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Coexistence of thin- and thick-skinned tectonics in Zakynthos area (western Greece): Insights from seismic sections and regional seismicity

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ABSTRACT

The structural style of the fold-and-thrust belt in the western part of the Hellenic foreland in an area offshore the Zakynthos Island is assessed on the basis of two lithoseismic profiles, earthquake data and field data in the broader area. The compressional deformation in the area is due to the late phases of subduction and collision between Aegean and African plates. Shortening is affecting a ~6–7 km thick sedimentary succession of Mesozoic carbonate rocks and evaporites, as well as Miocene to Pleistocene foredeep clastic deposits. Structures accommodating shortening in the Pre-Apulian and Ionian zones comprise a blind imbricate thrust system and fault-propagation folds that formed over a 10–11 km deep detachment. The thrust system comprises an array of forward-verging thrust faults, which sole out in a low-angle hinterland dipping décollement, separating the Mesozoic cover from the pre-Mesozoic basement. In this tectonic context, the Triassic evaporitic layers seem to play an important role in the vertical partitioning of the deformation style acting as either detachment horizons or lining the imbricate thrust surfaces. The concentration of recent seismic activity at a depth between 6 and 16 km implies that the pre-Mesozoic basement is also involved in the deformation in the form of upthrust basement slices, suggesting a combination of thick and thin-skinned deformation across this domain of the External Hellenides fold-and-thrust belt. The frontal part of the belt shows higher seismic activity along a low-angle blind thrust, as well as also in the areas where the individual thrust planes branch at depth to the basal décollement.

1. Introduction

Two major mechanisms, generally termed as *thin-skinned* and *thick-skinned* tectonic models, respectively, are commonly invoked to explain the formation of thrust fault systems in collisional settings (Coward, 1983; Pfiffner, 2006). The main concept of thin-skinned model is that the sedimentary cover is detached from the underlying undeformed basement along gently dipping thrust faults with ramp-flat trajectories (Butler, 1982; Chapple, 1978), while the thick-skinned model relies on the hypothesis that that basement is involved in the deformation with crustal-scale thrust faults which steepen at depth (Coward, 1983). Over the last three decades, thin-skinned tectonics have been considered more appropriate for the development of foreland fold-and-thrust belts but an increasing number of studies have argued that thin-skinned deformation coupled with hybrid or well-developed thick-skinned structures are also common in many natural fold–thrust complexes (Butler et al., 2004; Coward et al., 1999; McDowell, 1997; Mouthereau et al., 2007; Paton et al., 2006; Scisciani and Montefalcone, 2006). Uncertainties in the application of a particular model between these two end-members are compounded by the presence of pre-existing normal faults that play an important role in the style of compressional deformation through structural inversion. Additionally, examination of these issues requires information of the activity of the structures rather than simply on finite geometry alone.

The Hellenic foreland in western Greece, consisting of the Ionian and Pre-Apulian geotectonic areas, is commonly interpreted as a thin-skinned in style fold-and-thrust belt (e.g., Underhill, 1989), although synchronous thin-skinned and basement deformation has also been proposed for the evolution of the Ionian zone in NW Greece (Doutsos et al., 2006). Generally, thin-skinned deformation (Boyer and Elliott, 1982; Coward, 1983) has also been suggested for the evolution of the External Hellenides, including Gavrovo zone and Pindos thrust belt (Jenkins, 1972; Skourlis and Doutsos, 2003; Sotiropoulos et al., 2003; Xypolias and Doutsos, 2000). A critical feature of the thin-skinned model for the Hellenic foreland is a shallow gently hinterland dipping décollement horizon composed of rheologically weak Early–Middle Triassic evaporites, defining the base of the Mesozoic–Tertiary carbonate cover of the Ionian zone, as is indicated by structural mapping, deep boreholes, and some 2D seismic lines (Kamberis et al., 2000; Kokinou et al., 2005; Sotiropoulos et al., 2003; Underhill, 1988, 1989).

Although the area of western Greece has been considered for decades as a target area for petroleum exploration, there is still much ambiguity in the structure of the deep undrilled sections of the crust. Therefore, seismic reflection images are an excellent and valuable tool

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to study the subsurface structure. Regional seismicity is also important providing information for the style of deformation of the basement units. The Ionian Islands are traditionally recognized as a high seismicity region. In particular, Zakynthos Island is one of the most seismically active areas in the Eastern Mediterranean area. Devastating strong earthquakes over the recent past are quite common in the seismic history of the Zakynthos over the last three millennia including earthquake magnitudes up to 7.2 at short return periods (Hatzidimitriou et al., 1985; Papazachos and Papazachou, 1997).

In this paper, we present two newly released oil industry seismic reflection profiles, from the area south of Zakynthos Island (Ionian Sea), in order to describe the thrust system geometry and its relation to the recent seismic activity and especially during the April 2006 seismic sequence that occurred in the study area. The interpreted seismic lines enabled us to depict with greater clarity and in greater detail the sedimentary crustal section in this part of the fold-and-thrust belt. Particular attention is also given to the Mesozoic pre-existing structures, and present day syn-collisional structures in order to draw conclusions on the deformation style best applicable along this sector of the northwesternmost Hellenic Arc.

2. Tectonic framework

Western Greece is a geologically complex area since it comprises many tectonic events and records the initiation, development and final destruction of the southeastern margin of the Neo-Tethys Ocean. Mesozoic facies units recognized in the Hellenides orogenic belt deposited on a series of platforms (Pre-Apulian and Gavrovo zones) and deep basins (Ionian and Pindos zones) that formed the eastern rifted margin of the Apulian plate, bordering towards the east the Pindos Ocean (Papanikolaou, 2009; Robertson et al., 1991; Smith, 1977). These units, named as the External Hellenides, were developed during Tertiary times following the closure of the Pindos Ocean and the consequent continent–continent collision between the Apulian and Pelagonia microcontinents to the east (Doutsos et al., 1993, 2006).

A major structural feature of the Hellenides orogen in Zakynthos and surrounding areas is a crustal-scale thrust fault, called the Ionian Thrust, which brings the Ionian over the Pre-Apulian zone (Fig. 1a, b). Both zones consist of Mesozoic carbonates that passed into Tertiary Flysch in the Ionian zone and Miocene marls, sands and clays in the Pre-Apulian zone, respectively. The lithostratigraphy of the Ionian zone includes primarily a succession of syn- to post-rift Mesozoic limestones overlying pre-rift Early–Middle Triassic limestones and evaporites. Rift-related sedimentation is thought to be related to the opening of Neo-Tethyan oceanic strands (Kokinou et al., 2005; Pomoni-Papaioannou et al., 2004). The mean thickness of the carbonate sequence is slightly over 2 km, while the minimum evaporites thickness is 1.5 km. Data for the thickness of the evaporite are quite crucial for the interpretation of the upper crust deformation, but are scarce since none of the wells in western Greece has penetrated their entire thickness. Field observations from evaporite dissolution-collapse breccias, in close proximity with thrust faults and/or folds between the Pre-Apulian and Ionian zones, emphasizes the relationship between contractional structures and evaporites (Karakitsios and Pomoni-Papaioannou, 1998; Pomoni-Papaioannou et al., 2004; Underhill, 1988; Zelilidis et al., 1998).

Initial rifting and extension in the Apulian plate gave way to inversion tectonics of the Mesozoic passive continental margin, reactivating pre-existing pre-Alpine normal faults, as is the case of the Gavrovo–Arakynthos thrust system (Doutsos et al., 2006; Sotiropoulos et al., 2003) (Fig. 1). The thrust nucleation history and propagation history during this inversion, which were strongly influenced by the normal faults formed in the fore-bulge region of the Ionian–Gavrovo foreland basin, lasted until the Late Miocene, and based on litho-stratigraphic analysis of Ionian zone flysch sediments in the Aitoloakarnania area (Sotiropoulos et al., 2003). A series of ENE-dipping thrust faults and NNW-trending anticlines, that were formed from Upper Miocene to Pliocene–Quaternary, constitute the structural grain of the Zakynthos Island (Kamberis et al., 1996; Underhill, 1989). The structural control is important on the shelf-sedimentation on the flanks of anticlines and the evolution of Pliocene–Quaternary basins of the Zakynthos Island and their surroundings (Zelilidis et al., 1998). The geometries of thrust faults appear to be listric with a high angle portion close to the earth's surface and shallow dip at depth (Fig. 1c). The sedimentary sequence of Early Miocene to Quaternary rocks, which extends over most of the area, attains significant thickness of about 6 km at NNE-trending roughly elliptical depocenters (Brooks and Ferentinos, 1984; Kamberis et al., 2000; Monopolis and Bruneton, 1982). The Neogene basins' deformation has been complicated by the superposition of deformation associated with diapiric movements of Triassic evaporites which are active till present times (Kamberis et al., 2000; Zelilidis et al., 1998).

At present western Greece is characterized by ongoing subduction and mountain belt formation and consequent high seismicity. Its geological setting is fairly complex, with along strike changes in the subduction zone geometry and dynamics, comprising continent–continent collision in the north and ocean–continent subduction in the south (Doutsos et al., 2006; Hatzfeld et al., 1988, 1995; Hirn et al., 1996; Sachpazi et al., 2000; Underhill, 1989; Vassilakis et al., 2011). The change between continent–continent and ocean–continent subduction occurs at the Kefallonia transform fault (KTF; Fig. 1a; Kokkalas et al., 2006), which is characterized by a dextral strike–slip sense of shear at ~14 mm/yr (Louvari et al., 1999; Vassilakis et al., 2011). Focal mechanism solutions that exhibit reverse-faulting with a NW–SE strike have been reported for a number of moderate to strong-sized earthquake events located under the Ionian Islands (Sachpazi et al., 2000; Serpetsidaki et al., 2009). Some of the reverse faulting events are placed just above the plate interface, implying internal deformation in the overriding plate. Field observations made after the Lefkada earthquake (14/08/2003) of Mw = 6.2 (Benetatos et al., 2005), which occurred at a depth of 15 km, provided also indications of recent reactivation along an onshore NNE–SSW-striking thrust fault that can be considered as a strand of the Ionian thrust, with sinistral sense of motion (Kokkalas et al., 2003). East of the Ionian Islands, all types of focal mechanisms characterize the seismicity at various depths; however, the seismicity resumes in the Greek mainland with primarily normal and strike–slip focal mechanisms (Sachpazi et al., 2000).

3. Description and interpretation of the lithoseismic profiles

The presented NE–SW trending lithoseismic lines Z-141 and Z-207 are located south of Zakynthos Island and have a length of 23.5 and 36 km, respectively (Figs. 1b and 2). For the designation of shot points (SP) in both seismic lines, Syledis was used as a primary navigation system and satellite navigation as a secondary one (Fig. 2; inset map). The processing of the first line (Z-141) was performed by CGG (Compagnie General de Géophysique-France) in 1978. The seismic source was Vaporchoc, which fires a bubble of heated steam at 7 m depth under a shot point interval of 25 m. The data recorded with a sampling rate of 4 ms and a frequency range of 0.125–62.5 Hz. For the deconvolution, the operator length used was 240 ms. Velocity analysis was performed and the final product was of a 4800% stack (for data acquisition and processing sequence see Fig. 3). The processing of the second line (Z-207) was carried out by WGC (Western Geophysical Company of America) in 1980. The seismic source was Maxipulse, using a dynamite source at 10.7 m depth, with shot point interval of 25 m. The data recorded with a sampling rate of 4 ms and a frequency range of 6–87.5 Hz. Deconvolution was performed using a 248 ms operator length (two windows of 2.2 s each). Velocity analysis was also carried out and the final product was of a 4800% stack too (for data acquisition and processing sequence see Fig. 4). The interpreted seismic lines are shown in Figs. 3 and 4, while the original seismic lines are shown in Fig. 2.

Analysis of seismic sections allowed us to distinguish through the study area seven main litho-seismic sequences respectively labeled