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Comparison of gravimetric and seismic constraints on the structure of the Aegean lithosphere in the forearc of the Hellenic subduction zone in the area of Crete

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Abstract

The island of Crete is located in the forearc of the Hellenic subduction zone, where the African lithospheric plate is subducting beneath the Eurasian one. The depth of the plate contact as well as the internal structure of the Aegean plate in the area of Crete have been a matter of debate. In this study, seismic constrains obtained by wide-angle seismic, receiver function and surface wave studies are discussed and compared to a 3D density model of the region.

The interface between the Aegean continental lithosphere and the African one is located at a depth of about 50 km below Crete. According to seismic studies, the Aegean lithosphere in the area of Crete is characterised by strong lateral, arc–parallel heterogeneity. An about 30 km thick Aegean crust is found in central Crete with a density of about 2850 kg/m^3 for the lower Aegean continental crust and a density of about 3300 kg/m^3 for the mantle wedge between the Aegean crust and the African lithosphere. For the deeper crust in the area of western Crete two alternative models have been proposed by seismic studies. One with an about 35 km thick crust and another one with crustal velocities down to the plate contact. A grid search is performed to test the consistency of these models with gravimetric constraints. For western Crete a model with a thick lower Aegean crust and a density of about 2950 kg/m^3 is favoured. The inferred density of the lower Aegean crust in the area of Crete correlates well with S-wave velocities obtained by surface wave studies.

Based on the 3D density model, the weight of the Aegean lithosphere is estimated along an E–W oriented profile in the area of Crete. Low weights are found for the region of western Crete.
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Keywords: Hellenic subduction zone; Gravity field; 3D modelling; Seismic methods

1. Introduction

The Hellenic subduction zone (HSZ) is the seismically most active region in Europe (Fig. 1). It is a part of a convergent plate boundary between Africa and Europe. The tectonics of the region are dominated by the retreating active margin and extension of the Aegean domain. On a broader scale the kinematics of the region can be described

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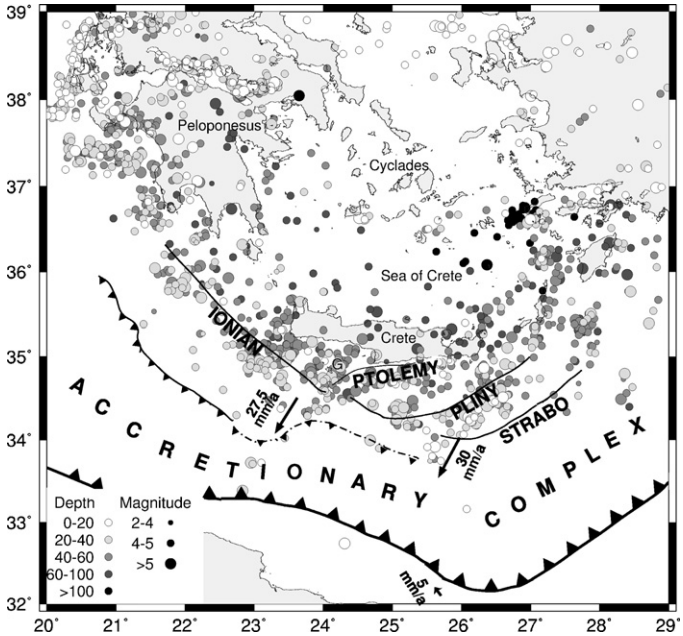


Fig. 1. Map of the Aegean region showing major tectonic units and earthquake hypocenters from the relocated ISC catalogue (Engdahl et al., 1998) for 1964 to 1998. The southern deformation front of the Hellenic subduction zone and the border between the central and inner units of the Mediterranean ridge (Lallemant et al., 1994; Huguen et al., 2001) are indicated. Horizontal GPS velocities relative to Eurasia (McClusky et al., 2000) are depicted by arrows. Additionally, trenches south of Crete are delineated.

by the counterclockwise rotation of the Aegean and the Anatolian plates and by an additional acceleration towards the active continental margin of the HSZ. These processes are determined by the northward migration of the Arabian plate that drives the motion of Anatolia along the North-Anatolian fault system and by the slab pull of the subducting lithosphere that causes the retreat of the continental margin. The relative velocity of the Aegean microplate, relative to the approaching African continent is about 35–40 mm/a (McClusky et al., 2000). In addition to the horizontal movements, geological and archaeological records yield evidence for the uplift of the Hellenic forearc in the area of Crete (see e.g. Lambeck, 1995; Meulenkamp et al., 1994). The maximum current rates of uplift are observed for western Crete (up to 4 mm/a) and the south-eastern part of the island (2 mm/a).

South of Crete a microseismically highly active zone between 20 and 40 km depth indicates a dipping of the plate contact of about 15–19 (e.g. Meier et al., 2004b; Delibasis et al., 1999). Further to the north, the arc-shaped Wadati-Benioff zone extends to a depth of 100–160 km with the maximum depth beneath the eastern volcanic arc (e.g. Papazachos and Comninakis, 1971; Papazachos et al., 2000).

Tomographic studies (Papazachos et al., 1995; Papazachos and Nolet, 1997) reveal a complicated lithospheric structure of the Aegean area with strong variations of crustal thickness and significant upper mantle velocity anomalies. The thickness of the crust varies from 40 km beneath the Peloponnese, 30 km beneath Crete, to about 20 km beneath the Cretan Sea. The analysis of seismic wave velocities from the same studies displays the subduction process. The African lithosphere appears as a higher-velocity dipping zone of either an amphitheatrical shape (Papazachos et al., 1995) or divided into two wings that form a more or less right angle (Papazachos and Nolet, 1997). The dipping angle of the western part of the slab varies from 10° at shallow depths to 25° at greater depths (Papazachos and Nolet, 1997). The dipping angle of the eastern part is steeper than that of the western part.

The crustal structure of the HSZ has been investigated by several active and passive seismic studies. Bohnhoff et al. (2001) and Brönner (2003) used wide-angle reflection and refraction seismic (WARP) to investigate the structure of Crete and its vicinity to about 35 km depth. They found the continental Moho below Crete at a depth of about 30 km. A high velocity layer underlying the continental crust was localised and interpreted by both authors as the subducted oceanic crust. Receiver function studies by Knapmeyer and Harjes (2000) and Li et al. (2003) differ from the WARRP interpretation. A strong signal from the Moho of the subducted African plate is clearly observed beneath Gavdos island (south of western Crete) and Crete at depths ranging from 44 to 69 km. The observed range of slab

depths is also in good agreement with results from surface wave dispersion measurements in the southern Aegean (Meier et al., 2004a). According to Li et al. (2003), beneath western and north-central Crete the Aegean Moho is located at 32 and 39 km depth, respectively, and behaves as a reversed crust–mantle velocity contrast, possibly caused by hydration and serpentinization of the forearc mantle peridotite. Knapmeyer and Harjes (2000) also found negative receiver function signals at a depth between 25 and 30 km beneath western Crete. They interpreted this as a thick fossil accretionary wedge with a characteristic en echelon fault sequence. Two-dimensional modelling of wave propagation suggests another explanation for the negative phases observed in receiver functions from northern Crete (Endrun et al., 2005). The data can also be reproduced by taking into account the strong variation in Moho depth between Crete itself and the Sea of Crete to the north and its effects on phase conversions.

Seismic surveys and seismological studies provided a set of local data about the structure of the HSZ. Gravimetric methods can be applied to fill the gap between local seismic models. Several interpretations of gravity data from the HSZ have been proposed so far. These density models are either two-dimensional (e.g. Truffert et al., 1993) or cover only part of the subduction zone. They are mainly concerned with the Aegean sea and continental Greece, so the actual arc lies on the edge of the models (e.g. Tsokas and Hansen, 1997; Tirel et al., 2004). Moreover, the data used in these studies did not cover the whole marine area (Tsokas and Hansen, 1997) or the satellite altimetry derived gravity field was used (e.g. Tirel et al., 2004). These (satellite derived gravity) anomalies (of the Cretan sea) differ significantly from those measured by marine surveys (ships) (Casten and Snopek, 2005). We present a new regional density model covering a central part of the HSZ with Crete in the middle of the modelled area. In our interpretation we use gravity anomalies from land, marine and satellite altimetry observations combined into one dataset. The model includes geological layers from 0 to 150 km depth, therefore non-filtered data can be used for the modelling.

The available geophysical studies of the region give a good general picture of the structure of the forearc. However, some uncertainties about the deep crust beneath Crete still exist. While the depth to the African Moho is well constrained by seismological studies, different models have been proposed for the Aegean lithosphere beneath Crete. Information on densities in addition to velocities of the rocks can help to define the structure of the Aegean crust in the forearc. A grid search is performed to obtain constraints on the density of the Aegean lithosphere in the area of Crete. The applied method allows to define a range of possible densities for rocks of the lower Aegean lithosphere. The resulting densities are compared with the velocities of seismic waves obtained from several studies. Finally, we investigate the weight of the Aegean lithosphere along an E–W oriented profile on Crete based on the 3D density model.

2. Seismic constraints

Receiver functions were used to image lithospheric discontinuities down to the Moho of the subducting African plate beneath Crete (Knapmeyer and Harjes, 2000; Endrun et al., 2004; Li et al., 2003). These studies found the African Moho at a depth between 45 and 65 km beneath Crete, dipping to the north and east. Above the slab, significant arc–parallel heterogeneity was outlined with pronounced differences between the western and central parts of Crete (Endrun et al., 2004). These differences are further supported by comparison with measurements of fundamental mode Rayleigh wave dispersion (Meier et al., 2004a; Endrun et al., 2004). Complementary to the receiver functions, which are primarily sensitive to the existence and sharpness of discontinuities in elastic parameters, but do not constrain absolute velocities, Rayleigh wave phase velocities were inverted for 1D S-velocity profiles to a depth of 250 km (Meier et al., 2004a). The subducting African lithosphere was detected as a high-velocity region with the slab interface at approximately 50 km depth in the area of Crete.

Due to the relatively larger wavelength of the teleseismic data used for analysis with receiver functions and surface waves, respectively, the vertical resolution of these measurements is limited in comparison to active seismic measurements. In the area of Crete, the detailed WARP profiles by Bohnhoff et al. (2001) provide a high-resolution picture of crustal P-wave velocities from the sedimentary cover down to a depth of about 35 km. They include an intracrustal discontinuity which agrees favourably with the receiver function results.

All of these structural information were combined to provide a three-dimensional structural model used to interpret the hypocenter distribution of local seismicity (Meier et al., 2004b). The seismicity also supports a plate interface depth of around 50 km beneath Crete. Furthermore, comparison with seismicity strongly suggests that the lower Aegean lithosphere beneath Crete between about 20 and 40 km depth is currently aseismic (Meier et al., 2004b). The southern, eastern and western shorelines of Crete correlate with the margins of this aseismic region.

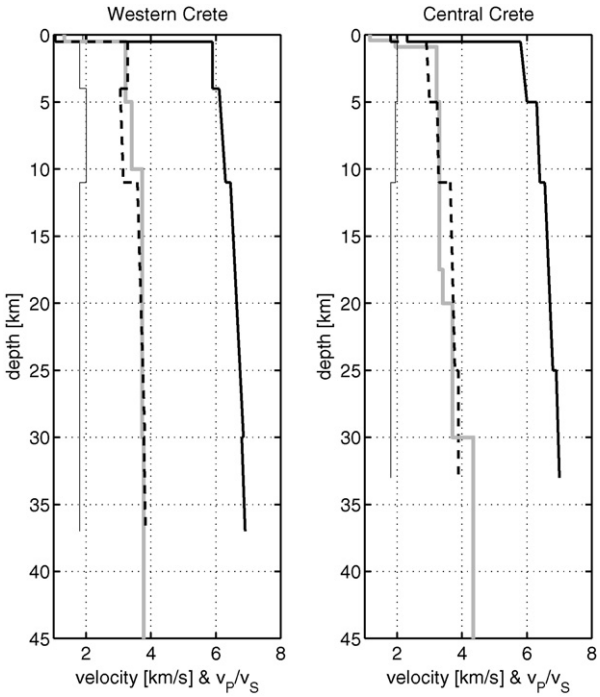


Fig. 2. Seismic constraints on elastic parameters beneath western and central Crete. Black lines denote parameter values from seismic profile lines, i.e. P-wave velocity (thick solid line), S-wave velocity (thick dashed line) and v_P/v_S ratio (thin solid line). Thick grey lines indicate S-wave velocity models from surface wave dispersion modelling.

For further comparison of the different seismic constraints, the WARP profiles, which include estimates of v_P/v_S (Bohnhoff et al., 2001), are converted to S-wave velocity profiles. In Fig. 2, they are plotted alongside the S-wave velocities derived from the modelling of Rayleigh wave phase velocity measurements on western and central Crete (Endrun et al., 2004). For the velocities from seismics shown here, average values of layer thickness, P-wave velocity and v_P/v_S ratio for western and central Crete were derived from the EW-profiles analysed by Bohnhoff et al. (2001). For central Crete, there is a significant difference in the v_P/v_S ratio of the upper 10 km between a NS profile and an EW profile in the same region (Bohnhoff et al., 2001), which might point to crustal anisotropy. For better comparison with the results from Rayleigh wave dispersion measurements, we chose the values of the EW-oriented profile as this is similar to the orientation of the Rayleigh wave path considered.

For western Crete, the P-wave velocity of the lower crust is on average 6.7 km/s. Together with the average v_P/v_S ratio of 1.8 for this part of the crust, this translates to a S-wave velocity of about 3.72 km/s, which is in excellent agreement with the results from dispersion analysis, in spite of the different wavelengths and wave types considered in both studies. The relatively high v_P/v_S ratio of 1.8 for the Cretan crust is further supported by micro-seismic surveys (e.g. de Chabalier et al., 1992; Delibasis et al., 1999) to a depth of at least 30 km.

Beneath central Crete, a slightly higher lower crustal P-wave velocity of 6.8 km/s was derived from the seismic profiles. This translates to an average S-wave velocity of 3.78 km/s, which is slightly larger than the value derived from surface wave studies (3.5 km/s). The difference is mainly caused by slower S-wave velocities between 11 and 20 km in the surface wave model. One possible explanation for this discrepancy could be the velocity dispersion caused by attenuation, as it would also lead to higher velocities for measurements at higher frequencies. However, for the dataset at hand, estimates indicate that a Q value of less than 60 would be necessary to even produce a difference of 0.1 km/s between the different models. This is well below of what is expected for the lower crust of Crete and besides, it is unlikely to find such highly attenuating behaviour confined to just a single middle-crustal layer. Of course, the error estimates for the respective measurements have to be considered as well before interpreting the differences between the two velocities. They are between 2.5 % (above 15 km) and 5.0% (below 15 km) for the seismic velocities, i.e. between 0.1 and 0.2 km/s. For the velocity models from dispersion curve modelling, uncertainties are around 0.15–0.2 km/s.

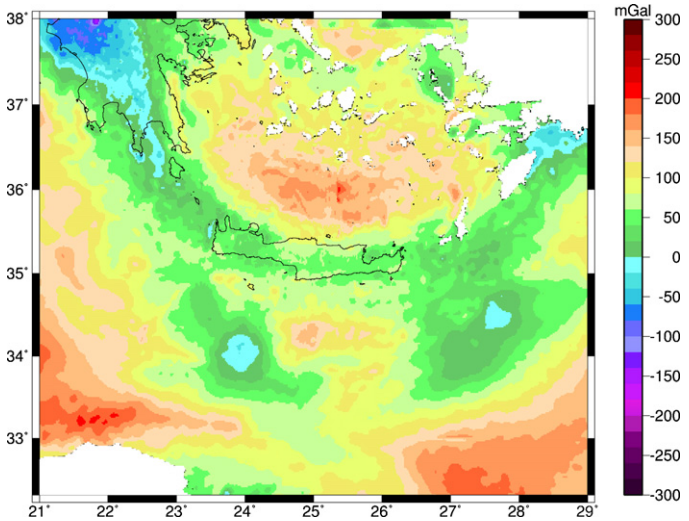


Fig. 3. Bouguer gravity field $\rho_B = 2670\,kg/m^3$ of the Helenic subduction zone.

Thus, the difference of on average 0.35 km/s in this depth interval between the various models cannot reliably be distinguished from measurement uncertainty.

3. Density model of the Hellenic subduction zone

3.1. Gravity field

In order to produce a new gravity map of the Aegean region several sources were used. Land, sea and satellite observations – linked to the IGSN71 gravity datum – were combined to form a uniform dataset covering the whole region (Casten and Snopek, 2005). The Bouguer reduction (with terrain correction) was calculated for the land and off-shore gravity field with a reduction density $\rho_B = 2670\,kg/m^3$. The resulting Bouguer gravity map is presented in Fig. 3. The characteristic feature of the Bouguer field is the arc-shaped zone of relative low gravity field (0–50 mGal, $1\,mGal = 10^{-5}\,m/s^2$) that extends, along the Hellenic arc, from the Peloponnese, through Crete and Rhodes to Anatolia. This zone is followed to the north by a quasi-arc-shaped zone of high gravity field (150–180 mGal). Below the volcanic arc gravity is again lower (≈100 mGal). The anomalies south of Crete are not so regular. High gravity anomalies are divided by two regions of a significantly low gravity field. Moreover, the anomalies show very high horizontal gradients. At the distance of less then 100 km the gravity field rises from −20 mGal (south of western Crete) to nearly +150 mGal (south of central Crete). The southernmost gravity field is characterised by very high anomalies (≈200 mGal) with a zone of lower gravity field (≈100 mGal) in the central part.

Table 1
Parameters of modelled layers

Layer	P-wave velocity (km/s)	Initial density (kg/m^3)	Modelled density (kg/m^3)
Seawater (SW)	–	2670	2670
Sediments (S)	2.5–4.9	1900–2600	1900–2600
Upper crust (UC)	5.8–6.4	2670	2670–2750
Lower crust (LC)	6.5–6.9	2850	2750–2950
Oceanic crust (OC)	6.9–7.1	2950	2950
African mantle (AM)	7.9	3350	3350–3400
European mantle (EM)	7.7	3300	3250–3350
Asthenosphere (A)	–	3350	3350

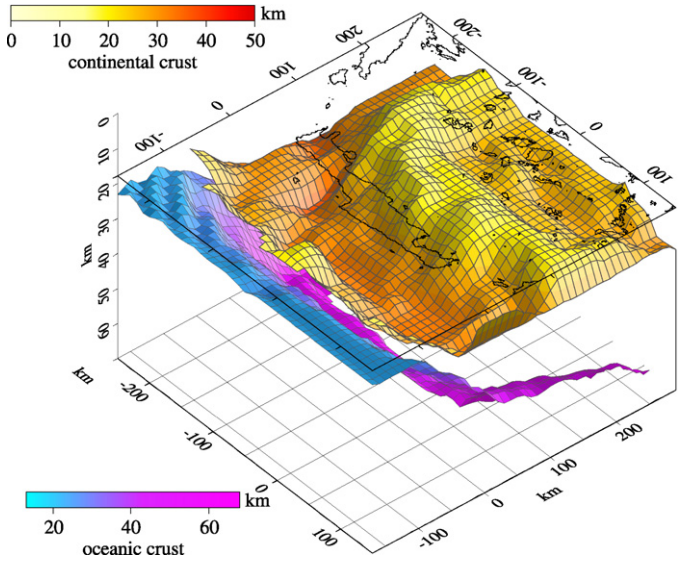


Fig. 4. Structure of modelled oceanic (African) and continental (Aegean) Moho (Model I).

3.2. Regional model (Model I)

A regional density model is constructed with a new software which uses rectangular blocks to calculate the gravity effect of a modelled structure (Snopek and Casten, 2006). The model covers an area of 300 km × 400 km with Crete in the central position. Because of a very high curvature of the Wadati-Benioff zone (e.g. Papazachos et al., 2000) and probably a significant gravity effect of the subducting African lithosphere the vertical size of the model is set to 150 km because the slab is observed at a depth of 150 km below the northern part of the region. The model consists of 78 000 blocks of constant horizontal size (10 km × 10 km) and variable vertical size (from 1 km at near surface depths to 10 km in the deepest part). The vertical resolution for crustal depths (5–50 km) is 2.5 km.

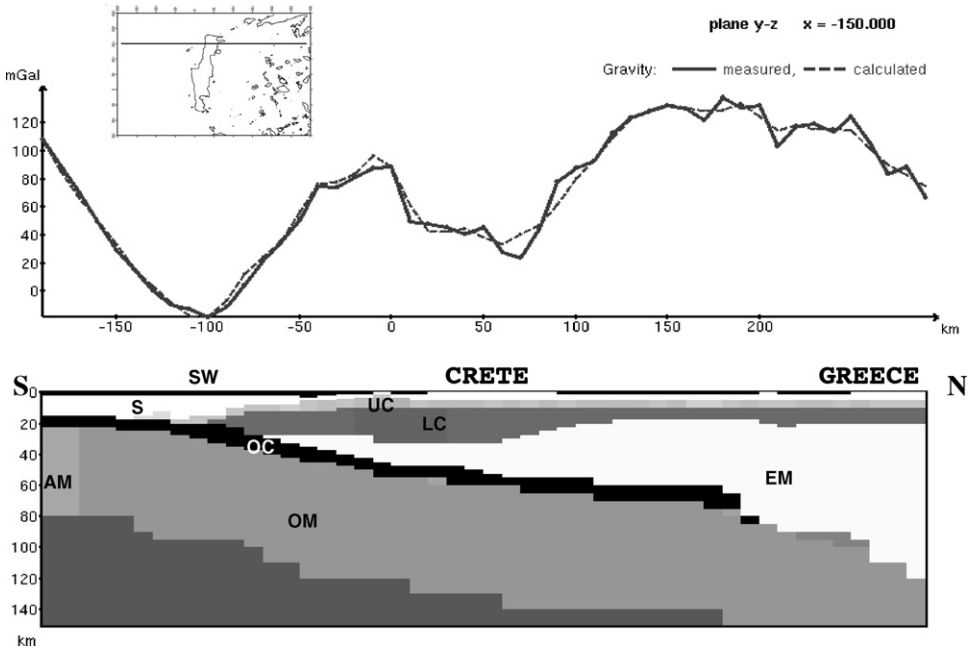


Fig. 5. SN cross-section of the density model through western Crete (Model I): (upper part) anomalies; (lower part) density structure.

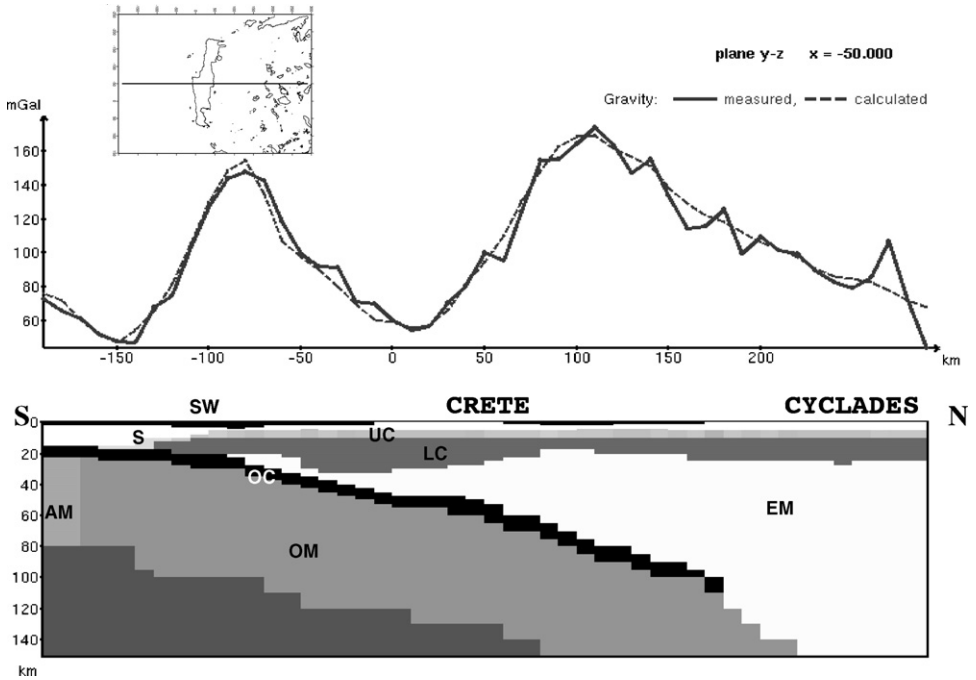


Fig. 6. SN cross-section of the density model through central Crete (Model I): (upper part) anomalies; (lower part) density structure.

Each block of the model can be assigned to one of the pre-defined layers: seawater, sediments, upper and lower Aegean crust (continental crust), African (oceanic) crust, upper mantle (divided into the African and Aegean one) and asthenosphere (see Table 1). Densities of the crustal layers are estimated from the WARRP profiles applying the Nafe and Drake relation between P-wave velocities and densities (e.g. Ludwig et al., 1971). Density values estimated from this relation are used as starting values for inverse modelling. As a reference density, the assumed density of the asthenosphere, i.e. 3350 kg/m^3, was used. The density model does not include deeper rocks of the Earth interior. In order to get the measured and modelled gravity field on the same level, constant shift value $g_{\text{shift}} = 780$ mGal was added to the calculated anomalies. This value is calculated as a mean difference between the measured and calculated gravity field.

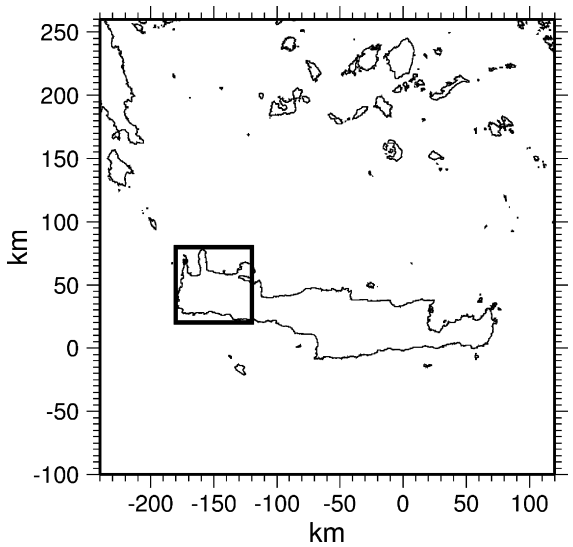


Fig. 7. Area defining part of model chosen for grid search analysis.

The interpretation is performed by means of manual inverse modelling and automatic inversion. The relief of the slab and Aegean Moho is not the object of the automatic inversion but is manually modelled. The contact between the sediments (S) and crystalline upper crust (C) is also modelled and the results from wide-aperture seismic are used as the constraining data. The densities of all layers are inverted to get an as good as possible match between the calculated and observed field. The evolution strategies (ES) algorithm is applied to optimise the densities of the blocks. ES is a global inversion method which uses principles of organic evolution in the optimisation of model parameters (e.g. Schwefel, 1981). It enables optimisation of the preset and constrained model. To minimise the number of parameters the blocks are grouped in columns. Therefore, the density of each layer varies only horizontally.

In the following we present two density models. Model I is characterised by constant density of the lower Aegean crust. Model II is a result of a detailed analysis of densities beneath western Crete. In this model, density of the crust

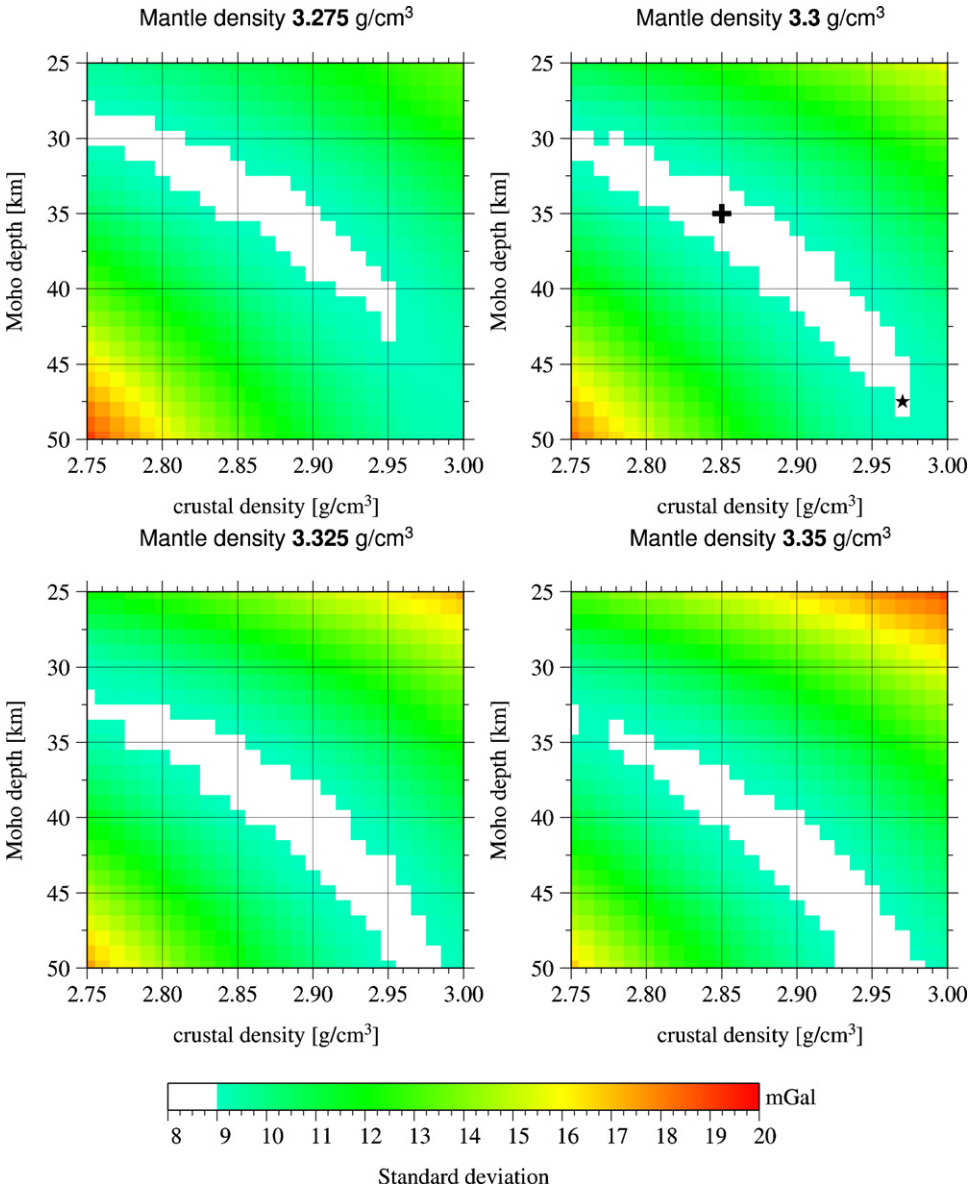


Fig. 8. Results of grid search for densities and Moho depth below western Crete. Symbols indicate parameters of Model I ((+) $\rho = 2850\,\mathrm{kg/m^3}$, depth = 35 km) and Model II ((⋆) $\rho = 2970\,\mathrm{kg/m^3}$, depth = 47.5 km).

beneath western Crete differs from the crustal density in the eastern part of the island. This alternative model will be presented in the next section.

The quality of the Model I is very good. The standard deviation of the misfits is about 7 mGal and is comparable with the accuracy of the observed anomalies. The structure of the Aegean Moho and the subducting slab is given in Fig. 4. Two SN cross-sections through western and central Crete are shown in Figs. 5 and 6. The depth of the slab varies from 15 km in the southern part of the model to 50 km below Crete and up to 150 km below the volcanic islands. The thickness of the Aegean crust is about 30 km below Crete and its proximity. Below the Cretan sea the thickness of the crust reaches its minimum of about 15 km and increases to about 20–25 km beneath the volcanic arc.

3.3. Grid search analysis of densities below Crete (Model II)

The regional model gives a good picture of the density distribution of the HSZ. However, on account of the non-uniqueness of gravity interpretation it cannot be treated as the ultimate one. We are particularly interested in the structure of the lithosphere in the forearc region beneath Crete. The density model beneath western Crete suggests existence of the Aegean Moho at a depth ≈35 km. The surface waves analysis presented here shows increased S-wave crustal velocities beneath western Crete. The seismologic studies, however, do not indicate an Aegean Moho beneath the western part of the island.

Gravity interpretation is characterised by depth–density equivalency. Hence, it is possible to create many models, with different depth and density of the crust beneath Crete, which would fit to the gravity observations. In order to investigate if any of the possible combinations of Moho depth and crustal density match the seismological interpretation, we perform a detailed analysis by means of a grid search. The gravity effect of all possible combinations of Moho depth and density of the lower crust in the area presented in Fig. 7 is calculated. The depth of the Moho is limited by the assumed depth of the inter-crustal boundary (15 km) and the modelled subducting oceanic crust (≈ 55 km). The standard deviation of the misfits between the observed and calculated field is plotted as a function of the two variable parameters. The calculations are repeated for four densities of the mantle wedge. The results are presented in Fig. 8. The regions marked in white indicate the values of the parameters which give the best match with the observed gravity field (the threshold value of 9 mGal is chosen arbitrary). The minimum depth of the Moho is about 30 km which correspond to decreased densities of the crust and the mantle below. For all plots, increased depth of the Moho is compensated by increased density of the crust. The corresponding density for the maximum Moho depth (i.e. 47.5 km)

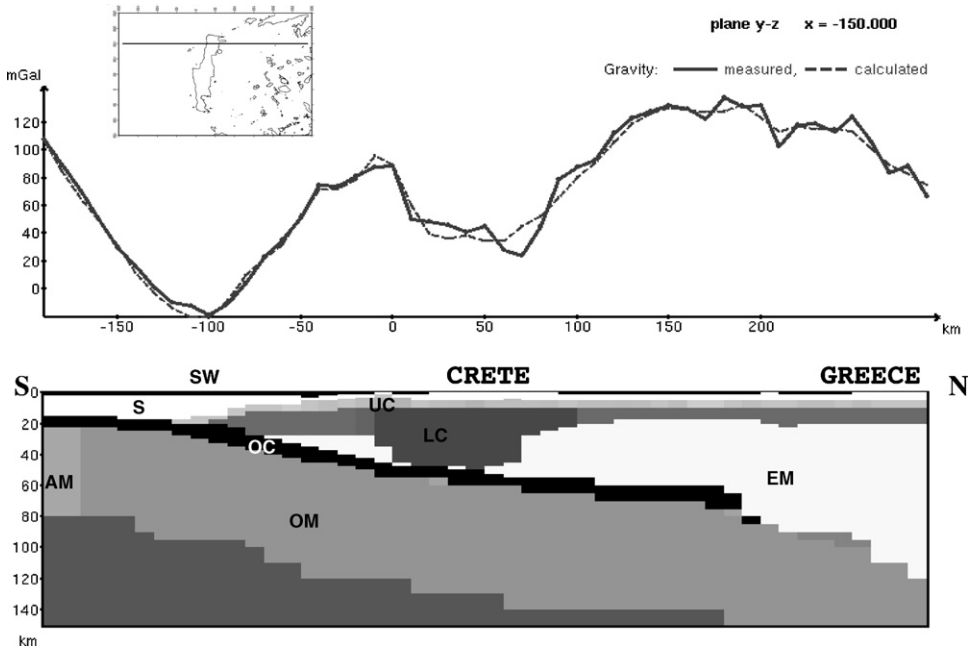


Fig. 9. SN cross-section of the density model through western Crete (Model II): (upper part) anomalies; (lower part) density structure.

is about 2970 kg/m^3. The maximum Moho depth means that there is no Aegean mantle beneath it, and the Aegean crust contacts the African one.

The model which is consistent with the seismological interpretation is marked with a star in Fig. 8. The density of 2970 g/cm^3 is characteristic for the oceanic crust and is comparable with the density of the oceanic crust in our model. In such a situation, the Moho, defined as contact between crustal and mantle rocks, vanishes and there is no significant density variation between 15 km (inter-crustal boundary) and 55 km (African Moho) (Fig. 9).

3.4. Sources of vertical crustal movements

The uplift of the forearc of the HSZ is confirmed by archaeological and geological records. According to Lambeck (1995) the whole Hellenic arc is subject to tectonic uplift with the maximum rates for Crete. The uplift rates for western Crete locally exceed 4 mm/a. The Lambeck (1995) interpretation based on archaeological data covers a period of the last 4–5 ka (e.g. Flemming, 1978). Meulenkamp et al. (1994) analysed geological data from late Miocene to present. According to this analysis, the maximum amount of uplift for Crete was locally in the order of 2000 m in the last 4.5 ma. It is assumed that western Crete was uplifted by 9 m during the magnitude 8.3 earthquake that struck Crete in 365 (e.g. Pirazzoli et al., 1982; Stiros, 2000). The uplift of Crete is confirmed by repeated microgravity observations (see Lagios and Hipkin, 1986). It may be possible that some of these movements are of local gravitational nature. Other sources may be related to the regional distribution of isostatic anomalies as investigated by Fleischer (1964) and Lagios et al. (1995) showing an over-compensation for the region of the Cretan Sea and under-compensation for the Hellenic trench region. Crete is situated between these two zones of different crustal load and may be affected by balancing movements additional to subduction related phenomena.

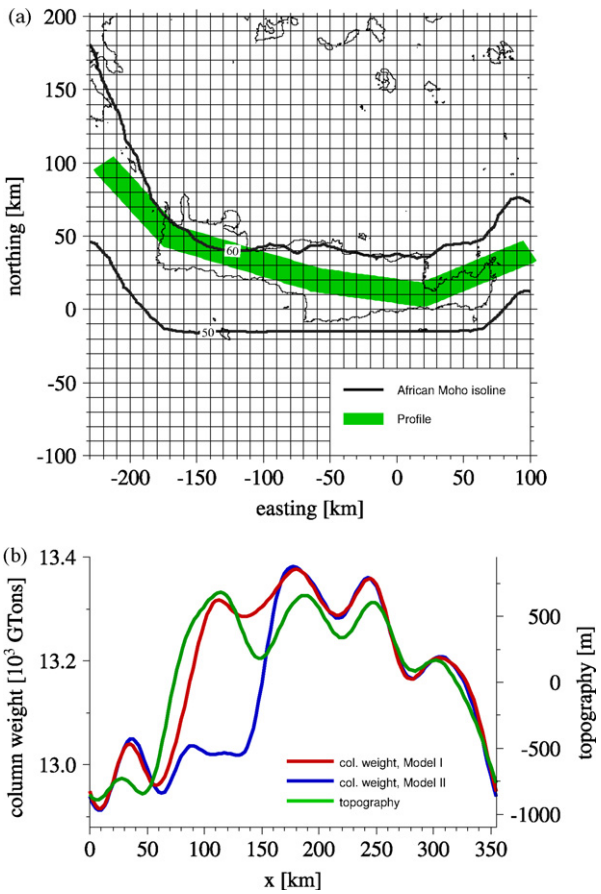


Fig. 10. (a) Profile of mass distribution analysis and (b) topography with calculated weight of columns along the profile.

The presented 3D density model allows to investigate the horizontal distribution of masses along the forearc and to interpret variations in the mass distribution with respect to buoyancy of the Aegean lithosphere beneath Crete. We consider a profile along the 55–60 km depth isoline of the African Moho which covers a land area of Crete. In geological time intervals of several million years, the plates may be regarded as decoupled in subduction zones (Wang et al., 1995; Wang, 2000). Therefore, only masses above the plate contact are considered. The weights of vertical columns along the profile shown in Fig. 10 are calculated. The width of the profile is 30 km (selected are blocks which are located on the profile as well as neighbouring blocks attaching the profile from south and north). Each column includes also the material above the gravity datum level (i.e. the sea surface) of a density $\rho_B = 2670\,kg/m^3$ (the standard Bouguer reduction density). In the off-shore columns, a water density of $\rho_{water} = 1040\,kg/m^3$ was used. The weight of columns is calculated according to Model I (with constant density of the lower crust) and Model II (with increased density and thickness of the lower crust beneath western Crete). The results are presented in Fig. 10b.

For Model I, an increased column weight (CW) of the land area of Crete is found. The fact that the column weight curve follows the topography curve fits in with expectation. However, for Model II, which is favoured by seismic studies, western Crete shows significant lower CW values than the rest of the island. This decreased weight of the western Crete block is suspected to contribute to the observed, relative uplift of this part of the island.

4. Conclusions

The presented interpretation of gravity data allows to define densities of the rocks constituting the crust and mantle of the Aegean lithosphere below Crete. Together with the velocities of seismic waves a geophysical model of the forearc in the area of Crete is constructed.

According to receiver function analyses the inter-plate contact between the subducting African lithosphere and the overriding Aegean lithosphere is located at about 50 km depth beneath Crete, which is in good agreement with results from surface wave dispersion measurements. Receiver functions additionally show significant differences in the structure of the Aegean lithosphere above the slab. According to these studies, the continental Moho is located at about 30 km depth beneath central Crete, while no Moho is found beneath the western part of the island. High reflectivity in the WARP cross-sections for western Crete at this depth points to a strong heterogeneous crustal structure. The difference in the crustal structure between central and western Crete is further confirmed by the velocity profiles derived from surface wave modelling. Surface wave studies show higher S-wave velocities for the lower crust beneath western Crete (3.75 km/s) than for central Crete (3.5 km/s). For western Crete, these crustal S-wave velocities extend down to the African slab in 50 km depth, while beneath central Crete, higher velocities are found below 30 km depth, indicating the Aegean mantle wedge. Active seismics provide additional independent information about the crustal velocity structure. For western Crete, there is an excellent agreement between the average S-velocity values of the lower crust determined by active seismic measurements and dispersion analysis. For central Crete, however, seismics indicate higher lower crustal S-velocities. Still, the results from phase velocity modelling and active seismics cannot be distinguished within the error bars of the respective measurements.

A 3D density model was constructed by means of manual inverse modelling of the main boundaries and automatic inversion of densities of selected layers. In the first step a constant density of the lower crust was used. According to this model the continental crust of the arc has a thickness of 25–30 km. Beneath western Crete crustal thickness reaches 35 km. This model, however, does not match the seismological interpretation. Therefore, applying a grid search analysis, we constructed an alternative model with varying density of the lower crust beneath western Crete. This model is consistent with the seismic studies. The main characteristics of this model are that no significant density variations between 15 and 55 km depth are found and the lower crustal density beneath western Crete is higher than for central Crete. The values of the parameters resulting from this study are shown in Table 2. The densities and velocities of the Aegean mantle beneath central Crete are characteristic for a hot and slow continental mantle. The results show a good correlation between density and S-velocity of the lower crust for Model II. The distribution of microseismicity beneath Crete (Meier et al., 2004b) points to ductile deformation of the Aegean lithosphere at depths between 20 and 40 km. The margins of this ductile, aseismic zone in the lower Aegean lithosphere correlate with the Cretan shore line.

Taking the weight analysis into account, Model II suggests that the mass deficit of the Aegean lithosphere beneath western Crete may contribute to the relative strong uplift of western Crete. In addition, buckling of the lithosphere due to inter-plate coupling and mass distribution at depth more than 45 km affect the uplift of Crete as well. This seems to

Table 2
Velocity and density values for Crete

Parameter	Western Crete[a]	Western Crete[b]	Central Crete
Depth to Aegean Moho (km)	32.5–35	45	30
Density of lower crust (kg/m^3)	2850	2950	2850
S-velocity of lower crust from surface waves (km/s)	3.75	3.75	3.5
S-velocity of lower crust from seismics	3.72	3.72	3.82
P-velocity of lower crust from seismics	6.7	6.7	6.8
v_P/v_S of lower crust from seismics	1.8	1.8	1.75
Density of Aegean mantle (kg/m^3)	3300	–	3300
S-velocity of Aegean mantle form surface waves (km/s)	–	–	4.35

[a] Model I
[b] Model II.

be the case for the eastern part of the island, where the uplift of Crete cannot be explained by buoyancy of masses of the Aegean lithosphere down to 45 km depth.

The models and hypotheses presented in this paper are well suited as a basis for further geodynamic modelling and testing.

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