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# **Intra-oceanic emplacement of the Comoros Archipelago through inherited fracture zones**

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## Highlights:

- The crust of the Comoros Basin is oceanic, resolving past controversies
- Basement roughness is typical of intermediate-spreading ridges
- Unloaded basement depth matches Cretaceous or Jurassic oceanic lithosphere
- Comoros volcanic islands emplaced through re-activated fracture zones
- The Comoros Archipelago aligns with a change in the direction of the fracture zones

32

33 **Abstract**

34 We shed light on the nature and structure of the crust surrounding the Comoros  
35 Archipelago, western Indian Ocean, offering insights into the region's geological history  
36 and volcanic island formation. Our comprehensive study encompasses the acquisition  
37 of new, deeply penetrating seismic data from the SISMAORE cruise (refraction and  
38 reflection seismic), and the subsequent analysis of the characteristics and structure of  
39 the crust surrounding the Comoros Archipelago. Both the reflection seismic imaging  
40 and the velocity structure using ocean bottom seismometers indicate that the crust of  
41 the Comoros Basin crust exhibits oceanic characteristics, thus resolving previous  
42 controversies about its nature. The thickness of the oceanic crust ranges from 5.8-  
43 6.6 km in the north of the archipelago to 6-7.2 km within the Comoros Basin. The  
44 estimated roughness of the top basement in the Comoros Basin ranges from 110 to  
45 200 m values typical of intermediate spreading ridges, such as the extinct spreading  
46 centre in the West Somali Basin.

47 The unloaded basement depth of the Comoros Basin closely matches the expected  
48 water-loaded subsidence for a Cretaceous or Jurassic oceanic lithosphere. In contrast,  
49 the West Somali Basin to the north of the Comoros Archipelago has shallower  
50 basement depths, potentially linked to recent volcanic activity along the archipelago.  
51 We propose that the pre-existing oceanic fracture zones in the West Somali Basin  
52 underwent reactivation, first during the Turonian period and later during the Late  
53 Eocene. These reactivated fracture zones may have acted as preferred pathways for  
54 the emplacement of the volcanic islands of the Comoros Archipelago. The EW trend  
55 of the archipelago appears to follow a marked change in the direction of these  
56 reactivated fracture zones, suggesting that the associated lithospheric weakening  
57 likely played a critical role in facilitating the formation of the Comoros Archipelago.

58

59 **1. Introduction**

60 The origin of the Comoros Archipelago (Fig. 1), in the northern Mozambique  
61 Channel, has been the subject of lively debate for some fifty years. It has long been  
62 interpreted as being due to the impingement of a mantle plume beneath a moving  
63 oceanic lithosphere (Emerick and Duncan, 1982; Hajash and Armstrong, 1972). While  
64 hotspot volcanism may be supported by the enriched mantle signature of the lavas on  
65 Grande Comore Island (Class et al., 2009), the general E-W trend of the Comoros

66 Archipelago is not consistent with the northward absolute motion of the Somalia plate  
67 (Morgan and Morgan, 2007).

68 In May 2018, a major phase of seismo-volcanic activity gave birth to the large –  
69 820 m high and 5 km in diameter - submarine Fani Maoré volcano, sitting at about  
70 3500 m water depth, east of Mayotte Island (Feuillet et al., 2021; Lemoine et al., 2020;  
71 Masquelet et al., 2022; Thinon et al., 2022). The present-day active volcanism both on  
72 Grande Comore Island and east of Mayotte Island together with widespread Holocene  
73 volcanic activity on Anjouan, Mohéli, and Mayotte Islands (Quidelleur et al., 2022;  
74 Rusquet et al., 2023) does not support a simple model of punctual and vertically rising  
75 mantle plume material piercing a moving lithosphere.

76 The recent seismo-volcanic crisis east of Mayotte Island together with previous  
77 sparse focal mechanisms along the Comoros Archipelago suggest mainly strike-slip  
78 faulting with a NE-SW normal component (Feuillet et al., 2021; Lemoine et al., 2020).  
79 This seismicity has been interpreted as resulting from a nascent plate boundary  
80 between the Somalia and Lwandle plates connecting the East African Rift System  
81 (EARS) to the Tertiary-Quaternary volcanism and rifting of Madagascar (e.g. (Famin  
82 et al., 2020; Feuillet et al., 2021; Thinon et al., 2022). However, there is no consensus  
83 over the offshore extent of the EARS. The deformation identified onshore and on the  
84 seafloor, is thought to be either focused within a 150-200 km-wide right-lateral shear  
85 zone along the Comoros Archipelago (Famin et al., 2020; Feuillet et al., 2021; Thinon  
86 et al., 2022) or distributed within a ~600-km-wide deformation zone positioned  
87 between the Rovuma microplate and the Somalian plate that extends over large parts  
88 of Madagascar and the Comoros islands (Stamps et al., 2021).

89 Although crucial to understanding the origin of the archipelago, the nature of the  
90 crust beneath the Comoros volcanic islands and in the Comoros Basin, between the  
91 archipelago and Madagascar, is poorly known. Fundamentally different interpretations  
92 have been issued. Oceanic crust, continental crust and a wide continent-ocean  
93 transition have been proposed for the Comoros Basin e.g. (Coffin et al., 1986; Lort et  
94 al., 1979; Nougier et al., 1986). In this paper, we first briefly review these  
95 interpretations, then present our newly collected seismic data set, show our results on  
96 the oceanic nature and structure of the crust around the Comoros Archipelago and  
97 finally discuss the implications of our findings for the emplacement of the magma that  
98 formed the Comoros volcanic islands. We show that the location of the islands is

99 controlled by the pre-existing segmentation of the oceanic crust of the West Somali  
100 Basin.

101  
102

## 103 **2. Geological Background**

104 North of the Comoros Archipelago, magnetic anomalies related to the spreading  
105 ridge of the West Somali Basin argue for a typical oceanic lithosphere e.g. (Cochran,  
106 1988; Davis et al., 2016; Sauter et al., 2016; Ségoufin and Patriat, 1980). There,  
107 magnetic anomalies, from M24 (~152 Ma, according to the time scale of Ogg, (2020))  
108 to M0 (~121 Ma) indicate slow to intermediate spreading rates (~30 to ~60 km/Ma full  
109 rates) until Aptian time when spreading ceased in the West Somali Basin (Davis et al.,  
110 2016). The nature of the crust south of the archipelago is less clear. Rabinowitz et al.  
111 (1983) interpreted magnetic anomalies up to the M25 magnetic anomaly (~153 Ma),  
112 while Davis et al. (2016) and Cochran (1988) were unable to correlate any significant  
113 magnetic anomalies on a few magnetic profiles. From an early sonobuoy experiment  
114 using only two instruments, Lort et al. (1979) speculated that the Comoros Archipelago  
115 marks the boundary between oceanic crust to the north and continental crust to the  
116 south. Later, with over a hundred sonobuoy records, Coffin et al. (1986) showed that  
117 both the West Somali and Comoros basins are oceanic in character.

118 However, the observation of sandstone xenoliths in Comoros lavas (Flower and  
119 Strong, 1969) and the presence of a km-scale quartzite massif on Anjouan Island have  
120 been used to suggest that the archipelago may be built on stretched continental crust  
121 (Montaggioni and Nougier, 1981). It was further proposed that magmatism along the  
122 Comoros Archipelago is controlled by deep lithospheric faults, linked to an inferred  
123 wide continent-ocean transition (Nougier et al., 1986; Tzevahirtzian et al., 2021).

124 Based on a single receiver function analysis at Mayotte Island, Dofal et al. (2021)  
125 proposed a continental nature of the crust beneath Mayotte. However, from a  
126 combined analysis of receiver functions and the joint inversion of receiver function and  
127 Rayleigh wave dispersion data, Dofal et al. (2023) could not discriminate the oceanic  
128 or continental nature of the crust beneath Mayotte Island. Recently, Roche and  
129 Ringenbach (2022) used seismic reflection profiles to show that oceanic crust is  
130 present at 13.5°S, to the southwest of Moheli Island. Wide-angle seismic data also  
131 showed the presence of oceanic crust on the eastern flank of the Davie Ridge, at  
132 14.5°S (Vormann et al., 2020; Vormann and Jokat, 2021).

133 The seismic stratigraphy of the Comoros and Somalia basins is based on the  
134 relatively well-known stratigraphy of the northern Mozambique Channel and the West  
135 Somali Basin (Coffin and Rabinowitz, 1992; Franke et al., 2015; Mahanjane, 2014;  
136 Mougénot et al., 1986; Roche and Ringenbach, 2022). Using IODP Site 1476 and  
137 DSDP Site 242 (Hall et al., 2017; Simpson et al., 1974) located on the Davie Ridge at  
138 the northern entrance of the Mozambique Channel, and seismic reflection profiles that  
139 intersect these sites, Masquelet (2023) and Masquelet et al.(submitted) identified the  
140 top Miocene and top Oligocene horizons (TM and TO in Fig. 3). These horizons were  
141 further extended towards the Comoros Archipelago, 500 km away. They also identified  
142 the Cretaceous-Paleogene unconformity (K/Pg, Fig. 3), which appears as a prominent  
143 event in the seismic profiles across the West Somali Basin and offshore Majunga Basin  
144 (Franke et al., 2015; Leroux et al., 2020; Mahanjane, 2014). They also interpreted the  
145 package of high amplitude and low frequency discontinuous reflectors beneath the  
146 K/Pg reflector as the Turonian volcanism previously observed in the Diego and  
147 Majunga basins offshore Madagascar (Coffin and Rabinowitz, 1992; Leroux et al.,  
148 2020).

149

### 150 **3. Data acquisition and processing**

151 A regional-scale acquisition of geophysical and geological data was conducted off  
152 the Comoros Archipelago aboard the R/V Pourquoi pas? in 2021 (Fig. 1), (SISMAORE  
153 cruise, Thinon et al., 2022, 2020). 1480 km of deep multichannel seismic reflection  
154 data (960 channels) and 6800 km of shallow seismic reflection profiles (48 channels)  
155 were acquired. In addition, three OBS (Ocean-Bottom Seismometers) recorded the  
156 seismic shots along the multichannel line M01 high-resolution profile, located East of  
157 Mayotte (Fig. 1C). Published seismic reflection profiles within the Comoros Basin  
158 complement this new data set (Fig. 1). We also incorporated the seismic reflection  
159 profiles of the R/V Sonne SO-231 cruise (Franke et al., 2015; Klimke et al., 2016) and  
160 the R/V L'Atalante PTOLEMEE cruise (Jorry, 2014; Leroux et al., 2020). We also used  
161 ION Geophysical™ seismic multichannel lines partly published by Leroux et al. (2020),  
162 Sinha et al. (2019), and Roche and Ringenbach (2022) on the eastern flank of the  
163 Davie Ridge and in the Comoros Basin.

164

#### 165 **3.1. Long-streamer 960-channel seismic reflection profiles**

166 We collected a set of 12 multichannel seismic (MCS) profiles (6000 m long  
167 streamer with 960 channels spaced every 6.25 m). The seismic source consisted of  
168 16 airguns triggered every 40 s with a total volume of ~82 L (4990 cu. in.). The record  
169 length was 20 s with a sampling rate of 2 ms. The distance between each common  
170 depth point (CDP) is 3.25 m. Super gather CDPs were built by merging four CDPs in  
171 a bin of 12.5 m during processing, allowing for a fold of 60. Processing involved a  
172 typical workflow using CGG GeoVation® software in the pre-stack domain (including  
173 trace editing, amplitude correction, normal move-out correction, FK-filtering, multiple  
174 attenuation, and predictive deconvolution). Masquelet et al. (2022), Masquelet (2024),  
175 Masquelet et al. (submitted) describe in detail the seismic processing of this data set.

176

### 177 **3.2. Short-streamer 48-channel seismic reflection profiles**

178 A series of 110 48-channel seismic reflection profiles was collected using a 600 m-  
179 long streamer (channels spaced every 6.25 m). The source consisted of two airguns  
180 with a total volume of 4.91 L (300 cu. In) firing every 50 m (recording length of 9 s and  
181 inter CDP is 3.125 m). CGG GeoVation® software was used for standard processing  
182 including 12.5 m binning (6-fold by stacking 2 adjoining traces), NMO correction, stack  
183 and a Kirchhoff migration at 1500 m/s.

184

### 185 **3.3. OBS data set**

186 A set of three OBSs recorded the shots during the acquisition of the 960 channel  
187 seismic reflection profiles (Fig. 1C). These three instruments were deployed along the  
188 ~290 km long M01 line, which is located east of Mayotte with a SSW-NNE orientation  
189 and crosses the new Fani Maoré volcano (Masquelet et al., 2022). The OBSs also  
190 recorded shots along cross lines, allowing the estimation of seismic velocities along  
191 different azimuths.

192 After clock-drift correction, the instruments were relocated using the direct wave  
193 picks from the shots within 10 km of their deployment location, the bathymetric grid,  
194 and averaged seismic velocities in the water column (1.505 km/s). The OBS records,  
195 organized as receiver gathers, allowed the picking of arrival times of several seismic  
196 phases in the different geological layers down to the uppermost mantle (Appendices A  
197 and B). We used a forward modelling program, RAYINVR (Zelt and Smith, 1992),  
198 following a top-down layer stripping strategy to generate velocity models that  
199 reproduce the picked data within their uncertainty range (Appendix C). We modelled 2

200 or 3 lines for each instrument. Here we present the velocity model along M01, which  
201 crosses all 3 instruments, although each instrument can be considered as an  
202 independent ~1D model since no seismic ray crosses the entire model (Fig. 2A). This  
203 issue implies that the obtained velocities and thicknesses may not be so well  
204 constrained, but this is counterbalanced by the use of strong constraints from the  
205 coincident high-resolution seismic reflection data. The initial velocity model was built  
206 using the picks at the main geological interfaces, where available, and a first estimate  
207 of (1) the interval velocities in the sediments along these seismic reflection profiles and  
208 (2) the apparent velocities of the corresponding refracted arrivals on the OBS records  
209 used to convert these interfaces from two-way travel-time to depth. RAYINVR  
210 modelling allows us to compute synthetic arrival times from ray-tracing through the  
211 model, which we updated to reduce the misfit between computed and picked arrivals  
212 (Appendix C). The final model is converted to two-way travel time to allow for a  
213 comparison with coincident seismic reflection data (Fig. 2B). The vertical velocity  
214 profiles (Fig. 2C) provide an assessment of the robustness of the velocity models. It is  
215 noteworthy that a vintage sonobuoy profile (Lort et al., 1979) located only 7 km away  
216 from the OBS3 shows comparable velocities in the sediment layers but is limited to the  
217 top of the crust (Fig. 2C).

218

#### 219 **4. Results**

220 The internal reflectivity of the crust imaged by seismic reflection shows a typical  
221 oceanic crustal layering as defined by the 1972 Penrose Conference (Fig. 4) (Bécel et  
222 al., 2015; Mutter and Carton, 2013; Sauter et al., 2021). It includes, from top to bottom,  
223 the following successive seismic facies: (1) a reflective top of basement with some  
224 short reflections in the shallowest part of the crust, (2) a transparent unit in the upper  
225 crust, (3) a more reflective lower crust with dipping structures and (4) a series of  
226 horizontal reflectors typically interpreted as the oceanic Moho, although various  
227 geological structures at the base of the crust may generate high-amplitude reflections  
228 (e.g. Collins et al., (1986)). We identified the seafloor multiple of the profile during the  
229 processing and took care of attenuating it using a demultiplex routine that filtered out  
230 low-velocity reflectors (i.e., velocities typical of shallow sediment layers rather than  
231 velocities expected near the Moho discontinuity). In addition, the velocity model  
232 requires two velocity gradients: a moderate velocity gradient below the basement and  
233 a low-velocity gradient above the Moho. We have therefore ensured that all the



234 reflectors at the base of the crust do not come from a multiple of the seafloor. This  
235 interpretation is confirmed by the OBS analysis, where geological layers were imaged  
236 down to the Moho or upper-most mantle (Fig. 2).

237

#### 238 **4.1. Nature of the crust in the Comoros Basin**

239 Both seismic reflection and refraction (OBS) data show evidence for the presence  
240 of oceanic crust south and SE of Mayotte (Figs. 4 and 5). Seismic reflection data show  
241 oceanic Moho reflections that are roughly parallel to the top of the crust (at tens of  
242 kilometres scale, Fig. 4). The oceanic Moho is generally characterized by a few strong  
243 and continuous reflectors (Fig. 4 A&B) at about 9.5 s TWTT. However, a package of  
244 laterally discontinuous reflectors up to 0.25 s TWTT thick can also be observed locally  
245 (Fig. 4C). It may also be faint or absent, especially towards the volcanic islands or  
246 submarine volcanic provinces of the Comoros Archipelago postdating the  
247 emplacement of the oceanic crust.

248 The results obtained from the OBS data support the interpretation of the oceanic  
249 nature of the crust. Indeed, the velocity models obtained for the three OBSs show a  
250 clear 2-layer basement, with velocities that are typical for the oceanic crust (Fig. 5).  
251 The uppermost basement layer shows velocities ranging from 5.25 to 6.70 km/s and  
252 strong vertical velocity gradients ( $0.3\text{-}0.55\text{ s}^{-1}$ ), which correspond to seismic velocities  
253 in basalts, i.e., the oceanic Layer 2. The lower crustal layer shows higher velocities  
254 ( $6.75\text{-}7.2\text{ km/s}$ ) with lower velocity gradient ( $\sim 0.1\text{ s}^{-1}$ ), typical of the gabbroic oceanic  
255 Layer 3 (Fig. 5). More than 700 wide-angle reflected arrivals were picked at the  
256 interface between Oceanic Layer 2 and Layer 3, and almost 2,300 at the Moho, thanks  
257 to the fact that incidence angle is much wider than in seismic reflection data (near-  
258 critical reflections, see modelling statistics for phases 6.2 and 7.2, respectively, in  
259 Appendix C).

260 The nature of the crust around the Comoros Archipelago has been debated,  
261 including around OBS3. Fig. 5A shows a comparison of crustal velocities beneath  
262 OBS3 with values reported in the literature. Lort et al. (1979) speculated that the crust  
263 might be continental based on velocities down to the top basement, whereas our study  
264 provides information down to the uppermost mantle. It is now clear that we observe a  
265 typical oceanic layering (Fig. 5A; White et al. (1992)). Indeed, the continental-type  
266 velocities for the extended crust are much lower than in the crust under OBS3 (Fig.  
267 5A; Christensen and Mooney (1995)). It has also been discussed whether the area

268 between the Madagascar margin and the ocean might be transitional (Nougier et al.,  
269 1986). Comparison with continental (Dean et al., 2000), and oceanic exhumed mantle  
270 zones (Van Avendonk et al., 2006)(Fig. 5A) shows that the lower crustal velocities  
271 beneath OBS3 are too slow to be consistent with a partially serpentinized mantle.

272

## 273 **4.2. Thickness of the crust**

274 We picked the top of the basement and the base of the crust to obtain the crustal  
275 thickness at places where the oceanic Moho is well marked by a series of highly  
276 reflective reflectors (Fig. 4). We then converted this TWTT thickness to kilometres  
277 using an average velocity of 6.56 km/s, which corresponds to the average seismic  
278 velocities we found for the crust from our OBS analysis.

279 Appendix D provides examples of crustal thickness measurements to the north  
280 (Table D1) and south (Table D2) of the Comoros Archipelago. Crustal thicknesses  
281 range from 1.8 to 2.07 s TWTT (5.9-6.8 km) north of Grande Comore. The crust is  
282 somewhat thicker in the Comoros Basin, ranging from 1.95 to 2.25 s TWTT (6.4-  
283 7.4 km) (see for example Fig. 3). This is consistent with our results from the OBS  
284 analysis in the Comoros Basin, which show a ~6.5-7 km thick crust beneath OBS3 and  
285 OBS2.

286 However, the crust is much thinner (~5 km) below OBS1 at the northern end of the  
287 profile. Velocity modelling shows that the thickness of the oceanic Layer 2 is  
288 homogeneously ~2-2.5 km, while the thickness of Layer 3 varies from ~2.5 km below  
289 OBS1 in the North to 4-5 km below OBS2 and OBS3 in the south (Fig. 5).

290

## 291 **5.3 Top of the oceanic crust**

### 292 *5.3.1 Small-scale oceanic crust roughness*

293 The variation of oceanic basement roughness at mid-oceanic ridges is primarily  
294 related to the spreading rate, which mainly controls the thermal state of the lithosphere  
295 (Goff et al., 1997; Sauter et al., 2018b). Estimating the roughness of the oceanic crust  
296 (i.e. the root mean square deviation of the residual basement relief) can thus provide  
297 insight into the spreading processes within the West Somali and Comoros basins. We  
298 selected three seismic reflection profiles in line with the inferred spreading direction to  
299 estimate the roughness of the oceanic crust. After removing the NS regional deepening  
300 of the basement, we obtained roughness values ranging from 110 to 200 m for the  
301 abyssal hill bathymetry along the three profiles (Fig. 6). Such roughness values are

302 typical for slow to intermediate spreading ridges (Goff et al., 1997; Sauter et al., 2018b)  
303 (Fig. 6).

304

### 305 *5.3.2 Regional depth variation of the top oceanic crust*

306 Mapping the basement depth along each seismic reflection profile (Fig. 7), we  
307 found that the top basement is much deeper in the Comoros Basin (mean depth of  
308 7 km) than in the West Somali Basin to the north of the Archipelago (mean depth of  
309 6 km). Using the sediment thickness derived from the seismic profiles and a  
310 conservative range for their mean density (2.0 to 2.2), the unloaded basement depth  
311 would be 5.4 km for the Comoros Basin and 5.0 km for the region north of the  
312 Archipelago (Fig. 8). Despite the large uncertainties inherent in such calculations, the  
313 unloaded depth in the Comoros Basin is not far from the asymptotic water-loaded  
314 subsidence inferred by Stein and Stein in their GDH1 plate cooling model (5560 m;  
315 Stein and Stein (1992)) (Fig. 8). Subsidence in the Comoros Basin is thus compatible  
316 with a Cretaceous or Jurassic age for the oceanic lithosphere. The West Somali Basin,  
317 north of the Archipelago, appears to be several hundred meters shallower (Fig.8).  
318 Seismic profiles image a rather sharp step in the topography of the top basement  
319 between 12.2°S and 12.5°S, along an EW trending line through the Comoros  
320 Archipelago joining Mohéli Island and Zélée Geyser banks (Figs 7 and 8). This  
321 shallower basement extends over a broad region, suggesting anomalous heating at  
322 mantle depth, possibly related to the recent volcanism discovered in the Mwezi and  
323 N'Droundé provinces to the northeast of Anjouan and Grande Comore islands and  
324 throughout the Comoros Archipelago (Rusquet et al., 2023; Thinon et al., 2022).

325

## 326 **5.4 Compressional structures**

327 Seismic profiles north of the Comoros archipelago document two distinct phases of  
328 compressional deformation. The first is Turonian in age and the second is Late  
329 Eocene.

330 The oldest event is recorded in the deepest sedimentary layers as a small dome  
331 bounded by reverse faults of opposite vergence (Figures 9 and 10). The dome is  
332 sealed by post-Turonian sediments, indicating an early phase of deformation. The  
333 basement faults are poorly imaged, but the sedimentary dome clearly coincides with a  
334 crustal basin of similar dimensions, suggesting tectonic inversion along basin-  
335 bounding faults (FZ1 in Fig. 9). Similar basins observed along the M02 profile (FZ2a

336 and b in Figs. 9 and 10) can be traced further north in the seismic lines (Fig. 10). They  
337 systematically correspond to elongated NNW-SSE trending gravity lows in the satellite-  
338 derived free-air gravity anomaly (Fig. 9) and are therefore interpreted as transform  
339 faults or fracture zones valleys that formed at the extinct seafloor spreading centre of  
340 the West Somali Basin (Davis et al., 2016) and were later reactivated.

341 The most recent event is associated with stronger compressional features that are  
342 well imaged north of Grande Comore Island along two intersecting EW and NS seismic  
343 profiles. The main episode of deformation is marked by a dense network of inverse  
344 faults in both the sedimentary cover and the crust. Oligocene sediments seal these  
345 structures. The amplitude of the uplifted areas reaches 0.3-0.4 s TWTT (300-400 m  
346 with an interval velocity of 2 km/s). The EW trending profile shows a series of strong  
347 reflectors occurring  $\sim 1$  s below the top of the basement, all located beneath the uplifted  
348 area and a former small basin within the crust. These reflectors connect the oceanic  
349 Moho both east and west of the deformed area. We thus interpret them as  
350 corresponding to the deformed oceanic Moho. The thin crust beneath the uplifted area  
351 is of the type typically observed at transform faults near their intersection with oceanic  
352 ridge segments. Both the top basement and the Moho are shallower to the east of the  
353 deformed area than to the west, which may indicate a younger-aged oceanic  
354 lithospheric compartment to the east, suggesting that the profile crosses a fracture  
355 zone. Note that the Turonian phase is also observed in the easternmost part of this  
356 EW profile with a small doming of the early sediments (Fig.11A).

357 Both the Turonian and Oligocene compressive events thus appear to have  
358 overprinted the original network of transform faults and fracture zones inherited from  
359 the paleo-spreading centre of the West Somali Basin. Since some small faults reach  
360 the seafloor, we further propose that such reactivation may have continued until recent  
361 times. Compressive structures show several strong reflectors within the crust above  
362 the oceanic Moho and locally within the uplifted area (Fig. 11 B). We interpret these  
363 reflectors as indicative of magma intrusions along faults, typically found in the lower  
364 oceanic crust (Sauter et al., 2021) and possibly up to the seafloor. As magma is unlikely  
365 to be present 120 Ma after the cessation of the spreading, these intrusions are more  
366 likely to be related to the active magmatic province around the Comoros Archipelago,  
367 which would then extend 200 km further north than previously mapped (Thinon et al.,  
368 2022; Tzevahirtzian et al., 2021).

369

## 370 **5. Discussion**

### 371 **5.1. Nature of the Comoros Basin**

372 We found evidence for oceanic crust in the Comoros Basin, consistent with  
373 previous studies (Coffin et al., 1986), and have thus resolved past controversies about  
374 the nature of the crust around the Comoros Archipelago.

375 The Moho seismic reflection is weak in some parts of the study area. This is not  
376 surprising, as there are many places, away from the axis of the mid-oceanic ridges,  
377 where the crust-mantle transition shows no distinguishable reflection (Mutter and  
378 Carton, 2013). Either the Moho interface is highly variable in structure, resulting in a  
379 gradual transition zone that is broad enough not to produce a vertical incidence  
380 response, and/or the scattering and attenuation properties in the crustal section lead  
381 to variable imaging conditions at the Moho level (Mutter and Carton, 2013). In our  
382 survey area, volcanic overprinting enhances the variability of both the sedimentary  
383 cover and the crustal structure (Masquelet et al., 2022). Locally absent Moho  
384 reflections have also been noted in the northern part of the West Somali Basin, where  
385 the variable seismic signature of the base of the crust has been attributed to faulting,  
386 deformation and/or intrusions (Sauter et al., 2018a; Sinha et al., 2019). Furthermore,  
387 no evidence for a clear oceanic Moho reflection was reported by Klimke et al. (2016)  
388 east of the Davie Ridge, where oceanic crust is inferred from wide-angle data  
389 (Vormann et al., 2020; Vormann and Jokat, 2021).

390 Because our data set does not include profiles right across the islands, we cannot  
391 rule out the possibility of some isolated continental blocks below the Comoros Islands,  
392 as has been hypothesized beneath some volcanic islands surrounded by oceanic crust  
393 such as Iceland or Mauritius (Foulger et al., 2020). Our dataset further shows that the  
394 young Fani Maoré volcano, which crowns the Mayotte volcanic edifice 50 km to the  
395 east of Mayotte Island, does not sit directly on the top of the basement but on a >2 km  
396 thick sedimentary unit above the oceanic crust (Masquelet et al., 2022). We conclude  
397 that such continental blocks, if present beneath the islands of the archipelago, would  
398 be of very small extent (<30-40 km) which seems unlikely when compared to typical  
399 microcontinents such as the undisputed examples of Jan Mayen and the Seychelles  
400 (Péron-Pinvidic and Manatschal, 2010).

401 The obtained 6-7 km crustal thicknesses in the Comoros Basin are in agreement  
402 with the 6.72 km average thickness for oceanic crust >125 Ma old (Christeson et al.,  
403 2019). Similar crustal thicknesses have been obtained from both seismic reflection

404 profiles and wide-angle seismic data on the eastern flank of the Davie Ridge between  
405 13.5 and 14.5°S (Roche and Ringenbach, 2022; Vormann et al., 2020; Vormann and  
406 Jokat, 2021). This range of thickness, together with the estimated roughness of the top  
407 basement (110-200 m), is typical for spreading ridges with a relatively high magmatic  
408 budget and minor contribution from tectonism (Goff et al., 1997). This also agrees with  
409 the intermediate spreading rate indicated by the magnetic anomalies for the extinct  
410 spreading centre of the West Somali Basin (Davis et al., 2016).

411 Locally the crust may be much thinner (e.g. ~5 km below OBS1). The velocity  
412 models show that while the oceanic layer 2 does not change along the M01 profile,  
413 Layer 3 is ~2 km thinner below OBS1. Because a thinner Layer 3 is often observed at  
414 the segment ends of slow to intermediate spreading ridges e.g. (Minshull et al., 2006),  
415 this variation in Layer 3 thickness may indicate the presence of a fracture zone at the  
416 northern end of the M01 profile, beneath OBS1. We note that the top of the crust along  
417 the M01 profile is shallowing northward with two topographic steps between OBS3 and  
418 OBS2 and below OBS1 separating sections of more or less constant depth (Fig. 2).  
419 We interpret this stepwise depth variation as indicating successive sections of oceanic  
420 lithosphere of different ages and thus the presence of fracture zones that offset  
421 segments of the extinct spreading ridge of the West Somali Basin. Moreover, a basin  
422 is well imaged beneath OBS1 on profile M02, which also corresponds to a gravity  
423 lineament revealing the paleo-segmentation of the extinct spreading centre (Figs. 9  
424 and 10). Some of these fracture zones have been locally reactivated.

425

## 426 **5.2. Pre-existing structures in the Comoros and West Somali Basins and** 427 **emplacement of the Comoros Archipelago**

428 Both reactivation of NS tectonic features and volcanic activity occurred along the  
429 Davie Ridge in the Late Cretaceous after the cessation of the southward drift of  
430 Madagascar (Bassias and Leclaire, 1990; Mascle et al., 1987). Reactivation of former  
431 extensional faults has also been documented for the onshore and offshore Majunga  
432 Basin during the Late Cretaceous (Razafindrazaka et al., 1999). Transpressional  
433 events have also been evidenced from the Late Cretaceous to the Paleogene along  
434 other NS-trending major fracture zones close to the East African margin, such as the  
435 Seagap (offshore central Tanzania) and Walu faults (i.e. the northern extension of the  
436 Davie Ridge (Roche and Ringenbach, 2022). Here we show that such a reactivated

437 fracture zone is also observed in the Late Eocene north of the Comoros Archipelago,  
438 away from these major transform faults.

439 Several islands of the archipelago, such as Grande Comore, Anjouan and Mayotte,  
440 show an elongated shape, several rift zones, numerous dikes and alignments of  
441 offshore volcanoes with a main NS to NNW-SSE orientation, along with some minor  
442 orientations (Famin et al., 2020) (Fig. 12). Similarly, the recently discovered N'droundé  
443 Ridge province (see Fig. 1 for location) and the Domoni ridge between Grande Comore  
444 and Mohélie islands show the same ~NS orientation (Thinon et al., 2022). (Famin et  
445 al., 2020) showed that the maximum horizontal stresses of the paleostress tensors  
446 obtained from fault analysis at Mayotte, Anjouan and Mohéli islands are consistently  
447 oriented NNW-SSE. We also note that several post-seafloor-spreading volcanic  
448 edifices of Late Cretaceous age have been found aligned with fracture zones of the  
449 extinct spreading centre both in the northern part of the West Somali Basin (Cruciani  
450 and Barchi, 2016) and on the eastern flank of the Davie Ridge (Klimke et al., 2016). It  
451 has also been suggested that deep lithospheric melt was channelized along shear  
452 zones up to shallow deformation corridors to feed volcanic edifices along pre-existing  
453 oceanic fracture zones of the West Somali Basin (Sauter et al., 2018a). We therefore  
454 propose that the emplacement of the volcanic edifices of the Comoros Archipelago is  
455 partly controlled by mechanically weak zones along pre-existing fracture zones of the  
456 oceanic crust.

457 Several analyses of the satellite-derived gravity data have resulted in maps of  
458 gravity lineaments related to fracture zones, all of which show a change in spreading  
459 direction at about 12°S from NW-SE to the south, to NS in the West Somali Basin  
460 (Davis et al., 2016; Phethean et al., 2016; Reeves, 2018; Sauter et al., 2018a) (Fig. 12).  
461 We note that the oceanic crust is thinner north of the archipelago and thicker south,  
462 and that the depth of the oceanic basement is shallower in the north than in the south  
463 (Fig. 7). The different thicknesses and depths could indicate a change in the magma  
464 budget related to the change of spreading direction. This change is also well defined  
465 by the striking bend in the Davie Ridge at 12-14°S, which shows diffuse extension  
466 compared to more focused extension further north and south, in the Kerimba and  
467 Lacerda Miocene to Pliocene narrow grabens, respectively (Franke et al., 2015) (Fig.  
468 10). The islands of the Comoros Archipelago show a general EW alignment at the  
469 same latitude (12-13°S) with NS gravity lineaments to the north and NW-SE ones to  
470 the south (Fig. 12). We compare this geodynamic context with the so-called Buried

471 Hills in southeastern Sri Lanka. These magmatic structures form an elongated NE-SW  
472 trending chain between the NS-oriented 85°E Ridge and the Afanasy Nikitin  
473 Seamounts (Altenbernd et al., 2020). They are located at a border between swaths of  
474 oceanic crust formed by seafloor spreading in two different directions. It was thus  
475 suggested that they may have been set up along a line of pre-existing structural  
476 weakness (Altenbernd et al., 2020). Reactivation of oceanic fracture zones has also  
477 been inferred to explain the formation of the Bathymetrists seamounts in the Central  
478 Atlantic (Van Der Zwan et al., 2023). There, a change in the motion of the African plate  
479 60-40 Ma ago, together with the reactivation of the fracture zones of the strongly  
480 sheared Equatorial Atlantic, created pathways in the lithosphere and possibly  
481 enhanced magmatism (Van der Zwan et al., 2023). We thus suggest that the  
482 segmentation of the extinct spreading centre of the West Somali Basin strongly  
483 controls the emplacement of the Comoros Archipelago. This does not exclude other  
484 controlling mechanisms, such as those suggesting a nascent plate boundary between  
485 the Lwandle and Somalia plates (Famin et al., 2020; Feuillet et al., 2021). The Late  
486 Eocene reactivation phase, coeval with the onset of the Arabia-Eurasia collision at the  
487 Eocene/Oligocene boundary (Song et al., 2023), may have triggered the weakening of  
488 the fracture zones in the Somali and Comoros basins. This phase would then result in  
489 the emplacement of volcanic edifices along a weakened area where these fracture  
490 zones change direction, favouring a nascent plate boundary.

491

492

## 493 **6. Conclusions**

- 494 1) The crust of the Comoros Basin is decidedly oceanic in character resolving  
495 controversies of the past.
- 496 2) The crust is thinner (5.76-6.62 km) north of the Archipelago and thicker (6-  
497 7.2 km) in the Comoros Basin and has a top basement roughness of 110-  
498 200 m. Such crustal thickness and top basement roughness are typical of  
499 crust formed at intermediate spreading ridges such as the extinct spreading  
500 centre of the West Somali Basin.
- 501 3) The unloaded basement depth of the Comoros Basin closely matches the  
502 asymptotic subsidence predicted by the Stein and Stein (1992) plate cooling  
503 model, indicating a possible Cretaceous or Jurassic age for the oceanic  
504 lithosphere. The West Somali Basin north of the archipelago appears to have



505 a shallower basement, possibly linked to anomalous mantle heating  
506 associated with recent volcanism in the Mwezi and N'Drounde provinces  
507 northeast of Anjouan and Grande Comore islands, respectively.

508 4) The pre-existing fracture zones of the extinct spreading centre of the West  
509 Somali Basin have been reactivated first in Turonian time and then in Late  
510 Eocene time. They may have served as mechanically weak zones controlling  
511 the emplacement of the volcanic islands of the Comoros Archipelago.

512 5) We suggest that the weakened area formed by the alignment of changes in  
513 direction of these fracture zones may have favoured the emplacement of the  
514 Comoros Archipelago.

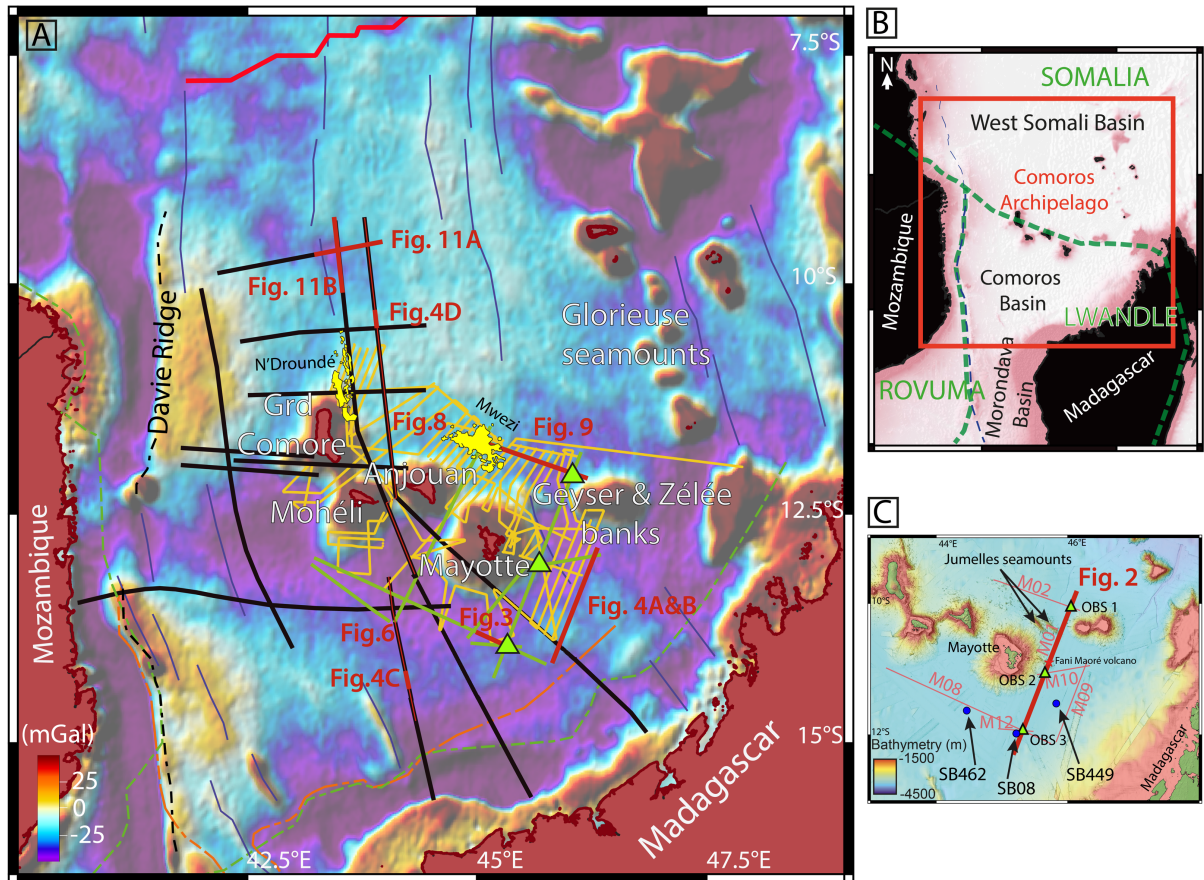
515

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531 seismic reflection profiles.

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536

537 **Figure 1:** Map of the study area offshore the Comoros archipelago (A). Orange

538 lines indicate the 48-channel seismic reflection profiles and green lines show the 960-

539 channel seismic reflection profiles collected during the SISMAORE cruise (Thinon et

540 al 2020). Black lines indicate the TGS lines available for the study. The green triangles

541 indicate the location of the OBSs (ocean bottom seismometers) of the SISMAORE

542 cruise. The N'Droundé and Mwezi volcanic provinces are shown in yellow. The blue

543 lines correspond to the fracture zones of Davis et al. (2016). The green and orange

544 dashed lines correspond to the COB (Continent-Ocean Boundary) of Roche and

545 Ringenbach (2022) and Sinha et al. (2019), respectively. The black dashed line

546 corresponds to the Davie Ridge. Thick red lines indicate the part of the profiles shown

547 in the following figures of this paper. The background image is the satellite-derived free

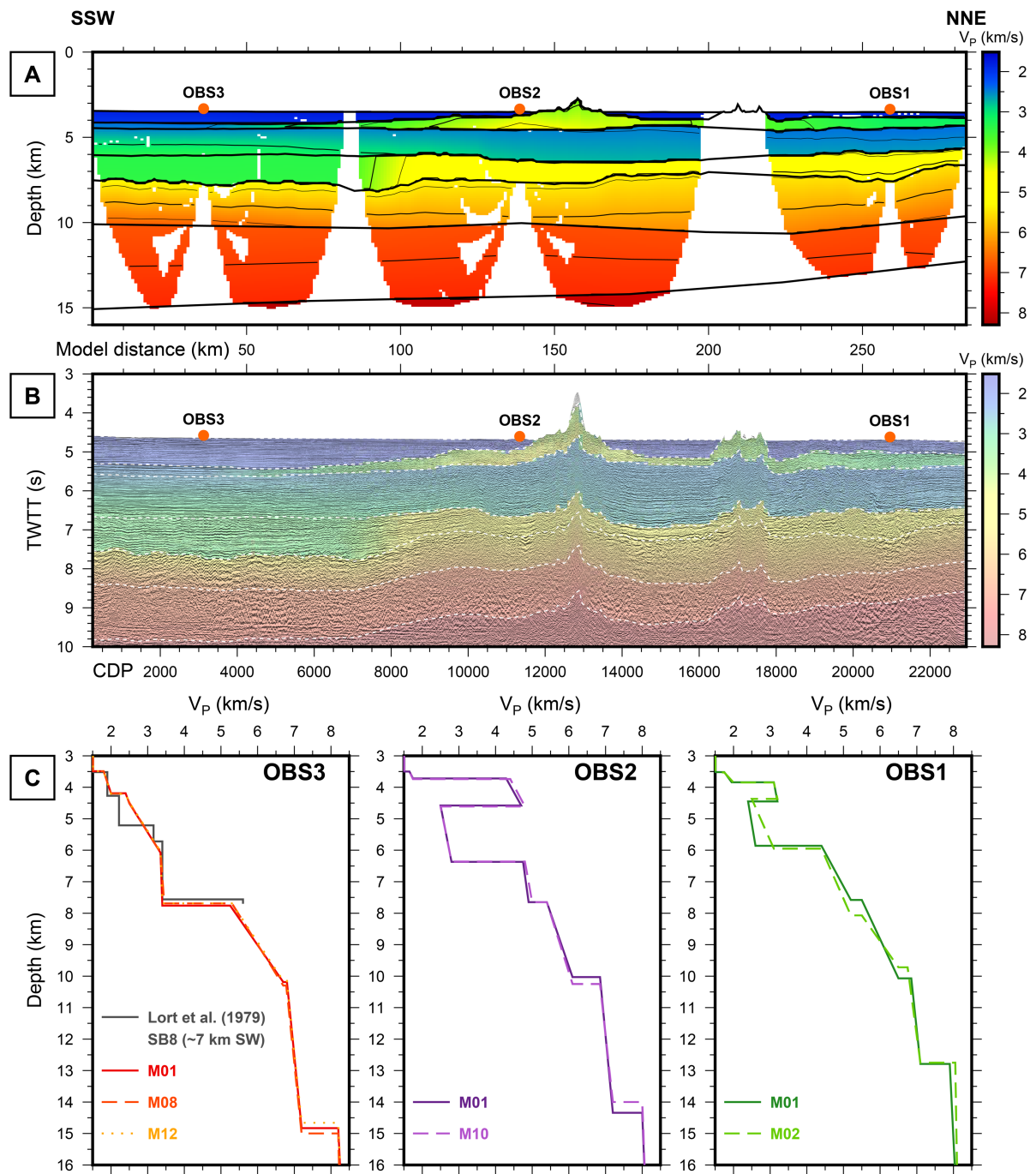
548 air gravity anomaly grid of Sandwell et al. (2014); (B) Location of the study area in the

549 western Indian Ocean. The dashed green lines indicate the inferred plate boundaries

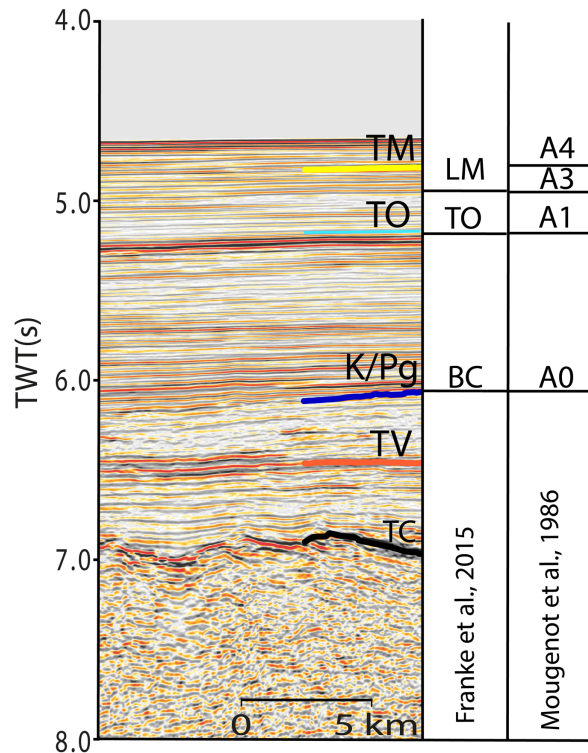
550 after Michon et al. (2022); (C) Location of the OBSs of the SISMAORE cruise (green

551 triangles) and the sonobuoys SB08 of Lort et al. (1979) and V449 of Coffin et al. (1986)

552 (blue points). The numbering of the seismic reflection profiles (M01, M08 etc...) is also  
 553 indicated. The background bathymetric grid is a compilation of several cruises after  
 554 Masquelet et al. (submitted).

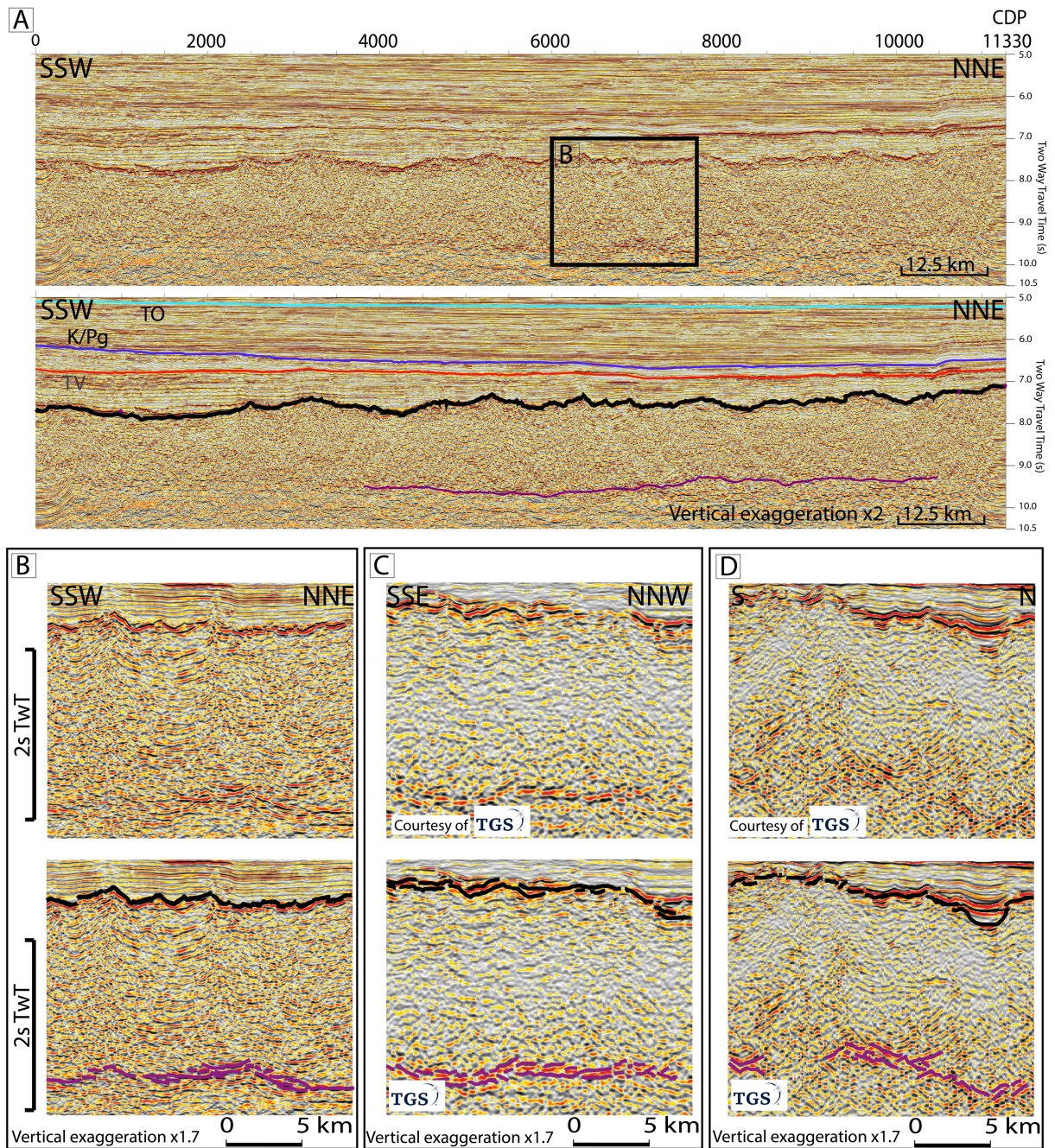


555  
 556 **Figure 2:** Results of the OBS data analysis. A: Velocity model along the M01 profile.  
 557 B: Seismic reflection data overlaid on the time-converted velocity model. C: Vertical  
 558 velocity profiles beneath each OBS for each crossing shot line.  
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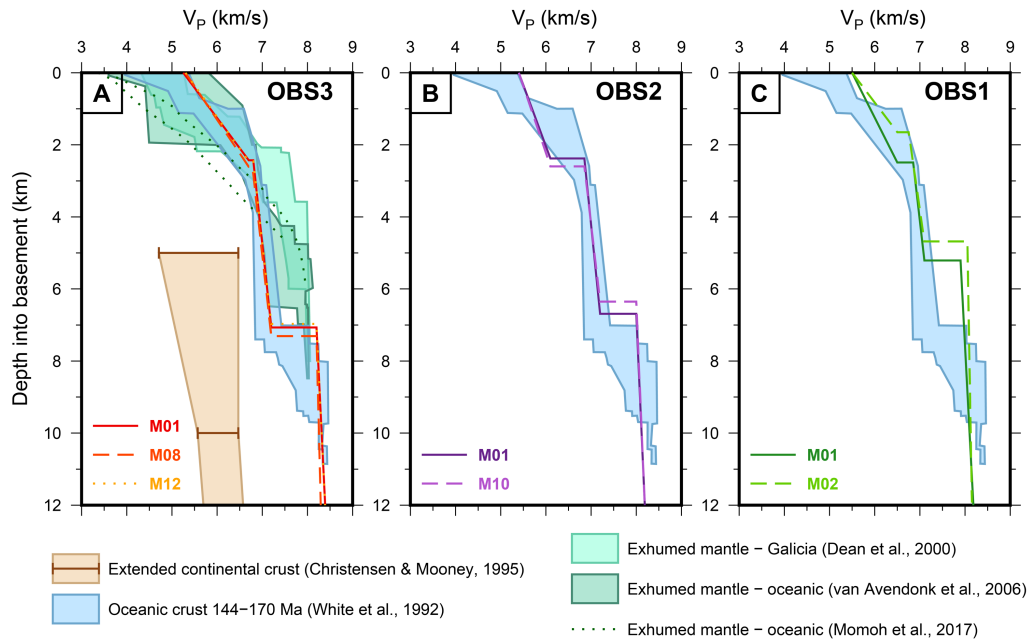
561 **Figure 3:** Seismic stratigraphy of the Comoros Basin from Masquelet et al.  
 562 (submitted). The seismic character of the interpreted key horizons is illustrated by a  
 563 960-channel seismic reflection profile; the Top Miocene (TM; 5.3 Ma), Top Oligocene  
 564 (TO; 23.5 Ma), Cretaceous-Paleogene unconformity (K/Pg; 66 Ma), Turonian  
 565 volcanism (TV; 89-93 Ma), and the top of the crust were identified in the Comoros  
 566 Basin and compared with the seismic stratigraphy of Franke et al. (2015) and  
 567 Mougnot et al. (1986): BC/A0: Cretaceous/Tertiary boundary; TO/A1: Top Oligocene;  
 568 LM/A3: Late Miocene; A4: Pliocene.



569

570 **Figure 4:** M09 960-channel seismic reflection profile east of Mayotte (A), and  
 571 interpreted profile (see location in Fig. 1). The black and purple lines correspond to the  
 572 top of the crust and the Moho, respectively. TO: Top Oligocene. K/Pg:  
 573 Cretaceous/Paleogene boundary, TV: Turonian volcanism. (B) Close-up view of part  
 574 of the M09 profile location shown in (A). (C) and (D) show other examples of typical  
 575 oceanic crust (see location of the profiles in Fig.1).

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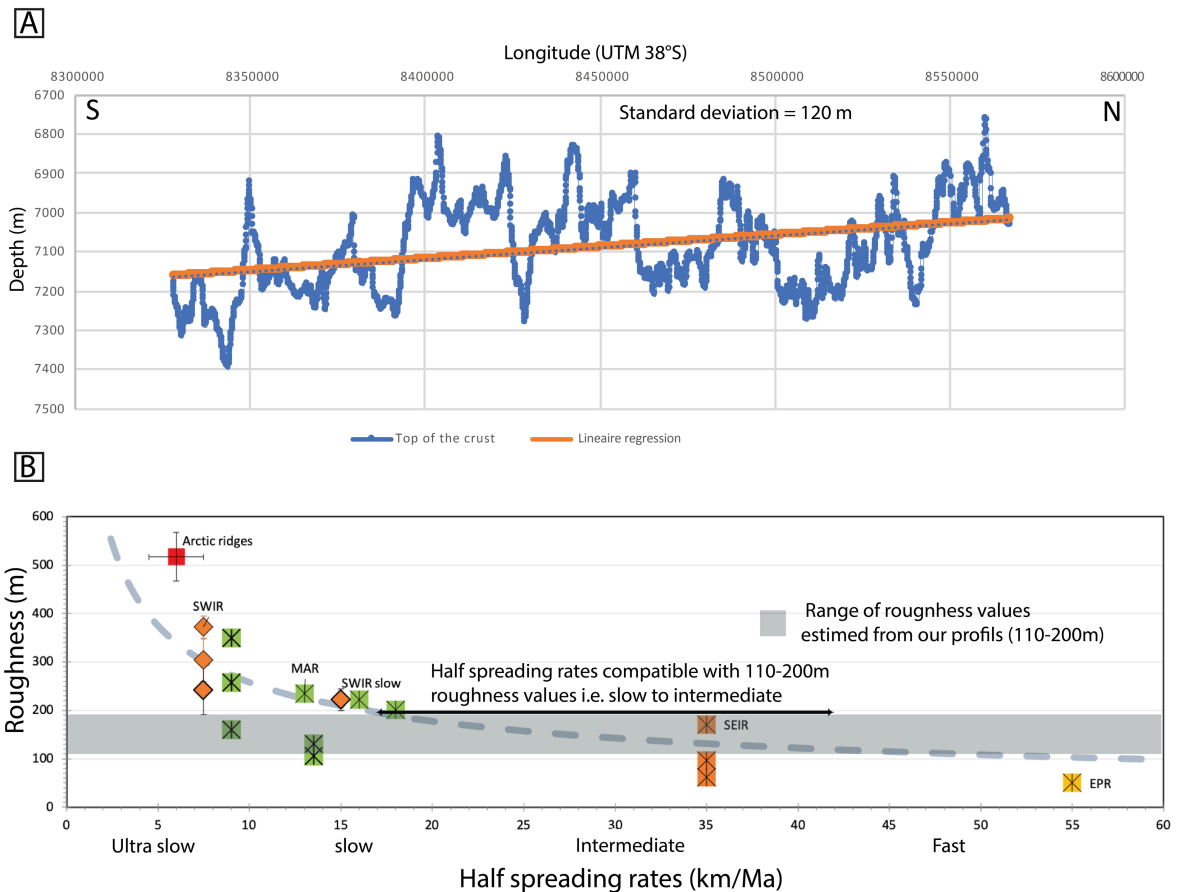


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578 **Figure 5:** Vertical velocity profiles in the basement and uppermost mantle for OBS3  
 579 (A), OBS2 (B) and OBS3 (C), compared with literature data for oceanic crust, extended  
 580 continental crust, sub-continental exhumed mantle and oceanic exhumed mantle.

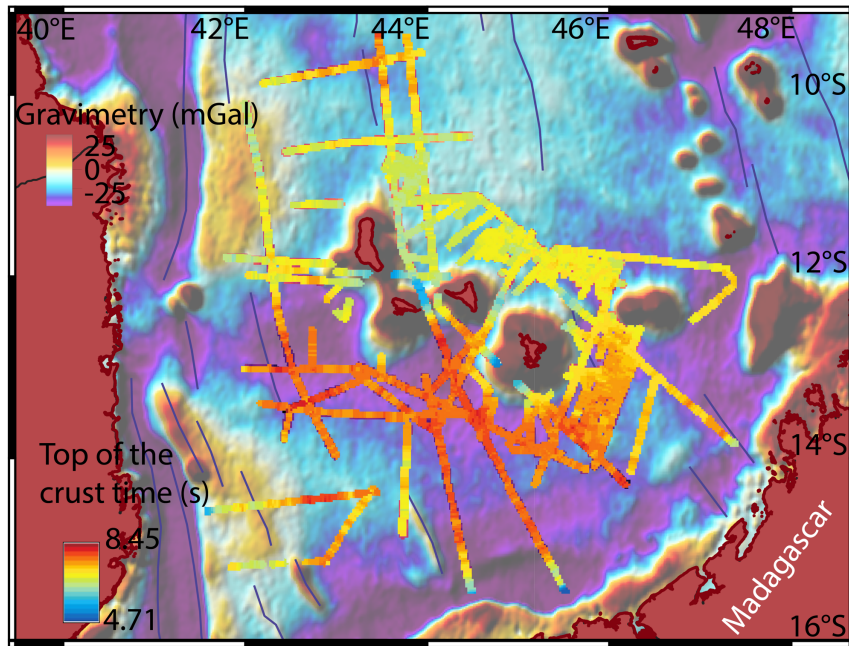
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584 **Figure 6:** Roughness estimate for a NS profile in the Comoros Basin (A; see  
 585 location in Fig. 1) and plot of the roughness versus spreading rate (B). Depth values  
 586 were obtained by converting TWTT using the velocity profile of OBS3. The standard  
 587 deviation of the detrended profile in A gives a roughness value of 120 m. Roughness  
 588 values for the Arctic Ridge are from Weigelt and Jokat (2001) and Ehlers and Jokat  
 589 (2009). Those for the southwest Indian ultraslow-spreading ridge are from Sloan et al.  
 590 (2012). Those for the slow-spreading Mid-Atlantic Ridge are from Goff (1991); Goff et  
 591 al. (1997); Minshull (1999); Neumann and Forsyth (1995). The roughness value for the  
 592 slow-spreading SWIR (before 24 Ma) is from Sauter et al. (2011). The roughness  
 593 values for the intermediate-spreading southeast Indian Ridge and fast-spreading East  
 594 Pacific Rise are from Goff et al. (1997). The dashed grey line is the power-law curve  
 595 fit of Malinverno (1991). The horizontal grey box shows the range of roughness values  
 596 measured in our study area.



597

598 **Figure 7:** Depth of the top of the crust (in seconds two-way travel time) obtained  
 599 along the seismic reflection profiles. The blue lines correspond to the fracture zones  
 600 from Davis et al. (2016). The background image is the satellite-derived free-air gravity  
 601 anomaly grid of Sandwell et al. (2014).

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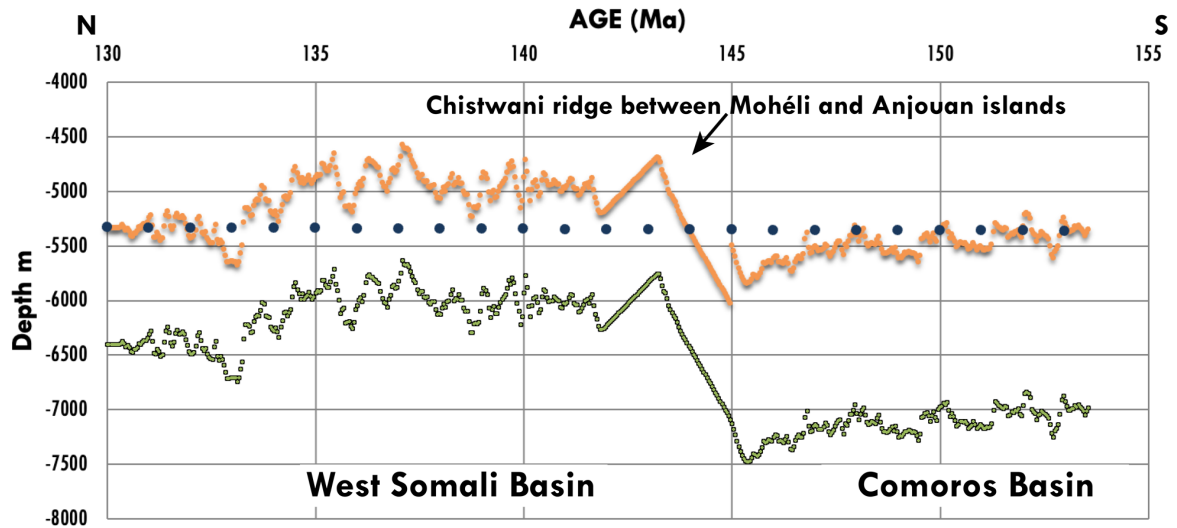
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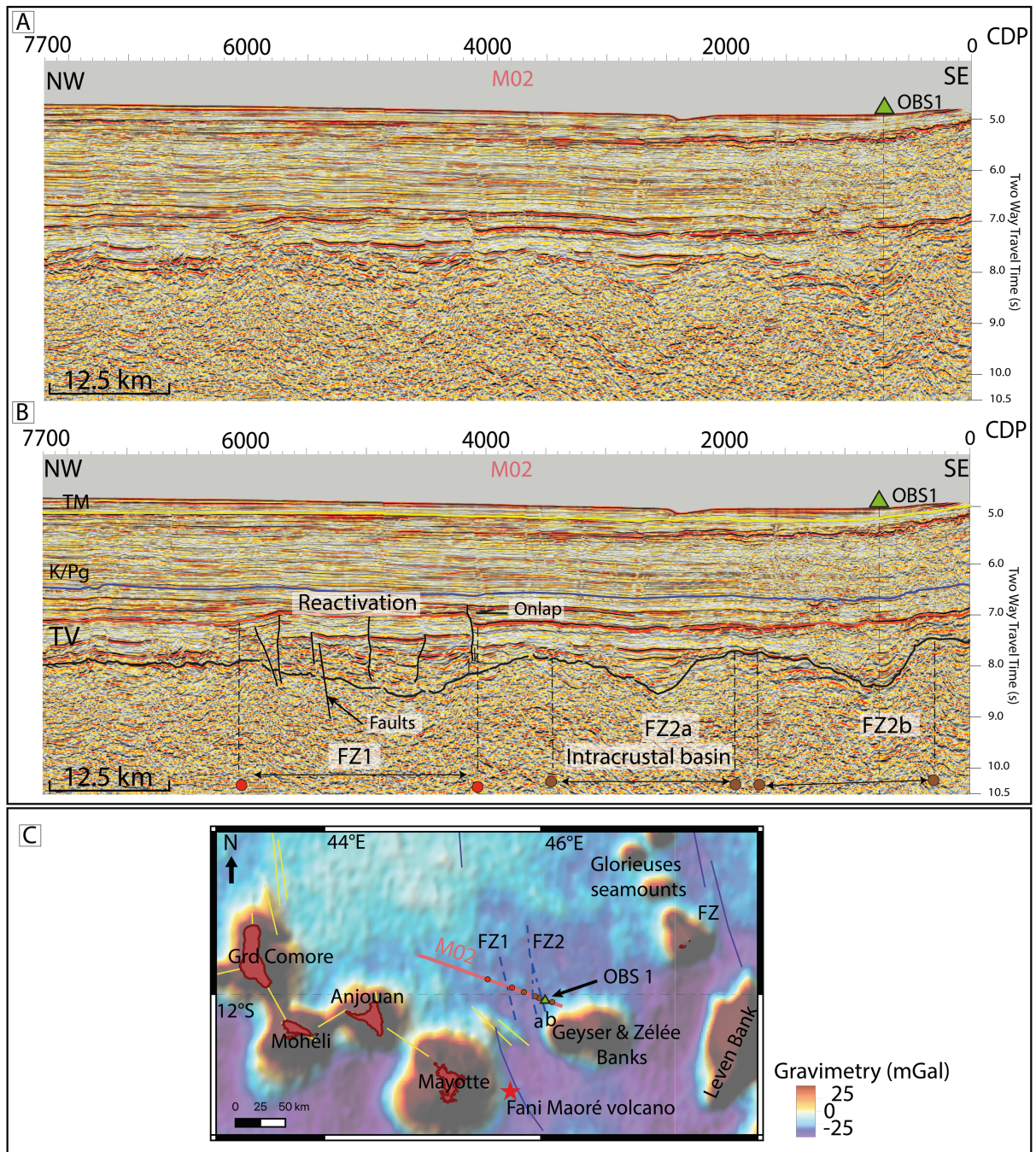
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610 **Figure 8:** Depth of the top basement obtained along a NS seismic reflection located  
 611 at 44°E longitude and passing between the islands of Anjouan and Mohéli (green dots)  
 612 (Figs 1 and 7 for location). Black dots follow the GDH1 cooling model of Stein and  
 613 Stein (1992) for a standard oceanic lithosphere. Orange dots show the depth of the  
 614 basement free from the weight of the sediments. Ages are calculated using the  
 615 distance from the extinct axis and an intermediate 25 km/Ma half spreading rate.



616

617 **Figure 9:** M02 960-channel seismic reflection profile north of Mayotte (A),

618 interpreted profile (B) and location map (C). TM: Top Miocene; K/Pg: Cretaceous-

619 Paleogene boundary; TV: Turonian Volcanism. In frame C, the dashed blue lines

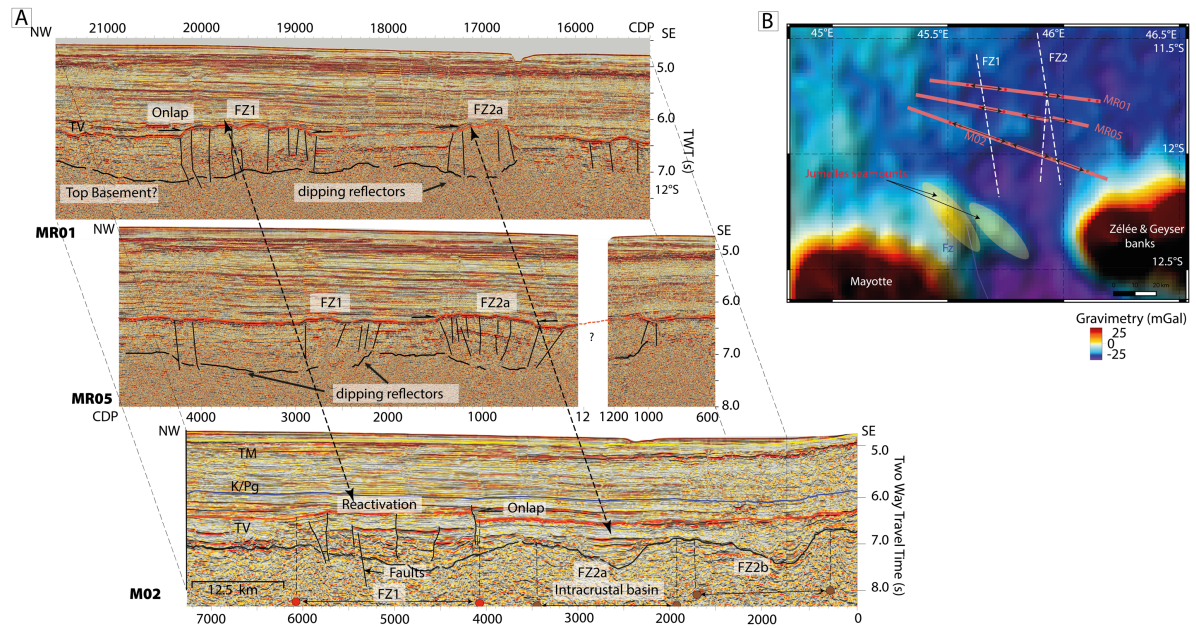
620 indicate the location of inferred fracture zones (FZ1 & FZ2) using the seismic profiles

621 M01 and M02. The background image is the satellite-derived free-air gravity anomaly

622 grid of Sandwell et al. (2014). Blue lines correspond to the fracture zones (FZ) mapped

623 by Davis et al. (2016). Yellow thick lines show the main orientations of some volcanic

624 structures.



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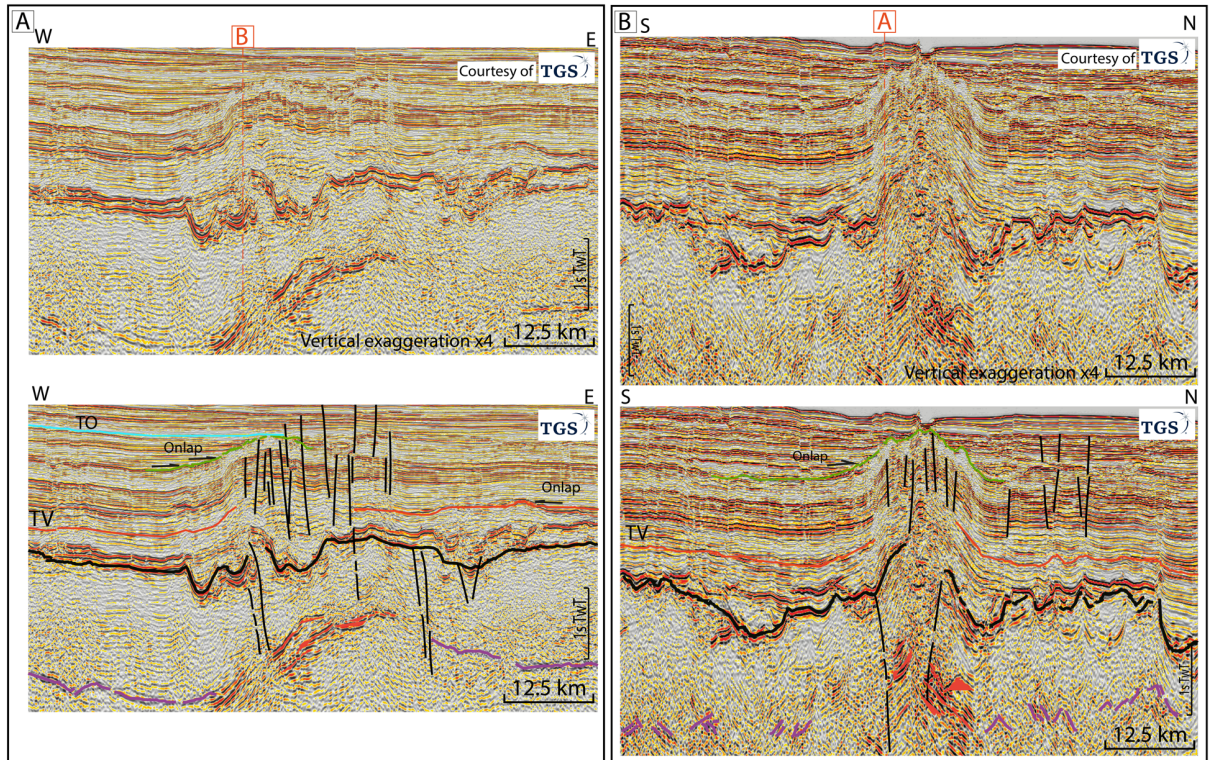
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**Figure 10:** Identification of reactivated fracture zones (Fz1, Fz2a, Fz2b) observed on three EW-trending seismic reflection profiles (A) and location map with fracture zones shown as white dashed lines (B). The dipping reflectors (black lines) in the 48-channel seismic reflection profiles (MR05 and MR01) are interpreted as the top of the oceanic crust.



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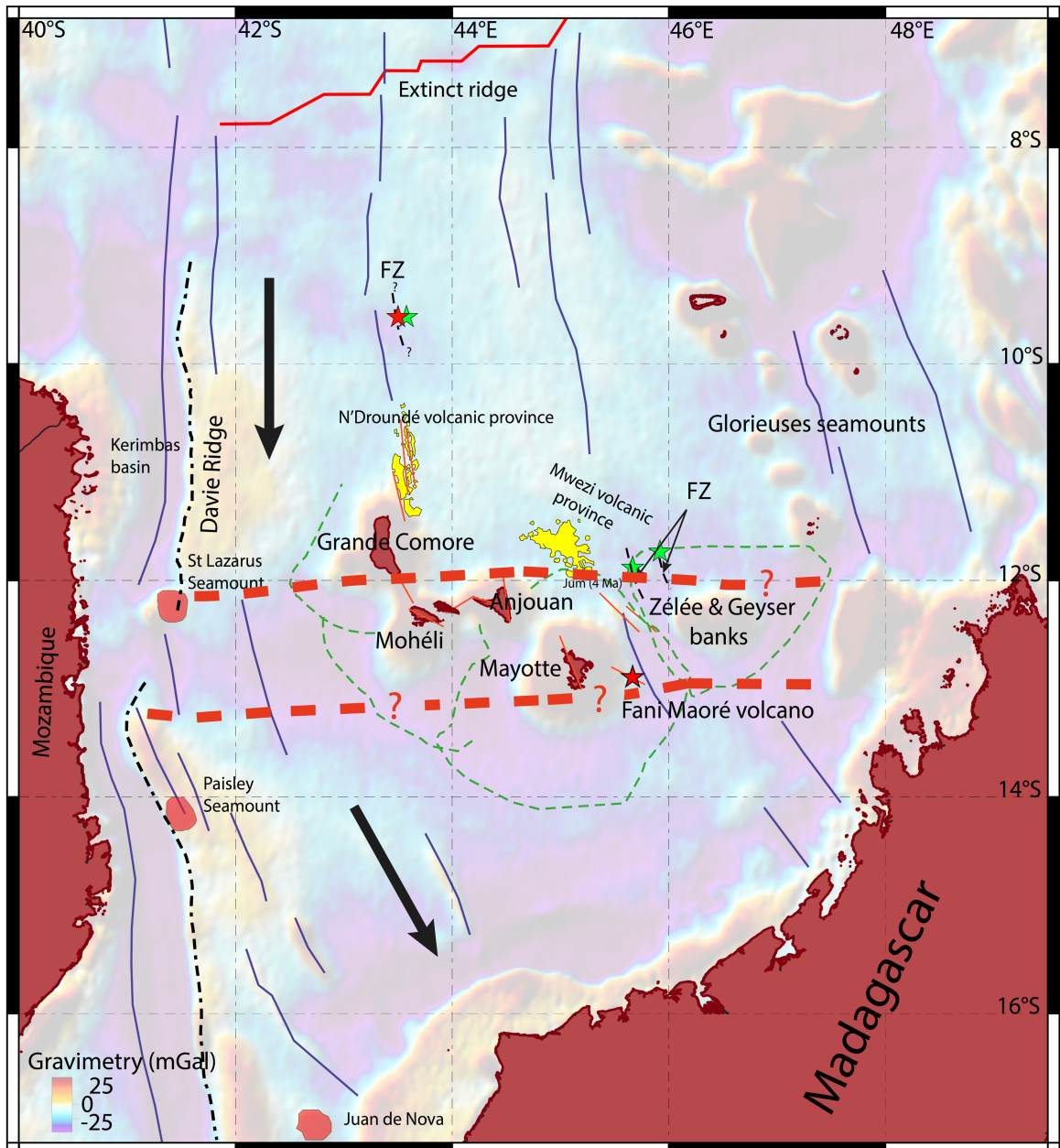
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**Figure 11:** Sections of an EW-trending seismic reflection profile (A) and an NS-trending seismic reflection profile (B) north of the Grande Comore Island (see location in Fig. 1). TO: Top Oligocene, TV: Turonian Volcanism. The green horizon shows the last sedimentary layer affected by the compressional deformation. The thick black and purple lines correspond to the top of the oceanic crust and the Moho, respectively. The thin black lines indicate faults.



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**Figure 12:** Schematic structural map of the West Somali and Comoros Basins. The black arrows show the paleo-spreading direction. The thin red line corresponds to the location of the extinct spreading ridge. The black dashed line corresponds to the location of Davie Ridge. The blue lines are the fracture zones identified by Davis et al., (2016). The dashed green lines indicate the extent of the volcanic zones around each island (Masquelet et al., submitted). The dashed red lines delimit the area where the fracture zones show a marked change in direction. The stars indicate the location of the reactivated fracture zones shown in Figs 9-11. The background image is the satellite-derived free-air gravity anomaly grid of Sandwell et al. (2014).

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