

Seismic velocity structure of the crust and the upper mantle beneath the island of Crete, Greece

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Abstract

Crete is located in the Southern Aegean, in the southernmost part of the Hellenic Trench. Given the elevated numbers of seismicity in the region generated by the convergence of the Eurasian and African tectonic plates, the research area is critical. More than 1400 manually revised events from 2018 to 2022 have been used in this work to construct both local and regional 1-D velocity models. These data were entered into Passive Seismic Tomography, which is using body-wave (P and S) travel-time data to analyze medium to large-scale (in km) anomalies linked to local neotectonic and regional tectonic processes. A complex shallow structure was revealed in Crete's central region, in contrast to a smoother and more continuous image at deeper slices ($> 40\text{km}$). This result was likely due to result of the subducted lithosphere triggering an upward migration of fluids. The final dataset analysis contributed to a better understanding of the research area's prevailing seismotectonic processes and the relationship between shallow and intermediate-depth earthquake activity.

Article Highlights

- More than 1400 manually revised between 2018 and 2022 have been used in order to obtain 1-D velocity models on both local and regional scale.
- An updated catalogue was constructed examining in detail aftershock sequences in the last two years of the study period (2021-2022).
- Earthquake Tomography was applied in order to examine local and regional tectonic elements in the study area.

1. Introduction

In Southern Aegean (Greece) lies the southernmost segment of the Hellenic arc which initiates in the region W-SW of Peloponnese, continues south of Crete, splitting into three main parts, Pliny, Ptolemy, and Strabo trenches, and terminating east of Rhodes island before its differentiation into Anatolia and South of Cyprus (McKenzie 1978; Le Pichon and Angelier 1979). This region is characterized by intense rates of seismicity, resulting from the convergence of the Eurasian and African tectonic plates, as shown by significant historic activity (e.g. Papazachos and Papazachou, 2003). The aforementioned motions are driven by the subduction of the Mediterranean lithosphere of oceanic origin under the Aegean continental one, with a rate of convergence $\sim 35\text{ mm/year}$ (LePichon and Angelier 1979; Jackson and McKenzie 1988; Le Pichon et al. 1995; McClusky; et al. 2000; Hollenstein et al., 2008; Reilinger et al. 2009 in Ganas and Parsons, 2009).

This study focuses on the island of Crete, which is located at the southernmost tip of the Southern Aegean (Fig. 1), at the centre of the Hellenic arc. This area appears to be one of the most seismically active areas in the eastern Mediterranean region, mainly due to the intense tectonic movements during the Late Quaternary, which led to a significant uplift in local scale and creation of extensional faults in approximately WNW–ESE and NNE–SSW directions (Ganas and Parsons 2009; Caputo et al. 2010; Shaw

and Jackson 2010; Stiros 2010). Crete's seismicity, in both the historical and instrumental periods, is mainly related to the geodynamics of the Aegean and Mediterranean plate convergence, and it is characterized by events of shallow depth (< 25 km; Papadopoulos 1982; Papanastassiou et al. 2001; Papazachos and Papazachou 2003; Kiratzi and Louvari 2003; Meier et al. 2004; Seismological Laboratory of NKUA Catalogue 2022).

Near the Hellenic Trench (HT), south of Crete, the planes for the available focal mechanism solutions of the offshore events indicate a direction of compression to be nearly NNE-SSW, while the onshore ones appear to have an almost ESE-WNW to N-S extension, associated with fractures along the main mapped faults (Fig. 2; Lyon-Caen et al. 1987; Taymaz 1990; Delibasis et al. 1999; Jost et al. 2002; Pondrelli et al. 2002; Kiratzi and Louvari 2003; Kassaras et al., 2020). Historical records indicate that three large shallow earthquakes occurred in this region (Papazachos 1990; Caputo et al. 2010, 2012; Papadopoulos 2010; Papadopoulos et al. 2022). The first occurred southwest of Crete ($M \sim 8.3$) in 365 AD, and the second occurred near Rhodes ($M \sim 8.0$) in 1303, causing a tsunami that devastated Alexandria, Egypt. The youngest was determined on February 16, 1810, in the area between Crete and Rhodes, with a moment magnitude (M_w) between 7.7 and 8.3. (Ganas and Parsons 2009; Shaw and Jackson 2010; Stiros 2010; Veliz et al. 2018; Nicol et al. 2020; Triantaphyllou and Papadopoulos 2022a).

Seismicity studies since 1990 have revealed active Quaternary faults, major compressional structures (i.e. thrusts), and other geologic features in the back-arc basin north of Crete Island. The existence of permanent stations of the regional Hellenic Unified Seismological Network (HUSN; Evangelidis et al. 2021), close to the most significant events, with the nearest stations being KNSS, PFKS (Institute of Physics of the Earth's Interior and Geohazards of the Hellenic Mediterranean University Research Center; IFEGG-HMURC), and temporary ones installed after large earthquakes, makes a significant contribution to the depth accuracy of the first hypocentral solution during the routine analysis in near real-time.

The lack of uniform and optimal coverage with seismic stations on the island and local velocity models is a major disadvantage in highlighting the earthquake distribution in areas near or along the HT. The latest important sequences led to an amelioration of the local station coverage, resulting in an improvement of the available catalogues. Improved 1-D regional models and well-localized aftershocks recorded by dense seismic monitoring networks provide a basis and baseline for more constrained seismic velocities in the region and detailed 3-D seismic imaging. Since the usage of manually selected high-quality arrival times (P and S) in 2018, a minimum 1-D velocity model of P and S wave velocities was developed in the regions of West, Central, and Eastern Crete. The recorded data improved the knowledge of the seismotectonic structures on the island and helped to provide a better 3-D body-wave velocity model through seismic tomography, examining the correlation between seismicity and the activation of the local fault pattern.

2. Geology-tectonics

The research area plays an important role in the understanding of recent orogenic processes and the development of the active geodynamic region of the South Aegean Sea. Crete consists of an autochthonous to parautochthonous rock system comprising the semi-metamorphosed unit of platy limestones and the underlying limestones, dolomites, with intrusions of shale, an allochthonous system consisting of various individual covers thrust onto the autochthonous and from the younger Neogene to Quaternary sediments.

The area consists of seven main nappes, all of which are exposed in the central part of the island, around the Psiloritis Mountains. Most of these rocks correspond to the great geotectonic units of the external Hellenic Plateau (H1) of mainland Greece. Specifically, the spine of the island is formed by the limestone of the Plattenkalk (Mani) unit, which corresponds to the metamorphic Ionian zone that formed from the Permian to the Oligocene. The Plattenkalk is either in direct tectonic contact with a mantle-forming group of rocks (Phyllite-Quartzites) that exhibits Plattenkalk-like Late Oligocene/Early Miocene high-pressure metamorphism, or it is the result of its disappearance. As in mainland Greece, the typical Pindos unit lies above the respective one of Tripolis in several places in the Psiloritis Mountains, but mostly on the southern ridge of the Asteroussia Mountains. Quaternary deposits are superimposed on all formations of both Alpine bedrock and Neogene deposits and consist of unconsolidated to slightly cohesive terrestrial, marine to lacustrine deposits, sands, cobbles, clays, and gravels (Fig. 2; Papanikolaou and Vassilakis 2010; Vassilakis et al. 2022).

From a morphological point of view, the Cretan Sea, north of Crete, consists of three main tectonic depressions, which are located between the Southern Cyclades and Crete, Karpathos and Rhodes. The main tectonic structures in the region are oriented NW-SE, to the west, and NE-SW, to the east, illustrating the curvature of the Hellenic Arc. The study area is characterized by complex geodynamics where transpressional tectonics dominates the southern part of Crete, in contrast to normal and strike-slip forms on the shallow part of the crust, producing a heterogeneous stress field (Kapetanidis and Kassaras, 2019; Vassilakis et al., 2020). In contrast, in the area to the north, between Santorini and Crete, extension is observed (Kokinou et al., 2006; Floyd et al., 2010; Kaviris et al., 2015; Papadimitriou et al., 2015). The latter shows extension in a general E-W direction, while N-forward shortening was observed in western Crete (Hatzfeld et al., 1990). Specifically, the faults in the area of Western Crete are divided into six categories, according to ten Veen and Meijer, (1998), whose directions are: a) N130°E, b) N100°E, c) N020°E, d) N075°E, e) N050°E, and f) N160°E. Angelier (1979), determined that of the fault categories mentioned above, we find mainly normal faults (a-d), while (e) and (f) refer to strike-slip faulting, with sinistral and dextral motion of the slip components respectively. On the contrary, in central and eastern Crete, extension is observed in NE-SW and NW-SE directions as mentioned by Armijo et al. (1992).

3. Phase Data And Earthquake Location

3.1 Data

The present study focuses on Crete and its neighboring regions to the west and east respectively. Several permanent stations of the Hellenic Unified Seismological Network (HUSN; Evangelidis et al. 2021) operate in this area, complemented by 4 temporary stations (CRE1–4) of the Geodynamics Institute of the National Observatory of Athens (GI-NOA), between September 27, 2021 and June 15, 2022, mainly contributing to the depth accuracy of the Arkalochori aftershock sequence for this time period (Fig. 3; Vassilakis et al., 2022; Vallianatos et al., 2022). The dataset comprises 1446 events from June 2018 to August 2022, with $M_L \geq 1.4$, recorded by at least 6 stations. Since 2007 continuous waveform recordings are being obtained from HUSN in real-time, comprising of both 1-component (1 Hz) and 3-component (either broad-band or short-period) stations, while manual arrival-time picking is being applied to obtain hypocentral locations using the SeisComP3 graphical user interface (Behr et al., 2016) and a custom regional 1-D velocity model for the region of Crete (Becker et al., 2010). In the first stage of the sequence analysis, hypocenters were located in near real-time, by employing the Hypo71 single-event algorithm (Lee and Lahr, 1975) and then they were relocated running the HypoInverse code (Klein, 2002).

3.2 1-D velocity models and location of the available data

In this study, the initial locations that were located using Becker et al. (2010) regional velocity model is presented in Fig. 4. Further filtering of this catalogue was applied in order to have more accurately determined events in the database for the minimum 1-D velocity model inversion using the mean travel-time residuals and location errors (RMS, ERX, ERY, ERZ) minimization method (Kissling et al. 1994, Chiarabba and Frepoli 1997). A selection of the initial phase data was applied, using earthquakes with at least eight (8) P- and five (5) S-wave arrival times and a maximum azimuthal gap of 180° . In total, we obtained a dataset of 635 events with an average that contained 14926 P- and 1078 S-phases.

The selected seismic phases were recorded by stations up to the distance of 200 km based on the Dmin-Dmax option provided by the HypoInverse code, targeting to exclude Pn and Sn phases. The respective model for S-waves was estimated by dividing the values of P-velocities to the V_P/V_S ratio obtained by the Chatelain method (1978). The final V_P/V_S ratio for the Cretan region was estimated to 1.78 (Fig. 5), yielding minimal RMS after several iterations for a range of different ratio values. This procedure started with two initial models: The model of Delibasis et al. (1999) and the respective one by Becker et al. (2010). Taking into account the final residuals, a minimal 11-layer one-dimensional P-wave velocity model ("P-S final model" in Table 1, black dashed line) was determined. The final V_P and V_S model results clearly converge with the ones of Becker et al. (2010) between 2.5 and 25 km in depth. At the shallowest layer (25 km), absolute velocities appear to be differentiated due to the decrease in dataset of Becker et al. (2010), resulting in fewer ray-paths of direct waves. Arrival times show a clear linear trend fit providing a V_P/V_S ratio value equal to 1.78.

Following, the initial catalogue (1446 events) was divided spatially into smaller regions using overlapping windows for further analysis and comparison with similar studies conducted in these regions (Delibasis et al. 1999; Meier et al. 2004; Nikolintaga et al. 2007; Becker et al. 2010). In total, four groups were considered: (A) Antikythera-Chania region (332 events), (B) Chania-Rethymno region (129 events), (C)

Heraklion region (445 events), (D) Agios Nikolaos-Zakros region (349 events), where the October 12th 2021, $M_L=6.2$ event took place (Fig. 4). The largest group is that of Heraklion, which contains the aftershock sequence of Alkalochori (Triantaphyllou et al. 2022; Vassilakis et al. 2022; Vallianatos et al. 2022a,b; Ganas et al. 2022). An estimation a region-specific 1-D velocity model for each group to further refine the initial hypocentral solution in each area (Table 2). This was performed after employing a subset of events with location uncertainties smaller than 2 km and having $RMS < 0.3$ s, as estimated with the obtained regional model from this study. The robustness each model was tested by several HypoInverse runs, estimating the obtained error as a change in number and the ceiling depth of the layers.

The majority of events in the final seismic catalogue lie between 0 and 18 km as shown in the map of Fig. 7. The superficial seismicity is distributed along the faults that intersect the shallow part of the crust (e.g. Ganas et al. 2017) while a significant portion of intermediate-depth seismic activity ($H \geq 40$ km) is located near Antikythera island (West), North of Heraklion (central part), and between Eastern Crete, Kassos and Karpathos islands (East). This observation shows a fair correlation between the rapid increase in velocity at depths greater than 32 km in these areas (Fig. 7) and recent tomographic studies in the regions along the Hellenic Arc. (Spakman et al. 1988; Papazachos and Nolet 1997; Li et al. 2003; Vallianatos et al. 2006; Kokinou and Vallianatos 2008; Kokinou et al. 2009; Sachpazi et al. 2016; Ventouzi et al. 2018). In the westernmost region (Antikythera-Chania), the velocity model of the crust varies between 5.8 and 6.7 km/s for approximately 32 km, while the respective values for the neighbouring area of Rethymno are between 5.6 and 6.8 km/s for crustal thickness of less than 31 km.

In the central part of the island, the velocity model is divided into more layers (6 layers) with approximately the same values of P velocity and crustal thickness, but a lower V_P/V_S ratio (1.75) (Fig. 8). Further to the east, in the region of Zakros-Karpathos, the velocity model is simplified to approximately four (4) layers and crustal thickness of 32–33 km, a result that converges to the respective ones of Nikolitanga et al. (2007) for the same territory. The updated catalogue shows a better concentration of the seismic events along main fault zones of the study area and smaller spatio-temporal errors (Fig. 8).

The velocity model of Zakros-Karpathos was tested in the recent earthquake sequence in the Eastern Cretan Sea (24 December 2022–15 January 2023) where more than 130 events were located with a better concentration near some of the most significant local fault sources (Caputo et al. 2010; Sboras et al. 2011; Caputo et al. 2012) and yielded improved error uncertainties than the initial model of Nikolitanga et al. (2007) (Figs. 10–11).

Table 1

Statistics of temporal errors (RMS), location uncertainties and mean depth for the regional and group-specific velocity models in Crete.

Model (This study)	Regional model	Antikythera- Chania	Rethymno	Heraklion (Central Crete)	Agios Nikolaos- Zakros
RMS (sec)	0.48	0.49	0.59	0.36	0.26
ERH (km)	1.44	1.74	1.79	0.99	0.93
ERZ (km)	4.74	3.88	2.82	2.15	1.74
Mean Depth (km)	33.70	15.33	11.45	25.23	12.67

4. Earthquake Tomography

Earthquake tomography is an efficient for the studies of the Earth's detailed interior. The main goal of this tomographic study is to determine the structure of the crust and the upper mantle of Crete's region, and neighbouring areas that were described in the previous section. Local seismic tomography has been successfully applied to reveal velocity structures associated with fault activation, fluid circulation along major asperities, magma transport and main boundaries on convergent plate boundaries (e.g. Karakonstantis et al. 2019; Matsubara et al. 2019; Koulakov et al. 2020; Gordeev et al. 2022; Yaroshenko et al. 2022). In this study, the inversion of body-wave travel-time data was based on LOTOS scheme as described by Koulakov (2009).

For the inversion procedure, 1208 manually revised events located on the island of Crete and its neighbouring regions were included during the first step. The dataset comprised 23544 P and 19451 S arrival-times, with at least 12 phases per earthquake and ratio of S to P residual smaller than 2.5 (Fig. 12). The inversion algorithm provides two different choices: a) inversion for both V_P and V_S (V_P-V_S scheme) using the respective travel-time residuals (dt_P and dt_S), and b) inversion for V_P and V_P/V_S ratio (V_P-V_P/V_S scheme) using dt_P and differential residuals, dt_S-dt_P . In the current study, tomographic inversion was applied for both schemes, in order to obtain more information on the body-wave anomalies (Koulakov, 2009; Jaxybulatov et al., 2011). Sensitivity analysis for the available dataset was performed by applying the checkerboard synthetic test (Humphreys and Clayton 1988). This method uses alternating anomalies of positive and negative velocity perturbations, on an initial one-dimensional gradient model, evenly distributed in a checkerboard pattern throughout the model.

The available tomographic software necessitates as input data the longitude and latitude of the available network stations and the arrival times from locally recorded seismicity. The coordinates of the hypocenter and the time of origin time are not essential, essential as they are determined during the implementation of calculations. However, if preliminary hypocentral temporary hypocenter locations are available, they can be utilized to reduce the processing time of the operations. In addition, a generic 1-D velocity model

and a set of input parameters for performing the convergence iteration steps, comprising parameterization, grid dimensions and damping parameter, were considered (Koulakov, 2009).

The model parameterization of the velocity field should be able to delineate, according to the local characteristics, the shape and perspective of heterogeneities. A nodal representation was used, since the velocity field reconstructed by a three-dimensional grid does not adopt any specific geometry of heterogeneities (Toomey and Foulger 1989). The interdistance of the grid nodes was kept significantly smaller than the expected size of the anomalies (< 5 km) to weaken the distortion of the resulting models due to the grid configuration. The optimal grid mesh has been determined considering the stations/events geometry. In addition, to further reduce the impact of the model parameterization on the solutions, the inversion was repeated utilizing several grid orientations (e.g. 0°, 22°, 45° and 67°). The results obtained were aggregated via a 3-D model of the absolute P- and S-wave velocities by simple averaging.

4.1 Resolution tests

Towards defining the fidelity area in this study, we used the checkerboard method (Humphreys and Clayton, 1988) as an indicator of the resolution and errors associated with the inversion. This method uses alternating anomalies of positive (fast) and negative (slow) velocity perturbations, evenly spaced throughout the model in a checkerboard pattern. Generally, ray-path distribution, model parameterization, and smoothing determine data resolution (Lees and Crosson, 1989). Among the limits imposed of such a test, is the fact that the extent of near-vertical alteration of structure is difficult to be fully recognized due to a structure (checkerboard), where the diagonal elements in the vertical plane are strongly associated with dominant ray directions (e.g. Rawlinson et al., 2014).

Synthetic models are initialized with cells that are the same size as those expected to exhibit anomalies. For the applied method, spiked regions are defined, with $\pm 10\%$ variability in velocity structure compared with a 1-D velocity reference model. Random noise of 0.25 s for P-wave and 0.40 s for S-wave arrival-time picks, derived by real data inversion, was added to the residuals. Subsequently, travel times for the paths between source and receiver were calculated. This procedural method corresponds to the real observation system that utilizes 3-D ray tracing following the principles of the bending algorithm. After the synthetic travel times are computed, the coordinates, source origin time, and velocity model are no longer available. The restoration of the synthetic model is conducted in the same fashion as actual data processing, such as the enhancement of the 1-D model and the absolute source position. The same free parameter values as in the real case are used to explore the area of fidelity for the values obtained. For the real data 3-D tomographic inversion, the set of variables that provides a greater area of confidence and recreates the model of checkerboard anomalies is used. Errors in the data, including mis-picks, mis-locations, and incorrectly determined ray paths, control the inversion variance. In order to characterize the limitations of our model, we utilized three different sets of anomaly dimensions for the horizontal tests (15x15 km², 20x20 km² and 30x30 km²). The variations (%) of body-wave velocity anomalies ($\pm 10\%$) and the V_P/V_S ratio distribution are presented in Figs. 13–15 and Figs.S8-S10 (supplementary material), at depths of 10,

40, 50, 60 and 70 km. The first group of anomalies ($15 \times 15 \text{ km}^2$) was apparently poorly resolved (depth slices of 60 and 70 km) or not resolved at all (depth slice of 40 km), while the other sets ($20 \times 20 \text{ km}^2$ and $30 \times 30 \text{ km}^2$) provide reasonable resolution at shallow depths throughout the rest of the study area (10 km) and it is fully resolved for depth ranges between 20–70 km. These models are reconstructed in the area between Cretan Sea and the southern shores of the island ($34.8^\circ\text{--}36.0^\circ\text{N}$, $23.6^\circ\text{--}26.5^\circ\text{E}$).

More particularly, in the significant part of the research region ($35.0^\circ\text{--}35.8^\circ\text{N}$, $23.7^\circ\text{--}25.75^\circ\text{E}$), the anomalies of the 15 km grid size are fairly resolved in the depth slices of 10–20 km depth for the P- and S-wave velocity models, while the respective ones of 20 and 30 km are reasonably restored in all depth intervals (10–70 km). The azimuthal gap of the available seismological stations and the relatively sparse epicentral distribution cause horizontal smearing in the NW and SE parts of the study area. In terms of vertical resolution, tests for the performed cross-sections are shown (Fig. S11).

A similar process was performed in order to examine the vertical resolution. In the vertical sections a separate checkerboard model was created with squared anomalies of $20 \times 20 \text{ km}^2$. In every cross-section, velocity sign is altered at the depths of 20, 40, and 60 km. In total, six cross-sections were performed. Each pair of profiles was performed from the western (A, B) to the eastern (F, G) part of the island (Fig. S11). The changeover from negative and positive velocity anomalies is fairly reconstructed for a great part of the research region in all the sections, for a depth range between 10 to 50 km, with minor horizontal smearing along ray routes. Sections B and E seem to be less resolved than section A, D and F, particularly in the edge of the profiles (Figs. S12–14). Previous studies have shown that poorer vertical resolution than horizontal one may be related to a trade-off involving source and velocity parameters (e.g. Kasatkina et al. 2014). Both horizontal and vertical synthetic investigations showed that the absolute amplitudes of the body-wave anomalies were up to 4% lower than the initial values of the checkerboard grid.

4.2 Tomographic inversion results

The inversion approach began with initial location and improvement of the 1-D velocity model, producing a mean RMS travel-time residual of 0.35 s after five trials. This is a 38% reduction over the best 1-D velocity model, which had an initial average RMS of 0.51 s. Furthermore, the average P- and S-residuals were lowered from 0.34 s and 0.45 s, respectively, to 0.24 s and 0.27 s (Table 2). The generated P- and S-velocity anomalies with respect to the 1-D starting model are depicted in six horizontal slices at depths of 10 to 70 km in Figs. 16 and 17. The equivalent V_P/V_S ratio measurements are shown in Fig. 18. The evaluation of the acquired values is restricted to model portions with sufficient reconstruction of the checkerboard model, as illustrated in Figs. 11–13. For P- and S-waves, velocity fluctuations in the shallow depth layers range between +10 and 10%, while the V_P/V_S ratio fluctuates between 1.58 and 2.05. According to synthetic modeling tests, the velocity structure is resolvable for depths ranging from 10 to 50 km. This fact, together with the presence of intense ray coverage down to a depth of 50 km, lends confidence to the 3-D inversion results and makes interpretation more viable.

Table 2

Average absolute values of P- and S-wave residuals and their cumulative reduction percentage during the inversion of experimental data.

Iteration	P-residual (s)	P-residual reduction (%)	S-residual (s)	S-residual reduction (%)
1	0.342	0.00	0.453	0.00
2	0.280	18.16	0.323	28.77
3	0.261	23.77	0.293	35.31
4	0.249	27.19	0.281	38.10
5	0.242	29.16	0.273	39.74

A ~ NE-SW trending discontinuity between slow and fast body-wave velocity anomalies is seen in the Kissamos and Chania gulfs in the western part of the island (Antikythera-Chania) at superficial depths, possibly linked to the local faults of N-S direction (Figs. 16 and 17). This shallow and broad slow V_P anomaly down to 20 km depth characterizes the collapsed zones. A similar pattern of body-wave velocity (P, S) anomalies is observed in deeper slices (10–40 km) south of Crete (east of Gavdos Island), with slow and fast deviations from the reference 1-D model to the north and south of the Ptolemy Trench respectively.

A quasi WNW-ESE trending zone of slow V_P anomalies and low V_P/V_S ratio values (1.58–1.62) emerges at shallow depth near Matala, on the southern tip of Crete (Figs. 16 and 18). This anomaly is consistent with the mean distribution of the Neotectonic faults that divides the Mesozoic Pindos Unit basement from the Neogene and Quaternary deposits. Further to the east, some strong lateral velocity perturbations are seen primarily at shallow depths (~ 10 km) in the central part of the study area (Heraklio, Malia) and in deeper depth slices (30 and 40 km). In particular, a discontinuity of fast to the north and slow to the south P-wave velocity anomalies is found at shallower depths (10 km) within the activated area during the Arkalochori earthquake sequence, which matches with the location of most of the concentrated aftershocks for this time frame (2021–2022). This is anticorrelated with the comparable fast velocity S-wave perturbations in the same area, leading to low values of the V_P/V_S ratio (~ 1.58–1.65), as shown in Fig. 16. This could be ascribed to the local high-density fault pattern of NNE-SSW to NE-SW direction of Central Crete (Caputo et al. 2010; Ganas et al., 2011, 2017) and the thick Neogene to Quaternary sediments (Bonneau 1984; Vidakis et al. 1989; Vassilakis 2006; Vassilakis et al. 2022). In the eastern part of the study area, in Mirambelos gulf, between Agios Nikolaos and Zakros region, strong E-W to ENE-WSW oriented negative P- and S-wave velocity anomalies predominate at depth slices of 40 to 70 km. This part of the research region contains intermediate-depth seismicity ($H \geq 40$ km), which can be attributed to subduction processes and upward migration of plunging lithosphere fluids. The model lacks fidelity south of Crete at depth exceeding 50 km and can only be regarded suggestive; therefore, it is not discussed.

To investigate the velocity anomalies in depth, Fig.S11 depicts six, vertical cross-sections performed at N-S and W-E directions, displaying V_P (Fig.S16 and V_S (Fig.S17) perturbations, as well as the distribution of absolute V_P/V_S ratio values (Fig.S18). The first two imply reveal slow velocity anomalies (P and S) in the shallow part of the crust (5–20 km) in the central part of the Chania region, while the fast ones are situated south and north, respectively (section B). The latter (Fig S18) has a significant V_P/V_S ratio (> 1.82) that goes down to 15–20 km. In sections C (S-N) and D (W-E), in the central region of Crete, show an anomaly of negative perturbations mostly in the shallow part of the profile (< 10 km), and the succession between fast (west) and slow (east) velocities exposes a major structural discontinuity of approximately N-S direction in section D. The main discontinuity between the body-wave (P, S) velocity anomalies and the V_P/V_S ratio distribution arises in the eastern end of the research area at depths shallower than 25 km in the last two cross-sections (E and F). In particular, at a distance greater than 65–70 km (section E), a negative perturbation exceeding + 10% and V_P/V_S ratio appear to be linked to the local fault pattern of NNE-SSW direction in Zakros and the bending of the HT towards the NE.

5. Discussion

In the framework of the present study, we've obtained local 1-D velocity models on both regional and local scales throughout the island as part of the current investigation, as well as a regional 3-D model from body wave travel time tomography (V_P , V_S , V_P/V_S). During the first phase of this procedure, we have determined a regional model for the region of Crete in order to proceed with relocation and extraction of the travel times in order to move on to the next phase of the Tomographic Inversion. Following that, the area was divided into four sub-regions (Antikythera-Chania, Rethymno, Heraklio, and Agios Nikolaos-Zakros) from West to East Crete in order to compare with previously available models of those regions and examine the local seismicity properties.

5.1 Velocity Models

In Figs. 5–8 best 1-D regional and local velocity models are displayed, along with their error statistics. The histograms depict the temporal (RMS) and spatial (ERX, ERY, and ERZ) error distributions of the available data between the study's starting and final models. The newly developed velocity models provide us with a better concentration of seismic events, more realistic hypocenter depth ranges, and a proper understanding of local structure activation.

The regional model (Fig. 5) utilized in this research shares several characteristics with those developed by Delibassis et al. (1999) and Becker et al (2010). The primary difference is that the first model produced substantially shallower events than the second one for the Arkalochori seismic series (Vallianatos et al., 2022). In particular, the stronger events of that sequence ($M_w = 6.0$ and $M_w = 5.3$) were located at very shallow depths ($H < 5$ km), which looked improbable in terms of earthquake physics and the area's history.

The local 1-D models in the first two areas to the west of the study area (Antikythera-Chania, Rethymno) have a quite simplified structure since they are composed of four and three layers respectively. The existence of a thick layer of recent deposits (Neogene to Quaternary) above the Alpine basement of different age and lithology, and the top of the low-angle subducting lithosphere (Fig. 2) may explain why the velocity structure in central Crete is divided into numerous layers. The model suggests a simpler layout with four strata in the eastern portion of the island. The V_P/V_S ratio of the 1-D model developed in the first two sub-areas of Western Crete is higher (1.80) than that of Central and Eastern Crete (1.75), which could be related to the closer location of the intermediate depth events reflecting an upward migration of fluids resulting from the dehydration of subducting sediments within the oceanic origin Tethys plate (Peacock, 2001; Hacker et al., 2003a, Hacker et al., 2003b; Kato et al., 2010; Pasten-Araya et al., 2018), like comparable cases (Husen and Kissling, 2001; Barnes et al., 2010; Collings et al., 2012; Bangs et al., 2015). As displayed in Fig. 17, the subduction of the Tethys oceanic lithosphere under the Aegean microplate results in velocities from both plates overlapping, making it difficult to address each plate individually in local one-dimensional models.

5.2 Earthquake Tomography

Variations in the velocity of body waves (V_P , V_S) and the V_P/V_S ratio have been mainly noticed in tomograms of 10 and 20 km in Figs. 14–16. The contrast between fast and slow anomalies is mainly linked to the local fault pattern, which produces a comparable distribution of V_P/V_S ratio (Fig. 16). These anomalies can be found across the area. This image is mostly influenced by local surface geology at shallower depths (e.g. Koulakov et al. 2010). The high V_P/V_S ratio reported in the eastern part of the depth slice of 10 km can be attributed to fluid-saturated sediments, non-consolidated soils, and/or cracked subducted crust. The results in sections E-F (Figs.S16-18) for depths ranging from 15 to 20 km may indicate a possible upward movement of fluids connected to this mechanism. The limited vertical resolution doesn't allow us to have much confidence on those results.

One of the key indicators that local faults influence the results at these depth slices is the shallow discontinuity between slow and fast body-wave velocity anomalies in the Kissamos and Chania gulfs (Antikythera-Chania region). One of the notable characteristics of collapsed zones is the broad slow V_P anomaly down to 20 km depth. A similar finding in Matala supports this hypothesis, where a similar finding supports this interpretation. At shallow depths near the southern tip of Crete, a WNW-ESE-trending zone of slow V_P anomalies and low V_P/V_S ratio values (1.58–1.62) arises (Figs. 14 and 16). We can see a significant contrast in velocity perturbations at shallow depths (10–20 km) in the central part of the research area in Central Crete (Heraklio, Malia). At shallow depths (10 km), a discontinuity of fast to the north and slow to the south P-wave velocity anomalies in scattered directions is discovered within the active area of the Arkalochori aftershock sequence. The formed anticorrelation between slow P- and fast S-wave velocity perturbations form low values of the V_P/V_S ratio (~ 1.58 – 1.65). This result supports the high-density fault pattern in this area as tectonic studies have shown. The unfractured sections of the crust, on the other hand, are clearly indicated as significant body-wave (P, S) velocity perturbations and a

low V_P/V_S ratio in depth slices of 10 and 20 km near the transition between Central and Eastern Crete (35.00°N-35.23°N, 25.85°E-26.08°E).

The steady increase in the V_P/V_S ratio ($V_P/V_S > 1.88$) below 40 km depth in the back-arc region north of Crete (Cretan Sea) confirms the notion that these observations are connected to subducting oceanic plate dehydration. Below 50 km, there is no sharp contrast between the body-wave anomalies ($\pm 4\%$), but the rapid increase of V_P in both 1-D and 3-D models can be indicative of the transition to oceanic Moho because the values are higher than 7.7–7.9 km/s, which can be associated with upper mantle velocities (Gailler et al. 2007; Font et al. 2013; Araujo 2016; León-Ríos et al. 2019). The most significant variations are in the area's southern edge and are mostly related to the HT and its change in direction from the central to the eastern part of Crete. We can notice the contrast between slow and fast deviations from the reference 1-D model to the north and south of the Ptolemy Trench in this area of the region. Finally, in depth slices of 40 to 70 km in the Agios Nikolaos-Zakros region, some substantial E-W to ENE-WSW directed negative body-wave velocity anomalies seem to be the dominant feature. This portion of the study area includes a significant part of the overall intermediate-depth seismicity ($H \geq 40$ km). This result can be attributed to subduction processes and upward migration of plunging lithosphere fluids.

6. Conclusions

Based on the travel times of chosen events between 2018 and 2022, we obtained local 1-D velocity models on both regional and local scales and performed tomographic inversion over the central part of the Hellenic Arc using data from the Hellenic Unified Seismological Network (HUSN). The resulting models are comparable on both local and regional scale on the structure of the crust and upper mantle in Crete and neighboring regions.

A new database of seismic events with well-constrained hypocentral parameters has been compiled based on the newly formed velocity models in order to perform the tomographic inversion. The distribution of P- and S-wave velocity anomalies, as well as the V_P/V_S ratio, supports the existence of a thin crust in the back-arc area, in contrast to a thicker one under Crete, which constitutes the accretionary prism in this portion of the Eastern Mediterranean Sea.

The regional distribution of seismic activity, body-wave velocity anomalies (P, S), and V_P/V_S ratio values revealed:

- A complex shallow structure in Crete's central region;
- A smoother and more continuous image in deeper slices (> 40 km).

Further work is needed in order to see the effect in Seismic Hazard of Crete, as well as to increase the dimensions of the research areas in order to examine in detail the geometry of the Wadati-Benioff zone in the Southern Aegean, the properties of the seismogenic layer, and have a comparison with studies conducted in the past and similar ones in other parts worldwide.

Declarations

Ethics approval and consent to participate

Research is not involving human participants, human material, or human data

Consent for publication

The authors consent for publication of our details under the Creative Commons Attribution License 4.0.

Availability of data and material

Materials described in the manuscript, including all relevant raw data, will be freely available to any scientist wishing to use them for non-commercial purposes, without breaching participant confidentiality. All dataset is freely available to the NKUA website:

http://www.geophysics.geol.uoa.gr/stations/gmapv3_db/index.php?lang=en

Conflict of interest

Professor Filippas Vallianatos (NKUA), the second author of this paper, is on the Editorial Board of the Journal of Seismology (JOSE). Other than that the authors declare that they have no other competing interests.

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Authors' Contributions

Andreas Karakonstantis have made the acquisition, analysis, and interpretation of data; wrote the main manuscript text, and prepared all the figures. Filippas Vallianatos designed the workflow and reviewed and edited the manuscript.

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Figures

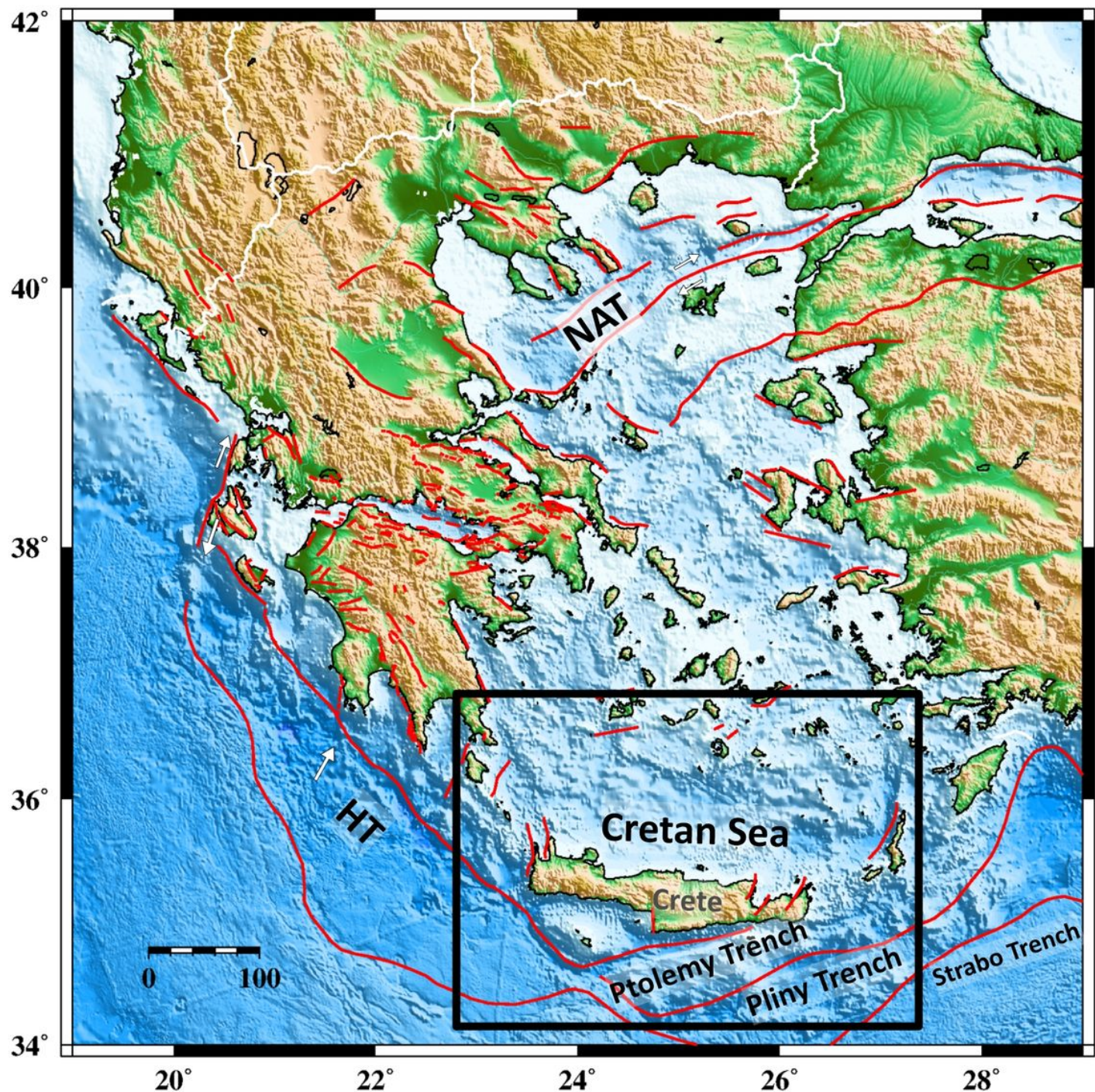


Figure 1

Main tectonic features in Greece and Western Turkey. The black rectangle contains the study area. Abbreviations-HT: Hellenic Trench; NAT: North Aegean Trough. Fault traces (red lines) derived by Pavlidis et al. (2010), Caputo et al. (2012) and Ganas et al. (2017).

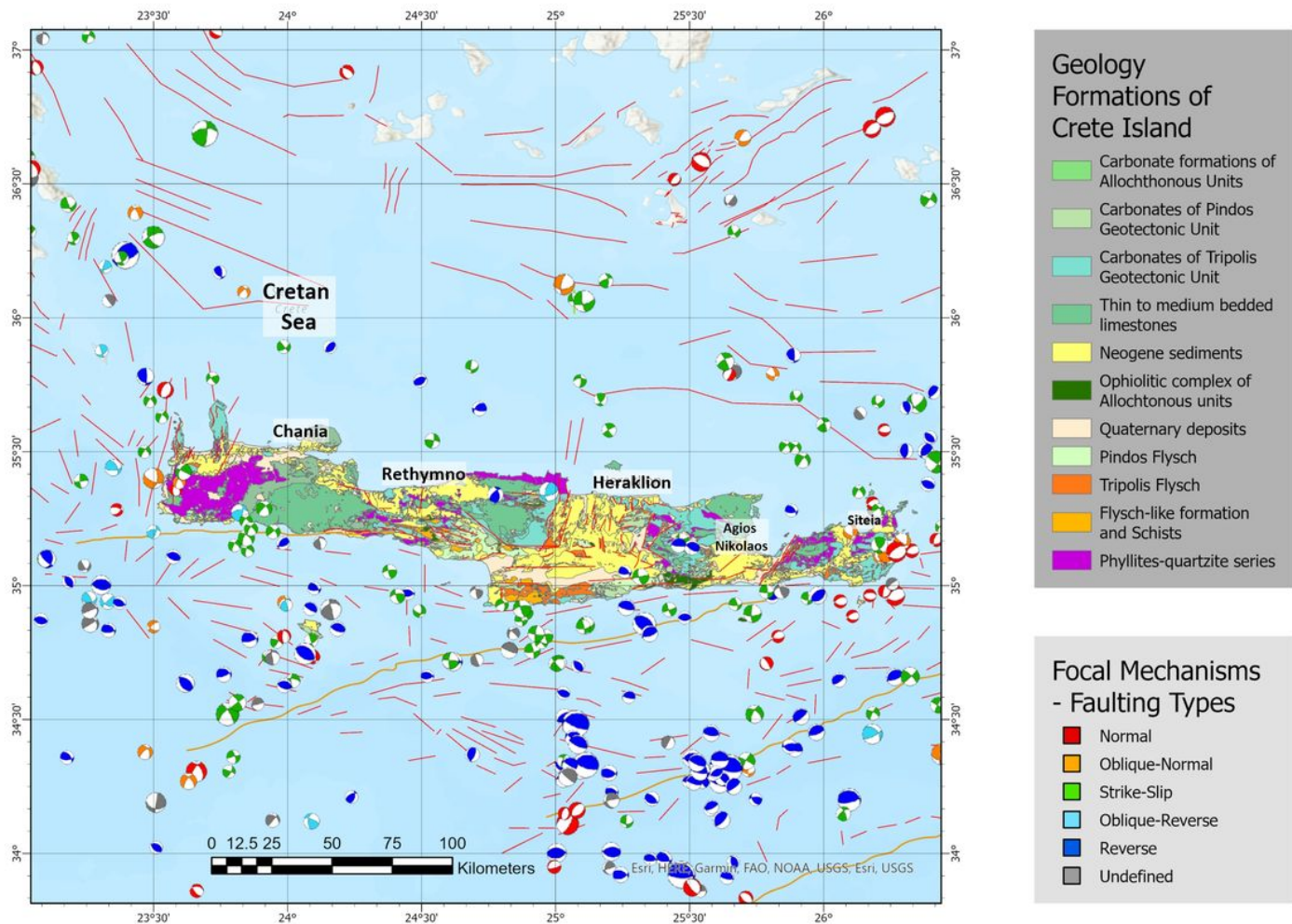


Figure 2

Geological features in the area of study (Crete and neighbouring regions). Main geologic and tectonic elements derived by the prefecture of Crete (geoportal.apdkritis.gov.gr), Papanikolaou and Vassilakis (2010), Vassilakis et al. 2022 and Ganas et al. (2017). Beachballs indicate focal mechanisms of earthquakes (Mw>4.0) from the compilation of Kapetanidis and Kassaras (2019) and sources referenced therein.

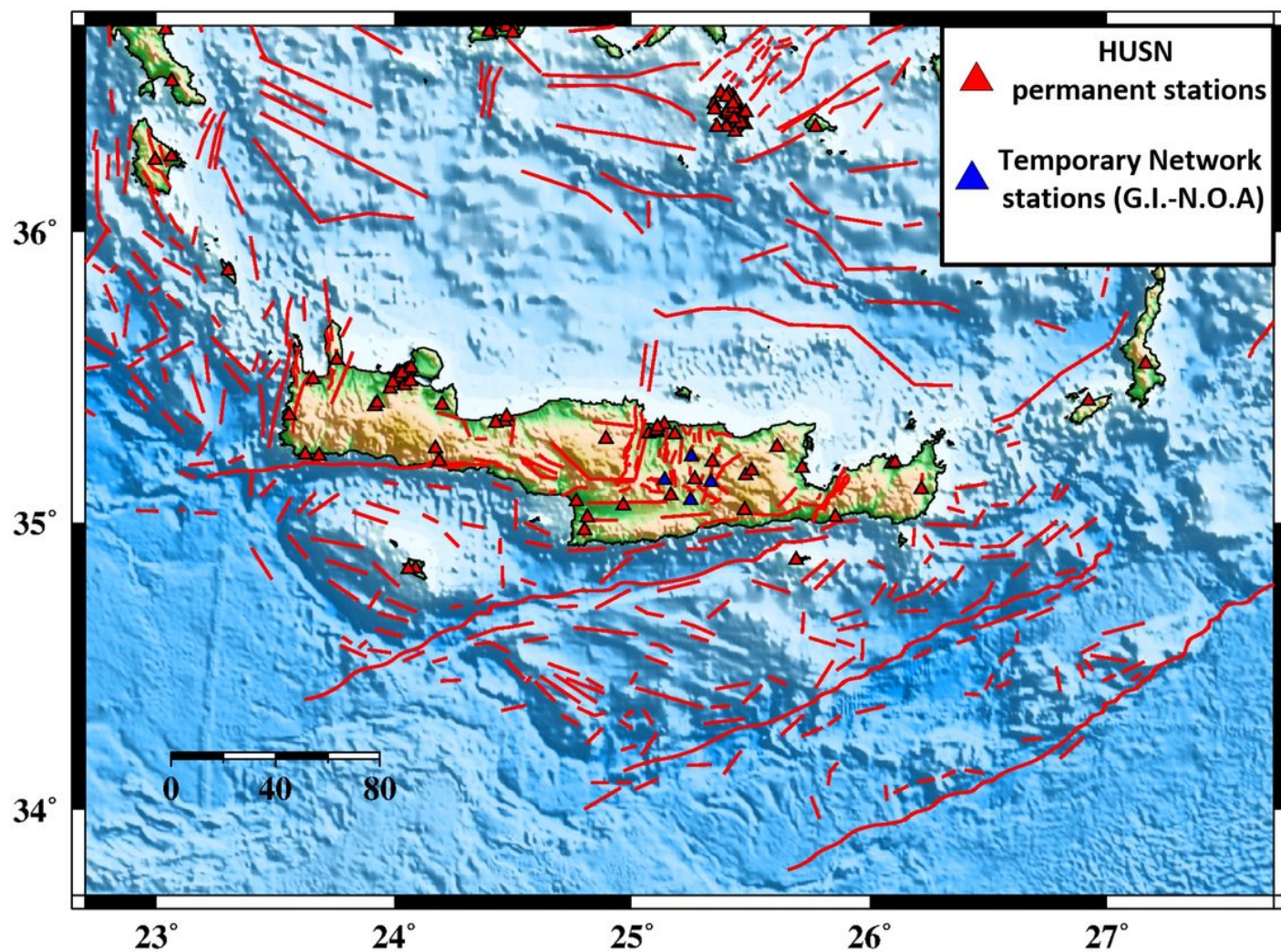


Figure 3

Location of the permanent (red triangles) HUSN stations and temporary installed stations (blue triangles) by the G.I.-N.O.A after the initiation of Arkalochori earthquake sequence (September 27, 2021).

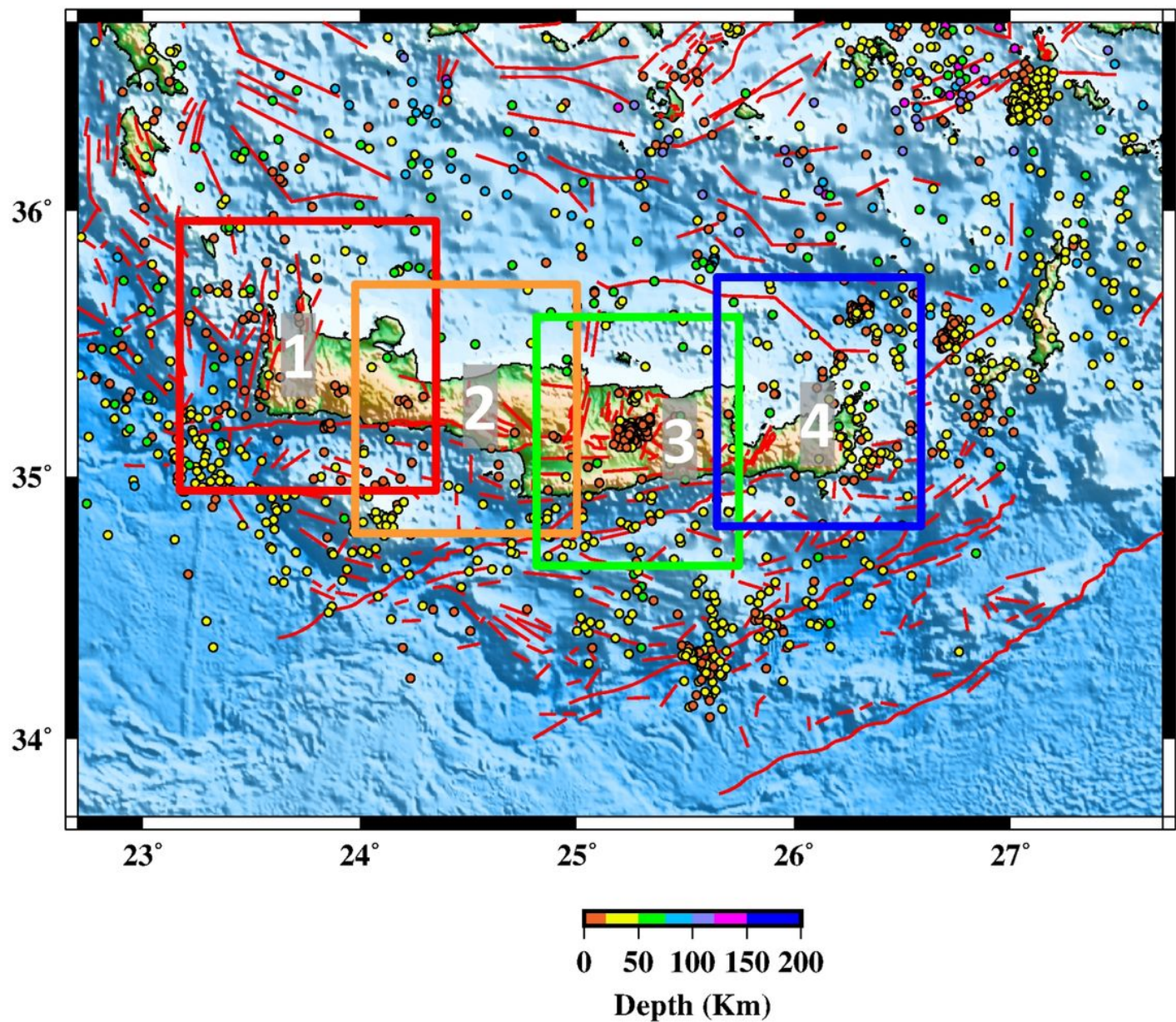


Figure 4

Located earthquakes using as initial 1-D velocity model the one derived by Becker et al. (2010). The area is divided into four (4) sub-areas: (1) Antikythera-Chania (2) Rethymno (3) Heraklion (4) Agios Nikolaos-Zakros.

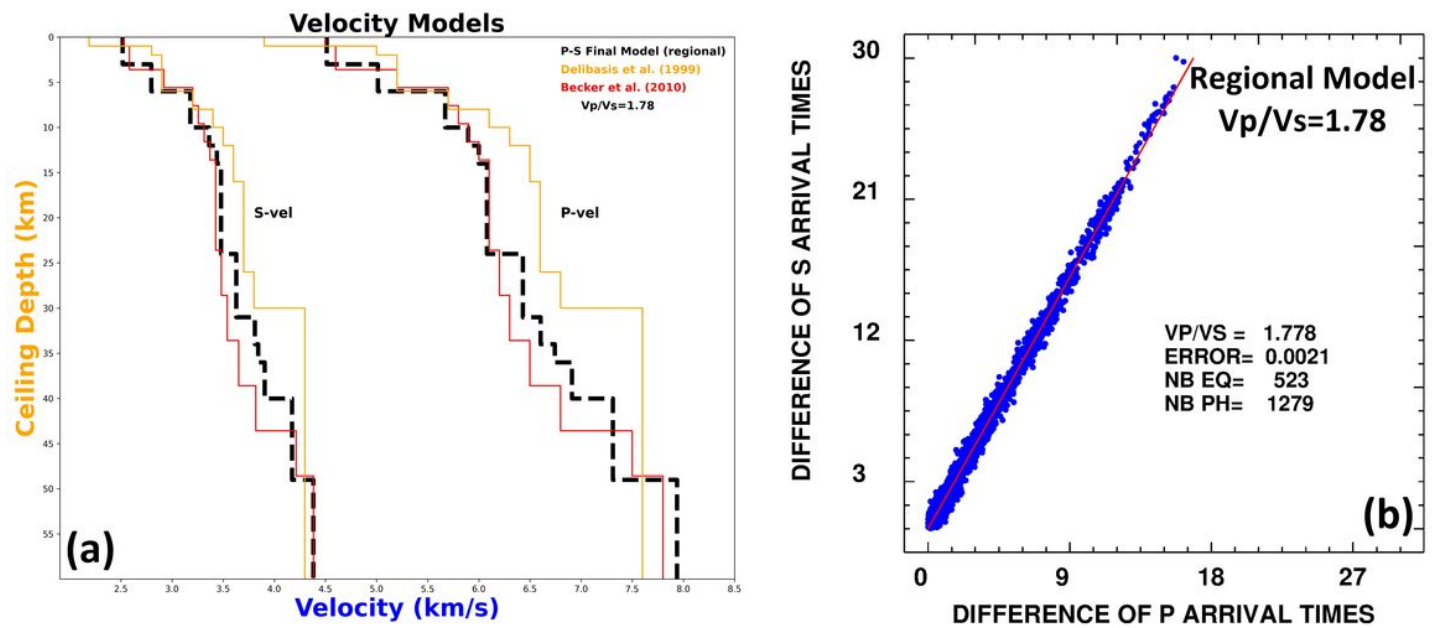


Figure 5

a) 11-Layer 1-D models of V_p and V_s determined by the mean travel-time residuals and location errors technique and the starting models (Delibasis et al. 1999; Becker et al. 2010) b) V_p/V_s ratio obtained by the Chatelain (1978) method.

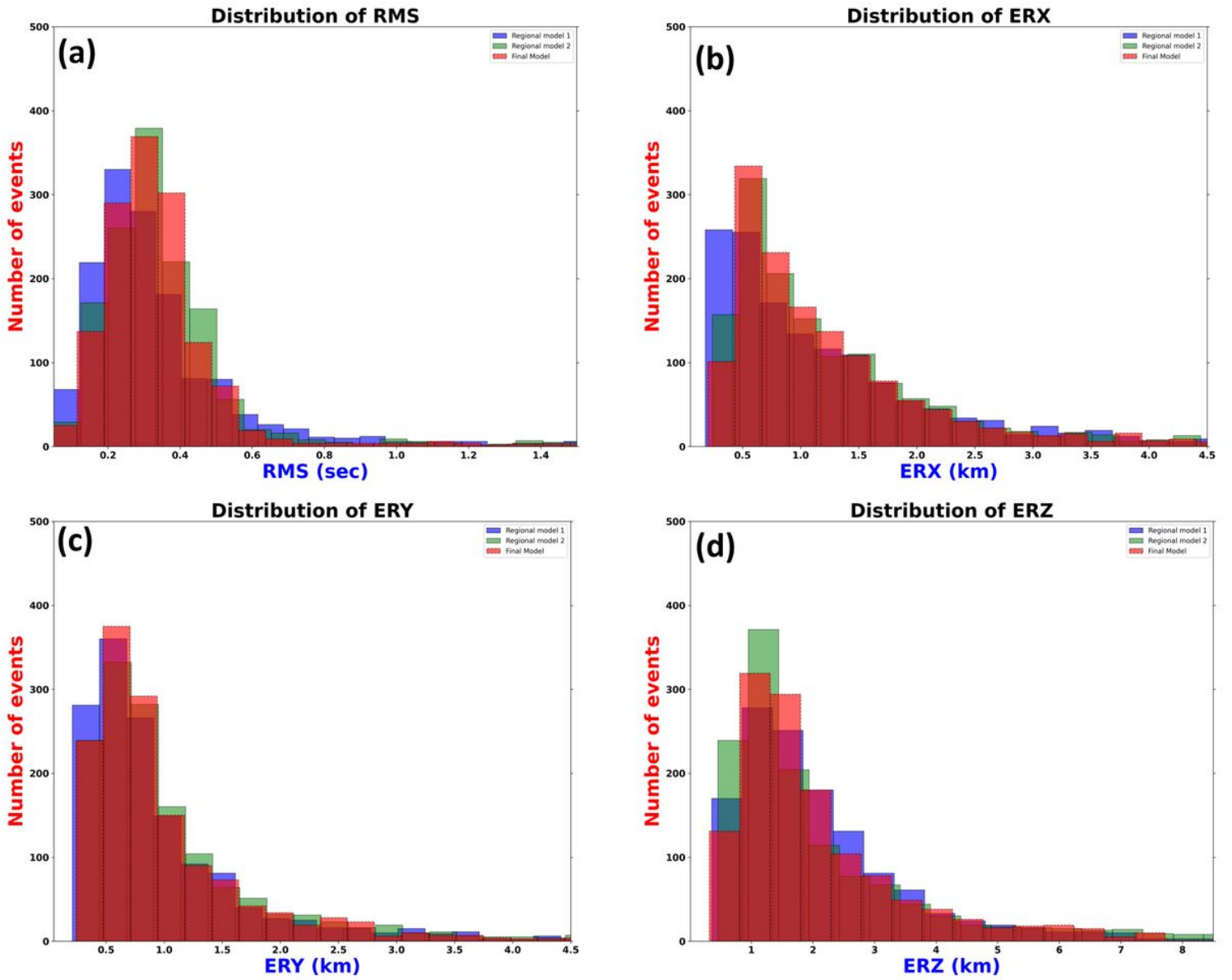


Figure 6

Histograms of the error statistics between the Model 1 (blue; Delibasis et al. 1999), Model 2 (green; Becker et al. 2010) and the final 1-D model (red) obtained in this study.

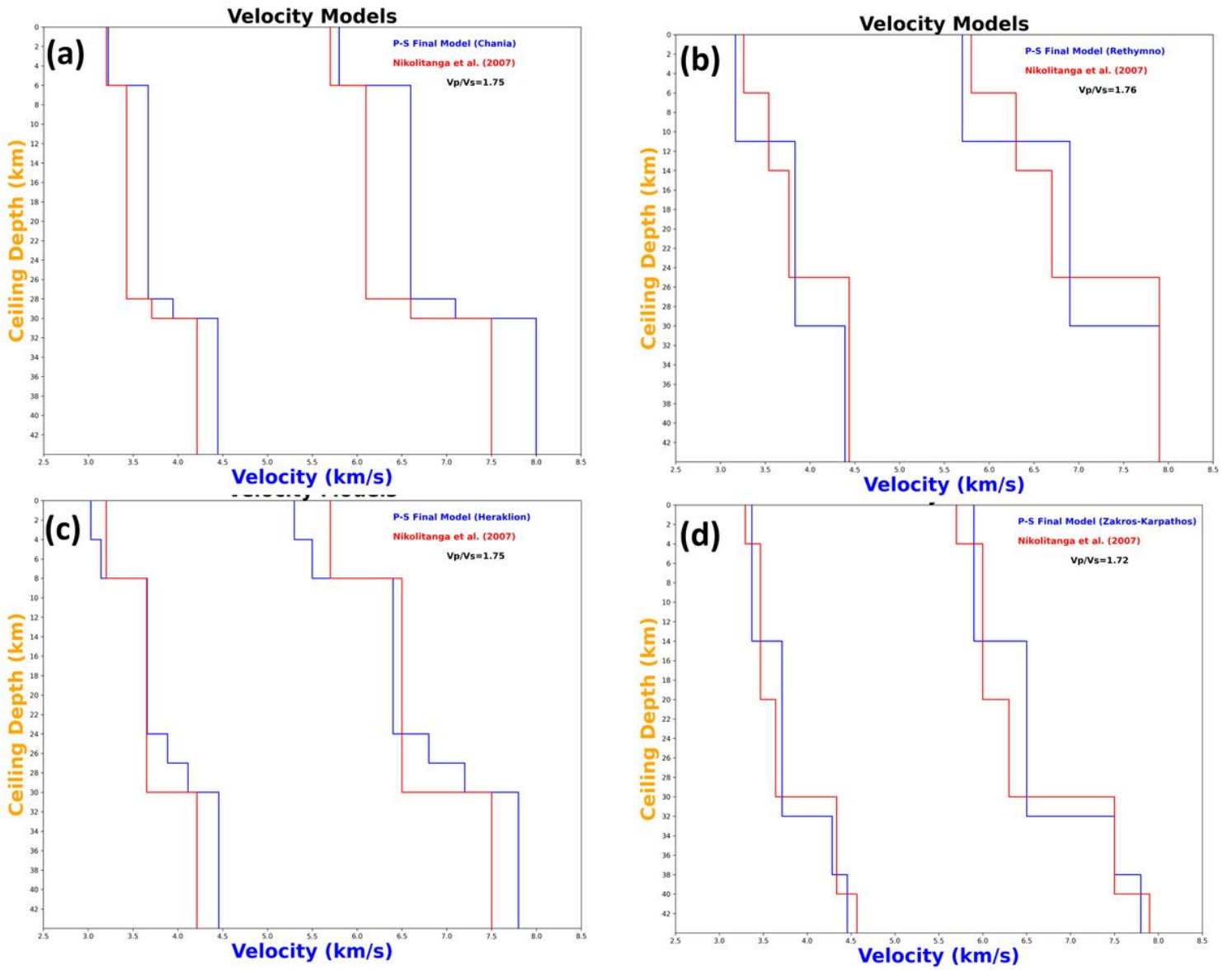


Figure 7

Local 1-D velocity models of V_p and V_s determined by the mean travel-time residuals and location errors technique in (a) Antikythera-Chania (b) Rethymno (c) Heraklion (Central Crete) and (d) Agios Nikolaos-Zakros (Eastern Crete), using as starting models the ones derived by Nikolitanga et al. (2007) study in red line and the final models obtained by this study in blue one.

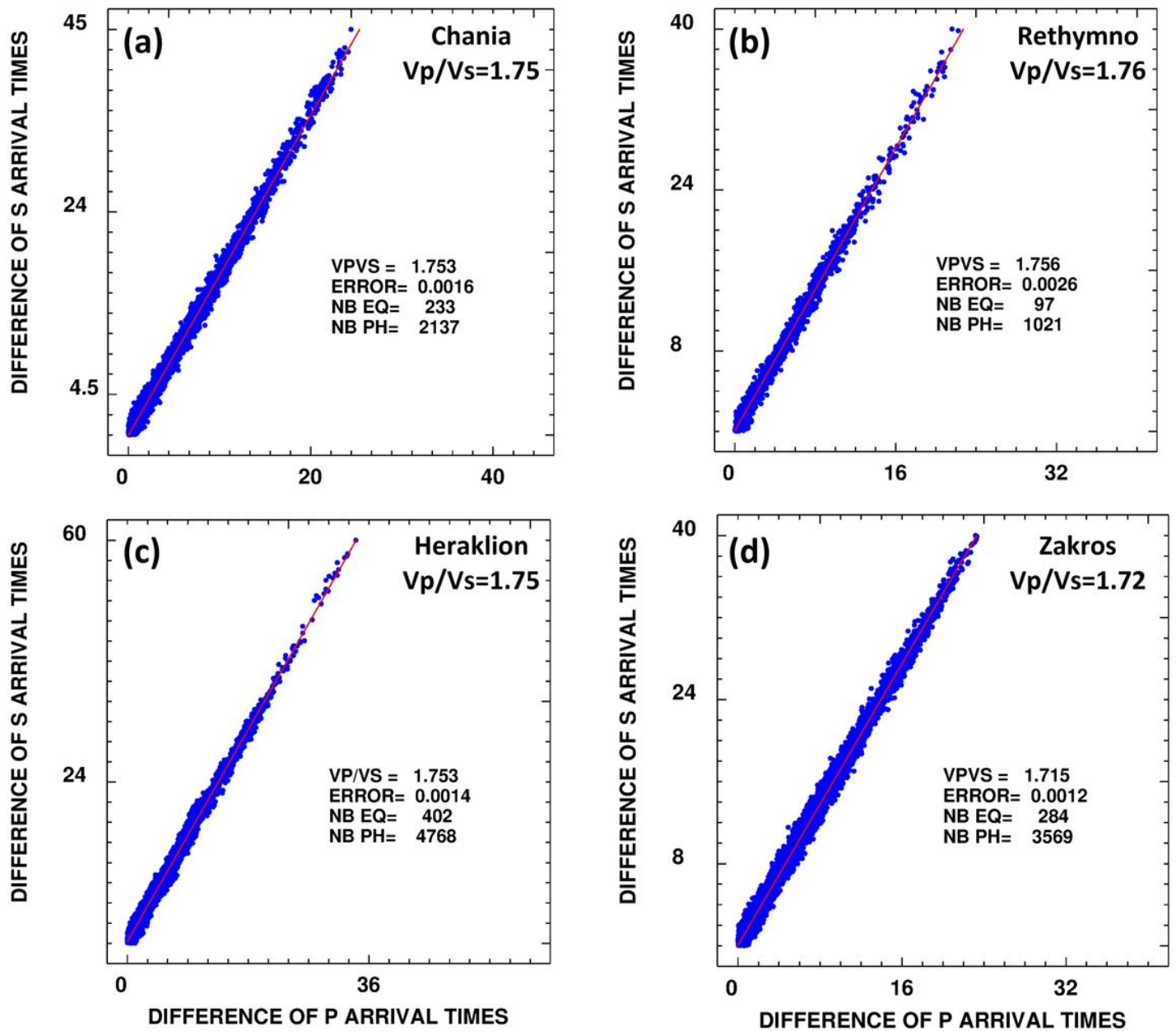


Figure 8

V_p/V_s ratio in (a) Antikythera-Chania (b) Rethymno (c) Heraklion (Central Crete) and (d) Agios Nikolaos-Zakros (Eastern Crete), obtained by the Chatelain (1978) method.

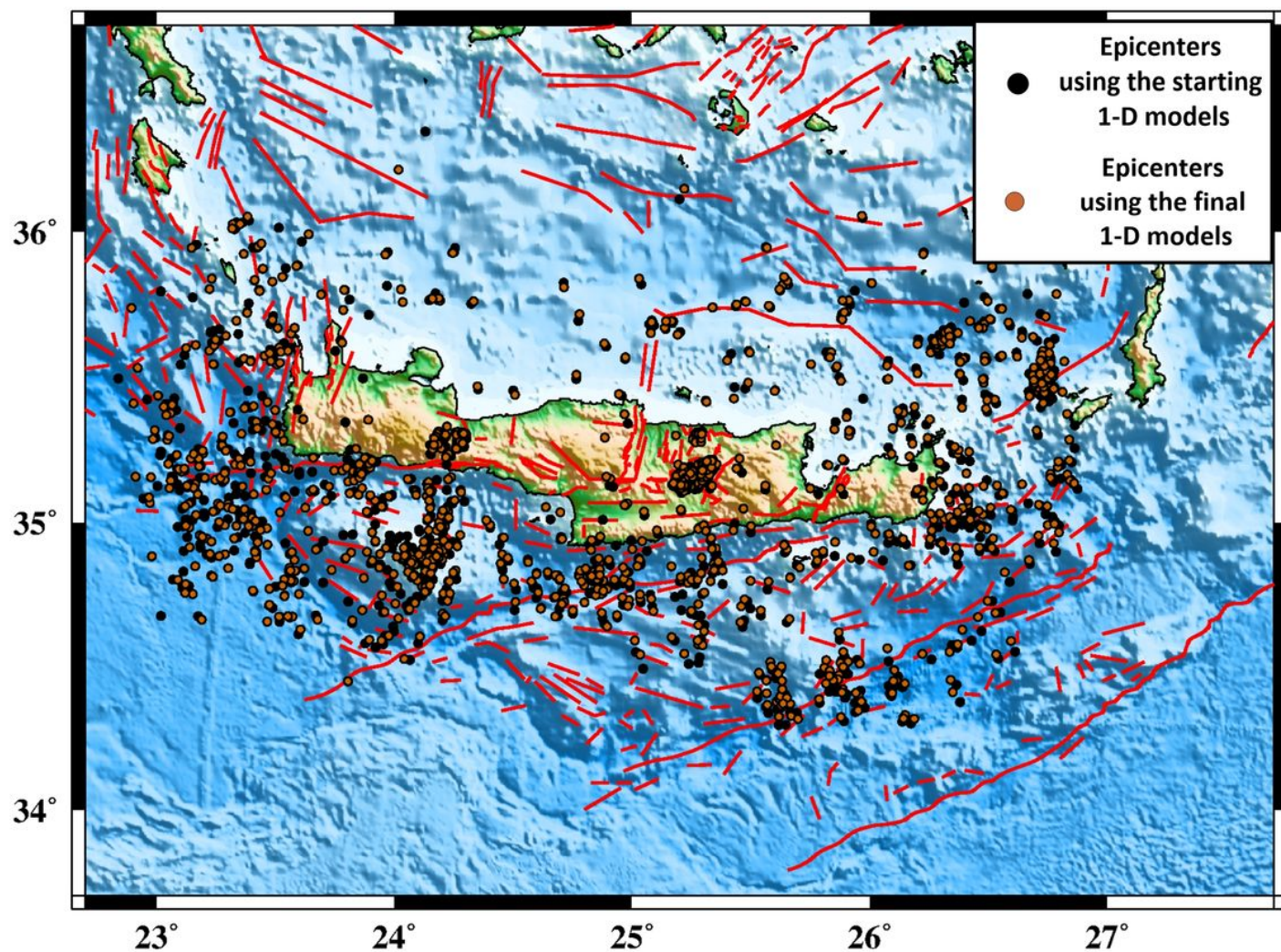


Figure 9

Preliminary (black circles) and final (orange circles) epicenters using the initial and final 1-D velocity models respectively.

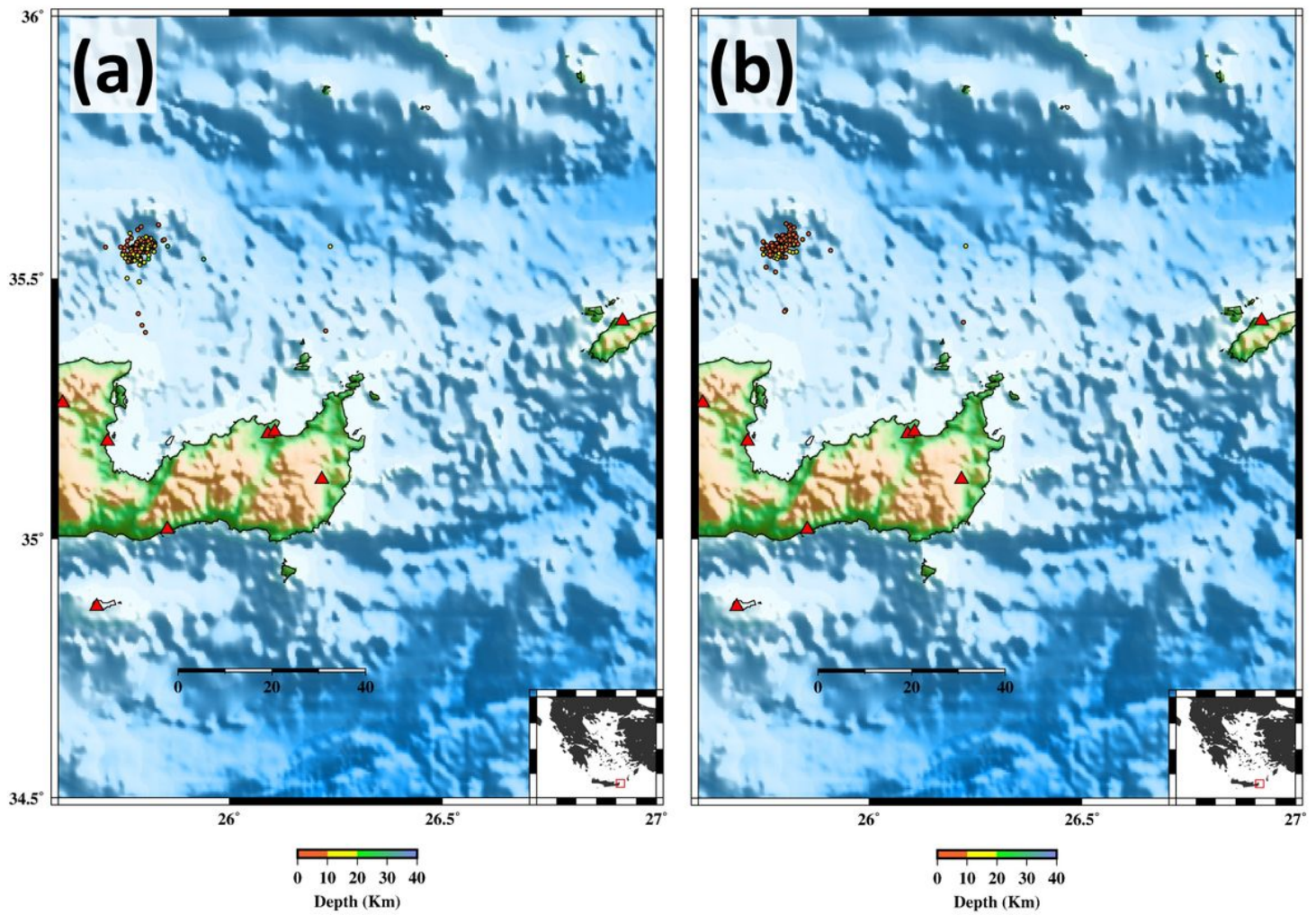


Figure 10

a) Preliminary and b) final epicenters using the initial and final 1-D velocity models respectively.

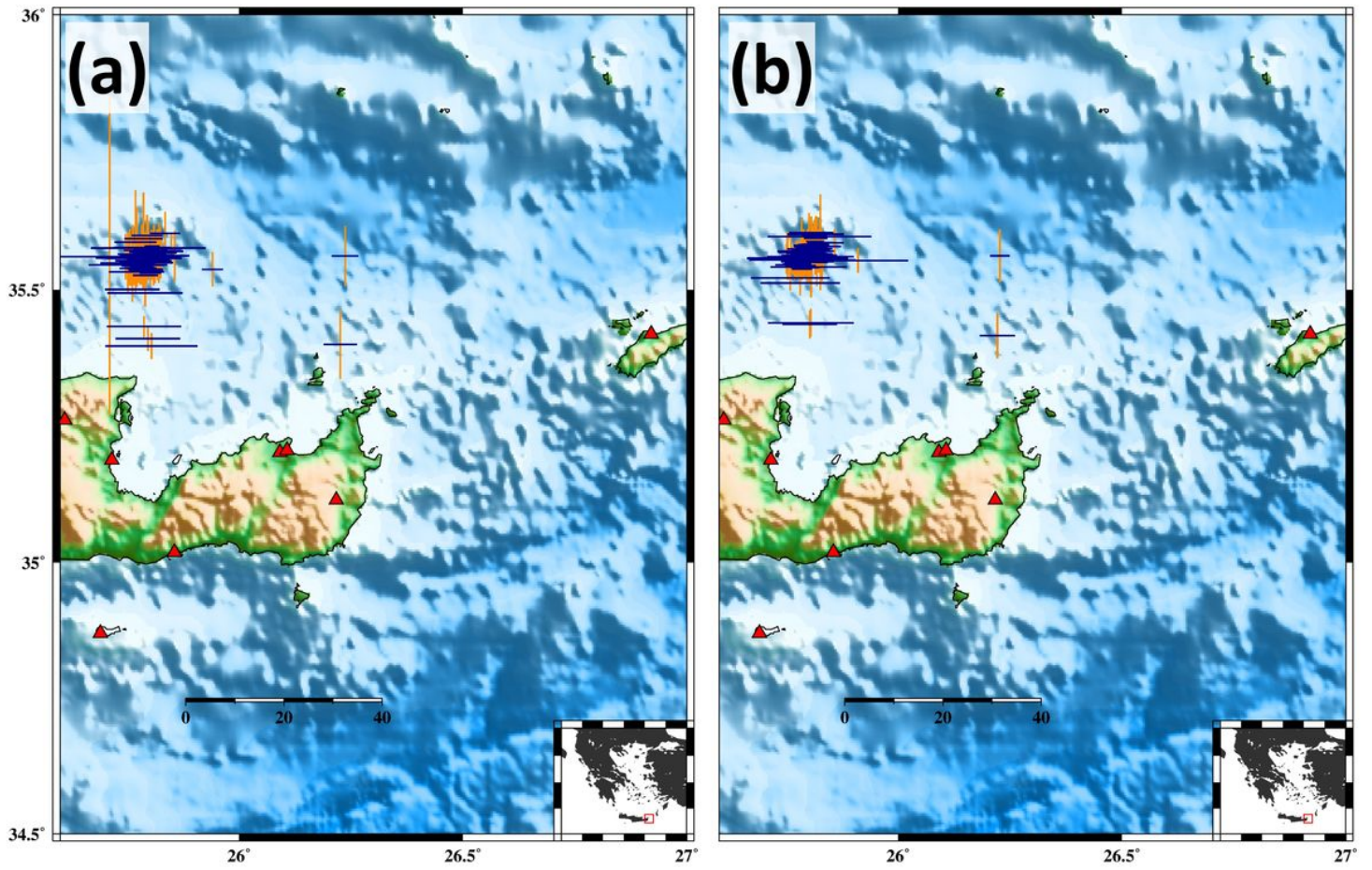


Figure 11

Spatial errors in x (blue) and y (dark orange) axis of a) preliminary and b) final catalogue using the initial and final 1-D velocity models respectively.

Rays Distribution

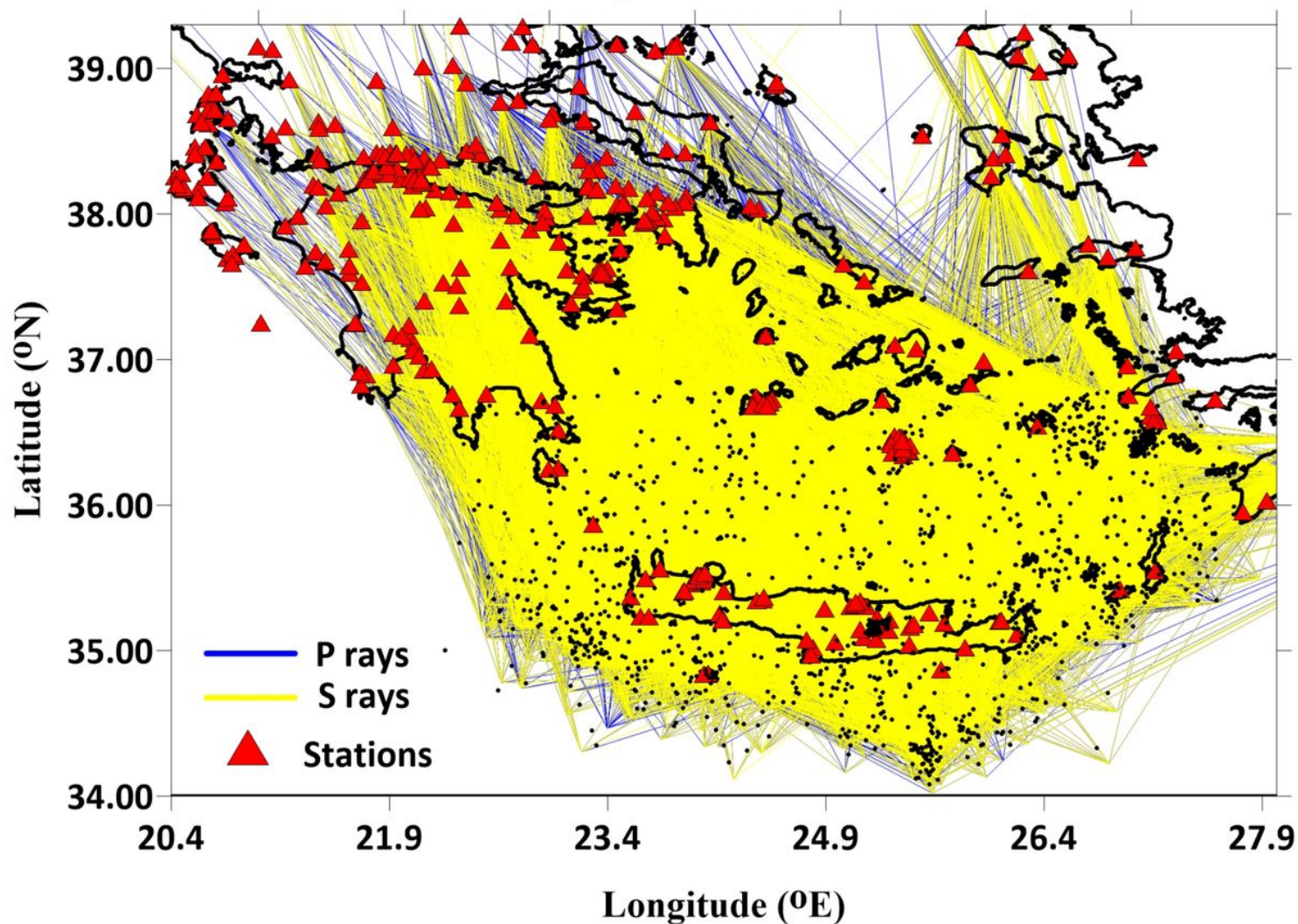


Figure 12

Total P- (blue) and S-ray (yellow) distribution. Red triangles indicate locations of the used stations. The selected seismicity used in this work ($M \geq 1.4$) during 2018-2022 is presented by black dots.

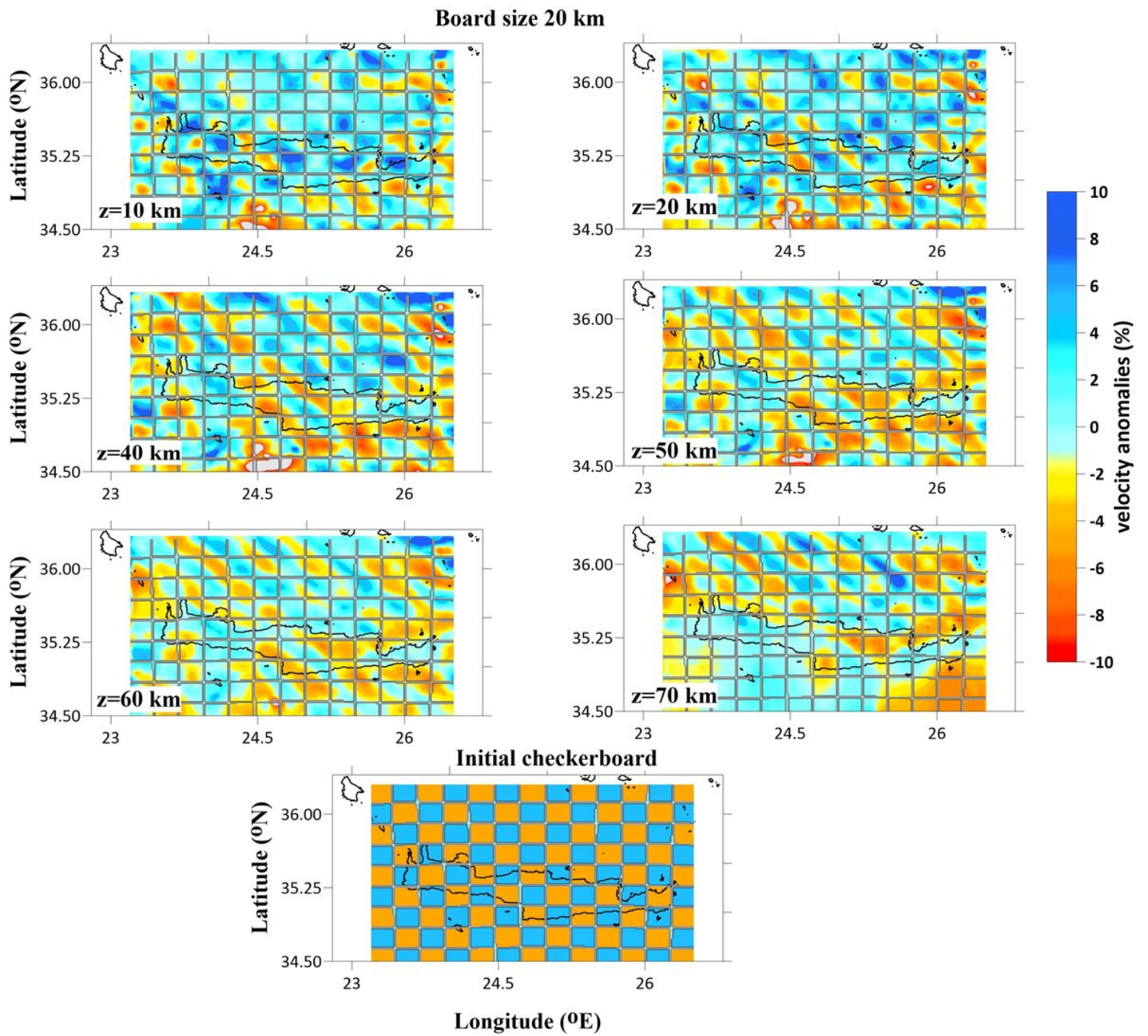


Figure 13

Initial checkerboard model (bottom row) with anomaly cell size of $20 \times 20 \text{ km}^2$ and reconstructed of P-wave velocity anomalies for the depth slices of 10, 20, 40, 50, 60 and 70 km.

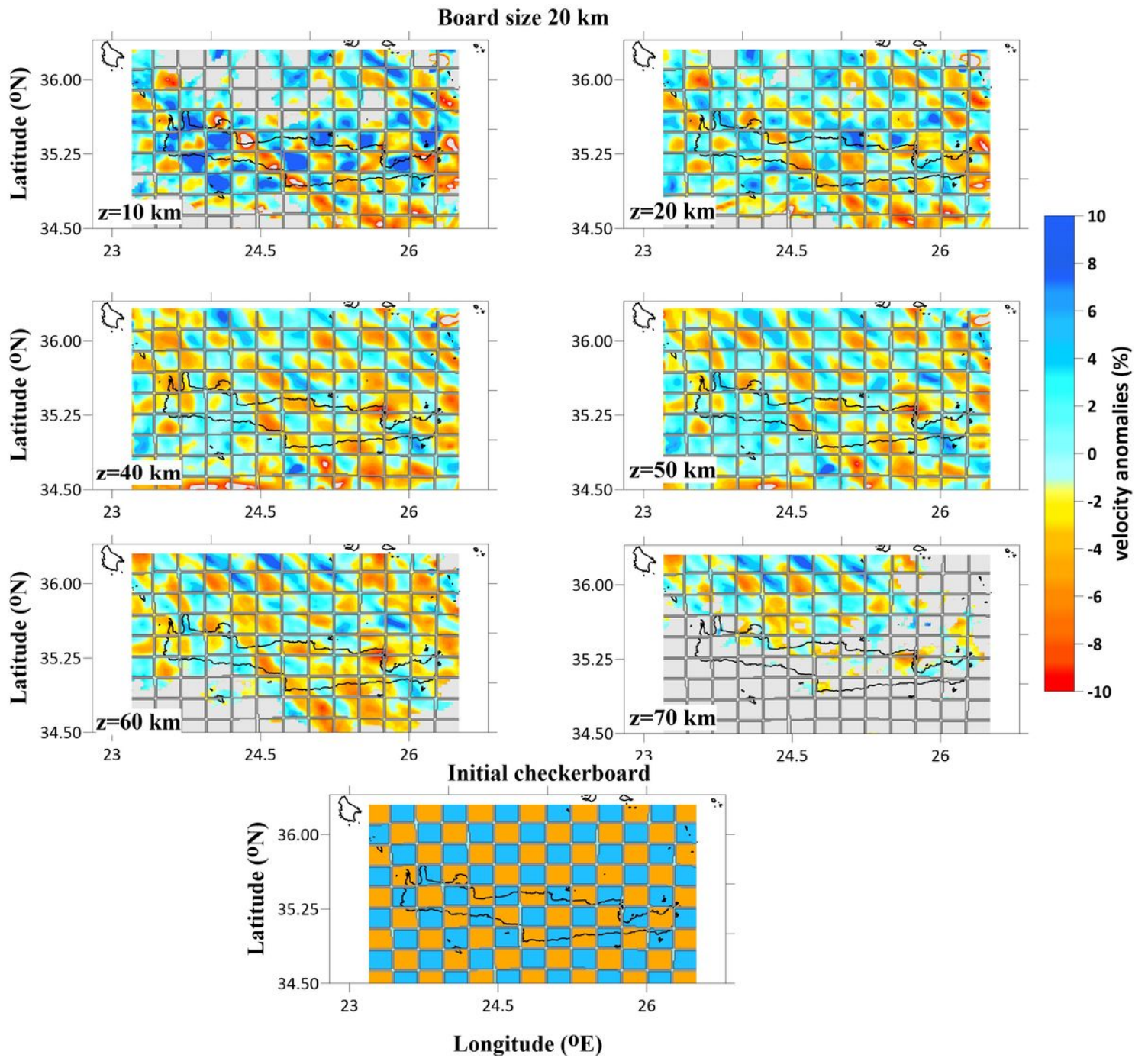


Figure 14

Initial checkerboard model (bottom row) with anomaly cell size of $20 \times 20 \text{ km}^2$ and reconstructed of S-wave velocity anomalies for the depth slices of 10, 20, 40, 50, 60 and 70 km.

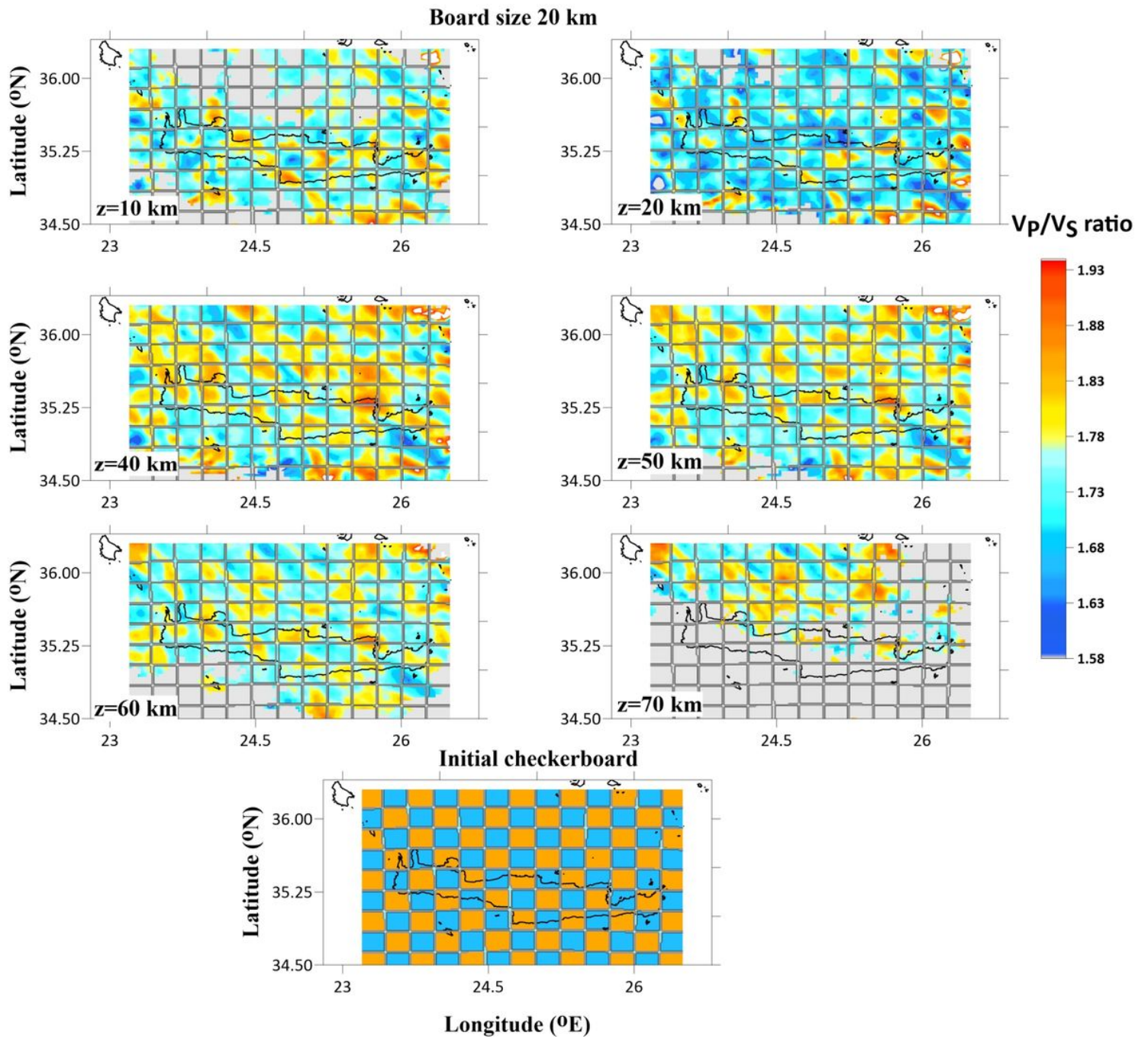


Figure 15

Initial checkerboard model (bottom row) with anomaly cell size of $20 \times 20 \text{ km}^2$ and reconstructed of V_p/V_s ratio distribution for the depth slices of 10, 20, 40, 50, 60 and 70 km.

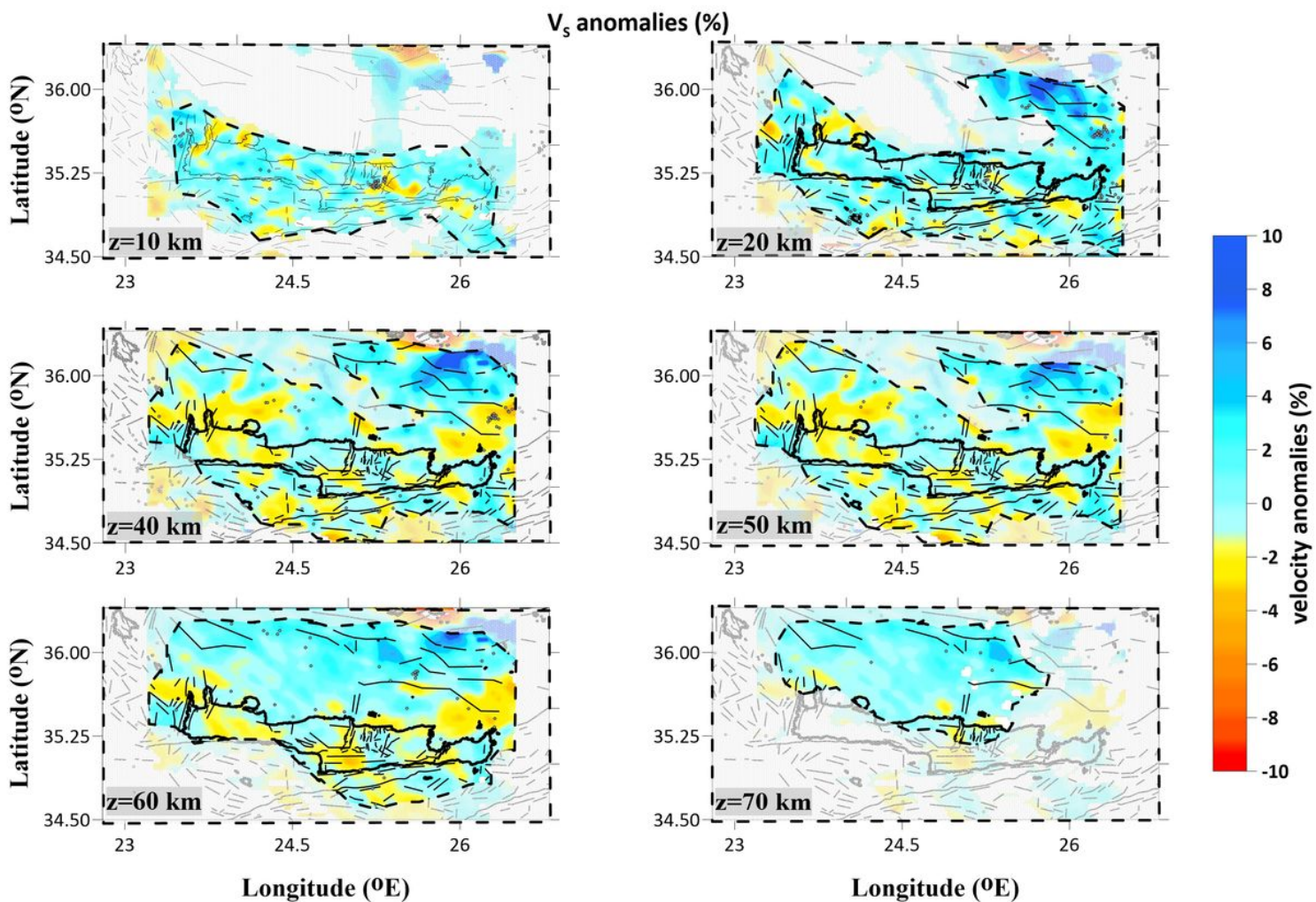


Figure 17

Tomograms of lateral V_s (%) variations at 10, 20, 40, 50, 60 and 70 km. Areas with lower resolution are masked. Black circles indicate the recorded seismicity. Toponyms and tectonic elements as in Figure 1.

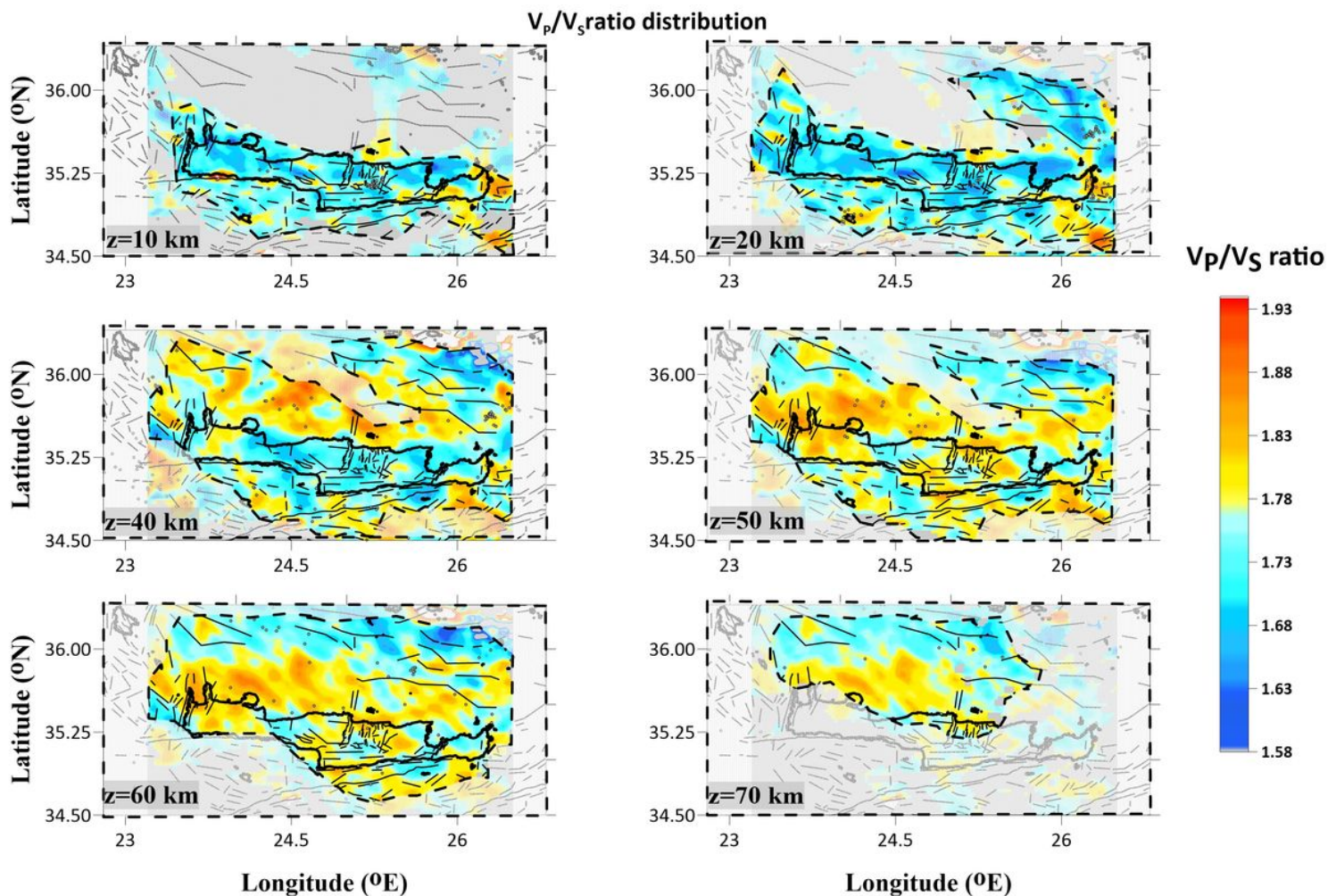


Figure 18

Tomograms of lateral V_p/V_s ratio distribution at 10, 20, 40, 50, 60 and 70 km. Areas with lower resolution are masked. Black circles indicate the recorded seismicity. Toponyms and tectonic elements as in Figure 1.

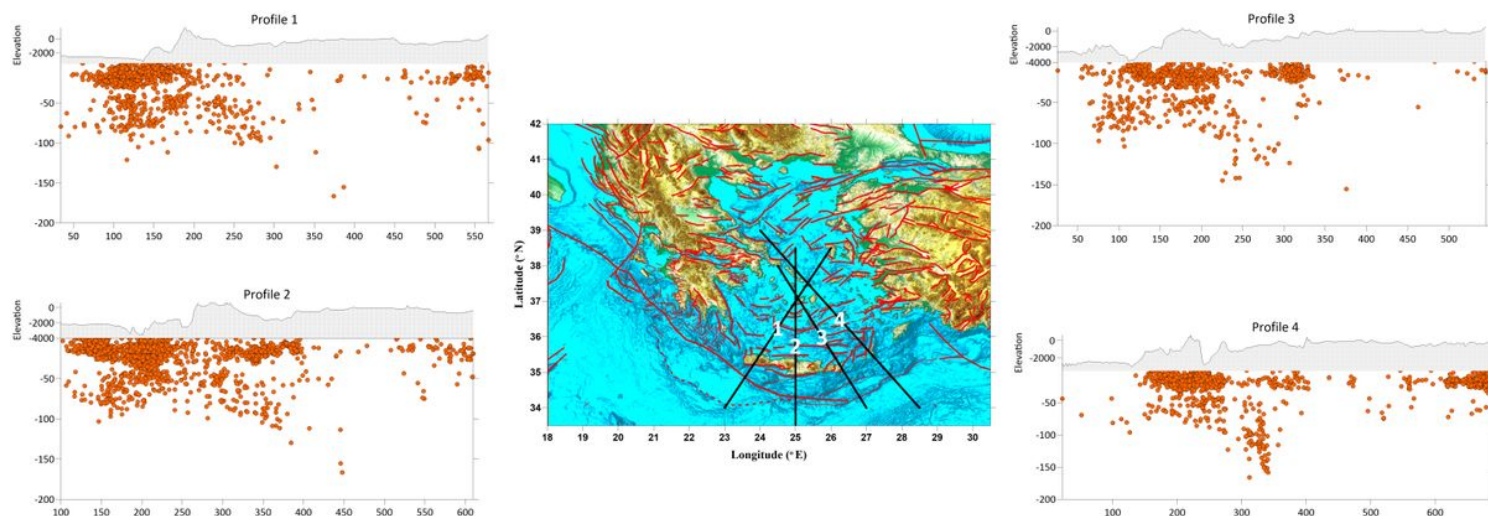


Figure 19

Hypocenter distribution of the 1996-2017 seismicity recorded in the Southern Aegean region in four (4) different cross-sections of SW-NE, S-N and SE-NW direction (Karakonstantis 2017).

Supplementary Files

This is a list of supplementary files associated with this preprint. Click to download.

- [appendix.zip](#)