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Geology of the Caucasus: A Review

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Abstract: The structure and geological history of the Caucasus are largely determined by its position between the still-converging Eurasian and Africa-Arabian lithospheric plates, within a wide zone of continental collision. During the Late Proterozoic–Early Cenozoic, the region belonged to the Tethys Ocean and its Eurasian and Africa-Arabian margins where there existed a system of island arcs, intra-arc rifts, back-arc basins characteristic of the pre-collisional stage of its evolution of the region. The region, along with other fragments that are now exposed in the Upper Precambrian–Cambrian crystalline basement of the Alpine orogenic belt, was separated from western Gondwana during the Early Palaeozoic as a result of back-arc rifting above a south-dipping subduction zone. Continued rifting and seafloor spreading produced the Palaeotethys Ocean in the wake of northward migrating peri-Gondwanan terranes. The displacement of the Caucasian and other peri-Gondwanan terranes to the southern margin of Eurasia was completed by ~350 Ma. Widespread emplacement of microcline granite plutons along the active continental margin of southern Eurasia during 330–280 Ma occurred above a north-dipping Palaeotethyan subduction zone. However, Variscan and Eo-Cimmerian–Early Alpine events did not lead to the complete closing of the Palaeozoic Ocean. The Mesozoic Tethys in the Caucasus was inherited from the Palaeotethys. In the Mesozoic and Early Cenozoic, the Great Caucasus and Transcaucasus represented the Northtethyan realm – the southern active margin of the Eurasiatic lithospheric plate.

The Oligocene–Neogene and Quaternary basins situated within the Transcaucasian intermontane depression mark the syn- and post-collisional evolution of the region; these basins represented a part of Paratethys and accumulated sediments of closed and semiclosed type. The final collision of the Africa-Arabian and Eurasian plates and formation of the present-day intracontinental mountainous edifice of the Caucasus occurred in the Neogene–Quaternary period. From the Late Miocene (c. 9–7 Ma) to the end of the Pleistocene, in the central part of the region, volcanic eruptions in subaerial conditions occurred simultaneously with the formation of molasse troughs.

The geometry of tectonic deformations in the Transcaucasus is largely determined by the wedge-shaped rigid Arabian block intensively indenting into the Asia Minor-Caucasian region. All structural-morphological lines have a clearly-expressed arcuate northward-convex configuration reflecting the contours of the Arabian block. However, farther north, the geometry of the fold-thrust belts is somewhat different – the Achara-Trialeti fold-thrust belt is, on the whole, W–E-trending; the Greater Caucasian fold-thrust belt extends in a WNW–ESE direction.

Key Words: Caucasus, convergence, collision, Eurasia, Gondwana, volcanism

Kafkasların Jeolojisi

Özet: Kafkasların yapısını ve jeolojik tarihini denetleyen ana unsur birbirine yaklaşan Avrasya ve Afrika-Arabistan levhaları arasındaki konumudur. Geç Proterozoyik ile Tersiyer arasında Kafkaslar, Tetis okyanusu ve bu okyanusun Avrasya ve Afrika-Arabistan kıta kenarları içermekteydi; bu sistem içerisinde yer alan ada yayları, yay-içi riftler, yay-ardı havzalar Kafkasların çarpışma öncesi jeoloji tarihinin bir parçasını teşkil eder. Erken Paleozoyik'te batı Gondwana'nın altına güneye doğru dalan bir dalma-batma zonu üzerinde gelişen yay-ardı riftleşme ile Kafkaslar, ve Alpin orojenik kuşak içinde yer alan diğer üst Prekambriyen–Kambriyen kristalen temel parçaları, Gondwana'dan ayrılmıştır. Kuzeye hareket eden bu Gondwana-çevresi (*peri-Gondwana*) mıntıkalarının güneyinde Paleotetis okyanusu açılmıştır. Kafkasya ve diğer Gondwana-çevresi mıntıkalarının Avrasya güney kenarını eklenmesi ~350 Ma'de tamamlanmıştır. Avrasya kıta kenarının altına kuzeye doğru dalan bir dalma batma zonu üzerinde yaygın mikroklinli granitoid plutonlarının yerleşimi 320–280 Ma aralığında gerçekleşmiştir. Tüm bu Variskan, Eo-Kimmeriyen ve erken Alpin olaylara rağmen Kafkasların güneyindeki Paleozoyik okyanusunun tamamen kapanmamış, ve Mesozoyik Tetis Paleotetis'ten miras kalmıştır. Mesozoyik ve erken Tersiyer'de, Büyük Kafkaslar ve Transkafkasya, Avrasya'nın levhasının güney aktif kıta kenarını, bir diğer ifade ile kuzey Tetis bölgesini temsil ediyordu.

Transkafkaya'nın dağ arası çöküntü bölgelerinde gelişen Oligosen–Neojen ve Kuvaterner havzalar bölgenin çarpışma ve çarpışma sonrası evrimini temsil eder. Bu havzalar Paratetisin bir kesimini temsil eder ve sedimanları kapalı veya yarı-kapalı havzalarda çökelmiştir. Afrika-Arabistan ve Avrasya levhalarının nihai çarpışması ve bugünkü kıtalararası Kafkaya dağ kuşağının oluşumu Neojen–Kuvaterner'de meydana gelmiştir. Geç Miyosen'den (9–7 Ma) Pleistosen'in sonuna kadar geçen zamanda Kafkafların merkezi kesimlerinde volkanik faaliyetler meydana gelmiş ve molas havzaları oluşmuştur.

Transkafkasya'daki tektonik deformasyonun geometrisini kontrol eden ana etken kama şeklinde sert Arabistan blokunun Anadolu-Kafkasya bölgesine saplanmasıdır. Buna bağlı olarak tüm yapısal-morfolojik çizgilerin, Arabistan levhasının kuzey sınırını yansıtan bir şekilde, kuzeye doğru içbükey bir geometri gösterir. Buna karşın daha kuzeyde kıvrım-bindirme kuşaklarının geometrisi farklıdır – Acara-Trialeti kuşağının yönü doğu-batı, Kafkaslar kıvrım-bindirme kuşağının uzanımı ise BKB-DGD'dur.

Anahtar Sözcükler: Kafkaslar, yakınlaşan, çarpışma, Avrasya, Gondwana, volkanizma

Introduction

The structure and geological evolution of the Caucasian segment of the Black Sea-Caspian Sea region (Figure 1) are largely determined by its position between the still converging Eurasian and Africa-Arabian lithosphere plates, within a wide zone of continent-continent collision. Problems of Late Proterozoic–Phanerozoic development of this area have been considered and discussed during the past decades in a great number of publications. According to some authors (Khain 1975; Adamia 1975; Adamia *et al.* 1977, 1981, 2008; Giorgobiani & Zakaraia 1989; Zakariadze *et al.* 2007), the region in the Late Proterozoic, Palaeozoic, Mesozoic, and Early Cenozoic belonged to the now-vanished Tethys Ocean (Prototethys, Palaeotethys, Tethys) and its Eurasian and Gondwanan/Africa-Arabian margins. Within this ocean-continent convergence zone, there existed a system of island arcs, intra-arc rifts, and back-arc basins etc. characteristic of the Late Proterozoic–Early Cenozoic pre-collisional stage of evolution of the region. During syn-collisional (Oligocene–Middle Miocene) and post-collisional (Late Miocene–Quaternary) stages of the Late Alpine tectonic cycle, as a result of Africa-Arabia and Eurasia collision back-arc basins were inverted to form fold-thrust belts in the Great and Lesser Caucasus and, in between, the Transcaucasian intermontane depression. Normal marine basins were replaced by semi-closed basins of euxinic type (Paratethys) and later on (Late Miocene) by continental basins with subaerial conditions of sedimentation (Milanovsky & Khain 1963; Gamkrelidze 1964; Andruschuk 1968; Azizbekov 1972; Geology of the USSR 1977; Jones & Simons 1977; Eastern Paratethys 1985; Vincent *et al.* 2007; Adamia *et al.* 2008; Okay *et al.* 2010).

Main Tectonic Units

The Caucasus is divided into several main tectonic units or terrains (Figure 2). There are platform (sub-platform, quasi-platform) and fold-thrust units, which from north to south are: the Scythian (pre-Caucasus) young platform, the fold-thrust mountain belt of the Great Caucasus including zones of the Fore Range, Main Range, and Southern Slope, the Transcaucasian intermontane depression superimposed mainly on the rigid platform zone (Georgian massif), the Achara-Trialeti and the Talysh fold-thrust mountain belts, the Artvin-Bolnisi rigid massif, the Loki (Bayburt)-Karabagh-Kaphan fold-thrust mountain belt, the Lesser Caucasus ophiolitic suture, the Lesser Caucasian part of the Taurus-Anatolia-Central Iranian platform, and the Aras intermontane depression at the extreme south of the Caucasus. The youngest structural unit is composed of Neogene–Quaternary continental volcanic formations of the Armenian and Javakheti plateaus (highlands) and extinct volcanoes of the Great Caucasus – Elbrus, Chegem, Keli, and Kazbegi.

Within the region, Upper Proterozoic–Phanerozoic sedimentary, magmatic, and metamorphic complexes are developed. Their formation occurred under various palaeogeographic and geodynamic environments: oceanic and small oceanic basins, intercontinental areas, active and

Figure 1. Physical map of the Caucasus and adjacent areas of the Black Sea-Caspian Sea region (Adamia *et al.* 2010).

passive continental margins – transitional zones from ocean to continents. The Late Proterozoic–Phanerozoic interval is divided into two stages: pre-collisional (Late Proterozoic–Early Cenozoic) and syn-post-collisional (Late Cenozoic). During the pre-collisional stage, there existed environments characteristic of modern oceanic basins and zones transitional from ocean to continent.

Geological Provinces

Existing data allow the division of the Caucasian region into two large-scale geological provinces: southern Tethyan and northern Tethyan located to the south of and to the north of the Lesser Caucasian ophiolite suture, respectively. During the Late Proterozoic, the Southern Province distinctly demonstrated Pan-African (Cadomian) tectonic events, and throughout the Palaeozoic, it was a part of Gondwana that accumulated mainly shallow-marine platformal sediments. In the Palaeozoic, the Northern Province is characterized by strong manifestation of tectonic events: supra-subduction volcanism, granite formation, deep regional metamorphism, deformation and orogenesis. The Southern and Northern provinces differ each from the other throughout the Mesozoic and Early Cenozoic as well. The boundary between them runs along the North Anatolian (İzmir-Ankara-Erzincan) – Lesser Caucasian (Sevan-Akera)-Iranian Karadagh ophiolitic suture belt (see Figure 2).

Pre-collisional Stage: Late Proterozoic-Palaeozoic Basement Rocks

Basement rocks are represented by regionally metamorphosed (eclogite, amphibolite, epidote-amphibolite and greenschist facies of high, moderate,

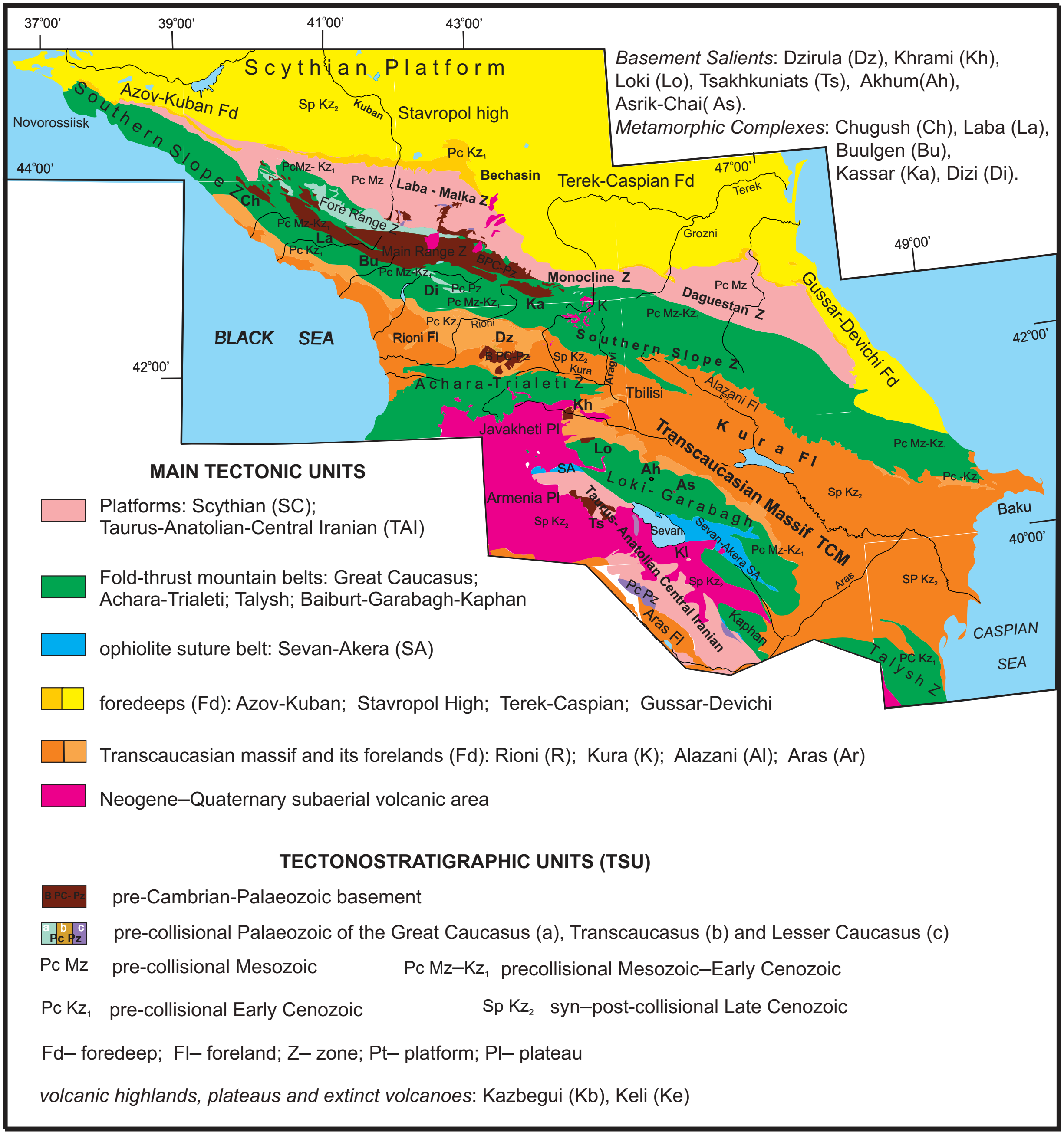


Figure 2. Tectonic map of the Caucasus (Adamia *et al.* 2010).

and low pressure) sedimentary, volcanic and plutonic rocks dated according to chronological and palaeontological data. Magmatitic rocks are represented by two main rock complexes of (1) ultrabasic-basic-intermediate and (2) acidic composition. The former is a carrier of information on the oceanic basins of Prototethys-Palaeotethys, it outcrops within almost the all main tectonic zones of the region (Abesadze *et al.* 1982), and is represented by relatively small, dismembered, strongly deformed and differently metamorphosed fragments of basic and ultrabasic rocks.

Rocks of pre-Mesozoic oceanic basins of the region, generally, are closely associated with granite-gneiss-magmatitic rock complexes of pre-Cambrian

(Southern Province) and Palaeozoic (Northern Province) basement, and represent rock associations belonging to continental crust as well as to transitional oceanic-continental crust. The rocks are strongly deformed: tectonic nappes, slices, tectonic and sedimentary mélanges (Lesser Caucasus, Dzirula massif), accretionary prisms etc (Great Caucasus) are frequent.

Southern Province

In the Caucasian region, the oldest, Pan-African (Cadomian–Neo-Proterozoic) basement of the Central Iranian platform crops out north of Erevan (Armenia, Tsakhkuniats massif, see Figure2). It includes two Pre-Cambrian complexes: (1) Arzacan (ensialic) and (2) Hancavan (ensimatic).

The Arzacan ensialic complex consists of paraschists (1500 m thick), which have undergone metamorphism in almandine-amphibolite facies, and of metavolcanics, phyllites, marbles, and schists (2000 m thick) metamorphosed in greenschist facies. The complex is intruded by granites whose Rb/Sr isochron age is 620 Ma and crust melt isotope initial ratio of $^{87}Sr/^{86}Sr = 0.7102 \pm 0.0006$ (Agamalian 2004).

The lower unit of the Hancavan complex (1900 m thick), which during the Pan-African events was obducted over the Arzacan complex, represents an oceanic-crust-type assemblage and is dominated by komatiite-basalt amphibolites with thin sedimentary intercalations, while the upper unit (1000 m thick) consists of metabasalt and metaandesite with beds of marble and quartz-mica schists. Both parts of the Hancavan complex contain tectonic lenses of serpentinite. The complex is cut by trondhjemite intrusions whose Rb/Sr isochron age is 685±77 Ma, with a mantle origin ratio of $^{87}Sr/^{86}Sr = 0.703361$ (Agamalian 2004).

Boundary Between Northern and Southern Provinces

The Sevan ophiolite mélange contains different exotic blocks represented by garnet-amphibolites (Amasia – the westernmost part of the Sevan ophiolite suture belt), amphibolites, micaschists (Zod, Adjaris and Eranos – the eastern part of the Sevan ophiolite suture belt) (Agamalian 2004), and redeposited metamorphic rocks of continental affinity: greenschists, marbles, limestones and skarns (Geydara and Tekiakay in Karabagh Figure 3; Knipper 1991). Layers of pre-Carnian breccia-conglomerates consisting of redeposited clasts represented by tectonized harzburgites, layered-, flaser- and isotropic gabbros, diabases, gabbro-diabases, altered basalts, tectonized other basic rocks, Upper Palaeozoic marbles, basaltic andesites, phyllites, carbonatic scarn have been found in the ophiolite mélange of the Lesser Caucasian suture directly to the east of the Lake Sevan (Zod and Ipiak nappes, r. Lev-chai; Knipper 1991). Available data show that throughout the Mesozoic at the northern edge of Palaeotethys occured destruction and erosion of obducted ophiolites, accumulation of redeposited ophiolitic clastics with admixture of continental ones.

According to Agamalian (2004), the Rb/Sr age of the Amasia amphibolites is 330±42 Ma, $^{87}Sr/^{86}Sr = 0.7051 \pm 0.003x \pm 0.000292$. The Rb/Sr age of a block of garnet gneiss (Zod) is equal to 296±9 Ma ($^{87}Sr/^{86}Sr = 0.705357 \pm 0.000292$), and for metamorphic schists from the same localities 243±13 ($^{87}Sr/^{86}Sr$ 0.706107±0.000156), 241±12 ($^{87}Sr/^{86}Sr$ 0.707902±0.000417) and 277±44 Ma ($^{87}Sr/^{86}Sr$ 0.704389±0.0003510).

The Sevan ophiolite belt represents an easternmost part of the İzmir-Ankara-Erzincan (or North Anatolian) ophiolite suture belt, interpreted by many authors as the main suture of the Paleotethys-Tethys (e.g., Adamia *et al.* 1977, 1981, 1987).

Northern Province: The Transcaucasian Massifs (TCM)

Loki, Khrami, Dzirula, Akhum-Asrikchai Salients of the Basement (see Figure 2)- All the above-mentioned salients of the basement, except the Akhum and Asrikchai ones, are mainly composed of Variscian granitoids. Relatively small outcrops of differently tectonized and metamorphosed basic and ultrabasic rocks are generally associated with pre-Cambrian–Early Palaeozoic gneissose diorites and plagiogranites.

The Loki salient outcrops in the territory of Georgia along the boundary with Armenia, within the Artvin-Bolnisi massif of the Transcaucasian intermontane

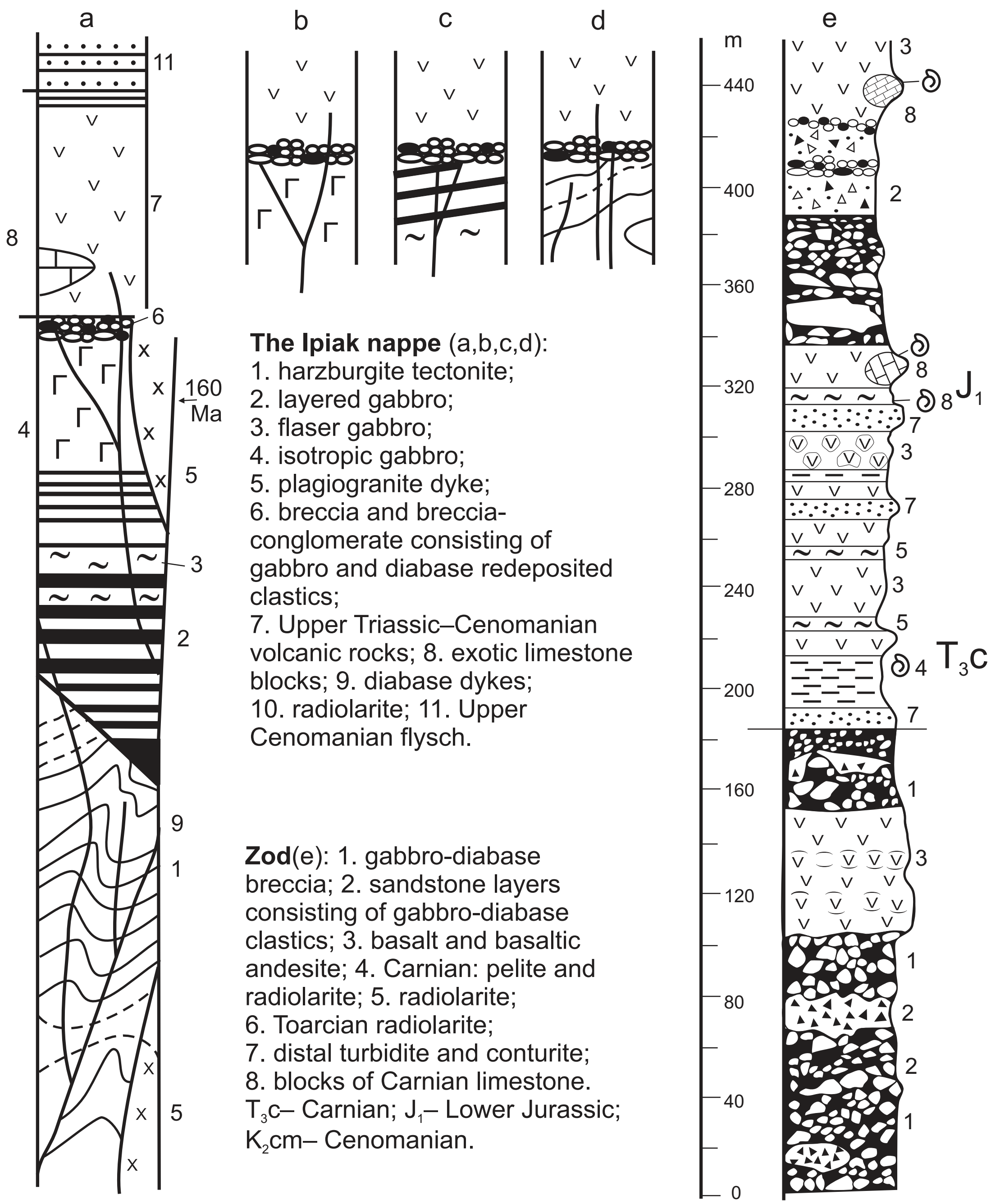


Figure 3. Pre-Upper Triassic and Upper Triassic-Jurassic sedimentary breccia in the ophiolite mélange of the Sevan-Akera suture belt (Knipper 1991).

depression. The structure of the Loki salient as well as of the other salients of the Transcaucasian basement seems to be more complicated than considered earlier (Abesadze *et al.* 2002). Most of the salient is composed of Upper Palaeozoic granitoids. Repeated tectonic displacement has caused interfingering of metabasites and metapelites and formation of tectonic sheets and slices (Figures 4 & 5).

The Loki basement, evidently, was formed during the Late Proterozoic–Early Palaeozoic. The metabasites, apparently, represent Prototethyan fragments. The metamorphic rocks along with

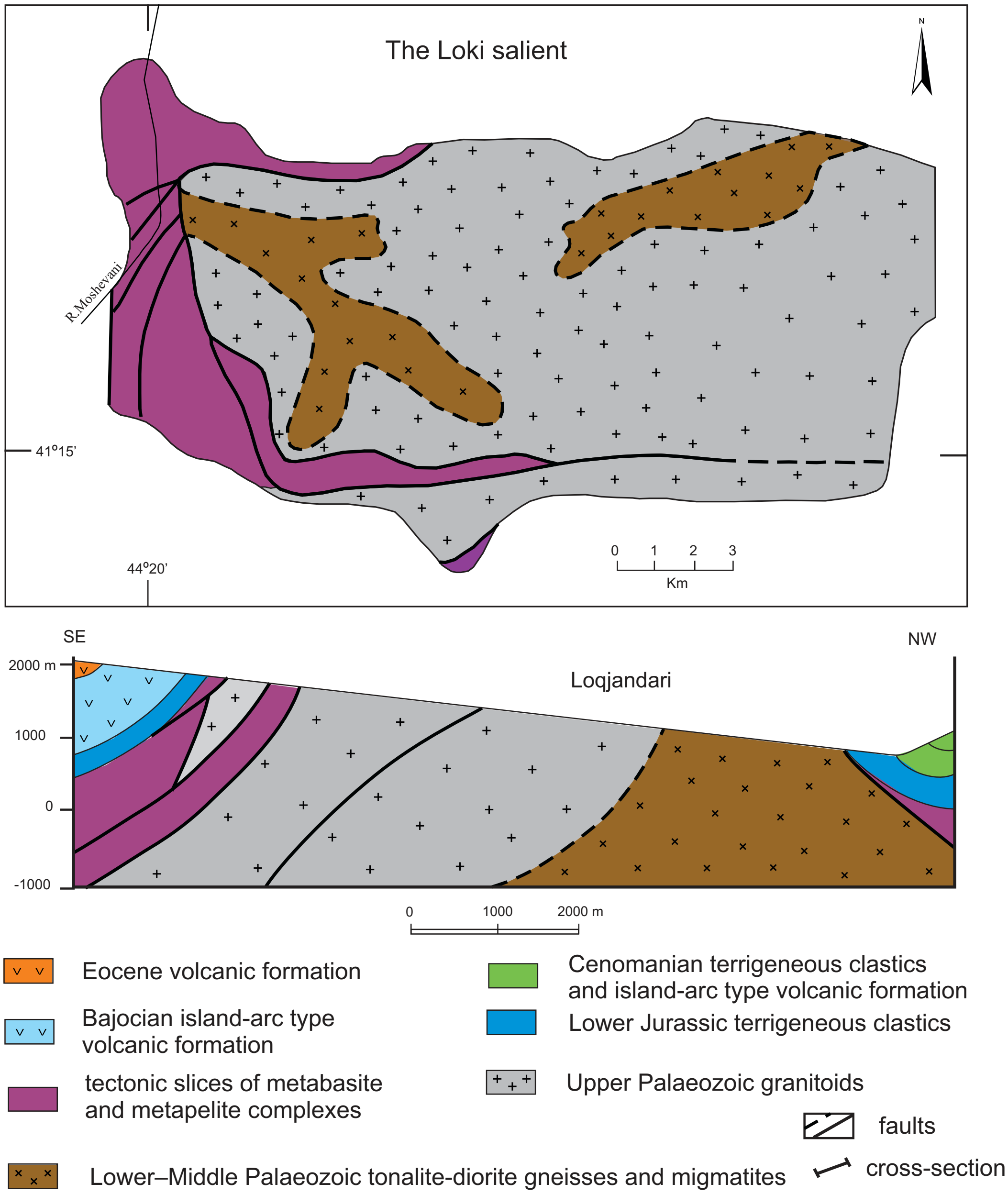


Figure 4. Simplified geological map and cross-section of the Loki salient (Abesadze *et al.* 2002; Zakariadze *et al.* 2007).

gneiss-migmatite complexes of the Transcaucasian massif bear resemblance to the immature continental crust of the Nubian-Arabian shield, and in the Early Palaeozoic, they were displaced to the southern edge of the East-European continent (Baltica). All the present-day geological structures were formed in the Late Palaeozoic–Cenozoic (Figure 4; Abesadze *et al.* 2002).

The Khrami salient situated to the north of the Loki (Figure 5), is made up mainly of Upper Palaeozoic (Variscan) granites. Plagiogneisses and migmatites occupy a limited part and bear small bodies of metabasites and serpentinites.

The Dzirula salient is dominated by pre-Variscan diorite-plagiogneiss-migmatite complex ('grey granites') and Variscan granitoids ('red granites').

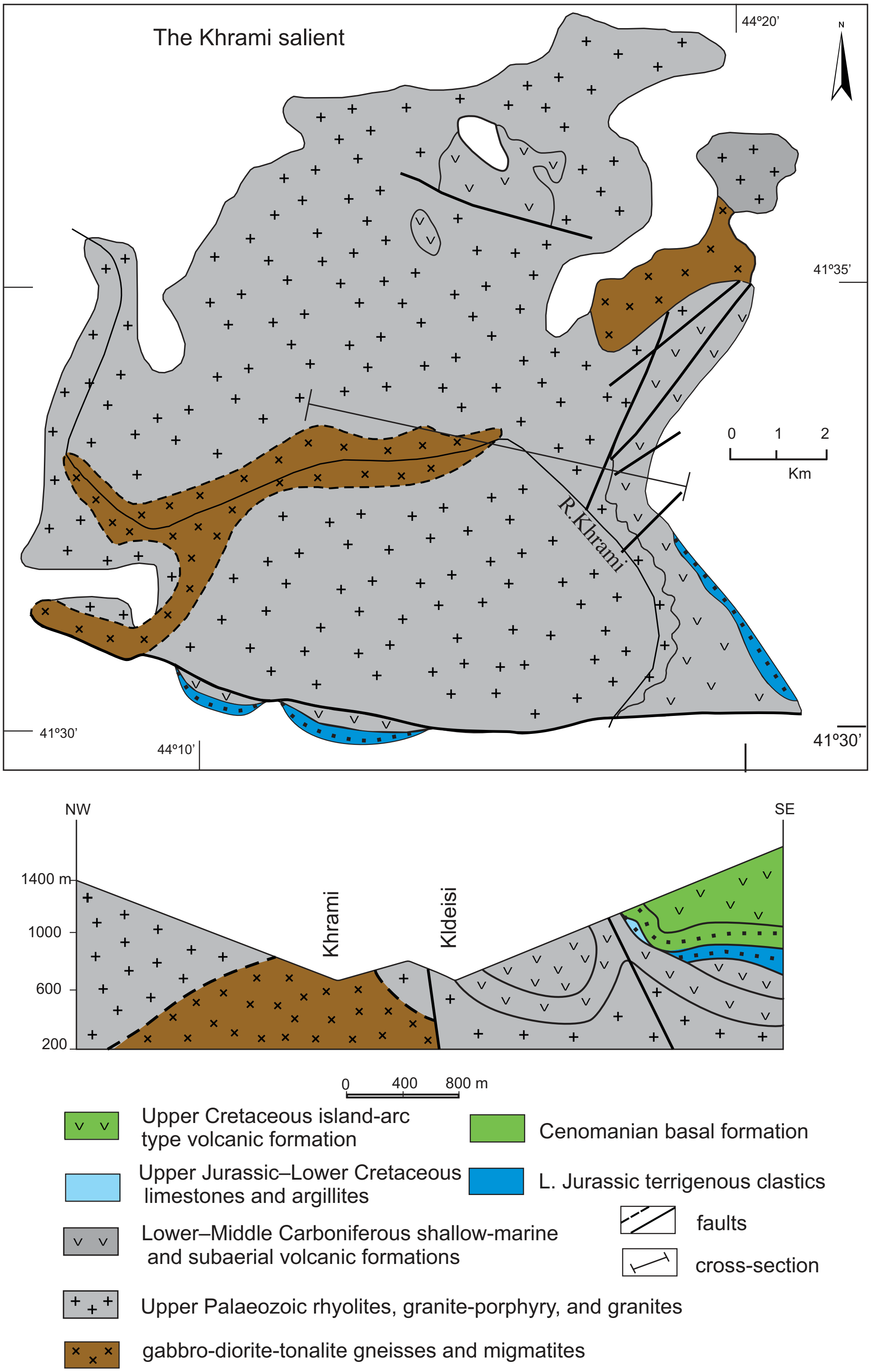


Figure 5. Simplified geological map and cross-section of the Khrami salient (Abesadze *et al.* 2002; Zakariadze *et al.* 2007).

Basic-ultrabasic rocks are found mainly within the field of diorite-plagiogneiss-migmatite complex, and also within the tectonic mélange of the Chorchana-Utslevi stripe cropping out along the eastern edge of the Dzirula salient (Figure 6).

The oldest basement unit consists of biotite gneisses, plagiogneisses, amphibolites, and crystalline schists subdivided into metabasic and metasedimentary successions presented either as irregularly piled-up tectonic slices or numerous inclusions in later gabbroic, dioritic and quartz dioritic intrusions. The metabasic succession includes various massive and banded amphibolites, garnet amphibolites, metadiabases, and subordinate subvolcanic bodies of metagabbro-metadiabase. Polymetamorphism of the metabasic succession did not exceed conditions of amphibolite and epidote-amphibolite facies of moderate pressure.

The Grey Granitoid complex incorporates associated basic, dioritic and quartz dioritic (tonalite) intrusions emplaced at various levels in the oldest basement unit. Basic intrusions include dykes and small stocks of diabases, gabbro-diabases, and gabbro.

The tectonic mélange zone consists of a number of allochthonous tectonic slices of different age and facies: phyllite schists, sheared Palaeozoic high-silicic volcanics, sheared Upper Carboniferous microcline granite, metabasite and serpentinite (Abesadze *et al.* 1980). Metabasic slices consist of massive and banded amphibolites, mylonitized amphibolites, metagabbro-diabases, metadiabases and metabasic tuffs. The ultramafic-mafic association of the mélange zone is interpreted as dismembered Upper Proterozoic–Lower Palaeozoic metaophiolite, and phyllite slices as fragments of hemipelagic cover of the paleooceanic basement.

Metabasic slices are associated with serpentine bodies in the tectonic mélange zone. All these ultrabasic-basic associations are considered as a unique ancient mafic basement of the massif. The ultrabasic rocks are composed mainly of harzburgite and dunite ubiquitously affected by polymetamorphism and extensive serpentinization.

Available data on geochronology and biostratigraphy of the *Transcaucasian massif (TCM)* basement unit, including three new Sm-Nd isochrons were obtained in the Vernadsky Institute of Geochemistry and Analytical Chemistry, Academy of Sciences of Russia, Moscow (Zakariadze *et al.* 1998, 2007). An attempt to date the mafic foundation of the TCM was based on the study of Nd-isotope variation for the metabasic series of the tectonic mélange zone, which despite narrow variation ranges for major elements (Mg#= 0.58 ± 0.01) show strong fractionation of incompatible trace elements and REE in particular (Sm/Nd= 0.41–0.27). The initial data on Neoproterozoic–Cambrian (c. 750–500 Ma) age limits of the dioritic intrusions of the Gray Granite Basement Complex were derived from U-Pb studies of zircon from gneissose granodiorite, quartz diorite-plagiogranite and migmatite of the Dzirula salient (Bartnitsky *et al.* 1990). New geochemical data, together with new La-ICP-Ms zircon and electron microprobe monazite age, are described from the Dzirula massif. U-Pb zircon ages of ca. 540 Ma, often with 330 Ma rims are obtained from deformed granodiorite gneisses; zircon and monazite data date the metamorphic age of HT/LP migmatites and paragneisses at ca. 330 Ma; some paragneisses contain relict 480 Ma monazites implying a previous thermal event (Treloar *et al.* 2009). Zircon and monazite dates of ca. 330 Ma from unfoliated calc-alkaline to high-K, I-type granodiorites, diorites and gabbros intrusive into the gneisses and migmatites according to Treloar *et al.* (2009) suggested that they represent the heat source that drove the metamorphism.

The whole stack of nearly vertical tectonic slices of the Dzirula metamorphic complexes is hosted by Late Palaeozoic granites. Middle–Late Carboniferous–Permian an emplacement ages of the potassium (red) granite is reliably constrained by stratigraphic and various K-Ar and U-Pb data from micas and zircons (Rubinshtein 1970; Zakariadze *et al.* 2007; Treloar *et al.* 2009).

Small salients of the basement complexes of the TCM are known to the north of the Sevan Lake (Akhum, Asrikchay). They are represented mainly by micaschists (Aslanian 1970; Azizbekov 1972). The Rb/Sr age of the quartz-micaschists according to Agamalian (2004) is 293±7 Ma (87Sr/86Sr= 0.7057 ± 0.0016).

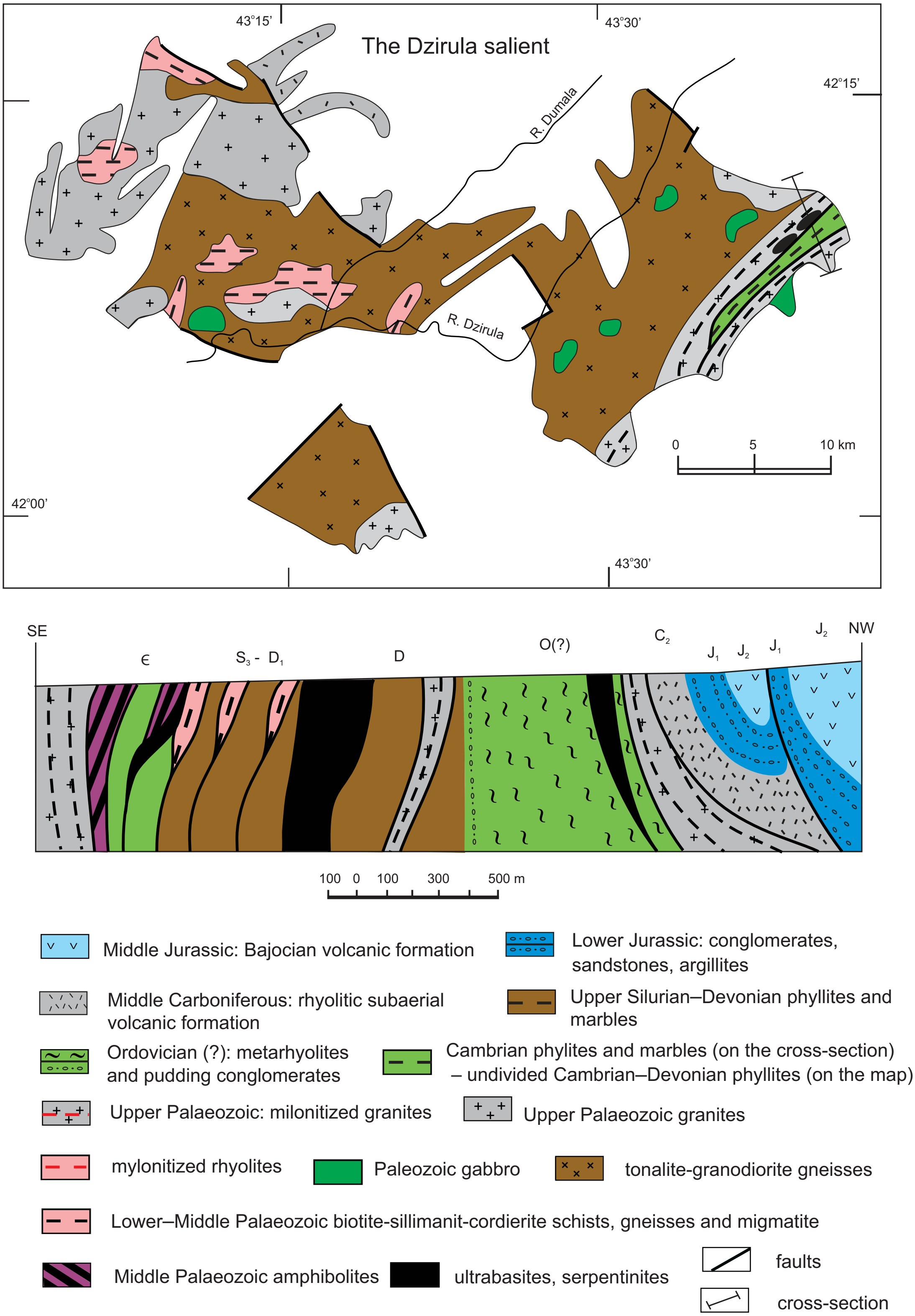


Figure 6. Simplified geological map and cross-section of the Dzirula salient (Abesadze *et al.* 2002; Zakariadze *et al.* 2007).

Great Caucasus, Main Range Zone, Ultrabasic-basic Complexes

Within the Great Caucasus, pre-Mesozoic basic and ultrabasic rocks crop out: (1) along the southern margin of its crystalline core (Chugush, Laba, Buulgen, and Kassar complexes), (2) at the western submergence of the crystalline core (Belaya and Kisha complexes), and (3) in the Fore Range Zone (Blib, Atsgara, Arkhiz, Marukh complexes) (Figure 7).

The southernmost strip of the crystalline core of the Great Caucasus is represented by thrust slices of metaophiolites (Adamia *et al.* 1978, 2004). They consist of ultrabasic rocks, gabbro-amphibolites, amphibolites and mica-schists, plagiogneisses, marbles (Laba, Buulgen, Chugush, and Kassar Groups, Figure 7). Mineral rock associations indicate amphibolite facies of high and moderate pressure metamorphism. The presence of crinoidea in marbles of the Laba Group indicates a post-Ordovician (Early–Middle Palaeozoic) age of the rocks. The composition of the Palaeozoic metaophiolitic complexes corresponds to that of oceanic spreading centers (T-type MORB) and suprasubduction zones (immature arc – back-arc association). Accretion of the fragments of the oceanic and transitional ocean/continent crust towards the eastern part of

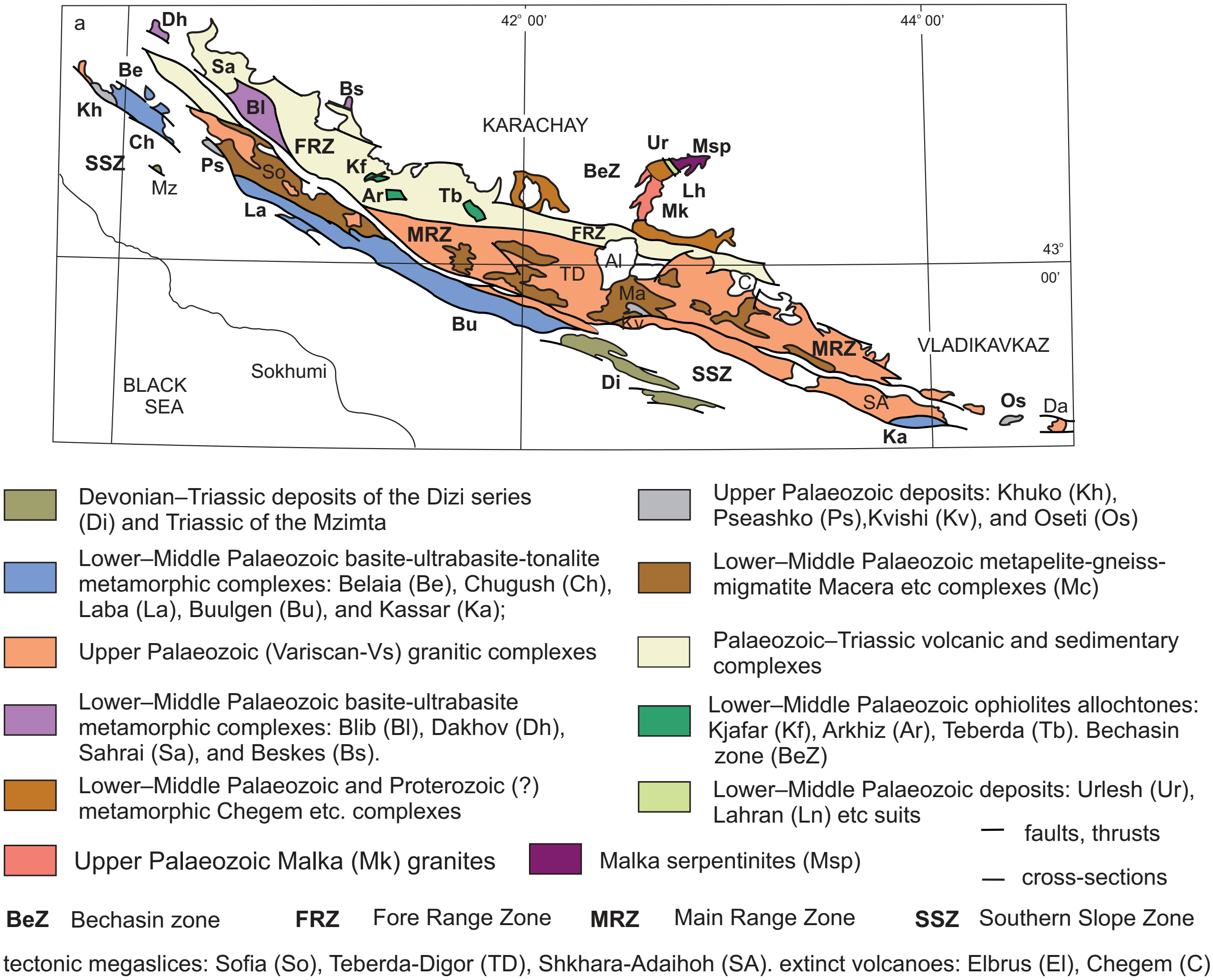


Figure 7. Schematic geological map of the pre-Jurassic complexes of the Great Caucasus. Abbreviations La, Bu and Kf indicate locations of the Figure 8a, b & c, accordingly.

the East European continent occurred during the Middle Palaeozoic, apparently, at the Early–Middle Carboniferous boundary (Adamia & Shavishvili 1982; Belov & Omelchenko 1986; Somin 2007b).

Thirty seven samples of the Kassar Group were studied in the palaeomagnetic laboratory of the Caucasian Institute of Mineral Resources. In amphibolites of this group, maghemite, magnetite and titanomagnetite proved to be NRM-positive. Calculated palaeo-latitude corresponds to 14° N, whereas an age of magnetization, apparently, can be attributed to Late Devonian–Early Carboniferous. That is in accordance with geological data on the position and age of the initial rocks of metaophiolites of the Laba and Buulgen Groups (Adamia *et al.* 2004a).

In the westernmost part of the Main Range Zone of the Great Caucasus under Upper Palaeozoic and Lower Jurassic deposits, there appear basic and ultrabasic rocks represented by tectonic slices of serpentinites, milonitized trondhjemites and gabbro-diorites, strongly deformed banded gabbro (with ultrabasic cumulatives), gabbro-diabases, and phyllites (Adamia & Shavishvili 1982). A pre-Mesozoic age for the basic-ultrabasic complexes of the Main Range Zone is confirmed by stratigraphical, palaeontological and isotopic geochronological data. Based on the stratigraphical position they are dated as pre-Middle Carboniferous (Andruschuk 1968; Ajgirei 1976; Belov 1981; Somin 2007a). According to palaeontological finds (bluish-green algae and crinoidea), the upper part of the Laba series is dated as post-Middle Ordovician (Potapenko & Stukalina 1971; Adamia *et al.* 1973; Somin & Vidiapin, 1989).

The absolute age results for the Laba- and Buulgen complexes agree well with biostratigraphical data. U-Pb data for zircon from microgneisses and amphibolite microgneisses of the Laba series yield 534±9 Ma and 520 Ma ages (apparently, an age of magmatic protolithes, Somin *et al.* 2004) and 345 Ma (an age of metamorphism, Somin *et al.* 2004). U-Pb (SHRIMP) dating for magmatic zircons from orthogneisses of the Buulgen complex yielded 381±3 Ma (Somin 2007a). The same study gave an age of metamorphism of 355–325 Ma. These data are in good agreement with a Sm-Nd mineral isochron age for garnet-biotite amphibolites of the Buulgen complex (287±33 Ma, Somin 1991). According to data by Hanel *et al.* (1993a, b), Gurbanov *et al.* (1995), an age of amphibolites of the Buulgen complex is 600±15 Ma (melanosome) and 500±20 Ma (leucosome).

Great Caucasus, Main Range Zone, Gneiss-migmatite-micaschist and Granitic Complexes

Rocks of the Granitic Complex, the granites themselves and various granitoids, granito-gneisses, migmatites and quartz-mica crystalline schists (amphibolitic, epidote-amphibolitic, and greenschist facies), and also granite blastomilonites and phyllonites are concentrated north of metabasites-metaophiolites of the Laba-Buulgen-Kassar accretionary complexes. Amphibolites are subordinate and marbles are very scarce. These complexes are made up of large tectonic slices in the Main Range Zone: Sophia, Teberda-Digori, Shkhara-Adaikhokhi (see Figure 7). Crystalline schists, granite-gneisses and migmatites, known in publications as the Makera, Gondaray and some other groups, represent the environment accommodating potassium-spar granites and associated granitoids of the crystalline core – pre-Alpine basement of the Great Caucasus (Gamkrelidze 1964; Andruschuk 1968; Somin 1971; Ajgirei 1976; Adamia *et al.* 1987; Somin *et al.* 2007a). According to the mineral and chemical composition, they are, generally, attributed to calc-alkaline series, S- and I-granites, and also to transitional from I-S-types of granites (Adamia *et al.* 1983; Potapenko *et al.* 1999; Somin 2007a).

The age of the Metamorphic-Granitic Complexes was determined geochronologically and stratigraphically (Gukasian & Somin 1995). Redeposited material of the metamorphic rocks as well as granites is abundantly present in Upper Palaeozoic molasse of the Great Caucasus. Geochronological dating of granite-gneisses-migmatites (Gondaray Complex) was obtained from detrital zircons (Somin *et al.* 2007a). Most often indicated values of 500±40 and 2000 Ma were obtained by the Pb/Pb evaporation method for zircons from orthogneisses (Hanel *et al.* 1993a, b). Dating of zircons from rocks of the same locality using the traditional U-Pb method provided an age of 400±10 Ma (Bibikova *et al.* 1991). An U-Pb age of 386±5 Ma was obtained for similar orthogneisses, and the U-PB dating of zircons from migmatite leucosomes yielded 305 Ma (Somin *et al.* 2007a). Data obtained for detrital zircons from paragneisses fall in the Proterozoic–Lower Palaeozoic interval (Somin *et al.* 2007b). Occurrence of detrital zircons dated back to 470–480 Ma suggests that the paragneisses are younger than Early Ordovician. 425 Ma SHRIMP dating of zircons from amphibolites, intruded by orthogneisses, shows that the rocks are not younger than Middle Silurian (Somin *et al.* 2007a). Rb/Sr isochron dating of potassium-spar granites (Ullukam complex – the central part of the Main Range Zone) gave values from 280 up to 300 Ma; K/Ar dating demonstrated an interval of 290–320 Ma (Gamkrelidze 1964; Andruschuk 1968; Potapenko *et al.* 1999).

Great Caucasus Fore Range Zone: Basic and Ultrabasic Rocks

Within the central (Tebaerd) and western parts (Great Laba) of the Fore Range Zone, outcrops of ophiolites are traced as strongly deformed allochthonous tectonic sheets and slices. They are underlain by Silurian–Devonian–Lower Carboniferous volcanogenic-sedimentary island-arc type deposits and overlain by Middle–Upper Carboniferous molasse (neoautochthone).

There are several allochthonous sheets consisting of ultrabasic rocks, gabbro-diabases, effusives (basalts and andesitobasalts), terrigenous, carbonate and volcanoclastic sediments (Grekov *et al.* 1974; Kropachev & Grekov 1974; Ajgirei 1976; Khain 1984; Zakariadze *et al.* 2009; see Figures 7 & 8). The ophiolitic association is subdivided into the following: mantle restites, cumulative complex, dyke complex, volcanic series, and volcanogenic-sedimentary series. According to rare finds of fossil fauna (corals), the age of the volcanogenic-sedimentary series is identified as Middle Devonian (Grekov *et al.* 1974). The K/Ar age of amphibole from gabbro-pegmatite is 457±13 Ma and 493±15 Ma (Khain 1984), and SHRIMP zircon age of gabbroid rocks of the Maruh ophiolite nappe is 416±8 Ma (Somin 2007a) that agree well with data for a Lower–Middle Palaeozoic age of the ophiolites of the Fore Range.

The Blib complex represents an isolated outcrop of metabasites and ultrabasic rocks tectonically underlying Middle Palaeozoic sediments of the westernmost part of the Fore Range Zone (Great and Small Laba). The complex represents an alternation of amphibolites, amphibolite-gneisses, plagiogneisses, and schists; garnet-epidote and albite amphibolites predominate. Massive and banded eclogites, and serpentinized ultrabasites alternating with the garnet mica schists crop out (Andruschuk 1968). Sm/Nd dating of eclogite, with exception of one pair of garnet rock, gave an age of 311±22 Ma, and Lu-Hf garnet chronometer showed an age of 322±14, 316±5 and 296±11 Ma, which is constrained to error limit of Ar-Ar age for phengite equal to 303±5 Ma (Perchuk & Phillipot 1997; Phillipot *et al.* 2001). Somewhat older mineral (amphibole, biotite, muscovite) ages (374±30 Ma) were obtained by the K-Ar method for amphibolites of the Blib complex (Somin 2007a).

Within *the Fore Range Zone* of the Caucasus, the Granitic Complex includes the upper part of the Blib Complex consisting of garnet quartz-muscovite schists, and granite-gneisses of the Armov ‘suite’ (Andruschuk 1968) that represents a tectonic sheet located between the lower, ultrabasic-basic (Balkan) part of the Blib Complex and the Middle Palaeozoic metavolcanics of the Fore Range Zone. Granitoids occupy a sizeable part of the complex and are termed the Urushten Complex (Andruschuk 1968). The complex is dominated by granodiorites, plagiogranites, plagioalaskites belonging to I-type of calc-alkaline series. Maximum values of K/Ar age reach 370 Ma (Potapenko *et al.* 1999).

Several small outcrops of the basement complexes termed as the Dakhov, Sakhrai and Beskes salients (see Figure 7) form the partly exposed northern margin of the Fore Range Zone of the Great Caucasus (Ajgirei 1976; Somin *et al.* 2007b). The Dokhov salient is mostly composed of granitoids intruded in metamorphic rocks. The U-Pb age of zircon grains from metaaplites has concordant ages of 354±3 Ma and 353±3 Ma and these values are considered to be the crystallization age of the aplites. Also according to K-Ar dating, hornblende from the unmetamorphosed granodiorite show an age of 301±10 Ma that constrains the upper age limit for the Dakhov salient (Somin *et al.* 2007b).

Scythian Platform, Bechasin Zone: Metamorphic and Granitic Complexes

Metamorphic and Granitic complexes outcrop in the central part of the Bechasin Zone (see Figures 2 & 7). The metamorphic rocks of greenschist facies are

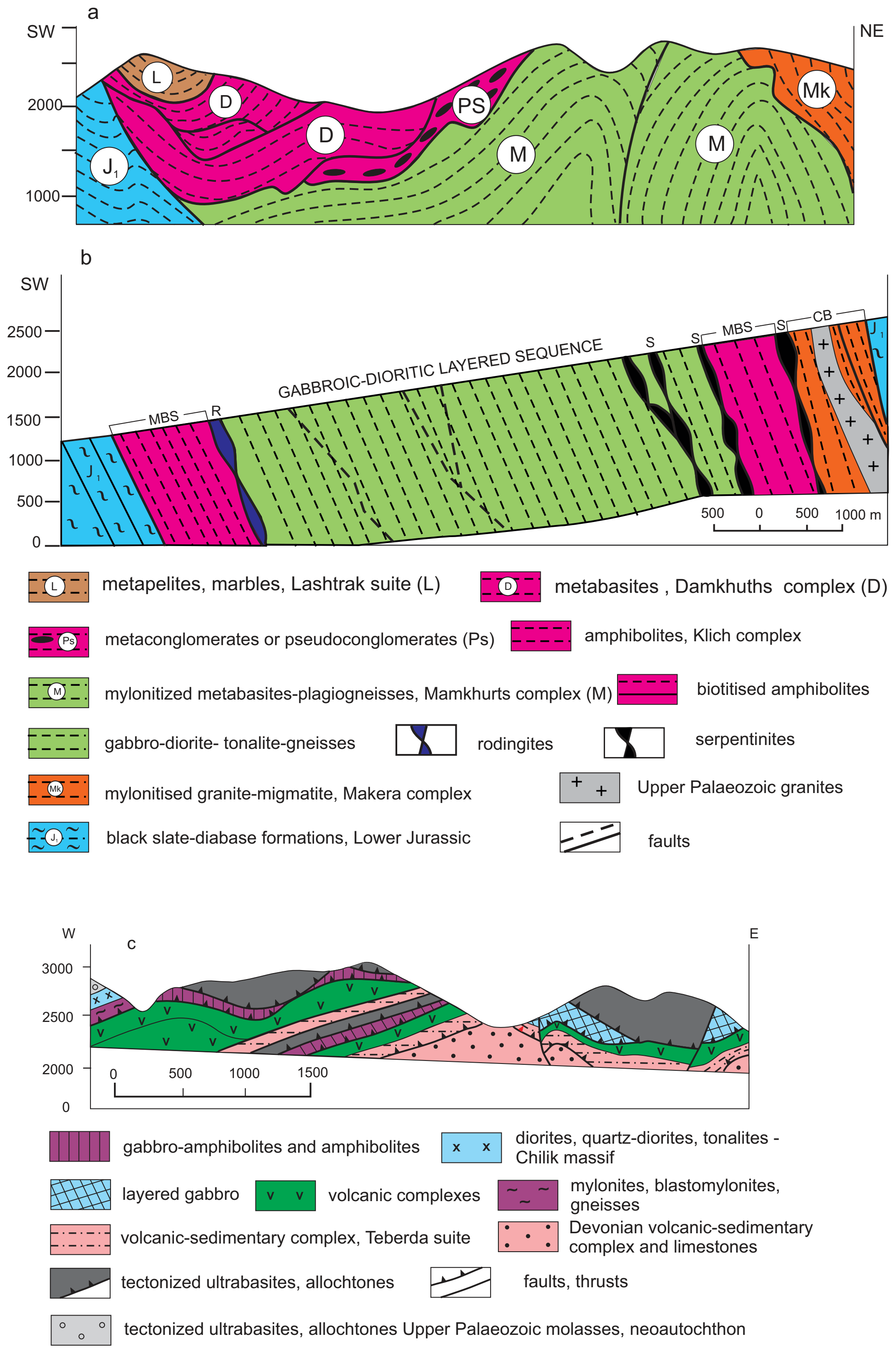


Figure 8. Cross sections of the pre-Jurassic complexes of the Great Caucasus. Abbreviations of La, Bu and Kh indicate location of the cross sections a, b and c, accordingly.

represented by micaschists, biotite-quartz schists, metavolcanoclastics, sericite-chlorite schists etc. The U-Pb SHRIMP average age of detrital zircons from metasandstones is 543 Ma (range= 573–579 Ma) and the age of zircons from orthogneisses is 530±8 Ma that indicate a Cambrian age of the complex (Somin 2007a). Metamorphic rocks of the Bechasin Zone are intruded by Upper Palaeozoic granitoids. Massive granites, granodiorites and quartz diorites, aplites and granite-porphyry (Malka- and Kuban-types of granitoids) are distinguished. Their petrochemical features indicate they are transitional from I- to S-types and I granites (Potapenko *et al.* 1999). Metamorphic and granite complexes of the Bechasin Zone are unconformably (tectonically) covered by Lower–Middle Palaeozoic sediments. In its turn, the latter are covered by a thick body of allochthonous ultrabasic rocks (serpentinites of the r. Malka, see Figures 2 & 7).

Palaeozoic Sedimentary Cover

Palaeozoic sedimentary cover is represented by various facies of terrigenous, carbonate, and volcanogenic deposits within almost all tectonic units of the Caucasus.

Southern Province

Within the Caucasian part of the central Iranian platform, the Palaeozoic sedimentary cover is slightly metamorphosed, deformed, and facially corresponds to shallow-marine type. Continuous Devonian–Carboniferous series of shelf deposits developed in Nakhchevan (Azerbaijan) and southern Armenia and mainly consist of coral-brachiopod limestones (often bituminous), quartzite sandstones, and argillites. The total thickness of the Devonian–Visean (Lower Carboniferous) deposits ranges between 3000–4000 m (Figure 9). There are no deposits of Middle and Upper Carboniferous age. Thick transgressive Permian bituminous algal foraminifera limestones (400–1000 m) containing corals, brachiopods, ammonoids, and conodonts unconformably overlies the Devonian and Carboniferous deposits (Aslanian 1970; Azizbekov 1972; Rustamov 2005).

Within the boundary zone, limestone blocks with Middle Carboniferous–Permian conodonts (Kariakin & Aristov 1990) are found in Karabagh within the mélange of the Lesser Caucasian ophiolite suture belt.

Northern Province

Within the *Transcaucasian massif (TCM)*, Palaeozoic deposits are known at the Khrami, Dzirula and Asrik-Chai salients of the basement. At the Khrami salient, they are represented by a shallow-marine and continental volcano-sedimentary formation with limestone lenses dated by corals, brachiopods, conodonts, and fossil flora as Upper Visean–Namurian–Bashkirian (see Figure 5). The formation is 600–800 m thick. With this formation are associated quartz porphyries and granite porphyries, which together with volcanoclastics form a volcano-plutonic formation. An analogous formation of subaerial dacite-rhyolitic lavas and volcanoclastics (1200 m thick) with rare andesitic and basaltic beds outcrops within the Dzirula salient (Figure 6). One more outcrop of a subaerial volcano-sedimentary formation represented by andesite-basaltic volcanics and containing remains of Upper Carboniferous flora (Gasanov 1986) is known north of Sevan Lake (Asrik-Chai). According to petrochemical features, Transcaucasian Upper Palaeozoic volcano-plutonic complexes correspond to formations of mature island-arc type.

Within the *Southern Slope* of the Great Caucasus, Palaeozoic deposits are known as phyllites of the Dizi series (1500–2000 m thick); they outcrop in the central part of the zone and are represented by strongly deformed rocks of relatively deep open marine basin located between island-arc morphostructural units of the Transcaucus and the Great Caucasus. In continuous succession there are alternations of pelite-siltstone and psammite-turbidite terrigenous deposits intercalated with chert, jasper, olistostromes, rarely volcanoclastics of andesite-dacitic composition, and lenses of recrystallized limestones (Figure 10). The presence of the Middle and Upper Devonian was established by conodonts from cherts, corals, crinoids and foraminifera from limestone (Kutelia 1983; Kutelia & Barskov 1983). Metamorphism of the Dizi series does not exceed greenschist-anchimetamorphism degree (Adamia *et al.* 2011).

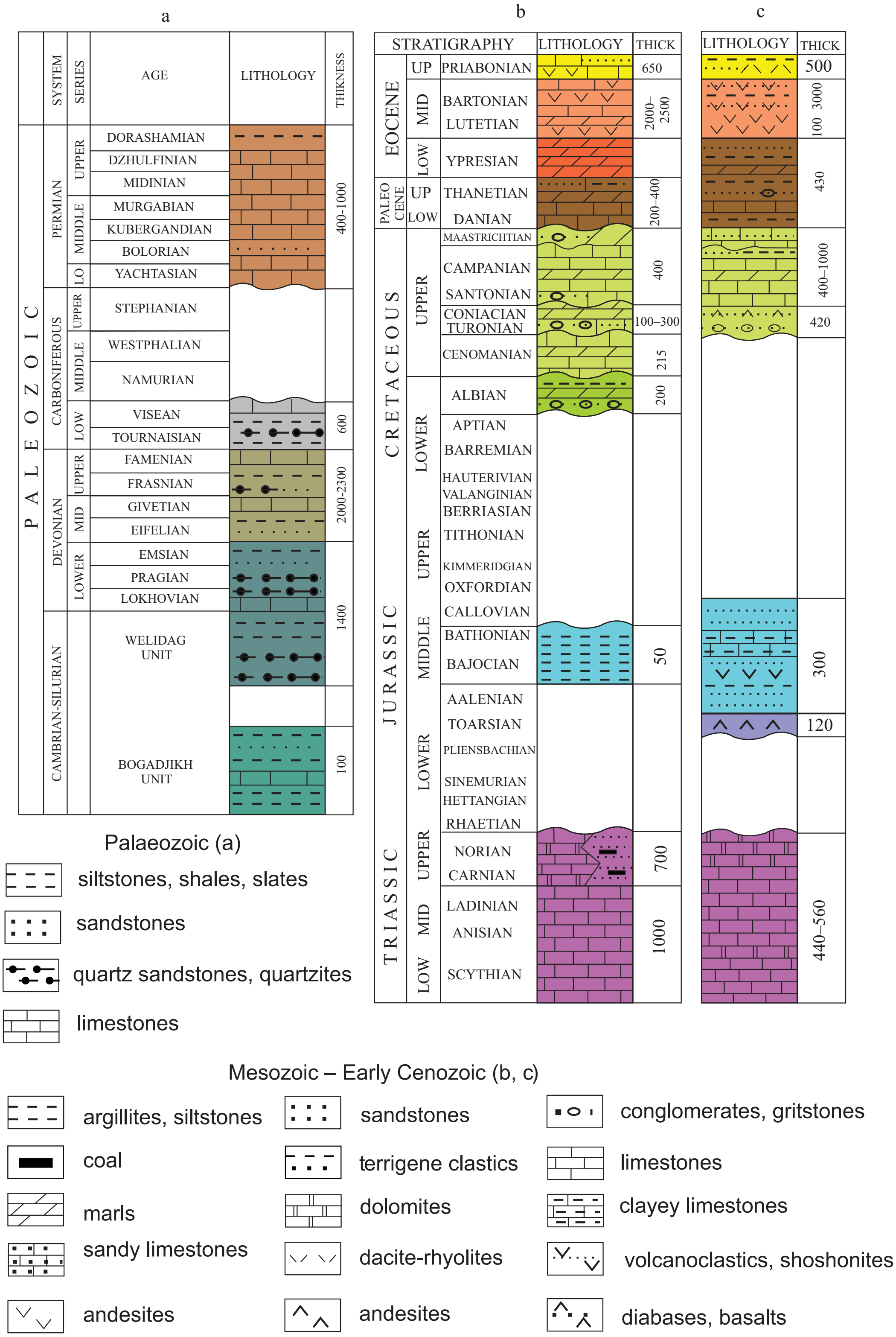


Figure 9. Generalized and simplified lithostratigraphic columns of **(a)** Palaeozoic (Belov *et al.* 1989) and **(b)** Mesozoic–Early Cenozoic pre-collisional units of the South Armenia and **(c)** Nakhchevan.

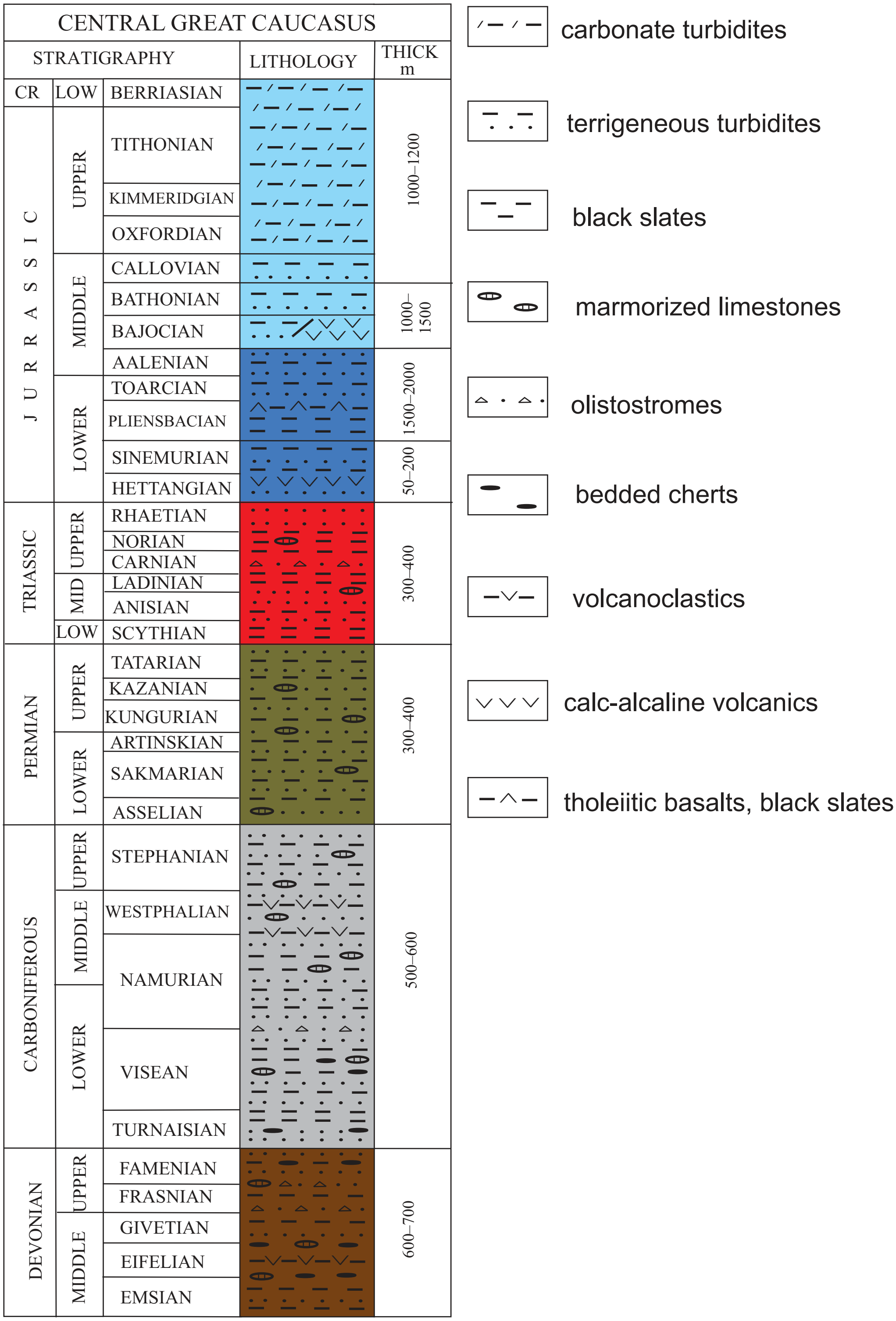


Figure 10. Generalised lithostratigraphic column of the Palaeozoic and Lower–Middle Mesozoic units of the central segment of the Great Caucasian Southern Slope Zone.

Within the *Main Range Zone*, Upper Palaeozoic sedimentary cover is composed of rocks accumulated under subaerial and shallow-marine environments. In the lower part of the Upper Palaeozoic section, there occur coarse clastic continental molasses, locally coal-bearing, with Middle Carboniferous floral fossils. Volcanics of quartz-porphyry composition are reported here. The higher horizons of the section constrained according to brachiopods to the Upper Carboniferous–Permian are represented by shallow-marine terrigenous molasses (Figure 11). The section is dominated by finely fragmented clastics and limestone lenses. Permian deposits, apparently, should also be attributed to marine terrigenous sediments. Terrigenous rocks alternate with organogenic limestones, the amount of which increases up-section. Fauna fossils in the limestones are represented by corals, brachiopods, gastropods, pelecypods, and foraminifera. The thickness of the deposits is 500–200 m (Andruschuk 1968; Somin 1971; Khutsishvili 1972; Belov 1981; Adamia *et al.* 2003).

The *Fore Range Zone* of the Great Caucasus confined from the south by the Pshekish-Tirniauz fault (see Figure 7) and representing a sub-lateral narrow trough is a depositary of polyfacial Middle–Upper Palaeozoic sediments. The Middle Palaeozoic of this zone (Silurian, Devonian and Lower Carboniferous) relates to facies of relatively deep sea, whereas the Upper Palaeozoic is represented by continental or shallow-marine molasses. The structural position of the middle Palaeozoic rocks is not quite clear: whether they are in autochthonous, paraautochthonous or allochthonous position. Many authors consider the Upper Palaeozoic as neo-autochtonous cover locally disrupted by Late Variscan and post-Variscan thrusts (Belov 1981; Belov & Omelchenko 1986).

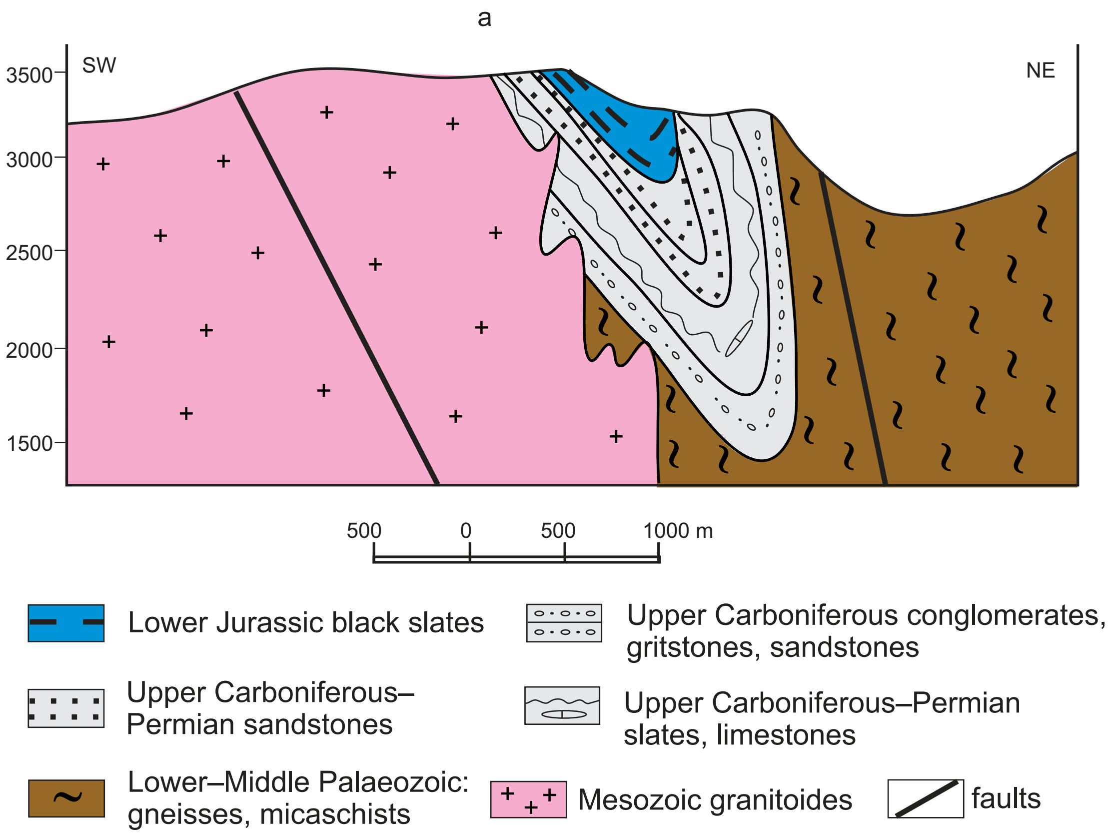


Figure 11. **(a)** Cross-section of the Kvishi formation (Khutsishvili 1972); **(b)** lithostratigraphic columns of the Upper Palaeozoic units, Great Caucasian Main Range Zone (Belov *et al.* 1989).

Within the middle Palaeozoic section of the Fore Range Zone, there is a distinguished deep-water hemipelagic graptolite-siliceous-volcanogenic formation with radiolaria (thickness more than 2500 m), argillite, siltstone, sandstone and siliceous schist (Silurian), terrigenous turbidites with olistostromes, and redeposited material of rocks of ophiolitic association (Devonian–Lower Carboniferous according to conodonts and other fossils). The Lower–Middle Devonian thick sequence plays an important role as a structure that according to its mineral-petrological features is attributed to formations of island-arc – intra-arc rifts (Belov *et al.* 1984; Adamia *et al.* 1987). The section is capped by a formation (up to 200 m thick) of terrigenous turbidites, olistostromes and limestones of Late Devonian–Early Carboniferous age (Figure 13).

Upper Palaeozoic molasses unconformably rest either upon Middle Palaeozoic deposits of the Fore Range Zone or upon ophiolitic allochthonous sheets of the zone (Belov & Kizevalter 1962; Andruschuk 1968; Somin 1971; Ajgirei 1976; Khain 1984). In the Fore Range Zone, they are represented by continental molasse of Middle–Late Carboniferous and Permian age (fine-clastics – Middle Carboniferous, coarse clastics – Upper Carboniferous, and red-coloured clastics – Permian). Island-arc type subaerial volcanic rocks are developed in the Middle Carboniferous (rhyolites) and in the Permian (sub-alkali andesite-dacites, thickness of 800 m) levels. The age of the deposits is established mainly by floral fossils (Andruschuk 1968; Belov 1981) belonging to the North Tethyan palaeobiogeographic province.

Upper Permian strata, besides continental molasses, comprises a carbonate-terrigenous formation (~300 m) of marine deposits with biohermal limestone bodies. The formation is dated on the basis of findings of brachiopods, corals, pelecypods, algae, sponges, gastropods, ammonoides, foraminifera etc.

Scythian Platform, Bechasin Zone– Within the Bechasin Zone, Palaeozoic deposits (Silurian–Middle Devonian) are represented by sandy-argillaceous carbonatic shelf facies (up to 2400 m thick) containing Silurian graptolites and Silurian-Devonian conodonts (Belov *et al.* 1984; Figure 14b). The Lower Palaeozoic, in particular Cambrian deposits, apparently, were represented by carbonate facies indicated by the presence of block limestones with Cambrian trilobites and brachiopods (Ajgirei 1976).

Late Proterozoic–Palaeozoic Development

The Gondwanan origin of pre-Cambrian crystalline massifs in the Central European Variscides has actively been discussed during recent decades (Paris & Robardet 1990; Tolluoğlu & Sümmer 1995; Stampfli 2000; Raumer *et al.* 2002). An interpretation in favour of a Gondwanan origin of Late Proterozoic Transcaucasian basement rocks was proposed by Zakariadze *et al.* (1998) and for the Tsakhkuniats massif (Armenia) by Agamalian (2004). The Arzakan and Aparan complexes (Tsakhkuniats) are considered as formations of the northern active margin of the Arabia-Nubian shield, which, at the end of the Proterozoic, experienced granitization related to the final stages of the Pan-African cycle of tectogenesis (Agamalian 2004). In contrast to the Tsakhkuniats, the Transcaucasian massif was not subjected to the process, since it broke away from the Arabia-Nubia shield and during Cambrian–Devonian time drifted deep into the Prototethys (Figure 15a) towards the northern continent (Zakariadze *et al.* 1998).

During the Early–Middle Palaeozoic, in the wake of northward migrating Gondwanan fragments the Paleotethyan basin was formed and, in the Ordovician, along its border with Transcaucasian massif, there occurred the subduction of oceanic crust accompanied by suprasubduction volcanic eruptions of rhyolites, denudation of the massif, and the formation of pudding conglomerates of the Chorchana-Utslevi Zone of the Dzirula salient (Abesadze *et al.* 1989).

Northward migration of the Transcaucasian massif throughout the Palaeozoic caused narrowing of Prototethys and its transformation into an oceanic back-arc (Dizi) basin (Figure 15b). Fragments of Palaeotethyan crust are met along the southern border of the TCM, within the accretionary complexes of the Sevan-Akera ophiolite suture, in the Pontides (Nilüfer formation, Pulur massif, Demirkent and Guvendic complexes – Yusufeli massif).

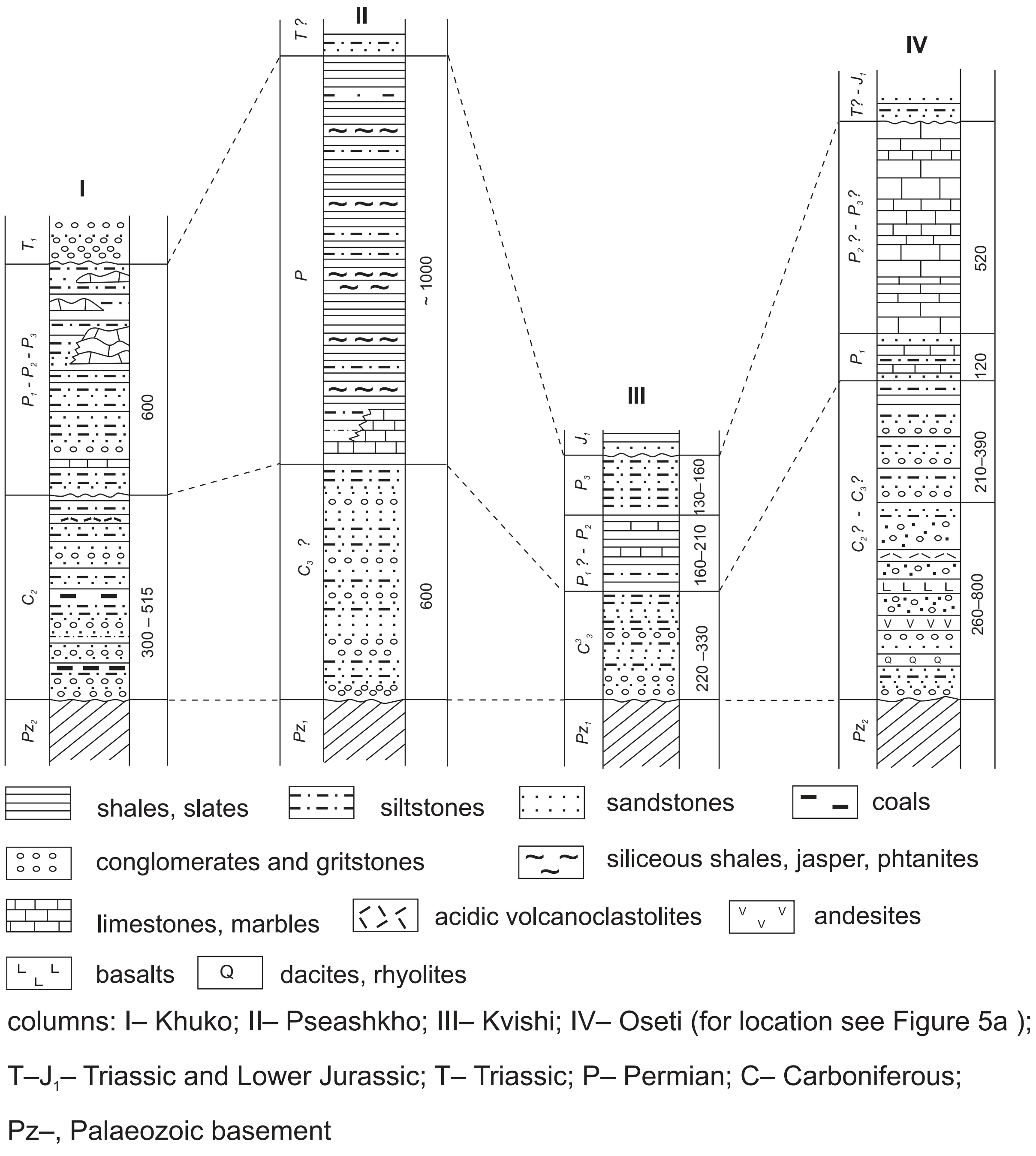


Figure 12. Lithostratigraphic columns of the Upper Palaeozoic units, Great Caucasian Main Range Zone (Belov *et al.* 1989).

At the western prolongation of the Sevan-Akera suture, in the Sakarya Zone, there occurs pre-Liassic, Upper Palaeozoic–Triassic series of metabasite-marble-phyllite low-grade metamorphic complexes, which are dated in the Pulur region as Late Permian (ca. 260 Ma, Topuz *et al.* 2004a), whereas the high temperature metamorphism in the same region is Early Carboniferous (ca. 330 Ma) in age (Topuz *et al.* 2004b). Metamorphic basic volcanic rocks in the Amasya region (Tokat group) which are believed to be equivalents of Agvanis and Yenişehir low-grade metamorphic rocks, can be interpreted as arc-related

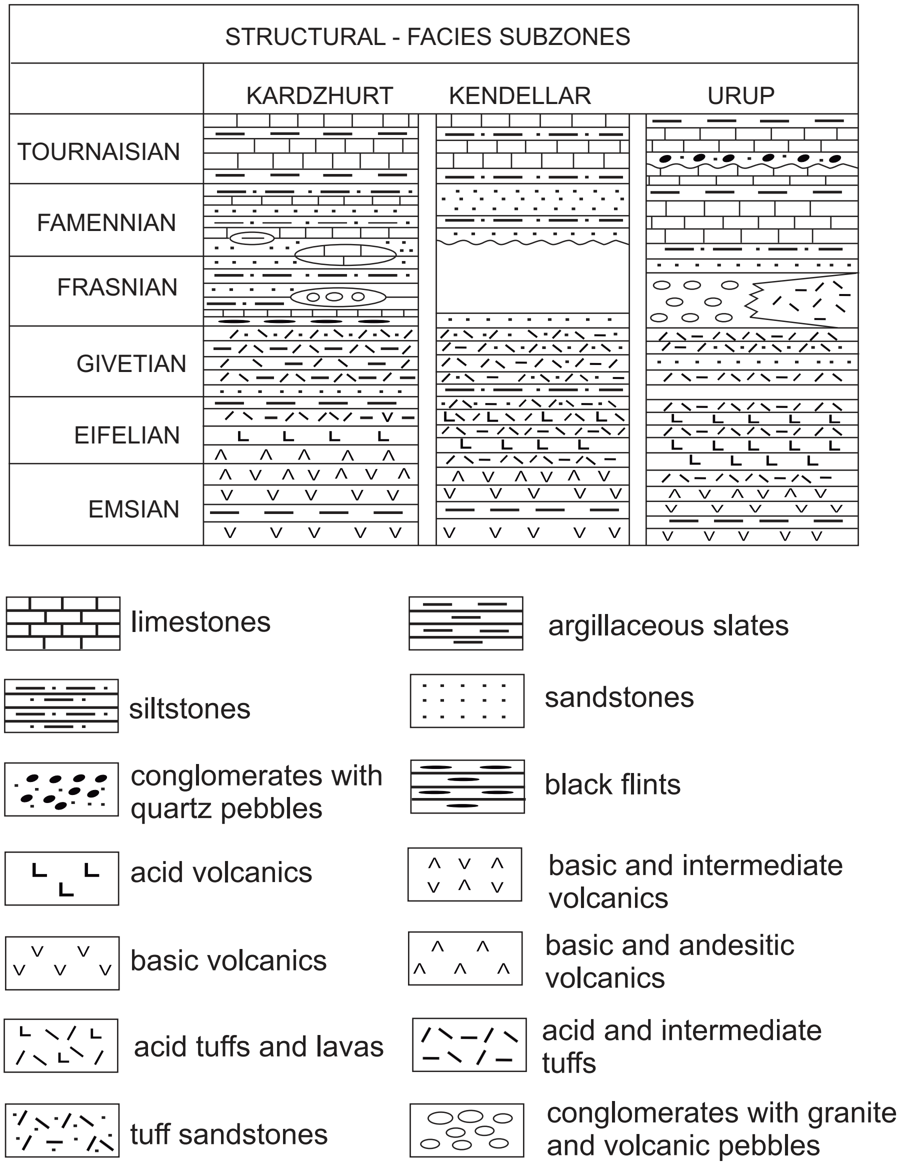


Figure 13. Generalised lithostratigraphic columns of the Palaeozoic units of in the Fore Range (Belov *et al.* 1989).

basinal sequences metamorphosed during the Late Palaeozoic (Göncüoğlu 1997; Okay & Şahintürk 1997). According to Topuz *et al.* (2009), the Kurtoğlu (Gümüşhane) metamorphic complex is the oldest rock assemblage known in the eastern Pontides, and it represents the vestiges of Middle Devonian oceanic subduction. The Çal Unit in Pontides consists of debris and grain flows with Upper Permian limestone and mafic pyroclastic flows, calciturbidites and radiolarian cherts dated as Late Permian. The unit may represent an oceanic seamount, which accreted to the Laurasian margin during the Middle Triassic (Okay 2000). Wall rock of the Demirkent dyke complex is composed of gabbros, microgabbros, and diabases. It can be supposed that some parts of these rocks are of Early–Middle Carboniferous age (Konak & Hakyemez 2001).

Relics of Palaeotethyan crust crop out in Iranian Karadagh directly to east of the Lesser Caucasian (Sevan-Zangezur) suture (Rustamov 2005).

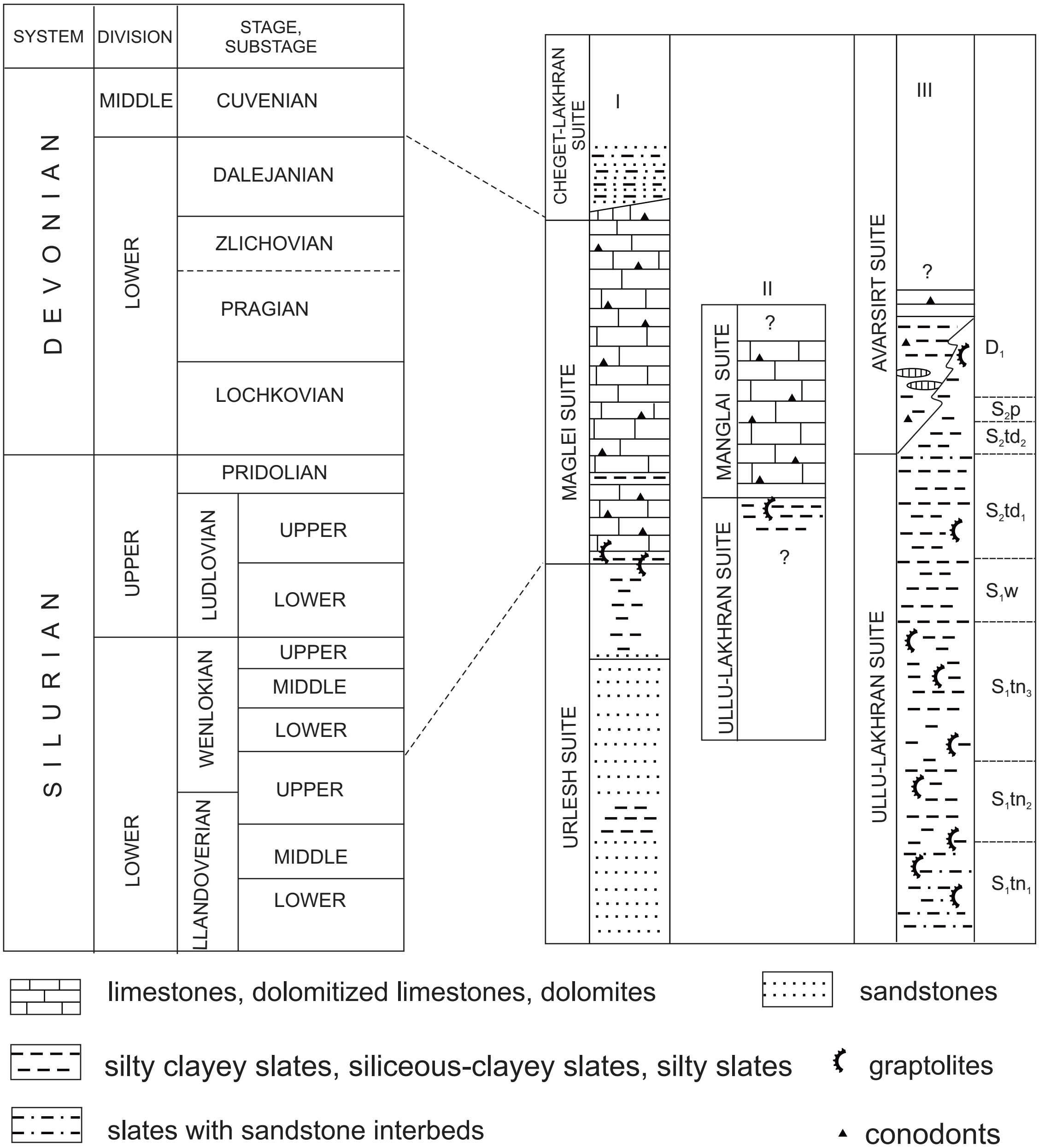


Figure 14. Generalised lithostratigraphic columns of the Palaeozoic units in the Bechasin zone (Belov *et al.* 1989).

Palaeozoic ophiolites are known also near Rascht (Majidi 1979; Davoudzadeh & Shmidt 1981), along Alborz, and as far as Binalud (Lyberis & Manby 1999; Zanchi *et al.* 2007).

Subduction of Palaeotethyan oceanic crust beneath the Transcaucasian island-arc system was going on. In the Late Palaeozoic, granite intruded the island-arc (TCM), there occurred lava extrusions and eruptions of volcanoclastics of island arc type under subaerial and shallow sea environments (see Figure 15).

The Variscan granitoids of the TCM (Loki and Khrami salients) correlate with leucogranites, granite-gneisses and plagiogranites of S-type of the Artvin salient, which are exposed in the easternmost part of the Pontides (Turkey), along the Çoruh river

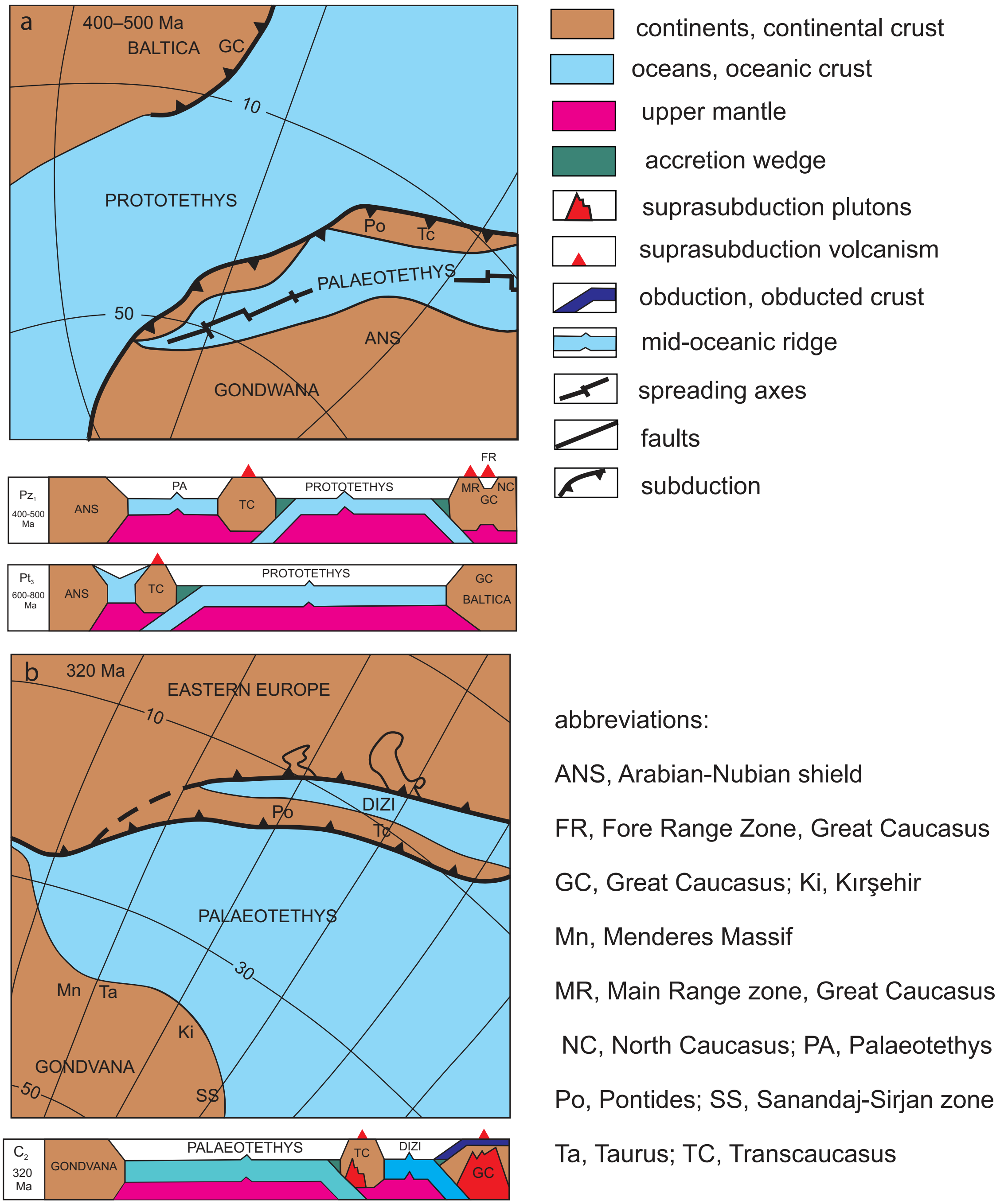


Figure 15. Late Proterozoic–Early Palaeozoic **(a)** and Late Palaeozoic **(b)** palaeotectonic reconstructions of the Black Sea-Caspian Sea region.

valley, not far from the Georgia-Turkey border. The Artvin granitoids are unconformably overlain by the Narlık anchimetamorphic formation of black slates, sandy-argillaceous turbidites, cherts (with Lower Jurassic radiolarians, Adamia *et al.* 1995), volcanic and volcanoclastic rocks (Yılmaz *et al.* 2001). The Upper Palaeozoic granitoids of the TCM may also be correlated with the Gümüşhane and other granites of the eastern Pontides (Çoğullu & Krummenacker 1967), which are overlain by Upper Carboniferous shallow marine to continental terrigenous and carbonate deposits alternating with dacite volcanics (Okay & Şahintürk 1997). The Gümüşhane granitoid has given zircon ages of 324 Ma (Topuz *et al.* 2010).

There is no evidence for Variscan metamorphic and magmatic events, folding and topographic inversion within the Southern Slope Zone of the Great Caucasus. Variscan and Eo-Cimmerian events did not lead to closure of the Dizi back-arc basin. During the Middle–Late Palaeozoic, within the Dizi basin, thick terrigenous turbidites, pelagic and hemipelagic deposits, andesite-dacite volcanoclastics, and carbonates were accumulated under deep-marine environment. The accretional prism consisting of tectonic slices of oceanic crust and island-arc was formed in a transitional back-arc to island-arc zone of the Great Caucasus. Obducted metaophiolitic sheets shifted from the root zone located along the southern margin of the island arc northward to the Fore Range Zone. Widespread emplacement of microcline granite plutons along the active continental margin of southern Eurasia during 330–280 Ma occurred above a north-dipping Paleotethyan subduction zone. Carbonate platform lasted in the southern province of the Caucasus (South Armenia-Nakhchevan) throughout the Palaeozoic (see Figure 9).

Mesozoic–Early Cenozoic Development

Division of the Caucasus into northern and southern provinces was also distinct throughout the Mesozoic–Early Cenozoic. In the extreme south of the region (Nakhchevan) and in the southern Armenia, a platformal regime remained with the accumulation of uniform, mainly carbonate shelf sediments. The ophiolite belt of the Lesser Caucasus is represented by rocks of the Tethyan oceanic basin; the Northern Province is represented by developed facies characteristic of active continental margins similar to Asiatic-margin-type of the present-day Pacific: island-arc, back-arc, continental shelf.

Southern Province

Triassic sediments of the Southern Province are dominated by shelf organogenic limestones, marls, and dolomites (south Armenia, Nakhchevan), while the northern part of the Province (Jermanis, Armenia) is represented by sediments of Upper Triassic shallow-marine terrigenous-coal-bearing formations (Aslanian 1970; Azizbekov 1972). Lower Triassic deposits are similar to shelf limestones of Permian age. The Late Triassic age of the terrigenous units was identified by molluscs and flora fossils (see Figure 9b).

Jurassic deposits in the Southern province are rare and are represented by shelf facies. In the near-Araksian part of the province (Julfa-Ordubad), Triassic limestones are overlain by the following formations: (1) diabasic porphyrites with interbeds of tuffs, thickness of 120 m (Lias–Dogger); (2) sandy-argillaceous deposits with interbeds of tuffogenic sandstones with pelecypods indicating an Aaleanian–early Bajocian age (50–100 m thick); (3) sandy-argillaceous-carbonate formation (over 200 m thick) with Upper Bajocian, Bathonian, Callovian ammonites (Azizbekov 1972). K-Ar age of gabbrodiabasic intrusions of the Julfa area varies between 175–186 Ma (Early–Middle Jurassic, Abdullaev & Bagirbekova 2007).

Data on Early Cretaceous facial environments are very scarce. Based on the features of Albian deposits known in the near-Araksian part of the Southern Province (sandstones, marls up to 200 m thick) containing abundant mollusc fossils, it can be considered that this domain during the Early Cretaceous was a marine basin; Upper Cretaceous deposits are widespread and also represented shallow-marine formations consisting mainly of organogenic-clastic limestones, sandy-argillaceous limestones, and marls. All Late Cretaceous stages are present. The sediments are several hundreds meters thick. Various levels bear conglomerates and sandstones; frequent stratigraphic unconformities indicate the shallowness of the basin. Fauna fossils are represented by gastropods, large pelecypods, rudistids and foraminifera (Aslanian 1970; Azizbekov 1972).

Paleocene (Danian, Thanetian) deposits represented by sandy-argillaceous-limestone shallow marine facies (thickness 200–400 m) are similar to Upper Cretaceous ones. Fossils are represented by echinoids and foraminifera. Relations with underlying formations are diverse: in some localities, the transition is gradual, though unconformity and basal conglomerates are also met. Unconformity is also evident in Lower Eocene rocks represented by shallow marine marls, organogenic limestones and calcareous sandstones; the thickness of the deposits ranges from 10's to 1000 m.

The environment of sedimentation within the basins of the Southern Province significantly changed in the upper part of the Eocene: from the end of Early – beginning of Middle Eocene there started a period of intensive submarine volcanic eruptions (see Figure 9b). Lavas, pyroclastics (volcanic tuffs), tuff-turbidites, more than 2000 m thick, accumulated locally alternating with terrigenous and carbonate deposits, which often occur in the base of the Middle Eocene forming together with conglomerates basal-transgressive part of the section. Besides nummulites, which allow confident dating of the rocks, these deposits contain large molluscs, echinoids, brachiopods, corals, and some other fossils indicating shallow sea environment (Aslanian 1970; Azizbekov 1972). Eocene volcanism of the Southern Province, as well as that of more northern zones of the Caucasus, represent a part of the vast andesite belt of the Middle East, running from Aegean Sea, through Turkey, the Lesser Caucasus, Iran, and as far as to Afghanistan. According to their petrochemical characteristics the volcanics are attributed to island-arc type. They are mainly represented by calc-alkaline series, however, rocks of tholeiitic and shoshonitic series are also present (Lordkipanidze *et al.* 1989; Vincent *et al.* 2005).

Volcanic activity sharply decreased at the end of the Late Eocene and under shallow-marine environments there occurred the accumulation of mainly sandy-argillaceous and carbonate sediments including coral limestones. Upper Eocene volcanic rocks are represented by thin andesitic bands and their clastics. The maximum thickness of the Upper Eocene interval does not exceed 500–600 m (Aslanian 1970; Azizbekov 1972).

Boundary Zone: Ophiolitic Suture Belt

Allochthonous ophiolites of the Sevan-Akera Zone include two types of series: tholeiitic and boninitic (Zakariadze *et al.* 1993). Each one shows a complete ophiolite sequence from ultramafic-mafic cumulates to massive plutonics, dyke swarms, and pillow lavas. An age of the boninitic plutons was reliably defined as Bathonian–Callovian (K-Ar, 168±8 Ma, biotite and muscovite from plagiogranite, Morkovkina & Arutiunian 1971; U-Pb 160±4 Ma, zircon from quartz-diorite, Zakariadze *et al.* 1983). Sm and Nd isotopes and obtained mineral isochrons for two gabbro-norites from representative tholeiitic gabbroic massifs (Lev-chai and Alticovshan-east of Lake Sevan) show an Upper Triassic age (Carnian –Norian) with rather low initial ε_{Nd} values T= 226±13, ε_{Nd}(T)+ 5.1±0.4 (Lev-chai) and T= 224±8.3 ε_{Nd} (T)= +4.0±0.3 (Altikovshan). Petrological and geochemical data show that the tholeiitic sequence of Sevan-Akera Zone has a supra-subduction origin (Bogdanovsky *et al.* 1992).

Mesozoic (?) and pre-Mesozoic sedimentary and magmatic rocks of the suture belt are met in the mélange as clastic-blocks of various dimensions. Among them are blocks of Upper Triassic limestones and calcareous sandstones with ammonites (Solovkin 1950), basaltic volcanic rocks, and radiolarite associations (see Figure 3) with Late Triassic, Lower and Middle Jurassic–Lower Cretaceous radiolarians (Knipper 1980, 1990; Zakariadze *et al.* 1983). The effusive-radiolarite part of the ophiolitic association, in some localities, includes lenses of Upper Jurassic–Neocomian reef limestones (Gasanov 1986). Lower, Middle and Late Jurassic as well as Lower Cretaceous radiolarians assemblages are known from several places of the Sevan-Akera ophiolite belt (Zakariadze *et al.* 1983; Knipper 1990). The younger ophiolitic association starts with the transgressive Albian–Senomanian, which is represented by flysch and olistostromes. At higher levels they are followed by basalts and Albian–Lower Coniacian radiolarites (Sokolov 1977).

Data from recent studies concerning ages and composition of Lesser Caucasian ophiolites of Armenia agree well with previous results. The investigation of Jurassic ophiolites from NW Armenia evidence the occurrence of several suites comprising rock representatives of slow-spreading ophiolite type, of a remnant oceanic plateau, and arc-type volcanic rocks of probable Upper Cretaceous age (Galoyan *et al.* 2007).

Radiolarians from radiolarites associated with ophiolitic volcanic rocks from three ophiolitic units cropping out in NW Armenia, east of Lake Sevan and in central Armenia suggest Middle–Upper Jurassic and Lower Cretaceous ages of the rocks (Danelian *et al.* 2007).

The $^{40}Ar/^{39}Ar$ age of amphibole-bearing gabbros of the Sevan-Akera ophiolites of north Armenia evidence a Middle Jurassic age (165.3±1.7 Ma) for oceanic crust formation (Galoyan *et al.* 2009). $^{40}Ar/^{39}Ar$ phengite ages obtained for the high-pressure assemblages (glaucophane-aegirine-clinozoisite-phengite) of the tectonic mélange (NW Armenia) range between 95 and 90 Ma, while ages of epidote-amphibolite retrogression assemblages are 73.5–71.0 Ma (Rolland *et al.* 2009). According to Rolland *et al.* (2009a), the Lesser Caucasus (Sevan, Stepanavan and Vedi) demonstrates evidence for a slow-spreading oceanic environment in the Early to Middle Jurassic. The oceanic crust sequence is covered by OIB alkaline lavas. $^{40}Ar/^{39}Ar$ dating of amphibole provides a Early Cretaceous age of the rocks (117.3±0.9 Ma).

The neoautochthonous complex of the ophiolitic belt starts with the transgressive formation of Coniacian–Santonian clastics; this indicates shallowing and closing of the deep-water basin and its replacement by a back-arc one. The younger Senonian–Lower Paleogene limestone formations were formed in the shallow basin, as well as sandy-argillaceous deposits of the Paleocene–Lower Eocene and calc-alkaline andesitic volcanic formations, terrigenous clastics and carbonates of the Middle and Upper Eocene.

Northern Province

Transcaucasian Massif– Within the Triassic–Eocene, the extreme southern tectonic unit of the Northern Province, i.e. the Transcaucasian massif (south Caspian-Transcaucasian massif according to Rustamov 2005), developed as a relatively uplifted structural-morphological island-arc type unit. Within its central part (Georgian, Artvin-Bolnisi and Azerbaijan massifs), the accumulation of mainly shallow-water, lagoonal-lacustrine, coal-bearing and salt-bearing deposits of variable thickness occurred accompanied by island-arc type volcanic eruptions (Figure 16).

At the Georgian massif, the Mesozoic sedimentary cover begins with subaerial volcanoclastics of rhyolitic composition (about 800 m) containing Upper Triassic flora of North-Tethyan palaeobiogeographical domain (Svanidze *et al.* 2000; Lebanidze *et al.* 2009). The Lower Jurassic and Aalenian are built up of arkosic terrigenous clastics and shallow-water organogenic limestones (‘red ammonitic limestones’) containing in addition to ammonoids, abundant crinoids and brachiopods, which also belong to the North-Tethyan palaeobiogeographical domain (Lebanidze *et al.* 2009). The Bajocian is almost fully represented by tuff-turbidites with rare bands of calc-alkaline andesite-basalts (Gamkrelidze 1964; Moshashvili 1982; Beridze 1983). The Bathonian is composed of freshwater-lacustrine coal-bearing sandy-argillaceous rocks followed by variegated lagoonal Callovian–Upper Jurassic deposits containing bands of coal, gypsum, and anhydrite. In the bottom of the salt-bearing variegated formation, there is a thick strata of sub-alkaline-alkaline basalts, which, according to its stratigraphic position, is attributed to the Upper Jurassic. The thickness of the Jurassic deposits varies, and locally reaches 3000–4000 m (Gamkrelidze 1964; Azizbekov 1972; Moshashvili 1982; Lordkipanidze *et al.* 1989).

During the Jurassic, along the southern and northern margins of the Artvin-Bolnisi, Azerbaijan and Georgian massifs, the environment of sedimentation has significantly changed: thickness increased, and Lower Jurassic–Aalenian limestone facies were replaced by deep-sea terrigenous turbidites. Bajocian tuff-turbidites were replaced by coarse volcanoclastics and lavas of andesite-basaltic and dacite-rhyolitic composition; folding of the deposits is more expressed. On the basis of these features the margins of the Artvin-Bolnisi, Azerbaijan and Georgian massifs are considered to be individual zones: Loki (Bayburt-Somkhiti)-Karabagh (in the south) and Gagra-Java (in the north) (Gamkrelidze 1964; Aslanian 1970; Azizbekov 1972). Within the Loki-Karabagh Zone, Bathonian–Callovian–Upper Jurassic facies also underwent significant changes. Carbonate and salt-bearing formations were replaced by normal-marine

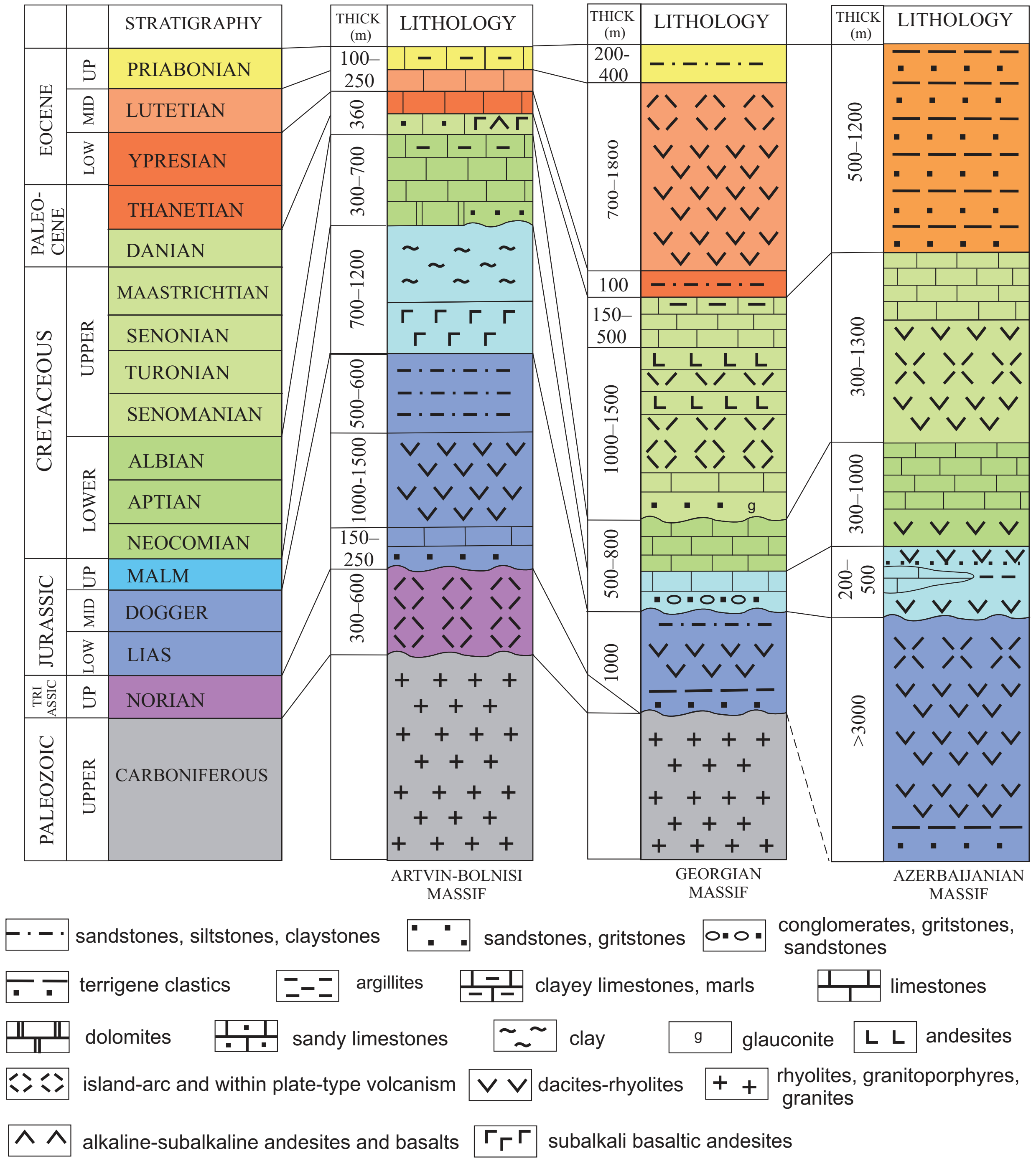


Figure 16. Generalised and simplified lithostratigraphic columns of the pre-collisional units of the western Transcaucasus: **(a)** Georgian and **(b)** Artvin-Bolnisi massifs and **(c)** Eastern Transcaucasus (Azerbaijanian massif).

terrigenous-carbonate (locally reef limestones) rocks and by thick volcanic rocks of basalt-andesite-dacite-rhyolite composition of calc-alkaline series (Aslanian 1970; Azizbekov 1972; Lordkipanidze *et al.* 1989). Upper Jurassic–Neocomian calc-alkaline series were replaced by tholeiites of MORB-type and by association of alkaline basalts in the immediate contact between the Bayburt-Karabagh Zone and

Lesser Caucasian ophiolitic belt; the Bajocian contains rocks of boninitic series (Zakariadze *et al.* 1983; Lordkipanidze *et al.* 1989).

Lower Cretaceous facies of the Trancaucasian massif are rather uniform (see Figure 16). Within the Georgian massif, there occurs a transgressive basal conglomerate-quartz sandstone formation, which is followed by shelf carbonate sediments (Eristavi 1962): dolomites, limestones of the Urgonian facies, organogenic limestones and marls (thickness from hundreds to 2500 m). In the Bayburt-Karabagh Zone, limestones are replaced by basalt-andesite-dacite-rhyolitic volcanic rocks of island-arc and intraplate types (Lordkipanidze *et al.* 1989).

Within the Georgian massif, Upper Cretaceous, Paleocene and Eocene deposits built up mainly of neritic organogenic limestones and marls are also facially uniform; their thickness in some places exceeds 2000 m. Formations of alkaline basalts and volcanoclastics stratigraphically constrained to the Turonian–Senonian stages are locally developed. Glauconitic sandstones predominate at the base of the Upper Cretaceous. However, to the south, within the Artvin-Bolnisi massif, the Upper Cretaceous section is dominated by volcanic rocks: basalts, andesites, dacites and rhyolites (lavas, pyroclastics) of calc-alkaline series. Their thickness reaches 3000–4000 m. Volcanic rocks are of shallow marine-subaerial type, ignimbrites are frequent. Volcanic eruptions have ceased in the Late Senonian, and throughout Maastrichtian and Early Paleocene at the whole territory of the Transcaucasian massif (as well as in the Sevan ophiolitic belt and Southern Province), uniform limestone-marly facies of shelf-sea have been formed.

In the Paleocene–Eocene, within the Transcaucasian massif (island-arc), north of the Early Paleogene andesitic belt of the Middle East, basaltic troughs (rifts) originated (Lordkipanidze *et al.* 1989): Talysh and Achara-Trialeti dividing the Transcaucasian massif into the Georgian (in the north) and Artvin-Bolnisi (in the south) massifs (Figure 17). Within these troughs, the Paleocene–Lower Eocene is represented by thick formation (up to 1500 m) of terrigenous turbidites (Borjomi Flysch and its analogues), which, in some localities, is associated with dacite-rhyolitic lavas and pyroclastics (thickness of 100–300 m). They are overlain by the Eocene formation of bimodal, calc-alkaline, sub-alkaline and alkaline volcanic rocks (andesites, shoshonites, basanites etc) whose maximal thickness in Achara is 5000 m (Lordkipanidze & Zakariadze 1974, 1986; Banks *et al.* 1997; Nadareishvili & Sadradze 2004). In Talysh, Eocene volcanism has more alkali tendencies (trachybasalts-trachyandesite basalts-phonolitic formation, Allen *et al.* 2003, 2004; Mamedov 1998; Vincent *et al.* 2005; Rustamov 2005) with which are associated intrusives of sub-alkaline basic-ultrabasic formation: sub-alkali peridotites (age 38–41 Ma) and gabbro-syenites (34–36 Ma). In the Late Eocene, volcanic activity gradually decreased. This time 'basaltic troughs', accumulated mainly terrigenous turbidites with admixtures of volcanoclastic material.

Great Caucasus– Mesozoic–Early Cenozoic basin of the Great Caucasus located behind the island-arc of the Transcaucasus was developing, at least, from Devonian, throughout Palaeozoic and Mesozoic–Early Cenozoic (Figure 18). In the upper part of the section cropping out within the central segment of the Southern Slope Zone of the Great Caucasus (Dizi series of Svaneti, river Mzymta gorge), there occurs terrigienous-turbiditic formation with intercalations of carbonate rocks whose Late Triassic age was established by corals and foraminifera (Andruschuk 1968; Saidova *et al.* 1988). In transitional Triassic–Lower Jurassic terrigenous turbidites E. Planderova reported the presence of marine microfossils of Rhaethian and Hettangian age (Adamia *et al.* 1990). Sinemurian terrigenous turbidites and volcanoclastics of the central part of the Southern Slope Zone conformably follow Hettangian levels. However, Liassic deposits transgressively overlie basement rocks along its southern and northern margins. Within the basin of the Great Caucasus, the Pliensbachian is generally represented by black slate formation locally with beds of diabase. Tholeiitic basalts of MORB-type – pillow lavas and agglomerates- are also confined to this level (Beridze 1983; Lordkipanidze *et al.* 1989). Within the entire basin, the Toarcian–Aalenian is composed of terrigenous proximal turbidites along its border with the Transcaucasian massif (island-

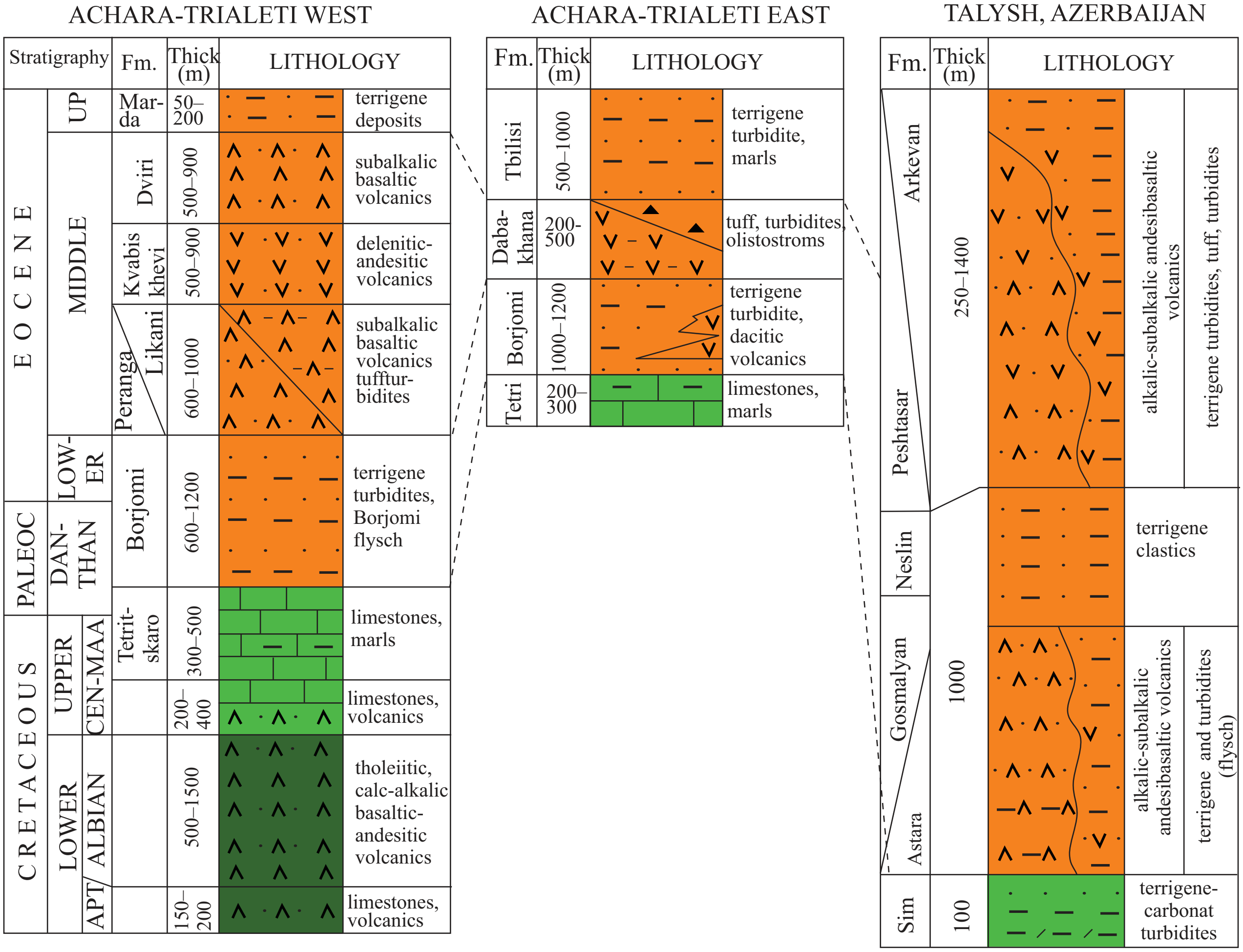


Figure 17. Generalised and simplified lithostratigraphic columns of the pre-collisional units of the Achara-Trialeti: west **(a)**, east **(b)** and Talysh **(c)** basins.

arc), and distal turbidites in the axial part of the Southern Slope Zone. Paroxysms of volcanic activity are periodically evident throughout the Liassic that is confirmed by the presence of admixtures, intercalations, and volcanic formations (basalts, keratophyres) in Sinemurian, Toarcian and Aalenian deposits (Lomize 1969; Beridze 1983; Ali-Zade 2003; Panov & Lomize 2007; Tuchkova 2007). During the Aalenian, volcanic eruptions under deep marine environment also bear the features of MORB-type; however, they also show resemblance to the island-arc type (Lordkipanidze *et al.* 1989).

It was mentioned that Bajocian lavas and volcanoclastics represented by calc-alkaline andesite-basalts are widely spread along the southern margin of the Southern Slope Zone, at its border with the Transcaucasian massif (Gamkrelidze 1964; Andruschuk). Northward, the Bajocian lava and pyroclastics grade into tuff-turbidites, and further on, into terrigenous turbidites (Beridze 1983). Black slate, terrigenous-turbiditic, and volcanogenic formations of the Southern Slope Zone (thickness more than 5000 m) are dated predominantly by ammonites, though, some limestone lenses host also pelecypods and brachiopods.

After the Bajocian, the environment of back-arc basin of the Great Caucasus has significantly changed: volcanic eruptions ceased, along the southern side of the basin there were deposited marine Bathonian, Callovian and Lower Oxfordian terrigenous sandy-

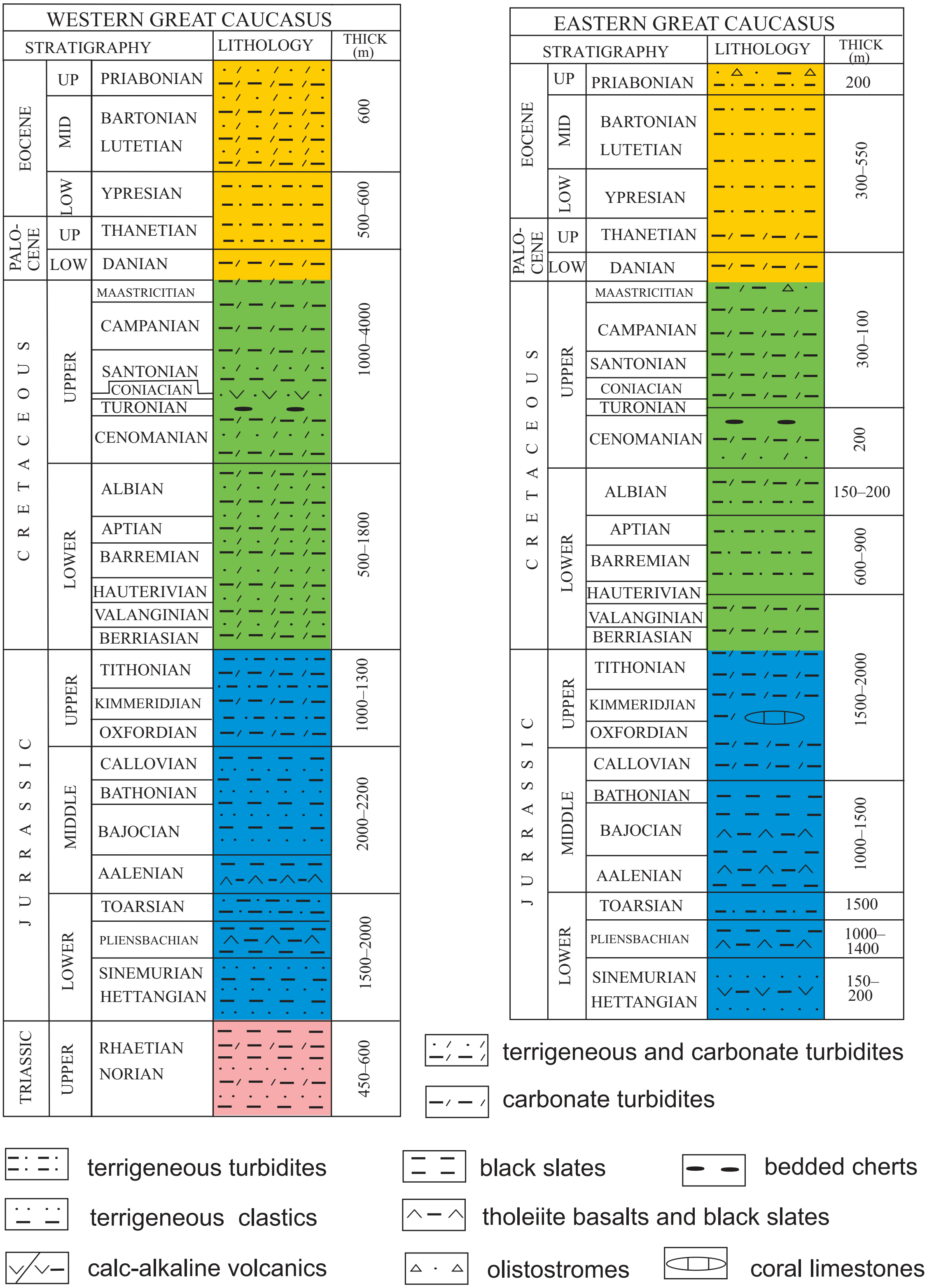


Figure 18. Generalised and simplified lithostratigraphic columns of the Mesozoic–Lower Cenozoic units of the Great Caucasian Southern Slope zone (western and eastern segments).

argillaceous sediments, which northward gain the features of terrigenous and carbonate turbidites (Adamia 1977, 1989). There were developing two, en échelon flysch basins: eastern and western ones, in which turbiditic sedimentation lasted almost uninterruptedly throughout Late Jurassic, Cretaceous, Paleocene, and Eocene. Here, thick formations of predominantly carbonate turbidites (Oxfordian–Valanginian, Turonian–Senonian) alternate with terrigenous turbidites (Hauterivian–Senomanian, Paleocene–Eocene) and olistostromes (Maastrichtian, Priabonian). The total thickness of the whole Jurassic–Cretaceous–Paleogene sedimentary complex, apparently, exceeds 10 000 m (12–15 km according to Saintot *et al.* 2006). Rocks are strongly deformed: isoclinal folds and steep south-vergent thrusts-overthrusts are common; cleavage and boudinage are strongly developed. Small-scale nappe structures are known only on the southernmost strip of the Southern Slope Zone.

The Northern Slope of the Great Caucasus – Southern Margin of the Scythian Platform: Laba-Malka Zone– The northern edge of the Great Caucasus, which is known as the Laba-Malka Zone (Milanovsky & Khain 1963), during the Mesozoic–Early Cenozoic, is featured by development of continental-shelf deposits. Triassic, Jurassic, Paleocene, and Eocene deposits are represented by shallow-marine and continental facies (Milanovsky & Khain 1963; Andruschuk 1968; Ajgirei 1976). Gaps, stratigraphic and angular unconformities are frequent (Figure 19). The Lower and Upper Triassic in the western part of the basin is represented by carbonate rocks with crinoids, corals, sponges, ammonoids, pelecypods, brachiopods, and foraminifera. The Middle Triassic is dominated by terrigenous clastics. Thickness of the Triassic ranges from 300 to 1500 m (Andruschuk 1968; Ajgirei 1976; Saintot *et al.* 2006). Palynomorph-bearing red-coloured continental molasse of the central part of the Northern Slope indicate a Triassic age.

Lower and Middle parts of the Jurassic series of the Northern Slope of the Great Caucasus are mainly composed of terrigenous clastic rocks. Shallow-marine facies with ammonites, brachiopods, and pelecypods alternate with continental coal-bearing facies. The Pliensbachian, Toarcian and Bajocian include thin horizons of crinoidal limestones. Gaps in the sedimentation are frequent; stratigraphic unconformities are present. The total thickness of the deposits within the Western Caucasus is several thousands metres. In the central part of the Northern Slope, their thickness decreases and, at the Pliensbachian level, there occur formations of calc-alkaline andesitic and dacitic lavas and volcanoclastics (thickness up to 200 m). The thickness of Lower–Middle Jurassic deposits increases in the eastern part of the Northern Slope – the Limestone Dagestan Zone (up to 10 000 m – Lower Jurassic, and up to 5000 m – Middle Jurassic; Kakhadze *et al.* 1957; Andruschuk 1968; Panov & Lomize 2007; Tuchkova 2007).

The next time-stage of the development of the basin of the Northern Slope of the Great Caucasus (Scythian platform) begins with the Callovian. Within the period starting in the Callovian and going on throughout Late Jurassic, Cretaceous, Paleocene, and Eocene, a shallow-marine environment ranged. The base of the Callovian is represented by the basal formation that marks a beginning of marine transgression. The Callovian is usually represented by terrigenous-carbonate shallow-water deposits with ammonites, bivalve molluscs, and gastropods. The Oxfordian and Kimmeridgian are dominated by dolomites and reef limestones. The Upper Jurassic section is capped with predominantly lagoonal facies represented by terrigenous-carbonate salt-bearing variegated formation. The thickness of the Callovian–Tithonian deposits ranges from some tens to 2300 m (Andruschuk 1968). Cretaceous deposits of the Northern Slope Basin are facially very diverse. Almost complete sections are replaced by the ones showing a lot of stratigraphic unconformities; their thickness varies from tens to some thousands of metres. The Lower Cretaceous is dominated by sandy-argillaceous rocks (Hauterivian–Albian), and only within its base, there are marls and limestones (Berriasian–Valanginian–Hauterivian) containing an abundant fauna. The Upper Cretaceous, at the extreme west of the Southern Slope, is represented by a thin carbonate formation unconformably overlying Lower Cretaceous rocks. The sections often show unconformities. The Upper Cretaceous is dominated by limestones and marls related to the facies of shallow marine basin. Terrigenous clastics

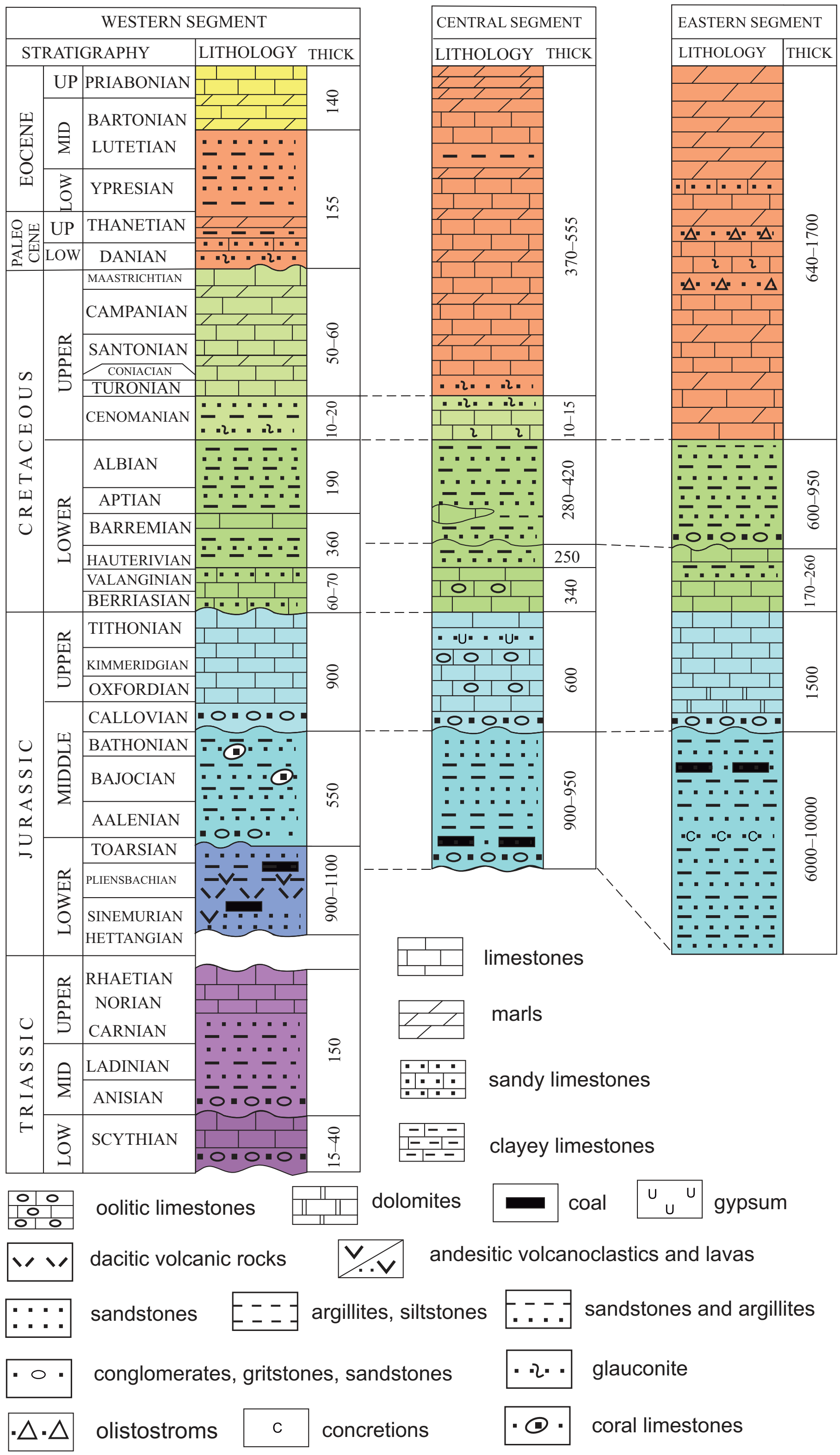


Figure 19. Generalised and simplified lithostratigraphic columns of the Mesozoic–Lower Cenozoic units of the Laba-Malka, Monocline, and Dagestan zones.

are characteristic of the Senomanian (glauconitic sandstones). Glauconite is, generally, characteristic of the Upper Cretaceous–Lower Paleocene deposits (Andruschuk 1968; Khutsishvili 1972).

In the Paleocene–Eocene, shallow marine sedimentation continued across a vast shelf along the southern margin of the Eastern European continent. Upon the shelf, there accumulated deposits represented by marly, argillaceous-marly, and sandy-marly facies with a thickness of some hundreds of metres; dated, predominantly, by foraminifera and echinoids fossils (Andruschuk 1968; Saintot *et al.* 2006).

Mesozoic–Early Cenozoic Development

In the Late Palaeozoic–Early Mesozoic (250–270 Ma), the oceanic basin separating the Africa-Arabian continent from the Tauro-Anatolian-Iranian platformal domain was gradually extending (Stampfli 2000; Babaie *et al.* 2005). However, according to the reconstructions proposed by some authors (for example, Stampfli 2000; Golonka 2004; Robertson *et al.* 2004; Barrier & Vrielynck 2008), only the Central Iranian terrain (CIT) was separated from Gondwana and was displaced northward and collided with the Eurasian continent in the Late Triassic. The Taurus-Anatolian terrains (TAT) separated from Gondwana later, in the Early–Middle Jurassic (c. 180–160 Ma). Neotethys was formed in the Middle–Late Mesozoic (Figure 20). Northward displacement of the Taurus-Anatolian terrains resulted in their gradual migration towards the Pontian-Transcaucasian-Iranian active continental margin, the narrowing of Palaeotethys-Tethys and its transformation into a back-arc basin, and the formation of the suture belt between the TAT and CIT (Barrier & Vrielynck 2008). The suture belt, apparently, is marked by fragments of ophiolite mélange of the Lake Van region.

The TCM during the Mesozoic represented an island-arc-type structure that accumulated mainly shallow-marine carbonate, lagoonal and lacustrine-continental gypsiferous and coal-bearing terrigenous clastics. Suprasubduction extrusive and intrusive activities lasted throughout the entire Mesozoic. Back-arc basins of the Great Caucasus separated the TCM from the southern shelf of the Scythian platform (Adamia 1975; Adamia *et al.* 1977, 1981, 1983; Dercourt *et al.* 1986, 1990).

At the Cretaceous–Paleogene boundary within the study area of the Great Caucasus and Transcaucasus, there existed the following basins (from south to north): shallow water, normal marine island-arc-type, Artvin-Bolnisi (and Loki-Karabagh) basin that accumulated calc-alkaline volcanics, limestones and marls; deeper Achara-Trialeti and Talysh troughs superimposed on the Transcaucasian island arc as a result of the Cretaceous–Eocene rifting (Adamia *et al.* 1974, 1977, 1981; Lordkipanidze *et al.* 1989; Kazmin *et al.* 2000); shallow-water, island-arc type, normal-marine basin of the Georgian massif showing predominantly carbonate sedimentation; the back-arc basin of the Great Caucasus demonstrating sedimentation of mainly deep-marine terrigenous and carbonate turbidites and shallow-marine basin of the pre-Caucasus (Laba-Malka Zone) representing a wide shelf at the southern edge of the Eastern European continent (Figure 21).

The Eocene is especially important as a time of intensification of riftogenesis in the Achara-Trialeti and Talysh basins and further evolution of intraarc deep-marine troughs with accumulation of thick sequences of basaltic volcanics, terrigenous and tuffogenous turbidites (Adamia *et al.* 1974).

Late Cenozoic: Syn to Postcollisional Stage

The Oligocene is traditionally considered as a beginning of syn-collisional (or orogenic) stage of development of the Caucasus (Milanovsky & Khain 1963; Gamkrelidze 1964; Andruschuk 1968; Aslanian 1970; Azizbekov 1972; Saintot *et al.* 2006; Vincent *et al.* 2007; Adamia *et al.* 2008). By this time, the palaeogeographic environment has significantly changed. First of all this became apparent in inversion of the relief: in the place of deep-water basins there were formed mountain ranges of the Great Caucasus, Achara-Trialeti, Talysh, and Lesser Caucasus (Somkhiti, Bazum, Halib, Murguz, Pambak, Shahdag, and Mrovdag ranges). The domains of shallow-marine basins of platformal type of pre-collisional stage (Scythian platform, Georgian massif, Artvin-Bolnisi massif, Caucasian parts of the Taurus-Anatolian and Central Iranian

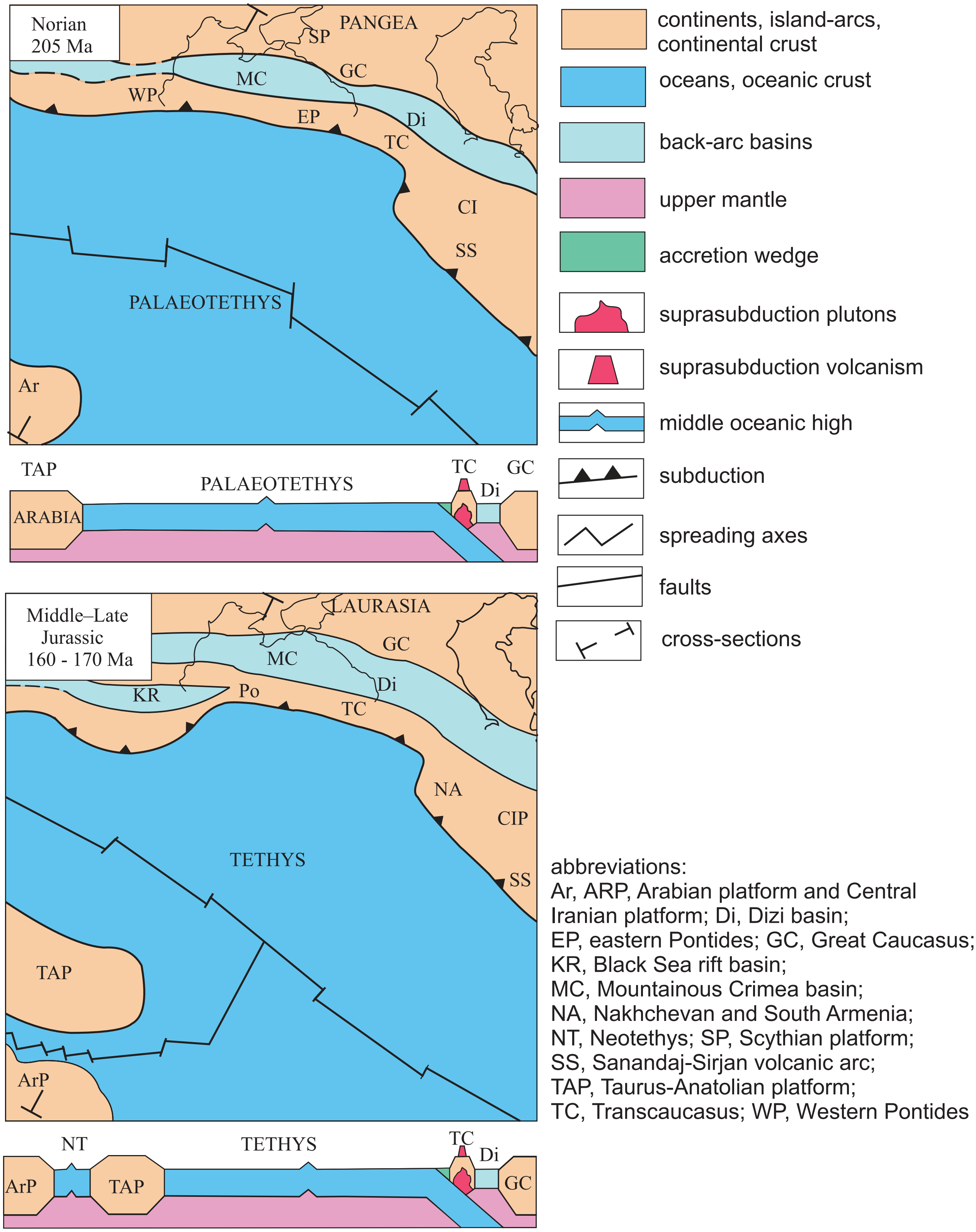


Figure 20. Norian and Late Jurassic tectonic reconstructions of the Black Sea-Caspian Sea region.

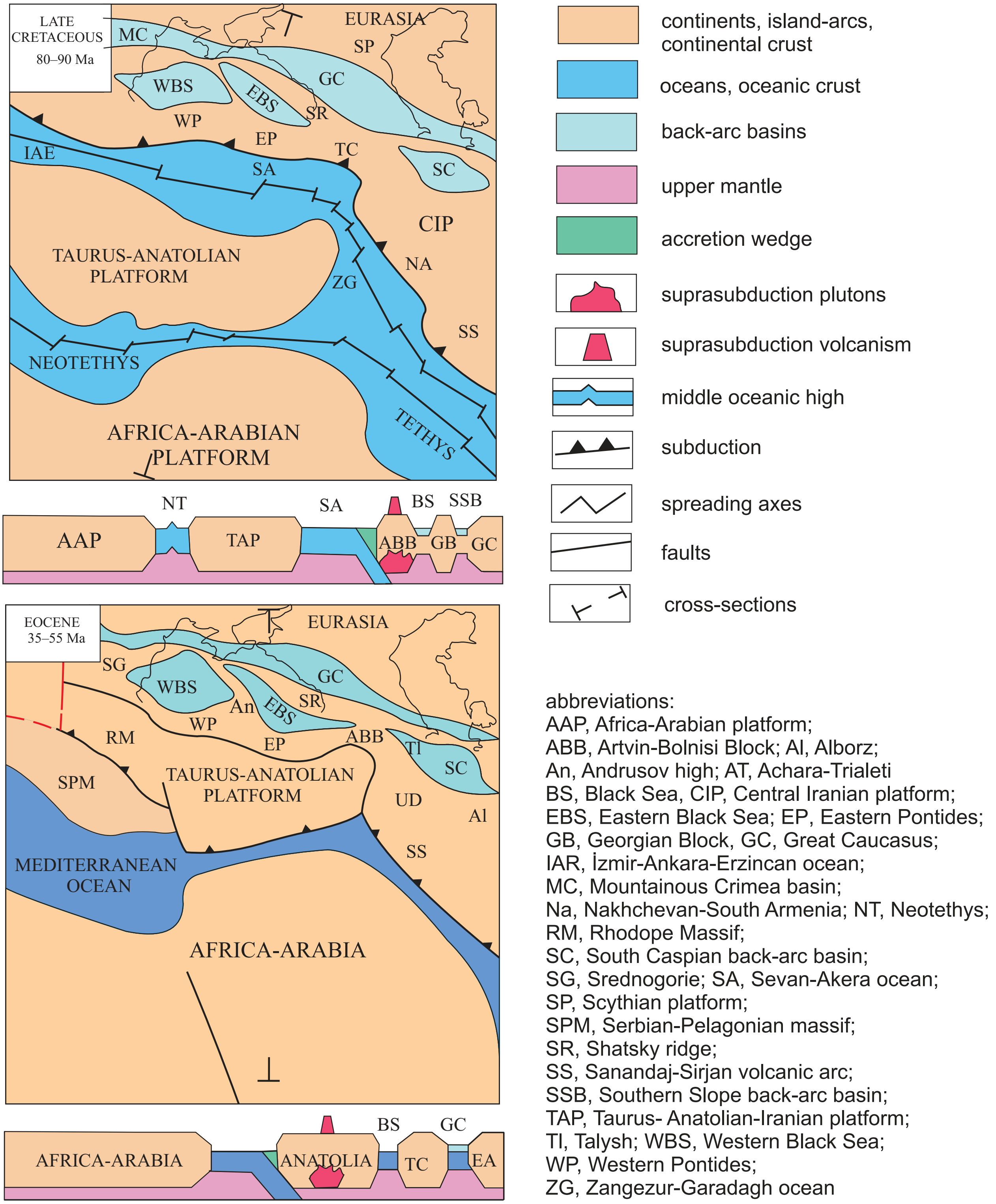


Figure 21. Late Cretaceous and Eocene tectonic reconstructions of the Black Sea-Caspian Sea region.

platforms) sank forming foredeep and intermontane depressions that accumulated molasse deposits – the products of denudation of arising mountain ranges (Gamkrelidze 1964; Andruschuk 1968; Azizbekov 1972; Geology of the USSR 1977).

Oligocene–Lower Miocene deposits resulted from accumulation in semi-closed euxinic basins of the Paratethys at the most part of the territory of the Caucasus are represented by uniform argillaceous-sandy gypsiferous facies termed the Maykopian series. The main characteristic features of these rocks are their brown colour, lack of carbonate content, scarcity of fossils (except for fish scales), abundance of jarosite on bedding planes, and intralayers and veinlets of gypsum. Disk-shaped concretions-septaria of spherosiderite, up to 0.5–3.0 m in diameter are very frequent, and they, as a rule, are located along the bedding planes; locally thin beds and lenses of hematite are observed. The Oligocene only locally is represented by other facies: siliceous (Dzirula massif), normal-marine terrigenous (Western Great Caucasus) continental coal-bearing, (Akhaltsikhe basin) volcano-sedimentary, normal-marine with corals and lagoonal multicolour terrigenous clastics (Southern Armenia, Nakhchevan).

The accumulation of molasse (maximum thickness several km), predominantly shallow-sea and lagoon-lacustrine terrigenous clastics with layers of organogenic carbonate rocks (coquina), lasted almost throughout the Miocene. Only at the end of the Miocene (Tortonian, c. 7 Ma), were shallow-sea environments replaced by subaerial surroundings and simultaneously clastic material became coarse. Marine environments of sedimentation remained only within the territories adjacent to the Black, Azov, and Caspian Sea basins.

Presented geological timescale (Figure 22) showing correlation of the Oligocene–Quaternary stages of the Caucasus with those of the Mediterranean area is based on data of longterm investigations.

Starting from the Late Miocene (c. 9–7 Ma) and as far as the end of the Pleistocene, in the central part of the region, occurred volcanic eruptions in subaerial conditions going on simultaneously with the formation of molasse troughs and the accumulation of coarse molasse.

PERIOD	EPOCH SERIES	AGE STAGE: MEDITER-RANIEN	BLACK SEA	CASPIAN SEA	AGES MA SIC
QUATERNARY	HOLOCENE Q_2				0.01
	PLEISTOCENE Q_1		NEOEUXINIAN	KVALINIAN	0.08
			KARANGATIAN	GIRKAN	0.41
			UZUNLARIAN	KHAZARIAN	0.70
			CHAUDIAN	BAKUNIAN	1.0
			GURIAN	APSHERONIAN	1.80
N E O G E N E	PLIOCENE N_2	GELASIAN	EGRISSIAN	AKCHAGILIAN	2.58
		PIACHENZIAN	KUYALNIKIAN		3.0
			KIMMERIAN		3.60
		ZANCLEAN			5.33
	MIOCENE N_1	MESSINIAN	P O N T I A N		7.11
		TORTONIAN	M E O T I A N		
			S A R M A T I A N		11.0
		SERRAVALIAN	K O N K I A N		
			K A R A G A N I A N		13.6
		LANCHIAN	C H O K R A K I A N		
			T A R K H A N I A N		16.4
		BURDIGALIAN	KOZAKHURIAN		
			SAKARAULIAN		19.1
		AQUITANIAN	UPHLISTSIKHEAN		23.8
P A L E O G E N E	OLIGOCENE E_3	C H A T T I A N			28.0
		R U P E L I A N	K H A D U M I A N		33.7

Figure 22. Correlation of the Upper Miocene–Pleistocene stratigraphic schemes of the Mediterranean and Black Sea-Caspian Sea regions (Haq & Van Eisinga 1987; Remane & Four-Muret 2006; Adamia *et al.* 2008).

Aras Basin

In the Aras basin, the Oligocene is represented by two facies accumulated in shallow-marine environment (Figure 23). The first facial type sandy-argillaceous fine- and coarse-grained terrigenous clastics (thickness ~1000 m) with coral limestones are exposed mainly in the W–NW part of the basin. The second facial type of the Oligocene, exposed mainly in the eastern part of the Aras basin, is represented by subalkali-alkali-basalt-andesitic and dacite rhyolitic lavas and volcanoclastic rocks (thickness ~1000 m). The Oligocene age of these formations was established using molluscs, foraminifera, and corals (Aslanian 1970; Azizbekov 1972). Lower Miocene deposits of the Aras basin, represented by lagoonal, salt-bearing multicolour sandy-argillaceous fine-, medium- and coarse-grained terrigenous clastics (thickness ~400 m), rest unconformably upon older (Oligocene and older) formations. Middle and Upper Miocene sediments in the Lesser Caucasian part of the Taurus-Anatolian-Central Iranian Platform (South Armenia and Nakchevan) are widespread mainly within the Aras basin and are represented by lagoonal gypsiferous-salt-bearing terrigenous clastics and shallow-marine terrigenous and carbonate rocks dated by molluscs, foraminifera, ostracods, corals, and plant fossils; thickness ~1000–1300 m (Aslanian 1970; Azizbekov 1972).

Upper Miocene and Pliocene sediments within the Aras basin (Sarmatian, Pontian, and Meotian stages) are represented by shallow marine, lagoonal and continental, mainly, fine-grained terrigenous molasse, also by carbonate and gypsiferous deposits (thickness ~1000 m). The thickness of the Oligocene–Quaternary sediments in easternmost part of the Kura basin, at the south Caspian seashore reaches ~2 km (Brunet *et al.* 2003). Their biostratigraphic subdivision is based on the data of marine molluscs and foraminifera.

Transcaucasian Basins

The formation of the Transcaucasian foreland started in the Late Pliocene; it was divided by the Dzirula massif that resulted in formation of the Rioni basin (-Black Sea) in the west and the Kura basin (-Caspian Sea) in the east (see Figures 2, 23 & 24).

Rioni-Kura Basin- The Oligocene–Lower Miocene (Maykopian series) is represented mainly by alternation of gypsiferous clays with sandstones (Gamkrelidze 1964; Andruschuk 1968; Azizbekov 1972).

Maykopian deposits in some localities conformably follow Late Eocene sandy-clayey sediments. Their maximal thickness is 2.5–3.0 km. In the ascending section one can observe the gradual transformation from more sandy varieties to more clayey rocks in the middle part and again more sandy rocks in the top of the section (Uplistsikhe, Sakaraulo, and Kotsakhuri stages).

Middle Miocene (Tarkhanian, Chokrakian, Karaganian, and Konkian Stages): Late Miocene (Early and Middle Sarmatian Stages)– At this time, a sublatitudinal marine basin was developing within the area between the Southern Slope of the Great Caucasus in the north and the Achara-Trialeti belt in the south. The maximal thickness of the Middle–Upper Miocene sediments achieves 1500–3500 m. In the axial parts of these subsided structures, deep-marine carbonate clays and sandstones with marls were accumulated; in their marginal zones, shallow-marine terrigenous-carbonate rocks of limited thickness were deposited (Gamkrelidze 1964; Azizbekov 1972; Eastern Paratethyan 1985).

The abundance of various marine fossils (molluscs, foraminifera etc) observed from the outset of the Middle Miocene (Tarkhanian stage) indicates the establishment of wide connections of the basin with the global ocean and the replacement of euxinic environments by normal marine conditions. However, later on, the connections with oceanic basins periodically either ceased (Chokrakian, Karaganian stage, Middle–Sarmatian times) or were impeded (the end of the Karaganian, the onset and the end of the Konkian; the end of the Middle Sarmatian) (Maisuradze 1971; Popkhadze 1983; Eastern Paratethyan 1985).

In the Late Sarmatian, there began the transformation of nearly all of the Caucasus into a dry land except for some small areas. The continuing development resulted in the further uplift of clastic source areas and subsidence of subaerial-sedimentation ones dominated by coarse molasses with subdued sandstones and clays. According

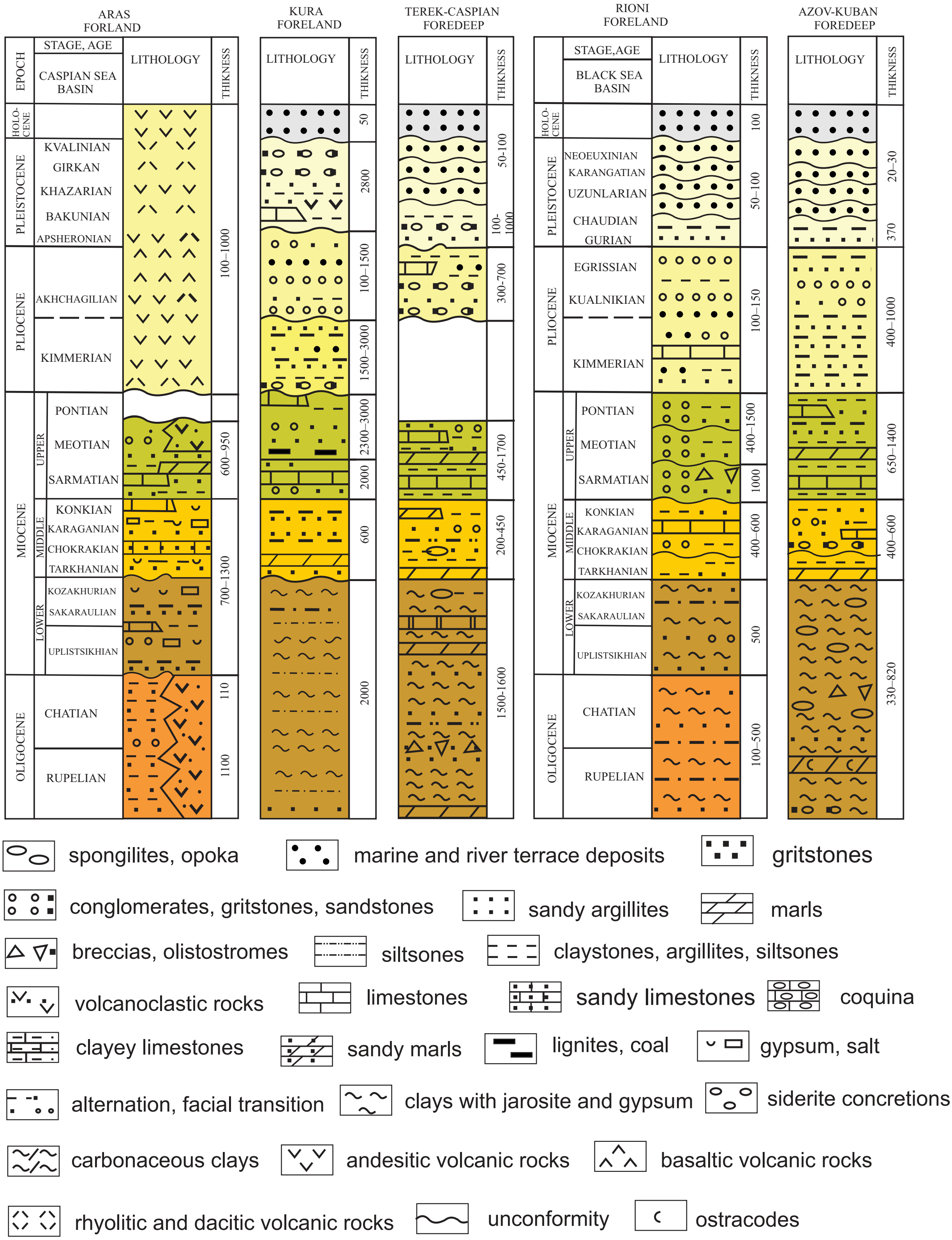
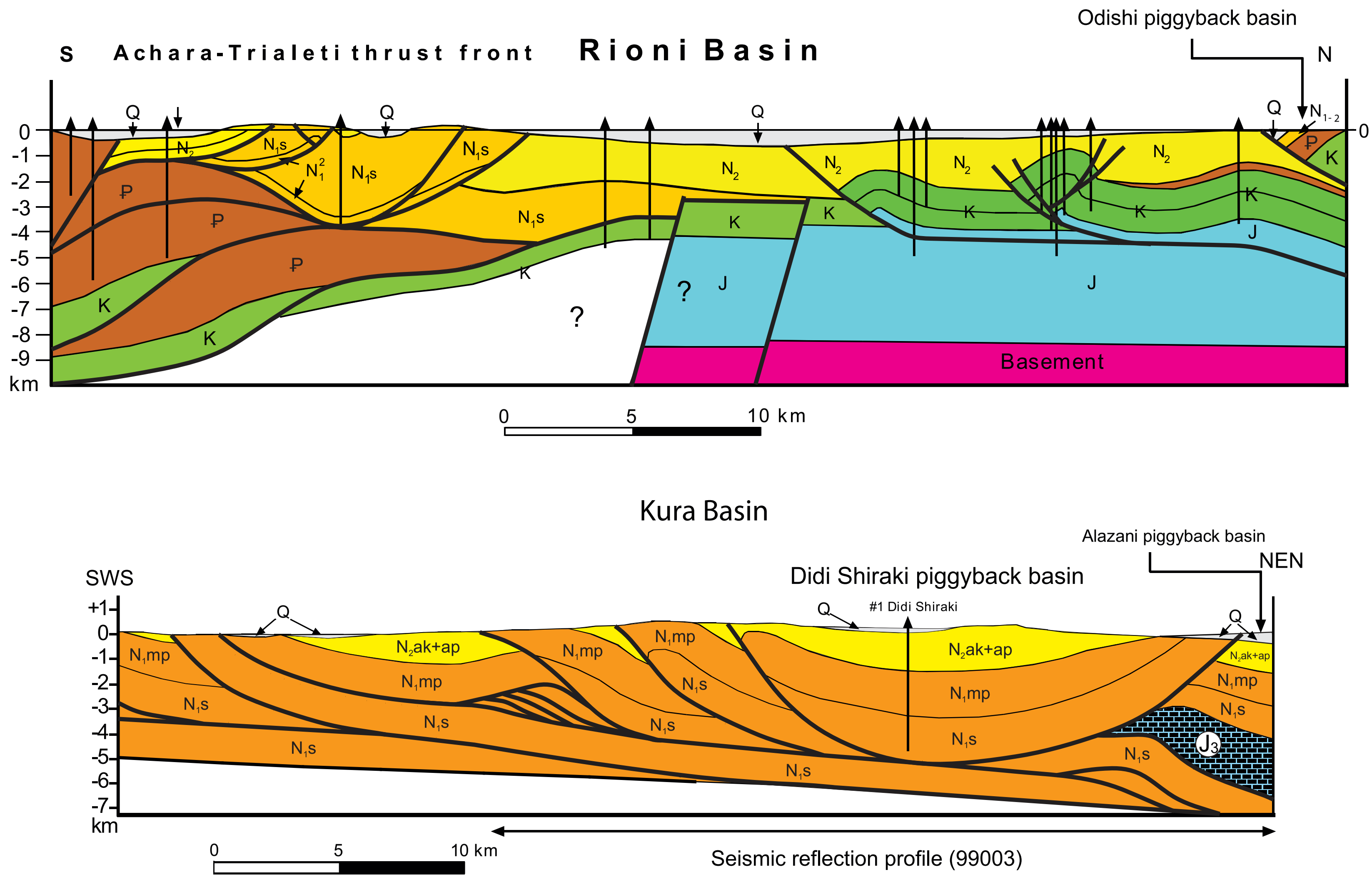


Figure 23. Generalised and simplified lithostratigraphic columns of the syn- and post-collisional units of the Caucasus.



J, Jurassic; J_3, Upper Jurassic; K, Cretaceous; P, Paleogene; $N_{1\text{-}2}$, Lower–Middle Miocene; N_1^2, Middle Miocene; N_1S, Upper Miocene-Sarmatian rs; N_1mp, Upper Miocene, Meotian, Pontian rs; N_2, Pliocene; N_2ak+ap, Upper Pliocene-Akchagylian rs and Lower Pleistocene-Apsheronian rs; Q, Quaternary. Red lines – faults; black vertical lines – wells

Figure 24. Cross-sections across Rioni and Kura basins (compiled by V. Alania).

to the composition of clastic material, one can easily recognize the northern (mainly Cretaceous–Paleogene flysch of the Great Caucasus) and southern and southwestern source areas (Mesozoic–Cenozoic volcanic rocks of the Lesser Caucasus).

In the central and southern parts of the Kura depression, the Upper Sarmatian (thickness ~200 m) is represented by variegated clays, cross-bedded sandstones, gritstones, containing pebbles of metamorphic, and crystalline rocks of the Lesser Caucasus. Upper Sarmatian rocks are highly gypsiferous. The rocks show no marine fauna and their age is determined on the basis of mammal relicts (Eastern Paratethyan 1985; Chkhikvadze *et al.* 2000).

Marine Meotian rocks are predominantly represented by coastal shallow water deposits. They transgressively overlap Sarmatian and older, Cenozoic and Mesozoic, rocks and, in turn, are overlain by Pontian and younger beds. Thickness of the Meotian and Pontian sediments varies from a few to several hundred meters sometimes achieving even 1500–2000 m. The most complete sections include rich and variable assemblages of molluscs, foraminifera, ostracods, and sponge spicules (Gamkrelidze 1964; Azizbekov 1972; Popkhadze 1975; Eastern Paratethyan 1985; Maisuradze 1988; Jones & Simmons 1997; Adamia *et al.* 2008, 2010).

The Pontian is regressive, its individuality is manifested by molluscs including not only forms common for the Black Sea realm, but also species, frequent in Dacian and Pannonian basins (Gamkrelidze 1964; Chelidze 1973; Eastern Paratethyan 1985). *Upper Pontian section* is represented by shallow water and relatively deep water facies. Meotian and Pontian ages of the continental formations have been established on the basis of mammal fossils (Chkhikvadze *et al.* 2000) and fossil tortoises (Gabashvili *et al.* 2000).

Pliocene (*Kimmerian Stage)*– Like the Pontian stage, the Kimmerian one has much in common with other coeval formations of the Black Sea area, although showing some individual features. The Kimmerian stage consists mainly of sandstones, claystones, and conglomerates (thickness ~200–300 m). Mollusc assemblage includes several endemic forms evidencing decreasing salinity in several parts of the Rioni Basin (Gamkrelidze 1964; Eastern Paratethyan 1985; Taktakishvili 1984, 2000).

Rioni Basin– Kuyalnikian stage (terrigenous clastics, thickness ~200 m) is characterized by complete profiles and richer brackish mollusc assemblages. Recent investigations have shown the greater presence of the western Georgian Kuyalnikian stage as compared to the other parts of the euxinian basin due to the presence of transitional, Kimmerian-Kuyalnikian beds that are almost completely absent in other parts of the Black Sea area (Gamkrelidze 1964; Taktakishvili 1984, 2000; Eastern Paratethyan 1985).

Pleistocene and Holocene: (Gurian, Chaudian, Uzunlarian, Karangatian, and Neoeuxinian Stages)

Rocks of the *Gurian (Eopleistocene) stage* crop out only in Guria (south-western part of the Rioni basin), although, it was penetrated by boreholes in other areas of the Rioni basin. These are sandy clayey rocks (thickness ~100–500 m) with homogenous but poor mollusc fauna (Gamkrelidze 1964; Taktakishvili 1984, 2000; Eastern Paratethyan 1985).

Chaudian, Uzunlarian, Karangatian (Pleistocene) and Neoeuxian (Holocene) marine deposits of the Rioni foreland are spread only within the areas adjacent to the Eastern Black Sea. They are mainly represented by sandy-argillaceous fine-grained clastics alternating with coarse-grained and carbonate rocks (according to well data). These deposits achieve their maximal thickness of ~600–800 m in the central part of the foreland. Fossil molluscs and ostracods indicate Middle–Late Quaternary age of these deposits (Gamkrelidze 1964; Kitovani *et al.* 1991). The age of Neoeuxinian marine clays and sands alternating with peat beds and containing brackish molluscs is equal to 31 300 years (Maisuradze & Janelidze 1987).

Kura Basin– Middle Pliocene productive series (Kimmerian) exposed in the easternmost part of the Kura (Caspian) basin and located mainly within onshore and offshore zones of Azerbaijan contain significant accumulations of oil and gas in fluvial-deltaic and lacustrine sediments. The maximal thickness of the productive series in the Kura basin reaches ~3500–4500 m (Gamkrelidze 1964; Azizbekov 1972; Abrams & Narimanov 1997; Devlin *et al.* 1999).

Akchagylian and Apsheronian stages are represented by continental and shallow marine molasse unconformably overlying older rocks (maximal thickness ~1500 m). Marine facies of the lower part of the section is made up of sandy-clayey rocks; it's upper part – of sandstones and conglomerates (Gamkrelidze 1964; Azizbekov 1972). Fossil fauna are represented by endemic brackish molluscs of Caspian type, and also by ostracods and foraminifera. Along with marine molluscs, there were found relicts of fossil mammals (Eastern Paratethyan 1985; Adamia *et al.* 2010).

Continental facies of the Akchagylian and Apsheronian stages are dominated by conglomerates with insignificant sandstones and sandy clays. Thin layers of volcanic ash are locally present (Gamkrelidze 1964; Azizbekov 1972).

Pleistocene and Holocene (Bakunian, Khazarian, Girkan, and Khvalinian Stages)– These deposits attributed to the Caspian Sea domain are represented by shallow marine fine-, medium-, and coarse-grained terrigenous facies (clay, sand, sandstone, gravelstone, conglomerate, volcanic ash), their thickness varies from ~150 up to 1000 m. The amount of coarse-grained clastics grows from the axial zone of the foreland towards its borders and marine facies are replaced by continental ones. Marine fossils are represented mainly by molluscs. Continental deposits of the Pleistocene–Holocene stages of the eastern part of the Kura foreland are dated mainly by remnants of mammalia (Gamkrelidze 1964; Azizbekov 1972; Maisuradze & Janelidze 1987; Kitovani *et al.* 1991).

Dzirula High– Within the Dzirula salient, the Oligocene–Lower Miocene is represented by a thin (200–375 m) manganese-bearing formation of spongilitic sandstones and clays with opoka and chalcedony. The formation unconformably rests on Cretaceous deposits (Gamkrelidze 1964).

The Middle Miocene, starting with the Tarkhanian stage, conformably follows the Lower Miocene and is represented by sandstones, clays, and marls (thickness 1–30 m). The Chokrakian deposits unconformably overlie older sediments. The Chokrakian stage is followed by Karaganian and Konkian ones having aggregate thickness 20–30 m. They are represented by shallow-sea, often variegated, sandstones, gravelites, clays, sandy, and oolitic limestones. Their Middle Miocene age was, generally, established using mollusc fossils (Gamkrelidze 1964; Ananiashvili & Sakhelashvili 1984; Eastern Paratethyan 1985) and foraminifera (Janelidze 1970, 1977).

Akhaltsikhe Basin– The Oligocene section of the Akhaltsikhe depression (the southern part of the central Achara-Trialeti basin) is complete and well-dated by fauna (mainly by marine molluscs). The Oligocene deposits of the Akhaltsikhe depression are subdivided into several suites (Yılmaz *et al.* 2001). The lowermost suite is represented by calcareous clays and sandy clays comprising interlayers and wedges of siltstones and marls. Lower Oligocene mollusc assemblages occur at all levels of the suite (Tatishvili 1965; Kazakhashvili 1984). Microforaminifera of Early Oligocene age are also met (Kacharava 1977) as well as fish fossils. The thickness of the suite is 600–700 m. Fish fossils, microforaminifera, and molluscs of Oligocene age were found in the upper part of the sandstones. The Lower Oligocene age of the suite was also proved by microforaminifera and fish fossils. The following suite is built up of clays, siltsones, conglomerates, and sandstones containing brackish molluscs (Kazakhashvili 1984).

This is followed by a coal-bearing suite up to 150 m thick represented by clays with interlayers and succession of sandstones, siltstones, and lignite with mollusc fossils. These deposits were formed under conditions of tropical-to-subtropical climate in the lagoon surrounded by swampy forests. The following suite is dominated by thick-bedded sandstones with fossil molluscs (Kazakhashvili 1984). The thickness is 25–50 m. The uppermost suite is built up of reddish-grey, grey and greenish-grey clays with sandstone beds. The thickness is 400–450 m. The sandstone succession includes terrestrial vertebrates (Gabunia 1964).

Great Caucasus– Syn- and post-collisional molasse formations exposed at the western and eastern periclines of the Great Caucasian anticlinorium are represented by facies analogous with those of the Transcaucasus foreland and pre-Caucasus foredeeps (Andruschuk 1968; Azizbekov 1972; 1:1 000 000 scale geological map (L36, 37) of the USSR, 1983; 1:1 000 000 scale geological map of Azerbaijan 1971). Outcrops of the Upper Sarmatian shallow-marine deposits are reported in the Shakhdag Mountains, on the crest of the Great Caucasus at about 3550 m above sea (Budagov 1964).

Scythian Platform– The pre-Caucasian part of the Scythian platform is almost completely covered by thick layer of Oligocene–Neogene and Quaternary molasse deposits that reach their maximal thickness in the axial zones of the Azov (Indol)-Kuban (~ 4400 m), Terek-Caspian (~4700 m), and Gussar-Devichi (~6000 m) foredeeps, while their minimal thickness is reported in Stavropol arch (~1600 m) separating the Azov-Kuban foredeep from the Terek-Caspian one (Milanovsky & Khain 1963; Andruschuk 1968; Azizbekov 1972; Ajgirei 1976; Ershov *et al.* 2003).

The Oligocene–Lower Miocene Maykopian series outcrops along the southern margin of these foredeeps (see Figures 2 & 23) and is termed after the Maykop city located in the westernmost part of the pre-Caucasus. Details of the Maykopian stratigraphy of the region have been discussed in various publications (see Milanovsky & Khain 1963; Andruschuk 1968; Jones & Simmons 1997; Mikhailov *et al.* 1999). The Oligocene–Early Miocene age of the series is chiefly based on foraminifera, nannoplankton, ostracods, and molluscs. The thickness of the formation varies from 800 to 1500 m in axial zones of the foredeeps up to some tens of hundreds metres at their borders (Andruschuk 1968; Azizbekov 1972).

Middle Miocene (Tarkhanian, Chokrakian, Karaganian and Konkian stages), and Upper Miocene (Sarmatian, Pontian, and Meotian stages) deposits of the pre-Caucasus are represented mainly by fine- and medium-grained sandy-argillaceous clastics with subordinate coarse-grained rocks, limestones, and marls. In some localities, dolomites, coquina, bioherms of bryozoan limestones, gypsiferous and bituminous terrigenous clastics are found. The thickness of these deposits, which are mainly shallow-marine and also lagoonal-continental, varies from some hundreds of metres within the periphery of the foredeeps up to 1000–2000 m in their axial parts (Andruschuk 1968; Azizbekov 1972).

Pliocene, Pleistocene and Holocene– Two foredeeps (Azov-Kuban and Terek-Caspian–Gussar-Devichi) separated by the Stavropol high were continuously developing in shallow-marine, lagoon-lacustrine continental environments during the Pliocene and Early Pleistocene.

Kimmerian, Kuyanlikian, and Gurian (Tamanian) stages (Western pre-Caucasus) and Kimmerian, Akchagylian, and Apsheronian stages in the Eastern pre-Caucasus are represented mainly by sandy-argillaceous clastics, marls, coquina, dolomites, brown iron ore, oolitic iron ore, and rare conglomerates. The age of marine deposits is determined by molluscs, ostracods; continental deposits were dated by mammal fossils. The maximum thickness of Pliocene–Eo-Pleistocene deposits within the axial zone of the foredeeps varies from 1100 m (western pre-Caucasus) to 800 m, (eastern pre-Caucasus) and 2700 m in the Gussar-Devichi foredeep (Andruschuk 1968; Azizbekov 1972).

During the Late Pleistocene and Holocene, division of the pre-Caucasus into the Black Sea and Caspian Sea domains remained. Marine deposits are found in the immediate proximity to the seashore. Within the Black Sea domain, the Late Pliocene–Holocene is subdivided into the Chaudian, Uzunlarian, Karangatian, and Neoeuxinian stages, while the Caspian Sea domain – into the Bakunian, Khazarian, Girkan, and Khvalynian stages. They contain sands, clays, conglomerates, limestones with molluscs of Mediterranean type. The Late Pleistocene–Holocene contains marine molluscs of Apsheronian type. This time, most part of the pre-Caucasus is represented by land (Andruschuk 1968; Azizbekov 1972; Avanessian *et al.* 2000).

Late Cenozoic Syn- and Post-collisional Magmatic Formations

Late Cenozoic syn- and post-collisional intrusive and extrusive formations are widespread in the Black Sea-Caspian Sea continent-continent collision zone. Oligocene, Neogene, and Quaternary ages of these formations are reliably dated on the basis of geomorphological, structural, biostratigraphic, geochronological, and magnitostratigraphic data (Adamia *et al.* 1961; Milanovsky & Khan 1963; Gamkrelidze 1964; Andruschuk 1968; Arakelyants *et al.* 1968; Aslanian 1970; Azizbekov 1972; Rubinstein *et al.* 1972; Vekua *et al.* 1977; Khaburzania *et al.* 1979; Maisuradze *et al.* 1980; Aslanian *et al.* 1984).

The Upper Cenozoic calc-alkali to shoshonitic volcanic belt runs from Turkey via Caucasus into

Iran. Outcrops of these magmatic rocks are exposed along the boundaries of the main tectonic units (terranes) of the region. In Turkey, they construct two branches. The northern branch roughly coincides with the İzmir-Ankara-Erzincan suture (Tethys); the southern branch – with the Antalya (Pamphylian)-Biltis (Southeast Anatolia) suture (Neotethys). In the Lesser Caucasus, syn/post-collisional magmatic rocks crop out along the Sevan-Akera ophiolite suture, and in the northernmost Iran, along the Karadagh-Rascht-Mashhad suture (Tethys), forming the Alborz magmatic belt. The southern magmatic branch of Turkey extends into Iran along the Zagros suture (Neotethys) and forms the Urmieh-Dokhtar magmatic arc (Figure 25).

All the above-mentioned sublatitudinal branches of syn/post-collisional magmatic formations are gathered in the region surrounding Lakes Van and Urmieh (Van triangle or Van knot). From here, the submeridional volcanic branch extended northward forming the East Anatolian, Armenia-Azerbaijan and South Georgian volcanic highlands and chains of extinct volcanoes of the Lesser Caucasus-Transcaucasus. The northernmost relatively short sublatitudinal (WNW–ESE) branch of syn/post-collisional magmatic formations located in the central segment of the Great Caucasus is connected to the boundary zone between the Great Caucasus fold-and-thrust mountain belt and the Scythian platform.

More or less intensive manifestations of syn/post-collisional magmatic activity are found within all tectonic units of the Caucasus, however, the most intensive magmatic occurrences show up within the rigid platformal units: (1) in the Lesser Caucasian part of the Taurus-Anatolian-Iranian platform

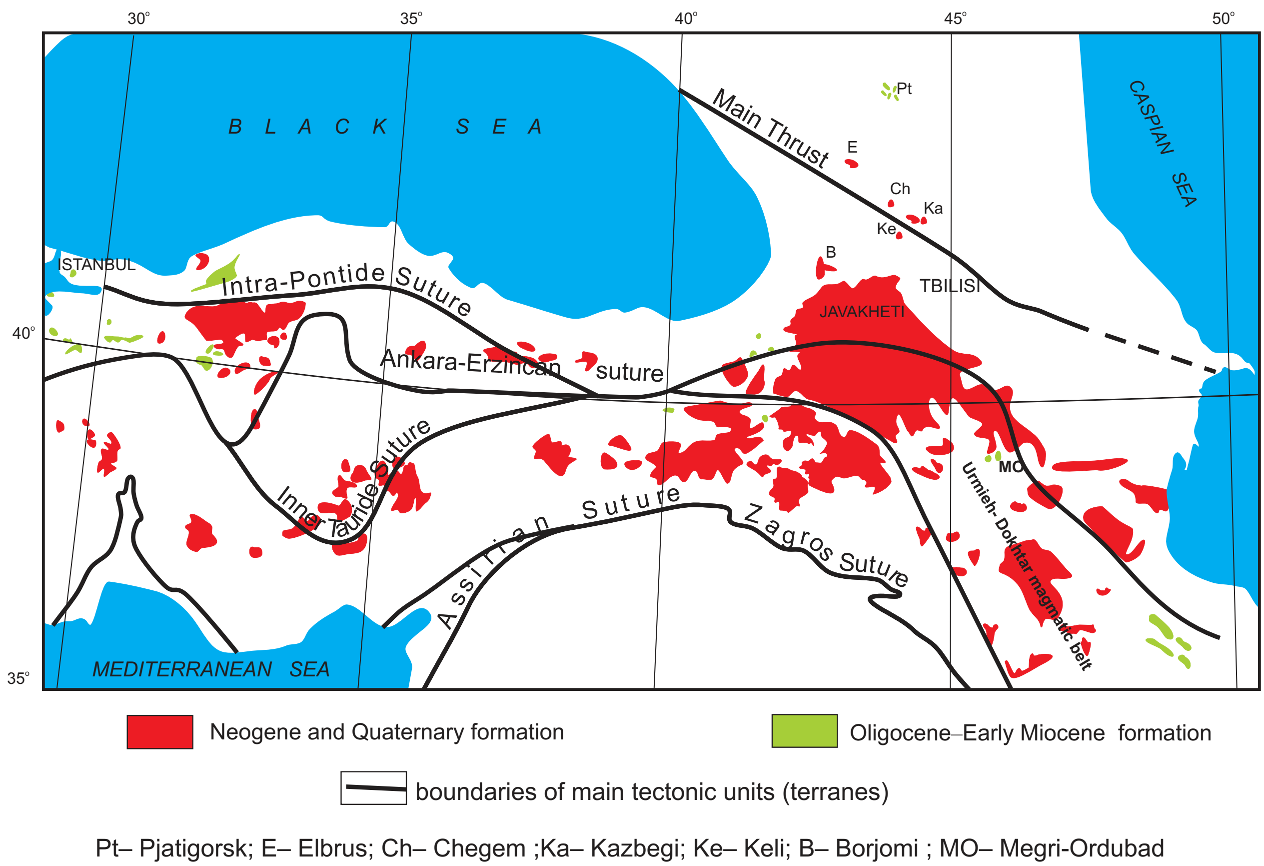


Figure 25. Late Cenozoic syn- and post-collisional intrusive and extrusive formations in the Black Sea-Caspian Sea continent-continent collision zone (modified from Okay 2000; Robertson 2000; Altıner *et al.* 2000).

(TAIP and Aras foreland), and (2) in the Artvin-Bolnisi rigid unit of the Transcaucasian massif – TCM (see Figure 2).

Three main stages of syn/postcollisional magmatism are distinct: Oligocene–Miocene, Miocene–Pliocene and Quaternary.

Oligocene-Miocene Magmatism– The southernmost part of the Lesser Caucasus and the boundary zone between the Great Caucasus and pre-Caucasus show evidence of syn-collisional intrusive activity. Oligocene–Lower Miocene intrusions (32–17 Ma) are widespread within the Aras foreland and Taurus-Anatolian-Iranian platform, along the border with the Lesser Caucasian (Sevan-Akera-Zangezos-Karadagh) ophiolite belt, as well as in the southernmost tectonic units of the TCM (Karabagh, Talysh). Intrusive bodies are composed of the following groups: gabbro, monzonite, syenite, diorite and granite (Megri-Ordubad – left bank of the river Araks, Tutkhun and other plutons and dykes; Aslanian 1970; Azizbekov 1972; Aslanian *et al.* 1984; Rustamov 1983, 2007; Nazarova & Tkhostov 2007; Azadaliev & Kerimov 2007).

In the southernmost part of the Scythian platform, at its border with the Great Caucasus, there occur the Beshtau (Pjatigorsk) group of the Oligocene–Lower Miocene alkali intrusive bodies of granite-porphyry, granosyenite-porphyry, and quartz syenite-porphyry composition that are intruded in Oligocene–Lower Miocene sediments and their redeposited material is found in Upper Miocene (Akchagylian) deposits (Andruschuk 1968). However, K-Ar data (c. 8–9 Ma) indicate a Late Miocene age of these intrusions (Borsuk *et al.* 1989).

Late Miocene–Quaternary Magmatism– Late Miocene–Quaternary volcanic activity in the region took place within a broad S–N-trending belt extending from Central Anatolia to the Great Caucasus-pre-Caucasus. The belt is related to transverse Van-Transcaucasian uplift (Lordkipanidze *et al.* 1989). In some localities, volcanic products exceed a kilometre in thickness and cover a wide compositional spectrum from basalt to high-silicic rhyolite. Two main stages of volcanic activity are present: (1) Late Miocene–Early Pliocene and (2) Pliocene–Quaternary. Lavas predominate over volcanoclastics, especially during the second phase of eruptions. According to their mineral-chemical compositions, rocks of the both stages are attributed to calc-alkaline, alkaline and subalkaline series. Data on absolute age demonstrate that the first stage of eruption in the Caucasus happened ~13–4.5 Ma while the second one c. 3.5–0.01 Ma (Rubinstein *et al.* 1972; Maisuradze *et al.* 1980; Camps *et al.* 1996; Ferring *et al.* 1996; Mitchel & Westaway 1999; Lebedev *et al.* 2004, 2008).

Upper Miocene–Lower Pliocene formations are known only in the Lesser Caucasus-Transcaucasus where they are represented by basalt-andesite-dacite-rhyolitic subaerial lava sheets and volcanoclastics. Basaltic lavas and pyroclastic rocks are represented in the lower, basal level of the formation. In some places, the formation contains economic diatomite deposits. The middle part of the section is represented mainly by volcanoclastic rocks. Pyroclastic rocks in the vicinity of the Goderdzi pass (Artvin-Bolnisi massif-ABM) contain remains of petrified subtropical wood, which date the rocks as Upper Miocene–Pliocene (Uznadze 1946, 1951; Gamkrelidze 1964; Uznadze & Tsagareli 1979). K-Ar dating of tuffs indicate a Late Miocene age (9.8 Ma, Aslanian *et al.* 1984).

Laminated and/or banded andesite and dacite lavas with volcanoclastic interlayers are common in the upper part of the formation. Andesite is a dominant rock unit. K-Ar ages of the andesites and dacites according to Aslanian *et al.* (1984), Lebedev *et al.* (2004) vary from 9.4 Ma to 7.0 Ma. The K-Ar age of calc-alkaline, sub-alkaline and alkaline basaltic, andesitic, dacitic, and rhyolitic volcanic rocks of the TAIP (Armenia) yielded c. 13–3.5 Ma age interval (Aslanian *et al.* 1984; Jrbashian *et al.* 2002).

Upper Pliocene–Holocene formations are widespread within the TAIP and ABM. Basaltic (doleritic) lavas are the dominant rock units in the lower part of the formation. In some places, they contain lenses of fluviatile to lacustrine and alluvial deposits, and also pyroclastic rocks; andesitic basalts are subordinate, more felsic rocks are rare. Because of its low viscosity, lava could spread over a large area. It covered an ancient relief forming an extensive flat plateau. The total thickness of the formation is approximately 100–300 m. The age of the lower part of the formations is identified by mammalia fauna as Late Pliocene–Pleistocene (Gamkrelidze 1964; Andruschuk 1968; Azizbekov 1972; Adamia *et al.*

1961; Gabunia *et al.* 1999; La Georgie 2002; Vekua *et al.* 2002).

Magnetostratigraphic investigations carried out during the last decade (Vekua *et al.* 1977; Khaburzania *et al.* 1979; Djaparidze *et al.* 1989; Sholpo *et al.* 1998) have recognized within the Upper Miocene–Quaternary sequence of volcanic rocks all the known standard palaeomagnetic chrons and subchrons: Brunhes, Matuyama (with Jaramillo, Cobb Mountain, Olduvai, Reunion 2 and Reunion 1 subchrons), and Gauss (with Kaena and Mammoth subchrons). These data place basalts of the ABM at the top of the Akchagylian stage (c. 1.8 Ma).

Andesites, andesite-dacites, and dacites crown the section of the Lower Pliocene–Quaternary volcanic formations of the TAIP and ABM. According to K-Ar dating, in Javakheti, the oldest rocks of dacitic composition are lavas (c. 760 000 a).

The younger rocks (400 000–170 000 yr) are found in the central part of the Javakheti highland (Abul mountain, lake Paravani, caldera Samsari etc). Volcanic activity in the region came to a halt, probably, at the end of the Pleistocene, about 30 000 a (Lebedev *et al.* 2004) – Holocene (Afanesian *et al.* 2000).

The central, mostly uplifted segments of the Great Caucasus contain only Pliocene–Quaternary volcanic and plutonic formations. Products of post-collisional volcanism of the Elbrus, Chegem, and Keli-Kazbegi centers of extinct volcanoes are represented mostly by lava flows of calc-alkaline-subalkaline andesite-basalt, andesite-dacite rhyolite composition (Tutberidze 2004; Koronovsky & Demina 2007). Neogene–Quaternary intrusives of the Great Caucasus also crop out in the same regions and are represented by hypabyssal bodies. A number of geochronological data indicate a Late Pliocene–Quaternary age of this volcanic-plutonic formation (Borsuk 1979; Chernishev *et al.* 2000). Two radiocarbon data (5950±90 a and 6290±90 a) were obtained from wood fragments collected from the lake beds near the volcano Kazbegi. These data indicate that the lake beds and lava flow may be attributed to Middle Holocene age (Djanelidze *et al.* 1982).

In the geological literature, migration of magmatic activity of the Javakheti highland is described as 'dominoes effect', i.e. attenuation of volcanism within one zone results in formation of another, 'shifted' in submeridional direction magmatically active extension zone (Lebedev *et al.* 2008). The geodynamical regime of collisional volcanism of the Caucasian segment is characterized by compression at the depth and uneven extension within the upper part of the Earth's crust. (Koronovsky & Demina 1996); petrochemical features of basalts of eastern Anatolia and the Lesser Caucasus points to propagation of rift volcanism from the Levant Zone northward. In other words, the rift does not exist yet, but a deep mechanism for its formation already exists (Koronovsky & Demina 2007).

Several geodynamic models have been proposed for the genesis of collision-related magmatism in continental collision zones, in particular, for the Eastern Anatolian Plateau (Pearce *et al.* 1990; Keskin *et al.* 1998; Şengör *et al.* 2008; Dilek *et al.* 2009; Kheirkhah *et al.* 2009), whose direct prolongations are represented by volcanic high plateaus of Southern Georgia. Some of them may be relevant to the Eastern Anatolian-Caucasian Late Cenozoic collision zone – for example, the detachment model (Innocenti *et al.* 1982) of the last piece of subducted oceanic lithosphere to explain the Late Miocene–Quaternary calc-alkaline volcanism of southern Georgia, and the lithosphere delamination (Pearce *et al.* 1990; Keskin *et al.* 1998) model for explanation of the Pleistocene–Holocene volcanism of the Central Great Caucasus.

Recent Geodynamics

The recent geodynamics of the Caucasus and adjacent territories is determined by its position between the still converging Eurasian and Africa-Arabian plates. According to geodetic data, the rate of the convergence is ~20–30 mm/y, of which some 2/3 are likely to be taken up south of the Lesser Caucasian (Sevan-Akera) ophiolitic suture, mainly in south Armenia, Nakhchevan, northwest Iran and Eastern Turkey. The rest of the S/N-directed relative plate motion has been accommodated in the South Caucasus chiefly by crustal shortening (Jackson & McKenzie 1988; DeMets *et al.* 1990; Jackson & Ambraseys 1997; Reilinger *et al.* 1997, 2006; Allen *et al.* 2004; Podgorsky *et al.* 2007; Forte *et al.* 2010).

Tectonic stresses caused by the northward motion of the Arabian Plate are adsorbed to a considerable degree in the Antalya-Bitlis (Periarabain) ophiolitic suture and in the Zagros fold-thrust belts (DeMets *et al.* 1990; Jackson & Ambraseys 1997; Allen *et al.* 2004; Reilinger 2006). North of these structures the stresses are propagated towards the Central Caucasus by means of a relatively rigid block (Van Triangle) whose base is located south of the Lakes Van and Urumiech along the PAOS and its apex lies in the Javakheti highland. Within the rigid block, there occur intensive eruptions of Neogene–Quaternary lavas in eastern Anatolia, Turkey, Armenia (Aragats, etc), Georgia (Javakheti, Abul-Samsar, Kechut ranges). Neogene–Quaternary volcanoes are also known in the central part of the Transcaucasian foreland and in the Main Range of the Great Caucasus (Elbrus, Chegem, Keli, Kazbegi). It is noteworthy that the strongest Caucasian earthquakes occurred within this area (Figure 26) – 1988 Spitak (Armenia) and 1991 Racha (Georgia).

A complex network of faults determines the divisibility of the region into a number of separate blocks (terrains) of different orders, varying one from another by their dimensions, genesis, and geological nature. Geological, palaeobiogeographical and palaeomagnetic data provide evidence that these terrains before being accreted together in a single complicated fold-thrust belt have undergone long-term and substantial horizontal displacements within the now-vanished oceanic area of the Tethys (e.g., Dercourt *et al.* 1986, 1990; Stampfli 2000; Barrier & Vrielynk 2008). The boundary zones between these terrains represent belts of the strongest geodynamic activity with widely developed processes of tectogenesis (folding and faulting), volcanism, and seismicity. As a result of continuing northward movement of the Africa-Arabian plate in Oligocene–post-Oligocene time, the region turned into the intracontinental mountain-fold construction. The process of formation of its present-day structure and relief (high-mountain ranges of the Caucasus foredeeps, and intermontane depression of the Transcaucasus, volcanic highlands) has especially intensified since Late Miocene (Late Sarmatian, c.7 Ma). Syn-, post-collisional sub-horizontal shortening of the Caucasus caused by the northward propagation of the Africa-Arabian plate is estimated at about hundreds km. Such a considerable shortening of the Earth's crust has been realized in the region through different ways: (1) crustal deformation with wide development of compressional structures – folds, thrusts; (2) warping and displacement of crustal blocks with their uplifting, subsidence, underthrusting (a process sometimes referred to as continental subduction) and (3) lateral escaping (Adamia *et al.* 2004c).

The geometry of tectonic deformations in the region is largely determined by the wedge-shaped rigid Arabian block intensively intended into the relatively mobile Middle East-Caucasian region. In the first place, it influenced the configuration of main compressional structures developed to the north of the Arabain wedge (indentor) -from the Periarabic ophiolite suture and main structural lines in East Anatolia to the Lesser Caucasus (Koçyiğit *et al.* 2001), on the whole, and its constituting tectonic units including the Bayburt-Karabakh and Talysh fold-thrust belts. All these structural-morphological lines have clearly expressed arcuate northward-convex configuration reflecting the contours of the Arabian Block (see Figure 1). However, further north, the geometry of the fold-thrust belts is somewhat different – the Achara-Trialeti belt has, on the whole, W–E trend (although, individual faults and folds are oblique, NE–SW-trending in regard to the general strike of the belt). The Great Caucasus fold-thrust belt extends in WNW–ESE (300°–120°) direction, while the chains of young Neogene–Quaternary volcanoes are oriented in submeridional (N–S) direction that is also in compliance with general NNE–SSW sub-horizontal compression of the region. Three principal directions of active faults compatible with the dominant near N–S compressional stress produced by the Arabian Plate can be distinguished in the region (Koçyiğit *et al.* 2001) – longitudinal (WNW–ESE or W–E) and two transversal (NE–SW and NW–SE). The first group of structures is represented by compressional ones – reverse faults, thrusts, overthrusts, and related strongly deformed fault-propogation folds. Unlike the compressional faults, the transversal ones are mainly extensional structures having also more or less considerable strike-slip component (see Figure 26). The tensional nature of these faults is evidenced by intensive Neogene–Quaternary volcanism related

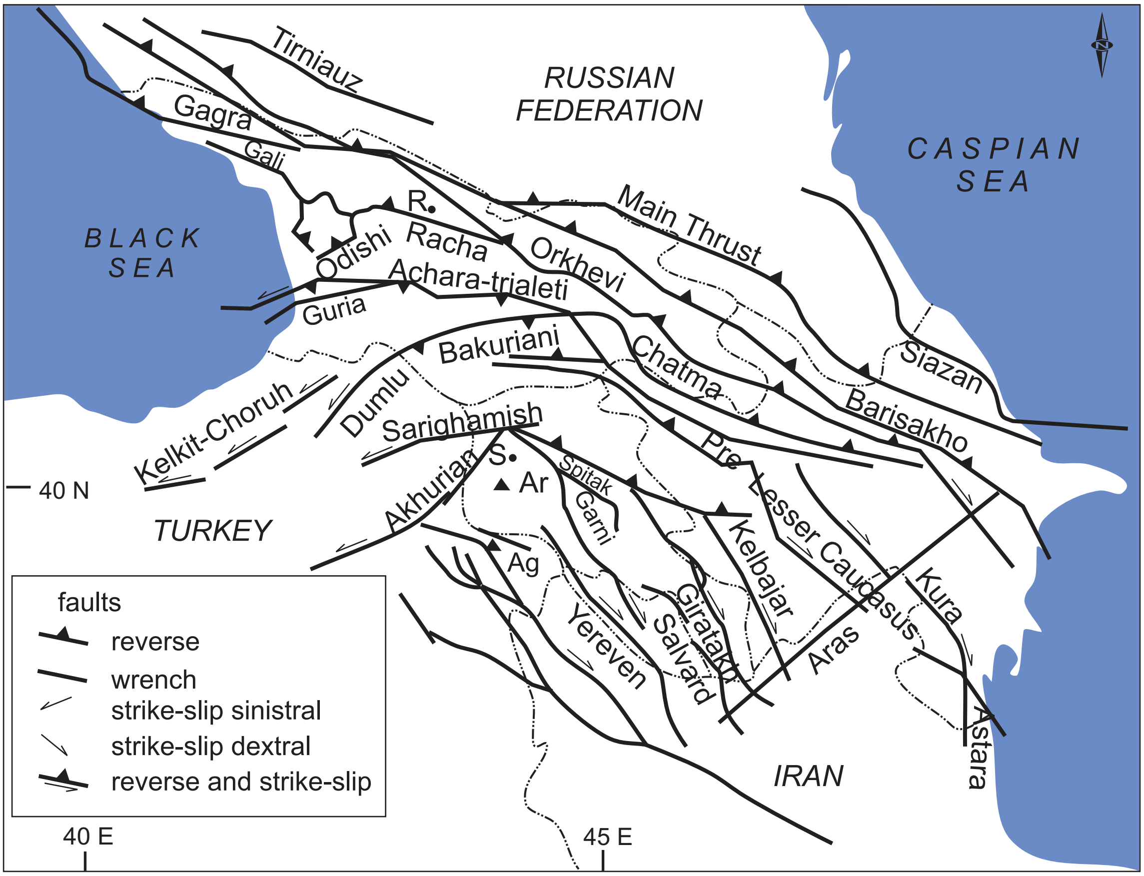


Figure 26. Schematic map of the seismic sources of the Caucasus according to A. Arakelyan, S. Nazaretyan, A. Karakhanyan (Armenia and adjacent areas of Turkey and Iran); B. Panakhi, T. Mamadly (Azerbaijan) and Sh. Adamia (Georgia, adjacent areas of Turkey and North Caucasus), Compiled in the frame of Internatuinal Scienece and Technology Center Project GA-651 'Caucasian Seismic Information Network for Hazard and Risk Assessment (CauSIN), 2005, Report. Epicenters: R– Racha, S– Spitak earthquakes; extinct volcanoes: Ar– Aragats, Ag– Ağrı Dağı.

to these faults in some places of the region – in Armenia, Azerbaijan, Southern Georgia (Javakheti highland), Transcaucasus and the Great Caucasus. NE–SW left-lateral strike-slip faults are main seismoactive structures in NE Turkey that borders on SW part of Georgia. Right-lateral strike-slip faults and fault zones are also developed in South Armenia, Nakhchevan and NW Iran. The analysis of focal mechanism of some strong earthquakes in the Caucasus shows that crustal blocks located to the west of submeridional line running across the Javakheti highland, volcano Aragatz in Armenia and Agridag in Turkey have experienced westward lateral escaping, whereas the crustal blocks east of this line evidence for ESE-directed displacement. These data are well corroborated by GPS measurements (McClusky *et al.* 2000).

Seismicity

Two large devastating earthquakes occurred in the Caucasus in the last 20–25 years. The first one was the magnitude 6.9 Spitak Earthquake on December 7, 1988 whose epicenter located within the Lesser Caucasus-Northern Armenia near the Georgian border. The earthquake became widely known due to the immense losses it caused – no less than 25 000 people were killed, some 500 000 left homeless, property damage was estimated at about 8 billion USD. The epicenter of the Spitak earthquake was related to the regional Pambak-Sevan fault, constituting a branch of the Sevan-Akera ophiolite suture. Another large seismic event was the magnitude 7.2 Racha earthquake on April 29, 1991. This earthquake, the strongest one ever recorded in Georgia, was located in Central Georgia in the southern foothills of the Great Caucasus at its junction with the Transcaucasian intermontane foreland (see Figure 26). The earthquake took about a hundred human lives and caused great damage and destruction within densely populated areas. The main shock was followed by numerous aftershocks the strongest one occurred on April 29 (M ~6.1), May 5 (M ~5.4), and June 15 (M ~6.2) causing additional damage.

Earthquakes have entailed many secondary effects in the region (landslides, debris flows, flash floods, and avalanches) that brought extensive damage to the country. The rate of risks associated with these hazards increases every year due to emergence of new complicated technological constructions, such as oil and gas pipelines (for example, Supsa and Baku-Tbilisi-Jeihan oil pipelines, Shah-Deniz and Russia-Georgia-Armenia gas pipelines), large dams, nuclear (Armenia) and hydropower plants etc.

Conclusions

Starting from the end of the Proterozoic and the beginning of the Palaeozoic, terrains detached from Gondwana were displaced towards the European continent (Baltica) that caused narrowing of Prototethys, which separated Gondwana and Baltica. Palaeozoic Paleotethys ocean was open in the wake of the Transcaucasian terrain. The Dizi basin represents a relic of Prototethys; it represents the northern passive margin of Transcaucasian island-arc system. The Dizi basin was continuously developing throughout the Palaeozoic. Variscan and Eo-Cimmerian tectonic events resulted only in its narrowing, but not closing, thus the Dizi basin development continued during the Mesozoic–Early Cenozoic.

In present-day structure, relics of the crust of Palaeotethys separating the Transcaucasian island arc system from Gondwana, are represented by Lesser Caucasian (Sevan-Akera-Zangezur) ophiolites. The Caucasian branch of Palaeotethys existed during the whole Mesozoic. However, east of the Lesser Caucasus, within Iranian Karadagh, Eo-Cimmerian tectonogenesis resulted in sharp narrowing of oceanic basin, which was replaced by flysch trough of Iranian Karadagh. Final collision of Africa-Arabian and Eurasina lithospheric plates closing of Neotethys branches and formation of the Caucasian zone of continent-continent collision happened in Oligocene–Early Miocene.

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References

Abdullaev, I. & Bagirbekova, O. 2007. About the age of stratal gabbro-diabasic intrusions in the Shazur-Julfa anticlinorium according to data of isotopic investigations. Problems of geodynamics, petrology and metallogeny of the Caucasus. *Proceedings of the Scientific Session*, 142–150 [in Russian].

Abesadze, M., Adamia, SH., Chkhotua, T., Kekelia, M., Shavishvili, I., Somin, M. & Tsimakuridze, G. 1982. *IGCP Project # 5. In:* Sassi, F.P. & Varga, I (eds), *Pre-Variscan and Variscan Metamorphic Complexes of the Caucasus*. Newsletter **4**, 5–12.

Abesadze, M., Adamia, SH., Chkhotua, T., Tsimakuridze, G. & Shavishvili, I. 2002. Structure and development of the Loki salient of the Transcaucasus. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **117**, 135–145 [in Russian].

Abesadze, M., Adamia, SH., Gabunia, G., Kopadze, T. & Tsimakuridze, G. 1989. Puddinga metaconglomerates of the Chorchana-Utslevi stripe of the Dzirula massif, their age and geological significance. *Bulletin of Academy of Sciences of Georgia* **135**, 129–132 [in Russian].

Abesadze, M., Tsimakuridze, G. & Planderova, E. 1980. New data on the age of metamorphic schists of the Dzirula massif (Georgia). *Geologycky Ustav D. Stura*, 137–143 [in Russian].

Abrams, M. & Narimanov, A. 1997. Geochemical evaluation of hydrocarbons and their potential sources in the western south Caspian depression, Republic of Azerbaijan. *Marine and Petroleum Geology* **14**, 451–468.

Adamia, SH. 1975. Plate tectonics and evolution of the Alpine system: discussion. *Geological Society of America Bulletin* **86**, 719–720.

Adamia, SH. 1977. Tectonics and geological history of Abkhazia. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **54**, 1–68 [in Russian].

Adamia, SH. 1989. Paleotethys, Mesotethys, Neotethys – various oceans or different stages of Tethys development. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **99**, 18–49 [in Russian].

Adamia, SH., Abesadze, M., Kekelia, M., Chkhotua, T. & Shavishvili, I. 1978. Metaophiolites, island-arc complexes and blastomilonites of the Main Range of the Great Caucasus. *Doklady of Academy of Sciences of the USSR* **241**, 1139–1142 [in Russian].

Adamia, SH., Abesadze, G. & Khotianovsky, P. 1973. On the stratigraphy and age of the metamorphic rocks of the Great Caucasian Main Range. *Doklady of Academy of Sciences of the USSR* **211**, 928–931 [in Russian].

Adamia, SH., Alania, V., Chabukiani, A., Chichua, G., Enukidze, O. & Sadradze N. 2004b. Evolution of the Late Cenozoic basins of Georgia (SW Caucasus). *In*: Sosson, M., Kaymakci, N., Stephenson, R.A., Starostenko, V. & Bergerat, F. (eds), *Sedimentary Basin Tectonics from the Black Sea and Caucasus to the Arabian Platform.* Geological Society, London, Special Publications **340**, 239–259.

Adamia, SH., Alania, V., Chabukiani, A., Kuloshvili, S., Maisuradze, G. & Gotsiridze, G. 2004c. Seismic source zone models of recent earthquakes of Georgia. *In*: Chatzipetros, A.A. & Pavlides, S.B. (eds), *Proceedings of the 5th International Symposium on Eastern Mediterranean Geology*, 545–548.

Adamia, SH., Alania, V., Chabukiani, A., Kutelia, Z. & Sadradze, N. 2011. Great Caucasus (Cavcasioni): a long-lived north-Tethyan back-arc basin. *Turkish Journal of Earth Sciences* **20**, 611–628.

Adamia, SH., Asanidze, B., Chkhotua, T., Kekelia, M., Shavishvili, I., Tsimakuridze, G. & Zakariadze, G. 2004a. Fragments of Paleo-Tethyan terranes in the crystalline core of the Great Caucasus. *In*: Chatzipetros, A.A. & Pavlides, S.B. (eds), *Proceedings of the 5th International Symposium on Eastern Mediterranean Geology*, 233–236.

Adamia, SH., Bayraktutan, S. & Lordkipanidze, M. 1995. Structural correlation and Phanerozoic evolution of the Caucasus-Eastern Pontides. *In*: Erler, A., Ercan, T., Bİngöl, E. & Örçen, S. (eds), *Geology of the Black Sea Region*. Mineral Resarch and Exploration Institute (MTA) of Turkey, Publication, 69–75.

Adamia, SH., Belov, A., Kekelia, M. & Shavishvili, I. 1987. Paleozoic tectonic development of the Caucasus and Turkey (Geotravers G). *In*: Flügel, H.W., Sassi, F.P. & Grecula, P. (eds), *Pre-Variscan and Variscan Events in the Alpine-Mediterranean Mountain Belts*, 23–50. Alfa Publishers.

Adamia, SH., Chabukiani, A., Chkhotua, T., Sadradze, N., Zakaraia, D. & Zakariadze, G. 2010. Geology of the Caucasus and adjacent areas, 1:250 000 scale geological map. *In:* Christofides, G., Kantinaris, N., Kostopoulos, D. & Chatzipetros, A. (eds), *Proceedings, XIX Congress of the Carpathian-Balkan Geological Association* **99**, 1–9.

Adamia, SH., Chkhotua, T., Kekelia, M., Lordkipanidze, M., Shavishvili, I. & Zakariadze, G. 1981. Tectonics of the Caucasus and adjoining regions: implications for the evolution of the Tethys ocean. *Journal of Structural Geology* **3**, 437–447.

Adamia, SH., Gamkrelidze, I., Zakariadze, G. & Lordkipanidze, M. 1974. Achara-Trialeti trough and problem of origin of the deep marine trough of the Black Sea. *Geotectonics* **1**, 78–94 [translated into English].

Adamia, SH., Kekelia, M. & Tsimakuridze, G. 1983. Pre-Variscan and Variscan granitoides of the Caucasus. *IGCP Project № 5, News Letter* **5**, 5–12.

Adamia, SH., Kutelia, Z., Planderova, E. & Khutsishvili, O. 1990. Raethian–Hettangian deposits of the Dizi series of Svaneti (Great Caucasus). *Doklady of Academy of Sciences of the USSR* **313**, 395–396 [in Russian].

Adamia, SH., Lebanidze, Z., Svanidze, Ts. & Shavishvili, I. 2003. Late Paleozoic–Triassic geodynamic evolution of Georgia (Caucasus). *Bollettino Società Geologico, Italy* **2**, 213–218.

Adamia, SH., Lordkipanidze, M. & Zakariadze, G. 1977. Evolution of an active margin as exemplified by the Alpine history of the Caucasus. *Tectonophysics* **40**, 183–199.

Adamia, SH., Matskhonashvili, K. & Khutsishvili, O. 1961. Geology of post-Paleogene continental volcanogene formations of the eastern part of South Georgia. *Proceedings of Geological Institute of Academy of Sciences of GSSR, Series Mineralogy and Petrology* **VI**, 73–106 [in Russian].

Adamia, SH., Mumladze, T., Sadradze, N., Tsereteli, E., Tsereteli, N. & Varazanashvili, O. 2008. Late Cenozoic tectonics and geodynamics of Georgia (SW Caucasus). *Georgian International Journal of Sciences and Technology* **1**, 77–107.

Adamia, SH. & Shavishvili, I. 1982. Root zone of the ophiolite nappes of the western Caucasus. *Doklady of Academy of Sciences of the USSR* **267**, 1424–1425 [in Russian].

Afanasiev, G., Borsuk, A., Kondakov, L., Lupanova, N. & Ploshkov, V. 1971. *Concrete Magmatic Formations of the North Caucasus*. Izvestia of Academy of Sciences of the USSR **7** [in Russian].

Agamalian, V. 2004. The Lesser Caucasus earth crust formation and evolution in the collision zone of Paleo-Tethys. *In*: Chatzipetros, A.A. & Pavlides, S.B. (eds*), Proceedings of the 5th International Symposium on Eastern Mediterranean Geology*, 17–20.

Ajgirei, G. (ed) 1976. *Geology of the Great Caucasus*. Moscow, 'Nedra', 1–363 [in Russian].

Ali-Zade, A. (ed) 2003. *Geology in Azerbaijan*. Baki, Economic Minerals **6**, 1–577 [in Russian].

Allen, M., Jackson, J. & Walker, R. 2004. Late Cenozoic reorganization of the Arabia-Eurasia collision and the comparison of short-term and long-term deformation rates. *Tectonics* **23**, 1–16.

Allen, M., Vincent, S., Alsoop, G., Ismaİl-Zadeh, A. & Flacker, R. 2003. Late Cenozoic deformation in the South Caspian region: effects of a rigid basement block within a collision zone. *Tectonophysics* **366**, 223–239.

Altiner, D., Özkan-Altiner, S. & Koçyiğit, A. 2000. Late-Permian foraminiferal facies belts in Turkey: palaeogeographic and tectonic implications. *In:* Bozkurt, E., Winchester, J.A, & Piper, J.D.A (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 83–96.

Ananiashvili, G. & Sakhelashvili, Z. 1984. Correlation of the Miocene deposits of Chekhoslovakia and Georgia on the basis of Mollusc fauna. *Geologicky Práce, Geologický Ústav Dionyza Stúra* **80**, 79–89 [in Russian].

Andruschuk, V. (ed) 1968. *Geology of the USSR. North Caucasus*. Moscow, 'Nedra' **IX**, 1–760 [in Russian].

Arakelyants, M., Borsuk, A. & Shanin, L. 1968. Youngest volcanoplutonic granitoid association of the Greater Caucasus based on potassium-argon dating. *Doklady of Academy of Sciences of the USSR* **182**, 1157–1160 [in Russian].

Aslanian, A. (ed) 1970. *Geology of the USSR. Armenian SSR.* Moscow, 'Nedra' **XLIII**, 1–464 [in Russian].

Aslanian, A., Bagdasarian, G., Gabunia, L., Rubinshtein, M. & Skhirtladze, N. 1984. Radiometric ages of Neogene volcanic formations of Georgian SSR, Armenian SSR and part of Nakhchevan ASSR. *Proceedings of Academy of Sciences of Armenian SSR, Earth Sciences* **35**, 3–24[in Russian].

Avanessian, A., Malkhassian, E. & Nazaretian, S. 2000. *Unique Geological Monuments of Armenia. Guidebook for Tourists*, 1–157.

Azadaliev, J. & Kerimov, F. 2007. Ore-bearing secondary-quartzitic formations in the Ordubad ore region in the south of the Lesser Caucasus (Nakhchevan AR of Azerbaijan). *Problems of Geodynamics, Petrology and Metallogeny of the Caucasus. Proceedings of Scientific Session*, 244–280 [in Russian].

Azizbekov, SH. (ed) 1972. *Geology of the USSR. Azerbaijanian SSR.* Moscow, 'Nedra' **XLVII**, 1–520 [in Russian].

Babaie, A., Babaie, H. & Arvin, M. 2005. Tectonic evolution of the Neyriz ophiolite, Iran: an accretionary prism model. *Ofioliti* **30**, 64–74.

Banks, C., Robinson, A.G. & Williams, M. 1997. Structure and regional tectonics of the Achara-Trialeti fold belt and the adjacent Rioni and Kartli foreland basins, Republic of Georgia. *In*: Robinson, A.G. (ed), *Regional and Petroleum Geology of the Black Sea and Surrounding Region.* AAPG Memoir **68**, 331–346.

Barrier, E. & Vrielynck, B. 2008. *Paleotectonic Maps of the Middle East.* Middle East Basins Evolution Programme. CGMW, Atlas, Maps, 1–14.

Bartnitsky, Y.N., Dudauri, O.Z. & Stepanyuk, L.M. 1990. Geochronology of Phanerozoic granitoides from folded areas of Eastern Europe. *5th Working Meeting, Isotopes in Nature, Leipzig, Proceedings*, 1–10.

Belov, A. 1981. Tectonic evolution of the Alpine folded domain in Paleozoic. *Proceedings of Geological Institute of Academy of Sciences of the USSR* **347**, 1–212 [in Russian].

Belov, A., Abesadze, M., Adamia, SH., Agamalian, V., Chegodaev, L., Letavin, A., Omelchenko, V. & Somin, M. 1989. Explanatory note of the stratigraphic correlation forms of the Caucasus (USSR). *Rendiconti Della Societa Geologica Italiana* **12**, 119–126.

Belov, A. & Kizevalter, D. 1962. Main features of the structure and history of development of the Late Hercynian structural stage of the central part of the North Caucasus (Middle Carboniferous–Lower Triassic). *Proceedings, Caucasian Expedition of Allunion Airgeological Department and Moscow State University* **3**, 275–314 [in Russian].

Belov, A. & Omelchenko, V. 1986. Late Variscan and post-Variscan thrusts in the pre-Upper Jurassic basement of the Northern Caucasus. *Izvestia VUZ (Institute of Higher Education), Series Geological and Mineral Resources* **9**, 14–20 [in Russian].

Belov, A., Omelchenko, V. & Chegodaev, L. 1984. Pjatigorsk-Uchkulan-Pjiatigorsk. Excursion 008 A+C. Pre-Mesozoic complexes of the Caucasus. *International Geological Congress, 27 Session, Georgian SSR, Guidebook, Tbilisi, 'Khelovneba'*, 114–126.

Beridze, M. 1983. *Geosynclinal Sedimentary Lithogenesis as Exemplified by Early Mesozoic Formations of the Greater Caucasus.* 'Metsniereba', Tbilisi, 1–321 [in Russian].

Bibikova, E., Somin, M., Krasivskaia, I., Gracheva, T., Makarov, V., Arakelianz, M. & Vidiapin, I. 1991. U-Pb age of the orthogneisses of the Main Range of the Caucasus. *Izvestia of Academy of Sciences of the USSR, Series Geological* **9**, 23–34 [in Russian].

Bogdanovski, O.G., Zakariadze, G.S., Karpenko, S.F., Zlobin, S.K., Pukhovskaya, V.M. & Amelin, Yu.V. 1992. Sm–Nd age of the gabbroids of a tholeiitic series of the ophiolites of the Sevan-Akera zone of the Lesser Caucasus. *Academy of Sciences Russia* **327**, 566–569 [in Russian].

Borsuk, A. 1979. *Mesozoic and Cenozoic Magmatic Formation of the Great Caucasus.* Moscow, 'Nauka', 1–299 [in Russian].

Borsuk, A., Ivanenko, V., Karpenko, M. & Chernishev, I. 1989. Precise K-Ar dating of Neogene intrusions of the Transcaucasian transversal zone and possible geodynamic consequences. *Doklady of Academy of Sciences of the USSR* **38**, 1188–1191 [in Russian].

Brunet, M-F., Korotaev, M., Ershov, A. & Nikishin, A. 2003. The south Caspian basin: a review of its evolution from subsidence modelling. *Sedimentary Geology* **156**, 111–148.

Budagov, B. 1964. About character of recent tectonic movement in the region of Shakhdag massif (SE Caucasus) in connection of finding of Upper Sarmatian fauna. *Doklady of Academy of Sciences of the USSR* **155**, 330–332 [in Russian].

Camps, P., Ruffet, P., Shcherbakova, V., Prevot, M., Moussin-Poushkin, A., Cholpo, L. Gogichaishvili, A. & Asanidze, B. 1996. Direction, intensity and Ar-Ar age of Pliocene geomagnetic field reversal or excursion recorded in Georgia. *Journal of Physics of the Earth Planet* **96**, 41–59.

Chelidze, G. 1973. Correlation of Pontian from Euxinian and Caspian basins. *Bulletin of Academy of Sciences of GSSR* **71**, 645–648 [in Russian].

Chernishev, I., Arakelyants, M., Lebedev, V., Bubnov, S. & Goltsman, J. 2000. Geochronology of eruptions of Quaternary volcanoes of the Kazbegi district of the Greater Caucasus (K-Ar method data). *Proceedings of Geological Institute of Academy of Sciences of Georgia* **115**, 81–91 [in Russian].

Chkhikvadze, V., Mchedlidze, G., Burchak-Abromovich, N., Bendukidze, N., Burchak, D., Gabelaia, Ts., Amiranashvili, N., Meladze, G., Kharabadze, E. & Chkareuli, N. 2000. Review of the localities of Tertiary Vertebrates of Georgia. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **115**, 153–163 [in Russian].

Çoğulu, E. & Krummenacker, D. 1967. Problems géochronométrique dans la partie NW del'Anatolie Centrale (Turquie). *Bulletin Suisse de Mineral, Petrographie* **47**, 825–831.

Danelian, T., Asatryan, G., Sosson, M., Galoyan, G., Avagyan, A., Sahakyan, L. & Rolland, Y. 2007. New paleontological (Radiolarian) age constraints for the sedimentary cover of ophiolites in the Lesser Caucasus (Armenia). *International Symposium on Middle East Basins Evolution.* Paris-France, p. 47.

Davoudzadeh, M. & Shmidt, K. 1981. Contribution to the paleogeography and stratigraphy of the Upper Triassic to the Middle Jurassic of Iran. *Neu Jährlich Geologisch Paläontologisch Abh* **162**, 137–103.

Demets, C., Gordon, R., Argus, D. & Stein, S. 1990. Current plate motions. *Geophysical Journal International* **101**, 425–478.

Dercourt, J., Ricu, L. E., Adamia, SH., Gsaszar, G., Funk, H., Lefeld, J., Rakus, M., Sandulescu, M., Tollmann, A. & Tchoumachenko, P. 1990. Anisian to Oligocene paleogeography of the European margin of Tethys (Geneva to Baku). *Evolution of the Northern Margin of Tethys.* Mémoire de la Sočiété Géologique de France, Paris, Nouvelle Série **154**, 159–160.

Dercourt, J., Zonenshain, L., Ricou L., Kazmin V., Le Pichon X., Knipper, A., Grandjacqet, C., Sbortshikov, I., Geyssant, J., Lepvrier, C., Pechersky, D., Boulin, J., Sibuet J., Savostin, L., Westphal M., Bajenov M., Laner J. & Biju-Duval, B. 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamir since the Lias. *Tectonophysics* **123**, 241–315.

Devlin, W., Gogswell, J., Gaskins, G., Isaksen, G., Pitcher, D., Puls, D., Stanley, K. & Wall, G. 1999. South Caspian Basin: young, cool, and full promise. *GSA Today* **9**, 1–9.

Dİlek, Y., Imamverdiev, N. & Altunkaynak, S. 2009. Geochemistry and tectonics of Cenozoic volcanism in the Lesser Caucasus (Azerbaijan) and the peri-Arabian region: collision-induced mantle dynamics and its magmatic fingerprint. *International Geology Review* **52**, 536–578.

Djanelidze, Ch., Klopotovskaya, N. & Maisuradze, G. 1982. Holocene volcanites and lake beds near Goristsikhe settlement. *International Union for Quarternary Research (INQUA), 11th Congress Guidebook for Excursion A-16*, 28–31.

Djaparidze, V., Bosinski, G., Bugianishvili, T., Gabunia, L., Justus, A., Klopotoskaya, N., Kvavadze, E., Lordkipanidze, D., Maisuradze, G., Mgeladze, N., Nioradze, M., Pavlenishvili, E., Schmincke, H., Sologashvili, D., Tushabramishvili, D., Tvalchrelidze, M. & Vekua, A. 1989. Der altpaläolitische Fundplatz Dmanisi in Georgian (Kaukasus). *Jahrbuch des Römish-Germanishen Zentralmuseums, Mainz,* (RGZM) **36**, 67–116.

Eastern Paratethyan Neogene Stratotypes and Key Sections in Soviet Georgia 1985. Excursion A3 *Field Guide, 8th Congress of the Regional Committee on Mediterranean Neogene Stratigraphy,* 4–14 September 1985. Hungarian Geological Survey, Budapest, 1–106.

Eristavi, M. 1962. Subdivision of Lower Cretaceous of the Alpine zone. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **10**, 1–113 [in Russian].

Ershov, A., Brunet, M.-F., Nikishin, A., Bolotov, S., Nazarevich, B. & Korotaev, M. 2003. Northern Caucasus basin: thermal history and synthesis of subsidence models. *Sedimentary Geology* **156**, 95–118.

Ferring, C., Swisher, C., Bosinsky, G., Gabunia, L., Kikodze, Z., Lordkipanidze, D., Tvalchrelidze, M. & Tutberidze, B. 1996. *Progress Report on the Geology of the Plio-Pleistocene Dmanisi and the Diliska Gorge*. Republic of Georgia, Paleontological Society, New Orleans, 5–6.

Forte, A., Cowgill, E., Bernardin, T., Kreylos, O. & Hamann, B. 2010. Late Cenozoic deformation of Kura fold-thrust belt, southern Greater Caucasus. *Geological Society of America Bulletin* **122**, 465–486.

Gabashvili, E., Amiranashvili, N. & Chkhikvadze, V. 2000. Fossil tortoises of Udabno locality (Late Miocene, Eastern Georgia). *Proceedings of Geological Institute of Academy of Sciences of Georgia* **115**, 177–180 [in Russian].

Gabunia, L. 1964. *Benara Fauna of the Oligocene Vertebrates.* Tbilisi, 'Metsniereba', 1–89 [in Russian].

Gabunia, L., Joris, A., Justus, A., Lordkipanidze, D., Muskhelishvili, A., Nioradze, M., Swicher, C. III, Vekua, A., Bosinski, G., Ferring, R. & Maisuradze, G. 1999. Neue hominidefunde des Altpaläolitischen fundplatzes Dmanisi (Georgian Caucasus) in kontext aktueller grabungsergebnisse. *Archäologisches Korrespondezblatt Jargang* **29**, 451–488.

Galoyan, G., Rolland, Y., Sosson, M., Corsini, M., Billo, S., Verati, C. & Melkonyan, R. 2009. Geology, geochemistry, and 40Ar/39Ar dating of Sevan ophiolites (Lesser Caucasus, Armenia): evidence for Jurassic back arc opening and hot spot event between South Armenian Block and Eurasia. *Journal of Asian Earth Sciences* **34**, 135–153.

Galoyan, G., Rolland, Y., Sosson, M., Corsini, M. & Melkonyan, R. 2007. Evidence for superimposed MORB, oceanic plateau and volcanic arc series in the Lesser Caucasus (Stepanavan, Armenia). *Comptes Rendus Geoscience* **339**, 482–492.

Gamkrelidze, P. (ed) 1964. *Geology of the USSR. Georgian SSR.* Moscow, 'Nedra' **X**, 1–655 [in Russian].

Gasanov, T. 1986. The history of development of the Sevan-Akera zone of the Lesser Caucasus. *Geotectonica* **2**, 92–104 [in Russian].

Giorgobiani, T. & Zakaraya, D. 1989. *The Folded Structure of the North-Western Caucasus and the Mechanism of its Formation.* Metsniereba, Tbilisi, 1–60 [in Russian].

Golonka, J. 2004. Plate tectonic evolution of the southern margin of Eurasia in the Mesozoic and Cenozoic. *Tectonophysics* **381**, 235–273.

Göncüoğlu, M.C. 1997. Distribution of Lower Paleozoic units in the Alpine terranes of Turkey: paleogeographic constraints. *In*: Göncüoğlu, M.C. & Derman, S. (eds). *Lower Paleozoic Evolution in Northwest Gondwana*. Turkish Association of Petroleum Geologists, Special Publications **3**, 13–24.

Grekov, I., Kropachev, S., Momot, S. & Koren, T. 1974. Marukh Paleozoic nappe of the Northern Caucasus. *Sovetskaia Geologia* **2**, 77–85 [in Russian].

Gukasian, R. & Somin, M. 1995. Rb-Sr isochron dating of metamorphic rocks of the Main Range of the Caucasus. *Proceedings of the VII Conference on Geology and Mineral Resources of the Northern Caucasus*, 239–240 [in Russian].

Gurbanov, A., Hanel, M., Lippold, G. & Kober, B. 1995. Paragenesis and age of the rocks of Makera and Buulgen series of the Main Range Zone of the Greater Caucasus. *Proceedings of the VII Conference on Geology and Mineral Resources of the Northern Caucasus, Essentuki*, 169–171 [in Russian].

Hanel, M., Lippold, X., Kober, B., Gurbanov, A. & Borsuk, A. 1993a. Isotop-geochronological reconstruction of initial nature of volcanics of metamorphic complexes of the Greater Caucasus. *Petrology* **1**, 171–187 [in Russian].

Hanel, M., Lippold, X., Kober, B., Gurbanov, A. & Borsuk, A. 1993b. On Early Paleozoic age of metagranodiorites of the Main Range Zone of the Great Caucasus. *Petrology* **1**, 487–498.

Haq, B.U. & Van Eisinga, F.W.B.1987. *Geological Time Table*. Fourth revised edition, Elsevier, New York, USA, printed in Netherlands.

Innocenti, F., Mazzuoli, R., Pasquare, G., Radicati F. & Villari, L. 1982. Tertiary and Quaternary volcanism of the Erzurum-Kars area (Eastern Turkey). Geochronological data and geodynamic evolution. *Journal of Volcanology and Geothermal Research* **13**, 223–240.

Jackson, J. & Ambraseys, N. 1997. *Convergence Between Eurasia and Arabia in Eastern Turkey and the Caucasus. Historical and Prehistorical Earthquakes in the Caucasus,* Kluwer Academic Publishers, 79–90.

Jackson, J. & McKenzie, D. 1988. The relationship between plate motions and seismic moment tensors, and the rates of active deformation in the Mediterranean and Middle East. *Geophysical Journal - Royal Astronomical Society* **93**, 45–73.

Janelidze, O. 1970. Foraminifers of the Lower and Middle Miocene of Georgia. *Proceedings of Paleobiology of Institute of Academy of Sciences of GSSR*, Tbilisi, 'Metsniereba', 1–172 [in Russian].

Janelidze, O. 1977. *Polymorphinids of the Miocene of Georgia*. Tbilisi, 'Metsniereba', 1–81 [in Russian].

Jones, R. & Simmons, M. 1997. A review of the stratigraphy of Eastern Paratethys (Oligocene–Holocene) with particular emphasis on the Black Sea. *In:* Robinson, A.G. (ed) *Regional and Petroleum Geology of the Black Sea and Surrounding Region.* AAPG Memoir **68**, 39–52.

Jrbashian, R., Karapetian, S. & Mnatsakanian, A. 2002. Late collision rhyolitic volcanism in the north-eastern part of Armenian highland. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **117**, 53–63 [in Russian].

Kacharava, M. 1977. *Stratigraphy of Paleogene Sediments of Achara-Trialeti Folded System*. Tbilisi, 'Metsniereba', 1–357 [in Russian].

Kakhadze, I., Tsagareli, A., Nutsubidze, K. & Zesashvili, V. 1957. Lower and Middle Jurassic deposits of the western part of the North Caucasus and its correlation with analogous deposits of Georgia. *Proceedings of Geological Institute of Academy of Sciences of Georgia, Geological Series* **10**, 197–295 [in Russian].

Kariakin, I. & Aristov, V. 1990. On the age and geological position of 'exotic rocks' of Turogachai zone (Lesser Caucasus). *Doklady of Academy of Sciences of the USSR* **311**, 1189–1193 [in Russian].

Kazakhashvili, J. 1984. *Paleobiological History of the Mollusc Fauna*. Tbilisi, 'Metsniereba', 1–132 [in Russian].

Kazmin, V., Schreider, A. & Bulichev, A. 2000. Early stages of evolution of the Black Sea. *In:* Bozkurt, E., Winchester, J.A & Piper, J.D.A (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 235–249.

Keskİn, M. 2003. Magma generation by slab steepening and breakoff beneath a subduction-accretion complex: an alternative model for collision-related volcanism in Eastern Anatolia, Turkey. *Geophysical Research Letters* **30**(**24**), 8046, doi:10.1029/2003GL018019.

Keskİn, M., Pearce, J.A. & Mitchell, J.G. 1998. Volcano-stratigraphy and geochemistry of collision-related volcanism on the Erzurum-Kars Plateau, North Eastern Turkey. *Journal of Volcanology and Geothermal Research* **85**, 355–404.

Khaburzania, I., Shubladze, R., Edilashvili, V. & Sholpo, L. 1979. Paleomagnetic section of volcanic rocks in the southern part of Georgian SSE. *Izvestia of Academy of Sciences of the USSR. Physics of Solid Earth* **15**, 269–273 [translated into English].

Khain, E. 1984. Ophiolites and Hercynian nappe of the Fore Range of the North Caucasus. *Proceedings of Institute of Geology of Academy of Sciences of the USSR* **328**, 96 [in Russian].

Khain, V. 1975. Structure and main stages in the tectono-magmatic development of the Caucasus: an attempt at geodynamic interpretation. *American Journal Sciences* **275-A**, 131–156.

Kheirkhah, M., Allen M. & Emami, M. 2009. Quaternary collision magmatism from the Iran/Tureky bordelands. *Journal of Volcanology and Geothermal Research* **182**, 1–12.

Khutsishvili, O.D. 1972. Stratigraphy and Fauna of the Kvishi Formation. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **33**, 1–92 [in Russian].

Kitovani, T., Imnadze, Z. & Topozov, R. 1991. Stratigraphy of the Antropogene deposits. *In*: *Georgia during Antropogene*. 'Sakartvelo', Tbilisi, 10–58 [in Russian].

Knipper, A. 1980. The tectonic position of ophiolites of the Lesser Caucasus. *In:* Panaiotov, A. (ed), *Proceedings of the International Ophiolite Symposium.* Cyprus, Geological Survey Department, Nicosia, 372–376.

Knipper, A. 1990. Pre-Late Jurassic tectonic events and their role on the formation of ophiolitic sequences in the Caucasus. *Symposium on Ophiolite Genesis and Evolution of Oceanic Lithosphere, Abstracts*, p. 112.

Knipper, A. 1991. Upper Triassic–Early Jurassic sedimentary breccias in the ophiolite suture of the Lesser Caucasus. *In*: Peters, T.J., Nicolas, A. & Coleman, R. (eds), *Ophiolite Genesis and Evolution of the Oceanic Lithosphere*. Ministry of Petroleum and Minerals, Sultanate of Oman, 705–713.

Koçyİğİt, A., Yilmaz, A., Adamia, SH. & Kuloshvili, S. 2001. Neotectonics of East Anatolian plateau (Turkey) and Lesser Caucasus: implication for transition from thrusting to strike-slip faulting. *Geodynamica Acta* **14**, 177–195.

Konak, N. & Hakyemez, Y. 2001. Tectonic units of the easternmost part of the Pontides: stratigraphical and structural implications. *Proceedings of the 2nd International Symposium on the Petroleum Geology and Hydrocarbon Potential of the Black Sea Region.* 22–24 September, 1996, Şile-İstanbul, Turkey, 93–103.

Koronovsky, H. & Demina, L. 2007. Disappeared Pliocene volcanoes of the Main Caucasian Range and grandiose explosion of the Kazbegi at the beginning of Pleistocene (North Caucasus) *Problems of Geodynamics, Petrography and Metallogeny of the Caucasus.* Proceedings of the Scientific Sessions, Baki, 92–103 [in Russian].

Koronovsky, N. & Demina, L. 1996. Collisional volcanism pattern of the Caucasian segment of the Alpine belt. *Doklady of Academy of Sciences* **350**, 510–522.

Koronovsky, N. & Demina, L. 2007. Late Cenozoic magmatism of the Great Caucasus. *In*: Leonov, Yu. (ed), *Alpine history of the Great Caucasus*. Moscow, 'GEOS', 251–285 [in Russian].

Kropachev, S. & Grekov, I. 1974. Horizontal movements during Middle Paleozoic within Fore Range of the Lesser Caucasus. *Doklady of Academy of Sciences of the USSR* **215**, 665–668 [in Russian].

Kutelia, Z. 1983. New data on the stratigraphy of the Dizi series (Svaneti). *Bulletin of Academy of Sciences of Georgia* **109**, 585–588 [in Russian].

Kutelia, Z. & Barskov, I. 1983. Conodonts of the Dizi series of Georgia (Southern Slope of the Great Caucasus). *Doklady of Academy of Sciences of the USSR* **270**, 394–396[in Russian].

La Géorgie 2002. *Berceau du Premier Européan*. Tbilisi, 1–47.

Lebanidze, Z., Koiava, K., Svanidze, Ts., Adamia, SH. & Asanidze, B. 2009. Position of the territory of Georgia during Late Paleozoic–Mesozoic. *2nd International Symposium of the Geology of the Black Sea Region, Abstracts, Ankara. Turkey*, 121–122.

Lebedev, V., Bubnov, S., Dudauri, O. & Vashakidze, G. 2008. Geochronology of Pliocene volcanism of Javakheti highland (Lesser Caucasus). Article 2. Eastern part of the Javakheti highland. *Regional Geological Correlation. Stratigraphy. Geological Correlation* **16**, 101–123 [in Russian].

Lebedev, V., Chernishev, I., Dudauri, O., Arakelyants, M., Chugaev, A., Golzman, Y. & Vashakidze, G. 2004. Geochronology of Neogene–Quaternary dacitic volcanism of the Javakheti highland (Lesser Caucasus, South Georgia). *Proceedings of Geological Institute of Academy of Sciences of Georgia* **119**, 535–544 [in Russian].

Lomize, M. 1969. *Volcanism of North-Western Caucasus and its Relations with Tectonics. Moscow. Publishers of Moscow State University*, 1–220 [in Russian].

Lordkipanidze, M., Meliksetian, B. & Djrbashian, R. 1989. Mesozoic–Cenozoic magmatic evolution of the Pontian-Crimean-Caucasian region. IGCP Project 198: evolution of the Northern Margin of the Tethys. *Memoires de la Societe Geologique de France, Paris* **154**, 101–124.

Lordkipanidze, M.B. & Zakariadze, G.S. 1974. Paleogene volcanism of Achara. Problems of Geology of Achara-Trialeti. *Proceedings of Institute of Geology of GSSR* **44**, 74–86 [in Russian].

Lordkipanidze, M.B. & Zakariadze, G.S. 1986. Evolution of the Alpine volcanism of the Caucasus and its relations to tectonics. *The 6 Colloquium on Geology of the Aegean Region. Piri Reis International Contribution Series, İzmir, Turkey* **2**, 415–435.

Lyberis, N. & Manby, G. 1999. Oblique to orthogonal convergence across the Turan block in the post-Miocene. *AAPG Bulletin* **83**, 1135–1160.

Maisuradze, G. & Djanelidze, A. 1987. Evolution of the Late Pleistocene environment in the Caucasus. *Paleogeography and Loess. Akademiai Kiado, Budapest*, 145–156.

Maisuradze, G., Smelov, S. & Tvalchrelidze, M. 1980. New evidence on the volcanics of Djavakheti (Southern Georgia). *Bulletin of Academy of Sciences of Georgia* **98**, 605–608 [in Russian].

Maisuradze, L. 1971. *Sarmatian Foraminifera of Western Georgia.* Tbilisi, 'Metsniereba', 1–131 [in Russian].

Maisuradze, L.1988. *Foraminifera of the Meotian of West Georgia*. Tbilisi, Metsniereba, 1–106 [in Russian].

Majidi, B. 1979. The ultrabasic lava flows of Mashhad, north-east Iran. *Geological Magazine* **118**, 49–58.

Mamedov, M. 1998. *Petrology of the Late Cretaceous and Eocene Magmatic Formations of the Lesser Caucasus and Talysh*. PhD Thesis, Institute of Geology of GSSR, Tbilisi, 1–72 [in Russian, unpublished].

McClusky, S., Balassanian, S., Barka, A., Demİr, C., Ergİntav, S., Georgiev, I., Gürkan, O., Hamburger, M., Hurst, K., Kahle, H., Kastens, K., Kekelidze, G., King, R., Kotzev, V., Lenk, O., Mahmoud, S., Mishin, A., Nadaraia, M., Ouzounis, A., Paradissis, D., Peter, Y., Prilepin, M., Reilinger, R., Sanli, I., Seeger H., Tealeb, A., Toksöz, M. & Veis, G. 2000. Global positioning system constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus. *Journal of Geophysical Research* **105(B3)**, 5695–5719.

Mikhailov, V., Panina, L., Polino, R., Koronovsky, N., Kiseleva E., Klavdiev, N. & Smolianinova, E. 1999. Evolution of the North Caucasus foredeep: constraints based on the analysis of subsidence curves. *Tectonophysics* **307**, 361–379.

Milanovsky, E. & Khain, V. 1963. *Geological Structure of the Caucasus*. Essay of Regional Geology of the USSR. Moscow State University Publishers, 1–358 [in Russian].

Mitchell, J. & Westaway, R. 1999. Chronology of Neogene and Quaternary uplift and magmatism in the Caucasus: constraints from K-Ar dating of volcanism in Armenia. *Tectonophysics* **304**, 157–186.

Morkovkina, V. & Arutiunian, G. 1971. On radiometric ages of ultrabasic rocks of the Sevan ridge. *Izvestia of Academy of Sciences of the USSR* **11**, 133–137.

Moshashvili, A. 1982. Evolution of the Kura basin's Earth crust and its connection with problem of the Paleotethys. *In:* Muratov, M.W. & Adamia, S. (eds), *Problem of Geodynamics of the Caucasus*. 'Moscow, 'Nauka', 64–71 [in Russian].

Nadareishvili, G. & Sadradze, N. 2004. Comparison of Paleogene volcanism of eastern Pontides (Turkey) and south-western part of Achara-Trialeti zone (Georgia). *Proceedings of Institute of Geology of Academy of Sciences of Georgia* **119**, 599–612 [in Russian].

Nazarova, N. & Tkhostov, T. 2007. The fluid regime of formation of granitoides of Nakhchevan zone of the Lesser Caucasus. *Problems of Geodynamics, Petrology and Metallogeny of the Caucasus.* Proceedings of Scientific Session, Baki, 122–127 [in Russian].

Okay, A. 2000. Was the Late Triassic orogeny caused by the collision of an oceanic plateau? *In*: Bozkurt, E., Winchester, J.A. & Piper, J.D.A. (eds), *Tectonic and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 25–42.

Okay, A. & Şahİntürk, Ö. 1997. Geology of eastern Pontides. *In:* Robinson, A.G. (ed), *Regional and Petroleum Geology of the Black Sea and Surrounding Region.* AAPG Memoir **68**, 291–311.

Okay, A., Zattin, M. & Cavazza, W. 2010. Apatite fission-track data for the Miocene-Arabia collision. *Geology* **38**, 35–38.

Panov, D. & Lomize, M. 2007. Early and Middle Jurassic of the Great Caucasus (stratigraphy and tectonics, volcanism and geodynamic evolution). *In*: Leonov, Yu. (ed), *Alpine History of the Great Caucasus*. Moscow, 'GEOS', 39–110 [in Russian].

Paris, F. & Robardet, M. 1990. Early Paleozoic paleobiogeography of the Variscan region. *Tectonophysics* **177**, 193–213.

Pearce, J., Bender, J., De Long, S., Kidd, W., Low, P., Guner, Y., Saroglu, F., Yilmaz, Y., Moorbath, S. & Mitchell, J. 1990. Genesis of collision volcanism in Eastern Anatolia, Turkey. *Journal of Volcanology and Geothermal Research* **44**, 189–229.

Perchuk, A. & Phillippot, P. 1997. Rapid cooling and exhumation of eclogitic rocks from the Great Caucasus, Russia. *Journal of Metamorphic Geology* **15**, 299–310.

Phillippot, P., Blichert-Toft, J., Perchuk, A., Costa, S. & Gerasimov, V. 2001. Lu/Hf and Ar/Ar chronometry supports extreme rate of subduction zone metamorphism deduced from geospeedometry. *Tectonophysics* **342**, 23–38.

Podgorsky, J., Hearn, E., McClusky, S., Reilinger, R., Taymaz, T., Tan, O., Prilepin, M., Guseva, T. & Nadariya, M. 2007. Postseismic deformation following the 1991 Racha, Georgia earthquake. *Geophysical Research Letters* **34**, LO4310, 1–5.

Popkhadze, L. 1975. Meotian from Western Georgia. *Bulletin of Academy of Sciences of Georgia* **80**, 381–384.

Popkhadze, L. 1983. The foraminifera and ostracoda of the Chokrakian sediments of Western Georgia from Western Georgia. *Bulletin of Academy of Sciences of Georgia* **110**, 541–544.

Potapenko, Y., Snejko, V., Somin, M. & Usikm, V. 1999. Hercynian granitoides in the structure and evolution of the Great Caucasus. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **114**, 148–167 [in Russian].

Potapenko, Y. & Stukalina, G. 1971. First finding of organic remnants in metamorphic complexes of the Great Caucasian Main Range. *Doklady of Academy of Sciences of the USSR* **198**, 1161–1165 [in Russian].

Raumer, J.F. von, Stampfli, G., Borel, G. & Bussy, F. 2002. Organization of pre-Variscan basement areas at the north-Gondwanian margin. *International Journal of Earth* Sciences **91**, 35–52.

Reilinger, R., McCLusky, S. & Souter, B. 1997. Preliminary estimates of plate convergence in the Caucasus collision zone from global positioning system measurements. *Geophysical Research Letters* **24**, 1815–1818.

Reilinger, R., McClusky, S., Vernant, Ph., Lawrence, SH., Ergintav, S., Cakmak, R., Ozener, H., Kadirov, F. & Guliev, I. 2006. GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions. *Journal of Geophysical Research* **111**, B05411.

Remane, J., Faure-Muret, A. & Odin, G. 2006. *International Stratigraphic Chart.* International Union of Geological Sciences, UNESCO.

Robertson, A.H. F. 2000. Mesozoic–Tertiary tectonic-sedimentary evolution of a south Tethyan oceanic basin and its margins in southern Turkey. *In:* Bozkurt, E., Winchester, J.A. & Piper, J.D.A. (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 97–138.

Robertson, A.H.F., Ustaömer, T., Pickett, E., Collins, A., Andrew, T. & Dixon, J. 2004. Testing models of Late Paleozoic–Early Mesozoic orogeny in Western Turkey: support for an evolving open-Tethys model. *Journal of Geological Society, London* **161**, 501–511.

Rolland, Y., Billo, S., Corsini, M., Sosson, M. & Galoyan, G. 2009b. Blueschists of the Amassia Stepanavan suture zone (Armenia): linking Tethys subduction history from E-Turkey to W-Iran. *International Journal of Earth Sciences* **98**, 533–550.

Rolland, Y., Galoyan, G. Bosch, D., Sosson, M., Corsini, M., Fornari, M. & Verati, C. 2009a. Jurassic back-arc and Cretaceous hot-spot in the Armenian ophiolites – implication for the obduction process. *Lithos* **112**, 163–187.

Rubinshtein, M. 1970. Regionale und Lokalale Verjungung des Argon-Alters am Beispeil des Caucasus. *Eclogae Geologicae Helvetiae* **63**, 282–289.

Rubinstein, M., Adamia, SH., Devnosashvili, D., Dobridin, V. & Rosentur, L. 1972. Dating of some Late Neogene and Quarternary effusives of Transcaucasia (based on geological, radiological and paleomagnetic data). *International Colloquium 'The Boundary Between Neogene and Quarternary',* Moscow, 168–174.

Rustamov, M. 1983. New data on evolution of Paleogene magmatism of the Lesser Caucasus. *Evolution of Magmatism of Main Structures of the Earth*. Moscow, 'Nauka', 65–66 [in Russian].

Rustamov, M. 2005. *South Caspian Basin-Geodynamic Events and Processes*. Baki, Institute of Geology NASA, 1–345 [in Russian].

Rustamov, M. 2007. Main tendencies of geodynamic evolution of the central segment of the Tethys. *Problems of Geodynamics, Petrology and Metallogeny of the Caucasus.* Proceedings of the Scientific Sessions, Baki, 14–35 [in Russian].

Saidova, Ch., Kazmin, V., Sborshikov, I., Matveenkov, V. & Ivanov, M. 1988. New data on stratigraphy of the Dizi series of Svanety. *Doklady of Academy of Sciences of the USSR* **302**, 407–410 [in Russian].

Saintot, A., Brunet, M.-F., Yakovlev, F., Sebrier, M., Stephenson, R., Ershov, A., Chalot-Prat, F. & McCann, T. 2006. The Mesozoic–Cenozoic tectonic evolution of the Greater Caucasus. *In*: Gee, D. & Stephenson, R.A. (eds), *European Lithosphere Dynamics*. Geological Society, London, Memoirs **32**, 277–289.

Şengör, A.M.C., Özeren, M.S., Keskİn, M., Sakınç, M., Özbakır, A.D. & Kayan, İ. 2008. Eastern Turkish high plateau as a small Turkic-type orogen: implications for post-collisional crust-forming processes in Turkic-type orogens. *Earth Sciences Reviews* **90**, 1–48.

Sholpo, L., Sologashvili, J., Kakulia, V., Pavlenishvili, E., Asanidze, B. & Khaburzania, I. 1998. *Paleomagnetism of Georgian Rocks.* Javakhishvili Tbilisi State University Publishers, 1–291.

Sokolov, S. 1977. Olistostromes and ophiolitic nappes of the Lesser Caucasus. *Proceedings of Geological Institute of Academy of Sciences of the USSR* **296**, 1–96 [in Russian].

Solovkin, A. 1950. On Triassic of the upper reaches of the r. Akera (Azerbaijanian SSR). *Doklady of Academy of Sciences of the USSR* **6**, 395–398 [in Russian].

Somin, M. 1971. *Pre-Jurassic Basement of the Main Range and the Southern Slope of the Greater Caucasus.* Moscow, Nauka Press, 1–245 [in Russian].

Somin, M. 1991. Geological characteristic of metamorphic complexes of the Great Caucasus. Zones of the Fore Range and Main Range. *Petrology of Metamorphic Complexes of the Great Caucasus.* Moscow, 'Nauka', 18–46 [in Russian].

Somin, M. 2007a. Main features of structures of the pre-Alpine basement of the Great Caucasus. *Rocks Complexes and Geodynamics of the Main Stages of Evolution of the Great Caucasus.* Moscow, 'GEOS', 10–52 [in Russian].

Somin, M. 2007b. *Structural Position and Geodynamic Conditions of Formation of Metamorphic Complexes of the Great Caucasus and Cuba.* PhD Thesis, Shmidt Institute of Earth Physics, RAN, Moscow, 1–56 [unpublished].

Somin, M., Baianova, T., Levkovich, N. & Lavrishev, V. 2004. Geological and geochronological investigations of metamorphic basement of the Main Range Zone of the Great Caucasus: new data and problems. *Proceedings of Institute of Geology of Academy of Sciences of Georgia* **119**, 521–534 [in Russian].

Somin, M., Lepekhina, E. & Konilov, A. 2007a. Age of high-temperature gneiss core of the Central Caucasus. *Doklady on Earth Sciences* **415**, 690–694.

Somin, M., Levchenko, O., Kotov, A., Makeev, A., Komarov, A., Lavrishchev, V. & Lebedev, V. 2007b. The Paleozoic age of high-pressure metamorphic rocks in the Dakhov salient, Northwestern Caucasus: results of U-Pb geochronological investigations. *Doklady on Earth Sciences* **416**, 1018–1021.

Somin, M. & Vidiapin, Y. 1989. The Damkhurts metaconglomerates of the Main Caucasus Range and their tectonic significance. IGCP Project 276. *In*: Papanikolau, D. & Sassi, F.P. (eds), *Paleozoic Geodynamic Domains and Their Alpidic Evolution in the Tethys.* Newsletter **1** Athens, 91–95.

Stampfli, G. 2000. Tethyan oceans. *In:* Bozkurt, E., Winchester, J.A. & Piper, J.D.A. (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area.* Geological Society, London, Special Publications **173**, 1–23.

Svanidze, T., Iakobidze, E. & Lebanidze, Z . 2000. Upper Triassic–Bajocian vegetation of the Western Georgia and stratigraphical significance of the flora. *Geologica Balcanica* **30**, 19–24.

Taktakishvili, I. 1984. *Biostratigraphy of the Pliocene of Western Georgia*. Tbilisi, 'Metsniereba', 1–135 [in Russian].

Taktakishvili, I. 2000. Peculiarities of Pliocene deposits of Western Georgia. *Proceedings of Geological Institute of Academy of Sciences of Georgia* **115**, 126–130 [in Russian].

Tatishvili, K. 1965. Saliniferous occurrence of the Maykopian deposits of Akhaltsikhe. *In*: Davitashvili, L. (ed), *Cenozoic Fauna of Georgia and its Geohistoric Importance*. Tbilisi, Metsniereba, 66–68 [in Russian].

Tolluoğlu, A. & Sümmer, E. 1995. An evolutionary model on Early Paleozoic of Anatolian microcontinent, northern margin of Gondwana land. *Geological Bulletin of Turkey* **38**, 1–10.

Topuz, G., Altherr, R., Kalt, A., Satir, M., Werner, O. & Schwartz, W.H. 2004b. Aluminous granulites from the Pulur Complex, NE Turkey: a case of partial melting, efficient melt extraction and crystallisation. *Lithos* **72**, 183–207.

Topuz, G., Altherr, R., Satir, M. & Schwartz, W.H. 2004a. Low-grade metamorphic rocks from the Pulur Complex, NE Turkey: implications for the pre-Liassic evolution of the Eastern Pontides. *International Journal of Earth Sciences* **93**, 72–91.

Topuz, G., Altherr, R., Schwarz, W., Zack, T. & Sunal, G. 2009. Middle Devonian blueschist metamorphism in the eastern Pontides, NE Turkey: vestiges of the Rheic Ocean? *2nd International Symposium on the Geology of the Black Sea Region, Abstracts.* 5–9 October, Ankara, Turkey, 203–204.

Topuz, G., Altherr, R., Siebel, W., Schwartz, W.H., Zack, T., Hasözbek, A., Barth, B, Satir, M. & Şen, C. 2010. Carboniferous high-potassium I-type granitoid magmatism in the Eastern Pontides: the Gümüşhane pluton (NE Turkey). *Lithos* **116**, 92–110.

Treloar, P.J., Mayringer, F., Finger, F., Gerdes, A. & Shengelaia, D. 2009. New age data from the Dzirula Massif, Georgia: implications for Variscan evolution of the Caucasus. *2nd International Symposium on the Geology of the Black Sea Region, Abstracts*. 5–9 October 2009, Ankara, Turkey. Abstract Book, 204–205.

Tuchkova, M. 2007. Lithology of Lower–Middle Jurassic deposits of Great Caucasus. *In*: Leonov, Yu. (ed), *Alpine History of the Great Caucasus.* Moscow, 'GEOS', 141–214 [in Russian].

Tutberidze, B. 2004. *Geology and Petrology of Alpine Late Orogenic Magmatism of the Central Part of Caucasian Segment.* Tbilisi University Publishers, Tbilisi, 1–340 [in Russian].

Uznadze, M. & Tsagareli, Ts. 1979. *Sarmatian Flora of the Dzindza Valley*. Tbilisi, 'Metsniereba', 1–164 [in Russian].

Uznadze, M. 1946. Flora of the Goderdzi suite. *Bulletin of Academy of Sciences of Georgian SSR,* volume **7**, 427–434 [in Russian].

Uznadze, M. 1951. Quisatibi fossil flora. *Proceedings of Institute of Geology and Mineralogy of Academy of Sciences of Georgian SSR*, 299–305 [in Russian].

Vekua, L., Dzotsenidze, N., Pavlenishvili, V., Asanidze, B., Sologashvili, J. & Kakulia, V. 1977. Paleomagnetism of the Neogene–Quaternary volcanogenic formations of the Djavakheti highland. *Izvestia of Academy of Sciences of the USSR*, *Physics of Solid Earth* **13**, 738–744 [translated into English].

Vekua, A., Lordkipanidze, D., Rightmire, P., Agnsti, J., Ferring, R., Maisuradze, G., Mouskhelishvili, A., Nioradze, M., Leon, M., Tappen, M., Tvalchrelidze, M. & Zollikofer, C. 2002. A new skull of early Homo Dmanisi. *Georgia Science* **297**, 85–90.

Vincent, S., Allen, M., Ismail-Zadeh, A., Flecker, R., Foland, K. & Simmons, M. 2005. Insights from the Talysh of Azerbaijan into the Paleogene evolution of the South Caspian region. *Geological Society of America Bulletin* **117**, 1513–1533.

Yilmaz, A., Engİn, T., Adamia, SH. & Lazarashvili, T. (eds) 2001. *Geological Studies of the Area Along Turkish-Georgian Border*. Mineral Research and Exploration Institute (MTA) of Turkey, Report, 1–388.

Zakariadze, G., Chkhotua, T., Adamia, SH., Shavishvili, I. & Kutelia, Z. 2009. Relics of Prototethyan and Paleotethyan Oceanic Lithosphere in the Caucasus. *2th International Symposium on the Geology of the Black Sea Region, Abstracts*. 5–9 October 2009, Ankara, Turkey, 238–239.

Zakariadze, G., Dİlek, Y., Adamia, S., Oberhänsli, R., Karpenko, S., Bazylev, B. & Solov'eva N. 2007. Geochemistry and geochronology of the Neoproterozoic Pan-African Transcaucasian Massif (Republic of Georgia) and implications for island-arc evolution of the late Precambrian Arabian-Nubian Shield. *Gondwana Research* **11**, 97–108.

Zakariadze, G., Karpenko, S., Basilev, B., Adamia, SH., Oberkhanseli, R., Solovieva, N. & Ljalikov A. 1998. Petrology, geochemistry and Sm-Nd age of pre-Late Hercynian paleooceanic complex of Dzirula salient of Transcaucasian massif. *Petrology* **6**, 422–444 [in Russian].

Zakariadze, G., Knipper, A., Sobolev, A., Tsamerian, O., Dmitriev, L.., Vishnevskaia, V. & Kolesov, G. 1983. The ophiolite volcanic series of the Lesser Caucasus. *Ophioliti* **8**, 439–466.

Zanchi, A., Zanchetta, S., Berra, F., Mattei, M., Garzanti, E., Villa, I., Molinex, S., Nawab, A. & Sabouri, J. 2007. The Eo-Cimmerian collision between North Iranian and the southern portion of Eurasia: new constraints from the Alborz-Kopet Dag regions. *International Symposium on Middle East Basins Evolution, Abstracts*, December 4–5, 2007, Paris-France, 17–18.