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Effects of mantle upwelling in a compressional setting: the Atlas Mountains of Morocco

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ABSTRACT

We discuss the implications of a lithospheric model of the Moroccan Atlas Mountains based on topography, heat flow, gravity and geoid anomalies, taking into account the regional geology. The NW African cratonic lithosphere, some 160–180 km thick, thins to *c.* 80 km beneath the Atlas fold-thrust belts, in contrast with the shortening regime prevailing there since the early Cenozoic. This fact explains several geological and geophysical features as high topography with modest tectonic shortening, the occurrence of alkaline magmatism contemporaneous to compression, the absence of large crustal roots to support elevation, the scarce development of foreland basins, and a marked geoid high. The modelled lithosphere thinning is related to a thermal upwelling constrained between the Iberia–Africa convergent plate boundary and the Saharan craton.

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Introduction

In plate-boundary orogenic belts, the work to elevate mountains is mainly done by crustal thickening because of tectonic shortening or magmatic addition, as the negative buoyancy of subducted lithosphere may counteract the crustal buoyancy pulling the earth's surface down. However, in continental interiors, mountain building by intraplate compressional stresses may be assisted by positively buoyant mantle upwellings that form part of convection circuits within the mantle. Mantle-derived surface effects are sometimes called dynamic topography, as the buoyancy sources that create them are moving (Lithgow-Bertelloni and Gurnis, 1997). However, on a scale of hundreds of kilometres, the reaction of the lithosphere on vertical mantle push is sufficiently fast to be considered always in a quasi-static force equilibrium (Zhang and Christensen, 1993; Davies, 1999), and can be modelled as isostatically compensated in the sublithospheric mantle.

The Atlas Mountains of Morocco are intracontinental chains with high topography that contrasts with modest crustal tectonic shortening and thickening. To explain this apparent contradiction, a mantle contribution to uplift was invoked (Teixell *et al.*, 2003). We now synthesize the results of lithospheric modelling that, in combination with a review of the known geological and geophysical evidence, define a mantle upwelling, largely independent of the local tectonic setting, that explains high topography and alkaline volcanism in the Atlas region.

Geological evidence

The Atlas chains of Morocco consist of two main branches, the E-trending High Atlas, with summits of 4000 m, and the NE-trending Middle Atlas, which reaches 3000 m (Fig. 1). These chains are fold-thrust belts formed during the Cenozoic from inversion of Triassic–Jurassic transtensional troughs in the interior of the African plate (e.g. Mattauer *et al.*, 1977; Warme, 1988; Giese and Jacobshagen, 1992; Frizon de Lamotte *et al.*, 2000; Piqué *et al.*, 2002; Arboleya *et al.*, 2004). Tectonic shortening across the Atlas Mountains is modest: Teixell *et al.* (2003) calculated total values ranging from 15% to 24% along transects of the High Atlas, whereas Gomez *et al.* (1998) estimated it in 15% in the central Middle Atlas.

High topography is not confined to the extent of the High and Middle Atlas. The chains are superposed on a broad topographic swell, so that the areas flanking them (Meseta, High Plateaux and Anti-Atlas; Fig. 1), still have a high elevation, with altitudes between 1000 and 2000 m, in spite of having largely escaped Cenozoic deformation. Foreland basins are poorly developed; synorogenic sediments are only locally preserved, and usually a few hundred metres thick.

There is abundant magmatism of Cenozoic age in southern Morocco. It is of alkaline to hyperalcaline character, and includes volcanic (basalts, phonolytes, trachytes), and minor plutonic rocks (nepheline syenites with carbonatite occurrences) (Agard, 1973; Harmand and Cantagrel, 1984; Berrahma and Hernandez, 1985; El Azzouzi *et al.*, 1999). The age and distribution of this magmatism is indicated in Fig. 1. It occurred synchronically with the Atlas compression, a fact that has intrigued previous authors. Available Sr–Nd isotopic data for Neogene to Quaternary lavas suggest a depleted asthenospheric mantle source for the magmas, which later interacted with lithospheric mantle (El Azzouzi *et al.*, 1999). A depleted mantle source was

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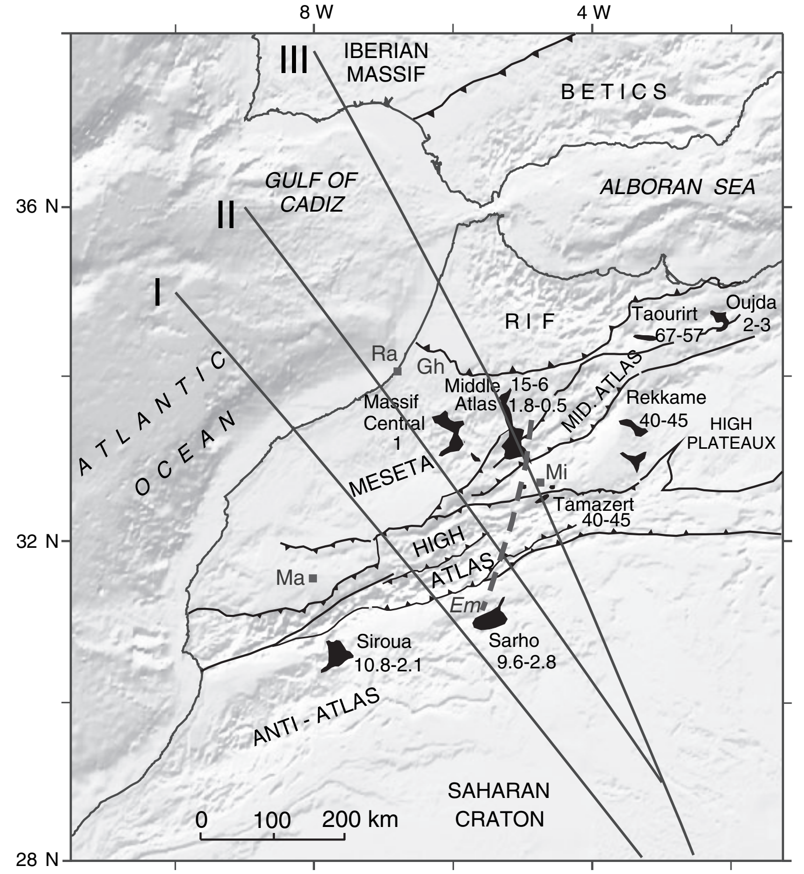

Fig. 1 Location map of the study area, showing topography, tectonic units and Cenozoic magmatism (black areas). Straight solid lines indicate the position of the modelled profiles of Fig. 2; dashed grey line marks the eastern limit of the mid Eocene marine limestone of the Atlas domain (*Em*). Shaded-relief from GTOPO30; absolute ages of magmatism after Tisserant *et al.* (1976), Harmand and Cantagrel (1984), Berrahma and Hernandez (1985), Berrahma *et al.* (1993), Rachdi *et al.* (1997), El Azzouzi *et al.* (1999), and Wagner *et al.* (2003). Localities: Ra: Rabat; Mi: Midelt; Ma: Marrakech; Gh: Gharb basin. Maps of volcanic ages of the Canary Islands, some 500 km to the SW of the Atlas coast, can be found in Anguita and Hernan (2000) and Carracedo *et al.* (2002). Detailed Bouguer and free-air gravity maps of the Atlas domain are presented in Ayarza *et al.* (2005).

also suggested on the same basis for the Eocene carbonatites (Mourtada *et al.*, 1997; Wagner *et al.*, 2003). This magmatism defines a separate district from that of the westernmost Mediterranean (Betics, Alboran, Rif), which shows a shift from calc-alkaline to alkaline character in Pliocene times (e.g. Duggen *et al.*, 2003, and references therein). On the contrary, the Atlas magmatism is roughly coeval and has the same petrological and isotopic signature as that of the Canary Islands, suggesting a common mantle reservoir (Anguita and Hernan, 2000).

Geophysical evidence

The knowledge of the crustal structure beneath the Atlas Mountains relies mainly on gravity and seismic refraction (Makris *et al.*, 1985; Tadili *et al.*, 1986; Wigger *et al.*, 1992; Ayarza *et al.*, 2005). The crust in the areas adjacent to the High and Middle Atlas is 33–36 km thick. A receiver function station near Midelt (Fig. 1) yielded punctual values of 36 and 39 km (Sandvol *et al.*, 1998; Van der Meijde *et al.*, 2003). The crust beneath the Middle Atlas has comparable thickness (*c.* 35 km; Wigger *et al.*, 1992). For the High Atlas, the published thicknesses are somewhat greater, in the range of 35–38 km. Localized maxima of up to 40 km are based only on gravity analyses (Tadili *et al.*, 1986; Ayarza *et al.*, 2005). These estimates indicate that the Atlas Mountains lack prominent crustal roots and therefore, their topography is only partially supported by the crust (Makris *et al.*, 1985), needing a buoyancy contribution from subcrustal levels.

Based on the analysis of teleseismic p-wave travel times, Seber *et al.* (1996) suggested that the lithosphere beneath the Atlas Mountains is thin or abnormally hot. Seismic tomography shows low velocities in the upper mantle beneath much of Europe and NW Africa (Hoernle *et al.*, 1995; Bijwaard and Spakman, 2000). Although the velocity structure of this zone is markedly heterogeneous, in the models the Atlas Mountains lie over a velocity low. Heat flow in Morocco varies from 40 to 100 mW m^{-2} (Rimi, 1999), although the Atlas Mountains lack a detailed sampling. On the basis of a punctual measurement of 86 mW m^{-2}, Ramdani (1998) suggested that the lithosphere beneath the southern High Atlas could be as thin as 50 km. In addition, the Atlas region is the locus of a strong positive geoid anomaly which, combined with a relatively weak negative Bouguer anomaly (Fig. 2), indicates that the excess of mass related to topography is compensated by a deep-seated mass deficiency. A recent integrated lithospheric model running from SW Iberia to the Saharan craton in a NW–SE direction reveals a prominent lithospheric thickening beneath the Gulf of Cadiz-Gharb basin region followed by a thinning beneath the Atlas Mountains (Zeyen *et al.*, 2005) thus supporting the interpretations inferred from tectonics, seismic tomography and potential fields.

Lithospheric modelling

Based on a 2D finite element algorithm developed by Zeyen and Fernàndez (1994) and Zeyen *et al.* (2005) we have modelled two regional lithospheric transects (profiles I and II), which together with profile III taken from Zeyen *et al.* (2005), have

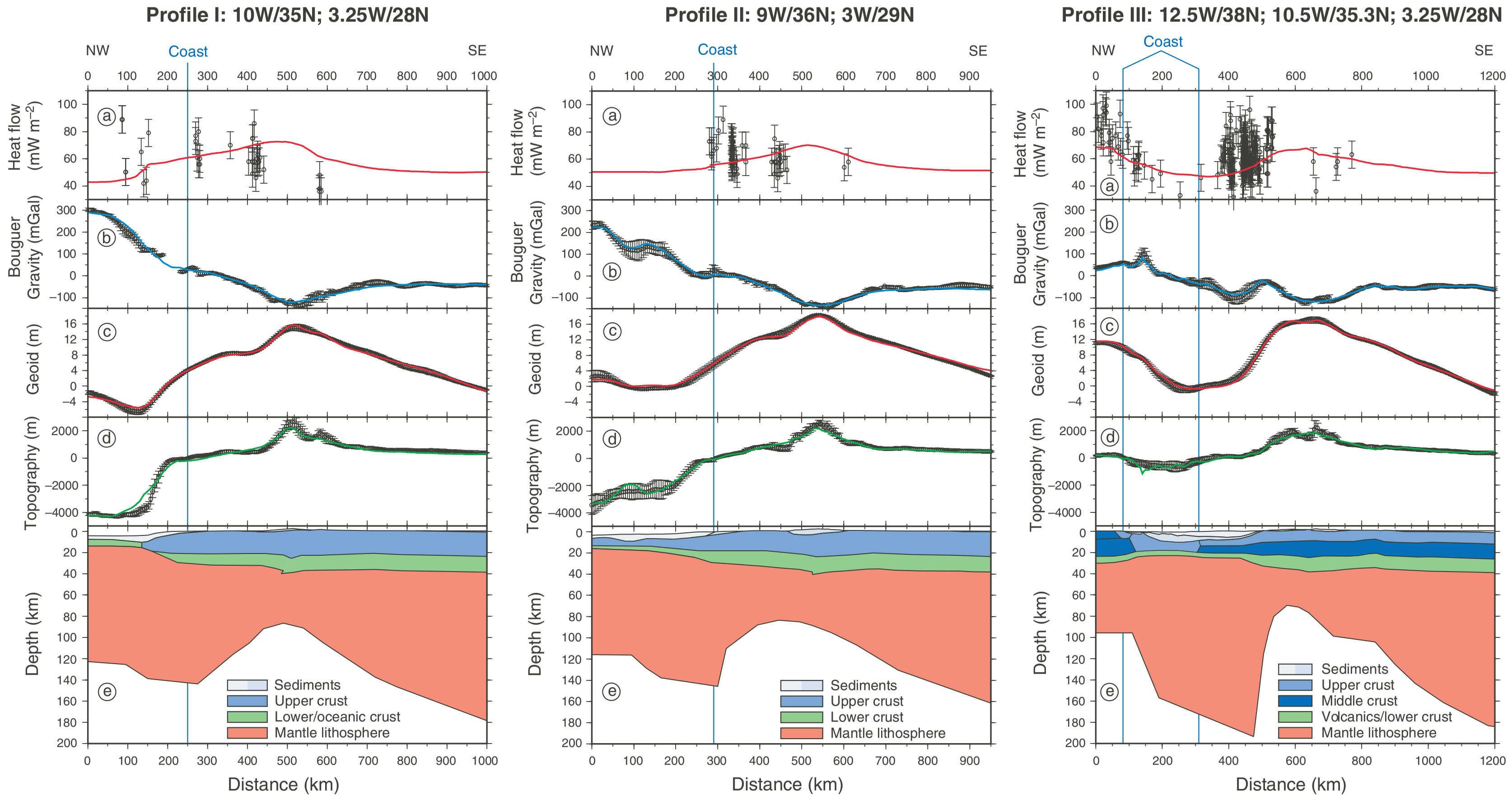

Fig. 2 Lithospheric models along the selected profiles. Panels (a), (b), (c) and (d) compare the observed values (open circles with error bars) to modelling results (continuous color lines). Panel (e) indicates the lithospheric structure. Profile III is after Zeyen *et al.* (2005) (see this publication for modelling details).

Table 1 Properties of the different bodies used in the lithospheric models

Description	Density (kg m^{-3})	Heat Production (μW m^{-3})	Thermal conductivity [W/(K*m)]
Neogene/ Quaternary sediments	2000–2200	2.0	2.0
Pre-Neogene sediments	2400–2650	2.0–2.5	2.4–2.5
Upper crust	2750	1.0	3.0
Lower crust	2930	0.2	2.1
Oceanic crust	2840	0.2	2.5
Lithospheric mantle	*	0.02	3.2

*Density of the lithospheric mantle is temperature dependent according to $\rho(T) = \rho_a \times [1 + 3.5 \times 10^{-5}(T - T_a)]$, where $\rho_a = 3200$ kg m^{-3} and $T_a = 1350$ °C are the density and temperature in the asthenosphere respectively.

allowed us to image the lithospheric structure in NW Africa (Fig. 2). The area of interest corresponds to the highest topography interval in the central part of the profiles, although we have extended the profiles to the Atlantic Ocean and the Saharan craton to avoid border effects and to model mantle depths. The modelling approach searches a good fit to heat flow, gravity, geoid and topography data and assumes conditions of thermal steady-state and local isostasy. The compensation depth is located at the deepest point reached by the lithosphere–asthenosphere boundary (LAB), which is considered to coincide with the 1350 °C isotherm. The density of the lithospheric mantle varies with temperature because of thermal expansion, whereas the area between the compensation depth and the actual base of the lithosphere is filled with asthenospheric material of constant density. Other physical properties for bodies forming the different crustal domains are listed in Table 1. For simplicity and due to lack of constraints, in profiles 1 and 2 we have considered a two-layer crust instead of three layers as used by Zeyen *et al.* (2005), although we kept a similar average crustal density so that the gravity response is the same and the mantle structure is unaffected.

Integration of the three models reveals a NE-trending, *c.* 400-km wide lithospheric thinning beneath the Atlas Mountains and surroundings, where the LAB reaches minimum depths of 70–80 km (Fig. 2). Towards the continental margin the three profiles show a noticeable lithospheric thickening which is most prominent in profile 3, where the LAB reaches 160–190 km depth (Zeyen *et al.*, 2005). In the oceanic domain, profiles 1 and 2 show a lithospheric thickness of 120–125 km in agreement with that expected for a Late Jurassic (*c.* 156 Ma) oceanic lithosphere. Towards the SE, the LAB deepens progressively until values of 160–180 km beneath the Saharan craton are in accordance with values expected for Precambrian terranes.

The depth of the LAB may change by about ± 10 km because of uncertainties in the crustal thickness and its average density. This is especially so in the continental margin, where the modelled lithospheric thickening could be substantially reduced by considering a thicker lower crust in the area. Nevertheless, the pattern of thinning in the Atlas region is a necessary result of our modelling. Actually, the LAB is mainly determined by the absolute elevation and the geoid height variations, whereas gravity better constrains the density distribution at crustal levels. Assuming that present topography was fully compensated by the crust would require crustal thicknesses beneath the Atlas Mountains of *c.* 45–47 km which, in addition to the tectonic shortening considerations, would not satisfy the observed gravity anomalies. According to our results, the crustal thickening in the Atlas Mountains (*c.* 5 km) explains about 50% of the topography whereas the other 50% is because of the buoyancy exerted by the asthenospheric upwelling (*c.* 60 km).

Discussion

The modelled asthenospheric high accounts not only for a topographical elevation exceeding that expected from the observed crustal shortening, but also for melting of depleted mantle and alkaline magmatism. Nepheline syenites and carbonatites, together with alkaline volcanic rocks, are traditionally associated to hot mantle plumes and rifts. However, the persistent state of compression of the NW African plate during the Cenozoic (Gomez *et al.*, 2000; Ait Brahim *et al.*, 2002) points against an extensional reason for lithospheric thinning. An inherited thinned structure is also unlikely because the Atlas lithosphere should be mostly thermally re-equilibrated from the Jurassic rifting episode. In addition, modest shortening in an intracontinental setting would not have favoured the formation of a large and heavy root prone to break-off or delaminate, as envisaged by Ramdani (1998). Therefore our modelled thinning can be viewed as the product of thermal erosion resulting from internal mantle dynamics. In fact, the low velocities detected in seismic tomography are roughly coincident with the locus of thinning. Hoernle *et al.* (1995) argued for hot upper mantle continuously underlying much of Europe and north Africa. Within this framework, our models document local, smaller-scale topography of the LAB that accounts for the particular geological features of the Atlas region. A clear age progression of volcanic centres is not observed. However, the width of the zone of lithospheric thinning argues against fracturing alone as a model to account for the volcanism. Rather than a simple hotspot, the Atlas appears as an elongate NE-trending upwelling, to a certain degree comparable with those described in other volcanic areas of W Africa as Cameroon (e.g. Meyers *et al.*, 1998; Marzoli *et al.*, 2000).

Cenozoic magmatism in the Atlas region has a complex spatial and temporal distribution, but it provides inferences about when and how thinning begun. A first stage of scattered magmatism is of early Tertiary age and appears concentrated in the East (Fig. 1). A comparable volcanic event is also identified in western and central Europe (Hoernle *et al.*, 1995; Michon and Merle, 2001), and in the Canary Islands (Anguita and Hernan, 2000). Notably, the location of the syenite-carbonatite intrusion at Tamazert and the volcanic outpours at Rekkame can be compared with the distribution of a mid-Eocene marine limestone that is the youngest available palaeoelevation indicator in the Atlas region. The limestone crops out in the western High and Middle Atlas (Fig. 1), and passes into more restricted water or terrestrial sediments to the east (Du Dresnay, 1988; Ben Brahim, 1994). Moreover, Schmidt (1992) presented geomorphic evidence for a differential subsidence in the foreland depression south of the High Atlas, with a relative high in the eastern segment that was actually experiencing erosion in mid-late Eocene times. These features suggest early uplift coinciding with the locus of magmatism in the eastern part of the Atlas region, thus documenting the earliest lithosphere thinning.

A second, more voluminous stage of volcanism started in mid-Miocene times, after a *c.* 30 Ma gap. Volcanism was more widespread (Fig. 1), and uplift progressed westward, raising the Eocene limestone to more than 1000 m a.s.l. in areas located outside of the Atlas deformed belts. This volcanism can be correlated in time with an alkaline volcanic surge of the same age in Europe (e.g. Michon and Merle, 2001) and in the Canaries (Anguita and Hernan, 2000), and appears clearly associated to the present-day area of lithospheric thinning defined by our modelling (possibly larger than the Eocene precursor). This second stage coexists with intense compression in the Atlas, which prevented the area to fail in extension, and attests for a robust lithospheric thinning process dominating the competition against thickening resulting from horizontal shortening.

With the available knowledge we cannot ascertain whether the Atlas upwelling relates to a deep-seated mantle plume, reflects only upper mantle flow, or a combination of both. Based on tomography, Goes *et al.* (1999) suggested a common lower mantle root for the European and Canarian volcanism, although the low amplitude of the velocity anomalies (=0.55%) is not conclusive. In any case, the Atlas and Canaries would represent distinct uppermost mantle features. Cenozoic European volcanism was spatially associated to rifting, but this is not the case in Morocco. As to the first phase of thinning and magmatism, we may speculate that erosive upwelling was probably induced by lateral flow of hot mantle at the SW tip of the Eoalpine subduction zone, which extended from the western Alps to the SE of Iberia (Michard *et al.*, 2002). The second stage of volcanism coexists with the accretion of the Rif wedge onto the African plate and westward roll-back in that area (Frizon de Lamotte *et al.*, 2000; Michard *et al.*, 2002, etc.), and then the upwelling may have also been guided by side-effects of the thickened lithosphere in that region. We thus propose that in NW Africa the regional tectonic scenario explains the spatial location of local lithosphere thinning, by focusing uppermost mantle flow. Flow may have been guided by impingement of the lithospheric roots at subduction zones, as envisaged for other areas of the Alpine belt (Merle and Michon, 2001; Faccenna *et al.*, 2004). However, the temporal coincidence of magmatic gaps and the Neogene magmatic reactivation across remote areas in Europe and Africa suggests that the timing of transcontinental magmatic surges was related to upwelling pulses from the deep mantle, independent of the regional tectonic setting.

Conclusions

Potential-field modelling of the NW African lithosphere defines the geometry of a NE-trending, 400 km wide asthenosphere high beneath the Moroccan Atlas intraplate belts. The lithosphere thins from 140–190 km in north Morocco and 160–180 km in the Saharan craton to *c.* 80 km in the Atlas system. This implies density compensation within the mantle, and explains singular features of the Atlas region as high topography compared with modest crustal thickening, the occurrence of alkaline magmatism synchronic to compression, and a scarce foreland basin record. The lithosphere structure deduced is, as yet, unique in an intracontinental compressional environment, like NW Africa during the Cenozoic. We interpret it as the result of a mantle upwelling, probably with a deep root, but focused by upper mantle flow influenced by the thickening of the lithosphere at the adjacent Iberia–Africa plate boundary.

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