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Neotectonics of the central parts of the Balkan Peninsula: basic features and concepts

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With 12 figures and 1 table

Zusammenfassung

Die neotektonischen Bewegungen der Balkan-Halbinsel begannen nach den letzten intensiven Überschiebungen (frühes Miozän) und nach der früh- bis mittelmiozänen Verebnung. Gesteuert wurden die Bewegungen durch den Dehnungskollaps des spätalpinen Orogens, der Dehnung hinter dem Ägäischen Bogen und den komplizierten vertikalen und horizontalen Bewegungen in der pannonischen Region. Die neotektonische Region des Zentralbalkans liegt zwischen den linearen, neotektonischen Morphostrukturen der Strara-planina und der Dinariden-Helleniden. Sie übernahmen die alpidischen Orogenzonen der Balkaniden und Dinariden-Helleniden und wurden zum Pannonischen-, dem Präkarpatischen- und dem Nordägäischen Trog geneigt.

Die Region zeigt einen komplizierten Blockaufbau (Horst- und Grabenstrukturen), der von den NNW-SSE streichenden Struma- und Vardar-Lineamenten, von den WNW-ESE verlaufenden Sava- und Marica-Lineamenten und der Mittelmesta- und der Nordanatolischen Bruchzone dominiert war. Die Serbo-mazedonische neotektonische Schwelle war von Bruchspaltenbildung und Absenkung parallel der Struma- und Vardar-Lineamente betroffen. Die Höhe der vertikalen Versatzbeträge erreichte ein Maximum von 3-4 km; an den Rändern des Pannonischen und Ägäischen Beckens sogar mit bis zu 6 km. Die allgemeine Aufwölbung der Region wurde durch isostatische Hebung der verdickten Krustenteile (Rhodopisches Massiv) am Südrand der Eurasischen Platte bedingt. Der Kollaps der komplizierten Domstruktur begann in dessen Zentralteil entlang der Hauptlineamente (Struma-, Vardar- und Marica-Lineament) und setzte sich, während des Pliozäns und Quartärs, in den peripheren Bereichen, parallel zu den äußeren Begrenzungen (Balkaniden, Dinariden-Helleniden) der linearen Morphostrukturen, fort.

Abstract

The neotectonic movements on the Balkan Peninsula occurred after the last intense thrusting (Early Miocene), and after the Early – Middle Miocene planation. They were controlled by extensional collapse of the Late Alpine orogen, and by extension behind the Aegean arc, and were influenced by the complicated vertical and horizontal movements in the Pannonian region. The Stara-planina and Dinarian-Hellenic linear neotectonic morphostructures inherited the Alpine orogenic zones (Balkanides and Dinarides-Hellenides) and bounded the Central-Balkan neotectonic region. The linear morphostructures were tilted towards the Pannonian and Euxinian basins and the North-Aegean trough.

The Central-Balkan neotectonic region has a complicated block structure (horst-and-graben pattern) dominated by the NNW-SSE Struma and Vardar lineaments, the WNW-ESE Sava and Marica lineaments, and the Middle-Mesta and North Anatolian fault zones. The dominating Serbo-Macedonian neotectonic swell was rifted, and subsided along the Struma and Vardar lineaments. The range of the vertical neotectonic displacements reached a maximum

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of 3 – 4 km, and even up to 6 km at the edges of the Pannonian and Aegean basins. The general doming of the region was controlled by the isostatic uplift of a thickened crustal lens (Rhodope Massif) in the southern margin of the Eurasian plate. The collapse of the complicated domal structure began along the main (Struma, Vardar and Marica) lineaments in the central parts of the dome, and continued in the Pliocene and Quaternary along a more external contour bounded by the Stara-planina and Dinarian-Hellenic linear morphostructures.

Résumé

Les mouvements néotectoniques dans la péninsule balkanique ont eu lieu après les derniers charriages d'âge miocène inférieur et la pénéplanation du Miocène inférieur et moyen. Ils ont été régis par l'affaissement extensionnel de l'orogène alpin tardif, par l'extension derrière l'arc égéen et par les mouvements verticaux et horizontaux complexes dans la région panonnienne. La région néotectonique centre-balkanique est située entre les morphostructures néotectoniques linéaires de Stara-Planina et des Dinarides-Hellénides. Celles-ci sont héritées des zones orogéniques alpines des Balkanides et des Dinarides-Hellénides et ont été inclinées vers les bassins panonnien, euxinien et nord-égéen.

La région possède une structure en blocs (horsts et grabens) compliquée, dominée par les linéaments NNW-SSE de Struma et du Vardar, les linéaments WNW-ESE de Sava et de Marica et les zones faillées de Moyenne Mesta et d'Anatolie du nord. La ride néotectonique serbo-macédonienne a subi rifting et subsidence au long des linéaments de Struma et du Vardar. Les déplacements néotectoniques verticaux ont atteint 3 à 4 km au maximum, et même 6 km dans les bordures des bassins panonniens et égéen. Le soulèvement en dôme de la région a été provoqué par la montée isostatique d'une portion épaissie de l'écorce (massif du Rhodope) dans la marge méridionale de la plaque eurasiatique. L'affaissement de cette structure en dôme complexe a commencé le long des linéaments principaux (de Struma, Vardar et Marica) dans les parties centrales du dôme et a continué pendant le Pliocène et le Quaternaire le long d'un contour plus externe limité par les morphostructures néotectoniques linéaires de Stara-Planina et dinaro-hellénique.

Краткое содержание

Неотектонические движения на Балканском полуострове начались после последнего интенсивного надвига (ранний миоцен) и ранне-, среднемиоценового пенеплена. Они были обусловлены экстенсивным коллапсом позднеальпийского орогена, растяжением в тылу Эгейской дуги и сложными вертикальными и горизонтальными движениями в Паннонским регионе. Центральнобалканский неотектонический регион расположен между Старапланинской и Динаро-Эллинической линейными неотектоническими морфоструктурами. Они унаследовали альпийские Балканиды и Динариды-Эллиниды и оказались наклоненными к Паннонскому и Эвксинскому бассейнам, а также и к Северо-эгейскому трогу. Регион имеет сложное строение в виде горстов и грабенов с преобладанием Струмского и Вардарского линеаментов в простирании NNW-SSO, Савским и Марицким линеаментами в простирании WNW-OSO и Среднеместенской и Североанатолийской зонами разрыва. Сербоско-Македонский неотектонический вал был затронут рифтообразованием и прогибанием вдоль Струмского и Вардарского линеаментов. Максимальные вертикальные неотектонические движания были амплитудой порядка 3–4 км, и достигали 6 км по краям Паннонского и Эгейского бассейнов. Появление куполообразных структурных элементов в данном регионе связано с изостатическим поднятием утолщенной части Земной коры на южной окраине Евразийской плиты. Коллапс сложной куполообразной структуры произошел вдоль главных (Струмским, Вардарским и Марицкий) линеаментов в центральной части свода, и продолжался в плиоцене и четвертичном периоде вдоль более внешнего контура, ограниченного Старапланинской и Динаро-Эллинической линейными морфоструктурами.

Introduction

Basic knowledge in the fields of geomorphology, Neogene and Quaternary stratigraphy, seismology, and brittle tectonics of the Balkan Peninsula had gradually accumulated since the beginning of this century (e.g. CVIJIĆ, 1902, 1924; BONČEV, 1910). However, the synthesis of this knowledge into neotectonics began about 30 years ago. Neotectonics deals essentially with the extensional tectonics following the last major compressive event. Therefore, the neotectonic stage began diachronously during the Middle or Late Miocene.

Alpine tectonics on the Balkan Peninsula had been dominated mainly by subduction processes at the eastern and northern margin of the Vardar – Izmir – Ankara Zone (Vardar Ocean) under the southern margin of the European continent, and intense folding and thrusting within the southern active margin of the Eurasion plate. At least since the Austrian (s.l.) phase, the central parts (Rhodope Massif) of the peninsula represented a thickened crustal lens in the southernmost part of this plate. Late Cretaceous and Paleogene history proceeded under the sign of local crustal thinning, formation of volcanic island arcs, and intra- and inter-continental collisions with folding and thrusting.

The last important compression events (folding and thrusting) in the central parts of the peninsula are referred to the Early Miocene – the Late Savian tectonic phase (BONČEV, 1936; JARANOFF, 1963; MERCIER, 1977; ANĐELKOVIĆ, 1982, 1986). They are observed as south- and southwest-vergent folds and thrusts at the boards of Despotovac – Zaplanje depression (folding of the Aleksinac "series" at the Egerian – Eggenburgian boundary), Bobovdol and Brežani grabens, as well as along the northern edge of the Rhodope Mountains. The offset along some of the thrusts was up to 3 – 5 km. In a number of cases, a bivergent compression occurred (BONČEV, 1971) but the amplitude along north- and northeast-vergent thrusts was of minor importance.

The Early Miocene compression phase was followed by a prolonged planation stage which covered most of the Early Miocene (Late Aquitanian, Burdigalian) and of the Badenian. The neotectonic stage was characterized by a dominant extension expressed in normal faulting and block disintegration. It began in the Sarmatian (JARANOFF, 1963) or at the end of the Badenian (KOJUMDGIEVA, 1987), and continued during the Late Miocene, Pliocene and Quaternary.

The young Alpine mountain chains (Balkanides and Dinarides-Hellenides) were uplifted as linear neotectonic morphostructures (VAPTSAROV et al., 1986) externally rimmed by the Pannonian, Precarpathian and Ionian basins. Together with the Euxinian and Aegean basins they surrounded the Central-Balkan neotectonic region which formed essentially at the expense of the thickened crustal lens (Thracian or Rhodope massif s.l.). This region is the subject of the present paper which aims to emphasize on some of the basic neotectonic principles and ideas.

Dating of neotectonic events

Neogene stratigraphy has considerably changed during the last 30 years due to extensive research mainly within the frame of IGCP Project 25 on the Neogene of the Tethys and Paratethys. The Central Balkan neotectonic region (Fig. 1) was surrounded during most of the Neogene by marine basins (Pannonian, Precarpathian, Euxinian, Aegean and Ionian). Correlation between marine and continental faunas is difficult, and continental mammal faunas are of paramount importance for dating the events. Thus *Anancus arvernensis* CROIZET et JOBERT has been earlier referred to the latest Pliocene (e.g. JARANOFF), 1963) but is now recognized over the whole Pliocene with first occurrence in the Middle Pontian. KOJUMDGIEVA et al. (1984) and NIKOLOV (1985) defined four Mammal associations in the Late Miocene of South Bulgaria: 1) Vallesian (Mein-Zones MN 9 – 10) with primitive hipparions and *Deinotherium bavaricum* MEYER (Sarmatian); 2) Early Turolian (MN 11) with *Gomphotherium angustidens* CUVIER and *Deinotherium giganteum* KAUP (Maeotian); 3) Middle Turolian (MN 12) with *Tetralophodon longirostris* KAUP, *Deinotherium giganteum* KAUP, *Hipparion longirostris* GERVAIS etc. (Early Pontian); 4) Late Turolian (MN 13) with *Anancus arvernensis* CROIZET et JOBERT, *Deinotherium giganteum* KAUP, *Zygolophodon borsoni* Hays, *Hipparion mediterraneum* GERVAIS etc. (Middle – Late Pontian, possibly passing into the Early Pliocene). As a result of this revision, many localities and formations earlier referred to the Pliocene are dated now as Upper Miocene. A full revision of Mammal faunas and host sedimentary sequences in Bulgaria and the neighbour countries has not yet been made but dating of the major neotectonic events is quite possible.

Neogene sedimentation

The Neogene sedimentation in the Central-Balkan region was controlled by the vertical block movements in relation to the development of the large marine basins of the Paratethys (Precarpathian, Pannonian, Euxinian) and of the Tethys (Aegean). The evolution of the Precarpathian and Pannonian basins was related to escape tectonics and extension due to late (Savian, Early Styrian, Late Styrian and Moldavian phases) thrusting and closure of flysch basins in the Carpathians (SANDULESCU, ROYDEN and others in ROYDEN & HORVATH, eds. 1988). Marine sedimentation was almost continuous (Fig. 2) in the Pannonian basin where transgression – regression cycles and volcanism were closely related to the principal tectonic events (e.g. BERCSI et al. in ROYDEN & HORVATH, eds., 1988). In the Pannonian basin the subsidence tendency passed from the limnic-alluvial sedimentation in the Late Oligocene

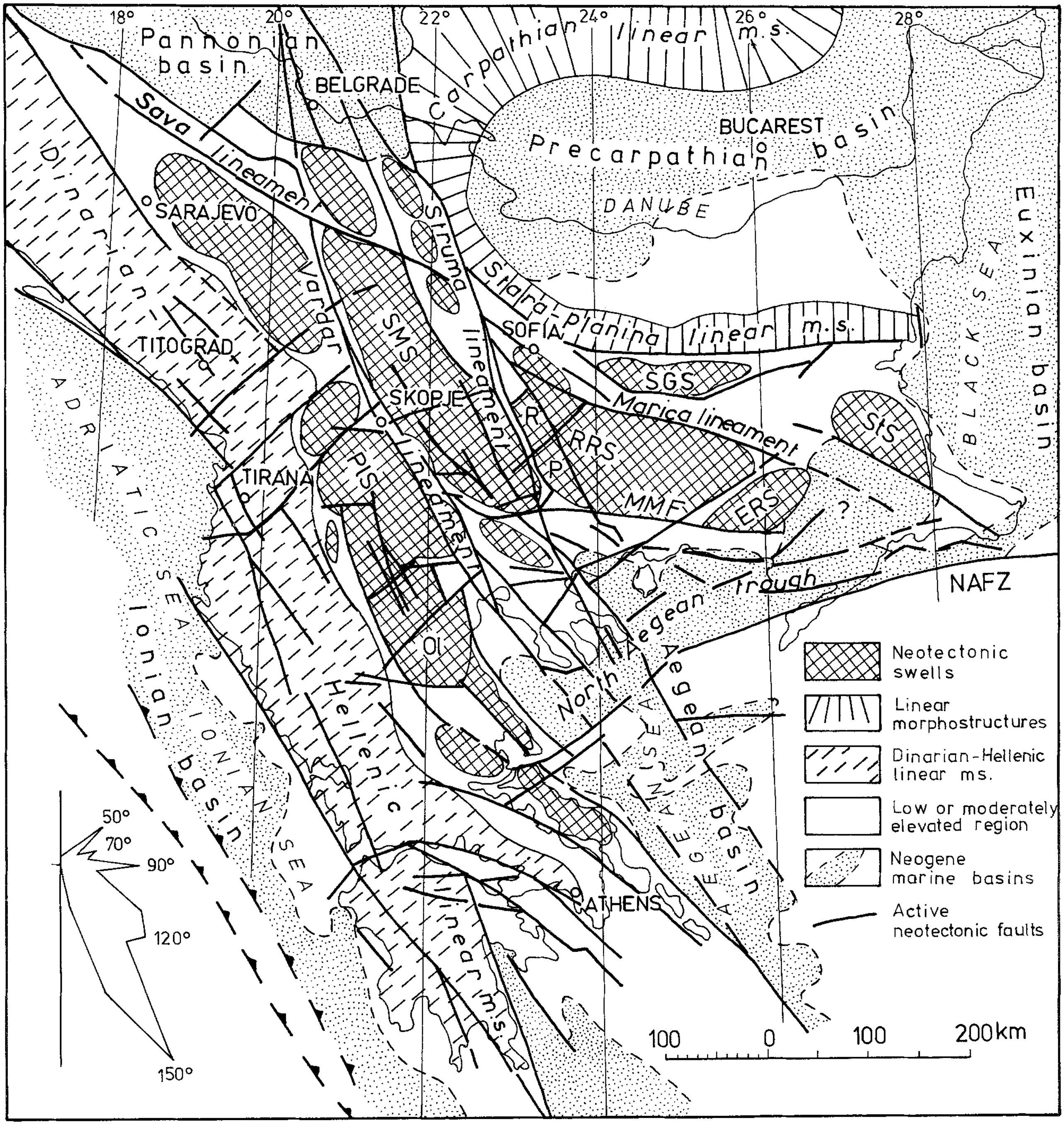

Fig. 1. Neotectonic sketch of the Balkan Peninsula. A part of the recent Hellenic arc is shown with thick dashed barbed line. Important neotectonic structures: SGS – Sredna-gora swell; PlS – Pelagonian swell; RRS – Rila-Rhodope swell; StS – Strandza swell; ERS – East-Rhodope swell; R – Rila swell; P – Pirin swell; Ol – Olympos structure; MMF – Middle-Mesta fault zone; NAFZ – North-Anatolian fault zone. Inset: rose-diagramme for the principal neotectonic faults (balanced on their lengths).

and Early Miocene (Egerian) after a slight manifestation of the Early Styrian ("Savian") phase into the Eggenburgian transgression which was followed by cyclic deposition of marine sediments in extensional environments up to the Moldavian phase in the Late Sarmatian. Then the marine sedimentation has been inherited by a dominant limnic sedimentation. In the other basins, however, the marine sedimentation began considerably later. Data shown on Fig. 2 refer mainly to the southern (next to the Balkanides) part of the Precarpathian basin, and to the westernmost part of the Euxinian basin, and are based on generalizations by KOJUMDGIEVA & POPOV (1989) and PARASCHIV (1983) as well as on other published in-

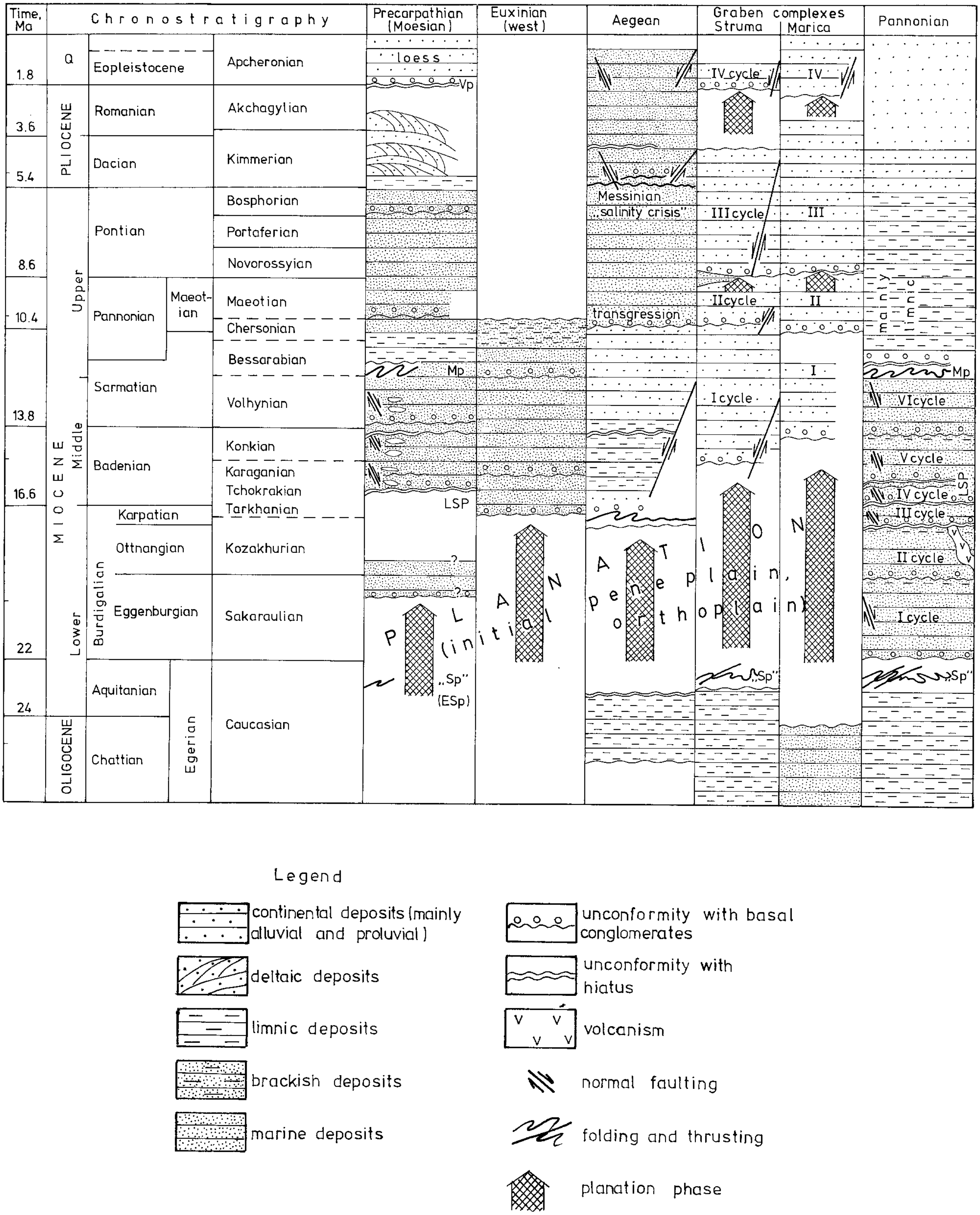

Fig. 2. Chart of the main neotectonic events. After data of ROYDEN & HORVATH, eds. (1988); PARASCHIV (1983); KOJUMDGIEVA & POPOV (1989); DERMITZAKIS (1990) and other sources. Abbreviations of tectonic phases: SP – Savian; ESP – Early Styrian; LSP – Late Styrian; MP – Moldova; VP – Valachian.

formation. After a short ingression at the end of the Early Miocene, the principal transgression of the Precarpathian basin occurred in the Badenian, and the sedimentation continued in the Badenian and Sarmatian in extensional environment with accumulation of thick olistostroms. The transgression within the Euxinian basin was of Tarkhanian and Tchokrakian age. The Aegean region had been occupied by limnic-alluvial systems in extensional environment since the beginning of the Middle Miocene to the end of the Sarmatian, the marine sedimentation beginning in the Late Miocene (KOJUMDGIEVA, 1987; SCHROEDER in JACOBSHAGEN, 1988; DERMITZAKIS, 1990). Therefore, the whole area of these basins as well as the territory later surrounded by them had been subject to planation during a period of 3 to 9 Ma. Long-lasting Early to Middle Miocene planation was the principal common feature of the whole region. It resulted in the formation of the initial peneplain (orthoplain, protopeneplain).

The neotectonic development of the Central Balkan neotectonic region was dominated by the deformation and block disintegration of the initial peneplain (orthoplain), its delevelling and fossilization by thick continental sedimentary sequences within the subsided blocks. The neotectonic depressions formed in several graben complexes (Fig. 2). The Vardar and Struma graben complexes were controlled by the most intense vertical movements along the Vardar and Struma lineaments. The Marica (related to Marica lineament) and Sub-Balkan (bound to the intense uplift of the Stara-planina linear morphostructure) graben complexes were seated between the Rila-Rhodope swell and the Stara-planina linear morphostructure. The latter isolated these graben complexes from the Precarpathian basin whereas direct relations with, and draining within the Pannonian basin (through the graben complex along the Sava lineament), the Euxinian basin, and the Aegean basin (through the Vardar, Struma and Marica graben complexes) existed. Neotectonic grabens south of the Middle-Mesta fault zone had a direct link to the North-Aegean trough and the Aegean basin.

As a rough approximation, four regional neotectonic sedimentation cycles can be distinguished. Three of them belonged essentially to the Late Miocene and the earliest Pliocene (e.g. KOJUMDGIEVA, 1987), and the fourth one, to the Late Pliocene – Early Pleistocene. Each cycle began usually with coarse proluvial – alluvial sediments. During some of the cycles coal-forming lake-swamp environment developed. At the end of each cycle a phase of comparative tectonic quiescence resulted in formation of imperfect planation surfaces (oroplains).

The Marica (KOJUMDGIEVA et al., 1979; DRAGOMANOV et al., 1981; NEDJALKOV & KOJUMDGIEVA, 1983; NENOV et al., 1990) and Struma (NEDJALKOV et al., 1988, 1990; KOJUMDGIEVA et al., 1984) graben complexes are the best studied. In both cases (Fig. 2) the Late Oligocene sedimentation (with a marine ingression along the Marica lineament) ended with the Savian phase and the Early Miocene planation. The first cycle covered the Late Badenian and most of the Sarmatian. The second cycle began with internal wash-outs and coarse deposits at the end of the Sarmatian or at the beginning of the Maeotian, i.e. at the time of the marine transgression in the Aegean region. It ended with a short planation episode and formation of the second planation surface (oroplain I). The third cycle began at the end of the Maeotian or in the beginning of the Pontian with intense contrasting vertical movements and formation of thick proluvial fans in the grabens. A short marine ingression took place in the southernmost part of the Struma graben complex (Serres graben) south of the Middle-Mesta fault zone. The third cycle continued during the Pontian and most of the Dacian, and the sedimentation ended usually in the beginning of the Romanian. A long planation with formation of one or two planation surfaces (oroplains II and III) followed to the end of the Pliocene when the fourth neotectonic regional cycle began. It continued in the Eopleistocene and the beginning of the Pleistocene with deposition of proluvial and alluvial sediments, and by exception, with limnic sedimentation in isolated grabens.

Local differencies are well visible when comparing (Fig. 3) the sequences within the individual grabens. Thus, the Marica graben complex consisted of two (Plovdiv and Zagora) distinct depressions with a different development (DRAGOMANOV et al., 1981). The thickness of the Neogene and Quaternary sediments rarely exceeded 400 m.

In the Zagora depression the sediments of the first regional cycle (Gledačevo Formation) followed partly with gradual transition an older coal-bearing Marica Formation which could be of Late Oligocene – Middle Miocene age. The covering Ahmatovo Formation began in the Plovdiv depression in the Maeotian and continued uninterrupted up to the Romanian (according to DRAGOMANOV et al., 1981, even in the fourth regional cycle) whereas in the Zagora depression it was clearly discordant and embraced only the third cycle (from the Late Pontian to the Early Romanian).

The Karlovo graben (ZAGORČEV et al., 1973; ANGELOVA et al., 1991) is a typical representative of the Sub-Balkan graben complex. The thickness of the Neogene (Upper Miocene – Pliocene) lacustrine and alluvial – proluvial sediments reached 300 m, and that of the Quaternary alluvial and proluvial sediments, up to 40 – 70 m. The Sofia graben (KAMENOV & KOJUMDGIEVA, 1983) developed at the junction between the Sub-Balkan and Marica graben complexes. If all age determinations are correct, the development of the Sub-Balkan graben complex, Sofia graben included, differs from that of the other graben complexes. The first sedimentation cycle (variegated terrigeneous formation in the Sofia graben; not preserved in the other grabens) is referred to the Maeotian, and the coal-bearing second local cycle, – to the Pontian and the beginning of the Dacian (Fig. 3).

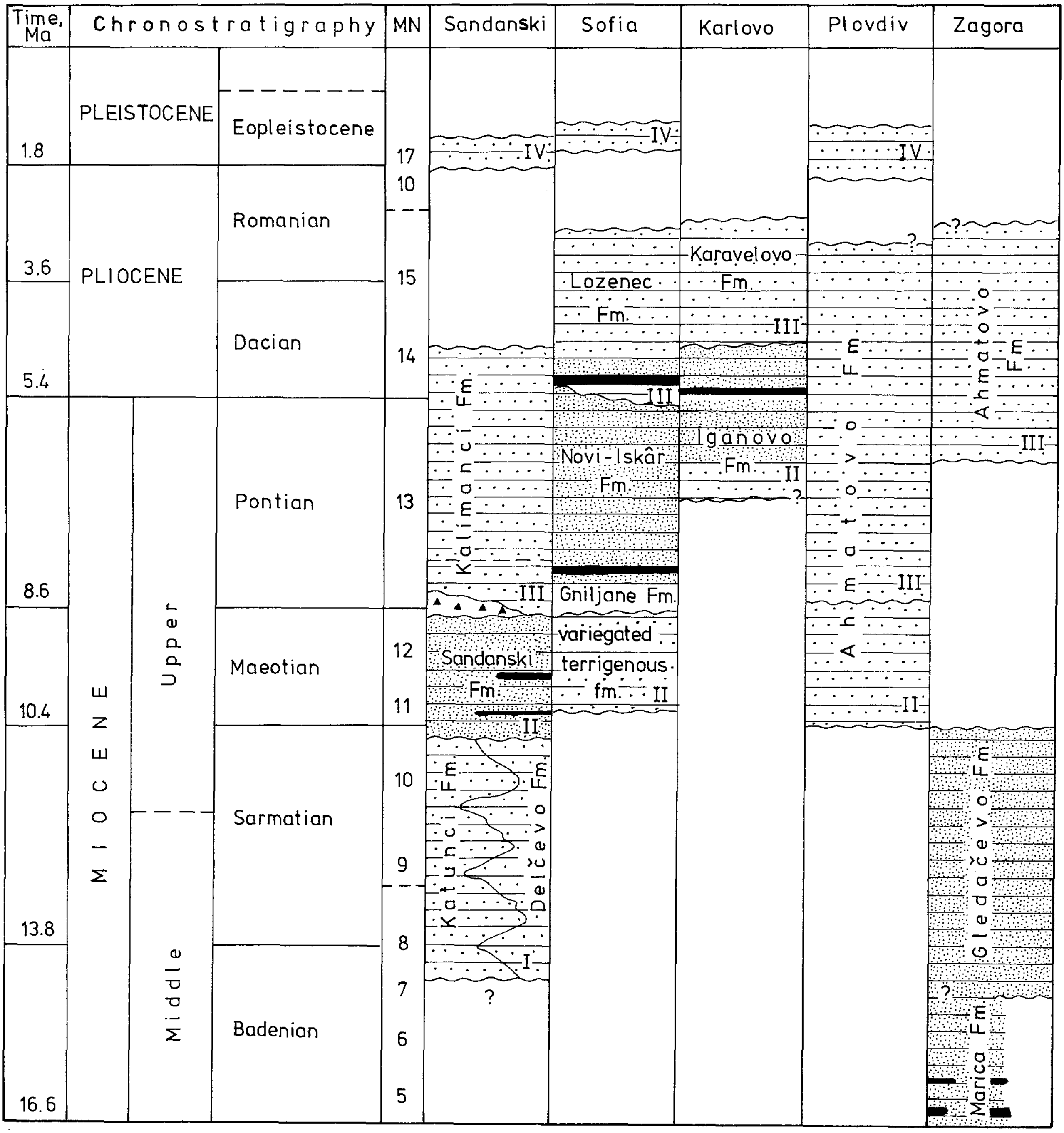

Fig. 3. Late Miocene to Pleistocene formations in some depressions. After data of NEDJALKOV et al. (1986), KOJUMDGIEVA et al. (1979, 1982), DRAGOMANOV et al. (1979), NEDJALKOV & KOJUMDGIEVA (1983), KAMENOV & KOJUMDGIEVA (1983), ANGELOVA et al. (1991). Coal-bearing limnic deposits densely stippled; coal seams black.

Lake and swamp environments favourable to coal formation have been diachronously created in the different graben complexes (Fig. 3), with a general younging tendency from the south to the north. As already mentioned, the age of the coal-bearing Marica Formation in the Zagora depression (Marica graben complex) is subject to controversy.

Planation surfaces and neotectonic morphostructures

The study of planation surfaces has a long history. The pioneer studies by CVIJIĆ (1902, 1924) were followed by many geomorphologists (e.g. GALABOV, 1982). Important neotectonic ideas about the deformation of older planation surfaces through long-radius doming and block movements were brought forth by JARANOFF (1963) and LILIENBERG (1966), and positive neotectonic morphostructures were recognized and described by ZAGORČEV (1969, 1970), ZAGORČEV et al. (1973), VAPTSAROV & DILINSKA (1980), PSILOVICOS (1984, 1986) and VAPTSAROV et al. (1986).

The initial peneplain (orthoplain) is of an Early – Middle Miocene age (Fig. 2). It is marked by intense weathering, sometimes with a weathering crust developed over different basement (from Precambrian metamorphics to Upper Oligocene – Lower Miocene sedimentary formations). The erosional (denudation) character is sometimes complicated with partial accumulation of a thin pebble gravel cover. Relics of older relief are preserved as isolated peaks. The present-day altitudes of the orthoplain vary within a broad range. They decrease gradually, and the subsided orthoplain in the neotectonic depressions is fossilized by younger (Badenian – Dacian) sediments (Fig. 4; Table 1).

The other planation surfaces (oroplains, pediments) are observed at different altitudes as mountain steps. They are usually inclined the dips varying from 0.5 to 3 – 5°. The age of oroplain I is determined as latest Maeotian because it is sealed by the Ilindenci Member of the Pontian Kalimanci Formation in the Sandanski graben (ZAGORČEV, 1970, 1991). Oroplains II and III are probably of Late Pliocene (Romanian, or latest Dacian – Romanian) age.

Younger planation surfaces often contain relics of the older ones. A complete set of well-developed planation surfaces may be observed only in those horsts that underwent the strongest differential uplift as e.g. the Pirin horst. In some neotectonic morphostructures the planation surfaces represent polyfacial and heterochronous surfaces (composite planation surfaces) which contain relics of the older truncated surfaces and of the older pebble gravel cover. The low contrast of the uplift movements in epochs of planation and of faster uplift lead to combination of two or more surfaces into a single composite surface thus rendering the geomorphological and neotectonic studies considerably more difficult.

The correlations between planation surfaces and depression-filled sediments as well as the analysis of the attitudes of planation surfaces and the dips of Neogene beds make possible to delineate the positive neotectonic morphostructures (linear morphostructures and swells). First data about these structures have been given in Bulgaria by JARANOFF (1963) but a more extended neotectonic analysis has been made by VAPTSAROV et al. (1986). The analysis of the planation surfaces allows for also to distinguish epochs of extension and graben tilting followed by planation, the new planation surfaces sealing the tilted sediments and the bounding faults.

The Stara-planina linear morphostructure (VAPTSAROV et al., 1986) almost coincides with the Stara-planina and Fore-Balkan zones of the Balkanides (cf. BONČEV, 1946, 1971). To the south it is limited by the Sub-Balkan neotectonic fault zone (amplitide of c. 1.5 – 2 km). The initial peneplain (orthoplain) was uplifted to an altitude of 1600 – 1800 m (GALABOV, 1982). It dips north, its altitude decreasing along a S-N profile (Fig. 4) from the Middle Stara-planina Mts. to the Danube (GALABOV, 1982; VELEV et al., 1988). The oroplains are also dipping north and east. Thus the Stara-planina linear morphostructure was acting as southern barrier for the Precarpathian basin, and was connected to the north (Fig. 1) with the Carpathian linear morphostructure. Its southern boundary (Sub-Balkan fault) was a zone of high-amplitude neotectonic normal faults (Figs. 4, 5) which became less prominent to the east (BONČEV, 1910).

The Dinarian-Hellenic linear morphostructure (Fig. 1) outlines the central parts of the peninsula from the southwest separating them from the Ionian basin. The linear character is less pronounced than the Stara-planina morphostructure. The altitudes of orthoplain and oroplains are comparable (Table 1) and planation surfaces and neotectonic faults become less distinct to the south. The abundance of Mesozoic carbonate rocks in the pre-Neogene basement controlled wide-scale Karst phenomena.

The backbone of the Central-Balkan neotectonic region is the Serbo-Macedonian swell (ZAGORČEV, 1969, 1970, 1991; Macedonian Swell – VAPTSAROV et al., 1986). Its central part (about 350 km long and 90 km wide) is elongated NNW-SSE and is limited

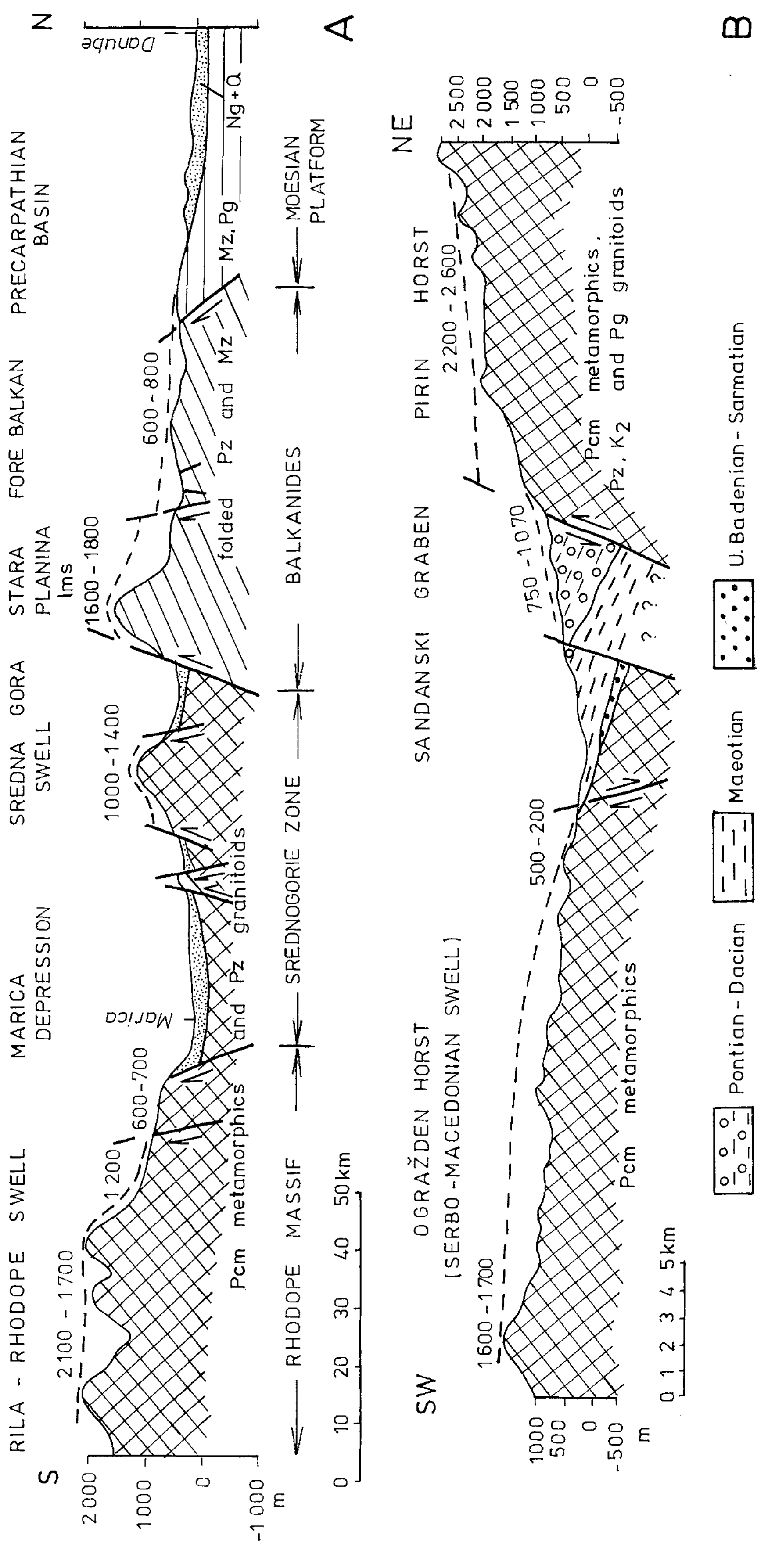

Fig. 4. Schematic sections for the deformation of the Early – Middle Miocene orthoplain (initial peneplain). The traces of the sections are shown on Figs. 7 and 9.

Morphostructure	maximum height, m	dipping gradually to m
STARA-PLANINA	1600 – 1800	east: 1000–800 (600–520) north: 1400–1000
Fore-Balkan (reconstructed)		1000–600
Danube depression (minimum)		–400
DINARIAN-HELLENIC	1800 – 2400	
SERBO-MACEDONIAN SWELL	1500 – 1800	east: 1200–200
in Sandanski graben		deeper than–1000
RILA-RHODOPE SWELL	1700 – 1900	east and south: 1520 – 1450 – 1000
PIRIN HORST	2200 – 2600	south: to 2000 m
PELAGONIAN HORST (Olympos, Pindos)	2000 – 2600	
SREDNA-GORA SWELL	1300 – 1400	north and south: to 700–800
in some grabens:		to 450 and less

Table 1. Altitudes of the initial peneplain (orthoplain) in some neotectonic morphostructures

between the Vardar and Struma lineaments. The orthoplain is situated in the highest parts (Fig. 5) at 1500 – 1800 m, and dips ENE and WSW in the eastern and western flanks, respectively. Neogene beds in the adjacent NNW-SSE grabens also dip outwards from the swell axis. In the most subsided Sandanski and Simitli grabens the orthoplain was sunken to about 1000 m below sea level. Oroplain I is at an altitude of 1200 – 1600 m in the highest parts of the swell, also gradually decreasing in height at the periphery, and being partially superimposed over the deformed orthoplain (Fig. 5). In the Sandanski graben, oroplain I is built over the Maeotian sediments of the Sandanski Formation at an altitude of ca. 400 – 520 m, and is sealed by the Pontian Kalimanci Formation (ZAGORČEV, 1970, 1991). Oroplain II reaches in the highest parts of the swell 900 – 1200 m, and oroplain III, 650 – 1000 m.

The Rila-Rhodope swell ("Rhodope meso-dome" – VAPTSAROV et al., 1986) is another important

Fig. 5. Planation surfaces in the highest parts of the Ogražden horst (Maleševska Mts., Serbo-Macedonian swell). The peak Kadiyca (1924 m) with the orthoplain gradually lowering from 1800 – 1850 to 1750 – 1700 m; the lower surface (oroplain I) at 1540 – 1590 m. View towards west.

Fig. 6. Planation surfaces in the central parts of the Pirin horst. The ridge Stražite with the orthoplain at about 2600 m, and oroplain I as two separated surfaces at 2350 – 2400 and 2200 – 2280 m, with the lakes Vasilaški ezera (glaciation relics).

domal neotectonic structure measuring about 200 x 90 km. It is elongated WNW-ESE, and is situated between the Marica, Struma, and Middle-Mesta lineaments (Fig. 1). The Rila and Pirin horsts represent secondary swells developed over the highest uplifted parts of the orthoplain. At the one hand, they are closely related to the Rila-Rhodope swell, and at the other, the attitude of the orthoplain (which dips east and south) corresponds to the eastern limb of the Serbo-Macedonian swell here uplifted towards the rifted and subsided central parts of the latter (ZAGORČEV, 1969, 1970). The orthoplain (Fig. 6) is uplifted to 2200 – 2600 m with a gradual lowering to about 2000 m. Oroplain I reaches 1900 – 2400 m, oroplain II, – 1500 – 1800 m, and oroplain III is lowering from 1100 – 1300 m to 750 – 900 m at the periphery of the horst where it seals the faults of the West-Pirin fault zone. Most characteristically, the planation surfaces in the highest horsts are monofacial, well differentiated and easily discernible.

The Pelagonian swell (200 x 90 km) is elongated NNW-SSE, and is situated between the Vardar lineament and the Dinarian-Hellenic linear morphostructure. Some of the highest mountains (Olympos, Pindos) belong here, and the orthoplain is elevated (Table 1) at 2000 – 2600 m. However, the neotectonic uplift of these mountains occurred relatively later, probably at the Pliocene – Quaternary and Early – Middle Quaternary boundaries (FAUGÈRES, 1977).

The Sredna-gora swell (150 x 40 km) is elongated W-E (Fig. 7), and is situated between the Marica lineament and the Stara-planina linear morphostructure (Sub-Balkan fault zone) (ZAGORČEV et al., 1973; VAPTSAROV et al., 1986). The orthoplain, as well as the Neogene sediments filling in the asymmetrical tilted grabens at the northern and southern peripheries, are dipping to the north and south, respectively, the dip angle of the deformed orthoplain reaching 5 – 10°.

All neotectonic swells are situated divergently around the Serbo-Macedonian swell and the Vardar and Struma lineaments. The altitudes of the orthoplain in the swells point at a possible rifting and relative subsidence of the Serbo-Macedonian swell (ZAGORČEV, 1970, 1991). Considered along a SW-NE cross section, the whole Central-Balkan neotectonic region may be regarded as a huge domal neotectonic structure which subsided at two contours: against the Stara-planina and the Dinarian-Hellenic linear morphostructures (external contour), and the Serbo-Macedonian swell against the most elevated horsts of the adjacent Rila-Rhodope and Pelagonian swells (internal contour).

This analysis is entirely based on observations upon the position of the orthoplain (Fig. 4), and therefore concerns the total neotectonic domal elevation. After the Maeotian (mainly during the Pliocene and the Quaternary), further disintegration into horsts and grabens, and second-order deformations within the horsts transformed some of the latter into second-order domal structures (swells). However, in most cases the very low-angle dips of the oroplains (of the order of 0.5 – 3°) towards the periphery of the swells might represent normal inclinations of these erosional surfaces towards the draining palaeorivers rather than a result of tectonic doming.

Fault neotectonics

Most of the meotectonically active faults and fault zones inherited pre-existing faults or fault sets. First-order fault structures possibly had a rhegmagenic origin (BONČEV, 1961, 1971). During the neotectonic stage, mostly steep and moderate-angle normal faults have been active although some very low-angle normal faults are recorded in the Struma graben complex. Strike-slip components have been inferred for some of the large fault zones (ZAGORČEV, 1969). Thus the post-Cretaceous right-lateral strike-slip along the NNW-SSE Struma fault zone is estimated at about 4 – 4.5 km (ZAGORČEV, 1969), and a right-lateral strike-slip component of unknown magnitude, – for the Middle-Mesta fault zone (remote-sensing analysis by GOČEV & MATOVA, 1989).

First-order faults (lineaments, fault bundles) may be classified on the basis of different criteria. Maxima on rose-diagrammes for the relative lengths of the major faults (Fig. 1) reveal the predominance of NNW-SSE striking faults as the Vardar and Struma lineaments followed by WNW-ESE (Marica and Sava lineaments), SW-NE (North-Anatolian fault zone) and W-E (Middle-Mesta fault zone). The orientation of the maximum normal stresses changed from NNE-SSW (ca.10°) in some stages, to approximately W-E, in the other (BONČEV, 1961, 1971; ZAGORČEV, 1969).

Another classification may be based upon the position of the faults as: 1) boundaries of neotectonic linear morphostructures; 2) first-order faults (fault swarms, lineaments); 3) boundaries of neotectonic blocks (horsts and grabens). Positive linear morphostructures are usually bounded by normal faults, their strike and dip changing rapidly in function of the curvature of the morphostructure. At the places of abrupt change of direction, transversal faults with a wrench component may develop, or both the strike and the dip of the master fault would change gradually producing a number of feathering faults. A typical example of this type is the Sub-Balkan fault zone (Fig. 7) which serves as a southern boundary to the Stara-planina linear morphostructure. As previously mentioned, the orthoplain within the latter is gradually dipping north into the Moesian platform. The vertical uplift along the Sub-Balkan fault zone reached 1.5 – 2.0 km (VRABLIANSKI, 1975). The uplift of the Stara-planina linear morphostructure was partially compensated by the subsidence of the Sub-Balkan graben complex (Figs. 7, 8) on the northern flank of the Sredna-gora neotectonic swell, and the formation of thick lacustrine and proluvial-alluvial deposits. Normal faulting proceeded locally along tilted surfaces (JARANOFF, 1960) of the older (Palaeogene) Stara-planina thrust.

The Marica lineament ("Marica suture" – BONČEV, 1946, 1971; BOYANOV & YOSIFOV, 1986) is a complex fault swarm traced at a distance of about 500 km. The most prominent neotectonic structure is the North-Rhodope fault zone with a maximum neotectonic upthrow of about 2000 m (in the westernmost part) decreasing to about 700 m in the eastern part (VRABLIANSKI, 1975). The uplift of the complex Rhodope neotectonic horst has been compensated by deposition of thick (up to 500 – 600 m) Neogene and Quaternary sediments into the Marica (Upper Thrace) graben complex. It consists of a number of grabens (the most important being the Plovdiv and Zagora depressions) and internal horsts (Fig. 7), some of the horsts being partially or entirely covered by Neogene and Quaternary deposits. At the northern edge of the Marica graben complex, a number of small horsts are bounded from the Sredna-gora swell by the South-Srednagora (Panagyurište) fault zone. Several small neotectonic grabens expressed as poljes (of Panagyurište, Strelča, Krasnovo, Hisar) are tilted to the south (Figs. 4, 7) against the Panagyuriste fault zone, and in accordance with the deformation of the orthoplain within the Sredna-gora swell (ZAGORČEV et al., 1973). The Marica lineament is apparently dying out west of the Sofia graben being displaced (right-lateral strike-slip) by the Struma lineament with a possible continuation in the Sava lineament on Yugoslav territory.

The Sava lineament is formed mainly by the Sava and Sprec-Kozarac dislocations (ANĐELKOVIĆ, 1982, 1986). It has a complicated internal structure. Faults of this lineament are bounding the Pannonian basin to the south, and neotectonic subsidence in some (Sava, Drava) grabens has been up to 6 km. The total length of the Marica and Sava lineaments exceeds 1000 km.

The parallel NNW-SSE Struma (Kraištid) and Vardar lineaments are also traced on more than 1000 km (BONČEV, 1961; ZAGORČEV, 1991). The Struma lineament (Fig. 9) consists mainly of normal faults and right-lateral strike-slip faults. Vertical neotectonic amplitudes along some of the faults (e.g. the West-Pirin fault zone) reached up to 3 – 3.5 km (JARANOFF, 1963), and the thickness of Neogene deposits in the Struma graben complex, up to 1.6 km (Sandanski graben) and 3 km (Serres graben). A general tilting of the grabens to the east (Fig. 4) is recorded (ZAGORČEV, 1970) corresponding to their position in the eastern flank of the Serbo-Macedonian swell. The Vardar lineament is located in the western flank of the swell, and accordingly, the orthoplain

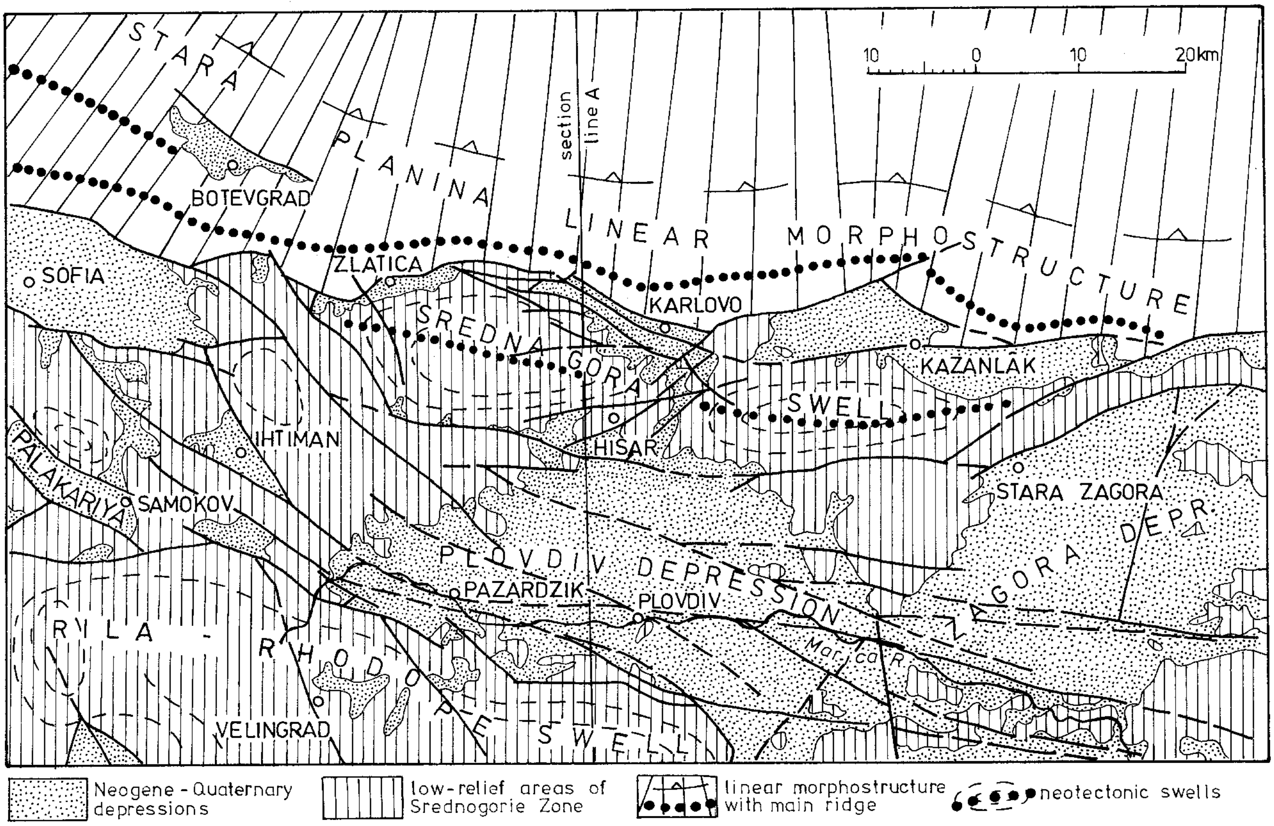

Fig. 7. Generalized neotectonic map of the Sredna gora swell and the Marica graben complex (Upper Thracian depression). The neotectonic grabens (Sofia, Zlatica-Pirdop, Karlovo and Kažanlak) south of the Stara-planina linear morphostructure and the Sub-Balkan fault zone are also shown.

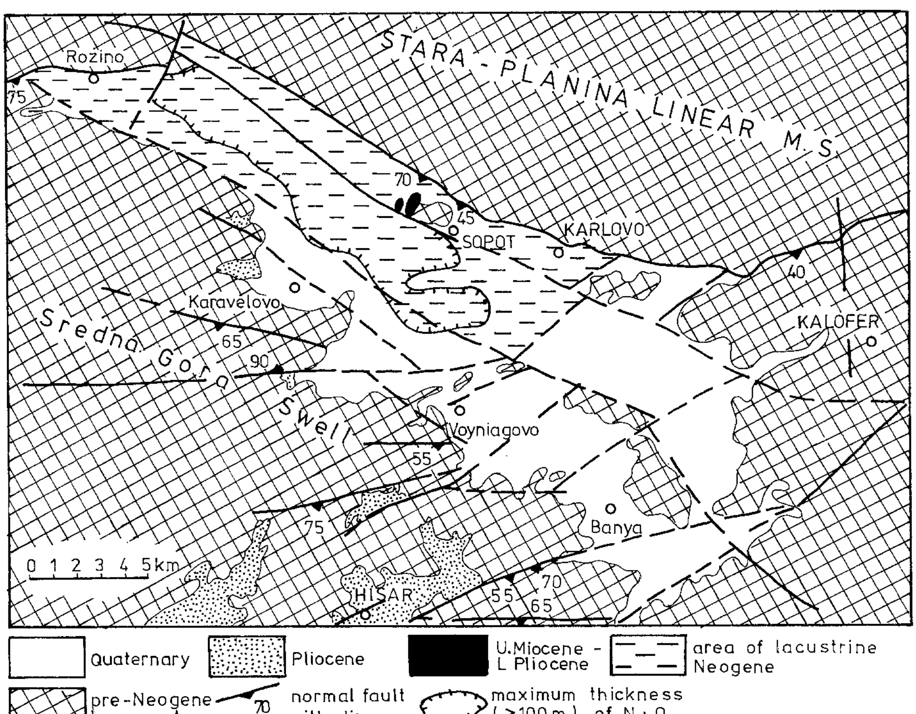

Fig. 8. Simplified neotectonic map of the Karlovo graben. After ZAGORCEV et al. (1973), ANGELOVA et al. (1991) and unpublished data of the author.

and the Neogene sediments are dipping west (ZAGORČEV, 1969). The thickness of the Neogene sediments in some of the depressions is up to 2 – 3 km (ARSOVSKI et al., 1975), and the vertical neotectonic amplitudes may be as high as 5 km. The Vardar and Struma lineaments and the Serbo-Macedonian swell form a complicated continental rift structure (ZAGORČEV, 1969, 1991) with maximum subsidence along the lineaments (graben complexes). Rift subsidence along the Vardar and Struma lineaments continued probably also north of the Sava lineament (in the Pannonian basin) and south of the North-Aegean trough (in the Aegean basin), and was sealed there by very young (Quaternary) sediments.

The Middle-Mesta fault zone (lineament) is traced at a distance of about 300 km in west-east direction (JARANOFF, 1960; ZAGORČEV, 1975; GOČEV & MATOVA, 1989). It is marked by differentiated neotectonic block movements most contrasting at the northern and southern slopes of the Belasica (Kerkini) horst at the intersection with the Serbo-Macedonian swell. The vertical amplitudes reach up to 2 – 2.5 km (JARANOFF, 1960, 1963). The Middle-Mesta lineament represented the northern boundary of Neogene marine ingressions from the North-Aegean trough. A right-lateral displacement along some of the faults is inferred from remote sensing (GOČEV & MATOVA, 1989). The lineament may be regarded as a branch of the North-Anatolian fault zone (e.g. MERCIER, 1977; SENGÖR & CANITEZ, 1982; SENGÖR et al., 1985).

The North-Aegean trough developed (Fig. 1) over the faults belonging to the North-Anatolian fault zone (lineament). It is a transversal rift structure characterized by considerable extension and normal faulting (including along low-angle normal faults), and a right-lateral strike-slip component. The thickness of the Neogene deposits reached up to 1.5 – 3.5 km (LALECHOS & SAVOYAT, 1977; LYBERIS, 1984), the total neotectonic amplitude being up to 5 – 6 km. By computation of load effects, the subsidence within the trough for the last 10 Ma has been estimated (LE PICHON et al., 1984) at 1 – 1.5 km reaching up to 2.5 km for the Thermaikos Gulf at the intersection of the trough with the Vardar lineament.

Transversal fault zones striking SW-NE are most important for the development of the Vardar and Struma lineaments and the whole Central-Balkan neotectonic region. They have been traced (e.g. Peć geofracture – cf. ANĐELKOVIĆ, 1982; Brežani fault zone – ZAGORČEV, 1969, 1991) at distances of about 100 km. Their intersections with major faults of the Struma and Vardar lineaments are characterized by most contrasting (3 – 4 km) vertical neotectonic movements, and represent seismotectonic knots (BONČEV et al., 1982) of very intense seismicity.

Although faults striking parallel to the main lineaments are predominant, faults belonging to the other sets are also frequent. The vertical movements along them created a complicated pattern of neotectonic block structures (horsts and grabens) which often deviated from the general tendencies described. Feathering and anostomosing faults made the pattern even more complicated, and transition from medium-angle to low-angle faults controlled additional block tilting and adjustments. For example, faults of the West-Pirin fault zone (separating the Sandanski graben from the Pirin horst) had a gradually decreasing westward dip from ca. 70° in the north to about 25 – 30° (Fig. 10) in the south. At the place where the angle of dip decreased to about 40° the fault splayed, and one of the branches (Melnik fault) took over a fraction of the vertical amplitude (ZAGORČEV, 1970, 1991) displacing the eastern part of the Sandanski graben. The adjustments between the adjacent blocks resulted in a counterclockwise rotation of the Pirin block estimated at between 3 and 11°.

In the cases of low-angle faults, the horizontal component across the fault strike may be greater than the vertical component by a factor reaching 2.5. In the case of the West-Pirin fault zone (vertical component about 3.5 km) the horizontal component may reach 8 – 9 km. An extension of about 40% may be calculated over the southern part of the Sandanski graben. However, when computed over the whole length of the section, and taking into account the curvature of the swells, the total WSW-ENE extension in that area is estimated at 3 – 11% (ZAGORČEV, 1991). In a similar manner, the extension factor for the faults of the North-Anatolian fault zone might be as high as 3.5 (LE PICHON et al., 1984) but the overall extension both longitudinally and transversally did not exceed 3 – 5%.

Neotectonics and crustal thickness distribution

The Earth's crust on the Balkan peninsula is highly heterogeneous due to former plate motions and continental collisions. Subduction of the African under the Eurasian plate is now located (Fig. 11) in the Hellenic subduction zone, and neotectonic movements are dominated by extension (MERCIER, 1977, 1979). Except for crustal thinning due to uplift and erosion, and to local addition of sedimentary Neogene cover, crustal thickness distribution was fairly constant throughout the neotectonic stage (at least from the Sarmatian onwards), the crustal thickness in a given region changing in time with about 1 – 3 km.

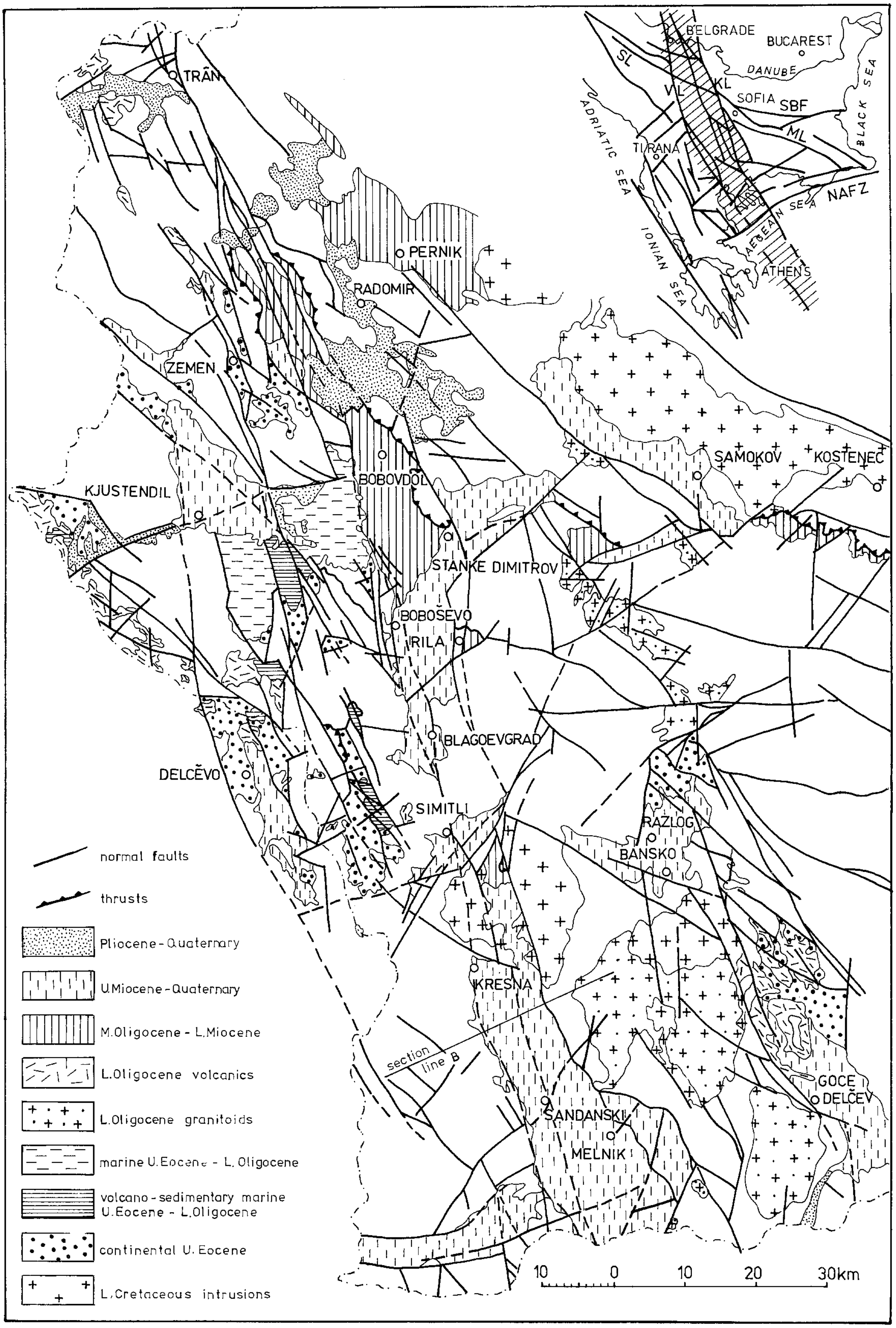

Fig. 9. Simplified Late Alpine and neotectonic map of the Struma lineament in Bulgaria. The pre-Paleogene basement is without ornament. On the inset: the principal lineaments on the Balkan Peninsula (SL – Sava lineament; VL – Vardar lineament; KL – Struma lineament; ML – Marica lineament; NAFZ – North-Anatolian fault zone).

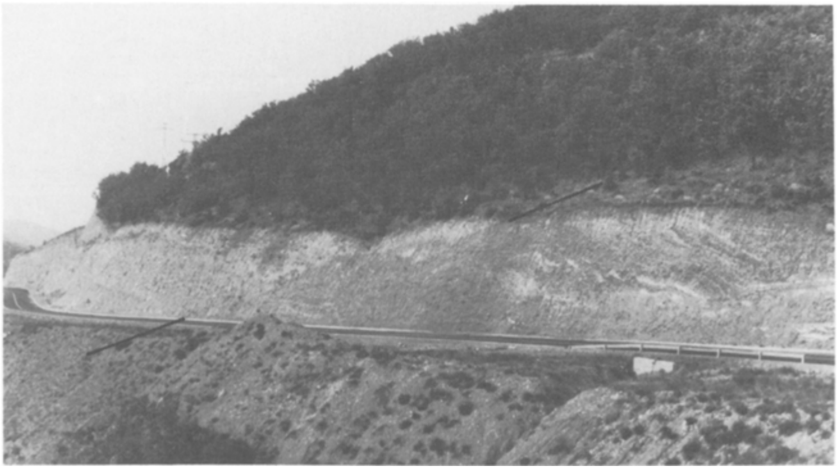

Fig. 10. The West-Pirin fault between the Neogene Kalimanci Formation (left) and the Precambrian Rhodopian Supergroup (right). On the road from the village of Gorno Spančevo to Goce Delčev. View towards north.

Linear neotectonic morphostructures are characterized by a thickened (from 35 to 50 km) crust. Therefore, isostatic uplift was most important in the neotectonic stage. At present, the westernmost part of the Stara-planina linear morphostructure is the most uplifted one, and almost reaches equilibrium: the positive recent vertical movements are almost negligible. In the central parts of the morphostructure, the most stable uplift tendency is recorded, and it is related to a stable isostatic anomaly. The lowest eastern part is involved into most rapid recent uplift (cf. data of DOBREV & STANOEV, 1985; MILEV & VRABLIANSKI, 1988). The other linear morphostructures have a similar behaviour.

The Inner Dinarides and Hellenides form also linearly oriented morphostructures but consist of a number of strongly elongated neotectonic swells east of the line Sarajevo – Athens (east of the Dinarian-Hellenic linear morphostructure). Except for a transversal thickened structure NW of Tirana – Skopje, this area is characterized by abnormally thinned crust (less than 30 – 35 km) and gradually passes into the thinned crust of the Pannonian and Aegean basins.

The thinnest crust occurs in the Ionian, Black sea, and Aegean basins, and in the central parts of the Moesian platform (Fig. 11). These areas are characterized by a stable neotectonic subsidence.

The crust of the Serbo-Macedonian swell and Struma and Vardar lineaments is of intermediate thickness: between 30 and 40 km. Subsided into an imperfect rift structure, this area is situated between areas of thickened crust corresponding to the Pelagonian and Rila-Rhodope swells.

The most thickened crustal lens of the Rila-Rhodope massif (40 – 55 km) coincides with a pronounced isostatic anomaly and the most intense (up to 5 mm/a) recent vertical uplift (cf. DOBREV & STANOEV, 1985). This area as well as the Sredna-gora swell, is the most saturated with granitoids and other light masses (JARANOFF, 1960), and the almost permanent vertical uplift tendency is characteristic for the whole neotectonic stage, recent movements included. The Rhodope massif (Rila-Rhodope swell) is characterized also by an important Paleogene (Late Eocene – Early Oligocene) tectonothermal event (rhyolitic volcanism, granitoid intrusions now exposed in neotectonically most uplifted blocks, and overall intense thermal flow). Its denudation after the Early Oligocene was probably of the order of 5 – 8 km.

Conclusions

Modern plate tectonic studies in the Eastern Mediterranean considered the interaction between the African and the Eurasian plates (MERCIER, 1979) as relative displacements between rigid microplates (e.g. MCKENZIE, 1970, 1972, 1978) or deformable microplates and plate boundaries (BRUNN, 1976; TAPONNIER, 1977). Overall extension in the Aegean region was also explained with a thermal dome, respectively hot plume or a mantle diapir (SCHUILING, 1972; MAKRIS, 1975). DEWEY (1988) argumented a post-collisional extensional collapse of orogens. This model could be also applied to the neotectonic development of the Balkan Peninsula.

Late Alpine collisions including Early Oligocene collapse ended with the Early Miocene (Savian) folding and thrusting. Crustal thickening was followed by isostatically compensated elevation and extensional

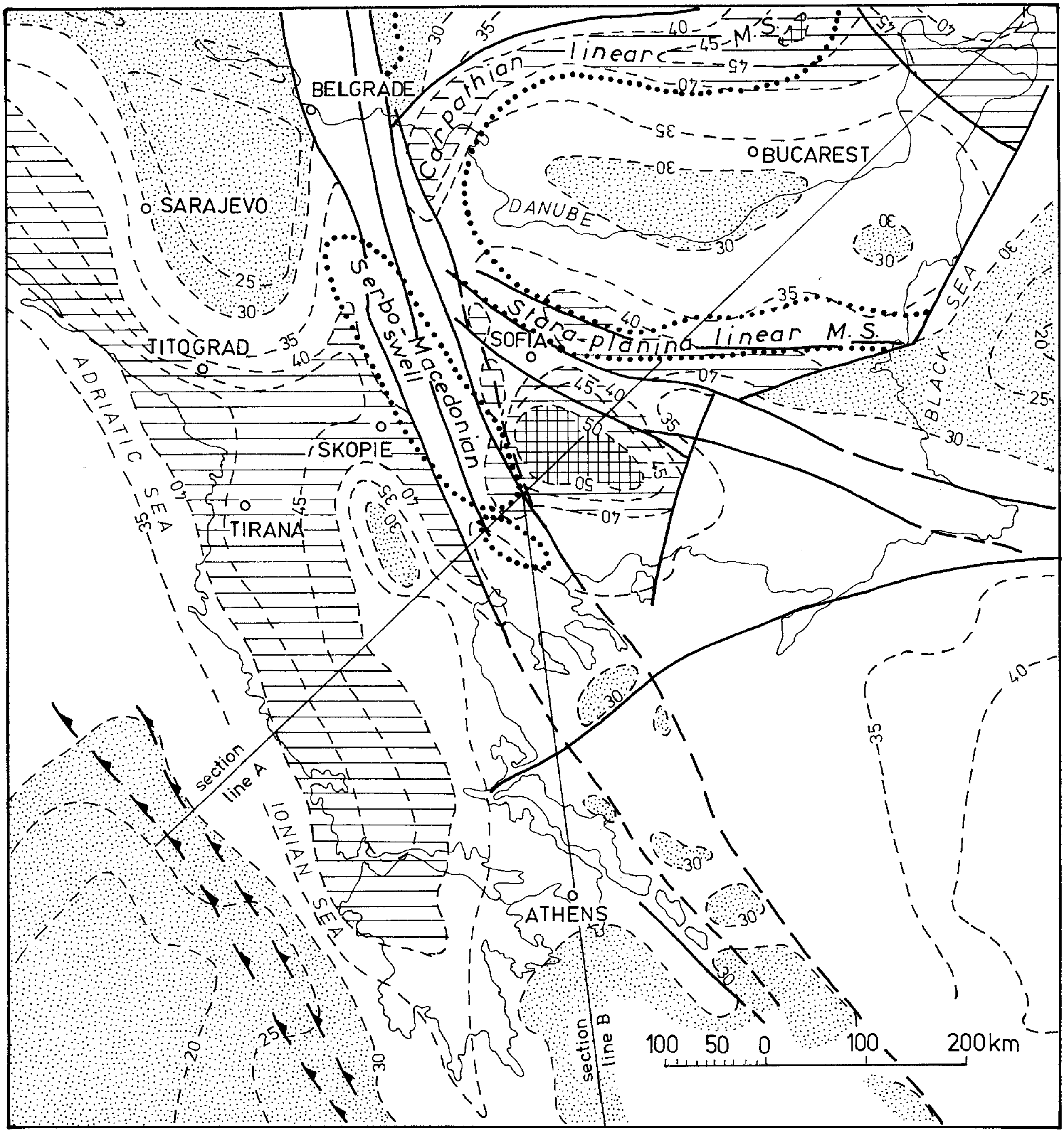

Fig. 11. Sketch for the relief of Mohorovicic discontinuity on the Balkan Peninsula. After data of many authors, summarized by ANĐELKOVIĆ (1982), DACHEV (1988) and MAKRIS (1976). The Hellenic arc and some of the principal neotectonic structures are also shown as well as the traces of the deep sections of Fig. 12. Areas with thin crust (less than 30 km) stippled; areas with thickened crust (more than 50 km) crosshatched.

orogenic collapse caused by body forces. The most thickened area (Central-Balkan neotectonic region) was bounded by the positive linear morphostructures which inherited the youngest fold and thrust belts (Dinarides-Hellenides and Balkanides), and the isostatic uplift maintained the general domal structure of the region. The latter was differentiated at several instances into partial swells which collapsed along first- and second-order lineaments (Fig. 12A), the Vardar, Struma, Marica, Sub-Balkan and Middle-Mesta lineaments included. The gradual lowering of the orthoplain towards the periphery of the region had its continuation into the surrounding Ionian, Pannonian, Precarpathian, Euxinian and Aegean basins.

The second factor in the neotectonic evolution was the development of the Hellenic subduction zone (MERCIER, 1977; MERCIER et al., 1979; ANGELIER, 1979) and the related Hellenic trough, South-Aegean arc, and Cycladic volcanic arc. The development of these structures south of the North-Aegean trough and the Middle-Mesta fault zone (as a prolongation of the North-Anatolian fault zone) resulted in a conspicuous Aegean microplate which was at least partially separated from the microplate of the Central-Balkan neotectonic region. It is characterized by a thinner crust (Fig. 12 B) which is strongly affected by the Hellenic subduction, and by marine sedimentation.

Thus crustal thickness distribution (Figs. 11, 12) may be regarded as one of the basic differences between the Central-Balkan neotectonic region and the Aegean neotectonic region (microplate), which controlled to a large extent their neotectonic behaviour. The isostatic uplift of the thickened crust within the Central-Balkan neotectonic region and its collapse isolated it from the surrounding regions through peripheral sea basins and inherited positive linear morphostructures. This isolation reflected also in the formation of a peculiar Central Balkan palaeobotanical sub-province (palaeophytohorion – PALAMAREV, 1990) as a most favourable area for development of endemic floras and preservation of relict species.

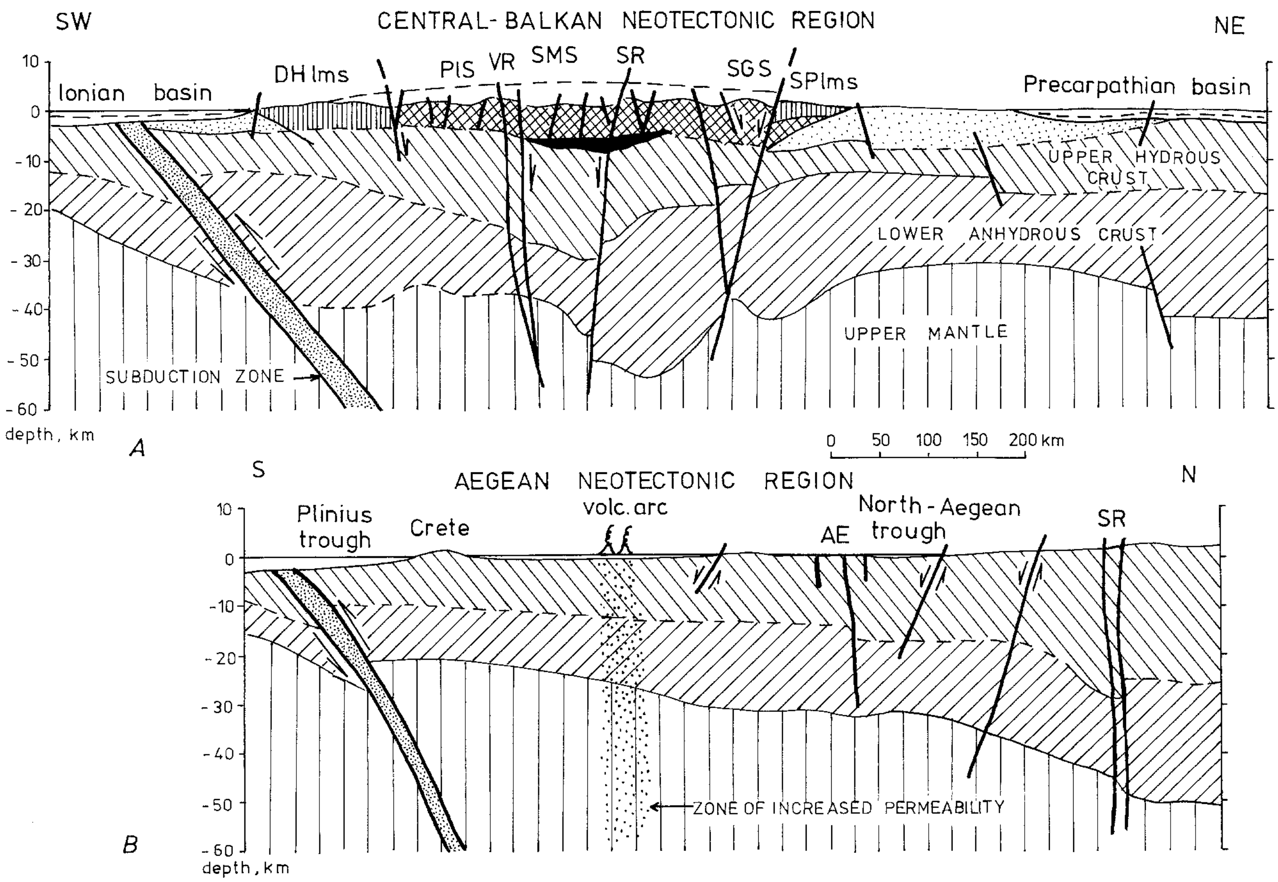

Fig. 12. Schematic sections through the Central Balkan neotectonic region (A) and the Aegean neotectonic region (B). Gently dipping sedimentary cover (stippled), folded Paleozoic and Mesozoic (dense vertical hatching), granitoid-metamorphic crust (crossed hatching), low-velocity layer (black). Principal neotectonic structures: DHlms – Dimarian-Hellenic linear morphostructure; PlS – Pelagonian swell; VR – Vardar rift; SMS – Serbo-Macedonian swell; SR – Struma rift; SGS – Sredna gora swell; SPlms – Stara-planina linear morphostructure; AE – Agios Efstratios fault.

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