Foredeep basins: the main features and model of formation

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Abstract

The comparative study of the foredeep basins of the northern Caucasus, Alpine and Apennine foredeeps revealed that there are no significant differences between outer and inner foredeeps situated at the front and the back of the crustal-scale thrust. So the driving mechanism should be essentially the same for internal and external foredeeps and elastic flexure should not be supposed as the only (sometimes as the main) mechanism of foredeep formation. The common feature of the foredeep evolution is a phase of slow subsidence lasting for 50 million years or more (passive margin stage sensu stricto) and then relative uplift followed by rapid subsidence. This twofold event of relative uplift (the duration is several million years) giving way to rapid subsidence may be repeated several times. The relative uplift or nearly equal to zero subsidence of the foredeep basins appeared to be synchronous with phases of external compression within adjacent mountain belts, as well as rapid subsidence took place between phases of external compression. Thus, the first stage of relative uplift could be recognised as the first stage of the foredeeps formation. To describe the development of the foredeeps at both sides of the compressional belts a model of evolution of the Earth’s outer shell has been employed. The model predicts that when the Earth’s outer shell has been once disturbed by external intraplate- or mantle-induced forces, small-scale convection within the asthenosphere arises. This convection causes deformation in the lithosphere of orogenic belts over a long period of time after the external forces stop. This manifests in uplift of belts and subsidence at their periphery. The suggested model shows a good agreement with data on the foredeeps structure and evolution. In particular, application of the model enabled us to explain the thickness of sediments in the foredeep basins and their shape, the formation of the foredeeps not only at the front but also at the back of asymmetric compressional thrust belts, uplift of foredeeps during compression in the belts and rapid subsidence after this compression stops. Numerical results for the Great Caucasus region showed that the first compressional event at the orogenic stage took place there before the formation of the Maykopian sediments, i.e. 39.5 Ma, in the end of the closure of the Arabian Ocean. The three further compressional events can also be recognised: one of them being between 16.6 and 15.8, the other between 14.3 and 12.3 Ma, and the last one between 7.0 and 5.2 Ma. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

Foredeep refers to a sedimentary basin closely related to the formation of a mountain chain and situated between its front and the adjacent craton (see for example, Allen et al., 1986). In the Perimediterranean belts foredeeps are well developed not only at the front of the mountain belts but also in the internal parts of the collisional belts (e.g. Pyrenees: Aquitaine basin; Alps: southern Alpine fore-
deep; Caucasus: Azov–Kuban and Terek–Caspian troughs).

Models of a crustal-scale thrust in a continent–continent collision zone were mainly employed for quantitative description of foredeep basins formation. These models describe a foredeep as a result of elastic flexure of underthrust plate that causes a depression at the front of the thrust. This model was described qualitatively by Price (1973); later numerical simulations were made for a uniform elastic plate (Jourdan, 1981), for a plate with temperature-dependent elastic thickness (Beaumont, 1981), for a visco-elastic plate (Karner and Watts, 1983), and for a nonuniform topography of underthrust and overthrust plates (Stockmal et al., 1986).

Problems arise when comparing these models with geological and geophysical crustal-scale images across mountain belts. For example, according to these geodynamical models, the higher a mountain chain the deeper has to be the foredeep. In many cases an opposite relationship is observed (e.g. northern Caucasus, the Apennines). To avoid this problem a ‘hidden load’ (i.e. positive density anomaly) was introduced inside the Alpine (Karner and Watts, 1983) and Apennine lithosphere (Royden and Karner, 1984). Such assumptions are reasonable when they are not conflicting with geophysical data. For the Great Caucasus region, for example, gravity anomalies show that the average density of the crust of the mountain chain is less than in the adjacent platform regions, so, positive density anomalies could not be proposed within the crust of the Great Caucasus ridge. The other problems which call for development of the flexural model are as follows:

1. The response of the lithosphere to long-term regional tectonic effects is not purely elastic (see, for example Ranalli and Murphy, 1987). The more realistic visco-elastic models predict that foredeeps should be uplifted and eroded through time due to viscous relaxation. Old foredeeps (for example the Urals) do not display such behaviour, but sometimes they are overlain by younger sediments.

2. The structure of the majority of mountain belts is asymmetric: the main thrusts are situated on one side of the belt. As a result the weight of overthrust rocks at the back of a mountain chain (back thrusts) is not enough to produce a foredeep by lithospheric flexure. The northern Caucasus foredeep especially in its western part (Azov–Kuban trough) is an example of a situation when thrusts are not well developed.

(3) The data on the evolution of many mountain belts have shown that the process of continental collision was not continuous, but included several stages of external compression. According to the flexural model, foredeep basins should subside as loading increases, i.e. periods of thrusting in mountain belts and periods of subsidence of the foredeep basins should be simultaneous. Detailed analysis of the data for the Carpathians, Urals and some other foredeeps led Artyushkov et al. (1996) to the conclusion that dependence between uplift of a mountain chain and subsidence in the foredeep is opposite: periods of uplift and thrusting in the mountain belts were accompanied by uplift or zero subsidence within the foredeeps, as well as subsidence of foredeep basins took place without external compression, i.e. when there was no thrusting and uplift in the mountain belts.

Analysis of the data on the structure and evolution of foredeeps situated at the front and at the back of mountain belts shows a strong similarity among them even if these belts are asymmetric. Thus, the main geodynamical processes responsible for the formation of a foredeep on each side of a mountain belt should be essentially the same. Crustal-scale thrusts that always accompanying formation of foredeep basins in a continent–continent collision zone undoubtedly rank among the main mechanisms of foredeep formation. But even in case they considerably change the inner structure of a foredeep basin they cannot be considered as being the only driving mechanism. In addition to such factors as structure and thermal state of the lithosphere (for example, ‘hidden’ load or non-steady distribution of temperature; Desegaulx et al., 1991) we propose to consider the processes which took place within the lithosphere and uppermost part of the mantle in the zone of the continent–continent collision.

Recent seismic tomography data (for the Alps and Apennines, see, for example Spackman, 1990) demonstrate that almost all large-scale lithospheric structures (in continent–continent collision zones as well as rifts and subduction zones) have deep roots stretching down as much as several hundred kilometres. These roots may have different origins, but whether they were formed as a result of disturbance
of the Earth’s outer shell by intraplate stress or due to mantle flows, there is no question that formation and evolution of large-scale lithospheric structures, such as orogenic belts, cannot be adequately described by simple plate kinematics and calls for incorporation of dynamic processes in the lithosphere, asthenosphere, and upper mantle and their interactions.

Instability of the Earth’s outer shell was discussed previously by Buck (1986), Housman and England (1986), Fleitout et al. (1986), Keen and Boutilier (1995) and others. Below we will consider the dynamics of interaction of the lithosphere, asthenosphere and upper mantle to investigate the processes of foredeep formation, after a short description of the main large-scale features of foredeep structure and formation. It will be shown that variations in depth and thickness of the asthenosphere and other density and/or rheological interfaces in compressional belts provoke a small-scale convection within the asthenosphere inducing in its turn movements in the lithosphere. The secondary-scale movements cause the alternation of compressional and extensional zones within large-scale lithospheric structures, which can hardly be explained by large-scale plate kinematics (see, for example Ziegler, 1992). Movements of this kind can produce zones of subsidence and relative extension on each side of a compressional belt.

The suggested model is based on the following assumptions (Mikhailov et al., 1996): density of the layers is supposed to be temperature-dependent, the top of the asthenosphere coincides with a specified isotherm or follows some P–T diagram, and sedimentation and denudation are incorporated in the boundary condition at the top of the model. To reduce the uncertainty of boundary conditions assessment at the base, the model was asymptotically linked to the equations of the mantle convection (Myasnikov et al., 1993). In the end the comparison of numerical results with data on foredeeps will be given.

2. The principal features of geodynamics of foredeep formation

The evolution of the northern Caucasus foredeep is taken as an example here and comparisons with the major states of evolution of the Alpine and Apennine foredeeps are given in order to stress the common features important for large-scale numerical modelling. For detailed analysis of the northern Caucasus foredeep structure and evolution see Mikhailov et al. (1999).

The northern Caucasus foredeep is located to the north of the Great Caucasus ridge which stretches over a distance of approximately 1200 km from the Azov Sea to the west to the Caspian Sea to the east. The northern slope of the Great Caucasus as a whole is commonly considered as the Great Caucasus foredeep. But it can be subdivided into three minor geological provinces that have a different structure and underwent partially different evolution, i.e. the Azov–Kuban trough in the west, the Laba–Malka monocline and the Stavropol high in the centre and the Terek–Caspian trough in the east (Fig. 1). As in many other foredeeps peripheral bulges occur to the north of the foredeep.

Current models suggest that the Caucasus was formed during several periods of regional compression when crustal-scale thrusting took place (see e.g. Khain, 1975). The main thrust plane is located to the south of the southern slope of the Great Caucasus and it dips to the north, so the northern Caucasus foredeep was formed at the back of the crustal-scale thrust. From this point of view it is comparable to the southern-(inner) Alpine foredeep (the northern part of the Po plain) or to the Aquitaine basin in the Pyrenees (Roure et al., 1989).

The sedimentary succession is not older than the Early Triassic (except for narrow troughs), and the thickness of the sediments varies from more than 10 km in the Terek–Caspian and Azov–Kuban troughs up to 1 km or less at the Stavropol high. Although the structure of the troughs has been studied by numerous wells and seismic profiles, there is a lack of data on the structure of the deepest part of it (for example in the Terek–Caspian trough, there are no reliable data on the layers older than the Middle Jurassic).

The Cenozoic compressional evolution of the separate parts of the foredeep and its junction to the mountain belt appear to be strictly dependent upon the extensional history of the region during the Mesozoic (Polino et al., 1996). For this reason modelling of subsidence at the peripheral part of the compressional belts is here mainly referred
Fig. 1. Schematic map of the region. Triangles show the position of wells used to calculate subsidence curves (Fig. 2). $K-B =$ Kanev–Berezan system of basement folds; $DN =$ Dagestan nappe; $T$ and $S$ represent, resp., the Terek and Sundja ridges.

to the eastern part of the belt (the Terek–Caspian trough) which structure is more typical for a thrust belt-related foredeep. Nevertheless, all the distinctive regional compressional events appear simultaneous along the belt, including areas where the foredeep basin developed at the front of a steep belt (Kuban) or onto a basement uplift (Stavropol high). For a comparative study of different parts of the foredeep see (Mikhailov et al., 1999).

Since the Triassic and up to the time when a foredeep basin was formed, this area was the transition zone between the Great Caucasus trough and the Scythian plate. The main regional events during this period are clearly seen on subsidence curves (Fig. 2) based on data from three wells located in the western part of the Terek–Caspian trough (for location see Fig. 1). The curves were chosen as representative of the evolution of the different parts of the Terek–Caspian trough (for details of the calculation of subsidence curves see Mikhailov et al., 1999).

The main common events of all the curves are as given below.

1. A slow subsidence during all of the Mesozoic, the Paleocene and the Early and Middle Eocene ($t > 39.5$ Ma). According to Koronovsky and Milanovsky (1987) the trough was a zone of subsidence, without any considerable deformations of sedimentary cover since the Early Jurassic. The long-term component of subsidence (Fig. 2) is similar to the thermal component (Mikhailov, 1993), whereas the short-term events are related to regional compression or extension as demonstrated in Mikhailov et al. (1999).

2. Late Eocene ($39.5 > t > 36.0$ Ma) was a period of relative uplift, followed by a rapid subsidence at Maykopian time ($36.0 > t > 16.6$ Ma). It should be mentioned that subsidence curves in Fig. 2 show the uplift since the Early Eocene because the sedimentary sequences of the Paleocene and the Early and Middle Eocene in these wells were not separated. In several wells where the data are more detailed, the uplift event appeared to be considerably shorter: only during the Late Eocene (for details see Mikhailov et al., 1999).
(3) The next stage of relative uplift started in the Middle Miocene ($t = 16.6$) and was also followed by subsidence beginning at $t = 15.8$ Ma. Another regional event of relative uplift or close-to-zero subsidence took place between 14.3 and 12.3 Ma. This event was followed by the formation of molasse conglomerates. After this at least one or two more events of relative uplift occurred. According to Mikhailov et al. (1999) one of the compressional events started at 7.2 Ma when the topography close to the modern one
was created. Sediments of this age at the northern slope of the Great Caucasus were eroded.

It should be emphasised that these features are not typical for the Terek–Caspian trough only; a similar trend and periods of evolution took place also in the Azov–Kuban trough and at the Stavropol high (for the last one up to \( t = 5.2 \) Ma) and they are essentially the same in the Po plain (Fig. 3) and the Molasse basin (see Homewood et al., 1986).

Comparing subsidence curves for the Terek–Caspian trough to those for the Azov–Kuban one, the Molasse, Alpine and Apennine foredeeps, the following common features emerged:

1. Subsidence curves analysis in the northern Caucasus foredeep did not reveal the migration of its depocentre that was pointed out for the Alpine and Apennine foredeeps, the following common features emerged: The phenomenon seems to be closely related to kinematics of the adjacent mountain belt and an apparent migration of the depocentre in these foredeeps depends upon movements along thrusts. In this particular case, the zone of maximum thickness of sediments do not coincide with the area of maximum subsidence. To estimate the location and possible displacement of subsidence maximum for the Alpine and Apennine foredeeps, the dimensions of thrusts and displacements along them should be taken into account.

2. Another characteristic feature mentioned in Allen et al. (1986) is that at the stage of rapid subsidence, curves are straight or convex upward. For the northern Caucasus foredeep this can be seen in Fig. 2. It seems obvious that the relatively short phase of rapid subsidence cannot be due to lithosphere cooling, but should be caused by other geodynamic process. On the other hand, it does not exclude existence of long-term thermal subsidence during the process of foredeep formation.

It is extremely important to state the relationships between the time of crustal-scale thrusting and the main events during the foredeep formation. All subsidence curves show an important event in the foredeep before the deposition of the Maykopian sediments (36 Ma), as revealed by relative uplift or slowing down of subsidence. During the Paleocene–Early Eocene there was a period of collisional volcanism in the Lesser Caucasus and folding in the Great Caucasus trough (Lomize, 1987). In addition, reworked nannofossils in the base of Maykopian series demonstrate that erosion was active in the area occupied by the present-day Great Caucasus (Lozar and Polino, 1997). This makes it possible to conclude that this event arose from regional compression.

The similar features can be recognised in the evolution of the Apennine and Alpine foredeeps. According to Ricci Lucchi (1986), in the Apennines
the phase of rapid subsidence also started at approximately 38 Ma and the terrigenous material was delivered from the north. The main events appear synchronous until the Messinian (6–5 Ma) when the Apenninic chain emerged. The Legnaro well (Fig. 3) shows no sediments for the period between 23 and 1.6 Ma, so it is not clear whether there were phases of uplift and subsidence during the Messinian phase of compression.

According to Pfiffner (1986) similar events took place also in the Molasse basin. Rapid subsidence with flysch sedimentation at (45 > t > 35 Ma) points to the existence of a moderate topography in the orogenic belt at that time, while further considerable increase in bulk and grain size of the terrigenous material is an indication of subsequent growth of the Alpine belt. So, for the Alps there are evidences that compression started earlier than the stage of rapid subsidence.

Thus, taking into account all points mentioned above we can conclude the following:

(1) In the process of foredeep formation a first phase of slow subsidence (passive margin stage sensu stricto) ends by the relative uplift followed by rapid subsidence (of 15–20 million years duration). This twofold event of relative uplift (of less than 10 million years duration) followed by a phase of rapid subsidence may occur several times. The first stage of relative uplift should, thus, be recognised as the first stage of foredeep formation.

(2) There are no significant differences between outer and inner foredeeps situated at the front and at the back of the crustal-scale thrust. So, elastic flexure is not the only driving mechanism in foredeep formation and, moreover, this driving mechanism should be essentially the same for internal and external foredeeps.

3. The model of subsidence at the peripheral parts of compressional belts

In the previous section we concluded that flexural deflection under the load of the thrust cannot be the only process responsible for subsidence on each side of collisional belts. One possible explanation is that part of the subsidence of foredeeps may be due to processes in the lithosphere and mantle induced by restoration of mechanical and thermal equilibrium disturbed during continental collision. To investigate these processes we discuss the interaction of the lithosphere and the upper mantle using the model of the Earth’s outer shell (Myasnikov et al., 1993; Mikhailov et al., 1996). In this model the sedimentary layer, the lithosphere, the asthenosphere and part of the upper mantle below the asthenosphere are considered as a boundary layer at the top of the convecting mantle. To describe the process of slow deformations of the boundary layer at the regional scale we used the model of compressible viscous fluid. The densities of all layers were functions of co-ordinates and temperature, the top of the asthenosphere was supposed to be a phase boundary and coincided with a specified isotherm or followed a specified P–T diagram, the process of denudation and sedimentation was incorporated into the boundary condition at the top of the model. To solve the mathematical problem, a small parameter, $\varepsilon_0 = 2/\sqrt{Ra} \approx 10^{-4}$ (where $Ra$ is Raleigh number), was introduced. This small parameter was used for expansion of the components of the velocity field and stress tensor into the series of power $\varepsilon_0$ within the main volume of the upper mantle. The viscosity of the asthenosphere and sedimentary layer was supposed to be $\varepsilon = \sqrt{\varepsilon_0}$ times less than that of the mantle, so $\mu_{\text{sed}} = \mu_{\text{asth}} = \mu_m \cdot \varepsilon$, and, on the contrary, the viscosity of the lithosphere was $\varepsilon$ times greater than that of the mantle so $\mu_{\text{lith}} = \mu_m / \varepsilon$. In the examples given below $\varepsilon = 10^{-2}$, $\mu_m = 10^{23}$ Pa s, $\mu_{\text{lith}} = \mu_{\text{asth}} = 10^{21}$ Pa s and $\mu_{\text{sed}} = 10^{25}$ Pa s. Within the boundary layer the extended vertical co-ordinate was introduced and the equations for evolution of the boundary layer were asymptotically linked to the equations of the mantle convection (see Appendix A).

The qualitative analysis of equations obtained as well as the results of the numerical calculations using the finite-difference method revealed the following principal features of evolution of the outer shell:

(1) The horizontal and vertical components of the velocity vector within the boundary layer (Eqs. 3–6 of Appendix A) include movements induced by external forces applied to the boundary layer at an active tectonic stage, $u_0(x, t)$, as well as movements caused by horizontal gradients of pressure, $u_1(x, z, t)$, and non-steady distribution of temperature, $Q(x, z, t)$.
Within the lithosphere and the boundary-layer’s part of the mantle below the asthenosphere, the main term of expansion of the horizontal component of velocity into the series of power \( \epsilon \) does not depend on a vertical co-ordinate. (For the lithosphere this statement is valid, unless it contains sublayers where the viscosity is of the order of that of the asthenosphere with a thickness comparable to that of the lithosphere.) As a result, the sedimentary layer and the asthenosphere are the only layers where the horizontal component of velocity depends on a vertical co-ordinate, i.e. decoupling can occur only in the asthenosphere. The extent of decoupling depends on the rate of deformation.

When active tectonic events are relatively short (5–10 million years) and the characteristic horizontal velocity caused by applied external forces, \( u_0(x, t) \), is of the order of 1 cm/year, the terms in Eqs. 3–6 (Appendix A) caused by the horizontal gradients of density, \( u_1(x, z, t) \), as well as the thermal process, \( Q(x, z, t) \), are small in comparison to \( u_0(x, t) \). Hence, at the short tectonic events (when the initial distribution of density and temperature are horizontally homogeneous) the horizontal components of the velocity vector in all the layers are approximately the same. And, thus, at the active tectonic stage there is no significant difference whether in the boundary layer external forces are applied to the lithosphere or to the mantle, i.e. between intraplate- and mantle-induced processes. (The distinctive feature of mantle-induced processes is that the heat flow at the base of the model can be considerably disturbed.) In the example below, a velocity component related to the external tectonic forces, \( u_0(x, t) \), was assigned within the mantle below the asthenosphere.

When we deal with large-scale spatial and time processes, the boundary layer is nearly in local equilibrium. The level of compensation is at the base of the boundary layer, i.e. below the asthenosphere. Deformations of the outer layer by external (intraplate or mantle) forces disturb the mechanical and thermal equilibrium here and induce circular flows (small-scale convection) within the asthenosphere. The surface manifestations of these convection movements are: further subsidence in regions of extension (sedimentary basins) and uplift in regions of compression (orogens) over a long period of time after the external forces ceased. The style of convection depends on the structure of the boundary layer, in particular, the thickness of the lithosphere and asthenosphere, distribution of density, dimensions of the areas of extension and compression, i.e. on the value of the horizontal pressure gradients within the asthenosphere. The convection is mostly sensitive to small vertical density gradients in the asthenosphere. Under certain conditions the impact of small-scale convection on the base of the lithosphere makes the extensional and compressional structures more complex, e.g. subsidence (foredeeps) at the peripheral parts of orogens. Let us consider the process of subsidence at the borders of a compressional area in more detail.

Let us assume that at the beginning (\( t = 0 \)) all the layers of the boundary shell were horizontal and their thicknesses were as follows: the sedimentary layer 1 km; the lithosphere 99 km (including 30 km of the crust); and the asthenosphere 100 km. (Thickness of the boundary-layer’s mantle below the asthenosphere is not important, because the horizontal component of velocity within it does not depend on a vertical co-ordinate; see Appendix A.) At \( t = 0 \), the density of the sediments was constant and equal to 2.4 g/cm\(^3\), density of the Earth’s crust increases from 2.7 g/cm\(^3\) at the top to 3.1 g/cm\(^3\) at its base, density of the lower lithosphere was constant and equal to 3.3 g/cm\(^3\), while the density of the asthenosphere slightly increased with depth from 3.30 at the top to 3.31 g/cm\(^3\) at the base. The density of the boundary-layer mantle below the asthenosphere was constant and equal to 3.31 g/cm\(^3\). We also assume that during the first 5 million years intraplate or mantle external forces formed a compressional belt of 200 km width. Horizontal velocity was assigned in the form of a trapezium (see Fig. 4b) with a maximum value of 1.6 cm/year. As a result, at the end of the ‘active’ stage (\( t = 5 \) million years) the lithosphere became 1.31 times thicker and an uplift of 2 km was formed. As horizontal pressure gradients within the boundary layer increased they induced small-scale convection in the asthenosphere (Fig. 4a). Two convective cells on each side of the compressional belt developed. The upwelling flows under the peripheral parts caused extension and subsidence, as well as downward flows behind the centre of the belt, resulting in compression and further uplift of this territory. Due
to these flows, the lithosphere moved to the centre of the compressional belt over a long period of time after the external compression stopped, and on each side of this belt sedimentary basins (Fig. 4a) with a depth of 3 km formed at $t = 30$ million years (i.e. over 25 million years after the external compression stopped). Fig. 5 shows the highly magnified structure of the foredeep at $t = 25, 45$ and 75 million years after the end of the active stage. The foredeep basin is not symmetrical, i.e. the outer slope (on the left) is more gentle than the inner one (on the right). When there are no external forces, subsidence of the foredeep slowly decreases with time. The characteristic feature of the state of stress within the foredeep is that during the ‘active’ stage ($t < 5$ million years) the whole foredeep is subjected to compression; after the end of this stage almost all foredeeps undergo extension except the inner slopes close to the orogen, being still under compression. Compression in the area of the inner slope of the foredeep supports formation of thrusts in this area.

Two subsidence curves, for the central part of the foredeep (curve 1 in Fig. 6) and close to its inner slope (curve 2 in Fig. 6), were constructed for two stages of compression: $0 < t < 5$ million years and $25 < t < 28$ million years (for the comparison with northern Caucasus subsidence curves, the beginning of the first compressional event was placed at 40 Ma and the horizontal axis at Fig. 6 was numbered as $\tau = 40\text{ Ma} - t$, where $\tau$ is time before present). Compressional events reflected on these curves by periods of uplifting, as well as rapid subsidence took place between these ‘active’ stages. The rate of subsidence appeared to be close to constant in the centre of the foredeep. At its inner slope it slowly decreased after the second compressional event, when this area became a part of the mountain belt. The comparison of typical subsidence curves for the Terek–Caspian trough (Fig. 2) with the theoretical curves in Fig. 6 made it possible to conclude that the first compressional event in the Great Caucasus took place in the Late Eocene (39.5 Ma), before the formation of the Maykopian sediments. The other
compressional events could be dated as 16.6–15.8 Ma, 14.3–12.3 Ma and 7.0–5.2 Ma. At present, the consensus regarding the beginning of the growth of the Great Caucasus, manifested in the formation of molasse conglomerates, appears to be the Sarmatian (13.7–9.3 Ma) (Koronovsky and Milanovsky, 1987, and many others). According to the results of the modelling the first compression took place considerably earlier, at 39.5 Ma. This time coincides with the end of oceanic subduction in the Arabic Ocean (Dercourt et al., 1993). As mentioned above, our conclusion was supported by recent studies of microfossils from the base of the Maykopian series, showing that at that time there were uplifted areas at the place of the Great Caucasus ridges (Lozar and Polino, 1997).

It is significant that the linear-viscous boundary layer model describes only principal features of evolution of the compressional belt–foredeep basin system. The main result of our modelling is that due to impact of small-scale convection on the base of the lithosphere, subsidence should take place on each side of a compressional belt. In general, deformation of the lithosphere, especially in its uppermost part, is vastly more complicated: deformations are mostly associated with thrust planes; elastic proper-
ties of the upper crust and possibly of the mantle below Moho took place at the beginning of the isostatic rebound process. All these factors make the structure of foredeeps more complicated than the model of the Earth’s outer shell predicts. But, anyway, when regional compression takes place, two foredeep basins, caused by small-scale convection within the asthenosphere, should appear on each side of a compressional belt. It could be speculated that for more realistic visco-elastic media, the subsidence after a short compressional event will be mainly due to elastic flexure. In the course of time the viscous relaxation will reduce the depth of the foredeep, as well as that small-scale convection will make it deeper, so the role of elasticity will reduce. To compare the numerical results with seismic data, the thickness of the crust and the location of foredeep basins were calculated at 40 Ma after one-step compression of the initially asymmetric model. The duration of the compression event was 5 million years, the lithosphere at the left side of the compressional belt was thicker than that at the right side. In Fig. 7 numerical results were combined with the DSS profile across the Pyrenees (Roure et al., 1989) to illustrate that also in the other cases than the compared ones the principal calculated large-scale features nearly match the observed ones.
4. Conclusion

If the Earth’s outer shell has been disturbed by external intraplate- or mantle-induced forces, small-scale convection within the asthenosphere arises. This convection makes for deformations in the lithosphere over a long period of time after the external forces stop. The small-scale convection amplifies uplift of orogenic belts and causes areas of subsidence at their periphery. We consider the small-scale convection to be one of the main mechanisms of foredeep basins formation. The suggested model showed a good agreement with the data on the foredeep structure and evolution. In particular, it is able to explain the thickness of sediments in foredeep basins and their shape, formation of foredeeps not only at the front but also at the back of compressional thrust belts, uplift of a foredeep during compression in the belt and rapid subsidence after this compression stops. Comparison of the numerical results with the observed data on the northern Caucasus foredeep made it possible to conclude that it was formed as a result of at least four compressional events. The first one took place before the formation of the Maykopian sediments, i.e. 39.5 Ma, and could be related to the closure of the Arabian Ocean and subsequent beginning of the continental collision in the Caucasus. The three further compressional events can also be recognised: one of them being between 16.6 and 15.8 Ma, the others between 14.3 and 13.7 Ma and between 7.0 and 5.2 Ma.

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Appendix A

To describe the slow deformation of rocks on a time scale of 10^6 years or more we employed the model of viscous compressible fluid. Application of this model to the lithosphere sometimes provokes objections, because deformations of the upper crust (about 10–20 km thick) are concentrated along fault zones. Since our model focuses on the description of regional long-term features of the deformations within the outer shell of 100–200 km thick, the effect of elastic interlayers can be neglected.

To describe the convective motions within the main volume of the mantle, we introduce the following characteristic parameters: \( Ra = g_0 R^2 \rho_0 (k_0 \theta_0) / \kappa \) – Rayleigh number, \( \theta_0 = R^2 / (Ra^{1/2} \kappa_0) \) – time of convective mixing, \( \nu_0 = R / \theta_0 \) – rate of heat transfer, where \( g_0 = 4 \pi \rho_0 G / 3 \) is the gravity acceleration, \( R \) is the Earth’s radius, \( \rho_0, \theta_0, k_0 \) are, resp., mean values of the density, viscosity and thermal diffusivity of the mantle, \( G \) is the gravitational constant. The values accepted for the mantle are: \( \rho = 3 \times 10^3 \text{ kg/m}^3 \), \( \theta_0 = 10^3 \text{ Pa s} \), \( k_0 = 10^{-6} \text{ m}^2/\text{s} \), which give \( Ra = 10^6 \) and \( g_0 = 10 \text{ m/s}^2 \) when \( R = 6.4 \times 10^6 \text{ m} \).

We introduce a small parameter \( e_0 = 1/\sqrt{Ra} \approx 10^{-4} \) and the following dimensionless quantities (the corresponding characteristic scales taken as units to generate dimensionless parameters are given in parentheses): \( x_0 \) are Cartesian co-ordinates, vertical axis is directed upwards \( (R) \), \( v \) are the velocity components of the vector \( V = [u, v, w](x_0) \), \( t \) is time \( (t) \), \( P \) is pressure \( (\rho_0 g_0 R) \), \( T \) is temperature \( (T_0) \), \( \rho \) is gravitational potential \( (\rho_0 R) \), \( Q_R \) is heat generation \( (Q_0) \), \( C_P \) is specific heat at constant pressure \( (C_{P_0}) \). Dimensionless density \( \rho \), viscosity \( \eta \), and thermal diffusivity \( \kappa \) were obtained by dividing corresponding dimension quantities into \( \rho_0, \eta_0, \kappa_0 \), respectively. The resultant nondimensional equations for the time-dependent velocity, density and temperature under the above assumptions in a 2D form are (\( \delta_{\alpha \gamma} \) is the Kronecker delta and repeated subscripts, as usual, imply summation):

\[
\begin{align*}
\frac{\partial \rho}{\partial x_0} + \frac{\partial \rho}{\partial x_0} &= \epsilon \frac{\partial}{\partial t} \left( \rho \left( \frac{\partial v}{\partial x_0} + \frac{\partial u}{\partial x_0} - 2 \delta_{\alpha \gamma} \frac{\partial u}{\partial x_0} \right) \right) \\
\frac{\partial \rho}{\partial t} &= \frac{\partial \rho}{\partial x_0} = 0; \\
\frac{\partial T}{\partial t} &= \epsilon \frac{\partial}{\partial x_0} \left( \frac{\partial T}{\partial x_0} \right) + Q_R \\
\alpha, \beta, \gamma &= 1, 2
\end{align*}
\]

Assume that the outer shell consist of four layers of uniform viscosity: a boundary layer’s mantle of viscosity \( \eta_0 = \eta_0 \), a low-viscosity asthenosphere \( \eta = \eta_0 \), a high-viscosity lithosphere \( \eta = \eta_0 / \epsilon \), and a low-viscosity sedimentary layer \( \eta = \epsilon \eta_0 (\epsilon = \sqrt{Ra} \approx 10^{-2}) \). The density of sediments is constant \( \rho_s \), in other layers it is a function of co-ordinates \( \rho(x, z, t) \). In a near-surface zone, an extended vertical co-ordinate \( Z = -R_0 = (Z - R_0 - Z) / \epsilon \) was introduced. Here, \( R \) is the radius of the Earth and \( R_0 \) stands for the Earth’s hydrodynamic radius, or, the same free mantle level. The origin of extended co-ordinates was placed in the mantle below the asthenosphere. The vertical component of the velocity vector in the extended co-ordinates is \( W = w / \epsilon \).

Functions \( u, W, P \) may be expanded into series of power \( \epsilon \), for example for \( u \):

\[
u = u^{(0)}(x, Z, t) + \epsilon u^{(1)}(x, Z, t) + \cdots
\]

The free boundary conditions are specified at the top of the boundary layer (i.e. at the top of the sedimentary layer \( Z = z_s \)) and the continuity conditions for normal and shear
stresses and for the velocity components are set at the internal model’s boundaries (top of the lithosphere, \(z_i\), and base of the asthenosphere, \(z_m\), and \(z_M\), respectively). Then we obtain the following equations for the layers. Note that the zero-order terms in the expansion of the horizontal component of velocity in the mantle and lithosphere are independent of the vertical co-ordinate \(Z\).

For the sedimentary cover, \(z_i \leq Z \leq z_1\):

\[
P_s^{(0)} = 0; \quad P_s^{(1)} = \rho_s(z_1 - Z); \quad \frac{\partial P_s^{(0)}}{\partial Z} = \frac{\partial u_s^{(0)}}{\partial x}(Z - z_1)
\]

For the lithosphere, \(z_1 \leq Z \leq z_m\):

\[
P_l^{(0)} = 0; \quad P_l^{(1)} = \rho_l(z_1 - z_m) + \int_{z_1}^{z_m} \rho \, dZ - \frac{Z}{2} \times \eta \left( \frac{\partial u_l^{(0)}}{\partial x} - 2 \times \frac{\partial u_l^{(0)}}{\partial Z} \right); \quad \frac{\partial P_l^{(0)}}{\partial Z} = 0
\]

For the asthenosphere, \(z_m \leq Z \leq z_M\):

\[
P_a^{(0)} = 0; \quad P_a^{(1)} = \rho_a(z_1 - z_m) + \int_{z_m}^{z_1} \rho \, dZ; \quad \frac{\partial P_a^{(0)}}{\partial Z} = \eta \frac{\partial^2 u_a^{(0)}}{\partial Z^2}
\]

Furthermore, the equations for density and temperature should be added for each of the layers. Note that the zero-order terms in the expansion of the horizontal component of velocity in the mantle and lithosphere are independent of the vertical co-ordinate \(Z\).

To match the mantle and boundary layer model we apply the method of asymptotic expansion matching, according to which any function \(f\), represented in terms of \(\varepsilon = \sqrt{\frac{z}{R_0}}\) in the boundary layer (\(f^{(i)}\)) and in terms \(\varepsilon_j\) in the main part of the mantle (\(f^{(j)}\)), must satisfy the relations of asymptotic equivalence in the vicinity of the model surface (\(z = R\)):}

\[
\left[ f^{(i)} + \varepsilon f^{(j)} + \cdots \right]_{z = R} \approx \left[ f' + \varepsilon f'' + \cdots \right]_{z = R}
\]

where \(R_0\) is the Earth’s hydrodynamic radius (free mantle level). This equation was obtained by expansion of each term of series Eq. 1 into Tailor’s series in the vicinity of \(z = R_0\). Coupling conditions for the global model leads to the following relationships:

\[
(1) \lim_{Z \to -\infty} u_m^{(0)}(x, Z, t) = u(x, R, t)
\]

hence, taking into account that \(\partial u_m^{(0)}/\partial Z = 0\), we obtain \(u_m^{(0)}(x, Z, t) = u'(x, R, t)\).

Thus, the boundary condition at the bottom of the boundary layer is the horizontal velocity component in the mantle below the asthenosphere as a function of \(x\) and \(t\); this function is denoted by \(u_0(x, t)\).

\[
(2) \int_{-\infty}^{z_m} \left( \rho - \rho_0 \right) dZ + \rho_l(z_m - z_1) + \rho_m(z_1 - R_0) = 0
\]

which is a generalised equation of the isotropy. According to Eq. 2 the isotropic balance (equality of pressure) attains at the bottom of the outer shell in any vertical section at any time. Moreover, a horizontal pressure gradient may exist at any level inside the model. This equation in particular, may be used for the determination of \(R_0\) when \(t = 0\).

\[
(3) \lim_{Z \to -\infty} W_m^{(0)}(x, Z, t) = -(Z - R_0) \frac{\partial u_0'}{\partial x}(x, R, t)
\]

\[
(4) \lim_{Z \to -\infty} \rho(x, Z, t) = \rho_0
\]

Assuming that the density of the layers depends only on temperature:

\[
\frac{d\rho}{dT} = -\alpha(\rho, Z, 0) \frac{dT}{dt}
\]

where \(\alpha = 3 \times 10^{-5} \text{ grad}\) is the thermal expansion coefficient, we then obtain the equations for the velocity component.

In the mantle:

\[
u_0 = u_0(x, t)
\]

\[
W_m = -(Z - R_0) \frac{\partial u_0}{\partial x} + Q(x, Z, t)
\]

In the asthenosphere:

\[
u_a = u_0(x, t) + u_1(x, Z, t)
\]

\[
W_a = -(Z - R_0) \frac{\partial u_0}{\partial x} - \int_{z_m}^{Z_m} \frac{\partial u_1}{\partial x} dZ + Q(x, Z, t)
\]

In the lithosphere:

\[
u_l = u_0(x, t) + \Pi(x, t)
\]

\[
W_l = -(Z - R_0) \frac{\partial u_0}{\partial x} - \int_{z_m}^{z_1} \frac{\partial u_1}{\partial x} dZ - \frac{Z}{2} - \eta \frac{\partial^2 u_1}{\partial Z^2} + Q(x, Z, t)
\]

In the sedimentary layer:

\[
u_s = u_0(x, t) + \frac{\rho_s}{2q} \frac{\partial c_s}{\partial x}(Z - \xi_s)^2 \bar{T} - \frac{\partial u_s}{\partial x} + \Pi(x, t)
\]

\[
W_s = W_l - \frac{\rho_s}{2q} \frac{\partial c_s}{\partial x}(Z - \xi_s)^2 \frac{Z - \xi_s}{3} (\xi_s - \xi_m)
\]

where

\[
Q(x, Z, t) = a \int_{-\infty}^{Z_m} \frac{\partial \bar{T}}{\partial Z} \left( \frac{\partial T}{\partial Z} \right) \rho \, dZ + \frac{Q_0(x, Z)}{\rho(x, Z, t)C_p}
\]

is the thermal subsidence, \(Q_0(x, Z)\) is the heat generation and

\[
u_1(x, Z, t) = \frac{1}{\eta} \int_{z_m}^{Z} \frac{\partial P_1^{(1)}}{\partial x} - \frac{dZ}{\partial x} - \frac{1}{\eta} \lambda_1(x, Z) \bar{T}
\]

the value of the last function at the asthenosphere–lithosphere boundary was denoted above by \(\Pi(x, t) = u_1(x, \xi_s)\) and \(\lambda_1(x, Z)\) is an unknown function.

The locations of the internal boundaries are determined by using the equation (below \(i = m, a, l, s\) for the mantle, asthenosphere, lithosphere and sedimentary layer, respectively, velocity components with index \(i\) stand for their values at the \(i\)-th surface):

\[
\frac{\partial z_i}{\partial t} + u_i(x, \xi_i, t) \frac{\partial z_i}{\partial x} = W_i
\]
which for the external surface is completed by the terms allowing for the sedimentation and erosion (the second and third terms in the right hand side of the equation):

$$\frac{\partial z}{\partial t} + u_\perp \frac{\partial z}{\partial x} = W_t + \psi(x, t) + \frac{1}{\lambda} \frac{\partial^2 z}{\partial x^2} \tag{8}$$

The sedimentation function can be arbitrary and may depend, for example, on the sea level, distance from the shoreline, and so on.

Now by using the above equations and the generalised equation of isostasy (Eq. 2), we can derive the equation for the unknown function $A(x, t)$. By applying the operator $\frac{\partial}{\partial x} + \frac{\partial}{\partial t}$ to Eq. 2 and using the equation for $\frac{\partial \rho}{\partial t}$ in the layers and equations Eqs. 7 and 8, we find after some manipulations:

$$\frac{\partial}{\partial x} \left[ \int_{z_0}^{\zeta} \left( u - u_0 \right) \rho \, dZ \right] = \left( \frac{\lambda^2 \zeta}{\lambda^2 z} + \psi \right) \rho_s$$

(where the lower limit of integration is $\zeta_0$, since below this boundary $u = u_0 = 0$). This equation shows that the total mass transfer across the vertical section, up to the constant, equals to the surface mass transfer due to sedimentation and erosion. For $\lambda = 0, \psi = 0$ this equation coincides with that derived by Turcotte and Shubert (1983) in their consideration of the problem of reflux in the asthenosphere. As a result, we find the equation for $A(x, t)$:

$$\rho_s \left( \frac{\partial^2}{\partial x^2} \left[ \int_{z_0}^{\zeta} \rho \, dZ \right] \right) = A(x, t) \int_{z_0}^{\zeta} P^{(1)}_{W} \, dZ + \int_{z_0}^{\zeta} \int_{\zeta}^{Z} \frac{\partial P^{(1)}_{W}}{\partial x} \, dZ' \, dZ$$

The set of equations is completed by the equations for temperature and density:

$$\frac{dT}{dt} = \frac{\partial}{\partial Z} \left( \chi \frac{\partial T}{\partial Z} \right) + Q_p(x, Z): \frac{\partial \rho}{\partial t} + \rho \frac{\partial Q(x, Z, t)}{\partial Z} = 0$$

This set of equations was solved numerically using the finite-difference method. For details see (Mikhailov et al., 1996).

References


