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The interplay between eustacy, tectonics and surface processes during the growth of a fault-related structure as derived from sequence stratigraphy: The Govora–Ocnele Mari antiform, South Carpathians

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ABSTRACT

Orogenic collision represents a moment when typical rules of thin-skinned development are at odds with the generalized interaction between the upper and buoyant lower plate. Deriving the kinematics of these collisional features is usually difficult because of the large amounts of subsequent erosion. Study areas such as those that exist along the South Carpathians foredeep, where the syn- and post-collisional sediments are still preserved, are important places to identify, record and quantify out-of-sequence deformation. Here, sequence stratigraphic techniques may be used in order to detect tectonic movements which are an order of magnitude higher resolution than those routinely identified by either standard structural analysis or isotope geochronology. This study proves that what has been previously interpreted as a one stage growing anticline situated at the contact with the core of the orogen is, in fact, a gradually evolving antiform structure with at least four quantifiable pulses of vertical movements associated with various amounts of orogenic uplift. During collision, the interplay between the orogenic core and the anticline uplift controls the depositional area, the effects of sea-level variations being subordinated. Transversal shear zones, such as strike-slip faults, are used as transport corridors by the progressive infill of large quantities of sediments from the source area towards the larger basins situated more to the foreland.

1. Introduction

During continental collision, out of sequence deformation is widespread across an orogenic belt. The simple foreland-breaking sequence of deformation characterizing the accretion of a sedimentary wedge during the subduction stage can be interrupted either by backward vergent deformation, syn-orogenic extension or back-thrusting/thickening near the orogenic backstop (e.g., Suppe, 1985; Platt and Vissers, 1989; Willett and Brandon, 2002; Smit et al., 2003). The timing and kinematics of these events may be described by analysis of the post-tectonic covers, particularly those in the frontal part of orogens. However, these post-tectonic covers may be affected by subsequent phases of deformation and can therefore be pre- or syn-tectonic sediments in relationship to these subsequent tectonic events. Numerical and analogue modeling indicates that overthickening during the collision is accommodated through significant deformation in the internal part of orogenic wedges (e.g., Beaumont et al., 1991; Sokoutis et al., 2005). The "soft"-collisional orogens (simply defined as being dominated by subduction, Royden and Burchfiel, 1989) generate reduced amounts of vertical movements, often below the resolution of low-temperature isotopic markers (U–Th/He and cosmogenic nuclides, e.g., Dunai, 2000; von Blanckenburg, 2006). Studying third order sedimentological base level variations (e.g., van Wagooner et al., 1990; Schlager, 1993) may provide a more detailed and precise method of deriving the kinematic history. In basins filled with gently dipping strata even small amounts of tectonic uplift can create regional tectonic unconformities (Nystuen, 1998). This type of geometry is frequently observed in the foredeep areas of low topography orogens that experienced significant subsidence in the post-collisional stages, such as the Apennines/Southern Alps (e.g., Bertotti et al., 2001) or the Carpathians (e.g., Matenco et al., 2007). Therefore, the use of sequence stratigraphy techniques to derive additional tectonically-induced depositional changes appears to be a relevant method.

The sedimentary architecture of foreland and thrust-sheet top (or "piggy-back") basins is closely controlled by tectonics due to flexural-derived subsidence (e.g., Puigdefàbregas et al., 1992), derived from the interplay of flexurally-triggered subsidence, tectonically-driven uplift and eustacy. During emplacement of thrust-sheets along low-angle decollements, the accommodation space is reduced and facies

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changes towards prograding fluvio-deltaic systems, potentially associated with cyclicity and basin axis shifts, while the sediment supply is spilled over, foreland-wards (e.g., Nijman, 1998; Clevis et al., 2004a). The controls studied so far are the balance between transversal influx of sediments and longitudinal redistribution towards basinal areas. In the latter, syn-tectonic (i.e. at the scale of the orogen) eustatic sea-level fluctuations result in retrograding and prograding fan and delta sediments bounded by flooding surfaces and type two sequence boundaries with frequent incised and axial channels (Clevis et al., 2004b). Changes in sedimentary environments as a record of eustatic signals on increased tectonic background during rapid collisional uplift and out-of-sequence deformations are still widely unknown. This type of deformation generates emerging anticlines in the proximity of the source area, creating frontal foredeep basins and thrust-sheet top basins with absent or reduced longitudinal redistribution. When collision is oblique, the latter can be replaced by transversal feeding of excess sediments throughout strike-slip fault corridors across uplifted areas.

The syn-kinematic sedimentation of a growing anticlinal fold is a function of its position with respect with the basin margin. Important parameters are the emergence of the anticlinal crest (e.g., Doglioni and Prosser, 1997), its potential in terms of source area and the geometry of growing anticlines, which may be transported in allochtonous positions (Giunta and Nigro, 1999; Vitale, 1998, see also Salvini and Storti, 2002). Apart from some tectonic geomorphology studies (e.g., Delcaillau, 2001; Garcia and Herail, 2005), much less is known on the interaction between the main sedimentary sources of compressional basins (i.e., the orogenic core) and growing anticlines situated in their proximity, which is often complicated by the subsequent erosion of syn-tectonic sediments.

One area which satisfies conditions required to study these processes is the partially buried Govora–Ocnele Mari antiform, which is located in the foreland of the South Carpathians orogen in a location where pre- and syn-tectonic sediments of the Getic Depression are still exposed (Figs. 1 and 2). Field sedimentological and structural mapping techniques have been combined with seismic stratigraphy calibrated by well information, in order to study the complex interactions between the uplifting Carpathians source area, local depositional environments, sea level changes and deformation along local structures.

2. The late stage evolution of the South Carpathians foreland

The contact between the South Carpathians and their lower Moesian unit is a foredeep deformed during the Miocene, locally known as the "Getic Depression" (e.g., Motaş, 1983; Motaş and Tomescu, 1983, Fig. 1b). Its sediments were deposited during the latest Cretaceous (Upper Campanian)–Late Miocene (Late Sarmatian) times as a post-tectonic cover relative to the Cretaceous phase of thick-skinned thrust-sheets stacking in the South Carpathians (e.g., Răbăgia and Matenco, 1999; Fig. 1b). It contains up to 5 km of mainly siliciclastic sediments deposited in a basin, which is less than 100 km wide and experienced poly-phase deformation during the Miocene (e.g., Dicea, 1996). These sediments were subsequently covered by essentially undeformed Uppermost Miocene (Upper Sarmatian) to Quaternary sediments which extend southwards overlying the Moesian foreland (e.g., Jipa, 1997; Figs. 1b and 3).

During the Cretaceous closure of the Ceahlău–Severin Ocean in the South Carpathians area, the foreland was coupled (sensu Ziegler et al., 2002) and part of Moesia was incorporated into the orogen along a crustal-scale duplex, known as the Danubian antiformal stack (e.g., Berza et al., 1983; Iancu et al., 2005; Fig. 1). Subsequently, the South Carpathians underwent large-scale clockwise rotations (~90°) around the Moesian foreland, presumably driven by the roll-back and subduction of the distal parts of the European margins (e.g., Wortel and Spakman, 2000; Schmid et al., 2008). Paleogene thrusting and transpression of the Balkans (e.g., Ivanov, 1988; Doglioni et al., 1996) was coeval and was subsequently followed by the latest Cretaceous–Eocene orogen-parallel extension and ~100 km dextral movements along curved faults systems (e.g., Fügenschuh and Schmid, 2005). Among the latter, the Early Miocene Timok fault accounts for 65 km of Early Miocene dextral offset (Krautner and Krstič, 2003). Its NE-ward prolongation is presently buried below Middle Miocene–Pliocene sediments and truncated by subsequent Middle–Late Miocene thrusts and strike-slip faults (e.g., Tărăpoancă et al., 2007; Fig. 1a). Seismic profiles suggest that the Timok fault is splaying into a transtensional fault system, with Up to 2.5 km normal offsets (Răbăgia and Matenco, 1999). Further to the east, the entire Palaeogene–Early Miocene rotation and dextral offsets are transferred to thrusting in the East Carpathians (Fügenschuh and Schmid, 2005). During the Middle and Late Miocene, the transtensional basin was obliquely inverted; the South Carpathians docked and were thrust southwards over the Moesian Platform (e.g., Dicea, 1996; Fig. 1b). Deformation was coeval with dextral translations of the South Carpathians with respect to Moesia (Matenco and Schmid, 1999).

Following the general trend of Paratethys basins separation from the larger Tethys realm (e.g., Senes, 1973), the final late Miocene orogenic uplift of the Carpathians has separated an Eastern Paratethys realm east of the orogenic chain (e.g., Rögl, 1999). This mostly semi-isolated brackish to fresh water realm has a separate endemic biostratigraphy (e.g., Papaianopol et al., 1995; Stoica et al., 2007; Vasiliev et al., 2005; Fig. 3). In the Carpathians foreland, this post-tectonic cover is part of the so-called "Dacian basin" (Jipa, 1997), spatially juxtaposed over a larger area including the Getic Depression.

2.1. Sediments of the Getic Depression

The Upper Campanian–Upper Miocene sediments of the Getic Depression are mostly siliciclastic with small intercalations of tuffs, salt and limestones. An Upper Campanian–Oligocene coarse-grained clastic molasse (Săndulescu, 1988) was deposited over the external South Carpathians units and Mesozoic/Paleozoic series of the Moesian platform (Dicea, 1996; Fig. 1b). Upper Campanian to Maastrictian coarse-grained clastics near the northern border of the basin (1–1.5 km, Szasz, 1975) are overlain by an up to 5 km of a Paleogene trangressive succession (Jipa, 1984), which contrasts with widespread erosion observed elsewhere (e.g., Paraschiv, 1997).

During the Miocene, two sedimentary cycles are documented in the Getic Depression, Lower–Middle Miocene and Upper Miocene (Paraschiv, 1979). The first one starts with Lower Burdigalian conglomerates which are less than 2 km thick and decrease in thickness northwards to 700 m (Moldovan, 1954). These were initially deposited in sub-aerial environments in elongated basins along coeval normal faults (Răbăgia and Matenco, 1999). Gradually they are fining upward to an Upper Burdigalian marine post-tectonic cover of 300–700m in thickness and are unconformably overlain by 200 m continental–lacustrine lower Badenian sediments (e.g., Motaş, 1983), changing southwards to a littoral/lagoonal facies (Răbăgia and Matenco 1999; Fig. 3). The overlying Badenian sediments contain two regional levels known at the scale of the entire Carpathians. The first one is an up to 100 m thick acidic tuff (see Seghedi and Szakacs, 1991), which is overlain by up to 200 m thick salt, coeval with the Middle Miocene Paratethys salinity crisis (Peryt, 2006; Fig. 3). Both sequences are overlain by a transgressive deep marine facies with variable thickness (Fig. 3). The Middle to Upper Miocene sediments (Lower–Middle Sarmatian) were deposited coeval with the main southward thrusting of the Getic Depression and are siliciclastic, continental to shallow marine (Dicea, 1996). During the Miocene, five depositional sequences were interpreted at the scale of the entire basin (three eustatic and two tectonic in origin) with local base-level variations observed near the main thrusts, usually in small piggy-back basins (Răbăgia and Matenco, 1999).