Lateral variations in mechanical properties of the Romanian external Carpathians: inferences of flexure and gravity modelling

L. Mațenco a,∗, R. Zoetemeijer b, S. Cloetingh b, C. Dinu a

a Bucharest University, Faculty of Geology and Geophysics, 6 Traian Vuia str., 70139 Bucharest, Romania
b Department of Earth Sciences, Vrije Universiteit, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands

Accepted 13 March 1997

Abstract

We present the results of two-dimensional flexure and gravity modelling of the subsidence of the Romanian Carpathian foreland based on twenty profiles through the Southern and Eastern Carpathians. The narrow spacing of the grid of profiles allows us to investigate the lateral tectonic variations transverse to the belt, but also along strike. The topographic elevation of the Romanian Carpathian mountain belt is modest and the minimum in the Bouguer gravity anomaly, characteristic for flexural control of subsidence, is located relatively far towards the foreland in respect to the mountain belt. As the contribution of topographic loading to the evolution of the Romanian Carpathian foreland system is small, a subduction (underplating)-dominated tectonic regime controlled the nappes emplacement and basin shortening in the external flysch and molasse basins during the Late Tertiary. The modelling results support the existence of important variations in effective elastic thicknesses (To) and plate boundary forces. High flexural bending stresses in the western part of the Southern Carpathians can explain low To values, whereas the increase of To to the east can be explained by changes in rheological properties of the Moesian Platform. Field observations indicate that variations in deflection along strike are probably also related to basement irregularities, stress field rotations and strike-slip movement along lateral ramps during the Tertiary. In the Eastern Carpathians the model results indicate a N–S trend of lateral back stepping of the subduction system. A distinct back step in the vicinity of the seismically active Vrancea area supports the involvement of the E–W crustal scale Trotuș fault, which forms the transition from the East European Platform to the Scythian and Moesian Platforms. Development of a possible N–S migration of plate boundary activity or slab detachment and successive intersection with the Trotuș fault can explain the large Pliocene subsidence in the Focșani depression and relative uplift of the East European Platform north of the mentioned fault.

Keywords: Carpathians; flexure modelling; gravity modelling; tectonics

1. Introduction

The Carpathian orogenic belt is a highly arcuate orogen, formed in response to continental collision of the Apulian and Eurasian plates during Cretaceous and Tertiary times (Sândulescu, 1984, 1988; Cson-tos, 1995). The outer Carpathian foreland system shows many similarities with other convergent systems in the Alpine/Mediterranean area, and it is also genetically linked. Therefore, the understanding and description of this foreland system is an important in-
gradient for obtaining better constraints on the overall tectonic evolution of the Alpine/Mediterranean area.

In this paper we present the results of the downflexing of the Eurasian plate (Moldavian and Moesian Platforms) under respectively the Eastern and Southern Romanian Carpathians (Figs. 1 and 2). The Romanian Carpathian foreland system is very interesting, because it contains the present-day seismically active Vrancea area at the transition between Southern and Eastern Carpathians. Whereas a number of studies describe the geological evolution of the Eastern Carpathian foreland zone (e.g., Burchfield, 1976; Sândulescu, 1984, 1988), relatively few publications have focused on the evolution of the Southern Carpathians (Sândulescu, 1984, 1988). The Southern Carpathians and their foredeep (Getic depression) are less studied, mainly because the thrust system is covered by Pliocene–Pleistocene sediments. Using the advent of new data sets based on (partly unpublished) seismic sections, the modelling of the lithospheric downflexing of the Moesian and Moldavian Platforms along a large number of profiles (Figs. 1 and 2) can now be pursued.

Lithospheric flexure and development of foreland basins adjacent to orogenic belts is caused by the response of the lithosphere to loading (Price, 1973). Flexural models (e.g., Beaumont, 1981; Jordan, 1981; Royden, 1993) and gravity models (Lyon Caen and Molnar, 1983; Lillie et al., 1994) show that the flexural behaviour of the lithosphere underlying a foreland fold-and-thrust belt is directly controlled by surface loading of sediments and stacked thrust nappes and plate boundary processes. Furthermore, the rheological properties of the lithosphere, previously inherited and modified during subsequent evolution, can play an important role in the evolution of foreland basins (Zoetemeijer et al., 1990).

In this study we focus on the lateral variations in the mechanical properties and plate boundary processes, represented in the effective elastic thickness ($T_e$), bending moment ($M_0$) and vertical shear force ($V_o$). Earlier studies have revealed that lateral variations of the rheological properties in the direction of convergence may be controlled directly by the flexural process as a result of lithosphere weakening by yielding due to pronounced flexural bending (e.g., Beaumont, 1978; Cloetingh et al., 1982; McNutt, 1984; Burov and Diamon, 1995). Furthermore, flexural analyses of other European foreland basins have demonstrated that lateral variations in $T_e$ may also result from inherited differences in strength, due to the presence of pre-orogenic sutures, as in the Ebro basin, Spain (Zoetemeijer et al., 1990), and Aquitaine basin, France (Desegaulx et al., 1991).

Analysis of these lateral variations in mechanical properties and processes along the Romanian Carpathians obviously requires detailed data sets. Therefore, twenty profiles are investigated, each based on seismic reflection lines (Figs. 1 and 2), supplemented by a detailed Bouguer gravity anomaly map (Fig. 3, modified after Mocanu and Rădulescu, 1994). When comparing model results of the Romanian Eastern and Southern Carpathians, differences in plate boundary forces and differences in $T_e$ are predicted. Not only regionally, but also when comparing the profiles within the Eastern and Southern Carpathians along the belt, lateral changes in $T_e$ and plate boundary forces are determined. These changes on a relatively small scale may result in the generation of lateral ramps and tensional fractures in the upper crust, as a result of brittle failure caused by pronounced down-flexing. Below, we will demonstrate that the model results are supported by recent field data (Maţenco et al., 1997), which have provided evidence for these secondary tectonic effects on the tectonic evolution of the Carpathian foreland basins during the Tertiary. The inferred lateral variations in the plate boundary processes and rheological properties along strike of the mountain belt are subsequently discussed in the context of plate tectonic scenarios such as lateral slab detachment (Wortel et al., 1993) or subduction of passive margin lithosphere (Lillie et al., 1994), in an oblique setting.

---

Fig. 1. Structural sketch of the outer Eastern Carpathians and location of the modelled profiles. Three different tectonic style zones are defined: (1) between the Bistriţa and Trotuș valleys (profiles XIV–XX), (2) between the Trotuș and Buzău valleys (profiles X–XIII), (3) between the Focșani depression and Dâmboviţa (profiles VI–IX). Contour lines of base to Neogene after Dumitrescu and Sândulescu, 1970. Note that the contour lines may not correspond in all the places with the depth-to-basement data used for the flexural modelling.
Fig. 2. Structural sketch of the Getic depression and location of the modelled profiles. Three tectonic style areas are defined: (1) western tectonic style area; (2) eastern tectonic style area; (3) intermediate zone. The zones are divided by the geographical positions of the Olt and Jiu valleys.
2. Tectonic setting and milestones in the evolution of the Romanian Carpathian system

Since variations in mechanical properties of the lithosphere can be largely controlled by inherited sutures as a result of pre-orogenic processes, it is important to have constraints on the evolution of the Romanian Carpathian system over a larger time frame.

The evolution of the Romanian East and South Carpathian foreland system took place in two main distinct events (Sândulescu, 1984, 1988; Matenco et al., 1997): (1) a Cretaceous–Palaeogene period of basin fill, with flysch and in some places molasse type sediments; and (2) a Miocene–Plio–Pleistocene period of compression, differentiated in several deformation phases with syn- and post-sedimentary basin cover.

Previously proposed tectonic models (e.g., Sândulescu, 1984, 1988; Royden, 1993; Csontos, 1995) are not fully compatible with the deformation mechanisms for both described events. Because the modelling presented further in the paper supports some of these mechanisms, a brief description of the tectonic events affecting the Romanian Carpathians is given in the next section, separately for the Eastern Carpathian foreland and for the Getic depression (Southern Carpathian foreland). Apart from the thin-skinned belt, the rheological properties and the crustal architecture of the underthrust platforms have significant influences on the flexural mechanism. A short review of previous researches will be further discussed.

2.1. Eastern Carpathian foreland

The study area is bounded by the Bistriţa valley in the north and the Dâmboviţa valley in the south
The undeformed foreland is represented by the Moldavian (southwestern East European) Platform from the Bistrița valley to the Trotuș valley and by the Moesian/Scythian Platform from the Trotuș valley southward.

The structure of the Moldavian Platform has been revealed by geophysical soundings. Deep reflection profiles show rough thicknesses of 10, 20 and 40 km for the base of sedimentary cover, Conrad and Moho (Răileanu et al., 1994) discontinuities, while seismological data show an average crust thickness of 43 km (Enescu et al., 1988, 1992). Magnetotelluric data (Visarion et al., 1988) show an important decrease in crust thickness, to roughly 35 km, west of the NW–SE-trending Solca fault (close to the frontal nappe contact). This feature is related to the Tornquist–Teissere lineament, whose NW–SE trend, hidden by the thrust nappe pile, has been recently revealed (e.g., Botezatu and Calotă, 1983; Guterch et al., 1986; Pinna et al., 1991). The sedimentary cover of the Moldavian Platform has thicknesses of around 6–12 km (Răileanu et al., 1994) and seems to decrease towards the east. The internal structure is characterised by three major sedimentation cycles, separated by major unconformities (Mutihac, 1990): Palaeozoic (Silurian–Carboniferous), Mesozoic (Upper Permian–Triassic), Tertiary (Eocene–Pliocene). Thickness distribution and detailed facies for the first two periods are still to be worked out. The Tertiary cover has thicknesses of around 2–6 km nearby the thrust front contact and slightly decreases towards the foreland, excepting the Focșani depression, where roughly 9 km of Neogene deposits were recorded (Figs. 1 and 4).

South of the Trotuș valley the autochthonous foreland is represented by the Moesian/Scythian Platform (Fig. 1), which extends farther into the undeformed foreland of the South Carpathians (Fig. 2). Deep refraction seismic profiles show crustal thicknesses of around 35–40 km (Rădulescu, 1988), while seismological data show an average value of 34 km (Enescu et al., 1992). The Vrancea (south-east bend) area has anomalous values of 40–47 km (Rădulescu et al., 1976; Cornea et al., 1981; Răileanu et al., 1994), while seismological data show an average of 44 km (Enescu et al., 1992).

The sedimentary cover has large thickness variations along the belt. The extreme value is placed in the Vrancea area, where values of 18 km were recorded (Cornea et al., 1981). Four sedimentation cycles were observed in the Moesian Platform (Mutihac, 1990): Palaeozoic (Cambrian–Westphalian), Palaeozoic–Mesozoic (Upper Permian–Triassic), Mesozoic (Middle Jurassic–Cretaceous) and Tertiary (Badenian–Pliocene). The Tertiary cover has thicknesses of around 2–6 km nearby the thrust front contact and slightly decreases towards the foreland, excepting the Focșăni depression, where roughly 9 km of Neogene deposits were recorded (Figs. 1 and 4).

Apart from the autochthonous foredeep, the external flysch and molasse nappes of the foreland fold-and-thrust belt of the outer East Carpathians form also part of the area of investigation. This area contains the external flysch and molasse nappes of the foreland thrust belt (Audia, Tarcău, Marginal, Subcarpathian) and the inner part of the Moldavian Platform. The tectonic evolution of this area has been the subject of many recent papers (Sândulescu, 1984, 1988, 1994; Royden and Baldi, 1988; Ellouz et al., 1994).

The evolution of the flysch basin starts in Early Cretaceous–Palaeogene time with deposition of large amounts of flysch type sediments on continental crust (Sândulescu, 1994). During the Miocene, the area was involved in contractional deformation (Royden, 1988; Sândulescu, 1988; Csontos, 1995), which propagated successively towards the foreland and deformed the flysch deposits. This deformation continued up to Pliocene–Pleistocene time in the Vrancea bending area (southeast corner of the Carpathians).

The Eastern Carpathian foreland can be subdivided into three zones with different particularities (Fig. 1). In the northern zone, between the Bistrița and Trotuș valleys, the thrust system consists of low-angle thrust faults (Antonescu et al., 1993). The basal detachments form staircase-trajectories along which large amounts of shortening took place. The structural shortening is about 100 km according to Burchfield (1976), about 70–100 km according to Ellouz et al. (1994), and 30–50% of the total width of pre-contraction sedimentary deposits according to Antonescu et al. (1993). The deformation of the hinterland dipping duplexes is mainly of Middle–Late Miocene age.
In the southern zone of the Eastern Carpathians, between the Buzău and Dâmboviţa valleys, the structural style is dominated by high-angle thrust faults, and low shortening with a minimum of 10-20% of the total width of pre-contraction sedimentary deposits (Antonescu et al., 1993). The frontal nappe system is always buried by Late Miocene–Pliocene–Pleistocene sediments.

The intermediate zone, between the Trotuș and Buzău valleys, forms the transition between the northern and southern zone. The structural shortening is high along nappe systems thrusting over the Moesian/Scythian Platform. Although classic interpretations have argued for a basal detachment unconformably buried under the Upper Miocene to Pleistocene sediments (Sândulescu, 1984, 1988), new interpretations (Matenco et al., 1997) suggest that in late stages (Late Miocene) shortening along the main detachment could be transferred back towards the hinterland along a backthrust in a frontal triangle zone (Figs. 1 and 4). Further to the east, in the foreland of the thrust system, the Focşani depression is located. This important depression contains more than 9 km of Plio–Pleistocene sediments (Fig. 1). The key problem for understanding the evolution of the southeastern Carpathian bending area is to explain the mechanism for this major subsidence and the tectonic evolution of the western flank of the Focşani depression.

2.2. Getic depression

The contact zone between the nappe pile (basement and sedimentary cover) which forms the South Carpathians and the Moesian Platform is the foredeep belt of the Getic depression (Fig. 2). In the north, this area is bounded by the Tertiary transgression limit, where post-deformational sediments cover the structures of the inner belt. In the south the area is delimited by the surface projection of the buried Miocene detachment front, the so-called 'Pericarpathian line'. The eastern limit of the foredeep belt is formed by the northern extension of the Intramoesian fault, which separates the Getic depression from the thin-skinned belt of the outer East Carpathians. The Danube River forms the western limit.

2.2.1. Uppermost Cretaceous–Palaeogene

The geological history of the Getic basin begins in the uppermost Cretaceous–Palaeogene through the deposition of a clastic sequence representing a marginal facies. This facies can not be called molasse nor flysch, because the deposition of the sedimentological sequences are synchronous with the flysch deposition in the Eastern Carpathians, but have molasse aspects in the Southern Carpathians (Jipa, 1980, 1982, 1984). Palaeogene deposits are transgressively deposited over the uppermost Creta-
ceous sediments. They are approximately 2000 m in thickness, but have important strike variability. The uppermost Cretaceous–Palaeogene sequence indicates an active subsidence, promoting the opening of the Getic basin.

Different hypotheses have been forwarded explaining the cause of the active subsidence. Classic theories (e.g., Sândulescu, 1984, 1988; Ştefănescu and Polonic, 1993) propose a contractional tectonic evolution of the Southern Carpathian belt which could explain the Getic basin as a pure flexural basin. The Moesian plate (southern passive margin of the Severin ocean) is separated into two main pieces: the Danubian nappe system, presently with crystalline pre-Mesozoic basement and Mesozoic sediments exposed at surface, and the Moesian Platform unit, which underthrusts the Danubian nappes. Mid- to Late Cretaceous deformations further develop the Getic basin in front of the Danubian nappes, a basin which is filled with a post-Laramic tectonic molasse (Dacide molasse), covering also the inner South Carpathians (Supragetic, Getic, Severin and Danubian nappes).

Ratschbacher et al. (1993) pointed out that the tectonic history may be not pure compressional, assuming that the Moesian Platform played an important corner effect in the Carpathian orogeny. They suggest a tangential stretching component along the northern margin of Moesia during late-Early to Late Cretaceous. This stretching in Moesia could explain the active subsidence started in Late Cretaceous time, but then must have prolonged to the Eocene, because of the great thicknesses of detritic material at that time.

3. Flexural modelling

Here we investigate the flexural response of downflexed lithosphere assuming a thin elastic plate overlying a viscous mantle (Hetényi, 1946). The flexural behaviour is controlled by loading as a result of sedimentation and nappe stacking and by plate boundary processes. For a detailed description of the flexural model we refer to earlier work (Zoetemeijer et al., 1990; Zoetemeijer, 1993). In addition to data of Zoetemeijer (1993), the model can take into account density contrasts within the topographic load and tectonic implications of pre-orogenic palaeobathymetry, such as crustal thinning and its contribution to the gravity field (Lillie et al., 1994) along the modelled profiles.

3.1. Flexural modelling results of the Southern Carpathian foredeep

The basement flexure in the studied area is reflecting the differential evolution of the foredeep between the main eastern and western zones, with different structural styles. The shape of the Bouguer anomaly along the southern foredeep shows the largest density contrast for the Romanian Carpathians. The large anomaly is caused by the relative young thick foredeep sediments, by the elastic bending of the downgoing Moesian plate combined with the crustal root/basement uplift observed in the inner South Carpathians. The Bouguer gravity anomalies have a
lateral minimum around −135 mGal in the transition zone between the Jiu and Olt valleys, and decrease towards the east and west into the two main tectonic zones (Fig. 6).

The modelling is constrained by detailed seismic sequence stratigraphy (Răbăgia and Fülop, 1994), kinematic reconstruction of deformation and palaeostress markers (Mațenco et al., 1997). The above-described Bouguer anomaly signal and kinematic markers suggest that, for the transition area, the loading mechanism is more complex than simple lithospheric down-flexing under frontal nappes emplacement. The basement data are compiled from seismic lines calibrated by deep wells in the external thrust belt. Inferred palaeobathymetric profiles were calibrated by seismic and surface facies interpretation, of which the positions are reconstructed using balanced cross-sections (Morariu et al., 1992).

Modelling results from the profiles in the western tectonic area (sect. I, II, III — Fig. 5) show a narrow and deep basin. The location of the plate boundary is in the vicinity of the surface contact with the basement units. To the east, the bending moments are increasing from 3.5 to 4.0×10^{16} N, whereas the vertical shear forces are relatively constant (0.8–1.0×10^{12} N/m). The southwestern bending corner is critically marked by the low value of effective elastic thickness ($T_e = 7.5$ km), which moderately increases to the east; profile II: $T_e = 8.0$ km and profile III: $T_e = 9$ km. These low values for the effective elastic thickness are characteristic for the areas under the deepest parts of the basin (Fig. 5c).

The eastern tectonic area of the Getic depression (profiles IV and V — Fig. 6) is very different from the western part. The basins are wider and shallower. This implies that the lithosphere compensates loading over a larger area and, therefore, larger effective elastic thicknesses are required. Whereas profile IV forms the transition zone through the eastern tectonic area, where $T_e$ is 10 km, the effective elastic thickness of profile V has increased to 15 km. Yet, the very clear differences in basin shape between the profiles in Fig. 5b and Fig. 6b are not so pronounced in the gravity signal (Fig. 5a and Fig. 6a). We propose a pre-orogenic palaeobathymetry for profiles IV and V whose gravity effects overprint the expected

---

**Fig. 5.** Flexure and gravity modelling results for profiles I–III through the western part of the Getic depression. For location of profiles see Figs. 1 and 3. X-axis indicates relative distance to predicted model plate boundary. For position of plate boundary see Fig. 12. Note the steep NNW-dipping basement of profile I.
smooth gravity signal. The boundary conditions for both profiles are similar \( M_o = 4.2 \times 10^{16} \) N and \( V_o = 0.8 \times 10^{12} \) N/m).

3.2. Flexural modelling results of the Eastern Carpathian foredeep

The profiles in the Eastern Carpathians are located between the Bistrița valley in the north and the Dâmbovița valley at the border of the Getic depression (Fig. 1). The estimates for the depth to basement are derived from the interpretation of the external flysch nappes by Antonescu et al. (1993). Similar to the southern foredeep profiles, the palaeobathymetry is based on simulated values obtained from the flexural modelling by trial and error. The results are in agreement with interpretations of surface facies and their reconstructed position using balanced cross-sections (Antonescu et al., 1993).

Although the minimum in the Bouguer gravity anomaly decreases slightly from the Southern Carpathians to the southern end of the Eastern Carpathians, the lateral continuity of the minimum is smooth and follows in the bending area the limit of the Tarcău and Marginal nappes (Fig. 3). Profiles VI, VII, VIII and IX are crossing this area and overlap very well in projection. In Fig. 7 the flexural modelling results for \( T_e = 12 \) km, \( M_o = 2.0 \times 10^{16} \) N and \( V_o = 1.2 \times 10^{12} \) N/m are given. The relative importance of the bending moment is shown in the comparison with the modelling result obtained for a zero bending moment (Fig. 7).

Further to the north the gravity pattern is less straightforward. Complications occur for the area covered by profiles XI, XII and XIII. The gravity minimum shows a narrowing relative to profile X (Figs. 3 and 4). The western flank of the gravity minimum coincides with the location of the seismically active Vrancea area. The general trend of the eastern flanks of the gravity anomalies in this area, however, can be compared very well with those of profiles IX and X. We have projected profiles IX, X, XI, XII and XIII in such a way that the eastern flanks of the gravity signal overlap (Fig. 8). Basement profiles of profile IX and X also overlap, but data from deep parts of the Focșani depression are lacking and,
therefore, we are not able to constrain the depth to basement for the profiles XI, XII and XIII. In spite of these limitations the projections of the profiles give a reasonable expression of both depth to basement and gravity signal. Flexural modelling predicts values of \( T_e = 12 \) km, \( M_o = 0 \) N and \( V_o = 1.2 \times 10^{12} \) N/m. The relative importance of the vertical shear force is shown in the comparison with the modelling result obtained for a zero vertical shear force (Fig. 8). Obviously these data are not sufficient to make any predictions for lateral changes of \( T_e \) in this area.

Similar uncertainties arise for the area further to the north. Whereas south of the Trotuș valley (Fig. 1), the minimum in the gravity anomaly coincides with the location of the young anomalous deep Pliocene–Pleistocene Focșani depression, north of the Trotuș valley the gravity minimum appears to jump to the west and follows the trace of the Audia nappe, or Tarcău nappe (Fig. 1). This jump in the Bouguer gravity signal in the vicinity of the Trotuș valley can be explained by sinistral movement along the important basement discontinuity, the Trotuș fault. However, the role of the Trotuș fault is not clear, since no large-scale horizontal displacement is traced in the overlying sedimentary record.

Since it is obvious that the gravity signal is disturbed by processes other than flexure, the gravity anomalies are not suitable for constraining the flexural modelling in this area for profiles XVI and XV. Nevertheless, the deflection data are fairly consistent with a flexural model with \( T_e = 14 \) km, \( M_o = 0 \) N and \( V_o = 0.9 \times 10^{12} \) N/m. Furthermore, the predicted gravity signal fits well the projection of profile XIII (Fig. 9). However, in that case the relative position of plate boundaries is shifted 10 km to the west relative to the model results presented in Fig. 8 which also fit profile XIII. This implies that also flexural modelling predicts a jump similar to the one observed in the gravity signal. According to the modelling a possible basement fault should have its trace somewhere between profiles XIII and XIV.

When comparing the eastern flank of the gravity minimum of the profiles further to the north (profiles XVI, XVII and XVIII in Fig. 10, and profiles XIX and XX in Fig. 11) with those of the profiles in the southern Eastern Carpathians (Fig. 7), the northern
profiles do not show the smooth curvature typical for foreland basins, but show much more irregular profiles. The effect can partly be explained by a lateral increase in effective elastic thickness from low values, close to the plate boundary, to larger values farther to the foreland as suggested for the profiles XVI, XVII and XVIII (Fig. 10) and XIX and XX (Fig. 11) (see for quantification the figure captions). Otherwise, because of the projected representation of the Bouguer gravity profiles, a trend is visible in the gravity signal north of the Trotuș fault. Relative to the predicted gravity anomalies, a limited area of increased gravity anomaly is recognizable (located between −20 and 20 km relative to the predicted plate boundary in Fig. 8a). The locally increased gravity anomaly migrates to the east (located between 20 and 80 km relative to the predicted plate boundary in Fig. 10a and located between 80 and 140 km relative to the predicted plate boundary in Fig. 11a). The amplitude of the feature decreases while the width increases to the north (Fig. 10a and Fig. 11a).

In the Eastern Carpathians the predicted plate boundaries generally follow the trace of the overall Bouguer gravity minimum, including the jump in vicinity of the Trotuș fault. $T_e$ values are in the range of 12–14 km. Bending moments ($M_o$) may be negligible, in contrast with vertical shear forces ($V_o$) which are most pronounced in the Vrancea area and decrease to the north from 1.2 to $0.7 \times 10^{12}$ N/m. Additionally, an anomalous second-order effect on the gravity signal, probably caused by processes other than flexure, is recognized north of the Trotuș valley. This gravity feature coincides with the seismically active Vrancea area and trends to the northeast. The combined modelling of flexure and gravity allowed us to trace this feature, which is difficult to recognize directly from the Bouguer gravity map (Fig. 3). Further analysis of the gravity anomalies and additional acquisition of deep basin data is necessary to better constrain the detailed quantitative analyses of the lateral changes in $T_e$ and plate boundary processes in the Eastern Carpathians.

Fig. 8. Flexure and gravity modelling results for profiles IX–XIII through the intermediate zone of the Eastern Carpathians. For location of profiles see Figs. 2 and 3. X-axis indicates relative distance to predicted model plate boundary. For position of plate boundary see Fig. 12. Note the effect of the incorporated vertical shear force ($V_o$ in N/m) and residual positive Bouguer gravity anomaly for profiles XI–XIII located between −20 and 20 km relative to the predicted plate boundary.
Fig. 9. Flexure and gravity modelling results for profiles XIII–XV through the transition zone of the Eastern Carpathians. For location of profiles see Figs. 2 and 3. X-axis indicates relative distance to predicted model plate boundary. For position of plate boundary see Fig. 12. Note the necessity of 10 km shift of projection profile XIII.

Fig. 10. Flexure and gravity modelling results for profiles XVI–XVIII through the northern zone of the Eastern Carpathians. For location of profiles see Figs. 2 and 3. X-axis indicates relative distance to predicted model plate boundary. For position of plate boundary see Fig. 12. Note the residual positive Bouguer gravity anomaly located between 20 and 80 km relative to the predicted plate boundary.
Fig. 11. Flexure and gravity modelling results for profiles XIX and XX through the northern zone of the Eastern Carpathians. For location of profiles see Figs. 2 and 3. X-axis indicates relative distance to predicted model plate boundary. For position of plate boundary see Fig. 12. Note the effect of lateral variation in $T_e$. Model results show response to $T_e$ of 12 km and variable $T_e$: 8 km at inner plate boundary and increasing to 20 km in the foreland.

4. Southern Carpathian foreland

4.1. Kinematic implications for Moesian Platform underthrusting

Earlier studies of the Mio–Pliocene evolution of the Getic basin (Sândulescu, 1988; Ştefănescu et al., 1988) have shown that the South Carpathians is a roughly south-vergent thrust system, with associated, secondary transpressional structures. Mațenco et al. (1997) show that the belt can be divided into two main tectonic areas, with distinct tectonic evolution during the mentioned time span: west of the Jiu valley, and east of the Olt valley, with a transition area between the two tectonic zones. Furthermore, the deformation changed direction in time. At least two main deformation directions are distinguished during the Miocene (ENE–WSW contraction and NW–SE to N–S strike-slip).

First, deformation starts in late Burdigalian time (Mațenco et al., 1997), with an ENE–WSW thrusting and SW-ward depocentre migration. In the west, the stratigraphic wedge is scooped out over the Moesian Platform along thrust faults with common basal detachment. West of the Jiu valley, the main deformation (thrusting over the Moesian Platform) is ending in late Burdigalian time. Later deformation is observed only on a second-order scale, by reactivations of thrusts, by folding and initiation of strike-slip structures. The eastern tectonic area east of the Olt valley is characterized by large-scale thrusting in the inner part and active subsidence in the outer part.

Second, deformation continues in the Late Miocene (intra-Sarmatian), when large-scale thrusting over the Moesian Platform took place east of the Olt valley. Further to the west, the displacement is gradually transferred into the inner part of Southern Carpathians through a series of strike-slip duplexes, placed between the Jiu and Olt valleys (Rábáigia and Fülop, 1994). This may generate a dextral rotation of the inner part along the almost rigid (at this time) part of the Getic depression (the western tectonic unit). This result is consistent with inferences from
previous studies (e.g., Ratschbacher et al., 1993; Mateńco et al., 1997). The deformation is migrating further to the east in Pliocene time with local thrust reactivations, like folding and small-scale uplift of the Moesian Platform.

We propose that the Moesian Platform in the north is flanked by thinned continental crust, probably resulting from latest Cretaceous–Palaeogene stretching. It can be traced along the foreland belt by ridges of uplifted Cretaceous–Palaeogene deposits, roughly in the middle of the foredeep, with a WSW–ENE strike (e.g., Stălpeni–Lăunele uplift, Fig. 2). These ridges can be explained by the inversion of the thick Late Cretaceous–Palaeogene syn- and post-extensional sediments and thrusting upon the margin of the Moesian Platform. Pliocene–Pleistocene sediments are covering the described features (Fig. 12).

4.2. Rheological interpretation of $T_e$ variations

A striking result of the flexural model is the retreated position of the model plate boundaries along all South Carpathian profiles (Fig. 13). When placing the trace of plate boundaries on a surface projection, it resembles the northern extension of the basin (the ‘Tertiary transgression limit’). This suggests that the Moesian plate hardly underthrusts the Southern Carpathian belt, which in fact, supports the idea of a South Carpathian foreland coupling during the compressional deformations. Late Cretaceous thrusting (or oblique convergence) can break down the Moesian plate into two main components: Danubian ‘autochthonous’ thrusting over the Moesian Platform, on top of which the flexural basin of the Getic depression formed.

Furthermore, all the profiles are characterized by a topographic load that is insufficient to explain the lithosphere flexural response, the process being almost entirely controlled by subsurface ‘loads’ acting on the underthrust plate. Even though there are indications for an early marine environment — and, therefore, an extra loading effect of sediments replacing the water column may be expected (Stockmal et al., 1986) — sediment loading is insufficient to explain the observed deflection. Plate boundary forces are necessary and can also explain the deep gravity minimum along the Southern Carpathians. As pointed out before for the Eastern Carpathians (Roden, 1993), the southern sector of the Carpathian belt can probably be included in the retreated subduction boundary type of orogen, with the subduction rate greater than the convergence rate. The plate boundary forces (vertical shear force, $V_o$, and bend-
ing moment, \(M_o\) may be associated with continental crust underthrusting.

Especially for the westernmost profiles of the Southern Carpathians and northernmost profiles of the Eastern Carpathians the models infer a rheological weak flexurally down-bent crust. Generally, weak crust generates narrow and deep basins (Angevine et al., 1990) with low flexural wavelength. Such low wavelengths can favour the creation of tensional down-flexing faults, similar to those observed in the Alpine foreland basin, during the Oligocene–Lower Miocene phase of rapid subsidence of the molasse basin (Ziegler, 1990). Evidence for similar tensional phenomena is described by Matenco et al. (1997). A number of antithetic normal faults, parallel with strike and dipping to the south, are outlined in the inner western part of the Getic basin. The offset of the described normal faults is decreasing towards the east, correlated with the increasing of \(T_e\) along strike, from 7.5 km at the western bending zone to 15 km in the eastern area. The low \(T_e\) value in the Moesian corner can be explained by flexural bending stresses but also by the high values of stretching associated with strike-slip in this area during latest Cretaceous–early Burdigalian time (Matenco et al., 1997).

The effects of changes in \(T_e\) values can be observed in the difference in basin shape in the two tectonic areas (west of the Jiu valley and east of the Olt valley): first, shortening of the stratigraphic wedge along the transition zone between the Jiu and Olt valleys; second, step-type shape of the basement along strike in the basin. The flexural response of
the two tectonic areas results in a different curvature of the foreland, in the west closer to the mountain belt than in the east. As shown in Fig. 13 the transition takes place between profile IV and V west of the Olt valley. Matenco et al. (1997) suggest that the associated lateral ramp can represent the margin of a pull apart basin, which opened in latest Cretaceous–Eocene time. Inversion of this ramp fault during nappe plate loading at Sarmatian time (Late Miocene) would result in strike-slip structures due to the advancing along this lateral ramp. This movement, however, is difficult to differentiate in the field or in a shallow seismic analysis from the possible dextral motion of the Carpathians in respect to the Moesian Platform. Whether the Miocene thrusting could be favoured by a stretched continental crust formed in the latest Upper Cretaceous–Palaeogene should be clarified by further deep seismic studies and extensional modelling.

5. Eastern Carpathian foreland

5.1. Rheological interpretation of $T_e$ variations

Similar to the Southern Carpathians, the Eastern Carpathian flexural modelling predicts low $T_e$ values (14 km). The modest topographic elevation implies a subduction (underplating)-dominated tectonic regime. The Bouguer gravity minimum, which is located relatively far to the foreland provides indications for a retreated position of the boundary conditions in respect to the orogenic belt. Also comparable to the Southern Carpathians, the low $T_e$ values can cause, because of its low down flexing radius, sets of tensional faults, which can actually be observed along the down-going Moldavian (Eastern European) Platform (Dicea, 1995). As a consequence, deformation concentrating in the faulted zones may further weaken the lithosphere. Flexural analyses of profiles through the Ukrainian Carpathians (Zoetemeijer et al., 1996) farther to the north, give indications for a situation close to complete yielding of the lithosphere. On the other hand, inversion of some of the tensional faults during late collisional events, Sarmatian (Late Miocene) or Late Pliocene, could explain basement penetration in the nappe structures as suggested by Ellouz et al. (1994).

5.2. Back stepping of the East Carpathian subduction system and the role of the Trotuş fault

In the Eastern Carpathians uncertainties in basement profiles and the irregular gravity signature do not deter us from drawing some conclusions about the tectonic development of this area. On the contrary, the projected representation of the data and model results allows us to determine the general trend in subsidence and predict plate boundaries to follow the trace of the overall Bouguer gravity minimum (Fig. 13).

The model results support a general back-stepping of the subduction system with its major step in the vicinity of the Trotuş fault between profiles XIII and XIV (Fig. 13). This crustal fault forms the transition between the East European Platform to the Scythian and Moesian Platforms. It influences the emplacement of the overthrusting Miocene nappes, and seems to be reactivated in the Sarmatian (Late Miocene) with an apparent sinistral component. Diffuse sinistral movement can explain the opening of secondary, small-scale basins in the overthrust belt, like the Comăneşti basin (Ciulavu et al., 1997; Matenco et al., 1997).

Model results predict a sinistral offset of about 10 km along the Trotuş fault. On the basis of the jump in the gravity signal, the offset of this fault would be greater. However, the combined modelling of gravity and flexure has revealed an anomalous second-order positive effect on the gravity anomaly (Figs. 9–11) which also partly explains the lateral jump. It is striking that the SW–NE trend of this feature coincides with the SW–NE trend in the spatial distribution of intermediate seismicity in the Vrancea area (Oncescu, 1984).

Model results predict a decreasing effect of the plate boundary force ($V_o$) to the north, which implies an increasing importance of plate boundary processes to the south. We suggest that the back-stepping of the subduction system can be attributed to a lateral migration of plate boundary activity, possibly by slab detachment (Wortel et al., 1993), from the Moldavian Platform (northern East Carpathian foreland) to the Moesian/Scythian Platform (southern East Carpathian foreland). Intersection with the Trotuş fault in the transition area may imply a complete uncoupling of the East European plate from
the slab fragment (Fig. 14) and consequently lead to differential uplift in the northern East Carpathian foreland. At the same time this uncoupling would result in an acceleration of subsidence in the southern East Carpathian foreland (Focşani depression), because the slab fragment would now only be carried by the Scythian/Moesian plate.

Although indications for denudation north of the Trotuş fault and accumulation of large amounts of Pliocene–Pleistocene sediments in the Focşani depression are obvious, validation of this hypothesis is necessary by detailed seismological investigations of the three-dimensional structure of the crust and lithosphere in the Vrancea area. Furthermore, more detailed stratigraphic and geomorphological data are necessary to provide the fine resolution needed to constrain detailed subsidence- and uplift scenarios for the tectonic evolution of the outer Eastern Carpathians.

**Acknowledgements**

The project was sponsored by Shell Romania Exploration B.V. and Vrije Universiteit, Amsterdam. Peter Ziegler is thanked for fruitful discussion and bringing new ideas into the modelling. Special thanks are addressed to Prospectiuni S.A. and to dr. Oprea Dicea for helpful comments and permanent support. NSG publication nr. 970162.

**References**


Wortel, M.J.R., Spakman, W., Yoshioka, S., 1993. Slab detachment and non-stationary processes in subduction zones: evi-