Development history of the foreland plate trapped between two converging orogens; Kura Valley, Georgia, case study

M. NEMČOK1,2*, B. GLONTI3, A. YUKLER4 & B. MARTON5

1Energy and Geoscience Institute at University of Utah, 423 Wakara Way, Suite 300, Salt Lake City, UT 84108, USA
2Energy and Geoscience Laboratory at Geological Institute of Slovak Academy of Sciences, Dúbravská cesta 9, SK-840 05 Bratislava, Slovakia
3Frontera Eastern Georgia, 15 Chavchavadze St, Tbilisi, 0179 Georgia
4Renaissance JMW Energy, Cevdet Pasa Caddesi No. 30/2, Bebek, Istanbul 34342, Turkey
5San Leon Energy, Mokotowska Street No. 1, 00–640 Warszawa, Poland

*Corresponding author (e-mail: mnemcok@egi.utah.edu)

Abstract: Structural, sedimentological, well and seismic data from the Kura basin show that the geometry of deformation has been largely determined by thick-skin structures occurring along margins of depressions filled with upper Oligocene–lower Miocene Maykop formation, flanked by highs without Maykop record. Thin-skin structures are detached inside the shaly Maykop formation and inside shale horizons of the Sarmatian–Pontian section. The main shortening took part during Sarmatian–Pontian, followed by subordinate shortening during Akchagylian–present. The thick-skin architecture formed first, reactivating pre-existing rift grain on the foreland plate that refused to flex underneath the load of advancing orogens. The thin-skin architecture developed subsequently, deforming thick-skin structures. During the Oligocene–early Miocene, the foreland basin behaved as a flexural basin, reacting by prominent fill asymmetry to Oligocene loading by the advancing Lesser Caucasus and earliest Miocene–early Sarmatian loading by advancing Greater Caucasus. Subsequently, the basin recorded only vertical movements responding to new orogen loading events. Each loading event was recorded by a shift of marine depositional environments northwestwards, up the SE-plunging Kura Valley. Conversely each quiescence period was recorded by their retreat southeastwards, towards the Caspian Sea.

Supplementary material: Discussion of the observations from geological and seismic cross sections, additional outcrop photos and data, lithostratigraphic charts of the Greater Caucasus, the Adjara–Trialet and the Eastern Pontides, heat flow map and of the Kura basin, depth distribution of earthquake hypocentres for the Kura basin region, and Apulia map and cross section are available at http://www.geolsoc.org.uk/SUP18626

Apart from thick-skin orogen cases such as the Pyrenees and High Atlas (e.g. Verges et al. 1995; Teixell et al. 2003; Ferrer et al. 2008), the most extreme examples of vast areas affected by thick-skin tectonic events come from narrow foreland plates trapped between two orogens of opposite vergences. Two existing examples come from the Apulian plate between the Dinarides, Apennines (Bertotti, pers. comm. 2000; Nieuwland et al. 2001) and Kura Valley basin plate between the Greater and Lesser Caucasuses (e.g. Buleischvili & Sepashvili 1960; Banks et al. 1997; Robinson et al. 1997; Adamia et al. 2010; Fig. 1), which is the study area for this paper. Eurasian stratigraphic classification used in the study is explained in Table 1.

The study area of the Upper Kura basin is located within the Kura basin, a NW–SE oriented foreland basin segment, which spans from eastern Georgia to central Azerbaijan (Fig. 1a). It is a southeastern portion of a foreland basin system, which runs along a NW–SE trend from the Black Sea to the Caspian Sea. The system developed on the foreland plate to both the Greater Caucasus orogenic system to the NE and to the Lesser Caucasus orogenic system to the SW, which were advancing towards each other (e.g. Banks et al. 1997; Adamia et al. 2010; Fig. 2). The northwestern margin of the study area is bordered by the Kartli foreland basin. Further west lies the Dziruli Massif, which separates the Kartli and Rioni foreland basins, all resting on the same foreland plate as the Kura basin (Fig. 1a).

Our first goal is to work with sediment thickness distribution data on the large-scale behaviour of the foreland basin to determine its response to the
Lesser Caucasus orogenic system, overriding it from the SE and the Greater Caucasus orogenic system overriding it from the NE. The second goal is to study the reflection seismic profiles tied to well data to define the structural architecture of the foreland basin and combine this study with an analysis of the structural and sedimentological outcrop data to define the role of thick- and thin-skin tectonics in the architecture and determine their timing. Our other goal includes an attempt to understand the key factors controlling the onset of thick-skin deformation in the foreland basin.

**Structural overview based on existing literature**

Pre-orogenic evolution of both orogenic systems and their foreland plate includes the back-arc extensional regime deforming the Eurasian craton, which was overriding the northward-subducting oceanic crust of Palaeotethys, and subsequently Neotethys (Dercourt et al. 1986; Golonka 2004; Saintot et al. 2006). It was during the Early Jurassic when the development of the Greater Caucasus basin started, which is indicated by Hettangian–Sinemurian basal unconformity underneath the syn-rift fill (see Nikishin et al. 2001). The rifting phase was followed by Cretaceous thermal subsidence. Further south a less extended crust, including the Shatsky Ridge of the Eastern Black Sea and its eastern continuation named the Georgian plate, occurred (Banks et al. 1997; Fig. 2). Even further south lay the depositional area of the Adjara–Trialet Zone, representing the Georgian portion of the Lesser Caucasus orogenic system. Because its frontal décollement was propagated

**Fig. 1.** (a) Topographic map of the Black Sea–Caspian region with the location of the study area inside the Kura basin (topographic map is taken from ESRI 2006). Different basins of the foreland basin system are indicated by light grey and their depocentres in a darker grey. (b) Geological cross section through the Lesser Caucasus and Greater Caucasus orogenic systems and their foreland plate between them in the Kartli basin (Banks et al. 1997). J, Jurassic strata; J?, probable Jurassic strata; J2, middle Jurassic strata; K, Cretaceous strata; K?, probable Cretaceous strata; Kvolc, volcanic arc rock suite; Pal-l. Eoc, Paleocene–lower Eocene strata; m. Eoc, middle Eocene; u. Eoc, upper Eocene strata; P2u, Eocene strata, oli-l. Mio, Oligocene–lower Miocene strata; N1?, probable lower Neogene strata; M. Mio, middle Miocene strata; Sar, Sarmatian strata; Meo-Pon, Meotian–Pontian strata. (c) Geological cross section through the Lesser Caucasus and Greater Caucasus orogenic systems and their foreland plate between them in the Kura basin (Schelling 1998).
Fig. 1. Continued.
inside the Upper Cretaceous and younger incompetent strata, observations of its structural architecture controlled by the pre-Late Cretaceous extension are lacking (Banks et al. 1997).

The Adjara–Trialet Zone contains numerous volcanic strata in its Aptian–Priabonian lithostratigraphic column (see Nikishin et al. 2001). Its thrust structures disappear into the post-rift fill of the Eastern Black Sea and its Palaeogene strata are similar to those of the Eastern Black Sea (see Robinson et al. 1995). All this evidence indicates that the Adjara–Trialet Zone must have been located somewhere in the transition between the back-arc region and volcanic arc, which is well documented for the Late Cretaceous–Eocene time period in the Eastern Pontides (e.g. Dercourt et al. 1986; Golonka 2000; Derman 2010; Okay 2010; Tuysuz 2010).

The Greater Caucasus basin was being inverted during the late Eocene–Oligocene, owing to the compressional stress developed on the retro-wedge side of the doubly vergent orogen associated with the subduction of the Tethys (Dercourt et al. 1986; Golonka 2000) and later, during the Neogene, owing to the convergence between the encroaching Arabian platform to the south and Eurasia in the north. Both cited plate reconstructions indicate the first major compressional event resulting in intensive mountain building, reaching the development of emergent land in the late Eocene. The outcrop evidence for this event includes the upper Eocene olistostrome sediments in front of the Greater Caucasus front in the Kartli basin (Banks et al. 1997) and folded upper Eocene flysch strata unconformably overlain by Oligocene sediments in the NW Caucasus (Saintot et al. 2006). At this time, the Greater Caucasus formed an intra-plate south-vergent thick-skin orogen divided by the Black Sea basin system from the mentioned subduction-related orogen whose south-vergent prowedge was formed by the Menderes, Kirsehir blocks and Eastern Taurides, and north-vergent retr wedge formed by the Western and Eastern Pontides.

Following the main, late Eocene, orogenic event, the predominantly thick-skin development of the Greater Caucasus continued until the earliest Miocene, when it started to make a significant SE-vergent advance that resulted in the northeastward shift of the depocentres in the Georgian foreland basins, including the Rioni, Kartli and Kura basins (see Buleischvili & Sepaschvili 1960). During the earliest Miocene–Sarmatian, the staircase trajectory of the basal décollement climbed from the ductile deformation levels in the crystalline basement, through Jurassic shale and evaporitic horizons, through a shaly Maykop formation, through various Neogene intra-formational shale horizons to the surface. The location of the décollement in the Maykop formation can be seen in reflection seismic sections through the Tuapse Trough that forms the northeastern flank of the Eastern Black Sea basin (e.g. Finetti et al. 1988). Here the frontal thrust belt of the Greater Caucasus has overridden the lowermost Maykop formation and the underlying Eocene sediments.

The utilization of the Maykop formation as the main detachment horizon for movement came only after the main inversion process in the Greater Caucasus reached a highly advanced stage. From this point, the Greater Caucasus could accommodate the ongoing shortening by the development
of a thin-skin fold-and-thrust belt along its southwestern flank. Although the work of Schelling (1998) reveals the development of a similar thin-skin thrustbelt, Dagestan fold belt, advancing into the Terek–Caspian foreland basin, on the northeastern side of the doubly vergent Greater Caucasus (Fig. 1c), we will concentrate only on the southwestern thrustbelt in this text.

During the Sarmatian, the southwestern thrustbelt advancing towards its foreland basins, including the Russian Tuapse Trough and Georgian Rioni, Kartli and Kura foreland basins, accelerated its development. At this time, its development started on top of the décollement propagating inside the shaly upper Oligocene–lower Miocene Maykop formation and progressively ramping up to...
shallower flats located inside shale horizons of the Sarmatian–Pontian section. The main shortening took part during the Sarmatian–Pontian, followed by subordinate shortening during Akchagylian–present, as documented by Banks et al. (1997), who found:

1. Sarmatian sediments of the frontal thrust displaced over the upper Miocene–recent sediments in front of the Rioni foreland basin;
2. Sarmatian–Meotian sediments in the hanging wall of the Tsaishi thrust next to the Rioni foreland basin; and
3. Pliocene syn-tectonic strata at the Greater Caucasus front in the Kartli basin.

Although the development of the doubly vergent orogen associated with the subduction of Tethys started much earlier, that is, in the Mesozoic, the main mountain building in its retro-wedge consisting of the Eastern Pontides and Adjara–Trialet Zone had middle–late Eocene timing. This main event followed after the late Maastrichtian initiation of the compressional regime that eventually finalized the modern architecture of the Pontides (Tuysuz 2010). This evolution was accompanied by a development of the volcanic arch in the depositional region of the Eastern Pontides, which became established during the Late Cretaceous period (see Dercourt et al. 1986; Golonka 2000; Derman 2010; Okay 2010; Tuysuz 2010).

One more important event recorded by the Pontides is the Paleocene–early Eocene event. Its timing is well constrained in the Kirkkale basin located between the Pontides and Kırşehir Massif (Nairn et al. 2010). It is a collisional event, which was characterized by northward thrust imbrication of the Pontides (Okay 2010). This imbrication took part within the framework of the Paleocene–Eocene shortening of a doubly vergent orogen with piggy-back basins dating the age of its deformation on both sides (Kaymakci et al. 2010). One of them is an important syn-Eocene piggy-back basin on the Pontides side of the orogen (Kaymakci et al. 2010). The Paleocene–early Eocene event was followed by the middle–late Eocene widespread deposition and volcanism, resulting in

Fig. 3. Surface geological map of the study area with location of main faults and oil seeps (modified from Siradze & Venon; 1999; Gujabidze; 2003). Stratigraphic codes: 1, Lower; 2, Middle; 3, Upper in cases of Lower/Middle/Upper division of the respective period; 1, Lower and 2, Upper in the case Lower/Upper division of the respective period; I, Jurassic; K, Cretaceous; P, Palaeogene; N, Neogene; Q, Quaternary; 11, lower portion of Lower; 12, middle portion of Lower; 13, upper portion of Lower; 21, lower portion of Middle; 22, middle portion of Middle; 23, upper portion of Middle; ap, Aptian; al, Albian; cm, Cenomanian; t, Turonian; st, Santonian; cp, Campanian; m, Maastrichtian; d, Danian; ns; vs, Lutetian; l, Lattorfian; t, Tongrian; c; kn; kz; t, Miocene; S, Sarmatian; m, Maeotian; pn, Pontian; ak, Akchagylian; ap, Apcheronian; cp, Pliocene.
turbidites, volcanics and limestones. For example, the Eastern Pontides close to Georgia contain volcanic rocks several kilometres thick. Based on our data from Georgia, from the flexural response of the Georgian foreland basins Rioni and Kartli, we see that the northward advance of the Adjara–Trialet orogenic system adjacent to the Pontides from the east did not stop after this event but continued in a significant manner until the end of the Oligocene, as indicated by the depocentre location adjacent to this orogen (see Buleischvili & Sepaschvili 1960). Furthermore, geological cross sections through the easternmost Western Pontides published by Gorur & Tuysuz (1997) document imbrication of a sedimentary section starting with the Lower Cretaceous Caglayan and Ulus flysch formations and ending with the Eocene Kusuri flysch formation. The respective thrust sheets are sealed by the Neogene elastic sediments lying on the thrustbelt over erosional unconformity (see also Akyol et al. 1974).

The evidence from the adjacent Adjara–Trialet orogenic system documents even younger northward thrusting. It includes (Banks et al. 1997):

1. folded Pontian sediments and Pontian syntectonic strata at the thrust front inside the offshore Rioni foreland basin;
2. Sarmatian and then Pliocene timing of thrusting and folding documented by cross-cutting relationships from the frontal portion of the thrustbelt in the proximal Kartli foreland basin;
3. upper Miocene–Quaternary strata thickening towards the Adjara–Trialet thrustbelt in the onshore Rioni foreland basin; and
4. Sarmatian sediments in the hanging wall thrust over the upper Miocene–recent sediments in the footwall of the frontal thrust in the proximal onshore Rioni basin.

Methods

Parallel to the list of goals declared in the Introduction, the methods used in this study were chosen to allow us to:

1. understand the flexural behaviour of the foreland basin from the asymmetry of its fill changing in space and time;
Fig. 5. (a) \(\sigma_1\) stress trajectories interpreted from earthquake focal mechanisms (modified from McClusky et al. 2000). Pink trajectory, Eurasian Plate; yellow trajectory, collision zone; blue trajectory, Arabian Plate; green outline, front of the indenting Arabian Plate and southern margin of the Eurasian Plate, which are both in darker grey; orange, dextral transpressional region of the collision zone, which is in light grey; dotted orange, axis of this region; thick purple, regional \(\sigma_1\) stress direction interpreted from local fanning stress trajectories; blue rectangle, area where one understands the regional \(\sigma_1\) stress direction and its deflection caused by the friction along the ‘suture zone’; v-shaped region of the
(2) define the structural architecture of the deformed foreland basin from the seismic imagery tied to wells;
(3) evaluate the role of thick- v. thin-skin tectonics in the structural architecture and their timing from subsurface imagery combined with surface structural and sedimentological data; and
(4) understand the role of key factors controlling the onset of the thick-skin deformation from combined methods listed in points 1–3.

In order to define the thickness distribution in a rigorous manner, we had to look at hundreds of sequence stratigraphic sections described in Buleischvili & Sepaschvili (1960) and synthesize these observations. It was a relatively complex task as they were described in different chapters, split according to stratigraphy and located with respect to anticlines and synclines. This task represents synthesizing sequence stratigraphic profiles from the Sarmatian, Maeotian–Pontian and Akchagylian–Apcheronian sedimentary sections for 95 anticlines and 20 synclines (Fig. 2).

A seismic interpretation tied to exploration and production wells was also employed during this study. The reflection seismic database was represented by 2D seismic images with a total length of 720 km and a 3D seismic volume in the Taribani area with areal extent of 230 km². Their acquisition was carried out by Frontera Resources during the last 10 years. The seismic images were interpreted using the Kingdom Suite software.

The remaining portion of the work flow included the interpretation of nine geological cross sections developed along the reflection seismic sections. Their interpretation started with the development of topographic sections and projection of bedding planes, fold axes, faults and stratigraphic tops into the section. Subsequently, the dip-domain technique was applied, defining the extent of domains characterized by the sub-parallel dip of seismic reflectors. Following that, profiles have been interpreted without using balancing techniques.

The fieldwork in the study area (Fig. 3) targeted 565 outcrops (Fig. 4). At each of them, if the respective features were available, measurements of styloliths, tensile fractures, shear fractures and sediment transport were made. Sediment transport measurements were made based on pebble imbrications, cross bedding and flute casts. The outcrop description was then completed with lithostratigraphic data, including dip direction and dip of bedding, stratigraphic position of strata and thicknesses of different lithologies.

Regarding styloliths, we were drawing from the knowledge, that, apart from styloliths resulting from compaction and pressure solution during diagenesis, which can be enlarged by subsequent groundwater flow, styloliths can form by contraction during shortening (e.g. Twiss & Moores 2007).

Styloliths appear as jagged discontinuities in sections, having their bottoms developed as relatively flat surfaces. Shortening direction can be measured as coincident with stylolithic lineation defined by axes of stylolithic columns and taken as proxy for $\sigma_1$ direction. Developed on pebbles, styloliths occupy small areas on the opposing sides of the pebble. They are formed by slow deformation facilitated by pressure solution (Jaroszewski 1972; Schrader 1988). Unlike classic styloliths, styloliths developed at pebbles usually form a cluster of solution pits. Sometimes, portions of the pebble surface between the two ends with solution pits can contain slickololiths or even slickenfibres, representing a deformation linkage between the pure styloliths at opposite ends of a pebble (Jaroszewski 1972; Hancock 1985; Schrader 1988). While the majority of pebbles were made from material resistant to stylolith development, stylolith measurements have been made on carbonate pebbles, which are relatively rare in Sarmatian and older Maeotian–Pontian gravel/conglomerate layers, progressively more abundant in younger Maeotian–Pontian layers and again very rare in Akchagylian–Apcheronian layers. To allow for reliable calculation of the mean $\sigma_1$ stress direction, a set of 50 measurements was attempted at each outcrop with carbonate pebbles. Most of the outcrops, however, owing to the lack of larger carbonate clast quantities, contain smaller numbers of data. Following the fieldwork, the data were visualized in the lower hemisphere Schmidt’s projection as pole diagrams.

Extension veins have been studied to indicate the $\sigma_3$ stress. They have been developed as tensile fractures, as a result of mechanical rupture, generally perpendicular to maximum longitudinal strain.
acting during their initiation (Ramsay & Huber 1983). During the following deformation increments, the fractures continued their growth either in a direction perpendicular to their walls or obliquely, depending on the orientation of the deformation ellipsoid of the respective deformation increment. Being the sites of reduced compression, tensile fractures have been filled with crystallized material. In the study area, all of them are filled with gypsum, precipitated from fluids migrating during the orogenic activity in the region. Observed gypsum cements were either massive or composed of fibrous crystal grains. While in the case of massive fill, the \( \sigma_3 \) was determined as pole to fracture wall, in the case of fibrous fill, the \( \sigma_3 \) was determined from fibre geometries.

Some 61% of the veins were sub-parallel to bedding, and 39% cut through the bedding. Some veins indicate multiple episodes of opening, having multiple median zones divided by cement zones with fibrous crystals or showing the cross-cutting relationship of fibrous zones within the same fill. The gypsum vein population frequently contains several generations of veins with different orientations. They have been separated based on the cross-cutting relationships of entire veins. Following the fieldwork, the data were visualized in the lower hemisphere of Schmidt’s projection as pole diagrams.

Shear fractures were measured in order to calculate palaeostresses from fault-striae data by methods published by Nemčok & Lisle (1995) and Nemčok et al. (1999). They differ from tensile fractures by having their surfaces striated. Some striated surfaces are frictional surfaces formed in the host rock, while others contain gypsum precipitated in pressure shadows along the displacement vector. The kinematics of shear fractures was determined as based on:

1. slickenside flutes, which look like gutters becoming deeper in the direction away from their initiation;
2. slickenside prod marks, represented by short depressions usually containing a preserved strong grain of rock that developed the feature ploughing the host rock being squeezed between the two moving blocks; and
3. idiomorphic cement crystals developed in the pressure shadow behind the shielding elevation of the host rock.

Fracture statistics shows that the shear fractures are rare. For example, out of 527 measured fractures in the Taribani area, only 3% are shear fractures, while 97% are tensile fractures. However, the studied shear fractures and faults allowed the geometry and timing of the main thrusting events to be determined. Most of the outcropping thrust detachment were characterized by the gypsum fill and the very shallow dip of the fault plane. Some outcrops allowed us to see that thrust sheets have been compartmentalized by strike-slip faults.

Palaeocurrent measurements were taken on flute casts, pebble imbrications and cross beds. Out of the various structures produced by the action of the currents, the ones produced in the marine environment for the Sarmatian stratigraphies, and fluvial and deltaic for the Maeotian–Pontian section, were studied most. Flute cast measurements were taken on five Maeotian–Pontian and 17 Akchagylian–Apcheronian outcrops in fluvial and deltaic environments. Measurements of the pebble imbrication were taken from 30 Akchagylian–Apcheronian and 105 Maeotian–Pontian gravel/conglomerate layers deposited in fluvial environments. To allow for realistic calculation of the mean direction, measurements at outcrops attempted to collect the data from at least 40–50 pebbles. Cross bedding measurements were taken at 10 Sarmatian and 360 Maeotian–Pontian outcrops, representing beach, river and sand-dune environments. To allow for reasonable calculation of the mean direction, measurements at outcrops attempted to collect about 50 readings. The data sets with less than 50 readings indicate less populated outcrops.

Our own data sets, together with pre-existing data sets, were compiled and interpreted using Arc GIS. Geological, geophysical and tectonic data existing prior to this study were added to the project as well, being originally gathered in different formats and scales. Registered and re-projected data sets, were compiled and interpreted using Arc GIS.

The version 9.2 Arc GIS software that was used does not have the ability to represent bedding, sedimentological and structural planar and vector data using standard geological symbols. Therefore, in order to represent readings of bedding planes, sediment transport vectors, styloliths-derived \( \sigma_3 \) directions and tensile fracture-derived \( \sigma_3 \) vectors, a symbol alphabet was developed using numerical data from the attribute table.

**Results**

*Regional sediment thickness distribution in the foreland basin*

Before we look at sediment thickness distribution data, it would be interesting to divide anticlines and synclines into those belonging to the Greater Caucasus orogenic system advancing from NE and those belonging to the Adjara–Trialet orogenic system advancing from the SE onto foreland represented by the Upper Kura basin in our study area.
Fig. 6. Continued.
Fig. 6. Continued.
We made a rough division, using fold and fault geometries as a guide. The division into regions, where both orogenic systems are either in contact or overlap, is not perfect, but is good enough to discuss the controlling mechanisms and constraining boundary conditions. Two relatively different groups of fold geometries allowed rough determination of orogen advance vectors (Fig. 2). Geometries of both fold groups and estimated advance vectors of both orogens correlate reasonably well to present-day horizontal movement component trajectories determined from GPS data (see Reilinger et al. 1997, 2006; Figure 5b) and the $\sigma_1$ stress pattern interpreted from earthquake focal mechanisms (see McClusky et al. 2000; Figure 5a).

Figure 2, being the present-day view of orogen advance vectors and fold geometries, then helps us understand that one can take thicknesses of sediments accreted by both orogens and develop rough sediment thickness distribution maps. They can be made for time intervals starting from the Oligocene time, when the foreland basin fill was in front of both orogenic systems, and finishing in the late Pliocene, when the foreland basin fill was accreted by the Greater and Lesser Caucasus orogenic systems.

The Oligocene fill of the foreland basin system (Fig. 6a) records prominent basin asymmetry. The depocentre associated with thickness maxima reaching about 2000 m was located along its SW border. The wedge-shaped fill thins out significantly in the NE direction. Thinning over the distance of the width of the Kartli basin is from 2000 to 200 m. Similar thinning can be found in three more areas within the Kura basin. The asymmetry...
Fig. 7. (a) Map of the pebble imbrications-based sediment transport vectors in Maeotian–Pontian sediments. (b) Map of cross bedding-based sediment transport vectors in Maeotian–Pontian sediments. Names and locations of individual anticlines and synclines in (a)–(b) are the same as those in Figures 3, 4, 9a, b, 10a, b.
helps indicate the distribution of the orogenic loading with respect to the flexing plate underlying the foreland basin system. The Oligocene must have been the time when the plate underlying the foreland basin system flexed in response to loading by the advancing Adjara–Trialet, that is, the Lesser Caucasus, orogenic system. It must have already been much closer to the study area than the Greater Caucasus system advancing from the north.

The lowermost Miocene–lower Sarmatian sediment asymmetry provides us with the opposite scenario (Fig. 6b). The depocentre was located along the NE margin of the foreland basin system and the thinning of its sedimentary fill took place in the SW direction. The thinning distance and thinning results are comparable to those listed for the Oligocene. The aforementioned scenario indicates that this time it was the Greater Caucasus orogenic system that had the dominant loading influence on the flexure of the plate underlying the foreland basin system.

The transition from the early to late Sarmatian brings a surprising change in thickness and facies distribution pattern in the foreland basin system (Fig. 6c). Instead of the classical foreland basin asymmetry, one sees a gutter-like foreland basin shape on both sides of the Dziruli massif (see Fig. 1 for location). The highest elevation of the foreland basin floor was in the Dziruli massif area. To the west of it, the Rioni and Kartli foreland basins plunged down to the NW towards the Black Sea. To the east of it, the Kura foreland basin plunged down SE towards the Caspian Sea. This overall geometry of foreland basins remained the same from the late Sarmatian to the present, although it underwent important modifications as discussed below. This kind of geometry of the Kura foreland basin, even in its later development stage involving accretion of its sediments into the Greater Caucasus orogenic system, explains the measured sediment transport vectors (Fig. 7a, b) and present-day river flow directions in the study area reasonably well. Such gutter-like geometry controls a system of short southward transversal transports from the advancing Greater Caucasus, eventually from local growing folds, and a longitudinal transport system funnelling sediments down the gutter southeastward to the Caspian Sea.

The upper Sarmatian facies and thickness distribution in the Kura foreland basin (Fig. 6c) reflect such geometry and transport patterns, having non-deposition areas in the very NW, continental facies a little southeastward and then marine environments in the SE part of the basin. The fact that marine conditions prevailed in the SE portion of the gutter-shaped foreland basin should indicate that the overall elevation of the basin floor was low during this period of time, most likely depressed by the loading from at least one of the advancing orogenic systems.

The early Pliocene scenario (Fig. 6d) represents purely continental conditions in the Kura foreland basin. The fact that the marine environments of the late Sarmatian time have retreated down the foreland basin towards the Caspian Sea can be interpreted as tectonic quiescence in the loading history of the Kura basin, which was already incorporated into the Greater Caucasus orogenic system in the study area. This must have been a time of intensive erosion of adjacent orogenic systems and resulting isostatic rebound of them and their shared remnant foreland basin. Recorded thicknesses at different locations down the gutter-shaped Kura foreland basin show a southeastward thickening trend (Fig. 6d), further supporting the southeastward sediment transport down to the Caspian Sea.

The late Pliocene scenario (Fig. 6e) shows a return of marine conditions into a large portion of the Kura foreland basin, which must have been caused by yet another loading event from at least one of the orogenic systems advancing against each other. Such loading created subsidence, resulting in a rapid surge of marine conditions from the Caspian region up the gutter-shaped Kura foreland basin already incorporated into the Greater Caucasus orogenic system in the study area.

Structural architecture in the geological map

Geological mapping documents that the central part of study area is cut by a NW–SE striking fault system. Fault traces of this system are frequently the sites of oil seeps (Fig. 3). The fault system divides the study area into two regions, the southern region being characterized by the occurrence of numerous NW–SE striking anticlines and synclines (the Fold and Thrust Zone of this paper). The narrow anticlines are the sites of deep erosion, providing outcrops of sediments sometimes as old as the early Sarmatian. On the contrary, synclines, broader than anticlines, contain sediments as young as the Maeotian–Quaternary. The Akhagyanian–Apcheronian–Quaternary sediments rest on folded older sediments over erosional unconformity. This is documented in several regions of the geological map (Siradze & Venon 1999; Fig. 3) and is recognizable at outcrops.

The region to the north of the aforementioned fault system lacks narrowly spaced fold structures (Fig. 3). This region is characterized by the presence of a broad structural high in the Dedoflitskaro area, called the Basin Edge West in this study, a broad Didi Shiraki syncline, and a broad structural high in the NE corner of the study.
area called the Basin Edge East. The Basin Edge West high contains a system of klippen on its top, formed by Jurassic carbonates, resting partly on the Maeotian–Pontian, partly on the Akchagylian–Apcheronian sediments. Most of the Didi Shiraki anticline, known from wells as including extremely thick sediment section, is masked by Quaternary sediments on the surface. The Quaternary and Akchagylian–Apcheronian sedimentary sections rest on the erosional unconformity, which postdates folding. The Basin Edge East contains the Maeotian–Pontian stratigraphy at the surface.

Two of the most important features of present-day topography are the Alazani and Iori rivers (Fig. 3). Both of them flow towards the SE. The Alazani river forms a portion of the northern boundary of the study area, and, as such, is not a focus of this study. The Iori river flows from the NW to the SE through the entire Fold and Thrust Zone. It is divided from the Alazani river drainage area by the NW–SE trending zone composed of the Basin Edge West, the Didi Shiraki syncline and the Basin Edge East. The geometry of the Iori river is composed of the NW–SE and NNW–SSE segments. NW–SE segments are parallel to anticlinal axes, preventing the river from flowing directly to the SSE. These river segments are located in synclines behind anticlines characterized by significant

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Fig. 8. (a) Profile through the 3D seismic volume, cutting through the Taribani structure. See Figure 4 for profile location. (b) Profile through the same volume imaging the Taribani structure. See Figure 4 for profile location.
crestal erosion. On the contrary, the NNE–SSW segments are located either between neighbouring anticlines or in places of less significant elevation of down-plunging anticlines.

**Structural architecture in seismic and geological profiles**

Profile C of the 3D seismic volume is cut through the buried Taribani anticline. The profile also images shallow anticlines and synclines above it (Fig. 8a). Shallow features occur at depths above 2.4 s TWT. They have a kink-band character, while the deeper anticline is relatively open. Reflectors imaging its sediments indicate multiple steep faults, documented by zones of reflector truncations. Profile D through the 3D volume shows that, although the buried anticline is deformed by multiple steep faults, it can be seen as it climbs on an old structural high with a Mesozoic carbonate cover (Fig. 8b). The profile images the frontal portion of the structure very well. It rests on the flat high and pinches out in a direction away from the anticlinal crest.

**Palaeostress orientation**

The $\sigma_3$ vectors calculated from veins hosted by the Akchagylian–Apcheronian and Maeotian–Pontian...
stratigraphies contain several generations (Fig. 9a). The Maeotian–Pontian vector pattern in the broader Kushiskhevi–Mkralikhevi–Mirzaani area contains vectors, which are sub-parallel to strikes of regional faults and fold axes. They have NW–SE to west–east trends. Others are roughly perpendicular to them, having NNE–SSW to NE–SW trends. Most of the outcrops contain representatives of both groups. The presence of extensional veins in some locations and their lack in others in this area allow us to see their occurrence along regional faults. The Akchagylian–Apcheronian outcrops do not occur in this area.

The Maeotian–Pontian $\sigma_3$ vector pattern in the Kila–Kupra area is characterized by several generations. Two vectors on the NE side are sub-parallel to strikes of regional faults and fold axes. They have NW–SE to west–east trends. Others are roughly perpendicular to them, having NNE–SSW to NE–SW trends. Most of the outcrops contain representatives of both groups. The presence of extensional veins in some locations and their lack in others in this area allow us to see their occurrence along regional faults. The Akchagylian–Apcheronian outcrops do not occur in this area.

The Maeotian–Pontian $\sigma_3$ vector pattern in the Iori–Baida–Chatma area, the $\sigma_3$ vector pattern is characterized by two generations. Some 90% of vectors from the Maeotian–Pontian sediments have a NNE–SSW trend. Vectors from the Akchagylian–Apcheronian sediments have a NW–SE trend.

The Chatma area also has several generations of vectors. Some of them have a NW–SE trend, and others are roughly perpendicular to them, having NNE–SSW to NE–SW trends. The Akchagylian–Apcheronian outcrops do not exist in this area as well.

The $\sigma_3$ vector pattern in the broader Taribani area is relatively complex. Vectors from the Maeotian–Pontian sediments are perpendicular to the Taribani structure, having a NE–SW trend. There are two vectors in the south, both determined from the Akchagylian–Apcheronian stratigraphy. Their trends are NW–SE. In the east we can see three more vectors from the Akchagylian–Apcheronian formation, having NW–SE to west–east trends.

$\sigma_1$ vectors (Fig. 9b) in the broader Kushiskhevi–Mkralikhevi–Mirzaani area are either sub-parallel to strikes of regional faults and fold axes, having NW–SE to west–east trends or roughly perpendicular to them, having NNE–SSW to NE–SW trends. Most of the outcrops contain representatives of

![Fig. 9. (a) $\sigma_3$ vectors calculated from extension veins at outcrops of Maeotian–Pontian and Akchagylian-Apcheronian formations. Different arrow grey shades indicate vectors determined from different vein generations. The mapped stratigraphies are identical to those shown in Figure 4. (b) $\sigma_1$ vectors calculated from solution pits at outcrops of Maeotian–Pontian and Akchagylian-Apcheronian formations. See caption (a) for further explanations. Names and locations of individual anticlines and synclines in (a)–(c) are the same as those in Figure 3, 4, 7a, b and 10a, b.](image-url)
both groups. The Akchagylian–Apcheronian outcrops do not occur here.

In the Taribani area, outcrops with styloliths represent the Akchagylian–Apcheronian stratigraphy, yielding the \( \sigma_1 \) vectors sub-parallel to strikes of regional faults and fold axes. Maeotian–Pontian outcrops do not occur here. In the Patara Shiraki–Kaladara area, \( \sigma_1 \) vectors are sub-parallel to strikes of regional faults and fold axes, having NW–SE to west–east trends.

**Local sediment thickness distribution in the fieldwork area**

The isopach map of the Sarmatian sediments deposited in marine environments, with the exception of the western Upper Kura basin during the late Sarmatian, which was occupied by continental environments (Buleischvili & Sepaschvili 1960), delineates two thickness maxima and two thickness minima (Fig. 10a).

The thickness maximum in the NW end of the study area, in the Tsitsmatiani–Mtsarekhevi area, is the largest, having values ranging from 1750 to 3000 m. To the SE, in the Iori area, lies the most significant thickness minimum, reaching values of 500–625 m. The second thickness maximum lies in the Taribani area. It has values of 1625–1750 m. The remainder of the study area, the Didi Shiraki–Patara Shiraki–Kaladara area, forms an areally extensive thickness minimum, where thickness ranges from 1000 to 1125 m.

The isopach map of Maeotian–Pontian sedimentary section deposited in various marine and continental environments contains thickness values ranging from 1000 to 3000 m (Fig. 10b). The thickness maxima are located in the NW, central and eastern regions, being surrounded by areas with reduced thickness. The NW maximum has a thickness in the 2000–3000 m range. It partially overlaps with the maximum known from the Sarmatian time, although it has now doubled in area, having expanded towards the SE and east. It occupies the Tsitsmatiani–Mtsarekhevi–Lambalo–Vashliani area. The central thickness maximum is in the Taribani area and reaches thickness values of 2000–3000 m. Its outline roughly overlaps with that of the Sarmatian maximum. The eastern thickness maximum is represented by 2000–2500 m thickness interval. It is located in the Didi Shiraki area, in the location of the NW portion of the Sarmatian thickness minimum.

The isopach map of the Akchagylian–Apcheronian lithostratigraphy deposited in both continental and marine depositional settings contains values ranging from 250 to 2500 m (Fig. 10c). Its comparison to the Sarmatian and Maeotian–Pontian maps documents that all three locations
with pre-Akchagylian thickness maxima did not continue evolving into the Akchagylian–Apcheronian time period with significance. The Taribani maximum practically disappeared during the Akchagylian–Apcheronian apart from very small remnant in its southern part. The maximum in the Didi Shiraki area has both areally shrunk and thinned from the Maeotian–Pontian to the Akchagylian–Apcheronian. The Tsitsmatiani–Mtsarekhevi–Lambalo–Vashliani maximum in the NW has mostly disappeared with the distinct exception of its south and NE portions. While the southern portion is insignificant, the NE portion forms the only more significant thickness maximum of the Akchagylian–Apcheronian time in the study area.

**Sediment palaeodispersal system**

The sediment transport vectors derived from pebble imbrication, cross bedding and flute casts were...
extensive for Maeotian–Pontian, intermediate large for the Akchagylian–Apcheronian and only a few were from the Sarmatian. The Maeotian–Pontian and Akchagylian–Apcheronian vector patterns have a relatively similar character. In order to save space, we will focus only on most robust Maeotian–Pontian data set (Fig. 7).

In the Iori–Baida-Chatma–Chatma area, pebble imbrication-based transport vectors (Fig. 7a) have directions to the south, WSW and SSE. The southward transports are prevalent. In the Taribani area, vectors point to the south, SSW, SE and NNE. In the Patara Shiraki–Kaladara area, the vector directions are chaotic, covering trends such as SSW, SSE, south, east and NE.

Most of the cross bedding-derived vectors in the Kila–Kupra area have south-southwestward directions (Fig. 7b). Other directions cover the NNE, south and SSE. The Taribani area is characterized by a single, south-trending, vector. The vector pattern in the broader Kushiskhevi–Mkralikhevi–Mirzaani area contains several directions. Most of them are perpendicular to strikes of regional faults and fold axes, having trends towards the NW to west. The subordinate group has directions towards the SSW, SW and SSE. The Patara Shiraki area can be characterized by the SSW and south direction. The southern margin of the study area is anomalous in the sense that its vector pattern is opposite to the generally southward transport determined in other areas, containing transports to the north, NE and NW. There is only a single transport towards the SW in this area.

Interpretation

The Sarmatian, Maeotian–Pontian and Akchagylian–Apcheronian thickness maps (Fig. 10) document deposition as reacting to changing topography controlled by the late Sarmatian onset of growth of anticlines and synclines, their Maeotian–Pontian development and Akchagylian–Apcheronian tectonic quiescence and limited growth. Thickness minima and maxima overlap with growing anticlines and synclines, respectively. Tsitsmatiani–Mtsarekhevi–Lambalo–Vashliani and Taribani thickness maxima allow us to see NW–SE elongation of the synclines capturing the sediments during cycles of marine and continental deposition.

The sediment transport maps (Fig. 7a, b) indicate that prevalent southward transversal sediment transport was from the Greater Caucasus encroaching from the north, represented by its front accreting sediments during Sarmatian to Pontian time. This transport was from orogen to its foreland formed by the Kura basin. This dispersal system indicates relatively subdued topography of the foreland basin that only slowly started to evolve into complex topography of the initiating thrustbelt. The longitudinal transport was represented by a NW to SE trend, down the Kura basin axis, plunging towards the South Caspian basin.

Local complexities in sediment transport vectors indicate that the growing anticlines formed local sediment provenances, sometimes causing local sediment transport in a direction opposite to the overall Greater Caucasus-foreland transport.
Transport vector patterns in Figure 7a, b further document that the present-day Iori river, with its NW–SE and north–south segments, must have been set during periods of continental deposition interacting with growing anticlines. Fold axis-parallel NW–SE segments represent sediment flow behind growing anticlines, which represent obstacles in transversal flow from the Greater Caucasus, and captured in pronounced syncline segments behind mature anticline segments. North–south Iori river segments represent more or less free transversal flow towards foreland, located either among anticlines or cutting through slow uplift segments of anticlines located near their ends.

Figure 10a–c allows us to see the Sarmatian–Pontian growth of the synclines in the Tsitsmatiani–Mtsarekhevi–Lambalo–Vashliani and Taribani areas. It also indicates a slightly different history of the Didi Shiraki syncline, which is, unlike other synclines, not elongated but circular. Furthermore, it started capturing sediments not during the Sarmatian but during the Maeotian–Pontian period, that is, starting later than other synclines in the area. While the first two were regular synclines of the

![Fig. 11. (a) Erosional unconformity formed after the thick-skin anticline development interpreted in profile E through the Taribani 3D seismic volume. (b) Erosional unconformities formed after the thick-skin anticline development and thin-skin anticline development interpreted in profile F through the Taribani 3D seismic volume.](image-url)
thrustbelt, the Didi Shiraki synclinal structure looks like one made by the interference of surrounding structures. It later became filled by sediments, which were very thick and prevented the syncline from undergoing distinct internal deformation. As a result, it remained a deep circular ‘hole’. Insufficient depocentres capturing sediments during the Akchagylian–Apcheronian time period, which post-date regional erosional unconformity observable in the outcrop, can be divided into two categories: (i) remnant depressions among anticlines left after the end of main contraction events, and (ii) synclines developing during subsequent, limited shortening.

The thrustbelt topography has been developed with the growth of thick-skin and thin-skin structures (Fig. 8a, b). Depending on orogen strike-parallel changes in pre-orogenic mechanical stratigraphy, one can see orogen strike-parallel changes from thick- to thin-skin structures and vice versa. Numerous areas with sequential development of thick-skin and then thin-skin structures indicate strike-parallel changes in mechanical stratigraphy with time during the orogen development.

With respect to lateral changes in structural style, the interpretation of cross sections, such as sections A and B, indicates that thin-skin structures developed as lateral deformation choice have been grown above detachment faults propagated inside thick upper Oligocene–lower Miocene Maykop formation, characterized as shale-dominant and rich in organic matter. This formation was
deposited in a system of depressions divided by highs. The highs are free of the Maykop formation owing to nondeposition or subsequent erosion. Such highs are usually characterized by Jurassic and Lower Cretaceous carbonates unconformably overlain by Sarmatian–Quaternary or Maeotian–Quaternary sediment over pronounced erosional unconformity or missing section.

On the contrary, thick-skin structures were developed along margins of local depressions with Maykop fill, flanked by structural highs, lacking the Maykop formation. With respect to their coexistence with thin-skin features, cross sections such as those through the Taribani area (Fig. 8) contain deeper thick-skin structures and shallower thin-skin structures. Zones of seismic reflector truncations inside thick-skin Taribani anticline allow one to interpret numerous strike-slip faults involved in their development. This fact is in accordance with numerous outcrop observations of roughly NW–SE striking dextral strike-slip and oblique-slip faults.

More frequent occurrence of dextral strike-slip faults can be found in outcrops towards the eastern end of the Upper Kura basin. Furthermore, deformation events in the study area are relatively young. These two observations indicate that present-day kinematics (Fig. 5b) and their controlling stress regimes (Fig. 5a) are an analogue for the Sarmatian–Pontian history. The present-day setting allows us to see that the collisional zone between Eurasia and Arabia has an overall dextral transpressional character. It is characterized by thrusting gaining more importance westward and dextral strike-slip faulting gaining more importance eastward. The appropriate use of the analogue is further supported by a match of present-day $\sigma_1$ trends (Fig. 5a) and Maeotian–Pontian ones determined in the study area, and frequent observations of oblique thrusts at outcrops.

The study of the buried Taribani anticline points to its transpressional origin. Its development was accompanied by numerous cycles of pore fluid pressure increase and release, as indicated by numerous polyphase extensional veins at outcrops in this area. We connect the cycles of pressure buildup and release with relatively fast deposition and coeval contraction of shale-rich low-permeability Sarmatian–Pontian lithostratigraphies.

The presence of numerous weak shale horizons makes them prone to development of local detachment faults, as one can see in the 3D seismic volume of the Taribani area (see Fig. 8a) where flats of the thin-skin system lie in Sarmatian–Pontian stratigraphies. They allow the development of local thin-skin anticlines and synclines, 'peeled' off the thick-skin structures at their shallow levels.

Extensional veins in the area (Fig. 9a) indicate several different development mechanisms. The veins parallel to the fold axes were caused by the release of flexural stresses in crestal regions. Those perpendicular to fold axes, usually younger than the former ones, were caused by lateral stretching of the folds undergoing extra shortening, which took place after their development to fully mature stage. This observation shows that most of the anticlines in the study area were developed by dextral transpression to their fully mature stage and then became further translated/deformed by continuing shortening.

With this understanding, one can interpret the deformational sequence in the Taribani area as the initial thick-skin anticline development and its subsequent deformation by thin-skin tectonics. If this interpretation is correct, it could also be applied to the seismic data.

The seismic imaging shows that the buried Taribani anticline is rather open (Fig. 8a) and thus unable to accommodate significant shortening. Some portions of the seismic volume (Fig. 8b) show that the anticline contains sediments that climbed from their original depositional position in the local depression to their present position on top of the adjacent structural high, climbing the inverted pre-existing normal fault. The seismic section in Figure. 8b images the frontal nose of the thick-skin structure resting on the structural high rather well.

When we interpreted a pre-existing normal fault in the entire 3D volume, it turned out that it is not a single fault but three normal faults connected with each other via two relay ramps. All three of these have a roughly NW–SE strike. As is shown by the more detailed interpretation of the easternmost fault, each of them consists of its own individual segments.

The reflection seismic imaging in the Taribani area is too shallow to allow us to see the detachment level of the thick-skin structure and the oldest incorporated sediments. The oldest sediments reasonably interpreted inside the structure have an Eocene–Oligocene age.

Profile E through the seismic volume shows that the development of the thick-skin anticline was followed by a period of peneplanation documented by erosional unconformity (Figure 11a). The unconformity is indicated by truncations of underlying seismic horizons parallel to bedding. The same figure shows that this erosional unconformity has been subsequently slightly folded and cut by a thrust fault, which is associated with a thin-skin anticline. Because this profile runs through the least mature portion of the younger thrust fault and associated structures, one needs to go along the fault strike to profile F (Fig. 11b), which allows us to see both thick- and thin-skin anticlines in a more mature development stage. Figure 11b shows that
the erosional unconformity postdating the thick-skin structure is cut and displaced by a younger thin-skin thrust. It underwent folding by a shear drag in its footwall. Being uplifted in the hanging wall and eroded off, it is missing in the hanging wall of the thin-skin thrust. The younger thrust is detached at
a relative shallow depth level along some intraformational shale horizon. The thin-skin event has been postdated by peneplanation, which resulted in erosional unconformity developed in both hanging wall and footwall of ‘the thin-skinned thrust fault. This erosional unconformity occurs at numerous outcrops in the area, overlain by the Akchagylian–Apcheronian sediments.

Discussion

The data discussed in the chapter on sediment thickness distribution in the foreland basin indicate that the foreland basin behaved in a classical way, that is down-flexing towards coevally advancing orogen, during the early stages of the Lesser Caucasus and Greater Caucasus orogens advancing against each other and overriding the plate hosting the Rioni–Kartli–Kura foreland basin system (Fig. 12a, b). This was the case for the Oligocene–early Sarmatian time (Fig. 6a, b). Outcrops in the study area do not indicate any distinct shortening in the study area, which was located in the Upper Kura foreland basin segment, during this time. This indicates that the basin must have been too far from both orogenic fronts to undergo sediment accretion into their frontal structures.

The foreland basin system stopped flexing classically when orogens came close to each other (Figs 6c–e & 12c). This took place during the late Sarmatian (Fig. 6c). Inability of the remnant foreland plate to flex under the load of the advancing respective orogen turned its late Sarmatian–Quaternary movements into cycles of subsidence and uplift without flexure. The remnant of the foreland plate had an uneven elevation, reaching the maximum in the Dziruli massif area and plunging down toward the Black Sea to the west and towards the Caspian Sea to the east. Analogous to the present-day thermal regime, we expect the late Sarma-
tian–Quaternary thermal regime to be relatively cold. Such a present-day regime is indicated by surface heat flow values ranging from 30 to 60 mW m$^{-2}$ in the study area and relatively deep bottom depth of the earthquake focal mechanisms progressively deepening from 40 to 70 km in the direction from the Dziruli Massif to the South Caspian region.

During the late Sarmatian–Quaternary, each episode of actively advancing orogen depressed the foreland plate under orogen weight. This resulted in a surge of marine conditions up the plunging remnant foreland basin towards the Great Caucasus orogen. Each subsequent tectonic quiescence period was characterized by unloading of the orogenic load owing to accelerated erosion and sediment transport down the plunging basin, resulting in the retreat of marine conditions away from the Dziruli massif.

Structural data described earlier (see Fig. 10a) indicate that the Upper Kura study area underwent initial shortening of the sedimentary fill of the remnant foreland basin into the Greater Caucasus orogen during Sarmatian. We believe that this coincides with the time when the foreland plate refused to flex down under the advancing orogenic load. This resulted in the increased buoyancy of the foreland plate, providing more resistance against orogen advance. We believe that this was the time when thick-skin tectonics (Figs 8a, b & 11a) started in the study area (Fig. 12c). It was eventually followed by thin-skin tectonics (Figs 9a & 11b), indicating later deformational response to ongoing shortening and attempt of the accreting sediments to accommodate a progressively increasing total amount of shortening (Fig. 12d) under the stress regime (Fig. 9b), which was relatively similar to the present-day regime (Fig. 5a).

The thick-skin event must have been the event of the initial internal shortening of the plate, which refused to flex during the Sarmatian and increased its resistance against further overriding by advancing orogens (Fig. 12c). However, as the orogens increased the angle of their taper in reaction to increased foreland resistance, they continued advancing in phases accompanied by the tightening
of the thick-skin architecture and its thin-skin modification (Fig. 11b), representing escalation of the Kura foreland basin deformation (Fig. 12d). The major events (Fig. 10a, b) were made prior to the development of the erosional unconformity overlaid by the Akchagylian–Aphabetonian formation (Figs 3, 5a & 11b), although the convergence continued during the Akchagylian–Aphabetonian (see Figs 9a, b & 10c) and continues until the present day (see Fig. 5a, b).

The only somewhat similar case area, characterized by the foreland plate refusing to flex, which we could find, was the Apulian Plate, located between the Apennines advancing eastward and the Dinarides combined with the Albagno–Hellenides advancing westward. The end of flexural behaviour of the foreland plate next to the Apennines, in the Bradanica basin, has been determined to take place some time during Pliocene (Bertotti, pers. comm. 2000). Using our own data, the stop of the flexural behaviour of the foreland plate in front of the Albano–Hellenides can be more precisely dated, studying a combination of well control and reflection seismic imaging, as taking place sometime at early Pliocene/late Pliocene boundary (Fig. 13).

Figure 13 indicates that significant flexing towards the Albano–Hellenides, which was associated with important orogenward thickening of the flexural basin fill and flexure-driven normal faulting in the circum-forebulge region, terminated some times prior to the Messinian. The Messinian–early Pliocene was a time period characterized by less important fill thickening and the lack of flexure-driven normal faulting. Furthermore, earlier studies, lacking the data on the Ionian Zone close to the forebulge, could find, was the Apulian Plate, located between the pre-existing NE–SW striking rift-related faults in the Ionian Abyssal Plain located between the Mediterranean Ridge and the Calabrian accretionary wedge;

(1) the Tortonian–early Messinian inversion of the pre-existing NE–SW striking rift-related faults in the Ionian Abyssal Plain located between the Mediterranean Ridge and the Calabrian accretionary wedge;

(2) post-Burdigalian development of the contractional architecture of the Pre-Apulian Zone and Cika Belt of the Ionian Zone;

(3) pre-Tortonian shortening of the central portion of the Ionian Zone;

(4) post-early Neogene shortening of the most proximal portion of the Cika Belt located right in front of the Kurveleshi Belt; and

(5) post-middle Neogene thrusting of the frontal thrust of the Ionian Zone in the area to the west of Corfu Island, Greece.

Our interpretation of Figure 13 indicates that the contact between the Ionian Zone and Apulian Platform reached a stage similar to the late Sarmatian stage of the Kura basin sometimes during middle Pliocene. This was the stage when both foreland plates refused to flex, and their fill was deposited in a form of a sub-horizontal body instead of sediment body thickening towards one of the loading orogens. Unlike the Kura case, the Apulian case shown in Figure 13 represents the stage just before the pervasive shortening of the foreland plate, resulting in a development of the thick-skin architecture.

Conclusions

The main outcomes of this study are the following:

(1) The study area underwent its main shortening events during the Sarmatian–Pontian period, followed by less intensive shortening from the Akchagylian until the present day.

(2) Thick-skin tectonics happened first, reactivating pre-existing rift grain. It was the initial deformational response of the foreland plate that refused to flex under advancing orogenic load and became more buoyant during the late Sarmatian, exerting increased resistance to the continuing convergence.

(3) Thin-skin tectonics took place a little later, accommodating the extra shortening of the already fully developed thick-skin structures.

(4) The overall deformation did not result from orthogonal thrusting but dextral transpression.

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