# Crustal versus asthenospheric origin of relief of the Atlas Mountains of Morocco

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[1] We investigate the respective roles of crustal tectonic shortening and asthenospheric processes on the topography of the High Atlas and surrounding areas (Morocco). The lithospheric structure is modeled with a direct trial-and-error algorithm taking into account gravity (Bouguer and free air), geoid, heat flow, and topography. Three parallel cross sections, crossing the High Atlas and Anti-Atlas ranges, show that the lithosphere is thinned to 60 km below these mountain ranges. An analysis of the effect of the lithospheric thinning allows us to conclude that the whole topography of the Anti-Atlas, which belongs to the Sahara domain, is due to asthenospheric processes. In the High Atlas the lithospheric thinning explains a third of the relief of the western High Atlas, 500 m for a mean altitude of 1500 m, and half of the relief of the central High Atlas, 1000 m for a mean altitude of 2000 m. At the scale of Morocco the domain affected by lithospheric thinning forms an elongated NE-SW strip crossing not only the main structural zones but also the Atlantic margin to the south and the Africa-Eurasia plate boundary to the north. This major lithospheric thinning is associated with Miocene to recent alkaline volcanism and seismicity. We propose that this thermal anomaly is related to a shallow mantle plume, emplaced during middle to late Miocene time, during a period of relative tectonic quiescence.

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## 1. Introduction

[2] In Morocco, the most elevated regions are situated in the Atlas System (including the High and Middle Atlas belts) and, farther south, in the Anti-Atlas domain (Figure 1). This topography is generally considered of tectonic origin exclusively, i.e., due to crustal shortening in the framework of the active convergence between the African and Eurasian plates [see *Frizon de Lamotte et al.*, 2000, and references therein].

[3] It has been known for a relatively long time that the crustal root beneath the Atlas Mountains is not thick enough to isostatically support the topography [*Tadili et al.*, 1986; *Makris et al.*, 1985; *Wigger et al.*, 1992; *Mickus and Jallouli*, 1999; *Ayarza et al.*, 2005]. This subject remains a matter of debate because of an apparent discrepancy: (1) Gravity modeling suggests a crustal root down to 50 km, i.e., a low-density body at depth [*Seber et al.*, 1996]. (2) Seismic surveys show a crustal root ranging between 35 and 39 km under the highest part of the High Atlas [*Makris et al.*, 1985; *Tadili et al.*, 1986; *Wigger et al.*, 1992; *Ramdani*, 1998].

[4] The major problem pointed out by this last observation is that such a small root cannot maintain an isostatically compensated belt with topographic highs above 4000 m. It is the reason why different authors [*Makris et al.*, 1985; *Qureshir*, 1986; *Van den Bosch*, 1971; *Gómez et al.*, 1998] propose that the Atlas Mountains are presently in an uncompensated isostatic state. *Schwarz and Wigger* [1988] suggest that the topography could have a deep origin, which, in other words, means that the compensation level would be deeper than the Moho discontinuity.

[5] Recent papers propose a solution invoking a thin and hot lithosphere. *Seber et al.* [1996], based on teleseismic *P* wave travel time tomography, show that an upper mantle low-velocity anomaly exists beneath the High Atlas and propose that this anomaly contributes to the relief. *Ramdani* [1998] explains the rapid uplift of the Atlas Mountains, the volcanism, and the intermediate depth of earthquakes by a delamination process subsequent to the tectonic thickening of the whole lithosphere. *Teixell et al.* [2003], studying the central High Atlas by means of three parallel balanced cross sections, put forward an anticorrelation between the relief and the tectonic shortening. These results also suggest an asthenospheric origin for at least part of the relief.

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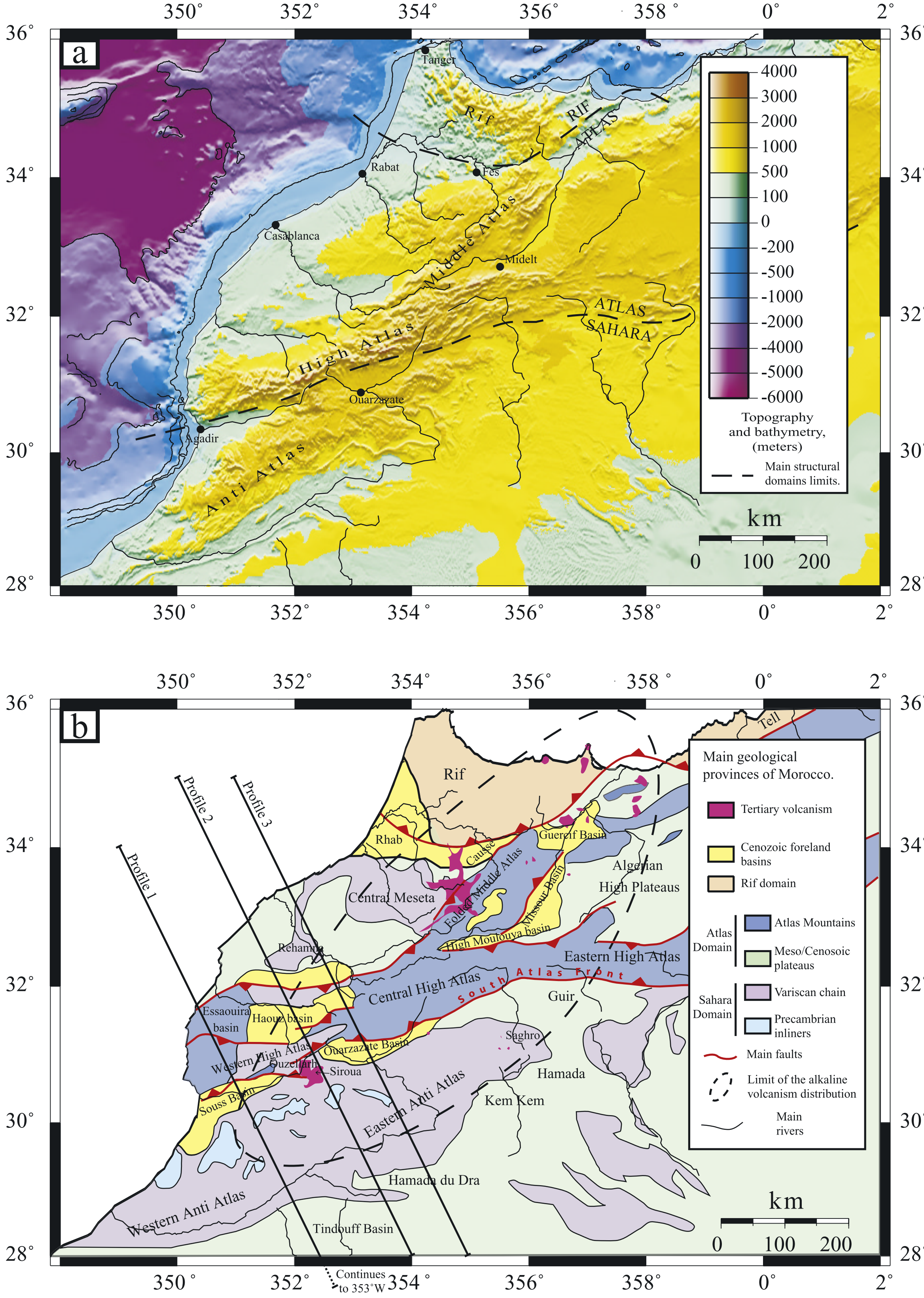


**Figure 1.** (a) Topographic map of northern Morocco (GEBCO data). The three major structural domains (Rif, Atlas, and Sahara) are indicated. Note that the intraplate High Atlas chain is higher than the Rif situated along the Africa-Eurasia plate limit. Relief follows two trends: one, WSW-ENE, is parallel to the structural boundary between Atlas and Sahara domains; the other, SW-NE, is oblique. (b) Structural sketch of the studied area. The position of the three lithospheric transects is indicated.

[6] East of the area studied in this paper, *Frizon de Lamotte et al.* [2004], *Zeyen et al.* [2005] and *Teixell et al.* [2005] modeled a thinned lithospheric structure beneath the Atlas system using gravity, geoid, heat flow, and topography data following the joint direct method fully described by *Zeyen and Fernàndez* [1994]. Finally, a first 3-D view of this zone of thinned lithosphere beneath the Atlas is given by *Fullea et al.* [2006], using a joint inversion method of geoid and topography anomalies without taking into account gravity anomalies.

[7] The aims of the present paper are (1) to identify the geometry of the lithospheric anomaly west of the beforementioned profiles and in the Anti-Atlas area [*Frizon de Lamotte et al.*, 2004; *Zeyen et al.*, 2005], (2) to quantify its influence on the present topography of the Atlas Mountains domain, and (3) to discuss its relationships with the geological history. After a short outline of the regional geology, we will present three new parallel lithospheric cross sections cutting through the lithospheric anomaly. Then, we will discuss the influence of the anomaly on the topography of the region. Finally, we will present its geometry in map view, its possible origin and the timing of the Atlas system uplift.

## 2. Geological Setting

[8] The Atlas system of Maghreb is an intracontinental orogenic domain running from Morocco to Tunisia over 2000 km. In Morocco, the Atlas system comprises the Middle Atlas and High Atlas belts and the Moroccan Meseta, which is a rigid block exhibiting a tabular thin meso-Cenozoic cover sandwiched between the Rif, the Atlas belts and the Atlantic Margin (Figure 1). East of the Middle Atlas, the Algerian High Plateaus are also a rigid core between the Tell, which is the eastern continuation of the Rif, and the Sahara Atlas. Then, at the scale of Maghreb, the Atlas system is bound northward by the Tell-Rif front and southward by the South Atlas Front (Figure 1).

[9] The Middle and High Atlas correspond to inverted Mesozoic intracontinental basins [see *Laville et al.*, 2004, and references therein]. The Middle Atlas is trending NE-SW while the High Atlas trend is WSW-ENE; both directions were inherited from the initial Triassic rifting. The highest peak of Morocco, and also of North Africa, is situated in the High Atlas (4165 m at Jebel Toubkal).

[10] South of the South Atlas Front and roughly parallel to the High Atlas is the Anti-Atlas domain (Figure 1). It is a foreland fold belt that belongs to the Variscan Appalachian-Ouachita-Mauretanides fold belt [see *Helg et al.*, 2004, and references therein]. It comprises Pan-African or older basement uplifts and a thick folded sedimentary cover including the uppermost Proterozoic and Paleozoic series. West, south, and east of the Anti-Atlas, the Upper Cretaceous and younger rocks rest unconformably on the Paleozoic rocks, forming large tablelands locally called ''Hamadas'' (Figure 1).

[11] The development of marginal foreland basins is expected along the borders of orogenic systems. So, for the purpose of our paper it is interesting to note the poor development and the absence of continuity of Cenozoic flexural basins fringing the Middle and High Atlas (Figure 1). This is particularly clear along the South Atlas Front (SAF), at the boundary between the High Atlas and the Anti-Atlas. At the westernmost end of the SAF, the Souss is a slightly subsiding flexural basin. East of it, the SAF cuts through a basement high, the Ouzellarh promontory [*Choubert*, 1942], which presents a mean elevation of about 2000 m and supports remnants of an Upper Cretaceous cover and a large Miocene volcanic complex, the Siroua [*Berrahma and Delaloye*, 1989; *De Beer et al.*, 2000]. East of the Ouzellarh, the Ouarzazate Basin is an inactive flexural basin, which is presently carried up to a mean altitude of 1300 m and undergoes erosion. Farther east, the Ouarzazate Basin disappears and the High Atlas directly overthrusts the flat and thin sedimentary cover of the Guir Hamada. Between the High and the Middle Atlas, the High Moulouya and Missour Basins are at a mean altitude of 1000 m (Figure 1). Situated west of the folded Middle Atlas, the Middle Atlas ''Causse'' corresponds to a tabular high plateau uplifted at a mean altitude of 1500 m. By contrast, the Haouz and Essaouira Basins, situated north of the westernmost High Atlas, are both located at low altitude (below 500 m). None of the available models are able to explain such differences in the topography of the undeformed zones.

## 3. Method and Data

[12] In order to explore the lithospheric structure of the Atlas and Anti-Atlas domains in Morocco, we used the trial-and-error algorithm of *Zeyen and Fernàndez* [1994], which determines the two-dimensional thermal and density structure of the lithosphere and therefore allows us to take into account the dependence of body density on temperature. This method, though it does not give a unique solution, allows us to greatly reduce the number of possible solutions through the combined use of different geophysical data sets.

[13] The modeling algorithm uses the fact that gravity, geoid and topography (calculated in local isostatic equilibrium) all depend on the density distribution but with different distance dependence [*Zeyen et al.*, 2005]. Topography reflects variations in average density of the lithosphere. The effect of density variations on gravity anomalies decreases proportionally to $r^{-2}$, whereas geoid undulations diminish proportionally to $r^{-1}$.

[14] Topography t is calculated as

$$t = \frac{\rho_a - \overline{\rho_1}}{\rho_a} H + t_0$$

where $\rho_a$ is the density of the asthenosphere (3200 kg/m$^3$); $\overline{\rho_1}$ is the average density of the lithosphere; $H$ is the thickness of the lithosphere (including the topography $t$); $t_0 = -2380$ m is a calibration constant that allows us to calculate absolute topography [*Lachenbruch and Morgan*, 1990] with respect to sea level. Variations of average density of the lithosphere and therefore of topography may be induced by density variations in the crust, crustal thickness variations or temperature variations in the lithospheric mantle, i.e., thickness of the lithosphere. The densities in the mantle are related linearly to temperatures through the formula:

$$\rho(T) = \rho_a * (1 - \alpha(T - T_a))$$

where $\alpha$ is the thermal expansion coefficient ($3.5 \times 10^{-5}$ K$^{-1}$) and $T_a$ the temperature at the lithosphere-asthenosphere boundary defined as the 1300°C isotherm.

[15] Gravity and geoid anomalies are calculated in two dimensions (see *Talwani et al.* [1959] and *Zeyen et al.* [2005], respectively). Gravity anomalies are mainly sensitive to density variations at shallow levels, i.e., in the crust, whereas geoid undulations and topography reflect also deeper-seated density variations and are rather sensitive to lithospheric thickness. Surface heat flow data, despite their significant lateral variability and inhomogeneous geographical distribution, constitute another parameter to control temperature distribution in the lithosphere depending on its thickness and the distribution of heat-producing elements.

[16] In order to calculate the in situ densities, the algorithm has to calculate first the temperature distribution. For this, a finite element grid is superposed on a predefined structure consisting of different bodies with constant heat production and thermal conductivity in each body (Table 1). The temperatures were calculated in steady state with the following boundary conditions: fixed temperatures at the top (10°C) and the bottom (1300°C) of the lithosphere and no horizontal heat flow at the vertical boundaries. The calculation in steady state is a simplification; however, this simplification minimizes the variations of lithospheric thickness: On the one hand, lithospheric thinning increases topography and vice versa. If, however, the thinning has occurred only recently, most of the lithosphere conserves the earlier lower temperatures and keeps topography at a lower level, so that one would need an even thinner lithosphere to adjust the data. We opted for the steady state calculations since the timing of thickness variations is too uncertain to confidently make calculations in nonsteady state.

[17] Surface heat flow has been compiled by *Rimi* [1990, 1999]. The data on Figure 2 (circles with error bars) are projected from a 100 km wide strip to each side of the profiles. All free air gravity and topography data used are taken from the worldwide data sets available from the site ftp://topex.ucsd.edu/pub [*Sandwell and Smith*, 1997]. Bouguer gravity data in Morocco have been taken from a gridded data set based on the data published by *Hildenbrand et al.* [1988]. The topography data have estimated errors of less than 20 m onshore and less than 200 m offshore (see, e.g., http://www.ngdc.noaa.gov/mgg/topo/report/s7/s7Bi.html). The uncertainty of the free air gravity data is estimated as 3–5 mGal offshore [*Sandwell and Smith*, 1997], whereas the Bouguer data in Morocco (see Figure 3 for point distribution) have errors of less than 1 mGal [*Van den Bosch*, 1971]. Most of these data were obtained with topographic correction of less than 0.5 mGal; locally those corrections may reach a maximum of 6 mGal. Geoid height (Figure 2d) is taken from the EGM96 global model [*Lemoine et al.*, 1998] with errors of less than 30 cm (see http://cddis.gsfc.nasa.gov/926/egm96/contents.html). In order to avoid effects of sublithospheric density variations on the geoid, we have removed the geoid signature corresponding to the spherical harmonics developed until degree and order 8 [*Bowin*, 1991]. Since our calculations were done in two dimensions, it is important to have a measure of the variability of the data in the direction perpendicular to the profile. To avoid small-scale local extrema, the different data sets were averaged every 5 km from a 50 km wide strip on both sides of the profile (100 km for the heat flow data) and the standard deviation of the data within these strips, that is everywhere larger than the data uncertainty at every single point, is considered to represent the uncertainty of our data. In Figure 2, those data are represented on each model by circles with error bars that correspond to the standard deviation of the data within the strip.

**Table 1.** Properties of the Different Bodies

| Body | Heat Production, $\mu$W/m$^3$ | Thermal Conductivity, W/(K × m) | Density, kg/m$^3$ |
|---|---|---|---|
| Sediments | $2 \times 10^{-6}$ | 2 | from 2300 to 2400 |
| Upper crust | $2.5 \times 10^{-6}$ | 3 | 2770 |
| Lower crust | $2 \times 10^{-7}$ | 2.1 | 2900 |
| Lithospheric mantle | $2 \times 10^{-8}$ | 3.2 | temperature dependent |

[18] Crustal thickness is constrained at some locations by seismic reflection data [*Makris et al.*, 1985] and these constraints are shown on Figure 2 with an assumed vertical uncertainty of ±2 km.

## 4. Description of the Lithospheric Cross Sections

[19] Three parallel cross sections have been modeled (Figure 2, see also Figure 1b). As much as possible, they are perpendicular to the belts, and long enough to avoid edge effects. The western section cuts through the south-eastern Essaouira Basin, the western High Atlas, the Souss Basin and the Anti-Atlas. The central section cuts through the highest part of the central High Atlas, the Ouzellarh promontory and the Anti-Atlas. The eastern section cuts through the central High Atlas, the Neogene Ouarzazate Basin and the eastern Anti-Atlas. The three cross sections end in the flat Tindouff Paleozoic Basin belonging to the Sahara domain.

[20] In our models, the base of the lithosphere lies at a depth of ~60 km underneath the High Atlas and the Anti-Atlas. Such a thin lithosphere is a necessary condition to obtain a model in local isostatic equilibrium satisfying all available data without chemical changes in the mantle. It is possible to change the lithospheric thickness by up to 10–15% and still obtain an acceptable fit of all data if we modify also crustal thickness and density distribution. However, a more important change than this will result in synthetic data, especially geoid and topography, that lie outside of the error bars for at least one data set. Figure 4 shows a numerical experiment where we increased the lithospheric thickness in the thinned area. The lithosphere-asthenosphere boundary is thus lying at depth between 75 and 85 km. The crustal structure has been modified in order to fit the gravity data. The curves obtained for the synthetic geoid and topography are now outside or at the limit of the assumed data uncertainties. The misfit reaches 2 m for the geoid and hundreds of meters for the topography. This exercise demonstrates that the effect of the lithospheric anomaly, particularly on the geoid, is too pronounced to be compensated by a shallower mass distribution.

[21] The modeled topography (Figure 2) is within the uncertainties of the data, except for the highest parts of the

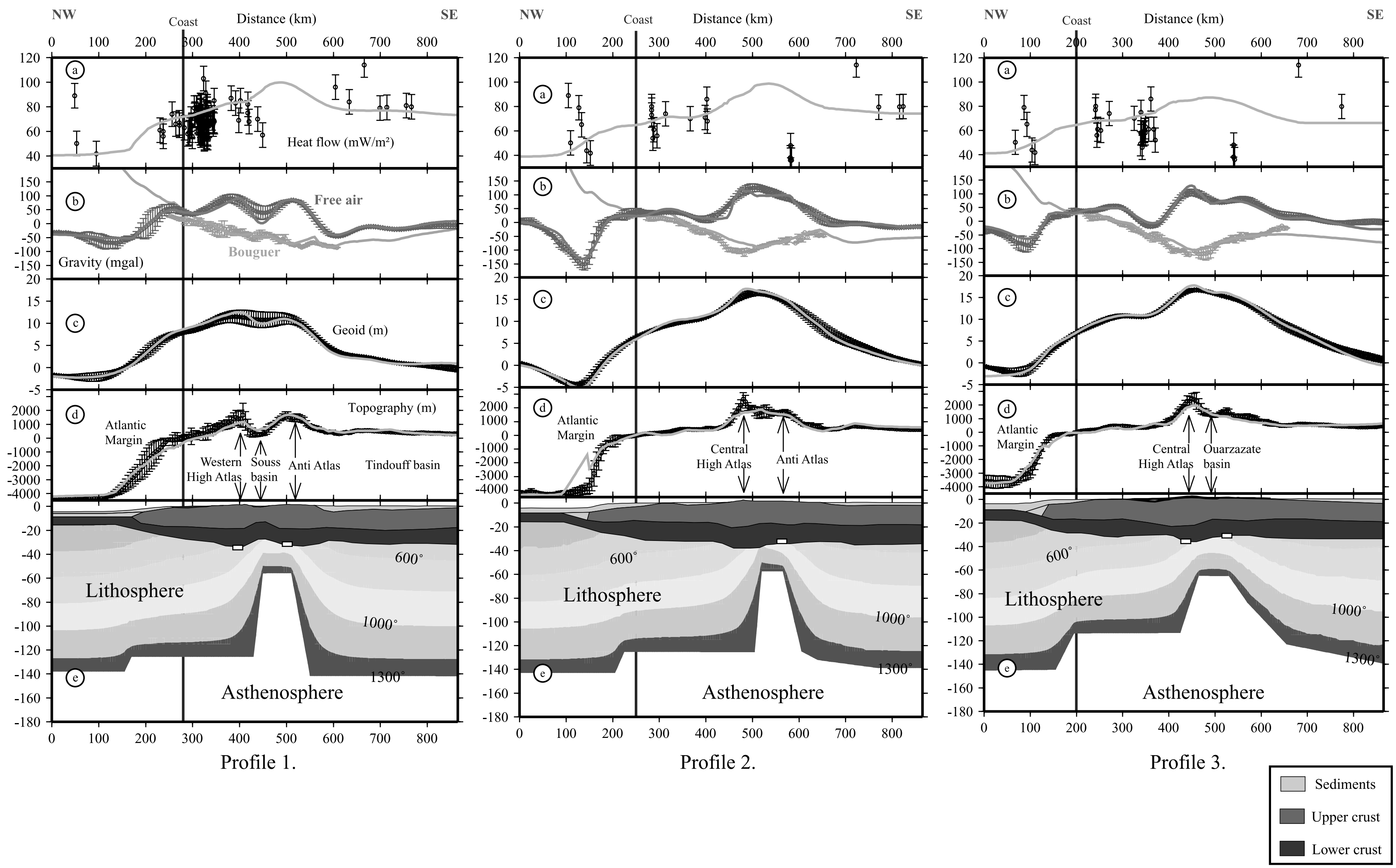


**Figure 2.** Lithospheric cross-section models through Morocco (see location on Figure 1b). For each profile, (a) heat flow, (b) free air and Bouguer gravity anomalies, (c) geoid, (d) topography, and (e) resulting model with isotherms in the mantle every 200°C. Circles correspond to data extracted from worldwide data sets with uncertainty bars and solid lines to calculated values. The model includes asthenosphere, lithosphere with temperature distribution, upper and lower crust, and sediments. The crustal structures are drawn wherever possible from available seismic data (white boxes). Short-wavelength discrepancies between measured data and modeling could be due to not fully compensated structures or small local variations of density distribution. The Atlantic Margin is beyond the scope of this paper and has only been modeled to avoid edge effects.

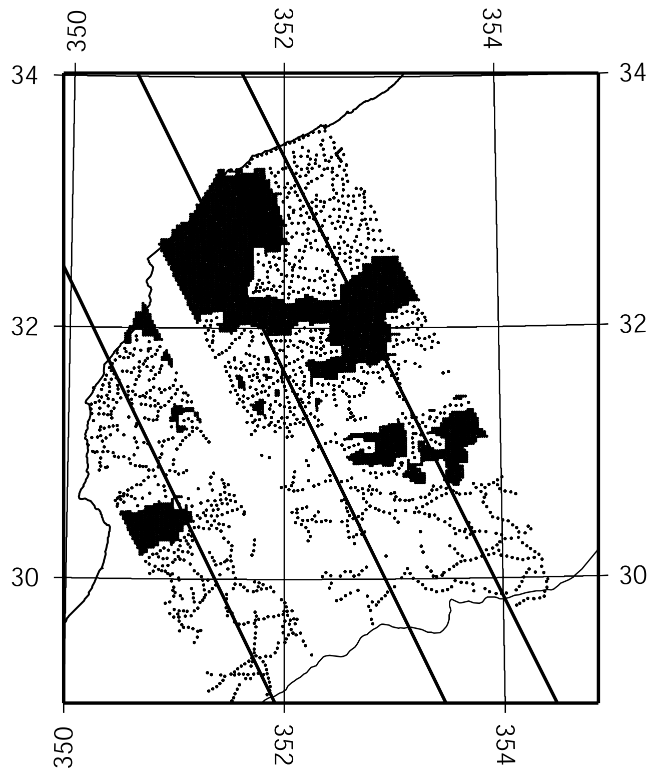


**Figure 3.** Distribution of the Bouguer gravity data used in this study. Data are from *Hildenbrand et al.* [1988].

High Atlas and the Atlantic margin. This last one has been modeled only to avoid edge effects and is beyond the scope of this paper. The other discrepancies may be explained by small-scale (<50 km) not fully compensated structures or small geological bodies not incorporated in the model.

[22] In order to test how much of the actual topography may be supported by an elastic plate we have filtered the topographic data in the Fourier domain with the following coefficients [*Turcotte and Schubert*, 1982, p. 123]:

$$\Omega(k) = \frac{\rho_{topo} \cdot g}{D \cdot k^4 + \left(\rho_a - \rho_{topo}\right) \cdot g} H(k)$$

where $\Omega$ and H are the Fourier coefficients of the plate deformation and the topography, respectively, k is the wave number, D is the flexural rigidity, g is the gravitational acceleration and $\rho_{topo}$ and $\rho_a$ are the densities of the material forming the topography (taken here as average crustal density of 2800 kg/m$^3$) and of the asthenosphere, respectively. The resulting deformation is then retransformed into the corresponding topography in local isostatic equilibrium by multiplication with the factor $(\rho_a - \rho_{topo})/\rho_a$ and compared with the measured topography. Figure 5 shows the results of this calculation for profile 3 (see Figure 1 for location). The thin dashed line corresponds to the measured topography, the thin solid line to the modeled one (Figure 2). The thick lines represent the topography filtered by a 5, 10 and 15 km thick elastic plate (dotted, solid and dashed lines, respectively).

[23] These calculations show that for the Anti-Atlas (between 500 and 600 km along the profile), the modeled topography corresponds to the filtered topography, so that the difference from the actual topography may effectively be explained by flexural effects, corroborating the need for a thin lithosphere. In the High Atlas, the modeled topography is in between the observed and the filtered ones. This may indicate, on the one hand, that we overestimated the topography and with this the crustal thickness and/or the lithospheric thinning. On the other hand it may be that the flexural effect has been overestimated, since implicitly it is supposed that all of the topography is due to additional mass on top of the plate. If, however, part of the topography is due to a mass deficit below the plate, the elastic strength of the lithosphere would also reduce this effect, implying that the lithospheric thinning obtained in our models would even be underestimated. Therefore elastic support is shown to have little effect on our results.

[24] Taking into account the proximity of the profiles, the thermal anomaly can be considered as continuous from one profile to the next.

## 5. Lithospheric Thinning Versus Topography of the Moroccan Atlas

[25] *Teixell et al.* [2005] make the first attempt to quantify the impact of the anomaly on the topography of the Moroccan Atlas system farther east; they estimate that the crustal thickening, i.e., the tectonic effect, explains about 50% of the topography of the High Atlas whereas the other 50% are due to the buoyancy exerted by asthenospheric upwelling. Concerning the topography of the Anti-Atlas, *Jacobshagen et al.* [1988] and *Frizon de Lamotte et al.* [2000] suggest a tectonic origin that could be related to a midcrustal fault propagating southward from the High Atlas during Neogene compressional phases.

[26] In order to constrain and quantify the effect of the thermal anomaly on the topography, we removed the anomaly of the model by defining a lithosphere/asthenosphere boundary gently dipping from 125 km at the Atlantic margin to 140 km under the West African Craton, thus maintaining the difference between oceanic and continental lithosphere of our model. The difference between the topography with and without the anomaly reflects the "thermal topography." Subtracting this "thermal topography" from the present topography gives an idea of what would be the topographic cross sections without lithosphere thinning (Figure 6).

[27] It appears on Figure 6 that the thermal anomaly has a major effect on the relief of the Atlas Mountains. Without the thermal topography, the Anti-Atlas lies at the same altitude as the undeformed Hamada, losing up to 1000 m of elevation. This indicates that even if recent deformation certainly occurs in the Anti-Atlas, as shown by the diffuse seismicity [*Hatzfeld et al.*, 1977; *Hatzfeld and Frogneux*, 1981], crustal deformation has not contributed significantly to the present-day relief. In other words, the high mean elevation of the Anti-Atlas seems to result almost entirely from processes occurring in the asthenosphere.

[28] In the High Atlas, the effect of the thermal anomaly on the topography is less important. On the western transect, crossing the western High Atlas (mean altitude, 1500 m), the altitude loss is approximately 500 m. To the east, the thermal effect in the central High Atlas (mean altitude,

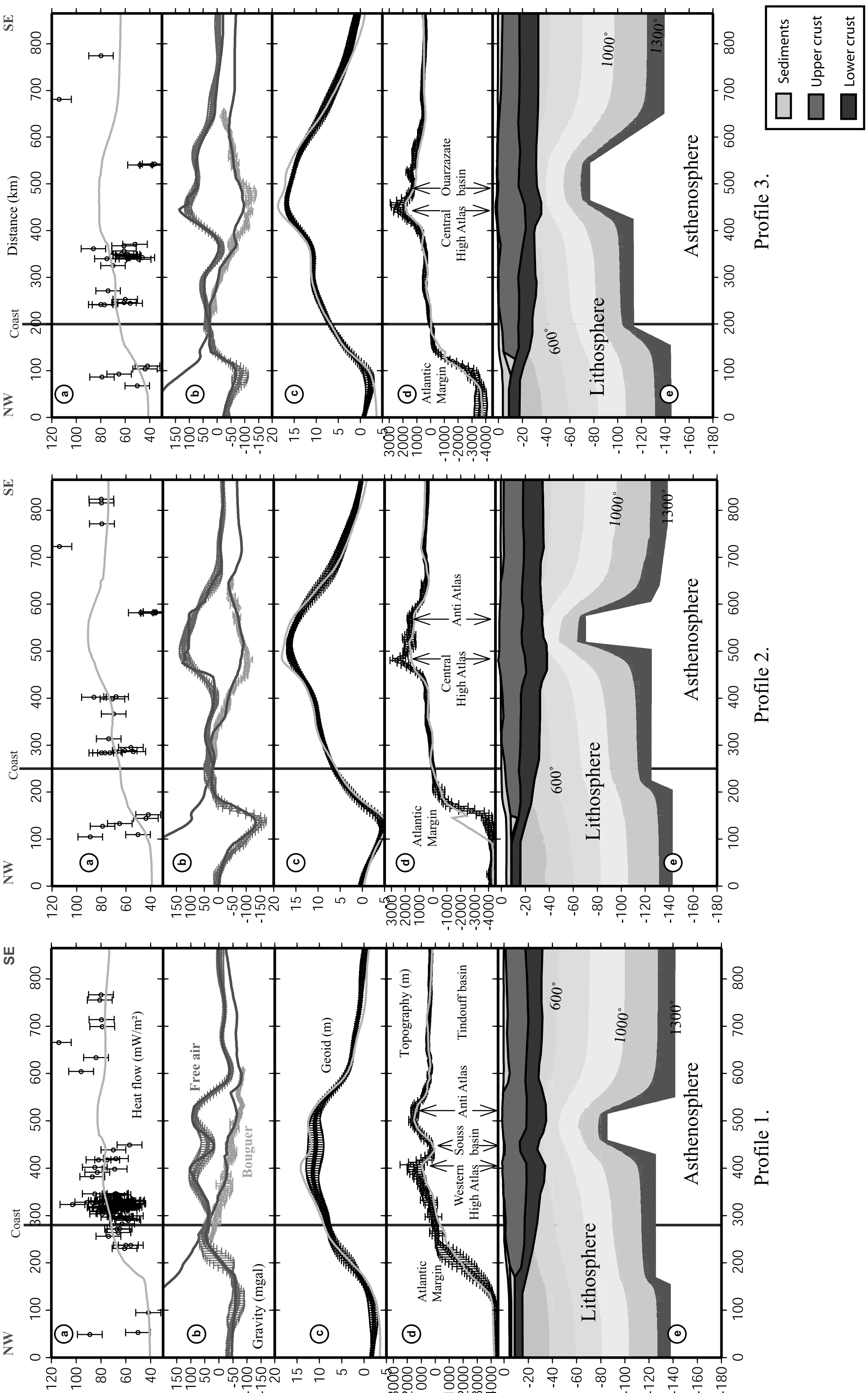


**Figure 4.** Alternative models, with same presentation of the lithospheric transects as in Figure 2. The solid curves correspond to the result of models with a deeper lithosphere-asthenosphere boundary (between 85 and 75 km depth) in the thinned regions (see Figure 2). Crustal thickness has been modified to fit the free air and Bouguer data. On all profiles the geoid curve reaches or exceeds the upper limit of the data uncertainty bars, whereas the topography reaches or exceeds the lower limit. The lithospheric thinning cannot be further reduced while fitting the measured data. It is therefore underestimated in these models. This demonstrates that topographic elevation cannot be compensated by variations of crustal mass distribution, and major lithospheric thinning as modeled on Figure 2 is needed to explain all data.

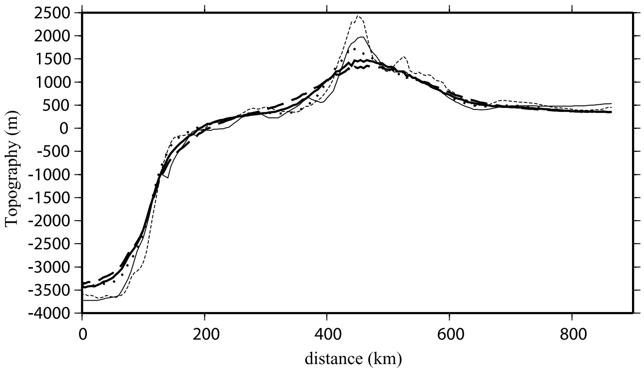


**Figure 5.** Support by an elastic plate. The thin solid line represents the topography calculated with our modeling of cross section 3 (see Figure 2) in local isostatic equilibrium; the thin dashed line represents the measured topography. The thick lines represent the topography filtered by a 5, 10 and 15 km thick elastic plate (dotted, solid, and dashed lines, respectively).

2000 m) increases up to 1000 m. Those lateral variations reflect changes of the position and depth of the lithospheric anomaly.

[29] The flexural basins bordering the High Atlas to the south are strongly affected by the lithospheric anomaly. Without the anomaly, the surface of the Souss Basin would lie below sea level, and that of the Ouarzazate Basin, which is presently at 1300 m, would be between 0 and 500m. By contrast, the Haouz Basin, situated along the northern front of the High Atlas, is almost not affected.

## 6. Discussion

### 6.1. Geometry and Origin of the Thermal Anomaly

[30] In order to constrain the geometry of the lithospheric anomaly, we have plotted on a map the three new profiles detailed above and added (1) a segment made in the frame of the Transmed transect I project [*Frizon de Lamotte et al.*, 2004] crossing the High Atlas farther east and (2) *Zeyen et al.* [2005] profile (Figure 7). For each profile, we have drawn the 70 km and 100 km lithospheric thickness isopachs and a line joining the points where the lithosphere-asthenosphere boundary is highest. Owing to the proximity of the profiles and the value of the thinning, we assume continuity between them. On the map, the thinned zone appears to be an elongated strip extending from the Anti-Atlas to the Middle Atlas and crossing the High Atlas (Figure 7). Its NE-SW direction is slightly but clearly oblique to the mean trend of the Atlas system, which is E-W to ENE-WSW [see *Frizon de Lamotte et al.*, 2000]. Moreover, outside the zone affected by the thermal anomaly the mean altitude of the High Atlas decreases rapidly to reach a value of about 1500 m, which is similar to the Sahara Atlas (Algeria), where relief is entirely tectonics-related [*Frizon de Lamotte et al.*, 2000].The whole Anti-Atlas, the Ouarzazate Basin, the central High Atlas, the folded Middle Atlas, the Missour Basin and the Middle Atlas "Causse" are situated within the zone where the lithosphere thickness is less than 70 km. This explains why the Ouarzazate and Missour Basins are at an unexpected altitude compared to the equivalent Haouz Basin, which is outside the lithospheric anomaly (Figure 1). The situation of the Souss Basin is more complex because it is lying on a slightly thinned crust because of its proximity to the Atlantic margin [*Makris et al.*, 1985; *Mustaphi*, 1997]. In the absence of the lithospheric anomaly, the whole Souss would be below sea level.

[31] At the scale of Morocco, we emphasize that the strip ignores the boundaries between the main structural domains and crosses not only the South Atlas Front but also the Rif Front farther north. This strongly suggests that the lithospheric anomaly is superimposed on the geological framework and independent from it. Even if we would consider large lateral uncertainties and suppose that the very thin lithosphere may extend to the 100 km isopach of our model, this observation remains valid. For this reason, the hypothesis suggesting that the thinning of the lithosphere is linked to a delamination event in response of the Africa-Eurasia collision [*Ramdani*, 1998] is rejected. Moreover, the tectonic shortening ratio measured in the Atlas system upper crust [*Beauchamp et al.*, 1999; *Frizon de Lamotte et al.*, 2000; *Teixell et al.*, 2003, *Arboleya et al.*, 2004] is less than 20% and is unlikely to generate a significant lithospheric root. For the same reason, i.e., independence between the orientation of surface structures and the crosscutting elevation anomaly, we do not favor the hypothesis that mantle upwelling is induced by lateral flow from Eurasia-Africa subduction [*Teixell et al.*, 2005].

[32] The plume hypothesis has recently been revisited. It was proposed that either it is a long-lived large mantle upwelling lasting since the Triassic [*Oyarzún et al.*, 1997; *Anguita and Hernán*, 2000] or as a small Cenozoic asthenospheric plume similar to those observed in the West European Alpine Foreland being a part of a deep mantle reservoir system (French Massif Central, German Eifel) [*Zeyen et al.*, 2005]. We favor this second hypothesis, which takes into account two important constraints: (1) the alkaline affinities and the middle to upper Miocene ages of volcanism along this major cross element and (2) its crosscutting relationships with the Eurasia-Africa plate boundary. Those constraints will be described in detail in section 6.2.

[33] The northeastern and southwestern extensions of the thinned lithospheric zone cannot be drawn from our 2-D models. According to *Fullea et al.* [2006], it extends northward into a zone known as the "trans-Alboran" volcanic corridor crossing the Alboran Sea from the Eastern Rif to the Eastern Betics [*Jacobshagen et al.*, 1988; *Gómez et al.*, 2000a and references therein]. To the south, the thinned lithospheric zone may connect with the Canary Islands volcanic domain [*Anguita and Hernán*, 2000, and references therein]. This NE-SW strip also corresponds to the zone where the Neogene to Quaternary magmatic activity [*Harmand and Cantagrel*, 1984; *El Azzouzi et al.*, 1999; *Anguita and Hernán*, 2000; *Teixell et al.*, 2005; *Zeyen et al.*, 2005] (Figure 1) and the seismicity (Figure 8), including intermediate-depth earthquakes (down to 150 km) [*Hatzfeld and Frogneux*, 1981; *Ramdani*, 1998] are concentrated. A similar concentration of seismicity above thinned lithosphere is also observed in the Baikal rift system [*Sherman et al.*, 2004]. In our case, the Africa Europe plate boundary is underlined by an E-W seismic

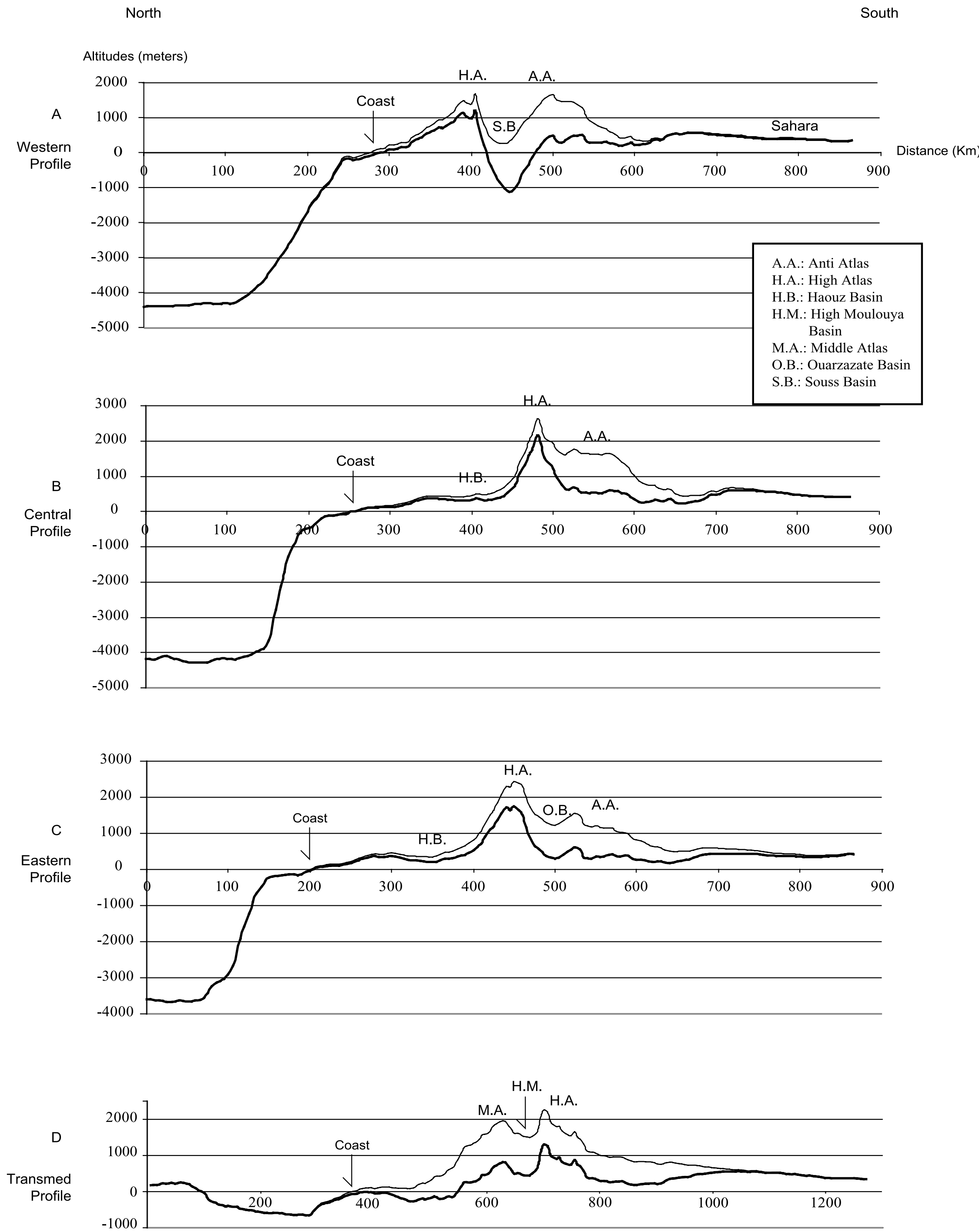


**Figure 6.** Comparison between actual topography (thin line) and topography without lithospheric thinning (bold line) calculated with the crustal model of Figure 2. The topography without lithospheric thinning only reflects the crustal-compensated structures and small-scale (<50 km) not fully compensated structures. Without thermal doming the Anti-Atlas relief is situated around 500 m above sea level, whereas crustal thickening supports the main part of the High Atlas relief. However, in the High Atlas the crustal compensation of the topography decreases from the eastern profile to the western one, while the thermal topography increases.

trend and crossed by a SW-NE trend above the thermal anomaly identified with cross sections. Consequently, it appears as a zone crossing not only the different structural domains of Morocco but also the Atlantic margin to the south and the Africa-Eurasia plate boundary to the north.

### 6.2. Timing of Onset of the Mantle Heating Under the Moroccan Atlas System

[34] We have shown that in the Atlas system both tectonic shortening and thermal anomaly contribute to the relief and we have a quite precise idea of the contribution of each

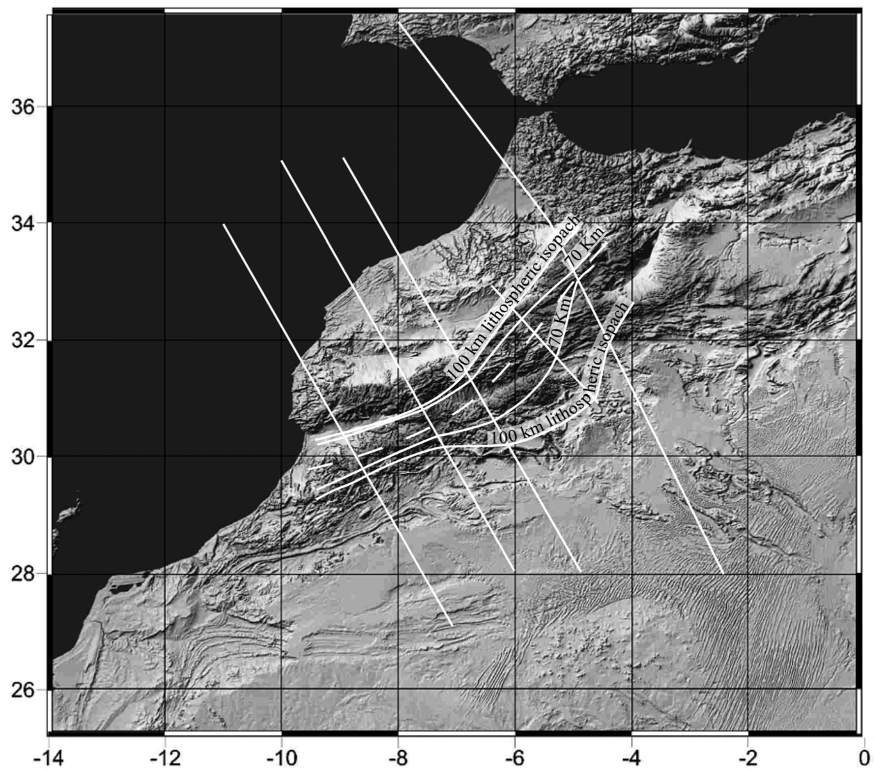


**Figure 7.** GTOPO30 (U.S. Geological Survey, GTOPO30—Global topographic data, http://edcdaac.usgs.gov/, 2000) topography and isopach map of the lithosphere in Morocco as deduced from correlation between (from west to east) the three lithospheric profiles, a segment of the Transmed I profile [*Frizon de Lamotte et al.*, 2004] and *Zeyen et al.* [2005] profile. The dashed line represents the position of the minimum lithospheric thickness. The lithospheric thinning is oriented SW-NE and crosses obliquely the different structural domains of Morocco (Figure 1).

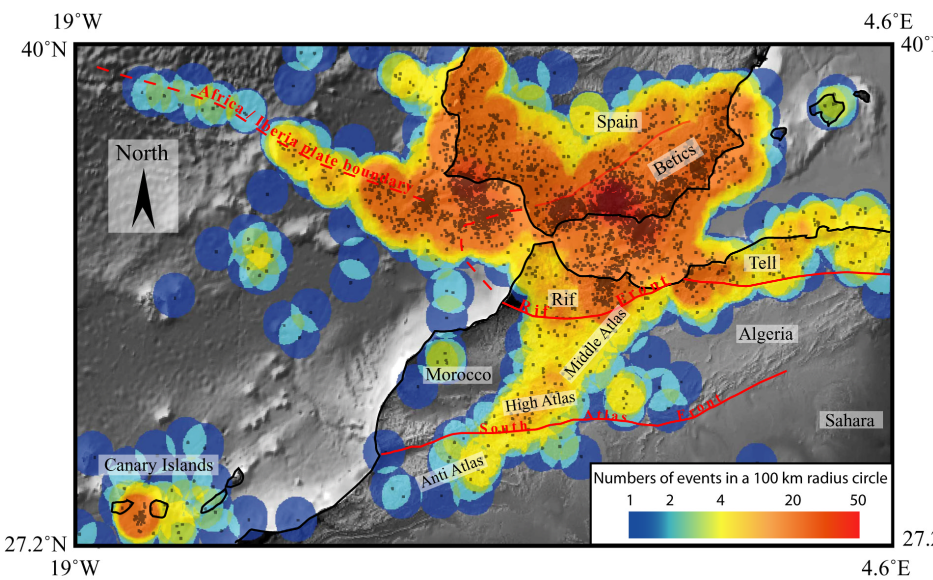


**Figure 8.** Density map of seismicity in the Mediterranean-Atlantic transition calculated from International Seismological Centre database from 1995 to 2000. Grey points indicate epicenters. An E-W trend clearly follows the Iberia-Africa plate boundary. It is crossed by a NE-SW trend from the Anti-Atlas to the Eastern Betics following the zone where the lithosphere is thin (see Figure 7).

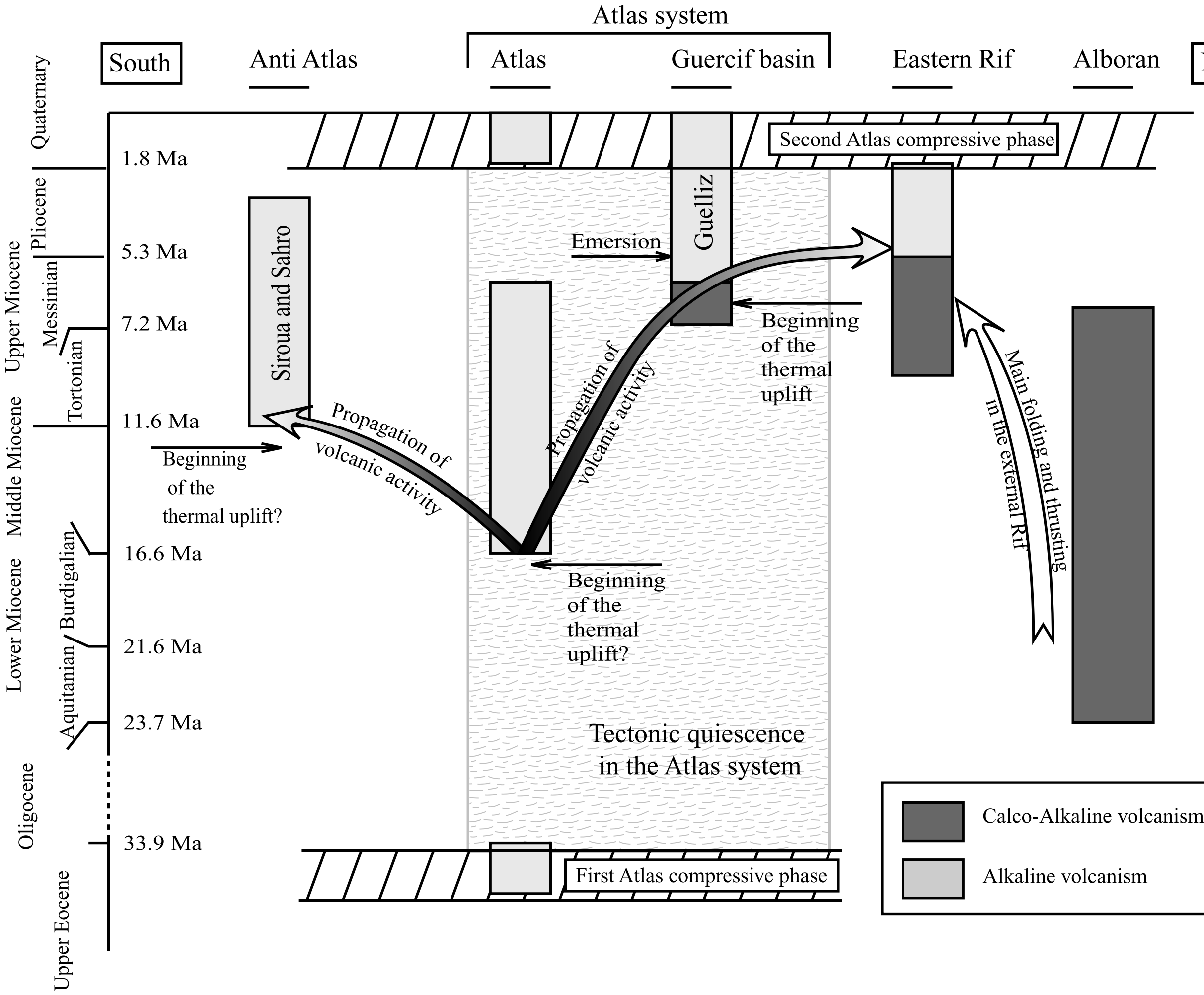


**Figure 9.** Sketch illustrating the timing of the tectonic phases and magmatic activity in Morocco. It appears that the alkaline volcanism began earlier in the Atlas domain. It propagated southward to the Anti-Atlas (Siroua and Sahro volcanoes) and northward to the Rif (Guelliz and Eastern Rif) where it replaces the calcoalkaline subduction-related volcanism. The well-constrained uplift of the Guercif Basin is taken from *Krijgsman et al.* [1999]. Note the good coincidence with the volcanic activity. Ages of volcanic rocks are from *El Azzouzi et al.* [1999] and *De Beer et al.* [2000].

component. The question of the timing is now addressed: Are the two processes successive or coeval? The sequence of tectonic events in the Atlas system is well established in Algeria, a region situated away from the thermal anomaly and where the whole relief is related to crustal shortening. Field and subsurface data from Algeria reveal phases of tectonic activity during late Eocene and Quaternary time [*Laffitte*, 1939; *Guiraud*, 1975; *Ghandriche*, 1991; *Frizon de Lamotte et al.*, 1998; *Khomsi et al.*, 2004] (see especially *Bracène and Frizon de Lamotte* [2002, Figure 10]). A thick siliclastic or marine sedimentary sequence was deposited during Oligocene-Miocene tectonic quiescence [*Frizon de Lamotte et al.*, 1998; *Bracène and Frizon de Lamotte*, 2002; *Khomsi et al.*, 2004]. It has been demonstrated by *Frizon de Lamotte et al.* [2000] that this sequence of events is representative for the whole Maghreb.

[35] We speculate that the episode of uplift and erosion, which began in the Atlas system of Morocco by the middle Miocene [*Görler et al.*, 1988; *Zouine*, 1993; *Chellaï and Perriaux*, 1996; *El Harfi et al.*, 1996; *Gómez et al.*, 2000a], is related to thermal doming. Moreover, this period is also characterized by intense magmatic activity. More precisely, in the Moroccan Atlas system and adjacent areas, the period of highest alkaline volcanic production is mid-Miocene in age even if it started earlier, i.e., during the Eocene (see review given by *El Azzouzi et al.* [1999]), and continued up to the Quaternary as in western and central Europe [*Hoernle et al.*, 1995; *Merle and Michon*, 2001; *Cloetingh et al.*, 2005]. A compilation of available ages (Figure 9) suggests that the magmatic activity migrated from the Middle Atlas toward the NE (Tertiary Guercif Basin) and the SW (Siroua), suggesting that related relief is also younger in these areas.

[36] Accordingly, in the Guercif Basin, located on the northeastern plunge of the Middle Atlas, *Gómez et al.* [2000b] and *Zizi* [2002] recognize the following events: (1) pre-Tortonian erosion; (2) Tortonian rifting characterized by a NW-SE extension; and (3) late Pliocene compression corresponding to the second tectonic Atlas event. Magneto-stratigraphic and biostratigraphic results allow *Krijgsman et*

*al.* [1999] to show that the basin opened at 8 Ma, underwent a rapid shallowing by at least 400 m between 7.1–7.2 Ma and 6 Ma, emerged at this date and reaches now a mean altitude of about 400 m above sea level. The present uplift rate is estimated at a minimum value of 2 mm/year for the last 5700 years [*Zarki et al.*, 2004]. It is worth noting that the uplift of the basin surface compared to sea level and related closure of the Guercif Basin is traditionally linked to the advancement of the Tell-Rif thrust front. An alternative hypothesis, proposed here, is that there is a relationship with the thermal dome crossing the plate boundary. Moreover, the Tell-Rif front exhibits an important reentrant between the eastern Rif and the western Tell and is in fact quite far from the Guercif Basin (Figure 1).

[37] Very few data are available to estimate the age of the uplift of the Anti-Atlas, at the southern part of the thermal dome. On the basis of sedimentologic analysis, *Görler et al.* [1988] propose that the Ouarzazate Basin was endoreic (internally drained) during Late Miocene time. *Stäblein* [1988], *Berrahma and Delaloye* [1989] and *Ibhi* [2000] also suggest late Miocene uplift during activity of the Siroua and Saghro volcanoes. However, uplifted Quaternary landforms indicate younger activity in the Anti-Atlas [*Görler et al.*, 1988]. For this very recent uplift, relative magnitude of tectonic versus thermal uplift cannot be distinguished.

## 7. Conclusion

[38] Our study quantifies the mixed crustal and asthenospheric origin of topographic relief in the Atlas system of Morocco, west of the region considered by previous studies [*Frizon de Lamotte et al.*, 2004; *Zeyen et al.*, 2005; *Teixell et al.*, 2005]. The good fit of topography, geoid, and gravity data to our model (Figure 2), and the poor fit to any model with thicker lithosphere (Figure 4), demonstrates that crustal isostatic compensation is insufficient to explain observations and requires density variations to be present in the deeper lithosphere.

[39] Some topographic relief has resulted from crustal thickening phases that occurred during late Eocene and Quaternary times. This intracontinental deformation partly accommodates Africa-Eurasia plate convergence during times of strong plate coupling and, at the scale of North Africa, produced the High Atlas belt [*Frizon de Lamotte et al.*, 2000; *Gómez et al.*, 2000a].

[40] Topographic relief that is related to low-density lithosphere and shallow asthenosphere has developed obliquely to the trend of the Anti-Atlas, central High Atlas, and Middle Atlas mountains and surrounding basins. This asthenospheric component of relief formed during a middle and late Miocene period of relative tectonic quiescence, during which time the Tell-Rif developed and subduction in the Mediterranean accommodated most Africa-Eurasia plate convergence [*Frizon de Lamotte*, 2000; *Faccenna et al.*, 2004].

[41] The occurrence of alkaline magmatic activity along a NW-SE strip crossing the main crustal structures in Morocco, and oblique to the plate boundary, leads us to conclude that the primary cause of elevated topography and alkaline magmatism is mantle upwelling, rather than Africa-Eurasia plate boundary geodynamics (e.g., subduction rollback). However, we suggest that the coincidence of mantle upwelling with the Miocene time of weak interplate coupling was significant in producing the observed geometry of the Atlas system. The highest topography of Maghreb is observed in the central High Atlas, where mantle upwelling coincided with the already elevated Atlas range. During Quaternary time, we suggest that plate convergence is again the main factor contributing to higher topographic relief across the Maghrebian domain.

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