Lithospheric structure in Central Eurasia derived from elevation, geoid anomaly and thermal analysis

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23

24 **Abstract**

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26 We present new crustal and lithospheric thickness maps for Central Eurasia from the
27 combination of elevation and geoid anomaly data and thermal analysis. The results are
28 strongly constrained by numerous previous data based on seismological and seismic
29 experiments, tomographic imaging and integrated geophysical studies. Our results indicate
30 that high topography regions are associated with crustal thickening being maximum below
31 the Zagros, Himalaya, Tien Shan and the Tibetan Plateau. The stiffer continental blocks
32 remaining undeformed within the continental collision areas are characterized by a slightly
33 thickened crust and flat topography. Lithospheric thickness and crustal thickness show
34 different patterns that highlight an important strain partitioning within the lithosphere. The
35 Arabia/Eurasia collision zone is characterized by a thick lithosphere underneath the Zagros
36 belt whereas a thin to inexistent lithospheric mantle is observed beneath the Iranian and
37 Anatolian plateaus. On the contrary, the India/Eurasia collision zone is characterized by a
38 very thick lithosphere below its southern part as a consequence of the underplating of the
39 cold and stiff Indian lithosphere. Our new model presents great improvements compared to
40 previous global models available for the region and allows us to discuss major aspects
41 related to the lithospheric structure and acting geodynamic processes in Central Eurasia.

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The Central Eurasia region contains two of the most prominent deformed regions on Earth: the Arabia/Eurasia and the India/Eurasia collision zones, forming most part of the well-studied Alpine-Himalayan belt. Defining the lithospheric structure of such large region including areas with very scarce information will provide important contributions to understand the relevant geodynamic processes occurring in collisional contexts.

During the last decades, the Himalaya-Tibet and Zagros-Iran regions have been the target of numerous geophysical surveys (mostly receiver functions, deep seismic and tomographic studies) to unravel their lithospheric structures. Crustal thickness data include 1D spotted estimations (e.g. Nasrabadi *et al.*, 2008; Chen *et al.*, 2010), 2D transects across the Himalaya-Tibet region (e.g. Kind *et al.*, 2002; Tian *et al.*, 2005; Nabelek *et al.*, 2009) or the Zagros-Iran region (e.g. Paul *et al.*, 2006, 2010) and 3D regional studies (e.g. Zör *et al.*, 2003; Pan & Niu, 2011). However, some other areas of Central Eurasia are less well known due to the lack of seismological and seismic studies. In fact, less than 19% of the region has crustal thickness data coverage better than 1 measure per 50,000 km² (i.e. one point every 2x2 arc-degree). Furthermore, results from these geophysical studies show a high scatter related to different surveys, instrumentation, recording periods and methodology. Some global crustal thickness models are proposed and they have provided general features of the crustal structures with a spatial resolution up to 1x1 arc degree (e.g. Nataf & Ricard, 1996; Mooney *et al.*, 1998; Bassin *et al.*, 2000; Laske *et al.*, 2013). Lithospheric thickness estimations are even scarcer and they are highly variable as a function of the used methodology. Regional tomography models have been developed to image the upper mantle structure (e.g. Ritzwoller and Levshin, 1998; Villaseñor *et al.*, 2001) and only few receiver functions studies are able to image the base of the lithosphere (e.g. Ramesh *et al.*, 2010; Zhao *et al.*, 2010).

Jiménez-Munt *et al.* (2012) presented a detailed image of the crust and lithospheric mantle variations in the Arabia-Eurasia collision zone based on the combination of geoid and

elevation data and thermal analysis. In the present study we use a similar approach with the aim to map the Crust-Mantle Boundary (CMB) and the Lithosphere-Asthenosphere Boundary (LAB) over the entire Central Eurasia region that we compare with available data. This study therefore incorporates an up-to-date compilation of crustal thickness data from seismological and seismic studies including 1212 estimations coming from more than 100 references used as main constraints. We discuss our resulting crustal and lithospheric thickness maps with previous global crustal and lithosphere thermal models (e.g. Laske *et al.*, 2013, Reguzzoni *et al.*, 2013; Artemieva, 2006; Goutorbe *et al.*, 2011) and we shortly discuss relevant geodynamic processes occurring in the Central Eurasia region.

Geodynamic context

During Cenozoic times, the Central Eurasia region has been affected by the India/Eurasia and Arabia/Eurasia continental collisions. In both cases, collision occurred between the strong and resistant Archean-to-Proterozoic shields of Indian and Arabian plates, and the weaker southern margin of Eurasia. The weakness of the Eurasian margin is interpreted as a result of the presence of major pre-existing structures as suture zones and/or large-scale fault zones between the different accreted Gondwana-derived continental blocks (e.g. Audet & Bürgmann, 2011). As a consequence of these collisions, two major topographic features arised: the Zagros and the Himalaya orogenic belts, which show different stages of development due to differences in amounts of convergence and ages of initiation of the collision (Hatzfeld & Molnar, 2010).

For the sake of simplicity, we differentiate the region into 5 major tectonic subdivisions: (1) the Arabian Plate, (2) the Indian Plate, (3) the Arabia/Eurasia collision zone, (4) the India/Eurasia collision zone and (5) the region to the North of the Tethysides (Figure 1).

The Arabian Plate

The Arabian Plate is composed of the Arabian shield (Late Proterozoic basement) unconformably overlapped to the East by the Phanerozoic Arabian platform and separated from Africa by young spreading centers located within the Red Sea and the Gulf of Aden (Al-Damegh *et al.*, 2005). The Arabian platform consists of thick Paleozoic and Mesozoic sediments dipping NE-ward and reaching more than 10 km thickness (Mokthar *et al.*, 2001). The Red Sea marks the boundary between Africa and Arabia and consists of both oceanic and thinned continental crust (Al-Damegh *et al.*, 2004) covered by up to 5 km thick Cenozoic sediments according to the Exxon Tectonic Map of the World (1994).

The Indian Plate

The Indian Plate is composed of several Early to Late Archean cratons delimited by rift zones containing Proterozoic and/or Phanerozoic sediments (Naqvi & Rogers, 1987). The massive basalts of the Deccan volcanic province erupted during Late Cretaceous to Early Tertiary and are considered as a consequence of the separation of India from the Seychelles microcontinent (a Gondwana-derived continental block) under the influence of the Réunion mantle plume (Mahoney, 1988). The break-up of Gondwanaland occurred around 140 Ma (earliest Cretaceous) and corresponds to the beginning of the northern drift of the Indian Plate towards Eurasia (Kumar *et al.*, 2007).

The Arabia-Eurasia collision zone

The Zagros orogen is the consequence of the closure of the Neotethys Ocean that was located between Arabia and Eurasia (e.g. Sengör *et al.*, 1988; Agard *et al.*, 2011). Though the timing of the continental collision is still controversial, the final closure of the Neotethys Ocean seems to occur during the Late Oligocene-earliest Miocene (e.g. Agard *et al.*, 2011,

McQuarrie & van Hinsbergen, 2013; Mouthereau *et al.*, 2012; Vergés *et al.*, 2011). The Zagros orogenic system continues to the South-East to Makran and its North-West continuation corresponds to South-East Anatolia where the collision started earlier than in the Zagros (Agard *et al.*, 2011). Subduction-related volcanism continues to occur throughout both the Iranian and the Turkish plateaus (Kazmin *et al.*, 1986) suggesting that, despite the change from subduction to continental collision, the tectonics of these plateaus is still driven by subduction processes within the upper mantle (Hearn & Ni, 1994; Agard *et al.*, 2011). The Alborz and Kopet Dag ranges developed along the Paleotethys suture zone (Sengör *et al.*, 1988; Robert *et al.*, 2014) and they correspond to the northern boundary of the Arabia-Eurasia collision zone. The Alborz is bounded to the North by the South Caspian Sea Basin, which is characterized by a sedimentary basin exceeding 20 km in sediment thickness (Brunet *et al.*, 2003) (Figure 2).

The India-Eurasia collision zone

The India-Eurasia collision zone hosts the largest and highest topographic feature on Earth: the Tibetan Plateau, which was built by sequential accretion of lithospheric terranes upon the southern margin of Eurasia since the end of the Paleozoic (Yin & Harrison, 2000; Pubellier *et al.*, 2008). From North to South, the Tibetan Plateau is successively formed by (1) the Kunlun terrane, (2) the Songpan Garze flysch complex related to the closure of the Paleotethys, (3) the continental Qiangtang terrane accreted at the end of the closure of the Paleotethys and (4) the Lhasa terrane accreted during Cretaceous times (Figure 1). The Indus Tsangpo suture zone, related to the closure of the Neotethys Ocean, is located to the South of the Lhasa Block and to the North of the Himalaya. The timing of the India/Eurasia collision is still subject to controversies but most authors agree that it occurred between 55 and 45 Ma during the Eocene (e.g. Rowley, 1996; Zhu *et al.*, 2005).

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144 North of the Tethysides

145 The North of the Tethysides is a region composed from East to West by the Ordos block, the
146 Altaid range (Figure 1), the Tien Shan range, the Kazakh terranes, the Turan platform and
147 the North Caspian Basin, the Black Sea and the East European Craton (Figure 1). The Altaid
148 range has a complex geodynamic evolution made of multiple accretions of terranes of
149 different origin, chiefly microcontinents and island arcs occurring from 600 Ma to 250 Ma
150 (latest Precambrian to earliest Triassic times) (Wilhem *et al.*, 2012). Numerous sedimentary
151 basins were developed in this area as the Junggar Basin or the North Caspian Basin, which
152 presents an up to 20 km thick sedimentary sequence.

153

154 **Methodology and input data**

155

156 Method

157 We calculate the crustal and the lithospheric mantle thickness by combining elevation and
158 geoid data together with thermal analysis in a 1D approach (Fullea *et al.*, 2007). This
159 method assumes local isostasy with a depth of compensation below the lithosphere-
160 asthenosphere boundary (LAB) and considers four layers: sea water, crust, lithospheric
161 mantle and asthenosphere. The crustal density increases linearly with depth between
162 predefined values at surface and at the base of the crust. The lithospheric mantle density
163 (ρ_m) is considered to be temperature dependent:

$$164 \quad \rho_m(z) = \rho_a(1 + \alpha(T_a - T(z)))$$

165 where ρ_a is the density of the asthenosphere considered constant everywhere, α is the
166 thermal expansion coefficient, T_a is the temperature at the LAB and $T(z)$ is the temperature

of the lithospheric mantle at a given depth z .

The geoid anomaly is calculated relative to a reference level (H_0), which in turn depends on a selected reference lithospheric column with a known crustal thickness and elevation. It can be demonstrated that knowing the crustal thickness and considering an average crustal density and heat production, we can calculate the lithospheric thickness that fits the elevation corresponding to the selected reference column (see Fullea *et al.*, 2007, for details). The lithospheric reference column for geoid anomaly has been chosen within the Indian shield where consistent crustal and lithospheric thicknesses are obtained from seismic and seismological studies (e.g. Bhattacharya, 2009; Kumar *et al.*, 2007; Ramesh *et al.*, 2010). This method was successfully applied to decipher crustal and lithospheric structures in the Gibraltar arc system (Fullea *et al.*, 2007) and in Arabia/Eurasia collision zone (Jiménez-Munt *et al.*, 2012).

Input parameters used in our modeling are summarized in the Tables 1 and 2. According to the large extent of the region considered, lateral variations of the surface density were used in our modeling (Figure 2). This 2D density distribution takes into account the occurrence of large sedimentary basins, as well as small differences between the considered tectonic subdivisions (Table 2). Because the crust is defined as a simple layer with a linear depth-density increase, the occurrence of thick sedimentary basins is associated with a decrease of the surface crustal density. First, we obtained the sediment thickness distribution by digitizing the Exxon Tectonic Map of the World (1994), complemented with some local studies (Singh, 1996; Sastri *et al.*, 1971; Rao, 1973; Karunakaran & Rao, 1979) (Figure 2). And then, we calculated the surface density according to:

$$\rho_{\text{surf}} = 2700 - 100 \times (1 - \exp(-H/5))$$

with ρ_{surf} being the crustal density at surface and H the sediment thickness expressed in km.

191 This relation has been chosen in order to simulate the lower average crustal density at the
192 surface with the increase of the sediment thickness. According to this equation, the surface
193 density is equal to 2700 kg.m^{-3} when there is no sediments and equal to 2602 kg.m^{-3} when
194 sediments are 20 km thick.

195 Certainly, the absolute density values used for the lithospheric mantle and the asthenosphere
196 are noticeably lower than the actual ones. This is because we are using a 'pure' thermal
197 approach in which the lithosphere mantle density is only temperature dependent and the
198 density of the asthenosphere is constant everywhere. In other words, temperature is the
199 dominant effect on density and no pressure effects are considered. Note that this approach is
200 widely used in describing variations of topography and potential fields since they mainly
201 depend on lateral density variations rather than on absolute density values.

202 Because of the occurrence of large magmatic provinces and numerous ophiolitic belts, the
203 crustal density at Moho depth in the Arabia-Eurasia collision zone has been increased to
204 2970 kg.m^{-3} . We use classical thermal conductivities of $2.7 \text{ W.m}^{-1}.\text{K}^{-1}$ for the crust and 3.2
205 $\text{W.m}^{-1}.\text{K}^{-1}$ for the lithospheric mantle. The thermal conductivities are considered as constant
206 in our modeling. In fact, the larger variability of the crustal thermal conductivity is restricted
207 to the uppermost part of the crust (up to 20 MPa) (Clauser & Huenges, 1995) and then, does
208 not affect much the average crustal thermal conductivity. The average crustal radiogenic heat
209 production is $0.5 \mu\text{W.m}^{-3}$ for continental crust and $0.3 \mu\text{W.m}^{-3}$ for oceanic crust (Vilà *et al.*,
210 2010). The heat production in the lithospheric mantle is considered as null because of its
211 very low contents in radiogenic elements.

212 Fulla *et al.* (2007) performed a sensitivity analysis for the calculation of CMB and LAB
213 depths induced by the typical RMS error of elevation and geoid anomaly databases. They
214 found that the inaccuracy for the crustal thickness in emerged regions is less than 2 km
215 whereas maximum error in lithospheric thickness is less than 10 km. The effect of varying

crustal and lithospheric mantle conductivity has been tested and the resulting lithospheric thickness is affected up to 6-8 km whereas varying the coefficient of thermal expansion (α) affects the lithospheric thickness up to ~12 km. In addition, the crustal thickness is not much affected (~1 km) by varying thermal parameters.

Considering this sensitivity analysis, we estimated the maximum misfits obtained in our modeling result to be up to 5 km for the crustal thickness and up to 15 km for the lithospheric thickness.

Input data

Topography and Bathymetry

Topographic and bathymetric data are extracted from the ETOPO1 database (Amante & Eakins, 2009) (Figure 3a). The most remarkable geomorphological feature of Central Eurasia is the Tibetan Plateau which covers an area of more than $2.5 \times 10^6 \text{ km}^2$ with a mean elevation of around 4500 m. High elevations are also found to the North of the Tibetan Plateau, along the Tien Shan or the Qilian Shan ranges. The Tarim basin presents a low relief with a mean elevation of around 1000 m. To the West, the highest elevations are located in the Zagros Mountains and they extend towards the Anatolian Plateau. The Alborz, Kopet Dag, Caucasus and Sistan ranges show also remarkable topographies with elevations up to 5000 m. To avoid unrealistic high frequency variations of the Moho and LAB depths we have filtered the elevation using a Gaussian filter with a wavelength of 100 km.

Geoid anomaly

The geoid anomaly was extracted from the EGM2008 global model (Pavlis *et al.*, 2008). The EGM2008 gravitational model is complete to spherical harmonic degree and order 2159, which determines the resolution of our modeling at 10 arc minute (~18.5 km). In order to

avoid the effects of sublithospheric density variations, the geoid was filtered to remove the signature corresponding to the lower spherical harmonics until degree and order 11 (Figure 3b). The degree and order of spherical harmonics filtering have been chosen in order to remove very large wavelength anomalies that probably result in sub-lithospheric density variations (see figure S2, supplementary materials).

The obtained geoid anomaly shows maximum values of up to 21 m located in two main regions: (1) along the Himalaya and the Tibetan Plateau with a northward decrease of the anomaly, and (2) in the Anatolian Plateau, in the Alborz and in the Sanandaj-Sirjan zones in Iran. Minimum values are down to -21.5 m coinciding with large sedimentary accumulations as occurs in the Tarim, Junggar and Ganga basins, the Caspian Sea, the Nile delta, the Persian Gulf and eastern part of Arabia.

Comparison with seismological and seismic studies

We carried out a complete compilation of previously published crustal thickness data (Figure 4, Table 3 and supplementary material S1) mainly from receiver function and deep seismic studies. To test the reliability of our modeling approach, we compared this compilation to crustal thicknesses obtained from our modeling (Figures 5, 6 and 7). The resulting CMB and LAB from our modeling approach are downloadable in the supplementary material and they are furnished in .xyz format. In contrast to the CMB, the LAB corresponds to a rheological rather than to a compositional contrast, which explains that its location and properties are more elusive (Eaton *et al.*, 2009). Shear-wave velocity anomaly within the upper mantle computed from global or regional tomography is often considered as correlated with the lithospheric thickness. We compare our results with the S40RTS mantle shear-wave velocity model (Ritsema *et al.*, 2011). The LAB is rarely imaged using receiver function methods but

we compared our modeling results with estimations from these studies where they were available (e.g. Kumar *et al.*, 2007; Zhao *et al.*, 2011).

The Arabian Plate

The crustal thickness of the Arabian Plate estimated from receiver functions, seismic refraction profiles and surface wave studies (see references in Table 3) ranges between 32 and 50 km with an average value of ~40 km, which is consistent with our modeling results (Figure 5a). Our map shows a smooth crustal thinning towards the North-East of the Arabian Plate, which could be interpreted as inherited from the passive margin structure though it is not well imaged by the scarce available seismic data. Both previous seismic studies and our modeling results image an abrupt crustal thinning along the boundaries of the Arabian Plate (the Gulf of Aden to the South-East, the Gulf of Aqaba to the North-West and the Red Sea to the West). Finally, beneath Oman and along the southern part of Red Sea coast, our model indicates that the crustal thickness may reach up to 43 km, whereas Al-Damegh *et al.* (2005) propose a crustal thickness up to 50 km.

The shear-wave velocity anomaly map at 100 km depth shows a large low velocity anomaly in western Arabia and along the Red Sea, whereas a positive anomaly is imaged beneath the Mesopotamian Basin (Villaseñor *et al.*, 2001; Ritsema *et al.*, 2011). Our model exhibits a thicker lithosphere in the eastern part of the Arabian Plate than to the western part, with a maximum thickness exceeding 200-220 km beneath the Persian Gulf and South-East Arabia (Figure 5b), which is in agreement with estimations from tomographic results (Villaseñor *et al.*, 2001; Shomali *et al.*, 2011). According to our results, this thickened lithosphere extends beneath the Zagros Belt, which is also confirmed by shear-wave modeling (Priestley *et al.*, 2012). Finally, our model indicates that towards the North-West of the Mesopotamian Basin, the lithosphere thickness decreases reaching minimum values of about ~140 km south of the

Anatolian Plateau.

The Indian Plate

Previous crustal thickness estimations of the Indian Plate vary from 30 to 44 km but most of these estimations are comprised between 36 and 39 km (see references in Table 3). Our modeling results fit very well with previously obtained crustal thicknesses for the whole Indian Plate except for the Shillong Plateau where our model predicts a thicker crust due to the fact that our approach is not strictly valid to image an uncompensated crustal pop-up structure as proposed by Mitra *et al.* (2005).

Previous studies using surface-wave dispersion data observed a shield-like lithospheric structure in most part of the plate (Singh, 1999; Mitra *et al.*, 2006; Suresh *et al.*, 2008; Bhattacharya, 2009; Prajapati *et al.*, 2011) and there are no major lateral variations in the shear-wave velocity anomalies across the Indian Plate (Villaseñor *et al.*, 2001; Ritsema *et al.*, 2011). According to previous geophysical studies, the lithospheric thickness beneath western India is ~155 km (Mitra *et al.*, 2006) and ~140 km beneath eastern India (Bhattacharya, 2009). Our model images a ~160-190 km thick lithosphere; though a lithosphere thickness reaching up to 230 km is imaged below the central part of the Ganga Basin (Figure 5b). In this region, the possible component of flexural support to the lithospheric bending, evidenced by large sedimentary thickness accumulation, could also affect the validity of our approach resulting in an overestimation of our modeled lithospheric thickness. However, the lack of other geophysical data in this area does not allow for firm conclusions.

The Arabia-Eurasia collision zone

In agreement to previous geophysical studies our modeling results show crustal thickness

values ranging between 35 and 45 km beneath the Mesopotamian Basin (see references in Table 3). Along the Zagros orogeny, seismic data indicate that the maximum crustal thickness is located below the Sanandaj-Sirjan Zone, whereas our model predicts the maximum crustal thickness values slightly displaced to the South-West coinciding with the highest elevations and directly below the Imbricated Zone and the Zagros Simply Folded Belt (Figure 5a). This shift of the maximum crustal root relative to maximum topography is common in collisional settings where crustal thickening is associated with underthrusting or down-flexuring at crustal scale as it might be the case of the Zagros orogeny (Motavalli-Anbaran *et al.*, 2011; Vergés *et al.*, 2011). Our 1D approach, based on the combination of elevation and geoid anomaly, is not able to reproduce these underthrusting structures and therefore to precisely locate the region with maximum crustal thickness. However, our modeling results indicate crustal thickness values between 35 and 54 km beneath the Iranian and Afghan blocks in good agreement with available seismic data in Central Iran (Paul *et al.*, 2006; Nasrabadi *et al.*, 2008; Paul *et al.*, 2010; Asfari *et al.*, 2011; Motaghi *et al.*, 2012) suggesting that the stiff lithospheric domains are related to relatively low crustal thickness values. The Alborz, Kopet Dag and Caucasus mountain ranges are characterized by thickened crust according to both, seismic data and modeling results (Figure 7). In the Anatolian Plateau our model shows crustal thickness values between 36 and 50 km in good agreement with most of the estimations from seismological studies, which in turn show a large scattering (up to ~20 km of difference).

Shear-wave velocity studies in the Arabia-Eurasia collision zone show a thick high shear-wave velocity anomaly within the upper mantle beneath eastern Arabia and Zagros whereas western Arabia, Central Iran and the Anatolian Plateau domains are characterized by low shear-wave velocity anomalies within the upper mantle (Villaseñor *et al.*, 2001; Priestley *et al.*, 2012). Low shear-wave velocities underneath the eastern part of the Anatolian Plateau

are interpreted as a consequence of a very thin or possibly removed lithospheric mantle (Gök *et al.*, 2008; Bakirci *et al.*, 2012), which is confirmed by S-receiver functions studies (Angus *et al.*, 2006) and low Pn wave speeds (Al-Lazki *et al.*, 2004). Mohsen *et al.* (2006) propose a lithospheric thickness value of 90 km beneath eastern Anatolia, which is in agreement with our model, highlighting a thin lithosphere beneath the whole Anatolian Plateau (from 100 to 130 km thick).

The India-Eurasia collision zone

Many geophysical studies have been carried out across the Himalayan range and the Tibetan Plateau to image its complex crustal and lithospheric structure. Most of these studies are based on deep seismic profiles (Zhao *et al.*, 2001; Galvé *et al.*, 2002; Ross *et al.*, 2004; Zhao *et al.*, 2006; Mechie *et al.*, 2012) and on passive seismological data (see the whole set of references in the Table 3). A general result is that, despite its low relief and high elevation, the Tibetan Plateau shows a highly variable crustal thickness ranging from 56 km up to 95 km and a heterogeneous lithospheric structure (e.g. Yin & Harrison, 2000; Zhao *et al.*, 2010). Collectively, the results suggest that the crust is thicker in the southern part (between 70 to 80 km thick) of the plateau than to the North (between 65 to 70 km thick) except for the easternmost part of the plateau (longitude > 95°E). In contrast, our results indicate that the crust in the southern part of the Tibetan Plateau ranges between 68 and 76 km thick whereas it predicts a thicker crust (more than 75 km thick) in the northern part of the plateau (Figure 5a). To the West of the Tibetan Plateau, seismological studies indicate a noticeable crustal thickening from ~50 km below the front of the Himalayan range in Pakistan (Soomro, 2009) to ~78 km below the Pamir (Li & Mashele, 2009; Mechie *et al.*, 2012). Our model is broadly in agreement with these estimations although it slightly underestimates the crustal thickness below southern Pamir.

Our model is able to reproduce the lithospheric thickness in the southern and western part of the plateau, in agreement with receiver function data (Owens & Zandt, 1997; Nabelek *et al.*, 2009; Zhao *et al.*, 2010, 2011), tomographic data (Villaseñor *et al.*, 2001; Priestley *et al.*, 2006; Replumaz *et al.*, 2013) and previous lithospheric modeling (Jiménez-Munt *et al.*, 2008) that suggest a ~200 to 240 km thick lithosphere. Most authors agree that this thick Tibetan lithosphere corresponds to the northern boundary of the underthrust Indian lithosphere below Tibet (e.g., Owens and Zandt, 1997; Nabelek *et al.*, 2009; Zhao *et al.*, 2010, 2011; Liang *et al.*, 2012). The main discrepancy concerns the northern part of the Tibetan Plateau where some studies suggest a very thin to inexistent lithospheric mantle (e.g. Meissner *et al.*, 2004; Jiménez-Munt *et al.*, 2008; Liang *et al.*, 2012) whereas other studies suggest a very thick lithosphere (e.g. Priestley *et al.*, 2006; Barron & Priestley, 2009). Our model results indicate a very thick lithosphere, up to 340 km thickness coinciding with the region where our model overestimates the crustal thickness. However, using the same geoid and elevation data plus gravity and thermal data, Jiménez-Munt *et al.* (2008) proposed a very thin to removed lithosphere. Unfortunately, our results do not allow discerning among the proposed geodynamic processes acting in this peculiar part of the Tibetan Plateau.

North of the Tethysides

According to seismological and seismic studies, the Tarim basin is characterized by a crustal thickness of ~40 km in its central part increasing up to ~60 km in its flanks (Kao *et al.*, 2001; Qiusheng *et al.*, 2002; Wittlinger *et al.*, 2004; Mi *et al.*, 2005; Zhao *et al.*, 2003, 2006; Chen *et al.*, 2010). Our model overestimates by ~ 5 km the crustal thickness in the Central Tarim basin and does not reflect noticeable crustal thickness variations beneath the basin (Figure 5a). According to seismological estimations the crustal thickness below the Tien Shan is 55-60 km (Bump & Sheehan, 1998; Zhao *et al.*, 2003; Vinnik *et al.*, 2004), whereas our model

predicts crustal thickness values between 60 and 68 km. In the Kazakh terranes, North-West from Tien Shan, the small number of available seismological estimations indicate a crustal thickness of about 45 km (Vinnik *et al.*, 2004; Bump & Sheehan, 1998), which is very close to the 40-44 km obtained from our model for the whole region. In the North-East part of the study area, North of the Altaids Ranges, the crustal thickness inferred from receiver functions (Zorin *et al.*, 2002) varies from 44 to 50 km in full agreement with our results. The lithospheric thickness have been estimated from S-wave receiver functions (Kumar *et al.*, 2005) indicating values of 90-120 km beneath the Tien Shan, ~180 km in the Tarim Basin, and ~130 km in the Kazakh Platform. Our results show a thicker lithosphere in all these regions varying from 240 km in the Tien Shan, to 220 km in the Tarim Basin and 180 km in the Kazakh Platform, being in agreement with S-wave tomography models, which indicate high velocities below the Tien Shan and the Tarim basin (Ritsema *et al.*, 2011).

Discussion

One of the goals of the present study is to provide new maps of both the Crust-Mantle-Boundary and the Lithosphere-Asthenosphere-Boundary in the Central Eurasia region and to compare these results with previous regional/global models. A second related goal is to discuss the implications on the geodynamic evolution of this large continental collision region based on our crustal and lithospheric thickness maps.

Our modeling approach is based on the assumptions of local lithospheric isostasy and thermal equilibrium. Whereas local isostasy is an acceptable approximation for wavelengths of hundred kilometers (e.g. McKenzie & Bowin, 1976; England & Molnar, 1997), the assumption of thermal equilibrium is not valid in regions of recent tectonic activity as the Alpine-Himalayan system. Therefore, the results of our model must be interpreted as

average physical conditions necessary to produce the required density distribution rather than the actual thermal boundaries. The calculation in steady state minimizes the variations in lithospheric thickness, since the modeling tends to underestimate the lithosphere thickness when thickening processes occur under transient conditions, and overestimate it for thinning processes (see Fullea *et al.*, 2007 and Jiménez-Munt *et al.*, 2011 for more details).

Comparison with global crustal models

The most widely used global datasets related to crustal thickness is the CRUST2.0 (Bassin *et al.*, 2000), which has recently been replaced by the CRUST1.0 that incorporates an updated version of global sediment thickness and presents a better accuracy (Laske *et al.*, 2013). Reguzzoni *et al.*, (2013) combined CRUST2.0 model with gravity observations from the GOCE satellite to propose the new global crustal GEMMA model that we compare with our results (Figure 7).

The 1x1° grid global crustal model CRUST1.0 is obtained from the analysis of seismic wave travel-times and seismic refraction data (Laske *et al.*, 2013). The resulting crustal thickness from the CRUST1.0 model (Figure 8a) in the study region is relatively smooth with maximum values exceeding 70 km in the Tibetan Plateau and minimum values of <15 km in the oceanic domains of the Red Sea and the Indian Ocean. Most of the continental domains show crustal thickness values of 40±5 km. The excessive smoothness of the resulting map makes difficult to relate crustal thickness variations with surface geological structures. Therefore, the crustal thickness derived from our model exceeds that from CRUST1.0 in more than 10-15 km along the main mountain ranges (i.e., Red Sea rift shoulders, Anatolia, Caucasus, Zagros, Alborz, Makran, Pamir, Qiangtang and Tien Shan).

The resulting crustal thickness from GEMMA Moho model (Reguzzoni *et al.*, 2013) ranges

from 15 to more than 100 km and presents undulations of ~40 km large over the whole region (Figure 8b). In contrast, our method allows for a better interpolation between regions with reliable measurements and then show a much better correlation with the major geological structures (Figure 8c).

Comparison with global lithospheric models

Concerning the lithospheric thickness, we compare our modeling results with 2 lithospheric models: the TC1 model (Artemieva, 2006) and the thermal lithospheric model presented in Goutorbe *et al.* (2011). In addition to these two models, there are a number of regional/global seismic tomography models and several regional studies based on the combination of different geological and geophysical data that can also constrain the crust and upper mantle structures as compared below.

The resulting lithospheric thickness from our model can be qualitatively compared to regional/global models based on seismic surface waves and thermal approaches. Regional and global S-wave velocity perturbations in the Central Eurasia region at 100 km depth (e.g. Villaseñor *et al.*, 2001; Priestley *et al.*, 2006; Ritsema *et al.*, 2011; Hatzfeld & Molnar, 2010 and references therein) show low velocities beneath the Red Sea, Eastern Anatolia, Central Iran and Afghan blocks, Eastern Tibet, and North to the Altaid ranges, with some differences among these models. High velocity perturbations are located beneath North India and West Tibet, and the Northern regions. Major differences between our model and S-wave tomography models are related to the lateral variations in lithospheric thickness from the Zagros to Central Iran and the Mesopotamian Basin. Some tomography models (e.g. Priestley *et al.*, 2006; Priestley *et al.*, 2012) propose a lithosphere thickness exceeding 230 km along the so-called Zagros core thinning rapidly towards Central Iran and the Mesopotamian Basin. In contrast, our model shows that the thick lithosphere affects the

Mesopotamian Basin, the South-East Arabian Plate and the Zagros until the Arabia-Eurasia continental suture with an abrupt thinning beneath the Sanandaj-Sirjan Zone and the Urumieh-Dokhtar Magmatic Arc. This lithosphere thinning extends from Anatolia, to Central Iran and to the Afghan Block, which is more in agreement with the tomography models by Villaseñor *et al.* (2001) and Ritsema *et al.* (2011) that also show a lithospheric thickening beneath the Mesopotamian Basin as in our model. Another conflictive region is the Eastern Tibet, where all tomography models show a relatively low velocity region at 100-125 km depth vanishing with depth and disappearing at 200 km depth. Finally, the high velocity anomalies occupying most of the northern regions of the study area are in agreement to the lithospheric thickness of 170-200 km proposed by our model. This fitting is not accomplished north to the Altaid ranges where tomography models consistently image a low velocity region in contrast to a rather homogeneous lithospheric thickness of 180-200 km inferred from our calculations. In any case, the interpretation of S-wave tomography in terms of thermal models must take into account that velocity anomalies of non-thermal origin may amount up to $\pm 3\%$ of V_s amplitude and attributed to chemical composition variations and/or to the presence of melts/fluids (Artemieva, 2009).

Global thermal models can be directly compared to our results since the approach used in the present work include thermal analysis. TC1 is a global scale continental lithospheric model that has been computed from available heat flow measurements supplemented with electromagnetic and xenolith data in cratonic domains (Artemieva, 2006) (Figure 8a). In Central Eurasia, TC1 model proposes lithospheric thickness ranging from less than 100 km below Anatolia and Iran, up to more than 200 km in the North-East Tibetan Plateau and W-Central part of the Indian Plate. Our model presents a roughly similar trend of the distribution of the lithospheric thickness but with a higher resolution and more precise lithospheric thickness estimations in the study region (Figure 8b). The lithospheric

thickening affecting the Eastern Arabian Plate and the Turan Platform are not evident in TC1 model and, conversely, our model does not image the apparent lithospheric thickening in W-Central India. Goutorbe *et al.* (2011) published a thermal model obtained from heat flow measurements combined with multiple geological and geophysical data and proxies. This model reproduces the lithospheric thickening affecting the eastern Arabian Plate and the Mesopotamian Basin, as well as the Tibetan Plateau and North India Plate. Lithospheric thinning is concentrated along the Red Sea and its shoulders, Anatolia and Central Iran and Afghan blocks. The proposed lithospheric thickness values are however unrealistically low especially in the thinned regions, where the LAB is less than 40 km thick.

Geodynamic implications

The new results indicate that crustal thickening in Central Eurasia is clearly associated with major frontal ranges as the Himalaya and Zagros, but also to distal ranges as the Alborz, Kopet Dagh, and Tien Shan. The more than 1200 km in width of the deformed area resulting from the Arabia/Eurasia and India/Eurasia collision zones is also highlighted by the extent of the area affected by crustal thickening (Figures 1 and 5a). As already suggested by many geophysical studies, the Tibetan Plateau is sustained by a very thick crust with an average thickness of ~75 km according to our model. Furthermore, the tectonic blocks remaining in the deformation zone, as the Central Iran or Tarim blocks, show a slightly thickened crust and uniform topography pointing out their rheological resistance and moderate deformation. A consequence of these features is that tectonic stresses are able to be efficiently transmitted hundreds to thousand kilometers far from the suture zones.

Contrarily to the crustal structures that are comparable between the two collision zones, their lithospheric structures differ substantially (Figures 5 and 6). The Arabia/Eurasia collision zone is characterized by a thick lithosphere underneath the Zagros belt and the Arabian

platform (between 180 and 220 km thick) whereas the Central Iran, the Anatolian Plateau and the northern part of the Arabian Plate are characterized by thin to very thin lithosphere (90-150 km thickness). In contrast, the India/Eurasia collision zone shows a thick to very thick lithosphere especially in its southern part thus evidencing the plate underthrusting of most part of the Indian lithosphere beneath the Eurasian lithosphere. The large differences in lateral thickness variations between the crustal and the lithospheric mantle indicate a strong strain partitioning with the CMB acting as the preferential detachment level as proved in Iran (Jiménez-Munt *et al.*, 2012).

Conclusions

We present a new crustal and lithospheric thickness model assuming thermal and isostatic equilibrium that fits well with existing seismic and seismological data. The method used permits to interpolate crustal and lithospheric thickness values between regions with reliable measurements and provides a full coverage for both the crust and the lithosphere of Central Eurasia. The presented work allows for drawing the following concluding remarks:

- Crustal thickening is correlated with major mountains ranges (Himalayas, Zagros, Pamir, Caucasus, and Tien Shan) whereas crustal thinning is restricted to the Arabian and Indian oceanic domains and to the South Caspian Sea, the Red Sea and the Black Sea. The Iranian and Anatolian plateaus do not show a significant thick crust whereas the Tibetan Plateau shows an extremely thickened crust up to 78 km thick. At regional scale, our resulting map matches well with geological features.
- Contrasting lithospheric structures are modeled between the India/Eurasia and the Arabia/Eurasia collision zones. Thickened lithosphere is obtained below the Zagros orogenic belt whereas the Anatolian Plateau and Central Iran are characterized by a

thin to very thin lithosphere (~100-130 km thick). In contrast, maximum lithospheric thickness, reaching up to 300 km thick, corresponds to the southern and western Tibetan Plateau and below the Pamir. Our model coincides well with S-wave tomography models but not in Eastern Tibet where our model cannot resolve the existence of a Low Velocity Zone down to 200 km depth, suggesting a lithospheric thinning related to deep geodynamic processes.

- Our resulting crustal map is roughly in good agreement with the global CRUST1.0 and GEMMA Moho models. However, the crustal thickness derived from our model exceeds that from CRUST1.0 in more than 10-15 km along the main mountain ranges, which better fits the crustal thickness in these tectonic domains based on geophysical studies. Our resulting model presents a similar trend of the distribution of the lithospheric thickness compared to regional and global models but it also presents better resolution and thickness estimations.

- India-Eurasia and Arabia-Eurasia collisional systems present comparable crustal structure but contrasted lithospheric structure. The Arabia-Eurasia collision zone is characterized by a very thin lithosphere that can be interpreted as the signature of the dominance of subduction processes. Oppositely, the more mature India-Eurasia collision zone shows a thick lithosphere in its western and central parts that highlights the importance of crustal underthrusting processes.

- Our new results indicate that crustal thickening is not restricted but extends hundreds to thousand kilometers away from the collision front, indicating an effective transmission of tectonic stresses, which is partly related to the presence of stiff lithospheric blocks that remain almost undeformed within the collisional systems.

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Tables captions:

Table 1: Constant parameters used over the whole Central Eurasia region in our modeling.

Table 2: Parameters used in our modeling, as a function of the zones defined on the figure 1.

Table 3: List of publications used for the compilation of crustal thickness data that are represented on the figure 4. Complete dataset of the crustal thicknesses is presented in the supplementary material (Table S1).

Table S1 (Supplementary Material): Compilation of crustal thickness data from publications that were used to be compared with our modeling results. These data are represented on the figure 4 and the Table S3 only indicates references used for this compilation.

Figure captions:

Figure 1. Simplified structural map of the Central Eurasia region modified from the Structural map of Eastern Eurasia (Pubellier *et al.*, 2008, modified) and publications (Sengör and Natal'in, 1996; Frizon de Lamotte *et al.*, 2013). a) this map highlights the accreted terranes in the region resulting from the long-lived convergence between

1125 Gondwana-derived blocks and the Southern margin of Eurasia since the end of the
1126 Paleozoic. b) Simplified map showing the different zones delimited in this study. The
1127 shaded relief was obtained from the ETOPO1 database (Amante & Eakins, 2009). The
1128 scale bars corresponds to 15° latitude.

1129

1130 **Figure 2.** Mean crustal density used as an input in our model. Contours denote
1131 sediment thickness in meters taken from EXXON Structural Map of the World (1994)
1132 and publications related in the text. Used parameters for each region are summarized
1133 in the Table 2.

1134

1135 **Figure 3.** a) Elevation map from ETOPO1 database (Amante & Eakins, 2009). b)
1136 Residual geoid height map resulting from the EGM2008 global model (Pavlis *et al.*,
1137 2008) after removing the lower spherical harmonics until degree and order 11. Main
1138 sutures zones have been represented. The scale bar corresponds to 15° latitude.

1139

1140 **Figure 4.** Compilation of previous crustal thickness estimations from seismological
1141 and seismic experiments (complete dataset in Table S1 and references in Table 3). The
1142 scale bar corresponds to 15° of latitude.

1143

1144 **Figure 5:** a) Crustal thickness map and b) lithospheric thickness map of the entire
1145 Central Eurasia region calculated from elevation and geoid anomaly inversion. The
1146 scale bar corresponds to 15° of latitude.

1147 **Figure 6:** a) Topographic map of the study and location of two lithospheric cross-

sections, b) Lithospheric cross-section across the India-Eurasia collision zone representing the topography and the depths of the crust-mantle boundary (CMB) and the lithosphere-asthenosphere boundary (LAB). c) Same as b) but across the Arabia-Eurasia collision zone.

Figure 7: Comparison between crustal thickness estimations from seismological and seismic experiments (circles and triangles corresponding to Figure 4) and the calculated values from inversion of geoid and elevation (background map corresponding to Figure 5a). Similar colors between filled symbols and background map indicate coincidence between seismic data and model results.

Figure 8. Crustal thickness maps from a) CRUST 1.0 model (1 degree grid resolution) (Bassin *et al.*, 2000), b) GEMMA Moho model (Reguzzoni *et al.*, 2013) and c) our results (0.16 degree grid resolution). Major sutures zones are plotted.

Figure 9: Lithospheric thickness maps from a) TC1 thermal model (Artemieva, 2006) and from b) our results. Major sutures zones are plotted.

Figure S2 (Supplementary Material): Comparison of the residual geoid anomaly resulting from filtering at different orders and degrees of spherical harmonics. These residual geoid anomaly indicates that large wavelength anomalies that probably result from deep sub-lithospheric density variations are filtered by removing the signal corresponding to degree and order above 10 of spherical harmonics.

Legend:

50°

Kazakh
terrane

Black Sea

Caspian Sea

Tarim Basin

Afghan block

Indian Plate

Arabian Plate

1000km

30°

40°

50°

60°

70°

80°

90°

100°

110°

a) Crustal thickness (in km) from CRUST 1.0 model

b) Crustal thickness (in km) from GEMMA crustal model

c) Crustal thickness (in km) from our modeling

a) Lithospheric thickness (in km) from TC1 thermal model

b) Lithospheric thickness (in km) from our modeling

Parameter	Value
Asthenosphere density (ρ_a)	3200 kg.m ⁻³
Sea Water density (ρ_w)	1031 kg.m ⁻³
Temperature at the LAB (T_a)	1300 °C
Surface temperature (T_s)	15 °C
Thermal expansion coefficient (α)	3.5 10 ⁻⁵ K ⁻¹
Crustal thermal conductivity (k_c)	2.7 W.K ⁻¹ .m ⁻¹
Mantle thermal conductivity (k_m)	3.2 W.K ⁻¹ .m ⁻¹

Parameter	Zone 1	Zone 2	Zone 3	Zone 4	Zone 5	Oceanic crust
Crustal density at surface <i>in kg.m⁻³</i>	[2608-2700]	[2605-2700]	[2601-2700]	[2605-2700]	[2601-2700]	[2602-2700]
Crustal density at CMB <i>in kg.m⁻³</i>	2910	2910	2970	2910	2910	2980
Crustal radiogenic heat production in mW.m ⁻³	0.5	0.5	0.5	0.5	0.5	0.3

Zone	References
Zone 1 – Arabian Plate	Al-Damegh et al. [2005]; Al-Lazki [2003]; Badri [1991]; Barton et al. [1990]; Brew [2001]; Gök et al. [2008]; Kumar et al. [2001]; Mohsen et al. [2006]; Sandvol et al. [1998]; Tiberi et al. [2007]
Zone 2 – Indian Plate	Gupta et al. [2003]; Kumar et al. [2001, 2004, 2007]; Mitra et al. [2005, 2008]; Radhakrishna et al. [2012]; Soomoro [2009]
Zone 3 – Arabia-Eurasia collision zone	Al-Damegh et al. [2005]; Angus et al. [2006]; Asfari et al. [2011]; Brew [2001]; Çakir et al. [2000]; Doloei & Roberts [2003]; Gök et al. [2008, 2011]; Gritto et al. [2008]; Magino & Priestley [1998]; Mellors et al. [2008]; Motaghi et al., [2012]; Nasrabad et al. [2008]; Nowrouzi et al. [2007]; Paul et al. [2006, 2010]; Radjaee et al. [2010]; Sandvol et al. [1998]; Sodoudi et al. [2009]; Tezel et al. [2013]; Yaminifard et al. [2012]; Zor et al. [2003]
Zone 4 – India-Eurasia collision zone	Bump & Sheehan [1998]; Chen et al. [2010]; Galvé et al. [2002]; Gao et al. [2013]; He et al. [2014]; Hetenyi [2007]; Jiang et al. [2006]; Kao et al. [2001]; Kind et al. [2002]; Kumar et al. [2005]; Li et al. [2001,2008]; Li & Mashele [2009] ; Lou et al. [2009]; Makarov et al. [2010]; Mechie et al. [2012]; Mi et al. [2005]; Mitra et al. [2005, 2008]; Nabelek et al. [2009]; Oreshin et al. [2008]; Pan & Niu [2011]; Qiusheng et al. [2002]; Robert et al. [2010a, b]; Ross et al. [2004]; Schulte-Peklum et al. [2005]; Shi et al. [2004, 2009]; Soomoro [2009]; Tian et al. [2005]; Tilmann [2011]; Tong et al. [2007]; Vergne et al. [2002]; Vinnik et al. [2004]; Wang et al. [2003,2010]; Wittlinger et al. [2004]; Xu et al., [2013]; Yue et al. [2012]; Zhang & Wang [2009]; Zhao et al. [2001, 2003, 2006]
Zone 5 – North of the Tethysides	Chen et al. [2010]; Magino & Priestley [1998]; Mellors et al. [2008]; Motaghi et al. [2012]; Pan & Niu [2011]; Radjaee et al. [2010]; Vinnik et al. [2004]; Wang et al. [2004]; Zhao et al. [2003]; Zorin et al. [2002]