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Subductions, obduction and collision in the Lesser Caucasus (Armenia, Azerbaijan, Georgia), new insights

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Abstract: In the Lesser Caucasus three main domains are distinguished from SW to NE: (1) the autochthonous South Armenian Block (SAB), a Gondwana-derived terrane; (2) the ophiolitic Sevan–Akera suture zone; and (3) the Eurasian plate. Based on our field work, new stratigraphical, petrological, geochemical and geochronological data combined with previous data we present new insights on the subduction, obduction and collision processes recorded in the Lesser Caucasus. Two subductions are clearly identified, one related to the Neotethys subduction beneath the Eurasian margin and one intra-oceanic (SSZ) responsible for the opening of a back-arc basin which corresponds to the ophiolites of the Lesser Caucasus. The obduction occurred during the Late Coniacian to Santonian and is responsible for the widespread ophiolitic nappe outcrop in front of the suture zone. Following the subduction of oceanic lithosphere remnants under Eurasia, the collision of the SAB with Eurasia started during the Paleocene, producing 1) folding of ophiolites, arc and Upper Cretaceous formations (Transcaucasus massif to Karabakh); 2) thrusting toward SW; and 3) a foreland basin in front of the belt. Upper–Middle Eocene series unconformably cover the three domains. From Eocene to Miocene as a result of the Arabian plate collision with the SAB to the South, southward propagation of shortening featured by folding and thrusting occurred all along the belt. These deformations are sealed by a thick sequence of unconformable Miocene to Quaternary clastic and volcanic rocks of debated origin.

The Mesozoic ophiolites of the Lesser Caucasus belong to the Tethyan ophiolitic suture zone (Knipper 1975; Adamia *et al.* 1981; Zakariadze *et al.* 1983; Knipper *et al.* 1986) (Fig. 1). Consequently the Lesser Caucasus mountain belt recorded Mesozoic and Cenozoic geodynamic evolutions related to the closure of the Northern Neotethys ocean (Sengör & Yilmaz 1981; Adamia *et al.* 1981; Dercourt *et al.* 1986; Ricou 1994; Nikishin *et al.* 1998; Yilmaz *et al.* 2000; Stampfi *et al.* 2001; Robertson 2002; Golonka 2004). In this context, the NW–SE trending Sevan–Akera suture zone extends on nearly 400 km, north of the Sevan Lake from Amassia to the Nagorno-Karabagh region. To the SE of this region it is covered by Quaternary deposits along the Araks valley (Fig. 2). The outcrops of this arched zone reach a maximum width of 25 km and are characterized

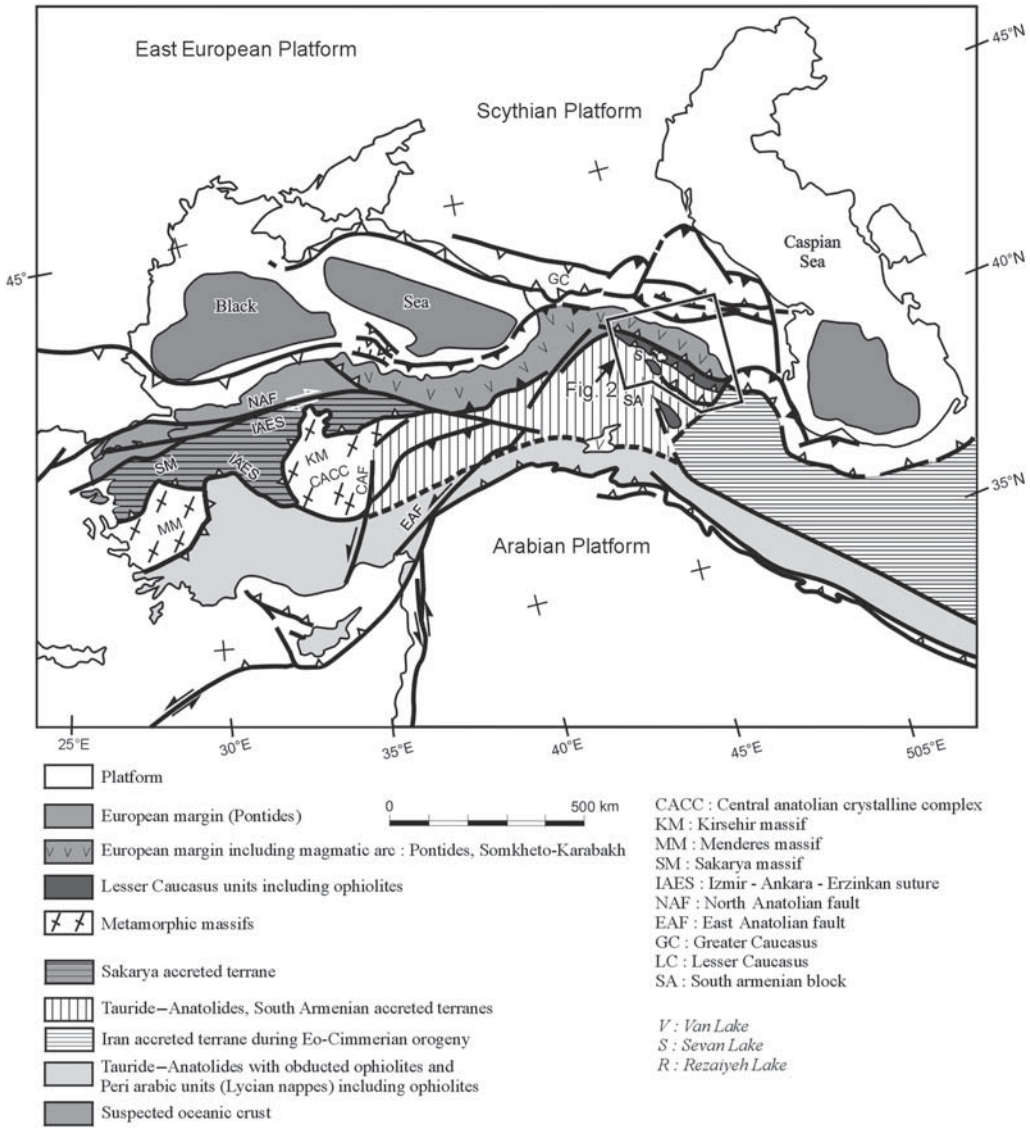

Fig. 1. Structural sketch map of the Tauride–Anatolides, Caucausus and Iranian belts. Location of Figure 2 is indicated.

by many discontinuous massifs, exposing ultrabasites, gabbros, basalts and pelagic sedimentary rocks (Knipper 1975; Gevorkyan & Gevorkyan 2003; Aghamalyan 2004; Melikyan 2004) (Fig. 2). The location of this ophiolitic zone between Eurasia and the Daralagez continental terrane or South Armenian Block (SAB; a continental terrane of Gondwanian origin, Knipper & Khain 1980; Monin & Zonenshain 1987) suggests either an allochthonous unit or a real suture zone within the Lesser Caucasus belt. According to Aghamalyan (1996, 2004), the Sevan–Akera ophiolitic zone corresponds to a suture zone where Eurasian and a microplate plates have joined together. However, other outcrops of ophiolites (the Vedi and Zangezur ophiolites) are known southwestward of the Sevan–Akera zone (Fig. 2). Consequently the ophiolites of the Lesser Caucasus could correspond to several sutures zones (Aslanyan & Satian 1977) or only one reflecting a complex west-Pacific type evolution of the Neotethys Ocean east of the Erzincan suture zone (Eastern Anatolia, Turkey) and west of the Iran ophiolitic ones (Knipper 1975; Adamia *et al.* 1981; Lordkipanidze *et al.* 1988; Yilmaz *et al.* 2000).

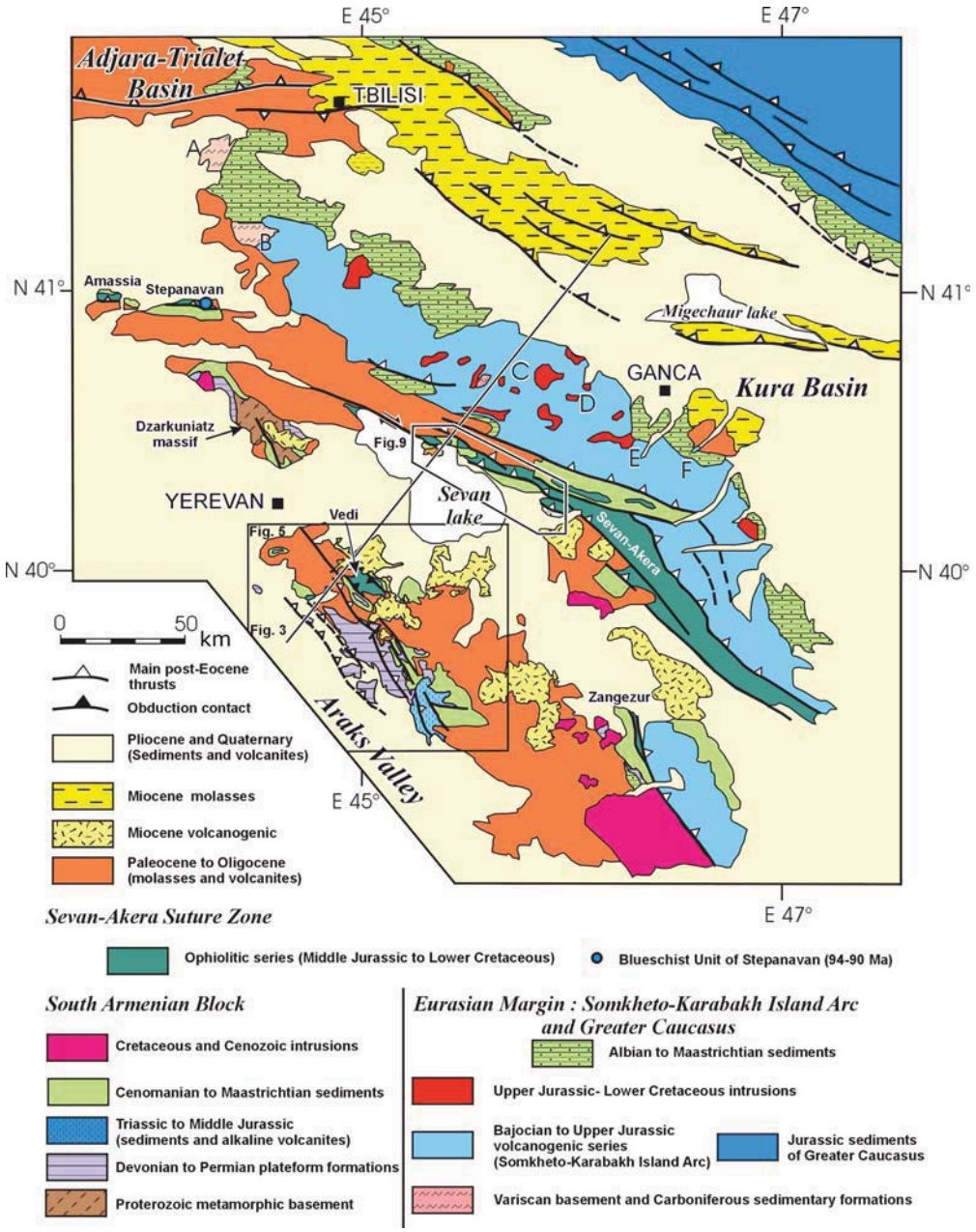

Fig. 2. Structural map of the Lesser Caucasus, modified from Nalivkin (1976) and including our observations. A, Variscan Khrami massif and its Carboniferous volcano-sedimentary cover (Georgia); B, Variscan Loki Massif (Georgia); C, Zayemchay Valley (Azerbaijan); D, Gochkarchay Valley (Azerbaijan); E, Yasamal Valley (Azerbaijan); F, Kurakchay Valley (Azerbaijan), all these locations are discussed in the text.

In the Lesser Caucasus the apparent southwestward obduction of oceanic lithosphere on the South Armenian Block (Knipper 1975; Knipper & Khain 1980; Monin & Zonenshain 1987) occurred during the Late Cretaceous (Sokolov 1977; Knipper & Khain 1980). It was classically admitted that the obduction was directly related to the collision between the SAB and Eurasia. As a result

the Eurasian plate overthrusted the ophiolites southwards (Adamia *et al.* 1977, 1981).

As for many examples around the world, in such structural setting, occurrences of ophiolites feature the obduction and collision stages of geodynamic evolution of mountain belts. Where the suture zone units have been preserved from metamorphism due to the collision, they may give significant information on the initial stages as the geodynamic setting of their formation and the subduction.

This belt was deformed by superimposed tectonics which makes it difficult to decipher each structure and consequently the geodynamic processes from which they resulted. Accordingly, the main debated subjects on the Lesser Caucasus are: 1) the occurrence or not of three main ophiolitic zones corresponding for some authors to three main sutures zones; 2) the modalities and timing of collision stages between the SAB and Eurasia, including the verging of continental subduction and age of its triggering; and 3) the origin of the widespread magmatic activity that occurred during Eocene time and then from the Oligocene–Miocene to the Quaternary. New investigations were deemed necessary to solve these questions all related to the geodynamic evolution of the Neotethys Ocean and its closure.

According to new field surveys (mapping and acquisition of structural data) and new analytic data (Ar/Ar dating, geochemistry, micropalaeontological dating: Tables 1 and 2) obtained thanks to the MEBE programme support, we discuss

Table 1. *Upper Cretaceous Nannofossils dating with (WGS84) GPS locations*

Location	Sample number	Latitude	Longitude	Elevation m	Age	Nannofossil assemblages
Vedi Valley	AR S 16	N 39°96551	E 44°92178	1392	Late Cenomanian	*Corollithion exiguum*, *Eiffellithus turriseiffeli*, *Eprolithus floralis*, *Manivitella pemmatoidea*, *Podorhabdus albianus*, *Prediscosphaera cretacea*, *Watznaueria barnesae*, *W. biporta*
Avazan Valley North Sevan Lake	AR 14 05	N 40°32581	E 45°70898	2291	Santonian	*Watznaueria barnesae*, *Micula staurophora* *Eiffellithus turriseffeli*, *Eiffellithus eximius* *Eprolithus floralis*, *Prediscosphaera cretacea* *Zygodiscus diplogrammus*, *Marthasterites furcatus* *Tranolithus orionatus*, *Reinhardtites anthophorus* *Quadrum gartneri*, *Lucianorhabdus cayeuxii*
	AR 17 05	N 40°32581	E 45°70898	2291		
	AR 21 05	N 40°32921	E 45°72668	2459		
	AR 22 05	N 40°32921	E 45°72668	2459		
	AR 94 05	N 40°32461	E 45°69376	2242		
	AR 100 05	N 40°32461	E 45°69376	2242		
Vedi Valley	AR 63 05	N 39°96291	E 44°94482	1385		
	AR 64 05	N 39°96291	E 44°94482	1385		
	AR 65 05	N 39°96291	E 44°94482	1385		
	AR 67 05	N 39°96291	E 44°94482	1385		
	AR 72 05	N 39°93349	E 45°01987	1915		
	AR 73 05	N 39°93349	E 45°01987	1915		
Avazan valley NE Sevan Lake	AR 20 05	N 40°32529	E 45°71949	2371	Late Santonian	*Watznaueria barnesae*, *Reinhardtites anthophorus* *Micula staurophora*, *Lucianorhabdus cayeuxii* *Eiffellithus eximius*, *Prediscosphaera cretacea* *Lithastrinus grillii*, *Marthasterites furcatus* *Quadrum gartneri*, *Broinsonia parca expansa*
Dali Valley North Sevan Lake	AR 39 05	N 40°48414	E 45°42111	2118		
Artanish Valley North Sevan Lake	AR 01 05	N 40°51028	E 45°37118	2021	Early Campanian	*Gartnerago obliquum*, *Eiffellithus eximius* *Eiffellithus turriseiffeli*, *Micula staurophora* *Prediscosphaera cretacea*, *Broinsonia parca*
	AR 04 05	N 40°51456	E 45°37415	2092	Late Campanian	*Arkhangelskiella cymbiformis*, *Broinsonia parca*, *Eiffellithus eximius*, *Quadrum gothicum*, *Q. trifidum*, *Reinhardtites anthophorus*

Table 2. *Paleocene–Eocene Nannofossils dating with (WGS84) GPS locations*

Location	Sample number	Latitude	Longitude	Elevation m	Age	Nanofossil assemblages
Artanish valley North Sevan Lake	AR 03 05	N 40°51343	E 45°37745	2048	Late Eocene, NP 18	*Coccolithus pelagicus*, *Sphenolithus radians* *Sphenolithus moriformis*, *Dictyococcites dictyodus* *Reticulofenestra umbilica*, *Cyclococcolithus formosus* *Cyclicargolithus floridanus*, *Chiasmolithus oamaruensis*
	AR 07 05	N 40°51812	E 45°37207	2236	Late–Middle Eocene to Late Eocene	*Coccolithus pelagicus*, *Sphenolithus radians* *Sphenolithus moriformis*, *Dictyococcites dictyodus* *Reticulofenestra umbilica*, *Cyclococcolithus formosus* *Cyclicargolithus floridanus*
	AR 08 05	N 40°52080	E 45°36939	2308		
Khosrov valley	AR 54 05	N 39°93931	E 44°86781	1127	Late Paleocene NP 5	*Coccolithus pelagicus*, *Ericsonia subpertusa* *Chiasmolithus danicus*, *Fasciculithus tympaniformis* *Cruciplacolithus tenuis*
	AR 101 05	N 40°00623	E 44°90870	1564	Early Paleocene NP 3	*Coccolithus pelagicus*, *Ericsonia subpertusa* *Cruciplacolithus tenuis*, *Zygodiscus sigmoides* *Chiasmolithus danicus*
	AR 107 05	N 39°98047	E 44°88994	1460		

here some key points allowing us to clarify the subduction, obduction and collision processes all along the Sevan–Akera suture zone. Moreover we present a new interpretation for the structure of the Lesser Caucasus belt along a NE–SW cross-section.

General structural setting

As presented on the structural sketch map (Fig. 2) and the NE–SW trending cross-section (Fig. 3) summarizing previous data (Nalivkin 1976) and our observations, three main lithostructural domains can be distinguished in the Lesser Caucasus, from NE to SW: 1) Eurasia; 2) the Sevan–Akera suture zone; and 3) the South Armenian Block (SAB). This structural framework was previously proposed by Milanovski (1968). The main difference between the Milanovski's section and ours is the occurrence of ophiolitic nappes (the Vedi ophiolites) in front of the Sevan–Akera ophiolitic zone (Fig. 3). Satian *et al.* (1979, 2005) consider the Vedi ophiolites as remnants of a small palaeo-ocean. Knipper & Sokolov (1976) and Sokolov (1977) provided evidence for an allochthon position of the ophiolitic unit on the base of structural data. Others differences concern the occurrence of post-Eocene reverse and thrust faults which accommodated shortening in the basement of the SAB and the Sevan–Akera ophiolitic zone (Figs 2 & 3).

Evidence for subduction processes

Subduction along the Eurasian margin. As documented by previous studies performed on the Lesser Caucasus the subduction of the Tethys is evidenced by a thick and mainly calcalkaline volcanogenic and volcanoclastic series dated as Bajocian to Santonian (e.g. Adamia *et al.* 1981 for a review). At this period of time the northern Lesser Caucasus was characterized by an island arc domain called the Somkheto-Karabakh Island Arc (Knipper 1975; Adamia *et al.* 1977, 1987; Ricou *et al.* 1986).

The basement formations are quite similar to those known all along the Eurasian margin. Indeed they are exposed in three Variscan metamorphic massifs in Georgia (Transcaucasian massif): the Khrami (A on Fig. 2), Loki (B on Fig. 2) and Dzirula salients (Kazakhashvili 1950; Adamia 1968; Kekelia & Khutsishvili 1980; Adamia *et al.* 1983, 1987; Adamia 1984; Shengelia *et al.* 1989; Abezadze *et al.* 2002). In the Khrami massif the Carboniferous volcanoclastic series unconformably overlie the Variscan metamorphic basement (Adamia 1968, 1984; Adamia *et al.* 1983). This volcanogenic-sedimentary formation is transgressively overlain by the Liassic, Upper Jurassic–Lower Cretaceous and Cenomanian sediments and volcanogenic series. Upper Palaeozoic metamorphic rocks are also described in the Azerbaijan part of the Northeast Lesser Caucasus flank (Nalivkin

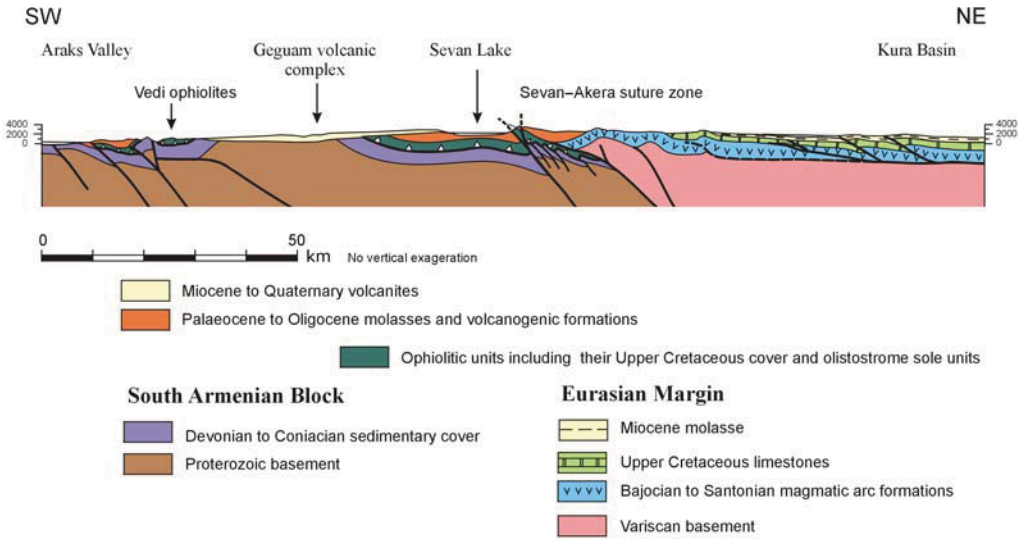

Fig. 3. Cross-section of the Lesser Caucasus. Location is shown on Figure 2.

1976) (see mark C on Fig. 2). We studied the Mesozoic volcanogenic series in four valleys near Ganca (Azerbaijan) (Fig. 2) from west to east: Zayemchay (C on Fig. 2), Gochkarchay (D on Fig. 2), Yasamal (E on Fig. 2), Kurakchay (F on Fig. 2). These observations allow us to present a synthetic stratigraphic log of the Eurasian margin of the northeastern part of Lesser Caucasus (Fig. 4). Bajocian–Bathonian formations ($\pm$2400 m) present a basal conglomerate on the Variscan basement (see mark A on Fig. 2) and upwards volcanoclastic series, sandstones, marl with Ammonites and Cancellophycus, basalts, volcanoclastic turbidites and andesite on top. This series indicates a major magmatic activity during the Middle Jurassic (Maghakyan *et al.* 1985; Lordkipanidze *et al.* 1988). The Oxfordian stage is composed of a transgressive sequence with conglomerate, siltstones, mudstones and reef limestones (Fig. 4). The Kimmeridgian features massive intrusions (granodiorites, gabbros, quartz-diorites, plagiogranites) (Melkonian 1976; Lordkipanidze *et al.* 1988). The gabbroic Dachkasan intrusion is responsible for the metamorphism of limestones into white marbles (Fig. 4). The intrusions caused many ore-deposits: 1) Middle and Upper Jurassic epigenetic copper and barite-polymetallic deposits; 2) Middle Jurassic gold-bearing porphyry copper deposits; 3) Upper Jurassic iron skarn deposits; and 4) Lower Cretaceous porphyry copper deposits (Kekelia *et al.* 2004; Melkonian & Akopyan 2006; Babazadeh *et al.* 2007). Coniacian formations unconformably overlie (unconformity U2, mark E on Fig. 2) the Upper Jurassic intrusions and Lower Cretaceous sedimentary rocks that are very much reduced in thickness in that part of the belt (Fig. 4). The Coniacian stage consists from bottom to top of a red conglomerate, sandstones and reef limestones indicating a shallow water environment at that time. A similar palaeoenvironment is also featured by the Santonian formations. They are characterized by a transgressive sequence (conglomerate, sandstones and limestones) including prismatic basalt flows. In some places, the series contains on its top pillowed and massive basalt lava flows. Consequently the magmatic arc was still active during this period of time along this part of the Eurasian margin. The end of magmatic activity occurred in the Campanian to Maastrichtian stages. The formations are made of thin layered pelagic limestones with some carbonates turbidites.

The compiled lithostratigraphic log (Fig. 4) characterizes an arc-type evolution of the active Eurasian margin from Bajocian to Late Jurassic all along the Lesser Caucasus northern flank. In this part of the Eurasian margin some ages of volcanogenic series (Late Cretaceous) are younger than in the northwestern part (Transcaucasian massif, Georgia). The southeastward younging of magmatic ages may result from migration of the magmatic arc during the Late Cretaceous.

On this basis we suggest that an oceanic plate subducted northward (if we exclude any major rotation since this time) beneath the Eurasian plate. At least it is well evidenced from Bajocian to Santonian (100 Ma), with younger ages present to the SE. This oceanic plate can be correlated to the northern Neotethys part or to one marginal basin related to an intra-oceanic subduction. In

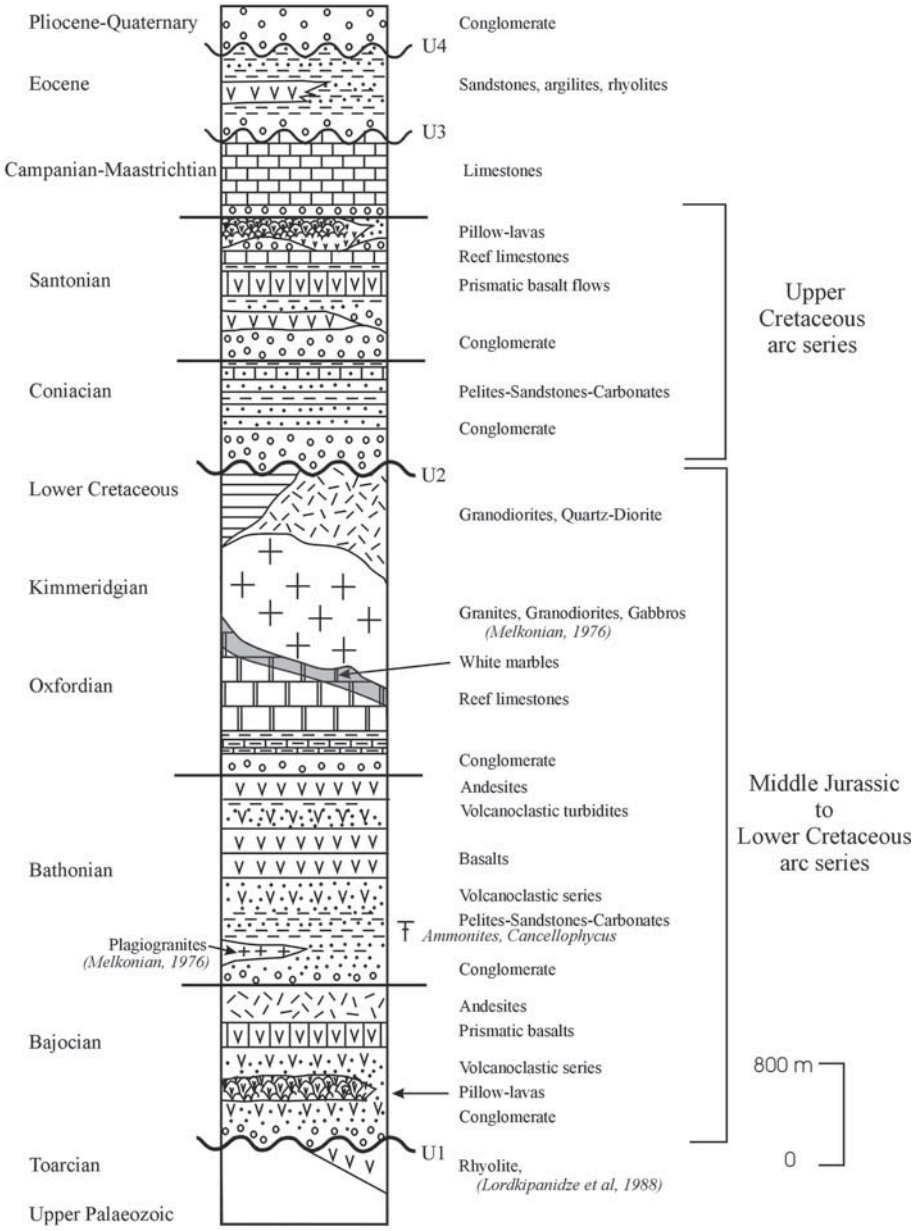

Fig. 4. Synthetic lithostratigraphic log of the Eurasian margin in the Lesser Caucasus (Armenia and Azerbaijan).

order to test these hypotheses we present below new evidences for an intra-oceanic subduction in the Neotethys domain during these times.

Evidence for an intra-oceanic subduction

According to petrological and geochemical data from volcanites, Lordkipanidze *et al.* (1988) suggested a Middle Jurassic Marianas-type subduction in the Lesser Caucasus region. New evidence for an intra-oceanic subduction is described in detail in a companion paper (Rolland *et al.* 2010), consequently we point out and summarize here the main results.

The northwestern part of the Sevan–Akera suture zone (SW of the Stepanavan town, Fig. 2)

is characterized by tectonic units outcropping within a nappe anticline in which the core is made of blueschist slices (Aghamalyan 1981, 2004; Galoyan *et al.* 2007; Galoyan 2008; Rolland *et al.* 2007). The blueschist units protoliths correspond to unroofed sediments along a subduction zone. The HP-LT metamorphism was dated 90–80 Ma by K–Ar methods (Aghamalyan 1998). Further, it was found to be a two-stage evolution using Ar/Ar dating technique (on phengites), with HP-LT stage at 94–90 Ma followed by a retrograde Green Schist/epidote amphibolite metamorphism during exhumation in the Late Cretaceous (71–74 Ma, Ar/Ar on phengites, Rolland *et al.* 2007, 2010). The blueschist unit is overthrusted by unmetamorphosed ophiolites, Late Jurassic in age (Danelian *et al.* 2007). In that area and in the other ophiolitic zones of the Lesser Caucasus (Sevan–Akera, Vedi), this geodynamical setting is also well documented by the magmatic evolution of the ophiolites. Petrologically and geochemically (major, trace, REE and isotopic geochemistry) these ophiolites are characterized by a slight island arc signature (Rolland *et al.* 2009). Such geochemical features are typical of oceanic crust formed in a back-arc setting (Galoyan 2008; Galoyan *et al.* 2009). Above the ophiolites, an Upper Cretaceous calc-alkaline volcanic series is evidenced, lying on a sedimentary unconformity marked by Albian conglomerates and limestones (Aghamalyan 1998; Galoyan *et al.* 2007). The geochemical data, including isotopic Sr–Nd compositions, obtained on these lavas indicate a volcanic arc-type series on top of the ophiolites (Galoyan *et al.* 2007, 2009; Galoyan 2008; Rolland *et al.* 2009). Moreover, the structural position of the Stepanavan blueschists unit in the core of a nappe anticline overthrusted by ophiolites and arc series suggests an intra-oceanic subduction context in the stage preceding collision of the SAB with the Eurasian margin.

According to these structural, petrological and geochemical results one can suggest that this part of the Neotethys Ocean was, at least from Middle Jurassic to Campanian in age, featured by an intra-oceanic subduction to its south and a subduction below the Eurasian continental plate to the north. These new results are significant constrains to the geodynamic processes evolution of the Lesser Caucasus at these periods of time.

Obduction modalities

As previously described by other works, several ophiolitic series are found in North and Central Armenia (Fig. 2). They have slightly different structures as a result of their lithologic and structural settings, but are thought to be remnants of a single obducted sequence (see Rolland *et al.* 2010). The ophiolitic series of the North Sevan Lake, Amassia and Stepanavan (Fig. 2) (Sevan–Akera ophiolites) (Knipper 1975; Adamia *et al.* 1977; Adamia *et al.* 1980; Maghakyan *et al.* 1985; Adamia *et al.* 1987) are found along the main suture joining the SAB and the Eurasian margin, while the Vedi ophiolites are an allochthonous nappe thrusted over the SAB (Knipper & Sokolov 1977; Sokolov 1977; Adamia *et al.* 1981). The Zangezur ophiolites located between the Sevan–Akera suture zone and the Vedi ophiolites correspond to some slivers of ophiolites unroofed along thrust faults during collision (Fig. 2) (Aslanyan & Satian 1977, 1982; Knipper & Khain 1980).

One of the main characteristics of these ophiolitic rocks is the very low grade and even, in most cases, the absence of any metamorphism related to subduction/collision stages. The main parts of the ophiolitic units are very well preserved excepted in some shear zones, especially at the base of the obducted sequence, in which some low greenschist facies metamorphic crystallisations occur. The ophiolites record a low metamorphic grade imprint related to heterogeneous hydrothermal alteration dating from oceanic stages (Rolland *et al.* 2010). Even in the Stepananvan area, the main ophiolitic body is unmetamorphosed. There, its rests on a tectonic mélange mainly composed of palaeo-accretionary prism blueschist facies rocks (Rolland *et al.* 2007).

The obduction event is well documented in the Vedi area (Southeast of Yerevan, Armenia) (Figs 2 & 5), where the frontal obduction contact is exposed (Figs 6 & 7). In that part of the Lesser Caucasus, ophiolitic rocks are thrusted over platform series of the SAB (Sokolov 1977). In order to describe the obduction modalities it is important to identify at first the nature of the autochthon. Is this autochthon a part of the Eurasian margin or does it correspond to an exotic terrane originating from a southerly position in the Neotethys domain?

The SAB autochthon

In the Lesser Caucasus, the SAB (also called Daralagez continental terrane) is mainly exposed in Armenia and corresponds of a Proterozoic metamorphic basement, an incomplete Palaeozoic sedimentary succession, Triassic and some Jurassic formations in Nakhijevan province (Azerbaijan), Upper Cretaceous sedimentary series, Paleocene to Oligocene detrital and volcanogenic rocks, Miocene volcanogenic rocks and Plio-Quaternary volcanites (Nalivkin 1976) (Figs 3 & 5).

The Proterozoic basement is made of metamorphic rocks characterized by gneisses, micaschists

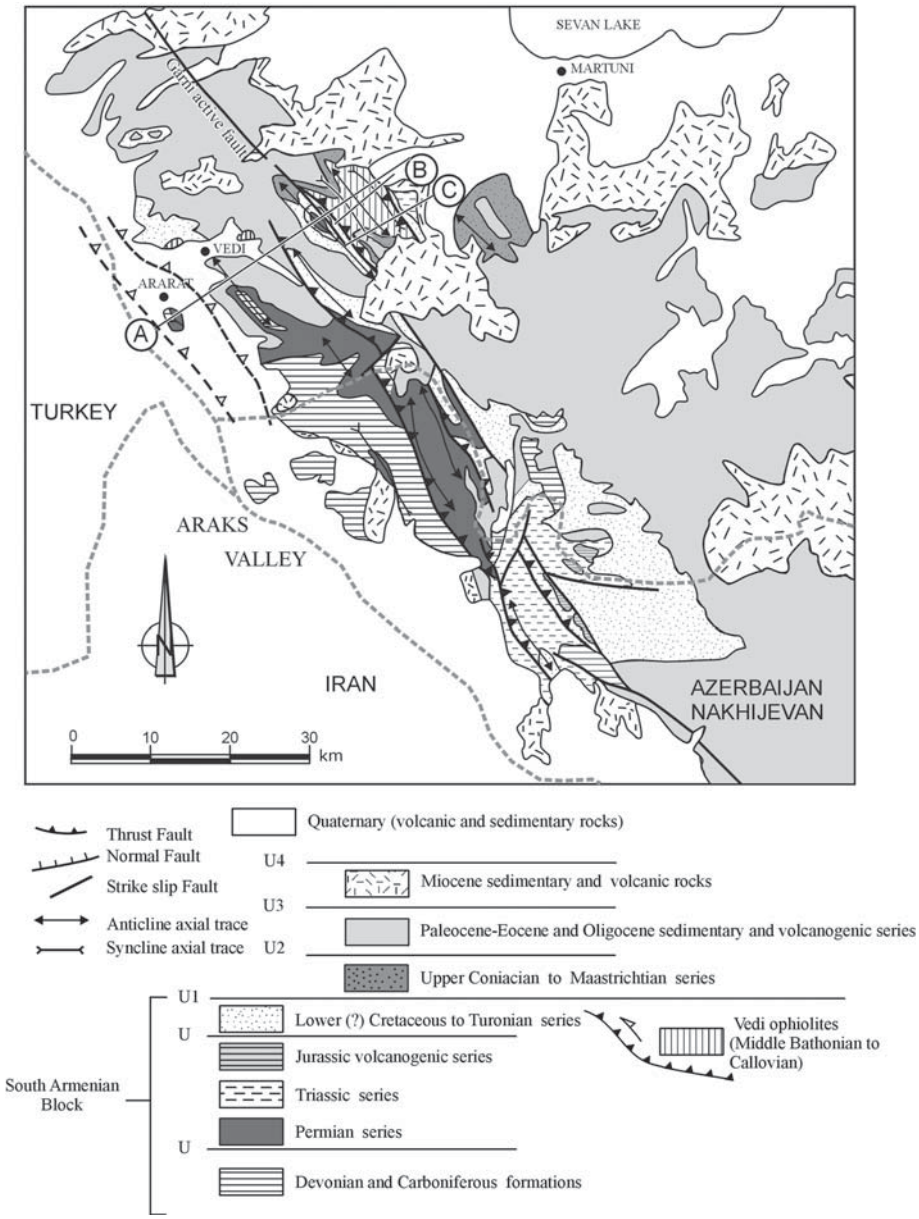

Fig. 5. Structural map of the Vedi area (Armenia).

and leucogranite intrusions. It is well exposed NE of Yerevan, in the Dzarkuniatz massif (Aghamalyan 1998) (Fig. 2). The Palaeozoic rocks of the SAB in Armenia are mainly localized in the southwestern part of the country (Figs 2 & 5). From bottom to top, the following succession of sedimentary rocks has been distinguished (Paffenholtz 1959; Karyakin 1989).

(1) Upper Devonian series (Frasnian–Famennian) composed of at least 1000 m thick detrital rocks including quartzites, sandstones and argillites.
(2) 500–700 m thick Carboniferous reef limestones.
(3) A Permian platform type section composed of 500–1500 m thick black limestones and

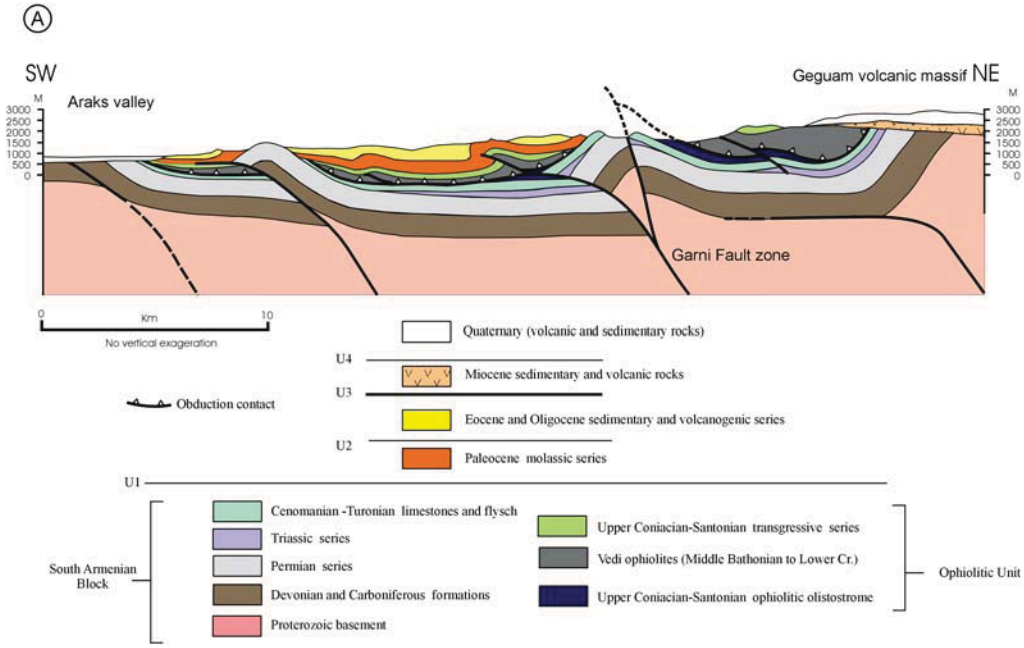

Fig. 6. Geological cross-section of the Vedi area (Armenia).

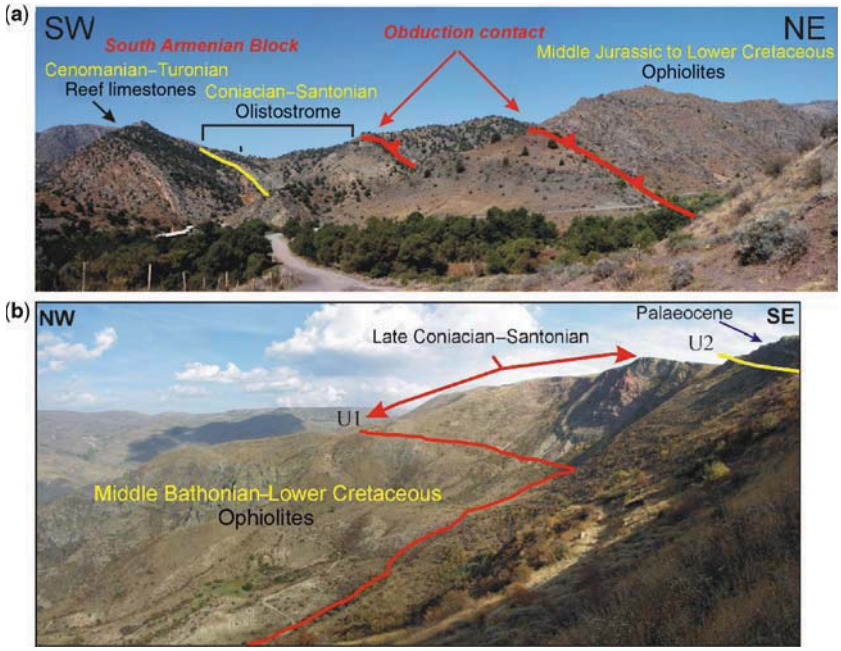

Fig. 7. Photographs of the obduction contact in Vedi area (**a**) and on the formations overlying the ophiolites (**b**). Unconformities U1 is related to the obduction event and U2 is related to the first collision stage.

marls with occurrences of bitumen. As the Upper Devonian formations they could represent a source rock for hydrocarbon. They disconformably overlie the Devonian and Carboniferous formations. Some syn-sedimentary normal faults occur in the sequence.

The contact of the sedimentary cover with the basement is unknown in these areas (Figs 5 & 6). The metasedimentary rocks exposed in the Dzarkuniatz massif could belong to the Proterozoic or to the Upper Devonian-Permian formations. Normal faults are suspected in the Devonian-Carboniferous formations as deduced by the geometry of the bedding within the Cenozoic structures (fold), (Fig. 5 and cross-section Fig. 6).

Lower Triassic rocks are conformably overlying on the Upper Permian limestones (Fig. 5). The series are composed of thin layered conodont rich limestones (Grigorian 1990), which evolve upward to detrital series composed of sandstones, black siltstones and mudstones of Late Triassic age. The thickness of the Triassic formations ranges from 700 to 900 m.

In Armenia, rocks from the Jurassic period are unknown in the southern part of the SAB. However, tens of kilometres southeastward in Nakhijevan (Azerbaijan) and in Iran (south to Araks valley) (Fig. 5), 500 m thick Lower and Middle Jurassic are described on top of the Upper Triassic section (Lordkipanidze *et al.* 1988). Palaeomagnetic data from Middle Jurassic alkaline basalts and from Aalenian to Bajocian sediments indicate a palaeo-latitude of $21.5^{\circ}\mathrm{N} \pm 3.7^{\circ}$ during this period of time (Bazhenov *et al.* 1996). The sedimentary sequence is characteristic of a platform containing Ammonites, Briachiopodes and Pelecypodes (Bazhenov *et al.* 1996).

Cenomanian and Turonian carbonates unconformably overlie the previous formations (Fig. 6). They are composed of thick reef limestones with some thin marl intercalations (Eghoyan 1955; Paffenholtz 1959; Sokolov 1977; Hakobyan 1978). In Vedi Area, the reef limestones are overlain upward by a Upper Cenomanian flysch (new dating by nannofossils, sample AR S 16, Table 1) thicker southward than northward (Sokolov 1977) and by a Upper Coniacian–Santonian olistostrome (Fig. 8) (Eghoyan 1955; Rengarten 1959; Hakobyan 1976, and ages we obtained from nannofossils, samples AR 63 05, AR 64 05, AR 65 05 and AR 67 05, Table 1) made of ophiolitic blocks contained in a muddy matrix (Fig. 4).

According to the stratigraphic and lithological data of the SAB, in comparison with the Eurasian characteristics (Fig. 4) the SAB clearly differs from the Eurasian margin. This is also evidenced by tectonic phases recorded in the SAB and the Eurasian margin basement.

The SAB metamorphic basement is Proterozoic (Aghamalyan 1969, 1978, 1998; Belov & Sokolov 1973), the Eurasian margin is mainly Variscan (Adamia *et al.* 1983; Adamia 1984, Zakariadze *et al.* 1998). The Palaeozoic series of the SAB mainly consist of platform-type carbonates whereas the sedimentary successions overlying the Eurasian crystalline basement are metamorphic schists dating back to the Visean and older. Moreover we do not find any sign of rifting all along the Eurasian margin series which could explain a drift to the south of the SAB after the Variscan orogeny. The palaeomagnetic data indicate a southerly palaeo-latitude of the SAB during the Early and Middle Jurassic (Barzhenov *et al.* 1996). All these data argue for a Gondwanian origin of the SAB located during Middle Jurassic, 2000 km south of its present location (Barzhenov *et al.* 1996).

Ophiolitic unit

This ophiolitic sequence outcrops within a folded klippe (Figs 2, 5 & 8). According to the palaeontological data (Radiolarians), the Vedi ophiolites are mainly Middle Jurassic (Middle to Late Bajocian) (Danelian *et al.* 2008, 2010) and Late Jurassic (Danelian *et al*. 2010) and probably Early Cretaceous in age (Belov *et al.* 1991). The geochemistry undertaken on the oceanic crust indicates an N-MORB origin (Galoyan *et al.* 2007, 2009; Galoyan 2008; Rolland *et al.* 2009, 2010) and some basalts located on top of the previous ones are signs of an alkaline component (Tsameryan *et al.* 1988; Galoyan 2008). The geochemical compositions and the fact that peridotites, gabbros and plagiogranites were on the ocean-floor together (the radiolarites cover all these rocks) (Galoyan *et al.* 2008) indicate a slow-spreading oceanic crust, put in place in a back-arc domain, and overlain by a hot-spot series (see Rolland *et al.* 2010 for a synthesis).

According to previous works, the whole ophiolitic sequence was interpreted as an ophiolitic melange (Sokolov 1977; Zakariadze *et al.* 1983). In contrast, we distinguish ophiolitic units comprising of preserved slices of oceanic crust thrusted over the Upper Coniacian–Santonian olistostrome of the SAB reworking the ophiolites (Figs 6 & 8).

On top of the ophiolites, we observed reef-type sediments, OIB lavas and arc-type volcanic rocks (andesites lava flows), which unconformably cover the pillow-lavas (Fig. 8). These intra-oceanic arc-type volcanic series have been evidenced from geochemical data (Galoyan *et al.* 2007, 2009; Galoyan 2008).

An Upper Coniacian to Santonian transgressive series disconformably overlies the ophiolitic Vedi unit. This transgressive series is characterized

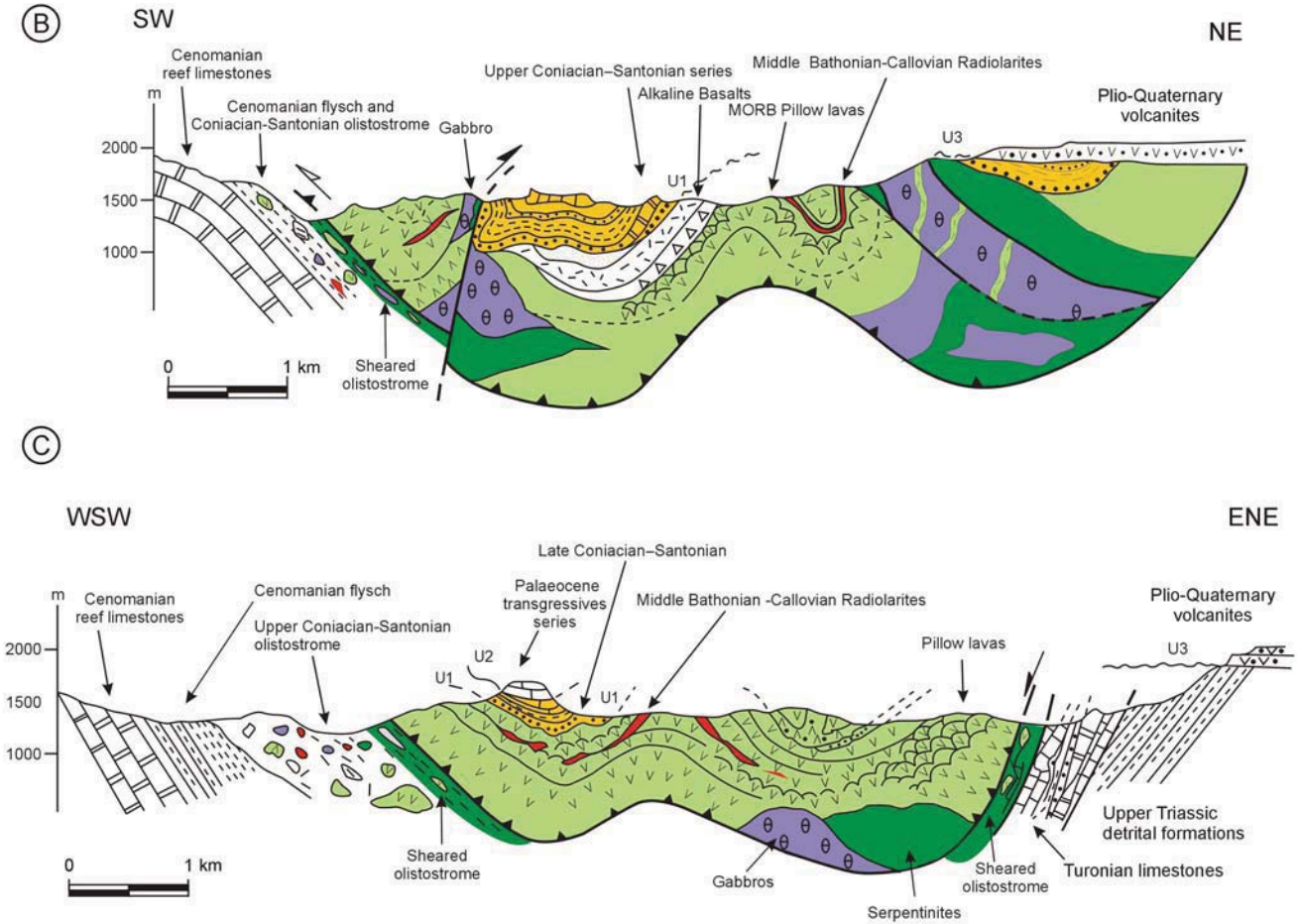

Fig. 8. Geological cross-sections of the Vedi ophiolites, locations are shown on Figure 5.

upwards by breccias and conglomerates, which rework ophiolitic rocks and the Cenomanian–Turonian autochthonous limestones (Fig. 7). The conglomeratic formation grades laterally into reef limestones composed of Upper Coniacian Hippurites fossils. Green mudstones, siltstones and thin turbiditic layered reef limestones with some pink levels overlie the reef limestones (Fig. 7). Late Coniacian–Santonian ages of the green mudstones were obtained by nannofossils (samples AR 72 05, AR 73 05, Table 1).

One can note that in the Vedi area the Campanian and Maastrichtian are rare or missing. This indicates erosion or non-deposition due to the obduction uplift starting during the Late Coniacian and ongoing during Campanian to Maastrichtian times.

Moreover, the HP metamorphism of the Stepanavan Blueschist unit is dated at 94–90 Ma and its exhumation at 74–71 Ma (Rolland *et al.* 2007). This exhumation is ascribed to the underthrusting of the SAB continental lithosphere still attached to the subducted slab below the oceanic crust section during the first stage of the obduction process. In this case, unroofing the Blueschist unit within the subduction channel could be the result of driven forces conducted by the low density of the continental crust.

In summary the back-arc basin is thrusted over the SAB during Late Coniacian–Early Santonian time, as evidenced by 1) the ophiolitic olistostrome age; and 2) the occurrence at the same time of reef limestones on top of the ophiolites which suggest a shallow marine environment indicating uplift of the oceanic crust along the obduction thrust. This obduction stage was still active during Campanian and Maastrichtian explaining the absence or the pauce of sedimentation in the Vedi area south of Lesser Caucasus.

Collision

The collision stages in the Lesser Caucasus are not well known. This is mainly due to superimposed tectonics which occurred in this region from Late Cretaceous to present. The first aim of our studies was to clearly identify the timing of collision triggering or more exactly that of the continental subduction of the SAB. Major unconformities and structures were dated allowing the reconstruction of the tectonic evolution of this collision event.

Beginning of the collision between SAB and Eurasia

Along the Sevan–Akera suture zone North of Sevan Lake, late-Middle to Upper Eocene nummulitic series unconformably overlie the ophiolite units, the Eurasian margin and the SAB (Unconformity U2) (Figs 2, 9 & 10). According to these stratigraphic

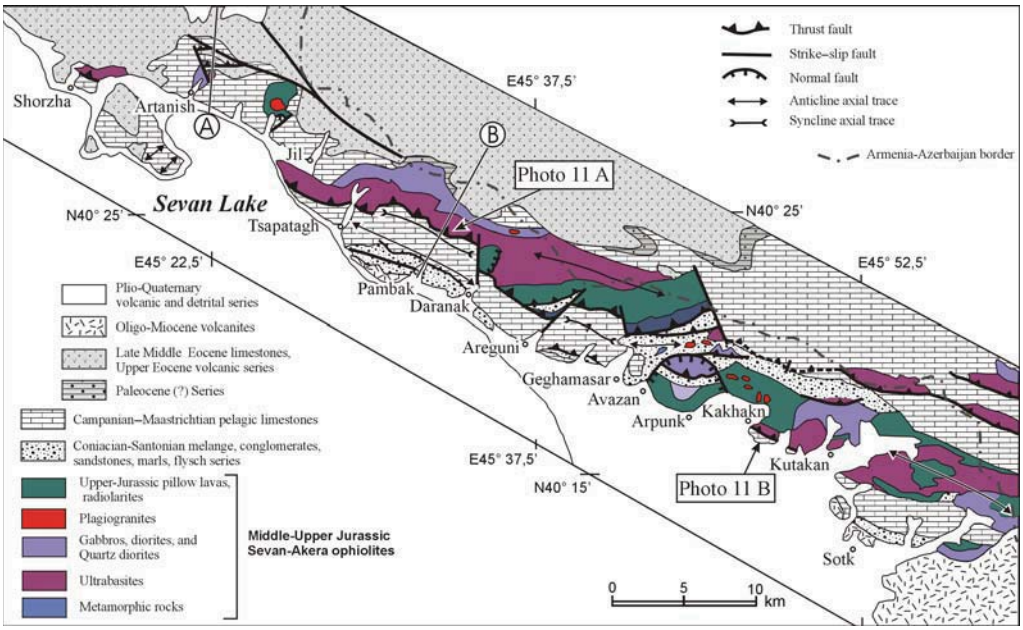

Fig. 9. Structural sketch map of the northern Sevan Lake area (Sevan–Akera suture zone), (Armenia and Azerbaijan). Modified from Nalivkin (1976), and including our observations.

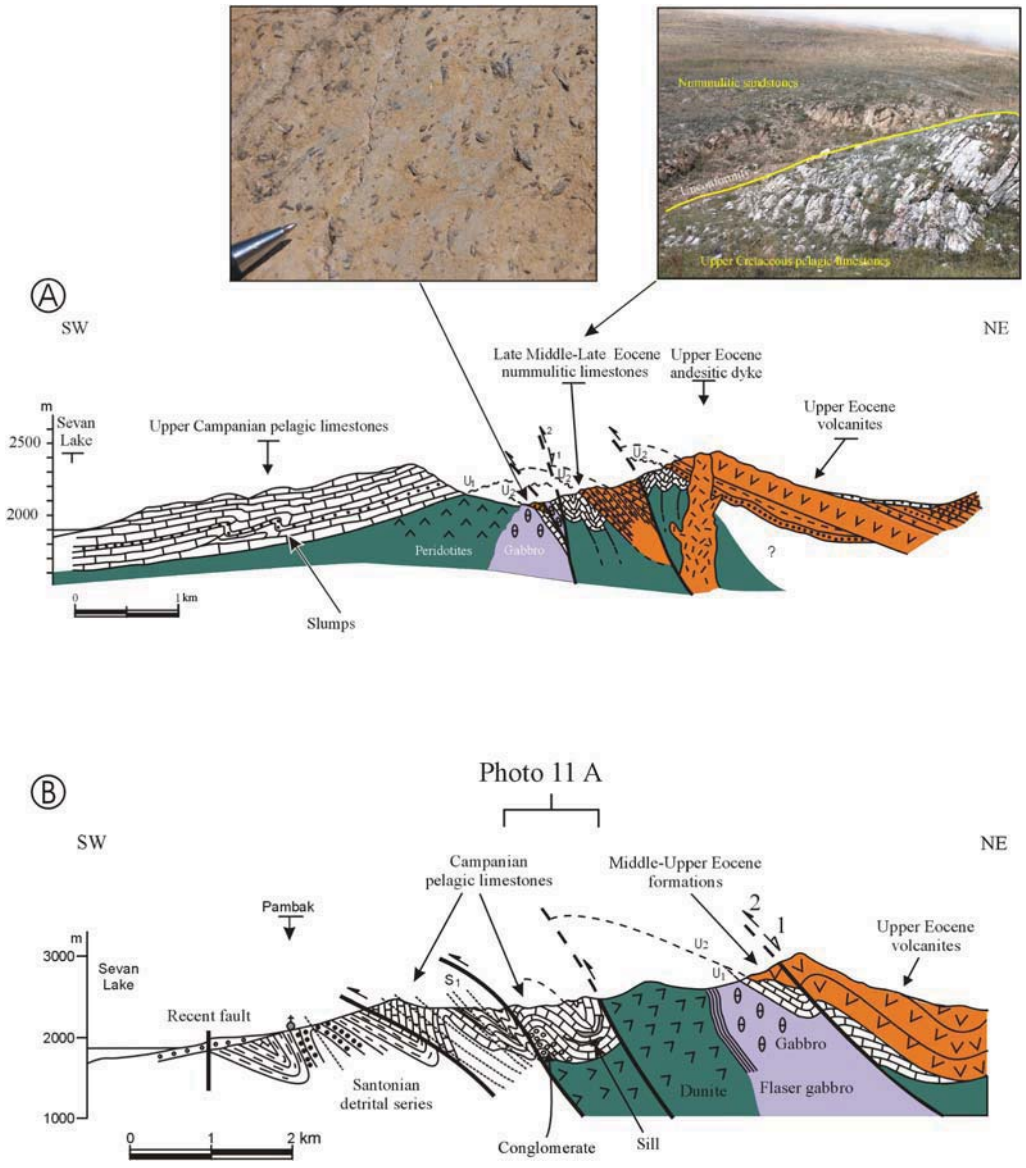

Fig. 10. Geological cross-section of the northern Sevan lake area (Armenia and Azerbaijan). Photographs represent the Nummulites of the Nummulitic sandstone of late-Middle Eocene age and the unconformity U2 of the late-Middle Eocene on the Campanian pelagic limestones.

data the onset of the collision between the SAB and Eurasia occurred before the late-Middle Eocene. In detail the unconformable sedimentary deposits are characterized upward by conglomerates reworking Upper Cretaceous pelagic limestones, ophiolites and Lower Eocene andesites. Sandstones and nummulitic limestones overlie the conglomerate. The sedimentary succession of few tens of metres thickness is interbeded with andesitic lava flows. Upwards, the series is characterized by thick volcanic layers made of andesitic and trachytic lavas (Fig. 10). The late-Middle to Late Eocene age (NP 18) of the base of this formation was obtained by nannofossiles (samples AR-03-05, AR-07-05 and AR-08-05, Table 2). In the western part of North Sevan area this volcano-sedimentary series unconformably overlies thick (maximum 2000 m thick) Lower to Middle Eocene calcalkaline volcanogenic

series mainly made of andesite, trachyte, rhyolites and sandstones (Lordkipanidze *et al.* 1988).

This is still well expressed on the Eurasian margin in Armenia and Azerbaijan where the late-Middle Eocene formations unconformably cover the Middle Jurassic to Upper Cretaceous formations (Fig. 2). Consequently the late-Middle Eocene deposition occurred after a period of erosion. The sedimentary products of this erosion event have to be found southwestward.

Syn-collisional basin

The Paleocene is not well expressed in the area of the Sevan–Akera suture zone. Danian sandstones have been described (Abovyan 1961; Sokolov 1977) but they are not clearly exposed all along the suture zone. In this part of the belt the Paleocene time is characterized by uplift, erosion and folding. Cross-sections and maps from the suture zone in North Sevan area (Figs 3, 9 & 10) illustrate well Paleocene folds unconformably covered by the late-Middle Eocene nummulitic sandstones and limestones. There, the Campanian pelagic limestones that overlie the ophiolites unit are folded (Fig. 10).

Nevertheless to the south in the Vedi area the Paleocene detrital series disconformably overlie the obducted ophiolites and the SAB. These formations are well developed and characterized by thick molassic deposits (more than 1000 m in thickness) (Figs 5 & 6). The detrital sequences are made, from bottom to top, of conglomerates, sandstones, greywackes, siltstones, marls and some algae limestones interlayers. The detrital series mainly reworked the ophiolites and the SAB autochthon. The Paleocene age of this formation was obtained by nannofossils (samples AR-54-05, AR-101-05, AR-107-05, Table 2) and are in agreement with the previous dating (Nalivkin 1976). The Paleocene basin to the south has to be considered as a foreland basin in front of a collisional belt located to the north (Figs 2, 5 & 6).

Structures and deformation related to the SAB collision

In Vedi area, N130°E trending folds deform the Palaeozoic to Santonian formations of the SAB, the ophiolites and the foreland basin (Fig. 5). Paleocene to Oligocene molasses covering the ophiolites are deformed by ramp anticlines (Fig. 6). As a result of the southwestward folding propagation, Eocene and Oligocene series are deposited with onlap relationships over the Palaeozoic–Mesozoic anticline cores (Fig. 6). The anticline cores are made of Palaeozoic formations from the SAB. For example, to the SE part of Vedi area the Oligocene formations unconformably cover the Permian. The corresponding tectonic phase is due to the southwestward propagation of the collisional deformation. These types of structures, featured by folds and ramps are related to décollement levels within the Devonian series and the ophiolites (serpentinites) (Fig. 6).

In contrast, the Miocene volcanic rocks seal the structures related to this deformation phase (Unconformity U3), as they appear to be monoclinal (Fig. 5). As a result the folding event can be bracketed between the Oligocene and Miocene periods. More precision on the age of these Neogene series is needed to accurately constrain the age range of the NE–SW shortening event responsible to folding. The disconformity (U4) of the Plio-Quaternary volcanites on the Miocene volcanic and sedimentary rocks is well exposed in this area (Figs 5, 6 & 8).

The structural interpretation on Figure 6 takes into account the occurrence of reactivated normal faults in the Palaeozoic series and a crustal-scale active fault (the Garni Fault: a currently dextral strike–slip fault, Rebai *et al.* 1993; Philip *et al.* 2001) precisely located at the anticline axis of the main structure (Figs 5 & 6). This interpretation is in favour of inherited structures within the SAB basement which have been reactivated during the Cenozoic, and still currently active faults (Philip *et al.* 2001; Karakhanian *et al.* 2004; Avagyan *et al.* 2005).

In the north area of the Sevan Lake two main unconformities have been observed. The first one (U1) corresponds to the deposition of the Upper Cretaceous pelagic series over the ophiolitic sequence (Fig. 10). We dated these formations by nannofossil assemblages from Late-Coniacian to Late Campanian (Avazan and Artanish valleys, Table 1). The Maestrichtian would be present in some parts (Nalivkin 1976). The second one (U2), described above corresponds to the deposition of the late-Middle to Late Eocene series over deformed ophiolites and Late Cretaceous formations (Figs 9 & 10). N50–80°E trending folds deform the Campanian pelagic limestones and are related to a shortening event that occurred before the deposition of Nummulitic sediments. Moreover, N120–130°E trending reverse faults and thrusts cross-cut the Upper Eocene series all along the North Sevan Lake from Artanish to Zod (Figs 9, 10 & 11). These structures also occur all along the suture zone from the Transcaucasus massif to the Karabakh (Fig. 2) and in front of the belt from Transcaucasus domain to the southeastward end of the Lesser Caucasus in Armenia and Nakhijevan.

Timing of the collision stages

According to the stratigraphical and structural data the collision between SAB and Eurasia started

during the Paleocene as evidenced by the formations of a foreland basin in the southeastern part of the belt and by the folding and uplift of the Sevan–Akera suture zone. Following this uplift and erosion phase it is very probable that during Early Eocene the north flank of the suture zone and the Eurasian margin subsided, resulting in deposition of Lower Eocene detrital rocks and magmatism. Nevertheless uplift to the South, along the suture zone, and subsidence to the north along the southern Eurasian margin must have been coeval, resulting in a huge erosional surface on which late-Middle Eocene shallow marine sediments were deposited. Some normal faults in the Upper Eocene volcanites were reactivated as post Eocene thrust faults are present in the North Sevan Lake area (e.g. Fig. 10, cross-section B).

During the Paleocene to late-Middle Eocene in the southern (Armenian) part, a flexural molassic basin covered the obducted ophiolite and related structures, and was progressively deformed along thrusts and décollement faults developing a fold-and-thrust belt type until the Miocene (Fig. 6). Some of the normal faults within the SAB Palaeozoic formations and the basement have been reactivated as reverse faults (Fig. 6). We presume that these faults were normal faults responsible for titled blocks of the north SAB passive margin.

The Miocene epoch corresponds to a drastic transition in the deformation style of the belt. As clearly evidenced by the Vedi area cross-section, folding is stopped and a huge erosion surface marks the transition (Figs 2, 3, 5 & 6). However, strain field evolved, and the shortening direction changed from NE–SW to NNW–SSE (Avagyan *et al.* 2005). This shift in strain direction produced reactivation of the main N130°E trending previous structures as left-lateral strike–slip faults with a reverse component. Since then deformation has remained with similar features, which resulted in the opening of NW–SE elongated volcanic clusters in the main shortening direction, and in a general uplift of the area (Karakhanian *et al.* 2004; Avagyan *et al.* 2005, 2010).

Geodynamic evolution and discussion

According to the new results presented above and with the help of numerous previous data mainly acquired during the Soviet period we can propose a new geodynamic model of the Lesser Caucasus evolution since the Late Jurassic (Fig. 12).

Middle Jurassic to Coniacian

According to palaeomagnetic data (Bazhenov *et al.* 1996) the SAB was located 2000 km south of its present position during the Middle Jurassic. Moreover from Middle Jurassic to Late Coniacian, two main subduction zones are responsible for: (1) the huge magmatic activity on the Eurasian active margin (Somkheto-Karabakh Island arc); (2) the calcalkaline magmatism on top of the oceanic lithosphere which have been obducted since Late Coniacian; and (3) occurrence of Blueschist unit (94–90 Ma). From the overall geometry, we can suspect a subduction towards the NW beneath the Eurasian margin, but the preceding stages of intra-oceanic subduction may be featured by different orientations. Based on stretching lineations, mineral HP-LT mineral lineations and senses of shear in the Stepanavan Blueschist unit, Rolland *et al.* (2007) suggest an east-verging intra-oceanic subduction zone. However this direction could have been rotated during collision. Whatever in this hypothesis the ophiolites of the Lesser Caucasus belongs to a marginal basin opened in back-arc position within the Neotethys Ocean (lherzolite ophiolite type [LOT] ophiolites, low spreading rate, Galoyan 2008; Galoyan *et al.* 2009; Rolland *et al.* 2009). All these results and the geochemical characteristics of the oceanic basalts evidence a SSZ type for the intra-oceanic subduction zone (Rolland *et al.* 2009).

On the Eurasian margin we have observed a pre-Coniacian erosional surface and disconformity of Coniacian and Santonian arc-type series characterized by a basal reddish conglomerate reworking the island arc series and their Upper Jurassic intrusions. Consequently during the Early Cretaceous an active erosion event took place, which resulted in the unroofing of plutons of the magmatic arc (Azerbaijan, mark E on Fig. 2). This erosion event is the result of significant uplift and denudation during the Early Cretaceous. The reasons for such a change in the Eurasian active margin strain field could be the subduction of the spreading ridge of the back-arc basin (see Fig. 12) or the entrance in the subduction zone of an oceanic plateau or an intra-oceanic ridge, which is suspected from the geochemical analysis of alkaline basalt on the ophiolites (Galoyan *et al.* 2007; Galoyan 2008; Rolland *et al.* 2009).

Coniacian to Paleocene

From the Coniacian to the Paleocene the intra-oceanic subduction (SSZ) evolved to a continental subduction of the SAB beneath the intra-oceanic arc and the marginal basin (Fig. 12). This event is supported by HP-LT metamorphism at 94–90 Ma of oceanic formations within an accretionary prism identified in the Stepanavan area (Fig. 2) (Rolland *et al.* 2007; Galoyan 2008). This step conducts around 5 Ma later to the obduction of the ophiolites over the SAB. This is well evidenced by: 1) the ophiolites uplift evidenced by the Upper

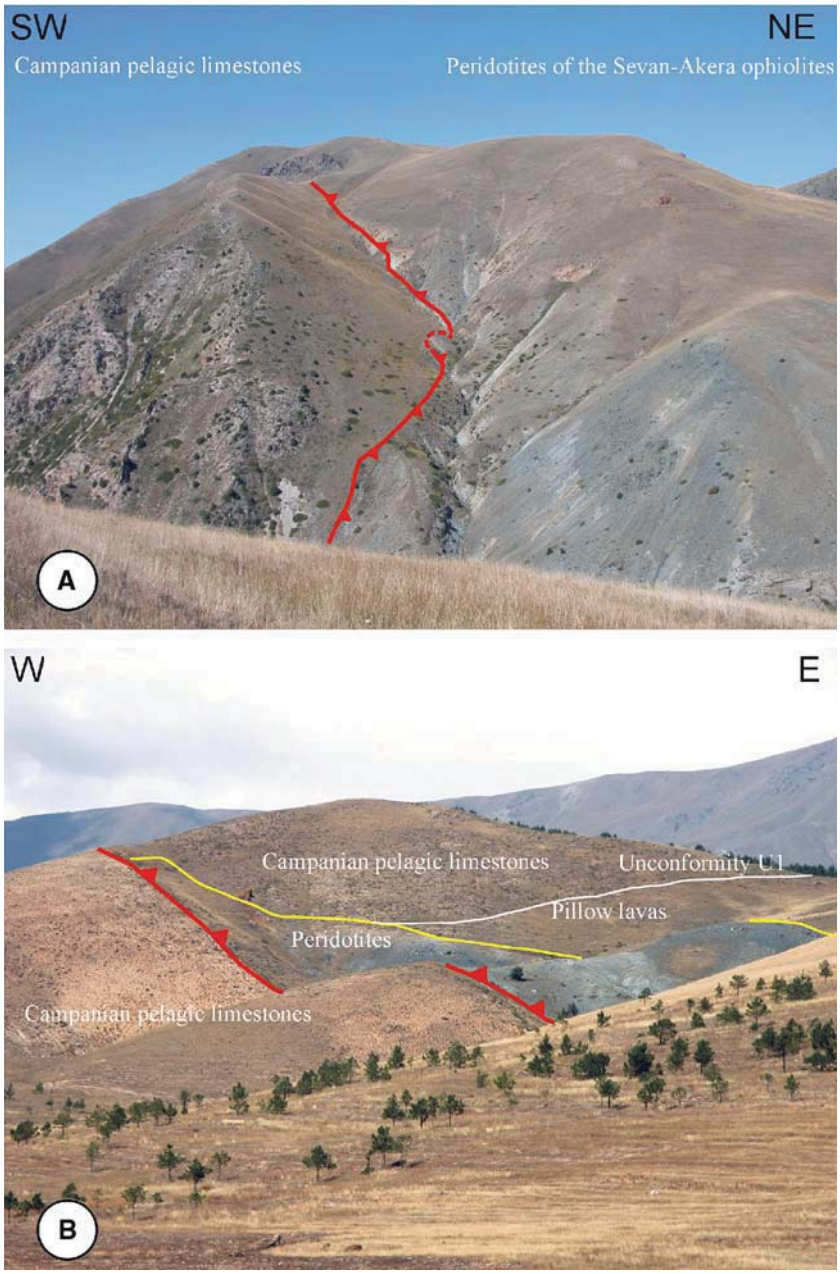

Fig. 11. Photographs of the post Eocene thrust faults in the northern Sevan Lake area; location of photo A shown on Figures 9 and 10, location of photo B shown on Figure 9.

Coniacian–Santonian reef limestone deposits on top of them in Vedi Area; 2) the coeval olistostrome formation beneath the ophiolitic thrust (Vedi and Sevan areas, Figs 8 & 9) and overlying Cenomanian–Turonian flysch of the SAB.

The Campanian–Maastrichtian formations are absent in the belt's frontal part, while they are present as pelagic sediments on the ophiolites of the Sevan–Akera zone. This lateral variation is ascribed to the southern frontal part's uplift

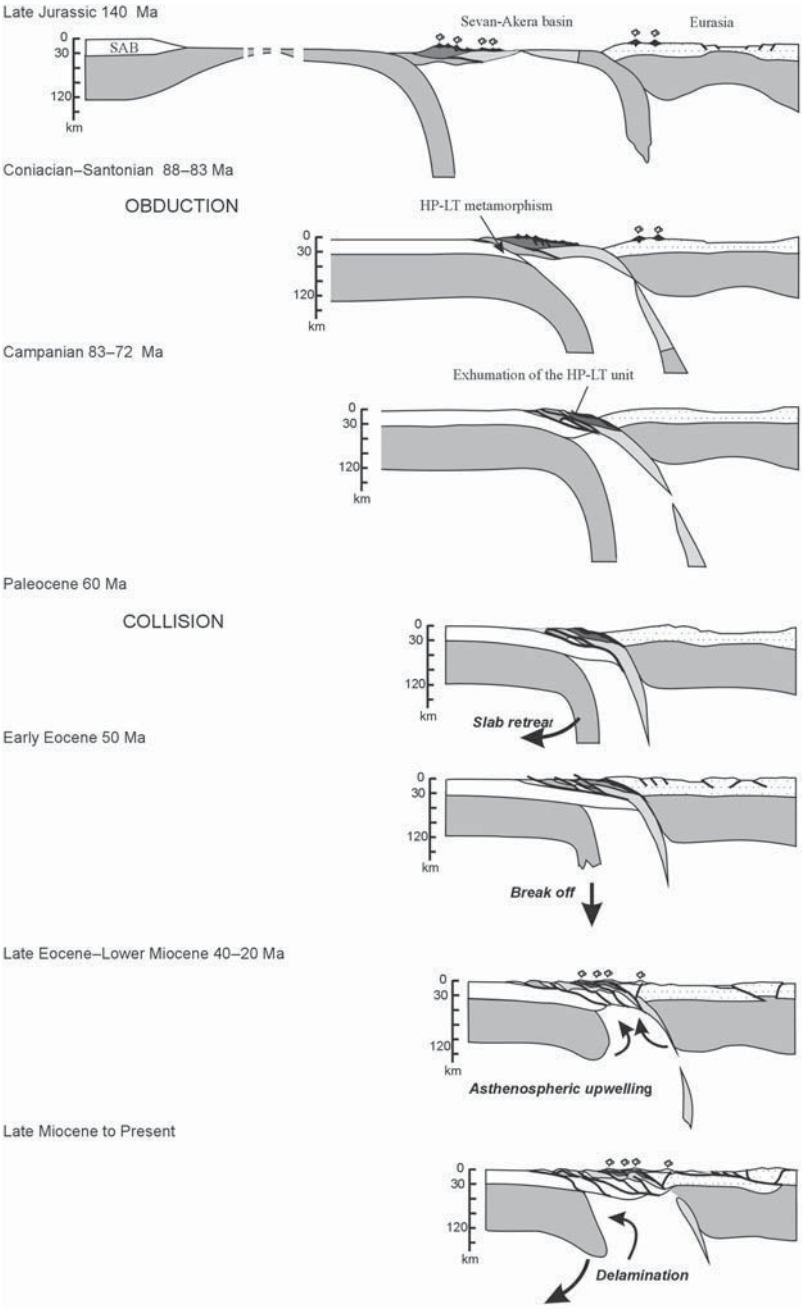

Fig. 12. Geodynamic model of the evolution of the Lesser Caucasus from Late Jurassic to present (see explanations in the text).

produced by the southward obduction thrust propagation and by a remnant oceanic crust covered by pelagic limestones to the north. The southern slope of this remnant oceanic crust (on the back of the obduction unit) was characterized by a northward dipping evidenced by north verging slumps within the Campanian pelagic deposits (Fig. 10 cross-section A). During the same time one notes

the exhumation of the Blueschist unit along the previous intra-oceanic subduction zone (Rolland *et al.* 2007).

Paleocene to Lower Miocene

The occurrence of a foreland basin in front of the belt and the folding and erosion of the Sevan–Akera ophiolitic zone suggest the entrance of the SAB in the subduction zone below the Eurasian margin since the Paleocene, pulled by the dense eclogitized oceanic slab to which it is still attached.

Due to the low density of the SAB continental crust entering in the subduction contact under the Eurasian margin, the continental subduction produces: 1) the uplift, shortening, folding and erosion of the obducted unit on the SAB (The Sevan–Akera zone); 2) the thrusting and folding of the Eurasian margin formations; and 3) the southwestward transport of the erosional products to the flexural basin in the belt front.

The Lower to Middle Eocene magmatism on the Sevan–Akera suture zone (Lordkipanidze *et al.* 1988) could correspond to the first stage of a slab retreat triggered by the continental subduction and the break off of the continental lithospheric subducted mantle. Thermo-mechanical physical modelling has evidenced this mechanism (Boutelier *et al.* 2003, 2004). But more studies on the geochemical analysis of this magmatism must to be performed in order to test this model for the Lesser Caucasus region.

One prominent result of this study is that the time of oceanic closure, indicated by the late-Middle Eocene unconformity on the SAB, the suture zone and the Eurasian margin clearly pre-dates this epoch. Therefore, the huge magmatic activity occurring from Late Eocene to Miocene (Lordkipanidze *et al.* 1988; Karapetian *et al.* 2001) has to be explained in this collisional context. We suggest that the slab retreat and break off processes lead to an asthenospheric upwelling below the suture zone (Fig. 12). This mechanism could explain the location of magmatism all along the Lesser Caucasus. However, some geochemical studies would be necessary to test this assumption.

Upper Miocene to present

The continental subduction seems to be stopped but deformations still occur. The Arabian plate collision with the SAB to the South (in the Bitlis region, Oberhänsli *et al.* 2010 for a review), is responsible for the strain field's reorganisation (Philip *et al.* 2001; Avagyan *et al.* 2005 and references herein). This produces the reactivation of the thrusts as strike–slip faults and the opening of volcanic clusters (Avagyan *et al.* 2005). Due to the collision, one may explain the widespread magmatic activity by partial melting of overthickened continental crust (Mitchell & Westaway 1999; Karapetian *et al.* 2001). Another possibility could be a slab break off process beneath Eastern Anatolia (Faccenna *et al.* 2006; Lei & Zhao 2007). However, this second hypothesis does not seem appropriate to explain all the magmatic activity known in the Lesser Caucasus and Greater Caucasus regions, which are relatively far from Eastern Anatolia. Another solution could be that of asthenospheric upwelling (if this one has actually existed) producing significant weakening of the SAB continental lithospheric mantle, which begins its delamination. This could explain the magmatism and uplift of the Lesser Caucasus since Pliocene time. The question if whether asthenosphere upwelling is related or totally unrelated to the lithosphere-scale plate tectonic processes remains open.

Conclusions

The results coming out from studies undertaken in the Lesser Caucasus region allow to clarify the main questions on its tectonic and geodynamic evolutions.

(1) The Sevan–Akera suture zone represents the only suture of the belt. The other ophiolitic units (Vedi and Zangezur, Fig. 2) correspond to ophiolitic nappes due to the obduction of oceanic crust over the SAB or to tectonic slivers reworked by the Upper Oligocene–Lower Miocene thrusts and reverse faults. The collision stage has significantly deformed the initial obduction contact by folding and thrusting and then more recently by strike–slip faulting. Consequently all the identified ophiolitic units in the Lesser Caucasus have been remobilized and outcrop today thanks to these collisional structures, as in the cores of anticlines (Sevan Lake, Stepanavan areas) (Figs 2 & 9) or in synclinal klippes (Vedi area) (Fig. 6) or as slivers along thrust faults (Zangezur and Amassia areas) (Fig. 2). In our interpretation of the structure of the Lesser Caucasus the ophiolites are continuous from Central to North Armenia, deep below the Sevan Lake (Fig. 3). The consequence of this structure is that the SAB is underthrusted below this huge ophiolite sliver, and may continue also below the Sevan Lake. This point is important in terms of natural resources because of the hydrocarbon potential of source rocks in the SAB Palaeozoic section. Indeed these source rocks could be capped by the allochthonous ophiolites, which could

represent an overscreen under the Sevan Lake and the neighbouring regions.

(2) The onset of collision or the continental subduction of the SAB below the Eurasian margin is dated as Paleocene. This process occurred around 20 Ma later than the obduction (Late Coniacian–Santonian, 88–83 Ma) of the marginal basin over the SAB. It led to the uplift of the Sevan–Akera suture zone, its folding, erosion and to the transfer of detrital materials in a flexural basin in front of the belt, above the obduction structures. From late-Middle Eocene to Miocene all the belt is deformed by thrusts and reverse faults in the internal part (Transcaucaus to Karabakh, Fig. 2) and by N130°E trending fold and thrust belt at the chain front. One can note an evolution from thin-skin to thick-skin tectonics in front mainly due to the reactivation of the previous normal faults in the SAB

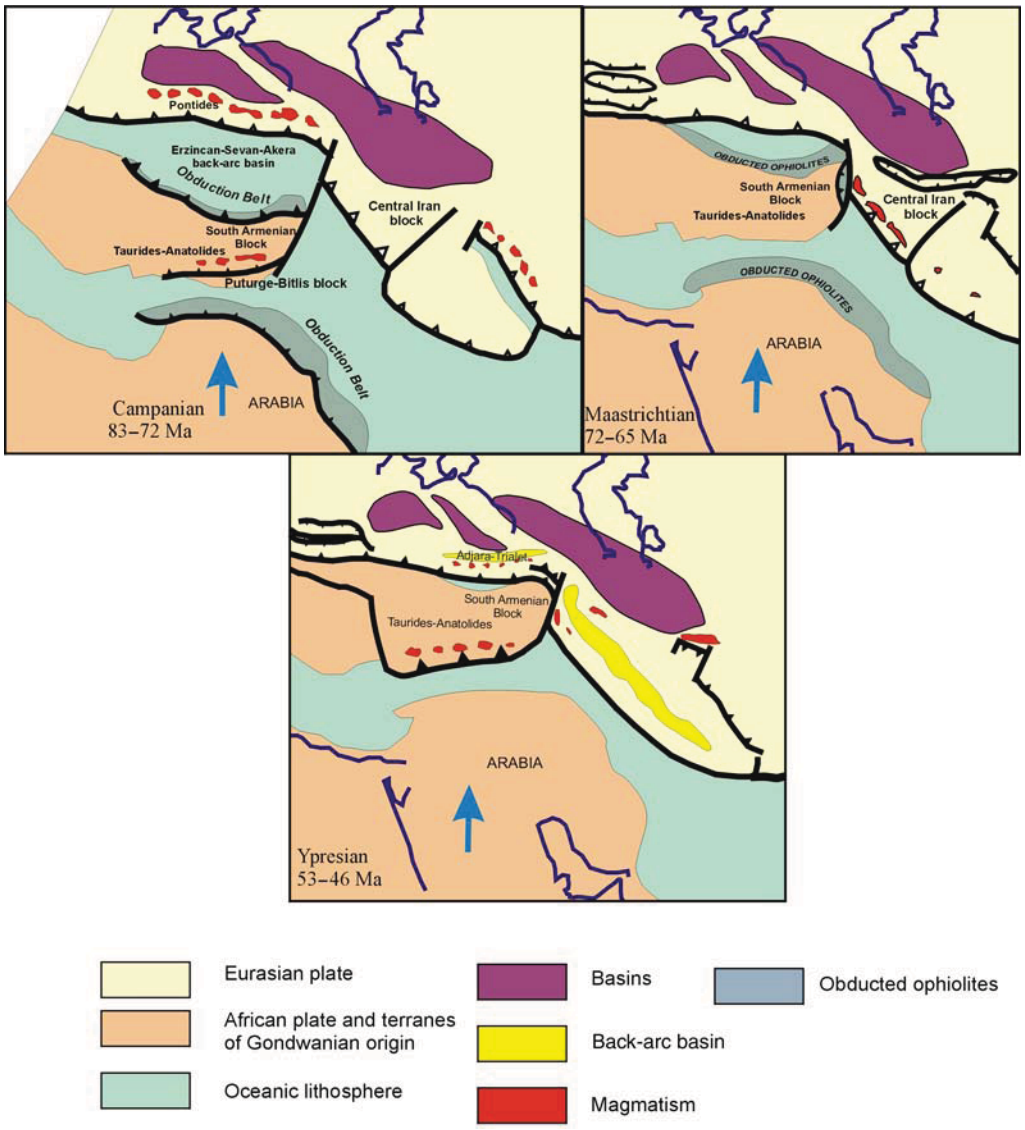

Fig. 13. Campanian, Maestrichrian and Ypresian palaeotectonic maps of the Lesser Caucasus region and its neighbouring areas (modified from Barrier & Vrielynck 2008).

basement at depth (Vedi and Southeast Armenia area, Figs 2 & 3).

(3) The widespread Eocene to recent magmatic activity is the result of various, still enigmatic processes. The Eocene magmatic activity occurred after the onset of the collision. Consequently it cannot be directly related to the oceanic subduction processes except for the Adjara-Trialet massif in Georgia considered as a back-arc basin due to a oceanic subduction of a small remnant oceanic lithosphere NW of the Sevan–Akera Suture zone (Adamia *et al.* 1981; Barrier & Vrielynck 2008). We suggest the possibility of a slab retreat and a break off allowing the rising of the asthenospheric mantle, which could have heated the overthickened continental crust (Fig. 12). The Miocene to recent magmatic activity seems to be related to 1) a possible asthenosphere upwelling beneath the suture zone after the break off and, 2) strain field reorganization at these times due to the Arabian plate collision with the SAB producing opening of volcanic clusters along main inherited faults (Karakhanian *et al.* 2004; Avagyan *et al.* 2005, 2010).

The Lesser Caucasus is a key area to understand lateral connections with the Taurides–Anatolides to the west and the Iranian regions to the East. Indeed, as demonstrated by Stöcklin (1974) and developed by Zanchi *et al.* (2009) the Eo-Cimmerian orogen is discontinuously exposed along the northern side of the Alborz Mountains of north Iran below the siliciclastic deposits of the Shemshak Group (Late Triassic–Jurassic). However, a thick comparable Shemshak-type basin is unknown in the Lesser Caucasus. In this belt the contact between the Upper Triassic detrital series and the Lower Triassic carbonates formations is not well exposed. A complete sequence has been described in the Nakhijevan part of the Lesser Caucasus (Barzenov *et al.* 1995). Nevertheless, no evidence of Eo-Cimmerian orogeny has been found in the Lesser Caucasus. These features conduct to expect in the study area a major stratigraphic and structural limit. Consequently, the oriental border of the continental blocks of Taurides–Anatolides and South Armenian Block of Gondwanian origin need to be clarified. This block collided with Eurasia during Paleocene Eocene times. According to the MEBE palaeotectonic map (Fig. 13) (Barrier & Vrielynck 2008) the passage between Lesser Caucasus and NW Iran (NW Alborz and Tabriz-Saveh) would be then a syntax similar to those delimiting India colliding with Asia, but it remains to be demonstrated. One way to explain the problems is the existence of NE–SW orientated Araks strike–slip fault, suggested as an oriental border of Lesser Caucasus (Fig. 1). But if this fault has a recent expression, it is not sure that it was present at the beginning of the collision between the South Armenian Block and Eurasia.

New interpretations allow some points to be studied in more detail, particularly the origin of the widespread collisional magmatism and the connections with the Iranian belt, which remain to this day a subject of debate.

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