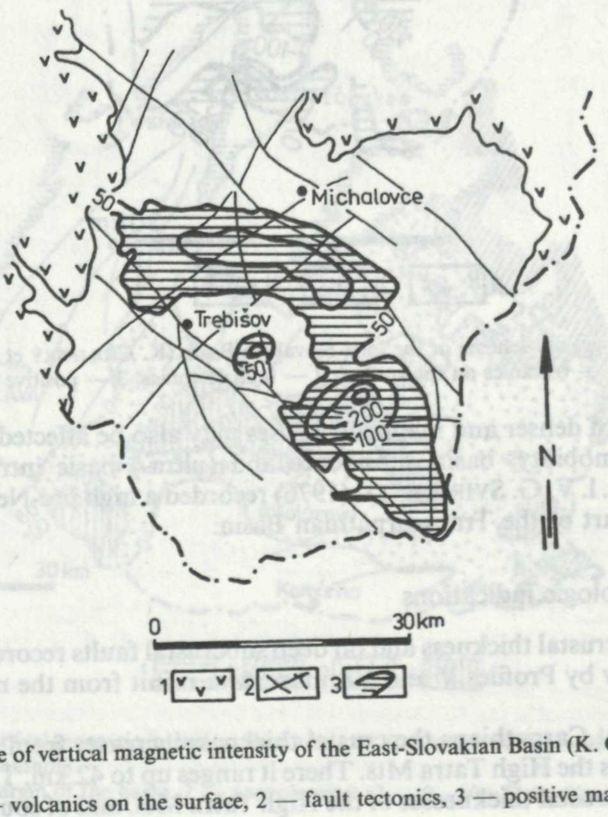



Fig. 49 The Scheme of vertical magnetic intensity of the East-Slovakian Basin (K. CIDLINSKÝ et al. 1984).
Explanations: 1 — volcanics on the surface, 2 — fault tectonics, 3 — positive magnetic zones.

It is impossible to characterize the oscillation of the depth of Moho discontinuity according to Profile V. There are two zonal areas and a deep fault between the Pannonian and the South-Slovakian blocks. The fault does not reach to Moho discontinuity and its root might have been liquidated by diapir. Significant is the constant Moho depth (about 30 km) up to the northern part of the Gemericum. According to B. BERÁNEK (1978) Moho discontinuity is dipping northwards and its dip is increasing beneath the Inner-Carpathian Flysch. There is a marked change in mantle depth NW of the Peripieninian lineament. It is difficult to explain it exactly. I suppose that the dipping mantle surface is a relic (Tertiary) diapir surface, extending up to the lineament to stop its expansion there.

The shallow-situated mantle extending northwards to the Levoča—Prešov line is in accordance with the northern positive gravity anomaly with its values up to 350 ums$^{-2}$, also seen in derived maps (mainly in residual anomalies maps — cf. J. IBRMAJER 1978). A northvergent lobate diapir is on the contact between the Slánske vrchy (hills) lineament and the Peripieninian lineament. The genesis of the intensely disturbed East-Slovakian Basin extending in the Carpathians is evidently associated with this deep elevation.

Profile III also proved the mantle elevation in South Slovakia. The transition of the elevation to depth is not exactly explained. In the Vienna Basin the seismic exploration tested the deep structure of the Neogene basement and proved the SE dip of the crystalline basement. Data on its structure are, however, missing. Only shallow levels of the Bohemian Massif slopes indicate the fault steps and partial grabens and horsts (e. g. the Hodonín—Gbely horst). Records made by the method of the common reflexion point indicate the gravitational slide tectonics on graben margins but not the existence of listric faults. The grabens have rather a character of graben-synclines, the elevations — of horst-anticlines. In the Moravian central depression and in the Hodonín—Gbely horst analogous structure is indicated in flysch sequences where fold structures may dominate over fault structures. More exact data on deep basement are missing.

G. PLÍVA (1981) interprets the structure according to seismic and DSS results and regards the large graben on the contact of the Bohemian Massif with the West Carpathians as aulacogen, presumably extending NE-SW-wards from the East Alps foreland to Ukraine. According to refraction seismic results the Neogene sediments thickness and that of the Mesozoic and Paleozoic basement are 22 km, the total crustal thickness is 35 km.

Data on deep structure also result from magnetotelluric measurements of O. PRAUS (1981). Besides vertical boundary on the Pericarpathian lineament there is a conductive zone elevation (perhaps of asthenosphere) indicated at the depth of 110—120 km. Another conductive zone is indicated at the depth of 60 km but its explanation is still missing.

The Danube Basin was only measured by several reflexion seismic sections and partly by the method of common reflexion point. A systematical research began in 1982. So far the dip of layers towards the basin centre was proved, and the horst-graben structure of the Neogene Basin and basement (Fig. 50), combined with flat undulate elevations and depressions (of the graben-syncline type) was indicated. In the western part — up to the central Pliocene depression — the seismic waves pattern is indicative of crystalline complexes, in the eastern part (Kolárovo, Komárno) it indicates consolidated sediments, perhaps of Mesozoic age. Thickness of the Neogene in the so-called central Pliocene depression situated in the extension of the Nitra Depression (between the Nitra and Levice horsts) — Fig. 46, ranges up to 6.100 m. This is the maximum known Neogene thickness in West-Carpathian Neogene basins.

The East-Slovakian Basin is sufficiently investigated by seismic refraction and reflex methods. Since the research was aimed at the structure of the Neogene filling, there are few data on the structure of the pre-Neogene basement. The measurements resulted in indications of the horst-graben structure, partly showing the character of horst-anticlines and graben-synclines (Fig. 51). Seismic investigations show that many geologically delimitated faults lack geophysical evidence.

### 3.1.3. Geothermic conditions

The area of Neogene basins with neovolcanic megastructures belongs among areas of an intense heat flow. The East-Slovakian Basin with the maximal temperatures in holes is close to the Pannonian Basin and its thermic values. Heat flow in the East-Slovakian Basin ranges up to 102.6—113.0 $mWm^{-2}$ (V. ČERMÁK 1975). Geothermal gradient ranges from 43 to 45 °C/km. Maximal value is 48 °C/km (I. MARUŠIAK — J. LIZOŇ 1976). The anomalous high temperature values exceed those in the Štiavnica—Kremnica area (heat flow density 108.9 $mWm^{-2}$).

The Danube Basin is another warmest area with a heat flow ranging up to 104,7 $mWm^{-2}$ and temperatures different from those in the East-Slovakian Basin (Tab. 9). By its temperature the area is closer to neovolcanic regions. Drilling in the basin was aimed at thermal waters prospection. In the Vienna Basin the lowest temperatures were measured. According to V. ČERMÁK (1981) the temperatures may be compared to those in areas of consolidated crust. It is, however, not confirmed by basin mobility and recent movements (see below). For a comparison I present the values measured in the foredeep i. e. the lowest temperatures (Table 9).

The three main Neogene basins also show a high geothermic gradient decreasing with the decreasing temperature in geologic units. So it cannot be associated with lithology or with hydrocarbons occurrences. The geothermic gradient is a value representative of thermal activity. The data quoted prove geothermic relation to the intense heat flow in the Pannonian Basin. The heat flow is associated to thinned crust and to mantle elevation.

I think the deep temperature values indicate still active endogenous processes and I do not explain them as reflecting different basin basements depth neither recent subsidence (I. MARUŠIAK — I. LIZOŇ 1976). Their relation to recent subsidence is indicative of a common endogenous cause.

Temperatures in °C in Neogene Basins at various depths beneath the Adriatic Sea level (after I. MARUŠIAK — I. LIZOŇ 1976)

Table 9

| Basin | Depth | | | | Geothermal gradient in °C/km |
|---|---|---|---|---|---|
| | −1000 m | −2000 m | −3000 m | −4000 m | |
| East-Slovakian | 58—76 | 102—125 | 149—167 | 170—192 on −3500 m | 43—45 max. 48 |
| Danube | 38—59* | 69—99 | 96—141 | | 32—35 max. 40 |
| Vienna | 32—59 | 65—98 | 97—110 | 114—121 | 25—32 max. 35 |
| Central-Slovakian neovolcanic region | 38—82 | 86—98 | | | 29—32 max. 34 |
| foredeep | 31—45 | 55—70 | | | 23—26 max. 27 |

Remark: *Minimum values denote anomalous low temperature measured in hole Dubové-2

The heat flow density 104,7—113,0 $mWm^{-2}$ (V. ČERMÁK 1975 and 1979) is also indicative of a relation to the Pannonian thermal structure. According to T. BOLDIZSÁR (1968) the structure is on its outer side bordered with the circum-Pannonian zone. Its heat-flow density is by 20,1 to 33,5 $mWm^{-2}$ lower. So the zone might be correlated to the lower-thermal circum-diapir zone — the rim syncline.

The heat-flow density of the East-Slovakian Basin, close to the Pannonian Basin is indicative of a partial upper mantle diapir (L. POSPÍŠIL 1980). Presumable oilparent rocks in the pre-Neogene basin basement may cause recent methanerelease at a depth of about 6 km. Migration might then be associated with deeper seated faults.

Changes corresponding to recent metamorphosis in the East-Slovakian Basin Neogene sediments are described by D. ĎURICA et al. (1979). According to these authors recrystallization of carbonate clayey shales commenced at the temperature of 150 °C and the pressure of about 50 MPa. The recrystallization limit reached by drilling in the depth between 2800—3000 m on an area of about several hundreds of sqkm is independent upon stratigraphic range of affected rock. Beginning with this depth the water vapour pressure increases as well as mineral grain size, and detritic muscovite changes into illite. Carbonization of organic matter and rock density increase, whereas porosity and rock water-content decrease.

### 3.1.4. Recent movements

Basins mostly represent areas of recent subsidence. Detailed recent movements measurements (P. VYSKOČIL 1981, P. MARČÁK 1978) prove, however, differenics, structures affecting rivers course, the course of the Leváre—Sološnica graben and other faults. Measurements of the recent movement horizontal component (P. VYSKOČIL 1981) recorded NW and SE movements along the basin margins, i. e. basin opening (Fig. 53).

In the Danube Basin the vertical movements field (P. MARČÁK 1978) is segmented to elevation- and depression zones (Fig. 52). There is an inversion between geologic structures and recent movements: morphologic elevations of the Inovec and Tribeč Mts. are in the field of maximum subsidence — in the Nitrianska pahorkatina upland up to 4,2 mm/year. The maximum Neogene subsidence area is in the recent elevations field up to 2,3 mm/year (Zlatná na Ostrove).

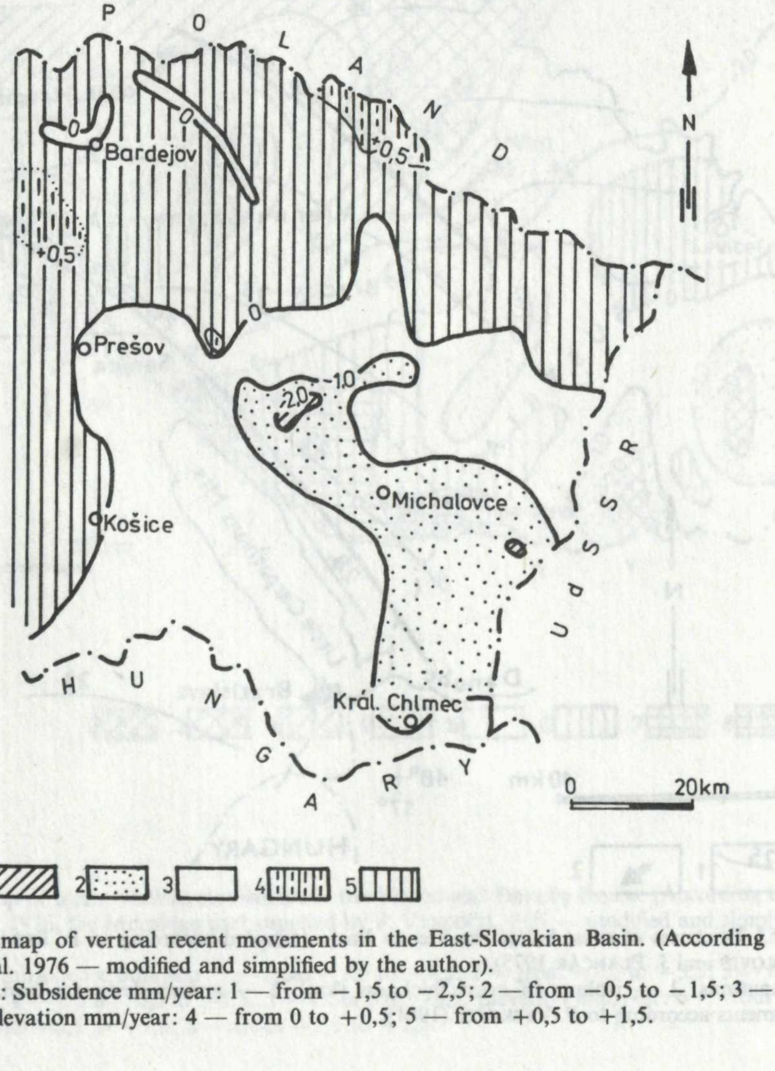



Fig. 54 The map of vertical recent movements in the East-Slovakian Basin. (According to P. MARČÁK et al. 1976 — modified and simplified by the author).
Explanations: Subsidence mm/year: 1 — from −1,5 to −2,5; 2 — from −0,5 to −1,5; 3 — from 0 to −0,5. Elevation mm/year: 4 — from 0 to +0,5; 5 — from +0,5 to +1,5.

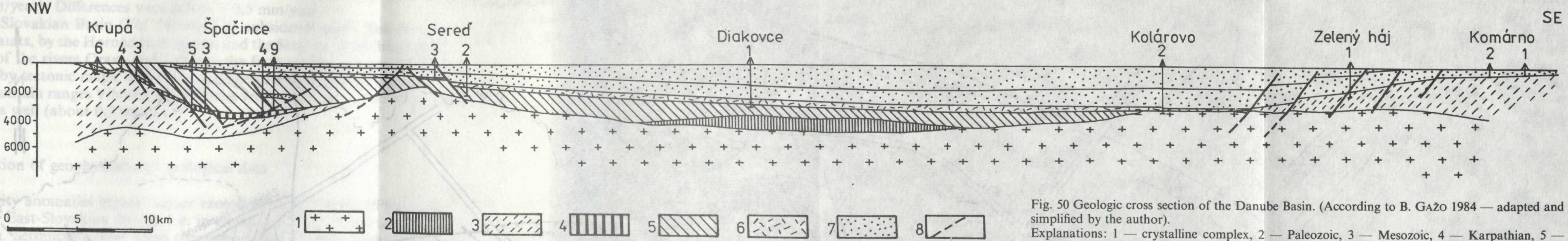



Fig. 50 Geologic cross section of the Danube Basin. (According to B. GAŽO 1984 — adapted and simplified by the author).
Explanations: 1 — crystalline complex, 2 — Paleozoic, 3 — Mesozoic, 4 — Karpathian, 5 — Badenian, 6 — Sarmatian, 7 — Pliocene + Pannonian, 8 — faults.

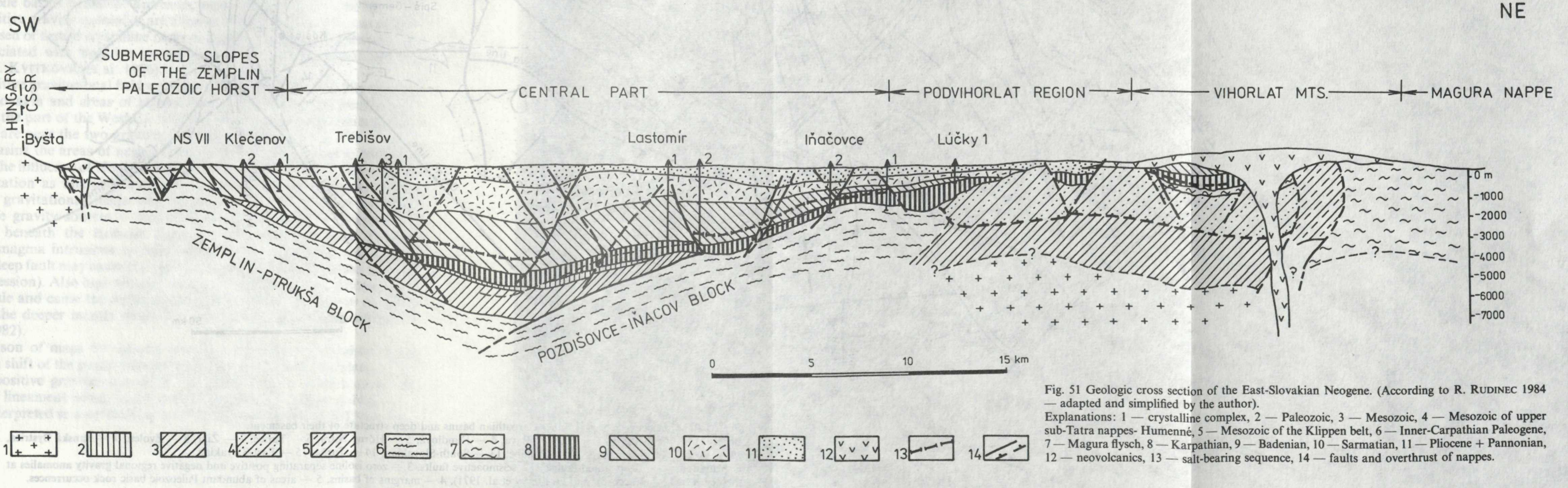



Fig. 51 Geologic cross section of the East-Slovakian Neogene. (According to R. RUDINEC 1984 — adapted and simplified by the author).
Explanations: 1 — crystalline complex, 2 — Paleozoic, 3 — Mesozoic, 4 — Mesozoic of upper sub-Tatra nappes- Humenné, 5 — Mesozoic of the Klippen belt, 6 — Inner-Carpathian Paleogene, 7 — Magura flysch, 8 — Karpathian, 9 — Badenian, 10 — Sarmatian, 11 — Pliocene + Pannonian, 12 — neovolcanics, 13 — salt-bearing sequence, 14 — faults and overthrust of nappes.

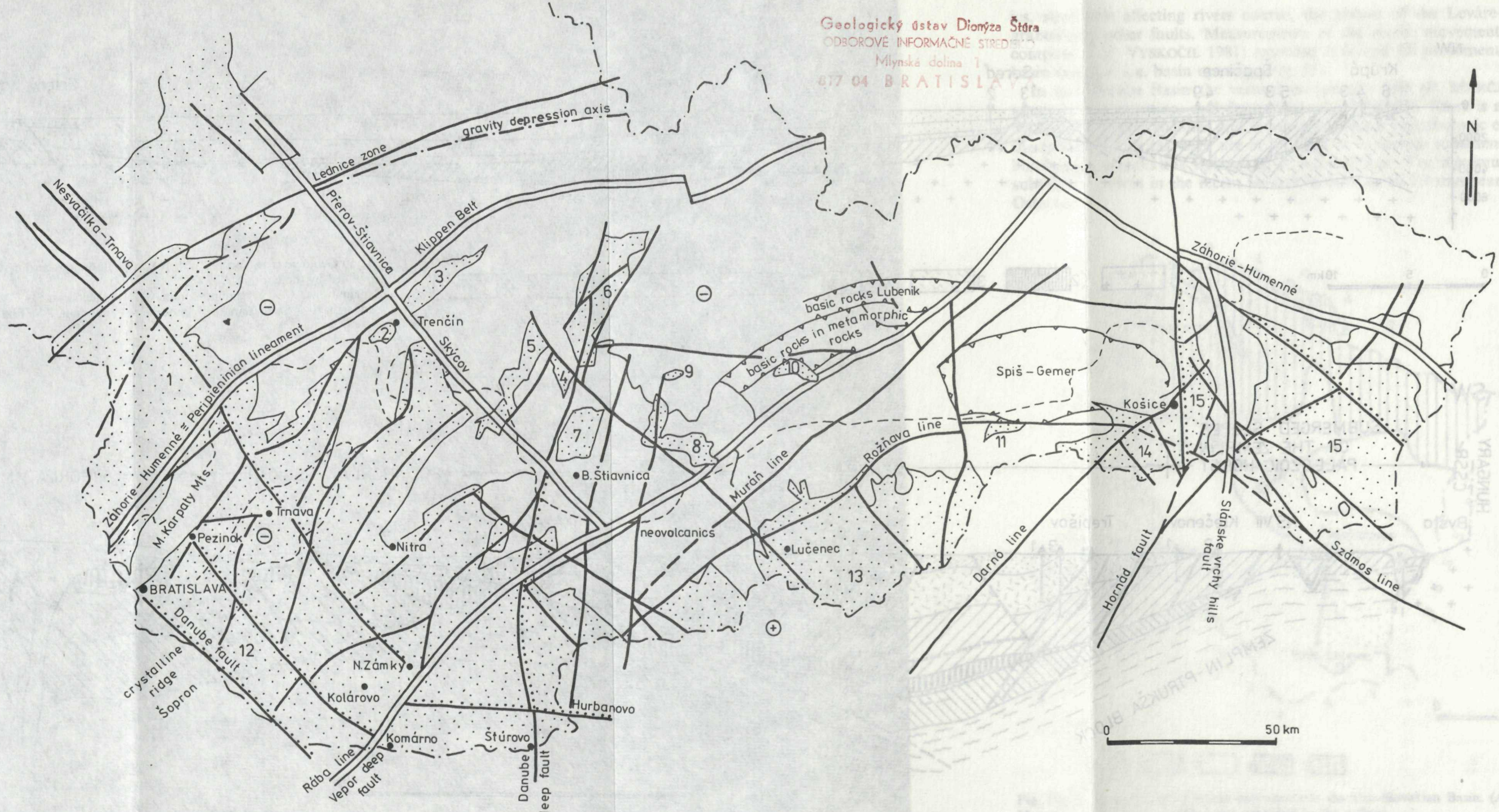



Fig. 55 Relation between Slovak Inner-Carpathian basins and deep structure of their basement.
Basins: 1 — Vienna, 2 — Trenčín, 3 — Ilava, 4 — Handlová, 5 — Horná Nitra, 6 — Turiec, 7 — Žiar, 8 — Zvolen, 9 — Banská Bystrica, 10 — Brezno, 11 — Rožňava, 12 — Danube, 13 — South-Slovakian, 14 — Turňa, 15 — East-Slovakian.
Explanations: 1 — deep-seated faults, 2 — seismoactive faults, 3 — zero isoline separating positive and negative regional gravity anomalies at the depth of 12 km (according to O. FUSÁN et al. 1971), 4 — margins of basins, 5 — areas of abundant Paleozoic basic rock occurrences.
Deep-seated faults and their seismicity according to O. FUSÁN et al. (1979).

In the South-Slovakian Basin the recent movements field is homogeneous, partly controlled by NW-SE faults. In the time of measurements the area was stable (0 mm/year). Differences were below $\pm 0,5$ mm/year.

The East-Slovakian Basin (Fig. 54) tends to subsidence controlled by submeridional faults, by the Hornád fault system and the Slánske vrchy (hills) fault. In the area of the rivers Orava and Laborec the fault system is more effective than shown by tectonic maps. The faults are likely to be youngest, still active. Volcanic mountain ranges Vihorlat and Slánske vrchy (hills) are involved in the subsidence as well (about 0,5 mm/year).

### 3.2. Correlation of geophysical and geological data

Positive gravity anomalies highest values exceed 300 ums$^{-2}$ beneath the Danube- and the East-Slovakian Basins, i. e. in connection with the positive Pannonian Basin. Seismically these areas correspond best to active faults, and — according to extrapolated DSS data — to the areas of thinned crust (between 28 and 30 km). There is a good correlation between gravimetrical and seismical data (J. IBRMAJER 1978). In the zone of increasing crustal thickness the depth of most mobile basins basement decreases, especially in the Danube Basin.

Local positive gravity anomalies are affected by basement elevations, mainly those composed of denser crystalline rocks or limestones, whereas granitoids are usually associated with negative gravity anomalies and with local isostatic anomalies (J. KVITKOVIČ et al. 1976).

In the positive gravitational field are largest Neogene basins, mainly those of the South Slovakia and areas of highest deep temperatures, whereas smaller basins are in the part of the West Carpathians with negative gravitational field. Most basins are near the two gravitational fields boundary (Fig. 55).

In large basins the areas of negative anomalies are interpreted as granitoid bodies or as the influence of Neogene filling (J. IBRMAJER 1978). I cannot regard this interpretation as unambiguous, since negative gravity anomalies are in transformed gravitational fields, for example in the East-Slovakian Basin (Fig. 43). The gravity sources — less dense masses — must be deeper than 2000 m, i. e. beneath the Košická kotlina (depression) basement. Possible hidden acid magma intrusions or other deep gravity sources on the Slánske vrchy (hills) deep fault may cause the uplift of the western margin of the Košická kotlina (depression). Also high temperatures may indicate partial melting of the crust or mantle and cause the negative gravitational field (SCHOLZ et al. 1971) or diapir in the deeper mantle which has not yet reached the lithosphere (J. NIKOLSKY 1982).

A comparison of maps of regional gravity anomalies in individual depth levels shows a shift of the positive/negative gravitational fields boundary, to the area of the positive gravitational field. The shift is more distinct along the Peripieninian lineament which in the gravity pattern appears as a NW overthrust or is interpreted as a subduction zone by some authors (e. g. J. LEXA — V.

KONEČNÝ 1979). The subduction may be testified by the Vienna Basin evolution and opening in the back region of subducting crust (H. D. KLEMME 1979). It is, however, not proved by its fault structure (see below). The shift of gravitational fields boundaries might also be indicated by the mushroom-shaped diapir.

### 3.3. Deep faults and block structure

O. FUSÁN et al. (1971) distinguished main basement blocks on the grounds of a correlation of geophysical and geological data. Neotectonic delimitation and internal segmentation were precised in 1979 by the same authors (Fig. 55). The Danube block, the South-Slovakian (Rudohorie—Pilich) fault and the Transtissia block have gravitational field mostly disturbed with positive anomalies. Largest Neogene basins are on these blocks. Smaller basins are on block margins, mainly on the contact of the above mentioned blocks with the Fatra —Tatry block. This rule also concerns the genesis of basins.

The Danube block and the East-Slovakian block show the most detailed internal segmentation into partial blocks. Most of them have differentiated gravitational fields and are limited by seismoactive faults. The latter are most frequent. These factors are in accordance with high mobility of basins and of their basement. On the Danube block is the deepest central Pliocene depression — the Gabčíkovo depression. It is partly controlled by the Rába line and the Hurbanovo fault. In my opinion, the older structural pattern manifested in N-E orientation of the maximal subsidence axis, was controlled by deep evolution of the Pannonian diapir.

In the East-Slovakian Basin, the most mobile pre-Neogene zones are on the contact with the Pannonian Basin, on the Peripannonian lineament in the sense of V. G. SVIRIDENKO (1976) — Fig. 56.

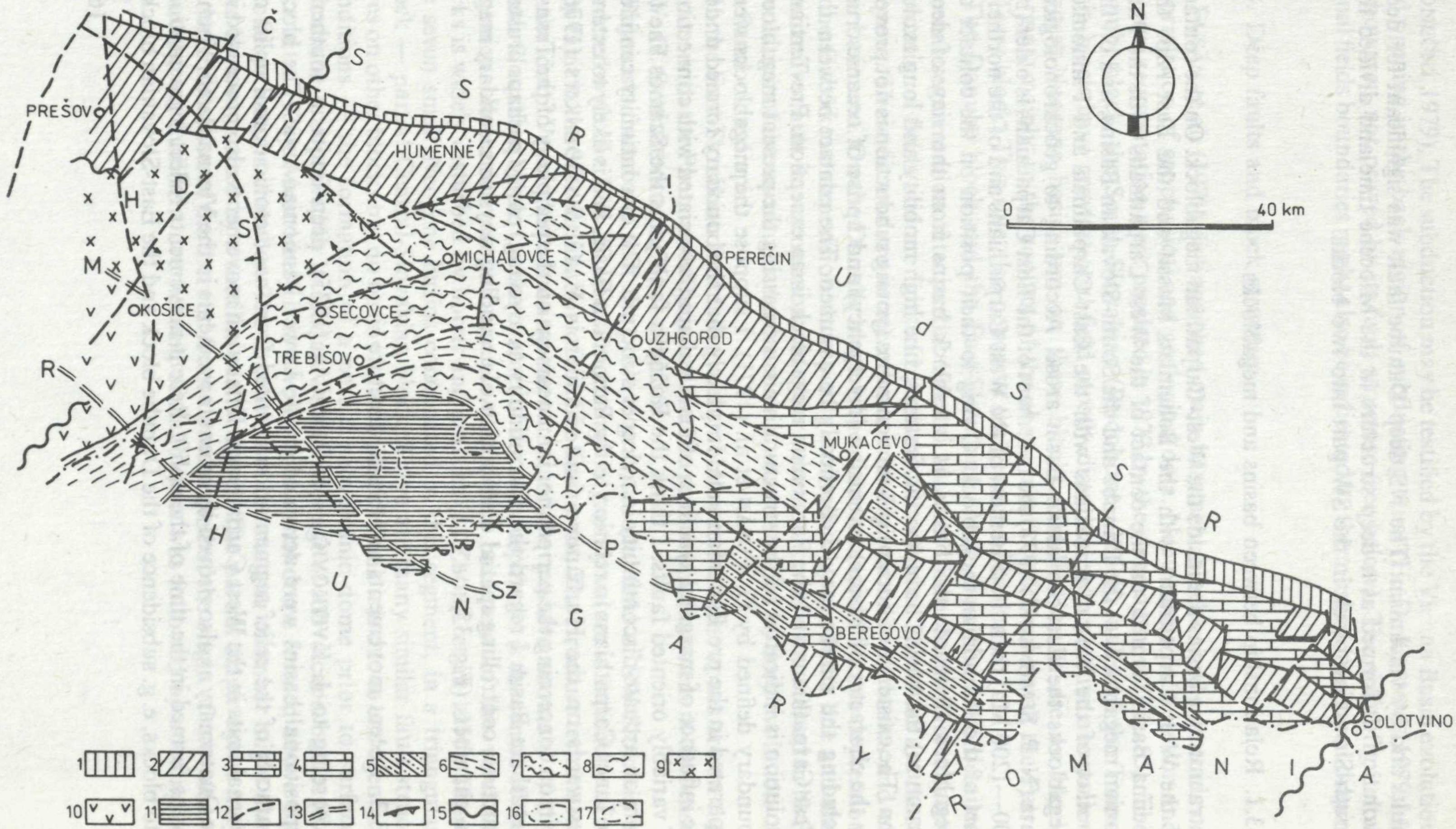
Physicists found out that the positive gravitational field of SW Slovakia is bordered with the NW-SE line. The change in gravity values is indicative of a deep fault in the SE continuation of the Láb lineament. The structure terminates on the deep Vepor fault limiting the extent of the positive anomaly of the South-Slovakian block. On the contact of the two deep faults is the centre of the Štiavnica—Kremnica neovolcanic elevation. The basement of the volcanic rocks is widely disturbed by horsts and grabens (L. ZBOŘIL et al. 1971). There are seven small Neogene basins concetrated in this segment, in a triangular block — partial Central-Slovakian block. There are many similar fault structures on other blocks but no basins formed there. I think it is due to endogenous conditions — the formation of a partial subvolcanic dome prior to surface volcanism — as proved by latest data on the older age of subvolcanic bodies (K. KAROLUS 1978). Most likely it is a volcanic-thermal dome formed above the secondary magma chamber. The formation of the magma chamber might also have been supported by a satellite partial diapir. The partial diapir uprise may be the explanation of the thermal front elevation and of the origin of subvolcanic intrusions. Sufficient overheating of sialic crust was followed by opening of faults and by volcanism. The NS deep Danube fault was significant for deep evolution. It formed as a deep structure in the Miocene time and divided the South-Slovakian block in the SW part into two blocks.

### 3.3.1. Relationship between basins and megablocks

Intramontane basins are inside the West-Carpathian megablock. On the contact of the West Carpathians with the Bohemian Massif and the East Alps the Vienna Basin formed, on the contact of the West Carpathians with the Pannonian megablock — the Danube and the South-Slovakian Basins, and on the contact of the West Carpathians with the East-Carpathian and Pannonian megablocks the East-Slovakian Basin arose. According to geochronological data (N. P. SEMENENKO 1977) the basement of the East Carpathians is older (by 700—1200 Ma) than the basement of the West Carpathians and of the northern part of the Pannonian megablock. Owing to their position on the contact of megablock (I denote the basins as interblock basins from the view of deep structure) the basins have specific features, like high mobility and long evolution. The existence of ancient subduction zones (young subduction is not proved by the character of basins), extension (divergency) and uprise of oceanic crust including the mantle may theoretically be assumed. The contact between the West Carpathians and the Pannonian megablock is an exception. Pre-Tertiary evolution is indicative of the structural units surpassing the present megablock boundary defined by O. FUSÁN et al. (1971). I suppose the megablocks were separated in the pre-Badenian time. It is likely that the boundary formed under the influence of marginal synclines of mantle diapir, associated with connection of variably oriented faults. I think the Rožňava fault joins the Számos line to form an arcuate discontinuity in a deeper structure. The discontinuity confines the Inner Carpathians in relation to the Pannonian Basin, and is likely to extend westwards on the Hurbanovo fault. Similar is V. G. SVIRIDENKO'S (1976) opinion concerning the peripannonian lineament on the periphery of the Transcarpathian Basin. I regard the discontinuity as a young, partly perhaps crustal structure, controlling spatial distribution of ascending ways or secondary magma chambers (Fig. 35).

### 3.3.2. Recent movements and block structure

According to J. KVITKOVIČ — J. PLANČÁR (1975), deep faults in southern depositional basins were activated in the Lower Pannonian time. The block character of the relief segmentation resulted from neotectonic domal uplift of internal units in the West Carpathians. Dilatation movements also proceeded in the Quaternary as also proved by recent movements in the Vienna Basin. Young basins formed in the time of the uplift. Recent movements reflect segmentation into blocks, e. g. subsidence of the Danube block and the East-Slovakian block.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The area of extensive subsidence is larger than the area of intense uplifts. The area of both divergently moving surfaces is approximately equal on the Slovak territory. The area of subsidence is larger than the area of uplifts in the entire Carpathian arc in relation to internal depressions.

The contour of subsidence in the East-Slovakian Basin correlates with positive gravitational field (J. KVITKOVIČ — J. PLANČÁR 1975).

Correlation of recent movements with crustal thickness and block structure proves deep activity. This indicated that young and recent movements may have their source in asthenosphere, in a different depth beneath uprising and subsiding West-Carpathian units. This might correlate with isostatic anomalies (J. KVITKOVIČ et al. 1976) if asthenosphere were an isostatic compensation horizon. If compensation is on the crust/mantle boundary — as assumed by J. KVITKOVIČ et al. (1976) — then the surface of mantle elevation may be regarded as tectonically active. Magmatic activity is indicated by geothermal calculations (see Chapter 3.1.3.).

## 3.4. Composition of basin basement

The basin basement was studied in detail by O. FUSÁN et al. (1971). A. BIELA (1978) revised later the basement rocks classification.

Ranging of a basement formation to a certain tectonic unit could inspire an idea of equal material composition of the two (the basement formation and the tectonic unit). Also the concept of a granite layer with slightly variable thickness in deep sections of O. FUSÁN et al. (1971) inspires the idea of a homogeneous crust with mature granite layer. M. MAHEĽ (1978b) pointed out to heterogeneous crust and slight granitization of the North-Gemeric zone. Explanation of

Fig. 56 Schematic geologic map of pre-Neogene basement of the Trans-carpathian Depression and East-Slovakian Basin. [Modified and simplified according to maps by D. ĎURICA (1982) and V. G. SVIRIDENKO (1976)].
Explanations: 1 — Klippen Belt, 2 — Paleogene (Inner Carpathian) in the U.S.S.R. of the Buda type, near Humenné in the basement with Mesozoic — mostly Carboniferous, 3 — Cretaceous sediments and diabases (deep-water evolution) and Paleogene, 4 — Cretaceous sediments (frequent pelites), in S-E part of the continental type; 5 — a) Jurassic — Triassic (deep-water with basics), non-distinguished Triassic partly of continental type, 6 — non-distinguished Mesozoic (in places with Paleogene?) mostly of flyschoid character, 7 — metamorphosed Paleozoic — Mesozoic and Paleogene (of continental type), 8 — Paleozoic metasediments, 9 — crystalline complex — Mesozoic of the Čierna hora unit and Mesozoic of the Choč unit, 10 — Gemericum, 11 — the Zemplín unit (Paleozoic, Mesozoic, Paleogene (in the U.S.S.R. of the Buda type), 12 — supposed faults, 13 — deep-seated faults, 14 — assumed overthrust, 15 — borders of transversal Pannonian-Volyn depression of V. G. SVIRIDENKO (1976), 16 — the main magnetic anomalies, 17 — main positive gravity anomalies in the czechoslovakian part. Significant faults: M — Margecany, R — Rožňava, D — Darnó, P — Peripannonian (according to V. G. SVIRIDENKO 1976), the direction NE-SW correlates with Zagreb-Kulcs deep-seated fault, H — Hornád, S — Slánsky, Sz — Sámos.

basin genesis only by thinned crust may lead to schematism. This is why I used A. BIELA'S (1978) data about deep holes, and revised rock composition in depression areas. I have also studied the relation of these areas to the basement with the envelope series or with overthrusts.

Material composition was also studied by O. FUSÁN et al. (1971) during his classification of positive anomalies. Their analysis shows that some anomalies may be caused by deep basic rock bodies or volcanic bodies. A larger part of the area is interpreted by relief elevations.

Smaller basins on the Fatra—Tatry block have a flysch basement or are underlain by Carpathian nappes. The position of basins is controlled by fault tectonics, frequently on a crossing of two faults of different course. The main longitudinal faults separate granitoid elevations with cover series from nappe units mostly concentrating in depressions.

Elevation units were involved in extensive subsidence in the northern part of the Danube Basin. Elevations sank to the depth of 1—2 km. In depressions with granite basement are downsagged cover series and overthrust sheets to the depth ranging to 3 km. Large basins are most complicated as for deep structure.

### 3.4.1 Vienna Basin

The Vienna Basin is on the contact of the West Carpathians with the Bohemian Massif. The composition of the pre-Mesozoic basin basement is not known and it has not been reached by deepest holes. Regionally the basin belongs to the Bohemian Massif (F. NĚMEC — G. KOCÁK 1976), to the Brunnia block with simatic crust according to J. ZEMAN (1978) or to an autonomous block in the sense of J. WEISS (1977). According to H. D. KLEMME (1978) the basin formed on transitional crust between the continental and oceanic types. Z. ROTH (1980) also placed the southern platform margin beneath the Vienna Basin and associated its genesis not with the platform margin but with sinistral rotation of the so-called Carpathian-Transylvanian block (beneath the Pannonian Basin) and of the Carpathians towards the Klippen Belt. Deepest holes in the basin — Šaštín — reached to the depth of 6505 m and the Lakšárska Nová Ves — 7 to the depth of 6405 m, both in the Triassic, most likely of the Choč nappe. According to F. NĚMEC and G. KOCÁK (1976), pre-Neogene basements consist of nappes of the Inner West Carpathians and Eastern Alps, downsagged nappes of the Ždánice and Magura Flysch, and Klippen Belt slices (Fig. 57). Reflex-seismic profiles show that the basement of autochthonous units of the Vienna Basin may be in the depth of 7—8 km. According to the DSS profile VI the autonomous crustal block on the slopes of the mantle elevation is dipping westwards. The crustal block in the sense of B. BERÁNEK et al. (1972) and G. PLÍVA (1981) forms a large graben.

**Basin-underlying crustal types.** The positive gravitational field of Brunnia is perhaps due to thin granite crust and I think that towards the east basic magmatites dominate in the crust (F. ČECH 1984, Fig. 44). Basing on geophysical data, J. KADLEČÍK et al. (1980) presume a crust composition beneath the basin different from that in orogenic regions. Also G. PLÍVA (1981) presumes-basic crust composition. Larger basic rocks bodies are indicated by magmatic anomalies, e. g. near Břeclav. The anomaly is likely to join the basic rock zone of the Brno Massif (Č. TOMEK — L. BUDÍK 1981) — Fig. 44. Elevation of the basalt layer is indicated by higher P-waves velocities along the Lednice deep fault (B. BERÁNEK — J. WEISS 1979). It may be due to ancient mantle elevation. The existence of basic-simatic crust may be explained from five viewpoints; namely:

a) the crust is a rest of non-subducted Mesozoic oceanic crust of Paleotethys (A. TOLLMANN 1978),

b) the crust is dragged out of the basalt layer by subduction (B. BERÁNEK and J. WEISS 1979),

c) the crust resulted from basification (oceanization) of platform crust during Alpine mobilization (Z. ROTH 1980),

d) may the crust have formed by basification during the ancient diapir intrusion in the Mesozoic?

e) the crust is a relic of perhaps original oceanic crust, slightly granitized and non-granitized, therefore mobilized by Hercynian events, and in the Mesozoic time. The lower crust may acquire physical (and material?) properties of the mantle (cf. I. A. REZANOV 1980).

I prefer the last interpretation (e), admitting possible pre-Tertiary diapirism associated with riftogenesis (cf. A. TOLLMANN 1978 — NNE-SSW Jurassic graben).

In the Vienna Basin the faults extending to the basement are crossing:

a) From the Bohemian Massif to N-S, NNE-SSW, ENE-WSW and W-E (the Leváre — Sološnica graben) and to NW-SE those running from the Nesvačilka — Vranovice graben (A. DUDEK — V. ŠPIČKA (1975); M. MAHEĽ (1978b) extrapolated the faults as far as the Pannonian Basin;

b) from the Outer Carpathians — NE-SW faults.

The NW-SE faults are regarded as Precambrian (P. GRECULA — Z. ROTH 1978) running from the subducted platform (of the Bohemian Massif) beneath the West-Carpathian unit. The W-E and N-S structural courses are regarded as oldest in the Czech part of the Bohemian Massif by J. ZEMAN (1978) and in the crust of the eastern part of the Bohemian Massif (Brunnovistulicum) by A. DUDEK (1980).

T. BUDAY — J. SENEŠ (1967) associated the location of the basin with the crossing of the Peripieninian and Labe lineaments (the so-called Sudetic fault of the Upper—Moravian valley). Latest geophysical and geological data prove the utmost significance of the Lednice and Peripieninian deep faults for the location of the Vienna Basin. The Láb fault system had only a subsidiary role in the northern basin margin.

Thickness of Neogene filling in deepest blocks (the Leváre graben, the Kúty graben) ranges to more than 5500 m, and the Vienna Basin together with the Danube Basin and the East-Slovakian Basin belong among the most mobile

Miocene basins on the periphery of the Pannonian mantle diapir. As for the deep structure it is interesting that the Vienna Basin and the Transcarpathian Basin join the Peripieninian lineament.

The structure of the Vienna Basin comprises NW-SE faults, parallel to but not identic with the Láb system. The faults join the Nesvačilka — Vranovice graben SE of Brno (A. DUDEK — V. ŠPIČKA 1975). O. FUSÁN et al. (1979) found the continuation of this structural strike in the Trnava — Nové Zámky line and in further two parallel faults in the Danube lowland (Fig. 58). The DSS data indicate a depth change in the crust/mantle boundary and the mantle elevation margin is running there. By extrapolation I complemented the mantle surface (Moho) isohypses under the Vienna Basin, using the data published by J. KVITKOVIČ — J. PLANČÁR (1975).

Correlation of the depths of Neogene basins basement, drilling and fault tectonics conceived by geophysicists show that:

a) The Vienna Basin and the Danube Basin have some common features like the deepest sagged blocks joining the zone of considerable changes in crustal thickness with a great horizontal gradient of thickness change — 60 m/km (F. ČECH 1980b). In the central part of the West Carpathians the gradient ranges to 80 m/km. The evolution of the two basins was also similar.

b) The most mobile blocks are delimited by longitudinal NNE-SSW deep faults and by transversal NW-SE deeper faults. The newly defined Nesvačilka — Trnava deep fault zone is the NE boundary of mobile blocks. The fault zone borders the mobile Danube — Záhorie sector (the Danube sector was mobile in the Pannonian and Pliocene times, the Záhorie sector in the Badenian and Sarmatian; F. ČECH 1980b, 1982). The maximum sedimentation (subsidence) rates in the Vienna Basin were determined for the Karpathian (8 cm/100 years) and the Badenian (5,5 cm/100 years) — D. VASS — F. ČECH 1983, Fig. 59. In the Pliocene the subsidence rate was low — 1,6—0,6 cm/100 years.

The elevation separating the two basins, longitudinal to their margins and transversal to the mobile zone, is kinetically associated with the Peripieninian lineament and with neotectonic movements of an elevation character in this segment of the lineament.

Structural analysis of the Vienna Basin basement and of the basement of the Danube Basin, as well as a correlation of geological and geophysical data have revealed a so far unknown regional-tectonic segmentation and structural elements integrating the evolution and deeper structure of two seemingly genetically different basins situated near the contact of two megablocks. This fact emphasizes significance of the Neogene restructure under the influence of the margin mantle diapir of the Pannonian Basin (F. ČECH 1980b).

#### 3.4.1.1. Basin evolution

It is likely that the area of the Vienna Basin was complicated. In the Tertiary a part of the basin was in a position of Alpine—Carpathian tectogen foredeep.

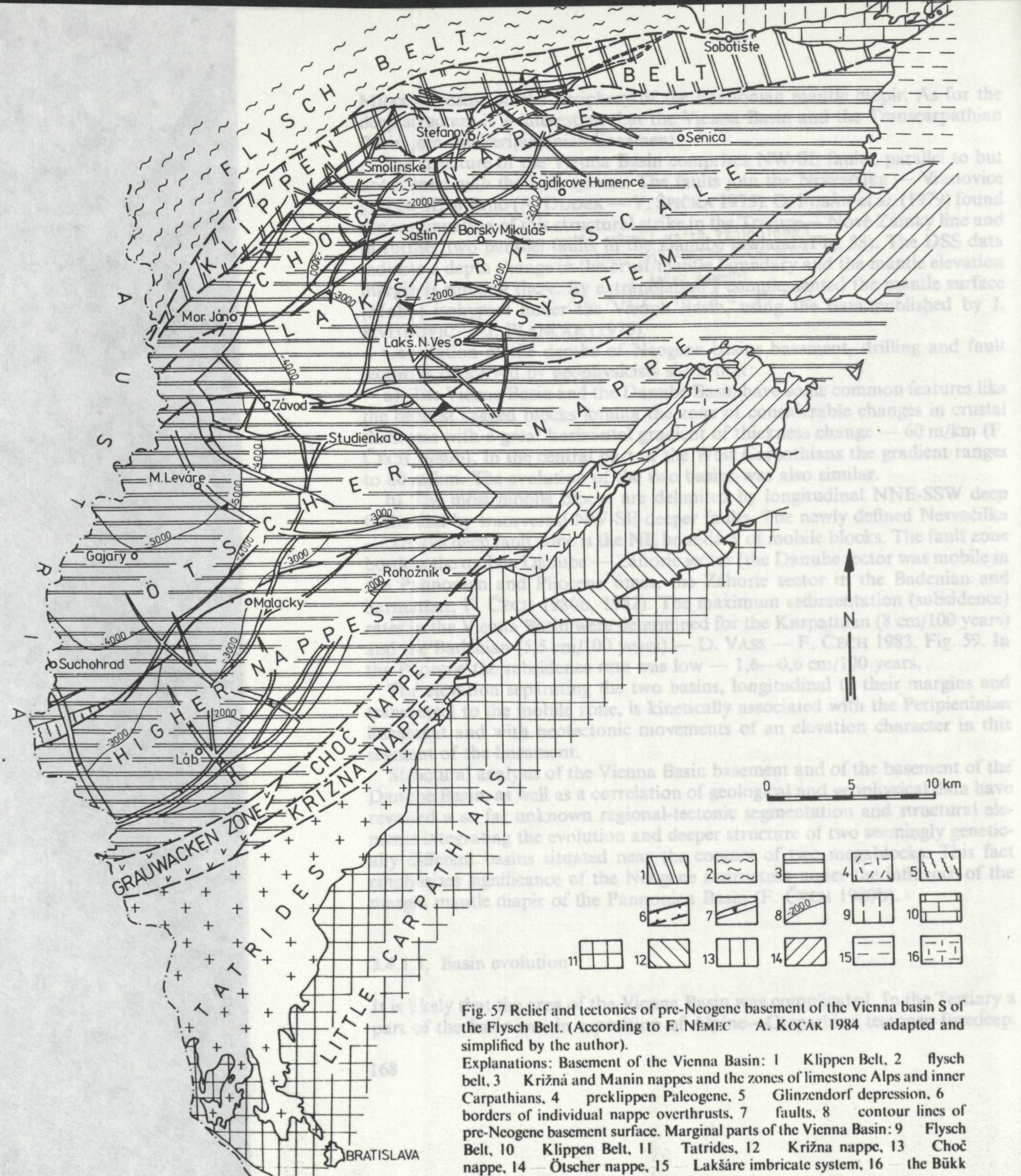


Fig. 57 Relief and tectonics of pre-Neogene basement of the Vienna basin S of the Flysch Belt. (According to F. Němec — A. Kocák 1984 adapted and simplified by the author).
Explanations: Basement of the Vienna Basin: 1 — Klippen Belt, 2 — flysch belt, 3 — Križná and Manín nappes and the zones of limestone Alps and inner Carpathians, 4 — preklippen Paleogene, 5 — Glinzendorf depression, 6 — borders of individual nappe overthrusts, 7 — faults, 8 — contour lines of pre-Neogene basement surface. Marginal parts of the Vienna Basin: 9 — Flysch Belt, 10 — Klippen Belt, 11 — Tatrides, 12 — Križna nappe, 13 — Choč nappe, 14 — Ötscher nappe, 15 — Lakšáre imbricate system, 16 — the Bükk depression.

The consequent nappes accumulation with their imbricate structure (F. NĚMEC 1984) affected formation of the complicated allochthonous structural level. Sedimentary sequences — except folds — also show a monoclinal dip. Sequences in higher parts of the level usually have a greater dip (55—85°), deeper

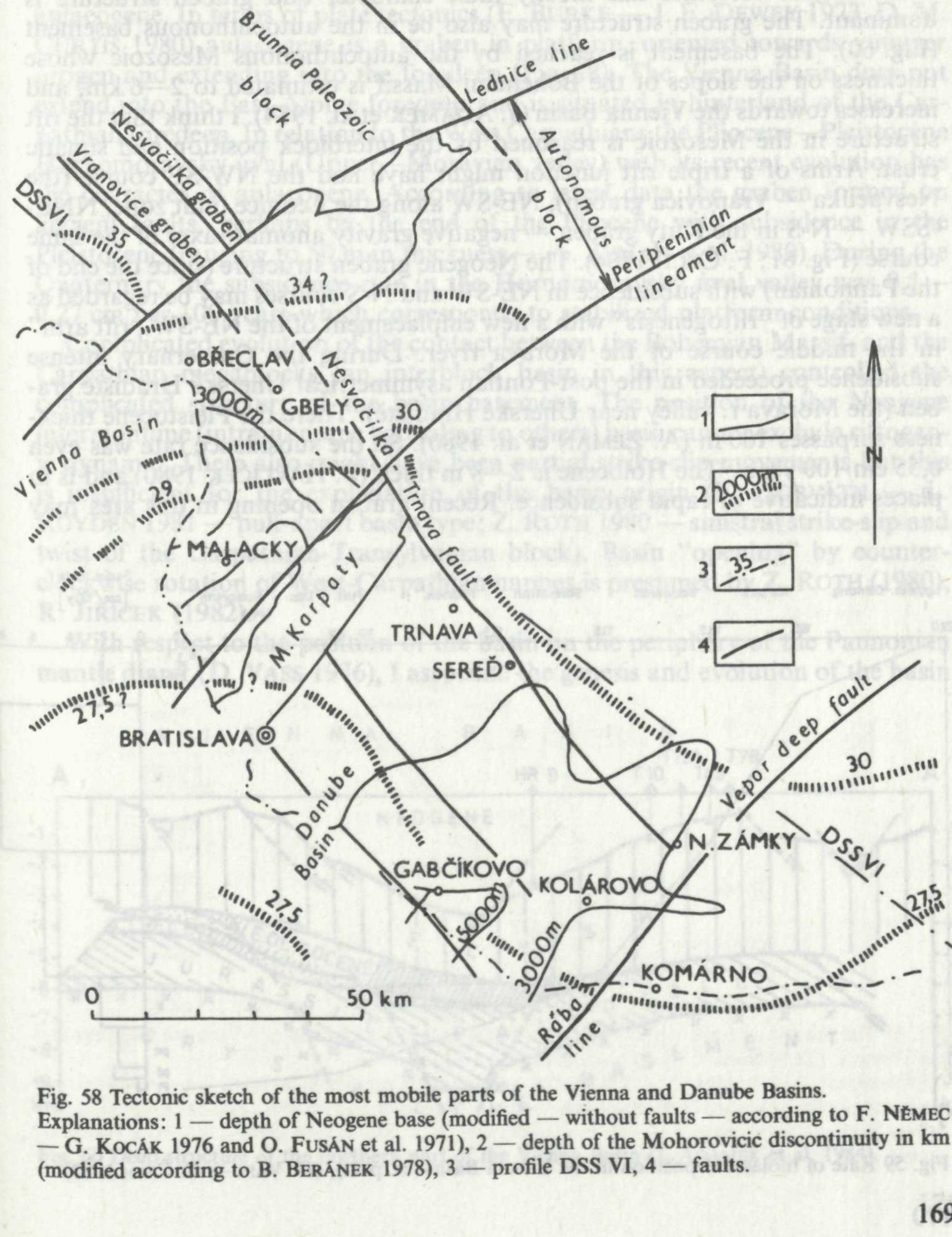


Fig. 58 Tectonic sketch of the most mobile parts of the Vienna and Danube Basins. Explanations: 1 — depth of Neogene base (modified — without faults — according to F. NĚMEC — G. KOCÁK 1976 and O. FUSÁN et al. 1971), 2 — depth of the Mohorovicic discontinuity in km (modified according to B. BERÁNEK 1978), 3— profile DSS VI, 4 — faults.

sequences dip at about 40°. The nappe pattern was modified by mechanism of nappe movements and Neogene subsidence. The autochthonous subhorizontal layering of Cretaceous sediments is likely not to have been affected by overthrust movements as shown by the hole Zistersdorf-1 in Austria. The nappes-underlying sediments are not folded and the basin was perhaps not compressed. The Neogene tectonics has mostly fault character and graben structure is dominant. The graben structure may also be in the autochthonous basement (Fig. 60). The basement is reached by the autochthonous Mesozoic whose thickness on the slopes of the Bohemian Massif is estimated to 2—6 km, and increases towards the Vienna Basin (J. ADÁMEK et al. 1984). I think that the rift structure in the Mesozoic is reasoned by the interblock position and simatic crust. Arms of a triple rift junction might have had the NW-SE course (the Nesvačilka — Vranovica graben), NE-SW along the Lednice fault zone, NNE--SSW — N-S in the Kúty graben — negative gravity anomaly axes of the same course (Fig. 61; F. ČECH 1984). The Neogene graben structure (since the end of the Pannonian) with subsidence in NE-SW and N-S courses may be regarded as a new stage of "riftogenesis" with a new emplacement of the NE-SW "rift arm" in the middle course of the Morava river. During the Quaternary intense subsidence proceeded in the post-Pontian asymmetrical Uherské Hradiště graben (the Morava r. valley near Uherské Hradiště). There the Pleistocene thickness surpasses 100 m (A. ZEMAN et al. 1980). So the subsidence rate was even 0,55 cm/100 years. The Holocene is 2—8 m thick (P. HAVLÍČEK 1980) and is in places indicative of rapid subsidence. Recent graben opening in this area may

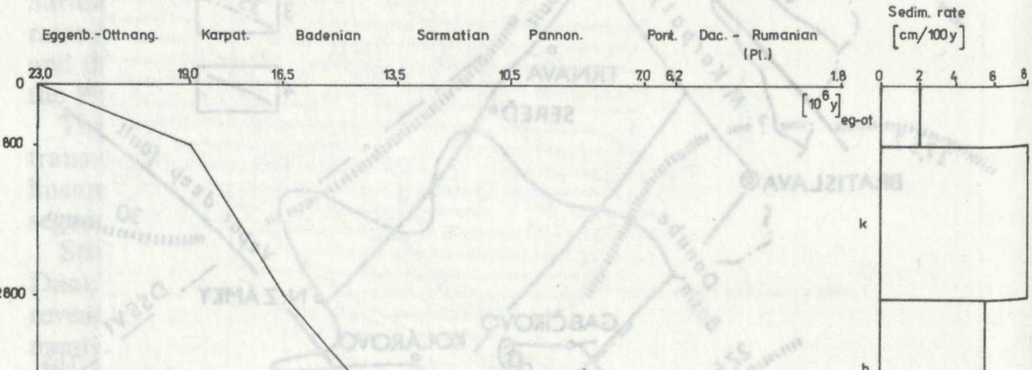


Fig. 59 Rate of molasse deposition in the Vienna Basin (cs. part), (D. VASS — F. ČECH 1983).

be deduced from measured divergent courses of strike-slip movements. For example the rivers Dyje and Morava have a triplet junction pattern, so a failed NW-SE arm might have the character of pseudoaulacogene in the sense of D. M. CURTIS (1980). I regard it as possible from genetic viewpoint.Denoting the Nesvačilka — Vranovice graben as aulacogene (F. PÍCHA 1979) or ranging the whole basin to aulacogene (G. PLÍVA 1981) is contradictory to the definition of aulacogene. In terms of plate tectonics (F. BURKE — J. F. DEWEY 1973, D. M. CURTIS 1980) aulacogene is a graben in platform, oriented towards younger orogen and extending into the foredeep (Fig. 18). The Vienna Basin does not extend into the East-Alpine foredeep and is situated in hinterland of the Carpathian foredeep. In relation to the West Carpathians the Pliocene—Pleistocene Hornomoravský úval (Upper—Moravian valley) with its recent evolution has the character of aulacogene. According to latest data the graben formed on ancient faults, perhaps by the end of the Pliocene with subsidence in the Pleistocene (ranging to 50 m in thickness — A. ZEMAN et al. 1980). During the Quaternary the subsidence rate in the Hornomoravský úval valley was 0,1—0,27 cm per 100 years which corresponds to stabilized platform conditions.

Complicated evolution of the contact between the Bohemian Massif- and the Carpathian megablocks (an interblock basin in this aspect) controlled the complicated structure of the basin basement. The position of the Neogene intermontane (intramontane according to others) basin cannot exclude riftogenic dynamic. There also might have been partial strike-slip movements but this is insufficient for the explanation of the basin origin (F. HORVÁTH — L. ROYDEN 1981 — pull-apart basin type; Z. ROTH 1980 — sinistral strike-slip and twist of the Carpathian-Transylvanian block). Basin "opening" by counterclockwise rotation of West-Carpathian nappes is presumed by Z. ROTH (1980), R. JIŘÍČEK (1982).

With respect to the position of the basin on the periphery of the Pannonian mantle diapir (D. VASS 1976), I associate the genesis and evolution of the basin

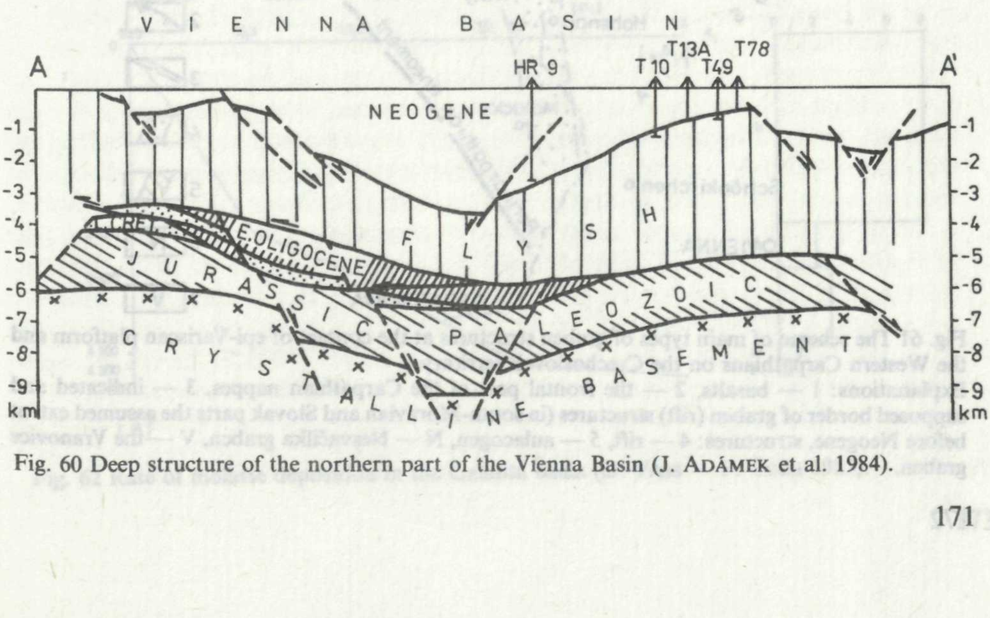


Fig. 60 Deep structure of the northern part of the Vienna Basin (J. ADÁMEK et al. 1984).

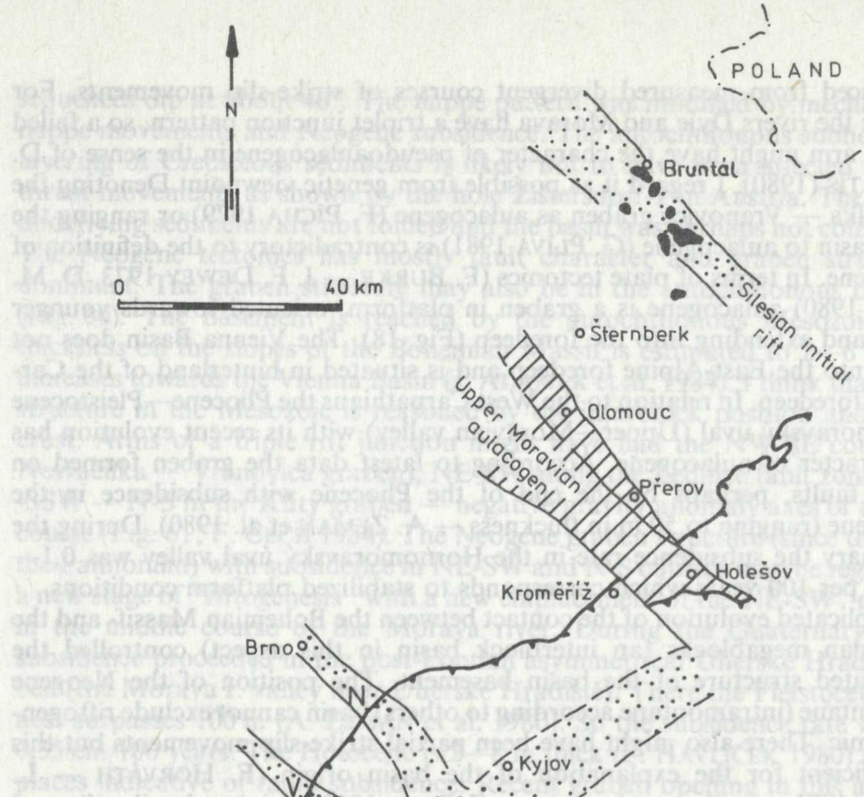


**Fig. 61 The scheme of main types of graben structures at the contact of epi-Variscan platform and the Western Carpathians on the Czechoslovak territory.**
**Explanations: 1 — basalts, 2 — the frontal part of the Carpathian nappes, 3 — indicated and supposed border of graben (rift) structures (in south-Moravian and Slovak parts the assumed extent before Neogene, structures: 4 — rift, 5 — aulacogen, N — Nesvačilka graben, V — the Vranovice graben.**

with dynamic and thermal regime of mantle diapir and its effects on its surroundings. I presume the Mesozoic riftogenic dynamics along the Lednice deep fault, possibly associated with partial mantle elevation on the (divergent?) contact between the megablocks. It is likely that the Pannonian diapir started to form by the end of the Mesozoic (D. VASS 1979, F. ČECH — J. ZEMAN 1982). The dense simatic crust sank into the marginal syncline, and Alpine and Carpathian nappes shifted into the thus forming depression. Partial subduction of dense crust beneath the diapir margin might have proceeded below the peri-Pieninian lineament. The subsidence into the marginal syncline might have caused repeated crustal downwarping beneath the present Vienna Basin. The lower crust in the zone of downwarp might have acquired physical properties of the mantle. This is why Moho-discontinuity is vague in geophysical records.

### 3.4.2. Danube Basin

Like the Vienna Basin, the Danube Basin is on a boundary of geologically different units with variable consolidation, namely the West-Carpathian and the Pannonian units. Subsidence in the Danube Basin commenced in the Early Miocene. The northern part of the basin belongs to marginal depressions of the Pannonian Basin. The maximum subsidence 7,8 cm per 100 years — proceeded in the Badenian (D. VASS — F. ČECH 1983, Fig. 62). The Miocene maximum subsidence proceeded in the northern part in the Trnava-, Galanta-, Komjatice-, and Dubník depressions. During the Pannonian the maximum subsidence centre shifted southwards. During the Pannonian the subsidence rate was

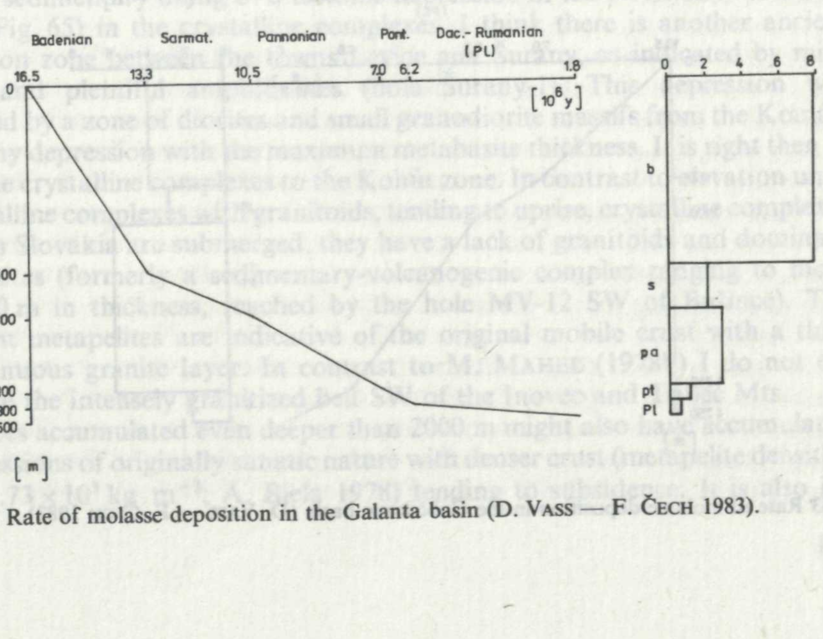


Fig. 62 Rate of molasse deposition in the Galanta basin (D. VASS — F. ČECH 1983).

4 cm/100 years, during the Dacian—Rumanian up to 6 cm/100 years (D. VASS — F. ČECH 1983) — Fig. 63. Thickness of the Quaternary ranges up to 400 m near Sládkovičovo which corresponds to the maximum subsidence rate 2.2 cm/100 years. Recent subsidence rate is 2 mm per year (P. MARČÁK 1978). The values are lower than in the Pannonian Basin and higher than in the Vienna- and East-Slovakian Basin. Dynamism of the Pannonian Basin also affected the Danube Basin definitely forming during the Pliocene (B. GAŽA 1984). The W-E axes of partial depressions were in accordance with the course of structural boundary between the two megablocks with courses of deeper structure indicated by geophysics (Fig. 64) and fault tectonics around Komárno (the influence of the Hurbanovo deep fault). Tectonic structure of the basin is inconspicuous. According to T. BUDAY'S classification the southern part of the basin represents the type of superimposed basins, its northern part represents the type of longitudinal basins with inherited structure (three NE-SW "embayments" due to the Inovec and Tribeč Mts. elevations). In the Danube Basin most faults terminate in Pliocene sediments or their throw decreases. This may be due to a great distance of the area from the diapir collapse and thus also from the centre of fault-controlled rapid subsidence. West-Carpathian crystalline units forming the higher basement were involved in subsidence to the depth of 6 km near the contact with the Pannonian megablock. Drilling results (A. BIELA 1978) and geophysical maps indicate that granitoids in the Danube block do not form any continuous granite layer. In the basin basement metapelites with basic rocks dominate and indicate the original simatic crust type tending to the formation of depressions (F. ČECH 1980b).

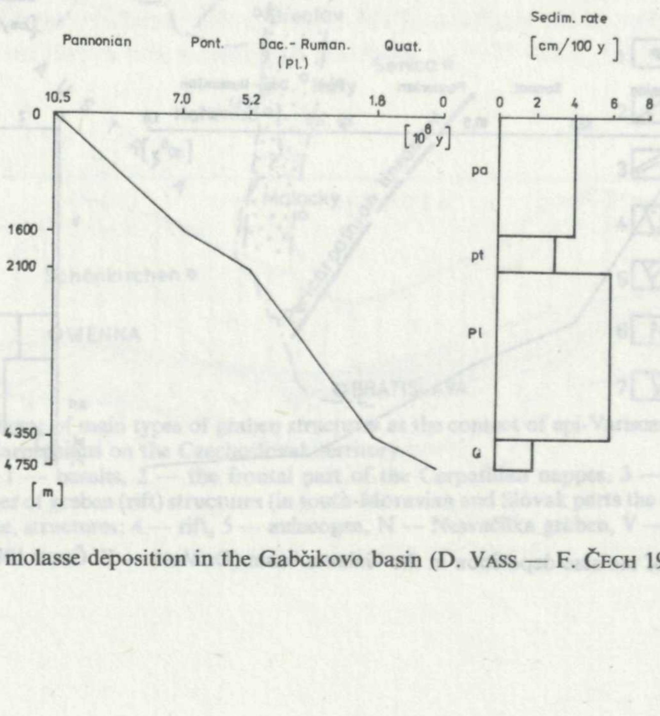


Fig. 63 Rate of molasse deposition in the Gabčikovo basin (D. VASS — F. ČECH 1983).

I have tried to precise the geologic-petrographic character of the submerged crystalline complexes according to the data from deep holes, and to correlate the data with gravity anomalies alternating there in NW-SE course. Their axes correspond to the basement structural courses.

Crystalline complexes reached by holes NW of the Vepor deep fault (the Danube block) consist in their northern part of mica schists and phyllites in the form of the mantle of the Tribeč granite massif. In the NW part the crystalline complexes consist of the Malé Karpaty Mts. granitoids. Like Tribeč, the Malé Karpaty Mts. are in the zone of positive gravity anomalies and are interpreted as massifs without deeper roots (B. BERÁNEK 1979). In my opinion, the negative gravity anomaly joining the Central-Slovakian neovolcanic region along the Rába line may indicate hidden intrusive masses causing stabilization or slight elevation of a block between the towns Komárno and Štúrovo, where coal-bearing sediments deposited.

With respect to crustal types it is important to know that granitoids do not reach greater depths and they are evidently separate bodies, not connected by any continuous thick granite layer. Here I cannot agree with O. FUSÁN et al. (1971) who assume that granitoids revealed by drilling near the villages Dedina Mládeže and Pozba form a continuous body. I prefer the idea of perisurficial pre-Miocene more isolated small bodies connected in depth by hypothetical Miocene acid magmatites (perhaps linear intrusions) and forming thus a granitized body with lesser gravity effects (Fig. 65). According to Beránek's map the less dense body extends to the depth of about 15 km and not more. The body may consist of anatexites which have not reached the surfase because of the Vepor fault compression.

Crystalline complexes in areas devoid of granitoids mostly consist of mica schists and two-mica paragneisses. Phyllites — partly brecciated, were only found in the area of Sereď—Sládkovičovo. I regard the phyllites as a relic of original sedimentary filling of a tectonic depression of the NNE-SSW and N-S strike (Fig. 65) in the crystalline complexes. I think there is another ancient depression zone between the towns Levice and Šurany, as indicated by mica schists and plentiful amphibolites (hole Šurany-1). This depression was separated by a zone of diorites and small granodiorite massifs from the Komárno—Šahy depression with the maximum metabasite thickness. It is right then to range the crystalline complexes to the Kohút zone. In contrast to elevation units of crystalline complexes with granitoids, tending to uprise, crystalline complexes in South Slovakia are submerged, they have a lack of granitoids and dominant metabasites (formerly a sedimentary-volcanogenic complex ranging to more that 600 m in thickness, reached by the hole MV-12 SW of Bušince). The prevalent metapelites are indicative of the original mobile crust with a thin, discontinuous granite layer. In contrast to M. MAHEĽ (1978b) I do not extrapolate the intensely granitized belt SW of the Inovec and Tribeč Mts.

Nappes accumulated even deeper than 2000 m might also have accumulated in depressions of originally simatic nature with denser crust (metapelite densities 2.71—$2.73 \times 10^3$ kg . $m^{-3}$; A. Biela 1978) tending to subsidence. It is also indirectly indicated by mica schists reached by the hole As-45 W of Zlaté Moravce. This dynamic concept of the effects of denser crust on nappes accumulation is also based on H. RAMBERG'S (1967) model.

I regard the northern boundary of the positive gravitational field as an approximate boundary of the crust. Its mobility increased most likely eastwards to the present Spiš—Gemer region. I. VARGA (1978), M. MAHEĽ (1978a, b) and R. MOCK (1978) pointed out to immature continental crust in this unit owing to weak Hercynian tectogenesis (an unfinished cycle) and its trend to mobility. Slight or no granitization was also proved by holes realized in this region: no migmatites and orthogneisses were revealed. The division into depressions and elevation zones running from the Inovec Mts. and the Tribeč Mts. copied the old segmentation of crystalline complexes. Elevation zones correlate with the presence of less dense granitoid rocks. The Danube basin offers an example of the crust preserving its different mobility even during its Neogene rebuilding owing to its differentiation into more or less granitized crust.

### 3.4.3 South-Slovakian Basin

The basin is situated on the narrowed part of the area of the positive gravity anomaly and of less intense segmentation by deeper faults. In its position and partly in subsidence evolution the basin resembles marginal depressions of the Pannonian Basin. Subsidence in the South-Slovakian Basin (Ipeľ depression) commenced in the Late Cretaceous and terminated in the Early Badenian, i. e. in the time of the maximum subsidence rate in depressions (D. VASS — F. ČECH 1983 — Fig. 66). Inversion was followed by stabilization and uplift of the basement, perhaps due to presumable less dense masses in the basin basement (D. VASS 1979). On the basis of petrographic character of crystalline rocks from deep well cores the northern suboceanic crust offspur (Fig. 65) may be interpreted. The cause of stabilization has so far not been explained exactly. The basin evolution shows that without deep effects no intense mobility occurs is suboceanic crust either. T. Buday (1961) ranged the South-Slovakian Basin among inherited structures and thus the absence of deep restructuring is emphasized. The basin has a time-limited evolution and lesser mobility. These factors were, however, favourable for the deposition of coal-bearing sediments. According to D. VASS (1979) the basin is in the position of a basin in a backdeep.

I shall try to explain a greater stability as follows: The uprising mantle diapir got in contact with tectonically variably disturbed crust, mainly by deep faults. The faults supported uprise of mantle varieties and melting of crust rocks. In the area of decreasing partial pressure on the mantle margin the cooling, followed by collapse, proceeded earlier than in the diapir centre. No collapse proceeded in the area without deep faults, and thermodynamic regime in mantle mass preserved on the depth level reached. Collapse was replaced by stagnation and/or elevations. These caused tensile stress in upper crust and dissection of basin filling by a system of smaller faults perpendicular to the extension of domed crust i. e. to NE-SW strike. This is why NW-SE faults prevail there.

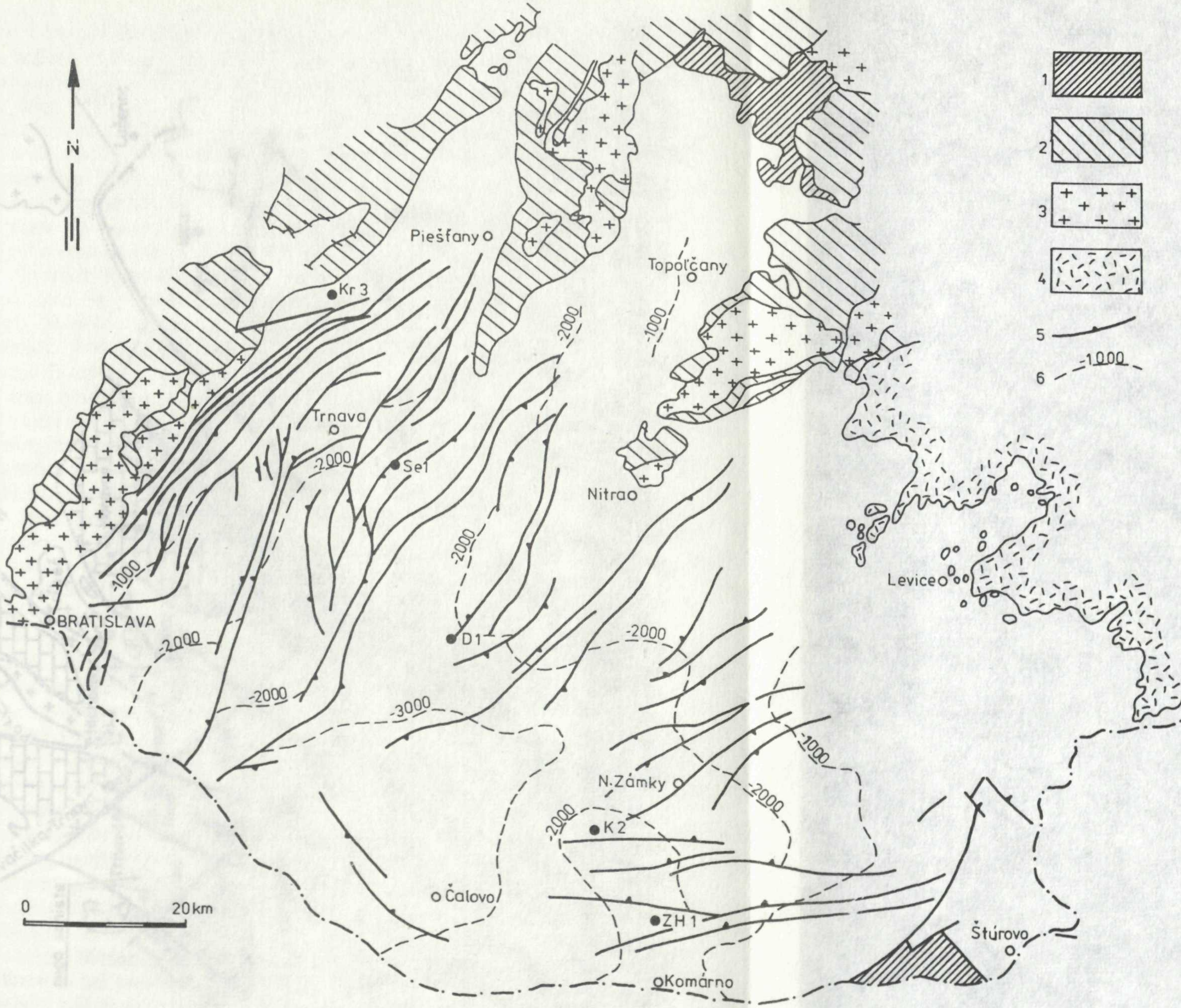


Fig. 64 The Danube Basin. (According to B. GAŽO 1984 — adapted and simplified).
Explanations: 1 — Paleogene, 2 — Mesozoic, 3 — crystalline complex, 4 — volcanics, 5 — faults, 6 — isohypses of the border Badenian — Sarmatian.

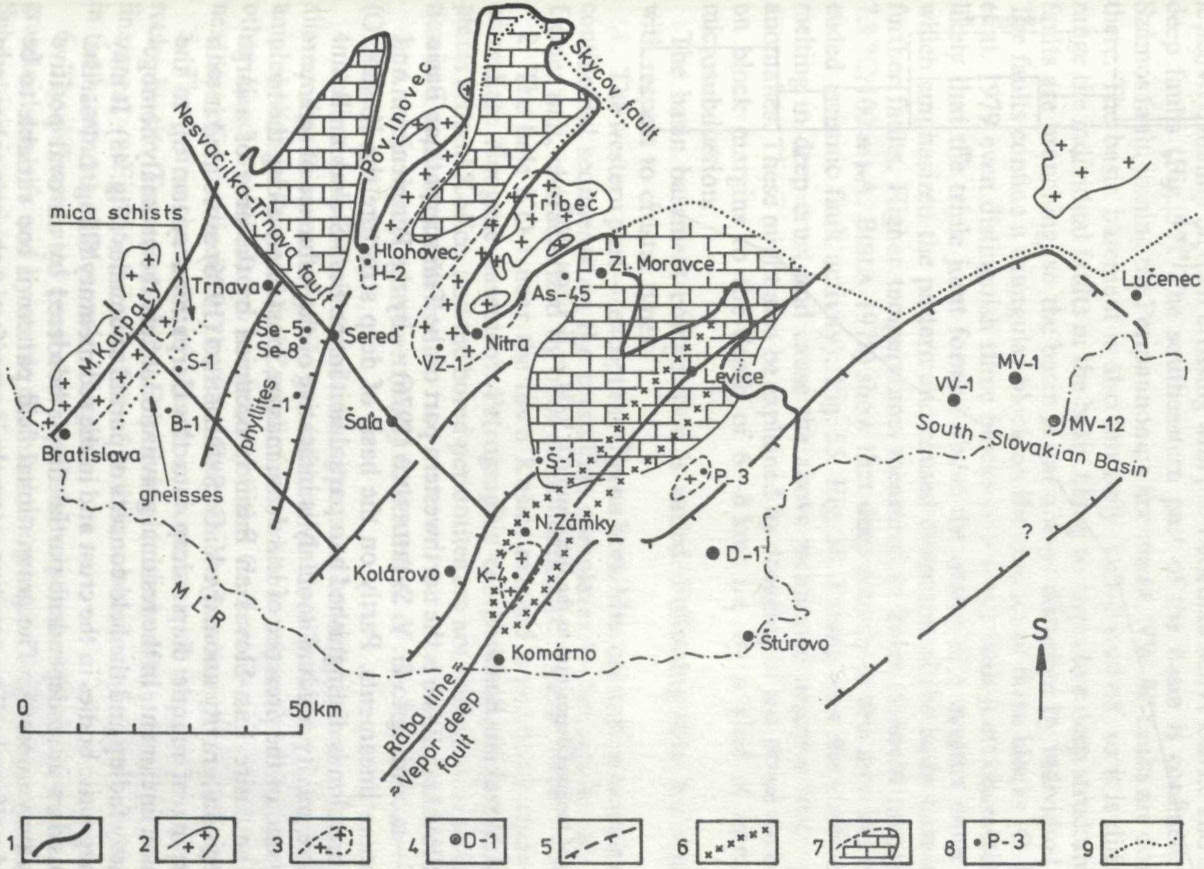


Fig. 65 Reconstruction of crust types in crystalline basement of the Danube Basin (a model).
Explanations: 1 — periphery of Neogene basin, continental crust, 2 — granitoid rocks on the surface, 3 — granitoid rocks in depth (assumed extent), suboceanic crust, 4 — drilled amphibolites, 5 — depression zones, 6 — hypothetic Neogene intrusions, 7 — pre-Neogene depressions filled with nappes, 8 — main drill holes, 9 — presumable boundary of continental and suboceanic crusts prior to Variscan metamorphism.

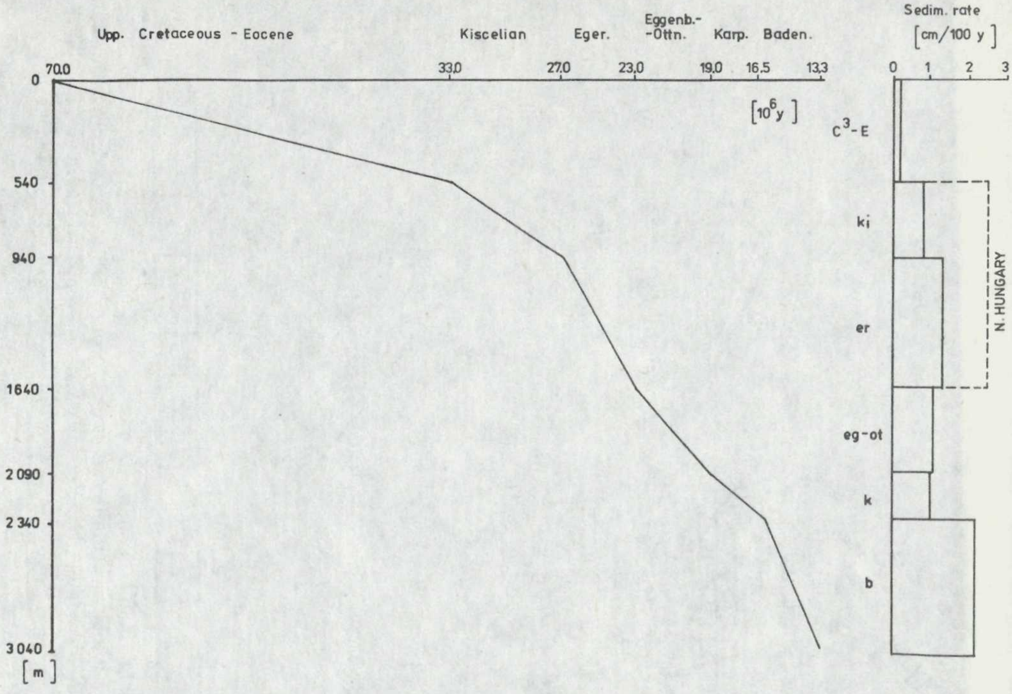


Fig. 66 Rate of molasse deposition in the Ipeľ basin (D. VASS — F. ČECH 1983).

### 3.4.4. East-Slovakian Basin

The East-Slovakian Basin is the northwestern part of the Transcarpathian Basin bordered — according to G. V. SVIRIDENKO (1976) — by the Peripieninian and Peripannonian lineaments. Partly on the basis of deep structure the Košická kotlina depression is distinguished as a partial unit in the East-Slovakian Basin. The negative gravity residual anomaly is indicative of the influence of sedimentary filling and of the presence of less dense masses, most likely along the basin margin. The entire East-Slovakian Basin is occupied by the area of a large positive regional gravity anomaly. V. G. SVIRIDENKO (1976) associated it with the upwarping of mantle diapir deep into the Carpathian system up to the Peripieninian lineament. In the residual gravitational field the anomaly homogeneity is partly fading and the field consists of partial anomalies (Fig. 48). It may indicate magmatic bodies in the crust and in the sedimentary filling. L. POSPÍŠIL (1980) considers an independent partial diapir bordered by an oval positive residual gravity anomaly. The gravitational field pattern is too variable to be interpreted uniformly. The negative anomaly is manifested as regional only in the Košická kotlina depression to the depth of 7 km. In geophysical maps of residual anomalies in greater depths this anomaly fades away (in maps by O. FUSÁN et al. 1971 — Fig. 43).

The western part of the basin is on the contact of the Hornád, Slánske vrchy (hills) deep faults and the Peripieninian deep fault with intense Neogene volcanic activity, mainly on their crossing with pre-Badenian active Darnó and Szamos deep faults (Fig. 55). The southeastern part of the basin is confined by the Szamos fault, joining the Peripannonian lineament. NW-SE faults are dominant there. The basin basement is also intensely fault-disturbed, so it is difficult to range the individual faults in the basin filling to respective deep structures. The faults are branching, so the basin is not always disturbed by individual faults. The faults confine a triangular block on the contact of three blocks (O. FUSÁN et al. 1979 even distinguish three blocks of a triple fault joint character). It is likely that the triple joint formed above the centre of a magma dome uprise which emphasized the pattern and caused dissection of the basin basement by further faults. Higher temperatures measured in holes (thermal gradient 5—7.8 °C/100 m; A. BIELA 1978) show that deep activity in this area has still not ended (seismic fault activity); Fig. 55. Fig. 34. I think that the rocks are still melting in deep crust and cause the above mentioned negative linear gravity anomalies. These might also be explained by dragging of less dense rock blocks on block margins to the depth of 6—8 km; i.e. by a kind of intracrustal microsubduction.

The basin basement composition is varied. Following data are significant with respect to crustal types:

1. The western part — except the Čierna hora Mts. crystalline complexes and continental sediments on the crystalline complexes — belongs to the Spiš—Gemer unit, characterized by poor consolidation (I. VARGA 1978) — Fig. 56.

2. The hole KO-1 near the town Košice reached a peridotite body in the immediate Neogene basement. Petrographic composition of crystalline complexes around Košice, and the bored peridotites are not indicative of sedimentation in a typical ensialic basin.

3. A basic body, gravimetrically indicated S of the line Košice—Hýľov (O. FUSÁN et al. 1979).

4. An intensely differentiated crust type is indicated by gravimetric maps with alternant positive and negative gravitational fields (Fig. 43), by mica schists and amphibolites of the so-called Zemplín islet, without migmatitization, and on the other hand by lithofacies evolution of the Carboniferous indicative of the nearness of a continent in the East Carpathians basement (I. VARGA 1978).

5. The pelitic character of Mesozoic sequence with frequent basic volcanic rocks in the Transcarpathian Depression (mainly along the Peripannonian lineament — Fig. 56) is indicative of mobile suboceanic (former oceanic) crust in the basement of Mesozoic troughs.

6. The differentiated crust character is also indicated by the absence of large granitoid massifs and by stable chemical composition of diabases and spilites. Subduction is excluded because of permanent volcanic fault activity since the Cretaceous till the Paleogene, in places till the Neogene (V. G. SVIRIDENKO 1976).

The most intense subsidence in the East-Slovakian Basin proceeded in the

Badenian (10.9 cm/100 years) and in the Sarmatian (8.6 cm/100 years). The minimum subsidence proceeded in the Pliocene (D. VASS — F. ČECH 1983, — Fig. 67).

Neogene sedimentation proceeded simultaneously with the folding of the Flysch in the basin in a foredeep position (J. ČVERČKO et al. 1984). Sediments have been repeatedly deformed since and after the Paleogene. According to V. G. SVIRIDENKO (1976) the maximum distance of overthrusts was 8—10 km. The basin has a block — imbricated structure, with faults prevailing. The basin is crossed by the broad transversal Pannonian — Volhynian Depression (Fig. 56). In the Slovak part of the basin the share of nappes in the basement structure is being permanently precised. Block structure is, however, dominant. It is difficult to explain the complicated tectonic evolution. It is likely, that the position of the basin on the contact of close geologic units with compression dynamics including diapir effects controlled compression deformations in pre-Neogene sediments and in crystalline complexes. The folded structural level is common with the Vienna Basin. The two basins have different folding and overthrust dynamics.

## 3.5. Correlation of main Neogene basins in West-Carpathians hinterland

I have selected three largest Neogene basins for correlation: the Vienna Basin, the Danube Basin and the East Slovakian Basin. Two of them are interesting with respect to economic geology. Prospectness of the Danube Basin concerning hydrocarbon deposits will be studied by deep exploration (B. GAŽA 1984).

The basins are situated along significant deep faults separating blocks and/or megablocks. The basins show the character of marginal depressions and indications of a position on divergent margins. The horst-graben structure including

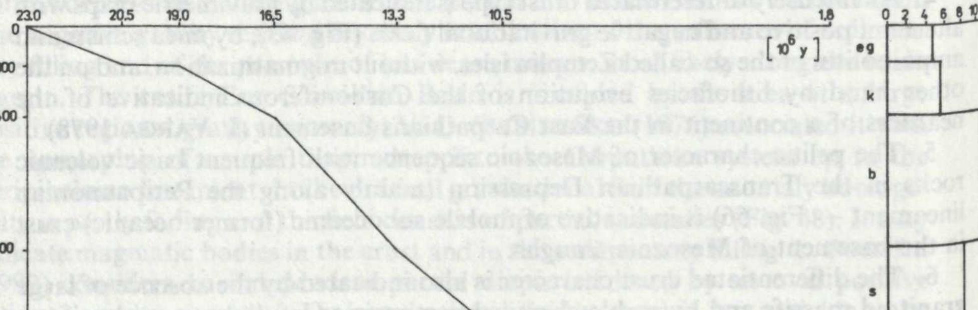


Fig. 67 Rate of molasse deposition in the East-Slovakian Basin (D. VASS — F. ČECH 1983).

integrating elements of fold structure (graben-synclines and horst-anticlines) is indicative of combined dilatation- and compression tectonics effects. The dynamics is associated with pulsation movements on megablock boundaries. The interblock position of the basins also controlled a complicated fault structure and intense fault segmentation of layers (Tab. 10). The basins differ from smaller intramontane basins in more fault systems, in their evolution and in endogenous causes of their evolution. The main deep phenomenon in intermontane basins is mantle diapirism typical of intermontane basins of the Mediterranean and circum-Mediterranean regions. In my concept strike-slips are not regarded as decising for the formation of basins as presumed by F. HORVÁTH — L. ROYDEN (1981) or L. POSPÍŠIL — D. VASS (1983).

Survey of main fault systems in Neogene basins

Table 10

| Basin | Fault strike | Deep-fault strike |
|---|---|---|
| Vienna | NE, NNE, NW, W-E | NE (NW) |
| Danube | NE, NNE, NW, W-E | NE |
| South-Slovakian | NW, N-S, NE, NNE | NE, N-S |
| East-Slovakian | N-S, NE, NW | N-S, NW (NE) |
| Ilava and Trenčín | NE (NW) | — |
| Turiec | NNE, NW | |
| Horná Nitra | NE, NW | — |
| Žiar | NE, N-S | — |
| Banská Bystrica and Zvolen | NE, (N-S), W-E | |
| Brezno | NE, NW | — |
| Handlová | NE, NW | |

Remark: It is evident that large basins are joined with deep structures and tectonically disturbed along more than two fault strikes. Small basins formed along one main longitudinal strike crossing with one transversal strike.

The basins are on the periphery of the Pannonian diapir. The maximum subsidence commenced in the Karpathian and stagnation in the Pliocene. The Vienna Basin and the East-Slovakian Basin show prominent inversion and subsidence polarity (D. VASS 1979, D. VASS — F. ČECH 1983, L. POSPÍŠIL — D. VASS 1983, — Fig. 68). The Danube Basin is a transitional area with the Badenian—Sarmatian subsidence intensity by about 30—50 percent lower (subsidence followed) than in other two basins. In the Pannonian—Pliocene time the subsidence intensity was higher. Its centre migrated from N to S and the evolution of the basin was similar like that of the Pannonian Basin. It was definitely formed as late as the Pliocene.

All basins belong to the peripheral circumdiapir zone (D. VASS 1979). The subsidence might be controlled by mantle diapir (D. VASS 1976, 1979, L. POSPÍŠIL — D. VASS 1983).

I associate the dynamics of basin evolution with the formation of a rim syncline of mantle diapir. Then the subsidence on the crust surface (cf. Chapter 1.10.2) was directly connected with outflow of low-viscosity masses in asthenosphere followed by crust downwarping into a vacated upper mantle „space". In the time of lateral expansion of the diapir top and the consequent collapse, the subsidence in peripheral basins stagnated. In the Danube Basin integration with the Pannonian Basin proceeded.

In the area of formation of thermoactive satellite diapirs (the Danube Basin, the East-Slovakian Basin — L. POSPÍŠIL 1980) sedimentary areas were differentiated during subvolcanic and volcanic processes. R. RUDINEC et al. (1981) denies the lower lithosphere cooling as a cause of rapid subsidence. They associate unstable lithosphere beneath the Transcarpathian Basin with rocks thrusts in the form of nappes into adjacent mobile zones with thermal plume in the Pannonian centre. So the crust thinning might have resulted from transport of rock complexes in combination with lithosphere stretching or subcrustal erosion. The authors do, however, not explain the spreading mechanism and regard it as an open problem.

Intense neotectonic movements and recent subsidence proceeded in the basin. N-S faults in the Vranov—Trebišov—Brehov zone were dynamically most active. The Quaternary sequence ranges to 50—70 m in thickness (V. BAŇACKÝ 1980). It represents the subsidence rate 0.3—0.4 cm per 100 years. Results of measurements of recent vertical movements in the basin show dominant sub-

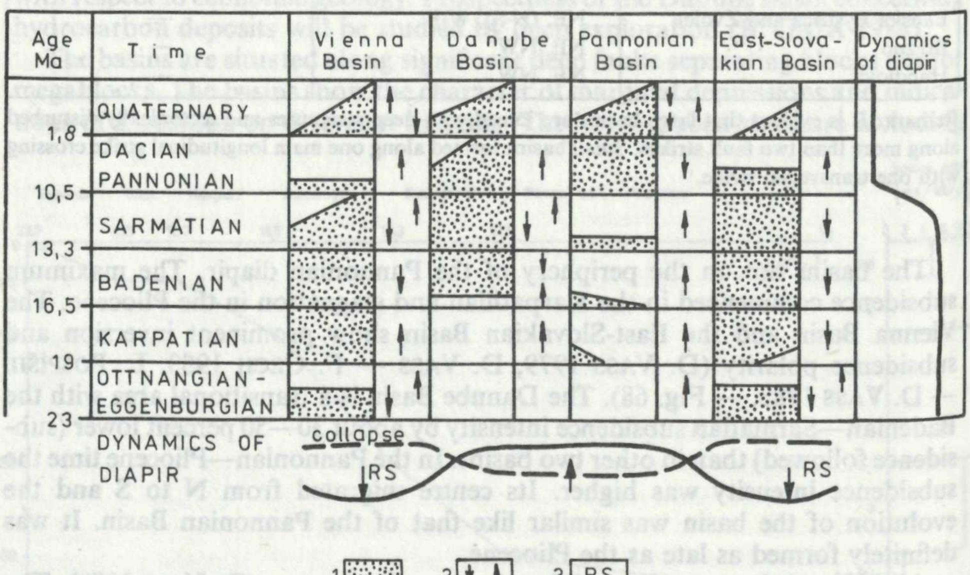


Fig. 68 Time of sedimentation is mainly related to depressions.
Explanations: 1 — sedimentation, 2 — subsidence-uplift (stabilization), 3 — rim syncline.

sidence rate 1.5—2.0 mm per year (P. MARČÁK 1978). The areas of maximum subsidence near the town Vranov nad Topľou and E of the town Michalovce represent young depressions active since the Pleistocene.

The three largest Neogene basins indicate a close relation among crust thickness and type, tectonic segmentation and dynamic effects of the close mantle diapir on their mobility. As for the crust thickness, the Vienna Basin is an exception. If we consider that beneath 3—5 km thick Miocene are 3—4 km thick tectonically dismembered sediments, then the original crust thickness must have been 21—28 km. All these factors enabled the thin suboceanic crust and intense deep fault disturbation to support upwarping of the mantle into the crust and destroy it. The collapse stage resulted then in a rapid, extensive and irregular subsidence controlled by crustal tectonic segmentation and chemical composition of the basins basement. The crust type evidently controls the areal extent of the basins.

### 3.6. Relationship between deep structure and paleogeographic conditions in Tertiary time

#### 3.6.1. Relationship to paleogeographic conditions in Paleogene

The data on the distribution of Paleogene facies are insufficient, so P. GROSS'S (1978) paleographic maps cannot inform exactly about the course of deep structures significant for the evolution of Neogene basins. They, however, indicate the influence of the Peripieninian lineament and of the deep Vepor fault upon differentiation of lithofacies- and paleogeographic units. In the eastern part (the Humenné branch) the lineament separated the hypothetical continent NE of the sea strait running from the Tatra Mts.

Paleogeographic data indicate extensive downwarping of continental- and subcontinental crust segments (mainly in the Priabonian and Lower Oligocene), and the connection of the Pannonian Basin with the South-Slovakian region — the Buda Paleogene (T. BUDAY — J. SENEŠ 1967). Extensive downwarping of pre-Paleogene units proceeded simultaneously with tectonic segmentation of the Pannonian Basin crust and with intense volcanism extending as far as North-Hungarian centres.

I think that the sinking of the mantle diapir periphery was associated with its uprise and formation of a subcrustal rim syncline.

#### 3.6.2. Relation to paleogeographic conditions in Neogene

Deep segmentation of Inner-Carpathian units managed already the disintegration of Paratethys into autonomously evolving basins. I have studied the relationship between deep structures mobility and paleogeographic evolution on the basis of paleogeographic maps compiled by J. GAŠPARIK, (1978), offering more information than similar maps of the Paleogene.

During the Eggenburgian — except the South-Slovakian Basin — the internal part emerged, and sedimentation — like in the Paleogene — concentrated around the Peripieninian lineament. Depression zones were parallel to the zone of the Přerov—Štiavnica (Skýcov) and Slánske vrchy (hills) deep faults. It is likely that the sea joined the Pannonian region via straits along these faults.

The Late Badenian deep structural rebuilding resulted in interruption of the connection of the relics of Paratethys with the Tethys region. The area of Eggenburgian sedimentation uprose inversely and became a source area — except the Humenné branch of the Peripieninian lineament, in the northern part of the Košická kotlina depression with the preserved sea.

During the Lower Badenian the inversion on main deep faults stopped: the area subsiding in the Eggenburgian uprose NE of the Skýcov fault zone and the NW area started to sink. Subsidence also started SE of the Vepor deep fault and E of the Slánske vrchy hills fault in the East Slovakian Basin. Mobility in the basin increased eastwards (T. BUDAY — J. SENEŠ 1967). The mobility passed from the Internal Flysch area southwards and culminated in the Pliocene. So the subsidence advanced towards the presumable mantle diapir, i. e. to the outer side — in the sense of above mentioned expansion indications — which means that it advanced against the expansion. This might indicate gradual cooling of diapir, starting from the periphery, and advancing collapse effects in the same direction, i. e. towards the elevation centre.

New paleogeographic data (J. GAŠPARIK 1978) confirm T. BUDAY'S (1961) opinion about the influence of deep faults upon basin evolution but contradict his conclusion concerning the connection between folding and stagnation or mobility in basins. Migration of mobility from W to E is confirmed, but basin expansion from internal parts of the Carpathians to their external parts is still not confirmed. Mobility migrated from external zones to internal zones of the Carpathians.

## 3.7. Basin dynamics

For the study of basins dynamics it is necessary to know their history. But the early stage of the basin evolution is hidden in great depths, so far inaccessible by drilling. Therefore dynamic problems are solved by models and speculations. The same concerns classification of basins and determination of basin types.

### 3.7.1. Application of plate tectonic model basin classification upon West-Carpathian Neogene basins

Classifications of basin types after D. M. CURTIS (1980), A. PERRODON (1980), H. D. KLEMM (1980), D. R. KINGSTON, C. P. DASHROON and P. A. WILLIAMS (1983b) were discussed in Chapter 1.9.3.2.2.

According to H. D. KLEMM'S (1980) classification, the West-Carpathian

Neogene basins may be ranged to type 2 or 7. No one of our basins represents one type. The Danube and the South-Slovakian Basins are close to the type 2 mainly to the type 2 A. This, however, is not indicative of prospectiveness in hydrocarbon deposits.

The East-Slovakian Basin is closest to the type 7, even with respect to the high heat flow. In the Tertiary time it also had some features of the type 6 B.

The Vienna Basin is most heterogeneous. F. HORVÁTH and L. ROYDEN (1981) ranged the basin to the „pull-apart" type because of its box form resemblance to this type of basins. But with respect to paleotectonic position (the basin was not on divergent margin of Atlantic type) and tectonomechanic reasons (e. g. unproved strike-slip on marginal faults — F. ČECH 1984a) the basin cannot be ranged among the „pull-apart" basins. The basin is closest to type 7 in its position in the foredeep zone. But in its lower level it is closer to type 2 A. In the respective crust segment in its pre-Neogene position even the type 3 can be considered (F. ČECH 1984b). The features close to divergent and convergent character of crustal blocks are indicative of a complicated evolution of the Vienna Basin and its complicated structure, mainly of its deeper parts.

F. HORVÁTH and L. ROYDEN (1981) also denoted the northern part of the Danube Basin (Fig. 40) as a pull-apart basin on strike-slip faults. Marginal faults of individual partial depressions show no evidence of strike slips; they only show indications of normal faults.

According to D. R. KINGSTON'S et al. (1983) classification the Vienna Basin should have an intraplate position with subsidence on normal faults. The basin might be ranged emong the IF type basins (basin in intraplate position) with graben and horst structure. This type of basin might have developed from the original MS type of basins on the divergent continent margin, near the contact between continental and oceanic crust.

In accordance with classification criteria the Vienna Basin had a complicated evolution if the hypothetic lower level formed on a continental margin as indicated by the horst and graben structure formed in the stage of a divergent cycle. The later evolution with the formation of nappes in the basement complex shows features of a convergent cycle [LL basin type in terms of D. R. KINGSTON'S et al. (1983) classification], i. e. basins along the convergent margin. In the plate tectonic model it might be explained by the fading out of the oceanic crust owing to subduction. According to a detailed classification the basin might show resemblance to the classification cycle LL 2 — LL 3, i. e. with commencing and continuous folding in an adjacent basin margin. The basin history may represent the IF or LL 1 type (formation of depressions in the tension field in association with strike-slip faults), later on the MS type with some features of LL 2 and LL 3 types (formation of nappes) and again of IF type. The author's remark: The LL 2 and LL 3 types are associated with active folding and formation of nappes in the basin region and not with overthrusts from an adjacent orogenic area. In all classifications the Vienna Basin and its basement show features of various basin types. The Vienna Basin is no pure type of a basin because of a complicated evolution of this megastructure.

The Danube- and the South-Slovakian Basins represent the IF type with some features of the MS type. The East-Slovakian Basin in the basement belonging to the lower level may be ranged to the LL 1 — LL 3 types, in the Neogene filling to the IF type with some features of the MS type.

Heterogeneous features of Slovak basins either represent their complicated evolution unperceivable by detailed classification either, or they may indicate schematic character of these classifications based on simplified plate tectonic models.

3.7.2. Classification according to tectonic criteria

The first complete classification of West-Carpathian Neogene basins according to their relation to the basement of the structure was presented by T. BUDAY (1961). Classification according to the history of subsidence and spatial relation to the Pannonian mantle diapir was presented by D. VASS (1979). My classification was based on crustal types (F. ČECH 1982).

Results of correlation researches in relation to the existing data on the origin and evolution of basins presented by T. BUDAY (1961) may be summarized as follows:

1. T. BUDAY was right in associating the basins in the Alpine-folded West-Carpathian arc with the deep structural pre-Neogene segmentation — hereditary basins conformable to deeper fault structure. Buday ranged there small Neogene intramontane basins in the West Carpathians, the Vienna Basin and the East-Slovakian Basin.

2. According to T. BUDAY the basins on the internal side of the West Carpathians are the above structurally different units so they cannot have uniform evolution and structural pattern. Buday ranged the Danube Basin and the South-Slovakian Basin among intermontane superimposed basins. He associated the origin of basins with migration and pulsation of folding. This definition should be complemented by new data (F. ČECH 1982).

1. Hereditary basins represent two types:

a) Basins on continental crust. They have a small areal extent and their mobility is spatially related with deeper faults of one or two structural strikes and with structural boundaries of pre-Neogene elevations and depressions — anticlinoria and synclinoria. Basins of this type are indirectly associated with intracrustal deep faults, they are close to the supracrustal faults. Their mobility is limited, and according to geophysical data the maximum depth of the basement surface is 2500 m (Žiarska kotlina depression). The basins are evidently associated with orogenic movements.

b) Basins on suboceanic crust or on thin continental crust are intensely segmented or confined by deep faults of several strikes. Some basins have a character of deep grabens with sediment thickness exceeding 5 000 m. (The maximum subsidence amplitude is twice so as in the preceding basin type.) Not all areas of maximum mobility are identic with the deep fault in the basement (Fig. 55), but they are spatially related to transversal separate blocks with their basement composed of mobile suboceanic crust. This group includes the Vienna Basin and the East-Slovakian Basin. Both basins show mixed features of hereditary and superimposed depressions. Owing to their position on the boundary of two different tectonic units (megablocks) with different pre-Alpine evolution, the basins have the character of intermontane basins or interblock basins according to depth structure criteria. There was no direct influence of folding upon the basin evolution; the basins are related to the margin of folded units.

2. Superimposed basins do not have such heterogeneous genesis and structure as assumed by T. BUDAY (1961) from the view of deep structure and their appurtenance to basement blocks. The basins are bordered by deep-seated faults and occupy most part of the deep sunken blocks area. The extent of their subsidence is the same as with the basins of the preceding category. The origin of the basins was affected by deep processes in mantle which markedly affected the poorly consolidated suboceanic crust or thin continental crust in the preceding evolution stage.

The genesis of superimposed basins was not controlled by variable folding intensity out of the basin. The basins evolved in the zone of regional subsidence and thin crust, the folding proceeded in the zone of elevations and crust accretion. The folding only proves that regional subsurface tectogenic processes have different forms of manifestation produced by a subcrustal homogeneous process.

With respect to tectonic development significant is the relation of folding migration from W to E along the Carpathian arc and radial migration from the inside of the arc to its outer part. Migration of volcanism had the same trend (J. LEXA — V. KONEČNÝ 1979) with reversible trends inside the individual units.

The basins developed in the course of the W-E migration but their radial mobility had a reversible growth tendency (since the end of the Paleogene). The evolution of molasse basins is variable in time. Early molasse (Egerian) basins mostly developed in the hinterland, main molasse (Eggenburgian—Sarmatian) basins — inside the Carpathians and the late molasse (Pannonian—Pliocene) basins — in the backdeep again. There is a correlation in spatial distribution between volcanism and subsidence but only in a wider regional extent.

Elements of deep factors are also in the Neogene basins classification by D. VASS (1979) trying to apply the plate tectonic model. Vass considering the structural resemblance of the South-Slovakian and Pannonian Basins in the Pliocene stage of their evolution, presumes their common origin. He regards the Buda Paleogene Basin as a result of culminating mobility in the time of final activity of diapir following the Late Cretaceous folding associated with subduction. Vass presumes the crustal thinning since the Paleogene diapir activity.

Vass also regards intramontane longitudinal basins as a consequence of diapirism inside the Carpathian Arc. Subsidence and volcanism culminated in the Late Badenian—Sarmatian. Vass associates the origin of volcanism and subsidence with subduction in the Outer Carpathians, followed by the Early Miocene diapir uprise. He regards the extension of diapir and its segmentation into partial diapirs (one of them is in the East-Slovakian Basin) as the cause of culminating volcanism and subsidence.

There are, however, contradictions and problems in D. VASS'S (1979) concept of partial diapirs causing volcanic activity and basin subsidence on the back side of the Carpathian Foredeep. The presence of partial diapirs in the Central—Slovakian neovolcanic region may be proved by the following factors: the deep faults contact, the circular shape of the volcanic region, magnetic field, concentration of tectonic grabens and horsts in the subsidence area above diapir. The diapir would form on the tectonic contact of variably thick crust with different composition (continental and suboceanic crusts). But the mantle elevation is not proved by gravimetric data. So there is a question whether the volcanic dome was due to mantle diapir or to crust melting and formation of subvolcanic granitoid bodies presumable according to gravimetric data (O. FUSÁN et al. 1971). If the elevation is pre-Badenian — as shown by older subvolcanic rocks (K. KAROLUS 1978) — then grabens and horsts mostly formed in this structure of crustal origin. The deep cause of their origin may be in crustal geothermal influx of the „hot spot" type, but without mantle elevation. In fact, it is not clear why volcanism caused by subcrustal erosion of the lower crust basalt layer should commence with acid effusives. If we accept J. LEXA — V. KONEČNÝ'S (1974) petrochemical data, although incomplete with respect to geochemic indication of the depth of the origin of volcanic rocks, then the formation of rhyolite magma chambers may hypothetically be situated inside the crust.

Volcanic centres are out of the basin — except the East-Slovakian and Danube Basins. Volcanism is in accordance with basin subsidence. There is a time coincidence between them in other regions. The distribution of neovolcanic centres shows that they are mostly on the contact between thin and thick crust. The contact is usually tectonic (Fig. 29, 31, 55). Basing on above mentioned indications I regard this contact as a boundary between a typical continental crust and mixed (suboceanic) crust, or prior to diapirism — with thin continental, very mobile crust. The boundary is usually represented by the block boundary, sometimes inside a megablock. In the West Carpathians this boundary is approximately the contact between hereditary and superimposed basins.

The evolution of hereditary basins on a thin crust was complex. The basins formed in areas of repeated subsidence. The Neogene subsidence was dynamically joined with the evolution of the Pannonian mantle diapir. The complicated basin evolution is manifested in heterogeneity of their features and difficult classification.

The Vienna Basin in the time of the flysch nappes shift was partly in the position of a foredeep. In the Neogene it changed into an intermontane depression with some features of an intramontane depression in its SW and NE parts (in relation to flysch units). May be its connection with the foredeep preserved with some breaks. Because of the lack of deep holes we still do not know the nature of the basin region prior to overthrust of nappes. Generally mobility of the region inherited mobility of formerly thin simatic crust and activity of deep faults on the boundary of megablocks. It is likely that the original structure of the autochthonous sedimentary cover was graben or rift-like, as I have already mentioned. The riftogenesis had transitional features of intra- and extra-continental positions (H. D. KLEMME 1980).

The early stage might correspond to the divergent type of crust margin which is not a part of the Bohemian Massif. The existence of nappes may seem to correspond to changing regime on the convergent megablock margin. So far we do not know a more detailed mechanism of the formation of nappes. When gravitational sliding dominated, then the overthrust movements might have also proceeded in a basin on a divergent margin in the dilatation zone. The same dynamic regime might last (since the Jurassic?) up to the recent opening trend. The basin may be ranged among mobile types on plate margins — F. L. SCHWAB (1976), (subsidence rate 0.4—0.7 cm/100 years) according to Neogene subsidence rates 5.5—8 cm/100 years (Fig. 59) and Pleistocene subsidence rate (0.55 cm/100 years) in the Uherské Hradiště graben. The basin has the same character — i. e. at the passive divergent margin — according to D. M. CURTIS'S (1980) classification (Tab. 4, Chapter 1.9.3.2.1.), emphasizing the position on the contact of two crustal types. Transversal grabens on the slopes of the Bohemian Massif may represent a failed arm of the original triplet rift with other arms beneath the present Vienna Basin (Fig. 61).

Divergent character of block margin with the extension regime would deny Jiříček's model of subduction on the contact between the Bohemian Massif crust and the West Carpathians crust (R. JIŘÍČEK 1981).

The East-Slovakian Basin in the evolution of its basement comprises an early dilatation stage with even deep-sea Mesozoic and partly Paleogene flyschoid troughs. Their existence may be indicative of suboceanic and oceanic crust beneath the troughs. If continental complexes of the Čierna Hora Mts. and of the Zemplín unit are allochthonous, then the former suboceanic crust may be beneath the whole East-Slovakian and other parts of Transcarpathian basins. The absence of granitoids in blocks with continental sedimentation is indicative of nonconsolidated and perhaps also thin continental crust. This may be confirmed by tectonic transport of sediments on the crust in the phase of Mesozoic troughs compression.

In the basin basement — like in the Vienna Basin — the dilatation phase with subsidence and following compression phase may be distinguished. Mechanism of overthrusts and their extent are not quite clear. Prevalent block structure (V. G. SVIRIDENKO 1976) is indicative of possible local compressions in the phase of otherwise dilatation regime on the contact of three megablocks.

T. BUDAY (1961) ranged the basin among hereditary longitudinal intramontane basins. According to D. ĎURICA (1982) the basin shows some features of a foredeep (the basin was connected with it), of a hereditary intramontane and an intermontane superimposed depression.

The Pannonian mantle diapir affected the basin dynamics indirectly, although it is presumed to have extended in the thin crust segment, i. e. up to the eastern branch of the Peripieninian lineament.

Shear stress might have occurred on tangential structures — as I mentioned with the Peripieninian lineament. Thus oriented stress might have caused the formation of straight NW-SE and NE-SW shear faults. Decreasing stress can theoretically be followed by relaxation and block subsidence along the faults. This explanation is acceptable for the Vienna Basin and the East-Slovakian Basin situated at the Peripieninian lineament. It is however, problematic with respect to other inner basins including NE-SW partial depressions in the northern part of the Danube Basin. The explanation based on presumable local partial diapir affecting the genesis of variably oriented small basins is preferred.

I have discussed the complex problem of genesis and dynamics of basins, better explicable by the mantle diapir concept. Insoluble problems occur concerning subduction on the contact between the Outer and Inner West Carpathians. These problems cannot be solved by partial methods. A complex approach based for example upon numerous detailed paleogeographic maps is inevitable.

At present the origin of Neogene basins is best explained by deep causes. Data on deep geology and geophysical results prove that Neogene basins in the Inner West Carpathians are on deep structures and the common factors of their evolution are not due to surface phenomena. This is also proved by correlation researches in depositional basins, mainly concerning relationship of their origin and intense subsidence to deep structures and volcanism. The origin of basins — like volcanism and other geologic processes — is only a form of the results of endogenous processes. According to the present geophysical data the causes of the processes should be searched in asthenosphere. The mantle diapir model, formerly indicated by geophysical data, now also by geological, mainly by the deep structure analysis, is preferable in accordance with modern trends of deep geology. The deep-geological aspects revealed new elements common to the basins or basin couples (the Vienna Basin — the Danube Basin), and enabled separation of specific — autonomous elements (the East-Slovakian Basin).

Studying the genesis of synorogenic and postorogenic basins we meet facts of rather a fixistic nature. No reliable proofs for indication of mobilistic models are offered by structural and paleogeographic data or by the relations to deep structure. The relations to crustal types and to the degree of crust consolidation represent a new element in genetic problems. The new approach was enabled by the latest publications, mainly of M. Mahel', proving crust heterogeneity and correcting the traditional concepts of a regenerated platform or accomplished pre-Alpine tectogeneses. The presented data emphasize hereditary mobility not only by structural predisposition, but also by crustal type preserving its mobility for a long time and resisting consolidation of platform character.

## 4. Coal- and hydrocarbon deposits in basins of Carpathian-Balkan region

### 4.1 Coal deposits in Carpathian-Balkan region in relation to deep structure of basins

Publications on geology of coal deposits mainly in Inner-Carpathian Neogene basins treated their genesis from the view of supracrustal tectonics and their relation to adjacent tectonic units. I tried to analyze their tectonic position and history in relation to deep structure and basins evolution. I also treated deposits in the Pannonian Basin and in its peripheral part. Deposits concentrations in the hinterland of Inner-Carpathian units are controlled by tectonic faults mobility in the time of stable coal-bearing sediments deposition. Whereas oil- and gas deposits concentrate in the Outer Carpathians and in the Foredeep, coal deposits are in the internal units of the Carpathian arc.

The deposits — if at all denotable as deposits — are in basins inheriting their structural position on ancient faults. This position, and persistent or fading out compressional dynamics of adjacent tectonic units predisposed a short living of coal-forming conditions. So there are mostly economically less significant or insignificant deposits with small coal reserves, mostly of ash coal.

I shall deal with Slovak deposits in a separate chapter. Here I only want to present results of the study of this problem in Hungary, Romania, and adjacent parts of Yugoslavia — except the Dinaric region. For the analysis I used synoptic maps of coal deposits, compiled by A. K. MATVEEV (1975) and his monography (A. K. MATVEEV 1966), and unpublished base maps of individual countries. I studied the relation of deposits distribution to deep faults, to their mobility in the respective geologic period, to block structure and crustal types. Basing on the results of the study I came to the conclusions concerning the relation of coal accumulations to endogenous structures and to processes controlling the origin of depositional areas in the Pannonian and circum-Pannonian regions. Generally, favourable conditions for coal deposits were on the contact between elevation and depression zones — near the dry land water reservoirs boundary. For a comparison I also present data on deposits in the Carpathian Foredeep and platform foreland.

#### 4.1.1 Main coal-forming belts

On the internal side of the Carpathian arc the coal deposits form four belts — coal accumulation centres: in NW Hungary on the Balaton block, on the Apuşeni Mts. block, on the transition from the East-Carpathian arc, mainly on the Serbian-Macedonian block and on the fore-Dinaric block. The Vardar zone, the Serbian-Macedonian Massif and the Kraishtide zone have a particular position with maximum coal deposit accumulation and repeated coal-formation tendency.

The Carpathians proper are poor in coal deposits when compared to the

Dinarides with their anomalously high concentration of Neogene coal deposits. For a comparison it is noticed that there is a good areal correlation between coal-bearing zones and ophiolite zone and so NW of Skoplje — where the Budva—Zuccali ophiolite zone is shifted WSW-wards — the coal-bearing zone extending from the Kičev Basin in Yugoslavia to Greece, into the basins Servia and Katerini. Coal-bearing basins formed on basic rock zones between elevations, supplying the material into basins. Elevations also controlled water streams supplying the coal-bearing basins.

### 4.1.2 Coal-bearing lineament

There is an analogous agreement between coal accumulations and basite crust in the Kraishtide part and the southern part of the Vardar zone. Even in the Balkan zone on the contact with the Rhodopes are the largest coal basins in depressions on thinned crust (the Marica district, the Balkan Basin).

In the Vardar-Kraishtide zone coal-bearing structures occurred since the Carboniferous through the Cretaceous, Paleogene until the Neogene when the formation of coal seams was only restricted to the southern part. It is a unique structure, because no other persistent coal-forming zone is known in Europe. The structure formed because of movements on a lineament (lineamentary geosyncline in the sense of E. Bontchev 1976) in the course of which there was an optimal relation between slow subsidence and organic matter accumulation. No other lineament in Central Europe was productive for such a long time — 200—230 Ma. Only in the carboniferous and Paleogene the coal-bearing sediments deposited repeatedly in parts of the Labe lineament between Mělník and Magdeburg. If the Nesvačilka and Vranovice grabens are parallel to the SE continuation of the Labe lineament (Přerov—Štiavnica deep fault in the sense of O. Fusán et al. 1979), the fault system may positively affect the origin of Upper Carboniferous black-coal seams reached by drilling near Němčičky in Moravia. Other deep fault structures also supported the formation of structures favourable for the coal-bearing sediments deposits, but only in the respective coal-forming period; for example the Odra deep fault in the Neogene time, the Moravian-Silesian deep fault in the Permian-Carboniferous time in the Rosice--Oslavany basin. Therefore I denote the Vardar-Kraishtide zone as a coal-bearing lineament. It is a typical structure indicative of an association among coal-deposits formation, deep structure and mixed crust type. Distribution of coal deposits is limited by the northern extent of the lineament — in contrast to V. N. Gergelchev's et al. (1977) opinion about a rift megastructure crossing the Pannonian Basin.

Generally 47 economically significant coal deposits with 50—450 mil. tons of coal reserves in single deposits formed on the Vardar-Kraishtide coal-bearing lineament, namely:

| | |
|---|---|
| in the Carboniferous | 7 coal deposits |
| in the Permian | 2 coal deposits |

| | |
|---|---|
| in the Jurassic | 11 coal deposits |
| in the Cretaceous | 6 coal deposits |
| in the Paleogene | 7 coal deposits |
| in the Neogene | 14 coal deposits. |

Since the Carboniferous to the Jurassic the coal-forming conditions persisted on the same line from Lupac in Romania to Pirot in Yugoslavia. In the Cretaceous the coal-forming conditions concentrated in the central segment, from the western part of the Vardar zone. In the Paleogene and Neogene the activity passed to the southern segment, and to the western margin in the northern part. No time-space relations of coal-forming structures to folding or to regional dilatations were found, so the lineament was autonomous during the tectonic evolution of the Balkan-Dinaric region. Association of mobile zones with the Serbian-Macedonian Massif supplying the rock material into depressions and forming continental conditions of sedimentation (vegetation areas), was paleogeographically favourable.

To compare the character and size of coal deposits controlled by tectonic factors in mobile and stable units, I shall present brief characteristics of coal deposits in the Carpathian foreland, in the foredeep, and in the internal parts of the Carpathian arc.

## 4.2 Deposits in platform foreland of Carpathians

Coal deposits in this area formed on the stable platform basement in the Carboniferous and Neogene times. The origin of Carboniferous deposits was associated with orogenic movements of the Moravian-Silesian Variscides, and with movements proceeding in their foreland. The foreland also comprised the tectonically reduced Carpathian paleoarc forming the basement of overthrust Outer Carpathians. I discuss the Paleozoic coal deposits also because they might had been a source of released methane during neoid processes.

### 4.2.1 Late Carboniferous coal basins

The Upper Silesian Basin and the Lublin (Galician—Volhynian) Basin belong among largest basins. The Lublin Basin formed on the mobilized margin of the East-European Platform in the area, rich in basic rocks in its Precambrian basement. According to J. Zeman (1977) the crust of the area has a simatic character. So the origin of the basin might also be controlled by a different crust type. Faults conformable to the lineament margin bordering the SW platform margin (the Baltic-Podolie lineament in the sense of H. Stille 1951) were tectonically significant.

The Upper Silesian Basin formed in the most part on the Upper Silesian block with the character of a marginal massif and represented the mobile foreland of the Variscides. The most part (central part of the basin) of the block has a crust with a thin granite layer and large thick bodies of basic rocks (J. Zeman 1977) — as proved by geophysical data and by deep holes. Marginal parts of the basin are on the mobile crust separated by deep faults from the Upper Silesian block where J. Zeman (1978) presumes the suboceanic crust and the simatic median massif character of the block. According to J. Zeman's (1977) analyses of crust types beneath coal basins, also Central-Bohemian coal basins in the hinterland of the Variscides rest upon simatic, poorly Variscan-consolidated blocks. The blocks caused more extensive subsidence and a larger coal-bearing area. Basins on sialic blocks are areally restricted to the strike of fault causing the formation of elongated basins with limited coal reserves. The same concerns Saxonian coal basins situated in the zone of the Central-Saxonian lineament.

Generally, the origin of coal basins is associated with the formation and/or with considerable increase of the granite layer and increasing thickness of continental crust. Tuffaceous layers in coal seams and in interseam layers are indicative of temporary volcanic activity. Thick beds of arcoses and arcose sandstones are indicative of denudation of acid rocks — mature continental crust. Elevations supplying rock detritus for compensated subsidence, continent with growing vegetation, and positive effects of volcanic activity upon plant vegetation were factors favourable for coal-bearing sediments deposition. The influence of volcanic activity upon coal-forming flora vegetation is undervalued. This concerns the increased $CO_2$ content, supply of alkali material, mainly K and Ca for vegetation soils in which volcanic products have the function of mineral fertilizer.

Coal-bearing sediments of the SE part of the Upper Silesian Basin extend along the Odra lineament deep beneath the Flysch Carpathians — as proved by deep holes in Poland in the San river region (near the town Wadowice).

Carboniferous coal deposits are not many and indicate poor tectonic-morphologic dissection of the surface and limited favourable paleogeographic conditions. On mobile blocks or zones the coal-bearing sequence ranges to more than 5 000 m and large areal concentration of coal reserves. On mobile blocks or zones the coal-bearing sequence ranged to more than 5 000 m in thickness and the coal seams concentrated on a large area. The relation between mobility, number of seams, their thickness and coalification is evident in all basins. The seams formation and thickness were controlled by tectonic dissection of basin basement. The basement was intensely dissected by faults. Some were active during coal-bearing sediments deposition. The segmentation caused the formation of thin seams with a very variable areal extent. More stable units were favourable for the formation of big spatially stable coal seams. In the Upper-Silesian Basin such seams developed in the Upper-Silesian block, i. e. out of mobile marginal zones. The seams are less frequent in internal Central-Bohemian molasse basins with a more consolidated basement owing to Carboniferous granitization resulting in the formation of blocks with original suboceanic crust.

### 4.2.2 Neogene basins in Carpathian foreland

Largest coal basins are out of the Carpathian foreland, deeper in the epivariscan platform. Their position in relation to the composition of the basement crust indicates that large basins formed on mobile crust with less granitoids and frequent basic rocks bodies. I will not associate the existence of basins only with more basic crust and form thus a dogmatic rule, but it is necessary to notice the fact. For example, in the North-Bohemian Tertiary basin the subsidence rate is lower in the block with the basement, composed of the Teplice quartz porphyry, ranging to more than 1000 m in thickness (so far it has not been bored—except marginal parts with the thickness exceeding 950 m in the area of the future coal open pit Barbora) than in the block with the Krušné hory Mts. paragneisses. The North-Bohemian brown-coal basin might be compared to the zone of linear basaltization along the Litoměřice deep fault, and to the basification zone and the origin of the basalt (?) — mantle (?) diapir which caused the formation of a rift structure — the Ohře rift in the sense of L. KOPECKÝ (in J. ZEMAN 1978).

In the area of the adjacent West-Carpathian foreland are only small low-quality coal deposits, like the worked-out small (xylitic) lignite deposit Uhelná in Silesia near the town Javorník. Coal seams were also found in the Polish Tertiary, and peat layers in the Quaternary.

In contrast to the West-Carpathian foreland there are more Neogene deposits in the platform foreland of the East Carpathians. There are two large basins: the peri-Dniestr and the Pericarpathian Basin. They are at the tectonic border of the East-European Platform (Tornquist line) separating the stable eastern basement with the Paleogene Dnepr basin from the mobilized platform margin in the front of the Carpathian Foredeep.

## 4.3 Neogene coal deposits in Carpathian Foredeep

There are less coal deposits in the Carpathian Foredeep than in the foreland. Coal basins are less extensive and less significant in coal reserves. It is due to tectonic events and to permanent foredeep migration, to extensive material transport from destroyed nappe fronts and to nappe movements towards the foreland. Therefore coal seams in the Moravian Foredeep only formed on the slightly subsided basement — the Bohemian Massif, on faults diagonal to the foredeep. Thin seams of low-quality lignite were bored on the outer margin of the foredeep in the Opava area near Dolné Životice. The coal-bearing sediments deposition had a foredeep character, and the tectonically controlled subsidence has intermittent cyclic/rhythmic movements. Because of the small area the coal seams have irregular development. The cyclic deposition of the sequence was controlled by variably rapid subsidence of the basement in the front of the advancing Carpathian Foredeep.

Three thin lignite seams at Lažánky NW of Brno do not belong tectonically to the foredeep (V. HAVLENA 1964) but the tectonic regime was analogous there. A reduced tectonic regime was already in the Vranov area (lignite seams mined at Langau in Austria). The basement stability was to the rigid granite basement of the Dyje massif.

More coal localities are in the East-Carpathians Foredeep. The deposits are also exploited there (the peri-Carpathian lignite basin), although with high mining expenses. Coal deposits are on faults bordering the graben structures of the foredeep basement. In contrast to the West-Carpathian area the coal-bearing beds are folded. The coal-bearing sequence has the cyclic layering. The mobile basement controlled the intermittent subsidence and changing depositional conditions, so some seams (1—3) range up to 1.6—5.0 m in thickness (the Doiceşti deposit W of Ploeşti). In the Romanian part of the foredeep are five localities of small coal reserves. The seams thickness is about 0.75 m.

The deposits are geologically interesting because in their foreland is not the East-European Platform but the Moesian Platform. It was more mobile and replaced tectonically the foredeep. The origin of coal-bearing basins is associated with the tectonic boundary of the platform and mobile part of the foredeep, which started broadening. Subjacent faults also controlled the submeridional deposit course and the course of the deposit group of the Subcarpathian basin in the Getic graben. A linear, prevalently subequatorial basin is in analogous tectonic position on the contact with the Moesian platform. It is, though, ranged to the Carpathian Foredeep but its tectonic position is the same like the platform marginal depressions, and with respect to crustal units the coal-bearing beds are analogous to coal deposits on the margin of a foredeep between the Moesian platform and the Rhodopes. This indefinite position gives specific features to „foredeep" deposits. The features are common to the foredeep and to the intramontane depression. They are related to the foredeep basins by extensive tectonic disturbance from the side adjacent to the internal Carpathian units and by numerous coal seams. The deposits are related to intramontane basins by a greater depth of coal seams and by their areal extent.

The southern contact of the Moesian platform with the Forebalkan tectogenic folded zone and its transition zone is not associated with coal deposits. The subequatorial Pliocene Lom basin is on the contact with the South Carpathians in Bulgaria. The Lom basin has a double position. Tectonically it belongs to the Moesian platform but its tectonic-historical position corresponds to the South-Carpathian Foredeep. The foredeep is reduced there and replaced by the subsided foreland.

Coal deposits in the foredeep zone are infrequent, have a small areal extent and are — except the double-positioned basins — economically insignificant. The coal deposits only occur in the zones of reduced width of the foredeep or in places of its total reduction, near the foreland fault structures, predisposing the origin of small coal basins. Coalification is low in most cases — brown-coal hemitypes and lignites. Coal deposits in the West-Carpathian and East-Carpathian parts differ in tectonic character of the foreland and in tectonic development of the foredeep. They have common genetic features, namely repeated tectonic activity and foredeep migration preventing the formation of coal seams in balance parameters on a larger area.

## 4.4 Coal deposits in Outer Carpathians

Economically significant coal accumulations are scarce in the Outer Carpathians. There, however, are many local occurrences of thin coal seams owing to repeated favourable conditions for the coal formation. Persistent tectonic activity markedly limited the duration of the coal formation. Since these occurrences are only interesting from the geological view, I shall discuss them in the chapter concerning Slovak coal deposits.

Such occurrences are also in the Polish and Russian Outer Carpathians but they were paid less attention in these countries than in our country having a thorough evidence of known occurrences.

Exploitable deposits in the West-Carpathian area are associated with the Vienna Basin, in the East-Carpathian area with superimposed basins of the Romanian flysch East Carpathians. In this part of my book I shall discuss these deposits, and Czechoslovak deposits will be treated in the chapter concerning the West-Carpathian region.

The elliptical Comăneşti basin is largest. It is slightly folded into five synclines, separated by faults and low saddles. The basin comprises more than 30 coal seams representing brown-coal orthotypes. It is a basin of the mixed type, inheriting the N-S course of the flysch basement. The basin is partly superimposed up NNW-SSE flysch structures. Inside the 10 km broad basin the Neogene thickness 350 m in the western part of the basin increases to 600 m in its eastern part. Intense movement differentiation during deposition also continued in the course of deformations resulting in the flysch thrusting over the Neogene near the western part of the basin. Occurrences of coal seams are variable owing to the basement mobility: there are 34 seams in the central syncline and no seams along the eastern basin periphery. There are two deposits at the town Fălticeni; the Rica deposit was exploited. Most basins have low balance reserves. The pulsating tectonic regime also limited a larger areal development of coal seams and their thickness.

## 4.5 Summarized data on mutual relations

Basins in the Carpathian foreland occur on the consolidated epivariscan platform or on the Precambrian East-European Platform. Coal deposits are infrequent but large. They are associated with older basic rock crust and selective mobilization, or with basified, secondarily thinned crust.

Like in the Outer Carpathians, the foredeep coal basins are scarce and significant deposits only occur on the contact with a mobile foreland composed of simatic crust (Moravian occurrences), or on a crust thinned in relation to surrounding orogenic zones (Moesian Platform).

Most deposits are on thicker crust where the origin of basins and their differentiated subsidence were controlled by tectonically active faults with the character of marginal deep faults. No deposits occur in folded areas with inactive faults. We only know geologically interesting coal occurrences indicative of paleogeographic conditions and rapid facies variability. Their relation to crust thickness and types is the same as in Variscan basins. The thick granite layer was not favourable for the formation of larger coal deposits. Basins formed on a tectonically dissected flysch basement are infrequent, small and mostly superimposed. The intense tectonic dissection caused variable occurrence of coal deposits.

### 4.6 Coal deposits in Pannonian Basin and in peripheral units

The structures of this category comprise most coal deposits, economically significant by their extensive coal reserves and favourable mining conditions. This is the only area in the Carpathian segment with repeated coal-forming conditions. The conditions repeated most frequently in the Balaton block and on the contact of the South Carpathians with the Serbian-Macedonian Massif — the northern part of the coal-forming lineament. No coal deposits are in the centre of the Pannonian Basin.

With respect to deep structure the coal deposits concentrate to the following structures on basin margins:

a) to the Peripieninian lineament on the contact with the East Alps,

b) to the mobile more simatic crust zone of the Balaton block,

c) to the Peripieninian lineament on the contact with its East-Carpathian arm,

d) to the Insubric line — Periadriatic lineament,

e) to the margins of the Apuseni Mts.,

f) to the above mentioned Vardar-Kraishtide lineament, mainly to the margins at the Serbian-Macedonian Massif,

The position of the Liassic basin in the Mecsek Mts. is particular, so far unexplained.

G. HÁMOR (1984) emphasized the association of coal deposits to mobile zones near the Pannonian Basin periphery, and associated the origin of coal with Ottnangian (the end of the Sáva cycle), Badenian (Styrian cycle) and Pontian (Rhodanian cycle) orogenic cycles.

#### 4.6.1 Basins at western branch of Peripieninian lineament

All deposits are Neogene, spatially related to both sides of the lineament separating the Pannonian Basin from the East Alps. The deposits are associated with the neotectonic stage of the evolving seismotectonic lineament, geophysically indicated and located. This concerns the Badenian Brennberg basin in

Austria and the larger Pliocene lignite Szombathely basin. Tectonic activity caused the formation of thin coal seams. There are small coal deposits Tauchen and Ilz along the lineament in Austria. They are ranged to the East-Alpine unit (Fig. 69).

4.6.2 Basins on Balaton block

The Balaton block is tectonically disturbed by faults parallel to the Rába line and is most productive in coal (Fig. 69). Favourable paleogeographical conditions were offered by the Balaton — Velence granitoid belt controlling the formation of the elevation zone. Coal seams started developing during the Late Cretaceous in the Ajka brown-coal basin. The basin is a typical graben structure. Coal seams show a lower coalification extent.

During the Eocene the same fault line activated in the NE extension of the Cretaceous basin — the so-called north-western coal district (its total area including the coal-less part is about 1200 sqkm). The Eocene is also associated with the Pilisvörösvár Basin near the village Nagykovács and with the largest basin of this region — the Dorog Basin (200 sqkm, with coal seams in the Oligocene as well) with coal-bearing beds extending to the town Štúrovo. There also are the Tatabánya Basin, the southern Oroszlány — Pusztavám Basin and a basin between Bakonycsernye and Zirc. Eocene coal deposits are spatially restricted to the area W of the Danube r. — W of the Danube deep fault. The stable eastern area uplifted in the Eocene. So only a small basin near Nagyszáló formed E of the Danube fault. Coal-bearing beds are intensely tectonically disturbed with subsidence ranging to 700 m (the Dorog Basin). The coal belongs to brown-coal ortho- and metatypes.

Tectonic disturbance as well as the horst- and graben system were limited by the extent of NW-SE coal-bearing sediments deposition. Longitudinal basin axes are parallel to the NE-SW deep fault structure. Intense tectonic disturbation of the Dorog and Tatabánya Basins is associated with repeated block movements and deep master faults — the Rába and the Čertovica faults, and faults parallel to the Hurbanovo fault. Faults extending deeper in the Eocene sediments transport methane exploited in this area. According to F. SÓLSOM (personal information) the source of methane in grabens with complete Eocene sedimentary complexes, from which methane ascends into horsts. During the Neogene (Ottnangian, Badenian) coal-bearing sediments deposition revived in this segment only in a limited extent near Várpalota (50 sqkm), Szentgál and Herenda. Besides the Várpalota Basin there is only one low-quality coal seam. Also this area is intensely tectonically disturbed. During the Pannonian peat layers about 8—10 m thick formed.

Eocene basins have a graben structure with more but thinner coal seams. Miocene basins have frequently the character of platform basins — one coal seam with a greater average thickness (5—8 m). The coal seams thickness varies in individual basins: in the Dorog Basin is a 20 m thick coal seam; in the Tatabánya Basin the coal seams thickness ranges up to 37 m.

The thickness of coal seams increases towards NE (a more stable block) whereas the seams fade away towards SW owing to the marine environment.

The distribution of Neogene basins E of the Danube fault proves disintegration of the Balaton fault into more stable western blocks and more mobile eastern blocks during the Miocene. The disintegration was affected by the N-S deep fault within the block. There are no indications of the influence of the Darnó line upon the basins distribution. Tectonic dissection of the basement was only associated with basin subsidence. The basins rather follow NW-SE faults running from the Danube Basin and from the South-Slovakian Basin — the Ipeľská kotlina depression fault system.

In the eastern part of the Balaton block are largest brown-coal deposits of Hungary on the area of 1200 sqkm. Eggenburgian paralic deposits form a zone in the NW part, and Lower Pannonian in the SE part of the block. The basins distribution is indicative of the Pliocene tectonic restructuring and displacement of the continent margin from the NW periphery of the Pannonian Basin to its SE periphery. The coal-bearing Neogene was partly folded there (undulation with dips to 10°).

The Ottnangian coal-bearing sediments deposition was limited by the coast line oscillating to SE and E. There the thickness of the coal-bearing sequence increased to 250 m. Coal seams fade out in the same course. The western part was stable with 50 m thickness of the Eggenburgian sediments. Repeating coal-forming conditions (7 seams in the Sajószentpéter area) in paralic environment remind of foredeep basins. There is, however, a marginal depression of the Tertiary Pannonian block. During the Ottnangian the subsidence rate increased as well as the thickness (450 m) of the sedimentary sequence. The Badenian sedimentary sequence ranges to 100—150 m in thickness. The deposion of sediments was associated with volcanism culminating in the Pliocene.

The Pannonian period is characterized by temporary stabilization of the formation of 1—3 lignite seams, 2—5 m thick. On the southern foot of the Mátra and Bükk Mts. Was a more mobile depression with oscillating basement movements. These cause the formation of 17 coal seams. The basin is on the Balaton line activated by neovolcanism in this place.

### 4.6.3 Basins at northern branch of Peripieninian lineament

Only Neogene basins occur on the internal side of the Peripieninian lineament and on the outer side of volcanic mountain ranges adjacent to the lineament. The basins comprise the Podvihorlatská panva basin (Fig. 69) inside the crossing of deeper and deep faults. Other Neogene basins are in Romania: Bicad, Borsec, Baraolt-Čapeni. All the basins are associated with neovolcanic units (which locally affects the intensity of coalification). The last two basins are on the eastern periphery of the Transylvanian Depression. Their position is analogous to the position of coal basins in the eastern part of the Balaton block. Their evolution is also similar. The basins are small, with few coal seams of which usually only one is thick.

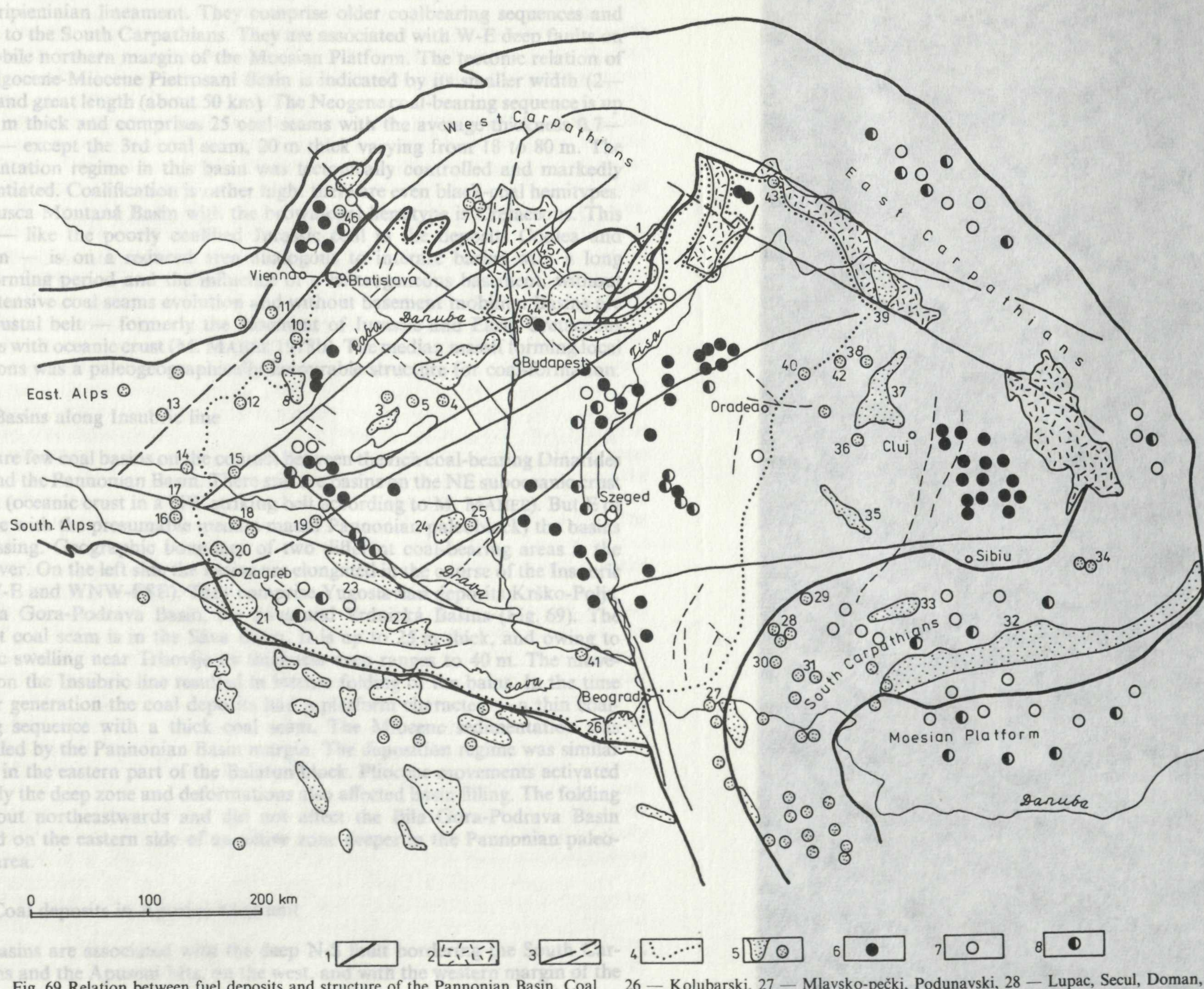


Fig. 69 Relation between fuel deposits and structure of the Pannonian Basin. Coal basins and smaller deposits.
Explanations: 1 — north-eastern b. (Nógrád, Borsodnádasd, Mátravidék), 2 — north-western b., 3 — Ajka, 4 — Várpalota, 5 — Szentgál, 6 — Hodonín, Dubňany, 7 — Handlová-Nováky b., 8 — Torony—Szombathely, 9 — Tauchen, 10 — Brennberg, 11 — Wiener Neustadt, 12 — Ilz, 13 — Koflach—Voitsberg, 14 — Lavanttal, 15 — Wiess, 16 — Zasavski, 17 — Velenje, 18 — Hrvatsko—Zagorje, 19 — Ludberg—Koprivnice, 20 — Krško Polje, 21 — Pokupsko-Vukomerički, 22 — Posavski, 23 — Bilogorskopodravski, 24 — Mecsek, 25 — Hidas, 26 — Kolubarski, 27 — Mlavsko-pečki, Podunavski, 28 — Lupac, Secul, Doman, 29 — Rusca—Montana, 30 — Anina, 31 — Rudaria, Mehadia, 32 — Subcarpatii, 33 — Petrosani, 34 — Codlea, Cristian, 35 — Brad, 36 — Ip-Zauani, 37 — Almasului, 38 — Saemasag, 39 — Bicsad, 40 — Popesti, 41 — Vrdnički, 42 — Bratca, 43 — Podvihorlatská, 44 — Nagyszál, 45 — modrokamenská, 46 — Kúty—Gbely.
Explanations: 1 — Inner Carpathians boundary, 2 — neovolcanic rocks, 3 — deep-seated faults, lineaments and faults, 4 — Pannonian Basin boundary, 5 — coal basins and smaller deposits, 6 — natural gas deposits, 7 — oil deposits, 8 — oil and gas deposits.

Basins Rusca Montană and Pietrosani are in a similar position — but out of the Peripieninian lineament. They comprise older coalbearing sequences and belong to the South Carpathians. They are associated with W-E deep faults on the mobile northern margin of the Moesian Platform. The tectonic relation of the Oligocene-Miocene Pietrosani Basin is indicated by its smaller width (2—9 km) and great length (about 50 km). The Neogene coal-bearing sequence is up to 300 m thick and comprises 25 coal seams with the average thickness 0.7—1.0 m — except the 3rd coal seam, 20 m thick varying from 18 to 80 m. The sedimentation regime in this basin was tectonically controlled and markedly differentiated. Coalification is rather high: there are even black-coal hemitypes. The Rusca Montană Basin with the brown-coal hemitype is Cretaceous. This basin — like the poorly coalified Jurassic coal in the deposits Codlea and Cristian — is on a reduced area analogous to interarc basins with a long coal-forming period and the influence of a heterogeneous basement, without any extensive coal seams evolution and without basement mobility. Basins are in a crustal belt — formerly the basement of Jurassic and Early Cretaceous troughs with oceanic crust (M. MAHEĽ 1978b). The median massif forming local elevations was a paleogeographically favourable structure for coal-formation.

4.6.4 Basins along Insubric line

There are few coal basins on the contact between the rich coal-bearing Dinarides zone and the Pannonian Basin. There still are basins on the NE suboceanic crust margin (oceanic crust in a NW striking belt according to M. MAHEĽ). But E of Danube with the presumable median massif Pannonian paleoblock) the basins are missing. Geographic boundary of two different coal-bearing areas is the Sava river. On the left side the basins are elongated in the course of the Insubric line (W-E and WNW-ESE). They comprise Yugoslavian deposits Krško-Polje, the Bila Gora-Podrava Basin, the Sáva and Vrdnická Basins (Fig. 69). The thickest coal seam is in the Sáva Basin. It is up to 25 m thick, and owing to tectonic swelling near Trbovlje its thickness even ranges to 40 m. The movements on the Insubric line resulted in intense folding of the bains. In the time of their generation the coal deposits had a platform character — a thin coal-bearing sequence with a thick coal seam. The Miocene sedimentation was controlled by the Pannonian Basin margin. The deposition regime was similar to that in the eastern part of the Balaton block. Pliocene movements activated intensely the deep zone and deformations also affected basin filling. The folding faded out northeastwards and did not affect the Bila Gora-Podrava Basin situated on the eastern side of an active zone deeper in the Pannonian paleoblock area.

4.6.5 Coal deposits in Apuseni Mts. unit

Coal basins are associated with the deep N-S fault bordering the South Carpathians and the Apuseni Mts. on the west, and with the western margin of the

Transylvanian Depression. The Marghita Basin is in a zone of presumable oceanic crust extending to ENE from the Szolnok trough basement. The Almássy Basin joins the N-S crust belt at the western margin of the Transylvanian block. The deposition of coal-forming sediments proceeded in the Oligocene here, and during the Miocene in the western part. The basement fault tectonics is not exactly known so the relation of individual basins to deeper structures cannot be explained. The distribution of basins (Popesti, Bratec and Ip — Zănasi) is indicative of a larger NW-SE fault dissecting the Apuseni unit into horsts and depressions (Fig. 69).

### 4.6.6 Coal deposits in northern part of Vardar-Kraishtide zone

The entire zone was characterized at the beginning of this chapter. The Permian-Carboniferous deposits are highly coalified, perhaps owing to the higher geothermal field on the deep zone. Intensity of intermittent subsidence movements inside the zone perhaps on the contact of plates with various crust types is indicated by plentiful thin coal seams. Tectonic movements resulted in intense folding, in many overthrusts (e. g. the Jurassic Anina deposit), in segmentation, crushing and mylonitization of Carboniferous coal (deposits Lupac, Secul). The deposit Baia Noire comprises a particular association of a Carboniferous coal seam with a serpentinite massif. The coal seam is on the contact of rocks of the oceanic type with continental quartz porphyries. Before the Miocene the generation of coal deposits in this area was mostly controlled by the Kraishtide zone and by the Getic and supra-Getic units of the South Carpathians. The units originated from a deep ancient continental crust belt. Among the units the Serbian-Macedonian Massif controlled decisingly the evolution of continental facies favourable for the coal-bearing sequences. Advancing mobilization and segmentation of continental crust were favourable for the formation of elongated narrow coal depressions. They mostly formed in the Jurassic time.

During the Cretaceous the subsidence movements shifted to the western margin of the Kraishtide zone with advancing elevations of the eastern part of the massif. Neogene coal basins in the area of the Vardar zone, resting on the oceanic crust, formed in the Jurassic and Early Cretaceous.

Regularities of the deposits distribution facilitate explanation of the evolution of this lineament zone. E. BONTCHEV'S (1976) opinion about the existence of the Vardar-Kraishtide lineament is to be precised with respect to the history of this zone.

The earlier pre-Mesozoic mobilization proceeded in the eastern part of the Kraishtide zone with adjacent Getic zones. During the Jurassic the Vardar lineament formed. It is younger than Paleozoic-according to B. SIKOŠEK (1976). The tectonic-historical evolution stages preserved and the structures seem a single broad structure comprising the Serbian-Macedonian Massif. The explanation of their evolution was facilitated by data on crustat types (M. MAHEĽ 1978b) and on the evolution of coal-bearing basins. The younger Vardar zone keeps its deep activity as a seismoactive zone.

### 4.6.7 Basin in Mecsek Mts.

The Liassic coal basin in the Mecsek Mts. is in a specific position among Mesozoic units. Its tectonic relation to surrounding geologic units is not clear because of so far unknown relations between the Mecsek Mts. and adjacent units (M. Maheľ 1978b).

The number of coal seams is anomalous in relation to other deposits of the Pannonian Basin basement. In the area of Pécs are 175 coal seams of variable thickness. The coal may partly be charred. Tectonic disturbance of coal seams was intensive. Thickness of Jurassic sediments increases from 2500 m to 4000 m from NE to SW (A. K. Matveev 1966) as well as the number of coal seams. Towards NE the lower coal-bearing beds fade away and the youngest beds show a transgressive character corresponding to the NE extension of the depositional area. A great sediment thickness in the SW part and increasing amount of alluvial facies indicate the existence of a denudation area on the SW, separated in the basin basement — perhaps tectonically — from the most mobile zone.

The depositional area rests upon mobile crust — upon a trough with the deposition proceeding since the Stephanian. Total thickness of Paleozoic-Mesozoic sediments is estimated to 8.5 km (A. K. Matveev 1966). The coalification degree increases from the SW towards the mobile zone on the NE.

The crust formation was decisingly controlled by Upper Cretaceous granitoid magmatism associated with migmatitization in surrounding rocks. The magmatism caused the formation of a NE-SW striking elevation zone supplying the depositional basin with material during the Jurassic. The zone is likely to have formed a transverse elevation restricting the area of the coal basin in SW. The mobile trough was partly consolidated by Hercynian events and its crust was thinned. During the Cretaceous there occurred alcaline volcanism and I regard it as a manifestation of rift structure.

During the Liassic the sea advanced gradually towards NE, the western part of the Pannonian paleoblock was uplifted and the sea concentrated in the Alföld-Szolnok trough. The NE-SW tectonic-paleogeographic segmentation commencing in the Liassic and continuing up to the Cretaceous shows that redistribution of tectonic units presumably adjacent to the Mecsek Mts. since the Late Cretaceous — according to I. Varga (1978) is problematic. It is more likely that M. Maheľ's (1978a) opinion about former connection with the Serbian-Macedonian Massif is right and the massif might have been the tectonic border of the coal-bearing basin. It is however, not excluded, that the sialic crustal unit of the western part of the Pannonian paleoblock was separated by the Insubric line from sialic cores of the NW Dinarides, and so it was genetically more related to the Dinaric area. In SW the coal-bearing basin might also have been tectonically bordered by faults parallel to the Insubric line.

The tectonic detachment of the gradually uplifting source area is proved by cyclical sedimentation (average sedimentation cycle thickness is about 4 m), and by rapidly fading-away sandstones towards NE over the distance of 1.5 km. According to exploitation data (unpublished) the transverse fault in relation to the basin, i. e. to the WNW-ESE fault formed as early as the Late Paleozoic. Formerly the southwestern block had a downwarp character with Paleozoic sedimentation on the basement. The northeastern block was exposed, and the Triassic sequences transgressed upon the crystalline complexes. The present tectonic division into a subsided NW block and an elevated SE block with granites is due to rejuvenation of pre-Alpine faults. The Mecsek coal-bearing Mesozoic Basin formed on the crossing of ancient faults. The NE-SW faults are dominant. Their course is identic with the main tectonic segmentation of the Pannonian paleoblock. It controlled the NE mobility of the basement blocks bordered by NW-SE faults. The transverse segmentation is also manifested in different tectonic structure and in the coal reserves in the basins (the Pécs, Komló and Nagymanek coal districts). Tectonic disturbance intensity increases to NE, i. e. in the course of gradual mobilization probably caused by the basin basement compression. Mobilization resulted in unfavourable conditions for deposition of coal-bearing sediments, extending after the uplift northeastwards up to SE elevations. The more stable Pécs block also controlled the formation of coal seams connected in one 20—25 m thick seam.

The end of the deposition of coal-bearing sediments corresponds to increasing crust consolidation. The crust was already continental in the Cretaceous, and the revived deep activity on faults resulted in structures resemblant to continental rifts.

## 4.7 Coalification in relation to tectonic structures

Basins of the Pannonian and circum-Pannonian regions comprise all coal-deposit types — from anthracites to lignites. The highest coalification degree is characteristic of Carboniferous deposits, mostly those along the western periphery of the South Carpathians. Jurassic coal deposits are highly coalified as well. The maximum coalification is associated with tectonically mobile zones, mostly near deep faults or on them. I associate coalification with a higher fossil heat flow on the Carboniferous structures but the flux is not associated with folding. The folding and/or tectonic pressure had a positive influence upon the coal quality, mainly its coking capacity, as far as coalification did not reach the highest metamorphism degree. Among Jurassic deposits the coal in the Mecsek Basin is most coalified as well as some smaller deposits in Romania (mainly the Pietrosani Basin). Coal in the deposits Codlea, Cristian and Rusca Montană regions is less coalified.

Among Tertiary deposits most coalified are those in the Comanesti Basin in the Outer Carpathians and the Slovak deposit Handlová. In the Pannonian Basin higher coalification is characteristic of the NW part of Hungary (Tatabánya) showing a higher coal metamorphism than the Cretaceous Ajka deposit.

Most deposits in the Pannonian Basin show low coalification. The coal represents the brown-coal hemitypes or lignites. Most lignites are Pannonian-Pliocene. Younger coal deposits seem less coalified but this is denied by differences between Mesozoic and Eocene deposits.

The Mecsek Basin shows that coalification is controlled by the Jurassic crustal mobility beneath the basin and increases with subsidence rate. Coalification is not controlled by the extent of tectonic disturbance increasing towards less coalified basin parts. There, however, is a striking agreement with the position of the Pannonian diapir and its tectonic activity. Whereas brown coal in Eocene and in some Early Miocene deposits is close to brown-coal metatype at a smaller submersion depth (sometimes in shelf environment) whereas the most frequent Pliocene lignites are poorly coalified even in the case of a greater submersion of coal-bearing beds (more than 300 m thick). Deposits with poorly coalified seams also comprise the deposit near the town Štúrovo, only it is older.

I explain differences in coalification by fossil geothermal activity of the mantle diapir in the time of the formation of coal seams and after their covering by younger sediments. I presume the maximum thermal activity in the Eocene and Early Miocene — in the time of diapir initial expansion. The Pliocene collapse and cooling of diapir resulted in decreasing temperature in the first hundreds of meters beneath the Earth's surface. This is why in spite of intense tectonic dissection of Pliocene basins and extensive subsidence of blocks with coal deposits no coal of a higher quality and metamorphosis degree formed, only lignite.

Spatial changes in coalification are controlled by tectonic exposition of deposits under different geothermic conditions in mobile and stable units. It is likely that tectonic stress had a catalyzation effects upon coalification, and so the maximum coalification is characteristic of coal deposits in the Romanian Carpathians.

Time variability in coalification are controlled by thermal evolution and mantle diapir thermal effects. Heat transfer due to radiation is slow because of a low heat conductivity of the crust, so the heat transfer must have been affected by the secondary heat chambers and by the uprising magma and hydrothermal fluids. During the Late Miocene then in these regions and around them the postvolcanic geothermal heat uprising flux proceeded and perhaps affected coalification. So the maximum coalification is associated with a close vicinity of the neovolcanic-active Balaton line, and with faults parallel to the line. The Danube faults represented a deep heat shield and separated the warmer western part from the colder eastern part. I explain the influence of volcanism upon the Hungarian NE striking coal region with lignites by the later generation of coal, late after the andesite and rhyolite volcanic activity. The basalt volcanism, limited in space was not associated by a sufficiently extensive thermal field. Melting centres were not in the crust anymore, they were deeper in the mantle, and the heat radiation was prevented by thick continental crust. For coalification most important were shallow secondary crustal magma chambers and gradually cooling subvolcanic bodies. Rapid uprise of less viscous basic lavas is excluded. Coal as a sensitive thermal indicator reacted by coalification changes upon short heat effects whereas the surrounding sediments were not affected by higher compaction.

The low coalification in Pliocene basins indicates that recent heat flow and vertical thermal gradient around 50 °C/km are not sufficient for higher coalification. The values should be much higher in the time of coalification. Temperature and pressure conditions of the rock metamorphosis in the East-Slovakian Basin do not produce the corresponding degree of metamorphosis of Neogene sediments (D. ĎURICA et al. 1979). Perhaps the time factor is important for the influence of pt conditions.

### 4.8 Relation of coal basins to crust types and to tectonic regime

I based my study of the relation of coal basins to basement structures and to crust types upon complex data on crustal types, presented by M. MAHEĽ (1978b). I tried to find such features in basin types which might serve as correlation characters of crust types. The existing ideas of these relations are based on the presumption about coal-bearing basins being continental structures. So it is presumed that they only formed upon mature continental crust as foredeeps, intermontane basins or basins on platform. Basing on data about Variscan basins (J. ZEMAN 1977) and on my own data about the North--Hungarian basin, I have tried to reveal elements indicative of a different type of continental crust — as far as the relation is manifested in the structure or evolution of basins.

There are relations among the extent of tectonic disturbance, the coal seams evolution and the basement mobility. The relations may, however, depend upon intense tectonic dissection of continental crust and suboceanic crust tending to higher mobility. The crust type cannot be determined only according to the extent of tectonic disturbance. So it is with the character of the coal seams evolution. The mobile basement is manifested in many thinner coal seams and with great facies variableness. Stable units represented by platform are characterized by few coal seams — mostly one but thick.

My evaluation of all elements has not led to a unambiguous conclusion. Coal basins on block margins inside orogenic belts or in mobilized median massifs show equal tectonic disturbance and coal seams evolution like coal basins on the former, probably suboceanic crust. So the crust type must be determined according to a complex of other geologic features. Coal basins also have some features of basins on a mobile basement. The Inner-Carpathian basins have the only feature in common: their limited areal extent.

Large coal basins are on block margins. The Salgótarján Basin and other northeastern Hungarian coal basins are on the contact between the Pannonian and the West-Carpathian blocks, the sub-Carpathian basin is on the northern margin of the Moesian Platform. The Lom Basin in the south is in a similar position. There are large basins out of block boundaries in the Dinarides and in the northwestern Hungarian coal district. The basins are probably on suboceanic crust; in the Dinarides on pre-Tertiary oceanic crust. They underwent intense tectonic disturbance but out of orogenic regions they are only disturbed by faults. Suboceanic crust may also be favourable for coal basins because of its tendency to repeated subsidence resulting in rejuvenating conditions for coal-bearing sediments deposition in case the subsidence proceeds under favourable paleogeographic conditions near elevations, with the uplift and subsidence compensation. Suboceanic crust is also favourable for coal basins because it forms extensive lowlands which cannot form in areas of growing uprising granite layer.

Continental environment is favourable for deposition of coal-bearing sediments and so mostly pre-Neogene basins occur near tectonic margins of ancient median massifs on a crust of continental type, consolidated more intensely by Hercynian events. Position and evolution of basins indicate that they have not formed in platform regime although some coal seams resemble in some features the platform basins in the Carpathian foreland. Nowhere coal-bearing layers formed on larger area. In this and in their genetic relation to disquiet tectonic regime they differ from platform coal basins.

Basing on these and other features (intense tectonic disturbance, variable coal seams evolution) I think that coal basins are in a closer relation to basins and/or segments of mobile zones. The limited basin extension is due to their predisposed position on deeper faults or lineaments separating the continental crust from suboceanic or oceanic. The limited areal extent of pre-Tertiary coal basins is indicative of a limited area of continental elevations, and of prevalent units covered with the sea. The tectonic regime disturbed compensation of subsidence by water reservoirs and this resulted in unfavourable paleogeographic conditions for coal seams. Therefore no coal depressions formed near the crust of oceanic types nor on blocks with suboceanic crust.

It is surprising in comparison to Variscan coal basins that in median massifs were no favourable conditions for larger coal deposits. It was because the median massifs were not surrounded by proper source areas — folding zones. The zones migrated to the external side of median massifs and the tectonic regime, causing the cyclical sedimentation and repeating conditions favourable for coal seams and their burial and protection by sediments (compensated basins) faded away.

So pre-Tertiary basins formed on continental crust and/or on its margins detached from thinned or suboceanic crust. A comparison of relations between the Permian-Carboniferous and Tertiary basins — as far as they are on thicker continental crust — may facilitate explanation of the limited areas and coal reserves of the basins by the limited subsidence on continental crust restricted to large faults. The Kraishtide zone seems to be an exception. But owing to intense tectonic activity a larger area of continental crust got mobile and inside the crust smaller basins with low coal reserves formed.

An actual exception is the Mecsek Basin with its character of a small intermontane depression, comparable to analogous Variscan structures. The basin is deeper in the Pannonian paleoblock which was not involved in the intense Liassic subsidence uncompensated by sedimentation. The intermittent variable subsidence was controlled by mobility of the basement transverse blocks. The Mecsek Basin represented mobile basins which is indicative of a generally labile regime also in consolidated units of continental crust of the Pannonian Basin.

Thus the data on intense Mesozoic activation and oceanization of crust complement the data on coal basins.

Paleogene basins, mostly Oligocene, are indicative of another regime. The basins formed on more mobile suboceanic crust (except the Almássy Basin in Romania with unidentifiable crust type) in the time of stabilizing tectonic regime and partial sea regressions. So in contrast to ancient coal deposits the basins are larger, more stable, with thicker coal seams. Their structure displays elements of intermontane depressions.

Miocene and Pliocene coal-forming areas are most frequent. Deposition of coal-bearing sediments commenced during the Early Miocene in the SW region around the Dinarides; during the Late Miocene the deposition continued on the Balaton block, and in the Pliocene the most intense deposition proceeded in Romania. So in the Neogene the deposition of coal-bearing sediments migrated along the Pannonian Basin margins from S to N and gradually to E. It was affected by volcanism in the northern and eastern parts. Basins formed on a block with suboceanic crust or on block boundaries with different crust. The basins formed near the margins of main depressions: the Pannonian, Transylvanian and the Moesian depressions.

Pliocene basins are different. They are small, associated to unextensive graben-formed depressions. Their origin and distribution are controlled by Pliocene block segmentation. The segmentation also affected ancient coal basins and resulted in an intricate block structure obstructing the mining-economical exploitation of coal reserves. The post-Miocene change of tectonic regime is reflected in the nature of coal basins. Owing to varidirectional mobility the Pliocene basins are chaotically dispersed in the eastern part of the Pannonian Basin. Since the block segmentation was unintense in the Balaton block, two largest lignite basins formed in the area of Mátra — Bükk Mts.

The analysis of data about Carpathian basins shows and agreement between stability and thickness of coal seams and the character mobility and tectonic segmentation of the crust. Intense segmentation (lineaments) and mobility are causes of plentiful coal seams, repeating coal-forming conditions as well as small thickness of coal seams due to interruptions in coal-forming conditions. There also is an agreement in tectonic disturbances intensified in earlier segmented zones. A mobile, poorly consolidated crust with poor tectonic segmentation due to inherited movements on persistent faults is optimal. The Balaton block is a typical example.

Summary: On the basis of the existing data about basin evolution in time we may state a relation between pre-Tertiary coal basins and Paleozoic continental crust, mainly its margins, whereas Tertiary basins are mostly associated with formerly suboceanic crust. Distribution of pre-Tertiary basins indicates heterogeneous basement of the Pannonian Basin, whereas Tertiary basins react upon the formation of the Pannonian block and of a large depression, and concentrate along the margins of the megastructure. They also react to the remnants of ancient block structure, to block disintegration and younger fault activity mainly since the Late Pliocene. In neovolcanic areas the basins react to

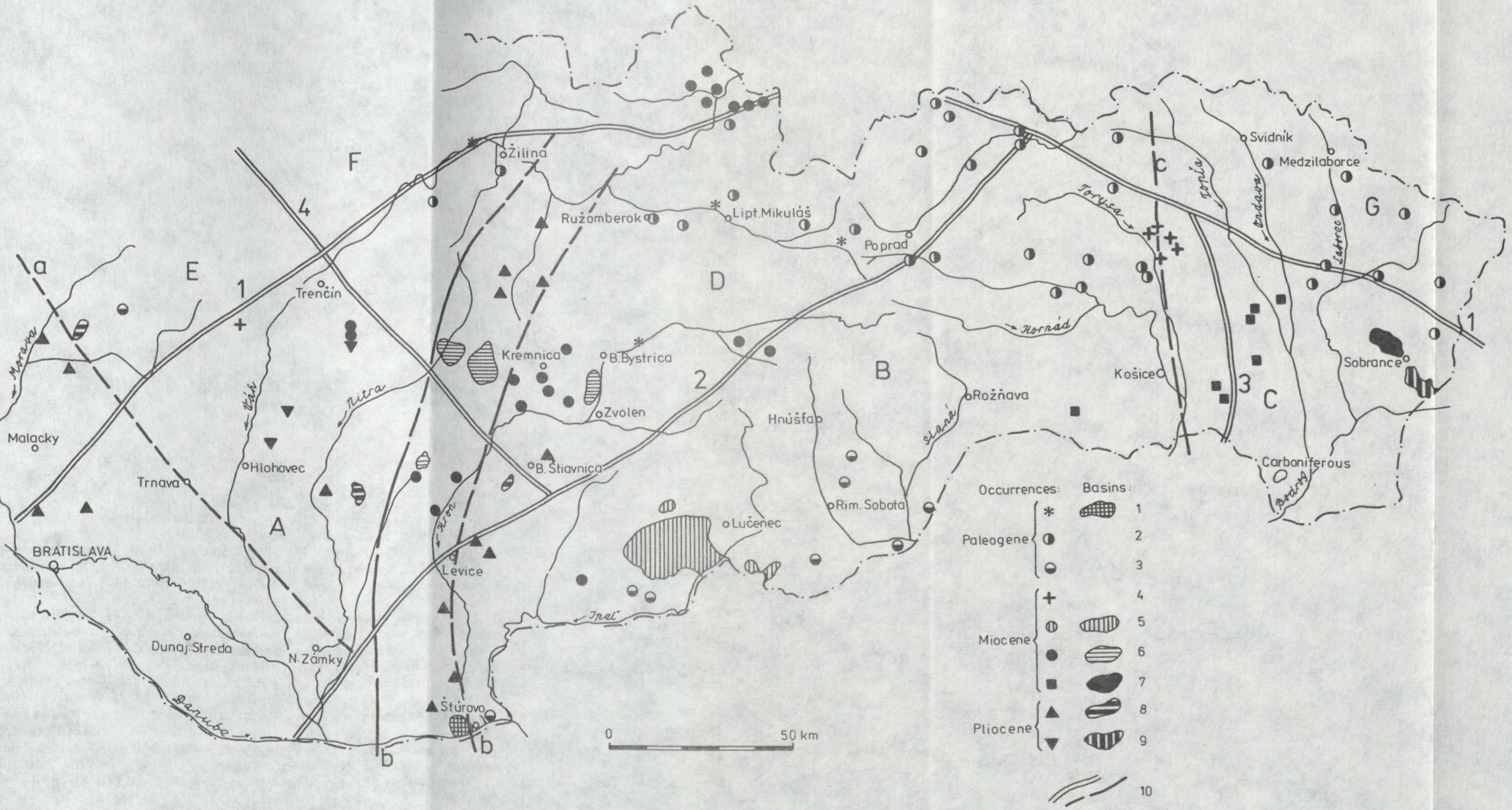


Fig. 70 Coal occurrences and coal basins of Slovakia in relation to deep structure.

Deep-seated faults: 1 — Peripieninian lineament, 2 — Vepor, 3 — Slánske vrchy deep-seated fault, 4 — Přerov—Štiavnica; a — Nesvačilka—Trnava, b — Inner-Carpathian, c — Hornád.

Blocks: A — Danube, B — South-Slovakian, C — East-Slovakian, D — Fatran-Tatran, E — Slovakian-Moravian, F — Slovakian-Silesian, G — Beskydy-Bukovec.

Coal occurrences and coal basins: 1 — Lower and Middle Eocene coal occurrences, 2 — Upper Eocene and Lower Oligocene occurrences, 3 — Upper Oligocene coal occurrences, 4 — Burdigalian (Eggenburgian) coal occurrences, 5 — Helvetian (Carpathian) coal occurrences, 6 — Tortonian (Badenian) coal occurrences, 7 — Sarmatian coal occurrences, 8 — Lower Pontian coal occurrences, 9 — Upper Pontian coal occurrences, 10 — deep-seated faults.

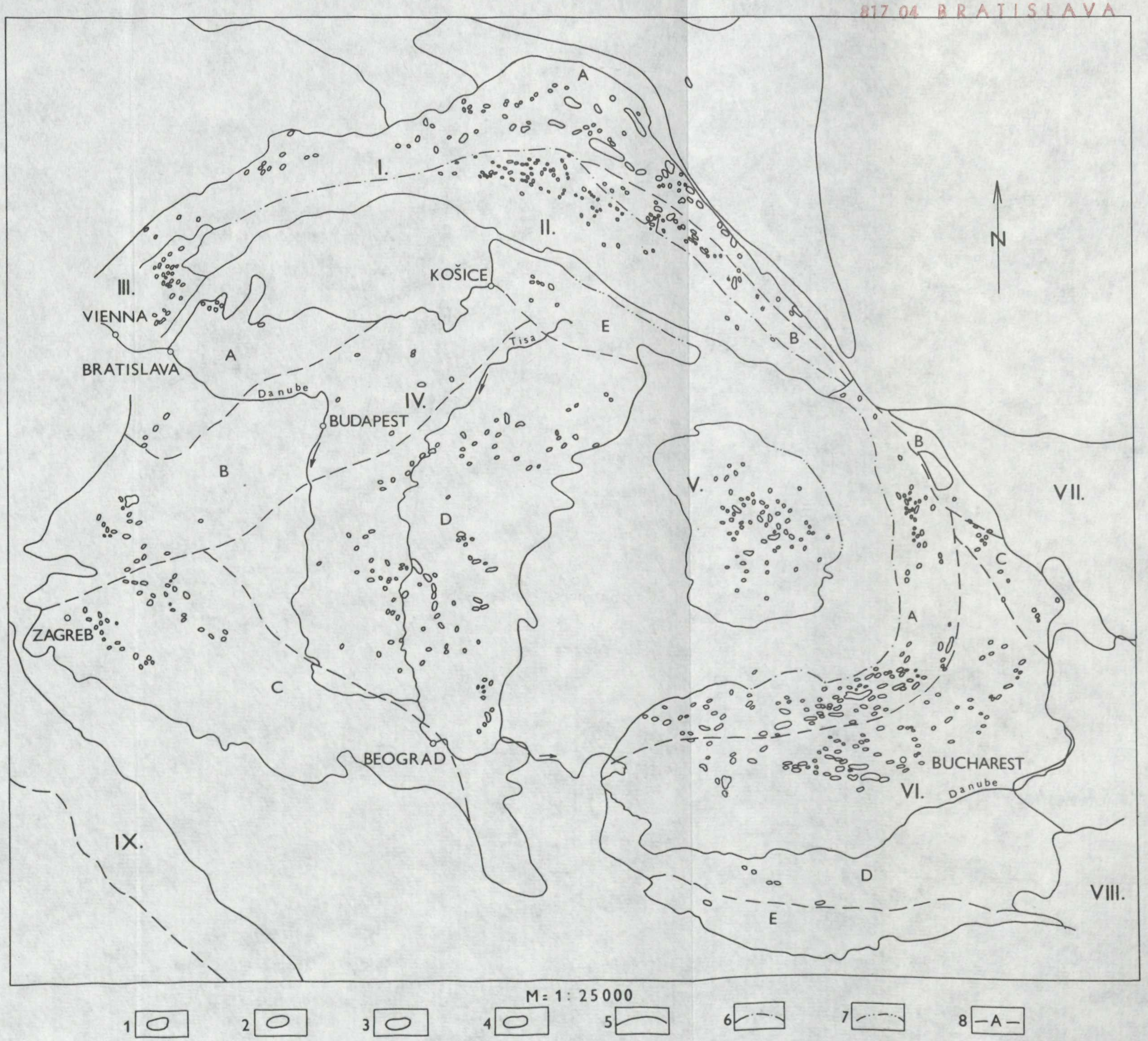


Fig. 71 Schematic map of hydrocarbon deposits in the Carpathian-Balkan region according to V. SEMENOVICH et al. 1977, adapted).

Explanations: I — Northern Fore-Carpathian oil-gas-bearing basin: A — dominantly gas-bearing area of the outer unfolded part of the foredeep, B — dominantly oil-bearing area of the inner folded part of the foredeep; II — Carpathian oil- and gas-bearing basin, III — Viennese-Moravian oil- and gas-bearing basin, IV — Pannonian oil- and gas-bearing basin: A — Little Hungarian-Danubian gas-bearing area, B — Midmountain Igal-Bükk oil- and gas-bearing area, C — Savian-Dravian oil- and gas-bearing area, D — Alföld-Banát-Bácska oil- and gas-bearing area, E — Transcarpathian, mostly gas-bearing area, V — Transylvanian gas-bearing basin, VI — Fore-Carpathian-Balkan oil- and gas-bearign basin; A — oil- and gas-bearing area of Carpathian Foredeep, B — gas-bearing area of Moldavian platform, C — oil- and gas-bearing area of Northern Dobrodgea, D — oil- and gas-bearing area of the Moesian platform, E — oil- and gas-bearing area of Fore-Balkan, VII — Northern Black Sea oil- and gas-bearing basin, VIII — Western Black Sea oil- and gas-bearing basin, IX — Adriatic oil- and gasbearing basin.

1 — oil deposits, 2 — gas deposits, 3 — oil and gas deposits, 4 — deposits of gas condensates, 5 — boundaries of oil-bearing basins, 6 — boundaries of oil-bearing basins beneath overthrust Carpathians Mts., 7 — boundaries of oil-bearing basins beneath Neogene volcanic rocks, 8 — boundaries of oil-beraing areas.

younger structures with volcanic activity or activated by volcanism — except volcanic centres. Basins on mobile blocks (the Balaton block) are largest and concentrate in coal-bearing belts. Basins on deep faults also concentrate in belts but the belts are narrow and their coal content is low.

Basins on large blocks, like on the Pannonian megablock are controlled by the heterogeneous structure resulting during regional subsidence (Pliocene) in numerous small block structures preventing concentration of coal-bearing sediments on a large area. Generally large deposits are associated with units — crustal blocks tending to subsidence with poorly consolidated crust, and with block margins with the crust reduced in thickness by secondary processes (basification?). Smaller deposits (frequent on thicker crust) are associated with faults or lineaments. The basin size was also affected by tectonic regime controlled by mantle events.

## 4.9 Relationship between basins and lithosphere dynamics

Data on the evolution of Neogene depositional basins in relation to mantle diapir, applied on regularities in the evolution of coal basins, support the mantle diapir conception. Tertiary coal basins, mostly Late Neogene, formed under a new structural pattern inheriting ancient structural lines controlling margins of blocks separated by post-Badenian events. Concentration of basins on the Pannonian Basin margin indicates that there were unfavourable conditions for subsidence in the central part of the basin in the time of tension caused by uprising mantle elevation. In the time of diapir collapse and rapid Pliocene subsidence mostly uncompensated large basins unfavourable for coal, formed. (I want to emphasize again that the term "diapir collapse" comprises rather the phase of downsagging thin lithosphere blocks into hot asthenosphere elevation, corresponding to volcanism in most basins, and not only the cooling phase). Favourable conditions were only along more stable margins or horsts, and on the contact with continent in uprising Carpathian internal units. Like the Eastern Alps, the Carpathians supplied the rock material and water (water flows) inevitable for vegetation. Vegetation was also supported by volcanism supplying carbon dioxide to coal-forming areas, and minerals — fertilizers to soils.

## 4.10 West-Carpathian coal basins in relation to block structure

Besides tectonic activity also paleogeographic character of the area is significant for coal-forming conditions. Paleogeographic data are not complete. The existing data on the Tertiary (P. Gross 1978, J. Gašparik 1978) show that the Peripieninian lineament, separating the subsidence area in the northern part of the West Carpathians from the southern part, participated in the lithofacial segregation of marine sediments already in the Paleogene. In the southern part, mainly in the Pannonian Basin, the continental crust matured by calc-alkaline volcanic activity, and elevation trend supported generation of coal-forming basins. In the Inner Carpathians the deposition areas were segmented by rejuvenation of fault structures of Paleogene basins basement, followed by differentiated contrast tectonic movements at the end of the Paleogene. The tectonic movements were favourable for deposition of coal-forming sediments and unfavourable for coal deposits accumulations. It is evidenced by plentiful economically insignificant Paleogene coal occurrences. The Paleogene occurrences are not related with block structure in Flysch zones. The correlation is better in back depressions, mainly in Hungary (see above).

A decising tectonic restructuring proceeded in the Miocene and the sea/continent boundary generally migrated southwards. The migration also comprised movements on block margins, mainly on NS faults along which the coal-forming conditions also returned migrating northwards inside the Carpathians. The Pliocene movements were markedly influenced by the evolution of blocks.

a) In the Carpathian Foredeep only local coal accumulations formed.

b) In the Vienna Basin favourable conditions formed NE of the Nesvačilka —Trnava line in the more stable basin part. Favourable conditions were on the contact of the line with the elevation zone around the Peripieninian lineament (the Gbely deposit) and on the contact near the Lednice zone — the Hodonín deposit.

c) In the East—Slovakian region are coal occurrences along margins of the mobile East-Slovakian block (the Slánske vrchy hills deep fault). Favourable conditions were along the Peripieninian lineament in the more stable Humenné block (Fig. 70). There is some analogy with the NW branch of the lineament. Favourable coal-forming conditions occurred repeatedly since the Paleogene through the Eggenburgian to the Sarmatian on the Hornád fault and the Slánske vrchy hills deep fault. In the Paleogene favourable conditions were on the Peripieninian lineament between Stará Ľubovňa and Vihorlat and farther in the Soviet Union. Because of intense tectonic segmentation — like along the margins of the East-Slovakian block — no favourable conditions formed for larger coal deposits concentrations. The only exception is the above mentioned Humenné block uplifting owing to Vihorlat volcanism following the deep magmatic elevation (perhaps of the secondary magma chamber).

The Vienna Basin and the East-Slovakian Basin show a common feature: large coal occurrences concentrations on a smaller area, including economically interesting deposits and concentrations of principal gas deposits of Czechoslovakia. Besides variable lithofacies due to intense tectonic differentiation of basin subsidence, the above situation is perhaps mostly caused by specific basin evolution controlled by their interblock position, tectonic exposition along the Peripieninian lineament on the contact with thin crust and dynamic effects of mantle diapir beneath the Pannonian Basin.

d) Further deposits ranged to small Inner-Carpathian basins are associated with the Central-Slovakian Fatra—Tatry block. The most important deposits are on the SW margin of the block at the Přerov—Štiavnica deep fault on its crossing with the N-S fault zone of the Central-Carpathian lineament. This lineament also influenced (cf. East-Slovakian region) concentration of coal deposits (Handlová, Nováky, Badín a. o.). The influence of the N-S faults may also be traced in the Danube block at the localities Obyce, Hronský Beňadik a. o., and in the South-Slovakian block — at localities of Lower Pliocene occurrences in the Hron r. valley, the Štúrovo deposit. The zone of the Central-Carpathian lineament, most likely formed in the Neogene, shows the features of a coal-bearing lineament.

The Fatra—Tatry block with intensely dissected continental crust with frequent alternation of elevation and depression zones comprises most frequent coal occurrences and economically interesting deposits. Coal-forming conditions may be traced there since the Early Eocene to the Middle Pannonian s. l. (Early Pontian — Fig. 70).

The Danube block is poorest in coal occurrences, and besides the Beladice locality the coal occurrences are only associated with the above mentioned N-S belt. High mobility of the Danube block caused its inundation by the sea margin in the Miocene. The paleogeographic configuration of the coast was favourable for the generation of coal seams. Spatially limited elevation extent was paleogeographically unfavourable for larger coal deposits in the Early and Late Pliocene. A greater areal extent of depressions caused long sea inundation.

The South-Slovakian block has the greatest areal extent. Coal deposits near the town Štúrovo and the South-Slovakian Basin are situated in the block. These Eocene-Pontian coal occurrences are indicative of partial uplifts, most extensive in Hungary. There also are the largest coal basins. The position of the South-Slovakian coal basin proves the association with margins and nearness of deep faults; in this case the continuation of the Přerov—Štiavnica deep fault which might have been active farther southeastwards in the pre-Badenian time. The post-Badenian block consists of six smaller blocks with different internal structure. Blocks distinguished by O. FUSÁN et al. (1979) correspond to variably mobile units with reversible vertical movements. Most coal occurrences and main deposits are in the Šahy block and — besides the above mentioned continuation of the Přerov—Štiavnica fault — also in the N-S part of the Central-Carpathian lineament.

A correlation of coal occurrences to coal basins with deep block structure shows that:

a) Coal deposits concentrate to block margins incorporated in the block structure formed in its present character at the end of the Miocene.

b) The greatest coal reserves concentrate in the Vienna Basin and in the East-Slovakian Basin. They are in the position of interblock basins.

c) Coal deposits in the South-Slovakian Basin are in a similar position on thin crust poorly consolidated by pre-Mesozoic events. The crust is a part of the Pannonian megablock, namely of its northern Balaton block. Coal deposits are associated with the thin/thick crust boundary in the Inner Carpathians; i.e. between two different crust segments.

d) Inner-Carpathian coal deposits ranged among intramontane basins on a typical continental crust are dynamically related to the Fatra—Tatry block margins or to deep margin structures, and to their crossings.

e) The role of the N-S Danube zone (Central-Carpathian lineament) in the coal deposits distribution has so far not been evaluated properly. It is the upper-crustal manifestation of a lineament activated during the Neogene and further evolving in the Quaternary. It has the character of a coal-bearing lineament like the Vardar-Kraishtide zone (F. ČECH — J. ZEMAN 1980).

f) With respect to deep structure the following factors were favourable for the formation of coal deposits: block margins with variable mobility (nearness of source area) tectonically differentiated paleogeographic pattern, slow subsidence mostly predisposed by thin crust unconsolidated by Paleozoic and Mesozoic events.

Basins formed along the Carpathian arc from W (oldest) to E (youngest), from the inside to the outside, but differentiated block movements also controlled the basins formation tendency in the reverse course. Mainly during the Pliocene, the subsidence got more intense in the zone around the Pannonian lowlands, and the mobility was most likely controlled by reducing tension in the mantle diapir-overlying crust. Coal-forming favourable conditions were mostly around more stable blocks near the diapir periphery, in a zone which was not involved in subsidence, not even in the extensive Pliocene subsidence.

Following were factors unfavourable for greater coal reserves: intense tectonic disturbance, rapid changes in positive and negative movements, poor tectonic segmentation of mobile (poorly consolidated) crust preventing differentiation of the area into favourable paleogeographic structures (F. ČECH 1980a).

### 4.10.1 Remarks on prospective areas

The above mentioned criteria may also be applied in prognosis but besides tectonic also paleogeographic factor must be considered. I shall only deal with criteria based on the analysis of relations between deposits and deep crustal structure, without the evaluation of paleogeography whose synthesis is still incomplete.

The area of the Foredeep and the Danube block among the internal units are least prospective. The N-S Central-Carpathian lineament delimitates the prospective area within the respective blocks to a belt between Levice and Krupina, and to a parallel belt between the rivers Nitra and Hron. Among other structures the Vienna Basin and the East-Slovakian Basin are the only prospective ones, mainly the zone along the peri-Pieninian lineament. In its eastern part it is the area around Kapušany, Vranov; in the western part — the zone between the Hornád fault and the Slánske vrchy hills deep fault. In most cases prospective areas are on block margins disturbed by uplifts and subsidences caused by movements on faults. No larger coal- deposits, only smaller occurrences may be expected there. More exact prognoses may only follow the evaluation of paleogeographic evolution. It is, however, necessary to consider tectonic data.

## 4.11 Oil and gas deposits in Carpathian-Balkan region and their relation to basin deep structure

There are many publications on oil and gas deposits in the Carpathian-Balkan regions but only few of them are synthetizing. Among them are works by G. N. DOLENKO (1962), V. V. GLUSHKO (1968), V. I. VYSOCKIJ a.o. V. V. SEMENOVICH et al. (1977) presented a survey of the deposits in a map and in a monograph (1981).

Since I deal with oil and gas deposits of the West Carpathians in a separate chapter, here I only present results of the study of deep structure to oil and gas occurrences, mainly in the adjacent parts of Poland, U.S.S.R., Romania, Hungary and marginal parts of Yugoslavia.

The oil and gas deposits distribution is irregular in the Carpathian system. The zones of occurrences are in accordance with the basical division of the Carpathian system, so we may approximately distinguish deposits

— in the Carpathian Platform foreland
— in the Carpathian Foredeep
— in the Inner Carpathians.

Geological zoning of oil-and gas-bearing areas done for the choice of proper methods of prospection and exploration represents one of the most complicated problems in oil geology.

I have taken over the division into oil- and gas-bearing basins and areas from the map by V. V. SEMENOVICH et al. (1977). The authors of the map distinguished nine oil-gas-bearing basins (Fig. 2). Most basins are divided into oil- and gas-bearing areas. In the basins are about 700 deposits of oil, gas, or mixed deposits. There also are a few deposits of gas condensates, mainly in the northern fore-Carpathian and fore-Carpathian-Balkan oil- and gas-bearing basins. Most deposits are in the fore-Carpathian-Balkan coal-bearing basin (214) and in the northern fore-Carpathian basin (116), i. e. in the Foredeep, in the Carpathian foreland, and in the Pannonian oil- and gas-bearing basin, i. e. in the Carpathian hinterland.

Almost all deposits known are in the depth to 3—3,5 km. Most of them are deposits small in size and reserves. Hydrocarbons mostly occur in structural types of traps. There are also combined traps — which does not exclude possible favourable conditions for non-structural traps.

The range of oil and gas occurrences is great: from the basement weathering crust (the area of Alföld-Banát-Bácska, the outer unfolded part of the Carpathian Foredeep with the overthrust Outer Carpathians on the Czechoslovak territory) to the sedimentary Neogene filling. Following are most productive sequences: Miocene (the Transylvanian, northern fore-Carpathian and Adriatic basins), Pliocene (the Moesian Platform, the Sáva-Dráva region), Cretaceous--Paleogene (the Carpathian oil- and gas-bearing basin, the area of Alföld--Banát-Bácska, Eocene (the inner folded part of the Carpathian Foredeep), Oligocene (the area of the Carpathian Foredeep in the fore-Carpathian-Balkan oil- and gas-bearing basin), Late Triassic—Late Cretaceous (the area of

Midmountains-Igal-Bükk Mts.) Middle-Triassic-Early Cretaceous (the area of the Moesian Platform). The origin and distribution of hydrocarbon deposits were controlled by the basement block structure in the foreland and intermontane depressions, and by fold- and nappe structure in orogenic zones. In these zones the deposits distribution was controlled by longitudinal and transversal deep faults.

### 4.11.1 Deposits on the platform foreland of the Carpathians

The origin of deposits on the platform was associated with movements of the Outer Carpathians and their displacement over the foreland. Hydrocarbon deposits are associated with platform structures inducting the structures in the sedimentary cover or deposits on structures resulting from vertical movements of the platform basement, and from the pressure of nappe fronts.

Among such areas also belongs the Lublin — Lvov step fold depression divided into two grabens (Lublin and Lvov). In the upper part of the basement and in Devonian-Carboniferous sequences are many NW-striking anticlinal zones. With such anticlinal structures, disturbed by local faults, three gas deposits are associated, namely: Minkowice, Komarów and Velikomostovskoe. The earth gas mostly consists of methane, the content of heavy hydrocarbons and nitrogen is low.

### 4.11.2 Deposits in northern fore-Carpathian oil- and gas-bearing basin

The basin extends on the Czechoslovak, Polish, Soviet and Romanian territories. It is bordered by the Bohemian Massif on the northwest. The northern border is conventionally placed on the distribution boundary of Neogene sediments filling the Carpathian Foredeep. On the NE the basin is bordered by the Rava Russkaya Zone of uplifted Kaledonides, on the SE — by the Bukovina transverse elevation, separating the basin from the fore-Carpathian-Balkan oil- and gas-bearing basin.

The northern fore- Carpathian oil- and gas-bearing basin corresponds to the northern Carpathian foredeep divided into the outer unfolded part and the inner folded part of the fore- Carpathian foredeep. The foredeep is completely covered by the marginal overthrust of the Carpathians on the Czechoslovak territory; it is on the most part covered on the Polish territory and slightly on the Russian territory.

According to tectonic position all hydrocarbon occurrences may be divided into the foredeep zone before nappe fronts and the foredeep zone beneath the overthrust nappes. The nappes served as protective cover and at the same time they caused a greater geostatic pressure on productive horizons. So they caused hydrocarbon migration on faults and fold limbs. Owing to that the hydrocarbons frequently escaped into intensely deformed nappe rocks. Their high permeability enabled degasification of bed complexes depletion or extinction of deposits. Migrated hydrocarbons are frequent in small structures or in small lithologic traps in nappe series or in the form of small natural escapes resulting in negative prospection for hydrocarbons in flysch. This is why also deposits preserved in autochthonous series are frequent but small. Because of the above mentioned facts, exploration is expensive. The conditions are more favourable in the East-Carpathian Foredeep with frequent oil and gas deposits (the area of Ivano—Frankovo).

In the northern fore-Carpathian basin they discovered 116 hydrocarbon deposits. More than a half of them are gas deposits. Most deposits are in Miocene molasse sediments and in Cretaceous Paleogene flysch sediments. Two areas are distinguished in the basin.

— a mostly gas-bearing area in the outer zone of the Carpathian foredeep partly covered by the Outer Carpathians,

— a mostly oil-bearing area in the inner zone of the Carpathian foredeep partly covered by the Outer Carpathians.

In the gas-bearing area of the outer zone the main productive horizons are represented by the Middle-Miocene molasse (mainly Sarmatian and Badenian sandstones) and underjacent terrigenous Late Cretaceous sediments, and Late Jurassic carbonate sediments. In the oil-bearing area of the inner zone most productive are Paleogene flysch deposits, Devonian and Carboniferous formations are less productive with their smaller gas deposits in the western part of the outer zone of the foredeep near the Upper Silesian Basin (for details see below). It is evident that conditions for the preservation of hydrocarbon deposits were unfavourable) there was a long emersion since the Late Carboniferous to the Jurassic, and in some cases to the Tertiary, in the foredeep area.

Following are largest gas deposits: Przemysl—Jaksmanice, Ugersko, Rudky, Bylcze—Wolica; largest oil deposits are Dolina, Boryslaw, Bytków—Babcze, Plawowice. Prevalent brachyanticlines are mostly disturbed by faults. Erosive-tectonic elevations of pre-Neogene formations are frequently associated with deposits in autochthonous Miocene sediments.

Deposits in the gas-bearing outer zone beneath overthrust nappes of the Outer Carpathians are screened by the overthrust Carpathian flysch nappes or by the folded Miocene of the inner zone. In the Boryslaw-Pokuly subzone most hydrocarbon deposits are in narrow elongated folds, usually reversed northeastwards. The subzone comprises productive horizons in two or three anticlinal reversed folds thrust over one another.

### 4.11.3 Deposits in Carpathian oil- and gas-bearing basin

The Carpathian oil- and gas-bearing basin extends on the Czechoslovak, Polish, Soviet and Romanian territories. On the north the basin contacts the northern fore-Carpathian oil- and gas-bearing basin, and its southern border on the Czechoslovak and Polish territories corresponds to the northern border of the Klippen Belt, and/or to the Peripieninian lineament, dividing the Carpathians into the Outer and the Inner Carpathians. The border on the Soviet and Romanian territory consists of the Transcarpathian deep fault. On the surface the zone of Marmarosh klippes and the crystalline Marmarosh Massif correspond to the Transcarpathian deep fault. The Outer Carpathians consist of a thick (above 6000 m) sequence of Cretaceous-Paleogene flysch sediments. The total thickness of the complex is estimated to 12—15 km, if the repeated burial with overthrust is considered. The flysch basement is not exactly known. In the central part of the Outer Carpathians the existence of a shallow platform depression filled with Devonian, Lower Carboniferous and perhaps Permian deposits resting Upon the Cadomian Caledonian basement, is presumed. Triassic and Jurassic sequences of the platform type are also ranged to the pre-flysch basement. The northeastern border of the Outer Carpathians (Poland, U.S.S.R.) is on folded Precambrian sequences forming the central anticlinoria of the ancient Galicia geosynclinal system.

On the Czechoslovak territory the pre-flysch fault-segmented basement of marginal NW zones consists of the Paleozoic and Mesozoic sequence of the platform type on SE slopes of the Bohemian Massif (for details see below).

In the Carpathian oil- and gas-bearing basin are 76 mostly oil deposits a half of which are well exploited. They are concentrated in Cretaceous-Paleogene flysch sequences mainly in the Silesian zone comprising largest deposits: Grabownica, Bobrka—Rogi—Ruwne, Kryg—Lipinky, Iwonicz—Zdróh. The area is characterized by smaller but plentiful oil deposits. Although hydrocarbons are in all flysch stratigraphic levels, their vertical distribution is extremely variable. The Ciezkowice and Istebnia sandstones are most favourable for hydrocarbon accumulations. The Menilite Group Eocene sandstones and the Albian Lhota sandstones are also favourable. The Oligocene Krosno sandstones are less important. In the Outer Carpathians the hydrocarbon deposits are usually in normal or asymmetric brachyanticlines or in recumbent folds disturbed by many faults. Traps of lithologic character are controlled by wedging out reservoir rocks in marginal parts of overfaults.

So the small hydrocarbon deposits mostly concentrate in the northern part of the Carpathian Arc. The deposits are less frequent on both wings, but they are larger so the eastern part is very rich in hydrocarbons. In is the richest part of the West Carpathians as regards oil. With respect to deep structure the greatest areal concentrations represented by the greatest total hydrocarbon reserves are typically situated on the contact between the West-Carpathian and East--Carpathian blocks, i. e. in a megastructure of an interblock character. In that place the platform got extensively broken and the so-called internal angle of the platform in the sense of N. S. SHATSKIJ (Z. ROTH 1962) formed there. In this area the Carpathian Foredeep expanded owing to the downthrown platform basement and formed thus favourable structural-deposit conditions for hydrocarbon accumulations. The foredeep expanded in the same position in the bending of the East Carpathians (oil deposits in the area of Ploiesti). So the interblock position controls the extent of fuels not only within the Inner--Carpathian basins but also in megastructures.

### 4.11.4 Fore-Carpathian-Balkan oil- and gas-bearing basin

The basin is situated between the East and South Carpathians on the north and the Balkan on the south. On the east the basin is bordered by the Dobrodgea Massif and the North-Bulgarian regional elevation, on the west by Banát. The basin is on the first place among European basins as regards the amount of exploited oil. Most deposits are in the northern part of the basin, mostly on the Romanian territory. On the south, in Bulgaria, only a few hydrocarbon deposits are known.

In the area of this basin are the Moesian Platform with the Precambrian basement, the marginal part of the ancient East-European platform (the Moldava Platform), Alpine elements (the Carpathians, the Forebalkan), and the Hercynian-folded system (the submerged prolonged part of the North Dobrodgea). A large part of the basin is occupied by the fore-Carpathian foredeep.

The economically interesting hydrocarbon accumulations are only associated with sequences of favourable vertical porosity. In the area of the fore-Carpathian foredeep the accumulations are associated with Paleogene and Neogene sequences; on the Moldavian Platform and in the North Dobrodgea — with Sarmatian, Badenian, less Pliocene, sequences; on the Moesian Platform — with Mesozoic and Tertiary sequences, and in the Forebalkan — only with Mesozoic sequences. There also is space and time zoning in the distribution of hydrocarbons with various phase state.

In the meridional part of the fore-Carpathian Foredeep the oil deposits are dominant. On the south, in the area of the Carpathian bending the oil- and gas-bearing deposits dominate over oil deposits. In the extended part of the foredeep the gas-and oil- and gas-bearing deposits prevail. It is explained by particularities in tectonic structure. Possible preservation of gas-bearing deposits is less probable in disturbed structures than the preservation of liquid hydrocarbons deposits.

In the eastern part of the Moesian Platform dominant are gas deposits associated with Pliocene, seldom Miocene sediments. In the central part the oil- and oil- and gas-deposits associated with Miocene and Cretaceous sediments prevail. In the western part are mostly oil deposits associated with Triassic and Jurassic sediments. So the gas- and oil-gas-bearing concentrations prevail in younger sequences — like in intermontane depressions.

In the oil- and gas-bearing area of the Carpathian foredeep are 117 hydrocarbon deposits, 42 of them are oil deposits, 51 oil- and gas deposits, and 24 gas deposits.

In the Paleogene flysch zone are 29 deposits, 20 of them are oil-bearing. Productive horizons are mostly Oligocene, partly Eocene. Reservoir rocks are mainly represented by sandstones of the Fusaru and Cliva formations (Oligocene). Some deposits even comprise four productive layers; mostly in the form of tectonically screened bedded domes.

There are 42 deposits concentrated in the Miocene-Pliocene subzone of the Neogene zone. Twenty of them are oil-bearing, eight are gas-bearing and the rest are oil- and gas-bearing deposits. Following are the largest ones: Moreni—Gura, Ocnita, Baikoi—Tintija—Floresti, Bodesti, a. o. Hydrocarbons are most frequent in Pliocene sediments. Levantinian, Dacian and Pontian sediments are less productive.

In the Getic depression the hydrocarbon deposits occur on tectonically intensely disturbed structures. There are 34 deposits known (20 gas deposits, 19 oil- and gas- deposits, 3 oil deposits). The Balteni and Ticleni deposits are largest. Meotian, Sarmatian, Karpathian and Ottnangian deposits are productive, Oligocene sediments show a lower productivity. Hydrocarbon deposits are scarce in the Pontian, Badenian, Eggengurgian and Eocene. The structures are layered domes, tectonically or stratigraphically screened.

There are 85 deposits on the Moesian Platform (27 oil-bearing, 36 oil- and gas-bearing, 21 gas-bearing, and 1 gas condensate deposit). Most deposits are in the northern and northeastern part of the Moesian Platform. There are only a few deposits in the southern part of the platform. Pliocene sediments comprise productive Dacian, Pontian and Meotian formations. They mostly contain gas deposits, only in the northeastern part are oil deposits. The structures are lithologically and stratigraphicaliy limited; layered domes are scarce.

There also are productive horizons in the Lower Cretaceous, in the Middle and Lower Jurassic. Productivity of the Paleozoic was only proved by one deposit (Bibesti—Bulbuceni) in the northwestern part of the Moesian Platform. Rock reservoirs are represented there by Givetian and Upper Devonian cavernous dolomites.

About 15 deposits are known in the area of the Moldavian platform and North Dobrodgea. On the Moldavian platform only Sarmatian sediments are gas-bearing. Oil- and gas-deposits in Dobrodgea are in the Sarmatian and Badenian, less in the Pliocene formations.

The Tchiren deposit is the only gas condensate deposit in the oil- and gas-bearing area of the Forebalkan.

### 4.11.5 Pannonian oil- and gas-bearing basin

The Pannonian oil- and gas-bearing basin is divided into five areas (Fig. 71): the Little Hungarian-Danube gas-bearing area, the Hungarian Midmountain Igal--Bükk, the Sáva—Dráva, the Alföld—Banát—Bácska oil-and gas-bearing areas, and the Transcarpathian, mostly gas-bearing area. Hydrocarbon occurrences are mostly associated with mobile zones left by subduction dynamics (G. HÁMOR 1984). The dynamics is, however, hypothetical, without geologic evidences.

The Alföld—Banát—Bácska oil- and gas-bearing area is largest and comprises most deposits. About 96 deposits were explored there. A half of them are on the Hungarian territory and the rest are in Yugoslavia and Romania. Most deposits are gas- and oil- and gas-bearing. The Pliocene formation is the most important oil- and gas-bearing formation. The Miocene is also productive in many deposits. In the northern part of the area is now explored the upper part of the Cretaceous-Paleogene flysch formation (the Hajduszoboszló deposit). In the central and southern parts of the area are hydrocarbon deposits in the Mesozoic (Palic, Novi Knezevac, Elemir) and in weathered basement rocks (Călacea, Turnu, Barias). The deposits Aldea, Kikinda, Velebit and Mokrin are largest.

In the Neogene the rock reservoirs are represented by sands and sandstones. Most structures are dome-formed, frequently tectonically or stratigraphically screened, and/or lithologically limited. Deposits in flysch, carbonate and in fissured metamorphosed rocks are in upper parts of the complexes forming the basement of Neogene sediments. The deposits have a massive character. The deposits contain both the high-gravity (960 kg . $m^{-3}$) and low-gravity paraffine-base oil sorts (815—830 kg . $m^{-3}$). The gas composition is extremely variable. In some deposits the $CO_2$ content in gas exceeds 50 percent.

In the Hungarian-Midmountain-Igal-Bükk area (36 deposits) the Paleogene and Neogene terrigenous sediments, and cavernous and fissure Mesozoic dolomites and limestones are productive. In the southwestern part of the area, in the Zalay district is the largest Nagylengyel deposit. It comprises oil-bearing Upper Triassic formations, Upper Cretaceous limestones and Badenian sandstones. Deposits in the Mesozoic carbonate formations are massive, the Paleogene and Neogene structures are layered. Mesozoic structures are tectonically screened; in the Neogene they are dome-formed.

Mesozoic deposits contain high-gravity oil (948 kg . $m^{-3}$) with 3.3 percent of S; in Neogene deposits is low-gravity oil (920—830 kg . $m^{-3}$) with a low S-content. In the central part of the area the hydrocarbon deposits are associated with Oligocene sandstones. There are 6 oil and gas deposits. The Mezőkeresztes deposit comprises productive Eocene sandstones and Triassic fissure dolomites. The deposits are on local elevations disturbed by faults. As a rule, they are tectonically screened. They are in the depth of 130—1 500 m. The low-gravity oil has a low S-content.

Thirty-five hydrocarbon deposits are known in the area of the Sáva and Dráva rivers. Most deposits are in Early Pliocene formations. Also Miocene formations are productive in the dome-like deposits Kloštar, Šumečani, Taran. They contain low-gravity oil with a small S-content in the depth 1 200—2 300 m.

Productive Miocene sediments are in the Little Hungarian-Danube gas-bearing area. The gas deposits are characterized by a high $CO_2$ content (for details see below).

Seven deposits are known in the Transcarpathian, mostly gas-bearing area on the Czechoslovak territory. Gas-bearing formations are represented by Upper Badenian and Lower Sarmatian sandstones in the depth 500—3 000 m (for details see below). Deep exploratory drilling to the depth 5 000—6 000 m in the East-Slovakian Basin (R. RUDINEC—B. LEŠKO 1984) revealed prospective deposit accumulation and favourable structural traps also in the Inner-Carpathian Paleogene adjacent to the Klippen Belt. The Inner-Carpathian Paleogene represents a new oil- and gas-bearing basin.

### 4.11.6 Gas deposits in Transylvanian Basin

All the known economically interesting gas accumulations in the Transylvanian Basin are associated with Badenian-Pliocene sedimentary complexes forming the so-called gas-bearing formation. Sixty-seven gas deposits are known in the Transylvanian Basin. Most of them are associated with brachyanticlinal-domal structures and salt domes of variable size and form. As a rule, there are more (10—20) or less (4—5) productive horizons, but there never are less than two. The average thickness of the productive horizons is 1—25 m, sometimes even 62—105 m — like in the deposit Craesti, Ulies a. o. Productive horizons are in the depth 200—400 m and 1 500—2 000 m. The Neogene formation is characterized by numerous sandy beds with the character of good reservoir rocks. Cap rocks are mostly represented by clayey rocks.

Gas in most deposits consists mostly of methane (80—90 percent) except the eastern part of the basin where its content decreases rapidly and the content of $CO_2$ and nitrogen increases. A great part of nitrogen, $CO_2$ and helium is associated with gases in Badenian, scarcely the Buglov sediments in the eastern part of the basin. It is likely that the deep faults represent migration ways for the above mentioned components. Besides gas deposits there are also smaller gas-condensates deposits in Badenian beds at the localities Sicai and Braid.

Genesis of gas is not exactly known. The structural position of deposits may be indicative of gas migration from deeper horizons, associated with the origin of traps. There is a lack of data on isotopic composition of gases and also other geochemical parameters are missing. They might explain the genesis of gas and reason two contradictory opinions:

— gas is in parent rocks,

— gas migrated from parent Oligocene rocks to Sarmatian and upper Badenian rock reservoirs (G. N. DOLENKO 1962).

According to the latest data the hydrocarbon accumulations of the so-called gas-bearing formation are syngenetic with the surrounding rocks. The rocks in the sedimentary filling of the basin show typical features of parent rocks (shales with Badenian radiolarians, Late-Sarmatian slaty marls with increased $C_{org}$.

The problem of possible existence of oil deposits in the Transylvanian Basin is not solved.

### 4.11.7 Relation between hydrocarbon deposits and block structure of Pannonian Basin

The Pannonian Basin is a complex fuel-deposit megastructure. The basement blocks structure pattern indicates the autonomous position of the megastructure, unconformable orogenic belts around the Pannonian Basin. At the end of the Tertiary the blocks integrated into the Pannonian megablock, newly segmented due to rejuvenation of ancient faults and the formation of N-S faults of the Danube and Tisza systems.

On the Miocene/Pliocene boundary the structural inversion was also followed by changed subsidence rate of peripheral basins and of the basin inside. The deep cause of the subsidence is in the evolution of Pannonian mantle diapir formed in the course of mass movements in asthenosphere from peripheral parts (rapid subsidence in basins during the Miocene) beneath the centre of the Pannonian crustal segment. There the movements stagnated. Pliocene collapse following the magmatic chambers depletion affected the distribution of maximum subsidence centres and block dynamics (Fig. 39, 69). U. DANK — J. BODZAY (1971) presented a different concept of the basin basement structure. These authors regard the fault lines as identic with overthrust boundaries of zonal orogenic units whose character and course are analogous to those of the Carpathian Arc. According to the two authors the basin structure is arcuate like the orogenic rim of the basin. This conception associates geologic units which cannot easily be compared to Carpathian tectogenic zones. The authors do not mention the relation of hydrocarbon deposits to the delimitated zones. A comparison of hydrocarbon occurrences to zones does not show any relation supporting the hypothetic scheme.

Most deposits are in the central part of the basin (Fig. 69). Least deposits are structurally associated with the Balaton block E of the Danube block. More deposits are concentrated close to the contact of the Balaton block with the Dinarides. The greatest areal concentration is on the Paleopannonian block E of the Danube fault, mainly in the area of the mobile Paleopannonian block s. s. If we regard the Danube block as the axis of the Pannonian Basin, then there are no hydrocarbon deposits around the axis. The areas devoid of hydrocarbon deposits correlate well with the areas of the thinnest crust and of the highest uprise of the Pannonian diapir. Also the heat flow is highest in these areas — as shown by data of L. STEGENA et al. (1975). With respect to hydrocarbon migration the area is unfavourable in relation to the areas of largest deposits accumulations.

Tectonic planes causing increasing porosity of rocks were favourable for migration. During the Pliocene inversion the Zagreb-Kulcs and the Rába lines blocked the passage of hydrocarbons into the stabilized Balaton block (with the dying out volcanism). Dynamic function of the Peripieninian lineament is not clear but the distribution of gas deposits shows that the hydrocarbon migration trhough the lineament was limited by shear movements and prevalent compression. In the case of the formation of favourable structural traps the lineament might also accumulate hydrocarbons (the hole Lubina-1). In the case of the Vienna Basin the lineament also might have functioned as a protective barrier against pressures from the inner side of the Carpathians. I regard the zone along the lineament as prognostic with respect to hydrocarbons occurrences at the depth exceeding 3 500 m.

The decompressed Pannonian Basin is favourable for preservation of hydrocarbons. Ancient blocks integrated into the Pannonian megablock differentiated partially the distribution of deposits in blocks with the most intense Miocene subsidence. So the subsidence might have caused squeeze-out of hydrocarbons from deeper levels and their preservation in shallower traps.

As regards crustal types, more deposits concentrate in blocks consolidated by pre Mesozoic events and regenerated by Mesozoic and Tertiary events (the regenerated platform type). Advanced consolidation limited intensity of tectonic segmentation.

So there is no apparent correlation between block structure and hydrocarbon deposits distribution. There only is a paleogeographically and tectonically conditioned regularity in hydrocarbon deposits absence in blocks with coal deposits. Exceptions are due to inversion movements of the respective blocks (Mátra, Bükk Mts.).

The hydrocarbon distribution was decisingly controlled by the mantle diapir, its dynamics and the respective geothermal field. Miocene subsidence of marginal depressions of the Pannonian Basin might have resulted in compression in deeper parts of the basin. There the crustal elevations margins might have thickened. In the Pannonian Basin migration and redistribution of hydrocarbon accumulations started in pre-Sarmatian sediments and structures. The collapse movement inversion caused change in the regional tension field. In marginal depressions decompression was associated with formation of horsts and grabens, and with migration and redistribution of hydrocarbon accumulations. In the Pannonian Basin, migration of hydrocarbons was controlled by Upper Pliocene and Quaternary subsidence and migration of hydrocarbons are likely to persist as late as the Recent. Variable subsidence rate above the diapir and its peripheral part controlled variable depth of productive levels and stratigraphic traps.

The distribution of gas deposits is indicative of their preferred concentration in the diapir limbs overlying crust and/or above its top peak. The gas deposits rather concentrate in the thick crust segments tending to elevation structures. More compression stress also cumulated at the diapir margin in thick crust, and decompression resulted in pressure affecting migration and increase of deposit pressure.

The heat flow whose density mostly increases in elliptical regions with thinned crust, was intensified in the thin crust.

### 4.11.8 Correlation with coal deposits distribution

The diapir uprise was favourable for subsidence stagnation in sedimentary basins and for the formation of coal deposits. Deposits mostly formed on a tectonically strongly semgented crust of the pre-Tertiary suboceanic type. A more intense sialization proceeding since the end of the Paleogene resulted in the continental crust accretion in the Balaton block. The origin of coal deposits from the period of sialization is analogous to the origin of Upper-Carboniferous Middle-Bohemian coal basins and to the historic-tectonic role in the crust evolution. The origin of hydrocarbon deposits is associated with the basin basement mobilization, most likely associated with crust basification.

This is why we miss the correlation of coal deposits with the pre- Neogene block structure, concerning hydrocarbon deposits (Fig. 69). They accumulated during the Pliocene inversion and restructuring of blocks. The deposits concentrate in large blocks of the Pannonian megablock, rapidly sinking after the inversion. Without respect to ancient block structure they subordinated to the new structure pattern mostly controlled by the Pannonian diapir dynamics during the Pliocene.

### 4.11.9 Conditions for oil- and gas occurrences in relation to deep structure

Investigations of other oil-bearing areas show that geothermic gradient is a significant phenomenon with respect to oil and gas occurrences. In the area of the Red Sea it was found that at the geothermic gradient 28 °C/km—50 °C/km oil is changing in gas in smaller depths (J. D. LOWELL et al. 1978). A high heat flow causes decreasing water content in sediments, causing their increased porosity. At the gradient 50 °C/km the porosity increases by even 50 percent in relation to porosity of rocks at the gradient of 30—33 °C/km. The temperature 120 °C (H. G. KLEMM 1978) is regarded as a critical temperature limit. Then the chemical reactions rate rapidly increaseas at the formation of hydrocarbons.

The high heat flow in the Pannonian Basin is unfavourable for the preservation of oil deposits in greater depths. The rate of chemical reactions at the geochemical gradient of 48—50 °C/km in the basin limitates depths of the origin of hydrocarbons to the depth of 1—2 km (according to H. D. KLEMME to 1100 m). The data also concern the Pannonian Basin with the gas deposits prevailing over oil deposits. A good quality oil occurrences range to the depth 1—1.3 km (localities Demjén, Battonya, Budafapuszta) with the asphalt content 0.2—0.34 percent and benzine with paraffine 62—73 percent. Lower quality oil deposits are in depth 1500—2500 m (Hahót, Nagylengyel) — G. KERTAI (1968). Owing to oscillating geothermic gradient in the Pannonian Basin, hydrocarbon deposits preserved in areas of lower temperature, and deposits are absent in areas of the gradient along the Danube river (up to 70 °C/km). The zones of thermal water ascents are unfavourable. The thermal conditions are indicative of favourable conditions in peripheral parts of the basin.

Hydrocarbon accumulations were unfavourably affected by variable tensions and tension fields. This is why hydrocarbon deposits mostly preserved in depression zones protected from contrast tensions. I think the highest tension is in the central part of the basin in the area of the thinnest crust. Among peripheral zones around mantle diapir the Vienna Basin has the most favourable tension, and for gas also the East Slovakian Basin. High temperatures prevented preservation of oil deposits there.

Faults oriented tangentially to the Pannonian diapir margin formed barriers to diapir expansion and to ascending heat flows. Destruction was also prevented by shear-compression conditions on the faults obstructing eventual hydrocarbons migration. In basins without basement compression no hydrocarbons migrated from deeper levels. So the basins are prospective for hydrocarbon deposits (the Vienna Basin). A. PERRODON (1980) regards the Vienna Basin as most prospective for hydrocarbon deposits in Central Europe.

The shear stress field around the Pannonian diapir might also have prevented hydrocarbons migration to the Carpathian hinterland — to SE in the Vienna Basin, to SW and SE in the East-Slovakian Basin. According to G. HÁMOR (1984) the generation of hydrocarbons is mainly associated with the Leithaian orogenic cycle proceeding in the Sarmatian and Pannonian. G. HÁMOR associates migration of hydrocarbons with the Rhodanian cycle in the Pontian.

On the basis of deep structure and dynamic role of the Pannonian mantle diapir I have made conclusions about the pre- Miocene role of the Klippen Belt in migration and accumulation of hydrocarbons, analogous to J. JANÁČEK'S (1976) opinion. In my conception the hydrocarbons accumulation is explained irespective of the time of their origin and I postulate mobilization of hydrocarbons from the deep Mesozoic or Paleozoic of the Vienna Basin at the temperature 250—300° during thermic mobilization kerogene. At the average thermal gradient 50 °C/km such temperatures may correspond to the depth 5—6 km. This depth is appropriate for prospection for hydrocarbons near the Klippen Belt. I also explain the highly productive structures in the Austrian part of the Vienna Basin by the positive role of the Peripieninian lineament on the Austrian territory, and by the W-E fault running from the Vienna Basin to Mistelbach, and by other W-E fault structures (see Chapter). The block was protected from deep tectonic-thermal processes and from orogenic pressure which was already missing on the Carpathian side. The N-S faults (of paleorift?) might have played the role of a drainage structure, as well as the Kúty Graben in the Pliocene. Structures of this course might also have had a screen function. The positive screen role was also played by elevations of the Malé Karpaty Mts. and the Sopron block. My conclusions are in accordance with F. CHMELÍK'S (1971) tectonic interpretation. JANÁČEK'S historical division of migration since the Middle Cretaceous to the Middle Miocene would in the Tertiary time be in accordance with supposed pressure and thermal effects of the Pannonian mantle diapir upon peripheral oil-generating parent beds in sedimentary fillings.

### 4.12 Oil- and gas- deposits in West-Carpathians

Hydrocarbon deposits in the West Carpathians. Largest deposits are in intermontane basins — the Vienna Basin and the East-Slovakian Basin.

#### 4.12.1 Deposits in Foredeep

Gas deposits are exploited in the Ostrava part of the Carpathian Foredeep. The deposits are associated with Badenian basal clastics and with structural elevations of the hidden Carboniferous mountain range. The deposits are in the foreland of the Beskydy Mts. nappes and beneath overthrust nappes. They are associated with Neogene morphological elevations without any relation to Alp-

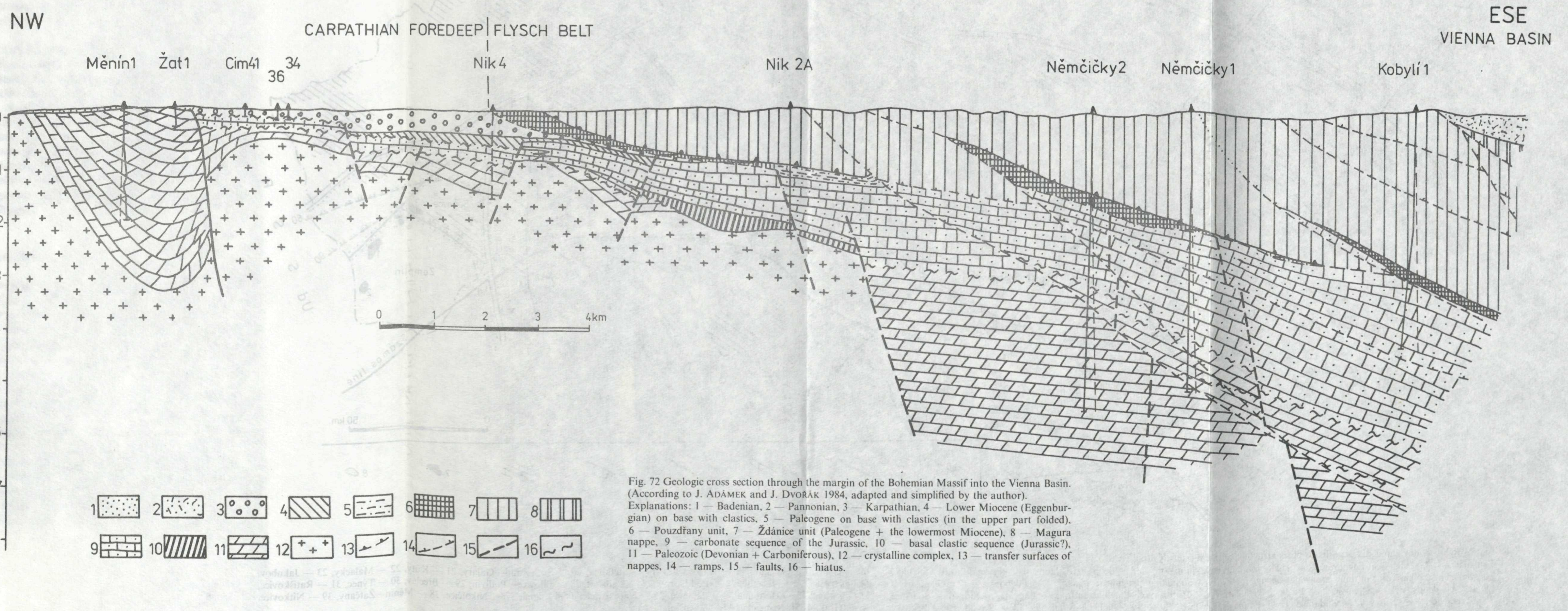


Fig. 72 Geologic cross section through the margin of the Bohemian Massif into the Vienna Basin. (According to J. ADÁMEK and J. DVOŘÁK 1984, adapted and simplified by the author).
Explanations: 1 — Badenian, 2 — Pannonian, 3 — Karpathian, 4 — Lower Miocene (Eggenburgian) on base with clastics, 5 — Paleogene on base with clastics (in the upper part folded), 6 — Pouzdřany unit, 7 — Ždánice unit (Paleogene + the lowermost Miocene), 8 — Magura nappe, 9 — carbonate sequence of the Jurassic, 10 — basal clastic sequence (Jurassic?), 11 — Paleozoic (Devonian + Carboniferous), 12 — crystalline complex, 13 — transfer surfaces of nappes, 14 — ramps, 15 — faults, 16 — hiatus.

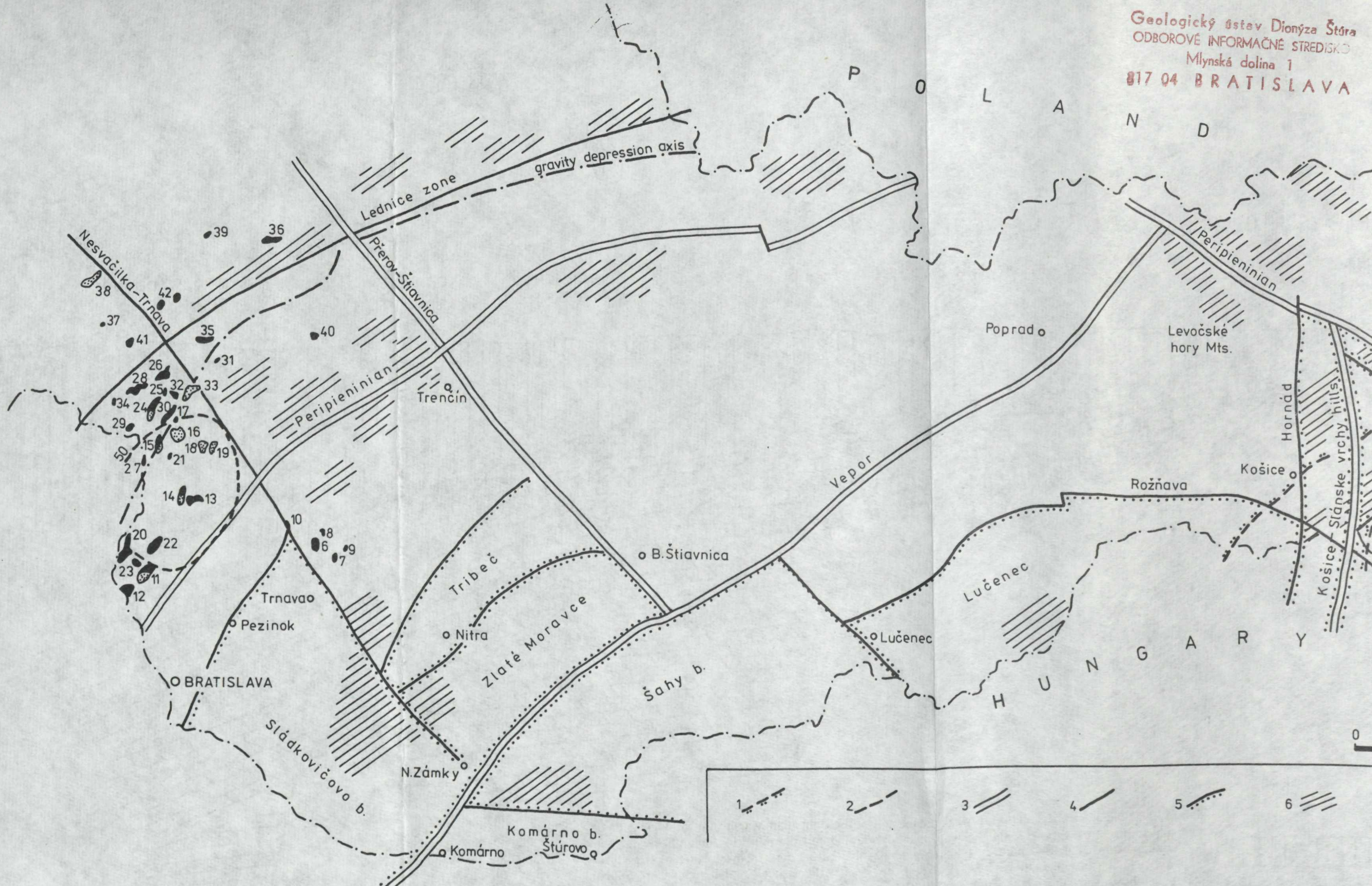


Fig. 74 Relation of oil and gas deposits to block structure of the Western Carpathians.
Explanations: 1 — high heat flow (in $mWm^{-2}$), 2 — low heat flow (in $mWm^{-2}$), 3 — deep-seated faults, 4 — regional faults, 5 — block boundaries in the basement of inner basins prospective for hydrocarbons, 6 — prognostic areas, 7 — gas deposits, 8 — oil deposits.
Oil and gas deposits: 1 — Stretava, 2 — Ptrukša, 3 — Trhovište—Pozdišovce, 4 — Lastomír, 5 — Bánovce n/Ond., 6 — Špačince, 7 — Trakovice, 8 — Nižná, 9 — Madunice, 10 — Krupá, 11 — Láb, 12 — Vysoká, 13 — Studienka, 14 — Závod, 15 — Brodské, 16 — Gbely, 17 — Cunín, 18 — Petrova Ves, 19 — Štefanov, 20 — Suchohrad—Gajary, 21 — Kúty, 22 — Malacky, 23 — Jakubov, 24 — Hrušky, 25 — Josefov, 26 — Poddvorov, 27 — Lanžhot, 28 — Bílovice—Podivín, 29 — Břeclav, 30 — Týnec, 31 — Ratiškovice, 32 — Lužice, 33 — Hodonín, 34 — Lednice, 35 — Vacenovice, 36 — Lubná, 37 — Nikolčice, 38 — Měnín—Žatčany, 39 — Nítkovice, 40 — Hluk, 41 — Němčičky, 42 — Žarošice.

ine fault tectonics or pre Alpine deeper structure. The genesis of gas is not exactly known sofar. Besides gas accumulations resulting from natural degasification of the coal-seams of coal — bearing Carboniferous deposits (Stonava, Mitrovice, Paskov, Příbor) there are smaller gas occurrences inside Badenian sediments in sandy or sandstone layers. They originate from Tertiary nearly parent rocks. Their generation was controlled by tectonic movements during the foredeep downsagging. Overthrust nappes had a screening function but they also supported escape of hydrocarbons into atmosphere through intensely disturbed rocks in shallower deposits. In other cases the gas escape was prevented by impermeable overlying clay rocks in the overlier of sandy horizons. The deposits were revealed during coal explorations, and the prospection is continued.

There is another prospective area in deeper levels of the foredeep beneath overthrust flysch nappes and in the foredeep basement. At present, autochthonous sediments are explored by deep drilling in the Slovak-Moravian and the Slovak-Silesian blocks (O. FUSÁN et al. 1971) NW and SE of the Lednice deep fault zone. According to unpublished results of geological researches most prospective are uplifted blocks still rising up — as shown by recent movements measurements. The data are not complete yet. It is likely that the distribution of deposits was also controlled by deeper block structure which was also copied in higher levels (Fig. 72). The deposits are at the localities Nikolčice, Měnín—Žatčany, Nitkovice, Lubna, Němčičky, Žarošice.

Other deposits in the Carpathian Foredeep in Poland were mentioned in the preceding chapter.

### 4.12.2 Vienna Basin

There are twenty-six oil- and gas-bearing deposits in the Czechoslovak part of the Vienna Basin. Both the Neogene sediments and the pre- Neogene basement of the Vienna Basin are productive. Neogene rock reservoirs are mostly in Badenian, Sarmatian and Pannonian terrigenous sediments, partly in facies — altered Eggenburgian, Ottnangian and Carpathian formations. Most Neogene deposits are associated with the Badenian Formation. The richest oil and gas deposits Hrušky, Závod, Vysoká pri Morave, Láb a. o. are associated with the so-called Láb horizon — an Upper Badenian complex of sandy sediments. Deposit traps are associated with fault zones and with local elevations of semi-brachyanticlinal type.

The basement exploration in this area was oriented on Mesozoic allochthonous units. Most positive data about economically interesting resources concern the locality Závod (P. OSTROLUCKÝ 1986). The hole Závod-72 revealed industrial hydrocarbon resources in the depth 4200 m in Late Triassic dolomites and at the base of the Neogene. Further holes Závod-73, 74, 76 and 77,78 in deeper levels of autochthonous nappes (4519—4645 m) revealed new productive horizons. The deepest Czechoslovak hole Šaštín-12 ranged to the depth 6505 m in the Mesozoic complex of the Vienna Basin. It is positive for gas from the depth about 6300 m. On the Czechoslovak territory the base of Mesozoic autochthonous sediments has so far not been reached. In Austria gas was revealed in the depth about 7500 m by the hole Zistersdorf 1 (A. KRÖLL et al. 1981; Fig. 73). Autochthonous Upper Cretaceous formations beneath Alpine nappes are also in our country regarded as very prospective. Their exploration will, however, be very expensive. At first sight the distribution of oil and gas deposits does not show any relation to deep structure. But Fig. 74 shows a relation to the deep fault structure. Most deposits are concentrated in a block between the Lednice deep fault zone, the Nesvačilka — Trnava deep fault and the Peripieninian lineament. In relation to coal deposits there is a negative correlation — hydrocarbon deposits are associated with the deeper downwarped block whose SW border is the Danube fault parallel to the Nesvačilka — Trnava NW-SE deep fault. The largest Austrian oil deposit Matzen and the largest gas deposit Vysoká — Zwerndorf are on this block. The deposits are adjacent to longitudinal faults in the basin, i. e. to the Láb — Šaštín and the Aderklaa faults. The two faults belong to the fault system parallel to the Peripieninian lineament.

The above mentioned block is the most productive one in the Vienna Basin. The downwarping of the block, rigidity of pre-Mesozoic basement slightly consolidated by pre- Mesozoic events, the screening properties of the Peripieninian lineament, and low-heat-flow density (about 50 $mWm^{-2}$ — according to V. ČERMÁK 1979) might prevent destruction of hydrocarbon deposits by pressure or temperature.

There only is a local heat-flow density increase on the Lednice deep fault (60 $mWm^{-2}$) around the towns Žarošice and Němčičky with gas occurrences, and SE of the Peripieninian lineament protecting the Vienna Basin from the effects of anomalous temperatures in the Pannonian region. Oil deposits occur in the area of low heat flow. The oil deposit near Gbely is highest with respect to structural position. It is on an elevated block NE of the Nesvačilka — Trnava fault. Interestingly, the axis of the maximum negative gravity anomaly in the area of the Vienna Basin is associated with concentrations of gas- and oil deposits. Then the anomaly may be interpreted in a different way like for example by Č. TOMEK (1979) who explains the source of the anomaly as the foreland more porous sediments which might under favourable conditions represent fluid- or gas-bearing rock reservoirs. In this sense also N-S and NE-SW paleorifts in the basin basement might be prospective.

The Vienna Basin and the area of prospection in the Carpathian Foredeep NE of the Vienna Basin are the only deposits accumulated on the external side of the Peripieninian lineament. The data on the relation of hydrocarbon deposits to block structure indicate that the Slovak-Moravian block NE of the Nesvačilka—Trnava deep fault is less prospective than that in the SW part of the Vienna Basin. The area rimming the Peripieninian lineament seems more prospective- as far as it has an elevation character and the lineament does not dip vertically beneath the Inner Carpathians.

### 4.12.3 East-Slovakian Basin

Gas deposits Stretava, Ptrukša, Trhovište—Pozdišovce, Lastomír and Bánovce nad Bebravou are associated with the East-Slovakian block SW of the Peripieninian lineament and/or to the area between the Slánske vrchy hills deep fault and the Számos deep fault. Mostly Late Badenian and Early Sarmatian sands and sandstones in the depth 500—3000 m are productive there. In the block structure the Zemplín block and its contact with the Sečovce block (Fig. 74) are most productive. The anomalous high heat flow in the blocks is highest in Slovakia. So there will be no oil deposits or larger gas accumulations in greater depths. With respect to deep structure the Zemplín block and the Sečovce block margins as well as the contact between the Sečovce block and the Košice depression are more prospective. Oil occurrences might also be expected in the northern part of the Košice depression but only with smaller hydrocarbon occurrences.

### 4.12.4 Correlation of basins to analogous structure

As I have already mentioned, the Vienna Basin and the East-Slovakian Basin are intermontane and interblock basins, and their features resemble the Transylvanian Basin. This basin is situated between sialic block and in its crust type it resembles the Vienna Basin. Also the crust of the Vienna Basin is rather thick whereas crustal properties of the East-Slovakian Basin are almost identic with those of the Pannonian Bain. Heat flow of the East-Slovakian Basin is also high. The East-Slovakian Basin seems less prospective with respect to areal concentration of hydrocarbon deposits in the Pannonian Basin. The Transylvanian and the East-Slovakian Basins with great deposit concentrations on a limited area are most prospective. The Pannonian and the East-Slovakian Basins are less prospective. They are disturbed by deep faults and show a variable crustal type.

### 4.12.5 Other Neogene basins and prospective structures

According to the existing data on the Neogene Basins the gas occurrences are only in the north-western part of the Danube Basin at the localities Špačince, Trakovice, Nižná, Madunice and Krupá. Gas accumulations are associated with the sediments on uplifted block NE of the Nesvačilka—Trnava deep fault, namely with the Trnava supracrustal depression. So far we have no positive data on other basins. But the $CO_2$, N or He outflows do not exclude deeper but smaller occurrences of further gas-bearing horizons. The basins structure lacks favourable elevation forms. Lithologic traps might form near faults.

In the Danube Basin is the Zlaté Moravce block and the SW submerged part of the Tribeč Mts. block near the Nesvačilka—Trnava deep fault. In the SW another fault parallel to this fault is running SW of the town Kolárovo up to

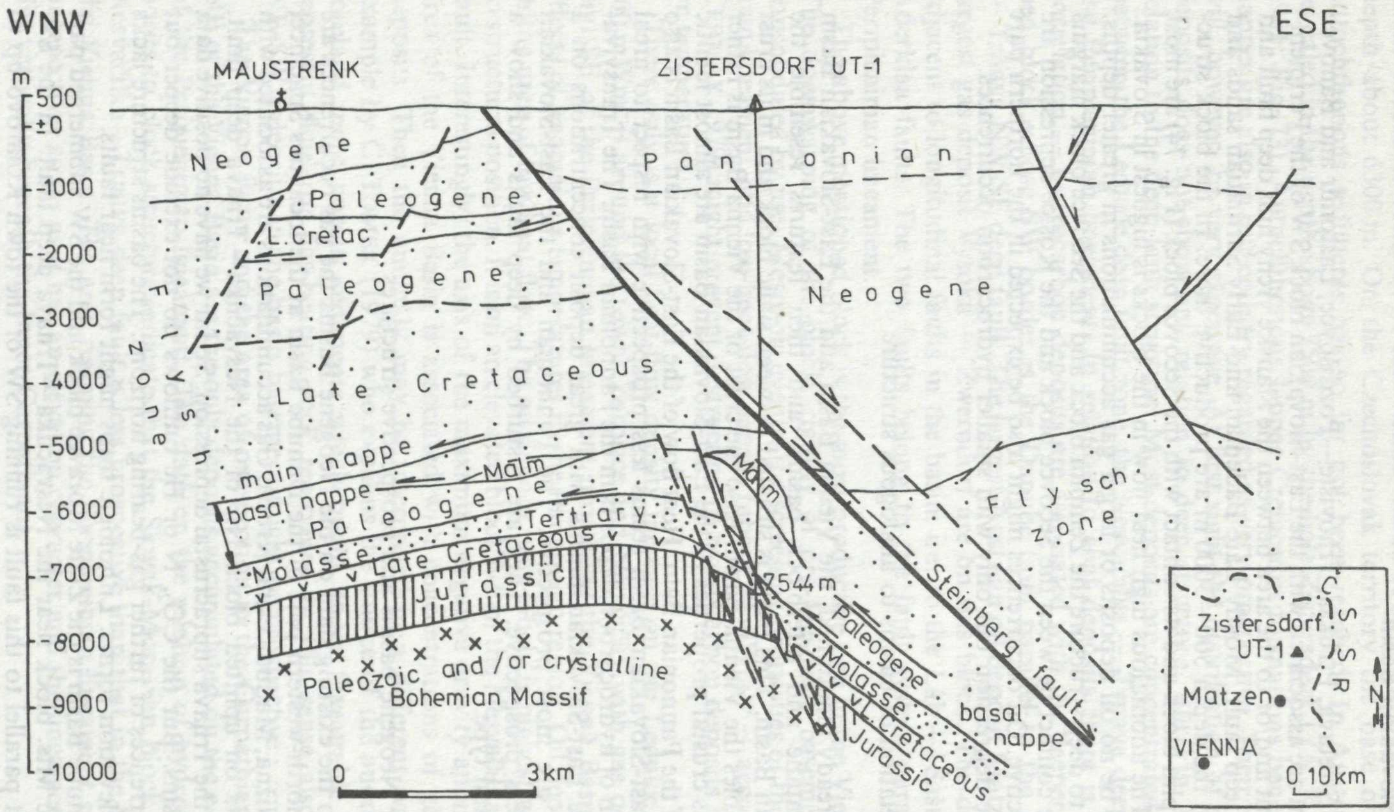


Fig. 73 Cross section of Zistersdorf UT-1 in Eastern Austria (After A. KRÖLL et al. 1981 — modified and supplemented by the author).

the town Pezinok. So far it is not sure whether the line is cutting the Peripieninian lineament and enters the Vienna Basin. In the area of towns Malacky and Veľké Leváre are faults possibly equivalent to the line representing the NW-SE fault system. The Danube Basin pre- Neogene basement gradually decreases towards the Danube river. If an analogous block were in the Vienna Basin then many gas- and oil deposits would occur there (for example Studienka, Závod, Brodské, Kúty a. o.). Because of certain analogy in productivity of the block NE of the Nesvačilka—Trnava deep fault also the block SW of the fault might be regarded as prospective. It is the central part of the Sládkovičovo block near the town Galanta and NE of the town Dunajská Streda. The subsidence migration from N to S migt have caused changing hydrocarbons migration trends and their displacement from the Gabčíkovo depression to shallower levels or elevations adjacent to the basin.

Small gas occurrences may also be in the Šahy block E of the town Nové Zámky (Fig. 74). But I do not range the Komárno block among prospective structures. I regard the Danube fault zone (the Inner — Carpathian lineament) with coal deposits as little prospective for hydrocarbon deposits. The Danube Basin with the intense heat flow comprises thermal water sources ranged among modern energy sources. It was the first area in Czechoslovakia with the exploration of the geothermal sources. Now they are being already locally exploited. The hole Diakovce-1 realized for the purposes of prospection for hydrocarbons as a source of geothermal energy was for the first time examined in 1958 (water temperature in the well orifice 41 °C). The hole Komárno M-1 realized for the same purpose in 1967 contained thermal water with the ascent temperature of 54 °C.

In 1971 the hole Dunajská Streda 1 was performed for agricultural purposes. The basin is prospective for further thermal springs.

In the East-Slovakian Basin the Košice depression and the Sečovce block, especially its contact with the Zemplín block are prospective with respect to deep block structure — deep hydrothermal ascents.

Prospectiveness of the area around the Peripieninian lineament with indications of structural elevations of the basement and favourable structural traps were revealed by deep holes, mainly Hanušovce 1 (6003 m deep; R. Rudinec — B. Leško 1984).

Block margins as zones of prospective structures are also indicated by further deep holes: Lipany, Smilno, Zboj. The parametric holes revealed the Inner-Carpathian Paleogene as a new oil- and gas-bearing province (R. Rudinec — B. Leško, l. c.). They confirmed thus my prognosis (F. Čech 1982) concerning fissure rock reservoirs in flysch and in it basement as well as structural traps at the flysch base.

It is not yet sure to what extent might the Számos deep fault be gas-bearing. It is SW of the Peripieninian lineament whose prospectiveness is still not exactly evaluated in the NE part of the Pannonian Basin. The Számos fault might have served as ascending way for neovolcanics which might have squeezed out eventual older hydrocarbon accumulations. The Számos line might have represented a favourable porous horizon and a communication way for migrating gas. Gas remained in depressions with less intense volcanism SW of the Peripieninian lineament. The gas might have migrated over the East-Slovakian Basin basement also to the southern part of the Košice depression (the most prospective area), and over the Slánske vrchy hills fault and the Hornád fault even farther southwestwards to the Levočské pohorie Mts.

There is the same relation between the size of structures and the amount of raw-material accumulations in Czechoslovak hydrocarbon deposits as with other sorts of raw-material. Large structures produce large deposits. Small-size basins and structures in the West-Carpathians are not prospective for large oil- and gas deposits. The Vienna Basin and the East-Slovakian Basin do not have the size of large depressions with great volumes of organic matter. According to H. D. KLEMME (1978) most productive are intracontinental structures comprising 75 percent of the world giant oil-gas deposits. When we range our interblock basins — the Vienna Basin and the East-Slovakian Basin among intracontinental basins, then — owing to their small size — they may be regarded as equivalent to giant deposits in our small-size geological units.

Both our basins lack regional elevation as a structural — lithological trap with concentrations of migrated hydrocarbons. Such structures control the origin of giant deposits up to 75 percent of the total basin reserves (H. D. KLEMME l. c.). When a gathering structure is missing, the deposits are dispersed, and this is the case of our basins. The Transcarpathian basin and the Transylvanian Basin have many elevation structures. Small structural-lithological traps concentrate maximum 10 percent of total basin reserves (H. D. KLEMME 1978). Basing on the empirically found relations, we may determine our prognoses by addition of difference between known reserves in individual horizons and production potential basin capacity (with subtraction of estimated migrated hydrocarbons). According to a new classification and productivity of individual basin types (H. D. KLEMME 1980) the lower structural level (autochthonous sediments) in the Vienna Basin should contain most traps and 3/4 of deposits should be in clastic sediments, 1/4 in limestones. If there were rift structures in the deep structure, the situation would be similar and clastic sediments might represent traps for deposits generated in Paleozoic or Early Mesozoic limestones. Prospects of the upper level of the intermontane basin type are better known and they are in accordance with productivity of basins of this type (Table 4).

There is a similar situation in the East-Slovakian Basin where most deposits are in clastic sediments. In both basins larger structural, lithostratigraphic and mostly combined traps may occur. In some features the South-Slovakian and the Danube Basin are similar to basins comprising large deposits (a great volume of organic matter, parent rocks) and large traps of all types but it is not proved by the existing data. Some features are positive in the Pannonian Basin.

## 5. General Conclusions

The problem of sedimentary basins genesis is being solved by many models based on crustal dynamics. Geologists solve the problem with respect to chemical, physical-chemical and physical changes in crust. In models explaining the basin origin by thermal or mechanical processes the physical changes are considered. In terms of geosynclinal or plate tectonic models the basin genesis is explained by tectonic processes.

At present the basins genesis is more interesting because of their importance for the origin of hydrocarbon-, coal- and nonmetallic raw material deposits. In the past decade several classifications in terms of plate tectonics were elaborated. The classifications are significant for their historical and dynamic approach to basin genesis. Hydrocarbons formed under paleogeographic and dynamic conditions different from those during migration and secondary hydrocarbon accumulations.

1. Subsidence is closely related to crustal type. Beneath basins with long and intense subsidence geophysical records revealed thinner crust than beneath surrounding (adjacent) geological units. This is typical of basins with orogenic rim — basins of intermontane type. Thinner crust usually has a reduced granite layer. In my book I present a schematic division of crust into a granite and a basalt layer, although I know that my division is not appropriate for a more complex crust composition as proved by the first superdeep holes in the Soviet Union (SG-3 and Saatly 1). In some basins with extensive subsidence the granite layer is missing.

The granite layer is either thin or absent. It is a granite layer of non-orogenic type. It may be ranged to primarily reduced granite layer.

2. Primary thinning of granite layer. In the crust of continents or along continental margins the relics of original basic oceanic crust are preserved, or the crust is discontinuously granitized. I denote this as simatic crust type. The terms suboceanic, paraoceanic, quasi-oceanic, melanocratic, femic or mafic crust may be used as synonyms s. l. The simatic crust type comprises variably metamorphosed and slightly migmatitized sediments or slightly granitized or non-granitized metasediments. Larger granite bodies of the orogenic pluton type are missing, and basic rocks with ultrabasic rocks of several generations represent 50 and more percent. In H. Stille's terminology the eugeosynclinal type of evolution and initial magmatism dominate.

3. The secondary thinning of the granite layer resulted from the alteration into a denser sinking crust, generally by basification (s. l.). Basification s. l., controlling the origin of subsidence, comprises eclogitization, deserpentinization, mantle diapirism, emplacement of mantle-asthenosphere asthenolith and/or invasion of hot asthenosphere into lithosphere resulting in its thinning. Asthenosphere may extend to crustal base and in the crust — as indicated e. g. by geophysical records from the Tyrrhenian Sea in the western Mediterranean Sea. Basification s. l. is a process tending to the change of continental crust into oceanic.

Basification ss. may proceed by basalt intrusions into crust (basaltization of crust), with sandwich structure resulting from repeated basalt sills, or by metasomatic alteration or by the formation of giant lava flows — traps. In the basalt layer basification may be represented by ultrabasic intrusions.

Basification processes are controlled by magmatism — a global mass migration from the Earth's deep to the surface. Basification models are accepted and also rejected. Recently they were accepted by plate tectonists because of genetical interpretations of some marine basins including the Mediterranean Sea basins.

In my opinion basification is a selective process proceeding preferably in zones and regions with relic oceanic and suboceanic crust primarily thinner than the adjacent typical continental crust. The thinner crust above mantle elevation as a region of an intense tension and weakening was easily invaded by mantle diapirs.

4. Mantle diapirism is evidenced global phenomenon of the transport of differentiated mantle matter present in the asthenosphere. Diapirism is also seismically indicated in the external core. The origin of basins is associated with diapirism.

Extinguished diapirism is in basins with anomalous low heat flow and residual thinned crust beneath the basin. Active diapirism is manifested by anomalous high heat flow. Mantle diapirism is closely related to asthenosphere elevation and thinned lithosphere.

The rapid subsidence mechanism is controlled by the collapse stage of cooling diapir with downsagging diapir top, by lateral expansion of the top part of diapir (associated with basification s.l.) or by downsagging of crust into hot diapir — asthenosphere elevation.

Mantle diapirism is associated with physical-chemical changes in crust i.e. crust enrichment with mantle derivatives, mostly basalts; with valume changes of thermoelastic nature and thermic changes resulting in secondary magma chambers. Fluids migration, and hydrocarbons migration out of parent rock into structural or lithologic-stratigraphic traps follows the endogenous processes.

5. Thermal changes may result in deserpentinization of serpentinized ultrabasics in basalt layer or in the mantle. I. A. REZANOV explains the increasing crustal density volume changes, followed subsidence and the genesis of basins and/or oceans by deserpentinization of serpentinized basic layers in the lower crust.

Serpentinization causing decrease of density leads to crust elevations. Deserpentinization has a reverse effect and might locally influence the formation simatic crust, both primary or secondary. Eclogitization of basic rocks — gabbroes may be controlled by thermal and pressure conditions. Eclogitization is spatially limited and cannot be ascribed a global significace — as done by E. V. ARTYUSHKOV (1981). Eclogitization may be accelerated by fluid, mainly water content. Then the ascent of juvenile water might be followed by serpentinization of ultrabasic rocks. It would prevent crust subsidence.

6. Diapirism is also associated with deep crust erosion and its thinning from below, or with surface denudation and erosion of upwarped crust whose thinning proceeds from the upper part. There is, however, no geologic evidence of the two models: a. the deep transport should then be followed by crust accretion around diapir, i. e. by thickening of the basalt layer due to the import of eroded mass — layer; b. the surface denudation should result in accumulation of thick complexes of coarse-clastic sediments around the denuded crustal elevation — but no such erosion products are known in the basin surroundings.

7. Mechanical concept of basin origin. Reduction of lithosphere associated with the mantle diapir uprise is at present frequently explained in the sense of D. P. Mc Kenzie (1978) by lithosphere expansion due to its horizontal movement in the hinterland of convergent contact of plates, or to the movements along transcurrent and transform faults. There, however, is no geological evidence, for example listric faults, paleosutures a. o. Subhorizontal crushed zones in the crust, revealed by the superdeep hole Kola SG 3 inspire the opinion that the mechanically-tectonically stratified lithosphere may be a product of strike-slip movements inside lithosphere over horizontal or subhorizontal dividing planes. This conception is discussed by geologists and geophysicists expecting its geophysical approval. Extension in the overriding plate in the subduction zone is used as explanation of the back-arc basin type genesis.

The opinions about the basin genesis being controlled by the Earth expansion are scarce. The expanding Earth hypothesis S is rejected. So is it with the Earth pulsation hypothesis, i. e. alternation of expansion- and contraction epochs in time and space.

8. In my book I prefer the model of mantle diapirism. Diapirs form and develop in the course of tectonic processes, before and after them. Diapirs are linear or isometric, they have variable size and duration. Crust type is important for the genesis of basins associated with diapirs. Diapirs invade more easily the simatic crust, i. e. the places of primary crustal thinning. Isometric or elliptical basins of non-rift character have linear marginal satelite basins — marginal depressions on their periphery. Rapid subsidence in marginal depressions preceded subsidence in the central basin above the mantle diapir. It is due to the rim syncline in mantle from which the products of mantle derivatives migrate into expanding diapir. Peripheral lithosphere sinks into the rim syncline-forming in the asthenosphere.

The accretion and expansion of diapir, particularly in the case of mushroom--formed diapir (e. g. the Alboranean Sea) are associated with subduction of lithosphere margin beneath the diapir periphery and with centripetal or asymmetric peripheral circumdiapir subduction. The cooler rigid mantle (lithosphere) subducts beneath the hot diapir and downwarps in the hot asthenosphere in the rim syncline.

Autonomous semiarcuate, so-called Benioff zones indicated in seismic focuses beneath the Tyrrhenian and the Aegean Seas reflect the contact between the uprising hot mantle and sinking cool lithosphere. Shear stress and fault zones tangential to basin margins form above diapir. So it is in active rifts, only the faults are parallel to the linear mantle elevation — rift margin.

9. Rifts are more frequent in lithosphere than presumed formerly. The rift or the initial riftoid stage is indicated geophysically in some basins without any sings of riftogenesis. The rifts indicate divergent plate margins and form generally on continent margins and inside continents (intracontinental or intraplate rifts). Aulacogens forming in the foreland of orogens are specific examples. Riftoid basins and rifts are also associated with asthenosphere elevations controlled by deep faults — lineaments, frequently much older than rifts.

10. Besides thinned simatic crust the basin generation is also predisposed by deep faults and their crossing. Both phenomena control the beginning or revival of subcrustal activity.

In plate tectonics the strike-slip faults are preferred in the explanation of the basin genesis — except divergent plate margins. According to plate tectonic models strike slips are enabled by transform and transcurrent faults extending far into continents. It is presumed that the origin of intracratonic basins is also controlled by paleolineaments and paleosutures. But the geological data on basin structure prove the decising role of normal faults existing on orogen margins, inside the orogens and in cratons.

11. Basin classifications are based on the geosynclinal model. In terms of the model, basins were classified according to the amount of sedimentary filling (compensated and uncompensated basins representing the oceanic type), according to crustal types (ensialic and ensimatic basins), and their position in geotectonic units (platform, marginal, geosynclinal and orogenic basins). Classifications were hard, mainly in the case of marginal depressions. The depressions were ranged among platforms and orogens, continent margins or ocean margins.

12. The plate tectonic models are associated with new classifications considering crustal type, position on plate margins or inside the plates, different plate margins dynamics. The classifications either have a general or economic-geological character. Basin classifications are mostly aimed at prospection and exploration of hydrocarbon deposits.

Classifications by A. W. BALLY — S. SNELSON (1980), D. M. CURTIS (1980, H. D. KLEMME (1980), A. PERRODON (1980), D. R. KINGSTON et al. (1983a, 1983b) and A. D. MIALL (1984) differ in the number of classification types, in complexity or simplicity of classification.

All classifications in terms of plate tectonics prefer mechanistic basin origin to thermoelastic and physical-chemical phenomena. Crustal changes and mobility persistence of simatic crust are mostly not considered. The models do not evaluate the role of normal faults and stress the lithosphere dynamics, i. e. horizontal movements.

I have tried to correlate basin types in terms of plate tectonics with the geosynclinal model. Some authors of basin classifications partly accepted the model and others rejected it as too fixistic. New models emphasize history and changing crust dynamics in basin evolution. It is a positive feature of classifications. There, hower, are many presumptions, speculations — like opening and closing of hypothetic fossil oceans. Only a thorough basin exploration and paleogeographic study may result in evidence of basin origin and evolution, and in acceptance or rejection of the respective sedimentary basins classification models.

13. Summary and evaluation of theoretic opinions about the basin genesis, regional geophysical data on deep structure of the Mediterranean and specially circummediterranean regions, the Carpathian Arc, the Pannonian Basin, and their correlation to geologic data on the structure and evolution of basins, the study of coal deposits, oil- and gas deposits resulted in new information about the relations between the deposits and deep structure of the basin underlying crust. The resulting prognostic criteria were complemented with regional prognoses. These are part of a complex of other criteria to be correlated with. So the prognoses are not decising — they only serve as a correction or complement of prognoses resulting from other methods.

The evaluation of data on deep structure was made on a new qualitative basis, i. e. on crust types evaluation according to petrographic composition and tectonics. In this sense the plate tectonics is highly progressive. Dynamic interpretations are still discussed and my dynamic conclusions result from a critical analysis of geological and geophysical data.

14. Like in other regions also in the Carpathians are ensialic and ensimatic basin types differing in evolution, areal extent and subsidence mobility.

a) Tertiary ensialic basins on continental crust consolidated by pre-Mesozoic and Paleoalpine events are genetically related with crust segmentation and differentiated mobilization in the course of advancing tectogenesis around basins, and especially with folding proceeding on the outer foreland of basin zones (the Outer Carpathians). The basins represent typical intramontane units with time-limited sedimentation duration and areal extent. They are associated with faults parallel to fold megastructures. Most extensive subsidence proceeded on the crossing or contact of these faults with transversal faults. With respect to block structure the basins are associated with block margins segmenting the formerly folded Inner-Carpathian units. According to T. BUDAY'S (1961) classification they belong among hereditary structure but with respect to crustal type and evolution stage — I range them among superimposed structures inheriting only the mobility of individual faults in narrow belts. The depositional process is an element superimposed upon the crust evolution stage but it does not terminate the evolution. Sedimentation in a mobile basin is an important process associated with this evolutionary stage.

b) Ensialic basins also comprise the Carpathian Foredeep with revived crust mobility. The crust was variably consolidated in different periods. Distribution of partial depressions was controlled by fault segmentation of the underlying platform and by its block structure. In the West-Carpathian Foredeep the NW-SE segmentation proceeded, in the East-Carpathian part the NE-SW segmentation course dominates. The faults were evidently revived by the pressure of the folded and overthrust Carpathians. In the northern part of the West-Carpathian Arc the W-E course was dominant. The South-Carpathian part of the foredeep was affected by the structures of the Moesian Depression (frequently denoted as platform) margins, so the foredeep evolution has some particular features differing it from other foredeep segments.

The Carpathian Foredeep is widest in the area of the contact of megablocks (plates): the East-European and the West-European Platforms, the East-European and the Moesian Platforms. Most hydrocarbon deposits concentrated there.

Intermontane basins have a great areal extent. They formed on heterogeneous crust where the poorly consolidated continental crust encloses relics of simatic crust, formerly perhaps oceanic. The crust tends to repeated subsidence, most intense in the Tertiary. The crust is thinner (about 24—30 km) than a typical continental crust. Before sedimentation the thickness ranged to 20—24 km and less. It was close to oceanic crust thickness. The crust was not affected by any extensive granitization. The basins may be ranged among inherited ensimatic basins, i. e. basins inheriting mobility associated with thinned crust and active crust-underlying mantle. The basin size is controlled by the size of simatic blocks. Ensimatic basins are larger than ensialic. Ensimatic basins are mostly on the periphery of the Inner Carpathians, inside the Carpathian Arc, and also in the Mediterranean and circumediterranean regions.

The division into superimposed and inherited basins with respect to crust type is different from Buday's division omitting the criterion of crust type.

Interblock basins on the contact of three megablocks (plates) have a specific position: the Vienna Basin is on the contact of the Bohemian Massif, the East Alps and the West Carpathians; the East-Slovakian Basin is on the contact of the West Carpathians, the East Carpathians and the Pannonian megablock. The position and boundaries of basins were controlled by the Peripieninian lineament and its transition or contact with megablocks. In the inner structure of both basins are many faults of different strikes, parallel to deep faults. The block segmentation was also affected by overthrust nappes. The deep basement of basins has a horst-graben structure, sometimes of rift character (the Vienna Basin).

15. The Pannonian Basin rests upon heterogeneous thin crust with block structure forming since the Late Paleozoic. At the end of the Tertiary the blocks integrated in one megablock — simatic median massif (F. Čech — J. Zeman 1980, 1982). Differences between sialic and simatic blocks were partly removed in the megablock, and the heredity of mobility in basic zones disappeared. The blocks controlled basin subsidence also in the Paleogene and proved the unaccomplished crust sialization. On the contact of both crust types in the Pannonian Basin (except the Flysch Belt) and along its margins, as well as along marginal depressions the Tertiary granite layer enlarged owing to neovolcanism. The crust accretion was associated with blocks integration. The process may be related to re-orientation of stress in crust above diapir. From the view the problem of crustal dynamics is to be solved at the lack of both direct and indirect evidence of hypothetic subduction in the Pannonian megastructure. Compression zones are replaced by dilatation. It is also proved by the rift character of depressions resembling the crust of unevolved ocean.

The block integration was associated with restructuring and new strikes of fault structures, mainly N-S.

The Panonian Basin underlying crust was variably consolidated owing to intense sialization. There the sialization was less intense than in crystalline elevations. Historical differences are also considered in the new block delimitation. The granite layer interpreted geophysically comprises various crust types and cannot be regarded as identic with typical continental crust. Formerly the suboceanic to oceanic crust was dominant. In the south and southwest it communicated with the same Dinaride and Mediterranean crusts.

In the deep structure the Pannonian mantle diapir was dominant. I think it extended horizontally up to the Peripieninian lineament. The diapir caused differentiated crust activation and affected it by different dynamics. The mobility was most intensive in zones (blocks) of thin suboceanic crust with possible rift character of basins. It is likely that diapirism commenced as early as the Late Cretaceous (banatites). Dynamic evolution may only be reconstructed roughly. The effects upon the outer sides of diapir are so far not exactly known, frequently contradictory to tectonic evolution and structure of basins in this region (intramontane basins). Extensive Pliocene subsidence is usually associated with diapir collapse preceding the cooling and resulting more-or-less from lithosphere downwarping into high thermal-asthenosphere elevation. (This opinion is supported by the existence of volcanism in most basins).

The collapse dynamics was modified by fossil diapirs. They also controlled the diapir differentiation into partial diapirs extending to the periphery of the central body. There are some indications of partial diapir beneath the East--Slovakian Basin. The role of diapirism beneath the Transylvanian Basin is not clear although a limited ancient diapirism cannot be excluded there like beneath the Moesian Platform. A zone of peripheral depressions around the Pannonian Basin was characterized by intense pre-Pliocene subsidence. I associate the origin of marginal depressions with the deepening of the rim syncline around the diapir due to the ductile mass runoff in the asthenosphere.

The Transylvanian Basin shows some features resemblant to those of marginal depressions. I regard the subcrustal erosion process applied on the explanation of thin crust beneath the Pannonian Basin as subsidiary or even excluded. Thinning or material changes of crust resulting in physical approximation of its lower part with the mantle have rather the character of basification s. l.

Mantle diapirism in the northern perimediterranean zone and in the Mediterranean region is probably more extensive, and its evolution and activity proceeded in various nonsynchronous phases.

16. The Peripieninian lineament is tangential-oriented to the Pannonian diapir periphery. It has the character of a shear zone downsagging beneath the diapir margin. The shear zone prevented horizontal expansion of diapir. Position, tectonophysical character, seismic activity of lineament and related parallel zones (faults above the diapir periphery) are in accordance with data on the relation of seismoactive shear zones to deep-seated elevations in analogous basins or rifts, and also correlate with model structures analogous to diapir. The

Sáva and Dráva lines are tangential-oriented to the SW part of diapir. In the northern part of the Pannonian Basin the shear and their role are replaced by segments of the Rába and Számos lines, most likely joined by the young Hurbanovo and Rožňava lines. Some Soviet geologists are thus inspired to a concept of the Peripannonian lineament joining the above mentioned deep faults.

17. Petrographic and geophysical data on the basement of the Danube, South-Slovakian and East-Slovakian-Transcarpathian Basins are indicative of poorly consolidated simatic crust tending to subsidence and to formation of paraoceanic paleodepressions — (meta) pelites with (meta) basic rocks. In the surroundings of the basins is indicated a typical granite layer with the elevation tendency. Transitions of the two crust types are associated with reduction of sediment thickness in basins.

The West-Carpathian Basins on the Czechoslovak territory are associated with faults and blocks accomplished in the Badenian. They are associated with thin denser crust characterized by an extensive positive gravitational field. Basins situated on the Pannonian diapir periphery are in a close genetic relation to the gravitational field. The basement of the basins has block structure. The interblock East-Slovakian and Vienna Basins show specific features, namely they are complex fuel deposit basins comprising coal and hydrocarbons deposits.

In its deep structure the Vienna Basin has some elements in common with the Danube Basin. The basins are segmented by the Nesvačilka—Trnava fault into NE stable and SW mobile blocks. The basin basement is divided by the Peripieninian lineament, joined with an elevation zone, prospective in fuel deposits. So the two genetically and positionally different basins are integrated by deep segmentation, probably of Badenian age.

I associate the origin of basins with endogenous subcrustal diapir processes. The influence of sediment load upon increasing mobility is denied by the fact that in some intramontane West-Carpathian basins whose filling also consists of dense rocks of downwarped nappes any increase of subsidence rate followed the loading of the crust.

18. T. BUDAY'S (1961) opinions about a relationship between basin genesis and folding or tension periods between folding phases are complemented by data

— on the relation of basins to crust type and mobility,

— on the fact that intermontane basins also evolved in the course of folding of the Outer Carpathians,

— on the fact that deep faults are not running beneath basin axes but they limit the area of basins and form their basin margins in accordance with block margins.

Inner differentiation of basins was controlled by faults of supracrustal character.

Basin migration proceeded from W to E on the Pannonian diapir periphery, in a reverse course in relation to folding,

— from the periphery to the inside of the Pannonian Basin (since the Paleogene to the Pliocene).

19. Coal deposits in the Carpathians and in the Pannonian Basin correlate well with the block structure. Most deposits concentrate on margins of opposite-moving blocks. Coal deposits form four coal-bearing belts on the internal side of the Carpathians. The Vardar-Kraishtide coal-bearing lineament is a specific newly defined structure. Forty-seven economically significant coal deposits formed in the course of 200—230 Ma, concentrated on the lineament. The Danube N-S fault zone as a part of the Inner-Carpathian lineament shows similar features. The Balaton block is most productive because there are the greatest coal reserves of the Pannonian Basin.

The simatic crust was related to optimal evolution of deposits under favourable (nonmarine) paleogeographic conditions. The ophiolite zone of the Dinarides and the simatic Balaton block are typical examples.

There are two ways of formation of coal deposits with respect to fault segmentation, namely:

— poor segmentation enables formation of extensive coal seams, mainly on the platform in the Carpathian foreland,

— intense segmentation, mainly in places of alternating elevations and depressions, supports formation of more coal seams of a smaller areal extent (coal-bearing lineaments).

There are fewest coal deposits in the foredeep, mainly in the West Carpathians. There are more coal deposits in the Inner Carpathians and their intermontane basins.

In the Pannonian Basin most coal deposits are in peripheral parts controlled by peripheral faults tangential-oriented to the Pannonian diapir. The faults of the Apuseni Mts. and margins of sialic massifs, mainly the northern part of the Vardar-Kraishtide zone are also ranged among productive structures. According to the distribution of endogenous ore deposits the Vardar-Kraishtide zone may be denoted as a complex coal-bearing lineament. The Inner-Carpathian lineament (the Danube zone) in the West Carpathians shows analogous features.

There were different conditions in the Mecsek Mts. The analysis of the coal-bearing layers evolution proves the existence of sialic crust bordered by suboceanic crust on the north. There possibly is a relation between the sialic crust and the Serbian-Macedonian Massif bordered by the Muresh fault on the north.

Coalification is differentiated. Thermal effects of the Pannonian diapir are evidenced in the higher coal metamorphosis. Differences and weakness of Pliocene deposits coalification may be explained by decreasing heat flow intensity or better by isolation properties of Pliocene sediments.

20. Oil- and gas deposits do not correlate with block structure of the Pannonian Basin basement.

Their distribution is controlled by restructuring resulting in migration of hydrocarbons. The distribution of hydrocarbon deposits correlates with a thicker crust and a lower heat flow. A high heat flow caused oil destruction. Deep faults, tangential-oriented to the periphery of the Pannonian Basin and of mantle diapir affected positively the hydrocarbon accumulations. They accumulated in marginal depressions near the faults. Depression margins on the contact with the elevation zone near the faults are prospective. In our country the eastern margins of the Vienna Basin and the autochthonous basement along the Peripieninian lineament are prospective. Basing on the association of deposits with elevated blocks near deep faults, I think that geophysically revealed structures along the Lednice deep-seated zone may be prospective for hydrocarbon occurrences.

It may also concern faults along the Peripieninian lineament in the East-Slovakian region. Conclusions based on the analysis of deep structure are usually in agreement with prognostic results of research and exploration, based on a greater complex of prognostic criteria.

Most hydrocarbon deposits concentrate in the foredeep and in the Carpathian foreland, mainly on the contact between the West and East-European Platforms in the foredeep bsement. The areas joined in depth with interblock basins are most productive if there are favourable elevation traps. Basins of this type seem most prospective for new (small) hydrocarbon accumulations.

On the Czechoslovak territory most prospective are the complex deposit structures — the Vienna Basin and the East-Slovakian Basin, as regards their position. Coal deposits distribution indicates a negative correlation.

The analysis of basinal deep structure shows that the largest basins are on boundaries of blocks and megablocks. The tectonic position also indicates their prospects because 60 percent of the world hydrocarbon reserves are in basins along lithospheric plate margins.

These conclusions show the importance of the study of basinal deep structure and of the relation of basins to block structure. The results of the study prove that practically all economically interesting hydrocarbon reserves in Czechoslovakia are in basins on blocks or megablock margins. So the structural boundaries are most favourable for further prospection for hydrocarbons in Czechoslovakia.

The prospective units and structures are, however, small. So we cannot expect discoveries of any giant hydrocarbon deposits. Some deep faults, mainly the Peripieninian lineament and the NW extension of the Számos line were denoted as prospective.

In Neogene Basins of South Slovakia are some structures suitable for formation of underground storages and there are important sources of geothermal energy which already is partly exploited in the Danube Basin.

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František Čech

# Dynamika neogénnych karpatských panví a ich vzťah k hlbinnej stavbe, typom kôry a ložiskám palív

Resumé anglického textu

Publikácia je rozdelená do štyroch častí. Prvé dve sa zaoberajú všeobecnou a regionálnogeologickou problematikou panví v karpatsko-balkánskej oblasti. Tretia časť sa týka hlbinnej stavby neogénnych panví vnútorných Západných Karpát a viedenskej panvy, štvrtá časť sa zaoberá vzťahom ložísk uhlia a prírodných uhľovodíkov k hlbinnej stavbe panví v karpatsko-balkánskej oblasti, so zameraním na ložiskové štruktúry v čsl. časti Karpát.

V poslednom desaťročí sa nazhromaždilo veľké množstvo geologických a geofyzikálnych údajov, vďaka intenzívne sa rozvíjajúcemu vyhľadávaniu a prieskumu ložísk uhľovodíkov a sčasti i ložísk uhlia. Tento rozvoj bol vyvolaný svetovou ropnou krízou, snahou štátov hľadať vlastné zdroje energie, ale tiež snahou kapitalistických monopolov nájsť nové zdroje ložísk uhľovodíkov v rozvojových krajinách a vytvoriť ich hospodársko-politickú závislosť na kapitalistickej ekonomike. V zákulisí týchto snáh je i tendencia obmedziť intenzívnu exploatáciu vlastných zdrojov a chrániť si svoje zásoby na budúce roky.

Postup prieskumných prác do hĺbky vyvoláva nutnosť preskúmať hlbokú stavbu panví a zároveň zodpovedať teoretické otázky príčin a podmienok ich vzniku. Novým impulzom v tejto problematike bola aplikácia modelov doskovej tektoniky. Vyskytli sa nové hypotézy o dynamike kôry a zakladaní panví. Odklon od fixizmu v geológii vyvolal potrebu revidovať klasické fixistické modely a tieto relativisticky aktualizovať (napr. R. W. van Bemmelen 1972). Na druhej strane pri mobilistických modeloch práve problém vzniku sedimentárnych panví viedol k hlbšiemu prepracovaniu pôvodne jednoduchých tektonických modelov a k hľadaniu spôsobov ako riešiť problémy dosiaľ vysvetľované z pozície fixizmu. Pri riešení genézy panví však bolo treba mobilistické modely modifikovať a tiež relativisticky upravovať, napr. akceptovaním existencie bazifikácie s. l. V súčasnosti trvá ešte diskusia, či možno geosynklinálny model korigovať, alebo či treba od neho upustiť.

## Procesy podmieňujúce vznik kontinentálnych panví

V prvej časti knihy sú uvedené a hodnotené najrozšírenejšie predstavy o dynamike kontinentálnej kôry. Problematika je zameraná na neoceánske panvy. Panvy pri okrajoch oceánov sú spomenuté len orientačne vzhľadom na možnosť, žeby niekto chcel riešiť niektoré dnes kontinentálne panvy v historickej retrospektíve ako panvy na okraji oceánov.

Vo vývoji a dynamike kôry má rozhodujúci vplyv magmatizmus, ktorý pohybom hmôt vyvoláva nerovnovážne stavy v kôre, mení ich chemické zloženie a mernú hmotnosť. Veľmi diskutovanou otázkou je existencia a účinky bazifikácie.

Bazifikácia s.l. je obecným procesom objemových a látkových zmien, prebiehajúcim v oceánskej i kontinentálnej litosfére, keď sa má ľahšia sialická kôra premeniť na ťažšiu kôru bázickú. Názory o globálnej bazifikácii sa zdajú byť nedoložené. Najčastejšie modely hľadali podstatu bazifikácie v metasomatických zmenách (príklad bazifikácie s. s.). Iné v naplňovaní kôry intrúziami a efúziami bázik — predovšetkým čadičov (ložné žily, lakolity, diapíry, trapové výlevy). R. W. VAN BEMMELEN (1972) rozlišoval atlantický, viac-menej mechanický typ bazifikácie (výlevy čadičov) a mediteránny, metasomatický typ bazifikácie. H. RAMBERG (1967) vysvetľoval bazifikáciu mohutnými výlevmi čadičov, ktoré spôsobili v kôre hustotnú inverziu. Ľahšie granitové horniny odtekali spod ťažkých bázik a diapiricky prenikali k povrchu. Hlavným súčasným predstaviteľom názorov o globálnej bazifikácii je V. V. BELOUSOV (1982), ktorý predložil nový variant modelu. Výlevmi a intrúziami bázik sa zvýšila merná hmotnosť kontinentálnej kôry a v dôsledku toho začala táto kôra klesať do horúcej astenosféry, ktorá mala nižšiu viskozitu než bazifikovaná kôra. V mieste bazifikovanej kôry má vznikať oceánska depresia.

V astenosfére sa predpokladá asi 15 %-ný podiel fluidnej fázy, ktorá je hustotne ľahšia o $0{,}1 \, . \, 10^3 \, kg \, . \, m^{-3}$ v porovnaní s rigidnými ultrabázikami plášťa. Fluidná fáza umožňuje odtekanie horninových más jednak z vrcholu astenosféry smerom k depresiám a jednak umožňuje pohyb ultrabázik v tekutom stave po hlbinných zlomoch do vyššej časti kôry. Takto sa laterálne rozširuje areál bazifikácie.

Presuny hmôt v astenosfére asi existujú. Dochádza pritom k uvoľňovaniu fluíd a vplyvom ich výstupu k narastaniu teploty v litosfére. Rozdielne teploty podmieňujú i diferencovanie viskozity a objemovej hmotnosti v litosfére. V areáloch vysokého tepelného toku sú geofyzikálne indikované elevácie astenosféry a plášťa pod stenčenou kôrou so subsidujúcou panvou. Pri ochladzovaní dochádza podľa E. V. ARTJUŠKOVA a M. A. BEERA (1983) k termoelastickej kontrakcii kôry, čo má byť jednou z príčin klesania v kôre. Aj A. E. SVJATLOVSKIJ (1984) predpokladá po skončení vulkanizmu v oceánoch kolapsový efekt, prehlbovanie oceánov a regresiu morí z okrajov kontinentov. SUZUKI YASUMOTO (1984) označil panvy na vnútornej strane vulkanických oblúkov ako kolapsové subsidenčné panvy.

S kolapsom chladnúceho plášťového diapíru je dávaná do súvislosti rýchla subsidencia panví na stenčenej kôre, napr. pliocénne klesanie v panónskej panve (D. VASS 1979).

Objemové a hustotné zmeny v bazifikovanej spodnej litosfére predpokladá model serpentinizácie a deserpentinizácie (H. H. HESS 1962, modifikácia I. A. REZANOVA 1977). I. A. REZANOV (l. c.) predpokladá, že Mohorovičičova diskontinuita je oscilujúcou fázovou hranicou serpentinizácie ultrabázik plášťa: klesanie Moho sa vysvetľuje serpentinizáciou zapríčiňujúcou dvíhanie jednotiek a naopak, elevácie Moho sú vysvetľované deserpentinizáciou, ktorá je príčinou vzniku depresií. Procesy serpentinizácie a deserpentinizácie môžu mať lokálny význam, nie však globálny (cf. V. V. BELOUSOV 1982).

Rozdielne zloženie kôry sa snažil vysvetliť metamorfnou diferenciáciou I. A. REZANOV (1981). Diferenciáciou sa oddeľujú bázické a kyslé frakcie v kôre, ktorá má spočiatku charakter pásikavého megamigmatitu. Kyslé diferenciáty postupne diapiricky stúpajú do vrchnej kôry a vytvárajú granitovú vrstvu, bázické zložky amfibolitového typu ostávajú ako restity po diferenciácii v spodnej kôre — bazaltovej vrstve. Metamorfná diferenciácia je druhom bazifikácie prebiehajúcej pri progresívnom vývoji kontinentálnej kôry. (Pozn. Jednoduché schematické označenie „granitová a bazaltová vrstva" používam vzhľadom na vžité modelové označenie. Pritom treba mať na pamäti, že zloženie oboch „vrstiev" je variabilnejšie a zložitejšie. Ani názvy granitovo-rulová a granitovo-bázitová vrstva asi plne nevystihujú skutočné zloženie kôry).

Model I. A. REZANOVA (1981) bol vypracovaný na základe výsledkov vrtu Kola SG-3, ktorým sa v hĺbke okolo 10 000 m zistil komplex svetlých granitoidných polôh v laminovaných amfibolitoch, typu sendvičovej textúry. Zatiaľ sa týmto najhlbším vrtom na svete zistili vo vrchnej kôre (granitovej vrstve) metasedimenty a metavulkanity, na archaickej kôre tiež s prevahou bázik (obr. 1). Z profilu vrtu by bolo možné usudzovať, že ide o perzistenciu kôry, ktorú by bolo možné označiť názvom simatická kôra.

Do sféry bazifikácie s. l. patrí i model eklogitizácie gabrových hornín. Proces má prebiehať v spodnej kôre pri zvýšení mernej hmotnosti o 0,5—0,6 . $10^3$ kg . $m^{-3}$. Model pôvodne obhajoval V. V. BELOUSOV, ale v súčasnosti ho opustil (V. V. BELOUSOV 1982). Model inovoval E. V. ARTJUŠKOV (1979), ktorý vznik panví vysvetľuje eklogitizáciou bázik v bazaltovej vrstve. Eklogitizované bloky kôry klesajú do pomerne ľahšieho plášťa (3,35 . $10^3$ kg . $m^{-3}$, na rozdiel od eklogitov s mernou hmotnosťou 3,5—3,6 . $10^3$ kg . $m^{-3}$). E. V. ARTJUŠKOV (l. c.) predpokladá, že za prítomnosti fluíd, najmä vody, sa proces urýchľuje a na eklogitizáciu stačí teplota okolo 500—600 °C a tlak 1 000—1 500 MPa. Bez prítomnosti fluíd by proces mohol nastať až pri teplotách 1 000—1 200 °C. Fázová premena gabra na eklogit môže nastať pri výstupe horúceho plášťa do chladnej kôry, alebo ochladením prv silne prehriatej (až natavenej) litosféry. Eklogitizácia by mala viesť k „erózii" — stenčovaniu bazaltovej vrstvy (napr. pod panónskou panvou), pretože odtrhnuté ťažké bloky eklogitizovanej litosféry sa ponoria do astenosféry. Preto čím väčšia bola hrúbka bazaltovej vrstvy, tým väčšiu časť profilu kôry mohla postihnúť eklogitizácia a tým väčšia bola i eklogitizáciou vyvolaná subsidencia v panve. Artjuškov vylučuje modely bazifikácie Belousova, Rezanova a i.

Eklogitizácia môže byť procesom s regionálne obmedzenými účinkami na výstup horúcich plášťových diapírov, napr. pod Tyrrhenským morom, azda aj pod panónskou panvou. Mohla by urýchľovať klesanie blokov litosféry do menej viskóznej astenosféry vystupujúcej v elevácii pod stenčenou litosférou. Celkom nepravdepodobný je globálny rozsah eklogitizácie.

Proces stenčovania kôry — litosféry sa okrem bazifikácie hľadá v hlbinnej a povrchovej erózii. Prvú hypotézu vyslovil J. GILLULY (1955); domnieval sa, že hlbinnou eróziou koreňov hôr ľahších než okolitý plášť a denudáciou hôr sa vytvorila stenčená kontinentálna kôra. Táto poklesla a na nej sa vytvorila panva. Vznik panónskej panvy hlbinnou eróziou kôry pod vplyvom vystupujúceho plášťového diapíru vysvetľoval L. STEGENA et al. (1975). Vznik Čierneho mora vysvetľoval R. BRINKMANN (1974) denudáciou rozsiahlej orogénnej elevácie kôry nad vystupujúcim plášťovým diapírom. Zvyšná stenčená kôra s odstránenou granitovou vrstvou začala ako ťažšia klesať. V modeli povrchovej denudácie veľkej elevácie kôry sa uvažuje o usadení horninového materiálu v okolí denudovanej elevácie. V okolí panví so stenčenou kôrou však chýbajú komplexy klastických hornín a neexistujú iné doklady pre skôr existujúce elevácie s výškou 2—3 km, prípadne i viac, ak by malo dôjsť k denudácii granitovej vrstvy.

V posledných rokoch sa akceptuje model roztiahnutia litosféry (D. P. MC KENZIE 1978), vypracovaný pôvodne pre vznik Egejského mora. D. P. MC KENZIE (l. c.) predpokladá, že vrchná časť litosféry je rigidná a správa sa ako krehká, zatiaľ čo spodná časť kôry je tvárlivá, duktilná. Táto časť litosféry sa rozťahovaním má stenčovať, zatiaľ čo vrchná časť sa zlomove trieští a jednotlivé bloky klesajú. Do litosféry stenčenej roztiahnutím môžu prenikať plášťové diapíry. Vysoké prehriatie má ďalej urýchľovať „roztekanie" duktilnej litosféry.

Problémom modelu je vysvetliť, kam sa pretiahnutá litosféra umiestni, ak v blízkosti diapíru-panvy nepredpokladáme subdukciu postupujúcu smerom od panvy. Pri horizontálnom pohybe litosféry by sa vo vrchnej rigidnej časti mali tvoriť listrické zlomy a antitetické nakláňanie klesajúcich blokov, pre ktoré nie sú geologické doklady v neriftových panvách (E. V. ARTJUŠKOV a M. A. BEER 1983, obr. 2). Na druhej strane model roztiahnutia litosféry podporujú indície tektonického „rozvrstvenia" litosféry, napr. zistenie subhorizontálnej drvenej zóny vo vrte SG-3, geofyzikálne údaje o horizontálnych diskontinuitách interpretované ako vznik čiastkových dosiek vnútri litosféry (S. V. RUŽENCEV a S. D. SOKOLOV 1984, A. L. JANŠIN et al. 1984, S. N. IVANOV 1984, C. FROIDEVAUX a L. FLEITOUT 1984 a i.). Roztiahnutie má vznikať mechanicky alebo termoplasticky. Posledný výklad sa prijíma pre vznik panví s diapírmi plášťa a zaoblúkových panví — konvergentne — s extencionálnym aktívnym okrajom dosiek (J. AUBOUINA et al. 1984). V súvislosti s koncepciou roztiahnutia v doske nasunutej nad subdukciou sa modely vracajú k pôvodnému výkladu Benioffa o aktívnom nasúvaní a výzdvihu kontinentálnej kôry.

Modely roztiahnutia litosféry vylučujú látkové zmeny v kôre, ako je bazifikácia s. l. Ani nepredpokladajú heterogenitu typov kôry v doskách. I. B. RAMBERG a P. MORGAN (1984) poznamenávajú, že jednoduchý model D. P. MC KENZIEHO (1978), používaný príliš špekulatívne, neplatí, ak pred roztiahnutím existovala anomálna stavba a zloženie kôry. Geofyzika neindikuje globálne, ale skôr miestne rozvrstvenie litosféry, t. j. nedajú sa preukázateľne navzájom spájať indikované diskontinuity. Okrem toho sú početné diskontinuity zvlnené, alebo niektoré prechádzajú do plôch subvertikálnych, a to v miestach, kde na povrchu nie je doklad pre výstup čiastkových dosiek, blokov a p.

V knihe sú preferované názory o primárnom heterogénnom zložení a o stavbe kôry, doložené geologicky a paleogeograficky. Početné panvy sa rovzíjali v mieste predošlej subsidencie a subsidencia na rovnakom mieste pretrvávala tektogénne epochy. Predpokladám, že zmeny hrúbky kôry sú často primárne. Pri formovaní kontinentálnej kôry z kôry oceánskej v tzv. protoorogénnom cykle (najnovšie A. V. PEJVE et al. 1984, A. L. JANŠIN et al. 1984) zostávajú v ranom štádiu vývoja kôry zachované heterogenity v kôre, t. j. zvyšky slabo sializovanej alebo nesializovanej kôry. V deutoorogénnom cykle na rozrušovanej kontinentálnej kôre, predtým konsolidovanej, vznikajú panvy predovšetkým v úsekoch ťažšej kôry, zafixovanej z doby protoorogénneho štádia formovania kôry. Tak možno vysvetliť odchýlky od typu panví definovaných v geosynklinálnom modeli, napr. znaky eugeosynklinály vo vnútrokontinentálnej panve. Modely uvedených sovietskych autorov i model M. M. LEBEDEVA et al. (1984) odmietajú koncepciu postupnej akrécie kontinentálnej kôry.

Pod panvami je stenčená kôra a plášť je podľa seizmológie buď rigidnejší, alebo naopak menej viskózny než normálny plášť pod kontinentmi. Podľa vrtných údajov alebo geofyziky je kôra bázickejšia, simatická. V súčasných názoroch sa existencia bázickej kôry pod panvami vysvetľuje štyrmi spôsobmi:

a) ako zvyšok nesubdukovanej kôry pri uzatváraní oceánu (J. F. DEWEY et al. 1973, A. TOLLMANN 1978, S. S. KARAPETOV 1984 a i.),

b) zachytením (uviaznutím) trosky oceánskej kôry pri okraji kontinentálnej dosky počas subdukcie alebo kolízie dosiek,

c) obdukciou na fosílnych kolíznych okrajoch dosiek (napr. A. V. PEJVE et al. 1972),

d) ako restit negranitizovanej kôry počas formovania kontinentálnej kôry (F. ČECH 1982, A. L. JANŠIN et al. 1984, F. ČECH a J. ZEMAN 1985).

Ťažšiu kôru s prevahou bázik, bez granitových plutónov orogénneho typu, možno označiť ako simatickú a synonymne ako bázickú, bázitovú, melanokrátnu, mafickú, suboceánsku, para- či kváziocéansku. Kôra je:

a) slabo granitizovanou (nesúvisle granitizovanou) alebo negranitizovanou kôrou, ktorá sa stala súčasťou kontinentálnej kôry;

b) bazifikovanou kontinentálnou kôrou, pričom bazifikácia selektívne prebiehala v miestach neorogénnej, slabo granitizovanej a konsolidovanej stenčenej kôry.

Kôra simatického typu predisponovala vznik opakovanej subsidencie, dlhodobej mobility platforiem a mobilných vnútorných — simatických masívov (F. ČECH a J. ZEMAN 1982).

V astenosfére existujú nerovnovážne stavy najpravdepodobnejšie vplyvom čiastočného natavenia hornín, hoci niektorí autori podstatu astenosféry vysvetľujú odlišne (napr. odlišným látkovým zložením — F. BIRCH 1969, O'CONNEL a JACKSON in T. H. JORDAN a W. S. FYFE 1976). Pre prvý výklad svedčí spojitosť elevácií astenosféry do hĺbky až 50 km pod oblasťami s recentným vulkanizmom. Do magmou vyprázdnených priestorov klesajú gravitačne bloky kôry a na povrchu vznikajú panvy. Do menej viskóznej astenosféry môžu klesať i bloky viskóznejšej alebo ťažšej (napr. eklogitizáciou) litosféry, a tak podmieňovať vznik panví bez vulkanizmu. Taveniny (čadičové) vytláčané z astenosféry spôsobujú bazifikáciu kôry, čo rozširuje areál subsidencie.

Podľa geofyzikálnych výskumov má astenosféra vrstevnatú stavbu, čo môže byť i odrazom nerovnomerného čiastočného natavenia, ale tiež odrazom rôzne hlboko ponorených blokov rigidnejšej natavenej litosféry.

Existujú geofyzikálne indície prepojenia aktívnej astenosféry pod bajkalským riftom „vertikálnym kanálom" znížených rýchlostí P-vĺn s vonkajším jadrom (G. N. BUGAJEVSKIJ 1977).

### Diapirizmus vrchného plášťa

Plášťové diapíry sú globálne sa vyskytujúcim fenoménom pod izometrickými alebo lineárnymi panvami s veľkou subsidenciou. Existujú odumreté diapíry bez vyššieho tepelného toku alebo s anomálne nízkym tepelným tokom (Čierne more, transylvánska panva) a aktívne diapíry s plytkou astenosférou a anomálne vysokým tepelným tokom. Sú buď bez súčasného vulkanizmu (panónska panva) alebo s recentným explozívnym vulkanizmom (Tyrrhenské a Egejské more — obr. 26).

Výstup plášťa umožňuje nižšia viskozita a merná hmotnosť hornín (čiastočne natavených) a znížený geostatický tlak (stenčená kôra). Mechanizmus výstupu plášťa riešia rôzne modely (J. NIKOLSKIJ 1982): advekcia, konvekcia, rádioaktívny rozpad, diferenciácia plášťa na hustotne rozdielne zóny a úseky i fyzikálno-chemické zmeny hornín podmienené stupňovitou gravitačnou diferenciáciou (P. M. SYČEV 1984), keď vplyvom hustotnej inverzie (H. RAMBERG 1967 — obr. 21) ľahšie diferenciáty stúpajú do pásiem nižšieho geostatického tlaku alebo do oblastí bez prídatnej kompresie. S. W. CAREY (1976) alebo E. E. MILANOVSKIJ (1981) vysvetľujú diapirizmus expanziou alebo pulzáciou Zeme. J. NIKOLSKIJ (1982) kladie vznik diapírov pod astenosféru do hĺbok 800—900 km a dobu výstupu odhaduje na 140—500 mil. rokov. Usudzuje, že v mediteránnej oblasti diapíry, založené vo vrchnom proterozoiku, prenikli najbližšie k povrchu, zatiaľ čo v Strednej Ázii sú ešte umiestnené v hlbšom plášti. Ich prejavom majú byť okrúhle záporné tiažové anomálie (obr. 23). Výstup diapírov údajne podmieňuje vysokohorský výzdvih stredoázijského pásu nadväzujúceho na Stredozemné more a Pacifik.

Množiace sa poznatky z družicových snímok o okrúhlych štruktúrach na kontinentoch by podporovali názor o globálnom rozšírení plášťových i kôrových diapírov. Kruhové štruktúry typu lunárnych morí z predgeologického štádia vývoja Zeme majú nielen impaktný, ale i endogénny pôvod. G. B. UDINCEV et al. (1984) dokladajú, že i vznik oceánskych riftov je podmienený diapirizmom plášťa.

Umiestnenie diapírov do „voľného" priestoru riešia rôzne modely. Najjednoduchší je výklad expanziou Zeme. F. HORVÁTH et al. (1981) umiestnenie panónskeho diapíru a iných diapírov rieši rozštiepením čela dosky a odchlopením rigidnej litosféry do duktilnej časti pri kolízii severoafrickej dosky s európskou. Umiestnenie diapírov v trieštiacom čele dosky predpokladá J. F. DEWEY et al. (1973), K. BIRKENMAJER (1976) alebo A. TOLLMANN (1978). Naopak, o umiestnení medzi odlamovanými klesajúcimi blokmi vplyvom ich bazifikácie uvažuje V. V. BELOUSOV (1982) alebo vplyvom eklogitizácie E. V. ARTJUŠKOV (1979 — cf. obr. 22, 24). O vzniku voľného priestoru rozťahovaním litosféry uvažuje D. P. MC KENZIE (1978). Jeho model na vznik panónskej panvy aplikovala L. ROYDENOVÁ et al. (1982) — obr. 40. Roztiahnutie litosféry vysvetľujú pohybom presunujúcej sa dosky smerom k subdukčnej zóne pod Východnými Karpatami (obr. 10). O konvergencii dvoch subdukujúcich dosiek pod panónskou panvou uvažoval L. STEGENA et al. (1975 — obr. 11).

Je pravdepodobné, že umiestnenie plášťa je predisponované primárne tenkou litosférou, v ktorej ľahšie nastala interakcia litosféry s aktívnou astenosférou (obr. 22). Diapír má v zárodočnom štádiu asi tvar plochej elevácie s postupne narastajúcou výškou, až dosiahne diapír tvar pňa. Vplyvom prehrievania spodnej litosféry a tlakom taveniny odštiepenej z astenosféry prenikajú do litosféry intrúzie bázik hydroruptúrnym mechanizmom ( P. M. SYČOV 1984) a diapír môže mať tvar lakolitu až lakolitu cédrového typu. Areál diapíru je pásmom bazifikácie litosféry. Chladnúci vrchol diapíru sa postupne prepadá do hlbšieho, menej viskózneho pásma a taveniny sú vytláčané do strán. Diapír dostáva hubovitý tvar a nad konkávnym vrcholom sa tvorí sedimentárna panva.

Expanzia plášťa (a jeho narastanie do strán) vyvoláva vo vrchnej litosfére vertikálne i horizontálne orientované strižné a ťahové napätie (obr. 3) a vznik poklesových zlomov orientovaných radiálne k centru diapíru. Nad obvodom (okrajom) diapíru vznikajú strihom prešmyky a horizontálne posuny. Oba zlomy sú orientované tangenciálne k okrajom diapíru (obr. 3, 5). V riftoch sa strižné zlomy stýkajú s pozdĺžnym nadložím lineárne pretiahnutého diapíru (obr. 4). Oblasť nad vrcholom diapíru je v pásme roztiahnutia bez významnejšieho uplatnenia tangenciálneho napätia (obr. 3, 6).

Na styku vystupujúceho horúceho diapíru s normálnym „chladným" plášťom sa vytvára rozhranie, ktoré má v pôdoryse oblúkový tvar. Toto rozhranie je sprevádzané ohniskami hlbokých zemetrasení (obr. 27); táto zóna býva označovaná ako zóna Benioffova. Je spájaná so subdukciou — v Tyrrhénskom a Egejskom mori so subdukciou africkej dosky (M. BOCCALETTI et al. 1976, D. P. MC KENZIE 1978). Vznik tejto zóny však s hypotetickou subdukciou kontinentálneho rozsahu nesúvisí (Benioffova zóna je najpravdepodobnejšie odrazom rozhrania rôzne temperovaného a rôzne viskózneho plášťa a jej vznik súvisí asi s klesaním chladnej litosféry pod diapír, a to do priestoru, ktorý vznikol v astenosfére ako okrajová lemová synklinála po odtoku hmôt do narastajúceho diapíru (cf. H. RAMBERG 1967, F. ČECH a J. ZEMAN 1985). Pri klesaní periférnej litosféry a plášťa stýkajúceho sa s aktívnou astenosférou vzniká čiastková peridiapírová subdukcia lokálneho rozsahu (obr. 20, 21). Môže mať centripetálny alebo len asymetrický jednostranný rozsah. Druhý prípad je u asymetrických diapírov, ktoré však môžu byť jedným z vývojových štádií diapirického rastu.

S výstupom diapírov je spojený vznik oválnych až izometrických (kruhových) panví medzihorskej pozície (F. ČECH a J. ZEMAN 1985). Sú typickým javom v mediteránnej a cirkumediteránnej oblasti.

S výstupom horúcich plášťových diapírov a s eleváciou astenosféry je na hlbinných zlomoch spájaný vznik riftov (N. A. FLORENSOV 1977, E. V. ARTJUŠKOV 1981, M. H. P. BOTT 1981, H. J. ILLIES 1981 a iní). Zlomová predispozícia určuje lineárny tvar diapírov. Počiatočné štádium však môže byť klenbové, keď nad oválnym diapírom vzniká tripletový systém zlomov.

Rifty sú v zmysle H. Y. ILLIESA (1981) lineárne grabeny, väčšinou spojené s vulkanizmom. Vulkanizmus nemusí byť súčasný so vznikom riftu, ale môže vznik riftu predchádzať až o 50 mil. rokov (rýnsky rift). Pod riftmi je geofyzikálne indikovaná tenká litosféra zapríčinená eleváciou horúcej astenosféry a vyvolávajúca vysoký tepelný tok i v litosfére. Rifty vznikajú na styku dosiek alebo paleodosiek, vnútri kontinentu i v predpolí orogénov, často na lineamentoch prekambrijského veku. Niektorí geológovia sa domnievajú (napr. A. D. MIALL 1984), že lineamenty môžu byť paleosutúrami dávneho konvergentného styku dosiek, a že teda rifty môžu vznikať ako divergentný fenomén na mieste bývalej konvergencie dosiek.

Kontinentálne rifty vznikajú na kontinentálnej kôre bez toho, že by v podloží riftu bola evidencia pre existenciu oceánskej kôry. Rifty však môžu vznikať na styku sialickej a simatickej kôry vnútri kôry kontinentálnej. Pred vznikom riftu dochádza k vytvoreniu klenby, dómu, antiformy (H. J. ILLIES 1981, S. BHATTACHARJI 1984 — obr. 6). V klenbe vzniká ťahové napätie, ktoré podľa M. H. P. BOTTA (1981) dosahuje hodnotu až 200 MPa, v závislosti na rýchlosti klesania v rifte. Pre hodnotu 200 MPa sa ráta s veľkosťou subsidencie až 5 000 m. I. B. RAMBERG a P. MORGAN (1984) uvádzajú, že pred vznikom riftu môže nastať i prehýbanie kôry. Pri mechanizme tvorby riftu sa všeobecne predpokladá, že roztiahnutie v litosfére má dvojaký charakter: vo vrchnej rigidnej časti pôsobia krehké deformácie, v spodnej duktilnej časti tvárlivé, viskózne deformácie (obr. 7). K vzniku riftu môže prispieť roztiahnutie litosféry i bazifikácia — eklogitizácia. V súčasnosti vznik riftov stále nie je uspokojivo vyriešený; je pravdepodobné že rifty nevznikajú jediným mechanizmom — procesom.

Znaky typické pre rift sú uvedené v tab. 1, rôzne názvy riftových megaštruktúr v tab. 2. Podľa výskumov, najmä v posledných 10 rokoch, je rozšírenie riftov väčšie než sa usudzovalo predtým (P. A. ZIEGLER 1982, V. B. SOLLOGUB a A. V. ČEKUNOV 1983 a i.). Elevácia plášťa zostáva zachovaná i v neaktívnych — odumretých riftoch (obr. 8). Smer a výraznosť prejavu riftu sa menia v závislosti na kompetencii či nekompetencii súvrství alebo geologických jednotiek, ktoré rift pretína (obr. 9).

Rifty majú v oceánoch i na kontinentoch tripletové usporiadanie, čo je následkom predchádzajúcej etapy vyklenutia kôry (K. C. A. BURKE a J. F. DEWEY 1973). V riftoch, kde nie sú aktívne všetky tri ramená, niektoré z ramien odumiera, býva to väčšinou rift smerujúci do kontinentu (obr. 17). Aulakogény bývajú niekedy stotožňované s riftmi. Aulakogény v zmysle definície K. C. A. BURKA a J. F. DEWEYA (1973) sú grabeny v kontinente smerujúce viac-menej kolmo k čelu orogénu (obr. 18). Charakter aulakogénu by mal hornomoravský úval (obr. 61). Grabeny, ktoré nie sú súčasťou riftového tripletu, tie ktoré smerujú do oceánu, navrhuje D. M. CURTIS (1980) označovať ako pseudoaulakogény. K pseudoaulakogénom by bolo možné zaradiť na Morave treťohorný nesvačilský a vranovický graben, ktoré označil F. PÍCHA (1979) ako aulakogény (obr. 61). O mechanizme vytvárania riftu sa stále diskutuje: roztiahnutie a stenčenie litosféry, klzanie litosféry do strán od osi riftu po svahoch astenosférovej elevácie, prepadanie segmentovanej litosféry do viskóznej astenosféry a iné modely. Diskutuje sa i o probléme, čo bolo skôr: astenolit spôsobujúci roztiahnutie, alebo roztiahnutie umožňujúce intrúziu astenolitu. Rifty sú dôležitými ložiskovými štruktúrami (S. P. MIKOUTSKIJ 1984).

### Vznik panví horizontálnymi pohybmi na zlomoch

V súčasnosti prevažná väčšina geológov vysvetľuje vznik sedimentárnych panví tektonickými príčinami. V modeloch doskovej tektoniky sa vznik panví viaže prevažne na pohyby dosiek a na charakter styku dosiek. Vplyvom subdukcie bol vysvetľovaný vznik panónskej panvy (L. STEGENA et al. 1975, L. ROYDENOVÁ et al. 1982 — obr. 10, 11) alebo viedenskej panvy (R. JIŘÍČEK 1981). Zlomový styk dosiek a horizontálne pohyby majú viesť k vzniku panví s rýchlou subsidenciou. R. A. GREEN (1977) označil tieto panvy ako strižné (obr. 12). Horizontálnymi posunmi vznikajú rôzne typy panví (obr. 14). Posunom na nerovnom zlome sa odlomí blok, odtrhne sa od susednej kôry a prepadáva sa. Tento typ panví bol označený ako „pull-apart basins". Panvy majú škatuľovitý tvar a sú spojené s divergentným okrajom dosiek (napr. pozdĺž atlantického pobrežia) alebo s okrajom dosiek tvoreným transformnými, prípadne transkurentnými zlomami (J. C. CROWEL 1974). Vnútri panví sa môže tvoriť i dno s vulkanitmi, pripomínajúce mikrooceán (obr. 16).

Vznik viedenskej panvy horizontálnymi posunmi na hypotetických zlomoch smeru SV-JZ, ako typu „pull-apart", riešil modelovo F. HORVÁTH a L. ROYDENOVÁ (1981). Vysvetlenie týmto modelom naráža na ťažkosti, pokiaľ ide o mechanizmus roztvárania a o to, že pre horizontálne pohyby (tiež predpoklad Z. ROTHA 1980) nie je geologická evidencia (R. JIŘÍČEK 1982, F. ČECH 1984). Ani charakteristické znaky pre panvy typu „pull-apart" (J. C. CROWELL 1974, obr. 16) nemožno vo viedenskej panve doložiť.

V doskových modeloch sa prikladá malý význam vzniku panví na poklesových zlomoch, hoci pre existenciu týchto zlomov prináša hlbinná geológia najviac dôkazov.

## Typy kôry

V modeloch riešiacich vznik sedimentárnych panví, najmä v modeloch odvodzovaných z geosynklinálneho modelu, nebol rešpektovaný typ kôry. V modeloch odvodených z doskovej tektoniky je síce typ kôry uvádzaný ako jeden z hlavných fenoménov, ale vznik i zánik kôry sa mechanicky riešia dynamikou dosiek. V modeloch sa podceňujú látkové zmeny vyvolané inými procesmi, napr. bazifikáciou s. l. Vylučuje sa i prvok perzistencie typov kôry, napr. restitový charakter simatickej kôry. Pritom pre perzistenciu sú doklady v dlhodobej alebo opakovanej subsidencii, keď subsidencia v panve presahuje jeden i dva tektogénne cykly. Na niektoré príklady upozornil F. ČECH a J. ZEMAN (1982, 1985). Ďalším príkladom je perikaspická depresia, kde hlbokomorská sedimentácia začala už vo vrchnom proterozoiku a potom sa s prestávkami (splytčenie či regresia) objavovala až do kvartéru (A. L. JANŠIN et al. 1984).

Výskyt simatickej kôry v kontinentálnej kôre (znova zdôrazňujem, že tento typ kôry nemožno mechanicky stotožňovať s oceánskou kôrou) sa vysvetľuje jednak bazifikáciou kontinentálnej kôry, jednak ďalšími modelmi, ktoré už boli spomenuté vyššie. Pre vývoj západokarpatskej kôry vypracoval M. MAHEĽ (1978) model, v ktorom predpokladá od paleozoika tenkú restitovú paraoceánsku až oceánsku nekonsolidovanú kôru medzi kôrou granitizovanou, s väčšou hrúbkou a vyššou konsolidáciou (obr. 15). Ťažšia paraoceánska kôra sa stala podložím mezo-kenozoických sedimentárnych panví. Vplyvom hustotnej diferenciácie sa subdukciou dostala pod ľahšiu granitickú kôru a sedimentárna výplň panví sa odlúčila od podložia, zvrásnila sa a stala sa súčasťou príkrovov.

Restitová simatická kôra môže vznikať pri postupnom narastaní kontinentálnej kôry. Nerovnomerný výskyt granitoidov v štítoch ukazuje, že granitizácia známa z nukleov kontinentálnej kôry nemala celokontinentálny, tým menej potom planetárny rozsah. Medzi nukleami a granitoidnými migmatizovanými komplexmi zostali pásy či areály slabo granitizovanej až negranitizovanej kôry (J. ZEMAN 1979a, 1980, F. ČECH a J. ZEMAN 1985).

Akrécia kontinentálnej kôry je preukázaná v štítoch (napr. T. N. SPIŽARSKIJ 1984), v ostrovných oblúkoch a riftoch (J. ZEMAN 1979b, N. A. BOŽKO 1984). Akceptovanie akrečného modelu nevylučuje lokálnu bazifikáciu ako regresívny proces vo vývoji kontinentálnej kôry. Tenšia simatická kôra môže byť konzervatívnym prvkom v heterogénnej kôre (restitový typ) alebo prvkom regresívnym, ak vznikala bazifikáciou. Bazifikácia v spodnej časti kôry, ale i okolo zón granitizácie vo vrchnej kôre, môže prebiehať pri progresívnom vývoji kôry, napr. procesom metamorfnej diferenciácie I. A. REZANOVA (1981).

Vplyv rozdielnej hrúbky sialickej a simatickej kôry sa prejavuje v rôznej mobilite. Táto je najvýraznejšia na okraji kontinentov, kde vzniká najväčší počet mobilných sedimentárnych panví s rýchlou subsidenciou (F. L. SCHWAB 1976). N. J. KUNIN (1984) označil okrajové panvy na prechodnej (simatickej) kôre ako deprogény. K ním patria i dlhodobe mobilné a klesajúce panvy.

Satlinský vrt, hĺbený v stredokurinskej medzihorskej depresii, s plánovanou hĺbkou 15 km, zastihol pod pliocénno-kvartérnou molasou a mezozoickými karbonátmi v hĺbke 3 530 m mezozoický vulkanodetritický komplex s bazaltovými, andezitovými a dacitovými lávami. Komplex bol od r. 1984 prevŕtaný zatiaľ v hrúbke 4 900 m. Tento vrt preukazuje taktiež dlhodobú subsidenčnú aktivitu na stenčenej kôre simatického charakteru.

Typ kôry je jedným z dôležitých fenoménov predisponujúcich veľkú subsidenciu. Dokladajú to i práce napr. J. AUBOUINA (1984), J. V. ARCHIPOVA (1984), M. A. BEERA (1984) a iných.

Na hrubej kontinentálnej kôre sa nevytvoria veľké panvy s rýchlou subsidenciou, pokiaľ sú v kôre orogénne plutóny a ich sprevádzajúce tektonické fenomény.

Veľké sedimentárne panvy vznikajú na oceánskej a suboceánskej kôre, s klesaním kompenzovaným alebo nekompenzovaným sedimentáciou. Hlbšou príčinou založenia panví je asi rozdielne zloženie plášťa pod kontinentálnou a oceánskou kôrou (D. H. GREEN a A. E. RINGWOOD 1969, B. G. LUTC 1975 a iní) a z toho vyplývajúca rozdielna dynamika v plášti a litosfére. V tenkej litosfére anomálnu dynamiku podmieňujú aktívne plášťové diapíry.

W. D. MAC DONALD (1972) podľa rôznej dynamiky a mobility rozlíšil tieto typy kôry:

a) normálnu oceánsku;

b) oceánsku s väčšou hrúbkou bazaltovej vrstvy, tzv. typ kôry platillo — kôra abysálnych plošín;

c) prechodnú kôru pri ostrovných oblúkoch a pod malými oceánskymi panvami (suboceánska kôra v mojom poňatí);

d) mladú tektonickú kôru — kontinentálnu kôru s malou hrúbkou;

e) kontinentálnu kôru štítového typu s mohutnými graniticko-metamorfnými komplexmi.

Zatiaľ sa ako najviac doložený javí progresívny model A. V. PEJVEHO et al. (1972) a W. D. MAC DONALDA (1972); podľa neho vývoj kontinentálnej kôry prebieha cez ostrovné oblúky (od typu *a* po typ *e*) a odkrajové moria (zaoblúkové či medzioblúkové panvy) vznikli na restitovej simatickej alebo azda oceánskej kôre. Pri mobilizácii okrajov platformy procesom bazifikácie s. l. nastávajú spätné premeny typov kôry smerom k oceánskej, ale štádium ostrovného oblúka sa nemôže vytvoriť.

V karpatsko-balkánskej oblasti vypracoval S. I. SUBBOTIN et al. (1972) model vzniku panví na kontinentálnej kôre, a to jej deštrukciou, riftogenézou. Extrémnym prípadom oceanizácie malo byť Čierne more (napr. V. E. CHAIN a V. I. SLAVIN 1972). Existujú však doklady pre vznik účinkami fosílneho plášťového diapíru (F. ČECH a J. ZEMAN 1985).

### Hmotnosť sedimentárnej výplne a subsidencia

Vznik subsidencie je podmienený endogénnymi príčinami. Vplyv hmotnosti sedimentov je maximálne prídatný (I. A. REZANOV 1977, E. V. ARTJUŠKOV 1981). Tento vplyv sa uplatňuje najmä v asymetrických lineárnych panvách, kde hmotnosť sedimentov pôsobí na úzkom páse zlomove členenej kôry. Hmotnosť sedimentov má zanedbateľný vplyv i na kompresiu kôry.

Pre otestovanie vyššie uvedených predpokladov o nepodstatnom účinku hmotnosti sedimentov na subsidenciu som vypracoval nasledujúcu úvahu a výpočet.

Základným faktorom v diferenciácii hustoty litosféry je jej vrstevnatá stavba s narastaním hustoty, a teda i hmotnosti horninových komplexov smerom do hĺbky. Zväčšenie hustoty prebieha vplyvom látkových zmien, t. j. zmenou petrografických typov hornín v rade žula — gabro — peridotit a aj fyzikálnych zmien, t. j. zmenou pórovitosti a hustoty hornín vplyvom stláčania narastajúcim geostatickým a tektonickým tlakom. Problémom je stanoviť strednú hustotu litosféry. Pri zvolenej hodnote = $3{,}15 \,.\, 10^3$ kg . m$^{-3}$ môžeme kalkulovať na báze litosféry, t. j. na hranici s astenosférou v hĺbke 100 km, s geostatickým tlakom $3{,}15 \,.\, 10^3$ MPa.

Ak výplň panvy tvoria sedimenty, podľa stupňa diagenézy s hustotou $2{,}2 \,.\, 10^3$ až $2{,}7 \,.\, 10^3$ kg . m$^{-3}$ (v priemere asi $2{,}5 \,.\, 10^3$ kg . m$^{-3}$), potom tlak vyvíjaný na fundament pod panvou, napr. v hĺbke 5 km, bude $12{,}5 \,.\, 10^1$ MPa.

Tlak vyvinutý rovnakým stĺpcom žúl bude väčší (granity majú hustotu až $2{,}65 \,.\, 10^3$ kg . m$^{-3}$) a v prípade metamorfovaných hornín, ktorých hustota je až $2{,}7 \,.\, 10^3$ kg . m$^{-3}$, bude ešte väčší.

Ak sa teda hlbšia časť kôry neprehýba pod hmotou žúl alebo až metamorfitov s nasunutým sedimentárnym plášťom, potom sedimentárna výplň panvy nemôže mať žiaden rozhodujúci vplyv na vznik prehybu hlbšej časti kôry alebo dokonca litosféry. Príčiny klesania panví nemôžeme teda hľadať v prostom vyplňovaní — kompenzácii panví. Treba ich hľadať — ako už bolo uvedené, v hlbších procesoch.

## Modifikovaný model vzniku panvy

Podrobný model vzniku panví nad plášťovým diapírom vypracoval R. W. VAN BEMMELEN (1972). Jeho koncepciu rozpracovali ďalší geológovia, najmä v súvislosti s riešením genézy riftov, mediteránnych panví, panvy Great Basin a ďalších (rozsiahla literatúra je sčasti citovaná v jednotlivých kapitolách tejto knihy).

Následne uvedený model je modifikáciou týchto modelov a pokusom vypracovať všeobecný model pre neriftové panvy, resp. panvy, ktoré mali rané štádium riftové a zmenili sa na panvy izometrické. Tento model vychádza z týchto fenoménov:

a) typ kôry,
b) mechanizmus tvorby a pretvárania diapíru,
c) predpoklad klesania litosféry do menej viskóznej astenosféry.

Pripúšťa i bazifikáciu s. l. v pásme aktívneho horúceho diapíru. Uvažuje zároveň i o kolapsových procesoch pri chladnutí diapíru.

Aktivitu diapíru podporuje primárne tenká simatická kôra uzavretá (obklopená) kôrou sialickou, ktorá je v dobe výstupu diapíru tektogénno-orogénne aktívna. Heterogénna rôzne hrubá kôra podmieňuje rozdielnu diferenciáciu v plášti, rôzny podiel restitov po vyplavení magmy a rozdielne napätia v plášti.

Eliptické, izometrické alebo trojuholníkovité panvy medzihorského typu sú lemované orogenickou kôrou väčšinou dvojnásobne hrubou, v porovnaní so simatickou kôrou nad diapírom. Pod tenkou kôrou je permanentná elevácia plášťa, vytvárajúca nerovnovážny stav v tlaku a napätosti v plášti. Tento stav zintenzívňujú hlbinné zlomy tvoriace sa na rozhraní tenkej a hrubej kôry. Niektoré zlomy pri vzniku horúceho diapíru podmieňujú jeho výstup a oddeľujú chladný rigidnejší plášť a hlbokú astenosféru od horúceho plášťa a elevácie astenosféry. Príkladom sú oblúkovité Benioffove zóny v Tyrrhenskom a Egejskom mori (F. ČECH — J. ZEMAN 1984); obr. 20.

V litosfére a astenosfére vzniká hustotná nerovnováha (obr. 21). Horúci diapír s bázickými diferenciátmi ľahšími než plášť sa snažia zo stavu hustotnej inverzie migrovať do vyššej úrovne litosféry mechanizmom pretavovania plášťovej elevácie. Taveniny stúpajú, nenatavené restity klesajú do astenosféry a hromadia sa pri jej báze. V litosfére sa diapír stabilizuje v pásme spodnej kôry v hustotne ekvivalentnom prostredí.

Pri tvorbe diapíru vznikajú okrajové synklinály (H. RAMBERG 1967). Tieto synklinály sa tvoria odtokom hmôt do diapíru. Nad nimi dochádza ku klesaniu kôry a na jej povrchu ku vzniku okrajových depresií. Sedimentácia v týchto depresiách začína skôr než v panve nad diapírom (F. ČECH — J. ZEMAN 1984); tab. 5.

Pri stúpaní diapíru môže vzniknúť krípové kĺzanie duktilnej litosféry, resp. prehriatej spodnej kôry po svahoch diapíru do okrajových synklinál. Kĺzanie môže nastať taktiež v astenosfére. Klesanie v plášťovej okrajovej synklinále ďalej pdmieňuje ohyb a zaklesávanie okrajov litosféry okolo diapíru. Vznikajú tak lokálne subdukcie modifikované okrajovými zlomami. Subdukcie označujem ako pasívne; vznik okrajových depresií na povrchu kôry je dôsledkom ťahove ohybových napätí. Pri konvergentnom styku litosféry alebo okolo diapíru môže vzniknúť hlboký koreň kôry, podmieňujúci orogénne zdvihy okolo diapíru pod panvou.

Počas výstupu sa diapír pretvára, rozpína do strán a vytvára hubovité alebo šikmé kužeľovité teleso. Príklady vývoja a rôzneho tvaru diapíru uvádza obr. 20. Rozpínanie diapíru a vytvorenie misovitého vrcholu podmieňuje začiatok klesania v panve nad diapírom pred jeho chladnutím. Litosféra pozdĺž hlbinných zlomov klesá do astenosféry. Subsidencia podporuje i vyprázdňovanie krbov u vulkanicky aktívnych diapírov. Nad okrajom diapíru vznikajú strižné napätia podporujúce vznik strižných zlomov alebo rejuvenáciu strižných pohybov na preddiapírových zlomoch. Nad diapírom vzniká ťahové napätie podporujúce vznik poklesových zlomov. Pri roztekaní rozpínajúceho sa diapíru sa môžu nad ním vytvárať i listrické zlomy, teda v iných podmienkach než predpokladá model D. P. MC KENZIEHO (1978).

Chladnutie a kolaps diapíru môžu spôsobiť zrýchlenie subsidencie. Pokiaľ sa nevytvorí diapír hubovitého tvaru, je možné, že klenbovité diapíry sú produktom expanzie a „roztekania“ diapíru, ktorý pôvodne mal kužeľovitý tvar (diapíry pod recentne vulkanicky aktívnymi panvami). Klenbovitý alebo hríbovitý diapír (alboránska panva) je finálna plášťová ireverzibilná štruktúra. Ostáva zachovaná i po znížení tepelného toku (depresia Viking, Čierne more).

## Klasifikácia sedimentárnych panví so zameraním na panvy s ložiskami uhľovodíkov

Problematika klasifikácie panví nezahrňuje staršie označovanie geosynklinálnych panví, ako ho vypracoval H. STILLE (1924) alebo M. KAY (1951).

Jedna zo starších obecných klasifikácií rozlišovala panvy kompenzované a nekompenzované sedimentmi (J. E. ADAMS et al. 1951). Kompenzované sú tie, v ktorých rýchlosť klesania je vyrovnávaná hrúbkou nahromadených sedimentov.

Prvú klasifikáciu panví podľa typov kôry vypracoval F. G. WELLS (1949), ktorý rozlišoval panvy ensialické a ensimatické. Klasifikácia sa týkala geosynklinál. Ensialické geosynklinály sa vyznačujú sedimentárnou výplňou, kde horninový detrit pochádza z kontinentu. V ensimatických geosynklinálach prevažujú bázické vulkanity, bridlice a droby s vulkanickým materiálom — všeobecne horniny s nízkym obsahom $SiO_2$. F. G. WELLS (l. c.) podotýka, že v sialickej kôre zostávajú zachované uzavreniny ensimatických geosynklinál.

Heterogénnym typom sú panvy, ktoré vznikali v ranom štádiu ako ensimatické a neskôr sa vplyvom sializácie kôry zmenili na ensialické. Príkladom sú niektoré andské panvy (J. JAROŠ 1975), okrajové geosynklinály (britsko-nórsko kaledónska geosynklinála F. W. DUNNINGA 1977) alebo niektoré paleorifty (J. ZEMAN 1979b).

Najčastejšie sa panvy klasifikovali podľa pozície v tektonických jednotkách. V súčasnosti existujú vedľa seba klasifikácie vychádzajúce z geosynklinálneho modelu a klasifikácie odvodené z doskovej tektoniky.

*Klasifikácie podľa geosynklinálneho modelu*

Najnovšie klasifikácie s použitím rôznych typov endogénnych režimov usporiadal V. V. BELOUSOV (1978b). Z tejto klasifikácie som vybral panvy s možným výskytom ložísk uhlia a prírodných uhľovodíkov.

1. V geosynklinálnom režime: flyšové trógy eu-, mio- a parageosynklinál, panvy na vnútorných masívoch mobilizovaných behom tohoto režimu (vnútorné alebo intermontánne panvy).
2. Panvy na styku geosynklinálneho a platformového režimu: okrajové depresie so zmiešaným vývojom, mobilnejším na strane priľahlej ku geosynklinálnej zóne. Pri migrácii geosynklinál dôjde k mobilizácii stabilnejších pásiem a ku prekrytiu a ponoreniu starších sedimentov. Tento proces je zo začiatku priaznivý, neskôr môže byť nepriaznivý pre kvalitu a zachovanie ložísk uhľovodíkov.
3. V platformovom režime: syneklízy a aulakogény (ich molasové štádium). Do tejto skupiny možno zaradiť i depresie atlantického typu, t. j. tie, ktoré nie sú na styku s vrásnenou geosynklinálou a nemajú alkalicko-vápenatý vulkanizmus.
4. V orogénnom epigeosynklinálnom alebo epiplatformovom režime (orogénna aktivácia platforiem, napr. Taň-Šaň): predhlbne, vnútrohorské a medzihorské depresie.
5. V riftovom režime — kontinentálne rifty.

Riftogenéza v geosynklinálnom modeli nemá presne vymedzené postavenie. Väčšinou sa spája s aktiváciou platforiem alebo s epiplatformovým orogénnym režimom, keď je lepšie depresie označovať ako tektonické grabeny. Rifty však môžu vznikať i pri geosynklinálno-orogénnom režime na okraji platforiem alebo v počiatočnom štádiu tvorby geosynklinál. Príkladom sú riftogeosynklinály (W. R. CHURCH — R. K. STEVENS 1971), ktoré boli častým typom rozvíjajúcej sa mobilnej zóny. E. BONČEV (1976) podobné štruktúry s priestorovo-fixovanou subsidenciou označil ako „lineament-geosynclinal zones". Tiež Z. P. ZONENŠAJN (1972), ktorý považuje geosynklinály za ensimatické, predpokladá, že počiatočným štádiom geosynklinály medzi dvoma kontinentmi bol oceánsky rift.

Problémom je i typ kôry pod geosynklinálami. A. V. PEJVE et al. (1972) uvádza, že eugeosynklinály boli vždy ensimatické alebo oceánske. Ensimatický charakter mobilných trógov (intrageosynklinál) je veľmi pravdepodobný. Pri existencii geosynklinálneho režimu sú však okupované i bloky alebo masívy sialickej kôry, najmä pri okrajoch starších kontinentov (teda i nukleov sialickej kôry — mikrokontinentov). Nie je známe, či mobilizácia okrajov kontinentálnej kôry nepostihuje predovšetkým pásy alebo bloky so simatickou kôrou (napr. v predpolí Západných Karpát). Pod vonkajšou časťou Východných Karpát predpokladal H. STILLE (1953) okrajovú depresiu v páse zelených bridlíc (depresiu možno zaradiť k ensimatickým panvám).

Klasifikácia panví geosynklinálneho modelu sa stretávala s problémami v časovom a priestorovom vymedzení jednotiek, s neúplnosťou geosynklinálneho cyklu a s početnými odchýlkami od tohoto cyklu. Model predpokladal založenie geosynklinál na kontinentálnej kôre, ktorá mala podľa mnohých geológov celoplanetárne rozšírenie (napr. V. V. BELOUSOV 1954, 1962 a i.). Model opomínal často historicky premenlivý vývoj kôry, alebo vývoju prisudzoval jediný trend: buď sializáciu, alebo bazifikáciu (V. ŠKVOR a J. ZEMAN 1976). Preto bol geosynklinálny model kritizovaný s návrhmi opustiť ho (P. I CONEY 1970, A. D. MIALL 1984), alebo zásadne revidovať. Diskusie o tomto probléme pretrvávajú (napr. J. CHÁB et al. 1984).

Kompromis geosynklinálneho modelu s doskovou tektonikou sa premieta do jednoduchej klasifikácie geosynklinál J. R. CURRAYA (1978), ktorý rozlišuje geosynklinály a s nimi spojené panvy:

a) Vo vnútri litosferických dosiek, kde subdukcia neexistuje. Panvy zaraďuje k nevyvinutému oceánu (cf. Z. P. ZONENŠAJN 1972 a A. V. PEJVE et al. 1972).

b) Na okrajoch dosiek, kde dynamiku a dobu trvania sedimentačnej panvy určuje subdukcia.

*Klasifikácia podľa doskovej tektoniky*

Tieto klasifikácie vychádzajú zo základných kritérií, ktorými sú: vývoj a typ kôry, pozícia v doskách, vzťah k okrajom dosiek a ich dynamike (W. R. DICKINSON 1971). V podmienkach neogénnych karpatských panví možno pri aplikácii klasifikácií riešiť vzťah panví k blokom či megablokom, pretože termín „doska" z dôvodu malých rozmerov geologických jednotiek neprichádza do úvahy (F. ČECH 1982).

Okrajové panvy sú viazané nielen na kontinentálnu, ale i na prechodnú kôru. Pásmo súčasných pasívnych okrajov zaberá 40 % alebo 200,7 mil. $km^3$ sedimentov, ak akceptujeme predpoklad L. E. LEVINA (1984), že sedimentárna výplň všetkých panví na zemi (bez miogeosynklinál) je celkove 510 mil. $km^3$. Prechodná zóna k fosílnym okrajom, tzv. Mezogea zaberá 6,5 % alebo 33 mil. $km^3$, prechodná zóna od pasívnych okrajov k recentným aktívnym okrajom zaberá 7,5 % alebo 38 mil. $km^3$ a na recentné aktívne okraje, vrátane kolumbijského (kolumbijský typ prechodnej zóny V. V. BELOUSOVA 1982) pripadá len 4,8 % alebo 24,3 mil. $km^3$. Panvy spojené s atlantickým typom okrajov kontinentov (divergentným typom v termínoch doskovej tektoniky) sú zastúpené najväčším podielom, pretože dynamické podmienky pre ich vznik tu boli najvhodnejšie.

Použitie klasifikácií vyžaduje znalosť hlbinnej stavby, paleogeografie a historického vývoja panví, keď sa menil typ panvy. Praktický význam je v poznaní procesov zmien akumulácie a migrácie uhľovodíkov a tvorenia produkčného potenciálu panví.

Obecnú klasifikáciu sedimentárnych panví vzhľadom na ich ložiskový potenciál vypracoval A. W. BALLY a S. SNELSON (1980). Klasifikácia rešpektuje rôzne typy fosílnych a recentných okrajov dosiek s oceánskou i kontinentálnou kôrou. Autori zdôrazňujú nutnosť študovať históriu panví a ich podložia, pretože v podloží mohli, alebo nemuseli byť príčiny subsidencie, napr. v mieste paleoriftov alebo paleosutúr (vrásových pásiem alebo lineamentov ako reliktu bývalej konvergencie dosiek). A. W. BALLY a S. SNELSON (l. c.) rozlišujú tri základné kategórie panví a v nich celkove 21 typov panví. Hlavným kritériom klasifikácie je ich pozícia vo vzťahu k okrajom dosiek a ďalej tiež k charakteru (rigidite) litosféry. V mobilných pásmach rozlišujú perisutúrové panvy na okraji mobilného pásma a episutúrové panvy nachádzajúce sa v mobilnom pásme. Rozlišuje sa, podobne ako v ďalších klasifikáciách, pri konvergentnom styku dosiek A-subdukcia a B-subdukcia. Subdukcia zasahujúca pozdĺž strmo uklonenej Benioffovej zóny až do hĺbky 600 km sa označuje ako B-subdukcia. Nehlboká príkrovová subdukcia vnútrokôrová alebo subdukcia vo vnútri litosféry pozdĺž plochej zóny je označená ako A-subdukcia (alebo alpínska subdukcia) na počesť O. AMPFERERA, ktorý rozsiahle presuny kontinentálnej kôry po prvý raz pre vznik Álp interpretoval.

Klasifikácia panví podľa A. B. BALLYHO a S. SNELSONA (1980):

1. Panvy na rigidnej litosfére, nespojené s formáciami megasutúr.

1.1. Vo vzťahu k oceánskej kôre:
rifty,
panvy spojené s transformnými zlomami,
oceánske abysálne plošiny,
panvy pri pasívnych okrajoch atlantického typu (ležiace na starších riftoch, transformných zlomoch a starších zaoblúkových panvách).

1.2. Nachádzajúce sa na predmezozoickej kontinentálnej litosfére, kratonické panvy (na starších riftoch alebo pôvodných zaoblúkových panvách).

2. Perisutúrové panvy na rigidnej litosfére (s formáciou kompresnej megasutúry).

2.1. Oceánske lineárne depresie a priekopy na oceánskej kôre alebo prechodnom kôrovom type pri B-subdukcii.

2.2. Predhlbne a podložné platformné sedimenty alebo grabeny pri A-subdukcii.

2.3. Panvy čínskeho typu viazané na zlomy vo zvrásnených pásmach bez A-subdukcie.

3. Episutúrové panvy spojené s kompresnými megasutúrami.

3.1. Spojené s B-subdukciou,
predoblúkové panvy,
cirkumpacifické zaoblúkové panvy (na oceánskej kôre nad B-subdukciou a na kontinentálnej kôre alebo kôre prechodného typu nad B-subdukciou.

3.2. Zaoblúkové panvy pri kolíznom styku na konkávnej strane oblúku A-subdukcie,
panvy typu panónskej panvy na kontinentálnej kôre,
panvy západomediteránneho typu na oceánskom alebo prechodnom type kôry.

3.3. Panvy viazané na episutúrový systém veľkých strižných zón panvy typu Great basin,
panvy kalifornského typu (tj. napr. typu pull-apart — pozn. autora).

Základné princípy klasifikácie panví A. W. BALLYHO a S. SNELSONA (1980) akceptoval do upravenej klasifikácie A. D. MIALL (1984). Klasifikácia A. D. MIALLA (l. c.) kladie väčší dôraz na dynamiku transformných a transkurentných zlomov, s ktorými spája vznik niekoľkých typov panví. Vychádza z modelu Wilsonovho cyklu, t. j. z intervalu vývoja litosféry od otvorenia po uzatvorenie oceánu. Pozícia panví vo vzťahu k okrajom dosiek sa ako klasifikačné kritérium na rozdiel od klasifikácie A. W. BALLYHO a S. SNELSONA (l. c.) dostáva až na druhé miesto. Prvým kritériom je správanie sa doskového okraja.

Panvy sú rozdelené do 5 kategórii s početnými typmi, ktoré si môžu byť po geologickej stránke podobné, ale vznikali v rôznych štádiách Wilsonovho cyklu. Klasifikačné kritériá W. R. DICKINSONA (1971, 1974), t. j. typ kôry, pozícia panvy vo vzťahu k okraju dosky a typ interakcie dosiek, ako aj vzdialenosť tejto interakcie od sedimentárnej panvy, sa podľa A. D. MIALLA (l. c.) v čase menili. Napr. panvy, ktoré vznikli na divergentnom okraji dosiek, sa v závere Wilsonovho cyklu pri kolízii dostanú do pozície konvergentnej, napr. predhlbňovej.

Klasifikácia panví podľa A. D. MIALLA (1984):

1. Panvy divergentných okrajov,
rifty: riftované oblúkovité panvy, okrajové depresie, okrajové oceánske panvy: „mladé“ typu Červeného mora a „zrelé“ atlantického typu,
aulakogény a odumreté rifty.

2. Panvy konvergentných okrajov,
oceánske priekopy a subdukčné komplexy,
predoblúkové panvy,
medzioblúkové a zaoblúkové panvy,
panvy na predpolí oblúkov.

3. Panvy na transformných a transkurentných zlomoch, panvy s pozíciou:
pri hraniciach dosiek tvorených transformnými zlomami,
pri divergentných okrajoch tvorených transformnými zlomami,
pri konvergentných okrajoch s transkurentnými zlomami,
pri sutúrach s transkurentnými zlomami;
typ panvy:
panvy v pásme spriahnutých zlomov,
panvy pri koncoch zlomov,
panvy typu „pull-apart" na kulisovitých zlomoch.

4. Panvy vzniknuté pri kontinentálnej kolízii a na sutúrach po kolízii;
panvy na predpolí (napr. na platforme — pozn. autora),
predhlbne a periférne depresie,
intrasutúrové zálivové panvy (zvyšky oceánskych panví),
panvy sprevádzajúce panvy na transkurentných zlomoch.

5. Kratonické panvy.

D. M. Curtis (1980) rozlišuje podľa tektonickej pozície panvy:

1. kratonické s vnútornou pozíciou alebo s pozíciou na pasívnych okrajoch kontinentov,
2. riftogénne,
3. okrajové na aktívnych okrajoch, spojené s A-subdukciou a s B-subdukciou,
4. panvy pri zlomoch s horizontálnymi posunmi.

Na pozícii panvy záviseli hodnoty tepelného toku, charakter cirkulácie povrchových a hlbinných vôd v dobe sedimentácie a obsah sulfosolí.

1. P a n v y v o v n ú t r i d o s i e k majú malú hrúbku sedimentov a veľmi nízky tepelný tok, čiže faktory nie vždy priaznivé pre vznik uhľovodíkov. Geomorfologická členitosť povrchu však mohla ovplyvniť klimatické podmienky v prospech akumulácie a zachovania organickej hmoty. Priaznivým faktorom je veľká plocha panví. Príkladom sú západosibírske roponosné panvy.

2. R i f t y sa nachádzajú vo vnútri i na okraji dosiek, v pásmach na prechode kontinentálnej kôry na typ prechodný alebo oceánsky. Vnútri dosiek sú kontinentálne rifty a tektonické grabeny.

Pri otváraní riftu dochádza k roztrhnutiu kontinentu a ku vzniku divergentného okraja. Po roztrhnutí zostávajú pri okraji kontinentu na zlomoch grabenové depresie, ktoré ostali oddelené (odtrhnuté) od aktívneho riftu a stávajú sa súčasťou mladého pasívneho okraja kontinentu (atlantického typu). Tieto panvy sa označujú ako „pull-apart basins" (obr. 16). V týchto panvách je obmedzená cirkulácia vody, niekedy i vysoký tepelný tok a panvy sú prevažne vhodné pre zachovanie a dozrievanie organickej hmoty. Množstvo organickej hmoty samozrejme závisí na komunikácii panvy s morom — oceánom.

3. P a n v y n a a k t í v n y c h o k r a j o c h d o s i e k — k o n t i n e n t o v. O panvách na divergentných okrajoch, riftoch a panvách typu „pull-apart" sme sa už zmienili. Ich okrajová pozícia je druhotná. Panvy na konvergentných okrajoch v kompresnej zóne sú viazané na kolízne zóny typu kontinent-kontinent (A-subdukcia) alebo oceán-kontinent (B-subdukcia). Okrem týchto sa rozlišujú panvy viazané na transformné zlomy.

Panvy v pásme A-subdukcie dvoch kontinentálnych dosiek reprezentujú aulakogény, predhlbne a medzihorské depresie.

Aulakogén v súlade s pôvodnou definíciou N. S. ŠATSKÉHO (1955) je graben buď prekrytý čiastočne orogénom, alebo ústiaci do predhlbne (obr. 18). Tým sa líši od „pseudoaulakogénu". Jeho konfigurácia s orogénom v poňatí konvergentného okraja doskovej tektoniky vyplýva z obr. 18. Podľa K. C. A. BURKA a J. F. DEWEYA (1973) aulakogén predstavuje „nevydarený pokus" o otvorenie riftu. Formuje sa ako subsidenčná panva hlavne počas uzatvárania kolízneho štádia alebo v ogorénnej etape (dvíhanie horstva). V tomto poňatí vychádzajúcom z definície K. C. A. BURKA a J. F. DEWEYA (l. c.) sa aulakogény nemôžu vyskytovať na divergentnom (atlantickom) okraji kontinentu.

Predhlbne v počiatočnom štádiu môžu byť miogeoklinálou so šelfovým základom. Ak ostal zachovaný pasívny okraj kontinentu atlantického typu, k zvrásneniu nedôjde. Ak sa zóna subdukcie presúva od oceánu ku kontinentu, alebo nastáva kolízia kontinent — kontinent, potom dochádza ku vrásneniu. Druhý prípad je z hľadiska dynamického pre zachovanie ložísk uhľovodíkov menej priaznivý, ale priaznivejší je v porovnaní s B-subdukciou, z hľadiska menšieho tepelného toku.

Medzihorské panvy sú konečným štádiom vývoja predhlbňových panví a ich rozpadu na postorogénne depresie. Tento typ panví je veľmi perspektívny z hľadiska výskytu uhľovodíkov, pretože panvy majú nízky tepelný tok, rýchlu subsidenciu a uzavretú cirkuláciu podzemných vôd. K tomuto typu panví možno zaradiť aj viedenskú panvu.

Panvy spojené s B-subdukciou sú pre ložiská uhľovodíkov väčšinou nepriaznivé a nemajú u nás analógie. Zvláštnosťou sú panvy pri pacifickom okraji kontinentu, ktoré mali v ranom štádiu ensimatický charakter a až neskôr sa stali panvami ensialickými.

4. Panvy na horizontálnych posunoch pri konvergentných okrajoch (pri divergentných okrajoch ide o typ „pull-apart"), majú podobný charakter ako pri B-subdukcii. Ich vznik je tiež s touto subdukciou spojený.

Podľa D. M. CURTISA (1980) sú pre ložiská uhľovodíkov optimálne riftovité grabeny, aulakogény, pseudoaulakogény, a to pre obmedzenú cirkuláciu vôd i uhľovodíkov a pre nižší tepelný tok. Veľké klesanie zaisťuje zachovanie sedimentov pred deštrukciou. Ďalej sú priaznivé tiež medzihorské panvy, miogeoklinály na pasívnych okrajoch dosiek, panvy typu „pull-apart" a niektoré typy predhlbní.

Klasifikáciu D. M. CURTISA (1980) v zjednodušenej modifikácii aplikoval A. PERRODON (1980). Podľa pozície v troch základných geoštruktúrnych provinciách rozlišuje panvy:

1. Platformovej provincie — panvy stabilnej platformy, panvy nestabilnej platformy, panvy karbonátových komplexov, vrátane rifov.

2. Provincie prelomov (grabenov) — riftové panvy, panvy pasívnych okrajov kontinentov (dosiek), delty na okrajoch kontinentov.

3. Orogénnej provincie — intramontánne panvy, panvy na horizontálnych posunoch, panvy na konvergentných okrajoch dosiek, vrátane predhlbní.

Klasifikácia D. N. CURTISA je výstižnejšia, lebo väčšmi rešpektuje škálu geotektonických podmienok, pri ktorých vznikajú kontinentálne panvy v rôznych endogénnych režimoch.

H. D. KLEMME (1980) vypracoval pre potreby prospekcie a prieskumu ložísk uhľovodíkov klasifikáciu s charakteristikou základných ložiskových parametrov. Základom klasifikácie je pozícia panví na kontinente a charakter sedimentárnej výplne, ktorá je rozhodujúca pre uhľovodíkový potenciál.

H. D. KLEMME (l. c.) rozlíšil 8 typov panví s intra- a extra- kontinentálnou pozíciou.

a) Intrakontinentálne panvy sú: vnútorné jednoduché (typ 1), vnútorné zložité (typ 2) s hiátmi alebo diskordanciami. Zvlášť zložitý je podtyp 2A s prepadlinovou stavbou spodnej etáže (viac grabenov alebo riftovitých štruktúr) a vrchná etáž má stavbu jednoduchú — platformovú (obr. 19).

Rifty — typ 3 majú analógie s podtypom 2A. Rifty však majú až 4 krát menšiu šírku.

b) Extrakontinentálne panvy, okrajové, sú najpočetnejšie: Zvláštnym typom (4) sú panvy „prelomové" (downwarp basins), ktoré sú na prechode k okrajovým moriam. Sú uzavreté (podtyp 4A), alebo majú charakter trógu (4B), alebo sú otvorené smerom k oceánu.

Panvy typu „pull-apart" (5) sú orientované paralelne alebo priečne k pasívnemu okraju kontinentu. Rané štádium môže byť riftové — charakteru satelitného riftu.

Panvy viažuce sa na subdukciu a ostrovné oblúky — typ 6, s podtypmi 6A — predoblúkové a 6B — zaoblúkové panvy.

Ďalšiu skupinu tvoria:

Vnútorné panvy (typ 7) vznikajúce počas orogenézy po skončení subdukcie; v podstate sú medzihorskými panvami.

Deltové panvy (typ 8) sú vlastne deltové kužele vrstvené na atlantickom type kontinentálneho okraja.

S divergentnými okrajmi dosiek sú spojené typy 3, 5 a 8, s konvergentnými typy 6 a 7.

Od typu 1 k typu 8 klesá vek panví (doba vzniku), ďalej tiež podiel štruktúrneho vplyvu podložia na sedimentárnu výplň, rastie podiel klastických sedimentov a klesá podiel karbonátových hornín (s výnimkou typu 8, ak je delta na šelfe). Klesá tiež veľkosť panví a rastie podiel príbrežnej plochy.

V panvách typu 1 a 2 sa nachádzajú veľké pasce s gigantickými akumuláciami uhľovodíkov. Pri ostatných typoch panví je malá nádej na objav veľkých ložísk. Vyšší stupeň zrelosti materských hornín majú typy 1 až 4B, nižší stupeň typy 4C—8. V panvách s pokračujúcou subsidenciou, napr. pri typoch 3, 6, 7, pokračuje i diagenéza a dochádza k premenám v migrácii uhľovodíkov a v cirkulácii vody. Hlavné údaje o faktoroch produktívnosti panví sú uvedené v tab. 3.

Klasifikácia vychádzajúca z doskovej tektoniky je väčšinou porovnateľná s typmi panví v geosynklinálnom modeli (tab. 4). Z porovnania vyplýva, že podrobnejšie členenie okrajových depresií rešpektovali klasifikácie v termínoch doskovej tektoniky.

Na rozdiel od vyššie uvedených jednoduchších klasifikácií panví s uhľovodíkmi vypracoval D. R. KINGSTON et al. (1983) globálnu klasifikáciu rešpektujúcu obecné i korelačné znaky. Podobne ako u A. D. MIALLA (1984) sú pri rekonštrukcii typu a histórie východiskom tri základné parametre:

a) tektonika podmieňujúca vznik panvy,
b) sedimentárna sekvencia,
c) tektonika modifikujúca panvu.

Klasifikácia je založená na modele doskovej tektoniky. Panvy môžu byť jednoduché s jedným alebo dvoma tektonicko-sedimentárnymi cyklami alebo ako panvy zložité môžu mať zložitú históriu s niekoľkými rozdielnymi cyklami a udalosťami — ako tzv. polyhistorické panvy.

V tejto klasifikácii je rozlíšených osem základných typov cyklov s ich sedimentárnymi komplexmi. Klasifikácia je založená na štúdiu dát z približne 600 identifikovateľných svetových sedimentárnych panví.

Pre stanovenie tektoniky podmieňujúcej vznik panvy treba analyzovať sedimentárno-tektonické cykly zaregistrované vo vrstevnom profile každej panvy. Jednotlivé cykly kontinentálnej, príbrežnej a morskej sedimentácie môžu byť oddelené diskordanciami. Ku kontinentálnemu počiatočnému štádiu cyklu sú zaradené i sedimenty oscilujúceho pobrežia, s limitom 50 % morských sedimentov. Vývoj 2. štádia cyklu ide cez morskú fáciu k nemorskému vrcholovému štádiu, s obsahom viac než 50 % nemorských sedimentov.

V tektonike podmieňujúcej vznik panvy sa rozlišuje:

a) Typ kôry pod panvou — kontinentálny alebo oceánsky; prechodný typ sa ťažko stanovuje, a preto sa do klasifikačných parametrov nezahŕňa.

b) Typ pohybu dosiek v minulosti pri formovaní panvy — konvergentný alebo divergentný. Na rozdiel od A. D. MIALLA (1984) nerešpektuje ako samostatný typ transformný. Konvergentné pohyby môžu byť prenášané do vnútra dosiek po horizontálnych posunoch.

c) Pozícia panvy na doske (vnútorná, okrajová) a primárne zlomové horizontálne posuny.

Tieto kombinácie sú základom k 10 teoretickým a 8 praktickým modelom jednoduchých panví s uhľovodíkmi. Z hľadiska prieskumu na ropu a zemný plyn sa hlavné produkčné panvy nachádzajú na kontinentálnej kôre vo vnútri i na okraji a sú viazané na zlomy. V tejto skupine boli tiež objavené v posledných rokoch najvýznamnejšie panvy s ložiskami uhľovodíkov.

V klasifikácii sa rozlišujú kontinentálne panvy (s prechodom k oceánu) a oceánske panvy na konvergentných i divergentných okrajoch.

1. Panvy (sedimentárne cykly) divergentných okrajov:

a) Vnútorné panvy — vklesliny (IS); osi panví voči kontinentálnemu okraju zvierajú vždy uhol. Majú oválny až kruhový tvar a malú hrúbku sedimentov. Vznikli synklinálnym prehnutím — vklesnutím kôry.

b) Vnútorné panvy na zlomoch (IF); môžu byť i na fosílnych okrajoch dosiek, ktoré sa teraz nachádzajú vo vnútri kontinentálnej dosky. V charaktere ich geologickej stavby prevažujú hraste a grabeny. Vznikajú tenziou v začiatkoch divergentných pohybov. IF typ sa môže zmeniť na MS typ.

c) Okrajové panvy — vklesliny (MS); osi panví sú paralelné s okrajmi kontinentu atlantického typu a sedimenty môžu prechádzať do oceánskej kôry. Panvy sú polyhistorické. Spodná etáž panví má hrasťovito-grabenovú stavbu (IF typ), vrchnú etáž tvoria napr. sedimenty kontinentálneho svahu alebo delty. Na niektorých zlomoch môžu intrudovať bazalty, ako štádium počiatočného otvárania oceánu.

2. Panvy (sedimentárne cykly) konvergentných okrajov: Do tejto katogórie sú zaradené panvy na okraji alebo blízko okraja dvoch divergujúcich dosiek. Väčšina panví má tenzné črty.

a) Panvy na zlomoch s horizontálnymi posunmi — strižné panvy (LL), vznikajú horizontálnymi pohybmi krýh na zlomoch. Majú rovnaký charakter ako panvy typu „pull-apart". Pre výskyt ložísk uhľovodíkov majú význam tie panvy tohoto typu, ktoré ležia na kontinentálnej kôre.

b) Panvy viazané na oceánske priekopy (T a Ta); panvy tohoto typu sú pre výskyt ložísk uhľovodíkov veľmi málo perspektívne.

c) Oceánske panvy — vklesliny na horizontálnom posune na okraji zlomov (OSLL); do tejto skupiny patria malé oceánske panvy a okrajové moria na oceánskej alebo prechodnej kôre, v pásme konvergencie dosiek.

Pozn. Skratky sú odvodené z anglického označenia typov panví: IF — interior fracture basin, MS — margin sag basin, IS — interior sag basin, LL — lateral — wrench basin atď.

Klasifikácie vychádzajúce z doskovej tektoniky sú väčšinou porovnateľné s typmi panví v geosynklinálnom modeli (tab. 4).

A. W. BALLY and S. SNELSON (1980) nepredpokladajú heterogenitu typov kôry vo vnútri kontinentálnej kôry, a tak vylučujú i hlavnú príčinu dlhodobej opakovanej subsidencie — dedičnosť mobility. Princíp dedičnosti vzťahujú len na prv alebo recentne aktívne okraje dosiek a menej berú do úvahy vnútrodoskovú mobilitu. Táto nemusí byť len následkom existencie zlomov, sutúr alebo megasutúr, ale i odlišného typu kôry. Uvedená klasifikácia neuvažuje ani o restitoch oceánskej či prechodnej kôry v dôsledku nedokončeného (embryonálneho) otvorenia, či nie úplne uzatvoreného oceánu (napr. odštiepnutej alebo obdukovanej oceánskej kôry). Kladom tejto a aj ostatných klasifikácií je historický aspekt vývoja príslušného úseku litosféry s panvou. Tento aspekt umožňuje riešiť podmienky a dynamiku vzniku a migrácie uhľovodíkov v produkčných panvách.

Ku klasifikácii panví vypracovanej A. D. MIALLOM (1984) možno mať podobné pripomienky ako ku klasifikácii A. W. BALLYHO a S. SNELSONA (1980). Ani Miallova klasifikácia neuvažuje o fyzikálno-chemických premenách kôry, a o interakcii litosféry s astenosférou len okrajovo alebo ako následku pohybu dosiek, ako je to obvyklé v doskových modeloch. Odmietanie termínu medzihorská a vnútrohorská panva môže viesť k omylom v paleogeografických rekonštrukciách, v oceňovaní ložiskového potenciálu panví a v stanovení prognóznych oblastí. Precenenie horizontálnych pohybov a aplikácia extrémnej dynamiky môže viesť k nesprávnemu oceneniu podmienok tvorby i migrácie uhľovodíkov a napr. i k vylúčeniu predpokladu existencie uhľonosných vrstiev. Nakoniec treba pripomenúť, ako obťažné, diskutabilné a často špekulatívne je riešenie charakteru fosílnych okrajov dosiek, kontinentov a existencie hypotetických uzavretých oceánov.

Panvy zahrnuté do klasifikácie D. R. KINGSTONA et al. (1983a) obsahujú nielen ložiská uhľovodíkov, ale i uhlia, solí a iných nerudných surovín. Väzba jednotlivých surovín na štádia cyklov je evidentná. Navyše, sedimenty jednotlivých štádií cyklov majú buď produkčné vlastnosti, alebo sa stanú behom vývoja panvy litologickými štruktúrnymi pascami. Historická analýza vývoja panví umožňuje oceniť podmienky vzniku uhľovodíkov, ich migrácie a akumulácie v pascách. Preto je dôležité študovať i nadväznosť súčasného typu panvy na jej geologickú minulosť.

V porovnaní so všeobecnou klasifikáciou panví A. W. BALLYHO a S. SNELSONA (1980) a A. D. MIALLA (1984) je klasifikácia D. R. KINGSTONA et al. (1983a) jednoduchšia. Pri panvách so zložitým vývojom lepšie podchytáva ich históriu, umožňuje zahrnúť do jednotnej nezložitej schémy všetky panvy na svete, pokiaľ ovšem generalizujeme ich špecifické rysy a pokiaľ vylúčime prechodné typy kôry. Vypustenie riftov z panvových typov nie je vydarené, pretože rifty sú významnou kategóriou panví so špecifickými rysmi hlbinnej stavby a litosferickej dynamiky. V klasifikácii iba ťažko nachádzajú svoje miesto panvy medzihorského typu a mediteránneho typu, u ktorých plášťové diapíry nemožno spájať s pohybmi po horizontálnych posunoch (strike — slip faults). Z dynamického hľadiska nie je správne, že medzi tektonickými modifikátormi nie sú vertikálne pohyby na zlomoch, ktorých evidencia je preukaznejšia než pre horizontálne posuny, najmä potom pre horizontálne posuny skryté pod mladšími sedimentmi.

## Neogénne panvy v alpínsky mobilnej Európe

Karpatské neogénne panvy sú súčasťou mediteránneho a cirkummediteránneho pásma, kde pásmový orogén obtáča medzihorské depresie oválneho až izometrického (kruhového) tvaru. Vo vnútri orogénu sú úzke väčšinou lineárne vnútrohorské panvy (obr. 55).

Kôra pod panvami má menšiu hrúbku (až o 100 %) v porovnaní s hrúbkou kôry pod susednými orogénmi. Morské panvy sú súčasťou Stredozemného mora (Tyrrhénske, Egejské, Iónske) alebo tvoria vnútrozemské moria: Čierne more a juhokaspická depresia. Nemorské panvy tvoria severné cirkummediteránne pásmo: panónska panva, transylvánska panva, moesijská platforma — depresia, perikaspická depresia.

### Problematika Stredozemného mora a periférnych panví

Vznik oboch typov panví súvisí so vznikom Stredozemného mora; rozvoj maximálnej subsidencie časovo zapadá do konca neogénu a do kvartéru. Vznik Stredozemného mora je stále predmetom rôznych modelov, v ktorých sa uvažuje o reliktnej alebo novotvorenej oceánskej kôre.

a) Reliktnú oceánsku kôru (zvyšok Paleotetýdy) predpokladá v podloží morí A. V. Pejve (1969), L. Glangeaud (1970); A. Faure-Muret s G. Choubertom (1975) predpokladajú, že došlo k odtrhnutiu sialickej kôry, jej driftu a na mieste dnešného mora zostala bazaltová vrstva.

b) Vznik novej oceánskej kôry predpokladá napr. D. Mc Kenzie (1972) alebo J. F. Dewey et al. (1973) a ďalší, a to v dôsledku rozpínania morského dna. Opakované otváranie a uzatváranie Tetýdy bez geologických dokladov predpokladá J. F. Dewey et al. (l.c.), K. Birkenmajer (1976) a A. Tollmann (1978). Dokladom novotvorenej oceánskej kôry majú byť mezozoické ofiolity pri pobreží (A. T. Aslanjan a M. A. Satjan 1984).

Vznik Stredozemného mora „metasomatickou bazifikáciou" predpokladal R. W. van Bemmelen (1972), bazifikáciou vplyvom intrúzií a efúzií bázických a metabázických hornín M. V. Muratov (1975) a V. V. Belousov (1982), eklogitizáciou E. V. Artjuškov a M. A. Beer (1983).

Špecifičnosť vzniku morských depresií, ktoré nemožno riešiť mechanizmom vzniku oceánov, zdôrazňujú i ďalší autori (I. V. Archipov 1984, M. A. Beer 1984 a iní). Vplyv plášťových diapírov na subsidenciu zdôrazňuje M. E. Artemjev (1975), B. Biju-Duval et al. (1976), J. P. Malovickij et al. (1982) — (obr. 24), J. I. Nikolskij (1982) — (obr. 22), F. Čech a J. Zeman (1985) — (obr. 20). Rozdiely medzi jednotlivými autormi sú v predpoklade existencie pôvodne kontinentálnej kôry, resp. pôvodne heterogénnej, viac-menej simatickej stenčenej kôry (F. Čech a J. Zeman l. c.).

Názory na riftogénny pôvod, najmä západomediteránnej oblasti, uvádzajú: B. Biju-Duval et al. (1976) — za spoluúčasti plášťových diapírov, K. J. Hsü (1977) a J. M. Lort (1977). Via Gian Battista (1984) predpokladá, že riftogenéza bola vedúcim procesom až do kambria, čo dokladá na základe geologickej stavby cirkummediteránnej oblasti.

Medzihorské panvy mimo Stredozemné more majú niektoré zhodné črty so Stredozemným morom, najmä tenkú kôru (obr. 28) a rýchlu subsidenciu od konca neogénu. Tieto panvy sú lemované okrajovými depresiami, kde maximálna subsidencia časovo predchádzala maximálnu subsidenciu v medzihorskej depresii (tab. 4, 5, obr. 36). K týmto depresiám treba zaradiť i tzv. moesijskú platformu, ktorá má všetky znaky depresie (moesijsko-pontická depresia — F. Čech a J. Zeman 1985) — (obr. 36, tab. 6).

Údaje o hlbinnej stavbe morských a nemorských medzihorských panví sa opierajú o nerovnomerné geofyzikálne výskumy. Tie ukazujú, že v stenčenej litosfére s aktívnymi plášťovými diapírmi je anomálne vysoký tepelný tok (tab. 6 a 7, obr. 26, 27, 29, 31, 32, 33). Rýchlosti seizmických P-vĺn sú nižšie než v okolitom normálnom plášti. Anomálny plášť je aseizmický. Tiažové pole pod panvami je kladné (obr. 29, 31). Panvy ležia v regiónoch porušených hlbinnými zlomami (obr. 30). Hlbinné zlomy môžu mať vplyv na intrúzie plášťových diapírov (V. B. Sollogub et al. 1984), alebo ich vznik môže byť diapírmi podmienený, a to najmä pod okrajom diapíru (zóny tangenciálnych strižných napätí). Zlomy ovplyvnili i stavbu, resp. tvar vnútrohorských depresií a panví eliptického (obr. 36), či izometrického tvaru (obr. 25, 34, 35). V. S. ŽURAVLEV (1972) perikaspickú a panónsku panvu uvádza ako príklad hexagonálnych panví (obr. 25).

## Indície hlbinnej stavby panónskej panvy

Karpatsko-balkánska oblasť je geofyzikálne pomerne dobre známa, ale hlbinná preskúmanosť je nerovnomerná (obr. 31, 32, 33). V panónskej panve je dobrá korelácia medzi kladným tiažovým poľom (s hodnotami i vyše 400 $\mu m^{-2}$) a hrúbkou kôry dosahujúcou i menej než 24 km (E. MITUCH — K. POSGAY 1972) — (obr. 29). Hrúbka kôry narastá smerom k okraju panónskeho plášťového diapíru, ktorý je lemovaný orogénnym prstencom. Jeho existenciu preukazuje regionálna i reziduálna záporná tiažová pásmovitá anomália. Podľa seizmických údajov (L. STEGENA et al. 1975, L. P. VINNIK et al. 1975) je hustota plášťa pod panónskou panvou menšia než pod orogénom a platformou v predpolí. Na bázy kôry je prechodná zóna typu zmesi kôra — plášť, hrubá 2—4 km (V. LĂZĂRESCU et al. 1983). V plášti je elevácia astenosféry, vystupujúca do hĺbky 60—50 km. Plášťové elevácie zasahujú i pod juhoslovenskú a východoslovenskú panvu, kde je tiež vysoký tepelný tok (90—110 $mWm^{-2}$, V. ČERMÁK 1979). Pri okraji panónskeho diapíru prebiehajú tangenciálne orientované zlomy s častou seizmickou aktivitou, ako to vyplýva z údajov O. FUSÁNA et al. (1979) alebo V. KÁRNIKA (1975) — (obr. 31, 34). Významné je zistenie tangenciálne orientovaného jz. pokračovania peripieninského lineamentu na línii Semmering — Verona (A. ZÁTOPEK — B. BERÁNEK 1974) — (obr. 30). Podložie panví je najlepšie známe v panónskej panve. M. MAHEĽ (1978) preukázal, že slabo hercýnsky konsolidované oblasti podliehali už od mezozoika klesaniu, niekedy spojenému i s bázickým vulkanizmom, pásma s granitoidmi boli stabilnejšie. Podložie je členené hlbinnými zlomami smeru SV—JZ s dominantnou balatónskou a záhrebsko-kulčskou líniou (G. WEIN 1969) oddeľujúcou silne mobilný priestor na SZ od stabilnejšieho na JV (obr. 34, 37).

V podloží panónskej panvy boli vymedzené bloky kôry líšiace sa typom kôry: bloky sialického a simatického charakteru s rôznym stupňom hercýnskej konsolidácie kôry (obr. 24). Neogénny vývoj viedol k prestavbe blokov, najmä novovymedzeného bloku paleopanónskeho. Konzervatívnym typom mobilného bloku je blok balatónsky (obr. 38, 39).

V západokarpatskej jednotke boli spresnené typy kôry blokov — podunajského, juhoslovenského a východoslovenského, kde kryhy s vyššou mobilitou majú v podloží indície simatickej kôry.

Pri prestavbe blokov sa menila dynamická funkcia hlbších a hlbinných zlomov. Predneogénne aktívne, štruktúrne dominantné zlomy SV—JZ smeru vo vnútri panónskej panvy strácali hlbinnú spojitosť koncom miocénu. Zachovali si ju len zlomy periférne — rábska línia a línia Darnó. Štruktúrne sa začali uplatňovať smery S—J, a to i v Západných a Východných Karpatoch. Tu však zlomy SV—JZ si zachovali aktivitu a funkciu hraničných zlomov blokov. Vznik panónskej panvy a okrajových depresií horizontálnymi pohybmi na zlomoch riešili F. HORVÁTH a L. ROYDENOVÁ (1981) — (obr. 40).

Pre vývoj neogénnych panví vo vnútri karpatského oblúka boli dôležité tieto udalosti:

a) Diferenciácia kôry sialického a simatického typu na rôzne mobilné pásma (paleozoikum), azda už hercýnsky členené na bloky.

b) Opakovaná mobilita v pásmach — jednotkách alebo hercýnsky konsolidovaných s kôrou suboceánskeho typu.

c) Terciérna, najmä neogénna štruktúrna prestavba s preferenciou klesania na simatickej kôre.

Dynamicky dôležité fenomény zistené výskumom hlbinnej stavby:

a) Koncentrácia zlomov strižného charakteru na okraje panónskej panvy — periférnu panónskeho plášťového diapíru.

b) Existencia dvojakého štruktúrneho plánu, paleozoicko-mezozoického a terciérneho, najmä pomiocénneho odrážajúceho vývoj panónskeho diapíru, najmä kolapsové štádium koncom terciéru. Príčinou kolapsu môže byť expanzia diapíru na vonkajšiu stranu (napr. východoslovenská panva) i uvoľnenie magmatických krbov po vulkanizme. Proti výkladu kolapsu chladnutím svedčí trvajúci anomálne vysoký tepelný tok.

c) Existencia seizmoaktívnych zlomov svedčí o pokračujúcej pohybovej aktivite v periférnych častiach panónskeho diapíru. Centrálna časť panvy má podložie takmer seizmické. V súlade s tým je sústredenie zdvihových pohybov v orogénnom prstenci panónskej panvy.

Existencia početných trógov pred vznikom panónskej panvy, ich spojitosť so simatickou kôrou (podľa M. Maheľa 1978 až oceánskou kôrou), slabá hercýnska konsolidácia ako celku a výskyty bázického vulkanizmu vedú k predstave, že kôra panónskeho segmentu bola oslabená, mala tendenciu vyššej mobility a bola už pred výstupom panónskeho plášťového diapíru tenká. Dnešná malá hrúbka kôry nie je len následok hlbinnej erózie alebo bazifikácie. Vznik diapíru kladiem do vrchnej kriedy a s jeho prvými prejavmi spájam banatitový vulkanizmus. Vývoj diapíru prispel k formovaniu a prestavbe blokovej stavby. Z geologických údajov vyplýva, že v terciéri hlavné udalosti vývoja kôry súviseli s dynamikou diapíru. V období eocén — oligocén dochádzalo k čiastkovej expanzii — satelitné prejavy vulkanizmu azda i vplyvom tvorby čiastkových diapírov v stredoslovenskej oblasti a Mátre. Obdobie báden — pliocén znamená vrcholenie vulkanizmu a prestavby blokov. Vyvrcholenie dynamiky nastalo v pliocéne, keď došlo k maximálnej subsidencii v panónskej panve a k stagnácii pohybov v periférnych panvách, v ktorých maximálna subsidencia bola predpliocénna.

Vulkanická epocha znamená ďalšiu akréciu kontinentálnej kôry a jej formovanie do panónskeho megabloku, ktorý ani po akrécii nezískal charakter orogénnej oblasti, ale stal sa simatickým vnútorným masívom (F. Čech a J. Zeman 1982). Orogénny pás obopínajúci panónsku panvu je na styku s tenkou kôrou sprevádzaný pásmom strižných, seizmoaktívnych zlomov, reprezentovaných hlavne peripieninským lineamentom. Tangenciálna orientácia zlomov k panónskemu diapíru koreluje s regionálno-geologickými a geofyzikálnymi údajmi iných analogických štruktúr a s poznatkami tektonofyzikálnymi získanými i z experimentov a modelov.

Niektoré prvky stavby panónskej panvy sa vyskytujú i u transylvánskej panvy (obr. 41). Táto sa však odlišuje štýlom tektonickej stavby sedimentárnej výplne, prevahou soľných diapírov a nízkym tepelným tokom. Z korelácie údajov o hlbinnej stavbe a fyzikálnych prejavoch panví vo vnútri karpatského oblúka, čiernomorskej a stredomorskej oblasti bolo možné zostaviť vývojový rad panví podľa intenzity endogénnej aktivity, typov kôry a hodnôt tepelného toku. Ak budeme vulkanicky aktívne depresie s plášťovým diapírom považovať za dynamicky najaktívnejšie a panvy s nízkym tepelným tokom za typy zrelé (Čierne more), neaktívne, potom môžeme zostaviť nasledujúci vývojový rad panví: Tyrrhenské more, Egejské more, panónska panva, transylvánska panva, Čierne more a juhokaspická depresia.

**Hlbinná stavba neogénnych panví vnútorných Západných Karpát a viedenskej panvy**

Podrobnejší výskum sa zameriaval na vzťah neogénnych panví Západných Karpát k hlbinnej, najmä blokovej stavbe. Podľa geofyzikálnych údajov panvy južného Slovenska nadväzujú hlbinne na panónsku panvu. Rozhranie kladného a záporného tiažového poľa Západných Karpát koreluje s odlišným typom kôry a s hranicami rôznych blokov kôry (O. FUSÁN et al. 1979). Podrobnejšie interpretácie geofyzikálnych údajov z hľadiska typov kôry viedli k čiastkovým spresneniam doterajších interpretácií rozšírenia granitov v podloží sedimentárnych panví. Na základe rekonštrukcie charakteru podložia panví boli vymedzené pásma sialickej kôry oddelené pásmami mobilnejšej simatickej kôry, a to v súlade s údajmi z hlbinných vrtov.

Vymedzenie blokovej stavby je prevzaté od O. FUSÁNA et al. (1971, 1979), pretože je v súlade s poznatkami o charaktere a vývoji panví. Panvy majú voči blokovej stavbe pozíciu vnútroblokovú a medziblokovú. Rovnaký vzťah je voči megablokom (Českého masívu, platformy atď.).

Vnútrohorské panvy ležiace vo vnútri megabloku Západných Karpát sa líšia rozsahom a stavbou výplne od medziblokových panví, ku ktorým patrí viedenská panva (na styku Českého masívu a Západných Karpát), východoslovenská panva (na styku Západných a Východných Karpát a panónskeho megabloku). Medziblokové panvy majú vysokú dlhodobú mobilitu a líšia sa i hodnotami recentných pohybov od panví vnútroblokových.

V hlbinnej stavbe podložia viedenskej (obr. 53, 57, 60) a podunajskej panvy boli zistené integrálne prvky (obr. 58) zbližujúce hlbinnú stavbu a mobilitu oboch geologicky odlišných panví (obr. 44, 45, 46). Výrazným zlomom je nesvačilsko-trnavský zlom oddeľujúci mobilnejšiu jz. časť od stabilnejšej sv. časti (F. ČECH 1980). Rozhranie krýh je významné z hľadiska akumulácie plynu a ropy. Štruktúrne, paleogeograficky, a tým i po ložiskovej stránke je významný peripieninský lineament, ktorý oddeľuje obidve panvy a svojím elevačným charakterom vytvára priaznivé akumulačné podmienky i v mobilných hlbokých kryhách oboch panví.

V podunajskej panve bolo na základe geofyzikálnych dát (obr. 46, 47) interpretované podložie panvy korigujúce staršie interpretácie O. FUSÁNA et al. (1971) — (obr. 65). Ide o koncepciu rozsahu skrytých telies granitoidov s preferenciou existencie malých telies a prevahy metabázitov v súlade s proponovanou suboceánskou kôrou v podloží panvy (F. ČECH 1980). Predpokladá sa existencia predmetamorfných depresií (Sereď—Sládkovičovo, Levice—Šurany), ktoré odrážajú slabú konsolidáciu kôry a jej predhercýnsku mobilitu. Pravdepodobne ani čiastkové depresie zaplnené ťažkými príkrovmi nevznikli hmotnosťou príkrovov, ale tvorili sa už skôr ako mobilné zóny na mobilnej kôre bez veľkých telies granitoidov. Panvy na mobilnej simatickej kôre dedia kôrovú mobilitu, a nielen zlomové členenie a pohyby. Ide o novú kategóriu dedičných panví odlišných od poňatia T. BUDAYA (1961). Dedičné panvy v tomto zmysle zahŕňajú i superponované panvy Budaya; príkladom je podunajská panva naložená na rôzne smery zlomov, ale dediaca mobilitu tenkej ťažkej kôry.

S panónskym diapírom a viac-menej simatickým typom kôry je spojená i juhoslovenská panva, ktorá však od bádenu prekonávala odlišný vývoj od panónskej panvy.

Typickou panvou dediacou mezozoickú mobilitu simatickej kôry, vrátane silného zlomového členenia, je východoslovenská panva (obr. 51, 56). Geofyzikálne indikovaný čiastkový diapír plášťa (L. POSPÍŠIL 1981) je v súlade s hlbinnou stavbou, medziblokovou pozíciou a ďalšími fenoménmi dokazujúcimi hlbinnú aktivitu (obr. 47, 48). Lokálnu mobilitu tu zvyšujú úseky kríženia hlbinných zlomov; v geologickej minulosti ju ovplyvnili i vyprázdňované čiastkové magmatické krby. Spojitosť vulkanizmu so sedimentáciou dáva tejto panve špecifické črty medzi slovenskými panvami tohoto plošného rozmeru. Štúdium neogénnych panví vo vzťahu k hlbinnej stavbe prinieslo nové genetické aspekty a impulzy pre klasifikáciu panví.

Panvy dedičné treba odlišovať podľa typov kôry. Ensialické panvy dedia štruktúrne smery podložia. Ensimatické panvy môžu byť rovnakého typu, ale môžu byť i naložené na štruktúry podložia a dedia len regionálnu mobilitu. Tieto panvy majú tiež veľké rozmery, prvý typ je rozmermi malý (vnútrohorské panvy). Ich podložie tvorí i niekoľko blokov simatickej kôry.

Superponované panvy v poňímaní T. BUDAYA (1961) sa zužujú len na panvy ensialické. Navyše ich vývoj nezávisel na vývoji intenzity vrásnenia, pretože vrásnenie prebiehalo mimo okraje panvy.

Migrácia panví prebiehala všeobecne od Z na V, s vratnou tendenciou (oživovaním) mobility v radiálnom smere — od vonkajšej časti k vnútornej, k panónskej panve. Naznačuje sa korelácia s migráciou vulkanizmu, nevýrazná je korelácia s postupom vrásnenia — panvy sa nesťahujú na vonkajšiu stranu s postupom vrásnenia k čelu Karpát. Veľkosť mobility v panvách nesúvisí s vývojom vrásnenia, ale predovšetkým s typom kôry, pozíciou v blokovej stavbe a na poslednom mieste tiež na zlomovej stavbe podložia. Pokiaľ sa všetky faktory zrátajú (panva leží na rozhraní blokov), dosahujú veľké klesanie (viedenská, podunajská, východoslovenská panva — obr. 52, 54, 59, 62, 63, 66, 67). Riešenie genézy panví z aspektu hlbinnej geológie pomohlo nájsť nové stavebné prvky spoločné pre viedenskú a podunajskú panvu (obr. 58), jednotky geneticky a tektonicky odlišné a ďalej nájsť prvky autonómne, ako je väzba na kríženie hlbinných zlomov a rozhranie rôznych typov v podloží východoslovenskej panvy.

Panvy klesajú i recentne (P. VYSKOČIL 1981, P. MARČÁK 1978). Pásma klesania korelujú vždy so zlomami (obr. 52). Vo viedenskej panve je napr. zaujímavá mobilita v smere Z—V. Vo východoslovenskej panve recentné klesanie určujú submeridionálne zlomy (obr. 54). V tejto panve je vysoký tepelný tok, ktorý môže byť pdmienený existujúcou hlbinnou aktivitou čiastkového diapíru (L. POSPÍŠIL 1981).

Na základe zhodnotenia geofyzikálnych a geologických poznatkov som sa pokúsil riešiť problém genézy panví. Vychádzajúc z medziblokovej pozície viedenskej panvy predpokladám, že sa tu v ranom predneogénnom štádiu, azda už v mezozoiku, mohol uplatniť riftogénny proces. Riftový triplet v kenozoiku by tvoril nesvačilský a vranovický graben smeru SZ—JV, grabenová štruktúra pozdĺž lednickej zlomovej zóny smeru SV—JZ a s.-j. grabenová štruktúra v mieste kútskeho grabenu (obr. 45). Viedenská panva režimom subsidencie patrí k okrajovým depresiám panónskej panvy a vznik subsidencie mohol súvisieť s vytváraním cirkumdiapírovej synklíny v astenosfére. Nie je vylúčené, že čiastkový diapirizmus plášťa existoval už v mezozoiku (riftový triplet). Počas kvartéru sa prvky grabenov obnovujú na s.-j. smeroch (obr. 61). Viedenská panva sa hypoteticky mohla vyvíjať v mieste pôvodnej riftogenézy (divergentný rys okrajov medziblokov Brunnie a Paleokarpat), potom sa stala predhlbňou, do ktorej sa sunuli príkrovy (obr. 57, 60) a napokon získala charakter skôr medzihorskej depresie medzi vonkajšími Karpatami, Alpami a orogénne aktivovaným okrajom Českého masívu (podľa doterajších názorov vnútrohorská panva). Vznik a vývoj panvy neovplyvňovali horizontálne posuny na zlomoch (typ pull-apart), ani A-subdukcia, ako predpokladá R. JIŘÍČEK (1982).

Podunajská panva sa vyvíjala pod vplyvom okraja panónskeho plášťového diapíru. Centrum maximálnej subsidencie bolo v severnej časti (typ okrajovej depresie), ale v panóne sa presunulo na juh (spoločný vývoj s panónskou panvou). V hlbšej stavbe má najmobilnejší úsek spoločné priestorové umiestnenie, ako najmobilnejší sektor vo viedenskej panve (obr. 58).

Juhoslovenská panva mala časove limitovaný vývoj a slabú mobilitu (obr. 66). Príčiny tohto vývoja mohli byť v špecifickom režime nad okrajom plášťového diapíru s neskorými vulkanickými produktmi.

Východoslovenská panva je súčasťou transkarpatskej depresie. Zložitý geofyzikálny obraz (obr. 48, 49) je podmienený nielen tektonickou členitosťou s typmi kôry, ale tiež prítomnosťou vulkanitov asociovaných s plášťovým diapírom. Tento charakter geologickej stavby je dôsledkom zložitého vývoja a typu panvy, Príčinou diferencovaného vývoja je i zložitý predneogénny vývoj a stavba podložia panvy (obr. 51, 56). Príkrovové fenomény a blokovo-šupinovitá stavba (J. ČVERČKO et al. 1984, V. G. SVIRIDENKO 1976 — (obr. 50) potvrdzujú premenlivú dynamiku oblasti od mezozoika, keď tu mohla existovať počiatočná riftogenéza. V porovnaní s viedenskou panvou majú spoločné rysy, napr. stenčenú kôru, zložitosť vývoja, inverziu a polaritu subsidencie, existenciu príkrovov, avšak mechanizmus vrásnenia a sunutia príkrovov bol odlišný. Líšia sa i geofyzikálnou charakteristikou a recentnou dynamikou.

Doterajšie poznatky o zlomovej stavbe a pohyboch na zlomoch dokladajú absolútnu prevahu vertikálnych pohybov. Koncepcie predpokladajúce horizontálne pohyby (F. HORVÁTH a L. ROYDENOVÁ 1981, Ľ. POSPÍŠIL a D. VASS 1983) nemajú geologickú evidenciu.

Slovenské panvy v tyle Karpát a viedenská panva svojím režimom a časovým vývojom subsidencie boli dynamicky zviazané s vývojom panónskeho plášťového diapíru, pričom sa nedá vylúčiť ani predpoklad existencie satelitných čiastkových diapírov pri okraji panónskeho diapíru. Problémom je spájanie maximálnej subsidencie v panvách s kolapsom diapíru. R. RUDINEC et al. (1981) preferuje ako príčinu subsidencie termálnu aktivitu. Predpokladám, že klesanie litosféry do menej viskóznej horúcej astenosferickej elevácie je opodstatnené, najmä v panvách s vysokým tepelným tokom. V klasifikačných schémach zodpovedajú neogénne panvy typu 1 až 2 v klasifikácii (D. M. CURTISA (1980) a typu 2 alebo 7 v klasifikácii H. D. KLEMMA (1980). Podunajská a juhoslovenská panva majú spoločné znaky typu 2, ale najmä 2A. Východoslovenská panva má najbližšie k typu 7, ale v terciéri mala znaky typu 6B. Najheterogénnejšia je viedenská panva. Má blízko k typu 7, spodná etáž má viac znakov typu 2A, pričom v predneogénnom vývoji sa nedá vylúčiť ani typ 3 (F. ČECH 1985). Zložitý vývoj signalizujú znaky inklinujúce k divergentnému i konvergentnému charakteru blokových (megablokových) okrajov.

V klasifikácii D. R. KINGSTONA et al. (1983) by viedenská panva mala mať vnútrodoskovú pozíciu, so subsidenciou na poklesových zlomoch. Bolo by ju možné zaradiť k typu panví IF (panvy s vnútornou pozíciou v doske na zlomoch) s grabenovou a hrasťovou stavbou. Typ, ktorý sa mohol vyvinúť z pôvodného typu MS panvy na divergentnom okraji kontinentu, pri styku kontinentálnej a oceánskej kôry. Viedenská panva má, v súlade s klasifikačnými kritériami, zložitý vývoj — hypoteticky spodná etáž sa tvorila na kontinentálnom okraji, súdiac podľa hrasťovitej a grabenovej stavby tvorenej v štádiu divergentného cyklu. Neskorší vývoj so vznikom príkrovov v podložnom komplexe má znaky konvergentného cyklu (typ LL panvy v klasifikácii D. R. KINGSTONA et al. (1983), t. j. panvy pri konvergentnom okraji, čo by mohlo byť v doskovom modele vysvetľované zánikom oceánskej kôry subdukciou. V podrobnejšej klasifikácii by v prípade tejto panvy mohla byť hľadaná podobnosť s klasifikačným cyklom LL 2 až LL 3, t. j. s počínajúcim a pokračujúcim vrásnením v priľahlom panvovom okraji. Históriu panvy by bolo možné vyjadriť typom IF alebo LL 1 (vznik depresií v tenznom poli za účasti horizontálnych pohybov na zlomoch), neskoršie MS, so znakmi LL 2 až LL 3 (vznik príkrovov) a znova IF. (Poznámka autora: Typ LL 2 až LL 3 je však spájaný s aktívnym vrásnením a vznikom príkrovov zasahujúcich do panvy, a nie s nasunutím príkrovov zo vzdialeného orogénneho priestoru). Vo všetkých klasifikáciach má viedenská panva a jej podložie znaky rôznych typov panví a nie je čistým typom, čo je odrazom zložitého vývoja tejto megaštruktúry.

Podunajská a juhoslovenská panva sú viac čistým typom IF s niektorými prvkami typu MS. Východoslovenská panva je v podloží prináležajúcom k spodnej etáži typom LL 1 — LL 3 a neogénna výplň typom IF s prvkami MS.

Heterogenita znakov slovenských panví značí buď zložitosť ich vývoja, nepostihnuteľnú ani podrobnými klasifikáciami alebo viac-menej schematičnosť týchto klasifikácií vychádzajúcich zo zjednodušených modelov doskovej dynamiky.

**Ložiská uhlia a uhľovodíkov v panvách karpatsko-balkánskej oblasti**

Vzťah ložísk uhlia k stavbe panónskej panvy a karpatskej oblasti je blízky k typom kôry a najmä rozhraniu kontinentálnej a suboceánskej kôry v preduhoľnom vývoji oblasti. Na tieto línie sú viazané najpočetnejšie výskyty ložísk alebo panvy s najväčšou mobilitou. V širšom regióne boli definované uhľotvorné pásy: balatónsky, apusénsko-transylvánsky a východokarpatský. Špecifické postavenie má vardarsko-krajštidný uhľotvorný lineament. Bola tak pomenovaná tektonicky predisponovaná zóna odlišných typov kôry s opakovanou uhľotvornou sedimentáciou od karbónu do neogénu. Okrem početných uhoľných výskytov tu vzniklo 47 ekonomicky zaujímavých ložísk so zásobami 40—450 mil. ton uhlia na jednotlivých ložiskách. V strednej Európe žiadny z lineamentov nebol produktívny po dobu 200 mil. rokov. Priaznivé podmienky tvorili diferencované, oscilujúce, zdvihové a poklesové pohyby spomaľované v dobe akumulácie organickej hmoty. Počas vývoja uhľotvorného lineamentu dochádzalo k priestorovej migrácii uhľotvorných podmienok; fixnú migráciu mali len ložiská od karbónu do jury na línii Lupac (RSR) — Pirot (Bulharsko). Autonómnosť lineamentu a tvorby uhlia zdôrazňuje nezávislosť na vrásnení okolitých jednotiek.

Orientačne sa korelovala tektonická pozícia ložísk uhlia vo variskej a západokarpatskej predhlbni. Zatiaľ čo hercýnska predhlbeň bola na uhlie veľmi produktívna, neogénna karpatská predhlbeň je takmer sterilná. V oboch prípadoch pozitívny vplyv na uhľonosnosť mali mobilné bloky, pravdepodobne so slabou granitovou vrstvou. Priaznivejšia uhľonosnosť je vo východokarpatskej predhlbni, kde sú však uhoľné sloje zvrásnené. Všeobecnou črtou umiestnenia ložísk uhlia v predhlbni je spojitosť s miestami čiastočnej alebo úplnej redukcie šírky predhlbne. Pozitívny vplyv mal pravdepodobne blízky okraj platformy v predpolí; táto platforma ovplyvňovala stabilizáciu klesania pri tvorbe slojov. Predhlbeň vplyvom rýchlej migrácie panvy na predpolie však bola z uhľotvorného hľadiska nepriaznivou megaštruktúrou.

V pásme vonkajších flyšových Karpát sú všeobecne ložiská uhlia vzácne. Veľmi časté sú úlomky uhlia vo flyši alebo tenké slojky obmedzeného rozsahu.

Najväčšia akumulácia uhlia je v panónskej panve a najmä v periférnych jednotkách (obr. 69). Z hľadiska hlbinnej stavby sa ložiská koncentrujú na okrajové časti panvy k týmto štruktúram:

a) peripieninskému lineamentu na styku s Východnými Alpami,
b) na pásmo mobilnej suboceánskej kôry balatónskeho bloku,
c) k peripieninskému lineamentu na styku s jeho východokarpatskou vetvou,
d) na insubrickú líniu — periadriatický lineament,
e) na okraje Mt. Apuseni,
f) na už skôr spomínaný vardarsko-krajštidný lineament, najmä na okraje srbsko-macedónskeho masívu.

Panvy na balatónskom bloku patria k balatónskemu uhľotvornému pásu. Tvorba uhlia začala už v kriede (Ajka) a končila pliocénom. Spojitosť s vulkanizmom a väzba na prelomovú sústavu mobilného balatónskeho bloku pripomína riftovú pozíciu ložísk uhlia severočeskej hnedouhoľnej panvy. Vo vzťahu k vulkanizmu, ako fenoménu tvorby granitovej vrstvy, majú uhoľné panvy pozíciu v utváraní kôry zhodnú so stredočeskými permokarbónskymi panvami. Na balatónsky blok sa viažu najväčšie uhoľné zásoby MĽR.

Panvy pri východnej vetve peripieninského lineamentu ležia na vnútornej strane, ale na vonkajšej strane vulkanických pohorí. V tejto oblasti sa vyskytuje i podvihorlatská panva a ďalej i ložiská rumunské. Lineament mal pravdepodobne funkciu hranice stabilnej a mobilnej kôry. Pri západnej vetve lineamentu sa vyskytujú ojedinelé ložiská v Rakúsku. V oboch prípadoch neogénne ložiská zdôrazňujú ložiskový význam pásma na vnútornej strane lineamentu.

Panvy pozdĺž insubrickej línie majú obdobnú tektonickú pozíciu ako panvy predchádzajúce. Sú vyvinuté pri okraji suboceánskej kôry, ktorá podmienila ich väčšiu plošnú rozlohu, na rozdiel od panví peripieninského lineamentu. Pri insubrickej línii sa tvorili i sloje hrubé až 25 m, miestami vplyvom vrásnenia nadurené až na 40 m. Pozitívne prvky sú spájané s mobilným typom kôry.

Kontinentálny typ kôry má podložie panví v apusénskom bloku. Panvy majú charakter vnútrokarpatských panví.

Špecifické postavenie má panva Mecsek. Narastanie hrúbky uhľonosných sedimentov k SV, spojené s vytrácaním slojov, sa dáva do súvislosti s prítomnosťou mobilnej kôry prechádzajúcej do podložia solnockého flyšového trógu, kde M. Maheľ (1978) predpokladá oceánsku kôru. Smerom na JZ možno v súlade s paleogeografickými údajmi predpokladať kontinentálnu kôru, ležiacu asi v pokračovaní kontinentálnej kôry srbsko-macedónskeho masívu. Tektonické oddelenie oboch typov kôry dokazuje zlomová stavba, ostré zmeny v hrúbke a vývoji uhľonosných vrstiev a zmena štýlu tektonického porušenia, ktorá narastá k SV do sféry mobilnej kôry.

Sloje sú nerovnomerne preuhoľnené. Slabo preuhoľnené sú i hlboko uložené sloje pri okraji panónskeho diapíru, silne preuhoľnené sú eocénne sloje bližšie k centru diapíru. Tepelné účinky možno vysvetliť týmto diapírom a ich reguláciu sedimentárnym obalom — izolátorom.

Ak aplikujeme poznatky o vývoji neogénnych sedimentárnych panví vo vzťahu k plášťovému diapíru na zákonitosti zistené vo vývoji uhoľných panví, potom dochádzame k záveru, ktorý podporuje diapírovú koncepciu.

Treťohorné uhoľné ložiská Západných Karpát sú všeobecne menšie v porovnaní s ložiskami uhlia na platforme v predpolí Karpát. Už toto jednoduché porovnanie dokazuje prvoradý vplyv intenzity tektonických pohybov a tektonického členenia podložia panví na veľkosť uhoľných zásob a na ich plošný rozsah. Pozícia uhľonosných panví v tektogéne môže byť preto jedným z kritérií ocenenia perspektívnosti panví a prípadného prognózovania nových skrytých ložísk uhlia.

Geologicky i geofyzikálne určil neotektonickú blokovú stavbu a základné hlbinné zlomy O. Fusán et al. (1976), J. Kvitkovič a J. Plančár (1975) a J. Kvitkovič et al. (1976). V práci sa vychádza z ich členenia na bloky s príslušným označením.

Vnútrokarpatské panvy majú dvojaký vzťah k hlbinnej stavbe:

1. Panvy na styku fatro-tatranského bloku s blokmi peripanónskymi (podunajský a juhoslovenský) sa vyskytujú vo väčšom počte, plošne sú však malé.

2. Panvy na blokoch s tenkou kôrou: podunajská, juhoslovenská a východoslovenská — majú veľkú rozlohu a vyznačujú sa väčšinou veľkou mobilitou.

Z korelácie výskytov uhlia a uhoľných panví s hlbinnou blokovou stavbou (obr. 70) vyplýva:

a) Ložiskové koncentrácie uhlia sa viažu na okraje blokov alebo krýh, ktoré boli včlenené do blokovej stavby formovanej v súčasnej podobe koncom miocénu.

b) Najväčšie uhoľné zásoby sa sústreďujú do viedenskej a východoslovenskej panvy, ktoré majú pozíciu medziblokových panví na rozhraní dvoch megablokov: Českého masívu a Západných Karpát, Západných Karpát a Východných Karpát. Obidve panvy, tradične zaraďované k vnútrohorským panvám, sú z hľadiska hlbinnej stavby panvy medzihorské.

c) Obdobné postavenie majú i uhoľné ložiská juhoslovenskej panvy, ktorá leží na tenkej, predneoidne slabo konsolidovanej kôre. Tá je súčasťou panónskeho megabloku, a to jeho severného balatónskeho bloku. Ložiská uhlia sú viazané na prechod medzi tenkou a hrubou kontinentálnou kôrou vnútorných Karpát. Ležia teda medzi dvoma odlišnými krustálnymi segmentmi.

d) Vnútrokarpatské uhoľné ložiská patriace k vnútrohorským panvám, na typickej kontinentálnej kôre, sú dynamicky viazané na okraje blokov, príp. hlbinných štruktúr.

e) V rozmiestnení ložísk uhlia má doposiaľ nedocenený význam s.-j. zlomové pásmo (obr. 70), ktoré je vrchnokôrovým prejavom asi neogénne založeného a ďalej sa vyvíjajúceho lineamentu. Ten má znaky lineamentu uhľonosného, akým je vardarská zóna (F. Čech — J. Zeman 1980).

f) Pre vznik uhoľných ložísk boli z hľadiska hlbinnej stavby priaznivé: okraje blokov s rôznou pohybovou tendenciou (blízkosť znosovej oblasti), tektonická členitosť diferencujúca paleogeografické podmienky, pomalé klesanie, ktoré predisponovala najmä tenká, paleozoicky a mezozoicky neskonsolidovaná kôra.

Nepriaznivé faktory pre vznik väčšieho objemu uhoľných zásob boli: silné tektonické porušenie a rýchle zmeny v pozitívnych a negatívnych pohyboch, malá tektonická členitosť mobilnej (slabo konsolidovanej) kôry.

Ak ložiská uhlia v panónskej panve a Karpatoch dobre korelujú s blokovou stavbou a koncentrujú sa najviac na okraje kontrastne sa pohybujúcich blokov, nebola u ložísk uhľovodíkov zistená korelácia s blokovou stavbou (obr. 69). Ich rozmiestnenie súvisí so štruktúrnou prestavbou, ktorá vyvolala i migráciu uhľovodíkov. V rozmiestnení ložísk uhľovodíkov je korelácia s hrubšou kôrou a s nižším tepelným tokom. Vysoký tepelný tok podmienil deštrukciu ropy. Pre akumulácie uhľovodíkov pozitívnu úlohu zohrali hlbinné zlomy orientované tangenciálne k okrajom panónskej panvy a plášťového diapíru. V miestach vzniku okrajových depresií sa pri týchto zlomoch akumulovali ložiská uhľovodíkov. Prognózne sú okraje týchto depresií na styku s elevačným pásmom pri týchto zlomoch. Na našom území ide o východné okraje viedenskej panvy a autochtónne podložie pozdĺž peripieninského lineamentu.

Zatiaľ bola zistená produktívnosť západného predpolia viedenskej panvy na svahoch Českého masívu pod karpatskou predhlbňou (obr. 72). Hlboké podložie viedenskej panvy môže byť prognózne pre výskyty zemného plynu, ako to už preukázal hlboký vrt Zisterdorf 1 v Rakúsku (obr. 73). Na území ČSSR nebola zatiaľ vo viedenskej panve séria príkrovov prevŕtaná. V hlbokých obzoroch dosahuje hĺbku 8—10 km. Perspektívnejšie budú preto východné časti podložia viedenskej panvy.

Zo vzťahu ložísk k vyšším kryhám pri okrajoch hlbinných zlomov možno usudzovať, že geofyzikálne zistené štruktúry pozdĺž lednickej zóny môžu byť prognóznymi pre výskyt uhľovodíkov.

Podobnú úlohu môžu mať i sprievodné zlomy pri peripieninskom lineamente vo východoslovenskej oblasti (obr. 74).

Maximálna koncentrácia ložísk uhľovodíkov je v predhlbni, najmä v pásme styku západo- a východoeurópskej platformy v podloží predhlbne (obr. 71). Najproduktívnejšími sú oblasti hlbinne viazané na medziblokové panvy, pokiaľ tu existujú priaznivé elevačné pasce.

Vo vzťahu k distribúcii uhoľných ložísk bola zistená záporná korelácia.

Na území ČSSR zhodne s touto pozíciou je pomerne najbohatšia panva viedenská a východoslovenská, komplexné ložiskové štruktúry.

Z rozboru hlbinnej stavby vyplývajú pre naše územie nepriaznivé prognózne závery. Perspektívne jednotky a štruktúry majú malé rozmery, a preto nemožno očakávať objav veľkých ložísk (gigantov) uhľovodíkov. Za perspektívny bol označený i peripieninský lineament a sz. pokračovanie sámošskej línie.

Ložiská zemného plynu, ktoré kvantitatívne prevažujú nad ložiskami ropy v panónskej panve, sú viazané na štruktúry v podloží neogénu a na pórovité horizonty neogénu (Gy. Kertai 1968). Z hľadiska hlbinnej stavby môžeme rozlíšiť ložiská uhľovodíkov viazané na periférnu časť panónskej panvy pozdĺž peripieninského lineamentu a na ložiská viazané na centrálnu časť panvy.

K periférnym štruktúram pozdĺž peripieninského lineamentu môžeme zaradiť i ložiská viadenskej panvy a ložiská na šopronskom bloku. Pozdĺž insubrickej línie sa sústreďujú ložiská drávskej a sávskej depresie.

V centrálnej časti panvy je najviac ložísk. Najmenší počet je štruktúrne viazaný na balatónsky blok východne od dunajského zlomu. Väčší počet sa koncentruje na predpolie styku balatónskeho bloku s dinaridami. Najväčšia plošná koncentrácia je na paleopanónskom bloku východne od dunajského zlomu, najmä v priestore mobilného paleopanónskeho bloku s. s. Miesta bez ložísk dobre korelujú s oblasťami najmenšej hrúbky kôry a najväčšieho výstupu panónskeho diapíru. V týchto oblastiach je aj najväčší tepelný tok, ako to vyplýva z údajov L. Stegenu et al. (1975). Z hľadiska migrácie uhľovodíkov je táto oblasť nepriaznivá v porovnaní s oblasťami najväčšej akumulácie ložísk.

Tektonické línie zvyšujúce pórovitosť masívu boli vhodné pre migráciu, pokiaľ boli orientované diagonálne až kolmo na predpokladaný západo-východný smer šírenia ťahového napätia od apikálnej časti rozpínajúceho sa diapíru (drávska alebo solnocká línia). Na druhej strane záhrebsko-kulčská a rábska línia pri pliocénnej inverzii ekranovali cestu uhľovodíkom do stabilizovaného balatónskeho bloku (kde mal okrem toho negatívny vplyv i doznievajúci vulkanizmus). Nejasná je dynamická úloha peripieninského lineamentu. Z rozmiestnenia ložísk zemného plynu však možno usudzovať, že strižné pohyby a kompresné tlaky limitovali migráciu uhľovodíkov cez lineament. Ak sa však vytvárajú vhodné štruktúrne pasce, lineament môže akumulovať zadržané uhľovodíky. V prípade viedenskej panvy lineament pravdepodobne mal i ochrannú funkciu, ako bariéra brániaca tlakovým účinkom z vnútornej strany Karpát. Priaznivé podmienky boli najmä juhozápadne od nesvačilsko-trnavskej línie (F. Čech 1980). Táto oblasť mimo dosahu flyšových príkrovov bola tlakovo menej postihnutá.

Panónska panva, v juhovýchodnej polovici napäťovo odľahčená, má z tektonického hľadiska priaznivé podmienky pre zachovanie uhľovodíkov. Existencia starých blokov spojených v panónsky megablok čiastočne diferencovala rozmiestnenie ložísk na kryhy maximálne klesajúce v pliocéne. Ponúka sa preto vysvetlenie, že klesanie bolo príčinou vytláčania uhľovodíkov a ich zachovania v pascách. Z hľadiska typu kôry závislosť nie je výrazná — ložiská sa koncentrujú ako na bloky predneoidne konsolidované a neoidne regenerované (typ regenerovaných platforiem), tak i na kôru slabo konsolidovanú. Pokročilejšia konsolidácia sa pozitívne prejavila tým, že obmedzila intenzitu tektonického triešťenia.

Medzi blokovou stavbou a distribúciou ložísk uhľovodíkov nie je teda zjavná korelácia. Existuje len paleogeograficky a tektonicky podmienená závislosť, že sa ložiská uhľovodíkov v panónskej panve koncentrujú hlavne v blokoch alebo kryhách, kde nevznikli väčšie koncentrácię ložísk uhlia. Pokiaľ existujú výnimky, tak sú dané inverznými pohybmi krýh (Mátra, Bükk).

Pre distribúciu uhľovodíkov mal z hlbinných fenoménov najväčší význam vývoj panónskeho diapíru a ním vytvorené termálne pole na styku rigidného podložia sedimentárnych panví s ich výplňou. Ložiská zemného plynu sa skôr koncentrujú na krídla apikálnych častí diapíru mimo oblasti najtenšej kôry. Tvar diapíru a heterogénna bloková stavba kôry modifikovali rozloženie napätia v kôre. Modelovo by mohla prichádzať do úvahy schéma, podľa ktorej vrcholová časť diapíru vyvoláva ťahové napätie, ktoré sa na krídlach diapíru v hrubšej kôre transformuje na tlakové pôsobenie. To umožňovalo migráciu uhľovodíkov a narastanie ložiskových tlakov. Nie je vylúčené, že i rôzna hĺbka produktívnych horizontov — nad diapírom menšia než pri okrajoch diapíru — bola podmienená napäťovým režimom s endogénnou príčinou.

Eliptické oblasti s najtenšou kôrou sú sprevádzané najväčším tepelným tokom. Tento okrem predpokladanej existencie ťahového napäťového poľa mohol podmieniť vytvorenie bezložiskových oblastí v už spomenutej osnej časti panónskeho diapíru.

V súčasnosti je vysvetlenie vzniku neogénnych panví hlbinnými príčinami najracionálnejšie. Poznatky hlbinnej geológie a geofyziky preukázali, že sa neogénne panvy vnútorných Západných Karpát nachádzajú na hlbinných štruktúrach a že ich vývoj mal spoločné faktory, ktorých príčiny nemožno hľadať v povrchových javoch. Takýmto javom však môže byť vrásnenie ako deformačný akt, ktorý nezasahuje do hlbinnej stavby.

Príčinou vrásnenia sú však hlbinné procesy. Dokázali to i korelačné výskumy sedimentárnych panví, najmä vzťahy ich založenia a periód intenzívnej subsidencie k hlbinným štruktúram a rozvoju vulkanizmu. Vznik panví, podobne ako vulkanizmus a ďalšie geologické procesy, sú len rôznymi formami endogénnych procesov. Súčasné geofyzikálne poznatky posúvajú príčiny týchto procesov do astenosféry.

Model plášťového diapíru indikovaného sprvu geofyzikálne, ale dnes i geologicky, najmä analýzou hlbinnej stavby, je v súlade s modernými smermi hlbinnej geológie najaktuálnejší. Hlbinný aspekt pomohol nájsť i nové prvky spoločné pre panvy alebo dvojice panví (viedenská — podunajská) a oddeliť prvky špecifické — autonómne (východoslovenská panva).

Súborne boli zhodnotené poznatky o hlbinnej stavbe po prvý raz v komplexe dostupných údajov a kvalitatívne boli postavené na nový základ, vychádzajúc z hodnotenie typov kôry podľa látkových a tektonických prejavov. V tomto smere treba považovať prínos globálnej tektoniky za vysoko progresívny. Otázka dynamických interpretácií zostáva na poli diskusií a v práci prijaté dynamické závery vychádzajú z kritického rozboru reálnosti geologických a geofyzikálnych údajov.

Zistili sa všeobecné vzťahy simatickej kôry k optimálnemu rozvoju ložísk, pokiaľ existovali vhodné (nemarinné) paleogeografické podmienky. Typickým príkladom je ofiolitové pásmo dinaríd alebo už spomenutý simatický balatónsky blok. Z hľadiska zlomového členenia sa toto uplatňuje dvojakým spôsobom:

— slabé členenie umožňuje vznik veľkoplošných uhoľných slojov, hlavne na platforme v predpolí Karpát;

— silné členenie, najmä na striedaní elevácií a depresií, umožnilo vznik väčšieho počtu uhoľných slojov, avšak plošne obmedzených.

Na ložiská uhlia je najchudobnejšia predhlbeň, hlavne v Západných Karpatoch. Bohatšie sú vnútorné Karpaty a v nich potom medzihorské panvy (obr. 74).

V panónskej panve sú na ložiská uhlia najbohatšie periférne časti a produktívnymi sa ukázali tiež k panónskemu diapíru tangenciálne orientované periférne zlomy. Z ďalších štruktúr sú priaznivé okrajové zlomy Apusén a okraje sialických masívov, hlavne severná časť vardarsko-krajštidnej zóny.

Stupeň preuhoľnenia je diferencovaný. Existujú doklady o termálnom ovplyvnení panónskym diapírom. Rozdiely v preuhoľnení a slabé preuhoľnenie pliocénnych ložísk možno vysvetliť alebo poklesom tepelného toku alebo skôr izolačnými vlastnosťami pliocénnych sedimentov.

### Niektoré závery pre prognózovanie ložísk prírodných uhľovodíkov

Ak sú perspektívy na objavenie uhoľných slojov minimálne a ich objav sa môže týkať len malých výskytov s malými uhoľnými zásobami na hranici, resp. pod hranicou ekonomickej rentability využitia zásob, potom je z hľadiska výskytu prírodných uhľovodíkov situácia perspektívnejšia. Vo viedenskej a východoslovenskej panve sú okrem klasických oblastí perspektívne hlboké obzory, a to pri okrajoch panví.

Výsledky doterajších prieskumov ukázali, že z ostatných neogénnych panví je na výskyty zemného plynu pozitívna len severozápadná časť podunajskej panvy — lokality Špačince, Trakonice, Nižná, Madunice a Krupá. Akumulácie zemného plynu sú viazané na vyzdvihnutú kryhu sv. od nesvačilsko-trnavského zlomu, a to na trnavskú depresiu. V ostatných panvách neboli zatiaľ dosiahnuté pozitívne výsledky. Navŕtané výrony $CO_2$, N alebo He však nevylučujú možnosť výskytu ďalších plynonosných obzorov vo väčšej hĺbke, avšak malého rozsahu. V štruktúre panví chýbajú vhodné elevačné štruktúry. Perspektívne by mohli byť litologické pasce pri zlomoch.

V podunajskej panve je to zlatomoravská kryha a jz. ponorená časť kryhy tríbečskej u nesvačilsko-trnavského zlomu. V tejto súvislosti by som chcel podotknúť, že paralelne s týmto zlomom prebieha na JZ ďalší zlom jz. od Kolárova k Pezinku. Zatiaľ nie je preukázané, či táto línia pretína peripieninský lineament a vniká do viedenskej panvy. V oblasti Malaciek a Veľkých Levár však existujú zlomy, ktoré by mohli byť ekvivalentom tejto línie, jednej zo systému zlomov smeru SZ—JV, podľa ktorých stupňovite klesá podložie podunajskej panvy smerom k Dunaju. Za predpokladu analogickej kryhy vo viedenskej panve by sa tu kumulovali početné ložiská zemného plynu i ropy, napr. na eleváciách Studienka, Závod, Brodské a iné. S ohľadom na určitú analógiu v produktívnosti kryhy sv. od nesvačilsko-trnavského zlomu by som považoval i kryhu jz. od tohoto zlomu za prognóznu. Ide o strednú časť sládkovičovskej kryhy okolo Galanty a maximálne sv. od Dunajskej Stredy. Za málo perspektívnu považujem oblasť gabčíkovskej depresie. Presuny subsidencie od S k J mohli zmeniť trendy migrácie uhľovodíkov a ich odmigrovanie z gabčíkovskej depresie do plytších úrovní alebo elevácií.

Malé výskyty zemného plynu by mohli byť i v šahanskej kryhe východne od Nových Zámkov (obr. 74). Naproti tomu komárňanskú kryhu nepovažujem za perspektívnu. Dunajské zlomové pásmo (centrálnokarpatský lineament) prognózne pre ložiská uhlia považujem za málo perspektívne pre výskyty ložísk uhľovodíkov. V podunajskej panve s vysokým tepelným tokom sa nachádzajú zdroje termálnych vôd, ktoré patria medzi zdroje energetické. Ako v prvej oblasti v ČSSR boli tu tiež preskúmané tieto zdroje, ktoré sú už využívané. Prvé využitie vrtu realizovaného za účelom vyhľadávania uhľovodíkov ako zdroja geotermálnej energie (vrt Diakovce-1) bolo v r. 1958 (teplota vody 41 °C). Prvý vrt realizovaný v rámci vyhľadávania geotermálnej energie bol vrt Komárno M-1 v r. 1967, s vodou teplou 54 °C. Poľnohospodárskym účelom slúžil v r. 1971 vrt Dunajská Streda 1. Panva je perspektívna pre objavenie ďalších termálnych prameňov.

Vo východoslovenskej panve je z hľadiska hlbinnej blokovej stavby perspektívna košická depresia a sečovská kryha, najmä jej styk s kryhou zemplínskou.

Perspektívnosť pásma okolo peripieninského lineamentu, kde sú indície výzdvihov i hlbšieho podložia a vytvorenia vhodných štruktúrnych pascí, overili hlboké vrty, najmä vrt Hanušovce-1, hlboký 6003 m (R. Rudinec — B. Leško 1984).

Okraje blokov, ako pásma perspektívnych štruktúr, indikujú i ďalšie hlboké vrty: Lipany, Smilno, Zboj. Pomocou parametrických vrtov bol objavený vnútrokarpatský paleogén (R. Rudinec — B. Leško l. c.) ako nová ropo-plynonosná provincia. Toto zistenie potvrdilo moju prognózu (F. Čech 1982), keď som predpokladal existenciu puklinových kolektorov vo flyši a v jeho podloží i možnosť existencie štruktúrnych pascí pri báze flyšu.

Ostáva ešte problém plynonosnosti sámošskej línie. Táto štruktúra nie je zatiaľ známa, ale leží jz. od peripieninského lineamentu, ktorý v sv. výbežku panónskej panvy nie je čo do perspektívy docenený. Po sámošskej línii vystúpili neovulkanity, ktoré vytlačili prípadné staršie akumulácie uhľovodíkov. Sámošská línia mohla však byť vhodným pórovitým obzorom a komunikačnou cestou pre migrujúci plyn. Zemný plyn zostal vo vulkanicky menej postihnutých depresiách jz. od peripieninského lineamentu. Po sámošskej línii mohol podložím východoslovenskej panvy migrovať i do južnej časti košickej depresie (najperspektívnejšia oblasť) a po slanskom a hornádskom zlome i ďalej na SZ do priestoru Levočského pohoria.

Pre československé ložiská uhľovodíkov platí rovnaká závislosť veľkosti štruktúr na množstve akumulácie surovín, ako u iných druhov surovín. Veľké štruktúry produkujú veľké ložiská. Malé rozmery panví a štruktúr v Západných Karpatoch nedávajú perspektívu pre objav veľkých ložísk zemného plynu alebo ropy. Ani viedenská a východoslovenská panva nedosahujú rozmery veľkých depresií kumulujúcich veľký objem organickej hmoty. Podľa H. D. Klemma (1978) sú najproduktívnejšie vnútrokontinentálne štruktúry, ktoré sú nositele 75 % svetových gigantických ložísk uhľovodíkov. Ak medzi vnútrokontinentálne panvy zahrnieme naše panvy medziblokové — viedenskú a východoslovenskú — potom sa vzhľadom k ich malým rozmerom môžu považovať za ekvivalenty ložiskových gigantov v našich malorozmerných jednotkách.

V oboch našich panvách však chýba regionálna elevácia ako štruktúrno-litologická pasca koncentrujúca migrované uhľovodíky. Také štruktúry sú podmienkou vzniku veľkých ložísk, až 75 % z celkových zásob panvy (H. D. Klemme l. c.). Pokiaľ zberná štruktúra chýba, ložiská sú dispergované, čo je tiež prípad našich panví. Naproti tomu východokarpatské panvy podobné štruktúry majú. Aj transylvánska panva je husto prestúpená elevačnými štruktúrami. Malé štruktúrno-litologické pasce zhromažďujú maximálne 10 % celkových zásob panvy (H. D. Klemme 1978). Ak vyjdeme z týchto empiricky zistených vzťahov, potom pre stanovenie prognóz stačí spočítať rozdiel medzi známymi zásobami v jednotlivých obzoroch a kapacitným produkčným potenciálom panvy (s odpočítaním odhadnutých odmigrovaných uhľovodíkov). Podľa novej klasifikácie a produktívnosti jednotlivých typov panví (H. D. Klemme 1980) by vo viedenskej panve spodná štruktúrna etáž (autochtónne sedimenty) mala mať najviac pascí, 3/4 ložísk by sa malo nachádzať v klastických sedimentoch a 1/4 vo vápencoch. Pokiaľ by v hlbinnej stavbe existovali riftové štruktúry, je situácia obdobná a klastické obzory by mohli byť pascami pre ložiská, ktoré vznikli napr. vo vápencoch paleozoika či raného mezozoika. Perspektívy vrchnej etáže typu medzihorskej panvy sú lepšie známe a sú v súlade s produktívnosťou tohoto typu panví (tab. 4).

Podobná je situácia vo východoslovenskej panve, kde sa prevažná väčšina ložísk nachádza v klastických sedimentoch.

V oboch panvách sú možnosti výskytu relatívne väčších pascí, a to štruktúrnych, litostratigrafických a najčastejšie kombinovaných. Juhoslovenská a podunajská panva majú znaky blízke panvám s veľkými ložiskami (veľkým objemom organickej hmoty, materských hornín) i s veľkými pascami všetkých typov, avšak doterajšie poznatky o tom nesvedčia. Niektoré znaky sa pozitívne uplatňujú u panónskej panvy.

Poznatky vyplývajúce z analýzy hlbinnej stavby panví ukazujú, že naše najväčšie panvy ležia na hraniciach blokov či megablokov. Z tejto tektonickej pozície vyplýva aj ich perspektívnosť, pretože 60 % svetových zásob uhľovodíkov je v panvách pozdĺž okrajov litosférických dosiek. Z uvedených záverečných poznámok vyplýva význam tohoto skúmania hlbinnej stavby panví a ich vzťahu k blokovej stavbe. Možno konštatovať, že prakticky všetky známe priemyslové zásoby uhľovodíkov v ČSSR sa nachádzajú v panvách medzi blokmi alebo megablokmi a že sú teda tieto štruktúrne hranice najperspektívnejšie pre ďalšiu prospekciu uhľovodíkov v ČSSR.

Vysvetlivky k obrázkom

Obr. 1 Schéma geologického profilu veľmi hlbokého vrtu Kola SG-3 v ZSSR
(Podľa E. A. Kozlovského et al. 1984, značne zjednodušené a upravené)
Vysvetlivky: 1 — metasedimenty, 2 — plagiogranity, pegmatity, granitizované ruly (migmatity), 3 — bázity a metabázity, metaandezity — čadiče a ich tufy, ultrabázity, bázické metatufy a tufity, 4 — zlomy (v archaických amfibolitoch), mylonity, kataklázity, brekcie (a — podľa originálneho obrázku, b — interpretácia autora).

Metamorfóza je progresívna od fácie pumpellyitovo-prehnitovej po amfibolitovú. V archaiku je naložená na staršej granulitovej fácii.

Obsah plynov sa vzťahuje na 1 $m^3$ horniny: na ľavej strane sú vyznačené len intervaly anomálnych obsahov plynov: He 4—15 . $10^{-3}$ $m^3/m^3$, $CH_4$ 0,5—2,5 . $10^{-1}$ $m^3/m^3$, $CO_2$ 1—3 . $10^3$ $m^3/m^3$.

Profil znázorňuje zloženie kôry simatického typu.

Obr. 2 Vznik panvy roztiahnutín litosféry
Vysvetlivky: 1 — plytkovodné sedimenty, 2 — hlbokovodné sedimenty, 3 — uhlová diskordancia.

Obr. 3 Rozloženie napätia nad diapírovým telesom — modelované podľa M. V. Gzovského (1960)
a) vznik tangenciálnych napätí ($\tau$),
b) schéma trajektórií tangenciálnych napätí,
c) schéma trajektórií normálnych napätí ($\delta$).

Obr. 4 Vývoj zlomov v rifte Červeného mora (zjednodušené podľa J. D. Lowela et al. 1978)

Obr. 5 Z-v. rez centrálnou Nevadou (podľa Ch. H. Scholza et al. 1971 upravené).
Vysvetlivky: 1 — seizmická zóna, 2 — nízkorýchlostná zóna, čiastočne natavená.

Obr. 6 Štádiá vývoja elevačnej a riftovej štruktúry
a) Termálny stĺpec podmieňuje pod kontinentálnou litosférou eleváciu hlbšej časti plášťa; b) kontinentálna litosféra sa prehrieva a stenčuje s následným izostatickým zdvihom a vznikom ťahových napätí vo vrchnej kôre; c) graben sa začína tvoriť, ak ťahové napätie dosiahne dostatočnú veľkosť; (podľa M. H. P. Botta 1981, upravené).

Obr. 7 Vznik riftového údolia na kontinente v dôsledku roztiahnutia a stenčenia silne prehriatej spodnej kôry. Táto deformácia spodnej kôry je sprevádzaná vznikom zlomov a klesaním vrchnej kôry, kde nemôže vzniknúť veľká viskózna deformácia (podľa E. V. Artjuškova 1981, upravené).
Vysvetlivky: 1 — časť granitovej vrstvy s krehkou deformáciou, 2 — časť granitovej vrstvy s viskóznou deformáciou (duktilná zóna), 3 — bazaltová vrstva, 4 — nízkorýchlostný plášť, 5 — normálny plášť, 6 — magma určujúca zlomy.

Obr. 8 Profil kôrou pod grabenom Viking (podľa P. A. Zieglera 1982, upravené)
Vysvetlivky: 1 — predriftové sedimenty, 2 — výplň riftu, 3 — poriftové sedimenty.

Obr. 9 Vznik riftu vyžaduje špecifické ruptúrne reakcie. Ak nastane uvoľnenie duktilnej deformácie, vzniknú podmienky stavu horniny ako by bola nekompetentná a graben nevznikne (podľa J. H. ILLIESA 1981, upravené).

Obr. 10 Model vzniku vnútrokarpatských panví subdukciou litosférických dosiek (podľa L. ROYDENOVEJ et al. 1982).

Obr. 11 Model vzniku panónskej panvy subdukciou litosférických dosiek (podľa L. STEGENU et al. 1975)

Obr. 12 Vznik tzv. strižnej panvy (upravené podľa A. R. GREENA 1977)
Vysvetlivky: 1 — sedimenty, 2 — kontinentálna kôra, 3 — vrchný plášť, 4 — výlevy čadičov, 5 — horizontálne pohyby.

Obr. 13 Predstava vzniku viedenskej panvy horizontálnych posunom — typ „pull-apart basin" a — originálny diskontinuitný zlom, b — pri horizontálnom posune v oddelenom bloku je ťah orientovaný v smere posunu, c — výsledný zlomový obraz v panve (s.-j. orientované ťahové štruktúry otvárané zlomami, by sa mali spájať s hlavným horizontálnym posunom). Schéma sub c sa líši od koncepcie ťahových napätí F. HORVÁTHA a L. ROYDENOVEJ (1981).

Obr. 14 Modelové predstavy vzniku panvy horizontálnymi posunmi (upravené podľa I. C. CROWELLA 1974)
a — posun na nerovnom zlome, b — vznik panvy typu „pull-apart" na diskontinuitnom zlome, c — posun na rozvetvenom zlome.

Obr. 15 Zjednodušený model vývoja paleozoickej kôry v Západných Karpatoch (upravené podľa M. MAHEĽA 1978)
Vysvetlivky: 1 — sedimenty, 2 — kontinentálna kôra, 3 — oceánska kôra, 4 — paraoceánska kôra.

Obr. 16 Vznik panvy typu „pull-apart" na transformnom rozhraní
a — pri divergentnom okraji, b — všeobecný model (zjednodušené a upravené podľa I. C. CROWELLA 1974).

Obr. 17 Vývoj riftového tripletu
a — vznik odumretého ramena riftového tripletu, b — pokročilé štádium vývoja a vznik miogeoklinály (s použitím schém F. BURKA — J. F. DEWEYA 1973).
Vysvetlivky: 1 — delta, 2 — miogeoklinála, 3 — kontinentálne sedimenty, 4 — oceánska kôra, 5 — oceán.

Obr. 18 Vznik aulakogénu z mŕtveho ramena riftového tripletu pri konvergencii dosiek (s použitím schém F. BURKA — J. F. DEWEYA 1973)

Obr. 19 Schematické profily jednotlivých typov panví (upravené podľa H. D. KLEMMA 1980)
Názvy typov panví sú uvedené v texte.
Vysvetlivky: 1 — kontinentálna kôra, 2 — oceánska kôra, 3 — sedimenty, 4 — vulkanický ostrovný oblúk, 5 — diskordancia.

Obr. 20 Modely plášťových diapírov a ich dynamiky
a — napätie v symetrickom diapíre (md — okrajová depresia, rs — okrajová synklina), b — rozpínanie vrcholu diapíru a vznik vejárovitého tvaru, c — model dynamiky litosféry Alboránskeho mora (upravené podľa M. LEMOINA 1978), d — napätie v asymetrickom diapíre, e — model dynamiky litosféry Tyrrhenského mora, f — napätie a dynamika v lineárnom diapíre (moesijsko-pontická depresia).
Vysvetlivky: 1 — bazifikácia, (zblíženie fyzikálnych vlastností s plášťom), 2 — simatická kôra, 3 — anatexia, 4 — chladný plášť, 5 — horúci plášť, 6 — povrch astenosféry, 7 — ťahové napätie, 8 — strižné napätie, 9 — klesanie.

Obr. 21 Model vzniku panvy nad plášťovým diapírom
Vysvetlivky: $\varrho_1$ — granitová vrstva, $\varrho_2$ — bazaltová vrstva, $\varrho_3$ — horúci plášť (ľahký plášť), $\varrho_4$ — normálny plášť.

Obr. 22 Schéma intrúzie plášťového diapíru (podľa J. I. NIKOLSKÉHO 1982)
Vysvetlivky: 1 — sedimentárne a metamorfné komplexy, 2 — „granitová vrstva", 3 — „bazaltová vrstva", 4 — plášťový diapír s intrúziami bázickej magmy, 5 — granitovo-rulová kupola, 6 — magmatické krby s kyslou magmou.
Diskontinuity: A — povrch kryštalinického podložia, K — Conradova diskontinuita, M — Moho-diskontinuita.
Izotermy: $T_1$, $T_2$, $T_3$. Vývojové štádiá diapíru: I, II, III. Šípky označujú pohyby vrchnej časti litosféry a pohyb hlbinných hmôt.

Obr. 23 Závislosť hodnôt a amplitúdy tiažových anomálií Stredozemného mora a Strednej Ázie na hĺbke umiestnenia vystupujúceho plašťového astenolitu
Vysvetlivky: 1 — kôra s vyznačením narastania hustoty od 2,8—4,0 . $10^3$ kg . $m^{-3}$ — na ľavej strane schémy, 2 — astenolit s hustotou 3,4 . $10^3$ kg . $m^{-3}$, 3 — krivka anomálie tiažového poľa.

Obr. 24 Schéma vzniku panví v západnom Stredomorí (podľa J. P. MALOVICKÉHO et al. 1982, upravené a zjednodušené autorom)
Vysvetlivky: 1 — sedimenty: a — spodnomiocénno-kvartérne, b — staršie, 2 — evapority (mesin), 3 — bloky podložia, 4 — hlbinné zlomy, 5 — vulkanity a magmatické krby, 6 — intrúzie plášťa astenolitu, 7 — horizontálne tlaky, 8 — zvrásnené a presunuté súvrstvia.

Obr. 25 Stavba podložia perikaspickej depresie (podľa V. V. BELOUSOVA 1982, zjednodušené)
Vysvetlivky: 1 — izohypsy bázy sedimentárneho komplexu v km hĺbky (podľa rýchlosti P-vĺn), 2 — zlomy (stupne v podloží), 3 — východy južného Uralu.

Obr. 26 Hrúbka kôry a hypocentrá zemetrasenia v Tyrrhenskom mori (upravené podľa P. GIESSE — C. MORELLI 1975)
Hĺbky hypocentier zemetrasení za obdobie od r. 1900—1970.
Vysvetlivky: 1 — do 60 km, 2 — od 60—100 km, 3 — vyše 100 km, 4 — izohypsy povrchu Moho-diskontinuity v km.

Obr. 27 Geologická stavba Tyrrhenského mora
Vysvetlivky: 1 — izolínie tepelného toku v $mWm^{-2}$, 2 — kabylidy a toskánsky chrbát, 3 — granitoidy, 4 — alkalicko-vápenaté vulkanity, 5 — oceánske toleity, 6 — príkrovy a prešmyky, 7 — smer tektonického transportu, 8 — neogénna predhlbeň, 9 — trógy, 10 — okrajové depresie mora, 11 predpokladané hlbinné zlomy. V pravom hornom rohu: približný tvar Benioffovej zóny. Upravené podľa A. CAIREHO (1973), M. BOCCALETTIHO et al. (1976). a F. HORVÁTHA et al. (1981).

Obr. 28 Mediteránne elevácie vrchného plášťa
Vysvetlivky: 1 — elevácie, 2 — vrcholy elevácií: a — 15—25 km pod povrchom, b — 30 km pod povrchom, 3 — hlavné strižné zóny, 4 — hranice vysokého (h) a nízkeho tepelného toku.

Obr. 29 Rozsah tiažovo kladne porušenej karpatsko-balkánskej oblasti vo vzťahu k vnútrokarpatským neogénnym panvám
Vysvetlivky: 1 — nulová izoanomála $\mu ms^{-2}$ obmedzuje kladnú plochu oproti zápornej izoanomále (podľa J. IBRMAJERA 1978), 2 — okraj neogénno-kvartérnych panví, 3 — neovulkanity, 4 — anomálie nad $+300$ $\mu ms^{-2}$, P — panónska panva, T — transylvánska panva, M — moesijská platforma.

Obr. 30 Schéma hlavných hlbinných tektonických jednotiek karpatsko-balkánskej oblasti (s použitím mapy M. MAHEĽA 1978b a V. E. CHAINA — V. I. SLAVINA 1972 — upravené a doplnené)
Vysvetlivky: 1 — vynorené vnútorné masívy, 2 — hlavné hlbinné zlomy, 3 — medzinárodné profily HSS, 4 — okraj sedimentárnych panví, 5 — okraje tektonických jednotiek, 6 — superhlboká plášťová zlomová zóna Sofia—Vrancea—Sumy ($\alpha$) a oďersko-kavkazská ($\beta$), 7 — ohniská hlbokých zemetrasení Vrancea (6,7 podľa V. B. SOLLOGUBA et al. 1984).
Vnútorné masívy: P — panónsky, T — transylvánsky, SM — srbsko-macedónsky, R — Rodopy, PB — Predbalkán, S — Strednogorie a Stará Planina.
Predpolia: BM — Český masív, V — vardarská zóna, WC — Západné Karpaty (O — vonkajšie, I — vnútorné), EC — Východné Karpaty (O — vonkajšie, I — vnútorné), SC — Južné Karpaty,

G — getická jednotka, EA — Východné Alpy, MA — Apuseni Mts., Mu — Murešský tróg, M — Moesijský megablok.
Hlbinné zlomy: 1 — peripieninský lineament, 2 — sámošská línia, 3 — rábska línia, 4 — insubrická línia, 5 — vardarsko-krajštidná zóna, 6 — labský lineament, 7 — balatónska línia, 8 — záhrebsko-kulčská línia, 9 — s.-j. línia dunajská, 10 — s.-j. línia hornádska, 11 — hurbanovský zlom, 12 — línia Darnó, 13 — murešský zlom.

Obr. 31 Hrúbka kôry, kladné tiažové pole a rozsah neogénnych vnútrokarpatských panví (karpatsko-balkánskej oblasti)
Vysvetlivky: 1 — seizmoaktívne lineamenty: (PL — peripieninský, IZ — insubrický, Va — vardarská zóna), 2 — izolínie hrúbky kôry v km (podľa B. Beránka 1978, upravené), 3 — okraj neogénno-kvartérnych panví (P — panva panónska, T — transylvánska, M — moesijská platforma).

Obr. 32 Geologicko-seizmický profil cez dinaridy—Panóniu—Západné Karpaty
Rez III—V (Dubrovník—Szeged—Warszawa) JJZ—SSV. Podľa J. Ilavského et al. (1979), upravené.
1 — prekambrium, 2 — diskontinuita Mohorovičiča, 3 — paleozoikum, 4 — diskontinuita Conrada, 5 — mezozoikum, 6 — hlboké transkrustálne zlomy, 7 — paleogén (flyš), 8 — regionálne krustálne zlomy, 9 — neogén až kvartér, 10 — neovulkanity (andezity až bazalty), 11 — ultrabázické vyvreliny, 12 — granitoidy, 13 — bazaltová vrstva, 14 — vergencia štruktúr a príkrovov.

Obr. 33 Pozdĺžny geologicko-seizmický profil VI a VI A (Brno—Budapešť—Bukurešť) (podľa J. Ilavského et al. 1979, upravené).
1 — diskontinuita Mohorovičiča, 2 — paleozoikum a kryštalinikum, 3 — diskontinuita Conrada, 4 — mezozoikum, 5 — hlboké transkrustálne zlomy, 6 — paleogén (flyš), 7 — regionálne krustálne zlomy, 8 — neogén až kvartér, 9 — neovulkanity (andezity až bazalty), 10 — ultrabázické horniny, 11 — granitoidy, 12 — bazaltová vrstva.

Obr. 34 Hlavné hlbinné zlomy a recentná seizmicita v panónskej panve
Vysvetlivky: 1 — hlbinné zlomy, 2 — hlbinné zlomy neotektonicky aktívne, 3 — vrchnokôrové zlomy, 4 — seizmický aktívnejšie oblasti (podľa V. Kárnika 1975, zjednodušené).
Hlbinné zlomy a lineamenty: 1 — peripieninský, 2 — sámošský, 3 — rábsky, 4 — periadriatický (insubrický), 5 — vardarský, 6 — hurbanovský, 7 — balatónsky, 8 — záhrebsko-kulčský, 9 — dunajský, 10 — hornádsky, 11 — tiský, 12 — Darnó.

Obr. 35 Tektonická stavba transylvánskej panvy (upravené podľa Tektonickej mapy Rumunska 1970 a V. V. Semenoviča — Ju. G. Namestnikova 1981)
Vysvetlivky: 1 — predneogénne komplexy, 2 — izohypsy povrchu podložia v m, 3 — zlomy, 4 — prešmyky.

Obr. 36 Moesijská platforma — depresia a jej členenie
Vysvetlivky: 1 — okrajové depresie, 2 — význačné tektonické jednotky, 3 — okraj moesijskej platformy.

Obr. 37 Hlavné lineamenty a okrajové depresie panónskej panvy (upravené podľa M. Boccalettiho et al. 1976, D. Vassa 1979, a F. Čecha — J. Zemana 1982)
Vysvetlivky: 1 — elevácie s predneogénnymi komplexmi, 2 — paleogénna budínska panva, 3 — okrajové depresie, 4 — silne mobilné úseky vo vnútri panvy s viac ako 2 km hrúbkou neogénno-kvartérnych sedimentov, 5 — hlbinné zlomy.

Obr. 38 Vzťah medzi silnejšou granitizovanou a slabo granitizovanou kôrou tvoriacou podložie panónskej panvy a neogénnou sedimentáciou. (Stupeň granitizácie — rôzna konsolidácia kôry je v zmysle M. Maheľa 1978b, oblasti maximálnej subsidencie v neogéne podľa L. Stegenu et al. 1975).
Vysvetlivky: 1 — silná paleozoická granitizácia (kontinentálny typ kôry), 2 — slabá až chýbajúca paleozoická granitizácia (suboceánsky typ kôry), 3 — veľká miocénna subsidencia (vyše 1500 m), 4 — veľká pliocénna subsidencia (vyše 2500 m).

Obr. 39 Bloková stavba panónskej panvy
Hlbinné zlomy a lineamenty: 1 — peripieninský, 2 — rábsky + veporský, 3 — balatónsky (3a — Darnó), 4 — záhrebsko-kulčský, 5 — solnocký, 6 — periadriatický (insubrický), 7 — dunajský (centrálnokarpatský), 8 — drávsky, 9 — sávsky (peridinársky), 10 — murešský, 11 — vardarský, 12 — nesvačilsko-trnavský, 13 — sámošský.
Bloky: A — šopronský s podunajským, B — balatónsky, C — paleopanónsky s. s., D — tiský (C + D paleopanónsky s. l.), E — apusénsky, F — transylvánsky, G — srbsko-macedónsky, H — preddinarský, I — východokarpatský.
Vysvetlivky: 1 — sialické bloky, varisky konsolidovanejšie a sializované zóny, 2 — simatické bloky a kryhy konsolidované koncom kriedy a v terciéri, 3 — hranice panónskej panvy, 4 — hlavné hlbinné zlomy a lineamenty.

Obr. 40 Schéma znázorňujúca systém horizontálnych posunov na zlomoch, ktoré určovali vznik vnútrokarpatských panví. Oblasti vyššieho roztiahnutia sú vyznačené šrafou. Transylvánska panva je asi výnimka; veľká subsidencia tu nastala bez väčšieho roztiahnutia (podľa F. Horvátha a L. Roydenovej 1981, upravené).
Hlavný zlom s horizontálnym posunom (insubrická línia je vyznačená silnou čiarou) a doplnená podľa L. Roydenovej et al. (1982). Hrubé šípky — smer pohybu jadranskej dosky.
Vysvetlivky: VB — viedenská panva, ES — východoslovenská (transkarpatská) panva.

Obr. 41 Tektonická mapa transylvánskej panvy
Vysvetlivky: 1 — izobaty pošvy bazálneho bádenského tufitu Dej v m, 2 — antiklinálne pásma diapírových vrás, 3 — kopulovité štruktúry so zemným plynom, 4 — predpokladané skryté zlomy, 5 — okraj bázitovej kôry, 6 — okraj neovulkanitov, 7 — os symetrie rozloženia ložísk zemného plynu — os ohybu panvy, 8 — kryštalinikum; s použitím tektonickej mapy Rumunska (1970).

Obr. 42 Rekonštrukcia predtriasových typov kôry v panónskej panve
Vysvetlivky: 1 — (sub) oceánska kôra, 2 — sialická — granitizovaná kontinentálna kôra — pásma sializácie, 3 — centrá prekambriovej granitizácie (nuklea), 4 — aktívne zlomy vo vrchnom paleozoiku, 5 — predpokladané predkambriové zlomy, A — apusénske nukleá, SM — srbsko-macedónske nukleá, SC — juhokarpatské nukleá, M — murešský tróg, 1 — peripieninský lineament, 2 — rábsky hlbinný zlom, 3 — balatónsky hlbinný zlom, 4 — záhrabsko-kulčský hlbinný zlom, 5 — sámošský lineament, 6 — murešský hlbinný zlom, 7 — vardarský lineament, 8 — periapusénsky zlom.

Obr. 43 Hlbinná stavba východoslovenskej panvy
Vysvetlivky: 1 — obmedzenie zápornej reziduálnej anomálie pre hĺbku 12 km, 2 — regionálna anomália pre hĺbku 8 km (hranica kladnej a zápornej anomálie), 3 — hlbinné zlomy, (podľa O. Fusána et al. 1979), 4 — seizmoaktívne zlomy, 5 — izohypsy podložia panvy v m.

Obr. 44 Schéma typov kôry v podloží viedenskej panvy
Vysvetlivky: 1 — granitoidy, 2 — tonality; predpokladané komplexy: 3 — bázická (metabázity) a ultrabázická, 4 — pararuly s metabázitmi, 5 — pararuly, miestami slabo migmatitizované, 6 — kladné tiažové pole v odkrytej tiažovej mape (Č. Tomek — L. Budík 1981) (granitoidné horniny upravené podľa S. Dudka 1980).

Obr. 45 Smerová orientácia úsekov s maximálnou hrúbkou neogénnych sedimentov vo viedenskej panve (rekonštruované podľa Č. Tomka — L. Budíka 1981 a R. Jiříčka — Č. Tomka 1981)
Vysvetlivky: 1 — hrúbky 0—4 km (3 km), 2 — hrúbky 4—5 km, 3 — hrúbky 5 a viac km, 4 — severná hranica kladného tiažového poľa v odkrytej tiažovej mape.

Obr. 46 Mapa reziduálnych tiažových anomálií podunajskej panvy so zakresleným priebehom hlavných zlomov (podľa K. Cidlinského et al. 1984 — upravené a zjednodušené autorom)
Vysvetlivky: 1 — okraj panvy, 2 — neovulkanity, 3 — kladné reziduálne tiažové anomálie ($\mu m \cdot s^{-2}$), 4 — maximálna hrúbka neogénnych sedimentov v centrálnej pliocénnej depresii. I — inovecká hrasť, II — nitrianska hrasť, III — levická hrasť, IV — centrálna depresia.

Obr. 47 Mapa vertikálnej magnetickej intenzity (podľa K. Cidlinského et al. 1984 — upravené a zjednodušené autorom)
Vysvetlivky: 1 — okraj panvy, 2 — neovulkanity, 3 — 0—60 nT, 4—60 a viac nT.

Obr. 48 Odvodené tiažové schéma východoslovenskej panvy (K. CIDLINSKÝ et al. 1984)
Vysvetlivky: 1 — vulkanity na povrchu, 2 — zlomová tektonika, 3 — tiažové kladné zóny.

Obr. 49 Schéma vertikálnej magnetickej intenzity východoslovenskej panvy (K. CIDLINSKÝ et al. 1984)
Vysvetlivky: 1 — vulkanity na povrchu, 2 — zlomová tektonika, 3 — magnetické kladné zóny.

Obr. 50 Geologický rez podunajskou panvou (podľa B. GAŽU 1984 — upravené autorom)
Vysvetlivky: 1 — kryštalinikum, 2 — paleozoikum, 3 — mezozoikum, 4 — karpat, 5 — báden, 6 — sarmat, 7 — pliocén + panón, 8 — zlomy.

Obr. 51 Geologický rez východoslovenským neogénom (podľa R. RUDINCA 1984 — upravené a zjednodušené autorom)
Vysvetlivky: 1 — kryštalinikum, 2 — paleozoikum, 3 — mezozoikum, 4 — mezozoikum vyšších subtatranských príkrovov (humenské), 5 — mezozoikum bradlového pásma, 6 — centrálnokarpatský paleogén, 7 — magurský flyš, 8 — karpat, 9 — báden. 10 — sarmat, 11 — pliocén + panón, 12 — neovulkanity, 13 — soľonosné súvrstvie, 14 — zlomy a násun príkrovov.

Obr. 52 Mapa vertikálnych recentných pohybov vo viedenskej a podunajskej panve (podľa P. MARČÁKA et al. 1976 s doplnením moravskej časti podľa P. VYSKOČILA 1981 — upravené autorom)
Vysvetlivky: Klesanie mm/rok: 1 — od −3,5 do −4,5; 2 — od −2,5 do −3,5; 3 — od −1,5 do −2,5; 4 — od —0,5 do −1,5; 5 — od 0 do −0,5.
Zdvíhanie mm/rok: 6 — od 0 do +0,5; 7 — od +0,5 do +1,5; 8 — od +1,5 do +2,5.

Obr. 53 Hrúbka kôry pod viedenskou panvou (upravené podľa B. BERÁNKA 1978 a J. KVITKOVIČA — J. PLANČÁRA 1975)
Vysvetlivky: 1 — izolínie hrúbky kôry (v km), 2 — šipky znázorňujú horizontálne recentné pohyby podľa P. VYSKOČILA (1981).

Obr. 54 Mapa vertikálnych recentných pohybov vo východoslovenskej panve (podľa P. MARČÁKA et al. 1976 — upravené a zjednodušené autorom)
Vysvetlivky: Klesanie mm/rok: 1 — od −1,5 do −2,5; 2 — od −0,5 do −1,5; 3 — od 0 do −0,5; zdvíhanie mm/rok: 4 — od 0 do +0,5; 5 — od +0,5 do +1,5.

Obr. 55 Vzťah slovenských vnútrokarpatských panví k hlbinnej stavbe ich podložia
Panvy: 1 — viedenská, 2 — trenčianska, 3 — ilavská, 4 — handlovská, 5 — hornonitrianska, 6 — turčianska, 7 — žiarska, 8 — zvolenská, 9 — banskobystrická, 10 — breznianska, 11 — rožňavská, 12 — podunajská, 13 — juhoslovenská, 14 — turnianska, 15 — východoslovenská.
Vysvetlivky: 1 — hlbinné zlomy, 2 — seizmoaktívne zlomy, 3 — nulová izolínia oddeľujúca + a − regionálne tiažové anomálie pre hĺbku 12 km (podľa O. FUSÁNA et al. 1971), 4 — okraje panví, 5 — oblasti s hojným výskytom paleozoických bázik. Hlbinné zlomy a ich seizmicita podľa O. FUSÁNA et al. (1979).

Obr. 56 Schematická geologická mapa predneogénneho podložia transkarpatskej depresie a východoslovenskej panvy (upravené a zjednodušené podľa máp D. ĎURICU 1982 a V. G. SVIRIDENKU 1976).
Vysvetlivky: 1 — bradlové pásmo, 2 — paleogén (vnútrokarpatský) v ZSSR budínskeho typu, pri Humennom v podloží s mezozoikom — prevažne karbonátovým, 3 — kriedové sedimenty a diabázy (hlbokovodný vývoj) a paleogén, 4 — kriedové sedimenty (časté pelity), v jv. časti kontinentálneho typu, 5 — a) jura-trias (hlbokovodný s bázikami), b) nerozlíšený trias s časti kontinentálneho typu, 6 — mezozoikum nerozlíšené (miestami s paleogénom?) väčšinou flyšovitého charakteru, 7 — metamorfované paleozoikum — mezozoikum a paleogén (kontinentálneho typu), 8 — paleozoické metasedimenty, 9 — kryštalinikum — mezozoikum jednotky Čiernej hory a mezozoikum chočskej jednotky, 10 — gemerikum, 11 — zemplínska jednotka (paleozoikum, mezozoikum, paleogén) v ZSSR budínskeho typu, 12 — predpokladané zlomy, 13 — hlbinné zlomy, 14 — predpokladaný násun, 15 – hranice priečnej panónsko-volynskej depresie (V. G. SVIRIDENKO 1976), 16 — hlavné magnetické anomálie, 17 — hlavné kladné tiažové anomálie v čsl. časti.

Vyznačené zlomy: M — margeciansky, R — rožňavský, D — Darnó, P — peripanónsky (podľa V. G. SVIRIDENKU 1976), vetva smeru SV—JZ koreluje so záhrebsko-kulčským hlbinným zlomom, H — hornádsky, S — slanský, Sz — sámošský.

Obr. 57 Reliéf a tektonika predneogénneho podložia viedenskej panvy na J od flyšového pásma (podľa F. NĚMCA — A. KOCÁKA 1984 — upravené a zjednodušené autorom)
Vysvetlivky: Podložie viedenskej panvy: 1 — bradlové pásmo, 2 — flyšové pásmo, 3 — križňanský alebo manínsky príkrov a pásma vápencových Álp a vnútorných Karpát, 4 — pribradlový paleogén, 5 — glinzendorfská mulda, 6 — hranice násunu jednotlivých príkrovov, 7 — zlomy, 8 — vrstevnice povrchu predneogénneho podložia. Okrajové časti viedenskej panvy: 9 — flyšové pásmo, 10 — bradlové pásmo, 11 tatridy, 12 — krížňanský príkrov, 13 — chočský príkrov, 14 — ötscherský príkrov, 15 — lakšársky šupinovitý systém, 16 — bukovská depresia.

Obr. 58 Tektonická skica najmobilnejších častí viedenskej a podunajskej panvy
Vysvetlivky: 1 — hĺbka bázy neogénu (upravené bez zlomov podľa F. NĚMCA — A. KOCÁKA 1976 a O. FUSÁNA et al. 1971), 2 — hĺbka Moho v km (upravené podľa B. BERÁNKA 1978), 3 — profil HSS VI, 4 — zlomy.

Obr. 59 Rýchlosť molasovej sedimentácie vo viedenskej panve (čsl. časť — D. VASS — F. ČECH 1983)

Obr. 60 Hlbinná stavba severnej časti viedenskej panvy (J. ADÁMEK et al. 1984)

Obr. 61 Schéma hlavných typov grabenových štruktúr na styku epivariskej platformy a Západných Karpát na území ČSSR
Vysvetlivky: 1 — čadiče, 2 — čelo karpatských príkrovov, 3 — indikované a predpokladané hranice grabenových (riftových) štruktúr (v juhomoravskej a slovenskej oblasti predpoklad rozsahu pred neogénom); štruktúry: 4 — rift, 5 — aulakogén, N — nesvačilský graben, V vranovický graben.

Obr. 62 Rýchlosť molasovej sedimentácie v galantskej panve (D. VASS — F. ČECH 1983)

Obr. 63 Rýchlosť molasovej sedimentácie v gabčíkovskej panve (D. VASS — F. ČECH 1983)

Obr. 64 Podunajská panva (podľa B. GAŽU 1984 — upravené a zjednodušené)
Vysvetlivky: 1 — paleogén, 2 — mezozoikum, 3 — kryštalinikum, 4 — vulkanity, 5 — zlomy, 6 — izohypsy rozhrania hranice báden-sarmat.

Obr. 65 Rekonštrukcia typov kôry v kryštalickom podloží podunajskej panvy (modelové riešenie)
Vysvetlivky: 1 — okraj neogénnej panvy; kontinentálna kôra: 2 — granitoidy na povrchu, 3 — granitoidy v hĺbke (predpokladaný .ozsah); suboceánska kôra: 4 — navŕtané amfibolity, 5 — depresné zóny, 6 — hypotetická neogénna intrúzia, 7 — predneogénne depresie zaplnené príkrovmi, 8 — hlavné vrty, 9 — predpokladaná hranica kontinentálnej a suboceánskej kôry pred variskou metamorfózou.

Obr. 66 Rýchlosť molasovej sedimentácie v Ipeľskej kotline (D. VASS — F. ČECH 1983)

Obr. 67 Rýchlosť molasovej sedimentácie vo východoslovenskej panve (D. VASS — F. ČECH 1983)

Obr. 68 Čas sedimentácie hlavne vo vzťahu k depresiám
Vysvetlivky: 1 — sedimentácia, 2 — subsidencia — dvíhanie (stabilizácia), 3 — okrajová synklína.

Obr. 69 Vzťah ložísk palív k štruktúre panónskej panvy; uhoľné panvy a menšie ložiská
Vysvetlivky: 1 — severo-východná p. (Nógrád, Borsodnádasd, Mátravidék), 2 — severo-západná p., 3 — Ajka, 4 — Várpalota, 5 — Szentgál, 6 — Hodonín, Dubňany, 7 — handlovsko-novácka p., 8 — Torony—Szombathely, 9 — Tauchen, 10 — Brennberg, 11 — Wiener Neustadt, 12 — Ilz, 13 — Köflach—Voitsberg, 14 — lavanttalská p., 15 — Wiess, 16 — p. Zasavski, 17 — p. Velenje, 18 — p. Hrvatsko—Zagorje, 19 — p. Ludberg—Kopřivnice, 20 — Krško Polje, 21 — p. Pokupsko--vukomerički, 22 — p. Posavski, 23 — p. Bilogorsko-podravski, 24 — p. Mecsek, 25 — Hidas, 26 — Kolubarski p., 27 — p. Mlavsko- pečki, p. Podunavski, 28 — Lupac, Secul, Doman, 29 — Rusca—Montana, 30 — Anina, 31 — Rudăria, Mehadia, 32 — Subcarpaţii, 33 — p. Pietroşani, 34 — Codlea Cristian, 35 — Brad, 36 — Ip—Zăuani, 37 — p. Almaşului, 38 — Săemăşag, 39 — Biscad, 40 — Popeşti, 41 — p. Vrdnički, 42 — Bratca, 43 — p. Podvihorlatská, 44 — Nagyszál, 45 — modrokamenská p., 46 — p. Kúty—Gbely.

1 — hranice vnútorných Karpát, 2 — neovulkanity, 3 — hlbinné zlomy, lineamenty a zlomy, 4 — hranica panónskej panvy, 5 — uhoľné panvy a menšie ložiská, 6 — ložiská zemného plynu, 7 — ložiská ropy, 8 — ložiská ropy a zemného plynu.

Obr. 70 Výskyty uhlia a uhoľné panvy Slovenska vo vzťahu k hlbinnej stavbe
Hlbinné zlomy: 1 — peripieninský lineament, 2 — veporský, 3 — slanský, 4 — přerovsko-štiavnický; a — nesvačilsko-trnavský, b — centrálnokarpatský, c — hornádsky.
Bloky: A — podunajský, B — juhoslovenský, C — východoslovenský, D — fatransko-tatranský, E — slovensko-moravský, F — slovensko-sliezsky, G — beskydsko-bukovecký.
Výskyty a uhoľné panvy: 1 — spodno- a strednoeocénne výskyty uhlia, 2 — vrchnoeocénne a spodnooligocénne výskyty uhlia, 3 — vrchnooligocénne výskyty uhlia, 4 — burdigalské (egenburské) výskyty uhlia, 5 — helvétske (karpatské) výskyty uhlia, 6 — torténske (bádenské) výskyty uhlia, 7 — sarmatské výskyty uhlia, 8 — spodnopontské výskyty uhlia, 9 — vrchnopontské výskyty uhlia, 10 — hlbinné zlomy.

Obr. 71 Schematická mapa ložísk uhľovodíkov karpatsko-balkánskej oblasti (podľa V. SEMENOVIČA et al. 1977, upravené)
Vysvetlivky: I — Severopredkarpatská ropo-plynonosná panva: A — prevažne plynonosná oblasť vonkajšej nezvrásnenej časti predhlbne, B — prevažne roponosná oblasť vnútornej zvrásnenej časti predhlbne, II — karpatská ropo-plynonosná panva, III — viedensko-moravská ropo-plynonosná panva, IV — panónska ropo-plynonosná panva: A — malomaďarsko-podunajská plynonosná oblasť, B — stredohorsko-igalsko-bukovohorská ropo-plynonosná oblasť, C — sávsko-drávska ropo-plynonosná oblasť, D — alföldsko-banátsko-báčska ropo-plynonosná oblasť, E — zakarpatská, prevažne plynonosná oblasť, V — transylvánska plynonosná oblasť, VI — predkarpatsko-balkánska ropo-plynonosná panva: A — ropo-plynonosná oblasť predkarpatskej predhlbne, B — plynonosná oblasť moldavskej platformy, C — ropo-plynonosná oblasť severnej Dobrudže, D — ropo-plynonosná oblasť moesijskej platformy, E — ropo-plynonosná oblasť Predbalkánu, VII — severopričiernomorská ropo-plynonosná panva, VIII — západočiernomorská ropo-plynonosná panva, IX — adriatická ropo-plynonosná panva.
1 — ložiská ropy, 2 — ložiská zemného plynu, 3 — ložiská ropy a plynu, 4 — ložiská plynokondenzátov, 5 — hranice roponosných panví, 6 — hranice roponosných panví pod nasunutými Karpatami, 7 — hranice roponosných panví pod neogénnymi vulkanitmi, 8 — hranice roponosných oblastí.

Obr. 72 Geologický rez z. okraja Českého masívu do viedenskej panvy (podľa J. ADÁMKA a J. DVOŘÁKA 1984, upravené a zjednodušené autorom)
Vysvetlivky: 1 — báden, 2 — panón, 3 — karpat, 4 — spodný miocén (egenburg) na báze s klastikami, 5 — paleogén na báze s klastikami (vo vrchnej časti zvrásnený), 6 — pouzdřanská jednotka, 7 — ždánická jednotka (paleogén + najspodnejší miocén), 8 — magurský príkrov, 9 — karbonátové súvrstvie jury, 10 — bazálne klastické súvrstvie (jura?), 11 — paleozoikum (devón + karbón), 12 — kryštalinikum, 13 — presunové plochy príkrovov, 14 — prešmyky, 15 — zlomy, 16 — hiáty.

Obr. 73 Rez vrtom Zistersdorf ÜT-1 vo východnom Rakúsku (podľa A. KRÖLLA et al. 1981 — upravené a doplnené autorom)

Obr. 74 Vzťah ložísk ropy a zemného plynu k blokovej a kryhovej stavbe Západných Karpát
Vysvetlivky: 1 — vysoký tepelný tok (v $mWm^{-2}$), 2 — nízky tepelný tok (v $mWm^{-2}$), 3 — hlbinné zlomy, 4 — regionálne zlomy, 5 — hranice krýh v podloží vnútorných panví, perspektívnych pre výskyty uhľovodíkov, 6 — prognózne oblasti, 7 — ložiská zemného plynu, 8 — ložiská ropy.
Ložiská ropy a zemného plynu:
1 — Stretava, 2 — Ptrukša, 3 — Trhovište—Pozdišovce, 4 — Lastomír, 5 — Bánovce n/Ond., 6 — Špačince, 7 — Trakovice, 8 — Nižná, 9 — Madunice, 10 — Krupá, 11 — Láb, 12 — Vysoká, 13 — Studienka, 14 — Závod, 15 — Brodské, 16 — Gbely, 17 — Cunín, 18 — Petrova Ves, 19 — Štefanov, 20 — Suchohrad—Gajary, 21 — Kúty, 22 — Malacky, 23 — Jakubov, 24 — Hrušky, 25 — Jozefov, 26 — Poddvorov, 27 — Lanžhot, 28 — Bílovice—Podivín, 29 — Břeclav, 30 — Týnec, 31 — Ratiškovice, 32 — Lužice, 33 — Hodonín, 34 — Lednice, 35 — Vacenovice—Mutěnice, 36 — Lubná, 37 — Nikolčice, 38 — Měnín—Žatčany, 39 — Nítkovice, 40 — Hluk, 41 — Němčičky, 42 — Žarošice.

ZÁPADNÉ KARPATY

SÉRIA
*geológia 12*

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Vydal Geologický ústav Dionýza Štúra vo vydavateľskom opravnení Vedy, vydavateľstva Slovenskej akadémie vied v Bratislave v roku 1988

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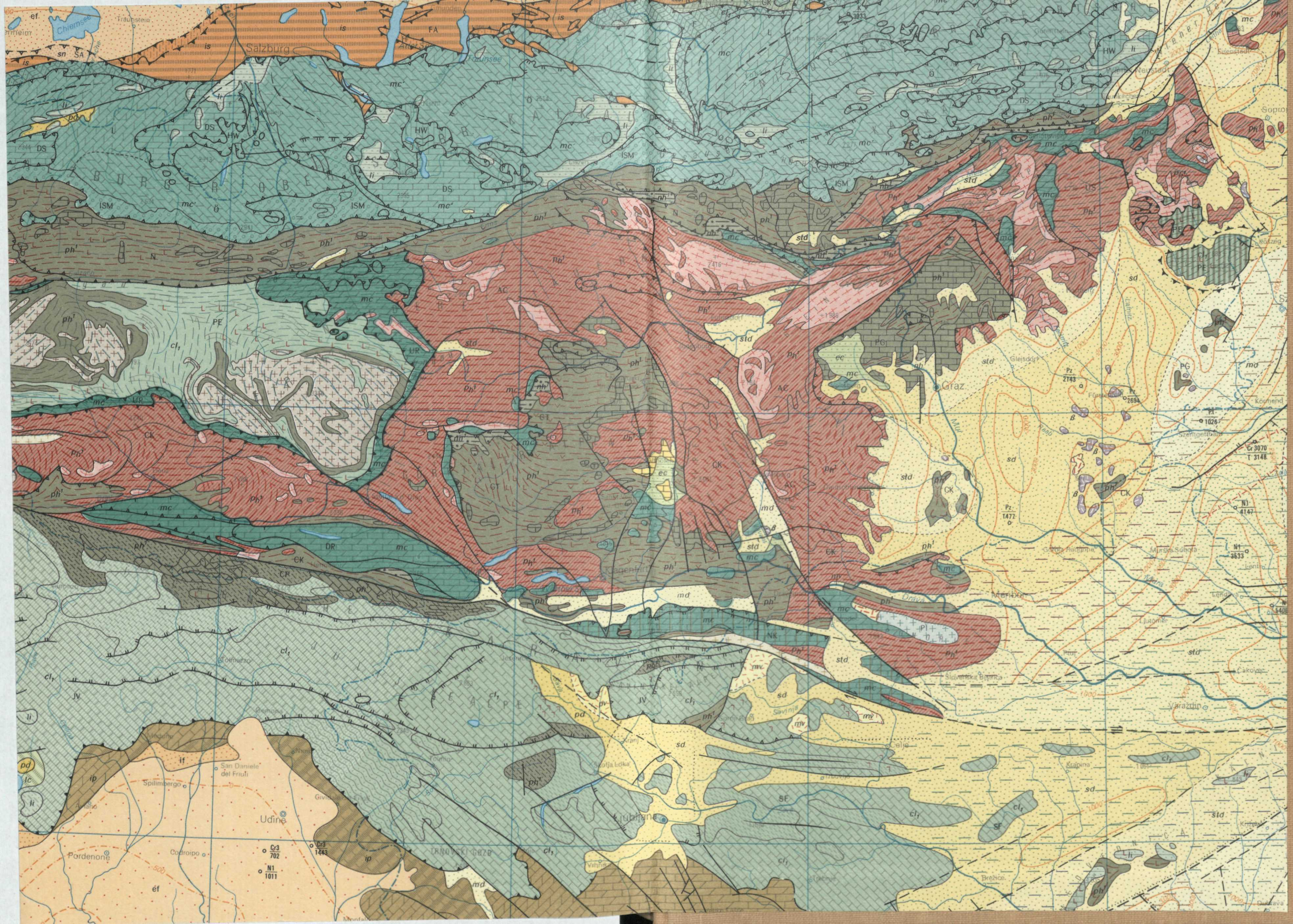
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