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Great Caucasus (Cavcasioni): A Long-lived North-Tethyan Back-Arc Basin

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Abstract: The Great Caucasus is a northwest–southeast-directed mountain range more than 1100 km long, located between the Black Sea and Caspian Sea. It represents an intracontinental tectonic system resulting from the Late Cenozoic structural inversion of a Palaeozoic–Mesozoic–Early Cenozoic back-arc basin (Dizi basin) in response to the convergence of the Africa-Arabian and Eurasian lithospheric plates. It is bounded to the south by the Transcaucasian massif, a palaeo-island-arc, and to the north by the Scythian platform. The Great Caucasus fold-and-thrust mountain belt is characterized by complete, intensive folding, mainly south vergent imbricated thrusting, close-joint cleavage etc. Structural relationships of the Great Caucasian fold-and-thrust mountain belt with the Transcaucasian massif and Scythian platform are, as a rule, tectonic overthrusts, but in some places the contacts are transitional.

The Great Caucasus basin has developed, at least from Devonian, throughout Palaeozoic and Mesozoic to Early Cenozoic, as established by marine palaeontological data. Late Palaeozoic (Variscan) metamorphic and magmatic events, folding and topographic inversion are not observed in the Southern Slope Zone of the Great Caucasus. Variscan and Early Mesozoic (Old Cimmerian) orogenic events did not lead to closure of the Dizi back-arc basin. The mountainous Crimea (Triassic–Liassic Tauric series) represents a similar basin of continuous deposition with no Variscan and Old Cimmerian orogenic events and is generally considered to be the western extension of the Great Caucasian basin. East of the Caucasus, such a basin characterized by continuous Triassic–Jurassic marine sandy-argillaceous sedimentation is the Great Balkans (Transcaspian), which was continuously developing since the Palaeozoic.

Throughout the whole Mesozoic, Palaeocene and Eocene, the Great Caucasus represented a domain accumulating thick terrigenous, carbonate, and volcanogenic marine deposits and only in Oligocene–Miocene it was transformed into a mountainous edifice between the Black Sea and Caspian Sea basins. At present, the Black Sea and Caspian Sea basins unconformably overlie different structures of adjacent land; their shoreline cuts several main tectonic units of the Caucasus and Crimea.

Key Words: Great Caucasus, Cavcasioni, Transcaucasus, pre-Caucasus, Tethys, back-arc

Büyük Kafkaslar: Kuzey Tetis'in Uzun Süreçli Yay-ardı Havzası

Özet: Büyük Kafkaslar, Karadeniz ile Hazar Denizi arasında kuzeybatı–güneydoğu yönünde 1100 km uzanan bir dağ silsilesidir. Bu dağ kuşağı Paleozoyik–Mesozoyik–Erken Tersiyer yaşlı bir yay-ardı havzanın (Dizi havzası) Geç Tersiyer'de Afrika-Arabistan ile Avrasya levhalarının yakınlaşması ve çarpışmasına bağlı olarak yapısal terslenmesi ile oluşmuştur. Büyük Kafkaslar kıvrım-bindirme kuşağı yoğun kıvrımlanma, genellikle güneye verjanslı bindirmeler ve sık eklem klivajı ile tanımlanır; güneyde eski bir ada-yayını temsil eden Transkafkasya masifleri, kuzeyde ise İskit Platformu ile sınırlanır. Büyük Kafkas kıvrım-bindirme kuşağının Transkafkasya masifleri ve İskit Platformu ile olan dokanakları genellikle bindirmeli fakat bazı bölgelerde geçişlidir.

Paleontolojik verilere göre Büyük Kafkasya havzası gelişmesine Devoniyen'de veya öncesinde başlamış, tüm Geç Paleozoyik, Mesozoyik ve Erken Tersiyer'de devam etmiştir. Büyük Kafkasların güney zonunda (*Southern Slope Zone)* Geç Palaeozyik'te (Variskan) metamorfik ve magmatik olaylar, kıvrımlanma ve yükselme gözlenmez. Daha kuzeydeki gözlenen bu Variskan ve Erken Mesozoyik (eski Kimmeriyen) olayları güney zondaki Dizi yay-ardı havzasının kapanmasına yol açmamıştır. Dağlık Kırım bölgesi de (Triyas–Liyas Tavrik serisi) Variskan ve erken Kimmeriyen olaylarının gözlenmediği, sürekli çökelim gösteren benzer bir havza oluşturur ve genellikle Büyük Kafkaslar havzasının batıya doğru devamını oluşturduğu kabul edilir. Büyük Kafkasların doğusunda benzer bir havza Triyas–Jura süresince sürekli kumlu-kili yaşlı sedimentasyonla tanımlanan, Paleozyik'te beri sürekli gelişim gösteren Büyük Balkan (Transcaspian) havzasıdır.

Tüm Mesozoyik, Palaeosen ve Eosen boyunca Büyük Kafkaslar kalın kırıntılı, karbonat ve volkanik malzemenin depolandığı bir havza oluşturur, Oligo–Miyosen'de bu havza Karadeniz ve Hazar denizi havzaları arasında yükselen bir dağ kuşağına dönüşür. Günümüzde Karadeniz ve Hazar Denizi çökelleri Büyük Kafkasların değişik birimlerini ve yapılarını uyumsuzlukla üzerlemektedir.

Anahtar Sözcükler: Büyük Kafkaslar, Transkafkasya, pre-Kafkasya, Tetis, yay-ardı

Introduction

Cavcasioni (Bolshoi Kavkaz in Russian and Great Caucasus in English) is a NW–SE-directed mountain range more than 1100 km long, located between the Black Sea and Caspian Sea. It represents an intracontinental fold-and-thrust mountain system resulting from the Late Cenozoic syn- and post-collisional structural inversion of a Palaeozoic–Mesozoic–Early Cenozoic pre-collisional back-arc basin (Khain 1974; Adamia *et al.* 1977, 1981; Dercourt *et al.* 1985; Stampfli *et al.* 2001) in response to convergence of the Africa-Arabian and Eurasian lithosphere plates (Figure 1). It is bounded to the south by the Transcaucasian massif, which is a Palaeozoic–Early Cenozoic island-arc, and to the north by the Scythian young platform (Figure 2). At present, this is a fold-and-thrust mountain belt characterized by intensive tight and subisoclinal folding, mainly south-vergent imbricated thrusting, close-joint cleavage etc (Figures 3 & 4). Structural relations of the Great Caucasian fold-and-thrust mountain belt with the Transcaucasian massif and Scythian platform are, as a rule, tectonic overthrusts, however, in some places the contacts are transitional (Milanovsky & Khain 1963; Gamkrelidze 1964; Shardanov & Borukaev 1968; Grigoriants *et al.* 1972; Gamkrelidze & Gamkrelidze 1977; Adamia *et al.* 1987a; Giorgobiani & Zakaraia 1989; Kengerli 2005).

The crust of the Great Caucasus is continental, up to 55–60 km thick; it covers an area up to 60–75 km wide. The 'granite' layer of the Great Caucasus is exposed within its Main Range Zone forming the crystalline core of the fold-and-thrust mountain belt. Sedimentary cover of the Great Caucasian fold-and-thrust mountain belt, which is widespread within the Main Range Zone and Southern Slope Zone, is dominated by deep marine terrigenous and hemipelagic deposits of Palaeozoic, Mesozoic, and Early Cenozoic ages. The oldest deposits of the Dizi series, dated as Devonian–Triassic, are exposed within the Southern Slope Zone (Figure 1).

The Palaeozoic–Triassic Formations

The Palaeozoic–Triassic sediments crop out within three regions of the Southern Slope Zone of the Great Caucasian fold-and-thrust mountain belt. They are: (1) the Mzimta gorge located within the western Southern Slope Zone, (2) Svaneti – within the central Southern Slope Zone (Dizi series), and (3) Speroza – within the eastern Southern Slope Zone.

The Dizi Series

The Dizi series crops out in the cores of two (Enguri and Tskhenistskali) alpine antiforms (Figure 5). The core of the former is more eroded; exposed formations range in age from Lower–Middle Devonian to the uppermost Triassic. Thickness of the series, as reported by various authors, is 1.5–2.0 km. Present knowledge on its stratigraphy is mainly based on the palaeontological data (corals, foraminifers, crinoids) described or reported by Sh. Adamia, G. Agalin, A. Belov, G. Chikhradze, V. Kazmin, Z. Kutelia, V. Slavin, and M. Somin. Lately, radiolarians and conodonts were found in this series (Z. Kutelia, I. Barskov) that promoted a better understanding of the stratigraphic setting of the series and changed existent notions on its structure.

The Dizi series is rather monotonous. The presence of Lower–Middle Devonian, Upper Devonian and Lower–Middle Carboniferous, and Triassic anchimetamorphic deposits (phyllites, slates) is proved on the basis of palaeontological findings. The deposits are represented mainly by fine-grained terrigenous and hemipelagic formations, characterized by an irregular alternation of siltstones,

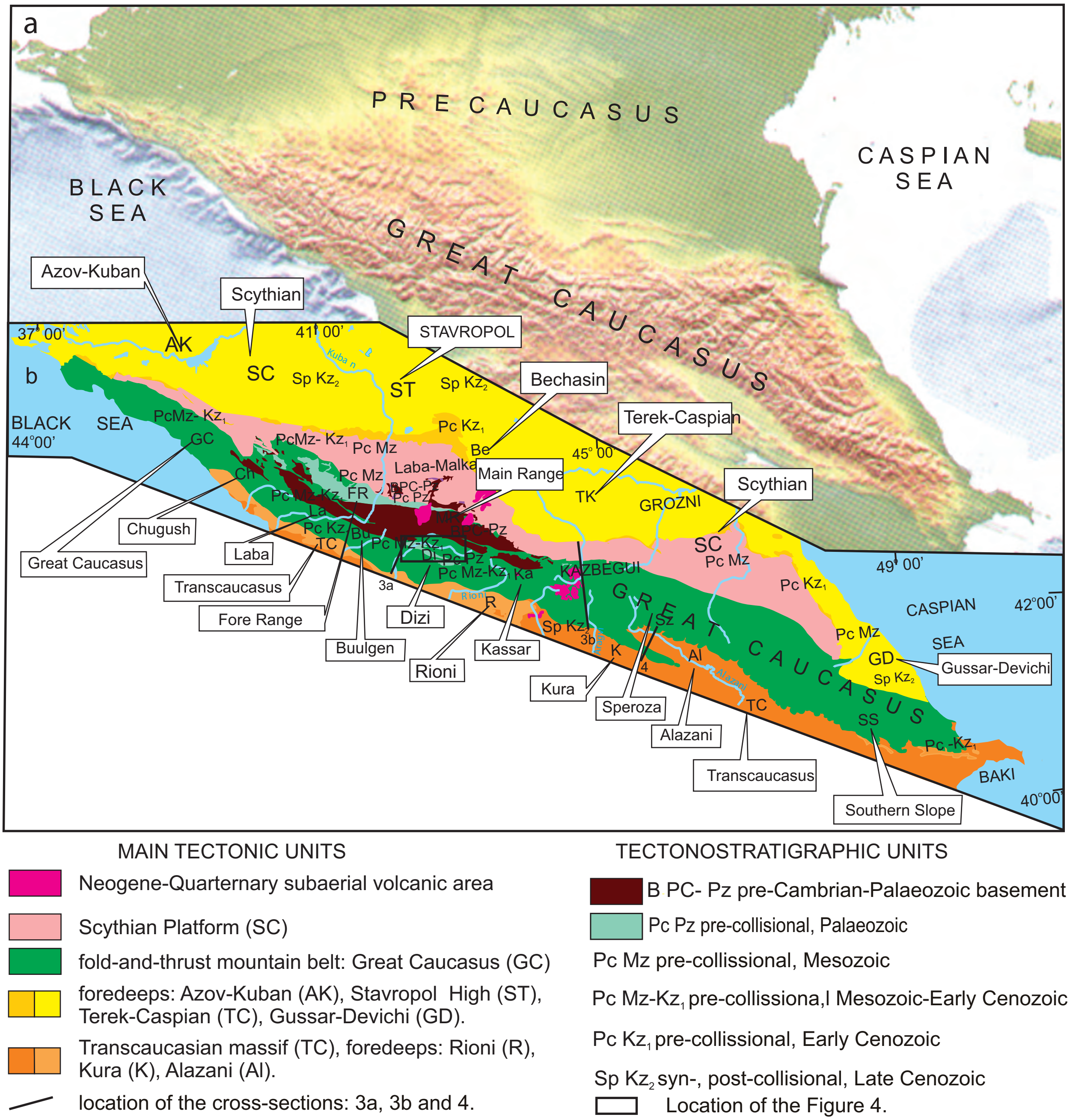

Figure 1. Physical **(a)** and tectonic **(b)** maps of the Great Caucasus.

sandstones, gritstones, rare conglomerates and olistostromes. Almost all levels contain thin (from decimeters to tens of meters) bodies of biohermal limestones and their breccias; however, fossils are not always well preserved because of recrystallization. The considerable components of the succession are cherts often bearing conodonts and radiolarians; they form thin intercalations, lenses, and single bands within terrigenous deposits. In the Middle Devonian, there occur tuffs and tuff breccias (up to 200 m thick) of andesite-basaltic composition. There may also occur another (Middle Carboniferous?) thin horizon of dacitic tuffites. Clastic sediments of the series are poorly sorted and usually are of

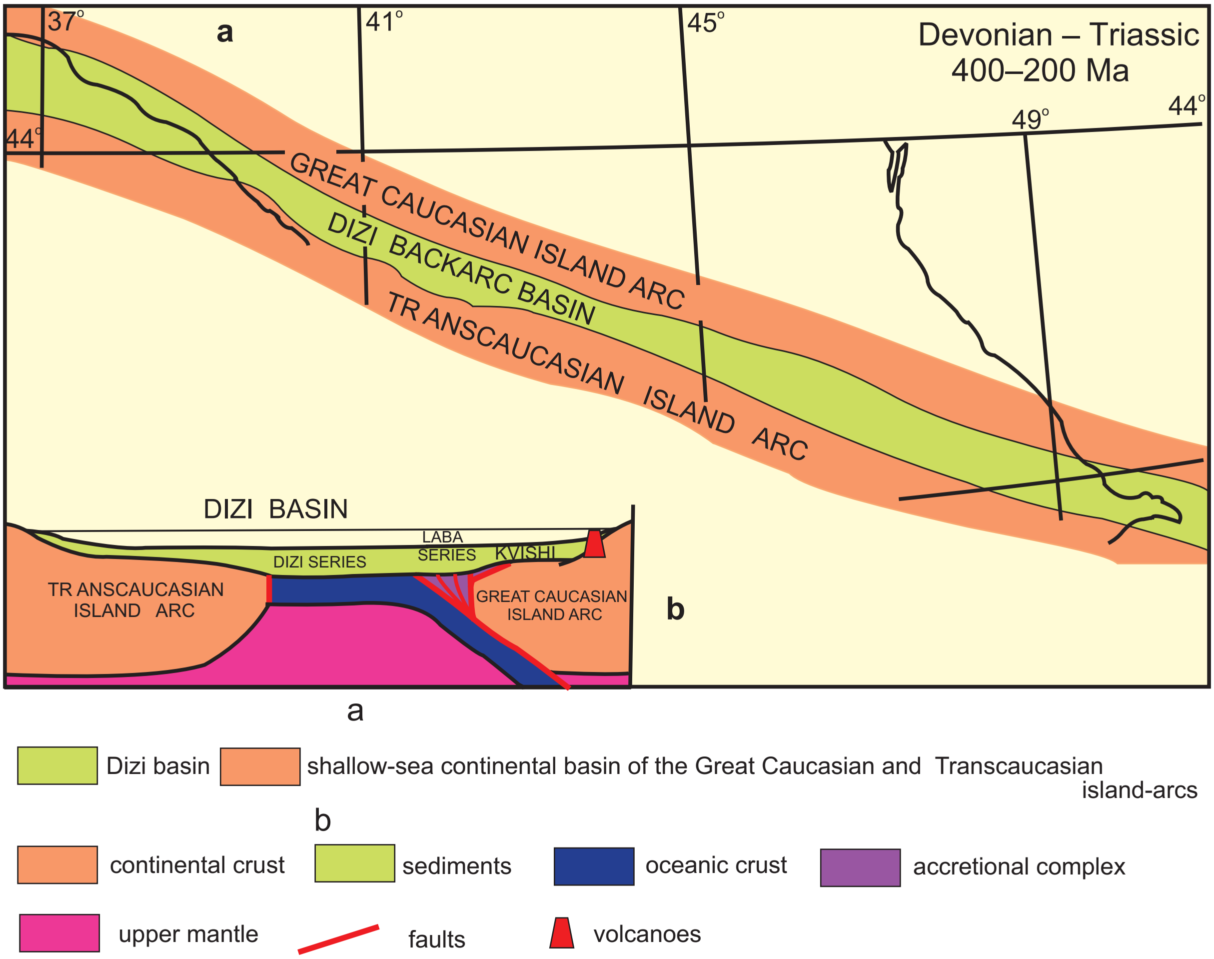

Figure 2. **(a)** Palaeotectonic map of the Great Caucasus for Palaeozoic–Triassic, without palinspatic reconstruction. **(b)** Palaeotectonic reconstruction across the central segment of the Dizi-Great Caucasus (not in scale).

quartz-polymictic composition. Some levels, e.g. Eifelian sediments, show the presence of subarkosic psammites: fragments of crystalline schists, quartzites and granitoids. The index fossils are: amphipors (Coelenterata (phylum), Stromatoporata (order), corals, and conodonts found in marbles and cherts (Figure 6).

The oldest Devonian rocks of the Dizi series crop out in Upper Svaneti. They are dated mainly on the basis of conodonts found in cherts and marble lenses (Kutelia 1983). Upper Emsian–Lower Eifelian age is established on the basis of conodonts (point 1; Figures 5 & 6); the Lower Eifelian – by conodonts of the Zone of Polygnatus costatus (point 2, 3); the Frasnian and Famennian – by conodonts from point 4. Limestone lenses of the same point contain Middle–Upper Devonian corals and crinoidea (Slavin *et al.* 1962; Somin 1971). Limestone lenses from the river Khumpreri (the right tributary of the river Enguri) bear also Ludlovian–Eifelian and Eifelian–Lower Givetian corals (point 5) (Adamia 1968; Belov & Somin 1964). Individual specimens of Lower Devonian conodonts were found in limestones cropping out along the river Leshnuri – left confluent of the river Khumpreri (point 6).

The presence of Tournaisian rocks within the Dizi series was established on the basis of conodonts. Siliceous schists of the point 7 include conodonts indicating Early Tournasian age of the rocks (Kutelia 1983).

Upper Palaeozoic part of the Dizi series begins with Visean–Namurian deposits containing lenses of

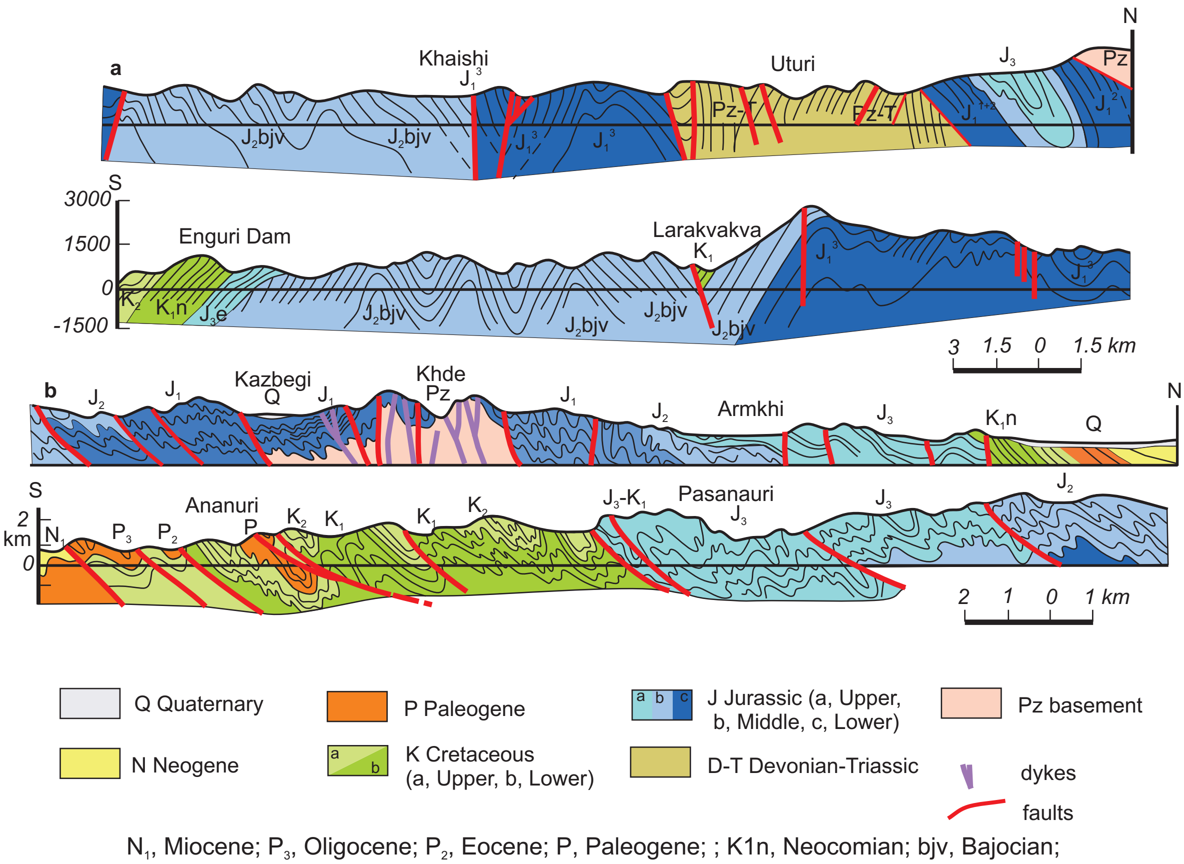

Figure 3. Simplified geological profiles of the Great Caucasus compiled on the basis of geological mapping: **(a)** across the central Great Caucasus (r. Enguri), **(b)** across the eastern Great Caucasus (Georgian Military Road, r. Aragvi and r.Tergi).

recrystallized limestones and cherts with conodonts and corals (points 8 & 9). Several forms of Visean conodonts and corals were found in the Kazakh-Tvibi gorge – the right tributary of the river Enguri (Adamia 1968; Somin 1971; Kutelia 1983). The presence of the Middle–Upper Carboniferous (Bashkirian and Moscovian stages) was confirmed by conodonts found in red cherts (point 10) and recrystallized limestones (point 11; Kutelia & Barskov 1983). In this part of the Dizi formation Somin & Belov (1967) found Middle Carboniferous corals and foraminifers (points 12 & 13). Nearly in the same level were found Middle–Late Carboniferous–Permian foraminifers (Adamia 1968), and also a microfauna (points 14 & 15) (Somin 1971). Within the Enguri section (points 16 & 17), Carboniferous–Lower Permian deposits of the Dizi series are dated on the basis of foraminifers. Permian microfauna are found in limestones (point 18) (Somin & Belov 1967). In Lower Svaneti, Middle and Upper Permian part of the Dizi series is dated by microfauna, foraminifers, and corals – points 19–21 (Slavin *et al.* 1962; Somin 1971), points 20, 21 (Adamia 1968; Somin 1971). Upper Triassic deposits of the Dizi series are dated by foraminifers and palynomorphs (Saidova *et al.* 1988; point 22). In the cross-section along the Enguri river, the Upper part of the Dizi series, sandwiched between faunistically characterized Norian turbidites and Sinemurian black slates (point 23), contain Rhaetian–Hettangian marine palynomorphs. According to Bukia (1959), Late Triassic palinomorphs were separated from the uppermost part of the Dizi series (point 24).

Mzimta. Within the Mzimta gorge, there are slates, siltstones, sandstones, and gritsones (terrigenous turbidites) alternating with fine-bedded limestones

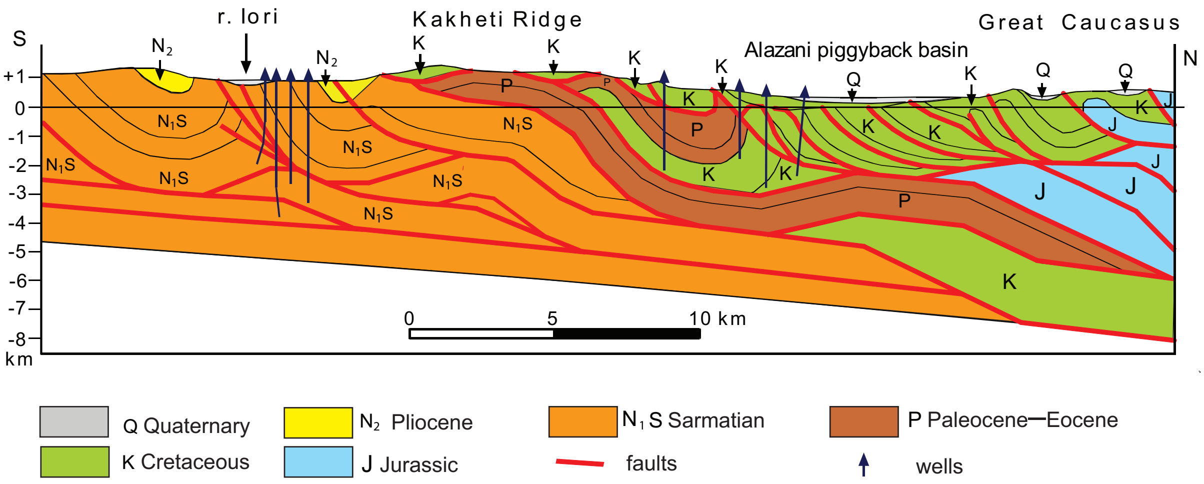

Figure 4. Geological profile across the Kakheti ridge compiled on the basis of geological mapping, seismic profiles and wells data.

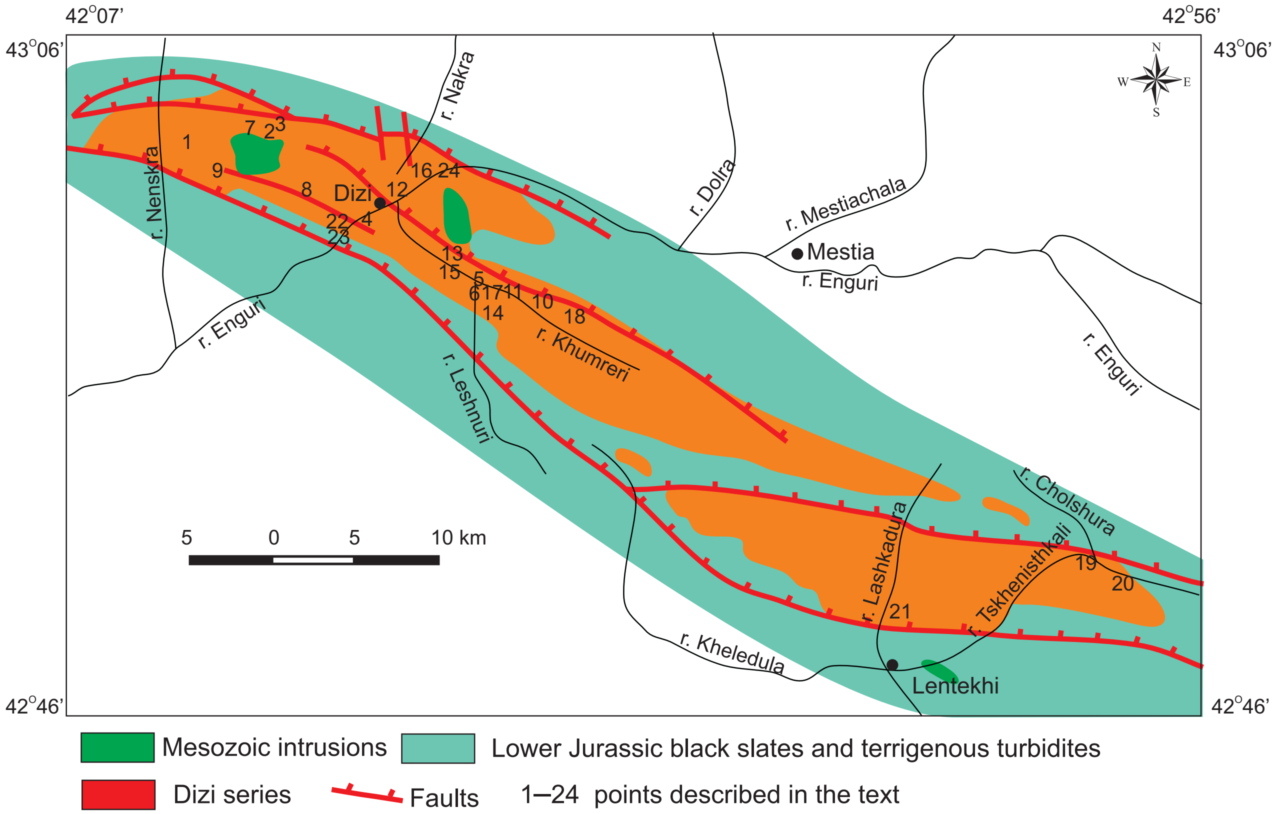

Figure 5. Schematic geological map of Upper Svaneti (Enguri gorge) and Lower Svaneti (Tskhenistskali gorge).

containing Norian corals (Slavin 1958; Baranov & Kropachev 1976). This formation was also penetrated by borehole as deep as 487 m (Prutsky & Lavrishchev 1988). Upper Triassic deposits are directly overlain by Lower Jurassic black slates, siltstones, and sandstones containing Sinemurian ammonites.

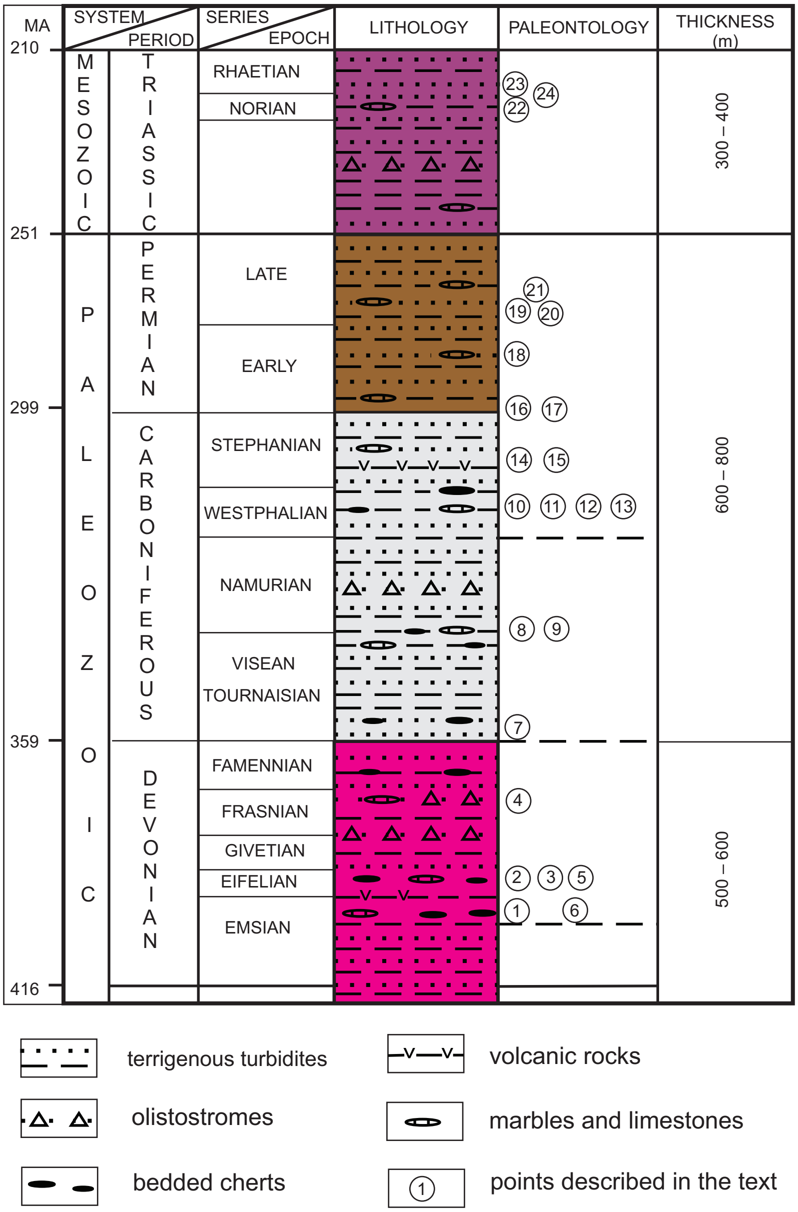

Figure 6. Generalized stratigraphic column of the Dizi series.

Speroza. Within the mountain Speroza, there are outcrops of recrystallized limestones spatially related to phyllites, black slates, siltstones and sandstones (see Figure 1b, Sz) bearing Late Permian (Tatarian= Upper Thuringian) brachiopods (Melnicov *et al.* 1973).

There is no evidence of Upper Palaeozoic (Variscan) metamorphic and magmatic events, folding and topographic inversion within the Southern Slope Zone. Variscan and Old Cimmerian orogenic events, which are strongly manifested to the south and to the north of the basin, did not lead to closure of the Dizi Basin (Adamia *et al.* 1980; Somin 2007). During the Middle–Late Palaeozoic and Triassic, the Dizi basin accumulated thick terrigenous turbidites, pelagic and hemipelagic deposits, andesite-dacite volcanoclastics, and carbonates under submarine environment on the northern continental slope of the Transcaucasian island-arc (Figure 2). Northward, along the northern margin of the Dizi Basin, the metamorphic complexes with tectonic slices of the Lower–Middle Palaeozoic oceanic crust and island-arc (Adamia *et al.* 1987a; Zakariadze *et al.* 2009) were formed (Chugush, Laba, Buulgen, and Kassar metabasite complexes). Obducted sheets of metaophiolites from the root zone located along the southern margin of the island arc were tectonically displaced north (Adamia *et al.* 2009; Somin 2007), towards the Fore Range Zone. Widespread emplacement and exhumation (Somin 2007) of microcline granite plutons and subaerial volcanism along the active continental margin of southern Eurasia at 330–280 Ma occurred above a north-dipping Palaeotethyan subduction zone (Adamia *et al.* 1987a). Early–Middle Palaeozoic age of the metabasic complexes of the accretional wedge is shown by the geochronological data (Somin 2007). On the basis of the stratigraphic position, they are dated as pre-Middle Carboniferous. According to palaeontological findings (bluish-green algae, corals, foraminifers, and crinoidea), the upper part of the Laba series is dated as post-Middle Ordovician (Kuznetsov & Miklukho-Maklai 1955; Adamia 1968; Potapenko & Stukalina 1971; Adamia *et al.* 1973; Somin & Vidiapin 1989).

The southern as well as the northern coastline of the Dizi back-arc basin are distinctly identified on the basis of transition of the deep-marine deposits to subaerial and shallow marine facies of the Great Caucasian and Transcaucasian island arcs (Figure 2, Adamia *et al*. 2003; Somin 2007). The Carboniferous along the northern margin of the Transcaucasian massif (Dzirula salient), is represented by the Chiatura suite composed of subaerial rhyolite-dacitic volcanic rocks while the Upper Triassic (Norian) by the Narula suite represented by basal conglomerates containing flora fossils and subaerial acidic volcanics (Svanidze *et al*. 2000; Adamia *et al*. 2003; Lebanidze *et al*. 2009). In the Main Range Zone of the Great Caucasus (northern coast of the Dizi Basin), the Carboniferous and Permian are represented by shallow-marine and continental molasse, the Triassic by variegated continental molasse (Gamkrelidze *et al*. 1963; Snejko 1968; Kruglov & Robinson 1968; Robinson 1968; Somin 1971; Khutsishvili 1972; Adamia *et al.* 1980, 2003; Belov *et al.* 1989).

The stratigraphic and structural relationships between the Liassic and Triassic of the Dizi series have turned out to be much more complicated than previously believed (Nutsubidze *et al.* 1963; Belov & Somin 1964; Adamia 1968). A detailed stratigraphic subdivision of the Dizi series on the basis of conodonts allowed to specify its structure more accurately. The widely accepted view that the Liassic is transgressive over various horizons of the Dizi series was not confirmed; in all cases the contacts seem to be reverse faults (see Figure 5). In many sections Sinemurian deposits conformably follow Upper Triassic ones (Adamia & Kutelia 1987; Adamia *et al.* 1990). In the majority of sections, clastic material is coarsening at the boundary of the Dizi series with the Sinemurian deposits: there appear coarse sandstones, gritstones and sometimes conglomerates. Clastic components are represented here almost exclusively by rocks of the Dizi series (Adamia 1968; Chikhradze 1979). It is characteristic that all over the Southern Slope Zone and in the Main Range of the Great Caucasus, composition of coarse-grained clastic deposits of the lower part of the Liassic is controlled by the composition of the directly underlying basement. Another important feature of the Dizi series seems to be different-polymictic composition of clastic components at similar levels, even in adjacent areas (Chikhradze 1979; Beridze 1983). This may be explained by local erosion of the upper level of islands – cordilleras (Kakhadze 1947; Slavin 1958; Adamia 1968).

The mountainous Crimea represents another basin of continuous deposition, where the Variscan and Old Cimmerian folding did not result in termination of sedimentation, and the Tauric series is generally considered as the western extension of the Great Caucasian basin (Kakhadze 1947; Muratov 1973; Slavin & Khain 1980; Adamia & Kutelia 1987). East of the Great Caucasus, a similar basin characterized by continuous Triassic–Jurassic marine sandy-argillaceous sedimentation was revealed by drilling in the region of the Great Balkans and Kopet Dag (Transcaspian). Thickness of the Mesozoic in this trough attains 13 km. This basin was continuously developing since the Palaeozoic (Amurski *et al* 1968; Slavin & Khain 1980).

Lower and Middle Jurassic Formations (Hettangian–Bathonian)

The Lower and Middle Jurassic in the axial part of the Great Caucasus back-arc is represented by black-slates, terrigenous turbidites, and basic volcanic formations (Gamkrelidze 1964; Andruschuk 1968; Azizbekov 1972). The presence of all Lower and Middle Jurassic biostratigraphic stages in this sequence is established on the basis of palaeontological (ammonites, belemnites, pelecypods, gastropods, brachiopods, corals, foraminifers etc) data (Kakhadze 1947; Bendukidze 1964; Nutsubidze 1964; Zesashvili 1964; Krimgolts 1968; Beridze *et al*. 1972; Khimshiashvili 1972; Nutsubidze *et al*. 1972; Paichadze *et al.* 1972a, b; Shikhalibeili & Agaev 1972; Panov 1976; Vashakidze 1985; Adamia *et al.* 1990; Topchishvili 2006; Topchishvili *et al.* 2006; Panov & Lomize 2007).

In the transitional Triassic–Lower Jurassic terrigenous turbidites E. Planderova discovered marine microfossils of Rhaethian and Hettangian ages (Adamia *et al.* 1990). Sinemurian terrigenous turbidites and volcanoclastics of the central part of the Southern Slope Zone conformably follow Triassic and Hettangian levels. However, Liassic deposits transgressively overlie basement rocks along the southern and northern margins of the Southern Slope Zone. Within the basin of the Great Caucasus, the Pliensbachian level of the Lower Jurassic is, generally, represented by black slate formation locally containing a major amount of diabase (Gamkrelidze 1964; Andruschuk 1968; Romanov 1968; Azizbekov 1972; Adamia 1977). Tholeiitic basalts of MORB-type (pillow lavas and agglomerates) are also present in this level (Beridze 1983; Chikhradze *et al.* 1984; Lordkipanidze *et al.* 1989). The Toarcian–Aalenian within the whole basin is composed of terrigenous proximal turbidites along its border with the Transcaucasian massif (island-arc) and distal turbidites in the axial part of the Southern Slope Zone (Beridze 1983; Tuchkova 2007). Paroxysms of volcanic activity were periodically evident throughout the Liassic that is confirmed by the presence of admixtures and intercalations of volcanic formations (basalts, keratophyres) in Sinemurian, Pliensbachian, Toarcian, and Aalenian deposits (Borsuk 1968; Romanov 1968; Abdullaev *et al*. 1972; Beridze *et al.* 1972; Adamia 1977; Beridze 1983; Agaev *et al*. 2003; Valiev 2003; Panov & Lomize 2007).

However, the Aalenian basalts show a slight tendency toward an arc-signature as indicated by lower Ti, Zr, Nb, Ni, and Cr contents and by Ti and Nb negative anomalies on the REE diagram (Lordkipanidze *et al.* 1989).

Bajocian lavas and volcanoclastics represented by calc-alkaline andesite-basalts are widespread along the southern margin of the Southern Slope Zone, at its border with the Transcaucasian massif (Gamkrelidze 1964; Andruschuk 1968). Northward, the Bajocian lavas and pyroclastics pass into tuff-turbidites, and farther on, into the terrigenous turbidites (Figure 7). Black slate, terrigene-turbiditic, and volcanogenic formations of the Southern Slope Zone (thickness more than 5000 m) (Gamkrelidze 1964; Andruschuk 1968; Panov & Lomize 2007; Tuchkova 2007) are dated predominantly on the basis of ammonites, pelecypods, and brachiopods (Kakhadze 1947; Balukhovski 1964; Nutsubidze 1964; Zesashvili 1964; Adamia *et al*. 1972; Beridze 1983; Topchishvili 2006; Panov & Lomize 2007).

The southern coastal facies of the Lower and Middle Jurassic of the Great Caucasus back-arc basin exposed in the Transcaucasian massif (Figure 8) unconformably rest upon Palaeozoic and Triassic rocks; they start with Sinemurian basal coarse-grained terrigenous deposits. Pliensbachian–Aalenian stages are represented by shallow-marine crinoidal limestones, Bajocian by island-arc volcanic formation, Bathonian by coal-bearing lacustrian terrigenous clastic rocks, Callovian by evaporites and alkali basaltic volcanic formation (Kakhadze 1947; Bendukidze 1964; Zesashvili 1964; Lordkipanidze *et al*. 1989).

The northern coastal facies of the back-arc basin are widespread in the Main Range, Laba-Malka, and other zones of the North Caucasus (Figure 1). Lower and middle parts of the Jurassic series of the Northern Slope of the Great Caucasus are mainly composed of terrigenous clastics (Ajgirei 1976; Tuchkova 2007). Shallow-marine facies containing marine fossils (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 (Krimgolts 1968; Panov 1976).

After the Bajocian, the back-arc basin environment of the Great Caucasus significantly changed: volcanic

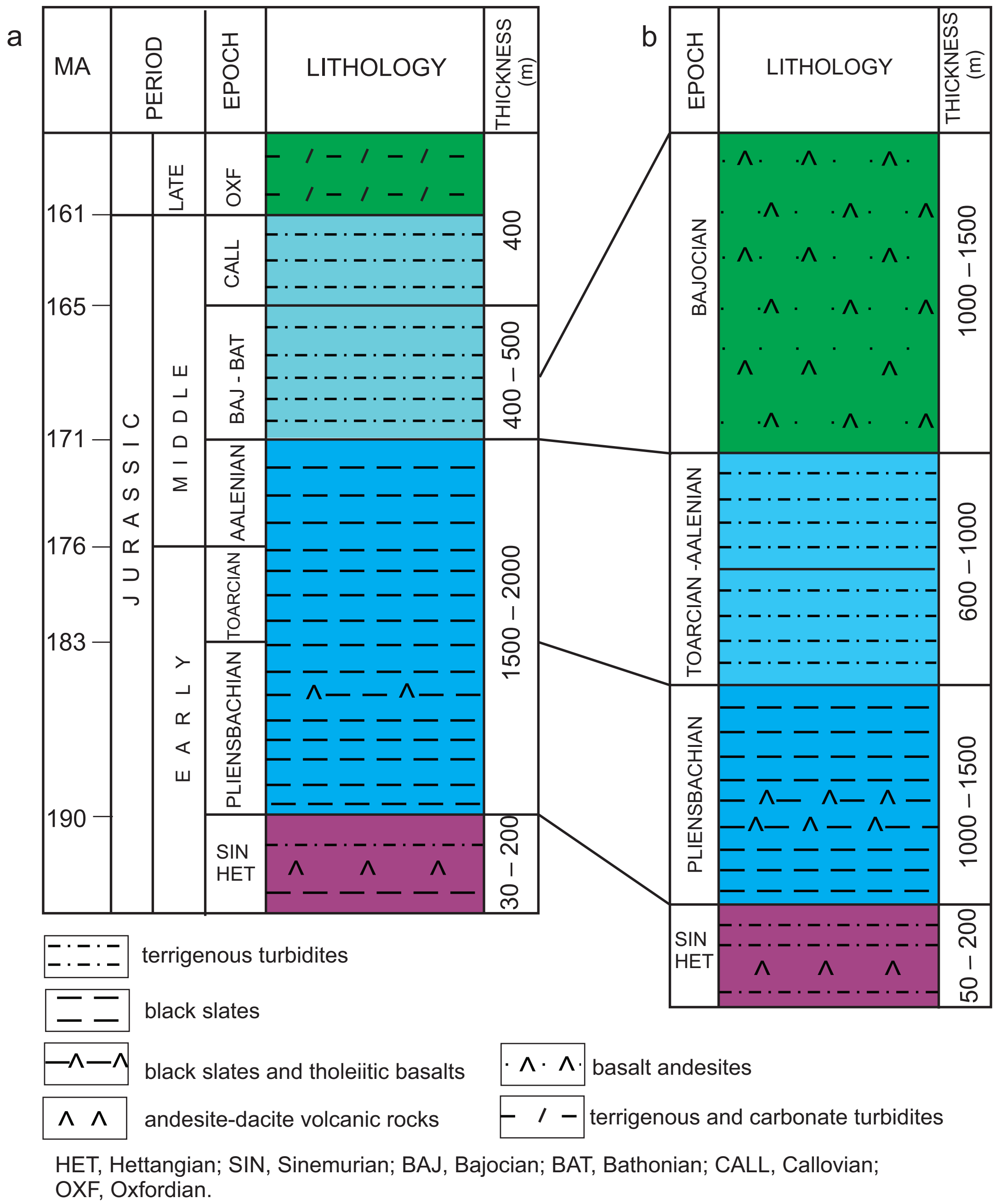

Figure 7. Generalised lithostratigraphic columns of Jurassic deposits of the central segment of the Southern Slope Zone: **(a)** northward and **(b)** southward of the Dizi series.

eruptions ceased and marine Bathonian, Callovian, and Lower Oxfordian terrigenous sandy-argillaceous sediments were deposited along the southern side of the basin , which pass northward into terrigenous and carbonate turbidites (Kikodze & Adamia 1972; Paichadze *et al.* 1972a, b; Adamia 1977). Two, en échelon-type flysch basins were forming: an eastern (Chiauri-Dibrar, Milanovsky & Khain 1963) and western (Novorossiisk and Abino-Gunai, Milanovsky & Khain 1963) ones, in which turbiditic sedimentation lasted almost uninterruptedly throughout the Late Jurassic, Cretaceous, Palaeocene,

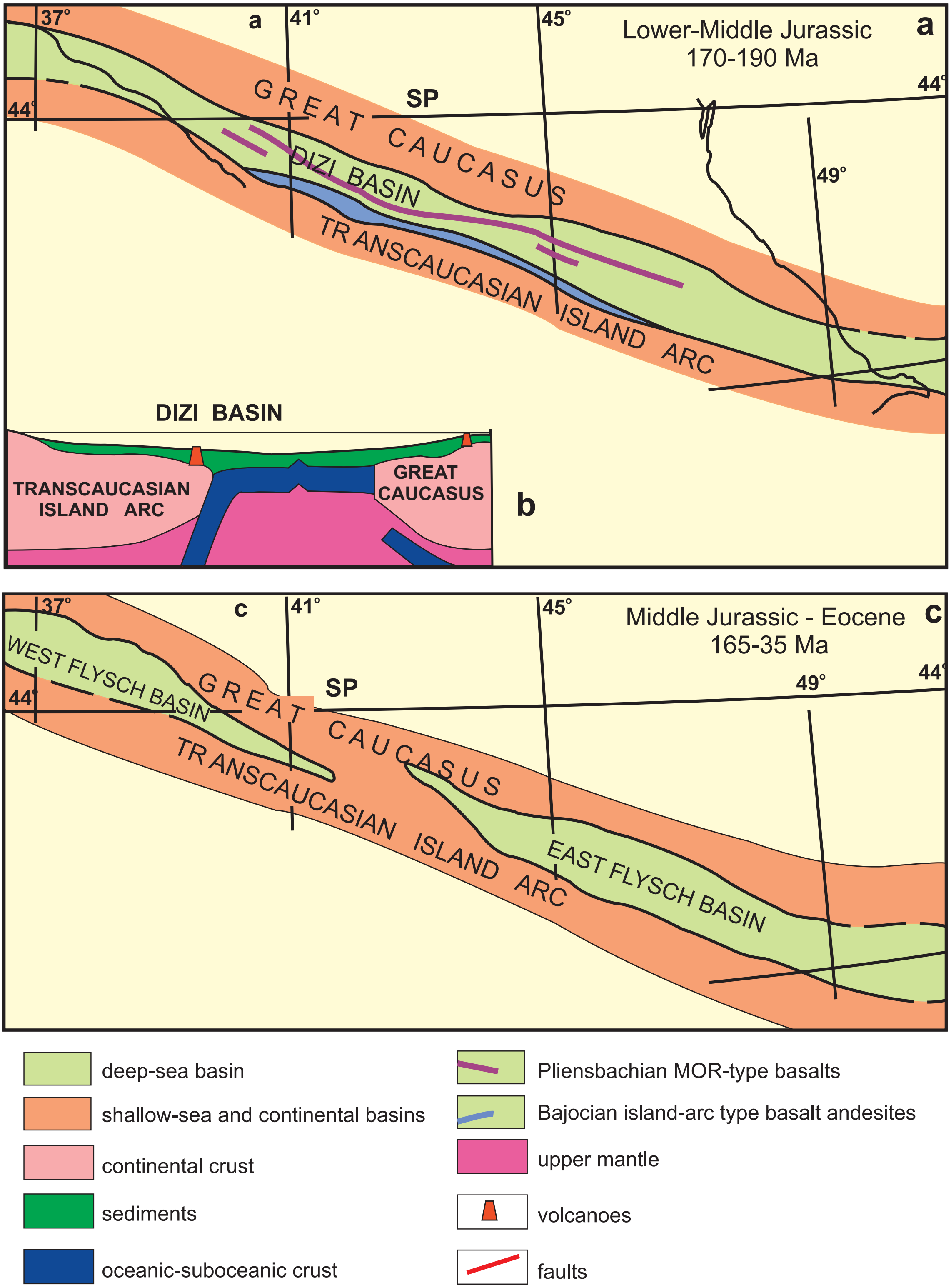

Figure 8. **(a)** Palaeotectonic map of the Great Caucasus for Early–Middle Jurassic, without palinspatic reconstruction. **(b)** Section across central segment of the Transcaucasus-Dizi-Great Caucasus (not in scale). **(c)** Palaeotectonic map of the Great Caucasus for Callovian–Eocene time, without palinspatic reconstruction.

and Eocene (Milanovsky & Khain 1963; Bendukidze 1964; Eristavi 1964; Kacharava 1964; Tsagareli 1964; Egoian *et al*. 1968; Jijchenko 1968; Krimgolts 1968; Alizade *et al*. 1972; Halilov & Aliev 1972; Shikhalibeili *et al*. 1972; Kokrashvili 1976; Sedenko 1976; Gambashidze 1979; Varsimashvili 2004; Lominadze *et al.* 2006). Here, thick formations of predominantly carbonate turbidites (Oxfordian–Valanginian, Turonian–Senonian) alternate with Hauterivian–Cenomanian, Palaeocene–Eocene terrigenous turbidites and Maastrichtian, Priabonian olistostromes (Mrevlishvili 1957; Leonov 2007). Total thickness of the entire Jurassic–Cretaceous–Palaeogene 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. Nappes with minor displacement are known only in the southernmost stripe of the Southern Slope Zone (see Figures 3b & 4).

During the Albian–Cenomanian, the eastern and western peripheries of the basin show weak, within-plate-type alkali-basaltic magmatism (Borsuk 1979; Lordkipanidze *et al.* 1989).

All stratigraphic stages from the Callovian to the Upper Eocene are shown on the basis of fossil molluscs, brachiopods, foraminifers, corals etc (*Jurassic*: Kakhadze 1947; Bendukidze 1964; Krimgolts 1968; Khimshiashvili 1972; Nutsubidze *et al*. 1972; Paichadze *et al.* 1972a, b; Shikhalibeili & Agaev 1972; Kokrashvili 1976; Lominadze *et al.* 2006. *Lower Cretaceous*: Eristavi 1964; Egoian 1968; Halilov & Aliev 1972; Kotetishvili 1986; Kvantaliani 1989; Kakabadze & Sharikadze 2004; *Upper Cretaceous*: Tsagareli 1963, 1964; Aliev & Aliev 1972; Gambashidze 1979; *Palaeogene*: Kacharava 1964; Jijchenko & Reznicov 1968; Mrevlishvili 1978; Salukvadze 2000; Alizade *et al.* 1972).

The Callovian–Eocene sequence of the southern coastal area of the Great Caucasus back-arc basin is exposed directly along the margin of the basin (see Figure 8). Callovian–Tithonian deposits are represented by evaporites, alkaline basaltic volcanic formation, and coral limestones (Bendukidze 1964; Khimshiashvili 1972; Lominadze *et al.* 2006); Cretaceous– by platformal limestones; Palaeocene–Eocene– by shallow-marine limestones and marls (Eristavi 1964; Kacharava 1964; Gambashidze 1979; Salukvadze 2000).

The next time-stage of the development of the northern coast of the Great Caucasian back-arc basin (the Northern Slope of the Great Caucasus Scythian platform) begins in the Callovian. The Callovian to Eocene period was that of a shallow-marine environment. The base of the Callovian is represented by the basal formation that marks a beginning of the tectonically induced sea transgression. The Callovian is usually represented by terrigeneous-carbonate shallow-water deposits containing ammonites, bivalve mollusks, and gastropods. The Oxfordian and Kimmeridgian are dominated by dolomites and reef limestones. The Upper Jurassic section is topped by the predominantly lagoonal facies represented by terrigeneous-carbonate salt-bearing variegated formation. Thickness of the Callovian–Tithonian deposits ranges from some tens of meters to 2300 m (Krimgolts 1968; Shikhalibeili & Agaev 1972).

Cretaceous deposits of the Northern Slope Basin show very diverse facies. Almost complete sections pass into ones showing a large number of stratigraphic unconformities induced by regression and transgression of the sea; their thickness varies from tens to some thousands 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 abundant fossils (Egoian 1968; Kancheli 1968; Halilov & Aliev 1972; Sedenko 1976). The Upper Cretaceous at the extreme west of the Northern Slope is represented by 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 a shallow marine basin. Terrigenous clastic rocks, including glauconitic sandstones characterize the Cenomanian. Glauconite is, generally, characteristic of the Upper Cretaceous–Lower Palaeocene deposits (Tsagareli 1963; Egoian *et al.* 1968; Aliev & Aliev 1972). The Palaeocene–Eocene deposits were formed in a vast shelf along the southern margin of the Eastern European continent. Marly, argillaceous-marly, and sandy-marly facies (thickness of some hundred metres) were accumulated on the shelf dated, predominantly, by foraminifers and echinoids (Jijchenko 1968; Alizade *et al.* 1972; Salukvadze 2000).

Palaeomagnetic data allow a Late Cretaceous palinspastic reconstruction for the Great Caucasus (Bazhenov & Burtman 1990, 2002; Asanidze *et al.* 2009). Width of the Great Caucasian Late Cretaceous basin, as reported by Bazhenov & Burtman (1990, 2002) is estimated as 900–400 km. Similar results (~1000 km) were obtained by Asanidze *et al.* (2009). On certain assumptions of equal-area lithospheric thickening and size of lithospheric root a Late Cenozoic shortening of 200–300 km is estimated in the Great Caucasus (Ershov *et al.* 2003).

Late Cenozoic: Syn- and Post-Collisional Stages

The Oligocene (or the latest Eocene–Oligocene epoch) is traditionally considered as a beginning of syn-collisional (or orogenic) stage of development of the Caucasus (for example, Milanovsky & Khain 1963; Gamkrelidze 1964; Laliev 1964; Andruschuk 1968; Azizbekov 1972). During syn- and post-collisional stages (Oligocene–Neogene–Quaternary) as a result of collision between the Africa-Arabian and Eurasiatic plates, inversion of relief took place and the back-arc basin was transformed into the fold-and-thrust mountain belt of the Great Caucasus.

The Oligocene–Lower Miocene deposits resulting from accumulation in hemi-closed basins of the Paratethys are represented by sandy-argillaceous gypsiferous facies known as the Maikopian series (lower- or fine-grained molasse). Accumulation of molasse with a maximum thickness several km, predominantly shallow-sea and lagoon-lacustrine terrigenous clastics with layers of organogenic carbonate rocks (coquina), lasted throughout most of the Miocene. Only at the end of the Miocene (the Tortonian, ~7 Ma), shallow-marine environment was replaced by subaerial one and simultaneously clastic material became coarser (upper- or coarse-grained molasse). Marine environment of sedimentation remained only within the territories adjacent to the Black, Azov, and Caspian Sea basins.

Maikopian deposits almost continuously surround the Great Caucasian fold-and-thrust mountain belt. In some places, they conformably follow the sandy-clayey Upper Eocene deposits, for example, in the Kinta and Thelathgori suites in the southernmost stripe of the Central Caucasian Southern Slope Zone (Milanovsky & Khain 1963; Laliev 1964), the Khadumian suite at the western and eastern terminations of the Great Caucasian Mountains (Jijchenko & Reznikov 1968; Alizade *et al.* 1972). Outcrops of the Upper Sarmatian shallow-marine rocks are reported in the Shakhdag mountains – on the crest of the eastern Great Caucasus at about 3500 m above sea (Budagov 1964).

The Great Caucasus in its central mostly uplifted segments includes Pliocene–Quaternary volcanic and plutonic formations. Products of the 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; Adamia *et al.* 2008). Neogene–Quaternary intrusives of the Great Caucasus also crop out in the same regions and are represented by hypabyssal bodies. This volcanic-plutonic formation is Late Pliocene–Quaternary in age (Borsuk 1979; Chernishev *et al.* 1999).

The Black Sea and Caspian Sea basins unconformably overlie different structures of adjacent land; their shoreline cuts several main tectonic units of the Caucasus and Crimea. According to marine geophysical research, submarine prolongation of the Southern Slope Zone of the Great Caucasus runs along the northern, pre-Caucasian coastal stripe (Terekhov & Shimkus 1989). The eastern Black Sea high– the Shatsky Ridge, is considered as submarine prolongation of the Georgian Block (Tugolesov *et al.* 1984; Afanasenkov *et al.* 2007).

Conclusions

The Great Caucasus Mountains represent an intracontinental tectonic system resulting from the Late Cenozoic structural inversion of a Palaeozoic–Mesozoic–Early Cenozoic back-arc basin in response to the convergence of the Africa-Arabian and Eurasian lithosphere plates.

Palaeozoic–Mesozoic–Early Cenozoic basin of the Great Caucasus was developing, at least, from Devonian. The presence of basinal deposits of all biostratigraphic stages beginning from the Lower–Middle Devonian and including Eocene are established on the basis of palaeontological data.

Variscan and Old Cimmerian events strongly manifested to the south and to the north of the Dizi Basin did not lead to closure of the basin. A similar basin with continuous deposition, where Variscan and Old Cimmerian folding did not result in termination of the basinal environment, is the mountainous Crimea generally considered to be the western extension of the Great Caucasian basin. East of the Caucasus, such a basin characterized by continuous Triassic–Jurassic marine sandy-argillaceous sedimentation is the Great Balkans (Transcaspian). This basin was continuously developing since the Palaeozoic.

Bazhenov & Burtman (2002) present Late Cretaceous palinspastic reconstruction for the Great Caucasus based on palaeomagmetic data. Width of the Great Caucasian Late Cretaceous basin according to these authors is estimated as 1000–400 km, on the basis of equal-area estimation of lithospheric thickening and size of lithospheric root of the Great Caucasus its Late Cenozoic shortening is estimated as 200–300 km. During syn- and post-collisional stages (Oligocene–Quaternary) collision between the Africa-Arabian and Eurasiatic plates initiated inversion of the relief and at the place of back-arc basin was formed fold-and-thrust mountain belt of the Great Caucasus. Maikopian deposits almost continuously surround the Great Caucasian fold-and-thrust mountain belt. In some places, they conformably follow the sandy-clayey Upper Eocene deposits. Neogene–Quaternary volcanic rocks of the Great Caucasus are represented by subaerial formations, volcanic eruptions occurred during Pliocene–Pleistocene and Early Holocene time. The Black Sea basin and also Caspian Sea basin unconformably overlie different structures of adjacent land; its shoreline cuts several main tectonic units of the Caucasus and Crimea. According to marine geophysical research, the western submarine prolongation of the Southern Slope Zone of the Great Caucasus runs along the northern, pre-Caucasian coastal stripe.

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