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Gravity modelling of the Hellenic subduction zone — a regional study

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Abstract

The Hellenic subduction zone is clearly expressed in the arc-shaped distribution of earthquake epicenters and gravity anomalies, which connect the Peloponnesos with Crete and Anatolia. In this region, oceanic crust of the African plate collides northward with continental crust of the Aegean microplate, which itself is pushed apart to the south–west by the Anatolian plate and, at the same time, is characterised by crustal extension. The result is an overall collision rate of up to 4 cm/year and a retreating subduction process. Recent passive and active seismic studies on and around Crete gave first, but not in all details consistent, structural results useful for supporting gravity modelling. This was undertaken with the aim of presenting the first 3D density structure of the entire subduction zone.

Gravity interpretation was based on a Bouguer map, newly compiled using data from land, marine and satellite sources. The anomalies range from +170 mGal (Cretan Sea) to −10 mGal (Mediterranean Ridge). 3D gravity modelling was done applying the modelling software IGMAS. The computed Bouguer map fits the low frequency part of the observed one, which is controlled by variations in Moho depth (less than 20 km below the Cretan Sea and extending 30 km below Crete) and the extremely thick sedimentary cover (partly up to 18 km) of the Mediterranean Ridge.

The southernmost edge of the Eurasian plate, with its more triangular-shaped backstop area, was traced south off Crete. Only 50 to 100 km further to the south, the edge of the African continent was traced as well. In between these boundaries there is African oceanic crust, which has a clear arc-shaped detachment line situated at the Eurasian continental edge. The subduction arc is open towards the north, its slab separates hotter mantle material (lower density) below the updoming Moho of the Cretan Sea from colder one (higher density) in the south. Subjacent to the upper continental crust of Crete is a thickened layer of lower crust followed by the subducted oceanic crust with some mantle material as intermediate layer. The depth of the oceanic Moho below Crete is 50 km. The presence and structure of subducted or underplated sediments remains uncertain.
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Keywords: African–Eurasian plate collision; Bouguer gravity field; 3D forward modelling; Density structure; Plate boundaries

1. Introduction

The Hellenic contact between Eurasian and African lithosphere is an active plate boundary with overall convergence rate of up to 4 cm/year. This is known from seismological and geodetic observations (Kahle et al., 1998). The results describe a process of collision and retreating subduction, with simultaneous extension in the highly mobile Aegean area. As shown in Fig. 1, the dominant direction of African plate movement is from south to north with a rate of 1 cm/year according to the global plate model NUVEL-1A, although recent investigations suggest 0.6 cm/year only (McClusky et

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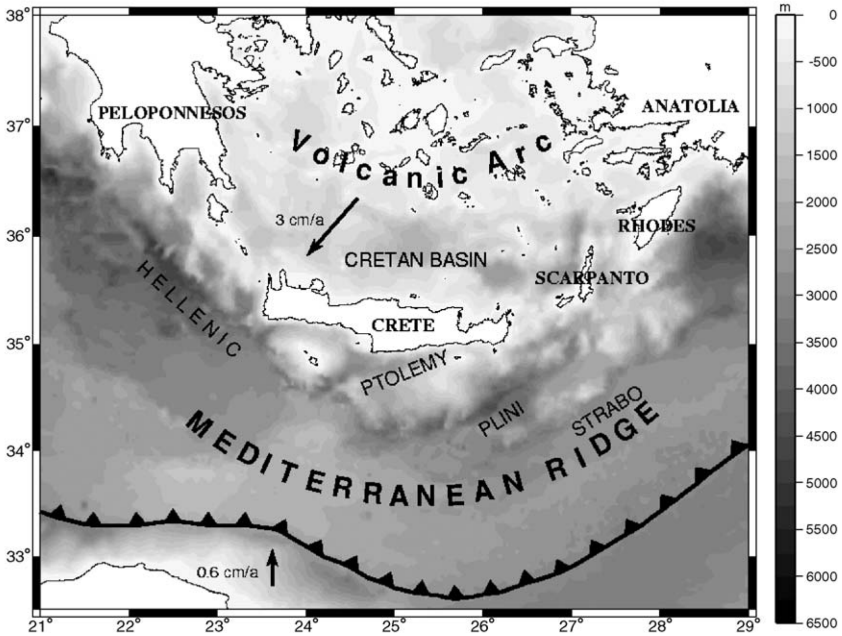


Fig. 1. Bathymetric view of the Hellenic plate contact between Eurasia and Africa, showing main plate movements and giving subductional forearc elements as basin (Cretan), island rise (Crete, Scarpanto, Rhodes), so called trenches (Hellenic, Ptolemy, Pliny, Strabo), and the accretionary complex (Mediterranean Ridge) with its southernmost front.

al., 2000). Part of the Eurasian plate – the Aegean microplate – moves from north–east to south–west with a rate of 3 cm/year (McClusky et al., 2000), pushed apart by the Anatolian plate with its dominant western drift (LePichon et al., 1995).

The plate boundary is defined by a Wadati-Benioff zone (Papazachos and Comninakis, 1971; Gregersen, 1977; McKenzie, 1978; LePichon and Angelier, 1979; Makropoulos, 1984; Taymaz et al., 1990). This northward dipping and laterally extremely bended zone of nearly amphitheatric-like shape (Papazachos and Nolet, 1997) shows a distribution of hypocentres down to a depth of about 200 km. Dipping starts south off Crete, and the zone extends north up to the volcanic arc (islands: Aigina–Milos–Thira–Nisiros). Tomographic models show a subducted slab with a velocity increase of about 2% (Spakman et al., 1988). On the basis of this result, the total length of the subducted plate is more than 600 km. The highly active seismic zone is sharply limited outside of its arc-shaped course. The shallow crustal seismic events mark a backstop area for the subducted crust. Following Truffert et al. (1993), this area defines the southern boundary of the Eurasian plate. A tomography of the crust and upper mantle in the Aegean area confirms strong lateral variations of crustal thickness and velocity structure, as well as the subduction of the African lithosphere under the southern Aegean (Papazachos et al., 1995). Furthermore, the velocity distribution directly reflects the stress field and the resulting tectonic regime. This picture changes in the deeper layers, where the tomographic images are affected by the subduction. Worth mentioning is the result of another tomographic study by Papazachos and Nolet (1997), in which a negative gradient indicates the existence of a low-velocity crustal layer below the Hellenic arc at a depth of about 10–15 km.

Within the collision area of the two plates, a forearc rise (Hellenic Rise: Crete–Scarpanto–Rhodes) is followed to the south by a system of structures misleadingly called “trenches” (Hellenic, Ptolemy, Pliny and Strabo trench). These deep-water features do not trace plate subduction, but are basins of the transitional zone between areas of extension and convergence within the forearc region. While the Hellenic trench strikes south–east (nearly perpendicular to the Aegean plate movement), the Ptolemy, Pliny and Strabo trenches strike north–east, oblique to the Aegean plate movement, but with a small relative outward component of less than 2 mm/year to the south–east demonstrating internal deformation. They have been interpreted as transform faults by McKenzie (1978) or strike slip faults by Chaumillon and Mascle (1997). Investigation of microseismicity by Kovachev et al. (1992) gives the picture of smaller Wadati-Benioff zones connected at least with the Pliny and Strabo trenches.

The trench system is followed to the south by the broad and arc-shaped Mediterranean ridge, which is squeezed in between the Eurasian and the African continents. To the west and also to the east the sediments of the ridge thin out into deep-sea basins (Ionian and Herodotus). Without additional structural information, the role of these sediments (Kastens, 1991) can be interpreted in different ways:

(1) The ridge is an accretionary wedge. Knowing that the oceanic basement has a thick cover of sediments, and taking into account a slab-length of 600 km, one is led to wonder where the total volume of the sediments could be. There is no room for a corresponding wedge. The sediments could have been partly subducted and then transformed under the higher p,T-conditions.
(2) Thinned-out continental crust (African) with a thin cover of sediments underplates the Aegean, with an uplift of the ridge through reactivation of older faults by collisional processes. At the same time the blocks of Crete were uplifted and extensional processes took place within the Eurasian crust.
(3) Following Angelier et al. (1982), the uplift of Crete could be a result of the accretion of sediments, a process which started 13 million years ago (LePichon and Angelier, 1979) and is still going on (Meulenkamp et al., 1988).

According to Chaumillon and Mascle (1997), the Mediterranean ridge is a huge accretionary wedge emplaced as a consequence of the African–Aegean plate convergence. The ridge shows a north–south structural arrangement with clear east–west lateral variations. The area includes three distinct major domains. The outer domain, appearing as western and eastern wings (not present opposite Libya) with accreted sediments, is bounded southward by the accretionary front. The central, or crestal, domain consists of accreted sediments with evidence of mud diapirs and mud volcano activity. It overthrusts northward the inner domain and is bounded southward opposite Libya by the accretionary front, which is already colliding with the African continental slope. The less deformed inner domain is bounded northward by the Hellenic trench system. On the western wing of the ridge, seismic velocities are characteristic of continental crust. In the east disconnected and uplifted basement blocks dominate. The inner domain is interpreted as a backstop area (Truffert et al., 1993; Chaumillon et al., 1996; Huguen et al., 2001; Jones et al., 2002; LePichon et al., 2002; Reston et al., 2002).

The process of forearc extension causes a thinning out of the continental crust in the area of the Cretan basin (Cretan Sea), and is also related to the subduction zone. In connection with this, Truffert et al. (1993) point to the "hot" mantle below the southern Aegean, while the positive gravity anomaly of the Cretan Sea can probably be attributed to the cold and consequently dense slab that goes down into the mantle (Papazachos and Comninakis, 1971).

2. Previous work

An early attempt at 3D gravity modelling of the suggested arc-shaped subducted slab of the Hellenic area was undertaken by Gregersen and Jæger (1984). Only in part does the calculated dipping plate anomaly yield the same features as the observed anomalies. The differences may be due to the effects of the crustal structure of the Hellenic arc, unknown in detail at that time. Tsokas and Hansen (1997) continued with a study of Bouguer anomalies covering the Greek region, but without the marine area south off Crete. An estimate of the variation in continental Moho depth resulted in a continental crustal thickness below Crete of 31 km, with slowly decreasing values towards the central Aegean Sea (20 km). The residuals at Crete have values of +20 to +60 mGal, and those of the Cretan Sea +120 mGal. After further subtraction of the gravity effect of the subducted slab, as given by Gregersen and Jæger (1984), the residuals still reflect the arc-shaped structure of the Hellenic rise, with negative values (−20 mGal) between Crete and the Peloponnesos, and positive ones (+60 mGal) for the Cretan Sea. The remaining anomalies are assumed to be related to near-surface features and in cases of minima to low-velocity material.

The Intergovernmental Oceanographic Commission (1989) published the mapped field of Bouguer anomalies (reduction density 2.67 g/cm^3, scale: 1:1.000.000) covering the whole Mediterranean area. Sheet 9 shows the area of the Hellenic arc with distinct field variations reflecting different crustal thicknesses (Makris et al., 1998). The Aegean area with shallow water depth has a thinned-out crust covered by a thin sedimentary layer. The arc-shaped gravity field around Crete connects the strongly negative area of the Peloponnesos with that of Anatolia. In these areas crustal thickness reaches more than 40 km. Further to the south follows the Mediterranean Ridge – also arc-shaped – with locally strong anomaly variations within the narrowed zone between Crete and North Africa. These may be due to thickness variations of the sedimentary cover and the crust. The ocean basins to the west and the east have positive anomalies of several hundred mGal. In these cases the oceanic crust is covered by sediments of slight thickness only. The interpretation concerning the thickness of crust and sediments is controlled by gravity modelling in a few 2D cases only (Truffert et al., 1993; Wang, 1995; Fruehn et al., 2002). Taking into account the arc-shaped structure of the subduction zone, 3D modelling is crucial.

The isostatic behaviour of the Hellenic area was investigated by Fleischer (1964) and Lagios et al. (1995), following the Airy model (local compensation). Isostatic anomalies (values varying from about −100 mGal to about +100 mGal) show a more or less compensated situation for the area of mainland Greece, but a non-compensated one for the area of the Cretan Sea (overcompensated) and the Hellenic trench (undercompensated). This special situation may be the source of tectonic stress anomalies and relevant processes like the uplift of Crete, which is confirmed by repeated microgravity observations (Lagios and Hipkin, 1986).

The structure of the Hellenic subduction zone has been investigated with manifold different approaches and at different sites, but without giving a regional outline of the geometry of the contact area between the two colliding plates. An improved understanding of forces, stress transmission and plate coupling will be achieved by better access to the actual physical state of the crust and upper mantle of this collisional area. Recent passive seismic studies on Crete (Knapmeyer and Harjes, 2000) and active ones on Crete and in the surrounding marine area (Bohnhoff, 2000; Bohnhoff et al., 2001; Brönner, 2003) gave structural results useful for supporting gravity modelling. Gravity interpretation is done with the aim: (1) of studying the density structure of the crust and upper mantle and the regional variations in the collisional area, (2) of outlining the southernmost edge of the Eurasian continent, the northernmost edge of the African continent and by this the width of the African oceanic crust in between, and (3) of defining the depth of the Moho and the geometry of the downbent and subducted slab at least for crustal and sub-Moho depths.

3. Database

Crete was surveyed in two field campaigns, one in 1997 and the other in 1998. The Geophysical Department of Bochum University in cooperation with Hamburg University covered the island for the first time with about 2000 gravity (and magnetic) stations with a spacing of about 2.5 km. Gravity observations were done using LaCoste-Romberg meters (G-type). Station coordinates and height were observed using Trimble GPS receivers applying differential observations. With this the mean accuracy of gravity values was 0.55 mGal.

The marine area north off Crete (Cretan Sea) had already been investigated during several surveys in the late 1980s. The shipborne observations were performed with the Bodenseewerke gravity meter system Kss 31 of Hamburg University, which gave cross-over accuracies of better than 1 mGal with ship-track distances of 1 sm in the central and 3 sm in the outer region. In 1999 a new survey was performed mainly around Crete (Ilinski et al., 2001). These shipborne observations, again with the Hamburg equipment, were tied to the land observations through a harbour point at Agia Gallini at the south coast of Crete. All the marine data and in Hamburg existing land data from the Peloponnesos could be used for this regional study. The distribution of all available observation points and ship tracks is given in Fig. 2.

Finally, gridded free-air anomalies, deduced from satellite altimetry, were used to get information about the gravity field of the marine area between Crete and North Africa as well as south–west and south–east of the Hellenic arc. These world-wide data with a grid spacing of 2′, one set from the Scripps Institution of Oceanography, La Jolla, CA (http://topex.ucsd.edu; Sandwell and Smith, 1997) and a second one from the Raytheon Company, USA (http://magus.stx.com; responsible: Y.M. Wang), have a theoretical accuracy about 5 mGal (Chapin, 1998). A quantitative comparison between both satellite data sets and the marine data in overlapping areas resulted in a clear misfit (standard deviation) of 14.6 mGal (Sandwell data) and 11.65 mGal (Wang data), which supports the often made observation that satellite data are not reliable in coastal regions. The better result obtained by applying the Wang data supported the decision to use these in the study instead of the Sandwell data. Satellite data were used to cover the area outside the directly observed region.

For the first time, land, sea and satellite data from the area of the Hellenic arc – linked to the IGSN71 gravity datum – were combined to a uniform data set covering the region from 21° to 30° East and 32.5° to 38° North (without the Turkish land area and the islands Scarpanto and Rhodes). With a newly developed software, taking into account digital elevation models in

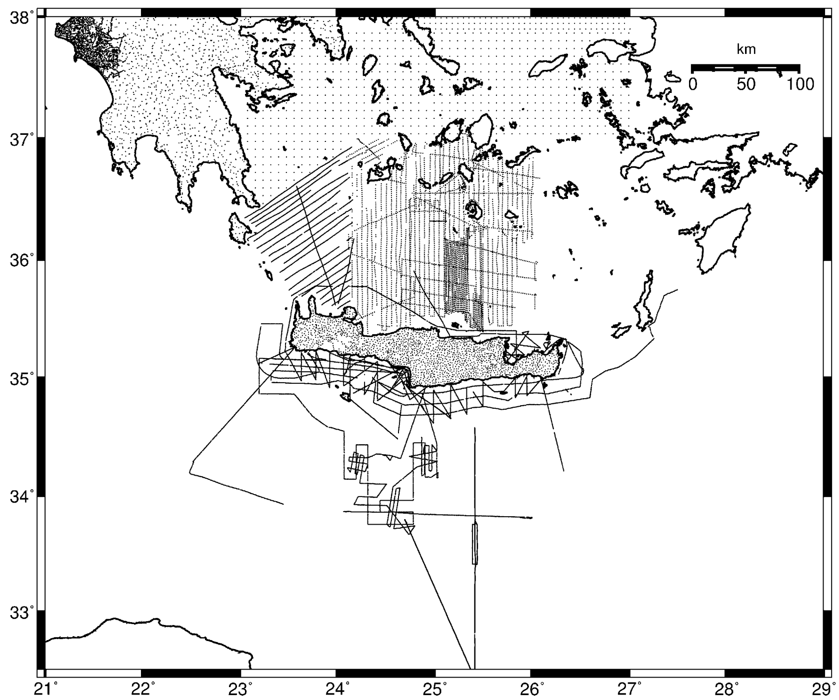


Fig. 2. Collected land and marine gravity data with outside coverage by gridded satellite based data (2′ grid, not shown).

gridded form and computing the gravity effect of vertical bodies, the free-air anomalies of observation points (land and shipborne) and grid points (satellite borne) were reduced for the mass effects of topography/bathymetry and Bouguer plate. The standard reduction density of 2.67 g/cm^3 for all continental masses and of 1.64 g/cm^3 for the water cover in marine areas were used, with this having in mind the comparability with the results of other gravity interpreters. The reduction radius was 150 km (with subdivision into three zones). The topographic data were taken from "The Global Land One-km Base Elevation (GLOBE) Project" (http://www.ngdc.noaa.gov) and the bathymetric data from the same source as the satellite ones given above.

The resulting Bouguer map, gridded with 1′ spacing, is shown in Fig. 3. Four regions dominate the gravity field from north to south: (1) the gravity high (up to 170 mGal) in the Cretan Basin — the forearc region between the volcanic arc and the island high; (2) an arc-shaped gravity low (relative, from 0 mGal to +50 mGal), which connects Crete with the Peloponnesos to the west and with Anatolia to the east — the island high of the forearc region; (3) a variation of anomalies (from −10 mGal to +100 mGal) outside the island high — the area of the arc-shaped Mediterranean Ridge; (4) gravity highs in the south–west and in the south–east towards the African continent. The two gravity lows of the ridge area – one south off Crete and the other south–east of the island – give rise to speculations on their origin, because they are local disturbances of the arc-shaped anomaly field.

4. 3D forward modelling

4.1. Seismic a-priory information

In order to control the parameter geometry in gravity modelling, the earlier results of several seismic experiments and of three gravity modellings mentioned above (Truffert et al., 1993; Wang, 1995; Fruehn et al., 2002) were used in this study as far as possible. Seismic structures at distinct positions will be given later together with the density structure of gravity modelling.

In the west, perpendicular to the Hellenic trench, a transect of marine expanded spread profilings ESP 8, 9 and 11 (Pasiphae cruise) give the velocity structure in the collisional area of the African and Eurasian plates and is confirmed by 2-dimensional gravity modelling of free-air anomalies (Truffert et al., 1993). The oceanic Moho dips below the Mediterranean Ridge with about 1° towards the collisional front. At ESP 8 Moho depth is about 22 km. The 11 km thick oceanic crust of the African plate is covered by water (depth of 3 km) and sedimentary layers with a total thickness of about 9 km.

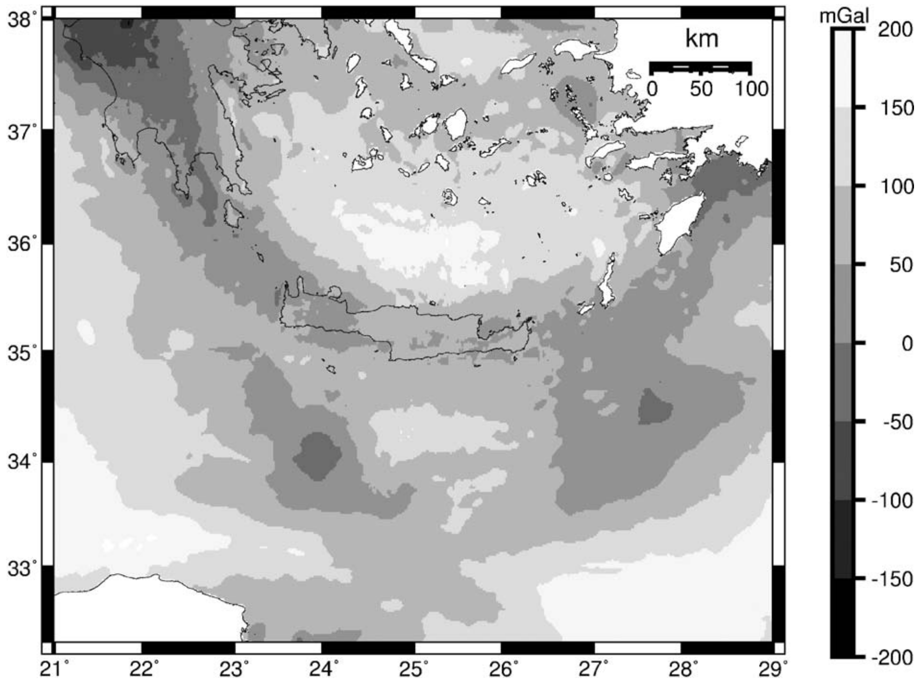


Fig. 3. Map of compiled Bouguer gravity field of the Hellenic subduction zone (reduction density 2.67 g/cm^3, gravity datum IGSN71 system). Amplitudes of anomalies are grey scaled (light colour — gravity high, dark colour — gravity low). Anatolia, Rhodes, Scarpanto and other small islands are not covered by gravity stations.

At ESP 9 Moho depth is about 23 km. Here the upper part of crustal section (thickness of about 8 km plus water depth of 3 km) is continental and contains mainly compacted sediments of the backstop area of the Eurasian plate (bodies with higher velocity and density) and the lower part the oceanic crust of the African plate (again 10 km thick). In between a sedimentary layer (about 1 km thick, away from ESP 9 at 3 km!) has been observed. Finally, at ESP 11 at the Hellenic trench, Moho depth is about 32 km. A thinned-out continental crust (11 km thick, water depth 5 km) covers the subducted oceanic crust (9 km thick, 20° inclination) with material of lower velocity in between (thickness of 4 km). In this structural model the Eurasian plate has an 8 km thick transition zone between crust and mantle, which is necessary to fit the gravity anomalies by modelling. Subduction and the collisional front between the two plates have been located about 150 km south–west of the Hellenic trench. More detailed information concerning the velocity structure of the sedimentary body is given by the results of the International Mediterranean Ridge Seismic Experiment (IMERSE) on a transect passing ESP 19 and ESP 9, as reported by Fruehn et al. (2002). It is possible to trace the dipping décollement to a depth of 6 km below the central ridge, and to identify the presence of evaporites (Messinian) on top of it in the outer domain. Controlled by gravity modelling total thickness of the sedimentary cover is 10 km and Moho depth is 18–20 km. Jones et al. (2002) investigated the velocity structure at the backstop front located between ESP 8 and ESP 9. The oceanic crust in this study is encountered at a depth of about 14 km.

The middle part of the study area, covering the arc region from the Cretan Sea to the Libyan Sea (Mediterranean Ridge), has been surveyed by seismic onshore–offshore experiments WARRP 1 to 3 (WARRP — wide aperture reflection and refraction profiling) on Crete and in the marine area around the island (Bohnhoff, 2000; Bohnhoff et al., 2001). The crust is of continental type and has a maximum thickness of 32.5 km below northern central Crete. It is subdivided into upper and lower crust. The eastern part of Crete shows a significantly thinner crust of 24–26 km. To the north, crustal thickness decreases to 15 km below the central Cretan Sea. To the south and south–west off Crete, the continental crust gradually thins out to a minimum of 17 km, and at approximately 100 km off the southern coast of Crete, it is in contact with oceanic crust below the Mediterranean Ridge. Below the continental Cretan crust, a 6–7 km thick layer was identified. Below western Crete it is still in contact with the overlying crust (decoupling may take place north of western Crete), while below eastern Crete it is decoupled and is interpreted as oceanic crust presently under subduction towards the NNE below the Aegean Sea.

In continuation of the Cretan seismic experiment, the Mediterranean Ridge was recently surveyed by three other wide aperture seismic onshore–offshore experiments WARRP 4–6 (Makris and Yegorova, 2002; Brönner, 2003). The western line (WARRP 5), nearly perpendicular to the Hellenic trench, clearly demonstrates the shortening of the oceanic crust between the continental ones of Africa in the SW and Crete in the NE. The continental approach is only 50 km and takes place below the Mediterranean Ridge. The 6 km thick oceanic crust is covered by about 10 km of sediments. Moho depth is 18 km, and increases sharply to 26 km below the African continental crust, which is 13 km thick and also covered by about 10 km of sediments. Between the collisional front and Crete – 150 km off Crete – the crustal structure has completely changed into a thin sedimentary cover (about 3 km) and a 10 km thick continental crust followed by the oceanic one. Moho depth becomes about 23 km and increases towards Crete, while the continental crust thickens. Subduction structures have not been detected. The central line (WARRP 4) gives similar crustal structures but with less approach of the two continents, now 110 km apart from each other. The eastern line (WARRP 6) resolves the crustal structure of eastern Crete and the marine area of the Plini and the Strabo Trench. The collisional front is located at the Pliny Trench. A subduction structure is given by the decoupling of the oceanic layer, which starts below the Pliny Trench and shows a mantle wedge below continental Crete with a crustal thickness of 22 km only. Several other profiles that cross the Mediterranean Ridge have mirrored the structure of its sedimentary body but without resolving the deeper crust (Chaumillon and Mascle, 1997).

Finally, a model must be taken into account which is based on receiver function analysis of data collected by temporary networks of seismic stations on western Crete. The crust below western Crete consists of a continental top layer of 15–20 km thickness above a 20–30 km thick subducted fossil accretionary wedge. The downbent oceanic Moho lies at a depth of 50–55 km (Knapmeyer and Harjes, 2000; Endrun et al., 2004). This model is not compatible with the above results of the WARRP experiments for western Crete, but the Moho depth has been proven by manifold interpretation. This value is also given in recent seismological studies, one about Rayleigh wave phase velocities (Meier et al., 2004a) and a second one about shallow seismicity observed by temporary networks of seismic stations in Crete and on the island Gavdos (Meier et al., 2004b).

4.2. Petrophysical parameters

The seismic structures as given above demonstrate a lateral variation of v_p-velocities. Especially the sediments have the expected wide range of values. For regional gravity modelling, which of necessity cannot take into account detailed sedimentary structures and the distribution and lateral variation of evaporites for the Mediterranean Ridge, a certain grouping of the structural units was necessary before the density values could be determined. The model region was subdivided into African continental, African oceanic, and Eurasian continental. The corresponding vertical structures were subdivided into sediments (with further subdivisions), upper crust, lower crust, and upper mantle. Table 1 summarises the structural units, the observed velocities and the estimated as well as the densities finally used. Density estimation was done applying velocity/density relations as given by the Nafe and Drake curve for sediments and crustal rocks (Ludwig et al., 1971; Fowler, 1990) and by Birch's linear relation for mantle rocks (Birch, 1964).

The largest variation of rock properties is given by the sediments from unconsolidated (marine, mean 1.7 km/s and 1.8 g/cm^3) to consolidated (backstop, mean 5.5 km/s and 2.6 g/cm^3) material. Evaporites (4.25 km/s and 2.2 g/cm^3) were not taken into account, because their presence and layer thickness has too strong local variations to become an element of the regional density model. Metamorphic limestone, as observed on Crete, has properties (up to 6.0 km/s and 2.7 g/cm^3) which come near the crustal ones. Continental crustal material is subdivided into upper (5.8–6.6 km/s and 2.65–2.85 g/cm^3) and lower crust (6.4–7.2 km/s and 2.8–3.0 g/cm^3), while oceanic crustal material is a single layer in WARRP interpretation but has a subdivision in ESP gravity interpretation. In the present study a two-layer oceanic crust is used for modelling (2.9 and 3.0 g/cm^3). Two properties are badly controlled. One is the density of 2.4 g/cm^3 for consolidated sediments of the African oceanic crust. This value is too low for sediments in greater depth, but is taken as a mean value. The other is the density of 2.45 g/cm^3 for limestone of the Eurasian continental crust. In this case the present study is not modelling the soft sediments on Crete with varying thickness, but rather soft and hard sediments together as one "limestone layer" with lower density.

Table 1
Petrophysical parameters of 3D model structure. Used density values are given in the cross sections of Fig. 7a–d also

Structural unit	Velocity (km/s)	Density (g/cm^3) (Nafe and Drake curve)	Density (g/cm^3) (Birch's relation I)	Density (g/cm^3) (used)
African continental				
Sediments	2.5–4.9	2.1–2.5		2.45
Upper crust	5.9–6.4	2.5–2.8	2.49–2.71	2.7
Lower crust	6.5–6.95	2.8–2.9	2.76–2.95	2.85
Mantle	7.9		3.36	3.35/3.4
African oceanic				
Sediments (soft)	1.7–	1.8–		2.25
Sediments (consolidated)	4.9	2.5		2.4
Crust 1	6.5–	2.8–	2.76–	
Crust 2	7.1	2.95	3.02	2.9
Mantle	17.9–8.0		3.36–3.39	3.0
				3.4
Eurasian continental				
Sediments—Libyan Sea	1.8–4.45	1.8–2.45		
Sediments—Crete (soft)	2.2–3.9	2.0–2.35		2.3
Sediments—Crete (limestone)	5.8–6.0	2.65–2.7		
Sediments—Cretan Sea	1.8–4.35	1.8–2.45		2.45
Sediments—backstop	5.0–5.5	2.55–2.6		2.3
Upper crust	5.8–6.6	2.65–2.85	2.45–2.80	2.6
Lower crust	6.4–7.2	2.8–3.0	2.71–3.07	2.7
Mantle	7.7		3.30	2.85
				3.3

The physical properties of the uppermost mantle can be separated into African and Eurasian ones. The African part has the higher (8.0 km/s and 3.39 g/cm^3) and the Eurasian the lower values (3.25 km/s and 3.30 g/cm^3). This is due to the "hot" and updoming mantle below the Aegean crust in accordance with the higher heat-flow density values as given by Cermák et al. (1977) and the temperature distribution models published by Makris and Stobbe (1984).

The one-dimensional velocity model BO97 (Engel, 1998), based on joint hypocenter determination of local seismicity of western Crete, was not taken into account. This model has a low-velocity layer (5 km/s) between 13–18 km, similar to that derived by Papazachos and Nolet (1997), and gives a velocity of about 8.2 km/s at a depth of 41 km. While the oceanic Moho of the subducted slab below Crete gives a clear signal in seismic interpretation, the lower continental crust and Moho are still a matter of discussion.

4.3. Modelling technique

The computer-aided gravity modelling was done using the software package IGMAS (for background information see: Götze, 1984; Götze and Lahmeyer, 1988). Fig. 4 illustrates the fundamentals of 3D model construction. As input (starting) model the density structure is given by nodal points (with x,y,z-coordinates and density values) within cross sections arranged parallel to each other and covering the study area. The nodal points represent lines separating the density bodies within a certain cross section. The nodal points of corresponding lines within neighbouring cross sections are automatically connected by triangulation, giving the surfaces of the density bodies. The gravity effect of each 3D (polyhedral) body is computed and added to the effect of the other ones, giving the anomaly value of the model structure for a given observation point. The structure, together with computed and observed anomalies, is displayed on the monitor. The computed anomalies are fitted to the observed ones by density and structural changes, performed interactively. Generally, the levels of measured and computed gravity fields are different. To enable the comparison between the two fields an automatically determined constant shift value is added to the computed anomalies.

The Hellenic subduction area was divided into 16 south–north oriented and 600 km long cross sections with a regular spacing of 50 km (Fig. 5). From east to

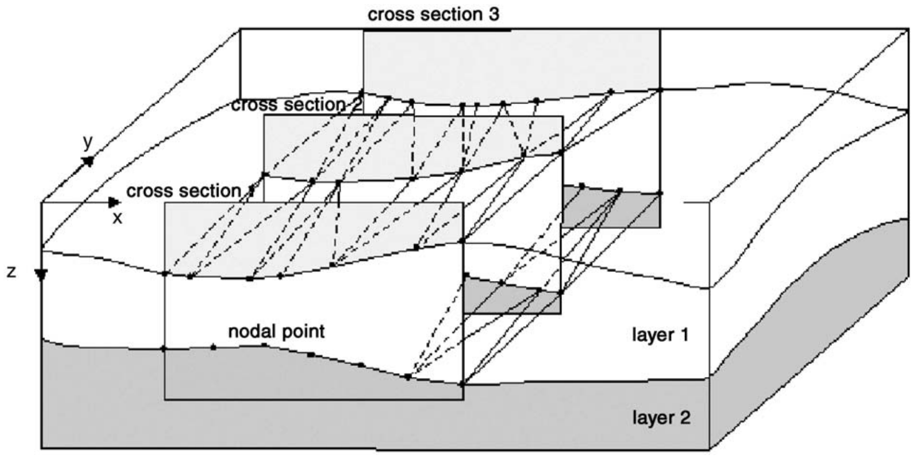


Fig. 4. Illustration of 3D model geometry for gravity computation with the software package IGMAS (for explanation see text).

west the model structure is 700 km wide. To suppress edge effects, the density model was extended in all directions with the last given values for up to 10,000 km. Depth extension is 100 km. None of the sections runs parallel to one of the seismic lines, but is crossed by them. From this it follows that the starting density model structure is taken from the seismic structure as given at these crossing points. It also follows that the best control is expected for the central part, the island of Crete and its surroundings. The eastern part is weakly controlled by the seismic line WARRP 6 only.

4.4. Modelling results

Fig. 6a shows the computed Bouguer map. The main anomalies, as described above, could be satisfactorily modelled. The standard deviation of misfits is about 10 mGal. The fit is achieved on the basis of the dom-

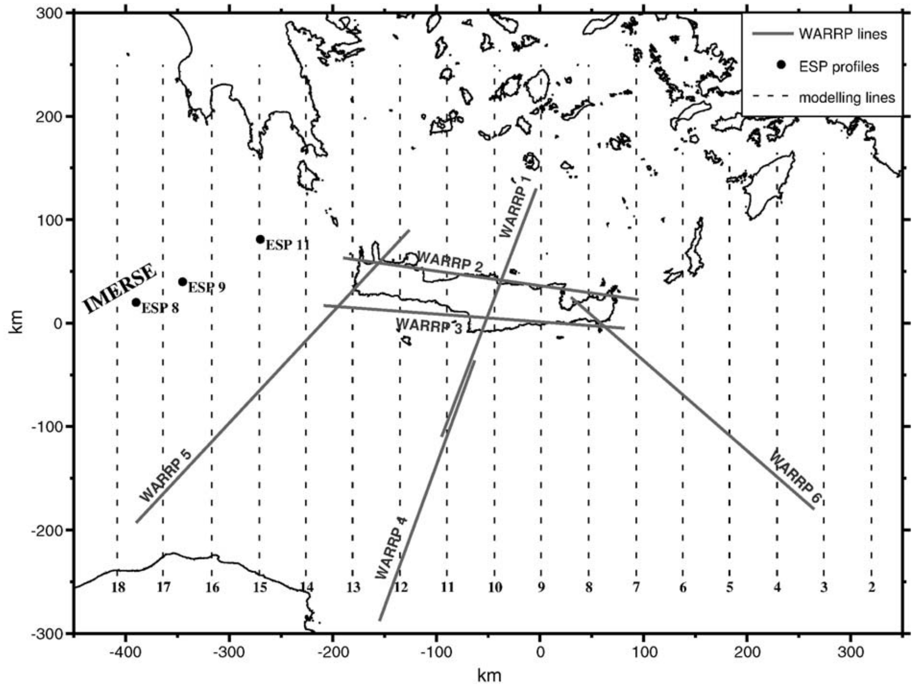


Fig. 5. Area of 3D modelling. Locations of controlling seismic deep soundings are given as points (expanded spread profilings ESP 8, 9 and 11 and part of the IMERSE transect by Truffert et al., 1993; Fruehn et al., 2002) and lines (wide aperture reflection and refraction profiling WARRP 1, 2, 3 by Bohnhoff, 2000, and WARRP 4, 5, 6 by Brönner, 2003). Parallel broken lines 2 through 18 give the cross sections for gravity modelling.

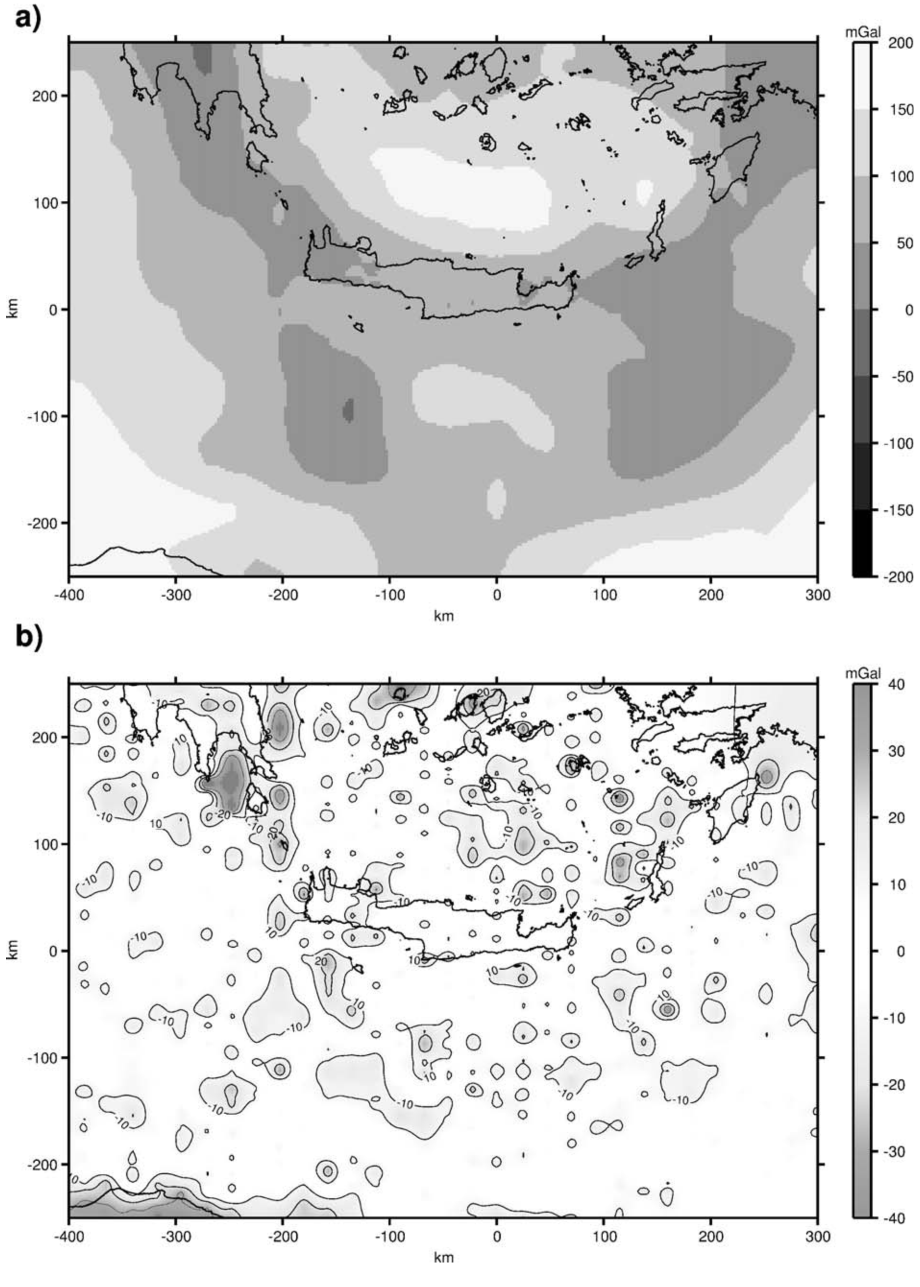


Fig. 6. Results of gravity modelling in map presentation. (a) Computed Bouguer gravity field giving the main, low frequency, anomalies of the subduction area. Amplitudes of anomalies are grey scaled (light colour — gravity high, dark colour — gravity low). For better comparison, scaling is the same as in Fig. 3. (b) Residual gravity field (observed minus computed) giving high frequency anomalies (positive and negative) mainly due to the incompletely modelled sediments and the insufficiently controlled structure of the African continent.

inant long wave length part of the field. This is mainly controlled by undulations of Moho depth. The shorter wave length part, mainly affected by the sedimentary cover, is modelled with less accuracy, as demonstrated by the residual map of Fig. 6b. This is evident throughout the region. The marine area around Crete has modelled values higher than the observed ones (negative residuals). In contrast, Crete and the marine area west of Gavdos have lower values (positive residuals). The negative residuals of the southern Peloponnesos, Cyclades and Rhodes reflect the insufficiently controlled model in those areas.

Selected cross sections shall be used as examples to illustrate the results of gravity modelling in detailed form

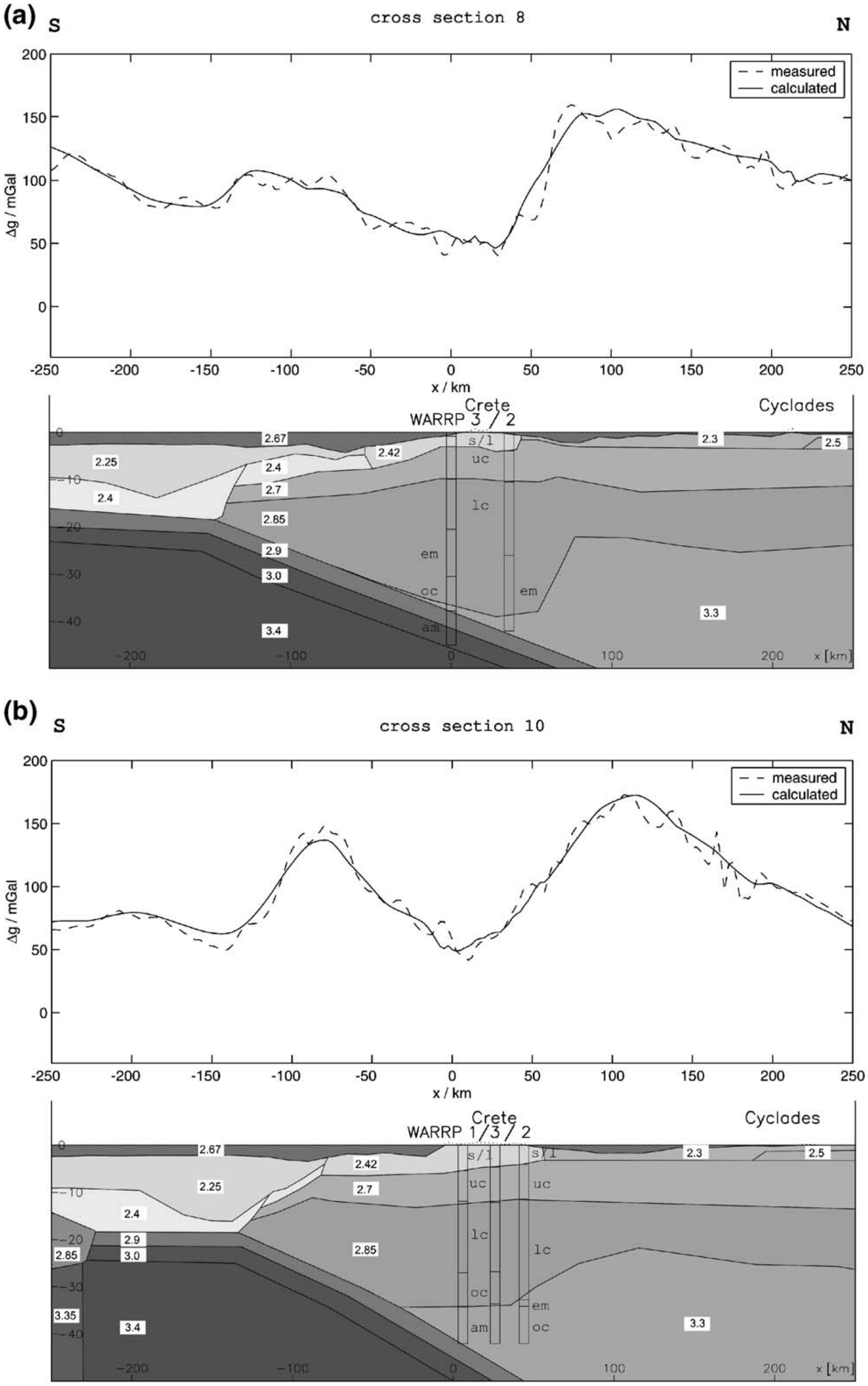
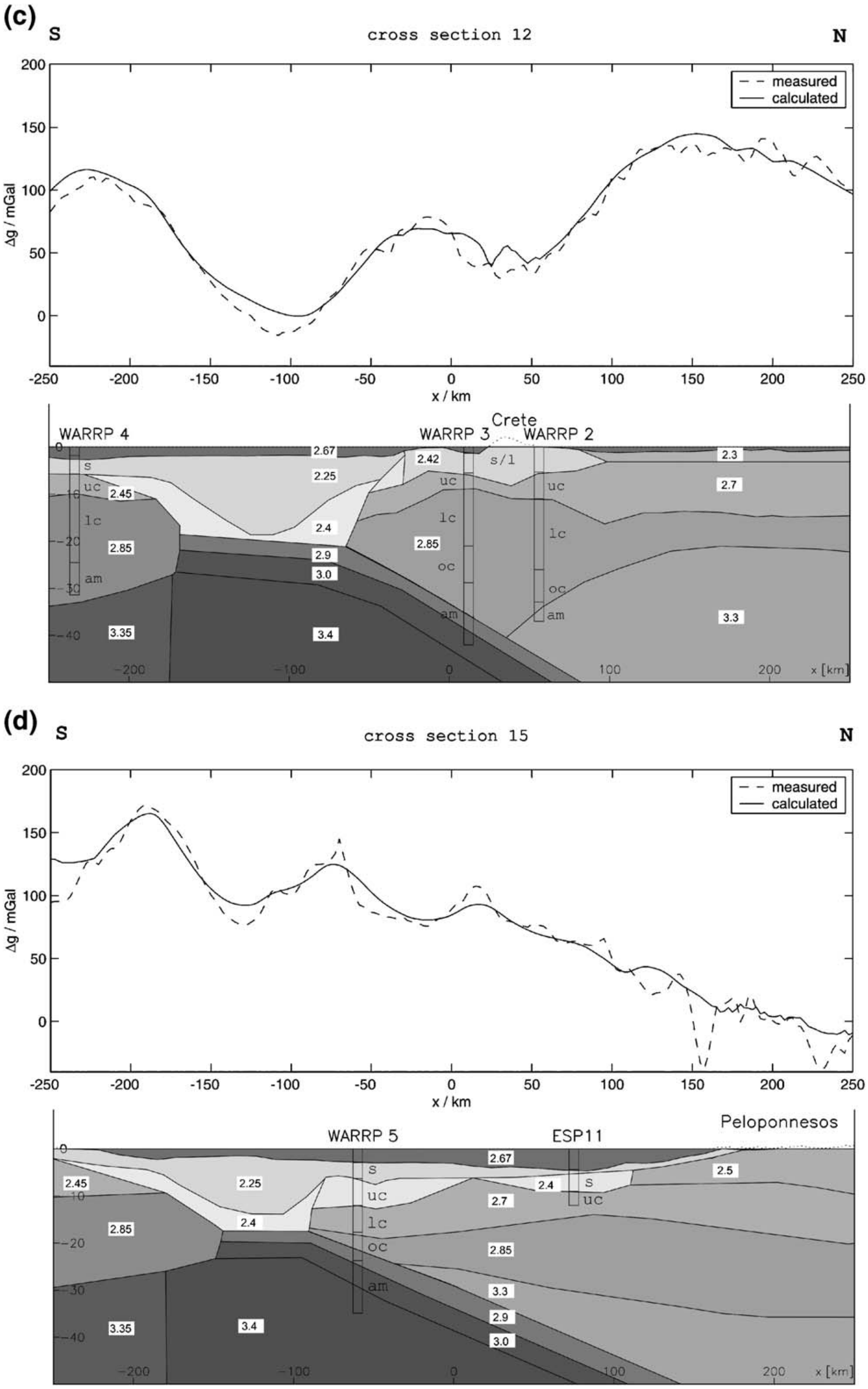


Fig. 7 (*continued*).

(Fig. 7a–d). The cross sections display density structure in the lower panel and computed as well as observed gravity values in the upper one. The different structural elements are grey toned following their density scaling. Seismic models for the above mentioned crossing points are added as structural columns.



Cross section 8 (Fig. 7a): The section passes eastern Crete, the gravity high of the Cretan Sea in the north, and the high south of the island. The section runs nearly perpendicular to the arc-shaped structure of the subduction zone, but oblique to the Plini and Strabo trenches. Its structure is crossed by the seismic lines WARRP 2 and 3, both on Crete.

The density model gives the lateral change in crustal type from African oceanic to Eurasian continental, with the subducted slab dividing the upper mantle into bodies with different densities. The 8 km thick oceanic crust, slightly dipping to the north, is covered by more than 15 km of sediments (consolidated and soft) and a water layer of about 2 km. Oceanic Moho here is at a depth of about 24 km. Collision with the Eurasian continent takes place at the backstop front 140 km south–east off Crete at the Strabo trench. The oceanic crust dips down at about 10°. Depth of the oceanic Moho below Crete is 50 km. Continental thickness below Crete exceeds 35 km. The continental body is subdivided into upper and lower layers, below Crete at a depth of about 10 km. A mantle wedge is just beginning to open below Crete. The crust of the continental slope area and of Crete is covered by backstop sediments and limestone. North off Crete, a steep rise of upper mantel to 21 km is realised. The crust there is covered by marine sediments.

Cross section 10 (Fig. 7b): This section runs through central Crete and passes the gravity highs in the south and in the north of the island. It crosses the subduction arc perpendicularly and the Ptolemy and Plini trenches obliquely. The structural model is crossed by the seismic lines WARRP 1, 3 and 2 on Crete.

The density model again demonstrates the change in crustal type from African oceanic to Eurasian continental with the collisional front 130 km off Crete. The subducted oceanic crust and mantle division seen in cross section 8 is evident as well. The trend of the gravity anomalies reflects Moho depth variations from 24 km below the oceanic crust, 34 km below continental Crete, 22 km below the Cretan Sea, to about 25 km below the Cyclades. Crustal subdivision below Crete takes place at a depth of 12 km. Depth of the sedimentary cover (soft and consolidated) is 18 km below the Mediterranean Ridge, about 6 km on Crete and 7 in the marine area south of the island (limestone and a small body of backstop sediments just at the collisional front), and 3 km below the Cretan Sea.

Cross section 12 (Fig. 7c): In this case the section passes western Crete, the gravity high of the Cretan Sea in the north, the low between Crete and Africa, and the high towards the African continent. The section runs nearly perpendicular to the subduction arc and oblique to the Hellenic and Ptolemy trench. The structure is crossed by the seismic lines WARRP 4 at the African continental edge and WARRP 2 and 3 on Crete.

The density model shows the change in crustal type, in this case from African continental to African oceanic and then to Eurasian continental, and the subducted slab. The oceanic crust is covered by up to 18 km of sediments and a water layer of about 2 km. Moho depth here is about 28 km, again with a slight dip to the north. Collision with the Eurasian continent takes place 90 km south off Crete, and the slab again dips down at 10°. Continental crustal thickness increases to 40 km below Crete, there without a mantel wedge. Subdivision of the crust below Crete takes place about 10 km. Further to the north, Moho depth decreases to 23 km (below the Cretan Sea). Striking is the up to 7 km thick limestone cover of Crete and the marine area south of the island, with an abrupt edge at the Pliny trench. Backstop sediments are not significant.

At the southern end of the section the crustal structure of the continental edge of Africa produces the gravity high. The structural change from continental to oceanic crust reflects in a Moho depth rise from more than 30 km up to 27 km.

Cross section 15 (Fig. 7d): This section is oblique to the western wing of the arc-shaped structure and the Hellenic trench. It passes the trench west off Crete, runs into the gravity low of the Peloponessos in the north, and has a southward continuation passing the ridge sediments and crossing the African continental edge. The structure is crossed by seismic line WARRP 5 and met by ESP 11.

Fig. 7. Selected cross sections from 3D gravity modelling (output from IGMAS) showing the density structure (lower panel) and the gravity anomalies (upper panel). Water cover has a Bouguer density of 2.67 g/cm^3. Depth of structures shown is 50 km. Seismic models are given as structural columns for relevant intersections (s — sediments, l — limestone, uc — upper crust, lc — lower crust, em — Eurasian mantle, oc — oceanic crust, am — African mantle). (a) Cross section 8. Main elements of anomalies are given by the southernmost edge of the Eurasian continent, the thick continental crust below Crete and the thin crust below the Cretan Basin. (b) Cross section 10. Main elements of anomalies are the same as given in (a) but more expressed at the continental edge. (c) Cross section 12. The African continental edge is producing a further anomaly in the south. The other elements of anomalies are a continuation from (a) and (b). (d) Cross section 15. Main elements of anomalies are given by the continental edges of Africa and Eurasia with a thick sedimentary cover of oceanic crust (African) in between. The negative trend of anomalies towards the Peloponnesos is affected by thickening of the Eurasian crust.

The density model contains both continental edges, here only 50 km apart, with a remnant of oceanic crust in between. The latter is covered by sediments having a maximum thickness of 15 km (water cover is 3 km). The Eurasian continental edge is located about 150 km south of the Hellenic trench, covered by a layer of consolidated continental sediments and one of overlaying marine sediments (further to the north). Depth of continental Moho is 18 km at the edge and increases continuously towards the Peloponnesos, reaching 36 km. The slab dip is less than 10° because of oblique geometry, giving space for a mantle wedge.

The African crust has a structure similar to the Eurasian one, but is poorly controlled and therefore uncertain, and is covered by marine sediments (soft and consolidated). Towards the Mediterranean Ridge the crust thins out, while sediment thickness is increasing.

One of the present tasks was to outline the course of both the African and the Eurasian continental edges in the area of the Hellenic collisional zone. This is now possible, following the margins of the crystalline crust in the density model from one cross section to the next. The result is given in Fig. 8a. The exposed width of the African oceanic crust between the two continents, here given by the shaded area between the depth contour lines at 15 km, varies from 50 km south–west off Crete to 100 km and more at the eastern wing of the subduction arc. Following the 15 km depth line, the Eurasian edge has a clear emplacement towards the north–east in the south of western Crete at the Ptolemy trench pointing towards eastern Crete. In the same area there is a change in covering material from consolidated continental sediments (in the west) to limestone (in the east), as given in the above sections. Both wings of the edge have a more linear than curved geometry. The steep local gradients of the contour lines at some cross-sections are well controlled by the density structure.

A second task was to follow the depth trend of the Moho, which controls the main part of the gravity anomaly field, i.e. the low frequency part, which varies between 40 and 150 mGal. This is demonstrated well in Fig. 8b, showing the arc-shaped structure of the continental Eurasian Moho, with its uplift below the Cretan Sea (Moho depth less then 20 km). Moho depth below Crete varies from about 35 km in the west to 32 in the center and again 35 in the east. The above mentioned crustal emplacement continues down to the Moho.

The third task, the structure of the subducted oceanic Moho, is answered in Fig. 8c. The Moho is clearly arc-shaped with a down-turn line situated below the Eurasian continental edge, which means below the Mediterranean Ridge. The arc is open to the north. The spacing of depth lines around Crete gives a dip angle of about 11°.

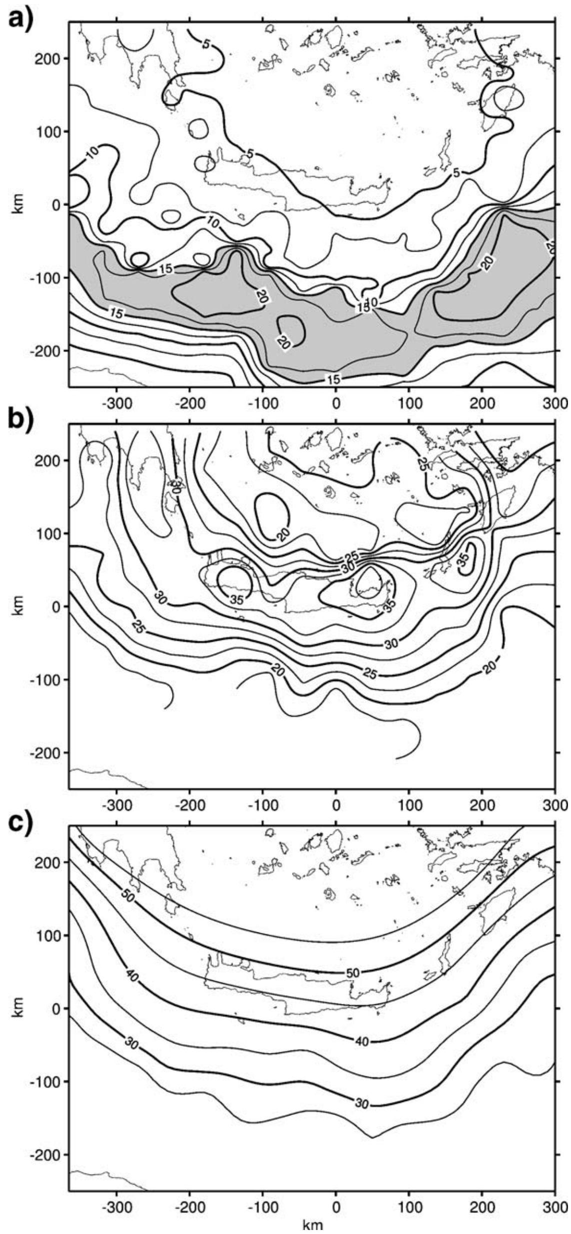


Fig. 8. Mapped model surfaces outlining the deep structure of the Hellenic subduction zone. The contour lines give depth in km. (a) Triangular-shaped southernmost part of continental crust (Eurasian) without sedimentary cover with emplaced edge south of western Crete, continental edge (African) and oceanic crust (African) in between as shaded area. (b) Arc-shaped continental Moho (Eurasian), thickened along the Hellenic rise and doming up below the Cretan Basin. (c): Arc-shaped and subducted oceanic Moho (African).

5. Discussion and conclusions

In general, the regional density model follows the main elements of the seismic structure, which was used to control the starting model as far as possible, at least at the intersections between the modelling sections and seismic lines. The areas between the seismic lines are, of course, the result of gravity modelling alone. As demonstrated by the above cross sections, the best fit between modelled and observed anomalies is given for the low frequency part, mainly the effect of the Moho surface. The misfit in the high frequency part is due to the fact that the sedimentary cover was not modelled in detail, and, as can be seen in cross section 15 above the continental edge, by faulty input data (here satellite data) also. While cross sections 8, 10 and 12 are oriented predominantly perpendicular to the arc structure, the course of section 15 is oblique to it, thus increasing the possibility of modelling misfits. It is necessary to mention that the cross sections shown depict a 2D figure of the 3D-treated gravity field, which can also result in local misfits between the modelled and observed anomalies.

The gravity high of the Cretan sea is the result of an updoming mantle in connection with forearc extension, and is not attributable to the slab at depth as assumed by Papazachos and Comninakis (1971). In the cross sections of Fig. 7 the minimum depth to Moho is quite variable. There is a steep transition to greater depth below northern Crete and a more steady increase towards the Cyclades. The sections shown do not intersect the actual maximum value of the gravity high, where Moho depth is less than 20 km. This value is the maximum on the 3D surface in Fig. 8b, as given by seismic interpretation of line WARRP 1. The thinning out of continental crust below the Cretan Sea is expressed in the positive isostatic anomalies (Fleischer, 1964; Lagios et al., 1995).

Concerning crustal structure below the gravity high north off Crete and the steep negative field gradient towards the island (see cross section 8), it can be stated that the density model verifies the velocity model quite well. The residual anomalies from Fig. 6b are within the noise level, while the previous modelling of Tsokas and Hansen (1997) resulted in clear positive residuals. Obviously, in that case the model was not sufficiently controlled.

The density model of Crete and its marine surroundings should be best controlled, compared with the other regions, by the results of active seismic experiments along the lines WARRP 1 to WARRP 5, and the seismic array observations on Crete. Continental crust (Eurasian) and the subducted oceanic crust (African) has been proven, but the final density model shows a structure different from the WARRP velocity and density model as given by Makris and Yegorova (2006-this volume). Independently executed gravity modelling applying different computer software has resulted in different models. Depth of sedimentary cover and upper crust could be verified, but not Moho depth. Fixing depth of upper crust at about 10 km and oceanic mantle at 50 km makes it impossible to verify a model with a 30 km (or less) deep reaching continental Moho and continental mantle in between. Gravity would become higher values than observed (about 35 mGal shift in cross section 12 between 0 and 50 km). Therefore, the crust/mantle region below Crete must consist of lower density. Without having further information this could be realised by filling out more of the volume with material having a density value of the lower crust. Consequently, continental crustal thickness increases up to 35 km with an east–west variation. Below eastern Crete (cross section 8) the thickness is 34 km with a small mantle wedge below. This model doesn't fit the WARRP 2 and 3 results. Crustal depth of 32 km and 10 km mantle wedge is met for central Crete (cross section 10). While in this case the WARRP 2 result is verified, the WARRP 1 and 3 structures are not met. Below western Crete (cross section 12) the thickness is reaching 34 km. While the WARRP 2 model is nearly verified the WARRP 3 again is not. Worth mentioning is the sudden crustal thinning from 34 to 31 km from cross sections 12 to 13, which limits the Cretean block in the west.

While the density model contains continental mantle material below Crete, the seismic continental Moho is directly followed by the African oceanic crust below western Crete, and the detachment starts below central Crete and leaves space for mantel material below eastern Crete (WARRP 2 and 3 in cross section 8). Depth to oceanic Moho below Crete, as given by line WARRP 3, becomes about 30 km below western Crete, 36 km below central Crete and 38 km (and more) below eastern Crete, again with a dip to the north. The obvious discrepancy between gravity modelling, as discussed here, and that one presented by Makris and Yegorova (2006-this volume) mainly illustrates different interpretations concerning slab detachment and depth to oceanic Moho. While the WARRP models are not taking into account lighter material in the crust/mantel region, its existence is essential for the present gravity modelling. No such low-velocity layer has been detected by active seismic experiments, which is in opposition to passive ones. Concerning the role of ridge sediments (interpretations 2 and 3, see Introduction), it is therefore likely that sediments are underplating the southern Aegean, at least the Cretan block. Another interpretation may be discussed in connection with the circulation of material in a subduction channel, according to Gerya et al. (2002), resulting in the accumulation of lighter material. Consequently, the negative residuals of previous gravity modelling given by Tsokas and Hansen (1997) seem to be related to low-velocity material. Seen in the light of isostatic behaviour, deep reaching lighter crustal material can be the source for the uplift of the Cretan block.

Following the present density model, the detachment of the African oceanic crust takes place south off Crete, and with this the southernmost edge of the Eurasian continental edge, acting as backstop front for the oceanic sediments, is defined. While the seismic experiments (lines WARRP 4–6) have proven this at three single positions only, gravity modelling of the anomalies at these and neighbouring positions gives the more rectangular course of the front line including the emplacement south off Crete, which may be linked to the Ierapetra Transverse Fault known from eastern Crete (LePichon et al., 2002). In the west the position of the backstop front corresponds with the position as given by ESP-IMERSE interpretations (Jones et al., 2002), and in the east it corresponds with the position given by Huguen et al. (2001). Continental thickness, with continuation of structural discrepancies, decreases southwards from Crete towards this front, and the high density sedimentary cover of the backstop body has varying thickness and different material, i.e. consolidated sediments in the west and uplifted basement blocks in the east, as demonstrated in the sections. This result fits the interpretations concerning the inner domain (Chaumillon and Mascle, 1997; Huguen et al., 2001). Subducted oceanic sediments are not present in the structure, as proven by the seismic experiments.

The African oceanic crust is deep-seated, with negative isostatic anomalies (Fleischer, 1964; Lagios et al., 1995), and is covered by sediments of extreme thickness overthrusting northward the continental sediments of the inner domain. This model supports the role of the Mediterranean Ridge as an accretionary wedge (interpretation 1, see Chaumillon and Mascle, 1997; Huguen et al., 2001). The décollement within this sedimentary body is not outlined in the density model. Oceanic Moho depth is reaching 27 km between Crete and Lybia (23 km in the west). Throughout the area, the Moho has a slight dip towards the north, which means that the remaining oceanic crust is affected by subduction in its full extension. One of the speculative gravity lows of the Mediterranean Ridge is situated just at the emplacement of the above crustal fault. The interpretation as an effect of deep reaching sediments is not controlled by seismic investigations. Its steep east–west gradient is mostly determined by the local gravity high in the east (uplifted limestone cover of the backstop area). The second gravity low, east of WARRP 6 and therefore also not controlled, again is the result of deep reaching sediments. The detachment and downbending of the oceanic crust follows the continental edge and gives the impression of an amphitheatre-like shape with an opening to the north. This is in full agreement with the Wadati-Benioff zone as given by Papazachos and Nolet (1997) and others. The smaller zones, given by Kovachev et al. (1992) and connected with the Pliny and Strabo trenches, could not be traced. The interpretation is not supported by the model of Makris and Yegorova (2006-this volume), who have presented a model, which is based on an oblique subduction directed to the north–east.

The course of the Eurasian continental edge is followed in the density model, controlled by seismic structures, but the course of the African edge remains uncertain. The WARRP soundings are about 200–400 km apart and the modelling sections do not cover the necessary range. As far as possible, gravity modelling of the edge followed the course of positive anomalies off the African coast (compare with cross sections 12 and 15). The question arises as to, whether the crust to the south–east of the Mediterranean Ridge is of continental type or not. The present density model incorporates the continental type, as given by the seismic velocity structure of WARRP 4 (cross section 12) and WARRP 5 (cross section 15). WARRP 6 could not resolve the crustal structure at its south-eastern end. While the thickness of sediments and upper crust was taken as given by the seismic models the Moho depth in the velocity structure had to be shifted down to 31 km in order to reach an approximate fit of the anomalies. Otherwise, only lower density values of lower crust (out of the velocity range) and upper mantle will fit gravity. In the presented model upper mantle density of the African continent was changed from 3.4 to 3.35 g/cm^3.

In the west the present gravity modelling largely follows the seismic structure of the collision area as given by WARRP 5 further in the south and by the ESP-IMERSE transect at the western edge of the model area. However, again a modification in layer thickness and interpretation concerning lower continental and oceanic crust was necessary (compare with cross section 15). Still, there is a remaining misfit in the anomalies of about +30 mGal at WARRP 5 demonstrating geometrical and or density effects, which are out of control. One reason may be lateral density changes, which could not be taken into consideration doing modelling with IGMAS. While the continental density model given by Truffert et al. (1993) for ESP 11 is met, the IMERSE 5a oceanic structure (Fruehn et al., 2002) is different. The oceanic Moho is placed in a depth of about 20 km (crustal density of 2.80 g/cm^3 and mantel density of 2.43 g/cm^3), while it is 23 km in the present study.

Summarising the results from the gravity modelling presented above it can be stated that the 3D interpretation has given more insight into the regional structure of the active African/Eurasian plate boundary at the Hellenic arc. Geometry and physical state of Eurasian and to some extent also African crustal elements and upper mantle have become outlined, thereby fixing the continental edges. Though the crustal and upper mantle structure below Crete still is a matter of discussion, the Moho depth generally is controlled in the collisional area, with some uncertainties concerning depth of the African continental part. The subducted oceanic crust separates plate material (upper mantle) with different densities. While the trend of the downbent plate (African) follows the arc-shaped geometry in accordance with the continental Moho (Eurasian), the upper continental crust (Cretan) together with its limestone cover has a more triangular shape, with its southern flanks parallel to the trenches. This difference in structural geometry correlates with the difference in relative plate movements, as interpreted by Papazachos et al. (1995) from tomographic images, leaving the question open as to where within the crust this change takes place.

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