Role of the 3-D distributions of load and lithospheric strength in orogenic arcs: polystage subsidence in the Carpathians foredeep

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Abstract

It has been widely documented that the depth of foredeeps does not always reflect the topography of the neighboring orogens. In many cases, the topographic load is insufficient to explain basin subsidence. Such is the case of the SE Carpathians where an anomalously deep (almost 13 km) foreland basin has evolved since the Middle Miocene (Badenian). A peculiar feature of this basin is its position relative to the orogen. In contrast to typical foredeeps, which deepen towards the belt, the maximum depth of this basin is 10–20 km out of the orogen. The subsidence in the Carpathians Bend foreland is characterized by two stages: the first is Middle Miocene (Badenian) in age and is related to NE-SW extension when fault-bounded basins were formed. Modeling shows that the foreland underwent small pre-orogenic uniform thinning. The modeling also predicts 100 m post-rift subsidence in accordance with the regional unconformity observed at the Middle/Upper Miocene (Badenian/Sarmatian) boundary. The second subsidence stage follows rifting and is caused by flexural loading of the Carpathians nappes. According to the planform flexural modeling results, the location of depocenter in front of the Carpathians Bend in the latter contractional stage can be accounted for by the present topography if lateral variations in lithospheric strength are taken into account. Since the depth of the predicted basin is half of what is actually observed, an extra-load is required, which is equivalent to 500–800 m in terms of extratopography. The modeling also suggests that the flexural fore-bulge of the Carpathians system is represented by the uplifted Dobrogea. Our explanation for the large subsidence recorded by the SE Carpathians foredeep highlights the control exerted by lateral changes in lithospheric strength on 3-D subsidence patterns in arcuate orogenic belts.

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1. Introduction

This study focuses on the tectonic processes controlling subsidence in the Focșani Depression, an almost 13-km-deep sedimentary basin developed in Miocene to Recent times in front of the Carpathians Bend Zone (Fig. 1), next to the Vrancea seismogenic area. The area is of great interest because of its pronounced, partly post-thrusting subsidence and because of the abnormal position of the depocenter located in front of the belt. To explain the anomalous subsidence, several models invoked the presence of a lithospheric slab sinking in the asthenosphere beneath the Carpathians Bend (also accounting for the Vrancea area seismicity). This slab results either from NW to SE roll-back and detachment of the former oceanic part of the lower plate [1–3] associated with tearing along the Trotuș fault [4] or from W to E mantle delamination [5].

The Carpathians belt (Fig. 1) represents a highly curved orogenic arc formed during Middle Cretaceous to Quaternary times [6,7]. Contractional deformations followed Triassic–Early Cretaceous extension. The emplacement of the outermost Carpathians nappe (namely Subcarpathian) onto the foreland occurred during the Late Miocene (Sarmatian) (e.g. [7]). The junction area between East and South Carpathians consists of out-of-sequence thrusts (e.g. [4]) related to the ESE-ward movement of the orogenic wedge since Late Sarmatian [8]. Over 40 km shortening was
estimated for the Bend orogenic wedge. The area is also the site of the youngest Carpathians contraction (Pliocene–Pleistocene), which was of limited magnitude, in the order of several kilometers [4]. Unlike the East and South Carpathians where major exhumation began at the end of the Middle Miocene, denudation in the Bend region started only at the beginning of Pliocene [9].

The Focșani Depression is a ~13-km-deep basin and is filled with Badenian to Quaternary sediments (Fig. 2). At present, the external margins of the foredeep basin strike NNW–SSE in the E and WSW–ENE in the S. The bottom of the basin is 6.5 km deep at the northernmost junction with the Carpathians structures and it is supposed to prolong beneath them as far N as the Trotuș fault. To the E–NE, the foreland gradually reduces its depth to 0.5–1 km on East European platform and North Dobrogea orogen. Toward the SE, the foreland becomes shallower and crops out in Dobrogea, East of the Danube River.

The SE Carpathians and the adjacent foredeep

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**Fig. 2.** Structural map of the Bend area of the Carpathians foreland at the Pre-Tertiary level (gray contours) and Upper Miocene (Sarmatian) base (black contours); modified after [8]. Depth in kilometers.
have been inferred to be affected by so-called hidden loads (e.g., [1]). Modeling studies showed that the topographic load of the Carpathians is insufficient to fit the observed foredeep geometry and an extra vertical force has to be applied to the deep end of the subducting plate [10]. These models have inferred large lateral variations of lithospheric rigidity along the Carpathians [11], possibly inherited from pre-compressional stages.

In a previous work we documented the 3-D architecture of the Focșani Depression and its evolution through time [8]. In that paper we used a large (>1000 km) seismic survey for mapping and several tens of well checkshots to convert the time structural maps in depth. Having established the basin structure, we make now the step toward numerical modeling in order to identify and quantify the basin forming mechanisms. We have demonstrated the existence of two quite different tectonic regimes. The first one, Middle Miocene (Badenian) in age, was characterized by NE-SW oriented extension and was followed by regional tilting and subsidence.

In this paper we will first model the subsidence associated with the extensional deformation stage (both during and following rifting). Our goals are: (1) to assess the amount of subsidence, which can be ascribed to extension, and (2) to determine the lithospheric characteristics following the extensional episode. We then model the foreland flexure due to the present topographic load using codes taking into account lateral changes of geometry and mechanical properties. Previous flexural modeling studies along vertical sections concluded that the topographical load alone is insufficient to account for the subsidence in the SE Carpathians Bend foreland [1,10,11]. The objective of our modelling study is to investigate whether the observed subsidence can be explained by 3-D loads. In doing so, we also take into account the presence of areas with different lithospheric strengths. This was not done in previous models.

2. Extensional modeling

2.1. The Badenian (16.5–13 Ma) extensional basin

One of the main results of the basin reconstruction presented in Tărăpoancă et al. [8] was the recognition along the eastern margin of the Focșani Depression of a set of NW-SE trending normal faults of Badenian age (Fig. 2). Badenian sediments are up to 4 km thick and mainly found along an area elongated in NW-SE direction. The thickness of Badenian sediments rapidly decreases from Focșani Depression to its eastern shoulder (on the North Dobrogea orogen). The prolongation of the basin to the NW and W is unclear since it is buried beneath the Carpathians nappes. The position and structure of the former western extensional margin of the Focșani Depression remain still unknown.

Figure 3: Cross-section 1 showing the Badenian basins. For location, see Fig. 2. The cross-section is obtained from maps derived from interpretation of industry seismic lines [8].
The Badenian extension is clearly visible to the SE of Focsani Depression where it produced NW trending basins bounded by ENE striking transfer faults (Fig. 2) [8,12]. Basins are narrow and reach a maximum depth of 2 km (Fig. 3). The Badenian sequence represents the syn-rift fill of the basins (Fig. 4). The Badenian/Sarmatian boundary is represented by an almost regional erosional unconformity. This horizon marks the end of extension and is covered by onlapping Sarmatian strata.

2.2. Modeling approach

To model the Badenian extension in the Carpathians Bend foreland we use a forward 2-D numerical modeling code [13] simulating the kinematic and thermal behavior of the thinning lithosphere. The lithosphere is divided in a series of vertical boxes having a given width (3 km in our case) to which thinning factors are assigned. The thinning factor is defined here as the ratio between the initial to final thickness of the box. Different thinning factors can be assigned to the crust and lithospheric mantle simulating a depth-dependent extension [14]. Subsidence is computed for the center of each box. Thinning can be either instantaneous or achieved in steps of finite duration.

To compute the thermal state of the lithosphere, the temperature of the asthenosphere and thermal diffusivity must be introduced. The model allows the computation of lateral heat flow. The syn-rift subsidence depends on the thinning factors and depth of necking [13]. From the thinning factors and thermal state of the lithosphere the unloaded subsidence is determined isostatically and the basement subsidence is computed by flexural loading filling the basin with water and sediments. Sediments fill the space between the imposed paleobathymetry and the basement subsidence. Surface processes, as the erosion of rift shoulders, are not considered. Sediment compaction is taken into account using a simple exponential porosity–depth relation [15].

The cross-section CS 1 (Fig. 3) has been chosen for the 2-D modeling because it best images the extensional basins. To obtain the basin architecture prior to thrusting and flexural loading, we have flattened the top Badenian and restored underlying features (Fig. 5). Since the top Badenian horizon does not extend over the North Dobrogea orogen this approach leads to an artifact in the easternmost rift shoulder overestimating its uplift during Badenian. Actually, most of the uplift occurred during Pliocene (note the thickening of the Pliocene sequence to the W of Peceneaga–Camena fault in Fig. 3).

2.3. Input parameters

Pre-thinning lithospheric and crustal thicknesses represent input values required by the model. The foreland of the Carpathians Bend area is characterized by a present day lithospheric thickness ranging between 170 and 190 km and a crustal thickness ranging between 30 and 40 km [16–18]. An unstretched lithospheric thickness of 180 km has been chosen as modeling assumption. Adopting these parameters, various thinning factors were tested in order to fit the basin geometry. It should be noticed that the model results are
tested only against the basin stratigraphy and not against features such as lithospheric thickness, which have been overprinted by subsequent tectonic events.

Zero paleowater depths have been used. This is consistent with the presence of evaporitic intercalations in the Badenian deposits on the outer part of the foreland. Post-Badenian sediments are mainly shallow water (e.g. [19]). The basin is assumed to be filled with sediments up to sea level and their compaction is taken into account using the values of 0.5 and 0.8 for the surface porosity (as fraction) and characteristic depth constant (km$^{-1}$), respectively, corresponding roughly to a silty sand lithology.

Rifting took place for a duration of 3 Ma during Badenian times [8]. Other parameters used are shown in Table 1.

The model requires the equivalent elastic thickness (EET) and depth of necking as input values, the former being defined either as a constant or associated with a specific isotherm. The modeled section is longer than the width of the basin to avoid boundary effects.

2.4. Results

The best fit model is shown in Fig. 5. Without taking into account the artifact visible in the eastern part of the section (see Section 2.3), a good fit is obtained for most of the profile. Only in the westernmost part there are some differences, probably related to the subsequent deformations along Intramoesian fault [8].

Our best fit model was obtained assuming uniform extension, i.e. the thinning factors are the same for crust and lithospheric mantle. Thinning factors are small and even the deepest central basin has a thinning factor of 1.1 (Fig. 5). The best fit model was obtained using a high EET value of 35 km, an intermediate to deep depth of necking of 25 km and a pre-extension crustal thickness of 32 km.

The thermal anomaly produced by rifting is small due to the low thinning factors and to the lateral dissipation of heat flow. In these circumstances, the post-rift subsidence (post-Badenian) is very low or absent (Fig. 6). The maximum predicted post-subsidence is $<100$ m.

Changes in initial crustal thickness (from 30 to 40 km) have little effect upon the model results. In contrast, a substantial influence is exerted by the EET and, especially, by the depth of necking. Assuming the same distribution of thinning factors

<table>
<thead>
<tr>
<th>Table 1</th>
<th>Parameters used as input data in the extensional modeling</th>
</tr>
</thead>
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<tr>
<td>Temperature</td>
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</tr>
<tr>
<td>Density at surface conditions</td>
<td>Crust $2800$ kg/mc</td>
</tr>
<tr>
<td>Sediment grain density</td>
<td>$2660$ kg/mc</td>
</tr>
<tr>
<td>Coefficient of thermal expansion</td>
<td>$3.4E^{-5}$$^\circ$C</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>$7.8E^{-7}$sqm/s</td>
</tr>
</tbody>
</table>
as in Fig. 5 and a crustal thickness of 32 km, basin geometries for four depths of necking and for four EET values, respectively, are shown in Fig. 7A, B. Due to the low magnitudes of the thinning factors, the changes in these two parameters affect especially the central and central-western part of the rifted area. It can be observed that the same basin and same amount of thinning are obtained by using a lower EET and a higher depth of necking or vice versa. That is, we cannot provide a unique solution in terms of EET, depth of necking and lithospheric thinning using only the basin shape. We note that our preferred values of EET and necking depth characterizing our best fit model are similar to those obtained by Spadini et al. [20] for the Mid-Cretaceous rifting in the Western Black Sea. This is consistent with the fact that both basins are underlain by the same Moesian platform.

An EET value considered as the depth to the 400°C isotherm has led to an almost identical shape of the basin as the elastic value of 35 km (for a depth of necking of 25 km).

2.5. Inferences from extensional modelling

The model predicts almost no post-rift subsidence. The corresponding time interval is encompassed in the unconformity at the base of the Sarmatian observed on the seismic lines (Fig. 4). This means that only a very little portion of the subsidence recorded during Sarmatian-Quaternary (Fig. 3) can be attributed to Badenian extension. As a result of the low amount of Badenian thinning, the reduction of Moesian platform strength following the rifting stage should be small, in order of few kilometers.

3. Flexural modeling

Previous modeling studies concerning the subsidence in the Carpathians have used a 2-D (cross-section) approach [1,10,11]. The main topic addressed by these modeling studies has been the very large subsidence in the Carpathians Bend region. All these investigations concluded that
the load exerted by the present topography of the Carpathians is too small (accounting for maximum 10% of foredeep subsidence) and that ‘hidden loads’ must be invoked. However, none of them took into account the 3-D nature of the Carpathian arc and the map view variations in strength of the lithospheric domains involved.

A topic that has not been addressed by these studies is the geographic position of the depocenter. The Carpathians foredeep in the front of the Bend Zone is peculiar because the depocenter does not lie under the thrust belt as predicted by simple models. A map of the foredeep base (Fig. 2) shows that it becomes shallower towards the belt. We investigate here the possibility that this feature is associated with the presence in the Carpathians domain of lithospheric blocks with significantly differing strengths.

Following an initial cross-section experiment we apply a planform modeling approach. Hereby we test: (1) the effect of 3-D load distribution and lateral variations of lithospheric strengths on basin position, and (2) whether the very large observed subsidence is the effect a 3-D loading rather than of a hidden load as shown by previous 2-D experiments.

3.1. The Sarmatian–Quaternary (13–0 Ma) foredeep basin

The overall architecture of the Focşani Depression is shown in Fig. 2. Three schematic geological sections crossing the basin in various directions are presented in Fig. 8A–C. The history of the entire East Carpathians foredeep basin began in the Sarmatian, in association with the emplacement of Carpathians nappes [6,7].

The northern foreland underwent important westward tilting (Fig. 8C) and Sarmatian sediments are wedge shaped, thickening from 0.6 km in the E to around 2.1 km in the W. This contrasts with the small and roughly constant thickness of the Badenian sequence. Only thin Meotian deposits are preserved.

At the beginning of the Pliocene, while subsidence was continuing in the central parts of the basin, the western margin of the Focşani Depression was tilted to the E and progressively ex-
humed. As a result of these events, around 8-km-thick Sarmatian–Quaternary sediments accumulated after the Badenian times in the Focșani Depression (Fig. 8A). The eastern basin margin is 1–1.6 km deep and dips gently towards the center. In contrast, the western margin is eastward dipping and the Upper Miocene strata crop out having roughly vertical position [21]. Removing this younger deformation, it can be concluded that during Upper Miocene time span the Focșani Depression was broader than at present. To the S of the Focșani Depression, the Upper Miocene (Sarmatian) base reaches ~3.7 km depth in front of the Carpathians belt and shallows southward (Fig. 8B). The Pliocene–Quaternary sequence has roughly the same thickness as the Upper Miocene one.

3.2. Lateral EET changes and the position of a foredeep depocenter: inferences from modeling along a vertical section

Flexural modeling studies of foreland basins show that the width and depth of the basin are dependent on the rigidity of the lithosphere expressed by its EET and on the shape and weight of the load [22,23]. A load imposed on a continuous plate with constant or very gradually vary-

Fig. 9. (A) Flexural deflection profiles obtained for two different EET distributions using the same topographic load. Constant EET value predicts maximum deflection under the load (dotted line), whereas an abrupt decrease in the foreland EET can induce maximum deflection in the weaker foreland (bold line). (B) Maximum of deflection as a function of the strength of loaded plate (black line) and the position of depocenter relative to the load (gray line). The maximum depth of the foredeep shifts out of the orogen when the difference in EETs becomes significant.
ing EET produces a basin centered underneath the load itself. The basin floor gradually shallows on both sides.

The predicted foredeep position and shape are, however, significantly changed when the loaded plate is formed by two adjacent blocks with very different strengths. The issue is relevant because the Carpathians fold and thrust belt overlies lithospheric blocks with very different characteristics, namely the East European and Moesian. The boundaries between these blocks are often formed by faults (e.g. the Trotuș fault) and are thus rapid.

To illustrate the effect of spatial changes in strength of the loaded plate on basin geometry, we have calculated the deflection of a 2-D thin elastic plate (e.g. [24]) under the weight of a rectangular load for different EET distributions (Fig. 9A,B). In these calculations, we impose that the topography after the lithospheric deflection is 2 km in the orogen and zero everywhere else. In this case, the maximum deflection tends to move away from its central position and shifts towards the weaker plate. The trend can continue and the site of maximum deflection can be detached from the load itself. From this point, the foredeep base becomes shallower towards the orogenic load. The shape of the basin also differs from the one predicted for a constant EET.

3.3. Planform modeling approach

To investigate the mechanism of subsidence in the Focșani Depression, we use a flexural model calculating the deflection corresponding to a laterally variable load. Details on the methodology are to be found in Garcia-Castellanos et al. [25]. The load is calculated from the observed topog-

![Fig. 10. Lithospheric domains used in the planform flexural modeling. Boundaries between them are represented by major faults (details in text).]
raphy in the Carpathians region and the deflection is filled with constant density material.

The model assumes loading of an elastic thin plate, a detailed description of the governing equations can be found e.g. in Van Wees and Cloetingh [26]. The loaded plate is actually formed by various blocks with different constant strength. A transition zone with variable strength exists in each case, linearly interpolated between the bounding blocks. The model geometry and strength/load distribution controls the deflection pattern and the site of maximum predicted flexure.

3.4. Input data for planform modeling

3.4.1. Strength of lithospheric blocks

The Focsani Depression lies onto the northern part of the Eastern Moesian platform which is surrounded by other lithospheric blocks with quite different tectonic histories and therefore, different strengths. We have simplified the structure of the Carpathian domain to 5 major lithospheric domains (Fig. 10): East European (inclusive of Scythian platform and North Dobrogea orogen), Eastern and Western Moesian platforms (the latter includes the South Carpathians as well), East Carpathians and Transylvanian (including the Apuseni Mts. and the easternmost part of the Pannonian basin). Major faults dividing various platform units represent the boundaries between the foreland blocks. The Carpathians are separated from the foreland by the Pericarpathian fault, which is the most external major thrust (e.g. [7]). The boundaries between Transylvanian block and the East and South Carpathians could be represented by a backthrust [9,27] and by a dextral shearing zone (e.g. [2]), respectively.

While the geophysical and main tectonic characteristics of the defined blocks are fairly well known (Table 2), a translation to EET values is far from straightforward (e.g. [30]). In our modeling strategy, we have initially used estimates from previous 2-D flexural studies [1,10,11] and from rheological predictions that took into account the thermo-mechanical age of various blocks [28] and Tertiary deformations in the Carpathians area. A compilation of the EET values obtained by previous works is given in Table 2. Except Royden’s estimate [1], the EET values obtained from flexural studies of the Carpathians/foreland system are much lower (at least half) than those inferred from rheological studies. Subsequently, EET values were modified to attain the best possible fit. Sets of EET were varied in such a way that: (1) strength distribution respects the same initial chronology as in Table 2, and (2) the absolute values are closer to the rheological estimates [28] rather than the flexural ones [11]. Low EET values are unrealistic for some lithospheric domains, such as the East European craton (see Table 2). On the opposite, previous Eastern Moesia estimates (both flexural and rheological) neglected the effects induced by the Badian extension, the presence of major crustal faults (structural heterogeneities, sensu Burov et al. [31]) and the increased curvature (bending stresses [30]) of the subducted subvertical foreland plate (e.g. [3]).

3.4.2. Initial and final topographic loads

The numerical model we have used calculates loads and deflections on a grid composed of 10000 points covering the entire Carpathians region.

We assume the beginning of the Sarmatian as the age for the foredeep base, with 0-m elevation over the entire model. This is justified by: (1) foreland sedimentation was either mainly evaporitic (on East European platform), fluvial (parts of Western Moesian platform) or absent (parts of Western Moesian platform and Eastern Moesian platform); (2) salt and shallow water deposits in Transylvanian basin; and (3) evaporitic and shallow water deposits in the Outer Carpathians.

As far as the load causing subsidence is concerned we have used the present day topography of the area (Fig. 10) with two modifications. We do not consider the topography in the Balkans because the last major deformations are older than Sarmatian [32]. For the same reason (e.g. [6]) the topography of the Apuseni Mts. is assumed as the average for Transylvanian basin (400 m). The load provided by the water from Black Sea is also neglected.

The densities of asthenosphere, load and infill-
ing material are 3200, 2700, and 2200 kg/m$^3$, respectively.

3.5. Results

3.5.1. Best fit model using the topographic load

The best fit model which we could get adopting topography as the only effective load is shown in Fig. 11. This deflection pattern is characterized by strong East European and Western Moesian platforms (EETs of 37 km and 30 km, respectively), intermediate Transylvanian and East Carpathians domains (EETs of 26 km and 29 km, respectively) and weak Eastern Moesian block (EET = 10 km).

Isobaths in Fig. 11A indicate that the deepest point of the basin is located in the inner part of the belt at the junction between Transylvanian block, East and South Carpathians. In front of the East Carpathians the predicted foredeep has maximum depths between 2.3 and 2.9 km that gradually decrease toward East to 1–1.2 km. Close to the Trotus fault, the predicted deflection fits the foredeep base to the W and underestimates it to the outer part (Fig. 11B), possibly due to the influence of the Sarmatian and younger tectonics of this fault. To the central-northern part of the East European platform, the foredeep base at the contact with the orogen has depths ranging between 1 and 3.5 km increasing from NW toward SE along the belt [4].

In front of the eastern part of the South Carpathians (CS 3 in Fig. 11A), the model predicts a shape for the foredeep base similar to the observed one, however, with a shallower depth. Around 1.8 km depth is predicted for the South Carpathians foredeep close to the contact with the Pericarpathian fault; the modeled basin shallows S-ward reaching ~0.3 km. Typically, the Western Moesian platform has a depth of 2–2.6 km in front of the Subcarpathian nappe decreasing in a S–SW direction [33,34]. A good fit between predicted and real deflections is obtained for most of this domain, taking into account that South Carpathians/Western Moesia region had a more complex evolution than the simple thrust loading observed elsewhere: this domain suffered additional Eocene orogen parallel extension [35], Burdigalian dextral transtension and Late Miocene strike-slip [36].

The most interesting result from this study is

### Table 2
Compilation of crustal/lithospheric thicknesses [16–18] and previous EET estimates obtained either from flexural [1,11,29] or rheological modeling [28]

<table>
<thead>
<tr>
<th>Lithospheric domain</th>
<th>Crust (km)</th>
<th>Lithosphere (km)</th>
<th>EET estimates (km)</th>
<th>Comments</th>
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<tr>
<td>E-Europ. (including Scythian and N-Dobr. orogen)</td>
<td>&gt; 40</td>
<td>&gt; 190</td>
<td>14 (plate boundary forces acting) [11]</td>
<td>oldest, coldest and strongest domain</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>28–32/38–79 (wet/dry rheology) [28]</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>40 (subsurface loads) [1]</td>
<td></td>
</tr>
<tr>
<td>Eastern Moesian platform</td>
<td>30–40</td>
<td>150–190</td>
<td>12 (plate boundary forces acting) [11]</td>
<td>weakest domain due to rifting+crustal scale shearing+bending</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>24/39 (wet/dry rheology) [28]</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>&lt;15 (plate boundary forces acting) [11]</td>
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<tr>
<td>Western Moesian platform (including South Carpathians)</td>
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<td>24–72/29–84 (wet/dry rheology) [28]</td>
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<td>East Carpathians</td>
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<td>120–170</td>
<td></td>
<td>the same as E-Europ. domain [11]</td>
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<td>~19/32 (wet/dry rheology) [28]</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>weaker than E-Europ. platform due to the bending+Neogene volcanism</td>
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the presence of a clear local depocenter in the predicted isobaths patterns to the SE of the Bend Zone (Fig. 11A) resulting in a ‘satellite’ basin, located away from the orogenic belt itself. This is due to the presence of a weak foreland block. The position of this satellite basin roughly coincides with that of the Focsani Depression. In a vertical profile, the predicted maximum depth is \( \sim 3.3 \) km, which corresponds to 40\% of the observed depth.

3.5.2. Sensitivity

The use of a forward modeling technique implies that the set of best fitting EET values is not unique. These values must be therefore considered as approximate with an estimate error of 20–40\%, depending on the region [25]. In order to concentrate subsidence out of the orogen, for example, a weak Eastern Moesian platform is clearly required, with an EET value less than 12 km. The EET value of the Western Moesian block, in contrast, is poorly limited between 24 and 35 km because we cannot quantify the changes in strength owing to the different tectonic events mentioned above. The high value for the East European platform is the most reliable because it is consistent with the thickest, oldest and coldest lithosphere (Table 2). In summary, the present topography explains satisfactory the flexural subsidence in the Carpathians foreland except the
Focșani Depression and Eastern Moesia. Although the shape of the flexural basin and its position relative to the belt are well predicted in the Focșani Depression, the modeled depth is less than half of what is observed. Therefore, an additional load should be added to the present topography.

3.5.3. The effect of additional vertical loads

Fig. 12A shows the contour of an additional load able to fit the data in the Bend zone. This load can be expressed in various ways, but we have chosen to apply an additional topographic load to the Bend orogenic region, with the same density distribution as explained before. Two situations give significant modeled results, from the variety of loads applied. The main effect for both scenarios to increase the depth of the satellite basin in front of the Carpathians Bend.

The first case corresponds to 500 m extra topography, the modeled deflection pattern is shown in Fig. 12B. The satellite basin reaches a depth of \(6 \text{ km} \) which accounts for 70% of the observed value, but the deflection pattern represents the best possible average of the basin shape (Fig. 12C), particularly south of the Focșani Depression (Fig. 12C, cross-section 3). The difference observed to the SE edge of cross-section 3 is probably related to the post-Badenian transtension [8,12]. At the southern edge of East European platform (Fig. 12C, cross-section 4) the difference between predicted and real basin is roughly the same as that obtained without an additional load (Fig. 11B).

The second case corresponds to 800 m extra topography, the satellite basin reaching 8 km in depth, accounting for 90% of the observed value. The deflection pattern would produce, however, a basin with a larger wavelength than observed (Fig. 12C).

3.5.4. The effect of intraplate stresses

In a further series of experiments, we have modeled the effects of intraplate stresses on the geometry of the Carpathian foredeep. The direction and type of intraplate stresses approximates the Pliocene to recent ones (e.g. [4] and references therein).

The results are shown in Fig. 12D. In the first case scenario (500 m extra topographic load), as a consequence of adding intraplate stresses the base of the Focșani Depression reaches \(6.3 \text{ km} \) depth which corresponds to 75–80% of the post-Badenian subsidence. In the second case (800 m extra load), intraplate stresses would produce a depocenter with equal depth and location as Focșani basin. However, similarly with the no intraplate stress scenario, the predicted wavelength is larger than observed. Modifying the magnitude of the intraplate stresses induced vertical variations in the order of few hundreds of meters in the depocenter area.

3.5.5. Discussion

Taking into account as novel features both the extensional and the contractional stages of subsidence in the Carpathians Bend foredeep, we can explain \(60\%\) of the basin depth using only the present topography as the load for the flexural period. A complete incorporation of the real planform topography predicts larger subsidence than the previous computations carried out along vertical sections (e.g. [1]). The most important result is that the 3-D distribution of load and lateral variations in lithospheric strength, particularly the presence of a weak domain in front of the Bend region, determines the localization of a basin out of the belt, the modeled shape being in good agreement with the observed data.

The incorporation of an additional vertical load together with the contribution of the compressive
intraplate stresses fit the observed Focșani Depression deflection. The two scenarios can either average the basin deflection and shape, the obtained values being still lower than the observed ones, or can entirely fit the maximum deflection, but not the wavelength. We cannot be sure on the cause of this additional load, but the extra topography might be equivalent to a sinking slab in the Bend region. For instance, the 800 m extra topography corresponds to a 86-km-long oceanic slab attached at the base of the underthrusted continental lithosphere with a density contrast of 25 kg/m$^3$. The surface projection of our additional load is roughly the same as that of the invoked slab [5].

Dealing with pure elastic plates loaded instantaneously, the modeling has considered neither the changes that occurred in time in the lithospheric strength due to thermal events, faulting or bending [30,31], nor those that affected topography. Since topography was diachronously created (the Bend region is $\sim$ 6 Ma younger than the East and South Carpathians), although the main thrusting event was synchronous [9], the present relief in the Bend region apparently reflects only the last contractions (Pliocene–Quaternary), which are minor by comparing with the Sarmatian ones. Therefore, a planform kinematic flexural modeling may better explain the subsidence in the foredeep and also the nature of this additional load. Further incorporation of lithospheric folding mechanisms, as previously suggested [37], can explain better the detailed pattern of the observed deflection.

4. Conclusions

The subsidence of the unusually deep SE Carpathians foredeep basin is mainly controlled by pre-orogenic extension followed by 3-D contractual flexure. NW–SE trending extensional basins are found to the S of Focșani Depression whose eastern margin underwent normal faulting as well. Extensional modeling for these basins reveals a crust–mantle lithosphere uniform thinning of maximum 1.1. The modeling also predicts $<100$ m post-rift subsidence after Middle Miocene (Badenian) in accordance with the regional unconformity observed at the Middle/Upper Miocene boundary.

For the second stage of subsidence (post-Badenian), a flexural model is used, taking into account the plan view distribution of topographic loads and the lateral variations in the lithospheric strength. Assigning realistic strengths to the domains comprising the Carpathians region, a basin is predicted in front of the Carpathians Bend because the weakened foreland (due to extension and crustal active faults) localizes the deflection.

Since the deepest modeled basin is only 3.3 km ($\sim$ 40% of the observed value), an additional load is required in the Carpathians Bend, corresponding to an extra topography in the order of 500–800 m. The latter is compatible with the suggested presence of an oceanic slab remnant at high depths. Depending on the magnitude of intraplate stresses, a basin can develop at the position and with the shape of the Focșani Depression accounting for its entire depth.

An alternative has been provided to models invoking a slab pull force to account for the entire subsidence observed in the SE Carpathians foredeep/foreland, particularly in the Focșani Depression. We demonstrated the importance of the previously neglected extension in the eastern Moesian foreland and that the 3-D distributions of the topography and lithospheric strength plus the action of compressive intraplate stresses can lead to major subsidence in front of the topographic load. The 3-D effect could be particularly important for foredeeps associated with highly arcuate orogenic belts.

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