1 Upper mantle structure of the northeastern Arabian Platform

2 from teleseismic body-wave tomography

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10

11Abstract

12 A thick Infra-Cambrian to Phanerozoic sedimentary sequence in eastern Arabia
13conceals a poorly known Neoproterozoic basement that once separated the currently
14exposed Arabian Shield (western Arabia) from NW India in Gondwana. In this study, the
15Arabian Platform (eastern Arabia) is targeted by relatively high-resolution passive-source
16seismic tomography for the first time, which aims to illuminate crust and upper mantle
17structure in order to make inferences on their nature and origin. This is facilitated by a
18new broadband seismic network of 55 stations in the United Arab Emirates (UAE), which
19has recorded sufficient teleseismic body wave data to allow for the determination of

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three-dimensional seismic structure. Using a grid-based eikonal solver and a subspace inversion technique implemented in FMTOMO, relative arrival-time residuals from teleseismic earthquakes are mapped as 3-D P-wave velocity perturbations in the upper mantle beneath the seismic network. The resultant tomographic images reveal a marked transition from higher velocities in the Gulf of Oman to lower velocities in the west that we interpret to indicate the lithospheric boundary between Cretaceous Tethyan oceanic lithosphere and the Arabian passive continental margin. The northern extent of Tethyan oceanic lithosphere appears to terminate against a possible offshore continuation of the Dibba fault, until now only identified onshore. Further inland, the tomographic model exhibits a number of low- and high- velocity anomalies with a predominant NW orientation. Significantly, one well-defined high velocity anomaly is situated along strike of the Neoproterozoic exposure of volcanic-arc granodiorites found in northern Oman. We relate this anomaly to an intra-oceanic arc created during the Tonian subduction of the Mozambique Ocean, which strongly support the idea of continental growth in the Arabian Platform via magmatic arc accretion.

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Key words: Teleseismic body-wave tomography, Continental Accretion, Oceanic Lithosphere, Semail ophiolite, Arabian Platform

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411 Introduction

42 The lithosphere beneath the Arabian Peninsula is typically divided into western and
43 eastern tectonic domains by a major magnetic feature (located in the proximity of the
44 Amar Arc - Figure 1b), but our current understanding is strongly biased towards the
45 western domain (Arabian Shield). Extensive basement exposures, geochemical analyses
46 and geophysical surveying over the Arabian Shield have greatly helped constrain its
47 origin and structure. In stark contrast, the composition, age and structure of the Arabian
48 Platform (eastern domain) are poorly known since it is concealed beneath Infra-Cambrian
49 to Phanerozoic volcano-sedimentary formations (> 10 km thick). This has significantly
50 impeded the ability of conventional surface mapping to make robust inferences into the
51 geotectonic history of the eastern segment. The only direct observations of the basement
52 in the eastern domain come from a few scattered Neoproterozoic inliers in southeast
53 Oman, in addition to isolated volcanic and exotic sedimentary clasts embedded in
54 intruded salt domes in the Arabian Gulf, which indicate late Ediacaran ages (560–545
55 Ma) (Thomas et al., 2015). However, no crystalline basement rocks of Cryogenian age
56 are recorded from the UAE salt domes (Thomas et al. 2015). Stewart (2016) reported that
57 two deep exploration wells, in Shaybah field and Lekhwair High, in Rub Al Khali, Saudi
58 Arabia, possibly penetrated the crystalline basement.

59 Even though the Arabian Shield and the Arabian Platform were both part of the
60 western flank of Gondwana, several lines of evidence indicate that their lithosphere is
61 fundamentally different (see Stern and Johnson, 2010 for a comprehensive review on the

62architecture of the Arabian Plate). Indeed, despite affinities in age and tectonic genesis, a
63simple Neoproterozoic tectonic link between the Arabian Shield and the Omani basement
64has so far achieved little consensus (Gas et al., 1990; Mercolli et al., 2006; Allen, 2007;
65Rantakokko et al., 2014; Whitehouse et al., 2016; Alessio et al., 2018; Blades et al.,
662020). As such, a complete understanding of the Arabian Plate's lithosphere cannot be
67achieved without gaining further insights into the crust and mantle of its eastern segment.

68 More than half a century of hydrocarbon exploration in the Gulf States, situated in
69the Arabian Platform, has yielded a rich multidisciplinary dataset that is targeted at
70constraining primarily the upper and mid sedimentary sections, with little information on
71the underlying basement. The United Arab Emirates (UAE)–Oman mountain belt, which
72contains the spectacularly exposed Semail ophiolite, has similarly received considerable
73attention since the first field study by Glennie et al. (1973). However, the basement
74record of the UAE's interior is entirely covered by a blanket of Quaternary sand dunes
75and up to 10 km of Phanerozoic sediments that mask any possible basement exposure.
76The thick sedimentary package has also prevented deep wells from intersecting the
77basement. The deepest well in the UAE, located in the southernmost part of the country,
78reached sedimentary rocks of Silurian age at a depth of 6609 m (Bloch, 2000). In order to
79decipher the nature of the tectonic structure beneath the sedimentary cover and
80submerged areas of the Arabian Platform, crucial inferences can be made by exploiting
81geophysical datasets. Interpretation of publicly available active seismic profiles is
82challenging on structures below the Permian formations (up to about 5 km depth),

83whereas potential field imaging and modelling (e.g., Ali et al., 2017) has been used to
84determine a depth-to-basement model and structural trends from the basement. While
85these methods offer constraints chiefly on the upper and mid crust, seismic tomography
86has the potential to better illuminate the deeper lithospheric structure.

87 Previous regional tomographic studies (e.g., Mooney et al., 1985; Şeber and
88Mitchell, 1992; Hansen et al., 2007; Wei et al., 2019; Kaviani et al., 2020) allowed first-
89order constraints to be placed on the Arabian basement architecture, but a consistent lack
90of data coverage in the UAE and Oman remains a large obstacle for the completion of a
91coherent lithospheric model at constant high resolution. The former Petroleum Institute
92(now Khalifa University of Science and Technology) deployed a new network of seismic
93stations in the UAE in an attempt to elucidate the crust and upper-mantle structure
94beneath the country and areas of Oman.

95 In this study, we report on new results from the inversion of teleseismic P-wave
96relative arrival time residuals that are mapped as 3-D P-wave perturbations. We discuss
97our findings in terms of lithospheric structure and attempt to reconcile the velocity
98anomalies with the known tectonic evolution of the study area.

99

1002 Tectonic Setting

101 The bulk of the Arabian basement includes rocks formed in the Neoproterozoic,
102although remnants of Paleoproterozoic to Neoarchean rocks are exposed in Saudi Arabia
103and Yemen. The Arabian Shield is characterized by the amalgamation of island-arcs,

104 associated ophiolitic remnants, microcontinental terranes and consequent suturing (see
105 Asir, Hijaz, Midyan, Afif in Figure 1b), successively accreted westwards onto the
106 Arabian–Nubian shield during a major compressional event from ca. 715 to 630 Ma
107 (Stoeser and Camp, 1985). Analogue rocks of the Arabian Shield have not been recorded
108 in the Omani Neoproterozoic inliers (e.g., Saih Hatat, Jebel Akhdar, Jebel Ja’alan in
109 Figure 1c). The basement of Al Jobah (Huqf supergroup in Oman) was formed by acid
110 volcanic rocks and granodiorite, which were dated at ~825 Ma (Bowring et al., 2007),
111 while the main unconformity between igneous basement and overlying sediments was
112 formed around 725 Ma (Allen, 2007). Timing of the eastern basement (Rayn microplate)
113 collision and suturing with the Afif Terrane is still debated, although it occurred along the
114 Amar suture. This marks the collision zone between the two blocks and formed N-
115 trending basement-controlled structures in the northern Arabian Platform (Al-Husseini,
116 2000). Based on detrital zircon ages of samples from the UAE, which show source
117 affinity to the western and eastern basements of Arabia, Thomas et al. (2015) claim that
118 collision and suturing of the two domains should have been accomplished by about 597
119 Ma. Major anticlines in the eastern Arabian Platform (e.g., Qatar Arch), created during
120 assembly of the western and eastern segments, are bounded to the south by the possibly
121 syn-collisional NW-trending Abu Jifan strike-slip fault (Al-Husseini, 2000). Further
122 south, a major and possibly younger sinistral fault system dislocated the Nabitah Suture
123 by ~300 km. This large displacement occurred along the NW-trending Infra-Cambrian
124 Najd shear system (570-530 Ma), which exhibits a width of approximately 300 km, and is

125exposed in the Arabian Shield for more than one thousand km (Al-Husseini, 2000).
126Although buried further east beneath Oman, potential field imaging suggests that it may
127extend beneath Rub Al Khali desert and southwestern Oman (Loosveld et al., 1996).
128 From the Neoproterozoic to Late Devonian periods (ca 610 to 364 Ma) the Arabian
129plate was situated in the continental interior of Gondwana, during which a number of
130intracratonic sag basins developed. Subsequently, from Late Carboniferous to mid-
131Permian, the Arabian Plate was located in a back-arc setting (Ruban et al., 2007). From
132mid-Permian until the Cenomanian, the northeastern margin of the Arabian plate was a
133rifted margin of the Tethys Ocean and experienced stable shelf carbonate deposition,
134outcrops of which can be observed in the northernmost Musandam Peninsula (Figure 1c).
135Stable sedimentary conditions on the Arabian passive continental margin endured for
136nearly 160 Ma and ended abruptly when approximately 15 km of allochthonous oceanic
137crust and upper-mantle thrust sheets (Semail ophiolite in the UAE-Oman Mountains)
138were progressively emplaced to the southwest onto the Arabian continental margin
139(Glennie et al., 1973; Searle, 1988, Searle, 2007). The Musandam Peninsula currently
140marks the transition from ocean-continent convergence in the Gulf of Oman to
141continental collision along the Zagros fold-and-thrust belt (Searle et al., 1983; Searle,
1421988). The earliest effects of continental collision in the Musandam Peninsula are seen in
143the Oligocene- early Miocene thrusting along the Hagab thrust, which records initial
144crustal thickening (Searle 1988).

145

1463 Data and Methods

147 Teleseismic data used in this study are sourced from 23 permanent stations
148 maintained by the National Center of Meteorology (NCM) of the UAE and from 32
149 temporary stations (Figure 1c) operated by the former Petroleum Institute (Pilia et al.,
150 2020). We select a total of 1687 teleseismic events (ISC catalogue) in a time window
151 spanning June 2014 to December 2016. The selection criteria were based on magnitude
152 and epicentral distance, with earthquakes of magnitude $M_b \geq 5.0$ located in the distance
153 range of $30^\circ \leq \Delta \leq 95^\circ$ retained. We predominantly exploit the direct P-phase, but to
154 improve azimuthal coverage, PcP phases from two events and PP phases from 17 events
155 are also included in the dataset. A bandpass filter with corner frequencies of 0.07 Hz and
156 5.0 Hz is applied to the vertical component data. Subsequently, all traces are aligned
157 using moveout corrections based on the ak135 reference model (Kennett et al., 1995) and
158 are cut 50 s before and after the predicted P-wave arrival time. We calculate the signal-to-
159 noise ratio (SNR) by using a 20 s record length before and after the P-wave onset, which
160 represent the noise and signal windows, respectively. The SNR is then calculated by
161 dividing the signal standard deviation by the noise standard deviation through the
162 following equation:

$$163 \left| R_{sn} = \sqrt{\frac{1}{N_s} \times \sum_{i=1}^{N_s} (s_i - \bar{s})^2} \right. / \left. \sqrt{\frac{1}{N_n} \times \sum_{i=1}^{N_n} (n_i - \bar{n})^2}, \quad (1) \right.$$

164 where $\overline{R_{sn}}$ is the SNR; $\overline{s}$ represents the signal and $\overline{m}$ the noise; $\overline{s}$ and $\overline{m}$ are the
 165 average values of signal and noise, respectively; $\overline{N_s}$ and $\overline{N_n}$ represent the total
 166 number of sampling points of the signal and noise, respectively. If the SNR is less than 1,
 167 which means the signal does not stand out from the noise, the seismogram is removed
 168 from our dataset. To ensure that only high-quality records are included in our
 169 calculations, we also visibly inspect the remaining traces. The relative arrival-time
 170 residuals are computed by using the adaptive stacking approach of Rawlinson and
 171 Kennett (2004a), which is a powerful and rapid procedure for estimating the residual
 172 patterns across a network of seismic stations (see an example in Figure 2). In order to
 173 avoid cycle skipping, the time-shift delay for alignment was limited to ± 2 s. Adaptive
 174 stacking involves stacking all traces, applying a time shift to each trace to optimally align
 175 it with the stacked trace, and then repeating the process until convergence (Figure 2).
 176 Following these procedures, a collection of 425 events with 9823 relative P-wave arrival-
 177 time residuals make up our dataset.

178

179 To enhance the efficiency of the forward calculation and reduce the influence of a
 180 heterogeneous source distribution that results in smearing along paths originating from
 181 regions of high-source concentration, we apply data binning (see details in Rawlinson et
 182 al., 2011). The relative arrival-time residual of each stacked source uses the average value
 183 of all sources in the same bin, as do the coordinates of the composite stacked source

184(Figure S1). If the bin size is too large, significant differences can occur in the relative
185arrival-time residuals of the sources in the cluster. In order to find the best bin size, we
186test across a horizontal range of 1° – 3° with a step of 0.5° and vertical length range of 50–
187300 km with a step of 50 km. We found that bins of $2.5^{\circ} \times 2.5^{\circ} \times 100$ km offered the best
188compromise (i.e., the relative arrival-time residuals recorded by each station are very
189close), with a meaningful reduction in the effective number of sources, from 425 to 207.
190Consequently, a total of 5526 relative arrival time residuals are used in the inversion.

191 In order to map the relative arrival-time residuals as 3-D P-wave perturbations, we
192use the multi-stage Fast Marching Tomography Method (FMTOMO) package
193(Rawlinson et al., 2006a; Rawlinson and Urvoy, 2006). FMTOMO solves the travel time
194prediction problem by employing a 3-D grid-based eikonal solver (Sethian, 1999;
195Rawlinson and Sambridge, 2004b). The velocity model is parameterized by a 3-D grid of
196nodes with cubic B-splines in spherical coordinates (Rawlinson et al., 2006b) used to
197produce a continuous, smooth, and locally controlled velocity medium. Layer interfaces
198of potentially variable geometry are parameterized by using a two-dimensional regular
199grid in latitude and longitude, with a smooth surface achieved by using cubic B-splines
200(Rawlinson et al., 2006a). A subspace inversion scheme subject to damping and
201smoothing regularization is implemented for the inversion step (Kennett et al., 1988),
202which enables non-linear relationships to be reconciled between the velocity field,
203interface depth, source location, and travel time perturbations. In this study, only velocity
204perturbations are inverted since teleseismic arrival time data cannot also constrain source

205 location or interface depth. To assess the appropriate trade-off among data variance,
206 model variance, and model roughness, we determine the damping and smoothing factors
207 required for inversion using so-called L-curves (see for example Rawlinson et al., 2006c).
208 A value equal to 4000 for the damping factor represents a sensible compromise between
209 minimizing model variance and optimizing data fit. The value that provides the best
210 trade-off between producing the smoothest model and data fit (smoothing factor) is 400
211 (see Figure S2 in the supplementary information).

212 The study area is discretized into two layers with three interfaces, including the free
213 surface, the Moho discontinuity and the bottom of the velocity model. The spatial
214 distribution of the velocity nodes for these two layers is identical, comprising 41 nodes in
215 latitude (19.5° to 29.5°N), 49 nodes in longitude (48.5° to 60.5°E), and 21 nodes in depth
216 (+4 km to -551 km). This yields a model space discretized into a total of 42189 nodes,
217 evenly distributed with a spacing of $0.25^\circ \text{ km} \times 0.25^\circ \text{ km} \times 27.75 \text{ km}$ in latitude,
218 longitude and depth, respectively. Similarly, the three interfaces incorporated in the
219 velocity model have the same number of grid nodes in latitude and longitude.

220 The presence of strong lateral velocity heterogeneities in the crust, largely due to the
221 contrast between obducted oceanic material and large, deep sedimentary basins in the
222 study region, can strongly affect the pattern of arrival-time residuals. Thus, for crustal
223 corrections of the teleseismic traveltime residuals we include as prior information in the
224 inversion the crustal model of Pilia et al. (2020), which is capable of resolving structures
225 as small as 50 km horizontally and a few kms vertically using ambient noise data (see a

depth slice in Figure 3). The dataset is described in Pilia et al. (2020), which make use of the same seismic network as the teleseismic dataset, with the inclusion of further stations from Iran. Phase velocities are measured from the empirical Green's function across a range of periods (2–40 s) as per in Pilia et al. (2016). Two novel transdimensional, hierarchical, Bayesian techniques are subsequently applied to obtain period dependent phase velocity maps in the first instance, followed by 1-D inversion for shear velocity profiles that are finally assembled in a 3-D shear-velocity model. While the teleseismic arrival-time residuals are generated from global P-phases, the ambient noise crustal model provides shear-wave velocities. Thus, the latter are converted into P-wave velocities according to the formulation of Brocher (2005), whereas the upper mantle velocity are defined by the ak135 global reference model (Kennett et al., 1995). Station elevations are included in the forward calculation to account for differences in arrival time due to topographic variations.

239 New constraints on crustal thickness have also been incorporated in FMTOMO
240 during the inversion. Moho depths are inferred from the combination of two receiver
241 function profiles and gravity modelling across the UAE-Oman mountain range (see Ali
242 et al., 2020). Although we do not have detailed Moho depths in the stable Arabian
243 Platform, peak variations are expected to be across the mountain belt. Crustal thickness
244 values found at both extremities of the profiles have therefore been extended in the Gulf
245 of Oman and Arabian Gulf (Figure 3). Similarly, Moho depths across the mountain belt

246are assumed to be similar and are extended to the north and south following the trend of
247the Semail thrust. The base of the 3-D model lies at a depth of 550.5 km.

248

2494 Results

2504.1 Resolution Tests

251 In order to evaluate the resolving power of our dataset and the quality of our
252inversion results, we carried out detailed resolution tests based on synthetic data,
253following a number of recommendations made by Rawlinson and Spakman (2016). This
254involves using multiple checkerboards with different scale length, a sparse distribution of
255spikes, synthetic structures that resemble those recovered in the final solution model (but
256are sufficiently different to avoid the issue of preconditioning), and using identical
257sources, receivers and phase types to the observational dataset. For any given 3-D
258synthetic structure, relative arrival-time residuals are predicted and inverted for P-wave
259velocity structure using the same regularisation parameters as the inversion of observed
260data. Ultimately, the difference between input synthetic structure and the recovered
261pattern of anomalies provides insight into which areas of the tomographic model are well
262constrained and which are not. All synthetic travel times are contaminated with Gaussian
263noise with a standard deviation of 100 ms in order to simulate the picking uncertainty
264associated with the field dataset.

265 We have conducted three checkerboard tests with alternating positive and negative
266anomalies (± 0.4 km/s velocity perturbation) of different size: i) a coarsely spaced

267 checkerboard with anomaly size $2^\circ \times 2^\circ \times 222$ km (Figure S3), ii) an intermediate
268 checkerboard with medium anomaly size $1.5^\circ \times 1.5^\circ \times 166$ km (Figure S4), and iii) a
269 finely spaced checkerboard with anomaly size $1^\circ \times 1^\circ \times 111$ km (Figure 4). Due to the
270 high concentration of earthquake sources from the east, vertical sections indicate that
271 there is a tendency to vertically smear the velocity anomalies in this direction (cross
272 sections in Figures 4, S3 and S4), a feature typical of teleseismic tomography studies
273 where azimuthal ray coverage is unevenly distributed. Nonetheless, our results
274 demonstrate that the path coverage is sufficient to accurately reconstruct the coarse and
275 intermediate checkerboards from the surface to about 350 km depth (Figures S3 and S4).
276 As expected, recovery of the fine checkerboard is less apparent at depths greater than 150
277 km (Figure 4).

278 When anomalies are smeared along predominant raypath trajectories, anomalies
279 with the same sign tend to merge with the diagonally adjacent cell of the checkerboard. In
280 this context, spike tests can be more useful for understanding directionally dependent
281 smearing (Rawlinson and Spakman, 2016). In Figure 5a, we show three synthetic vertical
282 spikes with 0.4 km/s maximum positive or negative velocity perturbations. A low
283 velocity spike is located in the northern Musandam Peninsula, two high velocity spikes
284 sit in the UAE-Oman Mountains and the foreland basin extending to a depth of 100 km.
285 While there is a degree of recovery in Figure 5 b, f and g, there is limited evidence of
286 spike reconstruction in Figure 5 h, presumably due to the sparser station distribution in

287the southwestern area of the model. Figure S5 shows a similar example where spikes are
288only limited to the crust.

289 The last synthetic test is designed to assess whether we can recover a high velocity
290anomaly in the Gulf of Oman, as would be expected by the presence of oceanic
291lithosphere. Thus, we construct a blocky, high-velocity synthetic structure that we
292attempt to recover with our dataset (Figure 6). The recovery of the synthetic structure is
293sufficient to allow us to feel confident that the lateral resolution of the pattern of
294anomalies in the Gulf of Oman is broadly correct. However, smearing in the vertical
295direction has an impact, as indicated by the recovered structure in Figure 6d and f, which
296needs to be taken into account during interpretation of the 3-D final tomographic model.
297

2984.2 P-Wave Tomographic Model

299 The final model is obtained after running six iterations of FMTOMO, which involves
300sequential application of the eikonal solver and subspace inversion method. Both crustal
301and mantle velocity nodes are included as unknowns in the inversion. Damping and
302smoothing parameters, calibrated after examination of trade-off curves, yield a final
303tomographic model that reduces the RMS travel time residual from 540 to 388 ms, which
304corresponds to a reduction of data variance from 0.29 to 0.14 s² (52% decrease) and $\sqrt{\chi^2}$
305from 42.78 to 17.49. Variance reduction is equal to 35% when the crust is not included in
306the inversion. We present our new P-wave tomographic model in Figure 7 via a series of

vertical and horizontal sections, with velocity anomalies expressed as perturbations from the laterally averaged version of the final model. Two horizontal slices are taken at 60 km and 120 km depth (Figure 7a and b); one east-west vertical cross-sections is taken at 24.9°N (Figure 7c), and two north-south vertical cross-sections are taken at 55.5°E and 56.4°E (Figure 7d and e). We also show two vertical profiles along great circles, as illustrated in Figure 7f and g (Figure S6 shows the tomographic model with no prior crustal model included in the inversion). Synthetic tests described in Section 4.1 demonstrate that we are capable of resolving velocity anomalies as small as ~110 km across above 150 km depth. If the size of the velocity perturbation is greater than ~170 km, then our dataset is capable of constraining such anomalies to a maximum depth of ~350 km. Anomalies greater than ~220 km in size are effectively recovered throughout the model region.

Although additional stations from Oman and southeastern Saudi Arabia could help widen the seismic network aperture and constrain deeper mantle structure, we believe that our tomographic model is an improvement when compared to global or continental scale tomographic images of the region (see a comparison of our model with a few selected models in Figure S7). In particular, larger scale models are typically dominated by the continental keel beneath the Zagros mountains, characterized by high velocity anomalies, and low velocity anomalies of the oceanic lithosphere in the Gulf of Oman. However, they are not able to detect the novel aspects highlighted here, an endeavour possible thanks to the new relatively dense data coverage we are equipped with.

Our new 3-D P-wave tomographic model reveals a number of intriguing velocity anomalies. A relatively high-velocity feature is located in the offshore Gulf of Oman (H1 in Figure 7a and c). H1 is bounded to the north by a relatively low velocity anomaly, while its western edge is nearly parallel to the strike of the UAE-Oman mountain range and runs along the coastline. This feature extends north-south for about 200 km (Figure 7a) and is visible up to a depth of 140 km in the upper mantle (Figure 7c), with a maximum velocity perturbation of about 0.4 km/s. In contrast, the northern Musandam Peninsula is characterized by a relatively low velocity anomaly, which is labelled as L1 in Figure 7a, b, e, f. L1 extends from the Moho discontinuity to the upper mantle, suggesting that this low-velocity anomaly is predominantly affected by the structure of the upper mantle rather than the crust. At depths greater than 150 km, however, this anomaly appears to be strongly smeared out. An important series of anomalies include the staggered high- and low-velocity channels that exhibit a northwest-southeast trend. They are roughly parallel to each other and are marked with labels L1, H2, L2, and H3.

Discussion

5.1 Extent of Tethyan oceanic lithosphere in the Gulf of Oman

Obduction of Tethys lithosphere initiated immediately after formation of the ophiolite above an active subduction-zone setting (NE dipping) during the Cenomanian (Pearce et al., 1981; McLeod et al., 2013; Rioux et al., 2013). The Semail ophiolite and

underlying thrust sheets (distal Tethyan sedimentary rocks of the Haybi and Hawasina complexes) were progressively emplaced from NE to SW onto the subsided Arabian passive continental margin in a process that persisted for around 20 Ma. Following emplacement of the main thrust sheets, it is likely that the continental margin of Arabia was pulled down to depths of about 80-100 km, as indicated by the presence of eclogites formed at PT conditions of 21-23 kbar and temperatures of 540-580°C at 79 Ma (Searle et al., 2004; Warren et al., 2003). Slab break-off may have occurred soon after, as implied by rapid exhumation due to the buoyant behaviour of the Permian carbonates and their eclogitic boudins. Remnants of Tethys Cretaceous oceanic lithosphere in the Gulf of Oman are presently being subducted beneath southeastern Iran and Makran (Pakistan).

359 The precise offshore extent of attenuated Arabian continental lithosphere beneath
360 Tethyan oceanic rocks is still unknown, although an extremity that runs roughly parallel to
361 the coastline in the immediate offshore has been suggested using a combination of several
362 geophysical methods (Ali et al., 2020). To the north, however, lack of data in the offshore
363 between the Musandam Peninsula and southern Iran makes the northern boundary of the
364 oceanic lithosphere even more cryptic. The Dibba zone is a structurally complex area
365 dividing the autochthonous Permian shelf carbonates from the Semail ophiolite. It
366 consists of a series of WSW-verging thrust sheets that are exposed onshore, but its
367 offshore continuation has been elusive thus far. Feature H1 in Figure 7a, which highlights
368 an E-W contrast from relatively high to low velocity anomalies in the Gulf of Oman,
369 exhibits a western limit that mimics the coastline, while its northern reach appears to

370spatially correlate with an offshore extension of the Dibba fault. We interpret this change
371to mark the approximate location of the western boundary of Tethyan oceanic
372lithosphere. The synthetic test in Figure 6 suggests that this is a laterally well-resolved
373anomaly but, as is typical of teleseismic tomography, is subject to vertical smearing
374effects that preclude us from imaging a potential slab break-off, continental
375underthrusting, or make meaningful estimates of lithospheric thickness. It is, however,
376apparent from Figure 7a and c that feature H1 is a significant feature that likely represents
377a distinct lithospheric boundary. Moreover, feature H1 appears to be underlain by a slow
378velocity anomaly, with a marked transition occurring at ca 150 km depth. Cretaceous
379oceanic lithosphere would be expected to be thicker than younger - and closer to a mid-
380ocean ridge - oceanic lithosphere, thus this transition in the upper mantle may hint at a
381change between mechanically weaker asthenosphere and overlying colder lithosphere.

382

3835.2 Evidence for continental growth in the eastern Arabian Platform?

384 Structural trend analysis on the regional gravity and magnetic dataset of the UAE
385(Ali et al., 2014, 2017) reveals several lineation trends, ranging from N-S in the
386westernmost UAE, to NW-SE west of the frontal Semail thrust, and NE-SW in
387southeastern UAE. A dominant feature of the residual gravity and magnetic data within
388the sedimentary sequence is the delineation of several NW-oriented left transpressional
389and transtensional faults that cut and offset N-S and NE-SW trending faults. Two well-
390defined lineaments (see Figure 1 for location), called Abu Dhabi Lineaments (Ali et al.,

3912014), run across the Abu Dhabi Emirate and have a strike comparable to the
392Neoproterozoic shear Najd Fault System and the Abu Jifan Fault. Low velocity anomaly
393L2 in Figure 7b appears to be roughly bounded by the Abu Dhabi Lineaments and reveal
394a strike akin to the Najd and Abu Jifan fault systems. Additionally, it has generally been
395observed that major strike-slip faults can be associated with the presence of low velocity
396anomalies, as initially interpreted for the San Andreas Fault (see Feng and McEvilly,
3971983), and later in other contexts such as the North Pyrenean Fault (Roure et al., 1989)
398and the Anatolian Fault (Papaleo et al., 2017). These similarities raise the possibility that
399feature L2 reflects the presence of a major, deep-seated lithospheric shear zone that may
400have been periodically reactivated to accommodate shear in the shallower levels of the
401sedimentary sequence. Alternatively, Ali et al., (2014) suggest that the northern Abu
402Dhabi Lineament could delineate the boundary between highly attenuated and
403unstretched continental crust to the north and south of the lineament, respectively. Crustal
404thickness suffers from poor constraints in the region and it is unclear from our
405tomographic model if low velocity anomaly L2 is coincidental or provides insights into
406the geometry of the Tethyan margin of the Arabian Plate. Another low velocity anomaly,
407which is NS-oriented, is located in the immediate offshore of the western UAE (centred
408at $\sim 24.5^{\circ}\text{N}$ and 52.5°E , Figure 7a). Although the sparser station distribution in this part of
409the tomographic model results in a blocky-like appearance of the anomalies, this feature
410overlaps with the tectonic trends (see Figure 1c) identified by Ali et al. (2017), from
411potential field (gravity and magnetic) and 3-D seismic reflection data. It is likely that this

low velocity anomaly in the upper mantle reflects original Neoproterozoic tectonic elements of the Arabian Plate, as suggested by Ali et al. (2014, 2017).

The absence of direct pre-Phanerozoic geological and geochemical observations in the UAE (except for the salt domes in the diapiric islands of the Arabian Gulf) makes it difficult to reconcile the location of our velocity anomalies with tectonic provenance or rock type. Several tectonic windows of Neoproterozoic basement in the adjacent Oman region can provide important clues, particularly those found in Jebel Akhdar and Jebel Ja'alan (Figure 1), which are along strike with the P-wave velocity anomalies found in Figure 7b. Alessio et al., (2018) performed geochronological analysis to constrain the age of magmatism and metamorphism from granodioritic samples of Jebel Ja'alan. They inferred that the volcanic-arc nature of the rock samples is related to the Tonian subduction of the Mozambique Ocean that divided Neoproterozoic India from the Arabia-Nubia Shield in Gondwana. Arc magmatism rocks of eastern Oman, including the Jebel Ja'alan granodiorites and possibly volcanic clasts of Jebel Akhdar, are interpreted to have similar geochemical signature to those found in the Arabian Shield but opposite polarity of accretion. In contrast, Tonian granitoid occurrences in NW India and Pakistan show geochronological affinities with those of eastern Oman, implying a connection in Gondwana between eastern Oman and NW India, rather than the Arabian Shield. High velocity anomaly H2 in Figure 7b and f is clearly juxtaposed against two low wavespeed perturbations and runs roughly parallel to L2 and H3. The anomaly is subvertical and laterally well resolved but affected by inherent vertical smearing below 200 km depth.

433Unfortunately, our tomographic model does not extend sufficiently far into eastern Oman,
434where Neoproterozoic basement occurrences have been recognized. However, we have
435no reason to believe that basement beneath eastern Oman does not extend beneath the
436UAE. Therefore, if we assume a continuation to the northwest of the rocks found in Jebel
437Ja'alan by some 50-100 km, anomaly H2 can be attributed to the signature of a volcanic
438arc formed during closure of the Mozambique Ocean in the Tonian (Whitehouse et al.,
4392016; Alessio et al., 2018; Blades et al., 2020). H3 could be another potential candidate
440for the arc magmatism signature in our tomographic model, although it lacks resolution
441since it is located in the peripheral area of the seismic network. The alternation of NW-
442trending low and high velocity anomalies observed in Figure 7b are potentially consistent
443with the idea of continental growth in eastern Arabia, possibly via magmatic accretion
444(Whitehouse et al., 2016; Alessio et al., 2018; Blades et al., 2020). While this model is
445yet to be fully tested, further high-resolution tomographic studies can help shed light on
446the tectonic evolution of the Arabian Platform.

447

4486 Conclusions

449 We have inverted body-wave arrival-time residuals to determine the lithospheric
450structure beneath a new network of 55 seismic stations in the eastern Arabian Platform.
451Assessment of solution robustness demonstrates that our tomographic model can
452generally discriminate features with a scale length of approximately ~ 110 km up to a
453depth of 150 km. Below this threshold, only structures larger than ~ 170 km can be

454recovered, although they can be affected by inherent vertical smearing. The 3-D
455tomographic solution model constrains the boundary between Tethyan oceanic
456lithosphere and the Arabian continental passive margin in the Gulf of Oman, and
457identifies the northern extent of Tethyan oceanic lithosphere as an offshore prolongation
458of the Dibba fault (Figure 8). With virtually no basement exposure, constructing a
459detailed terrane accretion history of eastern Arabia remains a major challenge for most of
460the Arabian Platform. Our results concur with recent suggestions that continental growth
461through magmatic accretion may have characterized eastern Arabia (Figure 8) but
462additional high-resolution lithospheric studies, particularly in Oman and eastern Saudi
463Arabia, are needed to address important questions regarding the nature and origin of the
464lithosphere in the region.

465

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474 Our 3-D tomographic model can be downloaded from the following digital object
475 identifier [10.6084/m9.figshare.12459665](https://doi.org/10.6084/m9.figshare.12459665).

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Figure 1: a) Location of Arabian Peninsula. Red rectangle highlights the study area shown in 686). b) Simplified tectonic map of the southern Arabian Peninsula, illustrating the main 687ructures and tectonic domains of the Arabian Shield (to the west) and Arabian Platform (to 688e east). Pink areas are Ediacaran salt basins. The western extent of the Rayn microplate 689yellow areas in eastern Arabia) is currently unknown. Modified from Al-Husseini, 2000 and 690tern and Johnson, 2010. c) Distribution of seismic stations displayed as blue circles. Yellow 691diamonds depict the location of exposed Precambrian basement. Red lines are major tectonic 692oundaries. Dashed black lines display the location of major tectonic trends derived from Ali 693a al., (2014, 2017). ADL: Abu Dhabi Lineaments; MP: Musandam Peninsula; ZFTB: Zagros 694old-and-thrust belt.

695

696

697 **Figure 2:** Adaptive stacking example using an earthquake located at 122.05°E, 24.68°N and
698 km depth. (a) Initial trace alignment achieved by using move-out corrections based on the
699 k135 global reference model; (b) Final alignment after application of the adaptive stacking
700 procedure; (c) Pattern of residuals obtained from the difference between the initial and final
701 stack. In both (a) and (b), flat traces indicate stations that did not record the seismic event.

703

704 **Figure 3:** a) Depth slice at 10 km extracted from the velocity model of Pilia et al. (2020),
705 converted to P-wave velocities using Brocher (2005). b) Moho depths from Ali et al. (2020)
706 used to constrain the base of the crust in the tomographic inversion. Both maps have a pixel
707 resolution equal to that used during the inversion (i.e., velocity and interface grids are
708 refined by 49 and 41 points in longitude and latitude, respectively).

709

710

711 **Figure 4:** Checkerboard resolution tests for our tomographic study, using alternating high-
712 and low-velocity anomalies of 111 km width and ± 0.4 km/s perturbation in amplitude.
713 Velocity perturbations are relative to a laterally averaged starting velocity model. (a) and (c)
714 are input models at 60 and 175 km depth. (b) and (d) are their respective output models. (e)

715 and (f) are input and output models from cross sections taken at 24.90°N. (g) and (h) are
 716 input and output models from cross sections taken at 55.90°E.

718

719 **Figure 5:** Synthetic test involving spike structures extending from the crust to the upper
 720 mantle. (a) and (b) are input and output models for structures taken from a horizontal slice at
 721 60 km depth. (c), (d) and (e) show vertical slices of input synthetic spikes, while (f), (g) and
 722 (h) are the corresponding output models.

723

Figure 6: Resolution test for a high velocity structure in the upper mantle simulating the presence of oceanic lithosphere in the Gulf of Oman. (a) and (b) are respectively input and output models at 60 km depth. (c) and (e) are input models for AA' and BB' cross-sections shown in (a), while (d) and (f) are their respective recovered models.

Figure 7: Horizontal and vertical slices taken from the final P-wave tomographic solution model. Velocity anomalies are relative to the ak135. (a) and (b) are horizontal slices at 60 and 120 km, respectively. (c), (d) and (e) are vertical cross sections with locations marked in (a). (f) and (g) are vertical cross sections taken from the great circle as shown in (b). Dashed black lines in a) and b) are tectonic trends as per Figure 1. Dashed red lines denote

736ansitions between anomalies of different sign, although we acknowledge that the zero
737ontour is not intrinsically significant, since only relative velocity values are constrained.
738

739

740 **Figure 8:** Schematic illustration of the interpretation that arises from this study. The black
 741 thick lines are possible tectonic boundaries as identified from Figure 7b. The dashed black
 742 line is the boundary of Tethyan oceanic lithosphere as inferred from Figure 7a. Black and
 743 white background is topography, while the colours are as per Figure 7b.

744