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► To cite this version:

Félicie Korostelev, Clémence Basuyau, Sylvie Leroy, Christel Tiberi, Abdulhakim Ahmed, et al.. Crustal and upper mantle structure beneath south-western margin of the Arabian Peninsula from teleseismic tomography. *Geochemistry, Geophysics, Geosystems*, 2014, pp.1-40. 10.1002/2014GC005316 . hal-01015496

HAL Id: hal-01015496

<https://hal.science/hal-01015496>

Submitted on 26 Jun 2014

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**¹ Crustal and upper mantle structure beneath
² south-western margin of the Arabian Peninsula from
³ teleseismic tomography**

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4 **Abstract.** We image the lithospheric and upper asthenospheric struc-
5 ture of western continental Yemen with 24 broadband stations to evaluate

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6 the role of the Afar plume on the evolution of the continental margin and
7 its extent eastwards along the Gulf of Aden. We use teleseismic tomography
8 to compute relative P-wave velocity variations in southwestern Yemen down
9 to 300 km depth. Published receiver function analysis suggest a dramatic and
10 localized thinning of the crust in the vicinity of the Red Sea and the Gulf
11 of Aden, consistent with the velocity structure that we retrieve in our model.
12 The mantle part of the model is dominated by the presence of a low veloc-
13 ity anomaly in which we infer partial melting just below thick Oligocene flood
14 basalts and recent off-axis volcanic events (from 15 Ma to present). This low
15 velocity anomaly could correspond to an abnormally hot mantle and could
16 be responsible for dynamic topography and recent magmatism in western
17 Yemen. Our new P-wave velocity model beneath western Yemen suggests the
18 young rift flank volcanoes beneath margins and on the flanks of the Red Sea
19 rift are caused by focused small-scale diapiric upwelling from a broad region
20 of hot mantle beneath the area. Our work shows that relatively hot mantle,
21 along with partial melting of the mantle, can persist beneath rifted margins
22 after breakup has occurred.

1. Introduction

23 The Afar triple junction, where the Red Sea, East African and Gulf of Aden rifts
24 intersect (Fig.1), is a key region to understand how continental breakup occurred under
25 the influence of a plume and abnormally hot mantle [e.g. *Bastow et al.*, 2011]. Many
26 seismological studies have been carried out in north-east Africa and Arabia to determine
27 the depths of the Moho [e.g. *Ahmed et al.*, 2013; *Mechie et al.*, 2013] and to image upper
28 mantle structure and understand regional geodynamics [e.g. *Bastow et al.*, 2005; *Benoit*
29 *et al.*, 2003, 2006; *Montagner et al.*, 2007; *Zhao*, 2007; *Sicilia et al.*, 2008; *Chang and*
30 *Van der Lee*, 2011; *Koulakov*, 2011] and how it is connected with global mantle flow [e.g.
31 *Montelli et al.*, 2006; *Boschi et al.*, 2007, 2008; *Forte et al.*, 2010; *Moucha and Forte*,
32 2011]. However no previous study has the resolution in continental Yemen on the Gulf
33 of Aden margin due to the lack of seismic stations. Surface wave studies [e.g. *Debayle*
34 *et al.*, 2001; *Sebai et al.*, 2006; *Montagner et al.*, 2007; *Sicilia et al.*, 2008; *Chang et al.*,
35 2011; *Chang and Van der Lee*, 2011] lack sufficient lateral resolution to image the detail
36 of lithospheric and uppermost mantle structures.

37 The YOCMAL project (YOung Conjugate MArgins Laboratory) deployed 23 broadband
38 stations in a network running from the Red Sea margin to Aden city, passing through the
39 Yemeni highlands and Sana'a city (Fig.2). Using classical teleseismic tomography [*Aki*
40 *et al.*, 1977] on recordings from these stations together with a permanent Geofon station
41 (DAMY), we image the relative velocity variations of P-waves in the crust and upper
42 mantle down to 300 km depth. We thus: (1) characterize the lithospheric structure of
43 the rifted margins of the Gulf of Aden and Red Sea system of western Yemen; (2) locate

44 the presence of asthenospheric upwellings in the region and their interaction with the
45 lithosphere.

2. Geodynamic setting

46 The Red Sea and Gulf of Aden rifts are connected to the East African rift at the Afar
47 triple junction, in the 'Horn of Africa'. In the Afar triple junction region, the presence of
48 flood basalts [e.g. *Baker et al.*, 1996; *Hofmann et al.*, 1997; *George et al.*, 1998] and an
49 abnormally low velocity upper mantle anomaly [e.g. *Debayle et al.*, 2001; *Bastow et al.*,
50 2005] could be due to the presence of a mantle plume (Fig.1). Around 35 Ma ago, the
51 current rift system began to open, under the influence of the Afar plume [e.g. *Leroy et al.*,
52 2012]. The flood basalts of Ethiopia and Yemen are the signatures of voluminous eruptions
53 produced during the Paleogene [*Ebinger and Sleep*, 1998] with highest eruption rates at
54 31 Ma [*Baker et al.*, 1996; *Hofmann et al.*, 1997; *George et al.*, 1998]. A renewed phase
55 of volcanism took place 24 Ma ago that corresponds with the synchronous appearance of
56 basaltic dikes and gabbros along the Red Sea, from Afar and Yemen to northern Egypt.
57 From 25 to 16 Ma, a series of basaltic, trachytic and rhyolitic dikes were emplaced along
58 the Red Sea margins [*Zumbo et al.*, 1995], at the same time as emplacement of large
59 granitic batholiths (Jabal Hufash and Jabal Bura, see Fig.2). These granites are oriented
60 north-south [*George et al.*, 1998] and located at the border of the Great Escarpment, a
61 1000 km long sudden change in altitude, from 200 m on the west in the Tihama Plain, to
62 more than 1000 m in the Yemeni highlands. In addition, syn-rift (30 to 16 Ma) basaltic
63 flows dipping towards the sea (seaward dipping reflectors - SDRs) have been imaged under
64 the Tihama Plain [*Davison et al.*, 1994]. In the Red Sea, a second phase of opening began
65 14 Ma ago synchronous with the formation of the Dead Sea transform fault to the north.

66 This phase is accompanied by increased extension rates and rift-flank uplift [*Courtillot*
67 *et al.*, 1999].

68 From 12 Ma to present, magmatic provinces developed within a 2500 km radius of Afar
69 and southwestern Yemen as far away as western Saudi Arabia, Jordan and northern
70 Syria, [*Zumbo et al.*, 1995; *Bertrand et al.*, 2003; *Coulié et al.*, 2003], see Fig.1. Recent
71 magmatism has occurred offshore in the ocean-continent transition of the Eastern Gulf of
72 Aden margin [*Lucazeau et al.*, 2009; *Autin et al.*, 2010; *Watremez et al.*, 2011], off-axis
73 Sheba ridge [*d'Acremont et al.*, 2010] and below the southern Oman continental margin
74 [*Basuyau et al.*, 2010] (Fig.1). The increased magmatism caused by extension above a
75 plume was first thought to stop at the Shukra al Sheik fracture zone [e.g. *Hébert et al.*,
76 2001], but these recent results indicate that the limit could be further east (Fig.1), as
77 proposed by *Leroy et al.* [2010a].

78 The Gulf of Aden is characterized by two stretched continental margins systems: non
79 volcanic margins in the east and volcanic margins in the west near the Afar triple junction.
80 The volcanic margins are associated with syn-rift SDRs up to 5 km thick [*Tard et al.*,
81 1991; *Leroy et al.*, 2012]. SDRs are also sparsely found in the east especially at the ocean-
82 continent transition [*Autin et al.*, 2010; *Leroy et al.*, 2010b] but no syn-rift volcanism has
83 been found east of the longitude 46°E, see Fig.1 [*Leroy et al.*, 2012]. The study region
84 is near the edge of the African superplume, as shown by large scale global tomographic
85 models [e.g. *Debayle et al.*, 2001; *Sebai et al.*, 2006; *Boschi et al.*, 2007, 2008] and by
86 petrologically derived temperature estimates of the mantle [*Rooney et al.*, 2012; *Rolandone*
87 *et al.*, 2013].

3. Data

88 Data have been collected from 23 temporary broadband stations deployed from March
89 2009 to March 2010 (YOCMAL project), and from one permanent station (DAMY, Geo-
90 fon). The network extends from the Red Sea margin to the Gulf of Aden margin, passing
91 through the Yemeni highlands (Fig.2). The sensors deployed were Guralp 40T (sampling
92 rate 50 sps, 30 s natural period), 6TD (sampling rate 40 sps, 30 s natural period) and
93 ESPD (sampling rate 40 sps, 60 s natural period). This network configuration allows
94 imaging structures with a high resolution, and down to 300 km depth (the surface extent
95 of our network). We selected 200 teleseismic events with clear first P-wave arrivals from
96 earthquakes with magnitude ≥ 5.5 . Among them, 142 events arrived as P, 35 events as
97 PP and 23 events as PKP (Fig.3), and were picked on consistent peaks in the first cycle.
98 We used 2456 delay times calculated relative to the IASP91 Earth model [*Kennett and*
99 *Engdahl*, 1991]. For each residual, a picking error was assigned within the range ± 0.05
100 to ± 0.2 s. As the events registered come mostly from east / northeast and have a range
101 of epicentral distances, the rays are crossing, and we expect our best resolution in this
102 direction.

4. Method

103 Our delay time residuals were inverted using the ACH teleseismic tomography method
104 first used by *Aki et al.* [1974], then developed by *Weiland et al.* [1995], *Zeyen and Achauer*
105 [1997] and *Jordan and Achauer* [1999]. As the inversion uses a 3D iterative ray tracing
106 at each step, the initial location of the parameterization nodes affects the inversion result
107 [e.g. *Calò et al.*, 2008]. To reduce this problem, we averaged the results of four inversions
108 with the same parameters of inversion, but with a meshing shifted by half a node (10 km)

109 towards the east, north, and northeast compared with our initial reference model. This
110 'Average Smooth model' technique smooths the local effects due to the meshing [*Evans*
111 *and Achauer*, 1993]. The resulting velocity model is presented in Fig.4 as depth slices and
112 in Fig.5 as depth cross-sections.

113 Our network's dimensions are 360 km from North to South, 260 km from West to East,
114 so our investigation depth (i.e. the depth until which we have resolution) is estimated
115 to be ~ 300 km [*Evans and Achauer*, 1993]. In our initial model, velocities are organized
116 in successive horizontal layers of nodes, with an interpolation gradient between each of
117 them [*Thurber*, 1983]. The minimum distance between two nodes horizontally is 20 km
118 (i.e. the minimum distance between two stations). The node spacing is 40 to 100 km at
119 the edges of the model. The initial model was the IASP91 reference Earth model. Ten
120 levels of nodes were placed in depth between 0 and 500 km, so we have nine layers (nodes
121 at 0, 30, 45, 70, 100, 150, 200, 250, 300 and 500 km depth). The thickness of the crust
122 is important in the inversion process [e.g. *Zhao et al.*, 1994, 2006]. In our study we use
123 the receiver functions results of *Ahmed et al.* [2013] to compute a time correction for each
124 station (Fig.6). This correction accounts for lateral variations in the crustal thickness,
125 which are difficult to constrain by our teleseismic data alone; seismic velocity within the
126 crust is treated as a free parameter in our inversion. The correction for each station is
127 computed on the absolute residuals and is in the range -0.611 s (for the station located
128 on the thinnest crust) to 0.160 s (for the station located above the thicker crust). This
129 allows a reduction in the propagation of crustal structure into the deeper layers of the
130 velocity model.

131 The smoothing factor, which limits the short wavelength velocity variations, and the ini-

132 tial standard deviation associated with each node of the initial model for each of the nine
133 iterations were chosen after a series of tests and are respectively 0.001 and 0.007 km.s⁻¹.
134 The smoothing value chosen is the same as *Tiberi et al.* [2008] and *Basuyau et al.* [2010].
135 Tests showed that the results are not significantly different for a standard deviation be-
136 tween 0.005 and 0.01 km.s⁻¹. We tested these inversion parameters in order to get a
137 stable model, choosing parameters that decrease the root mean square of residuals (RMS)
138 through the inversion's nine iterations. The overall decrease of the RMS is 55% in the
139 final model.

5. Results

5.1. Checkerboard test

140 We use a checkerboard test to assess the resolution of our inversion. The synthetic
141 checkerboard model consisted of rectangular velocity anomalies of +5% and -5% velocity
142 variations at depths of 70, 200 and 300km depth (Fig.7a). The size of the anomalies in-
143 creases around the edges as the method requires the nodes to be further apart at the edges
144 of the model. We use the same inversion parameters (smoothing, standard deviation) as
145 in our actual inversion.

146 The results of the checkerboard test (Fig.7b) show that at 70 km depth, due to the con-
147 centration of crossing rays under the stations, we have the best resolution, with ~ 20 %
148 recovered amplitude. At 200 km depth, the crossing rays cover a larger area, so the
149 anomalies are well retrieved. At 300 km depth, the eastern retrieved anomalies are much
150 better constrained east of 44°E, due to the concentration of ray paths coming from the
151 east. The Fig.7b shows the piercing points under our network at 300 km depth, at our

152 maximum investigation depth. Below 300 km depth, the rays are too dispersed to give
153 good resolution.

5.2. Crustal-scale structures

154 The thickness of the crust in our study area is estimated from receiver function analysis
155 to be between 14 km at the coast and 35 km inland [*Ahmed et al.*, 2013]. The crust
156 is represented by the first two layers of our model, and is characterized by the highest
157 anomaly contrasts in P-wave velocity, which have a range of $\pm 4\%$ (Fig.4 and 5). These
158 anomalies are related to geological structures observed at the surface. At 30 km depth, the
159 strong low velocity anomaly (-4%) beneath the center of our network (stations MAWI to
160 DALA) correlates with the high topography of the plateau (>2000 m above sea level). We
161 interpret this pattern as due to the thicker crust (>30 km) beneath the Yemeni highlands
162 [*Ahmed et al.*, 2013].

163 Under stations located near the Red Sea (north-westernmost part of the network) and Gulf
164 of Aden (southernmost part of the network) margins there are high velocity anomalies
165 above 30 km due to the thinner crust (<30 km). *Ahmed et al.* [2013] estimated the
166 thickness of the crust to be ~ 22 km in the coastal areas, and less than 14 km for the Red
167 Sea margin. These high velocity anomalies are located beneath SDRs, which is consistent
168 with the emplacement of sub-aerial volcanic material during rifting [*Tard et al.*, 1991;
169 *Davison et al.*, 1994; *Bastow and Keir*, 2011; *Leroy et al.*, 2012].

170 The high velocity anomalies under the stations ANID and SUGH could be due to a thin
171 crust as they are in the Red Sea coastal area, or to Tertiary granitic intrusions of Jabal
172 Hufash and Jabal Bura respectively [*Geoffroy et al.*, 2002], see Fig.2. The high velocity
173 anomaly corresponding to Jabal Hufash is imaged down to 45 km depth, which is slightly

174 deeper than that of Jabal Bura (30 km depth). This can be explained by significant
175 smearing due to the higher amplitude of the Jabal Hufash anomaly. There are lower
176 velocities under the stations UAYA and ZUWA, on the Tihama plain (Fig.5b), probably
177 due to ~ 4000 m of low velocity sediments [*El-Anbaawy et al.*, 1992; *Davison et al.*, 1994].

5.3. Upper mantle structures

178 The resulting upper-mantle P-wave anomalies are in the range of $\pm 2\%$. The most strik-
179 ing pattern is a low velocity anomaly located under the Yemeni highlands at 45 km depth,
180 apparently dipping northeastward down to 300 km. It reaches its maximum amplitude at
181 70 km depth (east of the DAMY station) beneath the volcanic field of Dhamar (Fig.2),
182 where there are two active volcanoes [*Manetti et al.*, 1991]. The northern part of this
183 anomaly is located, at 70 km depth, under the volcanic field of Sanaa, which is also still
184 active [*Manetti et al.*, 1991], see Fig.5.

185 There is a second low-velocity anomaly located under the southwestern corner of Yemen
186 and the stations MOKA and HAKI (Fig.5c and d). This low velocity anomaly is nearly
187 vertical and is recognized from the surface to 300 km. It is located just beneath the
188 volcanic area of Jabal An Nar (Fig.5d), which was active during late Miocene, around
189 10 Ma, [*Manetti et al.*, 1991].

190 Even if we corrected the residuals from the crustal portion, our results are likely to include
191 effects from Moho variations. This is because the corrections are based on receiver func-
192 tions which display a strong trade-off between Moho depth and crustal velocity model.
193 In addition, we took a 1D velocity model to be consistent with the receiver functions
194 study of [*Ahmed et al.*, 2013] and this can generate errors [*Martin and Ritter*, 2005]. The
195 lack of detailed 3D crustal information precludes us from going further with the crustal

196 corrections than we have done. To estimate the resolution of our models in both the crust
197 and the mantle, we proceed to synthetic tests. We test whether the low velocity anomaly
198 beneath the high plateau is dipping towards the northeast because of smearing along ray
199 paths (Fig.3) in the next section. We investigate by means of synthetic tests whether:
200 (1) the velocity variations observed in our resulting model are smearing downwards into
201 the mantle, (2) the low velocity anomalies under the southwestern corner of Yemen and
202 under the high plateaus are artefacts, and whether we can determine at what depth they
203 are located, (3) the dipping low-velocity anomaly is related to the presence of partial melt
204 or not.

6. Synthetic tests and presence of melt

6.1. Propagation of crustal signal

205 We test the smearing with depth of a crustal anomaly (0-30 km) without a crustal
206 correction by introducing a -5% anomaly around stations in the Yemeni Highlands and a
207 +5% anomaly under the two continental margins (Fig.8a). The rest of the input model is
208 laterally uniform. The results of the inversion of this synthetic model (Fig.8b and c) show
209 that the velocity variations are well recovered in location but with 40% lower amplitude.
210 Between 45 and 100 km depth, we can discern very low amplitude anomalies (<0.5%),
211 corresponding to a small amount of vertical smearing. At 100 km depth, we have a -1%
212 anomaly under the Yemeni highlands. Approximately half of this signal could be due
213 to a smearing of the crust signal. No significant anomaly can be seen beneath 100 km
214 depth. We conclude that crustal velocity anomalies do not propagate to deeper layers of
215 our model and that there is an authentic low velocity anomaly in the upper mantle. This

216 test shows that the use of a crustal correction is relevant and is necessary to limit the
217 extent of the propagation of crustal velocity structure into deeper layers of the model.

6.2. Low velocity anomalies beneath Southwestern Yemen

218 Using a synthetic input model, we simulate the resulting geometry of our final P wave
219 model (see auxiliary material). We computed a series of tests (available in auxiliary
220 material), and here we present the most relevant example. We first place a -5% anomaly
221 under the three Yemeni volcanic fields of Sana'a, Dhamar and Marib from the base of
222 the crust down to 200 km depth. The other -5% anomaly is placed from the base of the
223 crust to the base of our model, under the southwestern corner of Yemen and southern
224 Red Sea. Figure 9 shows the retrieved inversion image for a SW-NE profile (compare with
225 the observed results in Fig.9c). Although the anomaly amplitudes within the crust are
226 not retrieved, the dipping anomaly is quite well retrieved in the synthetic output model,
227 as well as the low velocity anomaly beneath Jabal An Nar volcanic field (Fig.9). This
228 tests shows that the dipping low velocity anomaly under the Yemeni volcanic fields can
229 be explained by a 220x260 km mantle upwelling between the base of the crust and 200 km
230 depth. Moreover, the low velocity anomaly beneath Jabal An Nar can be explained by a
231 large zone of hotter mantle.

6.3. Presence of partial melt

232 Traveltime tomography gives the present state of the upper mantle in terms of veloc-
233 ity variations but it precludes any direct interpretation concerning their origin. Indeed,
234 several factors, such as temperature, chemical composition or anisotropy can affect the
235 velocity of seismic waves [e.g. *Karato, 1993; Sobolev et al., 1996*]. *Karato [1993]* demon-

236 strates that a purely thermal origin leads to a linear relationship between P and S residuals
237 with a slope of 2.9. *Gao et al.* [2004] proposed that a slope higher than 2.9 between P
238 and S residuals for the same events highlights the presence of partial melting and several
239 authors used the relationship to explain the presence of negative velocity anomalies in
240 the upper mantle [e.g. *Bastow et al.*, 2005; *Basuyau et al.*, 2010]. This method considers
241 relative delay times, so that problems associated with amplitude recovery (e.g. due to
242 differing numbers of traveltime observations and regularization levels) and other artefacts
243 associated with the inversion procedure (i.e. parameterization and ray-path accuracy) do
244 not complicate the comparison of the data.

245 We thus selected the events coming from northeast (Russia, Japan, China), and picked the
246 S arrival on the transverse component for stations located above the Yemeni Highlands, so
247 that the rays chosen are passing through the northeastward dipping low-velocity anomaly.
248 In Fig.10, following *Gao et al.* [2004], *Bastow et al.* [2005] and *Basuyau et al.* [2010], we
249 plot S versus P relative travel-time residuals, and find out that our data are consistent
250 with a slope >2.9 , thus implying the presence of partial melt along the rays coming from
251 northeast. The presence of Holocene volcanoes (in Sana'a, Dhamar and Marib volcanic
252 fields) on the surface helps support the idea that there is partial melt being created in
253 the asthenosphere and intruding the lithosphere. Moreover, the isotopic signatures of the
254 melts from the three volcanic fields suggest a strong asthenospheric component [*Manetti*
255 *et al.*, 1991], which is consistent with our results.

7. Discussion

7.1. Crustal-scale structures

256 We produce a relative velocity model for the propagation of P waves down to 300 km
257 beneath western Yemen. The low velocity anomaly ($\sim -4\%$) located below the Yemeni
258 highlands between 0 and ~ 35 km corresponds to a region of 30-35 km thick crust [*Ahmed*
259 *et al.*, 2013]. At 30 km depth, we observe a dramatic transition from very low to high
260 velocities (-4% to $+4\%$) under the coast of the Red Sea and Gulf of Aden. We interpret this
261 sudden short length scale (<40 km for the Red Sea margin, <100 km for the Gulf of Aden
262 margin) variation as a lateral transition between a thick crust and a thinned, intruded
263 and stretched crust. In addition, our new seismic images showing lower mantle velocities
264 under the southwestern corner of Yemen is consistent with ongoing rifting above a thermal
265 anomaly in the underlying mantle, due to the Afar plume. Under these conditions, melt
266 generated by adiabatic decompression of the asthenosphere beneath thinned and stretched
267 lithosphere migrates upwards to intrude or underplate continental crust and extrude as
268 mostly basalt flows (Oligocene flood basalts) [*White and McKenzie*, 1989]. Our positive
269 V_p anomaly near the base of the crust under the Red Sea and Gulf of Aden margin are
270 consistent with melt produced from an abnormally hot mantle which enriches the melt in
271 MgO [*White and McKenzie*, 1989], capable of producing intrusions/underplate of up to
272 7.2 km/s rather than 6.8 km/s from melting normal mantle. These high velocity anomalies
273 are focused mainly into narrow zones of denser material, away from the most stretched
274 areas. That is because the mantle temperature was highest at the start of continental
275 break-up [*White et al.*, 2008]. Such lower crustal intrusions/underplating are a common
276 feature of volcanic margins such as the north Atlantic [*Geoffroy*, 2005; *Mjelde et al.*, 2005;

277 *White et al.*, 2008].

278 Placing constraints on the timing of the underplating is difficult. If emplaced before the
279 eruption of SDRs and continental break-up as has been proposed for the north Atlantic (at
280 least 10 to 15 Ma off Norway, *Gernigon et al.* [2006]) then it would have an influence on the
281 structural development of the margin and partly consist of high pressure granulite/eclogite
282 lower crustal rocks [*Gernigon et al.*, 2006].

283 At shallower depths of 0-30 km in our model, seismic anomalies are directly related to
284 the geological units observed at the surface. The granitic intrusions of Jabal Hufash and
285 Jabal Bura are associated with high velocity anomalies (up to +4%, see Fig.2). The depth
286 extent of the anomaly below the granitic intrusions of Jabal Hufash may indicate a deeper
287 root, as hypothesized by *Denèle et al.* [2012]. It could also be explained by a stronger
288 smearing effect due to the high amplitude of the Jabal Hufash anomaly.

7.2. Deep structure of the margins

289 Our synthetic tests show that deep anomalies cannot be explained by the smearing
290 in depth of crustal anomalies. We interpret the low velocity anomaly (-2%) between
291 ~35 km and 300 km depth under the highest topography as abnormally hot mantle up-
292 welling. This low velocity hotter mantle is located below the stations DAMY, RUSA and
293 YSLE (Fig.5), which are located on the thick Oligocene flood basalts and three more
294 recent volcanic fields (15 Ma to present) volcanic areas, e.g. Dhamar and Sana'a [*Davison*
295 *et al.*, 1994; *Pik et al.*, 2008; *Leroy et al.*, 2010b], see Fig.2. Additionally, we infer presence
296 of partial melt in the crust or uppermost mantle to be responsible for this low velocity
297 anomaly (Fig.10). A similar pattern of upper mantle off-axis upwelling has also been
298 found in the southern Red Sea rift of Afar and explained by small diapiric upwellings

299 (<100 km) [*Hammond et al.*, 2013]. We surmise that a similar mechanism is the best
300 explanation for our observations. This idea is supported by the large asthenospheric com-
301 ponent in the magma from the three Yemeni volcanic fields inferred from trace element
302 and isotope geochemistry [*Manetti et al.*, 1991]. Observations of recent dike intrusions at
303 Harrat Lunayirr in Saudi Arabia [*Pallister et al.*, 2010] show rifted margin magmatism
304 may be important in accomodating extension after breakup along the full length of the
305 Red Sea margin [*Ebinger and Belachew*, 2010].

306 Our new relative P-wave velocity model beneath western Yemen suggests the young rift
307 flank volcanoes on the margin of the Red Sea rift are caused by focused small-scale di-
308 apiric upwelling from a broad region of hot mantle beneath the area. Our work shows
309 that relatively hot mantle, along with partial melting of the mantle, can persist beneath
310 rifted margins after breakup has occurred. These findings have important implications for
311 interpreting the thermal history and deformation of volcanic rifted margins after breakup
312 is achieved since most breakup models assume rift margin volcanism ceases after seafloor
313 spreading starts.

314 Buoyant hot mantle may contribute to a dynamic topography of the Yemeni high plateau.
315 Almost all the topography in this region could indeed have a dynamic origin, because
316 the rift-flank uplift from flexure [*Daradich et al.*, 2003] is not sufficient to produce high
317 topographies over such a large area. Numerical modeling of the plume/lithosphere inter-
318 action predict an uplift of about 500 to 1500 m in less than 0.3 to 0.5 Ma after the plume
319 initiation [*d'Acremont et al.*, 2003]. Moreover, *White and McKenzie* [1989] explained that
320 an increase of about 150°C in the mantle leads to a dynamic uplift, but that the addition
321 of dense igneous material under a thinned crust produces an immediate subsidence of

322 more than 2 km in order to maintain isostatic equilibrium. We observe a similar pattern
323 on the Red Sea and Gulf of Aden margins, with high Yemeni plateau dynamically sup-
324 ported by a hot mantle (Fig.5a, b and d), and a subsiding area, for example the Tihama
325 Plain, underlain by seaward dipping reflectors, and ultra-mafic bodies accreted under
326 the crust. This hypothesis should however be tested for Yemen by gravity and isostatic
327 modeling.

328 The weak low velocity anomaly imaged under the southwestern corner of Yemen, beneath
329 the MOKA station, is underneath the Miocene volcanic area of Jabal An Nar. Our syn-
330 thetic test (Fig.9) shows that this low velocity anomaly may be explained by a large zone
331 of hotter mantle. This feature could be due to the Afar plume signal, located only ~ 300
332 km away, in the Afar depression.

8. Conclusions

333 We performed an inversion of P-wave teleseismic data to image lithospheric structure
334 beneath the SW of the Yemen, the southern Red Sea and western Gulf of Aden margins.
335 The crustal part of the model is dominated by a possible ultra-mafic underplating beneath
336 the Red Sea and Gulf of Aden margins, a sudden thinning of the crust for this volcanic
337 margin, and Tertiary granitic intrusions (Jabal Hufash and Jabal Bura). In the mantle,
338 we image a low velocity anomaly in which we infer partial melting just below the highest
339 topography, the thick Oligocene flood basalts and other off-axis volcanic regions (from
340 15 Ma to present). This low velocity anomaly could correspond to an abnormally hot
341 mantle and could be responsible for dynamic topography and recent magmatism in western
342 Yemen. Some hot material has also been inferred under the southwestern corner of Yemen
343 and may be due to the Afar plume signal.

9. Appendix : Other synthetic tests

344 **Acknowledgments.** This project was funded by the ANR YOCMAL (Agence Na-
345 tionale de la Recherche), CNRS-INSU-PICS Yemen, GSMRB Yemen and is in the frame-
346 work of the Actions Marges program. Seismic equipment from SEIS-UK is funded by
347 NERC under loan number 873. We thank David Hawthorn, Alex Brisbourne and Vic-
348 toria Lane for their efforts during the deployment and servicing of network, the French
349 Embassy in Yemen (J. G. Sarkis, J. Dechezlepretre and C. Bousquet), local governors and
350 the people of the Yemen governerates for their help during the field work.

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569 FIGURES

Figure 1. Geodynamic map of Arabia. Yellow triangles are for YOCMAL Network stations in southwestern Yemen. The magmatism older than 20 Ma is represented in purple, whereas the younger magmatism is in pink (modified from *Davison et al.* [1994]). J.: Jordan, UAE: United Arab Emirates, SS.: Shukra el Sheik Fracture Zone.

Figure 2. (a) Topographic map of Southwestern Yemen. The Yemeni highlands, above 1000 m high, are mainly constituted by basaltic trapps. These basalts are 3 km thick. The volcanic Pliocene to present day volcanic fields of Sanaa and Dhamar are represented in red. Jabal Hufash and Jabal Bura Tertiary batholiths are in blue (modified from *Davison et al.* [1994]). Black dots are for YOCMAL seismological stations. (b) Geological sketch of the northern part of our study area. Red stars are for Pliocene-Quaternary volcanoes. The batholithes are located along the Great Escarpment, which runs parallel to the Red Sea margin and the Tihama Plain. Left of the Great Escarpment, altitudes are below 200 m.

Figure 3. Azimuthal distribution of the events used on our study.

Figure 4. Final P-wave velocity model obtained from inversion. Seismic stations are located with the black triangles. Note the color bar change between the first two layers and the other slices of the model, and the scale is saturated for the first layer.

Figure 5. Cross sections in the final P-wave velocity model obtained from inversion. (a) Cross section along the Red Sea margin, (b) Cross section along the Gulf of Aden margin, (c) Cross section through the granitic intrusions of Jabal Hufash and Jabal Bura, (d) Southwest-Northeast cross section of the corner of Yemen.

Figure 6. Crustal correction applied for the stations. These corrections were computed from the crustal thickness obtained by receiver function analysis [*Ahmed et al.*, 2013].

Figure 7. Checkerboard test for the inversion of seismological data. (a) Synthetic input model for P-wave velocity, (b) depth slices at 70, 200 and 300 km through the retrieved velocity model, piercing points: impact points of the rays with the layer located at 300 km depth (c) North-South cross section in the retrieved velocity model.

Figure 8. Synthetic test for the propagation of crustal signal. (a) Synthetic output model, (b) depth slices through the retrieved velocity model. (c) Cross sections of the Aden and Red Sea margins in the synthetic model.

Figure 9. Synthetic test for two low velocity anomalies under the Southwestern Yemen. (a) SW-NE cross section in the input model. (b) SW-NE cross section in the output model. (c) SW-NE cross-section of the corner of Yemen from our final P-wave velocity model.

Figure 10. Plot of P-wave versus S-wave relative arrival-time residuals for all the stations and events coming from Northeast (Japan, China, Russia). The solid red line is the least square fit to our data (with the slope of the line), and the dashed line is a slope of 2.9 (thermal effect only). Standard deviation is 0.31 for the best-fit line and 0.36 for the 2.9 gradient.

Figure 11. Auxiliary material. Series of synthetic tests for constraining the low velocity anomalies under the Southwestern Yemen. (a) Location of the cross section. (b) SW-NE cross-section in the final P-wave tomography model presented Fig 5d of the paper (c to j) SW-NE cross-sections of a series of tests with low-velocity anomalies of distinct widths and depths.