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The backstop between the Mediterranean Ridge and western Peloponnese, Greece: its crust and tectonization. An active seismic experiment with ocean bottom seismographs

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ABSTRACT Five wide aperture seismic lines were recorded with ocean bottom seismographs offshore western Peloponnese, between the Island of Zakynthos and the Gulf of Messinia. Four lines were ENE-WSW oriented, and one N-S crossing three of them. Two of the ENE-WSW lines extended from western Peloponnese to the Mediterranean Ridge, from the continental domain of western Greece to the oceanic of the Ionian Sea. Seismic energy was generated by a 48 l (~ 2976 in^3) air gun array tuned at low frequencies. Data were compiled as Common Station Gathers (CStGs) and were evaluated by first break tomography and layer tomography combined with two-point ray tracing forward modelling. The Vp-velocity models obtained by this procedure were further used to depth migrate the CStGs. Line length of the profiles varied between 60 and 180 km, with seismic stations spaced between 2.5 and 5.5 km inline, and shots fired at 125 m. The crust from western Peloponnese to the Mediterranean Ridge is continental of variable thickness, while west of the backstop the crust is of oceanic origin, 5.5 to 6 km thick. South Zakynthos has Moho depth of nearly 28 km, thinning in the Kyparissiakos basin to 22 km, and south of Messinia Peninsula to 21 km. At the collision front between the backstop and the Ionian oceanic crust at the continent/ocean transition of the Mediterranean Ridge, the subducted oceanic crust was also mapped, adding 5.5 to 6 km to the crustal thickness. Sediments in most parts of the profiles range from 6 to 8 km in the Alpine domain and in the backstop between 4 and 6 km, having maximum thickness at the collision front of the Mediterranean Ridge. Vp velocities of soft sediments above the high velocity Alpine limestones have values of 1.8, 2.9, 3.8 and 4.8 km/s, while the Alpine limestones range from 5.6 to 5.8 km/s, and the pre-Apulia (Paxos) limestones and sediments, below the Alpine napes, have Vp wave speeds of 6.0 to 6.2 km/s. The two limestones are separated by Triassic evaporites with Vp ~ 4.2 km/s. Alpine limestones do not exist in the backstop domain, which is exclusively built by pre-Apulia limestones and sediments, overlying a stretched and thinned continental crust. Faults mapped in the Kyparissiakos basin are the continuation of onshore structures of western Peloponnese. We could identify the westernmost limit of the Ionian zone and the extend of the pre-Apulia backstop that widens from north to south, from 20 to 40 km west of Zakynthos and the Strophades Island to more than 80 km SW of Messinia Peninsula. Deformation of the sedimentary sequences at the collision front of the Mediterranean Ridge, between the Ionian oceanic crust and the

thin continental crust at the backstop, is very intense, and the backstop, particularly at its western side, is fractured by several major faults.

Key words: crustal structure, Hellenic backstop, Mediterranean Ridge, western Hellenides, SEAHELLARC.

1. Introduction

The western Hellenides located between the Ionian Sea oceanic domain and continental Europe, are tectonically one of the most active areas of the Mediterranean region. They are deforming by compression at the collision front between Europe and Africa (see Fig. 1), building an active margin where crustal shortening is progressing at a speed of more than 3,5 to 4 cm/y (i.e., Kahle *et al.*, 1998; Reilinger *et al.*, 2010). High seismicity with events of up to 7.0 M_S have occurred in historical times, destroying the coastal areas by seismic tremors and tsunamis (i.e., Galanopoulos, 1960; Ambraseys, 1962; Makropoulos and Burton, 1981; Papazachos and Papazachou, 1997). Consequently, understanding crustal deformation between western Peloponnese and the Mediterranean Ridge is essential for developing protection models and improving the existing building codes. This requires a clear picture of the tectonic processes, which in detail are still unclear.

Mapping the crust and sediments was to date limited to very few widely spaced locations between Crete and southern Peloponnese (Weigel, 1974; Makris, 1977; De Voogt *et al.*, 1992; Truffert *et al.*, 1993; Hirn *et al.*, 1996; Kamberis *et al.*, 1996; Kokinou *et al.*, 2005; Makris and Papoulia, 2009). Data obtained in few positions by two ships expanding experiments (DeVoogt *et al.*, 1992) produced good quality records that permitted the development of 1D velocity models only because their evaluation is based on the assumption of lateral homogeneity along the seismic profile. Older data, like those presented by Weigel (1974) or Makris (1977) are also of good quality due to the large volume of explosives used to generate the seismic signals. They

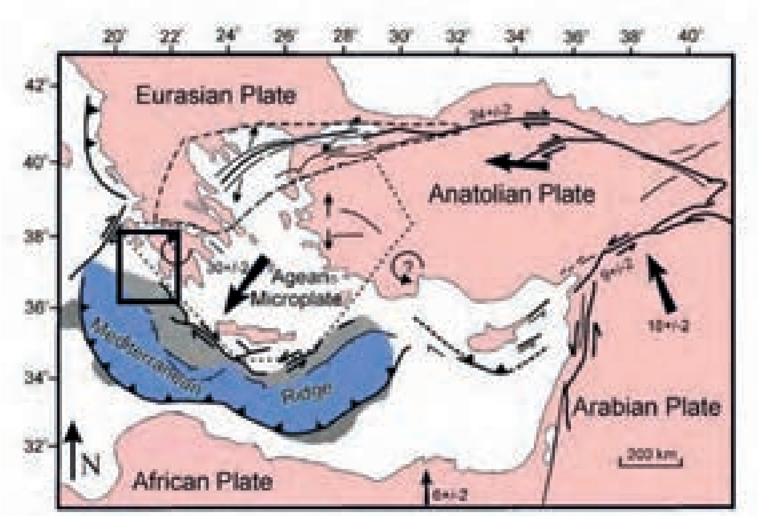

Fig. 1 - Simplified tectonic model of the Hellenides and the eastern Mediterranean Sea. The SEAHELLARC project area is located in the black frame.

were recorded however by only a few anchored buoys and the velocity models developed from these data are laterally smoothed. Finally, MCS streamer data, like those published by Hirn *et al.* (1996) or Kokinou *et al.* (2005) (STREAMERS project), were collected using a 6 km streamer. This streamer length is too short to permit reliable recordings for velocity modelling from the deeper parts of the seismic sections. Thus mapping deep reflections and interpreting their significance is highly ambiguous.

The MCS data collected on behalf of the Public Petroleum Corporation (PCC) of Greece in the 1980s in the Ionian Sea and south of Crete with a 2.6 km streamer and a small air gun volume failed to penetrate the sedimentary basins below the Triassic evaporites (Monopolis and Bruneton, 1982). Also the MCS regional seismic study of the Mediterranean Sea conducted by OGS from 1969 to 1974 using a 2.4 km streamer (Finetti and Morelli, 1973) failed to produce information on crustal structure or the deep sedimentary basins of the Mediterranean Ridge and the backstop. Finetti (1982) presented a regional interpretation of the central Mediterranean basin and the Mediterranean Ridge using the MCS profiles mentioned above. His interpretation of the crust and its nature however is speculative. The streamer used was too short and the energy source too small to produce signals able to map crustal thickness and reliably resolve the velocity structure.

In order to understand the nature of the crust offshore western Peloponnese and the backstop and delineate the sediments and tectonic development of the westward thrusted Alpine napes, we performed a seismic survey using densely spaced ocean bottom seismographs (OBS) and a large volume air gun source tuned at low frequencies. In the following, we present results of this active seismic experiment and discuss them in relation to the geology and tectonics of western Greece.

2. The OBS active seismic experiment

In the framework of the SEAHELLARC project (Papoulia *et al.*, 2014a), we observed five 2D OBS seismic lines (Fig. 2) using 20 to 40 OBSs per line, spaced between 2.5 to 5.5 km, except for Line P1_2006, where only 7 OBS positions were deployed. Shooting inline was generated by a 2960 in^3 air gun source with 51 bar•metre strength. Shots were fired every 125 m and the 8-gun array was tuned at low frequencies in order to obtain maximum seismic penetration from long offset observations. Dominant frequency of the shots was 8 Hz. Data were evaluated by applying first break tomography, layer tomography combined with two point ray tracing forward modelling (kinematic and dynamic) and depth migration. The evaluation procedure mentioned above is state-of-the-art, and is described in Ditmar and Makris (1996) and Pilipenko and Makris (1999). In the following, we will present the results obtained from the active seismic experiment and we will not elaborate on the methodology.

3. The N-S profile: Line P1_2006

This profile, striking parallel to the main structures, has a length of 120 km and was evaluated using 7 OBS positions. Velocity modelling obtained by the procedure described above was constrained from results obtained at the four ENE-WSW oriented lines that intersect this N-S oriented profile (see Fig. 2).

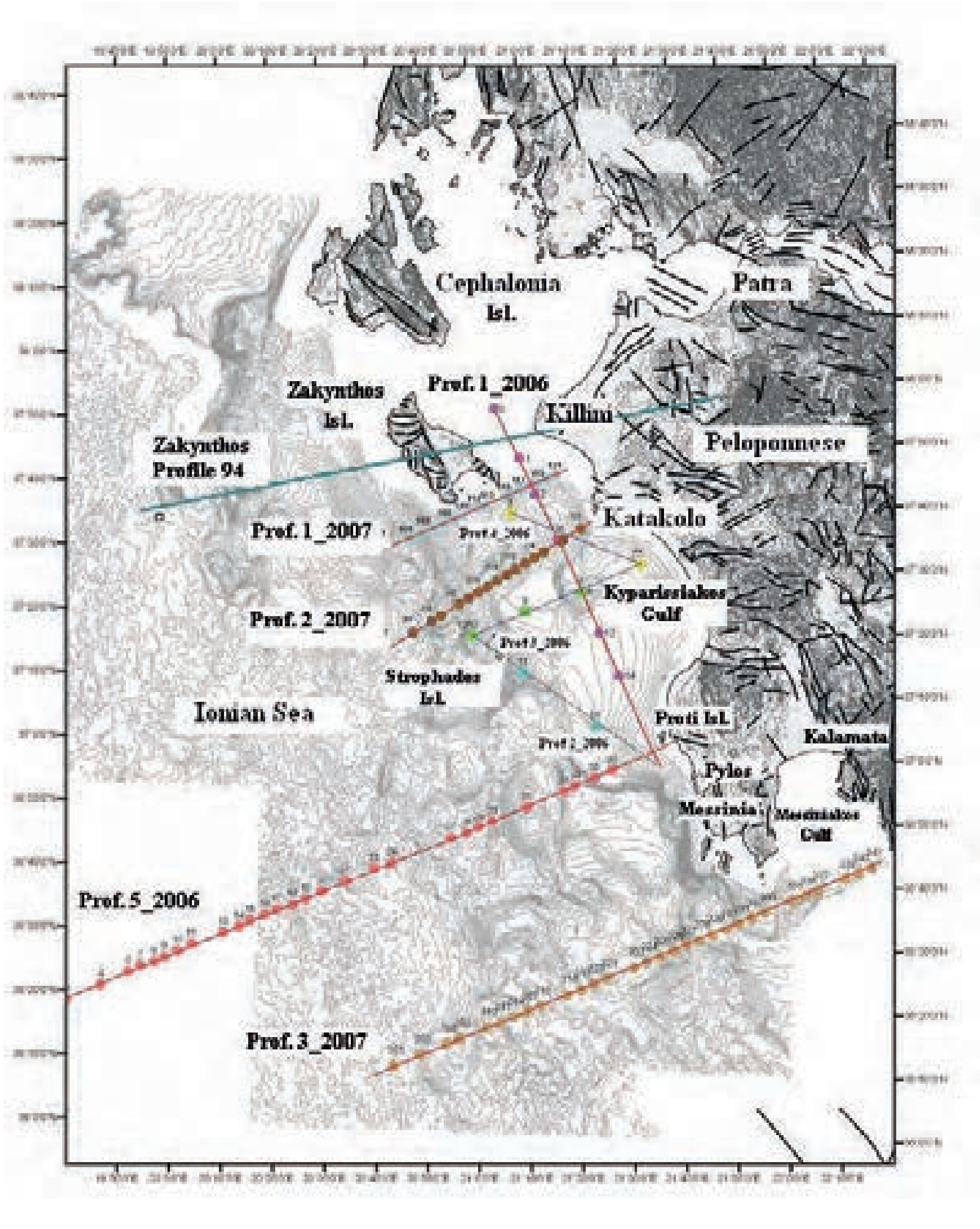

Fig. 2 - Location of seismic lines and OBS positions offshore western Peloponnese. The blue line shows the seismic profile published by Makris and Papoulia (2009) recorded by OBSs and land seismic stations.

A Common Station Gather (CStG) of position 6 is presented in Fig. 3. Seismic energy propagated efficiently and mapped the thickness of the continental crust by PmP reflections, as well as the basin geometry and velocity structure.

In Fig. 4 we present the velocity gradient model developed by first break tomography using first arrivals from all 7 OBS positions. This velocity model shows that the Kyparissiakos Gulf is separated in three sub-basins. These are limited to the north and south by two major fault

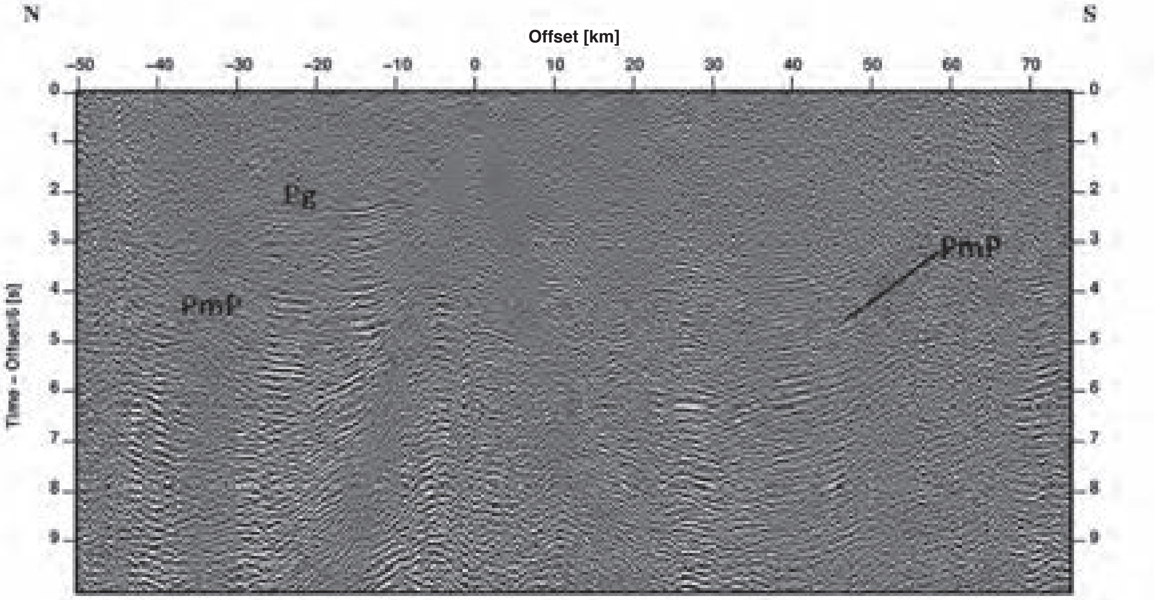

Fig. 3 - This CStG for OBS position 6, Line 1_2006 vertical geophone, is plotted with a linear move out using 6 km/s as reduction velocity. All CStGs in this paper are treated in the same way. PmP reflections from the crust/mantle boundary (Moho discontinuity) and Pg arrivals from waves propagating in the upper crust are indicated.

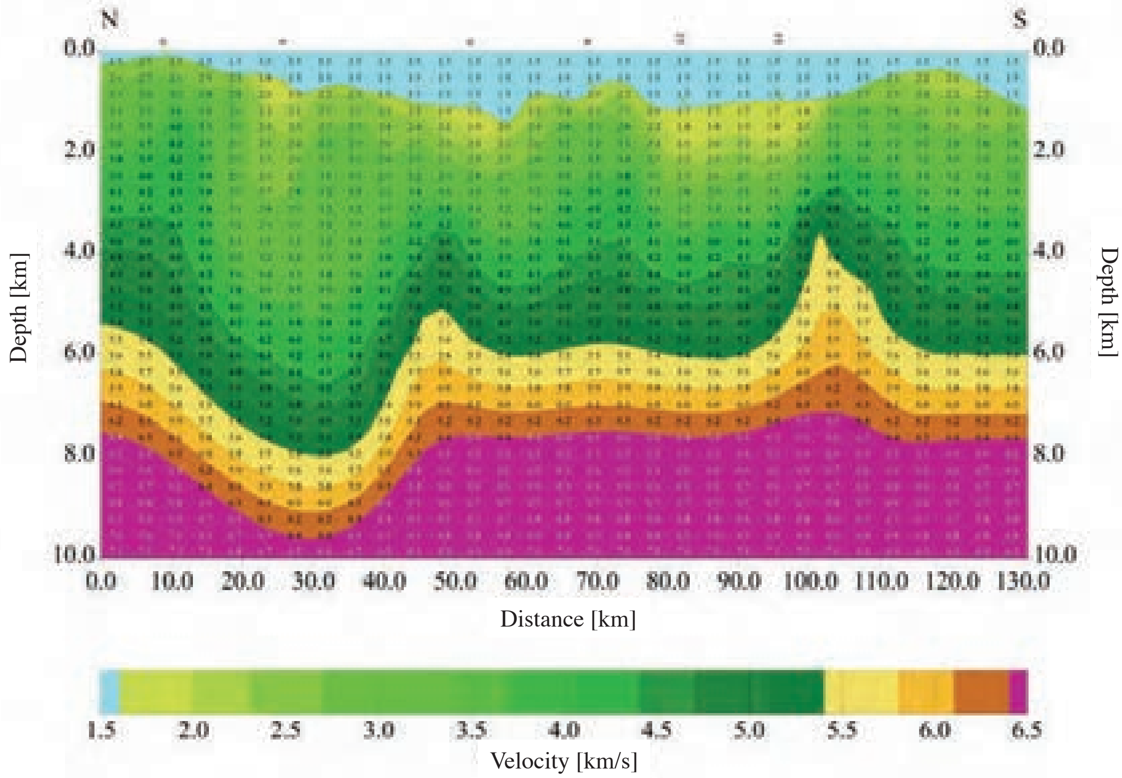

Fig. 4 - First break tomography from arrivals picked at 6 OBS positions shown on top of the figure, Line P1_2006 Velocity model is limited to about 8 km depth, in order to enhance the basin geometry. Numbers in the figure show velocity values. Result was obtained after 9 iterations, and the RMS value is 0.02544 s.

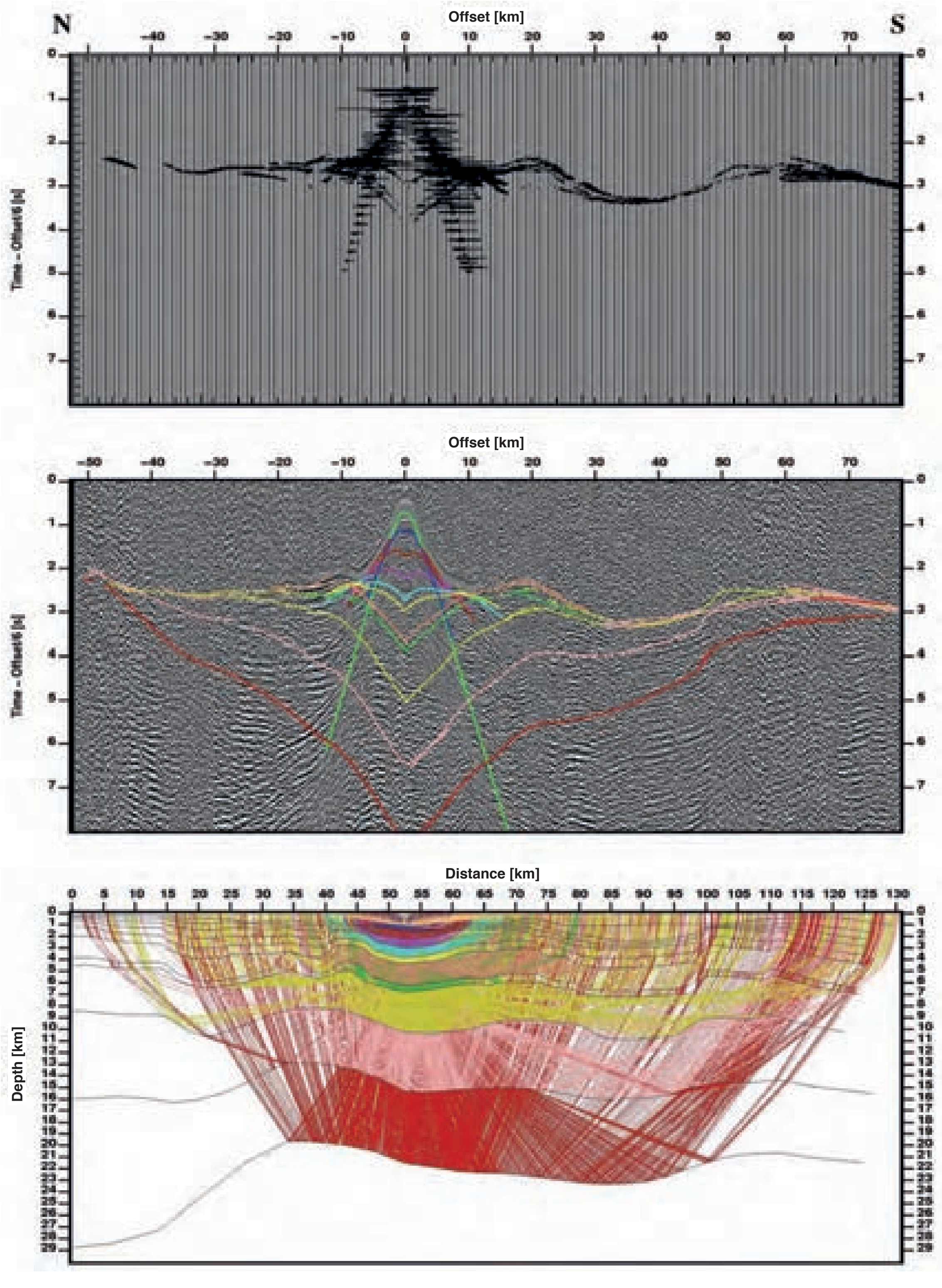

Fig. 5 - CStG of OBS position 6, Line P1_2006, vertical component with colour coded travel times, and corresponding travel paths. The upper part of the figure shows the synthetic amplitudes computed along the travel time paths.

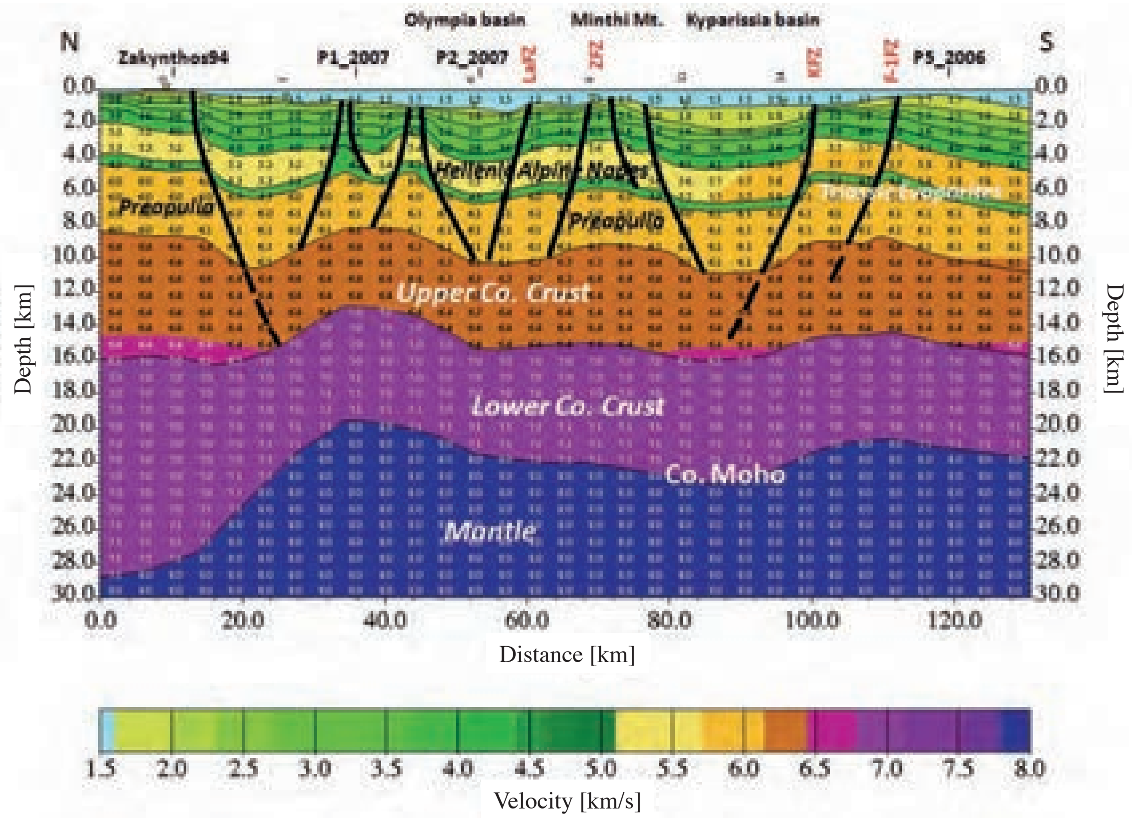

Fig. 6 - Vp-velocity model of Line P1_2006 with tectonic interpretation. Cross points with lines P1_2007 and P2_2007, P5_2006 and Zakynthos94 are indicated. LaFZ = Lapithas fault zone, KFZ = Kyparissia fault zone, and F1-FZ = Filiatra1 fault zone.

systems, one located south of OBS 0 and the other south of OBS 14. At the central part of the profile, sediments of 2.8 km/s velocity are elevated on either side of OBS 8, creating an anticline.

The first break velocity model was subsequently refined by layer tomography combined by two point ray tracing, using the SEIS84 algorithm (Cerveny and Psencik, 1981). An example of a ray traced CStG is presented in Fig. 5. Synthetic amplitudes were obtained by the Zoeppritz's (1907) equations. It is essential to generate kinematic as well as dynamic synthetic solutions for each observed CStG, in order to better constrain the results. Computation of travel times alone does not present the state of the art and wastes the important constrain posed by modelling also the energy distribution.

The final velocity model of Line P1_2006 with a tectonic interpretation is presented in Fig. 6. Accuracy of the model is approximately within ±5%. This refers to the definition of the Vp velocities and influences the depth of the various discontinuities, as shown in the ray density distribution presented in Fig. 7. The velocity model constrains the thickness of the continental crust (Moho depth) at the Kyparissiakos basin to range between 20 to 23 km, and 22 km towards south. The significant thickening of the crust towards Zakynthos to about 28 km has been adopted from a crustal velocity model published by Makris and Papoulia (2009). The faults mapped along the model, shown with thick black lines in Fig. 6, are the

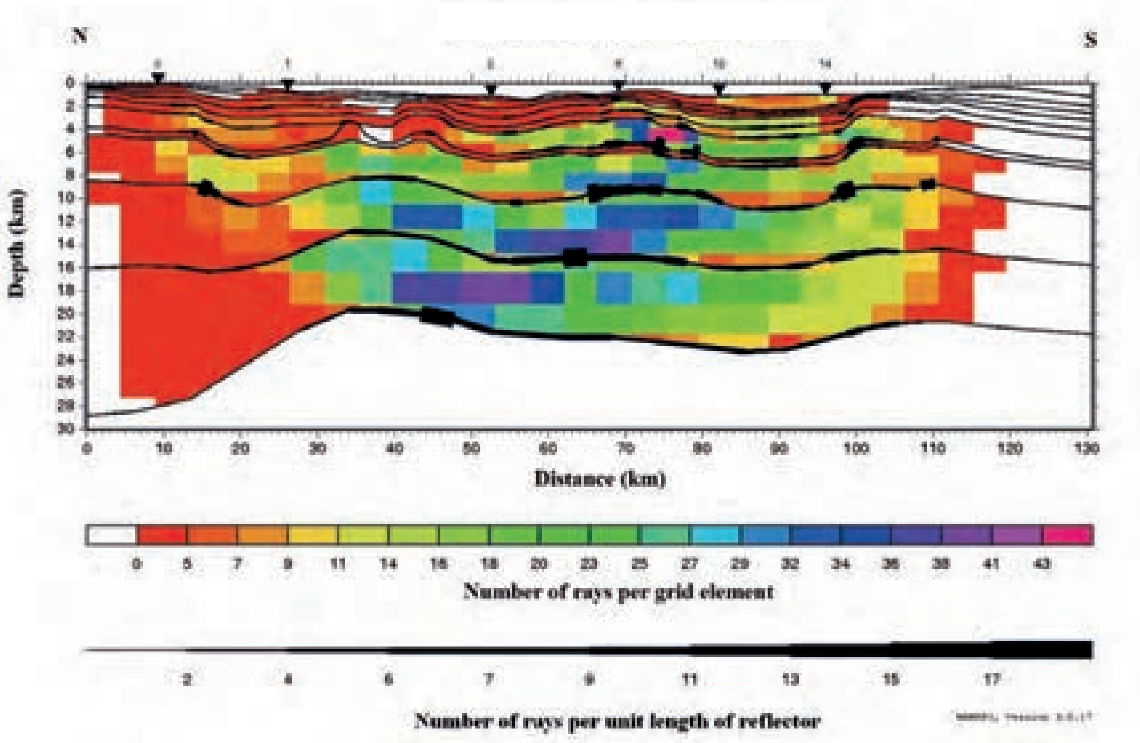

Fig. 7 - Ray density distribution showing the number of rays penetrating the different parts of the model and rays reflected per unit length of the model of Line P1_2006.

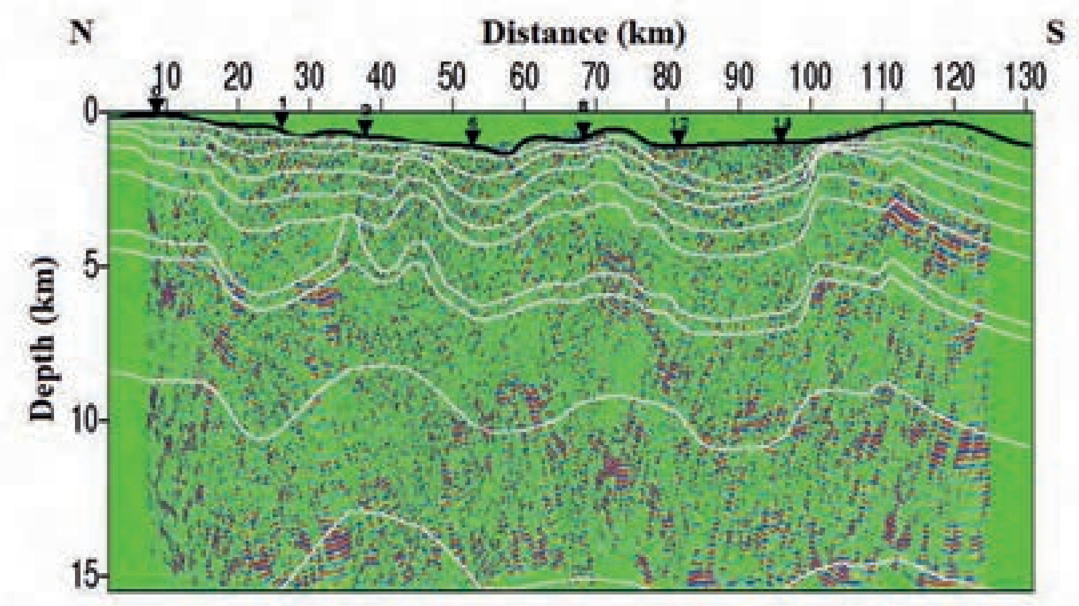

Fig. 8 - Depth migrated OBS data of Line P1_2006 superimposed with the layers mapped by the velocity model presented in Fig. 6.

westward extensions of onshore faults, mapped by IGME (1989), Mariolakos *et al*. (1998) and Papanikolaou *et al*. (2007).

The Kyparissiakos basin is separated by a series of faults in three extensional sub basins. Between OBS 0 and OBS 1, east of the Island of Zakynthos, a major fault was identified limiting a thicker continental block to the north from a thinner one to the south. This major fault at the north-western flank of the Kyparissiakos basin is seismically active (Papoulia *et al*., 2014b) and continues onshore, in the area of Killini. To the south, the basin is terminated by another major normal fault. It is located approximately 5 km south of OBS 14 and it is the westward continuation of the onshore Kyparissiakos Fault Zone (KFZ). This fault is also seismically active as revealed by local seismic observations (Papoulia *et al*., 2014b). South of this fault the sedimentary sequences are dipping southwards and the stretched crust is approximately 22 km thick. Fifteen km south of OBS 14 we identified a normal fault dipping northwards, which is the offshore continuation of an onshore mapped fault (F-1FZ), north of Filiatra (Papanikolaou *et al*., 2007). The Alpine limestones are separated from those of pre-Apulia by Triassic evaporites as identified also at the other three ENE-WSW oriented seismic lines.

In Fig. 8 we present the depth migrated section of Line P1_2006 superimposed by the first order seismic discontinuities defining Vp-velocity domains. The technique used for migrating wide angle reflections, and also diving waves, is presented by Pilipenko and Makris (1999). It is obvious that having more densely spaced OBSs inline the migration result would have been of much better quality.

4. The two ENE-WSW profiles between Zakynthos and Strophades: lines P1_2007 and P2_2007

These two profiles extend from the coast of Peloponnese for approximately 70 km to the west. Twenty OBS stations were placed along each line spaced at 2.5 to 3.5 km. Evaluation procedure for both lines is identical with that described for Line P1_2006. Examples of two CStG's of position 106, of Line P1_2007, and position 213, of Line P2_2007 are given in Figs. 9 and 10, respectively. In both positions energy propagated efficiently over the entire length of the CStGs. PmP arrivals were recorded at offsets between 25 and 40 km, which indicates that the crust along both profiles is fairly thin.

In Fig. 11 we present the first break tomographic model of Line 1_2007, which is restricted to the upper 10 km in order to have a better velocity resolution of the sediments and the upper crust. Sediment thickness to the ENE has maximum value of approximately 7.3 km. The basement is highly elevated at the central part of the profile, where thickness of sediments decreases to less than 3 km. At the western part of the profile, and at water depth of approximately 3000 m, the soft sediments are 2.5 km thick. Below this sedimentary sequence we have the high velocity metamorphic limestones and sediments of pre-Apulia extending to 7.5 km depth.

The first break tomographic model along Line 2_2007 is presented in Fig. 12. Thickness of sediments is very similar to that described along Line 1_2007. Basement geometry however is much more complex. This velocity model is used to constrain layer tomography and ray tracing for developing the final velocity model. Starting model for the deeper crust is constrained by PmP reflections and the apparent velocities of the Pn phase for the upper mantle, where identified.

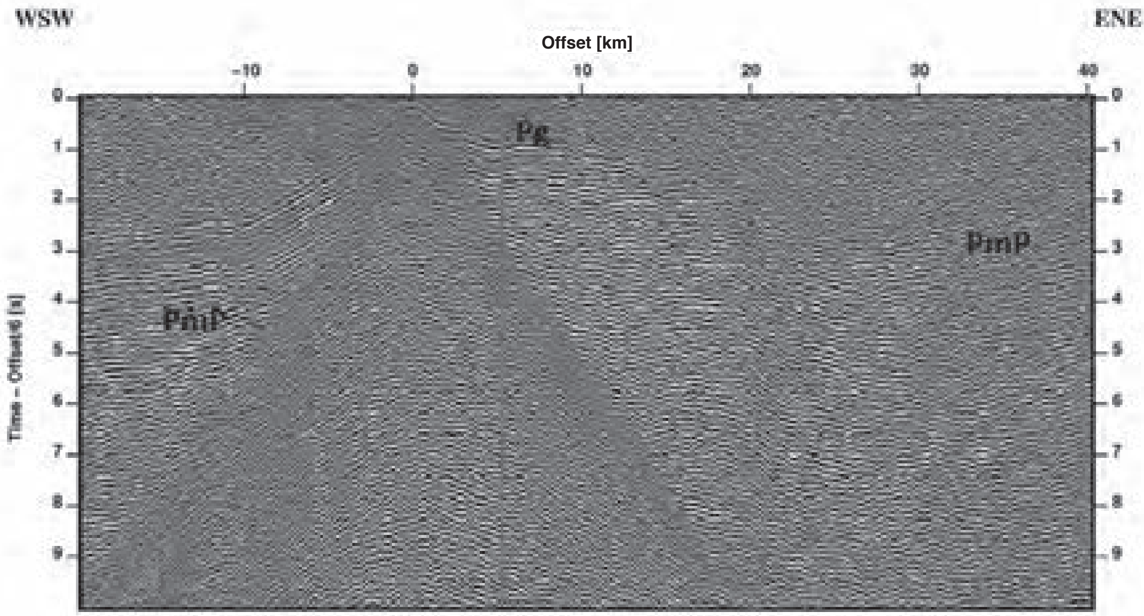

Fig. 9 - This CStG shows energy propagation over the entire length of Line P1_2007. Seismic waves propagating in the upper crust (Pg) and later arrivals of wide angle reflections from the Moho discontinuity (PmP) are marked. The complexity of the time section demonstrates the fragmentation intensity of the sequences offshore western Peloponnese.

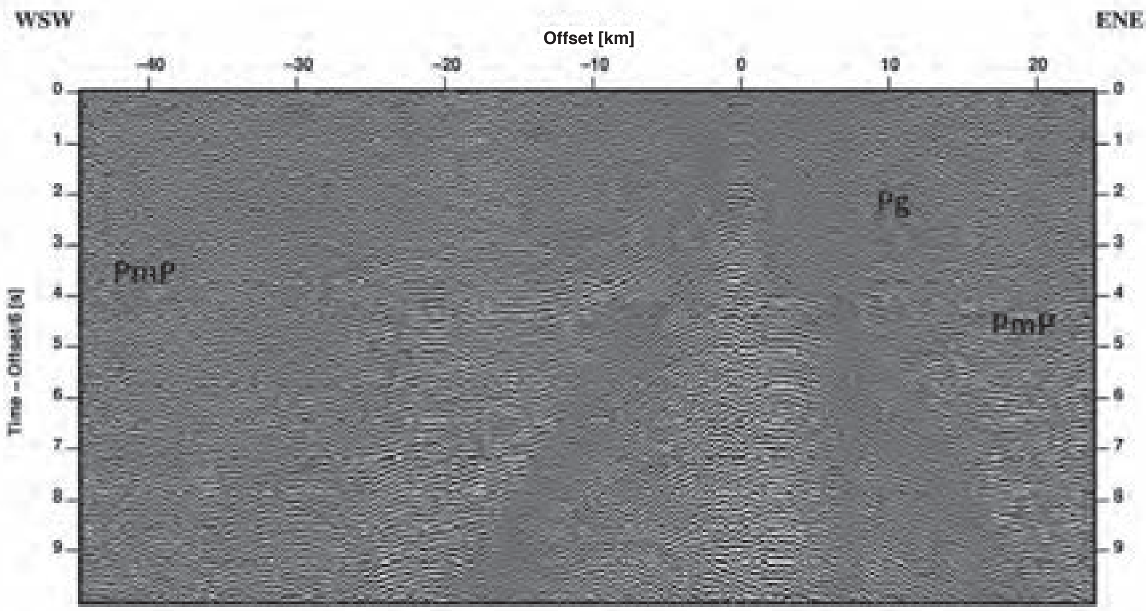

Fig. 10 - This CStG of Line P2_2007 shows a similar seismic behaviour as that presented in Fig. 9. To the WSW first arrivals disappear after 25 km from the OBS position due to the complex tectonism of the geological units. Deeper, delayed arrivals can be identified and permit to reconstruct velocity structure at depth.

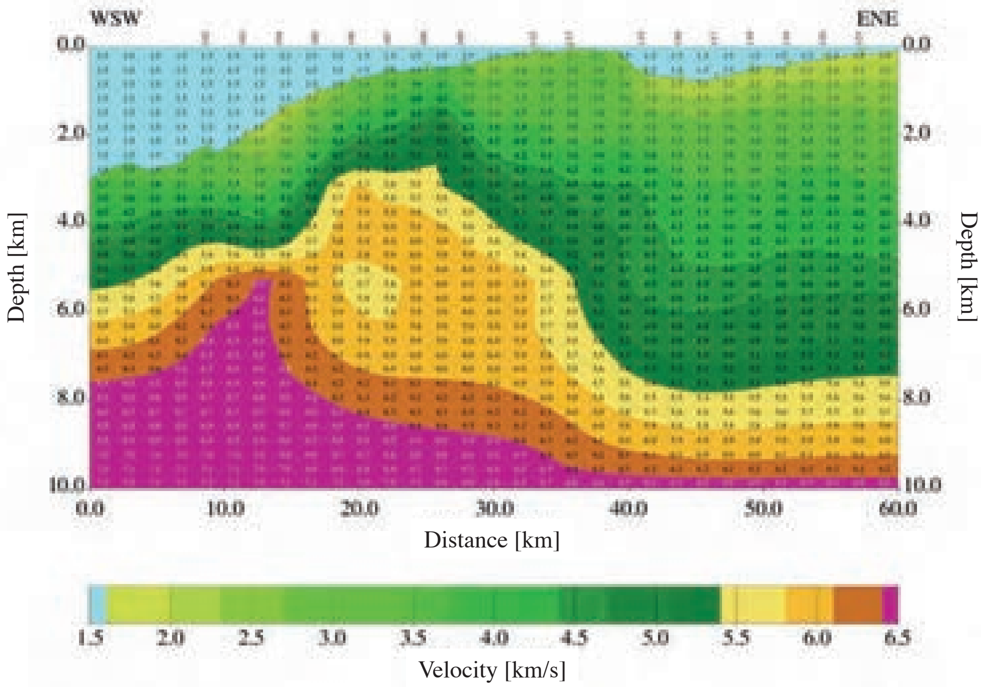

Fig. 11 - First break tomographic velocity model for the sediments of Line P1_2007. The pre-Apulia metamorphic limestone is highly elevated below the shallow part of the bathymetry. It is the southward continuation of the pre-Apulia limestones exposed at the western part of Zakynthos. Numbers in the figure show velocity values calculated after 10 iterations with an RMS of 0.06012 s.

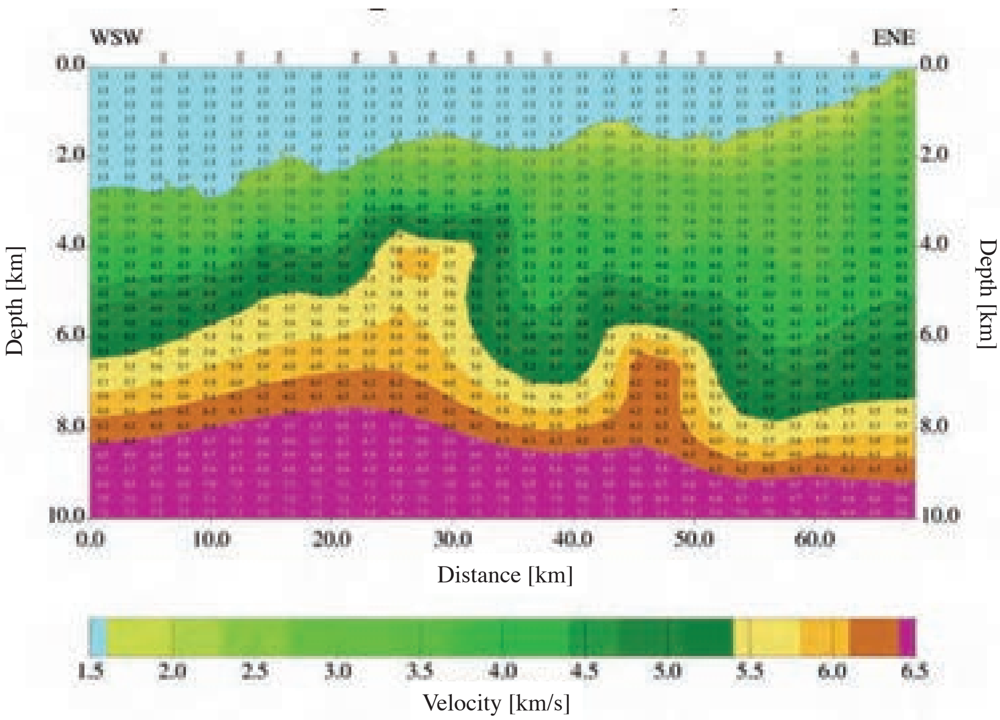

Fig. 12 - First break tomographic model along Line P2_2007. The RMS value is 0.07783 s after 10 iterations. Basement geometry is much more complicated than along Line P1_2007.

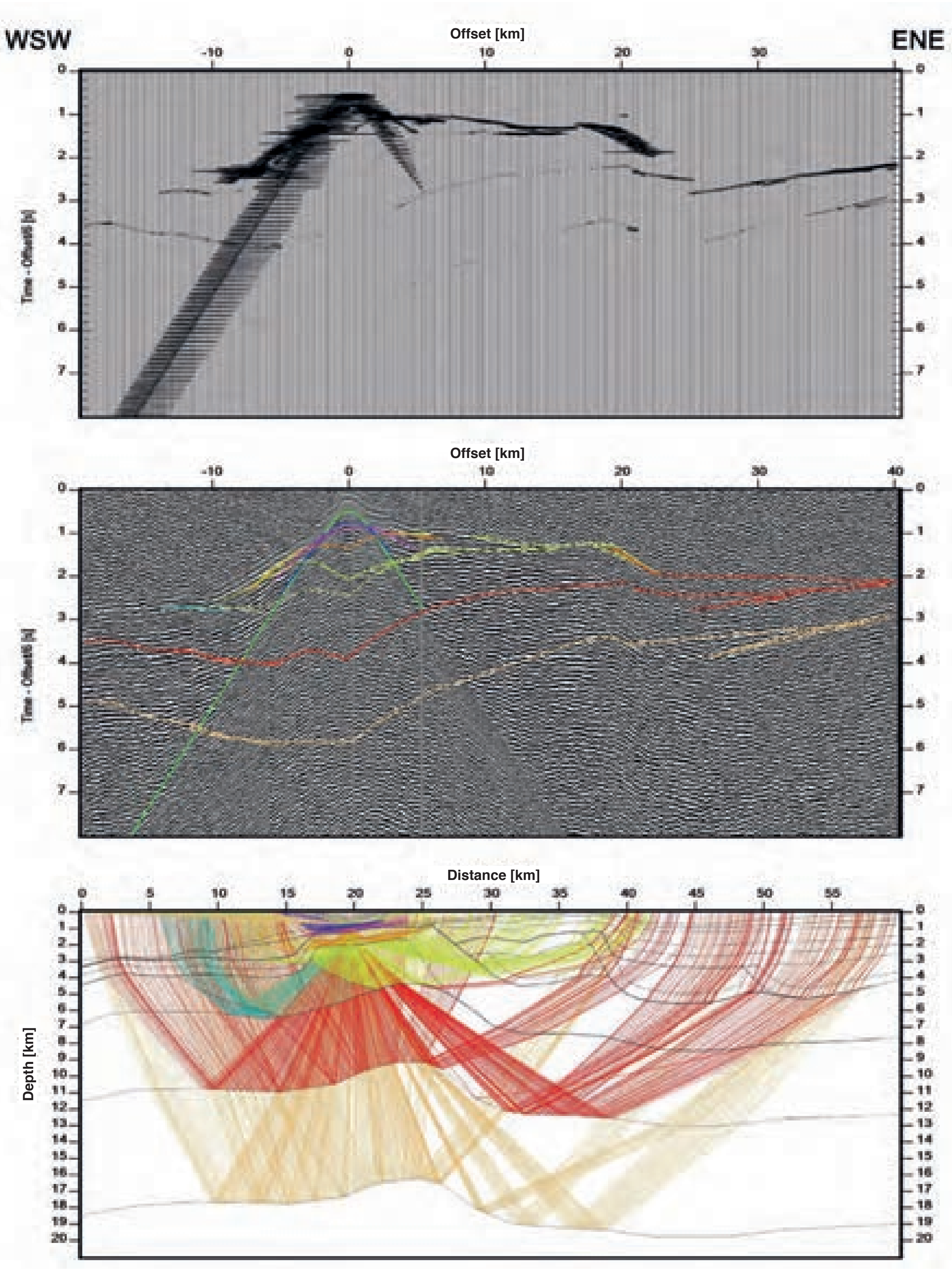

Fig. 13 - The ray traced arrivals of CStG 106, Line P1_2007 are presented in different colors for better identification of the seismic phases corresponding to different travel paths through the velocity model. In the upper part the synthetic amplitudes present the energy propagation along different parts of the velocity model.

The ray traced models for the CStGs 106 and 213, for lines 1_2007 and 2_2007, respectively, are presented in Figs. 13 and 14. In both cases Moho is constrained by reflectivity as was mapped by the synthetic calculations for travel times and amplitudes. We also present in Fig. 15 a synthetic example of calculated travel times and amplitudes from energy reflected at the upper and lower boundary of the Triassic evaporites. As seen in the synthetic amplitudes, the impedance contrast between the Alpine limestone and the evaporite is smaller than that between the evaporite and the pre-Apulia metamorphic limestone. In both cases however the contrast is strong enough to produce observable arrivals.

The Vp velocity models along the two northern ENE-WSW oriented seismic profiles are presented in Figs. 16 and 17. We also show the ray density distribution along Line P2_2007 (see Fig. 18) to demonstrate how the data constrain the velocity evaluation at depth. Both profiles south of Zakynthos have mapped a major fault, which is the western limit of the Ionian zone of the Hellenic Alpine units. Line P2_2007 shows a significant shift of the Ionian zone and the pre-Apulia thrust to the west compared to Line 1_2007. This is caused by the dextral strike slip motion along the Andravida fault. Vp-velocities of 5.6 and 6.1 km/s define two different metamorphic limestones, an upper and a lower one. The upper Vp-velocity limestone corresponds to the Alpine formations whereas the lower one is of pre-Apulia age. Very similar velocity values for these formations have been also observed at the seismic line reported across the Island of Zakynthos by Makris and Papoulia (2009). The two limestones are separated by evaporites that were identified in the OBS CStGs and are of Triassic age, as confirmed by Nikolaou (1986).

Crustal thickness in the Alpine domain ranges between 18 and 20 km, thickening towards Peloponnese. Towards the deep Ionian Sea, west of the pre-Apulia thrust, the continental crust is approximately 18 km thick at water depth of nearly 3000 m. The oceanic crust, which is subducted below the continental one, was not mapped since it was not identified in the CStGs. This is due to the limited profile length, which in both cases was less than 70 km.

Depth migrated CStGs superimposed to the Vp-velocity model (see Figs. 19 and 20) delineate the major fault systems, accentuating the tectonic processes. Basement geometry is clearly mapped and reveals major thrusting at the central part of both lines. To the west of the thrusted front between the Ionian and pre-Apulia zones the crust is tectonized and shortened by compression. At both lines, the Triassic evaporites separate the upper from the lower limestone that is the Alpine allochthonous napes from pre-Apulia basement. The evaporites act as "lubricants" for the westwards sliding of the Alpine napes.

5. The WSW-ENE profile offshore Messinia: Line P5_2006

This profile has a length of 180 km and extends from the Island of Proti (see Fig. 2), close to the Peloponnese - Messinia coast to the deep Ionian Sea. It crosses the backstop and the Mediterranean Ridge, and was occupied by 36 OBS positions. Data were evaluated as previously described. Tectonization along this profile is very intense as revealed by the observed CStG's (see Fig. 21).

As seen in Fig. 21, energy propagates efficiently to the ENE, while to the WSW, towards the deep Ionian Sea, the propagation is poor. This is due to the intense fragmentation of the crust

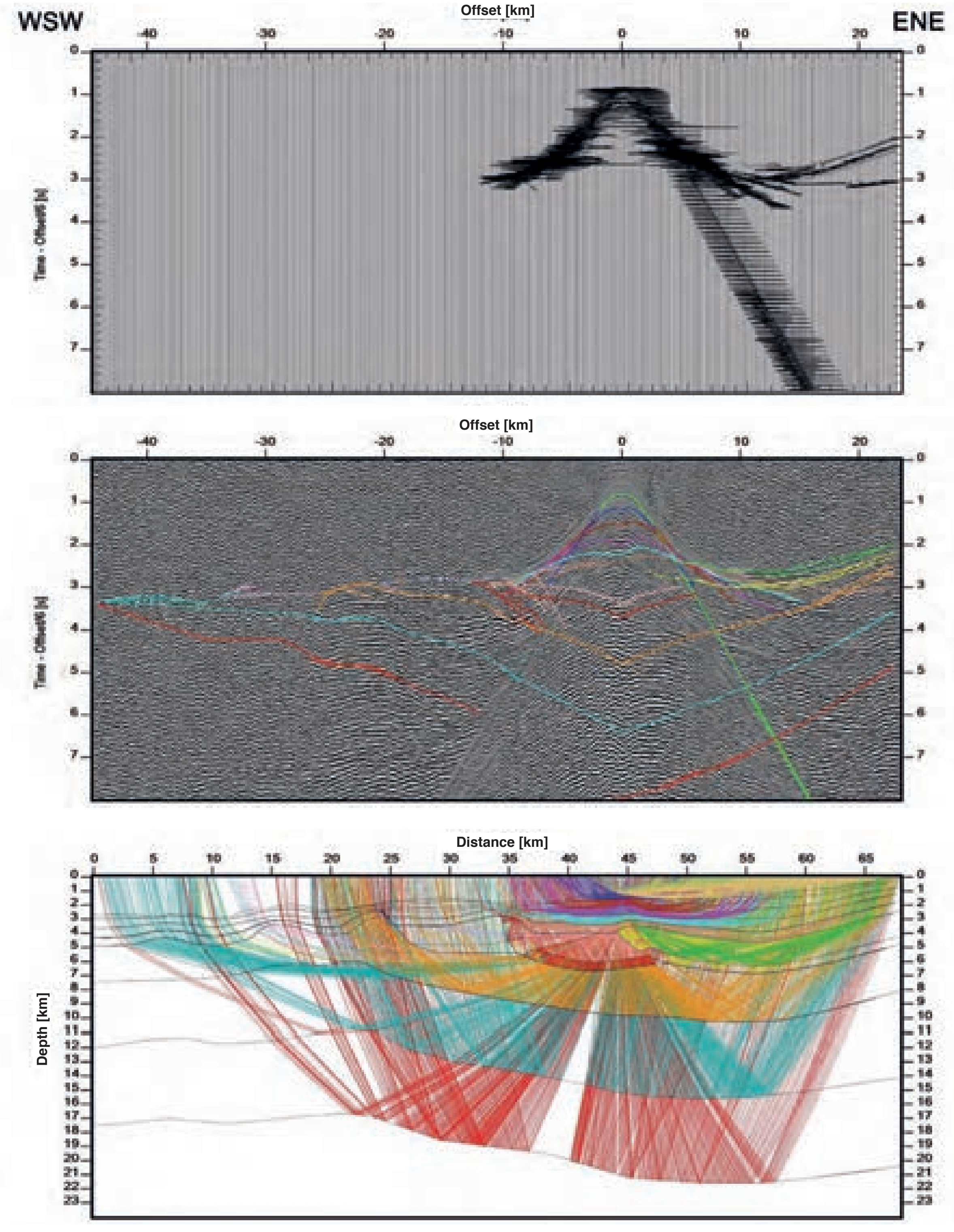

Fig. 14 - Synthetic travel times (middle part) and synthetic amplitudes (upper part) for the velocity model presented at the lower part for CStG 213, Line P2_2007. Colours indicate the correspondence between plotted travel times and travel paths along the seismic model.

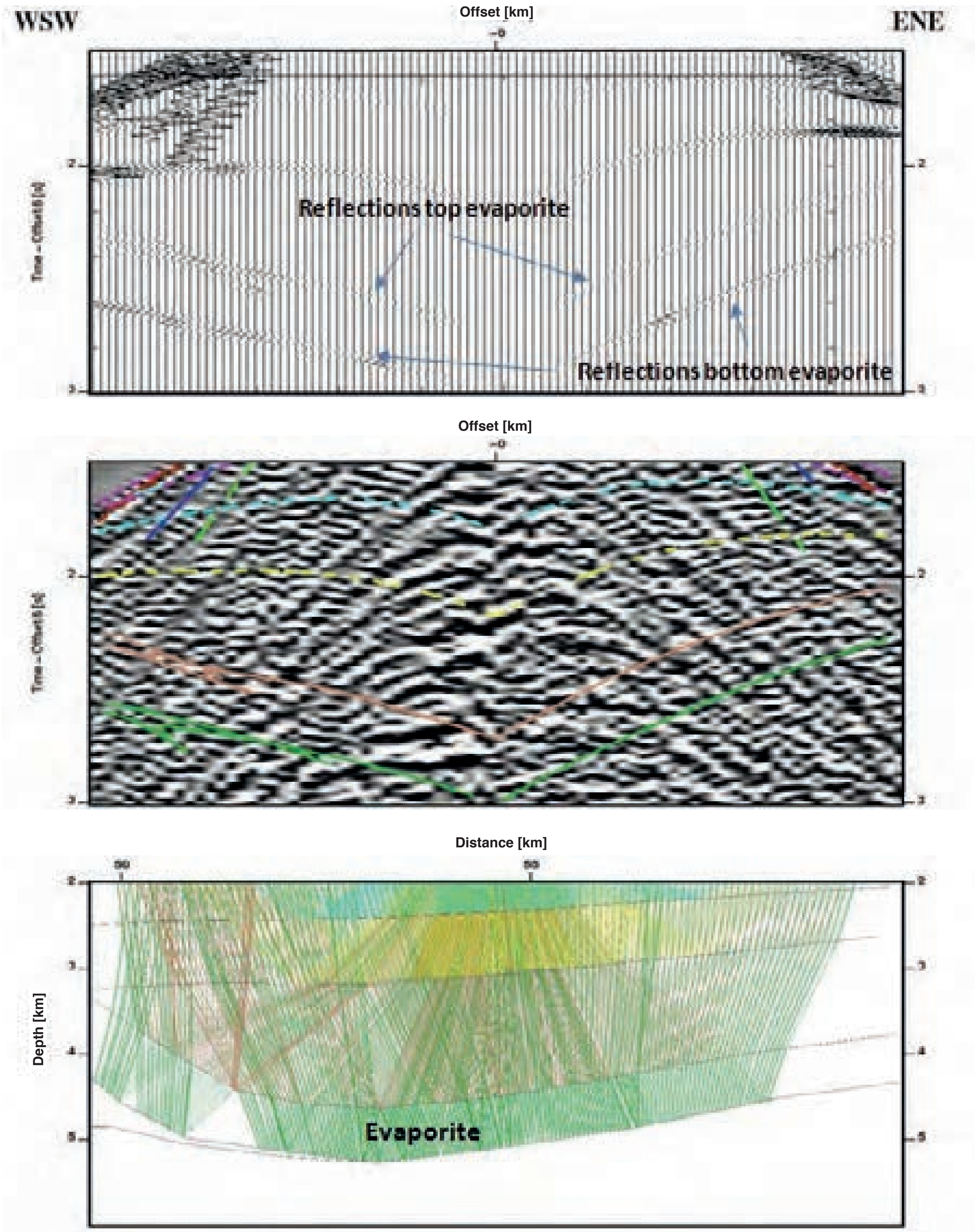

Fig. 15 – Wide-angle reflections from the upper (brown lines) and lower (green lines) boundary of the Triassic evaporite at OBS position 120 of Line P1_2007. Synthetic amplitudes are presented in the upper part of the figure.

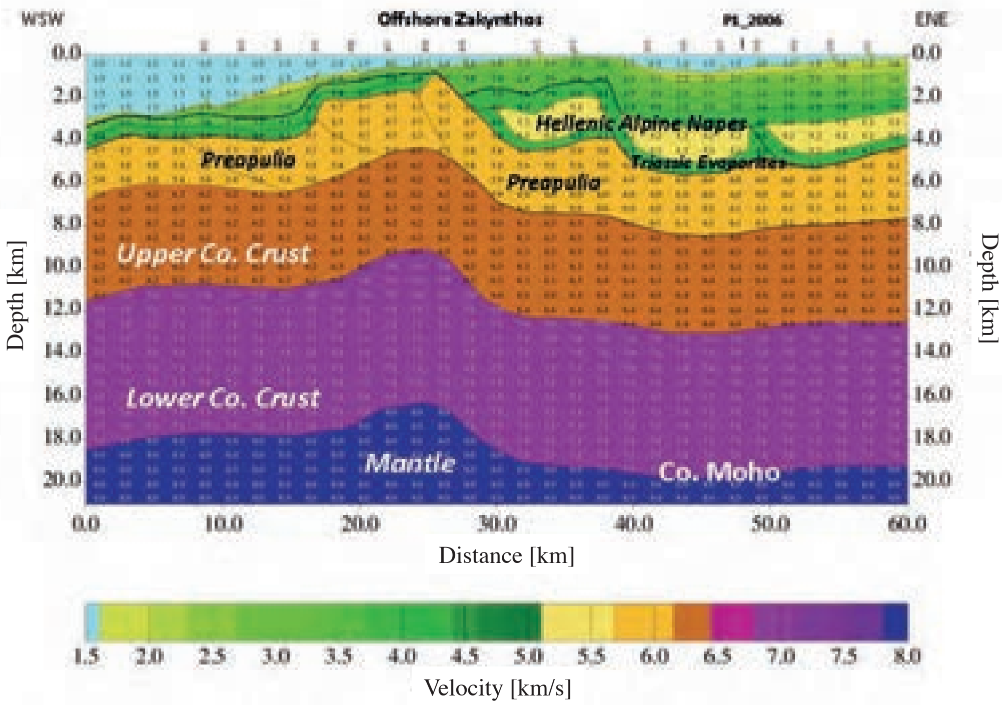

Fig. 16 - Final velocity model of Line P1_2007 south of Zakynthos. Crust thickens to the ENE to approximate 20 km below the Alpine units. Alpine napes glide over the pre-Apulia limestone over the Triassic evaporites. Intersection with Line P1_2006 is marked.

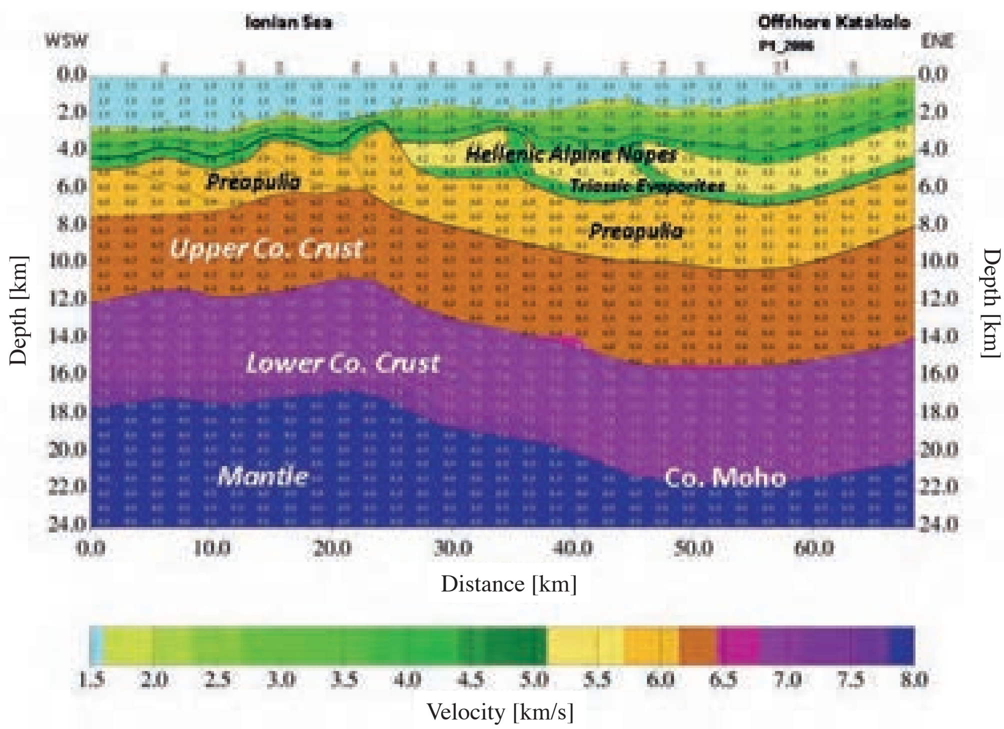

Fig. 17 - Final velocity model along Line P2_2007, south of P1_2007, between the islands of Zakynthos and Strophades. Triassic evaporites separate the Alpine from the pre-Apulia limestones and sediments. Intersection with Line P1_2006 is marked.

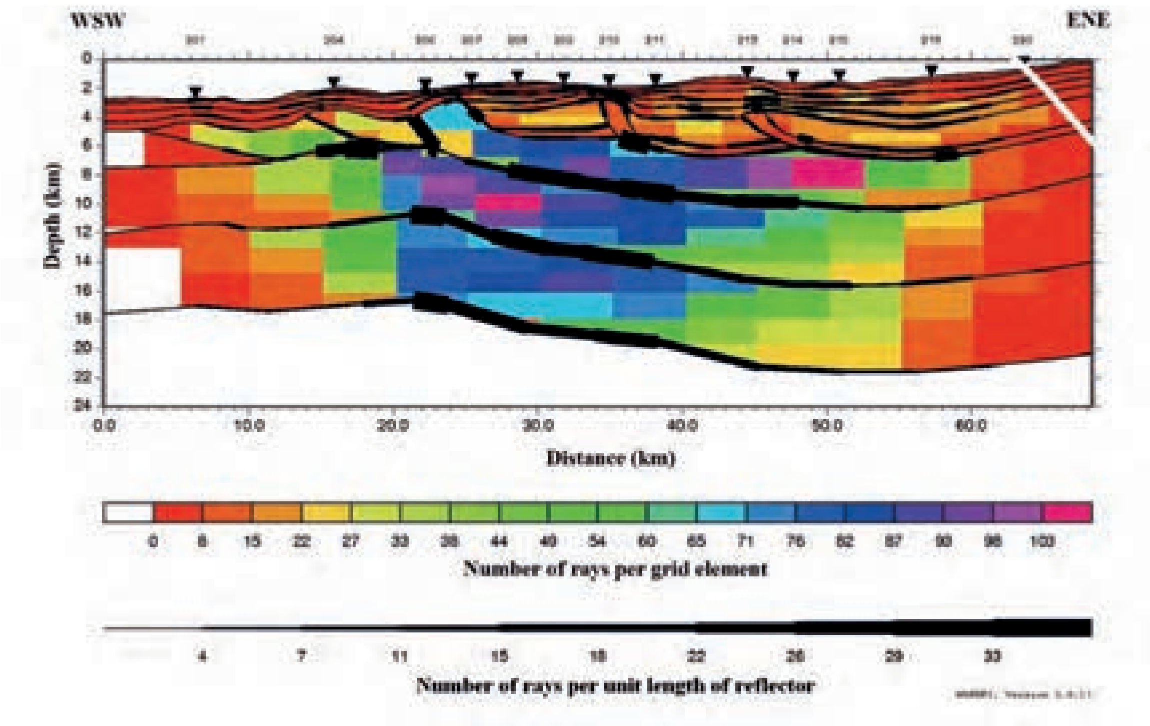

Fig. 18 - Ray density distribution showing the number of rays penetrating per grid element the different parts of the model and rays reflected per unit length of the model of Line P2_2007.

and the high velocity metamorpic limestones. Wide angle PmP arrivals from the Moho, although weak, were identified and constrained the crustal thickness.

The first break tomographic model along Line P5_2006 is presented in Fig. 22. The basement uplift at km 140 is the limit of the Alpine Hellenic napes, as will be demonstrated after the final velocity model evaluation.

The ray traced model in Fig. 23 shows the travel paths of the rays reflected from the top of the oceanic Moho at 21 km depth. In the CStG (middle part of Fig. 23) the synthetic travel times intersect the time axis at approximately 5.2 and 3.9 s at km 30, the ENE end of the line.

The final velocity model of Line P5_2006 is presented in Fig. 24. Low velocity sediments of Vp values ranging between 2.2 and 3.7 km/s thicken systematically from OBS position 4 to OBS position 19 that is from west to east (for better identification of OBS positions see also Fig. 25, where first order discontinuities are plotted in the depth migrated section). This indicates a gradual subsidence of the oceanic crust towards the collision front, at the Mediterranean Ridge. The western limit of the continental backstop is thrusted over sediments of the Mediterranean Ridge. The subducted oceanic slab as mapped follows the geometry of the continental Moho. Thickness of the continental crust at the backstop is approximately 16 km and that of oceanic Moho was mapped at a depth of 20 km. The oceanic slab below the continental crust was not always clearly seen by wide angle reflections; its continuation was assumed from geological considerations and microseismicity observations (Papoulia *et al.*, 2014b). It was also not possible to map directly the sediments, most probably subducted with the oceanic lithosphere below the continental crust. Resolution of velocity and structure obtained by the seismic data was not

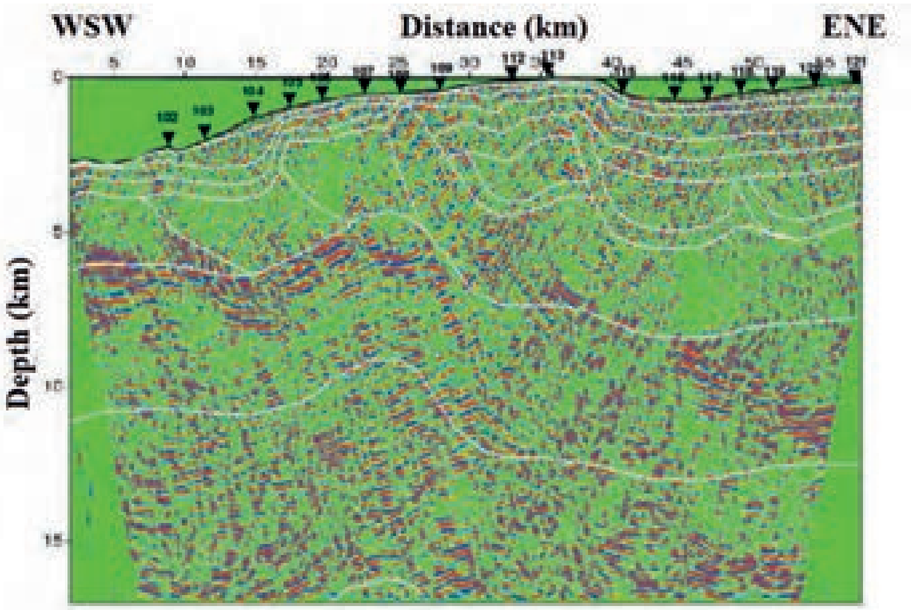

Fig. 19 - Layers of sediments and crust as developed by ray tracing velocity modelling (Fig. 13) superimposed with depth migrated data of Line P1_2007. We present the upper 16 km of the model, in order to resolve sediments more clearly and enhance the tectonic deformation. The agreement between depth migrated seismic arrivals and velocity structure is obvious.

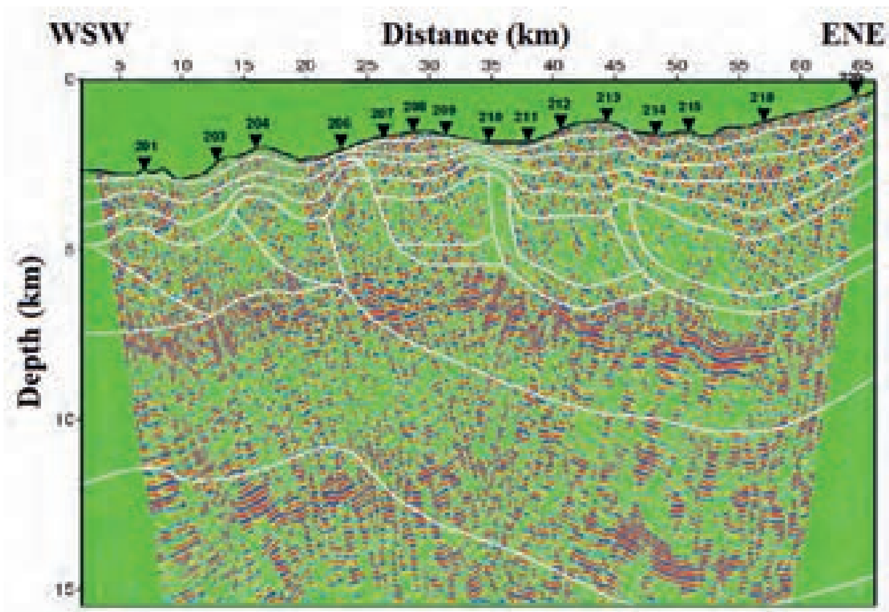

Fig. 20 - Upper part of the velocity model of Line P2_2007 and depth migrated section. Layers of crust and sediments are superimposed with the migrated seismic phases and are in good agreement. They demonstrate the efficiency of migrating wide-angle reflections even in widely spaced OBS stations. Migration resolution could be significantly improved by decreasing OBS stations spacing.

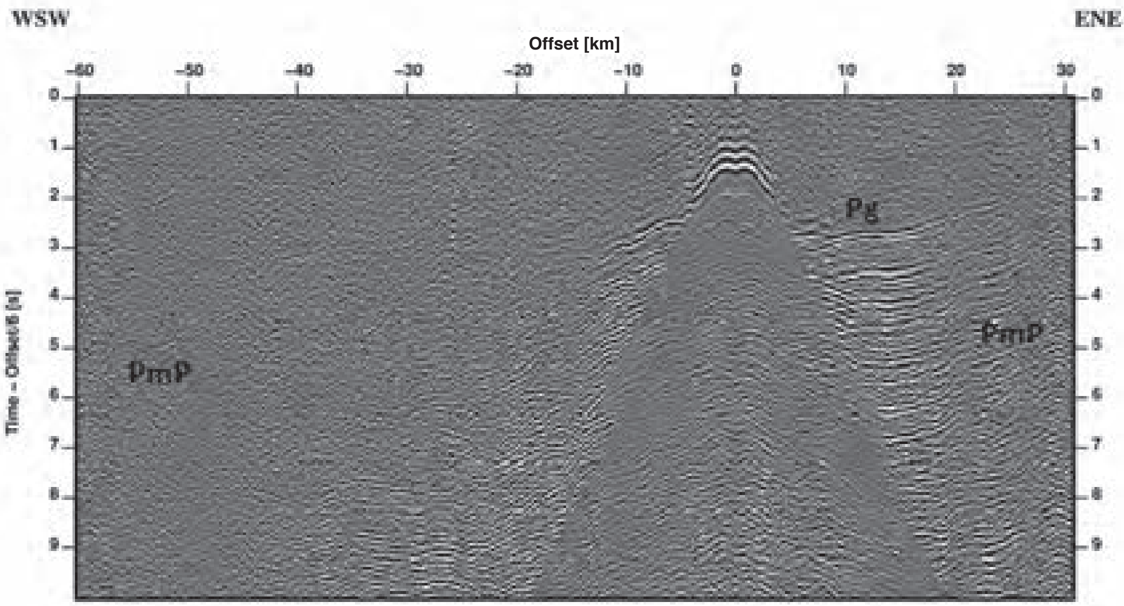

Fig. 21 - Example of one CStG of Line P5_2006 (OBS position 33), vertical component. Energy propagation was strongly affected by intense tectonization to the WSW. In spite of this shortcoming, it was possible to develop the velocity depth model by using wide-angle reflections. The velocity model could be resolved with sufficient resolution and permitted to reconstruct the tectonic elements.

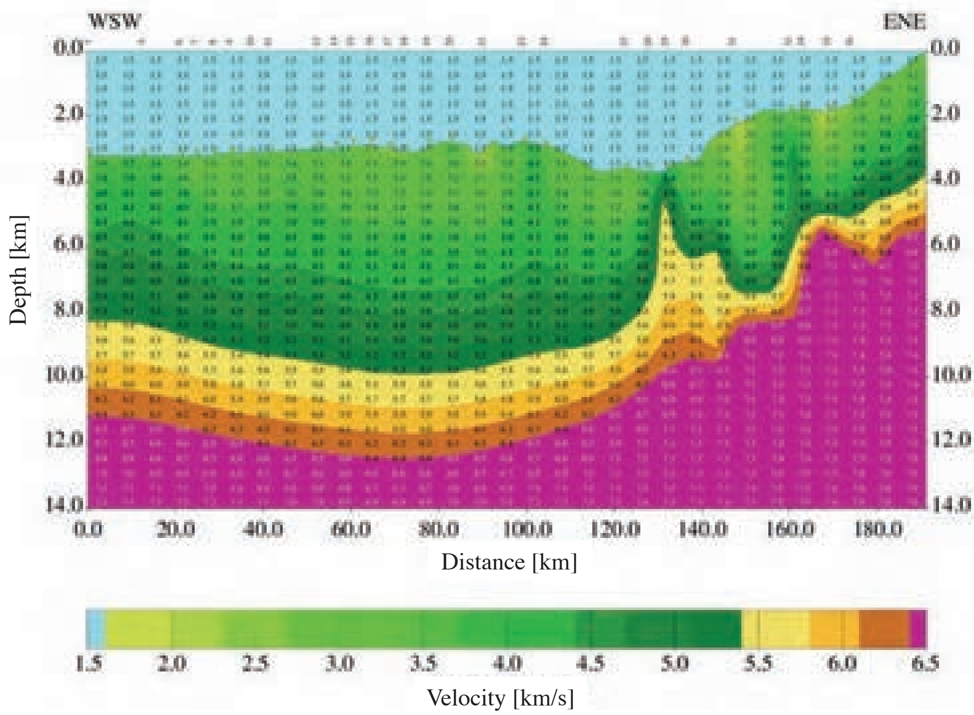

Fig. 22 - First break velocity model along Line P5_2006, extending from the south-western coast of Peloponnese (Proti Island) towards the deep Ionian Sea.

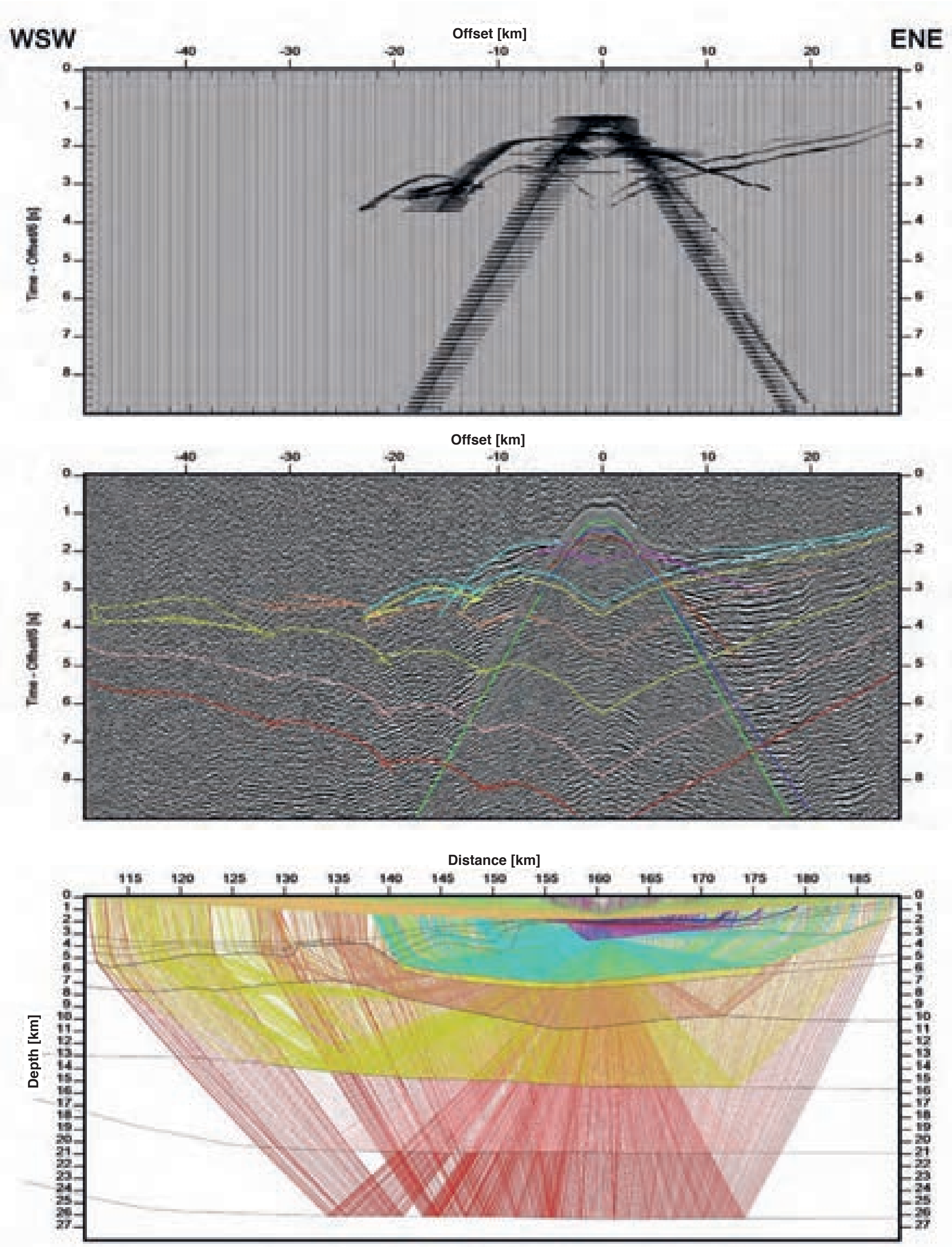

Fig. 23 - Ray traced model of OBS 33 of Line P5_2006. Ray paths are presented in the middle part and synthetic amplitudes at the upper part of the figure.

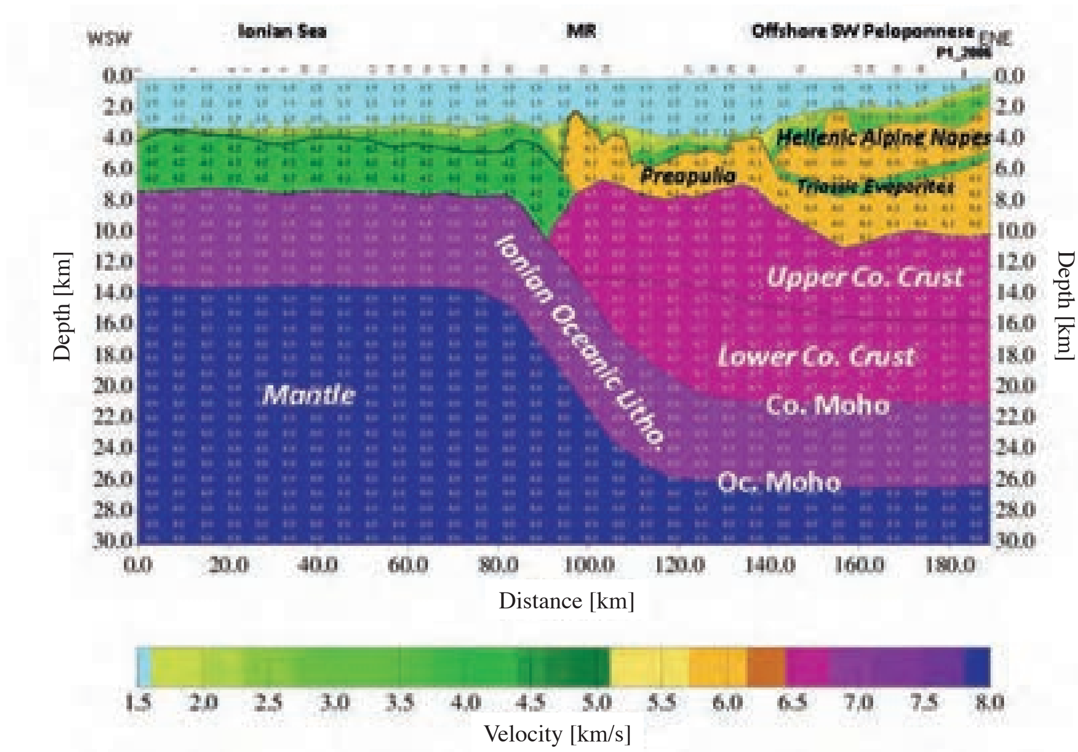

Fig. 24 - Crustal section from offshore Messinia (Proti Island) to the deep Ionian Sea. The eastern part of the profile is floored by continental crust and the subducted Ionian oceanic crust below it. The backstop is only 43 km wide. The western part of the line is floored by oceanic crust. Sediments thicken significantly at the Mediterranean Ridge (MR). Intersection with Line P1_2006 is indicated.

sufficient for this task, due to the wide spacing of the OBS stations and the intense fragmentation of the crust that limits energy propagation by scattering and absorption. A significantly stronger air gun array than that used could have been very useful.

The eastern part of the profile, floored by continental crust, is covered by very thick sediments of 8 to 10 km thickness. We also identified two high velocity metamorphic limestones separated by Triassic evaporites. The upper limestone, corresponding to the Hellenic Alpine napes, is terminated at OBS 30, where the uplifted and thrusted lower limestone of pre-Apulia is exposed nearly to the sea floor. This part of the section coincides with the deepest part of the bathymetry of about 4000 m, and is part of the Hellenic Trench. The backstop is covered by soft sediments with increasing thickness to the west, deposited directly on pre-Apulia limestone and sediments (see also Fig. 25 for OBS positions).

The migrated section, presented in Fig. 25, shows the upper 14 km of the model. The velocities and their limits defined in the velocity model of Fig. 24 are overlapped with the migrated section. Strongly tectonized upper crustal units are visible. Structural elements mapped by both techniques (Vp modelling and depth migration) are in good agreement and help identifying the tectonic deformation. It is obvious that the eastern part of the profile, east of the collision front at the Mediterranean Ridge, is strongly tectonized by three major thrusts and several normal faults. On the contrary, the deep Ionian basin is fairly homogeneously structured and the sediments are strongly deformed only at the Mediterranean Ridge (see also Finetti and Morelli, 1973; Hirschleber *et al.*, 1994; Hartmann, 1996).

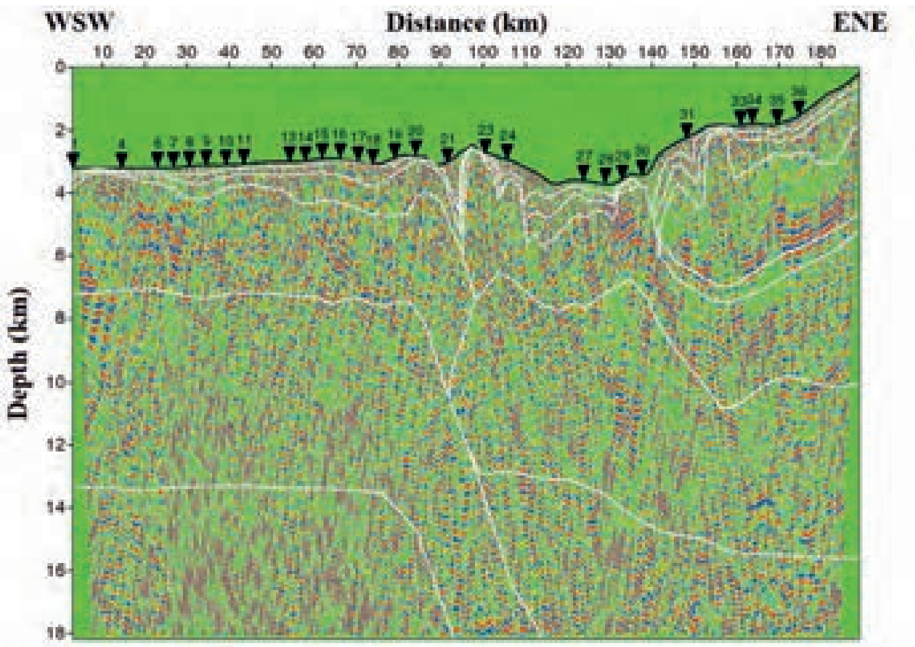

Fig. 25 - Depth migrated reflected and refracted arrivals of the CStGs, Line P5_2006. OBS positions at the sea bottom are indicated by triangles. White lines denote first order discontinuities mapped by the velocity model, as shown in Fig. 24.

6. The ENE-WSW profile from Messiniakos Gulf to the Mediterranean Ridge: Line P3_2007

This profile of 160 km length was occupied by 40 OBS positions, and crosses one of the deepest basins of the Mediterranean Sea exceeding 4500 m depth. Example of a CStG is presented in Fig. 26.

Energy along this profile propagated more efficiently than in the previous line (Line P5_2006) although tectonization is very intense and one would expect poor seismic efficiency, particularly for seismic energy observed to the WSW.

The first break tomographic model along Line P3_2007 is presented in Fig. 27. The basement uplift at km 100 coincides with extend of the Alpine Hellenic napes, as will be demonstrated after the final velocity model evaluation.

The ray traced model of OBS 330 of Fig. 28 shows travel paths of the rays reflected from the continental Moho at 22 km depth.

The velocity model along this line is presented in Fig. 29, while the ray density distribution demonstrating the model accuracy is shown in Fig. 30. At the western part of the profile, between OBS 341 and OBS 328, we have mapped two sequences of high velocity limestone, separated by Triassic evaporates (for better identification of OBS positions see Fig. 31). Two major thrusts were identified along this profile. The western one at OBS 335 is within the Hellenic Alpine units. It is the continuation to the south of the onshore structure mapped by IGME (1989) in Messinia. It is the limit of the Gavrovo to the Ionian zone in the offshore.

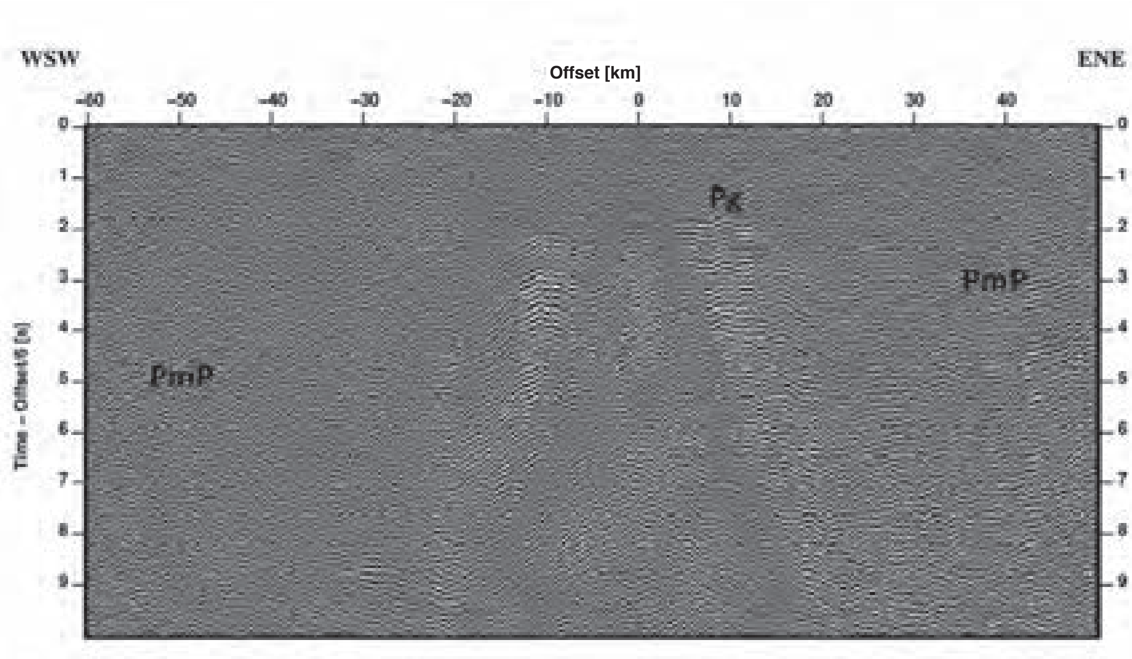

Fig. 26 - CStG of OBS position 330, vertical component, of Line P3_2007. The abrupt loss of energy propagation at about 18 km WSW of the OBS position is due to intense basement fragmentation. Energy was observed again at larger offsets as wide-angle reflections from the crust/mantle boundary (PmP – phase).

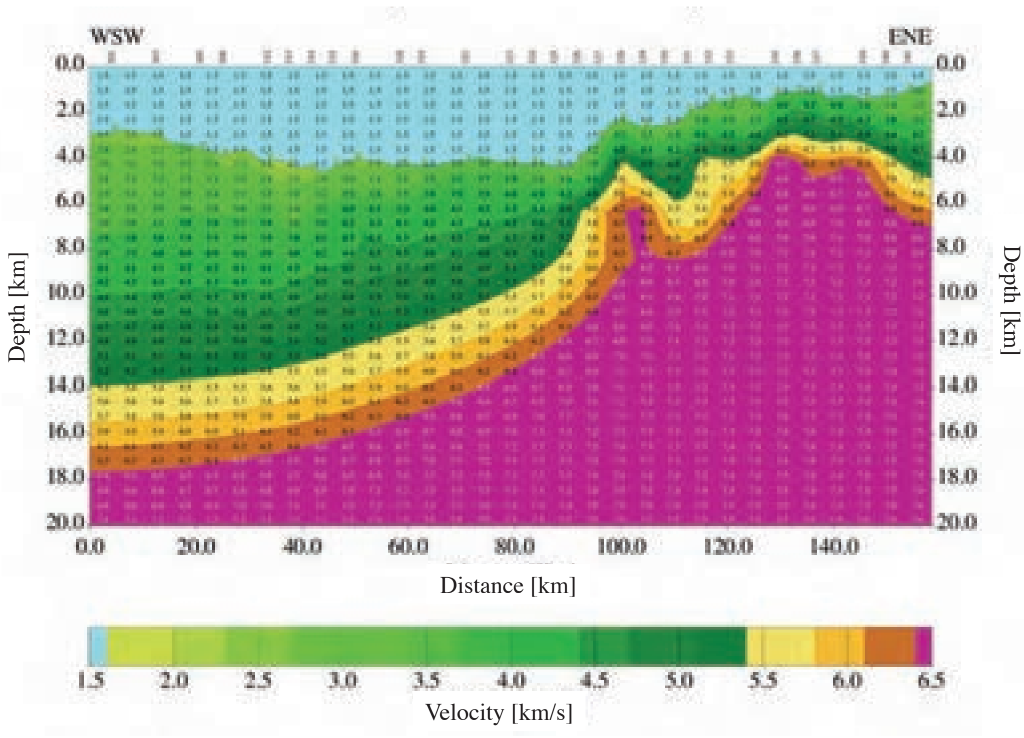

Fig. 27 - First break tomographic model along Line P3_2007, extending from the Messiniakos Gulf to the deep Ionian Sea.

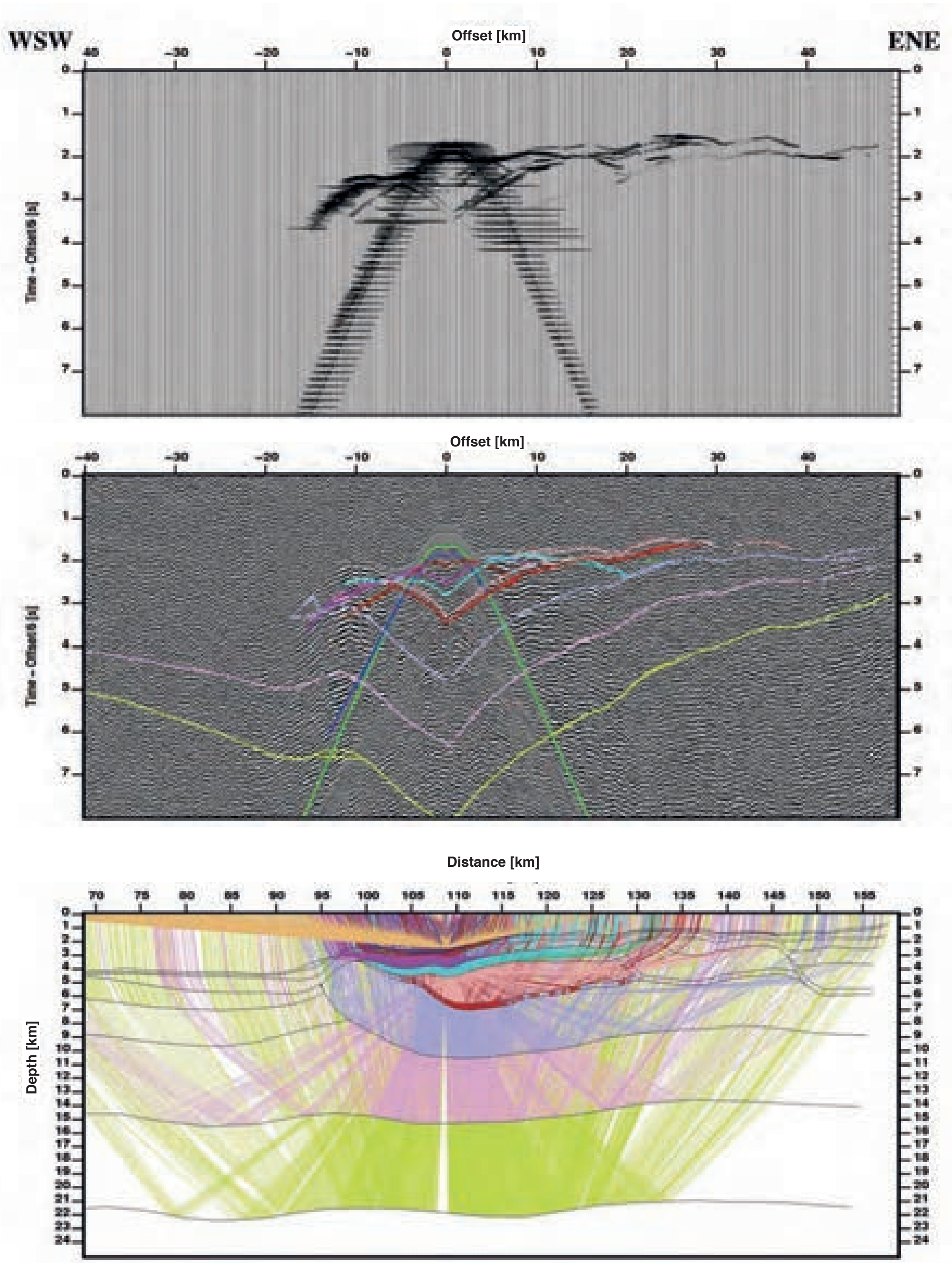

Fig. 28 - Ray traced model of OBS position 330, Line P3_2007. Ray paths are presented in the lower part and synthetic amplitudes at the upper part of the figure. In the middle part, we present the synthetic travel times superimposed on the CStG.

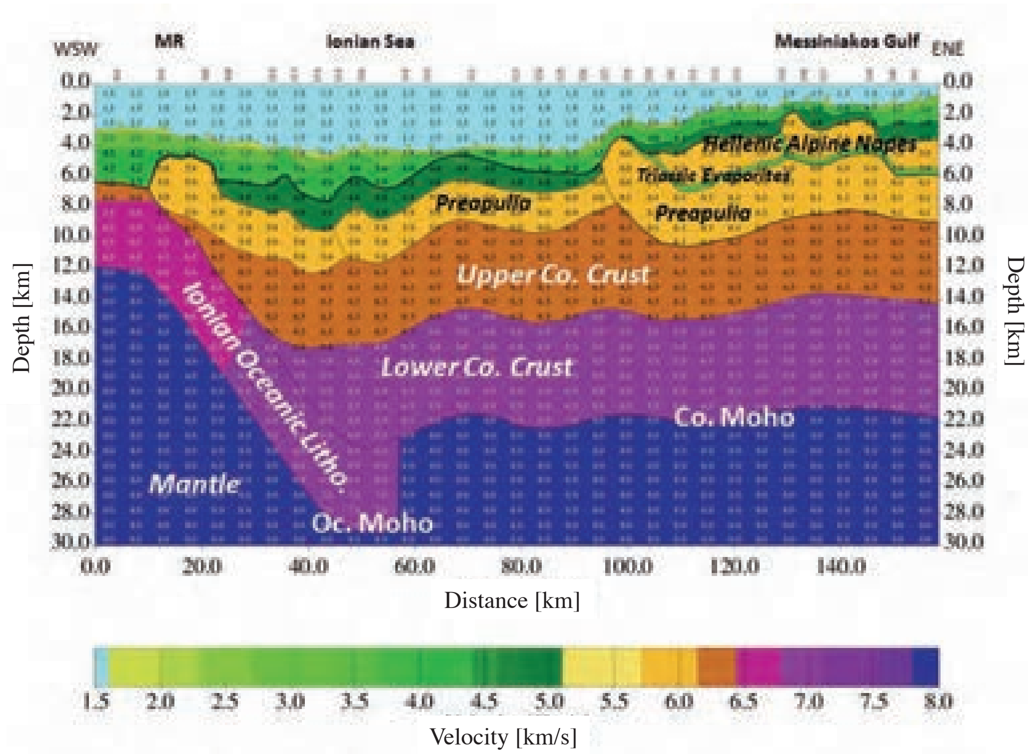

Fig. 29 - Velocity model of the backstop area west of the Messiniakos Gulf, Line P3_2007. The Alpine thrusts and the deformation of the backstop are well documented along this line. The backstop beyond the Ionian zone is more than 85 km wide.

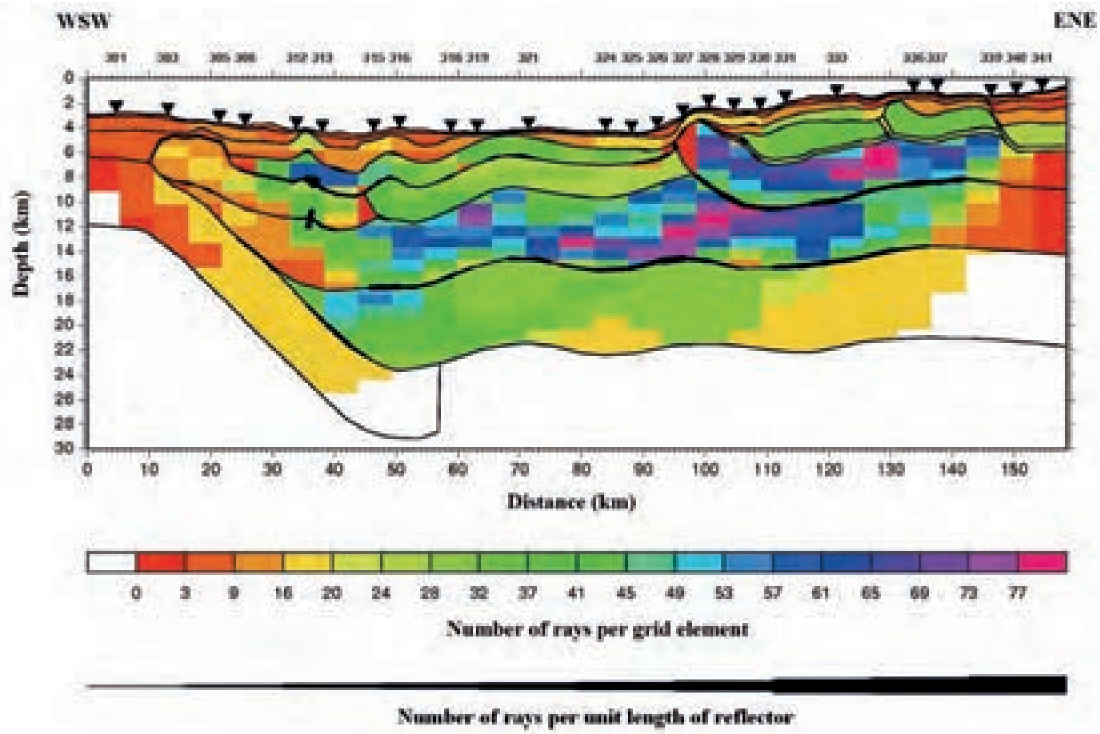

Fig. 30 - Ray density distribution showing the number of rays penetrating the different parts of the model and rays reflected per unit length of the model of Line P3_2007.

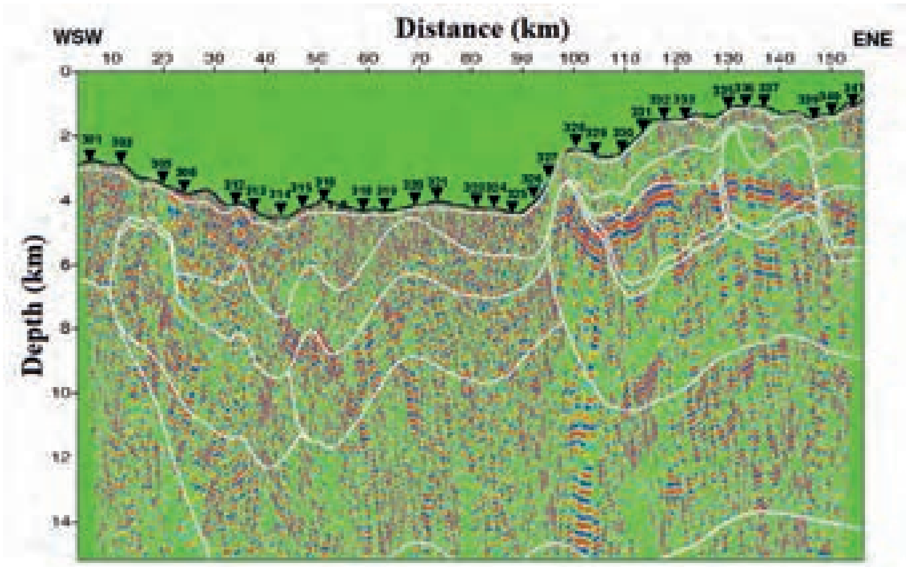

Fig. 31 - Superimposed depth migrated reflected and refracted arrivals and first order discontinuities from velocity model of Line P3_2007. Triangles show the OBS positions. Migrated data and velocity model resolve the tectonic structure of the area and accentuate the tectonic deformation.

The second major thrust is located at OBS 328. The lower limestone of pre-Apulia is nearly exposed to surface and defines the western limit of the Alpine napes. From this thrust-front until OBS 303, which is nearly at the western end of the profile, only one high velocity limestone has been identified. This part of the section is the continental backstop, built entirely by pre-Apulia limestone and sediments. The Alpine napes of western Greece do not extend to the west, in the backstop area, beyond this thrust. The Mediterranean Ridge is clearly marked to the west of OBS 303 by a bathymetric uplift and change of the sediments and their structure. The transition between the Mediterranean Ridge and the pre-Apulia backstop is deformed by intense fragmentation of the thin continental crust and by thrusting of high velocity limestones over sediments at the transition of the backstop to the oceanic crust of the deep Ionian Sea. The sediments at the backstop thicken from east to west. The thickest part of the low velocity backstop sediments is identified below OBS 314 and is about 5 km. Vp velocity of the sediments ranges between 4.0 and 4.85 km/s. We could not follow the oceanic slab below the backstop because energy failed to penetrate below the continental Moho, which was mapped at approximately 22 km depth. It is however obvious that below the continental crust the oceanic lithosphere is subducted and that sediments are intercalated between the two crustal oceanic domains.

We overlapped the velocity model with the depth migrated OBS data in Fig. 31. We present only the upper 15 km of the section in order to enhance the resolution of the tectonic elements. Two main zones of significant thrusting and crustal shortening were mapped. The eastern one, between OBS 328 and OBS 341, includes two major thrusts as were described in the velocity model. The other thrusted zone at the western part of the section is associated with the uplift of pre-Apulia limestones and sediments and their westward thrusting over the oceanic crust of the deep Ionian Sea. This again is linked with a second thrust belt close to the collision front below OBS 312 to OBS 316. It shows that compressional deformation is very intense and is spread over a significant part of the western backstop.

7. Conclusions and discussion

The active seismic experiment based on the efficient use of OBS technology and densely spaced airgun shots produced CStGs that permit to develop reliable crustal velocity models. The five OBS lines cover the complete offshore region of western Peloponnese, between Zakynthos Island and the Gulf of Messinia. Thus, the westward continuation of the Hellenic Alpine units in the offshore region was mapped, the width of the continental backstop was defined, and the main tectonic elements offshore were linked to the onshore structures.

The N-S oriented Line P1_2006 mapped a series of faults of nearly E-W orientation. The crustal thickness in the broader Kyparissiakos area is about 22 km thick and the crust is thin continental. Sediments range between 8 and 10 km and in their lower part are composed of two high velocity metamorphic units (Vp = 5.5 km/s, upper unit and 6.1 km/s lower unit). Triassic evaporites separate the two limestones, producing salt dome intrusions in several locations. The upper sedimentary sequences have Vp velocities ranging between 1.8 and 4.8 km/s. Their thickness varies between 2 and 4 km. The faults mapped along this profile defined the N-S extension of the Kyparissiakos basin. The master faults at the northern and southern flanks of the basin are also seismically active with moderate to intense seismicity (Papoulia *et al.*, 2014b) and truncate the entire crust. A series of faults were mapped within the basin that correlate with the onshore structures, separating the central part of the basin in a series of sub-basins. From south to north, we have the Kyparissia basin, separated from the Olympia basin by the offshore continuation of the Minthi Mts. The southern flank of this basin is seismically very active and associated with sub crustal seismicity that extends to 90 km depth (see also Papoulia *et al.*, 2014b). The northern flank of the basin is also linked to crustal seismicity. Fault plane solutions of locally recorded events have revealed mainly strike slip faulting of dextral orientation (Papoulia *et al.*, 2014b). We consider this fault system to be the offshore extension of the Andravida fault. Further to the north, offshore Katakolo, another sub- basin of significant downthrow has developed.

Lines P1_2007 and P2_2007 of ENE-WSW orientation are parallel to the southern coast of Zakynthos and have mapped the northern part of the Kyparissiakos basin, its crust and sediments, between the coast of western Peloponnese and the Zakynthos-Strophades uplift. Crustal thickness along both lines varies between 18 and 20 km, always thickening to the east. The thinnest part of the crust coincides with the highest uplift of the pre-Apulia metamorphic limestone and sediments, thrusted westwards. The continental crust is depressed and the water depth increases systematically to 3000 m, at the western edge of both profiles. To the east of the pre-Apulia thrust an upper and a lower high velocity sequence were mapped, separated by Triassic evaporites. The lower limestone belongs to pre-Apulia, and extends also west of the pre-Apulia thrust into the Hellenic backstop. The upper limestone is part of the external Hellenides and is limited to the east of the pre-Apulia uplifted block, exposed on Zakynthos Island. It is the limit of the westward extension of the Hellenides towards the deep Ionian Sea. It is, therefore, obvious that the backstop between the Mediterranean Ridge and the Ionian zone is composed exclusively by pre-Apulia continental crust and not by the western Hellenic napes as postulated Le Pichon *et al.* (2002).

The two southern crustal profiles, P5_2006 and P3_2007, have mapped the extension of the continental crust to the west, crossing the backstop and its interaction with the Mediterranean

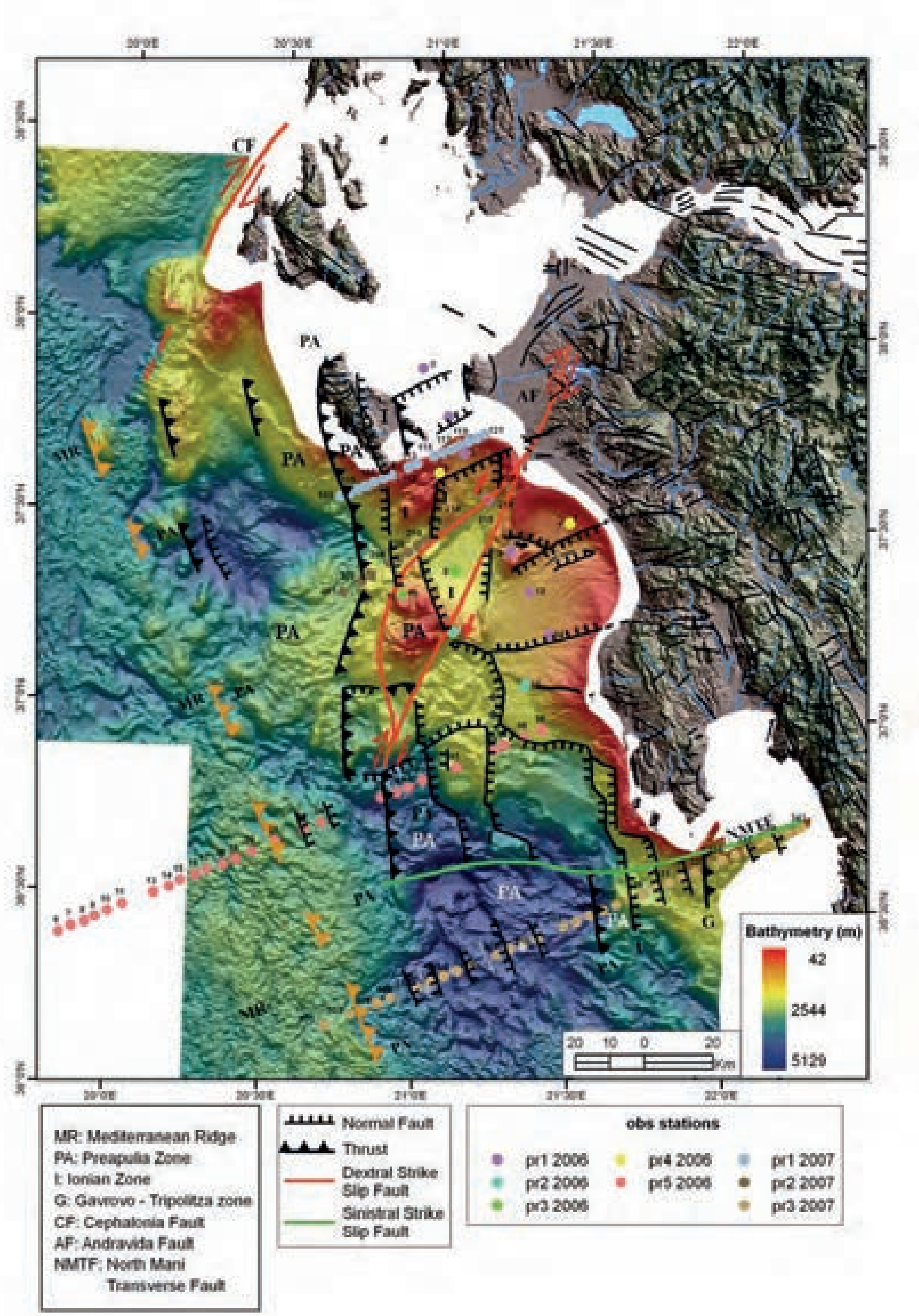

Fig. 32 - Simplified tectonic map offshore western Peloponnese derived from active seismic observations. The OBS positions used for velocity modelling and delineation of the faults are indicated in the map. Swath bathymetry was adopted from Camera *et al.* (2014). Thrust in orange color denotes the limit between pre-Apulia and the Mediterranean Ridge. Discussion of the map in text.

Ridge. The type of crust mapped in the deep Ionian Sea, west of the backstop is oceanic, and Moho is approximately 11 km deep, at water depth of 3000 m. Sediment thickness is approximately 4.5 km, increasing to the eastern limit of the Mediterranean Ridge. The igneous part of the crust with a Vp velocity ranging from 6.4 to 6.8 km/s is about 6.5 km thick and it is part of the Tethys Ocean, subducted below the Hellenic Arc. East of the collision front, between the continental domain of the backstop and the oceanic one of the Ionian Sea, the crust is thickening eastwards. Crustal thickness at the backstop area is about 17 to 18 km and sediments vary between 4.5 and 8.0 km, depending on the extend of deformation that has affected them. The two high velocity limestones mapped by all profiles at their eastern part are separated by Triassic evaporites. The lower limestone and sediments compose the pre-Apulia unit, while the upper one is the western extension of the Hellenic Alpine napes. Continental crust in this area is approximately 22 to 24 km thick and sediments and crust are strongly tectonized.

The limit between the Mediterranean Ridge and the backstop was located at 100 to 120 km west of the coast of Peloponnese. This front is systematically shifted westwards between the Zakynthos Island and offshore SW Messinia. It is interesting that the main difference between the two southern lines is the width of the backstop defined between the Mediterranean Ridge and the western limit of the Hellenic napes. Along the northern profile, the backstop is 40 km wide, while at the southern line it extends for more than 80 km. South of Line 5_2006, the pre-Apulia thrust is significantly shifted towards Peloponnese by more than 50 km (see Fig. 32). This is linked to the left lateral North Mani transverse fault (Lallemant, 1984), which is presently inactive.

Correlating the onshore geology of the Alpine units with that mapped offshore, we placed the limits of the Ionian to the pre-Apulia zones in the Kyparissiakos area, and the pre-Apulia, Ionian and Gavrovo zones south of Messinia. Nowhere beyond the uplifted pre-Apulia metamorphic limestone and sediments and its thrust off the coast of Peloponnese have we identified the existence of Hellenic Alpine napes in the backstop area. This observation does not agree with Aubouin *et al*. (1976) and Le Pichon *et al*. (2002), who place the limit of the western Hellenides deep into the Ionian Sea and even extend it to the Mediterranean Ridge.

The Kyparissiakos basin is strongly affected by dextral strike slip faulting linked to the Andravida fault, which is displacing the geological and tectonic elements westwards. This is also responsible for the development of transtensional basins and transpressional uplifts, like the rhombic basin NE of the Strophades Island and the Strophades uplifted block. The tectonic deformation of the Strophades Island is also described by Stiros (2005) and is in agreement with the general deformation pattern derived by the present observations. The fact that at the subduction zone we have not mapped subducted sediments is due to insufficient resolution of the OBS data at this depth. As Shaw and Jackson (2010) showed, the subducting slab is decoupled from the crust above it. It is, therefore, soft deformable sediments that account for the decrease of frictional forces between the two interacting rigid lithospheric systems. Only densely spaced OBS stations and shorter shooting intervals could map the deeper interfaces with sufficient resolution needed to identify the sediments involved in the subduction and the velocity inversion they would cause.

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REFERENCES

Ambraseys N.N.; 1962: *Data for the investigation of the seismic sea waves in the Eastern Mediterranean*. Bull. Seism. Soc. Am., **52**, 895-913.

Aubouin J., Bonneau M., Davidson J., Leboulenger P., Matesco S. and Zambetakis A.; 1976: *Esquisse structurale de l'arc Egéen externe: des Dinarides aux Taurides*. Bull Soc. Geol. France, **18**, 327-336.

Camera L., Mascle J., Wardell N., Accettella D. and the SEAHELLARC team; 2014: *The Peloponnese continental margin from Zakynthos Island to Pylos: morphology and recent sedimentary processes*. Boll. Geof. Teor. Appl., **55**, 325-342, doi:10.4430/bgta0092.

Cerveny V. and Psencik I.; 1981: *Two-dimensional two point ray tracing package*. Charles University, Prague.

De Voogt B., Truffert C., Chamot-Rooke N., Huchon P., Lallemant S. and Le Pichon X.; 1992: *Two-ship seismic soundings in the basins of the Eastern Mediterranean Sea (Pasiphae cruise)*. Geophys. J. Int., **109**, 536-552.

Ditmar P. and Makris J.; 1996: *Tomographic inversion of 2-D WARRP data based on Tikhonov regularization*. In: 66th Ann. Intl. Mtg. SEG, Denver CO U.S.A., Expanded Abstracts.

Finetti I.; 1982: *Structure, stratigraphy and evolution of central Mediterranean*. Boll. Geof. Teor. Appl., **24**, 247-315.

Finetti I. and Morelli C.; 1973: *Geophysical exploration of the Mediterranean Sea*. Boll. Geof. Teor. Appl., **15**, 261-341, 14 maps.

Galanopoulos A.; 1960: *Tsunamis observed in Greece from antiquity to present time*. Annali di Geofisica, **13**, 369-386.

Hartmann J.M.; 1996: *Geophysicalische Untersuchungen in der Jonischen See*. Berichte aus dem Zentrum für Meeres- und Klimaforschung- Reihe C, Band 10.

Hirn A., Sachpazi M., Siliqi R., McBride J., Marnelis F., Cernobori L. and STREAMERS – PROFILES group; 1996: *A traverse of the Ionian Islands front with coincident normal incidence/ and wide angle seismics*. Tectonophysics, **264**, 35-49.

Hirschleber J., Hartmann M. and Hieke W.; 1994: *The Mediterranean Ridge accretionary complex and its forelands—seismic reflection studies in the Ionian Sea*. In: Ansorge R. (ed), Universität Hamburg, Schlaglichter der Forschung zum 75, JahrestagHamburger Beitr. Wissenschaftsgeschichte **15**, Reimer Verlag, Berlin, pp. 491–509.

IGME; 1989: *Seismotectonic map of Greece with seismological data, 1:500.000 scale*. IGME, Athens, Greece.

Kahle H-G., Straub C., Reilinger R., McClusky S., King R., Hurst K., Veis G., Kastens K. and Cross P.; 1998. *The strain rate field in the eastern Mediterranean region, estimated by repeated GPS measurements*. Tectonophysics, **60**, 1-42.

Kamberis E., Marnelis F., Louckogiannakis M., Maltezou F., Hirn A. and the STREAMERS Group; 1996: *Structure and deformation of the External Hellenides based on seismic data from offshore western Greece*. EAGE Special Publications, **5**, 207-214.

Kokinou E., Kamberis E., Vafidis A., Monopolis D., Ananiadis G. and Zelilidis A.; 2005: *Deep seismic reflection data from offshore western Greece: a new crustal model for the Ionian Sea*. Journal of Petroleum Geology, **28**, 185-202.

Lallemant S.; 1984: *La Transversal Nord-Mani; etude geologique et aeromagnetique d'une structure transverse an l'Arc Egéen Extern*. Dr. Troisieme cycle, University Pierre et Maire Curie, 164 pp.

Le Pichon X., Lallemant S.J., Chamot-Rouke N., Lemeur D. and Pascal G.; 2002: *The Mediterranean Ridge backstop and the Hellenic napes*. Marine Geology, **186**, 111-125.

Makris J.; 1977: *Geophysical investigations of the Hellenides*. Geophysical Monographs, **34**, 1-124.

Makris J. and Papoulia J.; 2009: *Tectonic evolution of Zakynthos Island from deep seismic soundings: thrusting and its association with the Triassic evaporates*. In: Intl. Symposium on Evaporites, Zakynthos, pp. 47-54.

Makropoulos C. and Burton P.W.; 1981: *A catalogue of the seismicity in Greece and adjacent areas*. Geophys. J. R. Astr. Soc., **65**, 741-762.

Mariolakos I., Sabot V., Fountoulis I., Markopoulou-Diakantoni A. and Mirkou R.; 1998: *Filiatra. Neotectonic map of Greece 1:100,000*. Earthquake Planning & Protection Org., Athens, Greece.

Monopolis D. and Bruneton A.; 1982: *Ionian Sea (western Greece): its structural outline deduced from drilling and geophysical data*. Tectonophysics, **83**, 227-242.

Nikolaou K.; 1986: *Contribution to the study of the Neogene and the geology and boundaries of Ionian and pre-Apulian isopic zones in relation to petroleum geology observations mainly in the Islands Strophades, Zante, Cephalonia*. PhD Thesis, Univ. of Athens (in Greek).

Papanikolaou D., Fountoulis J. and Metaxas Ch.; 2007: *Active faults, deformation rates and Quaternary paleogeography at Kyparissiakos Gulf (SW Greece) deduced from onshore and offshore data*. Quaternary International, **171-172**, 14-30.

Papazachos B. and Papazachou C.; 1997: *The earthquakes of Greece*. Editions Ziti, Thessaloniki, 304 pp.

Papoulia J., Makris J., Mascle J., Slejko D. and Yalçiner A.; 2014a: *The EU SEAHELLARC project: aims and main results*. Boll. Geof. Teor. Appl., **55**, 241-248, doi: 10.4430/bgta0100.

Papoulia J., Makris J. and Tsambas A.; 2014b: *Microseismicity and crustal deformation of the Kyparissiakos Gulf - south-western Hellenic Arc using an "amphibious" seismic array and a 3D velocity model obtained from active seismic observations*. Boll. Geof. Teor. Appl., **55**, 281-302, doi: 10.4430/bgta0086.

Pilipenko V. and Makris J.; 1999: *Application of migration to the interpretation of WARP data*. In: Expanded Abstracts of the 69th SEG Meeting, Dallas.

Reilinger R., McClusky S., Paradisis D., Ergintav S. and Vernant P.; 2010: *Geodetic constrains on the tectonic evolution of the Aegean region and strain accumulation along the Hellenic subduction zone*. Tectonophysics, **488**, 22-30.

Shaw B. and Jackson J.; 2010: *Earthquake mechanisms and active tectonics of the Hellenic subduction zone*. Geoph. J. Int., **181**, 966-984.

Stiros S.; 2005: *Geodetic evidence for mobilization of evaporites during the 1997 Strophades (W Hellenic Arc) 6.5 Mw earthquake*. J. Geophys. Eng., **2**, 111-117.

Truffert C., Chamot-Rooke N., Lallemant S., De Voogt B., Huchon P. and Le Pichon X.; 1993: *The crust of the Western Mediterranean Ridge from deep seismic data and gravity modelling*. Geophys. J. Int., **114**, 360-372.

Weigel W.; 1974: *Die Krustenstruktur unter dem Ionishcen Meer nach Ergebnissen Refraktionsseismischer Messungen auf den Farthen 17 und 22 des Forschungsschiffes METEOR*. Heft 26, Hamburger Geophysikalische Einzelschrifften, S. 140.

Zoeppritz K.; 1907: *Uber Erdbebebnwelle*. II. Laufzeitkurven. Nachrichten Gess. Wiss,. Goettingen, Math. Phys. Klasse, S. 529-549.

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