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Crustal structure under the central High Atlas Mountains (Morocco) from geological and gravity data

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Abstract

Seismic wide angle and receiver function results together with geological data have been used as constraints to build a gravity-based crustal model of the central High Atlas of Morocco. Integration of a newly acquired set of gravity values with public data allowed us to undertake 2–2.5D gravity modelling along two profiles that cross the entire mountain chain. Modelling suggests moderate crustal thickening, and a general state of Airy isostatic undercompensation. Localized thickening appears restricted to the vicinity of a north-dipping crustal-scale thrust fault, that offsets the Moho discontinuity and defines a small crustal root which accounts for the minimum Bouguer gravity anomaly values. Gravity modelling indicates that this root has a northeasterly strike, slightly oblique to the ENE general orientation of the High Atlas belt. A consequence of the obliquity between the High Atlas borders and its internal and deep structure is the lack of correlation between Bouguer gravity anomaly values and topography. Active buckling affecting the crust, a highly elevated asthenosphere, or a combination of both are addressed as side mechanisms that help to maintain the high elevations of the Atlas mountains.

Keywords: High Atlas; Morocco; Bouguer gravity anomaly modelling; Crustal root; Topography; Isostatic compensation

1. Introduction

The High Atlas and its NE trending branch, the Middle Atlas, are intracontinental mountain chains located in the foreland of the Rif-Tell interplate orogen of North Africa (Mattauer et al., 1977; Pique et al., 2000; Frizon de Lamotte et al., 2000; Teixell et al., 2003). Previous knowledge of crustal thickness and internal structure of the High Atlas are based mainly on gravity modelling, seismic refraction and receiver function data (Makris et al., 1985; Tadili et al., 1986; Wigger et al., 1992; Seber et al., 1996;

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Sandvol et al., 1998; Mickus and Jallouli, 1999; Van der Meijde et al., 2003). The refraction data sets in which some interpretations are based have not been thoroughly presented and often have poor quality. In addition, the resulting crustal models have not been integrated with geological data. On the other side, receiver function data give only 1D velocity information that cannot be extrapolated without additional information. Few existing gravity models show varying results regarding depth and density of the different crustal layers. Therefore, the overall crustal structure and thickness across the High Atlas is partly unresolved.

In this paper we present two gravity-based crustal models along two geological transects across the central High Atlas of Morocco described by Teixell et al. (2003). These models are constrained by the available seismic refraction and local receiver function results, and are integrated with detailed geological cross-sections. The final crustal sections are in accordance with the pre-existing geophysical models regarding crustal thickness, and offer a more complete view of the crust across the High Atlas. The modelled crustal structure is compared with other intracontinental mountain belts where earlier rifts have been tectonically inverted by compression. Finally, we briefly discuss the mechanisms that support the high topography of the Atlas mountains.

2. Geological and geophysical overview

The intraplate Atlas mountain belt extends for more than 2000 km in an E–W direction from Morocco into Algeria and Tunisia (Fig. 1). It is composed of faulted and folded Paleozoic, Mesozoic and Cenozoic rocks with summits that in its western side reach more than 4000 m. To the north and northeast, the High Atlas of Morocco is bounded by the Meseta and the High Plateaux respectively, and to the south by the Anti-Atlas (Fig. 1), a mountain chain where topographic elevation is locally in excess of 2000 m, and by the Saharan Platform.

Previous works have shown that the structure of the High Atlas resulted from the tectonic inversion of a Mesozoic extensional basin, genetically related to the opening of the Atlantic and Tethys ocean (Choubert and Faure-Muret, 1962; Mattauer et al., 1977; Schaer, 1987; Jacobshagen et al., 1988; Beauchamp et al., 1999; Laville and Piqué, 1992; Pique et al., 2000). Compression related to convergence between Africa and Europe occurred from Cenozoic to present times (Laville et al., 1977; Dutour and Ferrandini, 1985; Görler et al., 1988; Fraissinet et al., 1988; Medina and Cherkaoui, 1991; El Harfi et al., 1996; Morel et al., 1993, 2000, etc.). The average compression direction was NNW–SSE according to Mattauer et al. (1977) and Gomez et al. (2000). Shortening was mainly achieved by thick-skinned thrusting and folding, affecting the pre-Mesozoic basement and a Mesozoic–Cenozoic cover. Thin-skinned thrusting usually played a minor role and appears related to basement underthrusting (Teixell et al., 2003). According to the recent work by these authors, the total shortening due to Cenozoic compression varies between 15% and 24% from west to east along the central High Atlas.

Rifting-related Triassic and Jurassic igneous rocks are widespread in the study area (Hailwood and Mitchell, 1971). In addition, Tertiary and Quaternary alkaline volcanism (Harmand and Catagrel, 1984) has been identified in the Middle Atlas, the Meseta, the High Plateaux, the northern part of the High Atlas, and in the Anti-Atlas. The tectonic context in which this volcanism appears is not fully understood yet and was explained by Harmand and Moukadiri (1986) as the result of upper mantle deformation which permitted partial melting in a compressive regime.

Seismic refraction studies provided the first estimations of crustal thickness across the High Atlas. Tadili et al. (1986) modelled refraction/wide angle data from an experiment carried out in 1975, concluding that the crustal thickness varies from 25 km along the Atlantic coast of Morocco to 40 km near Imilchil in the central High Atlas. To the east, these authors found values of 38 km in the Midelt-Rich area, descending to 35 km around Errachidia (see Fig. 1 for location). Part of the same data were also interpreted by Makris et al. (1985) who, in addition, modelled gravity data using densities (converted from velocities) of 2820 kg/m^3 and 2900 kg/m^3 for the upper and lower crust respectively. These authors concluded that crustal thickness of the western High Atlas, the most elevated part of the orogen (south of Marrakech), ranges from 34 to 38 km, depending on the assumed thickness of the sedimentary cover.

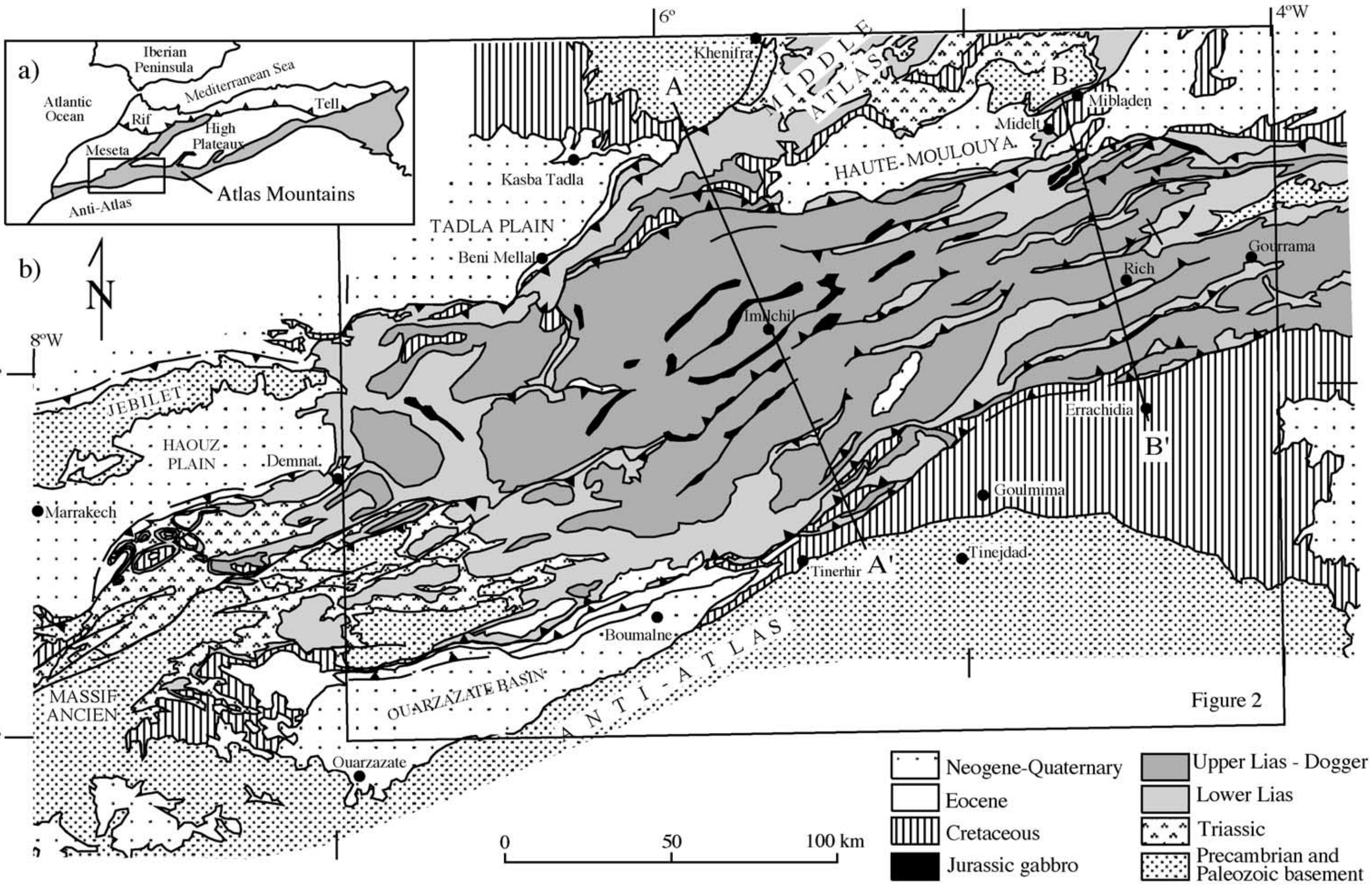

Fig. 1. (a) Location map of the Atlas mountain chains in the North African foreland. (b) Geological map of the central High Atlas (after Teixell et al., 2003), indicating the area corresponding to Fig. 2 and the profiles under analysis in Figs. 2–5.

Another seismic refraction experiment was carried out further to the east in 1986 (Wigger et al., 1992). The resulting model show that the crust is thickest underneath the northern border of the High Atlas near Midelt, reaching 38–39 km, and thins to the south and north to 35 km. A coincident magnetotelluric survey (Schwarz et al., 1992) revealed the existence of a north-dipping high-conductivity zone in the crust starting near the surface in the southern border of the High Atlas (Errachidia). According to these authors, the conductive zone admitted two alternative solutions: either it could flatten down at mid crustal levels or it could plunge continuously northwards to lower crustal depths.

Additional information about crustal thickness in the Atlas region has been recently obtained by receiver function studies. Their results show that the Mohorovicic discontinuity under Midelt (in the plains north of the High Atlas; see Fig. 1 for location) is located at 36 km according to Sandvol et al. (1998) or at 39 km according to Van der Meijde et al. (2003). Both results may be compatible if we consider that Sandvol et al. (1998) found in fact two velocity jumps, at 36 and 39 km. The shallowest one was interpreted by them as the crust–mantle boundary, whereas the deepest one remained uninterpreted. In general, receiver function results are in good agreement with those of wide angle experiments (see above).

These geophysical results, combined with estimations of tectonic shortening, suggest that the high topography of the Atlas cannot be supported by crustal thickening alone, and therefore the mantle may play an important role (Teixell et al., 2003; Arboleya et al., 2004). In fact, low-resolution gravity modelling carried out by Seber et al. (2001) over North Africa and the Middle East suggests that the crust underneath the High Atlas of Morocco is about 5 km thinner than expected if topography was to be compensated at the Moho level. On the other hand, Mickus and Jallouli (1999) suggested that the Saharan and Tunisian Atlas (Fig. 1a) are partially isostatically compensated, with regions of overcompensation in Algeria that may be due to low density material in the upper crust. Finally, seismic tomography seems to indicate a hot, low velocity mantle underneath the High Atlas (Seber et al., 1996). This interpretation is supported by heat flow data which, although scarce in the High Atlas, yields values of 54 mW/m^2, and much higher in the Middle Atlas (85 mW/m^2; Rimi, 1999).

3. Gravity data

The gravity data used for this study have been obtained from publicly available worldwide databases and from local sources, and have been complemented with new measurements acquired in the High Atlas. Bouguer gravity data (Figs. 2 and 3) come from the Moroccan gridded data sets (Hildenbrand et al., 1988). Topography and free air gravity data (Fig. 2) are taken from the TOPEX 2-min gravity data set (ftp://topex.ucsd.edu/pub; Sandwell and Smith, 1997). Additional gravity measurements have been acquired during 2002 and 2003 using a Scintrex CG-3 microgravimeter. Gravity sampling, carried out in the eastern and central parts of the High Atlas of Morocco (Fig. 4), aimed to improve the short wavelength component of the gravity anomaly along two NNW–SSE transects, coincident with geological cross-sections published by Teixell et al. (2003). These are the Imilchil profile (from Tinerhir to Khenifra) in the west, and the Midelt-Errachidia corridor in the east (sections A–A′ and B–B′ in Figs. 1, 2 and 4). An absolute gravity base station located at Ifrane (Middle Atlas), 100 km north of the High Atlas, was used to build a set of second-order base stations in the High Atlas. The sampling spacing of the new data set is around 1 km in the north–south direction and between 1 and 3 km in the east–west direction. This gives gravity information of higher resolution than the pre-existing data sets, where the spacing between samples was around 5 km in the plains and much larger in mountain areas (see points in Fig. 4 for an estimation of the spacing of sampling). The newly acquired data have been corrected for topography using the digital 0.5-min topographic database of Seber et al. (2001), and a reference density of 2600 kg/m^3, which is the average density of the outcropping rocks. Free air and Bouguer plate corrections have also been applied. Density used for the Bouguer plate correction was 2670 kg/m^3. Finally, the resulting values have been merged with the pre-existing Bouguer gravity anomaly data set, showing a good correlation with it (Fig. 4).

We integrate Bouguer and free air gravity anomaly maps with topography (Figs. 2 and 3) to make

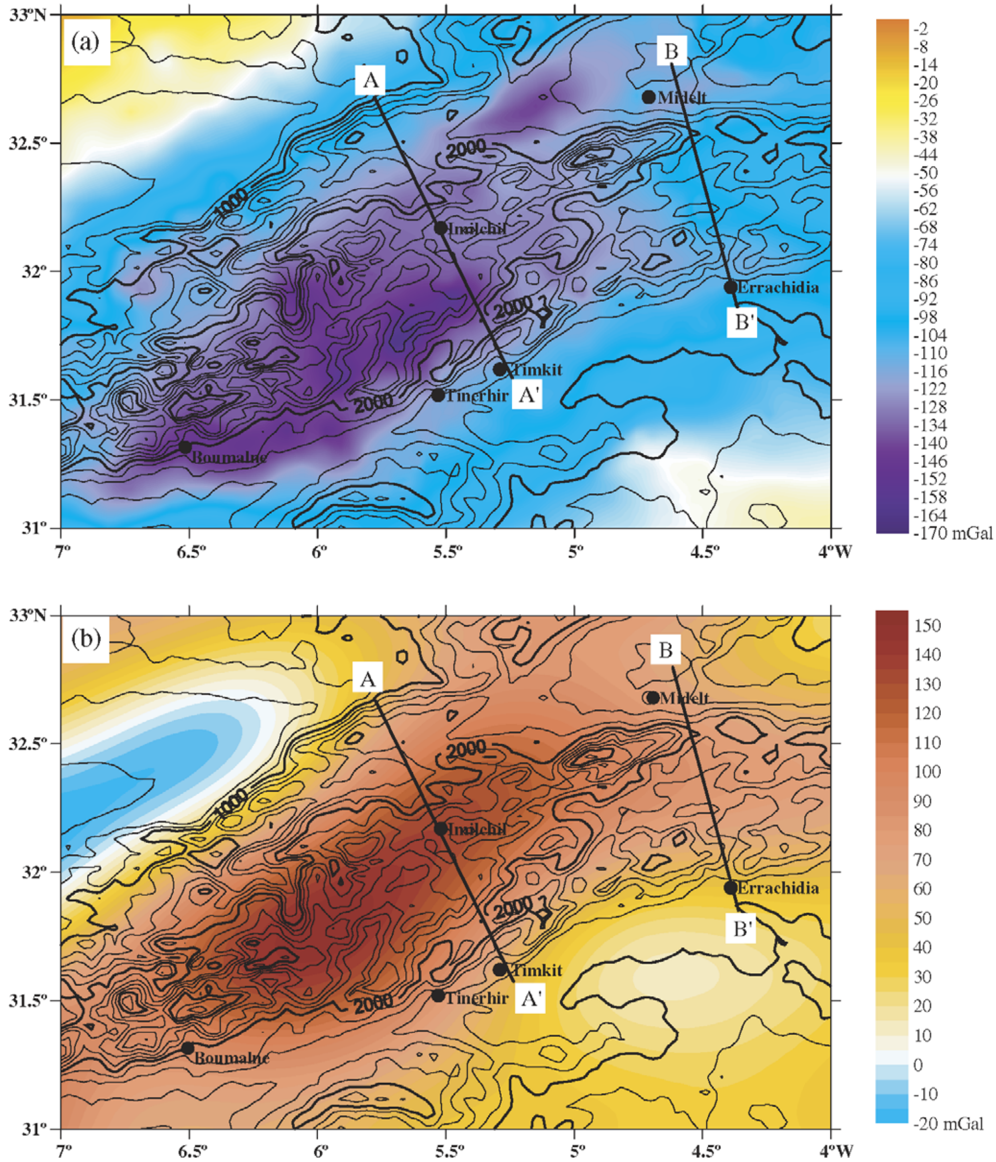

Fig. 2. (a) Combined topography contour and Bouguer gravity anomaly map of the central High Atlas. Contour interval is 200 m. Bouguer anomaly values in the colour scale to the right of the figure. Note that minimum Bouguer anomaly values (dark purple) do not always coincide with the highest elevations. (b) Combined contoured topography and free air anomaly map of the same area. Contour interval is 200 m. Free air gravity anomaly amplitudes in the scale to the right of the figure.

qualitative estimations of the degree of correlation between gravity and elevation, and accordingly, of the isostatic state. It is worth mentioning that the free air gravity anomaly data have a resolution of 2 min, whereas in the Bouguer gravity anomaly data set the resolution ranges from 1.8 to 3 min. Resolution of the TOPEX topographic data set is also 2 min, and therefore the three data sets may be integrated for a qualitative interpretation. The newly acquired gravity data set is used together with the public Bouguer anomaly map and with a denser topographic database (Seber et al., 2001) to undertake 2–2.5D forward

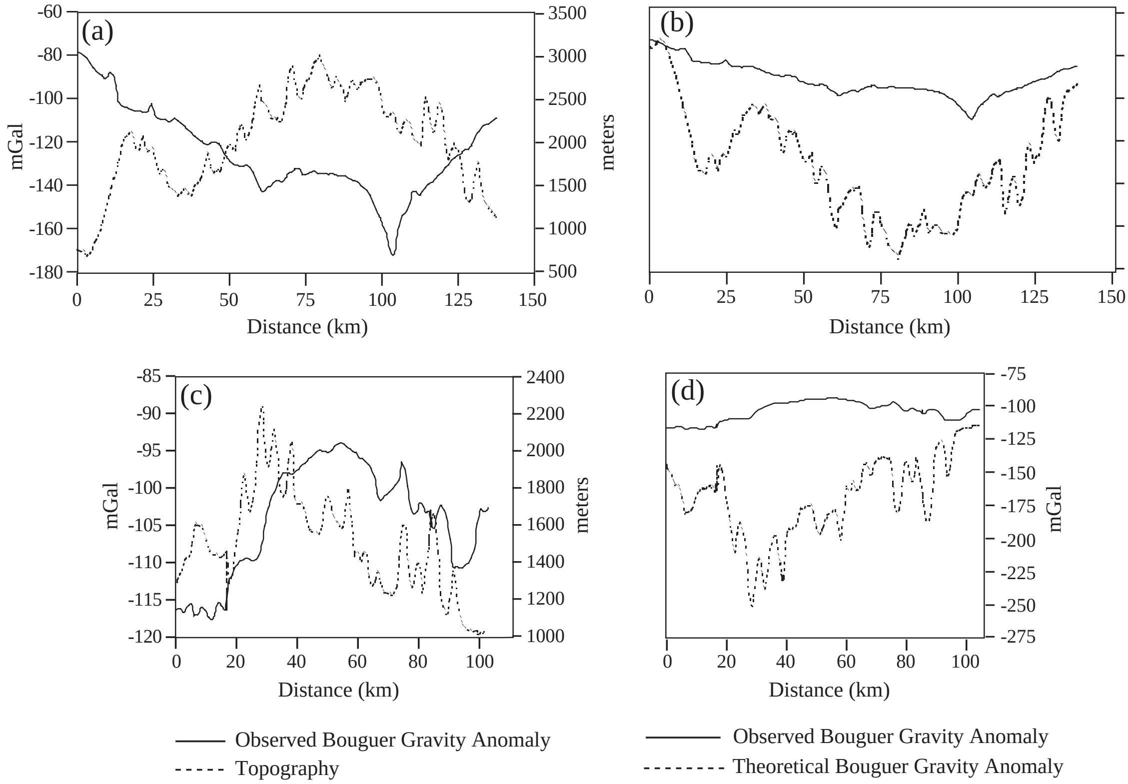

Fig. 3. (a) Elevation and Bouguer gravity anomaly values along profile A–A′ (see location in Fig. 2). In this case, the correlation between both variables is good and long wavelength minimum anomaly values coincide with highest topographies. (b) Difference between the observed and the theoretical Bouguer anomaly calculated for an isostatically compensated topography (Airy model) along section A–A′. Calculated values are more negative than the observed ones suggesting undercompensation at a crustal level. (c) Elevation and Bouguer anomaly values for profile B–B′. Highest topography is in an area of strong anomaly gradient and does not coincide with the minimum Bouguer anomaly values. (d) Difference between the observed and the theoretical Bouguer anomaly calculated for an isostatically compensated topography (Airy model) along section B–B′. Observed and calculated values show large differences mostly in the northern half of the section suggesting undercompensation and an anomalous Moho topography. See discussion in the text.

gravity modelling across the two transects of the central High Atlas (Fig. 4).

3.1. Relation between gravity anomalies and topography

Bouguer gravity anomaly values range from −75 to −160 mgal in the central High Atlas (Fig. 2a). Minimum values are found close to the southern border of the central and western parts of the central High Atlas, and in spite of being located in an area of high topography, they do not coincide with the highest summits, which lie further north. In the Midelt region, the highest topographic elevations do not coincide with the minimum Bouguer anomaly values either (Fig. 2a). Therefore, the expected correlation between topography and Bouguer anomaly is not straightforward in the High Atlas but shows misfits, suggesting a lack of local isostatic equilibrium at a crustal level.

Fig. 3a shows that along section A–A′ the correlation between high topography and Bouguer anomaly minima is reasonable. However, short wavelength minima reaching almost −180 mgal suggest the existence of very low-density material at shallow levels. On the other hand, if we calculate the theoretical Bouguer anomaly values along section A–A′ for an ideal Airy-type compensated state (Fig. 3b), it becomes evident that the thicker crust required by such a state would lead to anomalies more negative than the actual ones.

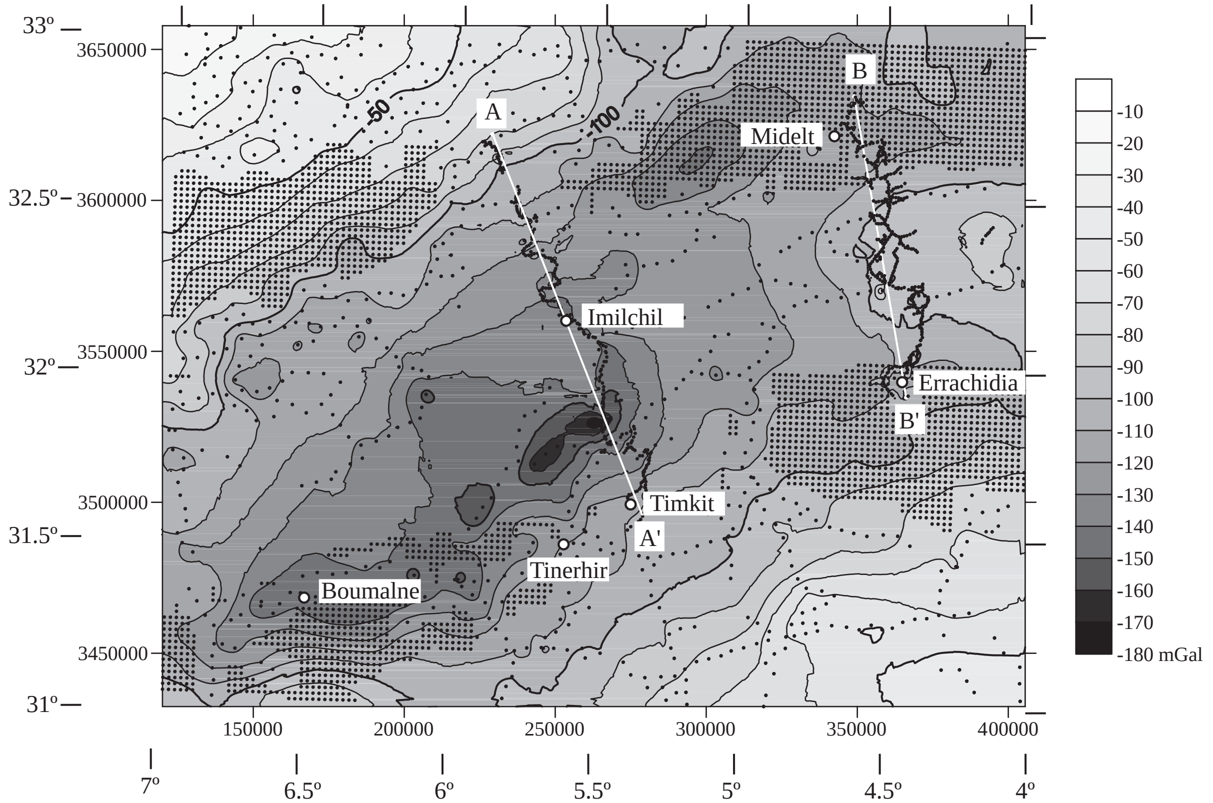

Fig. 4. Bouguer gravity anomaly map of the study area. Data show the integration of the Moroccan gridded data sets (taken from Hildenbrand et al., 1988) with additional data collected by the authors along the profile zones. Dots show the gravity sampling locations for both data sets.

In Fig. 3c, section B–B′ shows minimum Bouguer anomaly values to the north, where topography is lower. In contrast, highest anomaly values appear to the centre of the section, where the elevation is higher. This lack of correlation between gravity and topography is further emphasized by the important differences between the observed and calculated Bouguer anomaly values for this section, presented in Fig. 3d. Both curves differ not only in the value of the anomaly but also in the trend. The observed values reach a maximum of −90 mgal in the central part of the section (km 40–70) whereas the modelled ones show a minimum of −250 mgal very close to that area (km 30–40) where topography is highest (2300 m). These results suggest that the crust in this area (see Figs. 1 and 2) is thin, and that its greater thickness is off the area of highest topographic elevation.

Fig. 2b shows the relation between free air gravity anomalies and topography. In the central High Atlas, the contours of the free air anomaly strike approximately N55E, as does the most elevated area, and hence there is a good correlation between both parameters. However, this trend is slightly oblique to the general structural direction of the High Atlas, which is approximately N75E. Free air maxima of 150 mgal in the western part of the central High Atlas correlates with summits over 3000 m. In the lowlands to the N and S of the High Atlas, the anomaly decreases to −20 mgal and +15 mgal respectively. As a result, the integral of the free air anomaly with respect to horizontal distance does not add to zero, as is the requirement of Airy isostasy, which postulates that any excess or deficiency of load is compensated directly beneath the position of the load.

4. Crustal structure from gravity modelling

A quantitative interpretation of the Bouguer gravity anomaly along profiles A–A′ and B–B′ (Fig. 4) is reached through 2D (2.5D for round-shaped outcropping igneous bodies) forward modelling using GM-SYS 4.6 program. The results of modelling are shown in Figs. 5 and 6.

Rock densities used in our models are slightly different from those used by other authors in previous gravity models of North Africa (e.g., Makris et al., 1985; Mickus and Jallouli, 1999). Table 1 presents a comparison of densities used by previous authors and those used here. Variable densities have been attributed to the Mesozoic and Cenozoic sedimentary rocks according to their nature, as they range from limestones and shales to salts, and include gabbro bodies (Table 2). Great variability exists in the density attributed to the upper crustal basement. While Mickus and Jallouli (1999) used a density of 2720 kg/m^3, Makris et al. (1985) used a much higher density (2820 kg/m^3) which is probably too high

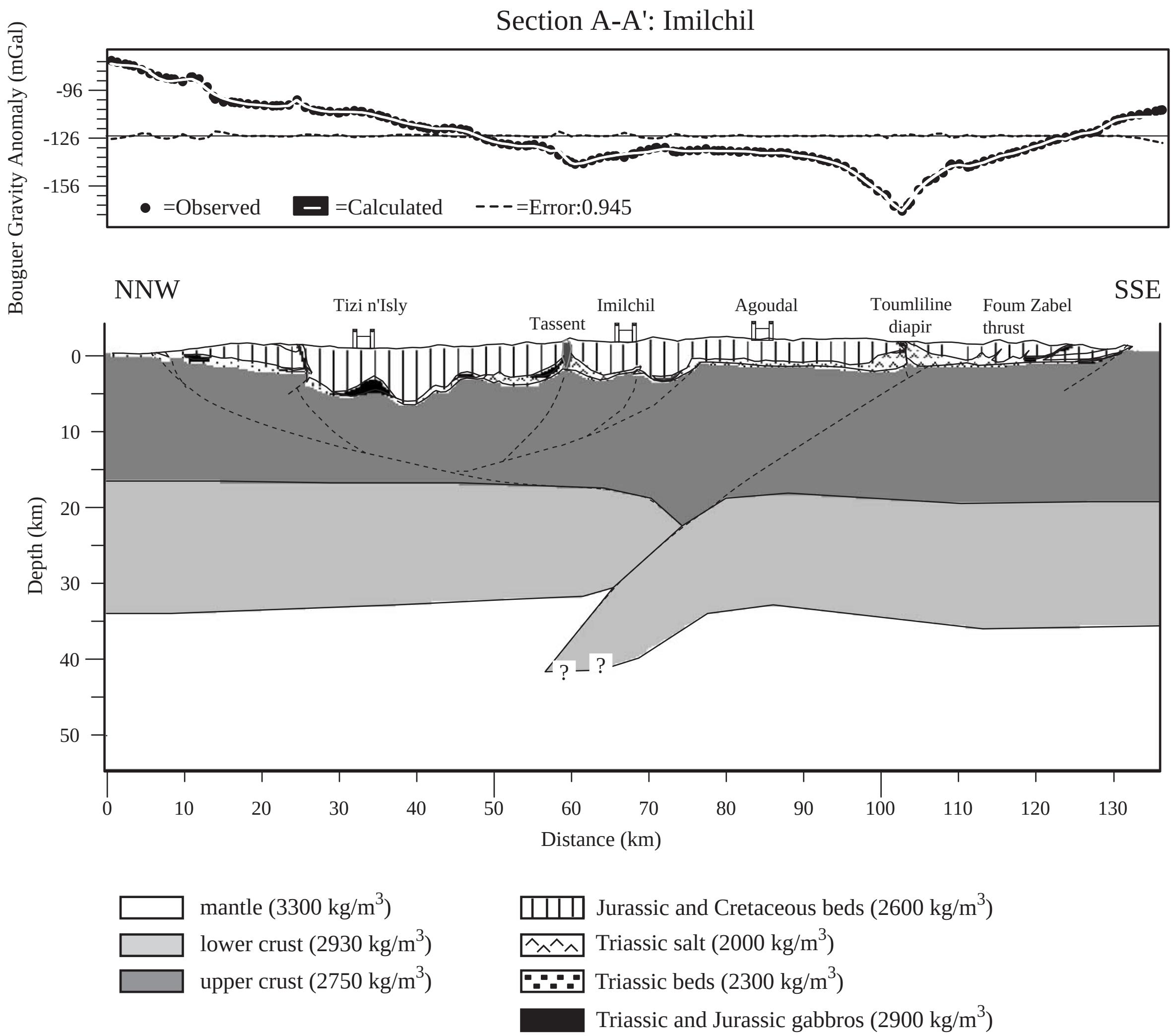

Fig. 5. Gravity model along profile A–A′ . See Table 2 for densities. Long wavelength Bouguer gravity anomaly minima are fitted with a crustal root in the Imilchil area, that has been modelled down to 41 km. Short wavelength minima are modelled as the response of a diapir of Triassic salt (e.g., Toumliline diapir). Dashed lines indicate thrust faults according to the integration with field data by Teixell et al. (2003), and are not a result of the modelling itself. Question marks indicate that, according to the estimated shortening, the crust might reach slightly higher depths than modelled (see text for explanation).

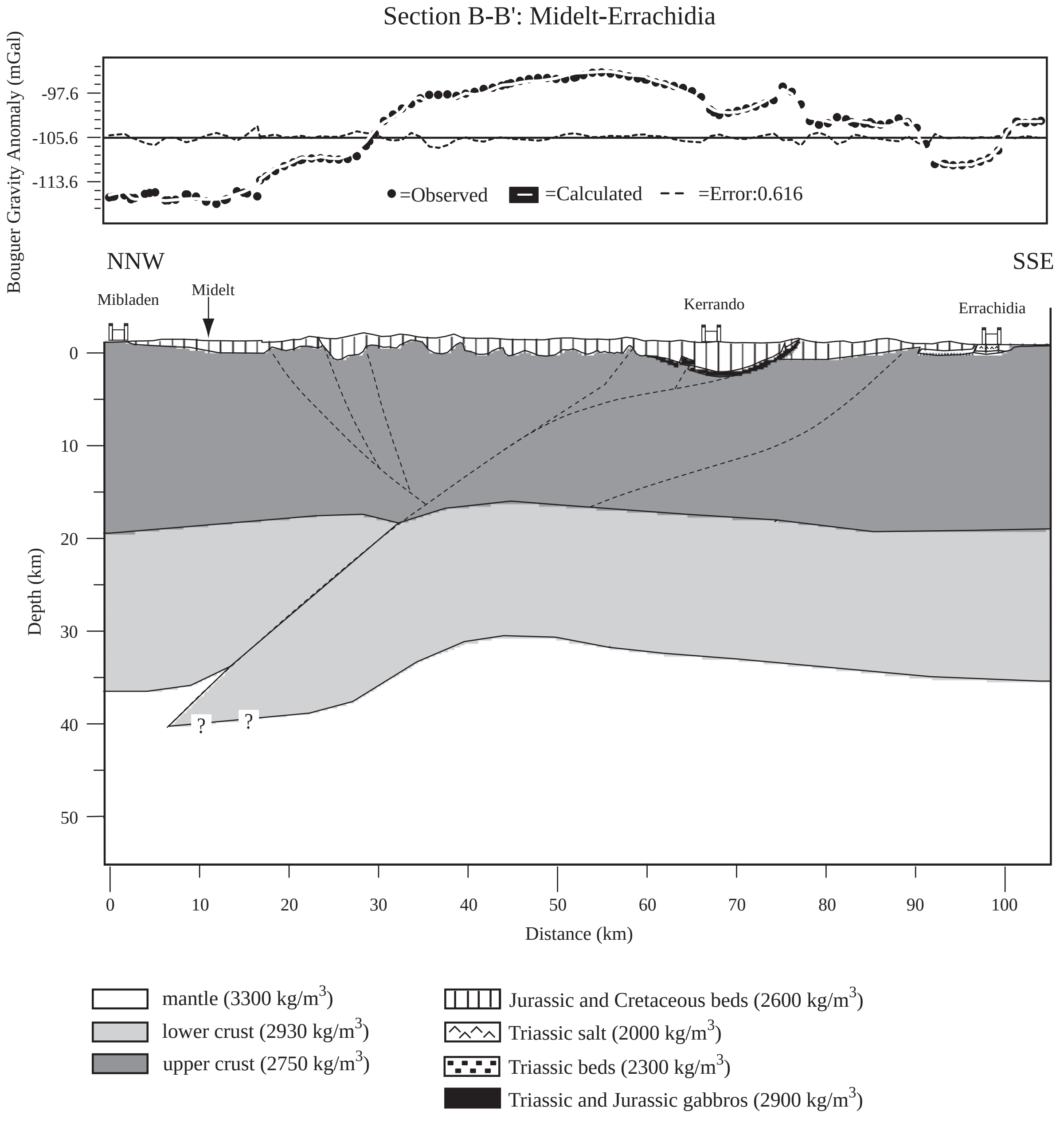

Fig. 6. Gravity model of section B–B′ . See Table 2 for densities. Long wavelength Bouguer gravity anomaly minima are fitted with a crustal root in the Midelt area that reaches 39 km. Short wavelength anomaly values are modelled as the gravimetric response of Mesozoic gabbros and low density Triassic sedimentary rocks. Dashed lines and question marks as in Fig. 5.

considering that this value would correspond to an average upper crust P-wave velocity of 6.9 km/s. Accordingly, we have used for the upper crust a density of 2750 kg/m^3, closer to that of Mickus and Jallouli (1999). The density values used for the lower crust range from 2900 kg/m^3 to 2960 kg/m^3 depending on the authors (see Table 1). Although the influence in the model is less important, we have

Table 1
Densities previously used in gravity models of the Atlas Mountains System compared with the ones used now for the models of Figs. 5 and 6

	Tunisian and Saharian Atlas (Mickus and Jallouli, 1999) (kg/m^3)	High Atlas (Makris et al., 1985) (kg/m^3)	Present model (kg/m^3)
Sedimentary rocks	2380	2400–2600	2000–2300–2600
Upper crust	2720	2820	2750
Lower crust	2960	2900	2930
Mantle	3300	3350	3300

used an intermediate value of 2930 kg/m^3. We have taken the usual density of 3300 kg/m^3 for the mantle. In summary, the densities we use correspond to an average of those used by previous authors, and show a correlation with P-wave velocities (Nafe and Drake, 1957) that matches the refraction data obtained by Wigger et al. (1992).

The upper part of the crustal models is constrained by the geological cross-sections of Teixell et al. (2003) and Arboleya et al. (2004), which are based on detailed field work. This allows minimum uncertainty on the geometry of the sedimentary cover and the depth to the upper crustal basement. The lower part of the sections has been modelled using available seismic refraction and receiver function results as constraints. These results helped to estimate the depth to the upper–lower crust boundary and to the Moho in precise locations, avoiding as much as possible the non-uniqueness of gravity models. The models presented here do not constitute a unique solution to the gravity pattern of the High Atlas mountains of Morroco, but they fit the available geological and geophysical constraints, being therefore admissible.

The Bouguer gravity anomaly modelled in profile A–A′ (Fig. 5) has a short wavelength component that correlates well with the Mesozoic structure deduced from geological surveying. Short wavelength gravity maxima coincide with basement highs where Paleozoic rocks stand close to the surface, or with Mesozoic dolerite and gabbro bodies (e.g., km 10 and 26). On the contrary, short wavelength gravity minima represent places where the Mesozoic cover is thickest or contains salt diapirs (e.g., km 61 and 105). Greatest sedimentary thickness is found north of Imilchil, (Tizi n'Isly area, Fig. 5), where limestones and shales occupy the interval down to 6 km. The Bouguer gravity minima found in this section vary between −170 and −142 mgal, and have wavelengths of less than 15 km (km 61 and 105 in Fig. 5). Based on geological evidence, both minima have been interpreted as the gravity signature of salt accumulations, the northern one being in the subsurface hidden below the gabbro of the Tassent massif and the southern one cropping out at Toumliline (Teixell et al., 2003). However, these low values are part of a longer wavelength minimum with an average value of −135 mgal that most probably has a deeper source.

Depth to the upper–lower crust boundary has been preliminary taken from a refraction-derived model by Tadili et al. (1986), who located this boundary at depths between 17 and 19 km, describing a very low wavelength high northwest of Imilchil. In our model A–A′ , the top of the lower crust has been also placed at 17–19 km, reaching 16 km in the northern part of the profile.

Once the depths to the pre-Mesozoic basement and to the lower crust are assumed, the depth to the Moho can be easily determined. Maximum depth to the crust–mantle boundary in section A–A′ is found in the Imilchil area, where crustal thickening brings the discontinuity to depths slightly over 41 km. From there, it rises to 33.5 km to the north and to 35 km to the south, following the increase in the Bouguer anomaly pattern. The resulting structure is in agreement with the depth to the Moho estimated on the basis of refraction and gravity data by Tadili et al. (1986): 32–33 km under Khenifra to the north and 35–36 km under Tinerhir to the south. The asymmetric crustal root that we model under Imilchil is the source of the long wavelength Bouguer anomaly minimum found between km 60 and 120. The concave-downwards shape of the Bouguer anomaly

Table 2
Densities used in the gravity models of Figs. 5 and 6 for the Mesozoic sedimentary and igneous rocks

	Sediments (kg/m^3)
Triassic sedimentary rocks	2300
Triassic salt	2000
Triassic and Jurassic gabbros	2900
Jurassic sedimentary rocks	2600
Cretaceous sedimentary rocks	2600

profile from km 70 to 110 suggests that the lower crust to the south of the root may be arched (Fig. 5), giving the low amplitude relative maximum measured between km 60 and 100. However, this could also be an effect of the bounding minimum values originated by the salt diapirs mentioned above. If the base of the crust was modelled as a flat boundary at −36 km, to fit the observed Bouguer gravity anomaly we would have to introduce important changes in the upper–lower crust boundary that are not supported by models deduced from refraction data.

The Bouguer gravity anomaly in section B–B′ (Figs. 2, 4 and 6) shows minimum values of −117 mgal located in the northernmost part of the profile, i.e., in the Moulouya plain north of the High Atlas. From there, entering into the High Atlas, the anomaly increases to −93 mgal in the central part of the profile, which surprisingly coincides with the area of highest topographic elevation. As for section A–A′, short wavelength anomalies are fitted with variations in sedimentary thickness and depth to the basement. Maximum sedimentary thickness close to 4 km is found in the Kerrando syncline (km 68 in Fig. 6), which corresponds to a succession of Jurassic shales with minor limestones carried by a thrust that reactivates an old basin-margin extensional fault (Teixell et al., 2003). Apart from this case, the basement along this section is located at a depth close to sea level, implying that the thickness of sediments is small. Very short wavelength maxima and minima located in the northern part of the section may be preliminarily interpreted as an effect of different degrees of mineralization of the Paleozoic basement, known to be highly productive in the Mibladen mining district. Other possible sources such as basic igneous intrusions cannot be ruled out (Tertiary and Quaternary basalts are known in the Moulouya plain). In any case, modelling this irregular and short wavelength pattern appears arbitrary, and hence, the associated density anomalies are left unmodelled.

The top of the lower crust is modelled at depths between 20 km in the northern part of profile B–B′ and at 18.5 km in the south, rising to 16 km near the central part of the section. These values agree in general with those of Tadili et al. (1986), although the data set of these authors has low resolution, probably due to wide station spacing, which leave information gaps (e.g., between Midelt and Rich, km 10 and 60 in Fig. 6 where our modelling suggests that the upper–lower crust boundary shallows by 2 km).

In our model, the crust–mantle boundary defines an asymmetric crustal root in the Midelt area (Fig. 6), reaching a maximum depth of around 39 km, coinciding with the Bouguer anomaly minimum. This is consistent with the interpretation of receiver functions by Van der Meijde et al. (2003). From there, the Moho rises to the north and south to 36 km in the northern edge of the profile (Mibladen area) and to 35 km in the southern end (Errachidia area), following the increase in the gravity anomaly. A local high in the Moho discontinuity appears around km 45. It is this high that fits the maximum values of the Bouguer anomaly in the central part of the section, and indicates that the lower crust might also be arched to the south of the root, as deduced for section A–A′. Alternative models with a flat Moho produce similar results to those for profile A–A′.

5. Discussion

5.1. Deep structure of the High Atlas

Although the surface structure of the High Atlas has been the subject of many studies (e.g., Laville, 1985; Schaer, 1987; Laville and Piqué, 1992; Beauchamp et al., 1999; Teixell et al., 2003, and references therein), large uncertainty remains about the structure of the deeper part of the crust. The models presented in this work give an image of the crust along two sections of the High Atlas whose shallow geological structure has been thoroughly described recently (Teixell et al., 2003).

Previous crustal models proposed the existence of a flat, intra-crustal detachment below the High Atlas at some −20 km, without crustal roots underneath (e.g., Jacobshagen et al., 1988; Giese and Jacobshagen, 1992). Beauchamp et al. (1999) also suggested that Atlas thrusts penetrated only to the middle crust, leaving the lower crust unfaulted. However, both Makris et al. (1985) and Wigger et al. (1992) modelled an asymmetric lower crustal structure beneath the High Atlas, with gradual thickness increase beneath the southern boundary of the chain, and a sharper change beneath the northern boundary. These authors did not provide an explanation for this finding, but actually Giese and Jacobshagen (1992) interpreted it as due to a small fault offsetting the Moho and dying out rapidly, without connection with the upper crustal thrust system. In our view, these features, considered together with the upper crustal structure and the gravity modelling results, suggest that thrust faults may cut the entire crust in a ramp fashion, defining a localized, small crustal root accounting for the pattern of the Bouguer gravity anomaly. In Figs. 5 and 6 we model the crustal root to derive from the displacement along a major low-angle, north-dipping thrust fault that effectively offsets the Moho discontinuity. This thrust fault, whose existence was already suggested by Arboleya et al. (2004) on the basis of structural data, is also compatible with one of the solutions for the north-dipping high conductivity layer modelled by Schwarz et al. (1992, their Fig. 6a). In this solution, the conductive layer, that can be interpreted as a zone of intense deformation with high fluid pressure, reaches Moho depths beneath the Moulouya plain near Midelt.

The crustal imbrication model presented here is in agreement with the structure of other inverted continental rifts that have been imaged by seismic reflection profiling, e.g., the Central Pyrenees (Choukroune et al., 1989; Daignières et al., 1994) or the Donbas fold belt of Ukraine (Maystrenko et al., 2003). These profiles showed the existence of crustal-scale thrust faults that offset the Moho and are associated to the underthrusting (subduction) of the lower crust (Muñoz, 1992; Teixell, 1998; Maystrenko et al., 2003). In the case of the High Atlas mountains, our model might also explain the two receiver-function velocity jumps at 36 and 39 km (Sandvol et al., 1998; Van der Meijde et al., 2003), which can be now interpreted as a Moho duplication.

We model underthrusting (subduction) of the southern (saharan) lower crust under the northern moroccan crust (Figs. 5 and 6). However, the modelled length of the imbricated lower crust implies slightly lower shortening at this level than that deduced for the upper crust (30 km for a cross-section following A–A′ profile and 26 km for B–B′ in Teixell et al., 2003). This might imply that the deeper parts of the root have been transformed into mineral phases that do not have a gravimetric and probably seismic signature clearly different from that of the mantle (i.e., possessing similar densities and velocities).

The position of the modelled underthrust root shifts to the north as we go from the western to the eastern section (i.e., the imbrication of the deep crust does not follow the High Atlas general direction but is oriented obliquely, that is NE–SW). This oblique trend is actually parallel to many internal faults within the High Atlas, and to the general direction of the Middle Atlas chain. This is also evident when observing the direction of the gravity anomalies and some of the topographic contour lines within the High Atlas, which also trend NE, obliquely to the general direction of the chain (Fig. 2). The oblique trend of the gravity minima cannot be easily explained by variations in sedimentary thickness. Minimum Bouguer gravity anomaly values in the southern border of the western part of the central High Atlas (Fig. 2) could be tentatively correlated with the Cenozoic Ourzazate basin (near Boumalne, Fig. 2). However, the free air gravity anomaly map does not show a minimum in that area, suggesting that the anomaly must have a deep source. The same applies for the Bouguer anomaly minimum found close to the Moulouya plain, near Midelt (Fig. 2), that appear in an area where the basement either crops out or is close to the surface.

The NE–SW strike characteristic of many of the internal thrust faults of the High Atlas (see Fig. 1) and of the Middle Atlas is inherited from normal faults created during the Mesozoic rifting phase perpendicularly to the extension direction at that time (Mattauer et al., 1977). However, the general orientation of the High Atlas trough was rather ENE, suggesting that its geometry conformed to an oblique rift (El Kochri and Chorowicz, 1995; Arboleya et al., 2004). Moreover, our results indicate that during Cenozoic inversion, the structure of the lower crust was also determined by the NE trend of the internal faults, rather than by the general orientation of the rift trough or by the regional compression direction.

5.2. How is topography supported?

The High Atlas does not show the expected correlation between topography and gravity (Fig. 3). Furthermore, where high topographic elevation coincides with minimum Bouguer gravity anomalies, the latter still seem insufficient to account for the high altitudes observed (Fig. 3b). Fitting of gravity anomaly profiles in the central High Atlas was achieved by modelling the response of varying crustal thickness. The non-uniqueness of the method left several possibilities open, but in the case of the High Atlas these have been sorted out taking into account previous wide angle-refraction seismic data (e.g., the Bouguer maxima could have alternatively been modelled with high density material at shallow or middle crustal levels, which would have resulted in a thicker crust than in our sections. However, this would not be supported by seismic data).

Fig. 7 shows a comparison between the modelled crustal thickness and that expected for a compensated crust following Airy local isostasy. Densities used for this calculation are 2670 kg/m^3 for the crust and 3300 kg/m^3 for the mantle. In section A–A′ (Fig. 7a), the overall shape of the base of the crust is markedly different in both cases: differences between modelled and expected crustal thickness are in excess of 5 km, and the modelled root does not coincide with the location of the expected maximum crustal thickness. Similar conclusions are obtained for section B–B′ (Fig. 7b). Neither in this profile, the location of the modelled crustal root, based on Bouguer gravity anomaly minima and receiver functions, coincides with its expected position, beneath the highest elevations if Airy isostasy would apply. Using higher mean crustal densities (e.g., 2800 kg/m^3) would make differences even more important.

In addition, we have modelled the gravity response of a crust as thick as that expected from the Airy rules (Figs. 8 and 9). Fig. 8a, for section A–A′ , shows that an Airy-compensated Moho topography would result in a Bouguer gravity anomaly that would not match the observed one, in terms of anomaly values and trend. Important variations in the upper–lower crust boundary should then be introduced to fit the observed anomaly, leading to an irregular lower crust as thick as 24 km in places (Fig. 8b). The same results are obtained for section B–B′ (Fig. 9). A crustal thickness

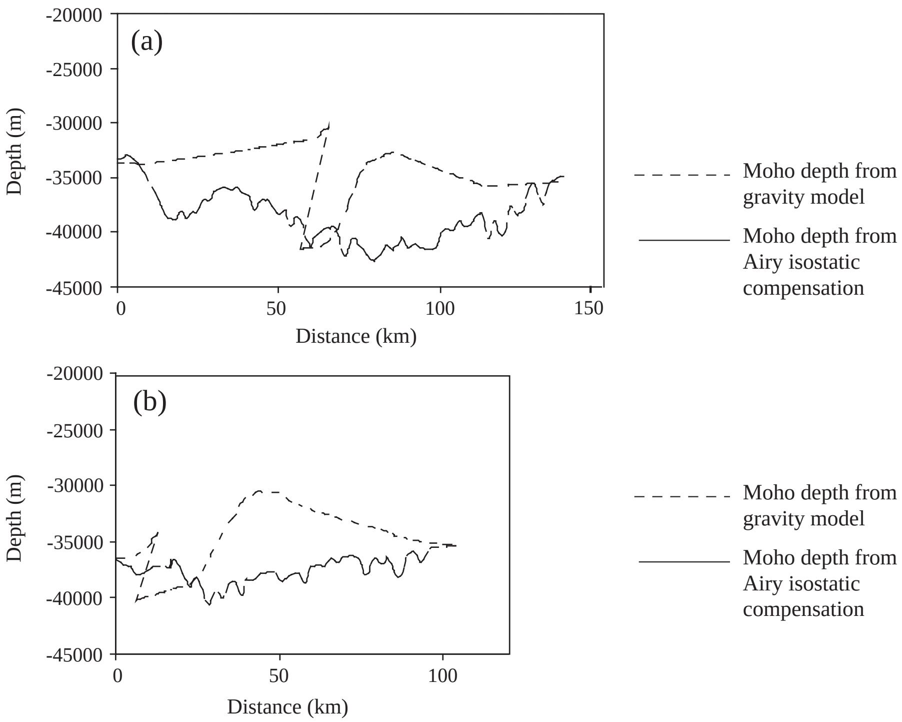

Fig. 7. Comparison between the Moho depth as deduced from our gravity models and that deduced for a crust isostatically compensated following the Airy model. (a) Case for section A–A′ : the crust as deduced for the Airy assumptions is thicker than our gravity-modelled crust. (b) Case for section B–B′ : the crust as deduced for the Airy assumptions is thicker except in the northern part of the section, where the Bouguer gravity anomaly minimum needs the modelled crust to be thicker.

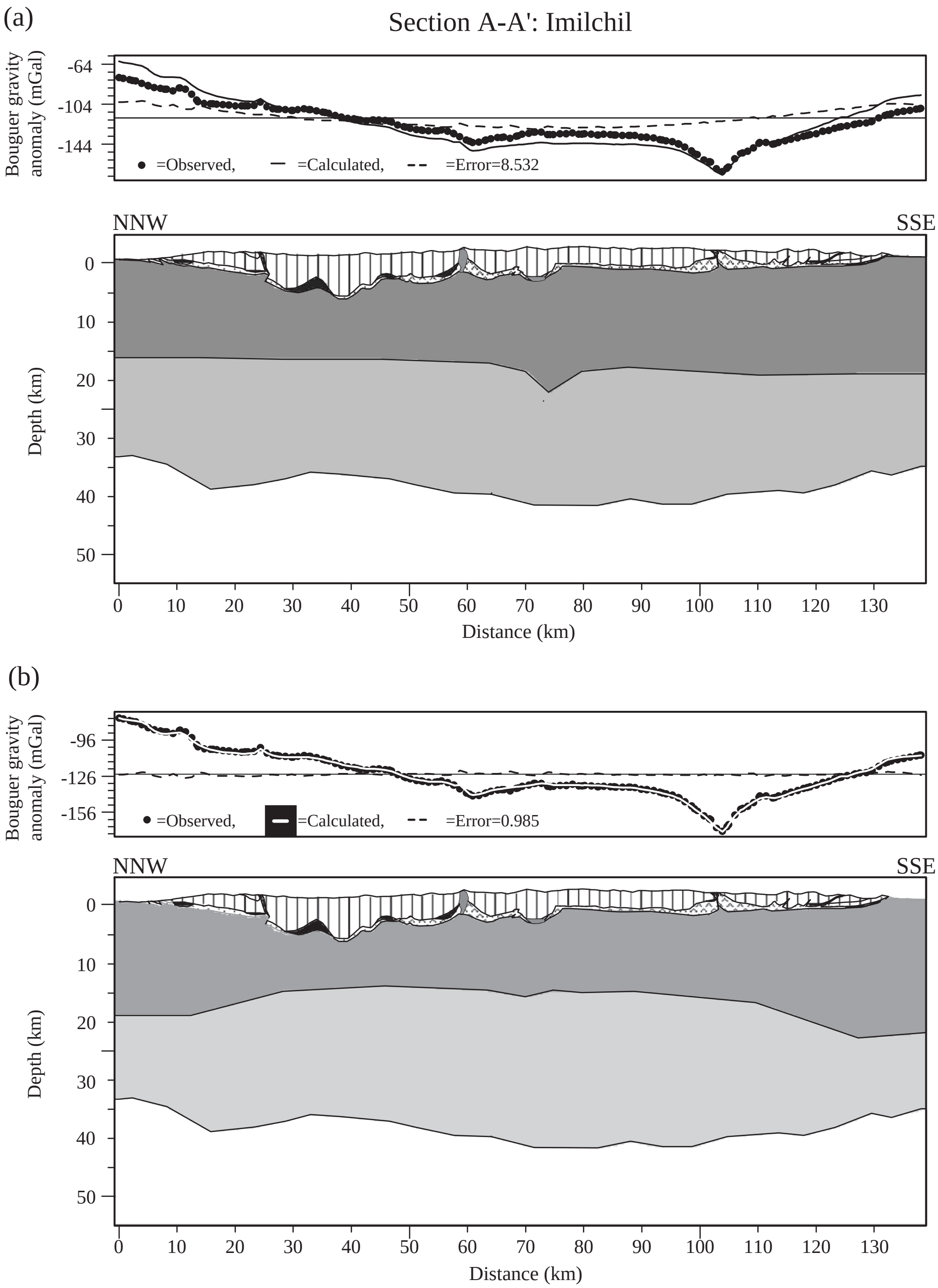

Fig. 8. Alternative gravity model solution for section A–A′ . (a) Bouguer gravity anomaly of a section with crustal thickness equal to that calculated in Fig. 7a (isostatically compensated crust-Airy model). The observed and modelled anomalies do not coincide. (b) Important variations in the upper–lower crust boundary are needed to match both anomalies. Legend as in Fig. 5.

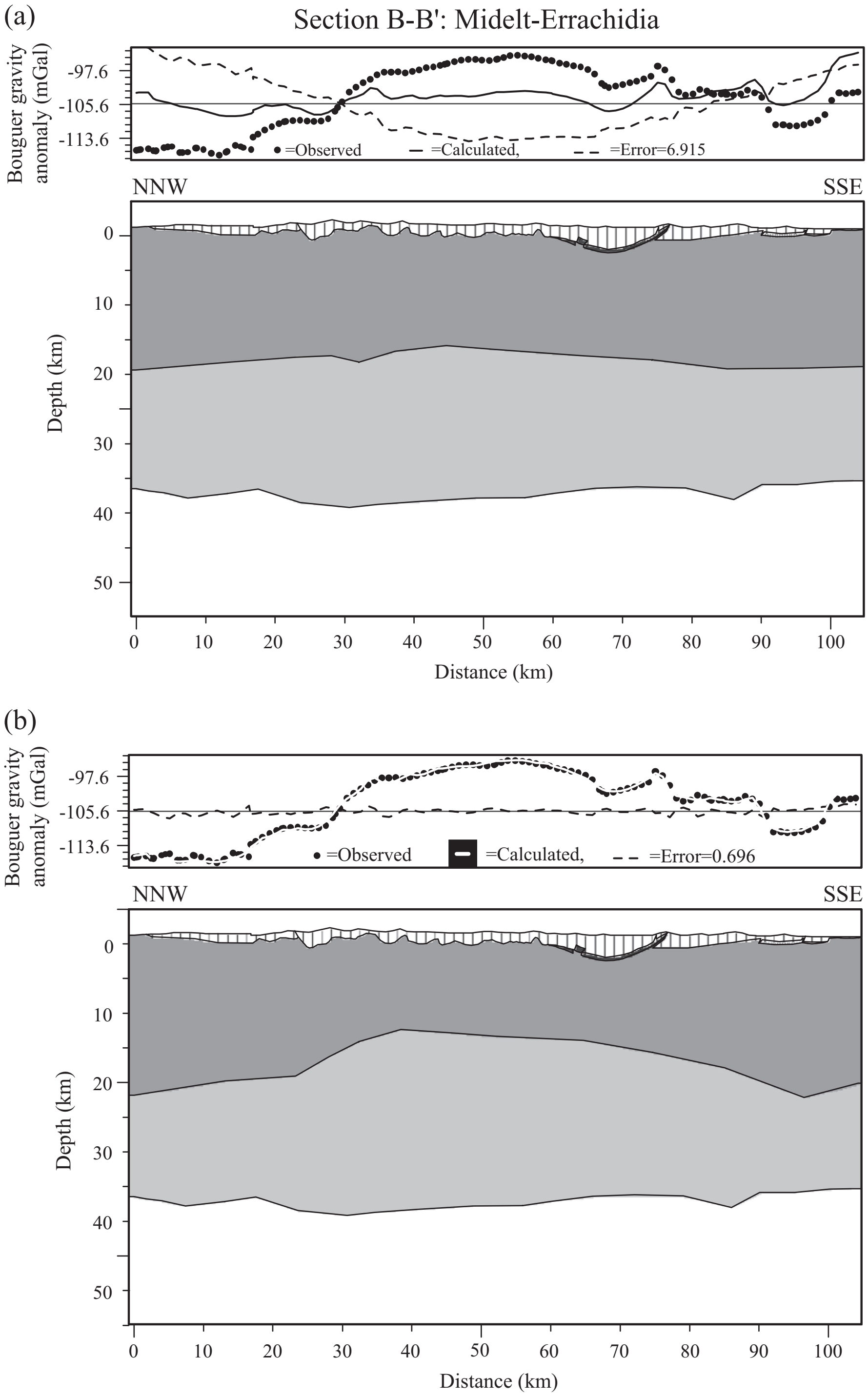

Fig. 9. Alternative gravity model solution for section B–B′ . (a) Bouguer gravity anomaly of a section with crustal thickness equal to that calculated in Fig. 7b (isostatically compensated crust-Airy model). The observed and modelled anomalies do not coincide. (b) Important variations in the upper–lower crust boundary are needed to match both anomalies. Legend as in Fig. 6.

like that deduced in Fig. 7b from the Airy calculations would result in a Bouguer gravity anomaly profile very different to the observed one (Fig. 9a), and only important changes in the thickness of the lower crust would help to fit the curves (Fig. 9b).

We thus conclude that the crustal thickness along the central High Atlas profiles cannot entirely explain the observed topography and therefore, that these mountains are not isostatically compensated. Crustal thicknesses of 40 and 36 km were deduced respectively for the Saharan and Tunisian Atlas by Mickus and Jallouli (1999), who consider these values as indicative of some degree of isostatic compensation. Comparatively, the central High Atlas of Morocco has higher topography but similar crustal thickness (see also models by Makris et al., 1985). It follows that isostatic compensation has not been reached to the same degree, and that there must be some additional mechanism to support the High Atlas elevation that is unrelated to shortening and crustal thickening (Teixell et al., 2003; Arboleya et al., 2004).

An explanation for this fact lies on topography being partly supported by a thin and hot lithosphere. Strong lines of evidence are the existence of abundant alkaline volcanism in the Atlasic domain and the low P-wave velocities detected by Seber et al. (1996) under the High Atlas. They suggest that the asthenosphere may occupy an anomalously high position helping to support the topography (Seber et al., 1996; Teixell et al., 2003; Zeyen et al., 2005). On the other side, the Bouguer gravity anomaly profiles shown in Figs. 5 and 6 exhibit a concave pattern in the areas where elevation is highest. This concavity required a slightly arched lower crust in the models, that might be the result of buckling during the Cenozoic compression or, alternatively, an inherited feature from the Mesozoic crustal thinning.

6. Conclusions

The crustal structure of the High Atlas, as deduced from integration of gravity modelling and other existing geophysical and geological data, is characterized by a doubly verging thrust system at upper crustal levels, which passes into the lower crust to a major, north dipping thrust fault that we interpret to offset the Moho discontinuity. An underthrust slab of saharan, dense lower crust defines a local crustal root that can be detected to a depth between 38 and 41 km in the central High Atlas. These results reinforce models of crustal undercompensation and Moho involvement in the deformation, defended by some authors, and have implications for the deep structure of intracontinental mountain belts. The High Atlas, together with other small orogens for which there is a wealth of geophysical data, show that intraplate orogens seem capable to develop underthrust (“subducted”) crustal roots in spite of not being considered as destructive plate margins, where roots are pulled by sinking oceanic slabs.

The crustal root deduced in this work does not follow the general direction of the High Atlas, but is oriented NE, parallel to internal, oblique faults reflecting a pre-orogenic, extensional basin configuration. A straightforward correlation between topography and Bouguer anomaly is not met in the High Atlas. Elevations of over 2500 m are supported by a relatively thin crust that seems insufficient for the observed mountain load. The existence of a thin crust in the High Atlas, together with the limited tectonic shortening reported (< 25%), corroborates that the crust does not entirely support the Atlas topography. Arching of the lower crust south of the crustal root suggests that active buckling may help to maintain the topography, but recent alkaline volcanism in the vicinity and low P-wave velocities at upper mantle levels suggests that a contribution from a thin, hot and less dense lithosphere is likely.

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References

Arboleya, M.L., Teixell, A., Charroud, M., Julivert, M., 2004. A structural transect through the High and Middle Atlas of Morocco: tectonic implications. J. Afr. Earth Sci. 39, 319–327.

Beauchamp, W., Allmendinger, R.W., Barazangui, M., Demnati, A., El Alji, M., Dahmani, M., 1999. Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect. Tectonics 18, 163–184.

Choubert, G., Faure-Muret, H., 1962. Evolution du domaine atlasique marocain depuis les temps paléozoïques. In: Soc. Geol. de France (Ed.), Livre a la Mémoire du Professeur Paul Fallot, Mem. Hors Sér. vol. 1, pp. 447–527.

ECORS Team, 1989. The ECORS deep seismic profile reflection data and the overall structure of an orogenic belt. Tectonics 8, 23–39.

Daignières, M., Séguret, M., ECORS Team, 1994. The Arzacq-Western Pyrenees ECORS Deep Seismic Profile. In: Mascle, A. (Ed.), Hydrocarbon and Petroleum Geology of France. Eur. Ass. Petr. Geosci. Spec. Publ. vol. 4, pp. 199–208.

Dutour, A., Ferrandini, J., 1985. Nouvelles observations néo-tectoniques dans le Haut Atlas de Marrakech et le Haouz central (Maroc): apports sur l'evolution récente d'un segment du bâti atlasique. Rev. Geol. Dyn. Geogr. Phys. 26, 285–297.

El Harfi, A., Lang, J., Salomon, J., 1996. Le remplissage continental cénozoïque du bassin d'avant-pays de Ourzazate: implications sur l'evolution géodynamique du Haut Atlas central (Maroc). C. R. Acad. Sci., Ser. II 323, 623–630.

El Kochri, A., Chorowicz, J., 1995. Oblique extension in the Jurassic trough of the central and eastern High Atlas (Morocco). Can. J. Earth Sci. 33, 84–92.

Fraissinet, C.E., Zouine, M., Morel, J.-L., Poisson, A., Andrieux, J., Faure-Muret, A., 1988. Structural evolution of the southern and northern central High Atlas. In: Jacobshagen, V. (Ed.), The Atlas System of Morocco. Springer-Verlag, Berlin, pp. 273–291.

Frizon de Lamotte, D., Saint Bezar, B., Bracene, R., 2000. The two main steps of the Atlas building and geodynamics of the western Mediterranean. Tectonics 19, 740–761.

Giese, P., Jacobshagen, V., 1992. Inversion tectonics of intra-continental ranges: high and Middle Atlas, Morocco. Geol. Rundsch. 81, 249–259.

Gomez, F., Beauchamp, W., Barazangi, M., 2000. Role of Atlas Mountains (northwest Africa) within the African-Eurasian plate-boundary zone. Geology 28, 775–778.

Görler, K., Helmdach, F.-F., Gaemers, P., Heissig, K., Hinsch, W., Mädler, K., Schwarzhans, W., Zucht, M., 1988. The uplift of the Central Atlas as deduced from Neogene continental sediments of the Ourzazate province, Morocco. In: Jacobshagen, V. (Ed.), The Atlas System of Morocco. Springer-Verlag, Berlin, pp. 361–404.

Hailwood, E.A., Mitchell, J.G., 1971. Paleomagnetic and radiometric dating results from Jurassic intrusions in South Morocco. Geophys. J. R. Astron. Soc. 24, 351–364.

Harmand, C., Catagrel, J.M., 1984. Le volcanisme alcalin Tertiaire et Quaternaire du Moyen Atlas (Maroc): chronologie K/Ar et cadre geodynamique. J. Afr. Earth Sci. 2, 51–55.

Harmand, C., Moukadiri, A., 1986. Synchronisme entre tectonique compressive et volcanisme alcalin: exemple de la province quaternaire du Moyen Atlas (Maroc). Bull. Soc. Géol. Fr. 8 (4), 595–603.

Hildenbrand, T., Kucks, R., Hamouda, M., Bellot, A., 1988. Bouguer Gravity Map and Related Filtered Anomaly Maps of Morocco. U.S. Geological Survey, Denver, Colorado.

Jacobshagen, V., Gorler, K., Giese, P., 1988. Geodynamic evolution of the Atlas System (Morocco) in post-Palaeozoic times. In: Jacobshagen, V. (Ed.), The Atlas System of Morocco. Springer-Verlag, Berlin, pp. 481–499.

Laville, E., 1985. Evolution sédimentaire, tectonique et magmatique du bassin jurassique du Haut Atlas, Maroc: modèle en relais multiples de décrochements. Thèse d'Etat, Montpellier.

Laville, E., Piqué, A., 1992. Jurassic penetrative deformation and Cenozoic uplift in the central High Atlas (Morocco): a tectonic model, structural and orogenic inversions. Geol. Rundsch. 81, 157–170.

Laville, E., Lesage, J.-L., Séguret, M., 1977. Géometrie, cinématique (dynamique) de la tectonique atlasique sur le versant sud du Haut Atlas marocain. Aperçu sur les tectoniques hercyniennes et tardi-hercyniennes. Bull. Soc. Géol. Fr. 7 (19), 527–539.

Makris, J., Demnati, A., Klusmann, J., 1985. Deep seismic soundings in Morocco and a crust and upper mantle model deduced from seismic and gravity data. Ann. Geophys. 3, 369–380.

Mattauer, M., Taponnier, P., Proust, F., 1977. Sur les mecanismes de formation des chaines intracontinentales. L'exemple des chaines atlasiques du Maroc. Bull. Soc. Géol. Fr. 7 (19), 521–526.

Maystrenko, Y., Stovba, S., Stephenson, R., Bayer, U., Menyoli, E., Gajewski, D., Huebscher, C., Rabbel, W., Saintot, A., Starostenko, V., Thybo, H., Tolkunov, A., 2003. Crustal-scale pop-up structure in cratonic lithosphere: DOBRE deep seismic reflection study of the Donbas fold belt, Ukraine. Geology 31, 733–736.

Medina, F., Cherkaoui, T., 1991. Focal mechanisms of the Atlas earthquakes and tectonic implications. Geol. Rundsch. 3, 639–648.

Mickus, K., Jallouli, C., 1999. Crustal structure beneath the Tell and Atlas Mountains (Algeria and Tunisia) through the analysis of gravity data. Tectonophysics 314, 373–385.

Morel, J.-L., Zouine, M., Poisson, A., 1993. Relations entre la subsidence des bassins moulouyens et la création des reliefs atlasiques (Maroc): Un exemple d'inversion tectonique depuis le Néogene. Bull. Soc. Geol. France, 93(1), 164, 79–91.

Morel, J.-L., Zouine, M., Andrieux, J., Faure-Muret, A., 2000. Déformations néogènes et quaternaires de la bordure nord haut atlasique (Maroc): rôle du socle et conséquences structurales. J. Afr. Earth Sci. 30, 119–131.

Muñoz, J.A., 1992. Evolution of a continental collision belt: ECORS-Pyrenees crustal balanced cross-section. In: McClay, K.R. (Ed.), Thrust Tectonics. Chapman and Hall, London, pp. 235–246.

Nafe, J.E., Drake, C.L., 1957. Variations with depth in shallow and deep water marine sediments of porosity, density and the velocity of compressional and shear waves. Geophysics 22, 523–552.

Pique, A., Charroud, M., Laville, E., Aït Brahim, L., Amrhar, M., 2000. The Tethys southern margin in Marocco; Mesozoic and Cenozoic evolution of the Atlas domain. Peri-Tethys Memoir 5: new data on Peri-Tethyan sedimentary basins. Mem. Mus. Natl. Hist. Nat. 182, 93–106.

Rimi, A., 1999. Mantle heat flow and geotherms for the main geologic domains in Morocco. Int. J. Earth Sci. 88, 458–466.

Sandvol, E., Calvert, A., Barazangui, M., 1998. Grid search modelling of receiver functions: implications for crustal structure in the Middle East and North Africa. J. Geophys. Res. 103 (B11), 26899–26917.

Sandwell, D.T., Smith, W.H., 1997. Marine gravity anomaly from Geosat and ERS1 satellite altimetry. J. Geophys. Res. 102 (B5), 10039–10054.

Schaer, J.P., 1987. Evolution and structure of the High Atlas of Morocco. In: Schaer, J.P., Rodgers, J. (Eds.), The Anatomy of Mountain Ranges. Princeton Univ. Press, Princeton, N.J., pp. 107–127.

Schwarz, G., Mehi, H.G., Ramdani, H.G., Rath, V., 1992. Electrical resistivity structure of the eastern Moroccan Atlas System and its tectonics implications. Geol. Rundsch. 81, 221–235.

Seber, D., Barazangui, M., Aissa Tadili, B., Ramdani, M., Ibenbrahim, A., Ben Sari, D., 1996. Three dimensional upper mantle structure beneath the intraplate Atlas and interplate Rif mountains of Morocco. J. Geophys. Res. 101 (B2), 3125–3138.

Seber, D., Sandvol, E., Sandvol, C., Brindisi, C., Barazangui, M., 2001. Crustal model for the Middle East and North Africa region: implications for the isostatic compensation mechanism. Geophys. J. Int. 147, 630–638.

Tadili, B., Ramdani, M., Sari, B., Chapochnikov, K., Bellot, A., 1986. Structure de la croute dans le nord du Maroc. Ann. Geophys. 4 (1), 99–104.

Teixell, A., 1998. Crustal structure and orogenic material budget in the west central Pyrenees. Tectonics 17, 395–406.

Teixell, A., Arboleya, M.L., Julivert, M., Charroud, M., 2003. Tectonic shortening and topography in the central High Atlas (Morocco). Tectonics 22, doi:10.1029/2002TC001460.

Van der Meijde, M., Van der Lee, S., Giardini, D., 2003. Crustal structure beneath broad-band seismic stations in the Mediterranean region. Geophys. J. Int. 152, 729–739.

Wigger, P., Asch, G., Giese, P., Heinsohn, W.-D., El Alami, S.O., Ramdani, F., 1992. Crustal structure along a traverse across the Middle and High Atlas mountains derived from seismic refraction studies. Geol. Rundsch. 81, 237–248.

Zeyen, H., Ayarza, P., Fernandez, M., Rimi, A., 2005. Lithospheric structure under the western African-European plate boundary: a transect across the Atlas Mountains and the Gulf of Cadiz. Tectonics 24, TC2001, doi:10.1029/2004TC001639.