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2 **Crustal density structure of NW Iranian Plateau**

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Crustal density structure of NW Iranian Plateau

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33 **Abstract**

34 We present a new 2D crustal-scale model of the NW Iranian plateau based on gravity-magnetic
35 modeling along the 500 km long CIGSIP seismic profile across major tectonic provinces of Iran
36 from the Arabian plate into the South Caspian Basin (SCB). The seismic P-wave receiver function
37 model along the profile is used to constrain major crustal boundaries in the density model. Our 2D
38 crustal model shows significant variation in the sedimentary thickness, Moho depth and the depth
39 and extent of intra-crustal interfaces. The Main Recent Fault between the Arabian crust and the
40 overriding Central Iran crust dips at $\sim 13^\circ$ towards the NE to a depth of ~ 40 km. The geometry of
41 the MRF suggests ~ 150 km of underthrusting of the Arabian plate beneath Central Iran. Our results
42 indicate the presence of a high-density lower-crustal layer beneath Zagros. We identify a new
43 crustal-scale suture beneath the Tarom valley between the South Caspian Basin crust and Central
44 Iran and the Alborz. This suture is associated with sharp variation in Moho depth, topography and
45 magnetic anomalies and is underlain by a 20 km thick high-density crustal root at 35-55 km depth.
46 The high density lower crust in Alborz and Zagros may be related to partial eclogitization of crustal
47 roots below ~ 40 km depth. The gravity and magnetic models indicate a highly extended continental
48 crust for the SCB crust along the profile. Low observed magnetic susceptibility at the Kermanshah
49 ophiolites probably implies that the ophiolite rocks only form a thin layer which has been thrust over
50 the sedimentary cover.

51 **Keywords:** 2D forward crustal modeling, Gravity and Magnetic anomalies, Receiver function,
52 Iranian plateau, Moho, Sediment thickness.

1. Introduction

The Iranian Plateau formed by long-lasting convergence and collision between the Arabian and the Eurasian plates. The interaction between these two major lithospheric plates and numerous microplates trapped between them created a broad zone of deformation with different type of structural units, similar to the tectonic style named Turkic-type (Şengör and Natalin, 1996). Tectonically, the region is divided into several major units, namely the Zagros fold and thrust belt, the metamorphic Sanandaj-Sirjan Zone (SSZ), the Tertiary Urmia-Dokhtar Magmatic Arc (UDMA), the Central Iran Zone, Alborz and the South Caspian Basin. **Figure 1** shows a proposed schematic cross section across the deformation zone, constructed based on commonly accepted views (e.g. Berberian and Berberian, 1981; Stocklin, 1968). The deformation style in this region is highly influenced by the evolution of the Tethyan oceans and collision of Arabian and Central Iran plates as evidenced by Tertiary magmatism associated with paleo-subductions and the presence of ophiolites, respectively. The different geotectonic units were mainly studied and categorized by numerous surface and near-surface geological surveys but the relationships between the units and deep crustal structures are not yet clear.

The geometry and density of intra-crustal layers of the Iranian crust are poorly studied and most of the seismic crustal studies focus on estimation of the Moho depth (e.g., Hatzfeld et al., 2003; Paul et al., 2010; Tatar and Nasrabadi, 2013). The thickness of the sedimentary cover is an important geophysical parameter. Previous geophysical interpretations of the lithosphere structure in Iran (Jimenez-Munt et al., 2012; Motavalli Anbaran et al., 2011; Tunini et al., 2015) have been carried out without exact knowledge of the geometry and composition of intra-crustal layers including the thickness of the sedimentary cover. These interpretations are along linear profiles and are mainly

75 based on integrated geophysical modeling of gravity, geoid, and topography together with the
76 sparsely distributed observations of surface heat flow. These profiles are unconstrained by
77 independent seismic observations from receiver function or active seismic sections. In the mentioned
78 models, much of the observed gravity signal is related to the shape of lithospheric-asthenospheric
79 boundary and, in some cases (Tunini et al., 2015), to assumed petrological heterogeneity in the
80 mantle lithosphere.

81 In this paper, we present a new high resolution crustal model of the NW Iranian plateau along a
82 profile coincident with the CIGSIP (China-Iran Geological and Geophysical Survey in the Iranian
83 plateau) seismic profile. The profile crosses all major tectonic units of the NW Iranian plateau
84 (**Figure 1**) providing an excellent opportunity to investigate the crustal structure of the tectonic units
85 and their relations. The 2D model includes sedimentary cover, and crustal interfaces and Moho as
86 constrained mainly by a receiver function seismic section (Chen et al., 2016). We model the density
87 and susceptibility distribution between the boundaries in order to improve our understanding of deep
88 crustal processes. In our adopted top-down modeling strategy, we calculate the effect of crustal
89 structure on the observed gravity signal and relate the residual to structure in the upper mantle.

90 Based on the modelled geometry, density and susceptibility of the crust, we discuss the following
91 questions: (1) How has the convergence between the Arabian and Eurasian plates been
92 accommodated in the crustal layers and how much shortening is accommodated in the lower crust?
93 (2) Has crustal thickening been accompanied by eclogitization of the lower crust beneath the deep
94 mountain roots? (3) What is the nature of the crust below the South Caspian Basin (i.e., oceanic or
95 continental)?, and (4) How thick are the exposed ophiolites along the suture between Arabian and
96 Central Iran plates ?

2. Geological Setting

The study region includes two orogenic systems, which developed in response to collision between the Arabian plate and the Eurasian plate. The Zagros-Makran orogenic system in the southern part of the Iranian plateau formed along the NE margin of the Arabian plate, whereas the Talesh-Alborz-Kopeh Dag mountains in the north of the plateau follow the boundary between the Central Iran Zone and the south Caspian basin of the Eurasia plate (e.g., Hatzfeld and Molnar, 2010).

The Zagros mountains preserve a record of the long-lasting convergence across the Neotethys from subduction to the present-day collision (~90 Ma to present, Agard et al., 2011). The Main Recent Fault (MRF) represents a Neotethian suture zone in the Zagros orogeny (Agard et al., 2011). The Zagros orogenic system from NE to SW includes three major parallel tectonic zones: the Urmia-Dokhtar Magmatic Arc, the Sanandaj-Sirjan metamorphic-magmatic zone, and the Zagros fold and thrust belt. (**Figure 2a**) (e.g., Manuel Berberian and King, 1981; Stocklin, 1968).

The Sanandaj-Sirjan metamorphic and igneous zone is a ribbon-like northwest trending belt, located immediately to the north of the Zagros suture (Stocklin, 1968) with a basement consisting of metamorphic, igneous and sedimentary rocks (e.g., Richards and Şengör, 2017). The initial subduction of the oceanic slab of Neotethys beneath Central Iran formed the Andean type Sanandaj-Sirjan zone. After a possible first slab break-off at ~60-55 Ma, the subduction of the Neotethys oceanic slab continued at a lower angle and formed the Andean type Urmia-Dokhtar magmatic arc to the north of the Sanandaj-Sirjan Zone. The low angle subduction and its associated volcanism was mainly active up to Oligo-miocene time (~35-25 Ma), the approximate collision time of the Arabian plate with Central Iran (e.g., Agard et al., 2011).

118 The Zagros fold and thrust belt is generally regarded as one of the youngest collisional belts on
119 Earth, hosting considerable oil and gas resources. There is general agreement that deformation is
120 mainly accommodated by folding in the sediments and thrust faulting in the crystalline basement
121 and sedimentary cover, with a high level of seismicity (e.g., McQuarrie, 2004; Tatar et al., 2004).
122 The Zagros fold and thrust belt is separated from the Mesopotamian foreland basin by a main
123 structural uplift of several kilometers along the Main Frontal Fault (MFF). Our profile crosses the
124 NW part of Zagros named the Lurestan arc (**Figure 1a**). It is located to the north of the high
125 resolution receiver function Zagros 2003 profile (Motaghi et al., 2017), which detected the presence
126 of a low angle seismic interface to a depth of about 17 km within the Lurestan arc. Motaghi et al.,
127 (2017) speculate that seismic deformation in the Lurestan arc is mainly accommodated by low angle
128 thrust faulting along this intra-crustal seismic interface.

129 The collision of the Cimmerian continental fragment with the Eurasian plate in the Late Triassic to
130 Early Jurassic formed the Talesh-Alborz mountain ranges along the paleo-Tethys suture zone
131 (**Figure 1a**). The Alborz mountain range is situated between the South Caspian Basin (SCB) and
132 the central Iranian block. The Alborz orogenic belt is narrow (120–150 km wide) and it is commonly
133 accepted that the orogeny is bounded by south and north-dipping thrust faults at its northern and
134 southern flanks (e.g, Allen et al., 2003; Guest et al., 2007; Şengör et al., 1984).

135 It is difficult to assess the active faulting along the northern margin of the Alborz Mountains due
136 to the poor basement exposure. The so-called Khazar fault is generally interpreted as a transition
137 zone between the Central Iran plate and the SCB. Tatar et al. (2007) interpret south dipping thrust
138 faulting on the Khazar fault and underthrusting of the South Caspian basin beneath the Alborz
139 Mountains during the 2004 Baladeh 6.2 Mw earthquake (Tatar et al., 2007). However, four seismic

140 receiver function sections and the seismicity pattern along the Caspian Sea have not identified
141 evidence for underthrusting of the SCB beneath central Iran along the Khazar fault (Ghods et al.,
142 2016).

143 Within the study area, the major zones of seismicity are in Zagros and Alborz. The deformation of
144 the upper crust in the study area has resulted in thrust and strike-slip faulting (Talebian and Jackson,
145 2002) (Figure 1). The most aseismic blocks are the Sanandaj-Sirjan Zone and the interior of SCB.
146 Seismicity in northern Iran is limited mainly to a belt along the Alborz mountain belt, where seismic
147 activity occurs primarily in the upper crust at depths shallower than 15 km depth. The South Caspian
148 basin has almost no seismicity and is considered as a rigid block (e.g., Engdahl et al., 2006; Jackson
149 et al., 2002)(**Figure 1a**). Relocation of seismicity in the western part of SCB does not support
150 underthrusting of the SCB beneath the Talesh mountains (Aziz Zanjani et al., 2013).

151 The Central Iran tectonic block is located at a roughly triangular area in the middle of the Iranian
152 plateau, bordered by the Alborz mountains in the north and the Sanandaj–Sirjan zone in the south-
153 southwest (e.g, Stocklin, 1968). It includes orogenic belts, metamorphic zones, magmatic arcs and
154 sedimentary basins. The block separated from Gondwana in the Late Paleozoic and was accreted to
155 Eurasia in the Mesozoic during the Cimmerian collisional events (e.g., Guest et al., 2007; Zanchi et
156 al., 2009). Aseismic zones surrounded by seismic zones are characteristic of this region (Engdahl et
157 al., 2006).

158 The origin of the South Caspian Basin is controversial and it is interpreted as a remnant of
159 Paleotethyan oceanic crust, a trapped remnant of early Mesozoic oceanic crust, a Cretaceous to
160 Paleogene strike-slip-related continental pull-apart basin, or a Middle to Late Jurassic marginal basin
161 of the Tethys ocean (Guest et al., 2007 and references therein). Seismic studies suggest a more than

15-25 km thick sedimentary cover above a 15-18 km thick mafic crystalline basement (e.g., Brunet et al, 2003).

3. Datasets

Our density and susceptibility model is based on geological and tectonic mapping, depth to magnetic basement and a P-wave receiver function model of the main crustal interfaces. The available geological and geophysical data are used in the modeling process to construct the geometry of the initial model, including thickness of the sedimentary cover, the depth to intra-crustal interfaces and the Moho.

P-wave receiver function model: The depth to the Moho is constrained by the seismic receiver function profile (Chen et al., 2016) along our profile. The broadband temporary CIGSIP seismic project deployed 63 stations, which operated from October 2013 to October 2014. The NE-SW trending seismic array includes three linear parallel seismic lines from the south of the Zagros Fold and Thrust Belt to the coast of South Caspian Sea. The main line of the array is ~550-km long and includes 46 stations with a station spacing of 10-15 km.

Teleseismic data recorded by the CIGSIP temporary linear array image the subsurface velocity interfaces. P receiver functions were calculated by analysis of teleseismic events recorded within epicentral distance of 25°-95° and magnitude ≥ 5.5 . The ZNE waveforms were windowed from 50 s before and 100 s after the onset of direct P phase and rotated into the ZRT coordinate system. A time domain maximum entropy deconvolution method (Wu and Zeng, 1998) was used to eliminate the source effects and isolate P to S conversions on the R component. A Gaussian parameter of 5 and a water level of 0.0001 were adopted in the deconvolution. The resultant receiver functions were

183 further bandpass filtered with corner frequencies of 0.03 Hz and 1 Hz to remove the high-frequency
184 noise.

185 Along the profile, the tectonic units are bounded by major thrust and strike-slip faults. The Main
186 Recent Fault (MRF) is the suture zone between the Arabian plate and the central Iranian micro-
187 plates (Berberian, 1995). Figure 3 shows all positive and negative convertor interfaces extracted
188 from the RF profile (Chen et al., 2016). The RF profile shows a strong northward dipping negative
189 impedance interface at a distance range of 150 - 300 km which continues from the surface trace of
190 the MRF to a depth of ~40 km, implying about 150 km of underthrusting of the Arabian crust
191 beneath Central Iran. We interpret the low velocity associated with the negative impedance contrast
192 as evidence for a shear zone of the MRF. The dipping interface is similar to the MRF fault zone
193 interpreted by Paul et al., (2010) along the Zagros 2003 seismic profile in the southern part of the
194 Lurestan arc and suggests underthrusting of the Arabian plate under the southwestern edge of the
195 Iranian Plateau. Similar negative conversion interface has been also observed in Himalaya (Nábělek
196 et al., 2009).

197 Two positive impedance interfaces are observed in the shallow crust of the Zagros belt. The first
198 near-surface seismic interface (8 to 10 km) is nearly flat beneath the entire Lurestan arc (**Figure 3**).
199 This boundary is interpreted as the base of the sedimentary cover, in agreement with geological
200 studies (Malekzade et al., 2016; Vergés et al., 2011b). Another flat intra-crustal seismic interface is
201 observed at ~17 km depth to the south of the suture zone (~150 km profile distance). A similar
202 feature has been reported along another profile (i.e., Zagros 2003 profile (Paul et al., 2010) to the
203 south of Lurestan arc (Motaghi et al., 2017) as a significant crustal interface between the upper and
204 middle crust. The recent 7.3 Mw Sarpolzahab (**Figure 1a**) low angle thrust earthquake (12

205 November 2017) was located at the same depth as this interface.

206 The RF section (**Figure 3**) shows a gradual crustal thickening from ~50 km in the southern end of
207 the profile to its maximum thickness beneath the MRF (~65 km), which demonstrates substantial
208 crustal thickening in the Zagros orogen. From the SSZ to the Alborz Mountains, the Moho depth
209 decreases gradually from ~55 km to ~45 km. In the middle of the Alborz Mountains or more
210 precisely at the northern boundary of the Tarom valley, the Moho boundary steps from ~45 km to
211 ~30 km (Chen et al., 2016) or to 40-55 km as indicated in our study. The sudden jump of the Moho
212 around a distance of 500 km may be related to the boundary between SCB and the Central Alborz
213 crust (Ghods et al., 2016). The RF section shows a deeper, weak, positive impedance boundary
214 beneath the Tarom valley.

215 **Sediment thickness:** A significant part of the Iranian plateau is covered by a thick sedimentary
216 cover in the Zagros, Kopeh Dagh and Makran (Teknik and Ghods, 2017). Aeromagnetic data can
217 provide a proxy for the thickness of sedimentary cover by estimates of the depth to the magnetic
218 basement based on the radially averaged power spectrum method (e.g. Maus and Dimri, 1995). This
219 approach assumes that the sedimentary cover is much less magnetized than the crystalline basement
220 (**Figure 2b**) (Teknik and Ghods, 2017).

221 The thickness of the sedimentary cover can also be estimated by the RF method. The basement is
222 expectedly the shallowest strong, positive impedance interface in the seismic receiver function
223 section, because sediments have lower seismic velocity than the basement. In our approach we
224 constrain the thickness of the sedimentary cover as the depth to the shallowest seismic interface in
225 the RF model (Chen et al., 2016; Paul et al., 2010), the depth to magnetic basement (Teknik and
226 Ghods, 2017), and the thickness of the stratigraphic column as derived from basin analysis studies

227 (Figure 2b) (e.g., Vergés et al., 2011a).

228 **3.3. Gravity data:** We use Bouguer anomaly data over the study area as retrieved from the world
229 database of BGI (Bureau Gravimétrique International), which has sparse coverage in Iran. The data
230 coverage is sufficiently dense for regional-scale tectonic studies. The main feature of the gravity
231 field is a wide, long wavelength negative Bouguer anomaly (minimum of -150 mGal) trending
232 parallel to MRF with a NW–SE trend and centered roughly 150 km NE of the MRF. At the northern
233 margin of the Central Iranian block, the gravity field increases smoothly reaching up to 50 mGal on
234 the northern side of the Alborz Mountains (Figure 2c).

235 **3.4. Aeromagnetic data:** To identify the major tectonic units of Iran, an aeromagnetic survey was
236 conducted by Aeroservice (Houston, Texas) during 1974-1977 under the auspices of the Iranian
237 Geological Survey. We use the composite aeromagnetic 1 km × 1 km grid calculated from the
238 original raw data (Saleh, 2006). The data are reduced to the pole using the variable RTP method
239 (Paterson Grant & Watson Limited (PGW), 1973) (Figure 2d).

240 Along our profile, the aeromagnetic anomaly shows insignificant variation over the Zagros fold and
241 thrust belt and sharp variation above the MRF fault zone, which probably is caused by strongly
242 magnetized ophiolites near the suture zone and/or surface displacement associated with MRF. The
243 pattern of magnetic anomalies in the SSZ unit may be caused by a relatively deep-seated igneous
244 intrusion. It is widely accepted that igneous rocks are present as intrusions in the sedimentary rocks
245 in the SSZ and Central Iran block (e.g., Verdel et al., 2011). Near Qorveh city, an anomaly named
246 here as the Qorveh anomaly (QA) indicates the presence of a possible intrusion of igneous rocks
247 into or below the sedimentary cover. We consider this anomaly to be caused by a single intrusion
248 in our susceptibility model. Magnetic anomalies show strong variation in the middle of the Alborz

249 Mountains, which we interpret as the border between the Central Iran and the South Caspian block
250 (**Figure 4**).

251 Depth variations in the Curie isotherm are usually identified by long wavelength magnetic anomalies
252 because of a deep source. Long wavelengths (greater than 150 km) magnetic anomalies reveal two
253 consecutive pairs of positive-negative anomalies, which may reflect Curie isotherm depth variations,
254 lateral crustal susceptibility variations, or a combination of both. The Tertiary magmatic arc near
255 the Alborz Mountains is associated with a broad positive magnetic anomaly, which suggests a
256 combined effect of near-surface high susceptibility magmatic rocks and the Curie depth variations
257 at depth. The origin of some of the observed anomalies may have alternative explanations. For
258 instance, positive magnetic anomalies PM1 and PM2 can be associated with either a slightly deep
259 Curie isotherm or the presence of a wide body of high susceptibility material at shallow depth. The
260 NM1 and NM2 negative magnetic anomalies can also be explained by the presence of a thin
261 magmatic body or by a shallow wide susceptibility rock body in the crust (**Figure 4**).

262 **4. Modeling**

263 We apply constrained forward modeling of the crustal density and susceptibility distribution along
264 a profile in NW Iran. The two dimensional (2-D) model is constrained by the geometry of crustal
265 layers derived from the seismic receiver function study (Chen et al., 2016) along the profile. The
266 best fitting model is determined by iterative fine-tuning of parameters within the frame of all
267 available geological and geophysical constraints until the model response optimally fits the observed
268 gravity and magnetic anomalies. In Appendix A, we present complementary tests to assess the
269 uniqueness of the model.

270 **Density Modeling:** The 2D forward density model was constructed by using additional constraints
271 in a stepwise manner to reduce the uncertainty of the model, and to assess the effect of each model
272 parameter on the misfit. The first step includes determination of the geometry of the expectedly
273 strongest density contrasts at the base of the sedimentary succession and at the base of the crust. The
274 basement depth is determined as described above. The geometry of the Moho is initially constrained
275 by the receiver function profile and was later fine-tuned by the long wavelength Bouguer gravity
276 signal in order to reduce uncertainties caused by depth conversion and migration of the RF section.
277 These uncertainties originate primarily from the velocity model that is based on surface wave noise
278 tomography (Jiang et al., 2016).

279 For the density of crustal units, we used a range of density values as obtained from global crustal
280 velocity studies and the Nafe-Drake relationship between velocity and density. Global and regional
281 crustal studies (e.g. Artemieva and Thybo, 2013; Rudnick and Fountain, 1995) indicate that
282 continental crust is highly heterogeneous in three dimensions. However, it is useful to divide the
283 crust into the sedimentary cover and upper, middle, and lower crystalline crust (e.g. Lyngsie et al.,
284 2006; Thybo, 2001). These layers have distinctively different densities. Average density, ρ , of
285 sedimentary cover, upper crust, middle crust and lower crust are $\rho < 2650 \text{ kg/m}^3$, $\rho < 2800 \text{ kg/m}^3$,
286 $2800 < \rho < 2900 \text{ kg/m}^3$, and $2900 < \rho < 3000 \text{ kg/m}^3$, respectively. Average density of high-velocity
287 lower crustal layers observed in some tectonic regions is $\rho > 3000 \text{ kg/m}^3$. Average density of the
288 upper mantle is larger than 3200 kg/m^3 (Artemieva and Thybo, 2013).

289 To construct our density model, we fixed the Moho interface at the base of the model and thickness
290 of sedimentary cover at the top. On this basis, the crust is divided into smaller blocks based on intra-
291 crustal boundaries observed in the seismic model. Finally, the density of each block is adjusted to

292 improve the fit. In our final model, we do allow for reduction of density with depth if it is required
293 by the RF section. The density modeling includes the following steps:

Step 1: Density is calculated for a simple model with three layers representing the sedimentary cover, crystalline crust, and upper mantle. The base of the sedimentary cover is defined as the shallowest positive interface in the RF section (Chen et al., 2016) (**Figure 3**). The depth of magnetic basement (Teknik and Ghods, 2017) is in overall agreement with the depth of sedimentary cover extracted from the RF section, although it is shallower over the Kermanshah ophiolite outcrops and considerably deeper in the western part of the Lurestan arc.

The initial density model has a homogeneous sedimentary cover, crystalline crust, and upper mantle with assumed densities of 2450, 2820 and 3270 kg/m³, respectively. The shape of the calculated gravity Bouguer anomalies is similar to the observed data but with a large misfit. The root mean square (RMS) difference between the observed data and the calculated response is ~85 mGal. The calculated gravity response of the model has a broad and strong negative gravity anomaly in the Zagros, the SSZ, and Central Iran, while in the northern part of the profile the calculated gravity is considerably larger than observed.

Step 2 (Figure 5): The RF model includes two flat and positive converters below the sedimentary cover beneath the Zagros part of the profile at depths of 8-10 km and 16-18 km. The seismic interface at 8-10 km depth is considered as the base of the metamorphosed sedimentary cover. In this step, we assume that the seismic interface at ~17 km depth separates the upper and middle crust. The upper and middle crust are assigned density values of 2750 and 2800 kg/m³. The positive converter in the RF model at depth ~35 km (profile distance 100 -250 km) is assumed to separate the middle and the lower crust. Assigning a density value of 3050 kg/m³ to the lower crust of the

Zagros fold and thrust belt (ZFTB) decreases the misfit between the computed and the observed gravity.

294 The RF model suggests two possible depths for the Moho boundary around the Tarom valley. The
295 depth of the shallower, high-amplitude boundary interface (shown by number 1 in **Figure 5**)
296 decreases sharply from 50 km to 30 km under the Tarom valley. The second, less distinct, converter
297 (shown by number 2 in **Figure 5**) represents an alternative converter for the Moho. We find that
298 assigning a high crustal density of 3100 kg/m^3 to the lower crust between converters 1 and 2 in this
299 region decreases the calculated anomaly to the level of the observations and, therefore, reduces the
300 misfit significantly. Thus, we consider the deeper boundary as the Moho. The structural differences
301 associated with the sharp variations in the topography and aeromagnetic anomaly (**Figure 4**)
302 suggests the presence of a transition zone between Central Iran and South Caspian Basin around the
303 Tarom valley.

The RMS misfit between the calculated and the observed gravity anomaly is reduced to around 24 mGal. In the central part of the profile, the calculated response is less than the observation and thus requires a denser lower crust in parts of Zagros, Central Iran and SCB.

304 Step3 (**Figure 6**): In this stage, we consider the compaction of the sedimentary cover with depth.
305 Sedimentary layers deeper than 5 km depth are expected to be highly compacted, leading to a low
306 density contrast at the sedimentary-crystalline rocks interface (e.g. Chakravarthi, 1995; Rao, 1986).
307 Therefore, we assign a density of 2400 kg/m^3 to sediments shallower than 5 km and 2550 kg/m^3 to
308 sediments deeper than 5 km. We also introduce a high density (2950 kg/m^3) lower crustal layer for
309 Central Iran and SCB.

310 We speculate that the underthrusting of Zagros lithosphere beneath the Iranian crust has produced

311 downward bending of the upper crust of Zagros to the extent that it is juxtaposed below the middle
312 crust of the Iranian crust. This scenario would explain the observed strong, dipping, negative
313 converter related to MRF (**Figure 3**). In this stage of modeling, we further modify our model to
314 consider the downward bending of the Zagros upper crust (**Figure 6**). We also model a possible low
315 velocity (and density) zone in a more straight way by introducing a low density area
316 (**supplementary material, Figure A7**), but because the width of the low velocity zone is small
317 (about 7 km), the gravity anomalies are not sensitive to the presence of the low density area.

318 Beneath the Tarom valley, the negative localized converter at around 500 km together with the steep
319 local gravity high indicate the presence of a shallow high density body, which may have its origin
320 as a magmatic intrusion or an ophiolite that is today hidden beneath the sedimentary cover. The
321 density of the shallow high density Caspian block is estimated to be 2950 kg/m^3 , which is higher
322 than normal upper crustal values. The Bouguer gravity map (**Figure 2c**) shows a high amplitude
323 short wavelength Bouguer anomaly along the south and southwest borders of Caspian Sea. We
324 speculate that this unusually dense body in the upper crust may be related to mafic rocks associated
325 with an ancient rift system. The three mentioned modifications of the model have reduced the RMS
326 misfit to $\sim 6.5 \text{ mGal}$.

327 In the above three steps, we did not include any lateral or depth variation of density in the upper
328 mantle and instead assigned an average density of 3270 kg/m^3 to the upper mantle. Our density
329 model has a very low RMS misfit which suggests that most of the Bouguer gravity signal may be
330 related to crustal anomalies. Joint teleseismic-regional travel time tomography of lithospheric
331 structure in NW Iran indicates that the pattern and amplitude of crustal Pn residuals are very similar
332 to those obtained from teleseismic P travel times ((Alinaghi et al., 2007; Bavali et al., 2016; Rahmani

et al., 2018). This implies that crustal heterogeneities are responsible for a significant part of the teleseismic P travel time residuals, thus supporting our modelling result.

Magnetic Susceptibility modeling: On crustal scale, the magnetic crust can be simplified into three main layers based on the generally accepted range of susceptibility values: 1) near-surface magmatic intrusions and ophiolites, 2) the sedimentary layer with insignificant susceptibility, and 3) the magnetic crust from the bottom of the sedimentary cover to the Curie Depth Point (CDP) with an average susceptibility of 0.035 SI (e.g., Clark and Emerson, 1991; Hunt et al., 1995). CDP is the depth at which magnetite, the dominant magnetic mineral in the deep crust, loses its magnetization at ~580°C. We apply forward modeling to estimate CDP instead of the widely used spectral methods (e.g. Bouligand et al., 2009; Maus and Dimri, 1995), which may lead to serious errors (Teknik and Ghods, 2017). The susceptibility of the sedimentary strata is not sensitive to compaction and thus the susceptibility contrast at the interface between sediment and crystalline rocks is not sensitive to the depth of the interface. The insignificant susceptibility of the sedimentary cover allows use of magnetic data to map the magnetic basement topography.

At the first step of forward magnetic modeling, we fix the thickness of sedimentary cover based on the results of the density modeling. A constant low susceptibility value of 0.01 SI is assigned to the sedimentary cover. We also assume a constant CDP at 40 km depth to detect the presence of significant near surface intrusions and ophiolites (**Figure 7**). The CDP depth only affects the long wavelengths of the observed magnetic anomalies whereas the shallow magnetized sources mostly control the short to medium wavelengths of the magnetic anomalies.

The high amplitude magnetic anomaly in the southern hills of Alborz (labeled as TMA in **Figure 4 and 7**) corresponds to the Tertiary (i.e., Eocene) Alborz magmatic arc belt. The shape of the magnetic

355 anomaly correlates with the ~150 km wide surface outcrop of volcanic rocks (**Figure 2a**) as mapped
356 in 1:100 000 scale geological maps of the Geological Survey of Iran (Nogole-Sadat, M.A.A. and
357 Almasian, 1993). A susceptibility value of 0.175 SI is assigned to this near surface volcanic unit.
358 The next short wavelength magnetic anomaly is within the Bijar sedimentary basin in the SSZ
359 (Teknik and Ghods, 2017) and suggests the presence of a hidden magnetized intrusion at the Qorveh
360 anomaly, beneath the sedimentary cover (**Figure 2d** and **7**). We assign a susceptibility value of
361 0.095 SI to this intrusion. A smaller amplitude magnetic anomaly near the MRF (**Figure 4** and **7**)
362 corresponds to the Kermanshah ophiolites, which we assign a susceptibility of 0.065.

363 The three anomalous bodies corresponding to the Kermanshah ophiolite, Qorveh intrusion, and the
364 Tertiary magmatic arc of Alborz (**Figure 7**) provide a close match to the general magnetic profile,
365 although some long wavelength misfit between the observed and the calculated anomaly remains
366 (highlighted with a thick gray line in **Figure 7**) with peak values between ~-100 and 150 nT. The
367 long wavelength of the residuals implies a deep source such as variations in the depth to the CDP
368 isotherm. The negative correlation between the residual curve and the long wavelength part of
369 magnetic anomalies (compare **Figure 4** with **Figure 7**) suggests that the residuals represent variation
370 in the depth to the Curie temperature, although there may be other sources involved. Because of the
371 uncertainties, we do not interpret the variations of CDP in full detail, and instead we mainly model
372 the lateral variation of the total susceptibility of the crystalline crust. The final magnetic model
373 shows a close correlation between the data and the calculated anomalies.

374 **Modeling uncertainties and non-uniqueness of the results:** The gravity and magnetic responses
375 of the model are sensitive to the geometry of the intra-crustal interfaces used in the model. A
376 combination of different density and magnetic susceptibility with different geometries of intra-

377 crustal interfaces can produce similar gravity or magnetic responses. Uncertainties in the geometry
378 and densities of intra-crustal layers generally increase with depth. We have reduced the geometry-
379 related uncertainties by using independent constraints on the geometry from a seismic receiver
380 function model and the observed near surface structures, including distribution of magmatic
381 intrusions, suture zones and thickness of sedimentary cover. Probably the most important
382 uncertainty of the 2D model is the presence of large 3D lateral and depth density and susceptibility
383 variations. However, our profile is perpendicular to the major tectonics trend, which reduces this
384 uncertainty, and we assume insignificant density and susceptibility variations normal to the profile
385 direction. In Appendix A, we present complementary tests to assess further the uniqueness of the
386 model presented in **Figure 6**. We test the sensitivity of the density model in relation to key
387 parameters, including the Caspian dense shallow body, the upper, middle and lower crust, Zagros
388 root and Alborz root. The sensitivity analyses were undertaken to understand how density variations
389 within plausible ranges may affect the response of the density model. The tests prove that the model
390 presented in **Figure 6** is the most reliable density model.

391 **5. Results and discussions.**

392 We present a 2D-crustal density and magnetic susceptibility model in NW Iran. The model includes
393 thickness of sedimentary cover, depth to intra-crustal interfaces and the Moho, Curie depth point,
394 and variation in crustal density and magnetic susceptibility.

395 **Sediment thickness:** The depth to magnetic basement is in overall agreement with the thickness of
396 the sedimentary cover interpreted from the RF section (**Figure 3**). Our results indicate that the
397 thickness of sediments in the ZFTB is ~8 km. This value is in agreement with estimates from

398 structural geological studies (e.g., Vergés et al., 2011a, 2003), but in sharp disagreement with the
399 depth to magnetic basement calculated by the fractal spectral analysis method (Teknik and Ghods,
400 2017)(**Figure 2b**). We speculate that the ~17 km depth to magnetic basement in the Lurestan arc
401 (distances 0-30 km in **Figure 2b** and **Figure 3**) is related to possible intense upper crust deformation
402 associated with the low angle thrusting in the region (Motaghi et al., 2017) which may have
403 destroyed the magnetization of the upper crustal layer.

404 Ophiolites are highly magnetized rocks compared to sedimentary rocks and one expects to have an
405 almost zero depth to magnetic basement over ophiolite outcrops. Depth of magnetic basement is ~
406 4 km over the outcrops of Kermanshah ophiolites (Compare **Figure 1a** with **Figure 2b**). Based on
407 this observation, we speculate that the Kermanshah ophiolites constitute a very thin layer of highly
408 magnetized rocks which has been thrust over a ~4 km thick sedimentary cover (Chen et al., 2016;
409 Teknik & Ghods, 2016). The recent 3D Pg velocity tomography map of NW Iran (Maheri-Peyrov
410 et al., 2018) also show slow Pg velocity (i.e., not in the range of velocity of upper crustal crystalline
411 rocks) of the upper crust within the region around Kermanshah ophiolites. This indirectly implies
412 that the high P velocity ophiolitic rocks are too thin to increase the average velocity of the
413 sedimentary cover.

414 To the NE of the MRF, the thickness of the sedimentary cover from the RF section (Chen et al.,
415 2016) is in agreement with the depth to magnetic basement (Teknik and Ghods, 2017). The 2D
416 density model (**Figure 6**) shows the presence of the huge Bijar sedimentary basin to the NW of MRF
417 and shows that this basin has a maximum thickness of the sedimentary cover of ~12 km thick. Based
418 on the map of depth to magnetic basement in Iran (Teknik and Ghods, 2017), the Bijar sedimentary
419 basin trends NW-SE and covers a 100 by 300 km large region (**Figure 2b**). From the margin of the

420 newly discovered Bijar basin to the Tarom valley (profile distance 350-600 km), the thickness of
421 the sedimentary cover decreases to less than 7 km (**Figure 2b** and **Figure 3**). The thickness of
422 sedimentary cover from the Tarom valley to the NE end of the profile gradually increases to a
423 maximum value of 15 km at the Caspian Sea, in agreement with the previous studies (Brunet et al.,
424 2003; Mangino and Prestley, 1998).

425 **Main Recent Fault (suture Zone):** The Main Recent Fault (MRF) is traced at depth as a
426 northeastward low angle dipping interface in the RF section (**Figure 3**). The negative converter in
427 the RF model indicates a low P-wave velocity below the fault zone implying the presence of a shear
428 zone at the MRF. The interpreted MRF has a gentle dip of $\sim 13^\circ$, and it extends to a depth of ~ 35
429 km over ~ 150 km horizontal distance away from its surface trace. The gravity data has insufficient
430 resolution to identify a possible low/density within the shear zone, and the magnetic data does not
431 indicate any magma migration along the shear zone and identified in other locations (Lyngsø and
432 Thybo, 2007).

433 **South Caspian basin crust:** The 2D density model (**Figure 6**) includes a thin SCB crystalline
434 crust (20 to 25 km) in the NE of the profile, which is overlain by a 15 km thick sedimentary cover.
435 The interior of the SCB has been proposed to be formed by an oceanic-like crust (Mangino and
436 Prestley, 1998). Our density model shows that the thin crystalline crust of SCB has an average
437 density of 2800 kg/m^3 and thus suggests a thinned and extended continental type of crust for the part
438 of SCB crust in NE of the profile. We find that assigning a typical oceanic crust density to the SCB
439 crust leads to a calculated Bouguer gravity anomaly, which is about 100 mGal larger than observed.
440 Furthermore, the low magnetic anomaly (**Figure 4**) over the SCB part of the profile and the low
441 susceptibility of the SCB igneous crust (**Figure 7**) also do not support an oceanic origin. Oceanic

442 crust is much more magnetized than continental crust (e.g. Clark and Emerson, 1991; Hunt et al.,
443 1995).

444 **Moho step beneath the Tarom valley:** The depth to the strongest converter abruptly decreases
445 from ~50 km in Central Iran to ~30 km in the SCB (distance profile of ~480 km) which could suggest
446 a steep change in Moho depth. Vertical Moho steps are generally expected to indicate major faults
447 or sutures (e.g., Allam et al., 2017; Schulte-Pelkum and Ben-Zion, 2012). The suggested abrupt 20
448 km vertical offset of the Moho is in agreement with the variations in magnetic anomalies and
449 topography (**Figure 4**) and the depth of seismicity (Aziz Zanjani et al., 2013; Ghods A. et al., 2016).
450 The 80 km long surface rupture of the destructive Rudbar-Tarom 7.3 Mw 1990 event (Berberian
451 and Walker, 2010) is located very close to the vertical offset of the Moho. We suggest that the sharp
452 vertical offset of the strong converter is related to a suture zone between the northern limit of Alborz
453 and the SCB crust, in contrast to previous studies which interpret the Khazar fault as the transition
454 zone between the South Caspian basin and the Central Iran blocks (e.g, Allen et al., 2003; Guest et
455 al., 2007; Şengör et al., 1984). The RF section (Chen et al., 2016) does not confirm any evidence of
456 major faulting along the proposed Khazar fault similar to the negative dipping converter observed
457 for the MRF (Chen et al., 2016; Paul et al., 2010).

458 We speculate that the converter step may indicate a Moho step which represents a Paleotethyan
459 suture zone. Remnants of the Paleotethys suture was first recognized in the NE of the Iranian plateau
460 (Alavi, 1991) and lithological similarities suggest that the suture continues towards central and
461 western Alborz near the Rasht ophiolitic unit (Alavi, 1996). Rossetti et al. (2017) suggest that the
462 metamorphic rocks of the Rasht ophiolites form part of the Paleotethyan suture in Iran and can be
463 interpreted as an exhumed subduction complex formed during Early Carboniferous oceanic

464 subduction and subsequent Permian–Triassic suturing. Additionally, our density model indicates the
465 presence of a body with intermediate density between normal lower crust and mantle at the step.
466 This indicates that the present Moho is at the deep, weak converter around 55 km depth, but does
467 not rule out that the original Moho has been offset in relation to tectonics during the Tethys
468 collisions.

469 **High density lower crust:** Based on seismological and petrological studies, the lower crust is
470 globally composed of high-density mafic rocks with a density range of $2900 < \rho < 3000 \text{ kg/m}^3$
471 (Christensen and Mooney, 1995; Rudnick and Fountain, 1995). Generally, the trade-off between
472 Moho depth and lower crustal density leads to non-uniqueness of these parameters. However, the
473 ambiguity for the density of the lower crust is reduced by the a' priori constrained Moho depth from
474 the receiver function model. The relatively large Bouguer anomaly above the thick crust in the ZTFB
475 requires the presence of an anomalously high-density lower crust (3050 kg/m^3). Likewise, high
476 crustal density of 3100 kg/m^3 is required below the Alborz Mountains (**Figure 5** and **Figure 6**). The
477 very strong positive converter at depths around ~40 km beneath Zagros and Alborz implies a lower
478 crustal layer with very high velocity. The high density layer may reflect partial eclogitization of the
479 lower crust as has been suggested for Himalaya (Hetényi et al., 2007; Schulte-Pelkum et al., 2005)
480 and other active and former orogenies (e.g., Abramovitz and Thybo, 2000; Sobolev and Babeyko,
481 2005),

482 The Zagros 2003 seismic profile (Paul et al., 2010) in the southern part of the Lurestan arc led to
483 similar speculations about the actual Moho depth. The strong positive converter extends a distance
484 of about 150 on the tip of the Arabian plate and extends to depths larger than ~40 km. The maximum
485 thickness of the high velocity lower crust is about 20 km which is very similar to our observation.
486 This could indicate that the high density lower crust may be present everywhere where the Arabian

487 plate descends to depths deeper than 40 km. However, the Zagros 2001 profile (Paul et al., 2010) in
488 Central Zagros does not have sufficient resolution for identification of the strong positive lower
489 crustal converter.

490 **Deeper density variations:** Different researchers using different methodologies suggest thicker
491 lithosphere beneath Zagros than the rest of the Iranian Plateau (e.g. Hafkenscheid et al., 2006; Paul
492 et al., 2010; Priestley et al., 2012) and others (e.g. Hafkenscheid et al., 2006) also suggest deeper
493 high velocity regions as relicts of the subducted Arabian plate. To investigate fully the style of
494 deformation due to the continental collision in Iran, we should calculate the depth to the lithosphere-
495 asthenosphere boundary (LAB) along our profile. However, our final density model explains the
496 entire gravity signal which precludes calculation of variation in depth to the LAB.

497 **6. Conclusions**

498 We have documented and discussed a new gravity and magnetic susceptibility model of the NW
499 part of the Iranian plateau along a recently acquired seismic profile across the Lurestan part of
500 Zagros, Sanandaj-Sirjan Zone (SSZ), the Tertiary magmatic arcs, the Central Iran block, the Alborz
501 Mountains, and the South Caspian Basin (**Figure 8**).

502 The thickness of the sedimentary cover is ~8 km in the Zagros fold and thrust belt (ZFTB) and the
503 central Iran block, ~12 km in the Sanandaj-Sirjan Zone (SSZ) including the recently recognized
504 Bijar basin (Teknik and Ghods, 2017), and ~15 km in the South Caspian basin. The magnetic
505 susceptibility model indicates that the Tertiary magmatic arcs at the northern edge of Central Iran
506 are extruded or thrust over the sedimentary cover, the Kermanshah ophiolites form only a thin layer
507 over a thick sedimentary cover, and the Qorveh intrusion is located within the Bijar sedimentary
508 basin.

509 Around the Tarom valley a ~20 km thick high density lower crust is detected at the base end of the
510 hitherto unidentified Palaeotethys suture. Also the lower crust below the Main recent fault has high
511 density, indicative of partial eclogitization of the lower crust or the presence of a magmatic
512 underplate. The crystalline crust of the ZFTB, the SSZ and the CI includes three layers, with a
513 uniform thickness of ~20 km in the middle crust. The thickness of the upper crust increases from
514 ~10 km in the ZFTB to ~15 km in the Central Iran block. The thickness of the high-density lower
515 crust varies from ~15 in the ZFTB to ~25 km at the MRF. The SCB has a ~38 km thick crust, which
516 include a ~15 km thick sedimentary cover, a ~18 km thick upper crust and a ~5 km thick lower crust.
517 Neither density nor magnetic susceptibility data support an oceanic type of crust in this region. We
518 propose the existence of a Palaeotethys related suture located ~50 km south of the Khazar fault
519 between the Alborz Mountains and the South Caspian Basin. Our model provides a new tectonic
520 framework for geodynamic understanding of the region and changes the widely accepted view about
521 the borders of South Caspian Basin along the Khazar fault.

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List of figure captions

Figure 1: a) Topography of the Iranian plateau based on the ETOPO 1 global elevation model (Amante and Eakins, 2009). The solid straight line (AA') shows the location of the profile along which we model the crustal density and susceptibility distribution. Major active faults (Hessami et al., 2003) are shown as solid black lines. The thick grey, solid lines show political boundaries. Colored dots show earthquake epicenters of earthquakes between 1964 and 2017 on the Iranian Plateau (Engdahl et al., (2006) and, for the period of 1918-2004, new data from Iranian Seismological Center website (irsc.ut.ac.ir/bulletin.php; last download March 2017)). To assure a reasonable location accuracy, only events with azimuthal gap less than 120 degree are selected. The red star shows the location of the recent 7.3 Mw Sarpolzahab low angle thrust earthquake on 12 November 2017. The beachballs are from CMT Global focal mechanisms solutions (Ekström et al 2012) for events larger than magnitude 6 that happened between 1970 and present in the study area. **b)** Schematic cross-section of the regional tectonic setting based on commonly accepted models (e.g. 2004; Berberian and King, 1981; Paul et al., 2010; Stocklin, 1968). The crustal scale interaction between the lithospheric blocks of the Zagros, Central Iran (CI) and South Caspian basin (SCB) is caused by large scale regional compression associated with the Arabian-Eurasian (EUR) collision. Abbreviations: MRF - Main Recent Fault, HZF - High Zagros Fault, MFF - Mountain Frontal Fault, TF - Talesh Fault, T - Tarom Valley, KF - Khazar Fault, Talesh M. - Talesh Mountains. MRF - Main Recent Fault, SSZ - Sanandaj-Sirjan Zone, MMA - Mesozoic Magmatic Arc, and TMA - Tertiary Magmatic Arc.

Figure 2: a) Simplified tectonic map of the NW part of the Iranian plateau (modified from the structural map of the National Geoscience Database of Iran, NGDIR; <http://www.ngdir.ir>). Black lines show the active faults; circles RF Moho depths in the study region: Talesh (Bavali et al., 2016), CIGSIP receiver function profiles (Chen et al., 2016), EUNASeis dataset (Artemieva and Thybo, 2013), Zagros 2003 profile (Paul et al., 2010), NW Iraq receiver function study labeled "G" shown by diagonal left hatches (Gritto et al., 2008) and other scattered stations (Hatzfeld et al., 2003; Nasrabadi et al., 2008). **b)** Depth to the magnetic basement (Teknik and Ghods, 2017), with superimposed sediment thickness based on a structural study across the NW Zagros belt marked by squares (Vergés et al., 2011) and sediment thickness determined from seismic receiver function studies (circles) including the CIGSIP profile (Chen et al., 2016) and the Zagros 2003 (Paul et al., 2010). The recently discovered Bijar basin (Teknik and Ghods, 2017) is marked by the hatched polygon. **c)** Bouguer anomalies in the NW part of the Iranian plateau. The small black dots indicate gravity stations from Bureau Gravimétrique International (BGI, <http://bgi.omp.obsmp.fr>). Major active faults (Hessami et al., 2003) are shown as solid gray lines. **d)** Reprocessed Aeromagnetic intensity map of Iran (Saleh, 2006). The map is reduced to pole using the continuous reduction to pole method (Paterson Grant & Watson Limited (PGW), 1973). Abbreviations: QA - Qorveh anomaly area (hatched polygon), ZFB - Zagros Fold Belt, CI - Central Iran block, AM - Alborz Mountains Tertiary Magmatic Arc. Positive magnetic anomalies PM1 and PM2 and negative magnetic anomalies NM1 and NM2 indicate magnetic anomalies with wavelengths longer than 150 km, which possibly are associated with deeper sources (see Figure 3). Abbreviations same as Figure 1.

Figure 3: density modeling Step 1: Top: Observed and computed Bouguer gravity anomaly along the profile (for location see **Figure 1a**). The observed anomaly is shown by a thick black line. Thin black line is the computed Bouguer anomaly and the thin red line is the misfit between the observed and calculated anomalies. Middle: All seismic interfaces extracted from the seismic RF model (Chen et al., 2016). Positive and negative interfaces are shown by black and red solid lines. The Moho interface is marked by a thick black line in the bottom of the model. Topography are shown by thick blue line. The depth to the first positive converter interface is shown by a green line. The depth to magnetic basement is shown as a thick pink line. Bottom: 2D density model with homogeneous density for the sedimentary succession, crystalline crust, and upper mantle. Vertical exaggeration (V.E.) is 1.7. ZFTB indicates Zagros fold and thrust belt and other abbreviations for the tectonic units are as in **Figure 1** and **2**.

Figure 4: Comparison of topography (solid black line), reduced to pole aeromagnetic anomaly (solid blue line), low pass 150 km filtered aeromagnetic anomaly (dotted blue line), terrestrial gravity anomaly (solid red line) and satellite gravity anomaly (dashed red line) along the modeling profile. Positive magnetic anomalies PM1 and PM2 can be associated with a depressed Curie isotherm or a high susceptibility material with a shallower CDP. Negative magnetic anomalies NM1 and NM2 can be explained either by a thin magmatic layer or a by low susceptibility in the shallow crust. TMA stands for location of the Tertiary Magmatic Arc. Abbreviations for the tectonic units are described in **Figure 1** and **2**.

Figure 5: density modeling Step 2: In order to reduce the difference between the computed and the observed gravity anomalies (**Figure 3**), the crust is subdivided into more layers. The boundaries between these layers are based on the observed RF boundaries within the crust (middle part). Numbers 1 and 2 in the RF section (the middle panel) show the two possible alternatives Moho depths around the Tarom valley. The numbers represent the density values for each layer in kg/m^3 . The dashed lines show the less reliable inferred boundaries.

Figure 6: Final density model (step 3): Top) Observed anomaly (bold line), calculated Bouguer anomaly (thin black line) and their difference (red line). Bottom) Crustal and upper mantle 2D density model. The crust primarily is divided into the sedimentary layer, the crystalline crust and the mantle. The sediment layer includes the upper, low density, and the lower, denser, part. The crystalline crust is divided into the upper, middle and lower parts.

Figure 7: Magnetic susceptibility models **a)** Susceptibility model with the Curie Depth Point (CDP) boundary at a constant depth of 40 km. The residual magnetic anomaly in the top panel (highlighted by thick gray line) includes long wavelength variations. **b)** The final magnetic susceptibility model in which the Curie isotherm boundary is allowed to vary. The model shows shallowing of the CDP beneath the SSZ and the Alborz mountains at the two interpreted sutures along the profile. Black lines on the susceptibility models are from the density model (**Figure 6**). Red lines mark the CDP and the green line shows the sediment-basement boundary. The numbers in the susceptibility models show susceptibility of a given layer in SI units. Susceptibility of the sedimentary layer is assumed to be 0.001 SI and is fixed to zero at depths below the Curie isotherm. The numbers 1, 2, and 3 on the susceptibility models represent the Kermanshah ophiolite, Qorveh intrusion, and the Tertiary magmatic arc of Alborz, respectively.

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850 scale faults, and solid dashed lines are the boundaries inferred from the density modeling.

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901 **Appendix A1: Sensitivity Analyze of density model.**

902 To address possible non-uniqueness of the gravity model presented in **Figure 6**, we test the
903 sensitivity of the density model in relation to key parameters, including the Caspian dense shallow
904 body, the upper, middle and lower crust, Zagros root and Alborz root. The sensitivity analyses were
905 undertaken to understand how density variations within plausible ranges may affect the response of
906 the density model. The density of the mentioned crustal units have been limited to the range of
907 density values derived from a compilation of seismic studies (i.e. Artemieva and Thybo, 2013).
908 Sensitivity of the model to density of a particular crustal unit is assessed systematically by varying
909 the density and calculating the response of the model. Crustal units with a high sensitivity on density
910 are strongly influenced by density variations and thus are less ambiguous than other parameters.

911 Figure A1 shows the response of the gravity model for different density values (Table A1) assigned
912 to the shallow high density Caspian body. The sensitivity analysis shows that a dense source in the
913 upper Caspian crust is necessary in the northern end of the profile. Figure A1 shows strong gravity
914 response with respect to density variations of the body and suggests a density in the range of 2950
915 to 3000 kg/m³ for the shallow seated body. The experiment vividly shows that a density value in the
916 range of normal upper crustal values (Test number 1 in the table A1) cannot reproduce the short
917 gravity wavelength. The Bouguer gravity map (**Figure 2c**) shows a high amplitude short wavelength
918 Bouguer anomaly along the south and southwest borders of the Caspian Sea. We speculate that the
919 high amplitude short wavelength Bouguer anomaly is due to shallow seated mafic rocks.

920 Figure A2 shows the response of the gravity model for different density values (Table A2) for the
921 Alborz root which is shown by purple color beneath Tarom valley in **Figure 6**. Figure A2 clearly

922 shows that the response of the crustal root of Alborz is strongly dependent on the density and give
923 the minimum misfit for the adopted density of 3100 kg/m^3 , the value used already in our density
924 model.

925 Figure A3 display the response of the gravity model for different density values (Table A3) for the
926 Zagros root which is shown by purple color beneath Zagros in **Figure 6**. As for Alborz, Figure A3
927 clearly shows that the response of the crustal root of Zagros is strongly dependent on the density and
928 gives the minimum misfit for the adopted density of 3050 kg/m^3 , the value used in our density model.

929 The models presented in Figure A3 shows that the gravity response of the model strongly changes
930 with density of the crustal root, also in the Alborz region. This observation casts serious doubt on
931 the possible existence of a tradeoff between the density of the Alborz and Zagros root. To investigate
932 this possible source of non-uniqueness, we calculate the gravity response of our model using 36
933 combinations for density of the crustal root of Alborz and Zagros as listed in Table A4. Figure A4
934 shows that the lowest misfit is for a density of 3050 kg/m^3 for the Zagros root and 3100 kg/m^3 for
935 the Alborz root, as already used in our model (**Figure 6**).

936 In the fifth stage of sensitivity analysis, we investigate possible tradeoff between density of the
937 upper, middle and lower crust and the dense Alborz and Zagros roots. Table A5 describes the tested
938 models. The results indicate that the differences in density of the two roots for Zagros and Alborz is
939 necessary and furthermore a relatively denser root for the Alborz mountains is required with respect
940 to the Zagros root. This indicates that the Bouguer anomaly over Zagros could not equally be caused
941 by increasing density through the whole crust.

942 In the final sensitivity analysis, we want to investigate the possibility of fitting the observed gravity

943 by changing the geometry of the lower crust while the density of all the shallower units are the same
944 as those used in Figure 6 but the density of the lower crust is fixed to 2950 kg/m^3 (i.e., no dense
945 crustal root is considered). With this test we would like to decipher if a small change in the depth of
946 the boundaries of the lower crust could remove the necessity of having the dense crustal root. If the
947 changes in the depth of boundaries are of the order of uncertainty in the RF image there would be
948 no need to include a dense lower crustal root. Using an inversion scheme (Rasmussen and Pedersen,
949 1979; Talwani et al., 1959) we derived the boundaries of the lower crust fitting the observed gravity
950 (Figure A6). The difference between the top boundaries of the lower crust in the two cases is minimal
951 (i.e., less than 3 km and in order of possible RF error) in Zagros but for Alborz it reaches to a
952 maximum difference of 5 km. The difference between the bottom boundaries of the lower crust in
953 the two cases are larger and could reach to a maximum value of 10 km for the Alborz root but for
954 the Zagros root the difference is about 4 km. This test shows that the change in the boundaries is
955 larger than the possible RF depth conversion errors especially in Alborz.

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Table A1 : The sensitivity of the model to the density of the Caspian shallow anomaly

Number of test	1	2	3	4	5	6	7
Caspian shallow anomaly	2800	2850	2900	2950	3000	3050	3100

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Figure A1: The sensitivity of the model to the density of the shallow high density Caspian body. The observed Bouguer anomaly (thick gray line) and calculated Bouguer anomalies (thin colorful lines) for different crustal density of the shallow high density Caspian body (the top panel). The numbers on the top panel corresponds to the models listed in the Table

A1. The bottom panel shows RMS misfit for different models. The RMS values are related to the difference between the calculated and observed gravity anomalies. The RMS misfit is minimum for density of 2950 to 3000 kg/m³.

TableA2: The sensitivity of the density model to the density of the Alborz root

Number of test	1	2	3	4	5	6	7
Density (kg/m ³)	2950	3000	3050	3100	3150	3200	3250

Figure A2: The sensitivity of the density model to the density of the Alborz root. The observed Bouguer anomaly (thick gray line) and calculated Bouguer anomalies (thin colorful lines) for different crustal density of Alborz crustal root (the top panel). The numbers on the top panel corresponds to the models listed in the Table A2. The bottom panel shows RMS misfit for different model. The RMS values are related to the difference between the calculated and observed gravity anomalies. The RMS misfit is minimum for density of 3100 kg/m³.

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TableA3: The sensitivity of the model to variations of the density of the Zagros root

Number of test	1	2	3	4	5	6	7	8
Zagros root	2900	2950	3000	3050	3100	3150	3200	3250

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Figure A3: Top panel shows Bouguer anomaly for different density values from 2900 kg/m³ to 3250 kg/m³ for the Zagros root. Numbers indicate index number is same in the table A3 and horizontal axes of the Root Mean Square (RMS) misfit figure. The RMS values indicate that the calculated anomaly is sensitive to the density of Zagros's root and the density value of 3050 kg/m³ produces the best fit between observation and calculation.

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Table A4 :The sensitivity of the model to the different combination of density for Zagros root and Alborz root

Number of test	1	2	3	4	5	6	7	8	9
Zagros root	2950	3000	3050	3100	3150	3200	2950	3000	3050
Alborz root	2950	2950	2950	2950	2950	2950	3000	3000	3000
Number of test	10	11	12	13	14	15	16	17	18
Zagros root	3100	3150	3200	2950	3000	3050	3100	3150	3200
Alborz root	3000	3000	3000	3050	3050	3050	3050	3050	3050
Number of test	19	20	21	22	23	24	25	26	27
Zagros root	2950	3000	3050	3100	3150	3200	2950	3000	3050
Alborz root	3100	3100	3100	3100	3100	3100	3150	3150	3150
Number of test	28	29	30	31	32	33	34	35	36
Zagros root	3100	3150	3200	2950	3000	3050	3100	3150	3200
Alborz root	3150	3150	3150	3200	3200	3200	3200	3200	3200

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Figure A4: The sensitivity of the model to the different combinations of density for Zagros root and Alborz root (Table 4). In this test, different density value for the Alborz and Zagros roots is considered to find best fit solution for the different combinations. Results indicates that the response of the model depends on the density of the roots and the best solution occur for density of 3050 kg/m³ for Zagros root and 3100 kg/m³ for Alborz root.

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Table A5: The sensitivity of the model to the different scenario of density of upper and lower crust.

Number of test	1	2	3	4	5	6	7	8	9
Upper crust	2650	2700	2750	2800	2800	2800	2800	2800	2800
Middle crust	2800	2800	2800	2850	2900	2900	2900	2900	2900
Lower crust	2900	2900	2900	2900	2950	3000	3000	3000	3000
Zagros root	3000	3000	3000	3000	3050	3100	3150	3200	3250
Alborz root	2950	2950	2950	2950	3000	3050	3050	3050	3050

Number of test	10	11	12	13	14	15	16	17	18
Upper crust	2800	2800	2800	2800	2800	2800	2800	2800	2800
Middle crust	2900	2900	2900	2900	2900	2900	2900	2900	2900
Lower crust	3000	3000	3000	3000	3000	3000	3000	3000	3000
Zagros root	3100	3150	3200	3250	3100	3150	3200	3250	3100
Alborz root	3100	3100	3100	3100	3150	3150	3150	3150	3200

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Figure A5: Sensitivity analysis of crystalline crust of the 2D density model where the properties of the upper, middle, and lower crust were altered according to scenarios given in Table A5. The numbers on the top panel corresponds to the models listed in the Table A5. The

bottom panel shows RMS misfit for different models. The RMS values are related to the difference between the calculated and observed gravity anomalies. The RMS misfit is minimum for scenario 8 (Table 5).

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Figure A6: Sensitivity analysis of the model to the shape of lower crust while assuming a constant density of lower crust of 2950 kg/m^3 (i.e., no dense crustal root is considered) and assigning the density of all the shallower units to those used in Figure 6. The top and bottom boundaries of the lower crust are calculated using an inversion method. The new calculated boundaries of lower crust are shown by a red polygon. This new model provides a solution with an RMS misfit around 5.58 mGal, which should be compared to the RMS misfit for the final model of 6.64 mGal. All the crustal boundaries from Figure 6 are superimposed on the figure.

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Figure A7: Density model of possible low velocity (and density) zone in a more straight way by introducing a low density area

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