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► To cite this version:

Abdulkhakim Ahmed, Cecile Doubre, Sylvie Leroy, Derek Keir, Carolina Pagli, et al.. Across and along-strike crustal structure variations of the western Afar margin and adjacent plateau: Insights from receiver functions analysis. *Journal of African Earth Sciences*, 2022, 192, 10.1016/j.jafrearsci.2022.104570 . insu-03691300

HAL Id: insu-03691300

<https://insu.hal.science/insu-03691300>

Submitted on 20 Oct 2022

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Journal of African Earth Sciences

Across and along-strike crustal structure variations of the western Afar margin and adjacent plateau: Insights from receiver functions analysis

--Manuscript Draft--

Manuscript Number:	
Article Type:	Research Paper
Keywords:	Africa; Continental margin; Afar margin; Southern Red Sea Rift; Receiver Functions; Crustal Structure
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respectively. Our estimated values are lower than $\beta > 3.0$ predicted from plate reconstructions, and we interpret that the variations are best explained by 2-5 km magmatic addition at the base of the crust. The crustal composition beneath the southern profile is more complex with elevated Vp/Vs ratios ranging between 1.79 to 1.85 beneath the western plateau and marginal graben. This is consistent with a greater mafic component and best explained by crust altered by intrusions due to significant pre and syn-rift magmatic activity. Abnormally high Vp/Vs ratios more than 1.90 are observed beneath the axial rift zone of central Afar, which most likely suggesting the localization of partial melt within the crust.

Across and along-strike crustal structure variations of the western Afar margin and adjacent plateau: Insights from receiver functions analysis

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Abstract

We used teleseismic receiver function analysis to image the crustal structure beneath 24 broadband seismic stations densely deployed along two profiles traversing different structural units across the western Afar margin. Our high-resolution receiver function results image pronounced spatial variations in the crustal structure along the profiles and provide improved insights to understand how strain is partitioned in the crust during rifting. Beneath the western plateau next to northern Afar, the crust is likely felsic-to-intermediate in composition (average V_p/V_s 1.74), with a step like thinning of the crust from an average of 38 km beneath the western plateau to an average of 22 km beneath the marginal graben. Consistently thicker crust is observed beneath the southern profile (central Afar), showing four distinct regions of uniform crustal thickness: 1) an average crustal thickness of 42 km beneath the western plateau; 2) 34 km beneath the foothills area; 3) 28 km beneath the marginal graben and the wide extensional basin and 4) 21 km beneath the central rift axis. We use crustal thickness results to estimate a stretching

27 factor β of 2.2 and 2.7 for central Afar and northern Afar respectively. Our estimated
28 values are lower than $\beta > 3.0$ predicted from plate reconstructions, and we interpret that
29 the variations are best explained by 2-5 km magmatic addition at the base of the crust.
30 The crustal composition beneath the southern profile is more complex with elevated
31 Vp/Vs ratios ranging between 1.79 to 1.85 beneath the western plateau and marginal
32 graben. This is consistent with a greater mafic component and best explained by crust
33 altered by intrusions due to significant pre and syn-rift magmatic activity. Abnormally
34 high Vp/Vs ratios more than 1.90 are observed beneath the axial rift zone of central Afar,
35 which most likely suggesting the localization of partial melt within the crust.

36
37 Keywords: Africa; Continental margin; Afar margin; Southern Red Sea Rift; Receiver
38 Functions; Crustal Structure

39 40 **1. Introduction**

41 The Afar Depression lies in northeast Africa where the Red Sea and Gulf of Aden
42 oceanic spreading ridges and the intra-continental East-African Rift meet, forming
43 a rift-rift-rift triple junction between the Nubian, Somalian, and Arabian plates
44 (McKenzie et al., 1972; Courtillot, 1982; Tesfaye et al., 2003; Doubre et al., 2017).
45 The Afar Depression is bound by the continental Danakil block to the northeast
46 (Fig. 1), and by the western and southern plateaus, to the west and south
47 respectively. The Afar Depression is thought to have formed shortly after the
48 impact of the Afar mantle plume and associated eruption of the bimodal Ethiopia-
49 Yemen Trap Series of basalts and rhyolitic lavas (Hofmann et al., 1997; Pik et al.,
50 1998, 1999; Pik, 2011). The extension of the proto-African lithosphere started in
51 the Oligo-Miocene (~29-26 Ma) shortly after the peak in Trap Series volcanism at
52 ~29-31 Ma (Wolfenden et al., 2004, 2005; Stab et al., 2016). In the present-day,
53 several magmatic segments are the locus of extensional deformation with
54 networks of normal faults formed above narrow zones of episodic dike intrusion
55 (Fig. 1; Wright et al., 2006; Grandin et al., 2009; Smittarello et al., 2016; Barnie et
56 al., 2016). Therefore, the continental margin of the Afar region is a unique location
57 where the transition from the stable continental plateaus to incipient seafloor
58 spreading within the Afar rift axis is subaerially exposed (Barberi and Varet, 1977;

1 59 Hayward and Ebinger, 1996; Beyene and Abdulsalam, 2005; Medynski et al.,
2
3 60 2016).
4
5 61 Despite reasonable consensus regarding the locus and mode of upper crustal
6
7 62 extension both at the stage of rift initiation and at the present day, few data allow
8
9 63 an informed interpretation of the structural evolution of the rifting between these
10
11 64 two periods. Continental margins underwater have been intensely investigated to
12
13 65 constrain the mechanisms leading to the continental break-up (e.g. Leroy et al.,
14
15 66 2012). Particularly, seismic experiments help to draw the geometry of the margin
16
17 67 units and quantify, in terms of finite deformation, amount of stretching and
18
19 68 thinning due to tectonic processes and also the role of magmatism (e.g. Leroy et al.,
20
21 69 2010; Clark et al., 2006; d'Acremont et al., 2005, 2006). However, due to the
22
23 70 subaerial exposure, passive seismic experiments and geological studies can be
24
25 71 more easily conducted on the Afar margin.

26
27 72 In addition to geologic studies (Wolfenden et al., 2005; Stab et al., 2016), the Afar
28
29 73 crust has been imaged along rift with seismic refraction survey (Makris and
30
31 74 Ginzburg, 1987). The crustal thickness variations and internal seismic structure of
32
33 75 Afar margin have to date been imaged with low resolution using receiver function
34
35 76 studies from seismic networks with sparse station spacing (~40-km) (Dugda et al.,
36
37 77 2005; Dugda and Nyblade, 2006; Hammond et al., 2011; Kibret et al., 2019). None
38
39 78 of the previous studies offer a sufficiently good image of the crustal structure
40
41 79 across the margin to unravel the extensional history and any along-margin
42
43 80 variations in the role of tectonics and magmatism during the continental breakup.
44
45 81 A detailed image of the finite deformation will help to constrain the structure and
46
47 82 evolution of the margin prior and coeval with the continental rifting and answer
48
49 83 some of the remaining questions: Does the stretching and the thinning of the
50
51 84 continental lithosphere occur over a narrow or broad region? Is the rifting smooth
52
53 85 or stepped? What are interactions between mechanical stretching and thinning of
54
55 86 the lithosphere, and magma intrusions (e.g. Keir et al., 2013)?
56
57 87 In order to resolve the high-resolution crustal structure of the western margin of
58
59 88 the Afar Depression for the first time we analyze data from 24 temporary
60
61 89 broadband stations distributed along two densely sampled profiles going from the

1 90 western plateau to central and northern Afar. We use P-to-S receiver functions
2
3 91 (Burdick and Langston, 1977; Langston, 1977; Ammon, 1991) to image the crust-
4
5 92 mantle transition and intracrustal discontinuities as well as to infer crustal
6
7 93 properties from Vp/Vs values. This allows us to understand the spatial
8
9 94 relationships between plate stretching, thinning and magma intrusion during
10
11 95 continental breakup. The comparison of the crustal structure along the two
12
13 96 profiles allows us to estimate the longitudinal variations of the rifting along the
14
15 97 margin and the respective role of tectonics and magmatism.
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17 98

18 99 **2. Background**

20 100 21 101 **2.1 Geological and Tectonic Settings**

22 102 Rifting along the Gulf of Aden initiated ~34 Ma ago (Leroy et al., 2012) followed by
23
24 103 the emplacement of voluminous Ethiopia-Yemen Trap Series on the Ethiopian and
25
26 104 Yemeni plateaus ~31-29 Ma ago (Baker et al., 1996a; Hofmann et al., 1997;
27
28 105 Ukstins et al., 2002). On the Ethiopian plateaus, they erupted and covered a thick
29
30 106 series of Mesozoic sedimentary rocks (Bosworth et al., 2005; Beyene and
31
32 107 Abdelsalam, 2005) currently outcropping on the western plateau, overlying the
33
34 108 Neoproterozoic crystalline basement rocks (e.g. Alemu et al., 2018; Le Gall et al.,
35
36 109 2018; Fig. 1). They also likely covered parts of the region that would become the
37
38 110 Afar Depression.

39
40
41 111 The initiation of rifting within the Afar is thought to have begun at ~29-26 Ma
42
43 112 postdating the main flood basalt event. Geological studies from the central and
44
45 113 southern parts of the western margin of Afar propose three main stages of strain
46
47 114 migration from the border faults to the current magmatic segments (Wolfenden et
48
49 115 al., 2005; Stab et al., 2016). In stage 1, starting just after the flood basalt deposits
50
51 116 of 29 to 26 Ma, narrow half-grabens initiate the rift, coeval with the emplacement
52
53 117 of rhyolites in these basins. Stage 2 (16-7 Ma) corresponds to the eastward
54
55 118 migration of extensional deformation towards the rift axis, and accompanied by
56
57 119 fissural basaltic volcanism. Finally, Stage 3 (<7 Ma) includes a further eastward
58
59 120 jump in the locus of extension and increase in localized magma intrusion and
60
61
62
63
64
65

1 121 voluminous basalts. Stage 3 is responsible for eruption of a ~1.5-km-thick Stratoid
2
3 122 Series lava pile of mostly basalts emplaced at 4.0-1.1 Ma that now covers 70% of
4
5 123 Afar (Varet, 1975; Stab et al., 2016; Fig. 1). After 1.1 Ma, the deformation and
6
7 124 magmatism became localized within discrete seismically and volcanically active
8
9 125 magmatic segments characterized by intense diking and volcanism (e.g. Barberi et
10
11 126 al., 1972; Abdallah et al., 1979; Hayward and Ebinger, 1996; Manighetti et al.,
12
13 127 1998; Wright et al, 2006; Grandin et al., 2009).

14 128 Extensional basins in southern and central Afar are filled with Pliocene–Recent
15
16 129 lacustrine and fluvial sedimentary rocks interbedded with basaltic and felsic
17
18 130 lava flows (Tesfaye et al., 2003). In northern Afar, the Danakil basin is known to
19
20 131 include at least 1 km of Holocene-Recent evaporites (e.g. Bastow et al., 2018; Le
21
22 132 Gall et al., 2018). However, the sedimentary basin likely extends to 3-5 km depth
23
24 133 with Pliocene-Pleistocene-Recent age evaporites and basalt flows (Bastow and
25
26 134 Keir, 2011; Keir et al., 2013).

27 135 The western Afar margin preserves the rift morphology since the onset of the
28
29 136 flood basalt till the initiation of the seafloor spreading (Tesfaye, et al., 2003;
30
31 137 Wolfenden, et al., 2005; Keir, et al., 2013; Ebinger, et al., 2017; Rooney, et al., 2018).
32
33 138 This margin is characterized by stratigraphic and structural changes from the
34
35 139 south toward the north (e.g. Keir et al., 2013). For example, south of latitude 13°N,
36
37 140 the transition from the western plateau, covered here by a thick sequence of Trap series,
38
39 141 to the extended crust in the Afar Depression corresponds to a zone of antithetic faults
40
41 142 forming a wide zone of marginal basins (Wolfenden et al., 2004, 2005; Stab et al.,
42
43 143 2016) . North of latitude 13°N, where the Trap series does not outcrop on the western
44
45 144 plateau, the margin is characterized by normal faults forming a narrow zone of
46
47 145 marginal grabens (Justin-Visentin and Zanettin, 1974; Kieffer et al., 2004; Wolfenden et
48
49 146 al., 2005; Rooney et al., 2018; Le Gall et al., 2018). The seismic activity is more
50
51 147 intense in this area with several events of magnitude greater than 5.0 (Ayele et al.,
52
53 148 2007; Illsley-Kemp et al., 2018). These changes along the Afar margin are spatially
54
55 149 associated with a northward thinning of the crust within the rift, and surface area of
56
57 150 Holocene volcanism (Barberi and Varet, 1977; Berckhemer et al., 1975; Makris and
58
59 151 Ginzburg, 1987; Hammond et al., 2011; Bastow and Keir, 2011).

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4 153 **2.2 Previous constraints on crustal structure**

5 154 Previous geophysical studies have contributed significantly to our knowledge of
6
7 155 the crustal structure of our study area. Estimates of the crustal thickness and
8
9 156 crustal composition beneath Afar and surrounding plateaus have been made using
10
11 157 controlled source seismic profiling (Makris and Ginzburg, 1987), inversion of
12
13 158 gravity data (e.g. Tiberi et al. 2005) and magnetotelluric imaging (e.g. Desissa et
14
15 159 al., 2013). Passive seismic techniques such as P-to-S and S-to-P receiver functions
16
17 160 (e.g. Hammond et al., 2011; Lavayssiere et al., 2018), surface waves tomography
18
19 161 (e.g. Gallacher et al., 2016) and ambient noise tomography (e.g. Korostelev et al.,
20
21 162 2015; Chambers et al., 2019).

22
23 163 Makris and Ginzburg (1987) used the data from an along strike seismic refraction
24
25 164 profiles to propose a four-layer crustal structure within the Afar Depression
26
27 165 underlain by an anomalous upper mantle (P_n velocity of $7.4-7.5 \text{ kms}^{-1}$). This model
28
29 166 consists of a 3-5 km sedimentary cover (a P-wave velocity, V_p of $3.35-3.95 \text{ kms}^{-1}$),
30
31 167 that overlies a basaltic layer ($V_p=4.5 \text{ kms}^{-1}$) on an abnormally thin upper crust (V_p
32
33 168 of $6.1-6.2 \text{ kms}^{-1}$) and a lower crust (V_p of $6.7-7.0 \text{ kms}^{-1}$). This study shows that the
34
35 169 crust thins from 26 km in the south and central Afar to 14 km in the north. Crustal
36
37 170 thickness inferred from gravity data (Makris et al., 1991; Tiberi et al. 2005), shows
38
39 171 thin crust beneath the Afar rift (14-23 km) and thick ($\sim 40 \text{ km}$) continental crust
40
41 172 beneath the western plateau that is partly intruded with high-density material.

42 173 Previous P-to-S receiver function studies (Dugda et al., 2005; Stuart et al., 2006;
43
44 174 Hammond et al., 2011; Reed et al., 2014; Kibret et al., 2019; Wang et al., 2021) and
45
46 175 a joint inversion of Rayleigh wave velocities and receiver functions (Dugda et al.,
47
48 176 2007) confirm these variations, showing $\sim 26-20 \text{ km}$ thick crust beneath the Afar
49
50 177 Depression and rapidly thinning to $\sim 16 \text{ km}$ beneath the Danakil Depression of
51
52 178 northern Afar (e.g. Hammond et al., 2011). The estimated crustal thickness
53
54 179 beneath the western and southern plateaus range from 36 to 45 km (Dugda et al.,
55
56 180 2005; Stuart et al., 2006; Cornwell et al., 2010; Hammond et al., 2011; Kibret et al.,
57
58 181 2019; Wang et al., 2021) . These studies also show a normal to elevated V_p/V_s
59
60 182 ratio for the plateaus and reaching very high values ($V_p/V_s > 2.0$) near the

1 183 magmatic segments where the crustal thickness is less than 26 km (e.g. Hammond
2 184 et al., 2011; Wang et al., 2021). A seismic anisotropy study (Hammond, 2014)
3 185 shows that the melt beneath the Afar Depression is stored in interconnected
4 186 stacked sills within the lower crust consistent with high Vp/Vs ratios more than
5 187 1.95 obtained by receiver functions, an interpretation supported by slow lower crustal
6 188 S-wave velocities imaged using ambient noise tomography (e.g. Chambers et al., 2019;
7 189 Korostelev et al., 2015). In addition, S-to-P receiver functions identify a velocity
8 190 decrease with depth at 65-75 km beneath the western plateau interpreted as the
9 191 lithosphere-asthenosphere boundary (LAB; Rychert et al., 2012; Lavayssiere et al.,
10 192 2018). The LAB is not imaged below Afar and interpreted as evidence for melt
11 193 percolation into the mantle lithosphere (Rychert et al., 2012; Lavayssiere et al., 2018).
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22 194 However, the seismic stations used in the previous studies (Dugda et al., 2005;
23 195 Stuart et al., 2006; Hammond et al., 2011; Lavayssière et al., 2018; Chambers et al.,
24 196 2019; Wang et al., 2021) were deployed at a spacing of several tens of kilometers
25 197 and those used by Reed et al. (2014) were located along a profile across the south-
26 198 eastern part of central Afar. Thus, these studies are limited in providing high-
27 199 resolution images of the crustal structure from the continental stable plateaus to
28 200 the active rift axes.

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202 ***3. Data and Methodology***

204 ***3.1 Data***

205 From May 2017 to September 2018, a network including 9 SEIS-UK broadband
206 seismic stations (7 Guralp CMG-3ESPD-60s and 2 CMG-40T-30s with Guralp CMG-
207 DCM datalogger) and 20 French Sismob-RESIF broadband seismic stations (Guralp
208 CMG-40T-30s with Nanometrics Taurus digitizer) were deployed along two SW-
209 NE oriented profiles, more or less perpendicular to the large fault scarps forming
210 the western Afar margin and the main axis of the Afar rift (Fig. 1). This network
211 has been configured to image the crustal structure across the transition from the
212 western plateau to the currently active rifts of the northern and central parts of
213 Afar (Fig. 1). For northern Afar we used 8 stations (N001 to N008) deployed along

1 214 an 86 km-long profile spaced by an average distance of 11 km (Keir et al., 2017;
2
3 215 Fig. 1). For central Afar we used 15 stations (S001, S003 to S009, S011 to S014,
4
5 216 S017, S018 and S020) deployed along ~225 km-long profile spaced by either ~20
6
7 217 km at the ends or ~10 km in the middle of the profile (Dobre et al., 2021; Fig. 1).
8
9 218 For improved continuity of the southern profile, we added data from FINE station
10
11 219 installed from March 2007 to November 2009 (Ebinger, 2007). All stations
12
13 220 recorded continuous data with a sampling rate of 100 Hz except FINE, which
14
15 221 recorded continuous data at 50 Hz.
16
17 222 For the receiver function analysis, the IRIS earthquake catalog was searched for
18
19 223 events with magnitude ≥ 5.5 Mb, occurring within an epicentral distance range of
20
21 224 30° - 95° from the center of the network. To increase the azimuthal coverage,
22
23 225 regional events between 20° and 30° were also included (e.g. Park and Lavin, 2001;
24
25 226 Salmon et al. 2011; Fig. 2d), with a specific treatment detailed in section 3.2.1. In
26
27 227 total, waveforms from 186 earthquakes including 11 regional events have been
28
29 228 selected.
30
31 229 SAC software (Goldstein, et al., 2003; Goldstein and Snoke, 2005), TauP toolkit
32
33 230 (Crotwell et al. 1999) and the 1D IASP91 velocity model (Kennett and Engdahl,
34
35 231 1991), were used to pick the P-wave arrival time and to prepare the raw seismic
36
37 232 data for receiver function analysis. The data were manually inspected and rated
38
39 233 based on the signal-to-noise ratio (S/N; the ratio between the amplitude of the signal
40
41 234 at P-arrival time to the maximum amplitude of the 5 s pre-arrival noise window), only
42
43 235 low-noise traces (S/N > 4) with clear P wave arrival recorded at each station were
44
45 236 kept for further analysis. This process resulted in the selection of ~126
46
47 237 earthquakes (including 11 regional events) for receiver function construction and
48
49 238 processing (Fig. 2d). After removing the mean and the first-order trends from the
50
51 239 selected waveforms, the data were filtered using a zero-phase Butterworth
52
53 240 bandpass filter with corner frequencies of 0.02–0.8 Hz and windowed 5 s before
54
55 241 and 35 s after the theoretical P arrival time. We then rotated the horizontal
56
57 242 seismograms into a great-circle path from the ZNE coordinate system to the ZRT
58
59 243 coordinate system.
60
61 244

1 245 **3.2 Methods**
2
3 246
4
5 247 **3.2.1 P-to-S receiver functions**
6
7 248 The receiver function technique is widely used to image seismic discontinuities
8
9 249 beneath a seismic station by deconvolving the vertical component from the radial
10 250 and/or transverse components. This aims to remove the instrument response, the
11
12 251 propagation path and the source mechanism effects from a broadband teleseismic
13
14 252 record (Langston, 1977; 1979). We used the iterative time domain deconvolution
15
16 253 technique of Ligorria and Ammon (1999) to compute the receiver functions with
17
18 254 200 iterations. We used a Gaussian width factor of 2.5 in the deconvolution of the
19
20 255 vertical trace to predict the radial one (Ligorria and Ammon, 1999), except for
21
22 256 noisy stations for which we used a Gaussian width factor of 2.0 to excluded high
23
24 257 frequency noise above ~ 1 Hz. The percentage of recovery was evaluated from the
25
26 258 rms misfit between the original radial waveform and the radial receiver function
27
28 259 convolved with the vertical component. If the final deconvolution reproduces less
29
30 260 than 80 % of the signal, then the event is discarded from further analysis. From
31
32 261 this together with a visual inspection for coherence and stability, the number of
33
34 262 receiver functions included in the final analysis varies between 15 and 72 per
35
36 263 station (Table 1), depending on the background noise and the state of health of the
37
38 264 station.
39
40 265 Extra care is required if using regional events in receiver function analysis. They
41
42 266 are susceptible to complex P-wave arrivals due to upper mantle triplications (e.g.,
43
44 267 Chu et al., 2012). To check for any artifacts due to triplicated phases, we only kept
45
46 268 the events that will not change the H and/or Vp/Vs beyond the error limits
47
48 269 determined from the teleseismic events alone. For example, station S005 has the
49
50 270 maximum number of 10 regional events, whereas for most of the other stations
51
52 271 they are less than six. Figures S1b,c shows the RFs and H-k plots from station S005
53
54 272 with and without the regional events included, respectively. The estimated crustal
55
56 273 thickness and Vp/Vs are approximately the same. Similarly, figure S1d, shows
57
58 274 binned and stacked receiver functions in epicentral distance of 10° for three
59
60 275 stations that have six or more regional events. It is clear that the results from

1 276 regional events (20°-30°) are comparable to the results from teleseismic events
2
3 277 (30°-95°) for each station.
4

5 278

6 279 3.2.2 *H-κ Stacking*

7
8 280 To determine the average crustal properties, we analyzed receiver functions for
9
10 281 each station using the H-κ domain stacking technique of Zhu and Kanamori
11
12 282 (2000). This method enables the determination of Moho depth (H) and the
13
14 283 velocity ratio of crustal P and S phases (Vp/Vs, or κ) by considering the crust as a
15
16 284 homogeneous, horizontal, isotropic layer over a half-space. The inherited trade-off
17
18 285 in the receiver function analysis between the crustal thickness (H) and the average
19
20 286 crustal velocity properties can be partially resolved using the H-κ stacking
21
22 287 algorithm (Ammon et al., 1990), assuming a good assumption can be made about
23
24 288 the average P-wave velocity from nearby refraction profiles (e.g. Makris and
25
26 289 Ginzburg, 1987). With this information, we sum amplitudes of the receiver
27
28 290 functions at predicted arrival times for the Moho P-to-S conversion phase Pms and
29
30 291 its multiple converted phases, PpPs and PpSs+PsPs using different weights and a
31
32 292 range of H and Vp/Vs values.

33 293 The stacking amplitude in the H-k domain is given by:

34 294

$$35 \quad s(H, \kappa) = \sum_{m=1}^n w_1 r_m(t_{p_s}) + w_2 r_m(t_{p_p p_s}) + w_3 r_m(t_{p_p s_s + p_s p_s}) \quad (1)$$

36 295

37 296 where n is the number of the radial receiver functions at each station, W_j is a
38
39 297 weighting factor that represents the contribution of the direct conversion from the
40
41 298 Moho and its multiples according to signal-to-noise ratio ($W_1 + W_2 + W_3 = 1$) and
42
43 299 $r_m(t)$ is the amplitude of the point on the receiver function at predicted arrival
44
45 300 time t (after the first P arrival) of the associated seismic phase corresponding to
46
47 301 the crustal thickness H and Vp/Vs. The stacking amplitude reaches its maximum as
48
49 302 the three phases stack coherently, and represents the best estimate for both H and
50
51 303 Vp/Vs (κ) beneath the station. Globally, we follow the same weighting factor
52
53 304 scheme $W_1=0.7$, $W_2=0.2$ and $W_3=0.1$ of Zhu and Kanamori (2000) that balance the
54
55 305 contribution of each phase proportional to low signal-to-noise ratio of multiples.
56
57 306 We modified the weighting factors in a few cases, by increasing those for multiples
58
59 307 when the Moho conversion phase amplitude is low ($W_1=0.5$, $W_2=0.3$ and $W_3=0.2$):
60
61 308

1 309 N008; Figs. 2 and 3) or when high-amplitude intra-crustal interface conversion
2 310 phases obscured the Moho Pms conversion ($W_1=0.4$, $W_2=0.3$ and $W_3=0.3$: N002;
3 311 Fig. 2), or by decreasing them when the multiples are less clearly identifiable
4
5 312 ($W_1=0.8$, $W_2=0.1$ and $W_3=0.1$: S018: Fig. 2).

6
7
8 313 The V_p/V_s ratio derived from the H- κ stacking method are related to the elastic
9 314 parameter Poisson's ratio (σ) through the simple relationship $\sigma = 0.5(1 - \{1/[(V_p$
10 315 $/V_s)^2 - 1]\})$ (Zandt and Ammon 1995; Ligorria 2000), and is typically used to
11
12 316 provide important constraints regarding the bulk crustal composition (e.g.
13
14 317 Christensen, 1996; Chevrot and van der Hilst, 2000). Poisson's ratio variations
15
16 318 mainly depend on mineralogical composition of the crust (felsic, mafic) and/or the
17
18 319 presence of fluids rather than on pressure or temperature conditions (Christensen,
19
20 320 1996; Watanabe, 1993).

21
22
23 321

24 322 **3.2.3 Results stability and Uncertainty Estimates**

25
26
27 323 Three important factors control the stability of the H- κ inversion results using the
28
29 324 receiver function method: 1) the average crustal P-wave velocity used to do the
30
31 325 inversion; 2) the back-azimuth coherency of the Moho converted phase Pms and
32
33 326 multiples PpPs and PpSs+PsPs and 3) the level of noise in the receiver function
34
35 327 waveform data.

36
37 328 Average crustal P-wave velocity (V_p) used in previous receiver function studies in
38
39 329 Ethiopia shows a large range of values (4.6 kms^{-1} - 6.5 kms^{-1}). For example, Dugda
40
41 330 et al. (2005) used constant average crustal velocity $V_p = 6.5 \text{ kms}^{-1}$. Stuart et al.
42
43 331 (2006) used three different V_p values constrained from the EAGLE controlled-
44
45 332 source profile of 6.1 kms^{-1} , 6.15 kms^{-1} and 6.25 kms^{-1} for the southern, central
46
47 333 Ethiopian rifts, and adjacent plateaus, respectively. Hammond et al. (2011) used
48
49 334 values of 6.15 kms^{-1} and 6.25 kms^{-1} for the stations in Afar and on the Ethiopian
50
51 335 plateaus again based on nearby refraction profiles (Makris and Ginzburg, 1987).
52
53 336 Finally, the average crustal velocity used by Reed et al. (2014) ranges from 4.8
54
55 337 kms^{-1} to 5.9 kms^{-1} based on the comparison of the stacking results using different
56
57 338 initial V_p values ranging from 4.5 kms^{-1} to 7.0 kms^{-1} and theoretical V_p versus κ
58
59 339 curve. In this study, we used $V_p = 6.3 \text{ kms}^{-1}$ for the stations located on the western
60
61 340 plateau and 6.0 kms^{-1} for the stations located within the Afar rift based on the

1 341 controlled-source experiments (Berckhemer et al., 1975; Makris et al., 1975;
2 342 Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006), and the
3 343 average crustal velocity constraints from the previous receiver function studies
4 344 (Dugda et al., 2005; Stuart et al., 2006; Hammond et al., 2011; Reed et al., 2014).
5 345 To estimate the standard deviation for both crustal thickness H and V_p/V_s , we
6 346 employ the bootstrap resampling technique (Efron and Tibshirani 1986). The
7 347 bootstrap analysis was done by repeating the stacking procedure 200 times with
8 348 random data subsets that are resampled versions of the original data set for each
9 349 station. The 2σ for each station is shown by error bars in the depth sections and
10 350 supporting information. The same technique was applied to estimate the errors
11 351 introduced to the results from the assumed average crustal P-wave velocity used
12 352 to determine the bulk crustal parameters (H and V_p/V_s ; Tiberi et al., 2007). We
13 353 recalculate H and V_p/V_s values using different average crustal V_p in the range 5.8-
14 354 6.8 km s^{-1} to test the errors introduced by assuming average crustal V_p (Fig. S5).
15 355 As suggested by several authors (e.g. Mohsen et al., 2005; Nair et al., 2006; Ogden et
16 356 al., 2019), for crustal thickness of less than 50 km, a V_p variation of 0.1 km s^{-1} will
17 357 affect the Moho depth by a value less than 1 km.

18 358

19 359 **3.2.4 Receiver Function Depth Migration**

20 360 The common-conversion point (CCP) technique is a robust method to transfer the
21 361 receiver functions from the time-domain to the space-domain (e.g. Kind et al.,
22 362 2002) and image lateral variations of the subsurface discontinuities along a
23 363 profile. In this study, we perform a common conversion point (CCP) stacking using
24 364 the Zhu (2000) technique, based on the back projection of the amplitude vector of
25 365 the receiver functions along the ray paths using a given 1-D velocity model. The 3-
26 366 D space beneath the profile is binned and the amplitudes of all the radial receiver
27 367 functions with rays passing through each bin are averaged to get the amplitude for
28 368 that bin. To create our depth migrated cross-sections along both the northern and
29 369 southern profiles, we use a 1-D V_p model based on the nearby refraction profiles
30 370 (Berckhemer et al., 1975; Makris and Ginzburg, 1987; Mackenzie et al., 2005;
31 371 Maguire et al., 2006) and the IASP91 velocity model (Kennett and Engdahl, 1991).

1 372 The Vs model at each station location is deduced from the Vp/Vs results of the
2
3 373 current study, constrained by the previous RF studies (Stuart et al., 2006;
4
5 374 Hammond et al., 2011; Reed et al., 2014) and Vs models of Chambers et al. (2019).
6
7 375 Therefore, the velocity model varies along both profiles and is constrained to
8
9 376 maintain the average crustal velocity Vp comparable to that used for H and Vp/Vs
10
11 377 calculations. To optimize for the resolution, the CCP depth images are obtained
12
13 378 with a bin width of 10 km (equivalent to the estimated width of the first Fresnel
14
15 379 zone for an interface at 30 km depth, assuming a signal frequency of 1.2 Hz and S-
16
17 380 wave velocity of 3.4 kms⁻¹; e.g. Salmon et al., 2011). The separation between the
18
19 381 centers of the bins is set to 2 km to preserve the spatial resolution and the vertical
20
21 382 bin width is set to 0.5 km less than the expected vertical resolution equivalent to
22
23 383 the half of the minimum wavelength.
24

25 384

26 385 **4. Results**

27 386 Most of the stations show a strong and easily identifiable Pms Moho conversion
28
29 387 phase on individual and stacked receiver functions (Figs. 2, 3, S1 and S2). The
30
31 388 corresponding multiple PpPs, is also identifiable in most cases, but not as
32
33 389 consistently visible as Pms, while the latter arrival multiples PpSs+PsPs are often
34
35 390 ambiguous and only clearly observed at a few stations. The Moho conversion
36
37 391 phase and/or its multiples may be disturbed by reverberations from near surface,
38
39 392 low velocity sediment layers and intracrustal and/or upper mantle interfaces at
40
41 393 some stations (Figs. 3 and S1). In this study, we present the results ordered and
42
43 394 grouped depending on them being on the northern or southern profiles (Fig. 1)
44
45 395 and in light of different geological and tectonic features (e.g. Keir et al., 2013).
46

47 396

48 397 **4.1 Receiver functions observations**

49 398

50 399 **Northern profile**

51 400 For the stations located on the western plateau (N001 to N005), the Pms phase is
52
53 401 observed 6 s after the direct P-wave arrival time. Before the Pms phase, i.e. within
54
55 402 1.0-4.0 s after the direct P-wave arrival time, we observe a strong and coherent
56
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59
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62
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64
65

1 403 negative intra-crustal phase and two positive intra-crustal phases on the majority
2
3 404 of the individual (Figs. 3 and S1) and stacked radial receiver functions (Figs. 2 and
4
5 405 S2e). The arrival time and amplitude of the first positive phase depends on the
6
7 406 back-azimuth (Figs. 3 and S1), and could be a P-to-S conversion at the bottom of a
8
9 407 low velocity shallow layer or could suggests anisotropy or dipping layers. The
10
11 408 second strong positive phase is interpreted as a conversion from a mid-crustal
12
13 409 interface at ~20 km depth. A clear positive and coherent conversion phase (Phs)
14
15 410 arrives behind the Moho P-to-S conversion phase (Pms) at ~9.5 s delay time (Fig.
16
17 411 S1e), that Levin and Park (2000) associated with the Hales discontinuity and was
18
19 412 reported by previous RFs studies beneath Nubian and Arabian continental shields
20
21 413 (Sandvol et al., 1998; Park and Levin, 2001; Ayele et al., 2004).
22
23 414 The amplitudes of the transverse components are much smaller than the
24
25 415 corresponding radial amplitudes (Fig. 2). The transverse receiver functions show
26
27 416 an intracrustal negative and positive energy that is varying over the back-azimuth,
28
29 417 which may suggest the presence of anisotropy (Levin and Park, 1997a).
30
31 418 As shown in Fig. 3b, at station N006 the results show a variation in the amplitude
32
33 419 and arrival time of the Pms-converted phase on the radial RFs as well as the
34
35 420 polarity reversal of the direct P and Pms-converted phases on the transverse RF as
36
37 421 a function of the back-azimuth. This suggests the presence of a dipping Moho. The
38
39 422 azimuthal RF variations of the reverberated latter phases show a complex pattern
40
41 423 which may indicate the presence of 3-D features such as dipping interfaces and
42
43 424 seismic anisotropy within the crust. Such complex pattern and the lack of a
44
45 425 homogeneous earthquake distribution with back-azimuth, means it is not possible
46
47 426 to do a detailed crustal anisotropy analysis.
48
49 427 The data from the stations located within the marginal graben at the eastern end
50
51 428 of the northern profile (N007 and N008) show a positive intra-crustal phase in the
52
53 429 radial receiver functions at ~ 1.3-1.5 s after the direct P-wave arrival time (Figs. 2,
54
55 430 3 and S1). The most prominent positive phase visible in the majority of both
56
57 431 individual and stacked radial receiver functions (Figs. 2, 3 and S1) arrives within a
58
59 432 delay time of 3.0-5.0 s after the direct P-wave arrival and is interpreted as the P-to-
60
61 433 S conversion phase (Pms) at the Moho discontinuity. At the easternmost station
62
63
64
65

1 434 (N008), we note a clear and strong negative phase arrives following the Pms phase
2
3 435 at ~4.0 s delay time after the P-wave arrival (Fig. 2).
4

5 436

6
7 437 ***Southern profile***

8 438 The two westernmost stations (S001 and S003) on the western plateau are
9 439 characterized by a clear Pms Moho conversion phase arriving with a 6.0 s delay
10 440 time after the first P-wave arrival and multiples within 18-22 s delay time. A clear
11 441 pre-Moho conversion phase arrives with a 2.9 s delay time after the first P-wave
12 442 arrival likely originating from an intra-crustal discontinuity between the upper
13 443 and lower crusts at ~20 km depth. This phase is particularly coherent and
14 444 stronger in the S003 individual receiver functions and on the stacked receiver
15 445 functions of the two stations (Figs 2 and 3). The transverse RFs of the stations
16 446 S001 and S003 show a significant energy associated with a group of positive and
17 447 negative pulses preceding the Pms Moho conversion phase. At the expected arrival
18 448 time of the Pms conversion phase the transverse RFs display small positive
19 449 amplitudes for the events approaching the station from easterly back-azimuths,
20 450 (Fig. S2a).

21 451 The four stations (S004, S005, S006 and S007) positioned within the basin of the
22 452 foothill area between the western plateau and the marginal graben are
23 453 characterized by a clear Moho conversion phase (Figs. 1, 2 and S2b). At station
24 454 S004, the Pms conversion phase arrives with a delay of 5.0 s after the arrival of the
25 455 direct P phase with PpPs and PsPs multiples arrives at 16.5 s and 21.0 s,
26 456 respectively. At stations (S005, S006 and S007) the Moho conversion phase Pms
27 457 and its multiples (PpPs and PsPs) arrive at ~4.0-4.5 s, ~14.8-15.2 s and ~18.5-
28 458 19.2 s delay time after the first arrival (Figs. 3, S1a and S2b). The transverse RFs
29 459 show significant energy from intracrustal interfaces and at the expected arrival
30 460 time of the Pms conversion phase, with the latter characterized by a polarity
31 461 change for the events approaching the stations from eastern (positive) and
32 462 western (negative) directions (Fig. S2b).

33 463 Stations sampling the marginal graben and Awra plain (S008, S009, S011, S012
34 464 and S013) show, a clear Pms conversion phase arrive within a time window of 3.8
35 465 to 4.0 s and its multiples PpPs and PsPs arrive at 12.0-12.5 s and 15-16.0 s after

1 466 the P-wave arrival time, respectively (Figs. 2, S1a and S2c). The transverse RFs
2 467 show a significant negative and positive energy between the direct P wave arrival
3 468 and the Moho converted phase characterized by reversed polarity in comparison
4 469 with the radial RFs (Fig. S2c).

5 470 At the five stations (S014, FINE, S017, S018 and S020), which are located near and
6 471 within the Afar rift axis, the radial receiver functions have two positive phases
7 472 within 4 s after the P-wave arrival time (Figs. 2, 3 and S1). The first low amplitude
8 473 phase arrival has a delay time of 1–1.5 s with respect to the direct P-wave, and is
9 474 more likely a conversion at the bottom of low velocity shallow layer. We interpret
10 475 the second strong and coherent phase arrival at a delay time of 3.5-3.8 s after the
11 476 direct P-wave with corresponding multiples at 10-10.5 s and 13.5-14.5 s, as the
12 477 conversion from the Moho (e.g. Figs. 2 and S1a and S2d). The transverse RFs
13 478 characterized by a clear negative and positive energy arrive between the onset of
14 479 the P-wave and the Pms Moho conversion phase. In addition, we observe a
15 480 significant positive energy at the expected arrival time of the Pms Moho
16 481 conversion phase for events approaching from east-southeast directions (Fig.
17 482 S2d).

18 483

19 484 **4.2 Crustal thickness and V_p/V_s ratio**

20 485 The H- κ stacking results show a well-defined peak on the H-k plots (Figs 4 and
21 486 S2). The crustal thickness H and V_p/V_s estimated from receiver function analysis
22 487 are shown on the topographic maps of Afar in Fig. 5a and Fig. 5b, respectively. The
23 488 estimated crustal thickness (H) and V_p/V_s ratio using the H- κ stacking technique
24 489 and associated errors from bootstrap method are shown in Table 1.

25 490

26 491 **Northern Profile**

27 492 Our crustal thickness estimates for the western plateau range from 37.6 ± 2.3 to
28 493 38.8 ± 1.9 km with an average crustal thickness of ~ 38.0 km. For station N006
29 494 located at the eastern edge of the western plateau results show two peaks on the
30 495 H- κ plot that correspond to discontinuities at ~ 28.0 km and 37.0 km, in addition to
31 496 another peak corresponding to an intra-crustal interface at ~ 20.0 km depth (Fig.

1 497 S3). By comparing the Pms piercing points of N006 (H ~28.0 and 37.0 km) with
2
3 498 the piercing points of N005 (H ~38 km) and N007 (H ~24 km) as shown by figure
4
5 499 S4, it is clear that a crustal thickness of ~37.0 km could be attributed to the
6
7 500 plateau side to the west of this station and ~28.0 km for the foothill area to the
8
9 501 east. We obtain 24.2 ± 1.5 km (N007) located at the foothill area and 19.4 ± 1.3 km
10
11 502 (N008) at the eastern edge of the marginal graben (Figs. 5a and S5).

12 503 The Vp/Vs ratio is relatively normal (1.73 ± 0.01 - 1.79 ± 0.05) for the stations on the
13
14 504 western plateau (N001-N005) and for the station (N007) at the western edge of
15
16 505 the marginal graben. At the eastern end of the northern profile at the riftward
17
18 506 edge of the marginal graben the N008 station shows high Vp/Vs ratio of $1.93 \pm$
19
20 507 0.02 .

21
22 508

23 509 ***Southern Profile***

24
25 510 The crustal thickness along the southern profile decreases from the W to E in
26
27 511 three steps, such that four different zones of uniform crustal thickness and Vp/Vs
28
29 512 ratio could be identified along the profile (Figs. 6 and S7). These zones broadly
30
31 513 correlate with the pre-rift crust of the western plateau, the foothill area between
32
33 514 the western plateau and the marginal graben, the stretched crust of the marginal
34
35 515 graben and Awra plain, and finally the thinned crust within and in the vicinity of
36
37 516 the Afar Manda-Harraro rift segment (MHRS) (Figs. 1, S7).

38 517 We estimate an average crustal thickness of 42.0 ± 1.0 km for (S001 and S003; first
39
40 518 zone) on the western plateau (Figs. 2, 5a and S7). Except for the S004 station
41
42 519 located on/or near the steep border faults with a crustal thickness estimated to
43
44 520 37.2 ± 4.0 km, the crustal thickness for (S005-S007; second zone) within the
45
46 521 foothill area is 34.0-34.6 km, with an average crustal thickness of 34.2 ± 1.0 km
47
48 522 (Figs. 5a, S7 and Table 1).

49 523 In the third zone of extended crust beneath the marginal graben and the Awra
50
51 524 plain, five stations (S008, S009, S011, S012 and S013) with a spacing of 5-12 km
52
53 525 were used to constrain the crustal thickness (Figs. S1, S7). The associated crustal
54
55 526 thickness ranges from 27.6 to 28.2 km (Table 1), with an average value of 28.0 km.

56
57 527 The H- κ estimates for the crustal thickness of (S014,FINE, S017, S018 and S020;
58
59

1 528 fourth zone) within and in the vicinity of the rift axis of central Afar range from
2
3 529 23.2 km to 19.6 km (Fig. S7, Table 1), with an average value of 21.0 ± 1.0 km.
4
5 530 Our Vp/Vs ratios for the southern profile could be grouped into two main groups
6
7 531 of the western marginal area and central rift area. Eleven stations located within
8
9 532 the western marginal area, eight of them associated with intermediate crustal
10
11 533 composition show elevated Vp/Vs ratios in the range from 1.79 to 1.87 and three
12
13 534 stations located within the second zone of the foothill area, show low to normal
14
15 535 Vp/Vs ratios in the range 1.72 ± 0.03 to 1.76 ± 0.02 . On the eastern end of the
16
17 536 southern profile, data from S014, FINE, S017, S018 and S020 stations located to
18
19 537 the west and within the MHRS show high Vp/Vs ratio of 1.95 ± 0.05 , 1.99 ± 0.02 ,
20
21 538 1.96 ± 0.02 , 1.92 ± 0.01 and 2.02 ± 0.01 , respectively (Figs. 5b and S7).
22

23 539

24 540 **4.3 Depth Migrated Profiles**

25 541 Figure 6 shows the migrated receiver functions along the northern and southern
26
27 542 profiles. The Moho depth obtained from the H-k stacking method of Zhu and
28
29 543 Kanamori (2000) and corrected for the station elevation is plotted with error bars
30
31 544 from the bootstrap technique of Efron and Tibshirani (1986).
32

33 545

34 546 ***Northern profile***

35
36 547 For the northern profile, the Moho interface appears as a strong positive phase
37
38 548 (indicated by the red color, Fig. 6a) and can be easily traced. It is almost horizontal
39
40 549 beneath the western plateau (N001-N006) at an average depth of ~ 36 km, and
41
42 550 then rises in two gradual changes to a depth of ~ 18 km below the eastern end of
43
44 551 the profile. The first crustal thinning (~ 14 km vertically over ~ 10 km horizontally)
45
46 552 occurs at the edge of the plateau between N006 station located on the western
47
48 553 plateau and N007 station located at the foot of the escarpment within the marginal
49
50 554 graben zone. The second crustal thinning occurs within the marginal graben zone
51
52 555 between the stations N007 and N008, (~ 5 km vertically over 15 km horizontally).
53
54 556 The crust continues to thin eastward to < 15 km towards Dallol within the Danakil
55
56 557 Depression (Makris and Ginzburg, 1987; Tiberi et al., 2005; Bastow and Keir, 2011;
57
58 558 Hammond et al., 2011; Illsley-Kemp et al., 2017).
59

60 559

1 560 ***Southern profile***

2
3 561 The SW-NE-migrated image along the southern profile (Fig. 6b) shows a clear
4
5 562 conversion from the Moho at a depth range from ~39-40 km at the western end to
6
7 563 ~19-20 km beneath the eastern limit of the profile. The first locus of the crustal
8
9 564 thinning occurs at the edge of the escarpment from 42 km beneath the western
10
11 565 plateau to ~34 km beneath the foothills zone (8 km vertically over 20 km
12
13 566 laterally). The second locus occurs between S007 and S008 stations located at the
14
15 567 foothills zone and the marginal graben zone, respectively, reaching an average
16
17 568 crustal thickness of 28 km (6 km vertically over 10 km horizontal distance). The
18
19 569 last locus of the crustal thinning is at the western edge of the Sullu-Adu range
20
21 570 located at the eastern half of Awra basin reaching an average crustal thickness of
22
23 571 ~21 km beneath the MHRS at the latitude ~12°N.

24 572

25
26 573 **5 Discussion**

27 574

28
29 575 **5.1 Comparison with previous estimates of the crustal structure**

30
31 576 ***Northern Profile:***

32
33 577 Our crustal thickness estimate of ~38 km for the western plateau of the northern
34
35 578 profile is in a good agreement with the previous receiver functions (~39.0 km;
36
37 579 Hammond et al., 2011; Ogden et al., 2019; Fig. 7), and from gravity inversion
38
39 580 (~35.0-38.0 km; Makris et al., 1991 ; Tiberi et al., 2001). Similarly, our Vp/Vs
40
41 581 values ~1.70-1.79 are consistent with the results of ~1.71-1.77 for the stations
42
43 582 ADYE and SMRE (Hammond et al., 2011; Ogden et al., 2019; Fig. 8). The crustal
44
45 583 thickness obtained from S-to-P receiver functions (Lavayssiere et al., 2018), and
46
47 584 ambient noise tomography (Chambers et al., 2019) is thinner than the crustal
48
49 585 thickness estimated by the current study, and likely the crustal thickness is
50
51 586 underestimated in their studies (e.g. see Lavayssiere et al., 2018 for a discussion
52
53 587 on this). At the eastern end of the northern profile, the crustal thickness and Vp/Vs
54
55 588 ratio of 19.4 ± 1.3 km and 1.93 ± 0.02 for N008 compare favorably with the values of
56
57 589 20.0 ± 2.0 km and 1.98 ± 0.1 for the nearby station HALE (Hammond et al., 2011;
58
59 590 Figs. 7, 8).

1 591 ***Southern Profile:***

2
3 592 Our estimated average crustal thickness of 42.0 km for the elevated western
4
5 593 plateau is in good agreement with the previous P-to-S receiver functions results
6
7 594 (38-44 km; Dugda et al., 2005; Stuart et al., 2006; Cornwell et al., 2010; Hammond
8
9 595 et al., 2011; Kibret, et al., 2019; Wang et al., 2021), with the crustal thickness
10
11 596 results from joint inversion of surface waves and Pms receiver functions (~40 km;
12
13 597 Dugda et al., 2007), with the estimates from seismic refraction profiles (40-45 km;
14
15 598 Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006) and
16
17 599 crustal thickness from gravity inversion (~40 km; Tiberi et al., 2005). For the
18
19 600 marginal area (marginal graben and Awra plain), our estimates of the crustal
20
21 601 thickness of $\sim 28.0 \pm 1.4$ km are consistent with the previous P-to-S receiver
22
23 602 function results (~ 25.0 - 29.0 ± 1.3 km; Hammond et al., 2011).

24 603 Our Vp/Vs results of ~ 1.79 - 1.87 ± 0.03 for most of the stations on the western
25
26 604 plateau and within the marginal area of central Afar (southern profile) are similar
27
28 605 to the results from the previous receiver function studies (~ 1.81 - 1.85 ± 0.05 ;
29
30 606 Hammond et al., 2011; Dugda et al., 2005; Figs. 8, S7). Regarding the stations at the
31
32 607 foot of the escarpment (half graben basin; southern profile), our Vp/Vs values of
33
34 608 ~ 1.72 - 1.76 ± 0.03 are consistent with the results obtained by Dugda et al. (2005)
35
36 609 and Hammond et al. (2011) for the nearby station DIYA.

37 610 In the easternmost zone within and in the vicinity of the rift axis of central Afar,
38
39 611 the crustal thickness and Vp/Vs results ($\sim 21.4 \pm 1.7$ and 1.99 ± 0.02) are similar to
40
41 612 the results obtained by Hammond et al. (2011) using the same data at the station
42
43 613 FINE. For the other stations our average crustal thickness value of 21.0 ± 1.0 km is
44
45 614 in a good agreement with result obtained by Hammond et al. (2011) and Reed et
46
47 615 al. (2014) (Fig. 7). Similarly, our high Vp/Vs values above 1.9 are comparable to the
48
49 616 results for the nearby stations obtained by Hammond et al. (2011) and less than
50
51 617 the values estimated by Reed et al. (2014) for the region to southeast of our
52
53 618 stations (Fig. 8). The higher values of Reed et al. (2014) may be the result of the
54
55 619 assumed average crustal velocity used in the receiver function analysis of ~ 5.1
56
57 620 kms^{-1} .

58 621

1 622 **5.2 Crustal thicknesses and architecture**

2
3 623 **5.2.1 Western Plateau**

4
5 624 For the stations located on the western plateau, observations of the high
6
7 625 amplitude coherent Moho converted arrivals and consistent estimated crustal
8
9 626 thickness beneath the western plateau stations of each profile indicate that the
10
11 627 Moho interface is mostly flat beneath the western plateau, while the intra-crustal
12
13 628 conversions are laterally variable in amplitude and depth at some stations. A
14
15 629 recent study of Ogden et al. (2019) show that the reliability of the high frequencies
16
17 630 RFs results are very low at the center of the flood-basalt province where the
18
19 631 gradational Moho thickness is >13.0 km and such reliability increases toward the
20
21 632 outside of the flood-basalt. However, they concluded that H- κ stacking maintains
22
23 633 reliability at low frequencies when the gradational Moho thickness is less than
24
25 634 13.0 km. Their study did not include results from the stations located nearby to
26
27 635 our stations. At the southern part of the western plateau, the wide-angle
28
29 636 reflection/refraction results at the latitude of the Main Ethiopian Rift (MER) show
30
31 637 a \sim 47.0 km thick crust. This study constrained the thickness of the high-velocity
32
33 638 layer at the bottom of the lower crust to be \sim 15.0 km with clear reflections from
34
35 639 the top and bottom of this layer and an apparent velocity of (7.4 kms⁻¹). Cornwell
36
37 640 et al. (2010), show that the high-velocity layer thins northeastward from 18.0 to
38
39 641 6.0 km with clear Pms conversions from the top and the bottom of the layer.
40
41 642 Our receiver function (calculated with both a maximum frequency of \sim 0.6 and
42
43 643 \sim 1.2 Hz; e.g. Cassidy, 1992; Julia et al., 2005) from the station located at the
44
45 644 eastern edge of the western plateau along the southern profile does not show
46
47 645 double peak Pms phase (Figs. 3a, d and S2a). Instead, it shows a large amplitude
48
49 646 intracrustal phase and Pms converted phase with a lag time of (6s), the latter
50
51 647 consistent with the Pms phase from the bottom of the high-velocity layer of
52
53 648 Cornwell et al. (2010), as well as a strong crustal reverberation (Fig. 3d). Our
54
55 649 results are consistent with a recent receiver function study, which yielded the
56
57 650 highest stacking amplitude values for the western plateau nearby to our stations
58
59 651 (Wang et al., 2021). This indicates that the Moho beneath the stations on the
60
61 652 western plateau is flat with a large change in seismic velocity contrast between the
62
63
64
65

1 653 crust and the mantle and does not allow us to constrain the thickness of the
2
3 654 intruded high-velocity layer. This could be due to a thin intruded layer (< 4.0 km)
4
5 655 below the vertical resolution owing to the wavelength of the incoming plane waves
6
7 656 and/or to a low velocity contrast between this layer and the overlying part of the
8
9 657 lower crust. However, if the crust-mantle transition is not a sharp boundary, our
10
11 658 estimated crustal thickness depends on the long wavelength of the teleseismic
12
13 659 waves we use, and more likely to select the center of the crust-mantle transition
14
15 660 gradient (e.g. Ogden et al., 2019). In that case, we estimate the crustal thickness
16
17 661 (include the magmatic additions) to be ~45.0-46.0 km in comparison with the
18
19 662 average crustal thickness of ~38.0 km for the plateau beneath the northern profile
20
21 663 (off-flood-basalt stations). In fact, the Jurassic sediments (pre-flood basalts
22
23 664 surface) lies at an average altitude of ~2200 m (Gani et al., 2007), and the higher
24
25 665 topography (~3700 m) of the western plateau at the latitude of the southern
26
27 666 profile is due to the emplacement of ~1500-1800 m-thick flood-basalts deposits at
28
29 667 the top of the Jurassic sediments. Hence, if the crust-mantle transition is a sharp
30
31 668 boundary, then this has the implication that the western plateau still under
32
33 669 dynamic support process as proposed by Sembroni et al. (2016). On the other
34
35 670 hand, if the crust-mantle is a gradational boundary of ~8.0 km thickness, the
36
37 671 implications are that the western plateau is locally compensated with magmatic
38
39 672 additions at the base of the lower crust as Tiberi et al. (2005) concluded from
40
41 673 interpretation of gravity and seismic data.

42 674 Most of the western plateau stations of the northern and southern profiles show a
43
44 675 clear intra-crustal conversion phase with a lag time of ~2.9-3.0 s from a mid-
45
46 676 crustal discontinuity at ~20.0-21.0 km depth, likely showing the upper crust-
47
48 677 lower crust boundary beneath the western plateau, similar to the discontinuity
49
50 678 mentioned by Hammond et al. (2011). Toward the southern part of the western
51
52 679 plateau at the latitude of the Main Ethiopian Rift (MER), a clear intracrustal
53
54 680 interface is modeled by the active seismic profile. This interface separates the
55
56 681 upper and lower crust with V_p of 6.1-6.4 kms^{-1} and 6.6-7.1 kms^{-1} , respectively
57
58 682 (Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006). A
59
60 683 similar interface has been reported and interpreted as the Conrad discontinuity in

1 684 previous receiver function studies of the eastern Red Sea conjugate margin of
2
3 685 Yemen and southeastern Gulf of Aden margin (Ahmed et al., 2013, 2014). It was
4
5 686 estimated at the same depth of ~21.0 km beneath the Yemeni high plateau covered
6
7 687 by the Cenozoic continental flood basalts (Ahmed et al., 2013). Previous seismic
8
9 688 refraction profiles across the Arabian shield confirm the existence of this interface
10
11 689 between the upper crust with an average Vp of 6.3 kms⁻¹ and the lower crust with
12
13 690 an average Vp of 7.0 kms⁻¹ (Mooney et al., 1985; Egloff et al. 1991). This
14
15 691 discontinuity suggests a mafic composition for the lower crust, likely altered by
16
17 692 magmatic intrusions or partial melt as a result of Miocene-Oligocene flood basalt
18
19 693 and the ongoing magmatic activity (Hammond et al., 2011; Ahmed et al., 2013).
20
21 694 Interestingly, this interface is almost imaged at the same depth of 20-21 km
22
23 695 beneath the Arabian shield, beneath Jordan (Mooney et al., 1985; Stern and
24
25 696 Johnson, 2010; and references therein), beneath Yemen plateau (Ahmed et al.,
26
27 697 2013) and beneath the western plateau in this study. Although the exact physical
28
29 698 meaning of this discontinuity is not clear, it is most likely separates the slower
30
31 699 brittle upper crust from the faster ductile lower crust (Stern and Johnson, 2010).
32
33 700 Our Vp/Vs values for the undeformed northern part of the western plateau with
34
35 701 1000-2000 m sedimentary cover are below 1.80 with an average value of ~1.74.
36
37 702 This is lower than the global average of 1.77 and best explained by Precambrian
38
39 703 crust of felsic to intermediate compositions (Zandt and Ammon, 1995;
40
41 704 Christensen, 1996), which is lacking significant magmatic modification before or
42
43 705 during the breakup (Christensen and Mooney, 1995; Thompson et al., 2010; Ogden
44
45 706 et al., 2019).
46
47 707 The western plateau at the latitude of the southern profile is characterized by
48
49 708 elevated Vp/Vs values which may indicate that the lower crust composition has
50
51 709 been modified due to mafic intrusions associated with the Cenozoic flood basalts
52
53 710 (Hofmann et al., 1997; Pik et al., 1999; Cornwell et al., 2010; Hammond et al.,
54
55 711 2011), partial melt in the lower crust (Hammond et al. 2011; Ahmed et al. 2013),
56
57 712 or be fluids circulation within the upper crust (Keir et al., 2009; Korostelev et al.,
58
59 713 2015; Chambers et al., 2019).
60
61 714

1 715 **5.2.2 Crustal thinning and composition across western Afar margin**

2
3
4 716 **Northern Afar**

5 717 For the first time we create a high resolution profile of the crustal structure across
6
7 718 the northern Afar margin into the Danakil Depression that offers an opportunity to
8
9 719 interpret the rifting evolution leading to the establishment of the currently highly
10
11 720 active segment of a divergent plate boundary. Taking into account the previous
12
13 721 constraints on the crustal thickness near Erta'Ale rift zone and the Dallol
14
15 722 Depression deduced by Makris and Ginzburg (1987) and Hammond et al. (2011),
16
17 723 the variations of the Moho depth estimated in this study can be extended from the
18
19 724 margin to the active rift axis.

20 725 The onset of the crustal extension occurs at the ~1400 m high escarpment
21
22 726 between N006 and N007 stations, (i.e. between 39.8°E-39.9°E; Figs. 1, S6) with
23
24 727 ~14.0 km vertical thinning over ~10.0 km horizontal distance. The crustal
25
26 728 thinning continues eastward in two more gradient changes before the crust
27
28 729 reaches a thickness of ~14.0 km at the rift axis (Berckhemer et al., 1975; Makris
29
30 730 and Ginzburg, 1987). We note that, the lateral change of the crustal thickness
31
32 731 across the western margin of northern Afar shows comparable crustal thickness
33
34 732 variations to that beneath Yemen which is the eastern Red Sea conjugate margin of
35
36 733 Ethiopia, with ~35.0 km beneath the Yemen plateau, ~22.0 km for the Yemen
37
38 734 coastal plain and a lowest value of ~14.0 km on the Yemen Red Sea coast (Ahmed
39
40 735 et al., 2013; Fig. 7).

41 736 Our close station spacing allows us to constrain the locus of the crustal extension
42
43 737 more precisely than previous studies. Assuming a pre-rift crustal thickness of 38.0
44
45 738 km beneath the western plateau, a crustal thickness of ~19.0 km beneath the
46
47 739 marginal graben and reaching the final thickness of ~14.0 km at the rift axis, we
48
49 740 estimate a stretching factor β of ~2.0 for the marginal graben and β of ~2.7 for the
50
51 741 northern Afar area between the western plateau and the rift axis. This factor
52
53 742 obtained for northern Afar is slightly higher than the maximum value obtained for
54
55 743 the Yemen Red Sea conjugate margin ($\beta \sim 2.5$, Ahmed et al., 2013).

56 744 There is a clear eastward change of the V_p/V_s ratio from low-to-intermediate
57
58 745 values, below the western plateau and the area at the foot of the escarpment (<

1 746 1.80) to high values for the area near the eastern border faults of the marginal
2
3 747 graben (> 1.90), and reaching very high values (>2.0) at the current rift axis (e.g.
4
5 748 Hammond et al., 2011). The high V_p/V_s ratio we found for the station N008
6
7 749 indicates an important change in the bulk composition of the crust due to mafic
8
9 750 intrusions along the eastern boundary antithetic faults of the marginal graben
10
11 751 (Illsley-Kemp et al., 2018; Zwaan et al., 2020b) or could be due to the existence of
12
13 752 crustal fluids. The reported eastward increase in the V_p/V_s ratios toward the rift
14
15 753 axis (Hammond et al., 2011) suggests more mafic/ultramafic material and/or may
16
17 754 indicate a partial melting within the crust increasing with proximity of the rift axis.
18
19 755 This is consistent with observations of recent magma intrusion related ground
20
21 756 deformation at the axis of the Danakil Depression (e.g. Nobile et al., 2012; Pagli, et
22
23 757 al., 2012).
24
25 758 The structure of the northern Afar margin has been primarily controlled by a
26
27 759 stretching phase in the early stages of continental rifting c.a ~ 26 -20 Ma (Ukstins et
28
29 760 al., 2002; Wolfenden et al., 2005), most likely related to the extension induced by
30
31 761 the divergence between Arabian and Nubian plates. It should be noted that during
32
33 762 this tectonic event the thickness of both the upper crust and lower crust decreases
34
35 763 by a factor of two. This is evident from the depth of the mid-crustal discontinuity
36
37 764 dropping from ~ 20.0 km beneath the western plateau to ~ 9.0 -10.0 km beneath
38
39 765 the eastern edge of the marginal graben (station N008; Figs. 6a, S2e). In the late
40
41 766 stages of continental rifting after 11 Ma the extension between the marginal
42
43 767 grabens and the rift axis is accommodated by thinning of the upper crust by a
44
45 768 factor of 2, while the lower crustal thickness remains constant. This is also evident
46
47 769 from the mid-crustal discontinuity at ~ 5.0 km beneath the Danakil Depression
48
49 770 (Makris and Ginzburg, 1987). This significant thinning of the upper crust over
50
51 771 short length scales is consistent with geological constraints suggesting that rifting
52
53 772 along the marginal area of northern Afar associated with upper crustal extension
54
55 773 and dominant tectonic deformation (Bastow & Keir, 2011; Keir et al., 2013; Bastow
56
57 774 et al., 2018). In contrary, an increase in rift magmatism associated with dike
58
59 775 intrusions confined to the rift axis has initiated relatively recently (Quaternary) in
60
61 776 the evolution of the rifting of northern Afar (e.g. Bastow and Keir, 2011; Nobile et
62
63 777 al., 2012; Pagli et al., 2012).

1 778
2
3 779 **Central Afar**
4
5 780 Unlike the northern profile, in central Afar the number of stations along our
6
7 781 southern profile allows us to continuously image the Moho depth and to constrain
8
9 782 the crustal composition variations from the western plateau to the rift axis. The
10
11 783 crustal thickness decreases eastward by ~ 8.0 km over ~ 20.0 km horizontal
12
13 784 distance between the western plateau and the second zone located at the foot of
14
15 785 the ~ 1700 m high escarpment. Then it decreases in two more steps by ~ 6.0 km
16
17 786 and ~ 7.0 km over ~ 10.0 km horizontal distance from the second to the third and
18
19 787 to the fourth zones, respectively (Fig. S7). The estimated β factors between these
20
21 788 zones are 1.24, 1.21 and 1.30 respectively. We estimate an overall β of 2.1 for
22
23 789 central Afar assuming an initial crustal thickness of ~ 42.0 km beneath the western
24
25 790 plateau and a final crustal thickness of ~ 20.0 km (S020) at the eastern end of the
26
27 791 profile. Our estimated stretching factor β for central Afar is comparable with the
28
29 792 estimate from the previous receiver function studies (Hammond et al., 2011) but
30
31 793 remains smaller than the β value of 3.0 and 2.7 predicted from plate
32
33 794 reconstructions (Wolfenden et al., 2005) and from cross section balancing (Stab et
34
35 795 al., 2016), respectively. The discrepancy between our estimated β and the
36
37 796 predicted β is best explained by magmatic addition to the crust during extension
38
39 797 (e.g. Thybo and Nielsen, 2009; Stab et al., 2016). This implies 4.0 or 5.0 km of
40
41 798 magmatic addition depending on the predicted β value used.
42
43 799 Most of the stations located within the western marginal area of central Afar
44
45 800 (southern profile) show elevated V_p/V_s ratios which suggest intermediate to mafic
46
47 801 composition except three stations that show low to normal V_p/V_s ratios. The latter
48
49 802 with low to normal values are located within the foothill area and reflect that the
50
51 803 Precambrian felsic crust has been poorly contaminated or intruded by mafic rocks.
52
53 804 These low V_p/V_s values could also result from the emplacement of felsic volcanic
54
55 805 strata coincidence with the onset of faulting along the central Afar margin (26-22
56
57 806 Ma; Wolfenden et al., 2005; Ayalew et al., 2019), and/or that the crust has been
58
59 807 intruded by granitic plutons (fast S-wave mantle material) such as along the
60
61 808 Yemen conjugate margin (Hughes and Beydoun, 1992; Davison et al., 1994;
62
63 809 Geoffroy et al., 1998; Ahmed et al., 2013). The slightly elevated values for the other

1 810 stations that are distributed in the marginal graben and the Awra plain may
2
3 811 indicate that the lower crust composition has been modified due to mafic
4
5 812 intrusions associated with the Cenozoic flood basalts (Hofmann et al., 1997; Pik et
6
7 813 al., 1999; Cornwell et al., 2010; Hammond et al., 2011), partial melt in the lower
8
9 814 crust (Hammond et al., 2011; Ahmed et al., 2013; Wang et al., 2021), fluids
10
11 815 circulation within the upper crust (Keir et al., 2009; Korostelev et al., 2015;
12
13 816 Chambers et al., 2019), or thick sedimentary deposits within the basin (Stab et al.,
14
15 817 2016).

16 818 Finally, at the vicinity and within the MHRS, the significant increase in the V_p/V_s
17
18 819 ratio is coincident with the thin crust (20.0 km) toward the central axis of the rift.
19
20 820 Such high values of V_p/V_s ratio above 1.90 are very rare (except for few minerals
21
22 821 like serpentine) and likely indicate a small amount of partial melt and/or fluids in
23
24 822 the crust (Watanabe, 1993; Christensen, 1996; Zandt and Ammon, 1995).
25
26 823 Hammond et al. (2011) used receiver function forward modeling for three stations
27
28 824 located west and northwest of MHRS (to the north of our stations) to show that
29
30 825 partial melt in the lower crust may exist beneath regions of thinner crust of central
31
32 826 Afar. Reed et al. (2014) used the theoretical velocity-melt relationship developed
33
34 827 by Watanabe (1993) for granitic rocks with rhyolitic melt, to estimate 4.2-11.4%
35
36 828 melt fraction at the location of their stations located in central Afar to the
37
38 829 southeast of our stations.

39 830 Additional observations supporting the existence of partial melt within the crust
40
41 831 at the vicinity and within the MHRS come from imaged slow shear wave velocities
42
43 832 within the crust and the uppermost mantle beneath the magmatic segments
44
45 833 (Hammond et al., 2011; Stork et al., 2013; Korostelev et al., 2015; Chambers et al.,
46
47 834 2019), magnetotelluric results that also point to a wide conductive zones within
48
49 835 the crust beneath MHRS interpreted as a large magma volumes that contains
50
51 836 ~13% melt fraction (Desissa et al., 2013; Didana, et al., 2014; Johnson et al., 2016)
52
53 837 and recent series of dike intrusions at MHRS (Keir et al., 2013). Finally, Hammond
54
55 838 (2014) invert $H-\kappa$ data at three stations within and in the vicinity of MHRS for melt
56
57 839 fraction and anisotropy and show that the melt is likely stored in the form of
58
59 840 stacked sills within the lower crust with a maximum thickness of 2.8 km and
60
61 841 connected by some melt inclusions in preferential vertical alignment. Hammond

1 842 (2014) shows that the anisotropy, and resultant amplitude dominance of the slow
2
3 843 shear wave in H-k stacking is a likely cause of the high V_p/V_s ratios for the average
4
5 844 crust of the MHRS.

6
7 845 The crustal structure variations across the western margin of central Afar are
8
9 846 consistent with the distinct geological and structural domains of this region. From
10
11 847 the west, the early stretching stage of the continental rifting (c.a ~26-23 Ma;
12
13 848 correspond to first step of the Moho interface between S003 and S004; second
14
15 849 zone; Fig. S7) occurs below the main escarpment. This stretching phase is
16
17 850 characterized by large offset normal fault, distributed over a wide area (Stab et al.,
18
19 851 2016) and accompanied with the formation of the half graben basin at the foot of
20
21 852 the escarpment and the emplacement of felsic magmatism (Wolfenden et al.,
22
23 853 2005). The following extension event occurs ~30.0 km eastward below the
24
25 854 western bounding faults of the marginal graben and corresponding to the second
26
27 855 step of the Moho interface from the second to the third zones (between S007 and
28
29 856 S008; Fig. S7). This extension event occurs after the emplacement of many
30
31 857 Miocene giant dykes within and in the vicinity of the marginal graben (Chorowicz
32
33 858 et al., 1999; Stab et al., 2016; Ayalew et al., 2019; Zwaan et al., 2020b). Zwaan et al.
34
35 859 (2020a, b) suggest that this event is related to the rotation of the Danakil block
36
37 860 and dated from geological studies as between 16-7 Ma (Wolfenden et al., 2005;
38
39 861 Zwaan et al., 2020a, b). During this extension event the stretching was distributed
40
41 862 over a wide area, most likely accompanied with the formation of antithetic normal
42
43 863 faults and NE tilted blocks were created (Stab et al., 2016). A third extension event
44
45 864 occurs below the western edge of Sullu-Adu range (between S013 and S014; Figs.,
46
47 865 1, S7). This event was accompanied with uplift of the Sullu-Adu area, subsidence of
48
49 866 the Awra plain and extensive magmatism (Bosworth et al., 2005; Stab et al., 2016).
50
51 867 However, the results deduced from our dense network of stations indicate a
52
53 868 differential extension between the brittle and ductile layers of the crust, such that
54
55 869 the upper crust thins from ~21.0 beneath the western plateau to ~7.0 km beneath
56
57 870 the fourth zone, whereas the lower crust thins from ~21.0 km to ~12.0 km. Hence,
58
59 871 these observations requires a décollement or ductile fault surfaces to
60
61 872 accommodate the upper crust lower crust differential extension and distribute the
62
63 873 upper crust extension over a wide area.

1 874
2
3 875 **5.3 Implications for the evolution of the rifting**
4
5 876 Continental rifting leading to continental breakup and oceanic spreading
6
7 877 encompasses a complex interplay of extensional faulting, thinning of the crust and
8
9 878 lithosphere and intrusive-extrusive magmatism (Buck, 2004; Geoffroy, 2005;
10
11 879 Lavier and Manatschal, 2006; Cannat et al., 2009; Leroy et al., 2010; Nonn et al.,
12
13 880 2019). Because of the geological observations that have been preserved in Afar
14
15 881 due to the subaerial context of the region and because the rifting is not complete
16
17 882 yet, the results on the crustal structure and the Vp/Vs ratios along our two profiles
18
19 883 help to discriminate between the possible mechanisms of extension involved
20
21 884 during the formation of the margin, over the specific period from the onset of the
22
23 885 rifting, i.e. ca. 26 Ma, to the time of the setting up the Stratoïd series, i.e. ca. 4-2 Ma.
24
25 886 Separated by only ~150.0 km with a major transfer zone (EATZ; Fig. 1), the two
26
27 887 sections along the western Afar margin show strong dissimilarities in their deep
28
29 888 structure, particularly regarding the geometry of the Moho depth across the
30
31 889 margin. The two sections differ primarily by the amount of the stretching and the
32
33 890 width of the extended area from the western plateau to the rift axis. In northern
34
35 891 Afar, the stretching factor β of the brittle crust and ductile lower crust is ~4.0 and
36
37 892 ~2.0, respectively, whereas in central Afar β is ~3.0 for the brittle crust and ~1.6
38
39 893 for the ductile lower crust. The sharp decrease in the crustal thickness between
40
41 894 the western plateau and the marginal graben during the early stages of extension
42
43 895 ($\beta \sim 2.0$ and 1.5 for the northern and central Afar, respectively) suggests different
44
45 896 rate of the mechanical extension that has been accommodated by large border
46
47 897 faults. During the latter stages of extension between the marginal graben and rift
48
49 898 axis, the differential thinning between the brittle and ductile crustal layers
50
51 899 suggests that the two layers were decoupled and associated with mechanical
52
53 900 extension in the brittle crust and ductile flow in the lower crust (e.g. Jolivet et al.,
54
55 901 2015). In northern Afar, the extension during these latter stages is accommodated
56
57 902 by the thinning of the upper crust only and distributed over a narrow extended
58
59 903 area (~40 km). This is consistent with the high Vp/Vs ratio, the thin seismogenic
60
61 904 layer (~5 km; Craig et al., 2011; Nobile et al., 2012) and the thin effective elastic
62
63 905 thickness ($T_e \sim 6$ km; Ebinger and Hayward, 1996; Pérez-Gussinyé et al., 2009;

1 906 Ebinger et al., 2017). The stretching and thinning across the western margin of
2
3 907 central Afar occurs on a wide distance (>100 km), accommodated by thinning of
4
5 908 both the upper and lower crust and associated with decrease in crustal thickness
6
7 909 by three successive steps from the stable continental western plateau to the active
8
9 910 rift axis. Whereas the Vp/Vs ratios do not show any correlation with the crustal
10
11 911 thickness and/or age, but well correlate to the mafic intrusions and increased melt
12
13 912 within the crust. These observations are consistent with a rifting history at the
14
15 913 latitude of central Afar dominated by a discontinuous riftward migration of the
16
17 914 extensional strain over time associated with intense magmatism (Wolfenden et al.,
18
19 915 2005; Stab et al., 2016).

20 916 Geological evidences across the margin of central Afar show that stretching and
21
22 917 thinning phases took place between extensive magmatic pulses, during the
23
24 918 Oligocene pre-rift period (main continental flood basalt; ~31-29 Ma), late
25
26 919 Oligocene-early Miocene (Miocene Basalts and Rhyolites; ~25-22 Ma), late
27
28 920 Miocene (Miocene Basalts and Rhyolites; ~9-7 Ma) and the Pliocene (Stratoid
29
30 921 series; ~5-2 Ma; Stab et al., 2016). Our results provide compositional and spatial
31
32 922 constraints with multiple crustal thinning steps consistent with the hypothesis of
33
34 923 episodic thinning and related magmatic cycles from the western plateau down to the
35
36 924 active central Afar magmatic segment.

37 925 The along-strike variations of the structure of the margin evidenced at depth by
38
39 926 our results but also at surface from several geological studies (Wolfenden et al.,
40
41 927 2005; Bastow and Keir, 2011; Stab et al., 2016; Le Gall et al., 2018) can be
42
43 928 interpreted as the result of a difference of stage of the continental breakup
44
45 929 evolution associated with a distinct conditions in which the continental extension
46
47 930 occurs, such as the local strain rate and/or the depth and the temperature of the
48
49 931 upper mantle which change the rheology of the crust and of the lithospheric
50
51 932 mantle. This different rheological behavior is marked by the major transfer zone
52
53 933 (EATZ) and on either sides of this transfer zone, we note the important variability
54
55 934 of the magma supply.
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57 935
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937 **7. Conclusion**

938 We use the receiver function technique and teleseismic data recorded by a
939 temporary broadband seismic array deployed along two profiles crossing the
940 western margin of Afar to constrain the tectonic processes that are involved in the
941 evolution of the rift. Our results provide constraints on the following
942 characteristics of the crustal thickness and composition:

943 (1) The crustal thickness beneath the western plateaus is 37-43 km, and decreases
944 eastward to its minimum ranges from 14-23 km beneath the central rift axis. The
945 decrease in the crustal thickness is more pronounced beneath the foothills and
946 marginal grabens, but the crust also thins elsewhere in discrete steps. This is
947 consistent with jumps in strain localization during rifting, rather than progressive
948 strain migration. The crustal thinning of northern Afar points towards a β of 2.7,
949 and in central Afar of 2.2. Both values are lower than a β of 3.0 predicted by plate
950 reconstructions, with the discrepancy best explained by 2-5 km of magmatic
951 addition to the crust. The amount of crustal thinning is higher in northern Afar,
952 pointing towards increased plate stretching in the north.

953 (2) The V_p/V_s ratios suggest that the crustal composition along the western
954 margin of Afar could be classified with unmodified crust with felsic to normal
955 crustal composition, (western plateau next to northern Afar), altered crust due to
956 magmatic intrusion and/or additions (western plateau and marginal area of
957 central Afar) and highly modified crustal composition associated with the
958 presence of partial melt at the central rift axis.

959 (3) There is a significant difference between the western margin characteristics of
960 the northern Afar (Danakil Depression) in comparison with the western margin of
961 the central Afar. In northern Afar, the crustal extension occurs over a short distant
962 associated with significant thinning of the upper crust and dominant tectonic
963 deformation. In central Afar, the crustal extension occurs on a wide area and
964 characterized by episodic crustal thinning associated with intense magmatic
965 activity.

1 966 (4) The elevated V_p/V_s ratios are generally associated with the crustal
2
3 967 thicknesses of less than 21 km, in or near the active magmatic segments showing
4
5 968 the role of the magmatic processes in the final stages of continental rifting.
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7 969

8 970 ***Acknowledgments***

9
10 971 Data used in this study were collected by the AFAR17 project in Ethiopia. The
11
12 972 instruments for the project were provided by the French National Research
13
14 973 Infrastructure SISMOB-RESIF (<http://www.resif.fr>) and SEIS-UK. RESIF is supported
15
16 974 by the Ministry of Higher Education and Research. The facilities of SEIS-UK are
17
18 975 supported by the Natural Environment Research Council (NERC) under agreement
19
20 976 R8/H10/64. Funding for fieldwork (deployments, data collection and
21
22 977 maintenance) is from Sorbonne Université, ISTEP, Institute des Sciences de la
23
24 978 Terre de Paris. AA is supported by the University of Sorbonne (iSTeP), the
25
26 979 University of Strasbourg (FACE program) and the PAUSE program. DK is supported
27
28 980 by NERC grant NE/L013932/1, and MIUR grant PRIN 2017P9AT72. We wish to
29
30 981 thank the Addis-Abeba University staff for help in customs clearance and for the
31
32 982 technical facilities, the Afar Regional Authorities and local people for their kind
33
34 983 cooperation and their help to make the fieldwork successful. We deeply thank,
35
36 984 Alex Nercessian, IPGP, the Centre Français d'Etudes Ethiopiennes (CFEE) in Addis-
37
38 985 Ababa for the help and support provided by the center during the experiment.
39

40 986

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Tables and Figures

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3 1530 **Table1.** Crustal thickness (H), Vp/Vs ratio for the seismic stations and associated
4 1531 uncertainty estimates deduced from the bootstrap analysis.

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6 1533 **Figure 1.** Topographical map of the study area showing the main tectonic
7 1534 structures and the location of the seismic stations (red triangles) along two
8 1535 profiles from the western plateau to the central rift axis. MHRS: Manda Harraro
9 1536 Rift Segment. Inset at the upper right corner is a structural map of the Afar
10 1537 Depression and surrounding highlands modified from Stab et al. (2016) and Le
11 1538 Gall et al. (2018), with a regional map show the location of the study area.

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13 1540 **Figure 2.** Stacked radial (dark/grey fill colour) and transverse (red/blue fill
14 1541 colour) receiver functions for the 23 seismic stations along two profiles. a) the
15 1542 southern profile at the latitude of the central Afar, and b) the northern profile -
16 1543 Danakil Depression, c) sketch illustrating the path of the converted phases and d)
17 1544 map showing the distribution of the ~126 teleseismic events selected and used for
18 1545 receiver functions analysis in this study centered at the network location.

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20 1547 **Figure 3.** a) Individual receiver functions for six stations. The receiver functions
21 1548 are organized by increasing backazimuth (value in red). The light vertical lines
22 1549 indicate arrival times for conversion phases (Pms and multiples) from the Moho
23 1550 for the maximum stacking amplitude; ; b) Ps radial RFs at the N006 seismic
24 1551 station for good quality data within an epicentral range 30°-95°, stacked in
25 1552 backazimuth into bins of 15° and with an overlap of 5°. The numbers of RFs in
26 1553 each stack are indicated above each trace to the left and the backazimuth ranges
27 1554 to the right; c) Ps transverse RFs at the N006 seismic station stacked with the
28 1555 same epicentral distance, backazimuth ranges and number of RFs in each stack
29 1556 as in b; d) Stacked Ps radial RFs with a maximum frequency of 0.6 Hz (red line)
30 1557 and 1.2 Hz (blue line) at the S003 seismic station located at the eastern edge of

1 1558 the western plateau southern profile (central Afar). The stacked transverse RFs
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3 1559 with a maximum frequency of 1.2 Hz is represented by a dashed black line.
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7 1561 **Figure 4.** Thickness (H) versus Vp/Vs ratio diagrams from the H-k stacking
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9 1562 method for the 6 stations presented in Figure 4. The white point indicates the
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11 1563 maximum stacking corresponds to the value indicated in Table 1. The scale bar is
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13 1564 the amplitude of stacking function.
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16 1566 **Figure 5.** Maps of crustal thicknesses (H) and Vp/Vs ratios calculated using the H-
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18 1567 k method; a) Topographical map with the crustal thickness variations across
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20 1568 western Afar margin and b) Topographical map with the Vp/Vs ratios across the
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22 1569 western Afar margin. MHRs: Manda Harraro Rift Segment.
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26 1571 **Figure 6.** Topography (vertically exaggerated) and migrated cross-sections along
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28 1572 the two profiles. Red colour indicates velocity increase with depth, and blue colour
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30 1573 velocity decrease with depth. Scale bar shows the amplitude of positive (red) and
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32 1574 negative (blue) polarities of arrivals. The Moho depth estimated from H-k stacking
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34 1575 method and corrected for the station altitude is plotted with the small circles and
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36 1576 the vertical bars represent the errors estimated from the bootstrap method. a)
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38 1577 Northern profile – Northern Afar and b) Southern profile – central Afar.
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41 1579 **Figure 7.** Regional map showing crustal thickness (H) based on the results of the
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43 1580 current study (white circles) and previous RF studies in Ethiopia (red triangles,
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45 1581 Hammond et al. 2011; inverted white triangles, Stuart et al., 2006; magenta
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47 1582 pentagon, Dugda et al., 2005), RF in Yemen (white stars, Ahmed et al. 2013). Active
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49 1583 seismic profiles (+ sign), in Afar (Berckhemer et al., 1975; Makris and Ginzburg,
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51 1584 1987), in the Red Sea (Egloff et al., 1991), in the Western Gulf of Aden (Laughton
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53 1585 and Tramontini, 1969) and in Djibouti (Ruegg, 1975).
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56 1587 **Figure 8.** Regional map showing Vp/Vs ratio based on the results of the current
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58 1588 study (white circles) and previous RF studies in Ethiopia (red triangles, Hammond
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1589 et al. 2011; inverted white triangles, Stuart et al., 2006; magenta pentagon, Dugda
 1590 et al., 2005), RF in Yemen (white stars, Ahmed et al. 2013).

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1592 **Table1.**

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Station	Latitude	Longitude	Elevation (m)	H (km)	Error H (km)	Vp/Vs	Error Vp/Vs	Number of RF
N001	39.3077	13.5545	2148	37.6	2.3	1.73	0.04	46
N002	39.4162	13.5453	2000	38.8	1.9	1.73	0.02	23
N003	39.5007	13.5580	2080	37.6	0.5	1.79	0.02	50
N004	39.5653	13.5775	2265	37.6	0.3	1.73	0.01	67
N005	39.6677	13.6334	2270	38.6	1.6	1.73	0.04	36
N006	39.7805	13.6509	2383	37.0	7.0	1.70	0.10	52
N007	39.8938	13.6922	970	24.2	1.5	1.79	0.05	31
N008	40.0066	13.8406	692	19.4	1.3	1.93	0.02	25
S001	38.9328	11.6591	2904	42.8	1.7	1.79	0.01	27
S003	39.3211	11.8511	3446	41.0	2.0	1.87	0.02	65
S004	39.4903	11.8274	2372	37.2	4.0	1.83	0.06	15
S005	39.5291	11.8387	1931	34.6	1.8	1.75	0.01	56
S006	39.5874	11.8472	2057	34.0	0.4	1.76	0.02	60
S007	39.6583	11.8478	1601	34.0	1.2	1.72	0.03	58
S008	39.7348	11.8621	1491	27.6	1.3	1.85	0.03	52
S009	39.7995	11.8001	1486	28.2	0.5	1.80	0.02	59
S011	39.9164	11.8203	1075	27.6	1.5	1.79	0.01	58
S012	39.9581	11.8717	1007	27.8	1.7	1.79	0.03	20
S013	40.0506	11.8574	914	28.0	2.0	1.79	0.03	27
S014	40.1831	11.8584	827	20.8	1.6	1.95	0.05	38
FINE	40.3100	12.0600	782	21.4	1.7	1.99	0.02	72
S017	40.7044	11.9856	613	23.2	1.5	1.96	0.02	60
S018	40.8044	11.8521	441	21.6	1.0	1.92	0.01	21
S020	40.9667	12.0019	355	19.8	0.6	2.02	0.01	65

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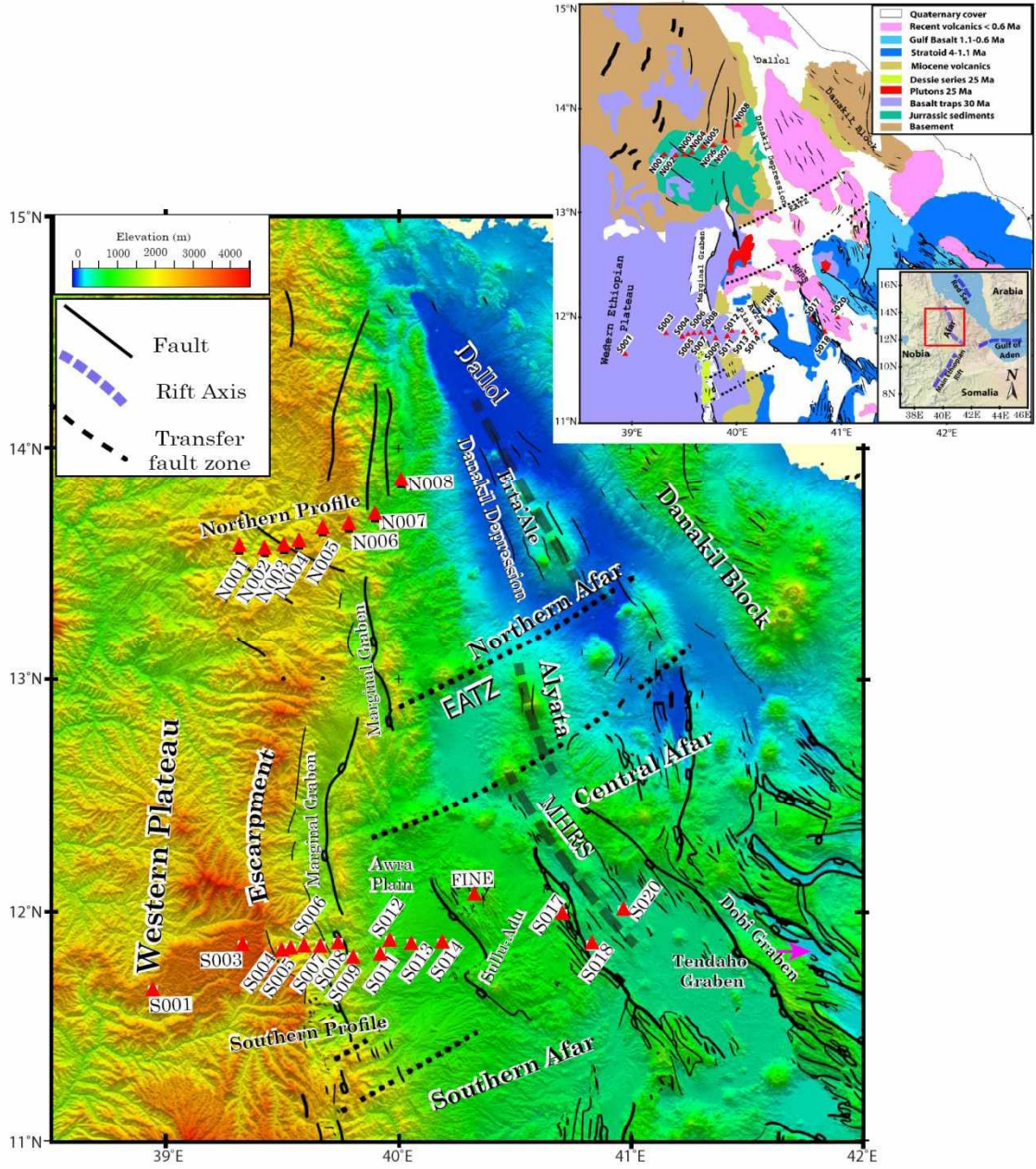
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Figure 1.



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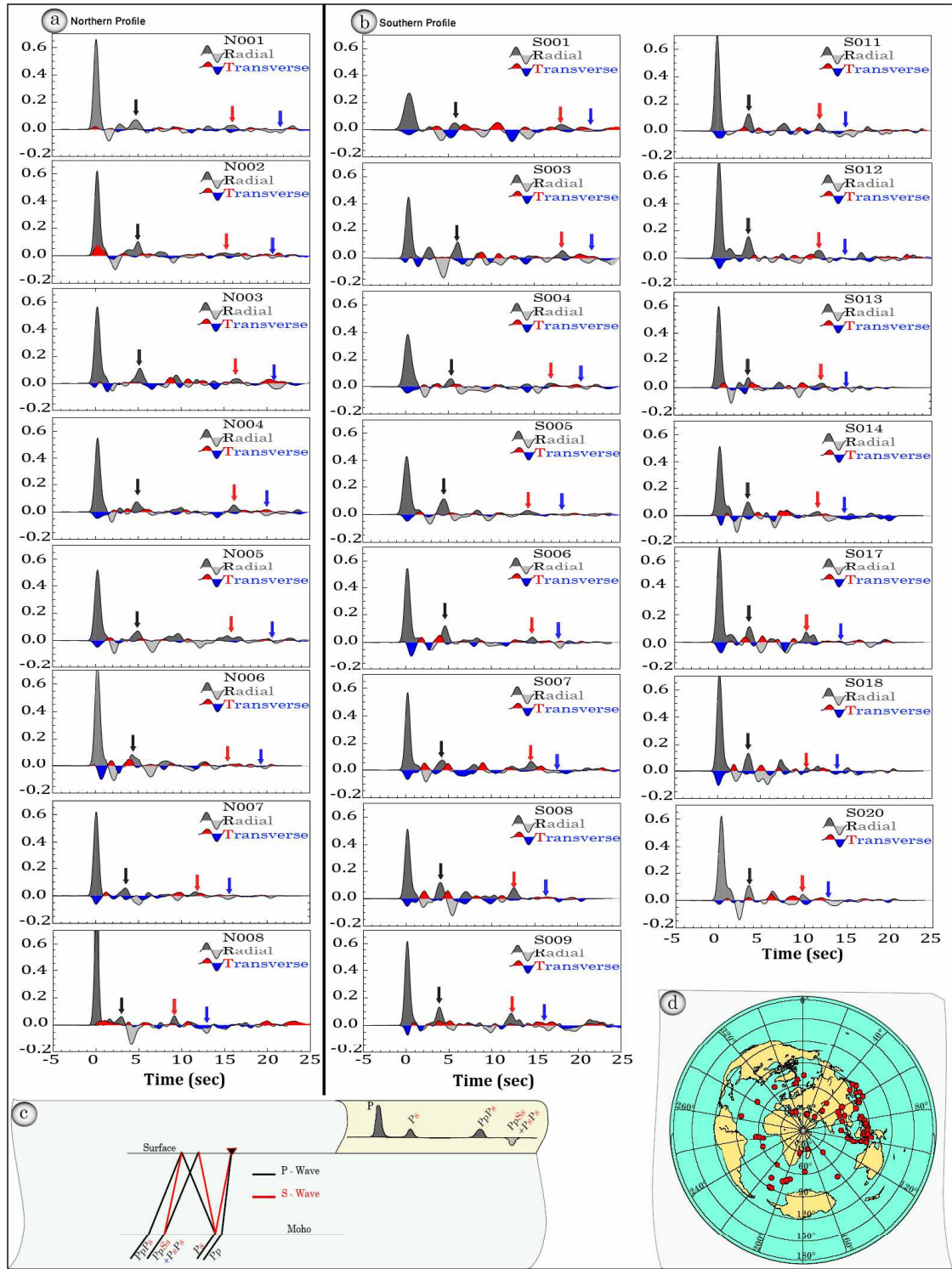
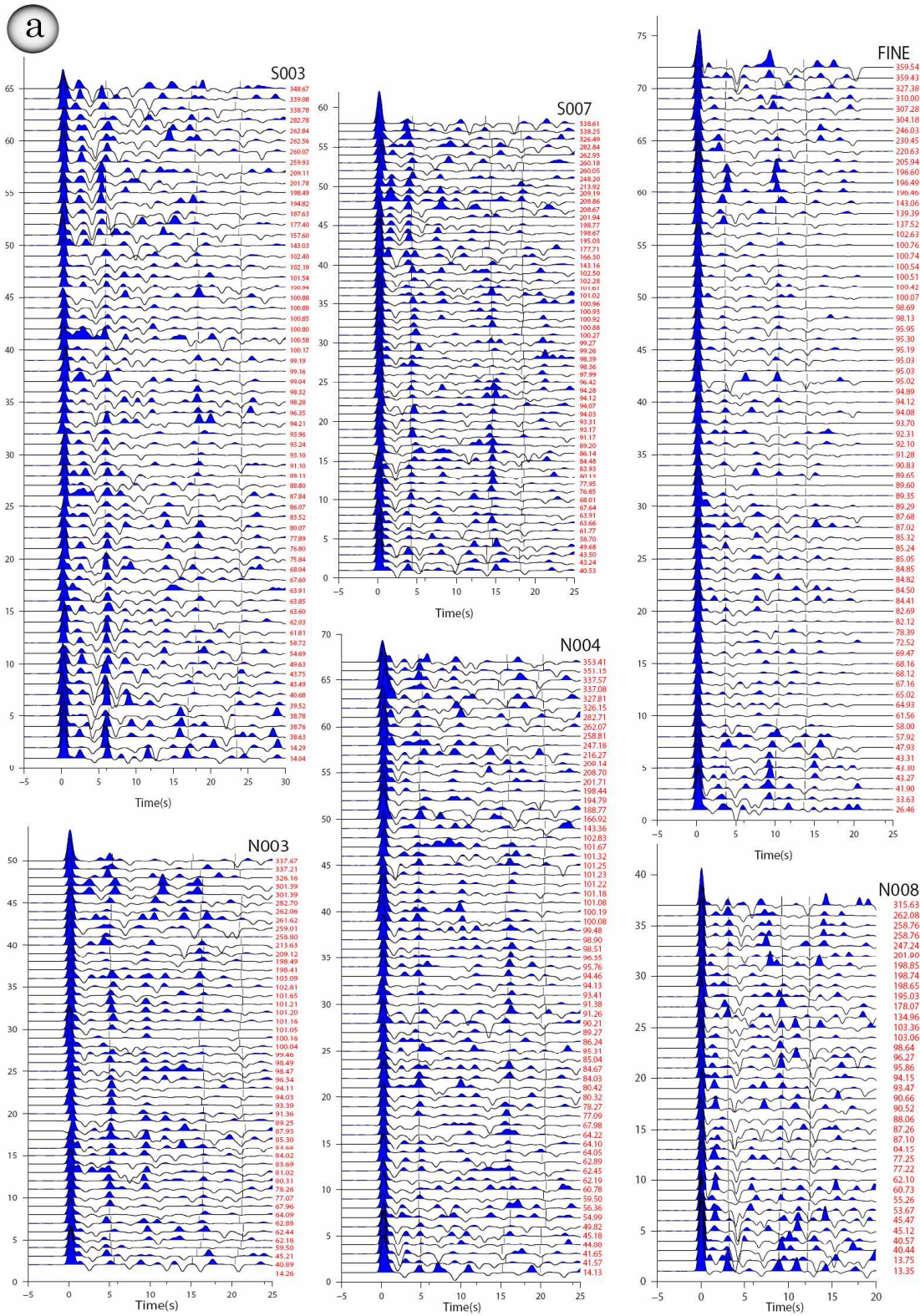


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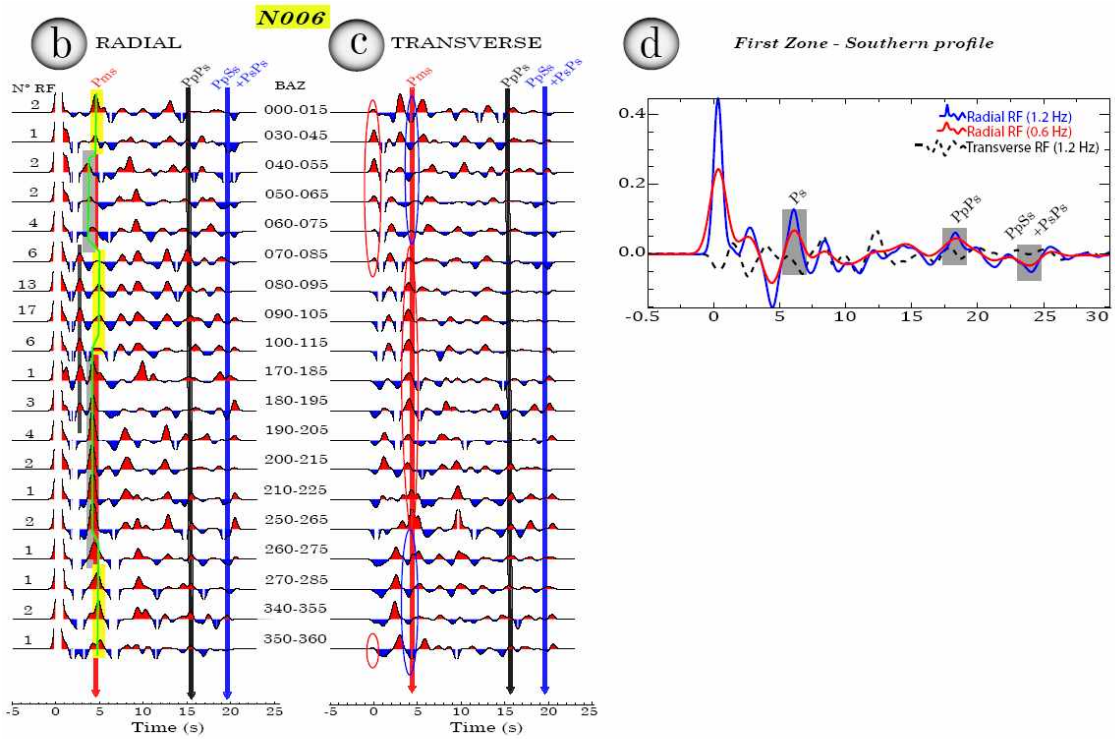


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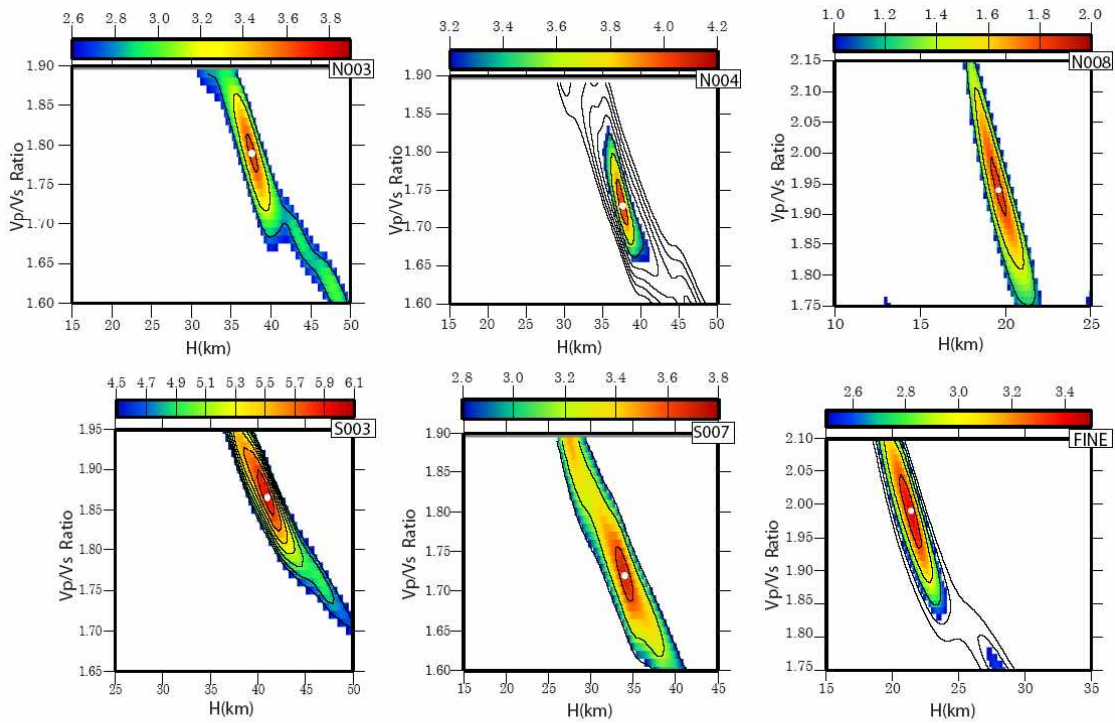
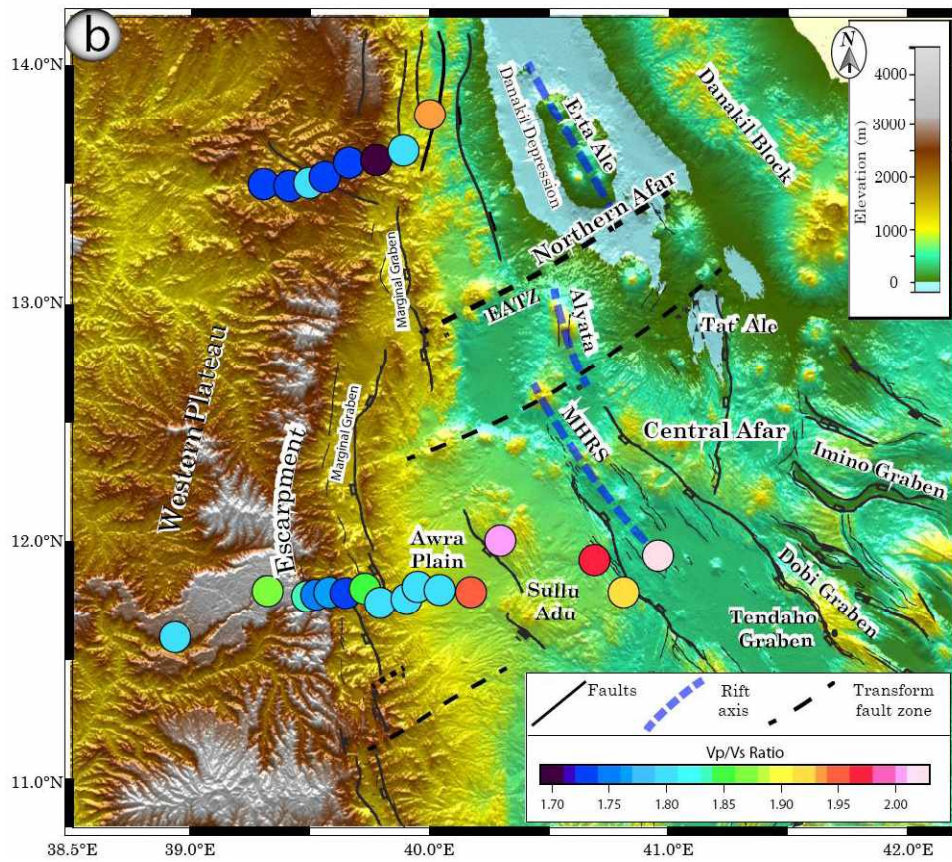
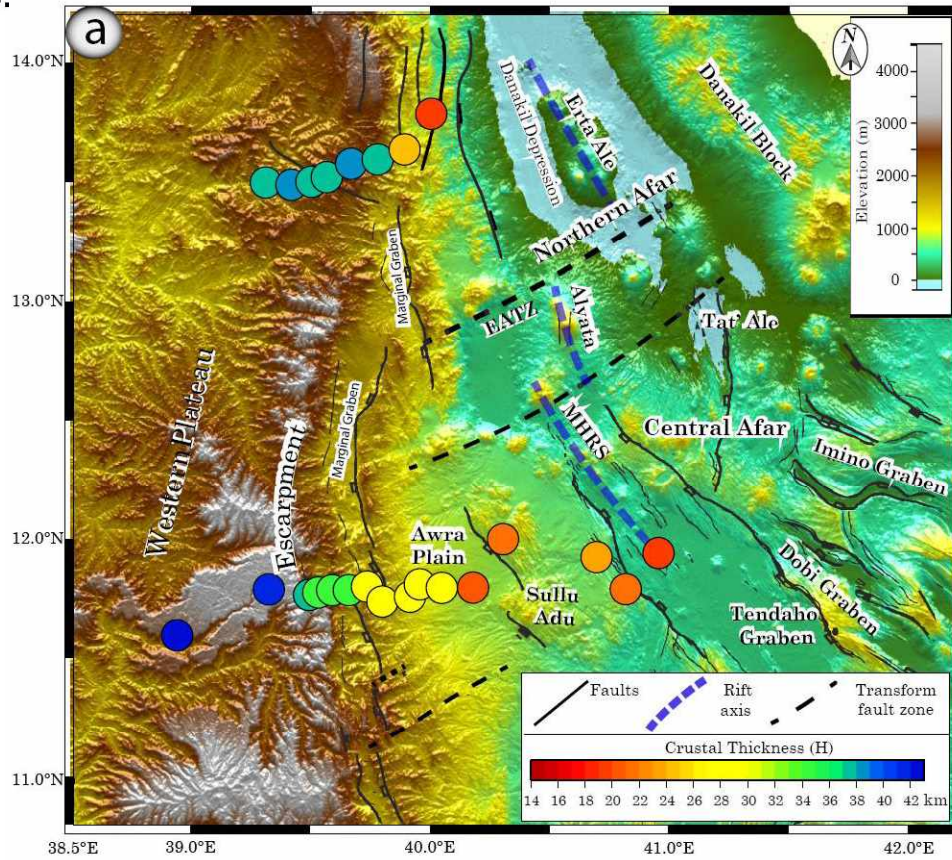
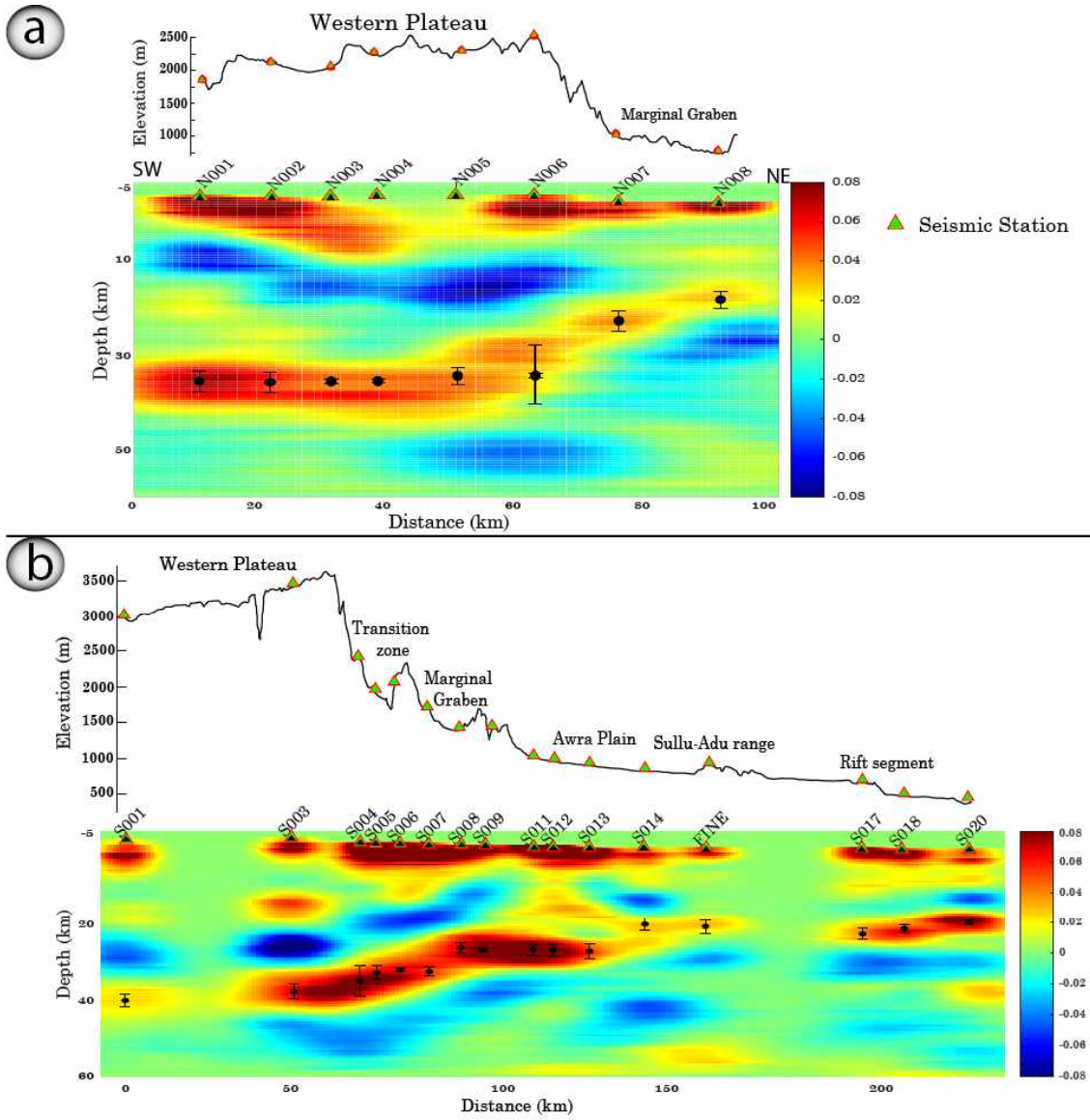


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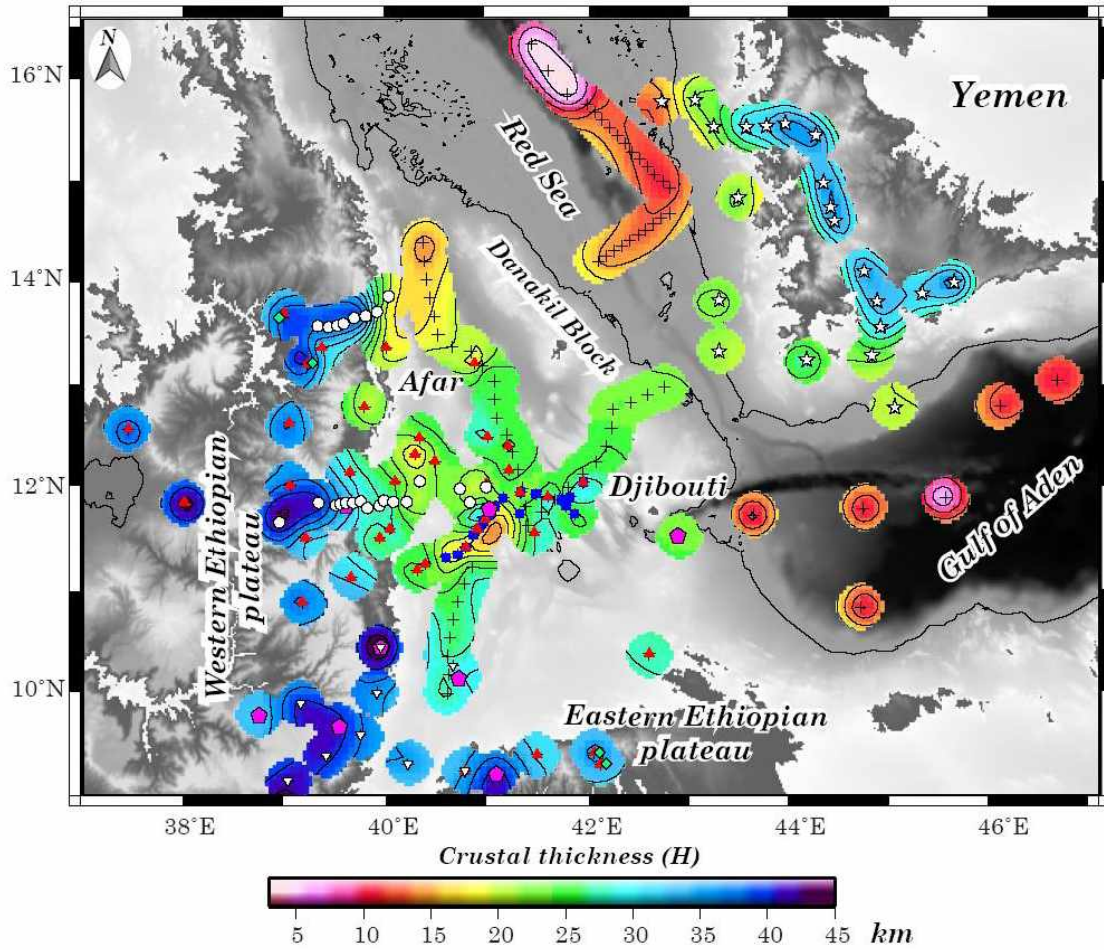
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Figure 6.



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Figure 8.

