The Mesozoic–Cenozoic tectonic evolution of the Greater Caucasus

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Abstract: The Greater Caucasus (GC) fold-and-thrust belt lies on the southern deformed edge of the Scythian Platform (SP) and results from the Cenozoic structural inversion of a deep marine Mesozoic basin in response to the northward displacement of the Transcaucasus (lying south of the GC) subsequent to the Arabia–Eurasia collision. A review of existing and newly acquired data has allowed a reconstruction of the GC history through the Mesozoic and Cenozoic eras. In Permo(?)-Triassic times, rifting developed along at least the northern part of the belt. Structural inversion of the basin occurred during the Late Triassic corresponding to the Eo-Cimmerian orogeny, documented SE of the GC and probably linked to the accretion of what are now Iranian terranes along the continental margin. Renewed development of extensional basin formation in the area of the present-day GC began in Sinemurian–Pliensbachian times with rift activity encompassing the Mid-Jurassic. Rifting led to extreme thinning of the underlying continental crust by the Aalenian and concomitant extrusion of mid-ocean ridge basalt lavas. A Bathonian unconformity is observed on both sides of the basin and may either correspond to the end of active rifting and the onset of post-rift basin development or be the record of collision further south along the former Mesozoic active margin. The post-rift phase began with deposition of Late Jurassic platform-type sediments onto the margins and a flysch-like unit in its deeper part, which has transgressed the basin during the Cretaceous and Early Cenozoic. An initial phase of shortening occurred in the Late Eocene under a NE–SW compressional stress regime. A second shortening event that began in the Mid-Miocene (Sarmatian), accompanied by significant uplift of the belt, continues at present. It is related to the final collision of Arabia with Eurasia and led to the development of the present-day south-vergent GC fold-and-thrust belt. Some north-vergent retro-thrusts are present in the western GC and a few more in the eastern GC, where a fan-shaped belt can be observed. The mechanisms responsible for the large-scale structure of the belt remain a matter of debate because the deep crustal structure of the GC is not well known. Some (mainly Russian) geoscientists have argued that the GC is an inverted basin squeezed between deep (near)-vertical faults representing the boundaries between the GC and the SP to the north and the GC and the Transcaucasus to the south. Another model, supported in part by the distribution of earthquake hypocentres, proposes the existence of south-vergent thrusts flattening at depth, along which the Transcaucasus plunges beneath the GC and the SP. In this model, a thick-skinned mode of deformation prevailed in the central part of the GC whereas the western and eastern parts display the attributes of thin-skinned fold-and-thrust belts, although, in general, the two styles of deformation coexist along the belt. The present-day high elevation observed only in the central part of the belt would have resulted from the delamination of a lithospheric root.

The Greater Caucasus (GC) belt forms a morphological barrier along the southern margin of the Scythian Platform (SP; contiguous with the southern East European Platform, EEP), running from the northern margin of the eastern Black Sea Basin to the South Caspian Basin (Fig. 1). It developed during several phases of deformation in Mesozoic–Cenozoic times (Milanovsky & Khain 1963; Adamia *et al.* 1977, 1981; Rastsvetaev 1977; Khain 1984; Muratov *et al.* 1984; Gamkrelidze 1986; Dotduyev 1989; Zonenshain *et al.* 1990; Nikishin *et al.* 1998*a*,*b*, 2001). The geology of the GC has been studied for at least 150 years and a significant volume of published literature deals with its evolution, although much of this is difficult to access for the international scientific community.

The GC is located in the Black Sea–Caspian Sea region, which is regarded as a mosaic of terranes of Gondwanan, Tethyan and Eurasian affinity that are sometimes controversial in origin (see discussions by Şengör 1984; Zonenshain *et al.* 1990; Dercourt *et al*. 1993, 2000). Accretion of these blocks along the SP occurred throughout the Phanerozoic and, accordingly, orogenic events developed in the GC as such: the Palaeozoic Variscan orogeny, the Triassic–Jurassic Cimmerian orogeny, and the Cenozoic Alpine orogeny. Structural styles of the GC belt are not yet unequivocally fixed and different proposed geometries exist in the literature, even for the major boundary faults separating fundamental tectonic units. It follows that there is still considerable disagreement regarding tectonic mechanisms, simply because there are insufficient diagnostic data. The GC orogenic events are also not well understood in terms of the driving mechanisms. There are major discrepancies concerning the rate of shortening and the nature of the Mesozoic–Cenozoic basement of the GC. Did oceanic crust and lithosphere form during this time or not? In other words, did a complete orogenic Wilson cycle from opening of an ocean to its consumption by subduction and collision of its margins take place along the GC during the Mesozoic–Cenozoic? There is rough agreement regarding the continuing Late Cenozoic pulses of mountain building and uplift, which have resulted from collision–accretion of the Transcaucasus continental block along the southern margin of the SP (Fig. 2). At a regional plate tectonic scale, this corresponds to the final stage of the Alpine orogenic cycle involving the collision between Eurasia and Afro-Arabia 'mega'-continental plates, with the main suture zone of the Tethyan Ocean running through Anatolia and the Lesser Caucasus (Fig. 1). The aim of this paper is first to assess and present the existing data, and then to describe and reinterpret them, as necessary, as well as to present some new data to constrain better the Mesozoic–Cenozoic orogenesis of the

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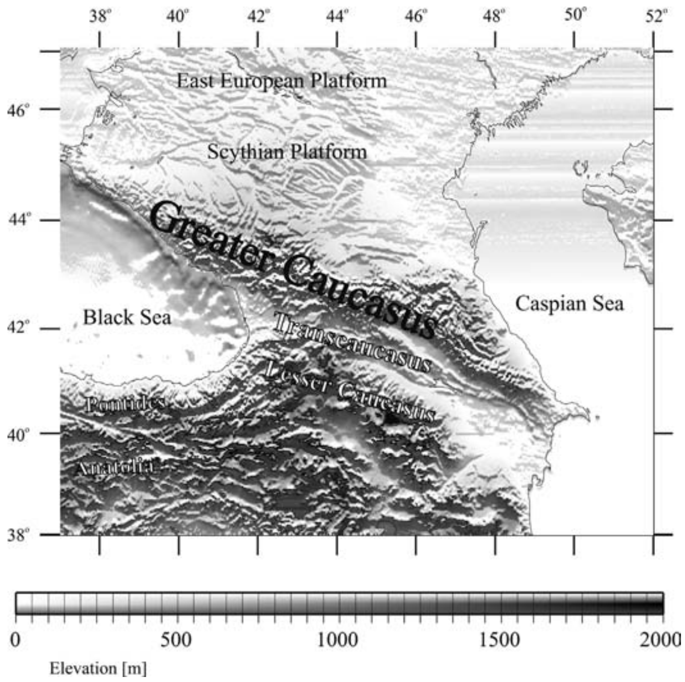

Fig. 1. Digital topographic map of the Greater Caucasus, from the global topography 2 minute database; illumination from N135.

GC: from the development of basins, oceanic or not, to their inversion and/or the collision of continental blocks subsequent to the consumption of an oceanic plate by subduction.

Structure of the Greater Caucasus

The GC belt comprises a basement-core containing strata as old as Proterozoic (Figs 2 and 3), with Jurassic to Eocene formations lying on its flanks.

The deep crustal structure of the GC and Transcaucasus are not known, as the resolution of deep seismic sounding (DSS) lines acquired more than 20 years ago and crossing the belt is insufficient to provide an accurate crustal image. Consequently, discrepancies exist regarding the published deep structure of the belt. The two main competing models of the deep structure of the GC are described below.

(1) The first model argues for a subvertical geometry of all the GC main faults, including the border faults. Somin (2000), for example, argued for a subvertical disposition of the Main Caucasian Thrust (MCT; shown in Fig. 3) at great depths, given its steep near-surface dip (65–80°) along all of its strike and at imaged depths to 3–5 km. Nevertheless, such a steep geometry could also have resulted from deformation of the fault during the final stage of collision. Both reprocessed old and newly acquired geophysical data (Shempelev *et al.* 2001, 2005; Grekov *et al.* 2004; Prutsky *et al.* 2004) were used to show a similarly deep inclination of the MCT at a depth of 80 km. Shempelev *et al.* (2001) and Rastsvetaev *et al.* (2004) proposed the same subvertical geometry of regional faults along profiles crossing different parts of the GC. It was concluded that the boundary of the western GC with the Black Sea is a steep (60–80°) and deep (80 km) major fault (Shempelev *et al.* 2001) linking with the Racha–Lechkhumy Fault Zone (RLFZ; see location in Figs 3 and 4) to the east (Yakovlev 2002, 2005).

(2) The alternative model, proposed by Gamkrelidze (1986), Dotduyev (1987), Giorgobiani & Zakaraya (1989), Baranov *et al.* (1990), Zonenshain *et al.* (1990) and Gustchin *et al.* (1996), and referenced by many others, differs strongly and favours instead flat-dipping thrusts at depth. This is a thick-skinned tectonic model for the GC (Fig. 3) with the orogen interpreted as a collage of two or three northward underthrust slabs. These authors have argued that the SP is thrust upon the Transcaucasus continental block. The MCT *sensu stricto* is thus considered to be a north-dipping flat thrust at depth (Fig. 3) along which the pre-Jurassic basement of the Main Range zone and its overlying Mesozoic cover were presumably displaced southwards some 100 km or more during the Cenozoic. (The MCT *sensu lato* comprises at least two parallel branches at the surface (see Figs 2 and 4) and the northern branch is the MCT *sensu stricto*, along which the crystalline basement thrusts onto the sedimentary succession (see Fig. 3). Similarly, units of the GC belt have been thrust southwards over the Transcaucasus along the RLFZ (Fig. 3). Accordingly, the GC is regarded as a large south-vergent fold-and-thrust belt with its northern limb forming a gently north-dipping monocline toward the SP. Some north-vergent thrusting could be locally present along the western and central part of the GC (Milanovsky & Khain 1963). Back-thrusting is more developed onto the Terek–Caspian foreland, where the northward propagation of the Dagestan nappes contributes to the fan-shaped structure of the eastern part of the belt (Fig. 4; see Ershov *et al.* 2003). Thick-skinned deformation is reported along a north–south profile cutting across the internal part of the GC, with thin-skinned deformation prevailing on the southern front in the Rioni and Kura basins. A north–south profile across the western GC (Fig. 5; Robinson *et al.* 1996) also shows thick-skinned deformation with imbricate structures involving the basement. It can be seen that the NW–SE and WNW–ESE faults parallel to the general grain of the belt are thrust faults flattening at depth whereas the NNW–SSE faults transverse to the belt are steeper (Koronovsky 1984; Giorgobiani & Zakaraya 1989; Philip *et al.* 1989; Giorgobiani 2004; Fig. 4).

The Fore-Caucasus region, which lies on the SP, evolved in conjunction with the GC. From Latest Eocene–Oligocene times, two flexural basins developed, separated by the elevated zone called the Stavropol High (Figs 2 and 3). This comprises a north–south elongated and anomalously thick crustal block (see Kostyuchenko *et al.* 2004) that from early Mesozoic times never significantly subsided. The Terek–Caspian foreland basin to its east and the Kuban foreland basin to its west developed during the Cenozoic, both showing a high subsidence rate during the Oligo-Miocene ('Maykop' facies). Thus, what is peculiar about the Fore-Caucasus area is that 'foreland type-like basins' developed in front of the more topographically subdued eastern and western parts of the belt but not in front of its topographically highest central part (see Ershov *et al.* 2003).

Several basins also developed south of the GC belt, in the Shatsky Ridge–Transcaucasus area, in Oligo-Miocene times. These are, from west to east, the Tuapse, Rioni and Kura basins (Figs 2 and 5). They are reported to be flexural in type and related to Eocene compression (Milanovsky & Khain 1963; Gamkrelidze 1986; Nikishin *et al.* 1998*b*). However, the history of the Kura Basin is more complex, as it is the western prolongation of the South Caspian Basin (Brunet *et al.* 2003).

A review of the Early Mesozoic tectonic evolution of the Greater Caucasus

The Triassic and Jurassic history of the area is not well constrained and is still a matter of considerable debate. The extent and age of Cimmerian orogenic phases, in Late Triassic or Early Jurassic, Mid-Jurassic and Late Jurassic times, as well as the successive rifting events, are not confidently known (see Nikishin *et al.* 1998*a*,*b*).

Early Triassic basin development and Late Triassic Eo-Cimmerian tectonics

Permo(?)-Triassic rifting and volcanism (and, probably, magmatism-related doming) are widespread in the Fore-Caucasus

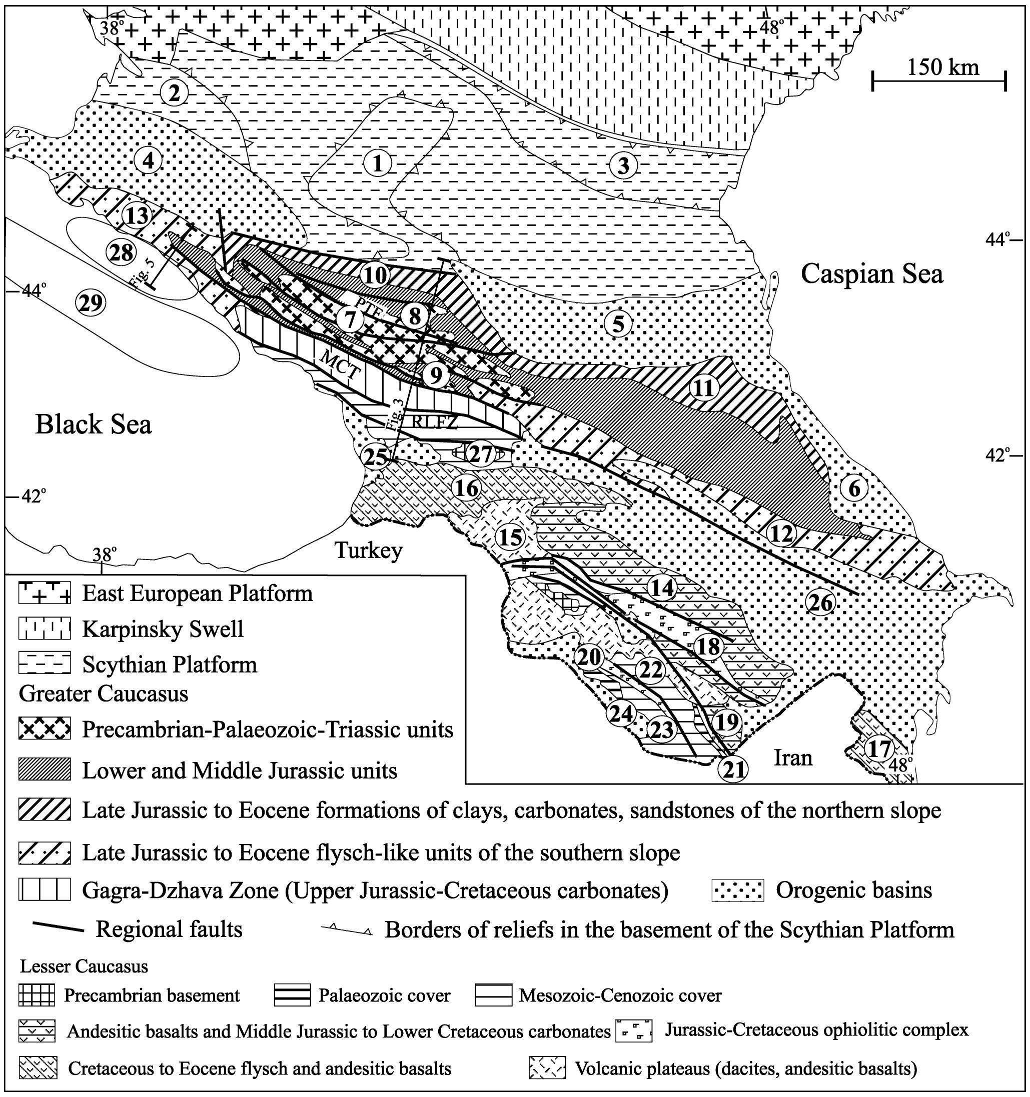

Fig. 2. Simplified geological map of the Greater Caucasus and Lesser Caucasus (from Milanovsky & Khain 1963) and locations of cross-sections shown in Figures 3 and 5. PTF, Pshekish–Tyrnauz Fault; MCT, Main Caucasian Thrust; RLFZ, Racha–Lechkhumy Fault Zone. Encircled numbers: 1–6 are zones of the Scythian Platform: 1, Stavropol High; 2, Azov–Berezan High; 3, Manych Basin; 4, Kuban Basin; 5, Terek–Caspian Basin; 6, Kussar–Divitchi Basin; 7–13 are zones of the Greater Caucasus: 7, Peredovoy Zone; 8, Betcha Anticline; 9, Svanetia Anticline; 10, Laba–Malka Monocline; 11, Dagestan Folded Zone; 12, Flysch Zone of southeastern GC; 13, Flysch Zone of north-western GC; 14–24 are zones of the Lesser Caucasus: 14, Somketo–Karabakh Zone; 15, Artvin–Bolnisi Zone; 16, Adzharo–Trialet; 17, Talesh; 18, Sevan–Akera; 19, Kafan; 20, Vedin; 21, Zangezur; 22, Mishkhan–Zangezur Massif; 23, Ararat–Djulfa Massif; 24, Araks Basin; 25–29 are intramontane zones of the Transcaucasus and Black Sea: 25, Rioni Basin; 26, Kura Basin; 27, Dzirula Massif; 28, Tuapse Basin; 29, Shatsky Ridge.

region and in the northern part of the GC (Nazarevich *et al.* 1986; Lordkipanidze *et al.* 1989; Tikhomirov *et al.* 2004). The geodynamic setting of such tectonics is still debated: was this a back-arc setting or not?

Throughout the area, including the northern GC, there was also a period of Late Triassic compression (Eo-Cimmerian tectonic phase) during which all the Permo(?)-Triassic basins were inverted (Nikishin *et al*. 1998*a*,*b*, 2001; Gaetani *et al.* 2006). The compressive event is probably related to the collision–accretion of Gondwana-derived blocks (which together form the composite Iran plate; Saïdi 1995; Besse *et al.* 1998) SE of the GC when the Palaeotethys Ocean closed along the

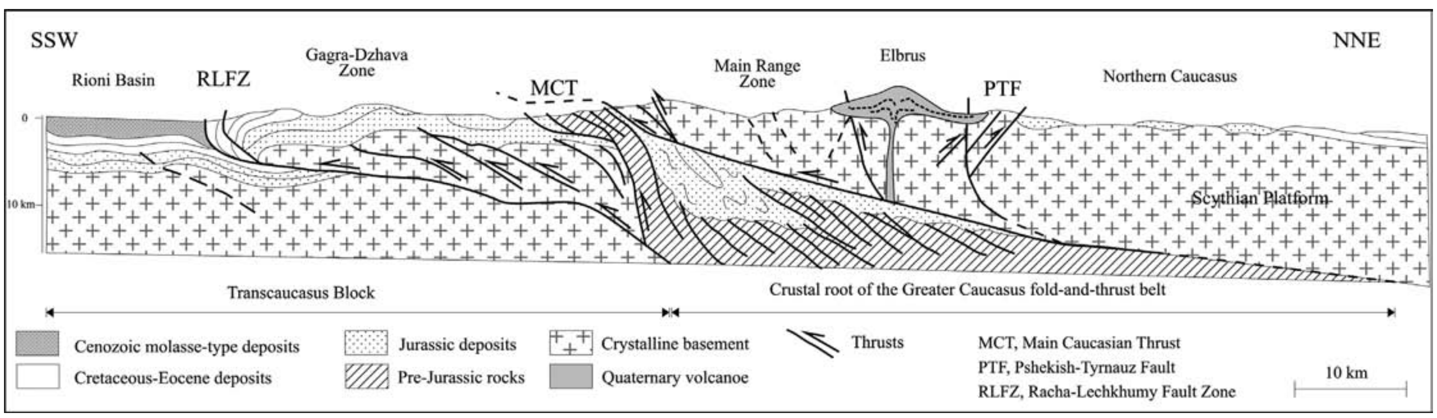

Fig. 3. Section across the central part of the Greater Caucasus showing the southward vergence of the whole belt and the major thrusting of the belt over the Transcaucasus (Dotduyev 1987). (Section location is shown in Fig. 2.)

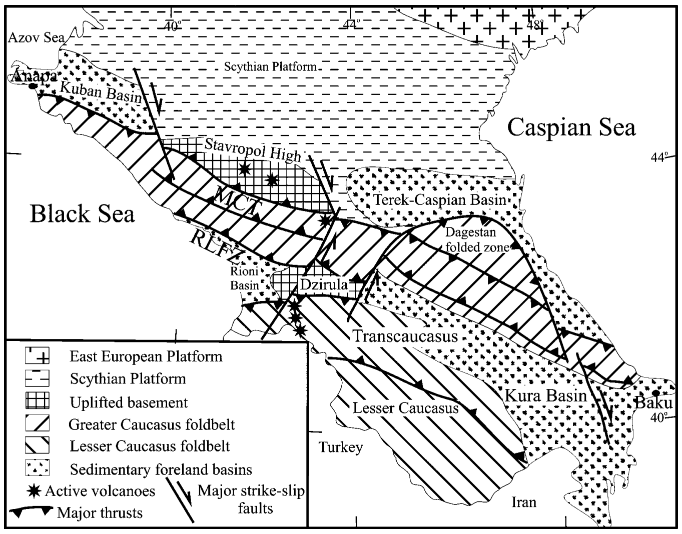

Fig. 4. Tectonic map of the Greater Caucasus area from Ruppel & McNutt (1990) (other sources: Milanovsky & Khain 1963; Kotansky 1978; Dotduyev 1987; Philip *et al.* 1989).

Talesh–Alborz–Aghdarband zones, in the area of the present-day South Caspian Basin (Stöcklin 1968; Davies *et al.* 1972; Şengör 1984; Alavi 1991; Ruttner 1993; Dercourt *et al*. 2000, and references therein). This Eo-Cimmerian orogeny has also been clearly identified as a major event in the Turkish Pontides (Okay 2000; Okay *et al.* 2006). It should also be noted that (1) the Black Sea was certainly not developed by the Late Triassic, and the Pontides were therefore close to the Transcaucasus–GC; (2) the Pontides–Transcaucasus–Talesh–Alborz–Aghdarband–GC zones probably together formed a contiguous part of the widespread Eo-Cimmerian orogenic belt.

Early Jurassic to Mid-Jurassic

A field study carried out in 2003 led to the postulation of a model of the GC in Jurassic times by Saintot *et al.* (2004). A new rifting phase occurred during the Early Jurassic (Zonenshain *et al.* 1990; Nikishin *et al*. 1998*a*,*b*, 2001, and references therein) under a transtensional stress regime with a nearly east–west-directed tensional stress axis (Sébrier *et al*. 1997; Saintot *et al.* 2004). Thus, this transtensional Early Jurassic rifting shares some similarities with the model of Banks & Robinson (1997) for the Black Sea region, which surmises that the Early Jurassic GC Basin corresponded to an en echelon set of rhomb-shaped depocentres. Early Jurassic rift activity is also reported in the Eastern Pontides, which were adjacent to the GC at that time (e.g. Okay & Şahintürk 1997) and in the South Caspian Basin (Early(?) to Mid-Jurassic times; Brunet *et al.* 2003). GC rifting continued through part of the Mid-Jurassic.

Extrusive magmatism (mainly rhyolitic) accompanied the GC rifting phase during Sinemurian–Pliensbachian times (Lordkipanidze *et al.* 1989). Sediments of this age are represented by deltaic(?) coarse sandstones NW of the belt, by deep marine mudstones–sandstones in the central part, and by shallower mudstones–sandstones to the south. In Toarcian times, from north to south, shelf to deep marine mudstones–sandstones were deposited, with no record of volcanic activity. In Aalenian–Bajocian units, sedimentary facies laterally vary from continental to deep marine. The Aalenian and Bajocian periods are also characterized by bimodal rhyolitic and basaltic extrusive rocks (from mantle and crustal sources) in a subaerial as well as a shallow-marine environment. In the model, the western part of the GC evolved during the Early and Mid-Jurassic as the western margin of the rift with shallow-water sedimentation and subaerial extrusion of lava flows. Deep-water sediments are encountered towards the present-day central part of the belt (crossing the inferred, north–south-oriented normal faults), associated with mid-ocean ridge basalt (MORB)-like tholeiitic basalt extrusion during the Aalenian. Not only partial melting of asthenosphere is implied, but also a high degree of extension, approaching that required for oceanic crust development in the present-day central part of the GC belt. The total thickness of the Lower Jurassic to Aalenian unit in some parts of the GC Basin is more than 5000 m, and it is mainly composed of black shales and deep-water sandstone turbidites (as well as the volcanic rocks and pyroclastic deposits). The Aalenian extensional phase has been well documented in the field with, for example, the presence of a large NW–SE-trending normal fault, east of the Kuban Basin. Toarcian to Aalenian units are tilted along this fault (Fig. 6) and the minimum downthrow should be of several hundreds of metres. The age of fault activity is constrained by overlying, sealing Upper Aalenian units.

In Bajocian times, a huge quantity of pyroxene-bearing basalts were extruded, and formed a subaerial to shallow-water volcanic chain on the southern margin of the basin (presumably accompanied by uplift at the rift margin). Synchronously with the formation of this relief, conglomerates (reworking the lavas) were deposited toward the depocentre of the basin to the north.

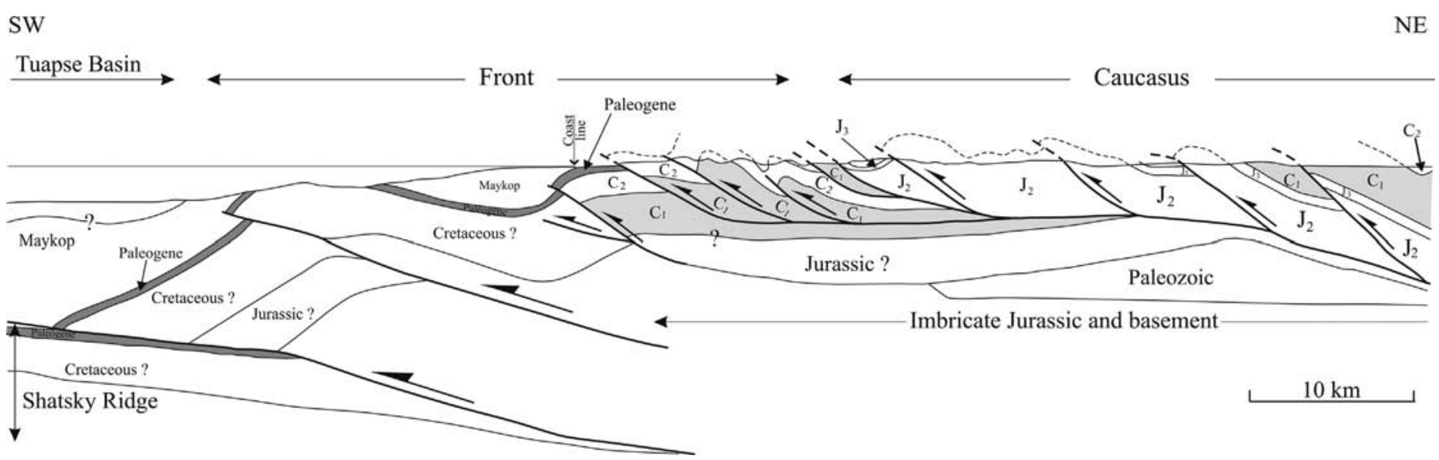

Fig. 5. Section across the western part of the Greater Caucasus showing the basement involved in south-vergent, flat thrusting. Offshore is shown an interpretation of the seismic line SU8040 (from Robinson *et al.* 1996). (Section location is shown in Fig. 2.)

Fig. 6. Photograph of a progressive unconformity created by a major synsedimentary normal fault within Aalenian deposits dipping to the NE. The normal fault, with a NW–SE strike, is located to the left of the photograph and may be followed for some 10 km along the upper rims of the Upper Kuban valley, on the northern side of the central GC. This normal fault testifies to the Mid-Jurassic extensional tectonics of the GC Basin and contributes to the along-strike variation of sedimentary thickness (photograph at 43°46.6′N, 42°11.12′E by M. Sébrier).

The calc-alkaline nature of the Bajocian lavas has been formerly interpreted as indicative of a subduction-related volcanic arc marking the incipient subduction of a large oceanic plate such as the Palaeotethys along the GC (see, for example, the southward subduction of Palaeotethys along the GC as described by Şengör (1984)). However, this hypothesis seems very unlikely because the above-mentioned Bajocian volcanic rocks are also spread over the Transcaucasus and there is no evidence for a subduction zone along the southern edge of the GC (i.e. remnants of an accretionary prism, high-pressure metamorphism, ophiolitic fragments, etc.). The calc-alkaline nature of the Bajocian lavas can also be explained by the GC rift being in a back-arc setting relative to a subduction zone located far to the south in the Lesser Caucasus (see Adamia *et al.* 1981; Gamkrelidze 1986; Panov 2004). The Artvin–Bolnisi zone lying between the Transcaucasus and the suture zone of the Lesser Caucasus (location shown in Fig. 2) is a good candidate for a subduction-related volcanic arc during the Early and Mid-Jurassic, with shallow-water to continental sediments and major calc-alkaline volcanism (Adamia *et al.* 1981; Gamkrelidze 1986; Panov 2004). It is not uncommon for lavas extruded in a back-arc rift setting, but close to the volcanic arc, to show such calc-alkalinity. Therefore, it cannot be excluded that the Bajocian lavas were extruded during what could still be considered as a synrift stage of basin evolution, continuing from the Aalenian. However, it is noted that, whereas structural constraints (e.g. synsedimentary normal faulting) clearly exist to define the Aalenian succession as synrift, there are no such structural constraints for the Bajocian units. Indeed, the widespread occurrence of Bajocian calc-alkaline volcanic rocks that can be encountered from the Lesser Caucasus to the MCT may also simply suggest an expansion of the subduction-related volcanic arc from some 50–100 km width in Aalenian time, restricted by the Artvin–Bolnisi zone, to nearly 200 km in the Bajocian, thus merging with the southern part of the GC Basin. Shallowing of the subducted slab could explain such an encroachment of the arc into the previously back-arc setting. If this were the case, rift activity in the GC Basin would have stopped (given that a rather flat-dipping slab does not favour the opening of a back-arc basin; eg. Lallemand *et al.* 2005, and references therein) and, therefore, the Bajocian volcanic rocks should be considered as occurring at the onset of the post-rift stage of GC Basin development. The available observations, relating to only the calc-alkaline character of the Bajocian lavas and their widespread occurrence, cannot discriminate between these two possibilities.

The Bathonian unit (where not absent) is composed of a greywacke siltstone unit into the basin and regressive coal-bearing terrigeneous sediments on its southern margin. The Upper Jurassic unit lies transgressively and discordantly on the Middle Jurassic unit. It reportedly lies conformably on the Middle Jurassic unit along the present-day southern slope of the central and eastern part of the GC Basin (Gamkrelidze 1986; Zonenshain *et al.* 1990; Nikishin *et al.* 1998*a*,*b*), although field observations made in the same central area (by M. Sébrier in 2004) revealed an unconformity between the Callovian deposits and underlying units. It is also worth noting that Cenozoic deformation is so intense in the so-called Flysch Zone (see Fig. 2) that no clear conclusion can be made regarding the detailed relationships between Mesozoic units. A compressional event has been proposed to have occurred in Bathonian times, resulting in the inversion of the margins of the basin (Adamia *et al.* 1981), although it may be that this unconformity is simply related to the cessation of rifting and the onset of post-rift basin development, such as recorded in many rift basins (see, e.g. Coward *et al.* 1987; Tankard & Balkwill 1989; Frostick & Steel 1993; Williams & Dobb 1993; Busby & Ingersoll 1995; Stephenson *et al.* 1996; Cloetingh *et al.* 1997; McCann & Saintot 2003). (Brunet *et al*. (2003) also pointed out that the regional Bathonian unconformity around the South Caspian Basin may be a ‘break-up unconformity’ marking the onset of sea-floor spreading rather than the occurrence of a compressive tectonic event.)

Nevertheless, in the southernmost part of the GC (in Georgia), Callovian strata overlie open folds in Middle Jurassic strata, constraining a gentle folding event to the Bathonian. In the central part of the GC, north of the MCT, highly folded Early Jurassic strata are overlain by subhorizontal layers of Upper Jurassic and Cretaceous platform-type deposits. According to Belov *et al.* (1990) and Somin (2000), they are in place and indicate that the Bathonian folding was significant and involved intense shortening. Other authors (e.g. Korsakov *et al*. 2001) have considered that the Upper Jurassic and Cretaceous strata are in an allochthonous position and, accordingly, that the thrust sheet and the folding developed together during Alpine orogenesis (implying the occurrence of a folding phase, followed by the development of an erosional surface and then thrusting of nappes along a décollement level). On the northernmost slope of the belt, the angular discordance between transgressive Callovian and older rocks disappears. Published cross-sections (e.g. Panov 2002, 2004) show south-vergent folds and thrusts affecting strata older than and including Bajocian, and no sealing by younger sediments (which are absent). In the northern part of each of these cross-sections lies a gentle monocline composed of Upper Jurassic and Cretaceous units underlain by Lower Jurassic units without any angular unconformity as might be expected to be related to a compressional phase during Bathonian times. (A Bathonian stratigraphic gap indeed exists locally, the Bathonian being a time of worldwide regression.) No important or diagnostic compressive structures (such as folds and thrusts) were observed in Middle Jurassic rocks sealed by the Callovian by the senior author during fieldwork in 2003 in the northern part of the belt (see Saintot *et al.* 2004). What was observed is a gently tilted unit (like the Aalenian unit) below the Callovian transgressive unit. Going southward across the belt, closely and tightly folded Lower and Middle Jurassic strata can be observed (Fig. 7). The same style of folding is observed some 10 km towards the Black Sea coast in Palaeocene rocks (Fig. 8). SE-vergent minor thrusts are also common in Lower and Middle Jurassic units, similar to the younger strata. The systematic analysis of brittle structures within the GC also strongly suggests that only one set of reverse faults developed in Jurassic and younger strata and that this set is related to the Cenozoic palaeo-stress field (Fig. 9; see discussion

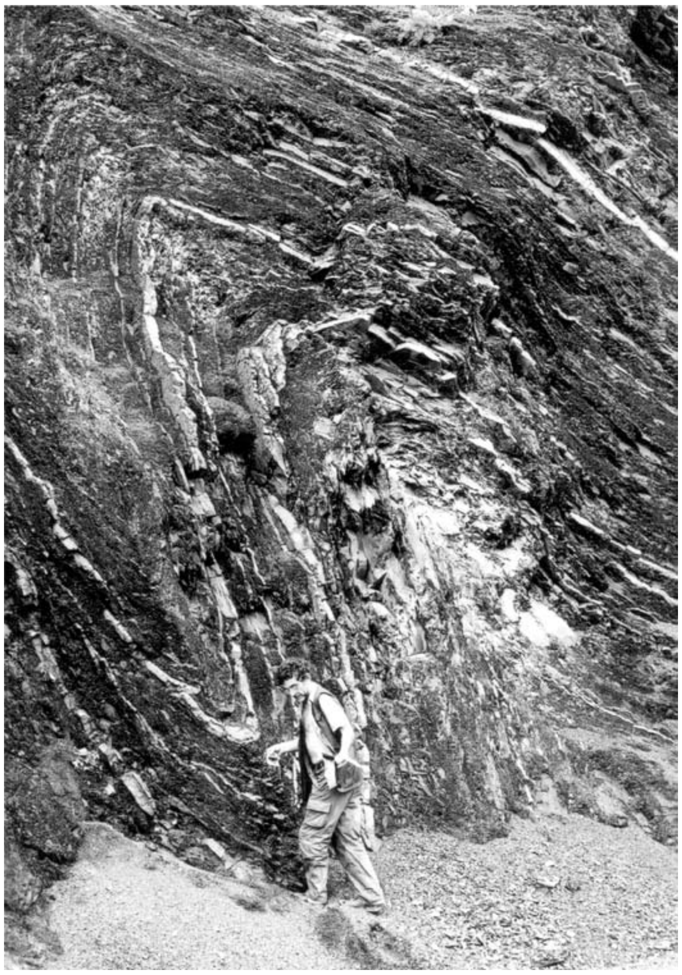

Fig. 7. Photograph of folded Aalenian–Bajocian unit (Pshish Formation) of the western Greater Caucasus (photograph by A. Saintot; S. Korsakov for scale). Fold axes strike NW–SE to WNW–ESE.

and analyses of structures related to Cenozoic shortening by Saintot & Angelier (2002)). The localized angular unconformity at the base of Upper Jurassic strata thus probably records not more than a phase of gentle compression of the GC Basin (or, indeed, only isostatic readjustments at the syn- and post-rift transition), affecting units from place to place, rather than the complete inversion of the GC Basin. (In Lower Middle Jurassic rocks there is no evidence of intense folding and thrusting that can be ascribed unequivocally to a Bathonian compressional event, most of the deformation being clearly Cenozoic in age). In summary, the pre-Callovian unconformity remains a matter of debate. It could record either the transition between syn- and post-rift phases in the GC Basin or, alternatively, a weak compressive event related to the accretion of crustal blocks along the active continental margin to the south.

The Callovian–Eocene Greater Caucasus Basin

The GC Basin evolved dominantly as a post-rift (thermally subsiding) basin from the Callovian until the Late Eocene following its Early to Mid-Jurassic episodes of rifting. A thickness of 6–8 km of calcareous, mainly Cretaceous, flysch-type sediments was then deposited and most of the Greater Caucasus Mountains corresponds to the so-called Flysch Zone of the southern limb of the GC (Fig. 2; Milanovsky & Khain 1963; Lordkipanidze 1980; Koronovsky 1984; Gamkrelidze 1986; Belousov *et al.* 1988; Adamia & Lordkipanidze 1989; Zonenshain *et al.* 1990). The nature of the underlying crust has not been established, although Ershov *et al.* (2003) estimated a crustal thickness of 15–17 km, suggesting that it corresponds to thinned continental crust. Such an interpretation is in agreement with the absence of oceanic crustal remnants in the belt. It follows that the basin was probably not floored by significant oceanic crust (see also the important discussion by Ershov *et al.* 2003, p. 102).

The Callovian conglomerates and calcareous sandstones clearly belong to the post-rift succession of the GC Basin. They unconformably overlie the oldest units on an erosional surface. Upward, the Callovian unit becomes marly, indicating platform subsidence. In Late Jurassic times, sandstones and clays filled in the sedimentary basin and reef limestones developed towards its margins. Kimmeridgian–Tithonian gypsum-bearing and lagoonal sediments were deposited on the northern (Laba–Malka zone) and southern margins (in Georgia). A very thick Cretaceous to Eocene greywacke siltstone flysch-like unit with clastic limestones in the Lower Cretaceous interval conformably overlies the Upper Jurassic sequence. (The Lower Cretaceous succession is 750–1600 m thick, the Upper

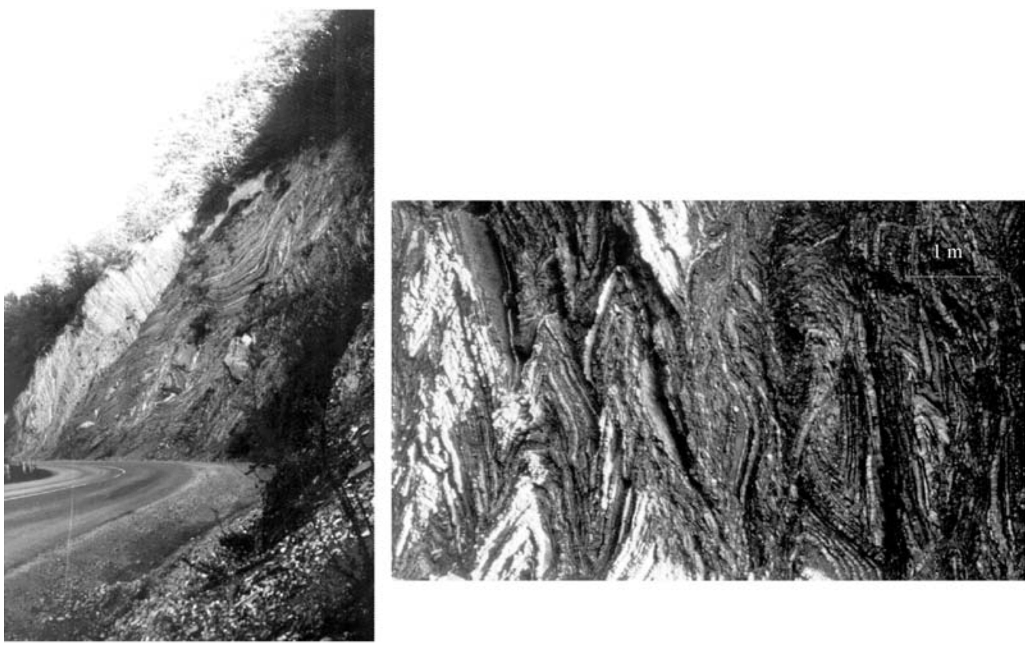

Fig. 8. Two photographs of the Lower Palaeocene flysch-like folded unit of the western Greater Caucasus along the Black Sea coast. Fold axes strike NW–SE to WNW–ESE (photographs by A. Saintot).

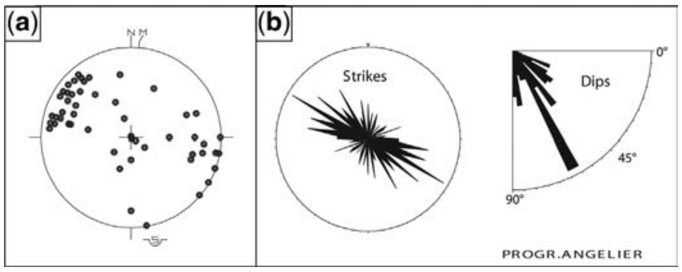

Fig. 9. (**a**) Stereoplot of 60 close and tight fold axes collected in rocks of Early and Mid-Jurassic age of the western Greater Caucasus (Schmidt's projection, lower hemisphere). (**b**) Attitudes (strikes and dips) of 215 bedding planes collected in rocks of the western Greater Caucasus from Late Jurassic to Eocene in age. The attitude of folds is the same in both stratigraphic intervals: the WNW–ESE close folds in Late Jurassic to Cenozoic rocks are also observed in Early and Mid-Jurassic rocks. Most of the compressive structures and tectonic contacts (such as thrusts) measured in Early and Mid-Jurassic rocks are consistent with a NNE–SSW Cenozoic compression (see Saintot & Angelier 2002).

Cretaceous 500–900 m thick, and the Palaeocene–Eocene, 600–850 m thick).

Restorations by Yakovlev (2002, 2005) along several profiles crossing the GC clearly show that the RLFZ (Figs 2 and 3) was the southernmost normal fault bordering the GC Basin, involved in controlling the northward increase of sedimentary thickness from the Mesozoic–Cenozoic Transcaucasus stable platform to its subsiding part, the Gagra–Dzhava zone (see Fig. 3). They also show that the cumulative primary normal displacement along the RLFZ was much larger than the secondary reverse one. The most important normal fault at this time, however, was the Utsera fault, limiting the Gagra–Dzhava zone and the flysch zone, in which the thickness of Mesozoic and Cenozoic deposits reaches 12–15 km.

It was also during this time that the eastern Black Sea Basin developed, close to the GC Basin and south of the Shatsky Ridge (a western prolongation of the Transcaucasus; see Fig. 5), although the precise timing is still matter of debate: Late Cretaceous and Palaeocene according to Finetti *et al.* (1988), Eocene according to Lordkipanidze (1980), and Late Palaeocene–Eocene according to Robinson *et al.* (1996) and Shreider *et al.* (1997). Analysis of kinematic data by Saintot & Angelier (2002) revealed that a transtensional stress field affected the GC Basin during the Eocene (with an east–west trend of extension), which those workers considered to be a far-field effect of rifting (or, at least, of rift reactivation) in the eastern Black Sea Basin. In any case, during Palaeocene–Eocene times, prior to the main shortening event, the GC Basin was a deep-water basin, with a sedimentary infill of 10 km on average (Borsuk & Sholpo 1983), similar to the eastern Black Sea and South Caspian basins (Zonenshain & Le Pichon 1986; Nikishin *et al.* 1998*b*) and linked with the latter but apparently separated from the former by the Shatsky Ridge (see Fig. 2).

Cenozoic to present-day shortening along the Greater Caucasus

The main orogenic phase is considered to extend from Late Eocene to Early Oligocene time by Shardanov & Peklo (1959), Beliaevsky *et al.* (1961), Milanovsky & Khain (1963), Grigor'yants *et al.* (1967), Khain (1975, 1994), Milanovsky *et al.* (1984), Muratov *et al.* (1984), Giorgobiani & Zakaraya (1989), Robinson *et al.* (1996), Lozar & Polino (1997), Robinson (1997), Nikishin *et al.* (1998*a*,*b*, 2001) and Mikhailov *et al.* (1999), with pulses of orogeny encompassing the rest of Cenozoic to the present day. According to other workers, the orogeny did not begin prior to the Miocene, in Sarmatian time (e.g. Dotduyev 1987; Shcherba 1987, 1989, 1993; Zonenshain *et al.* 1990; Kopp 1991; Kopp & Shcherba 1998). In general, according to the Russian literature, there are two distinct orogenic processes recorded in the GC. The first of these is understood to have occurred in Late Eocene–Early Oligocene times, and involved folding with significant crustal shortening but without major uplift of rocks at the surface. The second is considered to have occurred in Miocene times, and is characterized by limited shortening (reported to be 5–10%) and no significant folding but rather by significant uplift, thus explaining the deposition of Sarmatian conglomerates.

Late Eocene

What follows is a summary of the main arguments used by authors to demonstrate that the inversion of the basin began in Late Eocene times.

(1) An angular unconformity of Maykop (Oligo-Miocene) on deformed older units is regionally observed in the field (Milanovsky & Khain 1963; Khain 1975, 1994; Borukaev *et al.* 1981; Rastsvetaev & Marinin 2001; Banks, pers. comm.) and on seismic lines (Tugolesov *et al.* 1985; Robinson *et al.* 1996; Banks, pers. comm.) on both sides of the GC. On the southern slope of the central GC, there is (a) a Late Eocene olistostrome unit (10–400 m thick) coeval with southward thrusting of the GC Basin (Khain 1975, 1994) and (b) a deltaic, southward prograding sandy facies in the Oligocene unit coeval with the uplift–emergence of part of the GC (M. Sébrier's field observation). Sharafutdinov (2003) dated the folding event as latest Eocene–Early Oligocene on the northern slope of the GC and in the Fore-Caucasus, and he reported Early Oligocene folds, olistostromes (confirming the existence of back-thrusts), angular unconformities and the tectonic removal of a large part of the section with everything being overlain by flat-lying strata of Mid-Oligocene age. The youngest unit prior to the onset of deformation is Late Eocene in age (Khadum Formation). The reported features, including the folds (especially if they are related to slumping), imply synsedimentary deformation in a foreland developing at the front of a propagating back-thrust.

(2) A high rate of tectonic subsidence occurred at the beginning of the Oligocene in the Indolo–Kuban and Terek–Caspian basins as shown by burial history modelling (back-stripping analyses of wells and numerical modelling of lithospheric deformation) by Nikishin *et al.* (1998*a*), Ershov *et al.* (1999) and Mikhailov *et al.* (1999). Those workers assumed that the Indolo–Kuban, Terek–Caspian and Tuapse, Kura and Rioni troughs developed as flexural foreland basins in response to lithospheric compression from the south during Late Eocene times (resulting from the closure of Neotethys and collision south of the Transcaucasus area). Ershov *et al.* (2003) discussed mechanisms other than foreland flexure for the formation of these basins, including the role of mantle processes occurring at the cessation of shortening, related to the underthrusting of thinned continental crust beneath the basins.

(3) Lozar & Polino (1997) carried out a study based on nannofossils occurring in Maykop sediments of the Kuban Basin and of Upper Cretaceous rocks on the northern slope of the western GC. The base of the Maykop group is inferred to be Late Eocene–Early Oligocene in age and its lowermost part contains a reworked assemblage (80% of the total assemblage) of Late Cretaceous and Palaeogene nannofossils. These are very well preserved (with, for example, intact spines), implying that they were not transported over long distances. The sediment source was the area of the present GC where, indeed, Late Cretaceous and Palaeogene sediments were eroded. However, although there is general agreement on the Late Eocene uplift of the central part of the GC, this area was not yet actually above sea level by this time according to Kopp & Shcherba (1985) and Ershov *et al.* (1999, 2003). The types of Maykop nannofossils found *in situ* suggest restricted environmental conditions, leading to the interpretation that environmental changes occurred during

Late Eocene–Oligocene times, either with the onset of a cooler climate or as a result of the isolation of the Paratethys domain by the uplift and emergence of an orogenic belt acting as a barrier along the Pontides–Lesser Caucasus.

The Late Eocene compressional palaeostress field responsible for the inversion of the GC Basin has been determined through tectonic analysis by Saintot & Angelier (2002). It was oriented NE–SW to NNE–SSW, leading to the development of NW–SE and WNW–ESE dip-slip thrusts in the GC. The main features and chronology of this tectonic phase have been established as follows: (1) on the northern slope of the GC, where the regional structure is a monocline, the Palaeocene strata are clearly affected by the inferred compressional palaeostress field (see details of measurements and site numbers given by Saintot & Angelier (2002)), whereas no related reverse and strike-slip faults can be observed affecting the overlying Miocene rocks studied by Saintot & Angelier; (2) the palaeostress field has also been recorded in Middle Eocene rocks along the southwestern coast (see details of measurements and site numbers given by Saintot & Angelier (2002)); (3) this palaeostress field is the only one recorded during pre- (e.g. Fig. 10), syn- and post-folding phases (Saintot & Angelier 2002).

Miocene to present day

From Sarmatian times (Mid-Miocene) until the present, pulses of compressional deformation have affected the GC (Belousov 1940; Shardanov & Peklo 1959; Beliaevsky *et al.* 1961; Milanovsky & Khain 1963; Shcherba 1987, 1989, 1993; Giorgobiani & Zakaraya 1989; Kopp 1989, 1991, 1996; Rastsvetaev 1989; Zonenshain *et al.* 1990; Milanovsky 1991; Khain 1994; Kopp & Shcherba 1998; Nikishin *et al.* 1998*b*). However, it appears as though the present-day structure of the GC is inherited mainly from the Sarmatian compressional pulse. The Sarmatian sedimentary unit surrounding the belt comprises syndeformational conglomerates reflecting the growth of topography at this time (Mikhailov *et al.* 1999) and, indeed, the emergence of the GC belt as a whole, the central GC having already been uplifted since the latest Eocene (Khain 1994; Lozar & Polino 1997; Ershov *et al.* 1999, 2003).

The present-day displacement of Arabia relative to Eurasia by several centimetres per year is recorded throughout the GC. The indentation of Arabia occurs at Bitlis–Zagros and deformation propagates towards the GC. This indentation has produced large strike-slip faults along which the Anatolian block escapes westward. In the GC, both strike-slip faults and thrusts actively accommodate deformation. Earthquake focal mechanisms reveal that the whole Caucasian area is under a north–south compressional stress regime (Gushtchenko *et al.* 1993; Gushtchenko & Rebetsky 1994; Mikhailov *et al.* 2002), a continuation of the inferred Sarmatian palaeostress regime (Saintot & Angelier 2002). The 'Caucasian' NW–SE and WNW–ESE faults act as oblique reverse faults. The depth distribution of earthquakes is limited to the crust and the overlying sedimentary succession; no deeper earthquakes are observed, nor has a Benioff Zone been imaged. Earthquakes at depths of 10–15 km are related to strike-slip faults, whereas deeper hypocentres are along thrust faults. Also, it is observed that along single focal zones, the depths of hypocentres increase northwards (Gamkrelidze 2005) along gently north-dipping planes. In particular, the identified focal plane of the 29 April 1991 Racha earthquake (Mw = 7) exhibits a dip angle of 20–40° north (Triep *et al.* 1995). The distribution of seismicity also indicates the propagation of the GC front southwards to the offshore Shatsky Ridge (a western prolongation of the Transcaucasus; Fig. 5), and to the Rioni and Kura basins (Figs 2 and 4). On seismic lines crossing the offshore western GC (Finetti *et al.* 1988), it can be observed that, with the continuing compression, the Tuapse Basin as a whole overthrusts the Shatsky Ridge with a southward propagation of the GC deformation front. Active thrusting of the GC also affects the Rioni and Kura basins. The Oligocene–Early Miocene sedimentary infill of these two basins has been incorporated into the south-vergent fold-and-thrust belt during the Mid-Miocene compressional phase.

Fig. 10. Photograph of reverse faults developed prior to the tilting of beds under a NE–SW compression (Late Cretaceous flysch-like unit of the western GC) and stereoplots of the related stress tensors (calculated for both attitudes of beds: present-day and restored to horizontal stress tensors obtained from inversion of the fault slip data as given by Saintot & Angelier 2002). (Photograph by A. Saintot; J. Angelier for scale.)

The faults transverse to the GC belt have been invoked as conduits for Quaternary volcanism (Milanovsky *et al.* 1984; Giorgobiani & Zakaraya 1989; Lordkipanidze *et al.* 1989; Koronovsky *et al.* 1997). These faults, which were very active during the Cenozoic, have segmented the GC and Transcaucasus area as well as the WNW–ESE 1250 km long south-vergent frontal thrust of the GC (Giorgobiani 2004). Similarly, a large NE-striking left-lateral fault, with a reported offset of 90 km (Philip *et al.* 1989), was proposed as the conduit for the Kazbek volcano (see location of volcanoes shown in Fig. 4). However, the evidence for large strike-slip displacements along such structures in the GC belt and Transcaucasus area remains very speculative.

Some characteristics of the inversion of the Greater Caucasus Basin

Using simple area-balancing restoration of cross-sections, Ershov *et al.* (2003) estimated the amount of shortening along the GC to have been 200–300 km (as also reported by Khain 1982; Zonenshain & Le Pichon 1986; Shcherba 1993; Nikishin *et al.* 1998*b*). Such an estimate is in agreement with the inferred plate kinematics of the area, which suggests a 400 km displacement of Arabia northwards to (fixed) Eurasia from Oligocene times and takes into account the amount of shortening in the Lesser Caucasus area. However, field observations (M. Sébrier) of the structural relationships between Mesozoic GC formations indicate that the shortening accommodated by the MCT *sensu lato* is of the order of some tens of kilometres and that each of the few other major thrusts should accommodate some 2–5 km (e.g. along the eastern part of one of the MCT branches, the Lower Jurassic units are thrust over themselves). It follows that the shortening across the GC as a whole could be much less, as little as 100 km.

The central part of the GC belt, which has the highest elevation and the highest rate of Neogene to present uplift, corresponds to the thinnest part of the Aalenian GC rifted lithosphere. Thus, the anomalously high elevation in this area could be a consequence of the subduction of highly thinned continental lithosphere (if not partly oceanic, as mentioned earlier). The lithospheric root might also be comparatively more important in the central part of the GC because collision and shortening was concentrated there, directly in front of the indenting Arabian plate. The Quaternary and still active uplift of the central part of GC could

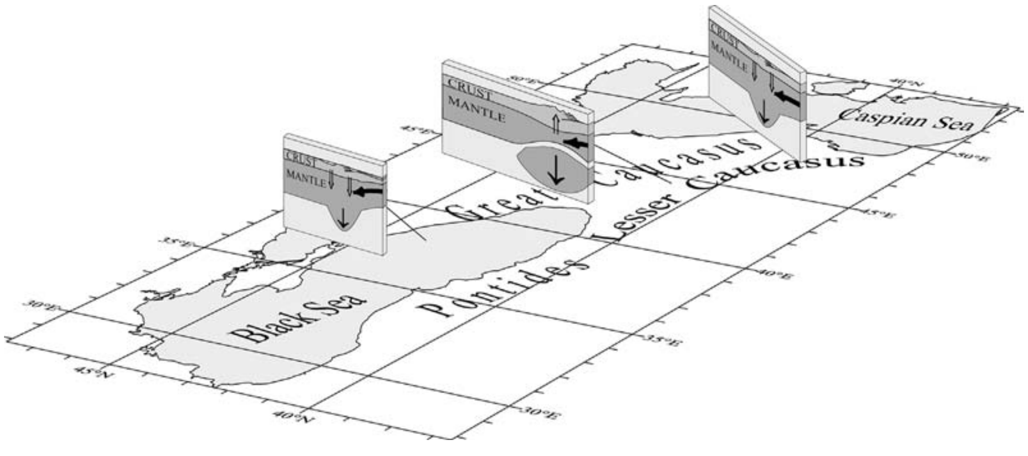

Fig. 11. Main subsidence-driving mechanisms for the foreland stage of basin evolution and uplift of the central part of the GC as a result of delamination of the root (from Ershov *et al.* 2003).

be the result of delamination of a lithospheric root, as suggested by numerical modelling (Ershov *et al.* 2003; Fig. 11) and tomography (Brunet *et al.* 2000) results, with the Quaternary volcanism being linked to this deep process. Lithospheric roots would not have been so well developed in the western and eastern GC because there was less shortening there and it was accommodated differently. In the eastern GC, shortening is symmetrically accommodated by the fan-shaped development of foreland structures. To the west of the central GC, no large lithospheric root is expected because the cumulative shortening there is significantly less because of the western escape of Anatolia.

The structural style of the GC belt agrees with the inversion of a deep basin developed on very thin continental crust, perhaps similar to what Gamkrelidze & Giorgobiani (1990) referred to as 'intraplate subduction' in an intraplate setting. As such, the GC can be viewed as a Pyrenees or Atlas Mountains analogue (see, e.g. the overview of the Pyrenees by Grup de Geodinàmica i Análisi de Conques 2005). No lateral escape during shortening and consequent development of large nappes (and rootless nappes), such as in the Alps, occurred. The absence of any remnants of an ophiolitic suture supports such a model. Furthermore, there is no obvious record of any subduction zone along the GC

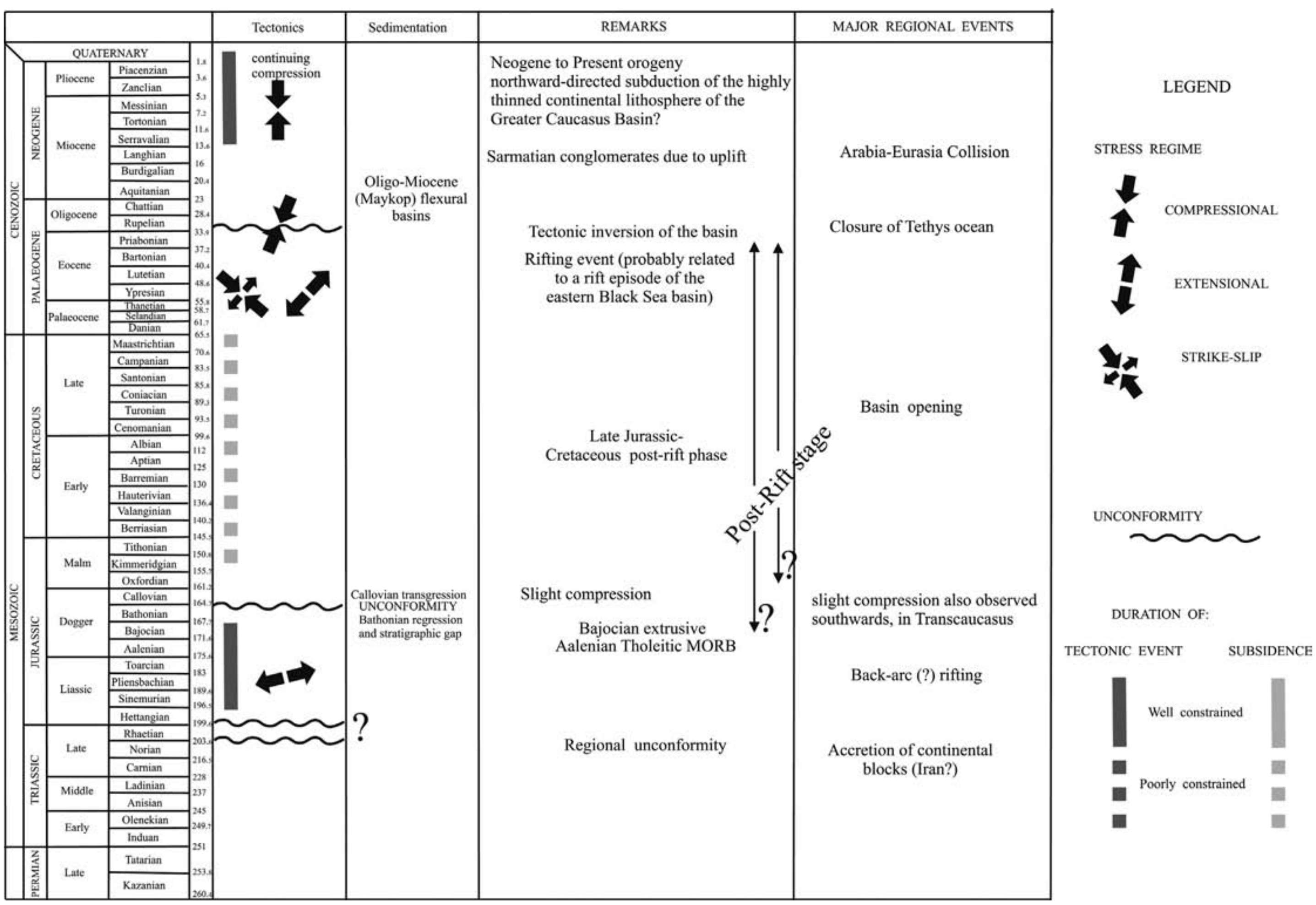

Fig. 12. Summary of the tectonic evolution of the Greater Caucasus. Absolute ages are from Gradstein *et al.* (2004).

during Mesozoic and Cenozoic times. There is no volcanic arc or blueschist and high-grade metamorphic rocks, and no accretionary complexes are present. (It is noted, however, that older, Palaeozoic, ophiolites and associated high-grade metamorphic rocks do crop out in the central part of the belt.)

Conclusions

The crustal structure of the Greater Caucasus remains a matter of debate, and two different models have been postulated. One model considers the GC belt as a former deep marine Mesozoic basin that was subsequently squeezed between steep crustal faults, these faults separating the GC from its adjacent tectonic units, the Transcaucasus and the SP. The alternative model considers the GC as a south-vergent, crustal-scale, imbricated fold-and-thrust belt with the SP thrust over the Transcaucasus massif along north-dipping planes, which flatten at depth. More and better geophysical data are needed to discriminate between these two models. However, the latter appears in general to satisfy better the available data, although some interpretations remain questionable (such as the geometry of the boundary fault zone between the GC and the SP and the amount of shortening in the GC belt). The tectonic evolution of the Greater Caucasus during Mesozoic and Cenozoic times can be summarized as follows (see Fig. 12): (1) Permo(?)-Triassic rifting; (2) Eo-Cimmerian shortening related to collision of the Iranian Block with Europe; (3) development of Early–Mid-Jurassic rift basins, possibly related to north-dipping subduction south of the Transcaucasus (i.e. in the Lesser Caucasus); (4) development of a Bathonian (Mid-Cimmerian) unconformity related either to the syn- to post-rift transition or to a collisional event at the active margin; (5) Mid–Late Jurassic to Eocene post-rift subsidence; (6) Late Eocene basin inversion related to the final closure of the Tethys oceanic domain; (7) a second shortening phase from Late Miocene time to the present accompanied by uplift and magmatism and corresponding to the final stages of Arabia–Eurasia collision.

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