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Preface

The geology of vertical movements of the lithosphere: An overview

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Defining and quantifying horizontal movements of the Earth's crust has been a common activity for geoscientists during the last decades due to the success of structural geology in measuring magnitudes of shortening or extension via balanced cross-sections and strain analysis, coupled with the popularization of continental drift and plate tectonics theory on the larger scale. On the other hand, vertical movements of rocks or of the surface of the Earth, although perceived for over two centuries, have been more elusive, because of the smaller magnitude of translations, the difficulties in finding unambiguous markers, and because more often their characterization requires the concurrence of integrated approaches, some of them recently developed. Downward movements, i.e., subsidence, associated with the formation of sedimentary basins are better quantified and understood but often not coupled with the areas experiencing upward vertical movements.

Over the past decades, considerable progress has been made in the understanding and quantification of vertical motions of the lithosphere and, in particular, of the Earth's surface, thanks to the implementation of new tectonic, geomorphologic, geochemical and geophysical techniques, largely motivated by the strong implications of uplift and subsidence for landscape and environmental issues. In addition, there is an increased perception that areas experiencing upward and downward movements (mountains and basins) should be looked at as parts of a single tectonic, geomorphic and sedimentologic system. A significant number of challenges remain to be addressed. An inspection of the recent literature clearly indicates that the geology of vertical movements is a field of very active multidisciplinary research. The ESF-sponsored TOPOEUROPE initiative (with its national counterparts TopoIberia, TopoScandia, etc,) is an obvious demonstration of the increased attention of the geoscience community towards processes controlling the development of continental topography.

With this in mind, on 28-31 September 2007 we convened a workshop entitled “The geology of vertical movements: uplift and subsidence, mountains and basins” in the framework of the International Lithosphere Program's task force on Sedimentary Basins. The workshop was hosted by the 1st Convention of the Moroccan Association of Petroleum Geologists (MAPG), and the venue was the town of Marrakech, in the frame of the superb Moroccan mountains and basins (e.g., Fig. 1), which have been the recent objective of many studies. Thus, the workshop intended a particular emphasis on African geology. A total of 81 oral and poster communications were presented; this special issue of Tectonophysics is an outcome of the Marrakech conference. The objectives of this introductory article are to review the causal mechanisms and methods of study of vertical movements of the lithosphere, including a highlight of the relevance of the geology of Morocco for the subject, and to briefly introduce the papers in the special volume.

1. Basic concepts

Broad, long-wavelength vertical movements of the lithosphere, usually referred to as “epeirogenic” movements in early times, have long challenged geologists (see an extensive review in Şengör, 2003). We all recognize that Earth landscape and outcrop geology are the product of the interactions between tectonic processes of uplift and subsidence on the one side and surface processes of erosion and sedimentation on the other. Lithosphere dynamics, development and destruction of mountains, sinking and infilling of sedimentary basins are increasingly looked at as interlinked phenomena, addressed in the research fields of basin analysis and tectonic geomorphology. Vertical movements result in sedimentary basins that host important hydrocarbon and water resources, and in mountain belts and high plateaux that dramatically influence Earth environments and climate.

Vertical movements of rock particles or the Earth's surface are usually defined relative to an appropriate marker, i.e., the sea level or the geoid. Since the publication of the benchmark paper of England and Molnar (1990), it became of common knowledge that vertical movements of the crust and the lithosphere are translated into actual movements of the surface depending on the amount of erosion or sediment aggradation (a relationship simply expressed as surface uplift = bedrock uplift – exhumation, where subsidence would be defined as uplift of negative sign and exhumation is the movement of rocks relative to the surface, and as such may be substituted by burial in the equation; the equation could then be rewritten as surface subsidence = bedrock subsidence - burial). Further discussion on the implications of this well-known relationship can be found in the textbooks of Burbank and Anderson (2001) and Bull (2007). True surface uplift can only be considered as such at the regional scale, referring to changes of the mean elevation of a region. Two components should at least conceptually be distinguished when considering bedrock (or “crustal”) uplift: the component arising from tectonic forcing (“tectonic uplift”), and the isostatic rebound resulting

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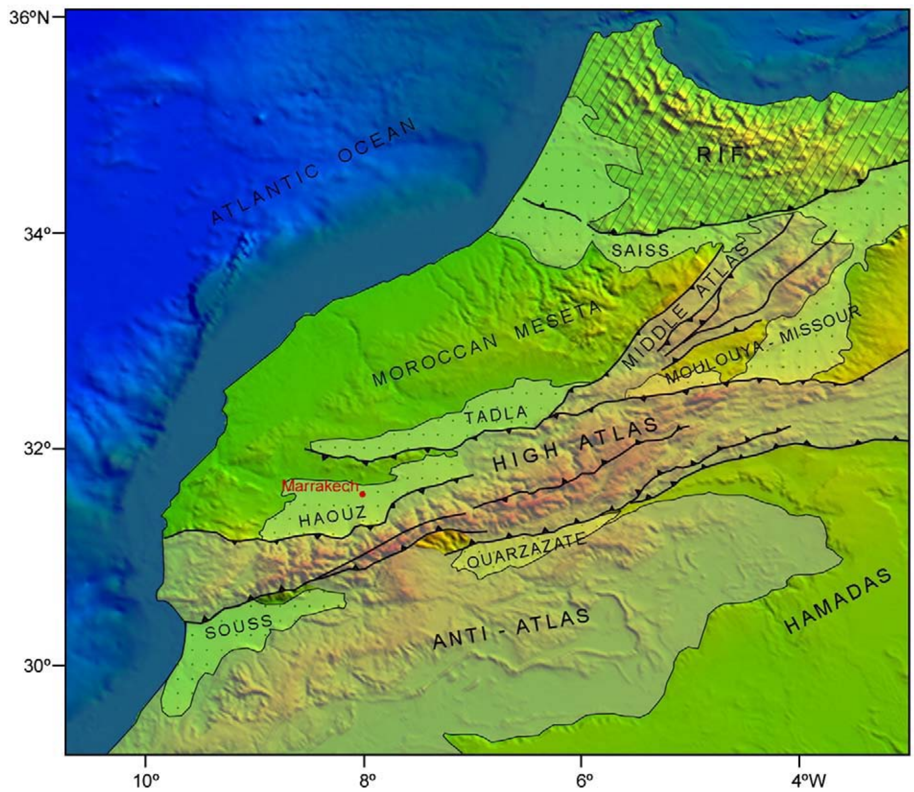

Fig. 1. Digital topography and tectonic sketch map of Morocco, showing the main mountain and basin systems. Oblique lines indicate the Rifean plate-boundary orogenic belt, the grayed areas represent the Atlas chains, and the extent of the Cenozoic basins is indicated by the dotted pattern.

from erosional unloading ("isostatic uplift"). As isostatic rebound will never increase the mean elevation of a region, it follows that positive uplift of the surface is solely the result of tectonic processes, and as such is one of the main goals in tectonic studies of mountain regions. Similarly, the subsidence of the surface or of sedimentary rocks below the sea level or a regional base level requires some sort of tectonic forcing which will be amplified by sediment loading.

It is increasingly recognized that tectonic forcing and surface processes are interconnected phenomena, linked by a series of interactions and feedbacks. Tectonic uplift raises topography and by this enhances denudation rates, which tend to lower the mean elevation. Erosional unloading not only drives isostatic rebound, but also the associated redistribution of mass influences the tectonic development, which may respond to maintain equilibrium wedge dynamics (Davis et al., 1983; Willet, 1999). Redistribution of mass driven by interactions between tectonic and erosional processes is of great importance also in sedimentary basins, in particular in extensional settings. Here, the transfer of material eroded from the rift shoulder and deposited in the rift axis cause change in the lithostatic pressure in the crust, horizontal flow of lower crustal material and, thereby, can be a source of vertical movements (e.g., Burov and Poliakov, 2003). As erosion rates increase with growing topography, the interaction between uplift and erosion can drive the topography of an active mountain region towards a steady state (Willet and Brandon, 2002). However, disequilibrium states and transient landforms such as migrating knick-points in river profiles are particularly interesting as they indicate non-linear tectonic histories including specific uplift events or changes in base level (e.g., Wobus et al., 2006). Thick accumulations of homogeneous sedimentary facies are indicative of steady states in basins, where subsidence and compaction are balanced by sedimentation.

There are also notable feedbacks between these processes and climate: tectonic uplift of mountains and plateaux influences climate at a local scale, increasing precipitation and opening the possibility of glaciation, which in turn enhance rates of erosion and mountain range destruction (but also increase isostatic rebound and the local altitude of summits). The formation of mountains at high angle to the trend of dominant winds can cause the development of rain shadows thereby focusing precipitations and creating a highly humid area contrasting with the arid conditions of the leeward side of the orogen. In turn the development of a rain shadow impacts the way crustal material is accreted to the orogen (e.g., Koons, 1990). It has been argued that tectonic uplift can also influence climate at the global scale, exemplified by the widespread development of mountain belts and climate variations during the late Cenozoic, although the directionality relation between tectonics, erosion and climate during this epoch is still a matter of debate.

2. Causes of uplift and subsidence

In essence, vertical movements of the lithosphere result from vertical loads (changes in the weight of the lithospheric column) or horizontal loads (associated to a state of horizontal compression), both not being mutually exclusive. Mechanisms that alter the weight of the column and move up and down the rocks and the Earth surface fall in one of these categories: crustal deformation, mass loading (or unloading), and thermal changes in the lithosphere or sublithospheric mantle. Similarly to the case of the horizontal movements of the tectonic plates, where gravity converts the internal energy of the Earth into plate driving forces, gravity is also crucial in the understanding of vertical movements, which are ultimately governed by the buoyancy and rigidity of the upper layers of the Earth.

Uplift occurs at different spatial scales. Short-wavelength (10^2-10^4 m) uplift is usually related to active or growing structures (such as fault-related mountain fronts, etc.), and tends to be transient and superimposed on a more general regional uplift. The resulting topography is supported by the strength of the crust. It is the regional or long-wavelength uplift (tens to hundreds of km) that obeys to entire crustal or lithospheric-scale processes and is the focus of the following discussion.

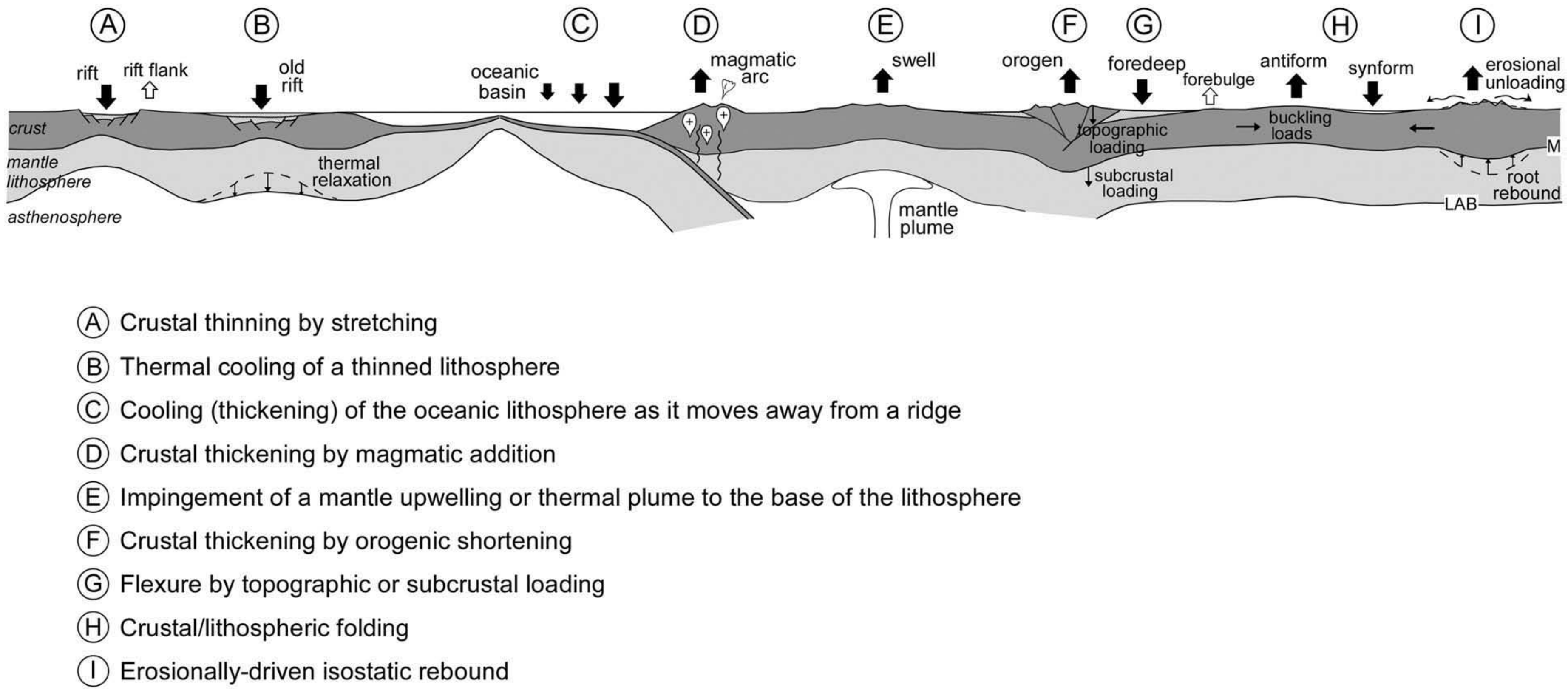

Fig. 2. Vertical movements of the lithosphere: mechanisms of subsidence and uplift at the scale of the tectonic plates (see text for discussion). M: base of the crust; LAB: lithosphere-asthenosphere boundary. Vertical and horizontal scales are not the same, crust and lithosphere thickness not to scale.

The cartoon of Fig. 2 shows the diverse mechanisms that produce large-scale uplift or subsidence at the lithospheric scale. Vertical movements occur at plate boundary regions, in close association with plate kinematics, and also in intraplate regions, partly independently from the interplate tectonic processes. In the figure, areas marked with A, B, C, G and H experience subsidence, leading to the development of sedimentary basins, whereas regions D, E, F, H and I illustrate the main mechanisms of uplift in mountains and plateaux.

Lithospheric stretching in rift zones thins the crust, causing the surface to subside creating extensional basins (McKenzie, 1978) (A in Fig. 2). Normal faults guide extension and sediment accommodation space in the brittle upper crust, whereas the lower crust may accommodate necking by ductile mechanisms. It is the replacement of crustal material by denser upper mantle which imparts negative buoyancy and negative uplift to the surface. However, away from the tectonically extended area in the crust, rift flanks may experience subsidiary uplift in response to larger wavelength lithospheric necking and to regional isostatic compensation. After the initial stretching episodes, rift zones may experience slow, thermal subsidence as the thinned lithosphere relaxes and recovers its original thickness (B in Fig. 2). Because of its larger wavelength, the perturbation of the thermal boundary at the base of the lithosphere imparts subsidence over an area wider than the originally stretched crust, producing the characteristic geometry where expansive post-rift sediments overlap the rift flanks (White and McKenzie, 1988). A similar mechanism produces deepening of the ocean basins away from mid-ocean ridges, as the newly created lithosphere cools and subsides while drifting away from the ridge (C in Fig. 2). In such a frame, the depth of the ocean floor can be expressed as a direct function of the distance from the ridge, that is, of the lithosphere age (Parsons and Sclater, 1977; Stein and Stein, 1992).

Thickening of the crust uplifts topography, by addition of differentiated igneous bodies derived from the mantle (like it happens above subduction zones, Fig. 2 D, or in other magmatic provinces), or as a direct consequence of horizontal shortening in a compressional orogen. The density of the lighter crustal material compared with the mantle is of major relevance in determining the magnitude of surface uplift and the thickness of the crustal root necessary to achieve isostatic equilibrium (e.g., Fischer, 2002).

The entire lithosphere is flexed by the load of a large mountain belt, a subsidence mechanism that creates foreland basins adjacent to orogens (Fig. 2 G) (Beaumont, 1981; DeCelles and Giles, 1996). Both the loads of the excess mountain topography and subcrustal loads related to a dense lithosphere root may in principle contribute to the total subsidence. The lithosphere also responds by flexure to other imposed loads as large volcanic edifices or ice caps. Flexural subsidence is strongly controlled by the lithospheric rigidity, which governs the size and actual depth of the resulting depressed area. On the other hand, flexure of an elastic plate may bend upwards the distal foredeep to create a forebulge (Fig. 2 G).

Impingement of a mantle plume or any sort of thermal upwelling bends up the surface by thermally eroding the lower lithosphere and imparting positive buoyancy (Fig. 2 E). This is the mechanism responsible for the typical hotspot uplift, and is the most characteristic source of vertical movement in plate interiors (Crough, 1983; Sleep, 1990). In a way, the buoyancy mechanism is analogous to the replacement of dense lithosphere by hot asthenospheric material via lithospheric detachment or delamination in an overthickened zone from a compressional boundary (Kay and Mahlburg Kay, 1993).

As mantle plumes are part of the mantle convection system, the resulting elevation is sometimes referred to as dynamic topography, because the buoyancy sources are moving (Lithgow-Bertelloni and Gurnis, 1997). However, at a scale of hundreds of kilometres and at a given increment of time, the system can be viewed as isostatically compensated at the level of the sublithospheric mantle. Conversely, subduction zones that represent downwellings in the mantle convection system are sites of negative dynamic topography.

The importance of horizontal loading of the crust and lithosphere in producing large wavelength folding and creating areas of uplift and subsidence has been demonstrated by several numerical and analogue models, and is increasingly recognized in nature. (Fig. 2 H) (Cloetingh et al., 1999). In most cases, folding is assisted by areas of more localized deformation (faults), thus creating a complex pattern of vertical movements. Applying these models to real world situations is less straightforward as the lithosphere is not homogeneous everywhere, and is rather characterized, for instance, by the presence of inherited zones of weakness which can localize deformation and cause changes in the thickness of crust and/or lithospheric mantle.

As a mountain range or plateau is eroded, it rebounds to a certain degree to compensate for the mass unloading (Fig. 2 I). We recognized that the process does not lead to positive surface uplift, although it is clear that at the local scale, summits or interfluves where erosion is less efficient may gain altitude during the process. This must be considered as bedrock or crustal uplift, because the mean elevation of an eroding, inactive mountain range decreases with time: isostatic rebound does not compensate for the entire mass loss but only for a fraction of it, which is a function of the density ratio between crust and mantle.

Often the uplift or subsidence history of a given region cannot be simply explained by one single mechanism. They may either succeed in time, occasionally with opposite effects (e.g., the mountain belts resulting from inversion of extensional basins), or may act in combination at a given interval of time. Documented cases of the latter include the central Andes, where crustal thickening leading to uplift is a combination of tectonic shortening and magmatic addition (e.g., Allmendinger et al., 1997; Haschke and Günther, 2003), the Aegean subduction complex, where negative dynamic topography adds to back-arc crustal thinning (Husson, 2006), or, close to the conference venue, the Moroccan High Atlas, where a combination of crustal shortening and a thermal mantle upwelling appear to be responsible for the total topography (e.g., Teixell et al., 2003, 2005; Zeyen et al., 2005; Missenard et al., 2006).

3. Measuring vertical movements

Measuring vertical movements of the lithosphere, typically an order of magnitude smaller than horizontal displacements, is being appreciated as an important task in tectonic studies to complete the full understanding of the plate kinematics picture (Willet et al., 2001).

A major objective of such studies consists in determining the timing and rate of vertical movements. The measure of surface uplift or subsidence rates may be fundamental in order to understand the deep tectonic mechanisms operating (e.g., distinguishing between crustal thickening and lithospheric root removal in mountain systems, Garzione et al., 2006; or between subsidence drivers in sedimentary basins). While in sedimentary basins the resolution of vertical movements may be high due to the presence of sediments, markers of upward movements in the continental domain are of very diverse kind but often rare or poorly accurate. Many of the available tools require an analysis of the stratigraphic record or geomorphic history combined with geochemical or geophysical techniques.

The popularization of digital topography in the past decade is both an instrument and a reflection of the increasing interest in the relief of the Earth surface and its quantification. As a testimony, we can think about the shift evident from many recent tectonic papers, in which the first of all illustrations is a digital elevation model, where we used to have a geological map. A major issue, beyond pure illustration, is how good we are at inferring surface processes or uplift histories from the analysis of topography. DEM's can be useful to calculate mean elevations of wide regions or to estimate eroded volumes and erosionally-driven rock uplift (e.g., Small and Anderson, 1998). Another important application consists in extracting river profiles (e.g., Wobus et al., 2006). Once the effects of the drainage area and lithological parameters (bedrock strength) have been eliminated, river steepness may be a proxy for uplift. In general, there is a direct relationship between steepness and uplift rate, where gradient breaks (knick-points) may indicate active fault zones or propagating waves of erosion related to recent uplift events. However, other factors as channel morphology, erosion parameters or climate, whose influence remains difficult to account for, allow arriving only to semi-qualitative results.

Vertical movements can be inferred from stratigraphy and burial geothermometry in basins, and from paleoaltimetry, geodesy, thermochronology and landform dating in mountain regions. In sedimentary basins, geohistory analysis is routinely applied to evaluate vertical motions (including subsidence and uplift) of a particular stratigraphic interval from a column, based on stratigraphic thickness, age and lithology of horizons, and paleobathymetry in marine successions (van Hinte, 1978). With this, detailed histories of vertical movements can be retrieved. Thermal data may be included if the maturation history is among the objectives. Paleothermometry is also useful in the analysis of eroded basins: low-temperature geothermometers as vitrinite reflectance, illite crystallinity index, kerogen alteration and level of low-grade metamorphism can provide data on burial and subsidence. These indicators need however a knowledge of the geothermal gradient to be converted into depths, and may be altered by intrabasinal fluid flow or magmatic underplating in deep basins.

A number of techniques has been recently developed to estimate ancient elevations and to infer uplift (paleoaltimetry). Paleoelevation indicators currently used include 1) paleobotany (e.g., physiognomic characteristics of floras in function if altitude, or stomatal index in leaves as proxies for PCO_2; Forest et al., 1999; McElwain, 2004), which may have limitations concerning paleoclimate estimations and atmospheric physics, 2) stable isotope analysis (i.e., based on the empirical relationship between oxygen isotope composition and altitude, which can be recorded in terrestrial carbonates and paleosoils; Poage and Chamberlain, 2001; Rowley and Garzione, 2007); 3) analysis of atmospheric pressure proxies (e.g., size and shape of gas vesicles in basaltic flows; Sahagian and Proussevitch, 2007). Although they all have given valuable results, the uncertainty of these three methods may be large. Finally, 4) dated uplifted marine deposits or terraces may constitute more precise markers for surface uplift, provided that they have not been dissected by erosion. In fact, all the described tools inform about the values of uplift of the rocks that contain them. To convert them into values of real surface uplift needs careful consideration of their regional significance and of the amount of isostatic uplift by erosional unloading (England and Molnar, 1990).

Geodetic techniques as repeated levelling, GPS tracking and radar interferometry are widely used to monitor changes in elevation or ground deformation. GPS surveys yield short-term results and, as the other methods, only local bedrock uplift data. Radar interferometry may inform about wider regions, but it is used in the assessment of seismic, volcanic or landslide hazards rather than in long-term uplift studies.

Thermochronological methods are also widely used in uplift studies, while in the best cases they inform about exhumation of rocks. Fission track and (U-Th)/He techniques are popular because they inform about the low-temperature spectra, i.e., while they rock voyages through the upper crust and paleotemperature is more likely to reflect proximity to the Earth surface (Gleadow and Brown, 2000; Ehlers and Farley, 2003; Reiners and Brandon, 2006). With the combined use of different thermochronometers, steps in the cooling history of a sample can be reconstructed. In good quality samples, the fission track length distribution within crystals can be used to model a dated cooling path, in the range from 110 ºC to 60 ºC in the case of apatite. The obtained cooling histories are used to calculate exhumation rates, on the basis of an assumed geothermal gradient. Sampling along a vertical profile of a mountain slope or a well is also used to plot fission-track or (U-Th)/He age vs. elevation, and extract exhumation rates. Thermochronologically-derived rapid exhumation rates are frequently correlated with pulses of tectonic uplift. This may be valid in certain cases but the correlation should be submitted to a critical exam because: 1) we know little about past geothermal gradients and their variation with time; even in the upper crust, cooling ages and paths may be influenced by non-exhumative processes as post-rift or post-magmatic cooling (e.g., Bertotti et al., 1999), and 2) in the cases where the methods safely inform about exhumation rates, there is a component of denudation induced by isostatic rebound, which may be enhanced by climate factors.

At the Earth surface, exposure dating with cosmogenic nuclides has proven to be of great importance to calculate timing and rates of short-term erosion and landform development (Zreda and Phillips, 2000; Cockburn and Summerfield, 2004; Walker, 2005). As only a few cm of the exposed bedrock or boulder surface are valid for the analysis, the results are likely to be altered by processes of weathering, erosion or re-deposition. The half-life of the unstable cosmogenic isotopes makes the technique in most cases valid only for scales of 10^5 years or a few million years, although only very rarely landforms have been kept pristine over these time scales anyway. Thus, the technique is essentially used in Quaternary geology studies. As for the study of vertical movements, cosmogenic nuclide dating has been applied to stepped fluvial terraces, and from this river incision rates have been calculated (e.g., Repka et al., 1997; Arboleya et al., 2008). Incision rates can only be converted into actual bedrock uplift rates provided that incision is not induced by a base-level drop and the rivers' longitudinal profiles are in steady-state. Alternatively, in continuously eroding areas, cosmogenic nuclides are used not to resolve specific exposure ages, but to quantify rates of denudation, either based on sampling of eroding surfaces (punctual data) or river sediments (spatially averaged) (Cockburn and Summerfield, 2004; von Blanckenburg, 2005).

4. Cenozoic vertical movements in the Atlas mountains and basins of Morocco

The Atlas belts of Morocco constitute a good natural laboratory to examine the causes and effects of vertical movements of the lithosphere, as could be appreciated by the many presentations and field excursions developed in the frame of the ILP Marrakech meeting. The Atlas Mountains formed during the Cenozoic in the interior of the African plate but in close association to the Europe-Africa plate convergence (Mattauer et al., 1977). The history of vertical movements of the Atlas system is long and complex and results from a combination of crustal and mantle processes (e.g., Frizon de Lamotte et al., this volume).

The Atlas mountain belts consist of two thrust- fold chains (the NE-tending Middle Atlas and the ENE-trending High Atlas) and a wide arch with modest post-Paleozoic deformation (the Anti-Atlas) (Fig. 1). Adjacent to the High Atlas is a discontinuous system of foreland basins filled with Cenozoic sediments (the so-called Souss, Ouarzazate, Tadla-Haouz and Moulouya-Misour basins) (Fig. 1). The High and Middle Atlas thrust-fold belts were created by the tectonic inversion of pre-existing extensional basins of Triassic-Jurassic age (Mattauer et al., 1977; El Kochri and Chorowicz, 1995; Gomez et al., 1998; Frizon de Lamotte et al., 2000; Piqué et al., 2002; Teixell et al., 2003, Arboleya et al., 2004; Laville et al., 2004; Frizon de Lamotte et al., 2008). The structure of the Atlas belts is relatively simple, and the amount of orogenic shortening is moderate (<25% in the High Atlas, <10% in the Middle Alas). Cenozoic shortening is very small in the Anti-Atlas, whose post-Paleozoic structure conforms to a large (100 km-scale or lithospheric-scale) domal fold cut by minor faults. As a compressional belt, the High Atlas exhibits a small crustal root that contributes to the mountain topography, as revealed by gravity and low-resolution seismic surveys (Tadili et al., 1986; Wigger et al., 1992; Ayarza et al., 2005). However, the total elevation of the Atlas system (where the mean altitude of the mountain belts exceeds 2000 m over large areas and some of the undeformed foreland basins stand above 1200 m) is only partially explained by shortening and crustal thickening, being the system in a state of isostatic under-compensation at the crustal scale (Van den Bosch, 1971; Makris et al., 1985, Teixell et al., 2003; Ayarza et al., 2005).

The causes for the remaining elevation of the Atlas have been sought at the level of the mantle. The modeling of topography combined with gravity and geoid data has suggested that at least half of the total altitude must be attributed to a pronounced lithospheric thinning beneath the Atlas mountains and basins which imparts extra buoyancy (Zeyen et al., 2005; Teixell et al., 2005; Missenard et al., 2006). The domain affected by this lithospheric thinning forms a NE-SW strip following the Middle Atlas but cutting through the High and Anti Atlas. This phenomenon explains the high altitude and poor preservation of foreland basins, and is consistent with the occurrence of alkaline magmatism of Cenozoic age. The reasons why this thinning occurred in an area that has been subject to compression during much of the Cenozoic remains unclear, but it has been attributed to a thermal upwelling or plume tip more or less independent from the regional tectonic setting.

A major challenge in the tectonic understanding of the Atlas mountains and basins has been the relative timing and rates of compressional mountain building and thermal uplift. There is little consensus as regards to the chronology of shortening and uplift, partly because of the difficulties un finding appropriate markers and partly because of the discontinuous and poorly dated sedimentary record of mountain building.

In an attempt of synthesis, we can conclude that most works devoted to the timing of orogenic deformation place the Atlas orogeny in the Cenozoic, with the time bracket from the mid Eocene to the Quaternary (Görler et al., 1988; Fraissinet et al., 1988; Morel et al., 2000; Frizon de Lamotte et al., 2000; Teson and Teixell, 2008). The attributions are based on tectonics-sedimentation relationships at the mountain fronts and adjacent foreland basins (notably the Ouarzazate basin, in the southern side of the High Atlas). As expected from the relatively simple structure, shortening rate estimates after balanced cross-sections are very low (~0.3 mm/yr averaged for the southern High Atlas thrust belt after Teson and Teixell, 2008, and probably not exceeding 1 mm/yr for the Atlas system as a whole).

The timing of long-wavelength uplift related to the mantle structure is still more elusive. It has been indirectly inferred from the chronology of associated magmatic events (last 15 My for the later phase; Teixell et al., 2005; Missenard et al., 2006), or more recently from stratigraphic paleoelevation/paleohorizontality markers (last 5 My at ~0.2 mm/yr; Babault et al., 2008). Recent low-temperature thermochronological investigations have shown that Cenozoic exhumation in the Atlas Mountains and adjoining regions is limited: pre-Tertiary apatite fission track ages are recorded almost everywhere except for a narrow upthrust block with Miocene ages in the axial part of the High Atlas of Marrakech. (Barbero et al., 2007; Ghorbal et al., 2008; Missenard et al., 2008; Balestrieri et al., 2009).

In conclusion, the Atlas mountains and basins of NW Africa provide a good example of how different geodynamic processes can combine and contribute to uplift in nature, and also constitute a good illustration of the challenges posed by mountain and basin systems in terms of timing and rate of vertical movements, and of evolution of surface processes and form of relief. A significant amount of ongoing research in the region lets us expect that exciting new insights will be gained in the future.

5. Contents of this issue

The special issue contains twelve original or review papers that we present below following a geographic order, from Maghreb, which was the main target of the meeting, to the rest of the world. Seven papers concern directly continental or off-shore Maghrebian regions. Three others deal with the European crust or lithosphere and the last two with America.

An overview of the Mesozoic and Cenozoic vertical movements at the scale of the whole Maghreb (Morocco, Algeria and Tunisia) is presented by **Frizon de Lamotte et al.** The authors emphasize the present asymmetry of the Maghreb topography, with highest elevation in Morocco than in Tunisia, and show that this configuration is inherited from a long history in which thermal processes play a major role. Presenting the results of a fission-track study on Variscan granites from the Western Meseta (Morocco), **Sadiqqi et al.** show that this domain, situated in the core of the Atlas system, experienced since the Permian a succession of up and down movements. The fact that AFT ages are not the same in Rehamna and Jebilet adjacent massifs is interpreted as a consequence of differences in age and depth of crystallization of the sampled granites. **Khomsi et al.** use surface and subsurface data to analyze the Cenozoic evolution of eastern Tunisia. They emphasize the existence of a Mid to Late Eocene compressive event followed by erosion and then by a strong subsidence during the Oligocene-Early Miocene, which is interpreted as the result of lithosphere flexure in front of the Maghrebide system. Following the authors, this tectonic agenda should be relevant at the scale of the whole Maghreb.

To explain the apparent contradiction between seismological and geological observations of the Al Hoceima (Morocco, Rif) 2004 earthquake ($M = 6.3$), **Galindo et al.** propose a decoupled tectonic model with a crustal detachment separating a deep brittle crust from shallow levels undergoing uplift, folding and normal faulting. **Iribarren et al.** present a quantification of sedimentary volumes supplied during the Neogene to the basins fringing or superimposed to the Betic-Rif orogen. These authors show that spatial and time variability of the sediment supply from the Betic–Rif orogen to basins is closely related to the morphologic and tectonic evolution of the system.

Based on a compilation of existing studies, **Gutscher et al.** present depth to basement, sediment thickness, depth to Moho and crustal thickness maps of the Gulf of Cadiz accretionary prism, west of the Gibraltar Arc. Using morphological arguments, compared with the results of analog models, the authors propose that the accretionary wedge and underlying subduction system are still active at the moment, and discuss the implications for the 1755 Lisbon earthquake. Thanks to a new marine data base (MARADJA 2003 cruise) the offshore part of the Algiers margin as well as a new tectonic framework is presented by **Yelles et al.** These authors show that the main south-dipping offshore faults are the driving fault system, as also found further east in the Boumerdès (M 6.8) 2003 rupture zone. They suggest that this fault system marks an inception of subduction of the Algerian basin beneath Africa plate.

In their contribution, **Bertotti and Mosca** use the architecture of the sedimentary basins inside the arc of the Western Alps to infer patterns of subsidence and coeval uplift in the mountain belt. Differential paleostress estimates from mechanically-induced calcite twins are used by **Lacombe et al.** to constrain paleo-burial and subsequent uplift by folding in the Ionian Zone the outer Albanides (Albania). Using this methodology a ~4 km paleo-burial of Cretaceous limestones from the Saranda anticline is estimated, just before they were uplifted by folding during the Cenozoic. **Krzywiec** uses high-quality seismic reflection data to constrain vertical movements along the Teisseyre–Tornquist Zone, one of the most fundamental lithosphere boundaries in Europe. The author focuses, in particular, on problems of Late Carboniferous and Late Cretaceous–Paleogene basin inversion and uplift. In order to test the hypothesis of an asthenospheric diapir explaining Cenozoic uplift of Norway, **Pascal and Olesen** present an integrated gravity and thermal model for the region. The results permit to exclude this hypothesis but do not rule out the possibility that thermal processes in the lower mantle could have contributed to the uplift.

Roure et al. show that after the Late Cretaceous-Paleocene development of the east-verging Sierra Madre Oriental thrust belt (Mexico), it is necessary to advocate mantle dynamics to explain post-orogenic uplift and regional tilting of the basement. Numerous paleo-thermo-meters (Tmax, Ro), paleo-thermo-barometers (fluid inclusions), PVT and coupled forward kinematic and thermal modeling have been used to constrain this scenario. Finally, **Feinstein et al.** analyse a 1.15 km deep apatite fission track (AFT) thermochronology profile from the southwestern Canadian Shield. The authors show two Phanerozoic heating and cooling episodes indicating significant Phanerozoic heat flow variations. The periods of variations in geothermal gradient are coeval with vertical movements that are tentatively attributed to far-field effects of orogenic processes occurring at the plate margin.

6. Further directions

The geology of vertical movements of the lithosphere will continue to be an area of very active research in the years to come. In short term, we need to expand the global database of timing and rate of surface uplift, erosion and landform development to complete the picture from different mountain and basin systems of the world. Sedimentary basins are quite well studied and depositional processes well known, while significant progress is being made to understand the rise and destruction of mountain topography. Time is mature for integrated, source-to-sink studies accounting for the connections between tectonics, climate, erosion and drainage development in mountain areas, and sedimentary accumulation in depositional sinks. This will require further exploration of the feedbacks between tectonic uplift and erosion, and to resolve quantitative uplift data from geomorphic features, issues that are already being addressed from analytical and modelling perspectives. In parallel, at the deep level and beyond the refinement of the knowledge of crustal structure, significant improvements are to be expected in the understanding of the role of mantle processes in the vertical motions we observe at the surface of the Earth.

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