

Imaging early oceanic crust spreading in the Equatorial Atlantic Ocean: Insights from the MAGIC wide-angle experiment

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Abstract :

During the MAGIC (Margins of brAzil, Ghana, and Ivory Coast) experiment, five combined wide-angle, and reflection seismic profiles were acquired in the Pará-Maranhão/Barreirinhas/Ceará basins northern Brazil. This is a pull-apart passive margin, with two strike-slip borders. The equipment deployed includes 143 sea-bottom seismometers (OBS), a 4.5-km seismic streamer, and a 7587-in3 airgun array. In this paper, we focus on the distal parts of three profiles, and one entire transverse NW-SE profile, located on the presumed Cretaceous oceanic crust.

Forward modelling of these wide-angle data sets reveals an E-W lateral evolution of the oceanic crust spreading initiation with: 1) just after the so-called intermediate domain, 60 km-wide domain that consists of a 5-km-thick crystalline crust. The basement presents two layers characterized by high acoustic velocity. This domain is bounded to the NW by a NW-SE volcanic line (Volcano Alignment), and 2) a 5-km-thick oceanic crust consisting of two layers characterized by “normal velocities” spanning between the two main fracture zones that fringe the Pará-Maranhão-Barreirinhas-Ceará segment. Despite a similar thickness, these two sub-domains present different velocity distribution in their two layers. They are both overlain by 5.5 km of sedimentary deposits. Forward wide-angle modelling confirms that the seafloor spreading process was progressive, with firstly the emplacement of a proto-oceanic crust, and then a thin oceanic crust. The “proto-oceanic” crust presents a similar seismic velocity with the intermediate domain interpreted as exhumed lower continental crust except for the lower part where the intruded lower crust

gives place to a very sharp Moho at the base of the proto-oceanic domain. By contrast, the thin oceanic crust domain has a lower velocity structure in its upper layer, that is interpreted as basalt and is absent in the proto-oceanic crust. This eastward evolution, as also observed in the Provençal Basin, and the Santos Basin, suggests the involvement of the lower continental crust in the first steps of seafloor spreading.

Highlights

► The Pará-Maranhão-Barreirinhas-Ceará passive margin in the Equatorial Atlantic segment presents a strong E-W-segmentation with: ► A transition to oceanic crust very complex, with the presence of an about 60 km-wide band of proto-oceanic crust before a “more typical” but still thin oceanic crust eastward. ► The passage between an intermediate domain and a typical oceanic crust is not abrupt in terms of composition but takes place in stages. ► This eastward evolution suggests the involvement of lower continental crust and/or continental mantle lithosphere erosion over time until the inception of « typical » oceanic crust.

Keywords : Equatorial Atlantic Ocean, Brazil, Pará-Maranhão-Barreirinhas-Ceará basins, Wide-angle seismic, Oceanic crust, Intermediate domain

54

1 Introduction

55 The equatorial Brazilian margin, which is the focus of the MAGIC (Margins of
56 brAzil, Ghana, and Ivory Coast) research experiment, represents a unique natural
57 laboratory for addressing fundamental questions on strike-slip margins. The MAGIC
58 experiment is a joint project of the Department of Marine Geosciences (IFREMER:
59 Institut Français de Recherche pour l'Exploitation de la MER, France), the Laboratory
60 of « Oceanic Geosciences » (IUEM: Institut Universitaire et Européen de la Mer,
61 France), the Faculdade de Ciências da Universidade de Lisboa (IDL, Portugal), the
62 Universidade de Brasilia (Brazil), and PETROBRAS (Brazil). The main goals of the
63 MAGIC experiment are (i) to investigate the deep structure of the Pará-Maranhão-
64 Barreirinhas-Ceará basins, N-NE Brazil, (ii) to characterize the segmentation and the
65 nature of the crust in the different domains of this passive margin, between the

66 unthinned continental crust and the true oceanic crust, and (iii) to understand the
67 fundamental processes that lead to the thinning, and finally to the breakup of the
68 continental crust in the specific context of a pull-apart system with two strike-slip
69 borders. This paper presents the results of P-wave velocity modelling on coincident
70 near-vertical reflection multi-channel seismic (MCS) and, wide-angle seismic data
71 sets in an area that is supposed to be of oceanic nature.

72

73 **2 Geological setting and previous work**

74 The Pará-Maranhão-Barreirinhas-Ceará margin (Figure 1) is located in the southward
75 second 600-800-km-wide segment of the Equatorial Atlantic Ocean, between the São
76 Paulo Fracture Zone to the north, and the Chain Fracture Zone to the south. The
77 Equatorial Atlantic Ocean can itself be interpreted as a "transfer zone" between two
78 main oceans resulting from two ruptures during two global geodynamic revolutions
79 (Moulin & Aslanian, 2010; Leroux *et al.*, 2018): the opening of the Central Atlantic
80 Ocean that started by Sinemurian time (Sahabi *et al.*, 2004), and the opening of the
81 South Atlantic Ocean in Hauterivian time (Rabinowitz & LaBrecque 1979; Austin &
82 Uchupi 1982; Curie 1984; Moulin *et al.* 2010a), which started about 60 My later and
83 shifted about 30° to the east respect to the Central Atlantic.

84 Despite the lack of magnetic lineaments due to its position close to the equator, the
85 beginning of seafloor spreading on this portion of the ocean is dated to Cretaceous
86 time, separating Africa from South America. Besides the exact date still being debated
87 between Aptian (112 Ma) (Blarez, 1986; Mascle and Blarez, 1987; Gouyet, 1988;
88 Azevedo, 1991; Matos, 1992), and Upper Albian (100 Ma) (Oliveira Marinho, 1985;
89 Gouyet, 1988; Basile *et al.*, 2005; Torsvik *et al.*, 2009; Heine *et al.*, 2013; Granot &
90 Dymant, 2015), the major unknown on this segment is the formation and evolution of
91 the conjugate system (Pará-Maranhão-Barreirinhas-Ceará and the Deep Ivory Basin-
92 Ghana Platform). This fact is primarily because, until the MAGIC experiment, all data
93 were collected on the African margin (Mascle *et al.*, 1988; Basile *et al.*, 1993; Mascle
94 *et al.*, 1995; Sage, 1994; Pierce *et al.*, 1996; Edwards *et al.*, 1997; Sage *et al.*, 1997;
95 Antobreh *et al.*, 2009). Studies based on seismic refraction data published in the
96 1990s only concern the southern strike-slip boundary of the African system, and not
97 the entire Pará-Maranhão-Barreirinhas-Ceará/Ghana-Ivory coast pull-apart system.

98 These previous results based on refraction and IODP (Integrated Oceanic Drilling

99 Program data provided an image of the deep structure of the southern limit of the
100 African side of this pull-apart system, where continental crust thickness never exceeds
101 20 km (Figure 2). This layered continental crust thins abruptly southwards, across the
102 Ghana-Ivory Coast (GIC) Ridge, reaching directly the younger oceanic crust from the
103 next segment, south of the Romanche Fracture Zone. In the divergent part of the pull-
104 apart system, the initial position of the South American plate respect to the African
105 one, before any horizontal movement, the position and the age of the oldest oceanic
106 crust are still a matter of debate (Torsvik et al., 2009; Moulin et al., 2010; Aslanian
107 and Moulin, 2010; Heine et al., 2013; Muller et al., 2016). Whilst the Landward Limit
108 Oceanic Crust (LaLOC) defined by Heine et al. (2013) is supposed to represent the
109 first inset of oceanic crust, neither of the two flow-lines parallel wide-angle profiles
110 from the EQUAREF experiment shows clear evidence of oceanic crust (Figure 2).
111 Whilst on the EQUAREF-1 profile, Pierce *et al.* (1996) concluded that the extreme
112 western end of the profile exhibits a 10-11-km-thick crust with a velocity typical of
113 thinned continental crust (3.2 to 3.6 km/s and 6.8 km/s), Sage (1994) proposed the
114 presence of oceanic crust only westwards of IODP site 961, at the westernmost end of
115 the profile, between their OBS-3 and OBS-4 on the EQUAREF-7 profile. This
116 position is 50 to 100 km westwards of the LaLOC (Figure 2).
117 On the Brazilian side, no refraction data were available before the MAGIC
118 experiment, but this margin benefits from a good coverage of industrial ultra-deep
119 high-quality seismic lines. The interpretation of these industrial profiles suggests a
120 huge discrepancy in the position of the presumed oceanic crust (Figure 1b), and the
121 LaLOC of Heine *et al.* (2013). Aslanian et al., 2021, who presented two another
122 MAGIC P-wave velocities modelling, reveal distinct structural domains from onshore
123 Brazil towards the Atlantic Ocean characterized by variations of the crustal
124 thicknesses and velocities: (1) an unthinned continental crust below the São Luís
125 Craton, where the crust is 33 km thick, (2) a 60 km wide necking domain; (3)
126 offshore, east of the continental slope, a 10km-thick deep sedimentary basin (basin I
127 and II); (4) eastwards, the limit of the previous domain is marked by NW-SE aligned
128 volcanoes and the inception of the oceanic domain.

129

130 **3. Data acquisition, quality and processing**

131 During the MAGIC experiment, five coincident Multi-Channel Seismic (MCS)

132 reflection and wide-angle seismic profiles, sub-bottom high-resolution (CHIRP)
133 profiles, calypso cores, and bathymetry data were collected on the Pará-Maranhão-
134 Barreirinhas-Ceará basins (Figure 1). The seismic reflection data were acquired using
135 a 4.5 km, 360-channel digital streamer, and a tuned airgun array of 7587 in³, towed at
136 a depth of 18–25 m. A total of 143 ocean bottom seismometer/hydrophones
137 (OBS/OBH) from Ifremer (Auffret *et al.*, 2004), and the University of Brest were
138 deployed, spaced every 7 nautical miles (~13 km). Three seismic profiles,
139 perpendicular to the margin, were extended onshore by land seismic stations (LSS)
140 (Figure 1) (see Aslanian *et al.*, 2021). The seismic source consisted of a tuned array of
141 18 airguns ranging from 250 in³ G-guns to 9 L Bolt airguns, with main frequencies
142 centred around 10-15 Hz. Shots occurred at a constant time interval of 60 s (firing
143 rate), resulting in intervals of about 150 m between shots. A total of 12382 shots
144 (profile MC-1: 3032, MC-2: 1741, MC-3: 2834, MC-4: 1145, and MC-5: 3801) were
145 fired by the air gun array, recorded simultaneously on MCS reflection and wide-angle
146 seismic profiles.

147 This paper focuses on profile MC1 and the distal parts of crossing profiles MC2, MC3
148 and MC4. The MC1 profile (Figure 1) is about 360 km long and is located entirely on
149 the presumed oceanic-floored basin, except on its south-easternmost part. It spans the
150 entire domain between the two main fracture zones that fringe the Pará-Maranhão-
151 Barreirinhas-Ceará segment: the São Paulo Double Fracture Zone (FZ) to the north
152 and the Romanche FZ to the south. Twenty-five OBS were deployed along this profile
153 (Figure 1). Shots were acquired from MC1OBS25 until MC1OBS02 where the water
154 depth shallows to less than 100 m. MC1OBS01 is hence not corrected from its
155 deployment position, since the water arrival does not provide sufficient constraints to
156 be able to relocate it.

157 Pre-processing of the OBS data included calculation of the clock-drift corrections to
158 adjust the clock in each instrument to the GPS base time. Instrument locations were
159 corrected for drift from the deployment position during their descent to the seafloor
160 using the direct wave water arrival. The drift of all instruments did not exceed 200 m.
161 All the instruments were recovered and provided useful data on all four channels.
162 Data quality along the profile was generally highly satisfactory with clear arrivals to
163 offsets over 100 km between the ship and the seafloor instrument. Picking of the onset
164 of first and secondary arrivals was performed without filtering where possible (mostly
165 between offsets of 0 and 40 km). Further processing of the data to facilitate picking at

166 larger offsets included deconvolution, application of a bandpass-filtering (1-6-48-64
167 Hz), and trace normalization.

168 All the seafloor instruments provide equally good quality of data and seismic record
169 sections in the presumed oceanic basin are remarkably similar, suggesting that the
170 crustal structure has a minor lateral variation (Figure 3). OBS 9, 11, 18 and 24 were
171 selected to show the quality of the data used to constrain the forward modelling
172 (Figures 4 to 7). Useful arrivals could be picked up to a 40-160 km offset, including
173 arrivals reflected from the Moho (PmP) and refracted in the shallow mantle (Pn)
174 (Figure 4).

175 Processing of the multichannel seismic data was performed using the Geovecteur
176 processing package. The processing sequence of the reflection seismic data was
177 composed of geometry (including streamer feathering), CMP binning at 12.5 m
178 intervals and sorting, bandpass filter (2–16–64–96 Hz), re-sampling from 2 to 4 ms.
179 After velocity analysis, true amplitude recovery was applied plus normal move-out,
180 multiple attenuation, time-variant bandpass filter (from 2–16–48–64 at sea bottom to
181 2–16–32–48 Hz 3 s below), inside and outside mute, stack and post-stack time-variant
182 bandpass filter and Kirchhoff time migration.

183

184 **4. Forward Modelling**

185 The data were modelled by using an iterative procedure of two-dimensional forward
186 ray-tracing followed by a damped least-squares travel-time inversion with the
187 RAYINVR software (Zelt & Smith, 1992). The wide-angle modelling proceeds in a
188 top-to-down strategy of arrival time fitting of the reflected and refracted phases
189 identified in the sedimentary section. For each sedimentary sequence, we correlated
190 the two-way travel times of its base, from the MCS section, with the arrival times of
191 the reflected and refracted phases identified in the OBS data.

192 An iterative procedure of velocity and depth adjustment, with control of the depth-twtt
193 (two way travel-time) conversion against MCS data, was then applied. This procedure
194 was applied to all sedimentary layers down to the basement. For the basement, we
195 only used arrival times from the OBS and LSS data set. The iterative procedure stops
196 when an appropriate fit of the arrival times from both OBS and MCS data is reached,
197 i.e., the stopping criteria is a normalized chi-square of 1.

198 To constrain velocity gradients and velocity contrasts at the interfaces of the model,

199 we performed amplitude modelling (Zelt & Ellis, 1988), which is a trial and error
200 procedure. Amplitude modelling is based on the variation of the angle of incidence of
201 the transmitted and reflected seismic energy across boundaries; therefore, synthetic
202 record sections should reproduce the observed offset variations of amplitude.
203 Furthermore, for each incidence angle, the fractions of transmitted and reflected P-
204 wave energy (relative to the total incident energy) depend on the impedance contrast
205 across interfaces; thus, synthetic record sections should reproduce the relative
206 amplitudes amongst all the arrivals. Another important constraint provided by the
207 amplitude modelling is the thickness and velocity gradient in each layer.

208 The final velocity model of the MC1 profile images the sedimentary and crustal layers
209 to a depth of around 20 km (Figure 8). The model is composed of five sedimentary
210 layers (S1 to S5), 5.5 km thick in total, with top and bottom seismic velocities of 1.8
211 and 2.2 km/s (S1), 2.4 and 2.75 km/s (S2), 2.9 and 3.0 km/s (S3), 3.75 km/s (S4), and
212 4.15 and 4.2 km/s (S5). Near the intersection with the Romanche FZ (at a distance of
213 280 km from the model), the velocity increases in layer S4 (4.5 to 4.75 km/s) and
214 layer S5 (5.0 to 5.5 km/s). The velocity increases gradually inside the sedimentary
215 column and no velocity inversion is identified in this profile, in contrast to what is
216 observed, westwards, in the deep basin of the Pará-Maranhão/Barreirinhas/Ceará
217 margin (Aslanian *et al.*, 2021). The model comprises a two-layered (upper and lower)
218 crust above a lithospheric mantle layer (Figure 8). The upper crustal layer was set to
219 less than 1.5 km thickness, with top-bottom velocities of 5.7-5.9 km/s (on the
220 northwestern part), smoothly increasing from 5.9-6.0 km/s (on the southeastern part).
221 The presence of this thin crust was added to the model to correctly predict time
222 arrivals of the lower crust and to be coherent with MC3 P-wave velocities model.
223 Lower crust velocities range from 6.2-6.8 km/s in the northwestern part of the profile
224 to 6.2-7.2 km/s in the southeastern part. The thickness of the lower crust is ~4 km. All
225 OBS seismic record sections show clear crustal refracted and reflected arrivals, from
226 the lower crust (Pg2) and the crust-mantle boundary (Pm1P) (Figure 7). Finally, the
227 lithospheric mantle velocity, constrained by Pn arrivals, varies from 8.2 to 8.7 km/s at
228 45 km depth. This high velocity in the upper mantle (8.7 km/s) is essential to be able
229 to explain the high apparent velocity and strong amplitude of the Pn, visible up to
230 100-170 km offset on several OBS (Figures 4, 5 & 6).

231

232 **5. Evaluation of the MC1 model**

233 We have digitized a total of 35629 events and interpreted their respective phases.
234 Travel-time uncertainty was estimated on the MC1OBS record and generally fixed at
235 0.100 s. The model fits the travel-time and phase of 34656 events, i.e. 97 % of all
236 picks, with a global RMS travel-time residual of 0.060 s. Given the individual
237 uncertainty of our events, the model results in a normalized chi-square of 0.356 (Table
238 1). Individually, the MC1 model explains the MC1OBSs with a chi-square lower or
239 equal to 0.848 and RMS lower or equal to 0.095 s (Table 2). During the forward
240 modelling, the most difficult element to constrain was the velocity of layers S3 and S4
241 due to the reduced number of picks (Table 1). Nevertheless, the RMS value for each
242 phase varies from 0.017 s (Pw) to 0.097 s (Ps4), and the normalized chi-square varies
243 from ~0.1 (short offset range of observation Ps2P), and ~0.875 (Ps4) (Table 1). Most
244 interface nodes in the MC1 model produce a hit count larger than 200 rays (Figure
245 10b). The number of rays that constrain each velocity is always above 100 (Figure
246 10b). The resolution of the interface depth and velocity nodes is build from the
247 diagonal terms of the inversion kernel, and is a measurement of the spatial averaging
248 of the true earth structure by a linear combination of model parameters (Zelt, 1999).
249 Typically, resolution matrix diagonals greater than 0.5–0.7 are said to indicate
250 reasonably well-resolved model parameters (e.g. Lutter & Nowack 1990). The
251 resolution is generally very good, above 0.9 (Figure 10d). The Spread-Point Function
252 (SPF) suggests that smearing is low (Figure 10c). The SPF is build from the off-
253 diagonal terms of the inversion kernel, and suggests that smearing is low (Figure 10c).
254 In summary, the MC1 model is well constrained from arrival times, except for the
255 velocity-gradient of the second crustal layer, which was set by amplitude modelling,
256 considering the typical features of seismic wave propagation in a layered medium,
257 namely the fit of cut-off and critical distances, as well as triplication of Pg2, Pm1P
258 and Pn (Figure 4).

259

260 We have built a 2-D gravity model consisting of 170 homogeneous density blocks, by
261 converting seismic velocity to density according to Ludwig *et al.* (1970). The density
262 conversion of our velocity model can predict the main trend of the gravity anomaly,
263 with a density of 3430-3460 kg/m³ in the deep lithospheric mantle to remove the
264 regional trend. The density ranges from 2200 to 2500 kg/m³ in the sedimentary

265 basins, 2550 to 3000 kg/m³ in the igneous crust and 3430 kg/m³ to 3460 kg/m³ in the
266 mantle (Figure 9a). Where the MC1 model is not covered by seismic rays, the largest
267 differences among observed (red dotted lines in Figure 9b) and calculated gravity
268 anomalies (black line in Figure 9) occur close to the continental slope with a
269 maximum difference reaching 50 mGal. Generally, the observed and calculated
270 gravity anomalies are similar (Figure 9b).

271 It worth to note that the thickness of the unthinned continental crust, set to ~21 km, on
272 the southeastern end of the model, is based on gravity modelling only (Figure 9) and
273 the information provided by the MC5 modelling.

274 **6. Comparison to reflection seismic data**

275 The MAGIC1 MCS profile starts in the middle of the continental slope with ~3 s twtt
276 sedimentary layers (Figure 11). These sediment layers thicken to 3.5 s twtt into the
277 distal part (Northwards). The four shallowest layers are well individualised and
278 stratified, but the absence of well-imaged reflectors in the fifth one makes it difficult
279 to define its top. Nevertheless, the top of the acoustic basement and reflections from
280 the Moho are well-marked by a strong reflector at 8 s, and a discontinuous one at 9 s
281 twtt, between 160 and 280 km model distance. This strong and discontinuous reflector
282 disappears between 0 and 160 km model distance. It is important to note that the base
283 of the first crustal layer does not correspond to a seismic reflector nor a change in the
284 seismic facies, as noticed on the MAGIC3 profile (Aslanian et al., 2021) but needed
285 by wide-angle modelling. In the northern part of the profile, crustal arrivals (top of
286 basement and Moho) are interpreted from the seismic wide-angle data only.

287 To verify the accuracy of the wide-angle velocity model, we also performed a Pre-
288 Stack Depth Migration (PSDM) of the MCS data by using the MC1 final P-wave
289 velocity model (Figure 12). The pre-processing sequence is identical to the MCS data
290 time processing and includes geometry, wave-equation multiple attenuation, shot-
291 gather predictive deconvolution, time-variant bandpass filter, and random multiple
292 attenuation. Hence, if the velocity model used for the depth migration is close to the
293 true medium velocity, all common offset migrated panels map the recorded seismic
294 events to the same reflector depth. If the velocity is under/over estimated, the residual
295 move-out from near to far offset at selected common-mid points along the MCS
296 profile can be estimated through semblance analysis and translated into an interval
297 velocity correction (Liu & Bleistein, 1995). This provides discussion as to the real

298 geometry of the main features imaged by the MC1 profile. The PSDM of MC1
299 strongly improves the seismic imaging of the basement and sediments (especially in
300 the centre of the profile and at its eastern extremity) (Figure 12). The PSDM MC1
301 profile shows a rather continuous seismic character with strong well-stratified and
302 high-frequency reflectors and confirms the interpretation of the crustal structure
303 (basement and Moho) (Figure 12).

304 Our interpretation of the ION-GXT lines (courtesy of Petrobras) provides additional
305 information. In the oceanic domain, the ION-GXT 7000 line equivalent to MAGIC3
306 (Figure 1b) shows two sets of very strong deep reflectors in the typical oceanic
307 domain at around 9 s twtt (around 12 km depth), 4 km below what appears to be the
308 top of the basement, and the second set of deep reflectors at 11.5 s twtt, around a
309 depth of 24 km below sea level, rising slowly oceanwards to 18 km at the end of the
310 profile. Nevertheless, these deepest reflectors at 11.5 s twtt are seen neither on MCS
311 MC1 profile nor on the crossing, perpendicular ION-GXT profiles, and their
312 significance (out-of-plane, artefact, anisotropy or unidirectional reflectors?) and
313 interpretation (intra oceanic, Moho or intra-mantellic reflectors?) need to be explored.

314 Deep reflectors in the oceanic crust are mostly interpreted as the manifestation of
315 serpentinization along faults (Carton *et al.*, 2014) as the result of magma emplacement
316 forming sills in the lower crust (Canales *et al.*, 2009) or frozen melt bodies
317 representing relicts of a paleo melt channel system (Nedimovic *et al.*, 2005; Sauter *et al.*,
318 2016), a fossil melt-rich crust-mantle transition zone (Moho Transition Zone -
319 MTZ) (Sauter *et al.*, 2016), or large shear zones that tapped a connection to deep
320 lithospheric melt channels leading to magmatic intrusions within the crust and
321 emplacement of post-seafloor-spreading volcanic edifices within deformation
322 corridors (Sauter *et al.*, 2018). Wide-angle data are therefore crucial for the
323 interpretation of deep crustal/mantellic reflectors on the MCS section. The modelled
324 Moho, at 12.5 km depth, of the wide-angle seismic model of MC3 (Aslanian *et al.*,
325 2021) fits well with the first set of deep reflectors (Figure 13). At the base of the
326 Mantle 1 layer, in the deep Basin I and II, the MC3 wide-angle seismic data recorded
327 a very deep intra-mantle reflection (Pm2P) at a depth of 25 km. This reflection is
328 associated with a very small step in the P-wave of 0.05 to 0.1 km/s. The flow line-
329 oriented deep reflectors lying at <17 km, visible on the IONGXT7000 profile, may
330 correspond to the prolongation of this intra-mantle reflector. P-wave velocities within
331 this deepest Mantle 2 layer are 8.2-8.4 km/s from the continental domain to the

332 oceanic domain.

333 **7. 1D-Vz profiles and crustal nature**

334

335 The interpretation of E-W wide-angle MAGIC MC3 profile has shown the presence
336 of a deep basin interpreted as exhumed lower continental crust, with two different
337 sub-basins I & II, followed by two different oceanic domains A & B (Figure 1;
338 Aslanian *et al.*, 2021). The NW-SE MC1 profile is inside the Oceanic A domain,
339 parallel to the external limit of the deep basin I highlighted in the north by a series of
340 aligned volcanoes, and crosses the three MAGIC MC2, MC3 & MC4 profiles. To
341 characterize the P-wave seismic velocity variations along this domain, 1-D velocity-
342 depth (Vz) profiles were extracted from the velocity models at 10 km intervals
343 (Figures 14 & 15). Figure 14 reveals a strong segmentation and abrupt passage at the
344 two limits of the oceanic A domain (purple lines) respect to the intermediate domain
345 (yellow and green lines) and to oceanic B domain (red lines). In oceanic A domain,
346 the deepest crustal layer present in the deep Basin I (yellow lines in Figure 14) is
347 absent whilst the Moho is characterized by a larger jump in velocity of about 1 km/s,
348 at the same depth: the transition between the upper mantle and the crust is thus
349 sharper. The 1D-Vz profiles of MC1 (Figure 15b) also homogeneously present a less
350 than 1-km-thick upper crustal layer with a velocity of 5.8-6.0 km/s, which is not
351 observed on the other profiles, except for MC4 (Figure 14a). Figure 15a presents the
352 1D-Vz profiles at the crossing points of MAGIC profiles. Despite an overall similarity
353 in gradient and steep velocity variation, some discrepancies can be observed. While
354 the MC1 profile, which samples a single homogeneous domain along the profile
355 clearly shows a thin top layer at the top of the basement, this layer is not always
356 identified on the MAGIC E-W profiles, probably due to the smaller sampled portion
357 of this domain by these profiles. The increase in thickness of the upper crustal layer (2
358 km instead of 1 km) on the MC4 profile may thus explain its identification and may
359 be related to the vicinity of the Volcano Alignment to the west (Figure 1).

360 The second steep change (Oceanic B – red profiles in Figure 14) is imaged by the
361 MC3 and MC4 profiles, with a very strong lateral variation in the upper basement
362 layer where velocity decreases towards the east to 4.8-5.1 km/s. On the one hand, the
363 oceanic B domain has 1D-Vz profiles that match, even at a lower thickness, the shape,
364 gradient and thickness ratio of the 1D-Vz profiles of 'normal' oceanic crust compiled

365 by White et al. (1992) and Christeson et al., 2019. In contrast, in the oceanic A
366 domain, the 1D-Vz profiles of the basement do not match the profiles of either
367 continental or typical oceanic crust (Figures 14 and 15b). Note that crustal
368 architecture has similar velocity structure than oceanic crust defined by Watts *et al.*,
369 (2009) in another segment of the Equatorial Atlantic Ocean (segment I in Figure 1):
370 their wide-angle data present a thin layer with velocity 5.0 – 5.5 km/s overlying a high
371 velocity layer of 7.0-7.2 km/s, with the Moho interpretation at circa 14-16 km.

372 Figure 16 presents several basement 1D-Vz profiles of intermediate domains of the
373 South Atlantic margins and the western Mediterranean Sea, showing the evolution
374 from an exhumed middle-lower continental crust to a thin oceanic crust. The Santos-
375 São Paulo system (SSPS) (Figure 16a), offshore the southeast Brazilian margin,
376 immediately north of the Florianópolis Fracture Zone, is interpreted as a kinematic
377 buffer zone (Moulin *et al.*, 2012) and represents an entire conjugate passive margin
378 system (Evain *et al.*, 2015). Whilst several studies have previously hypothesized the
379 presence of either an aborted oceanic propagator (Mohriak, 2001; Mohriak and
380 Szatmari, 2008; Mohriak *et al.*, 2010) or a failed spreading ridge (Demercian, 1996;
381 Karner, 2000; Meisling *et al.*, 2001; Mohriak, 2001; Gomes *et al.*, 2002, 2009) in the
382 SSPS surrounded by thinned continental crust, the acquisition of wide-angle seismic
383 data has shown that the central part of the SSPS has never reached a typical oceanic
384 crust (Evain *et al.*, 2015). Instead, in the central part of the SSPS, the southern profile
385 of the Sanba survey shows 1D-Vz profiles that are very consistent with the
386 surrounding domains and are interpreted as the result of exhumation of the
387 middle/lower continental crust (green lines, Figure 16a) with two thinner crustal
388 layers and a well-marked Moho. Only 50 km to the north, the northern wide-angle
389 profile (yellow lines, Figure 16a) shows important lateral variations with the lack of
390 the strong Moho step, replaced by a progressive increase in velocity in the lower
391 layer, which can be interpreted as mantle intrusion or altered mantle (Evain *et al.*,
392 2015).

393 In the Jequitinhonha margin, the Salsa wide-angle experiment has shown that the
394 typical oceanic crust is never imaged along two approximately 200-km-long profiles
395 (Loureiro *et al.*, 2018) as shown on the 1D-Vz profiles (Figure 16b). Instead, this 150-
396 km-wide transitional domain is interpreted as an exhumed lower continental crust
397 with shearing of the deepest parts, underlain by an anomalous velocity zone. The
398 seismic propagation velocities and geometry of the reflectors suggest the existence of

399 a boudin-like feature probably linked to the crustal thinning, with an oceanward creep
400 of the lower crust, or ultramafic intrusions in the lower crust (Loureiro *et al.*, 2018).
401 Furthermore, the Provençal basin is a young, aborted rift system where both conjugate
402 margins are accessible, with, in the central part, a domain assumed to be of oceanic
403 nature (Burrus, 1984; Le Douaran *et al.*, 1984; De Voogd *et al.*, 1991; Pascal *et al.*,
404 1993; Rollet *et al.*, 2002; Gailler *et al.*, 2009) and characterized by a magnetic
405 anomaly pattern that sharply differs from the quiet adjacent domains (Bayer *et al.*,
406 1973; Galdéano and Rossignol, 1977) and a concomitant low-gravity anomaly. The
407 Sardinia wide-angle experiment (Moulin *et al.*, 2015; Afilhado *et al.*, 2015) has shown
408 the strong symmetrical segmentation of the basin, with intermediate domains on both
409 sides interpreted as exhumed lower continental crust and a central part of the 4-5-km-
410 thick atypical oceanic domain with strong variations between the conjugate profiles.
411 Figure 16c presents the 1D-V_z profiles of the Gulf of Lion profile (purple lines) and
412 the Sardinia profile (red lines). Whilst the comparison between the 1D velocity
413 structure of typical oceanic crust (White *et al.*, 1992 and Christeson *et al.*, 2019) and
414 the Sardinian 1D-V_z profiles shows a thinner crust consistent with oceanic crust
415 (Afilhado *et al.*, 2015), the Gulf of Lion 1D-V_z profiles present strong similarities
416 with the 1D-V_z profile of the adjacent domain interpreted as exhumed and intruded
417 lower continental crust. As quoted by Afilhado *et al.* (2015), this observation raises
418 the question as to the role of the lower continental crust “flow” which was maybe
419 gradually recrystallized to build the first atypical oceanic crust (Bott, 1971; Aslanian
420 *et al.*, 2009; Sibuet *et al.*, 2012). A similar evolution is observed on the segmentation
421 of the Pará-Maranhão-Barreirinhas-Ceará segment from its Oceanic domain A to
422 Oceanic domain B (Figure 14 & 16d).

423 Figure 17 compiles the evolution of the 1D-V_z profiles and compares them with new
424 wide-angle and MCS seismic results acquired on the North Colorado margin (Shuck
425 *et al.*, 2019; Becel *et al.*, 2020). The Colorado 1D-V_z profiles (black lines in Figure
426 17), between the two Magnetic anomalies ECMA (East Coast Magnetic Anomaly) and
427 BSMA (Blake Spur Magnetic Anomaly), are interpreted as a proto-oceanic domain
428 before the establishment of normal steady-state oceanic spreading ridge, which
429 according to these authors started at the time of the Blake Spur with an approximate
430 age of ≈ 170 Ma, which may correspond to a strong change in the spreading rates (an
431 increase from ≈ 0.8 cm/y to ≈ 1.7 cm/y) and direction, from NNW–SSE to NW–SE
432 (Sahabi *et al.*, 2004; Labails *et al.*, 2010) or an eastward ridge jump (Vogt, 1973; Bird

433 *et al.*, 2007).

434 These 1D-Vz profiles present top velocities between 5.7 and 6 km/s, which are much
435 higher than the velocity structure of a normal oceanic crust (figures 14 to 17 ; White
436 *et al.*, 1992; Christeson *et al.*, 2019). These 1D-Vz profiles fit better with the 1D-Vz
437 profiles of the central part of the Santos Basin (green lines in Figures 16 & 17) or the
438 1D-Vz profiles of the Jequitinhonha Basin, with a smaller thickness. In these two
439 basins, the sampled domain is interpreted as exhumed/intruded lower continental crust
440 (Evain *et al.*, 2015; Loureiro *et al.*, 2018), which took place before the inset of true
441 oceanic crust spreading.

442 For the evolution of the Central Atlantic margin, Becel *et al.* (2020) propose a new
443 model in which the complete continental breakup of the conjugate margins system
444 was delayed by a slow extension rate and finally achieved at Blake Spur Magnetic
445 Anomaly time (early Bajocian), about 20 My after a first breakup at Sinumerian time
446 (Sahabi *et al.*, 2004).

447 Further north, in the conjugate Mesozoic Tarfaya–Laayoune and George Bank basins
448 system, Labails *et al.* (2009) highlighted strong asymmetry of this system for the first
449 phases of seafloor accretion until the BSMA, which they attributed to the existence of
450 the unaffected Precambrian Reguibat Craton on the African side: for these authors, the
451 main thinning process seems to have occurred along the Baltimore Canyon margin
452 and within the lower continental crust.

453 Towards the north, a similar observation of asymmetrical process is observed: on the
454 one hand, on the western margin, the SMART 1 Profile seems to show the presence of
455 a thin continental crust above a layer of serpentized mantle off the East Coast
456 Magnetic Anomaly (Funck *et al.*, 2004) while on the eastern conjugate margin, a true
457 oceanic crust is interpreted on the MIRROR profile (Biari *et al.*, 2015) off the
458 conjugate West African Coast Magnetic Anomaly (Sahabi *et al.*, 2004).

459 Whatever the nature of this American intermediate domain between the ECMA and
460 the BSMA, asymmetry is visible, always on the same side and throughout the entire
461 Central Atlantic Ocean, at least during the first phases of seafloor accretion.

462 It is problematic to define the difference between a proto-oceanic crust and
463 exhumed/intruded lower continental crust, as short lateral variations may often occur
464 in this mixing area as in the Santos Basin or the Provençal Basin, and it may become a
465 subject of interminable debate. The evolution of 1D-Vz profiles every 10 km provides
466 a favourable image of the evolution along one profile, with rather homogenous

467 segments and strong velocity steps in between (see for instance Moulin *et al.*, 2015 or
468 Aslanian *et al.*, 2021). As shown in the Provençal, Santos, Angola, Jequitinhonha and
469 Pará-Maranhão-Barreirinhas-Ceará basins, the evolution from a passive margin
470 towards a true oceanic crust was progressive, and probably involved the lower
471 continental crust in the proto-oceanic crust construction, as previously suggested
472 (Bott, 1971, Aslanian *et al.*, 2009; Sibuet *et al.*, 2012).

473

474 Contrary to Aslanian *et al.* (2009), who described a stress field that narrows by steps
475 from more than 500 km to less than 50 km in the South Atlantic margins, Becel *et al.*
476 (2020) claim that continental rifting between eastern North America and northwest
477 Africa did not lead to rapid localization of extensional strain as previously thought.
478 They suggest that the full continental breakup was delayed with apparent variations in
479 the spreading rate of proto-oceanic crust possibly highlighting the continuous thermal
480 erosion of the continental mantle lithosphere by mantle melts over time up until the
481 formation of the BSMA. The pinch-out of continental mantle lithosphere at the apex
482 of the end of the proto-oceanic crust in their model may correspond to the deep
483 reflectors observed on the ION-GXT 7000 line, which also seem to pinch out at the
484 apex of the inset of thin oceanic crust (Figure 13A).

485 **8. Conjugate margin, asymmetry and age of the oldest oceanic** 486 **crust**

487 Any margin is only a part of the jigsaw, half part of a margins system, and palinspastic
488 reconstructions are needed to understand the evolution of the entire jigsaw. Figure 18
489 presents a palinspastic reconstruction of the Brazilian-Ivory Coast-Ghana system at
490 100 Ma, with the Euler pole of Moulin *et al.* (2010). Due to the lack of magnetic
491 information during this period, this position the position of the South American plate
492 respect to African plate is extrapolated from the intermediate pole at C34 of Campan
493 (1995) and the fit of Moulin *et al.* (2010) at 112 Ma, assuming a constant spreading
494 rate. This extrapolation identifies a good fit between the small circles and the fracture
495 zones, which indicates that no change in movement direction occurred between C34
496 and the fit, but the timing of the breakup remains unknown. If, as in the Central
497 Atlantic Ocean, the rate of extension was low during the initial development of the
498 continental rift (Labails *et al.*, 2010; Becel *et al.*, 2020), the reconstruction presented

499 in Figure 18 is younger than 100Ma.

500

501 Knowing the exact position of the first proto-oceanic or oceanic crust on the African
502 side, which would imply an asymmetric or symmetric system, is of major importance
503 in understanding its evolution. Both supposed African oceanic limits from Heine et al.
504 (2013) and Sage et al. (1997) are represented of Figure 18. Whatever the age of this
505 reconstruction, as this margins system is strongly influenced by the NW-SE direction
506 as demonstrated on the Brazilian side, the rounded shape and position of LALOC
507 (Heine et al., 2013; Mueller et al., 2016) raises issues. Moreover, as the Brazilian side
508 is now well constrained and described, this would imply a strong asymmetry between
509 the two sides of the system. These issues are less significant but still present with the
510 Continent-Ocean Boundary (COB) from Sage et al. (1997). Present knowledge and
511 data do not provide the age of the first oceanic inset. We would need to conduct
512 specific experiments on the Ghana-Ivory Coast margins to clarify this question. It is
513 maybe worth to note that In contrast Figure 18 shows a striking coincidence to the
514 alignment at that time (100Ma) of the Brazilian and African coasts and the inner limit
515 of the Brazilian proto-oceanic crust.

516

517 **9. Conclusion**

518 The Pará-Maranhão-Barreirinhas-Cearà margin shows an E-W segmentation with, in
519 the presumed oceanic crust, two different domains. Along the MC1 profile, which
520 entirely crosses the margin, the first A domain appears homogeneous, with strong
521 affinity with the deep basin interpreted as exhumed lower continental crust (Aslanian
522 *et al.*, 2021). Further to comparison with similar studies in the Angola Margin
523 (Aslanian *et al.*, 2009), the Provençal Basin (Moulin *et al.*, 2015; Afilhado *et al.*,
524 2015), the Santos Basin (Evain *et al.*, 2015), the Jequitinhonha Basin (Loureiro *et al.*,
525 2018) and the eastern North American Margin (Becel *et al.*, 2020), we interpret this
526 domain as a proto-oceanic crust preceding the inset of true oceanic crust and
527 following the inset of an exhumed/intruded middle/lower continental crust, following
528 the idea of Bott (1971) and Aslanian *et al.* (2009).

529 As suggested by Becel *et al.* (2020), this evolution may suggest a continuous thermal
530 erosion of the continental mantle lithosphere by mantle melts over time until the inset
531 of true oceanic crust.

532 In 2009 and 2018, Aslanian *et al.* highlighted the crucial role of the lower continental
533 crust and mantle intrusions in the passive margin formation and proposed an evolution
534 in three stages: 1) a first rift stage, with small or no crustal thinning but mafic
535 intrusions which produced overloading and subsidence (Tozer *et al.*, 2017; Shulgin *et*
536 *al.*, 2018; Moulin *et al.*, 2020; Leprêtre *et al.*, 2021); 2) an exhumation phase, which
537 is the main thinning phase and produces a basement made of exhumed/intruded lower
538 continental crust, proto-oceanic crust (Contrucci *et al.*, 2004; Funck *et al.*, 2004;
539 Moulin *et al.*, 2005; Aslanian and Moulin, 2012; Sibuet *et al.*, 2012; Moulin *et al.*,
540 2015; Afilhado *et al.*, 2015; Evain *et al.*, 2015; Loureiro *et al.*, 2018; Shuck *et al.*,
541 2019; Becel *et al.*, 2020), and in some rare cases, exhumed mantle (e.g., Péron-
542 Pinvidic and Manatschal, 2008); and 3) the oceanic spreading stage. This scheme may
543 be progressive, with lower continental material flow inside a first proto-oceanic crust
544 and as suggested by Becel *et al.* (2020) continental mantle lithosphere erosion before
545 the inset of true oceanic crust.

546

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568 **Author contributions**

569 The MAGIC Project was imagined by D. Aslanian and led by D. Aslanian, M. Moulin
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584 (aviana@petrobras.com.br).

585 **Figures**

586 *Figure 1: a) Kinematic reconstruction at C34 (Campanian, 84 Ma) of the Equatorial Atlantic*
587 *Ocean, after Moulin et al. (2010), showing segmentation of the Equatorial Segment and*
588 *localisation of Figures 1b and 2a. The figure shows, on each plate, the gravity data from*
589 *Sandwell and Smith (1997) between the coast and the anomaly C34. The West African plate is*
590 *fixed. b) Bathymetry map of the Pará-Maranhão/Barreirinhas margin acquired during the*
591 *MAGIC experiment. White circles represent the OBS deployed during the MAGIC experiment.*
592 *GXT lines are indicated by dotted red lines and labels. Blue rectangle shows the location of*
593 *the OBS presented in this study. Location of basin I, II and volcanic alignment, issued from*
594 *Aslanian et al., 2021 is also indicated.*
595

596 *Figure 2: a) Gravity map of the Ghana-Ivory Coast margin, see general location in Figure*
597 *1a. The position of kinematic fit from Moulin et al. (2010) is in blue. The dotted red line*
598 *represents the supposed True Oceanic Boundary resulting from refraction studies (Pierce et*
599 *al., 1996; Sage et al., 1997; Edwards et al., 1997) whilst the pink line represents the position*
600 *of Landward Limit Oceanic Crust from Heine et al. (2013). Positions of the Equasis, Equaref*
601 *and Darwin cruise 55 experiments and some industrial lines available are in black; thick*
602 *yellow lines represent the part of the profiles shown in Figure 2b. White dots represent the*

603 *OBS, which register data. Brown area represents the West African craton. b) Final velocity*
 604 *model of 2 seismic lines located on the Ivory Coast Basin (yellow lines on the map below)*
 605 *from Pierce et al. (1996, EQUARE-1) and Sage (1994, EQUAREF-7). The basement and the*
 606 *Moho are indicated by thick black lines, the anomalous velocity zone of EQUAREF-1 profile*
 607 *by a thick red line, and the supposed oceanic crust by a blue area. The positions of the*
 608 *landward Limit Oceanic Crust from Heine et al. (2013) and the fit from Moulin et al. (2010)*
 609 *are also indicated by pink and blue lines, respectively.*

610

611 *Figure 3: Comparison between MC1OBS06, MC1OBS12 and MC1OBS24 on the MAGIC1*
 612 *profile – southeast direction to the right and northwest to the left. a) Colour-coded observed*
 613 *travel-times reduced by a velocity of 4 km/s overlain by predicted times in black. b) Seismic*
 614 *rays. c) MCS time migrated section and colour-coded model interfaces.*

615

616 *Figure 4: MC1OBS09 on the MAGIC1 profile – southeast direction to the right and northwest*
 617 *to the left. a) Seismic record. b) Synthetics. c) Colour-coded synthetics. d) Colour-coded*
 618 *observed travel-times overlain by predicted times in black. e) Corresponding ray tracing of*
 619 *black line in Figure 4d. f) MCS time migrated section and colour-coded model interfaces. On*
 620 *a, b, c, and d, travel-time is reduced by a velocity of 7 km/s. Same legend and colour code as*
 621 *Figure 3.*

622

623 *Figure 5: MC1OBS11 on the MC1 profile - southeast direction to the right and northwest to*
 624 *the left. Same legend and colour code as Figure 3.*

625

626 *Figure 6: MC1OBS18 on the MC1 profile - southeast direction to the right and northwest to*
 627 *the left. Same legend and colour code as Figure 3. Note that the horizontal scale of this*
 628 *profile is different from the previous one.*

629

629 *Figure 7: MC1OBS24 on the MC1 profile - southeast direction to the right and northwest to*
 630 *the left. Same legend and colour code as Figure 3.*

631

632 *Figure 8: Velocity model of the MC1 profile. Thick black lines mark layer boundaries from the*
 633 *modelling. Coloured areas are constrained by seismic rays. Inverted black triangles mark*
 634 *OBS positions and thin blue lines cross points with other MAGIC profiles and the ION-GXT*
 635 *profiles. Vertical exaggeration 1:5. Thin blue lines mark the intersection with MAGIC and*
 636 *ION-GXT datasets.*

637

638 *Figure 9: Gravity model for the MAGIC1 profile overlain by interfaces from wide-angle*
 639 *modelling. a) Density model up to a depth of 30 km. b) Free-air gravity anomaly observed*
 640 *(Pis et al., 2012) along the MAGIC1 profile (red dotted) and calculated (black line). c) Load*
 641 *anomaly.*

642

643 *Figure 10: a) Distribution of interface depth nodes and top and bottom velocity nodes of the*
 644 *final P-wave interval velocity model along the MAGIC1 wide-angle profile. Interfaces where*
 645 *reflections have been observed on OBS data are highlighted in blue. b) Hit count of rays*
 646 *traced in the model during inversion. Depth interface nodes are plotted with squares scaled to*
 647 *the hit count through the node. c) Spread-Point Function (SPF) in the model. Depth interface*
 648 *nodes are plotted with squares scaled to the SPF at the node. d) Resolution in the model.*
 649 *Depth interface nodes are plotted with squares scaled to the resolution at the node.*

650

651 *Figure 11: Two-way travel-time record section of MCS data along the MAGIC1 profile A) un-*
 652 *interpreted and B) overlain by time-converted interfaces of the wide-angle model.*
 653 *Intersections with MAGIC and ION-GXT datasets are indicated by red and grey arrows,*
 654 *respectively. OBS locations are indicated in blue at the base of the profile. Vertical*
 655 *exaggeration at seafloor is 1:12.5.*

656

657 *Figure 12: Pre-stack depth migrated record section of MCS data along the MAGIC1 profile*

658 A) *un-interpreted and B) Model's interfaces are represented with continuous lines.*
 659 *Intersections with MAGIC and ION-GXT datasets are indicated by red and grey arrows,*
 660 *respectively. Vertical exaggeration is 1:5.*

661

662 *Figure 13: A) Line drawing of the depth-converted profile ION-GXT 7000, coincident with the*
 663 *MC3 profile (Aslanian et al., 2021). Thick blue lines indicate where wide-angle reflections hit*
 664 *the interface. B) MC3OBS11 (indicated in blue triangle in figure 13A) on the MAGIC3*
 665 *profile – east direction to the right and west to the left. a) Seismic record; b) Synthetics; c)*
 666 *Colour-coded synthetics; d) Colour-coded observed travel-times overlain by predicted times*
 667 *in black; e) Seismic rays; f) MCS time migrated section and colour-coded model interfaces.*
 668 *On a, b, c, and d, travel-time is reduced by a velocity of 7 km/s.*

669

670 *Figure 14: a) Distribution of 1D- velocity/depth profiles extracted every 10 km from part of*
 671 *the final P-wave velocity model showing the segmentation (colour code in legend) along MC4*
 672 *(a - top), MC2 (b - middle) and MC3 (c - bottom) profiles, crossing Basins IIA and I and the*
 673 *presumed oceanic crust. Velocity profiles for oceanic crust are a compilation of White et al.,*
 674 *1992 (black lines) and Christeson et al., 2019 (orange lines). Grey area on profiles indicates*
 675 *the High Velocity Layer (H.V.L.) revealing by Aslanian et al., 2021.*

676

677 *Figure 15: Top: Total 1D-Vz profile (including water and sediment layers) of the three*
 678 *crossing points between the MC1 profile and the MC2, MC3 and MC4 profiles, within the*
 679 *presumed oceanic A domain. Bottom: distribution of 1D- velocity/depth profiles extracted*
 680 *every 10 km from part of the final P-wave velocity model showing the segmentation (colour*
 681 *code in legend) along MC1, crossing the supposed oceanic domain. Light Purple area:*
 682 *compilation of velocity profiles for the oceanic crust (White et al., 1992 and Christeson et al.,*
 683 *2019).*

684

685 *Figure 16: Compilations of the Basement 1D-Vz profiles extracted every 10 km. Light grey*
 686 *area: compilation of velocity profiles for the oceanic crust (White et al., 1992 and Christeson*
 687 *et al., 2019). a) SANBA experiment – the central part of the Santos-São Paulo system: Green:*
 688 *south profile; Yellow: north profile (Evain et al., 2015); b) SALSA experiment: deep basin of*
 689 *the Jequitinhonha margin in orange (Loureiro et al., 2018); c) SARDINIA experiment in the*
 690 *Provençal Basin: evolution from the proto-oceanic crust of the Gulf of Lion (purple, Moulin et*
 691 *al., 2015) to the thin oceanic crust (red, Afilhado et al., 2015) on the Sardinian side; D)*
 692 *MAGIC experiment: proto-oceanic crust (purple) and thin oceanic crust (red) (This study).*

693

694 *Figure 17: Compilation of the Basement 1D-Vz profiles of figure 17 (with the same colour*
 695 *code) and the Basement 1D-Vz profiles of the Colorado Basin (Shuck et al., 2019).*

696

697 *Figure 18: Palinspastic reconstruction of the Pará-Maranhão/Barreirinhas—Ghana-Ivory*
 698 *Coast system. Euler pole from Moulin et al. (2010). The south American plate is fixed. Note*
 699 *that, due to the lack of magnetic information during this period, this position is extrapolated*
 700 *from the intermediate pole at C34 of Campan (1995) and the fit of Moulin et al. (2010) at*
 701 *112 Ma, with a constant and regular spreading rate. The segmentation of the Pará-*
 702 *Maranhão/Barreirinhas margin issued from this study and Aslanian et al. (2021) (in blue*
 703 *lines) is represented together with the location of the LaLOC, the African COB from Sage et*
 704 *al. (1997) and different propositions for the initial fit (Moulin et al., 2010 ; Heine et al., 2013*
 705 *and Muller et al., 2016)*

706

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MC01

Phase	npts	Trms	chi-squared
Pw	8883	0.017	0.028
Ps2P	1264	0.035	0.126
Ps3	707	0.066	0.434
Ps3P	1396	0.050	0.246
Ps4	652	0.093	0.875
Ps4P	1458	0.036	0.132
Ps5P	2721	0.074	0.542
Pg1/Pg2	6392	0.076	0.583
Pg1P	1953	0.058	0.333
Pm1P	3032	0.051	0.262
Pn	6198	0.079	0.627
All phases	34656	0.060	0.356

Table 1: Reflected or refracted phase name, number of explained events, residual mean-square, and normalized chi-squared value for all phases and the complete model of profile MC1 profile.

Instrument	shot	dir	npts	Trms	Chi-squared
MC1OBS01	299.213	-1	717	0.066	0.434
MC1OBS02	286.339	1	143	0.048	0.234
	286.339	-1	580	0.052	0.269
MC1OBS03	274.119	1	740	0.053	0.279
	274.119	-1	491	0.092	0.848
MC1OBS04	261.817	1	650	0.062	0.390
	261.817	-1	633	0.056	0.313
MC1OBS05	249.396	1	805	0.068	0.467
	249.396	-1	762	0.048	0.232
MC1OBS06	236.906	1	923	0.055	0.306
	236.906	-1	598	0.050	0.254
MC1OBS07	223.935	1	793	0.044	0.192
	223.935	-1	750	0.057	0.325
MC1OBS08	211.952	1	939	0.055	0.308
	211.952	-1	718	0.035	0.125
MC1OBS09	199.476	1	966	0.055	0.297
	199.476	-1	670	0.033	0.112
MC1OBS10	186.967	1	1169	0.057	0.327
	186.967	-1	694	0.036	0.132
MC1OBS11	174.216	1	715	0.062	0.387
	174.216	-1	666	0.057	0.326
MC1OBS12	162.041	1	696	0.075	0.570
	162.041	-1	705	0.046	0.213
MC1OBS13	147.385	1	693	0.077	0.590
	147.385	-1	832	0.048	0.233
MC1OBS14	137.046	1	534	0.046	0.215
	137.046	-1	679	0.056	0.314
MC1OBS15	124.667	1	573	0.050	0.250
	124.667	-1	851	0.069	0.470
MC1OBS16	112.174	1	680	0.060	0.360
	112.174	-1	966	0.090	0.810
MC1OBS17	99.701	1	518	0.068	0.463
	99.701	-1	829	0.061	0.368
MC1OBS18	87.222	1	559	0.042	0.178
	87.222	-1	897	0.057	0.321
MC1OBS19	74.701	1	496	0.049	0.243
	74.701	-1	552	0.071	0.502
MC1OBS20	62.283	1	670	0.062	0.386
	62.283	-1	530	0.061	0.369
MC1OBS21	49.807	1	619	0.061	0.370
	49.807	-1	755	0.064	0.415
MC1OBS22	37.415	1	688	0.056	0.312
	37.415	-1	1084	0.095	0.911
MC1OBS23	24.958	1	493	0.052	0.273
	24.958	-1	773	0.068	0.467
MC1OBS24	12.542	1	727	0.044	0.194
	12.542	-1	818	0.047	0.221
MC1OBS25	0.000	1	630	0.050	0.247
	0.000	-1	687	0.045	0.204

Table 2: Instrument name, distance along model, direction code, number of explained events, residual mean-square, and normalized chi-squared value for each OBS of the MC1 profile.

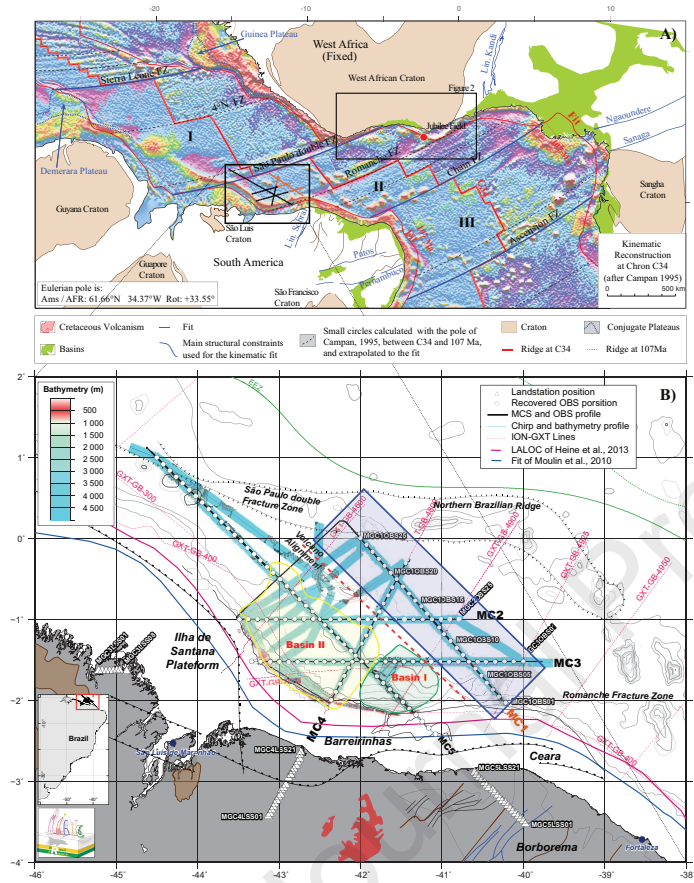


Figure 1 - Moulin et al.

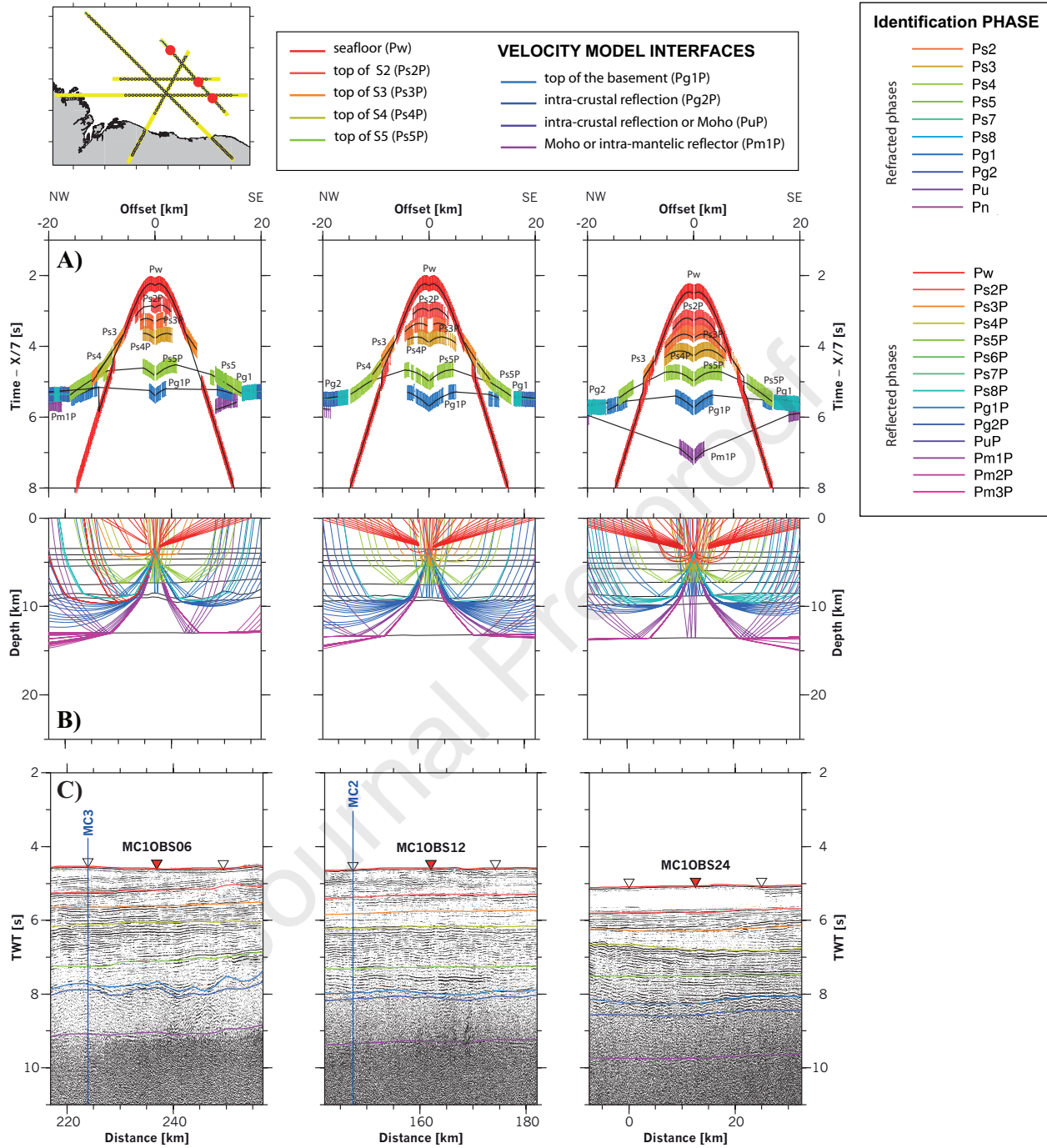


Figure 3 - Moulin et al.

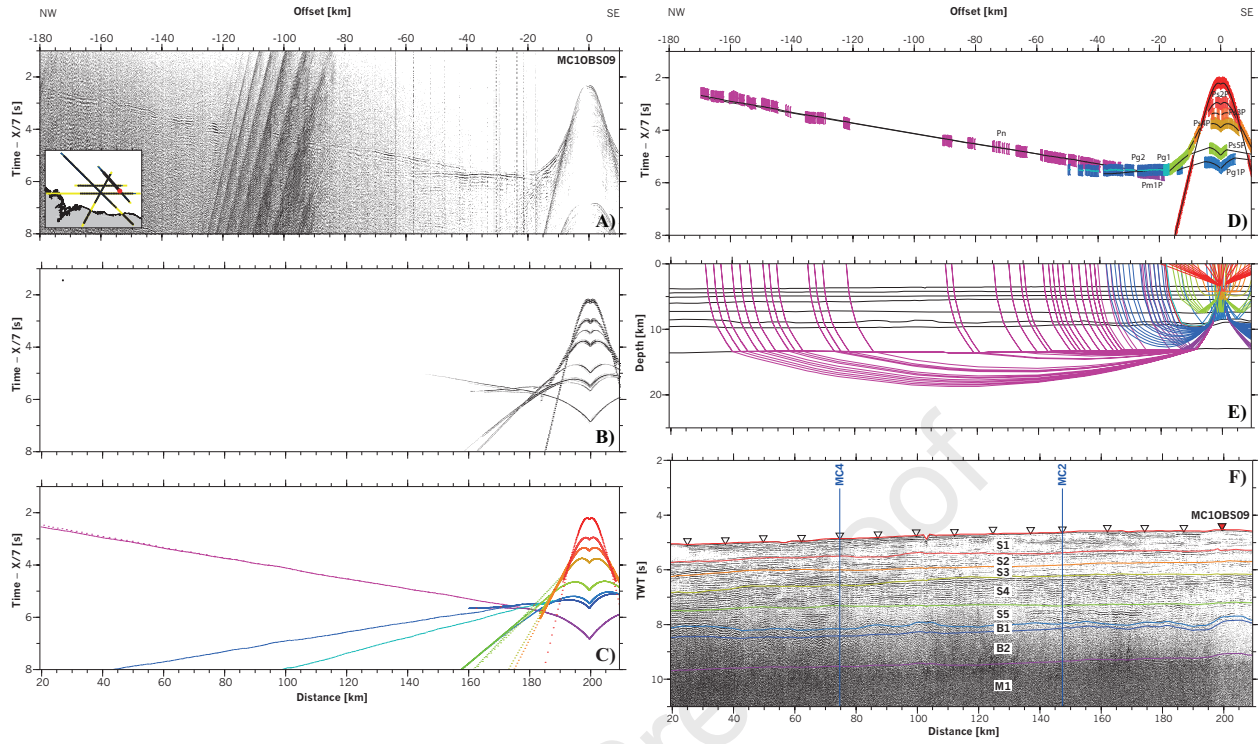


Figure 4 - Moulin et al.

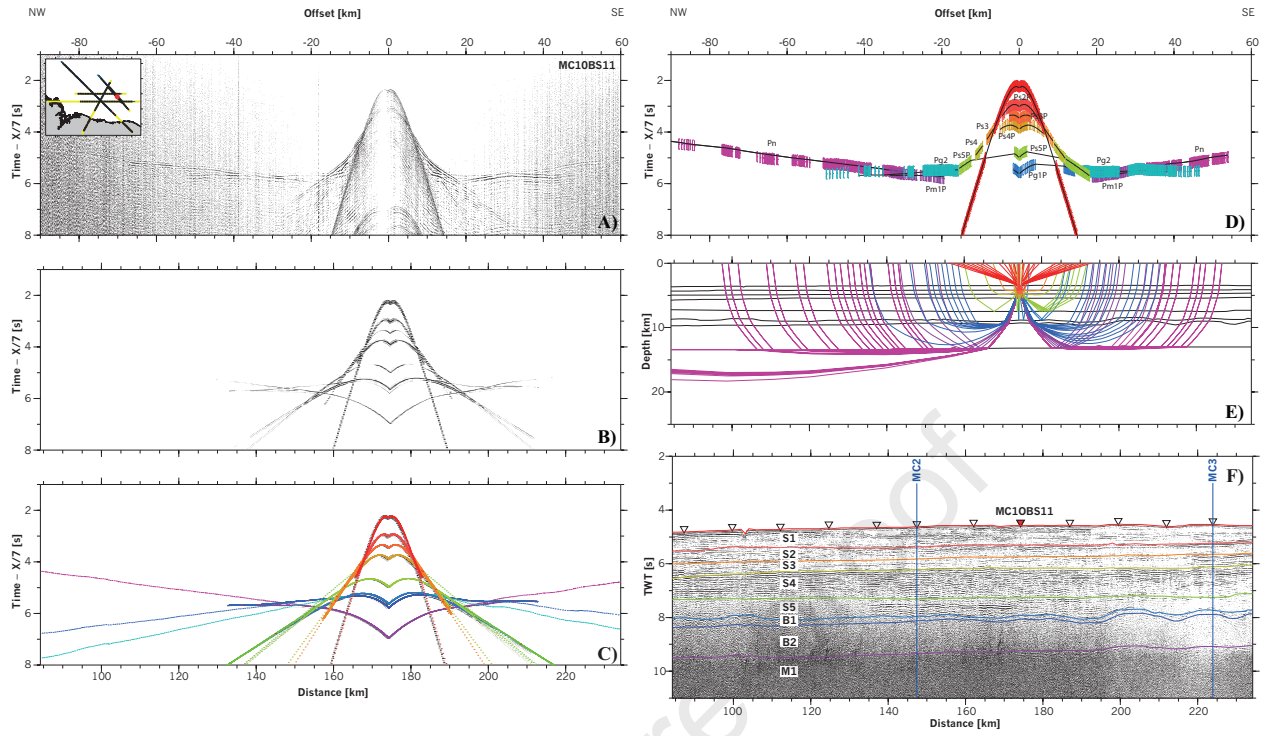


Figure 5 - Moulin et al.

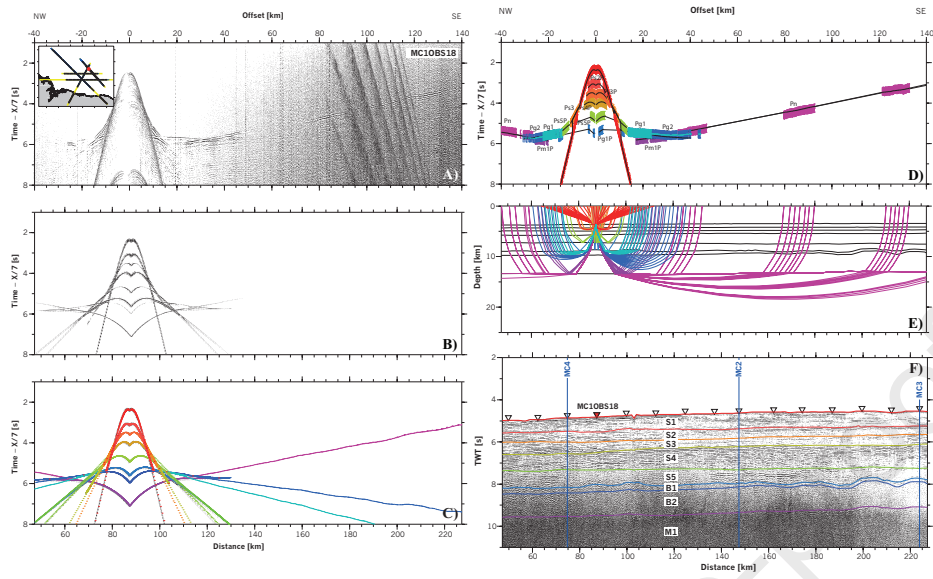


Figure 6 - Moulin et al.

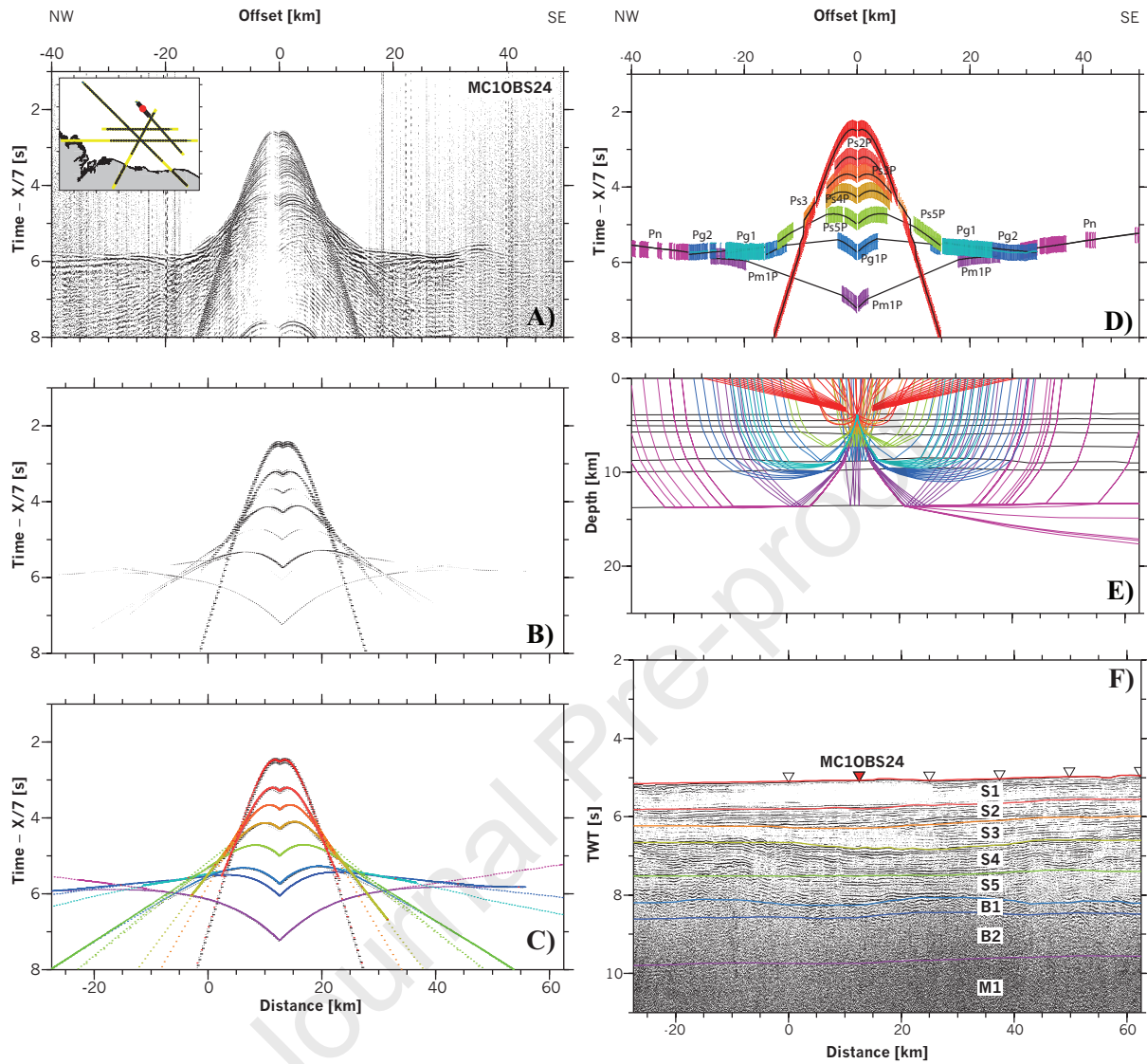


Figure 7 - Moulin et al.

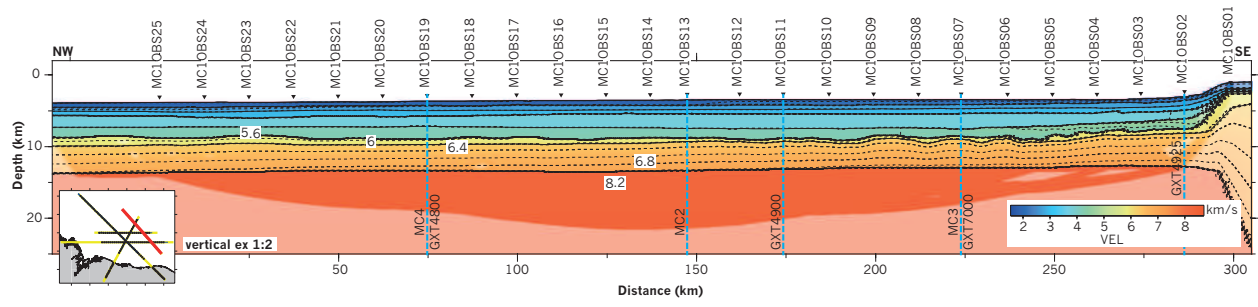


Figure 8 - Moulin et al.

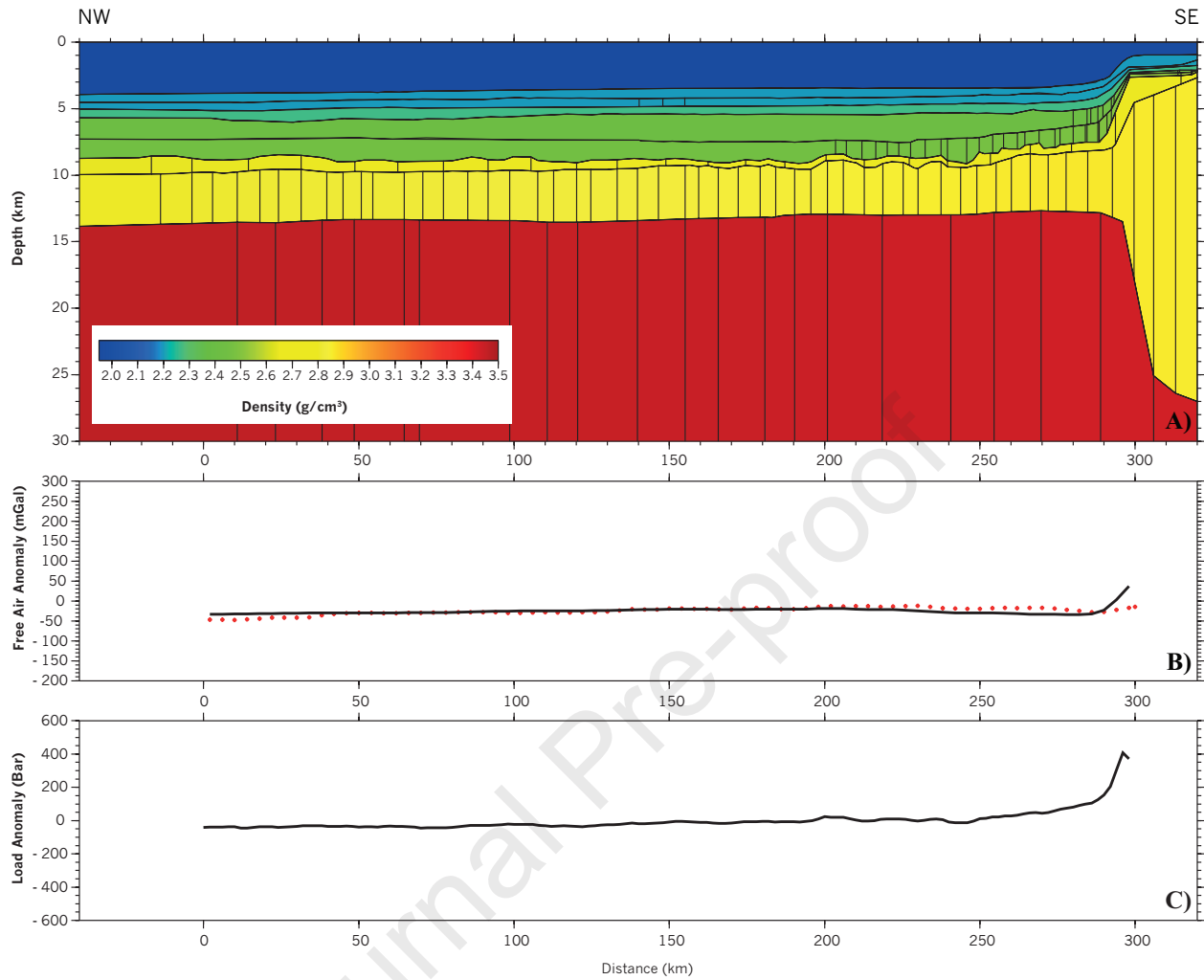


Figure 9 - Moulin et al.

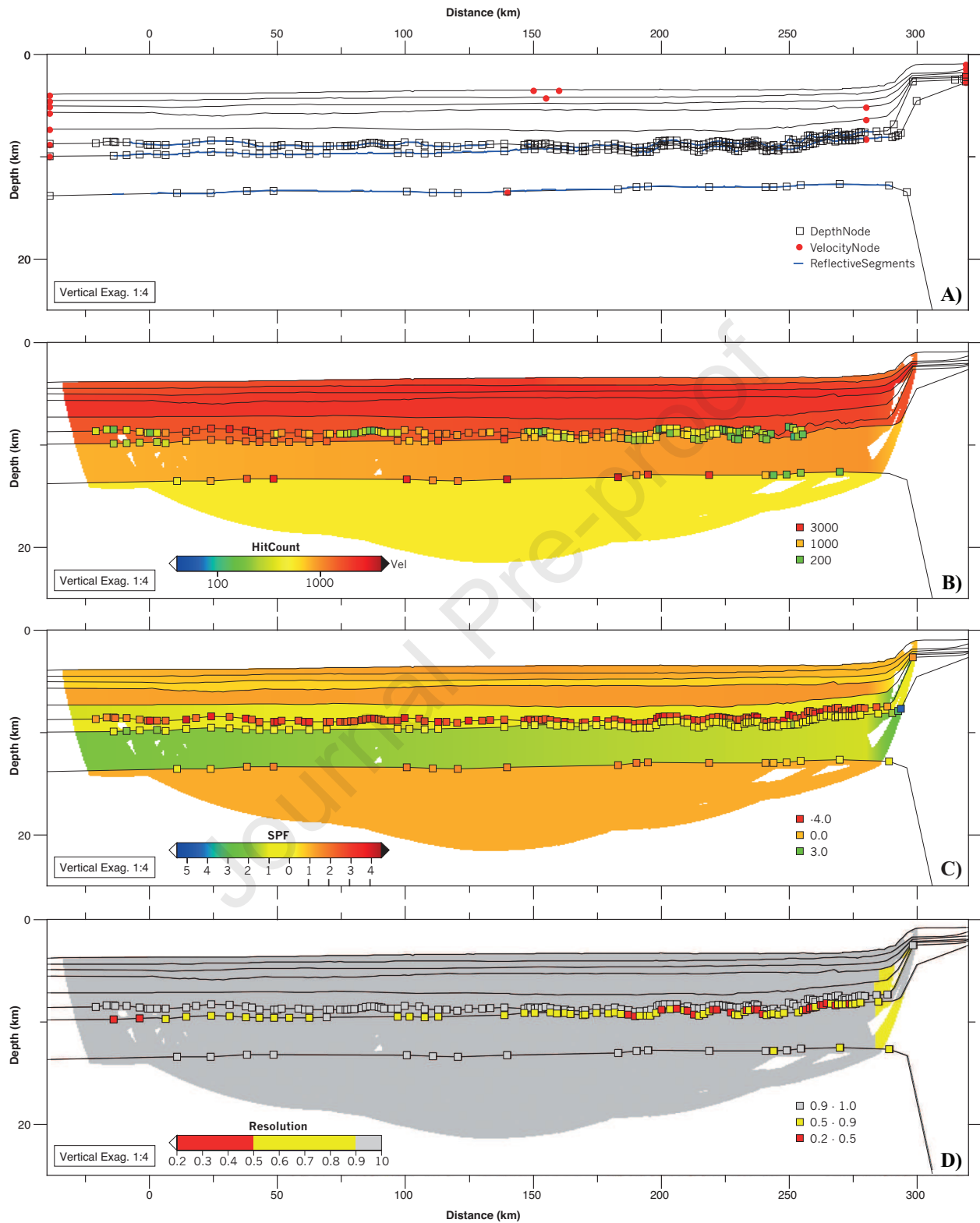


Figure 10 - Moulin et al.

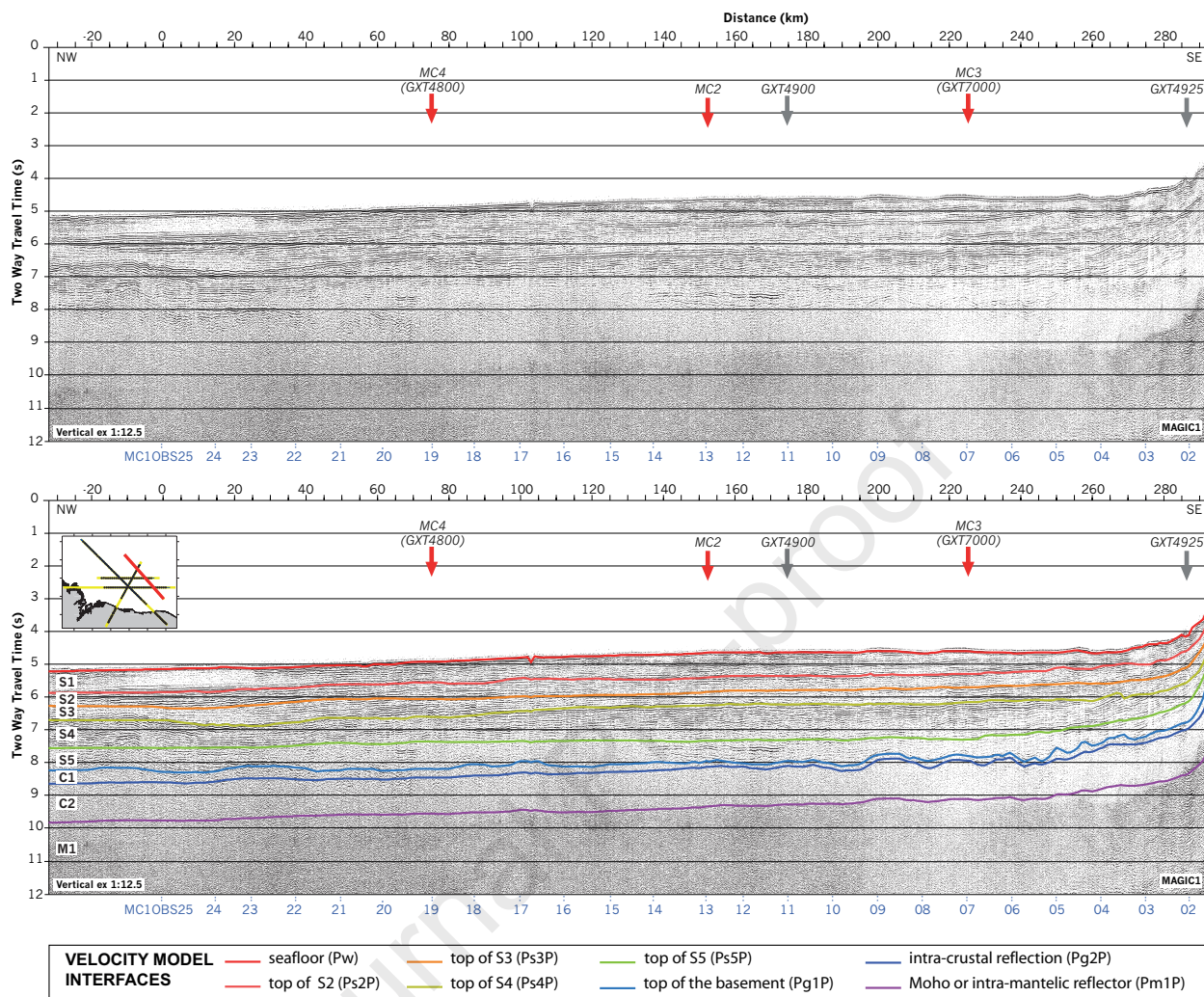


Figure 11 - Moulin et al.

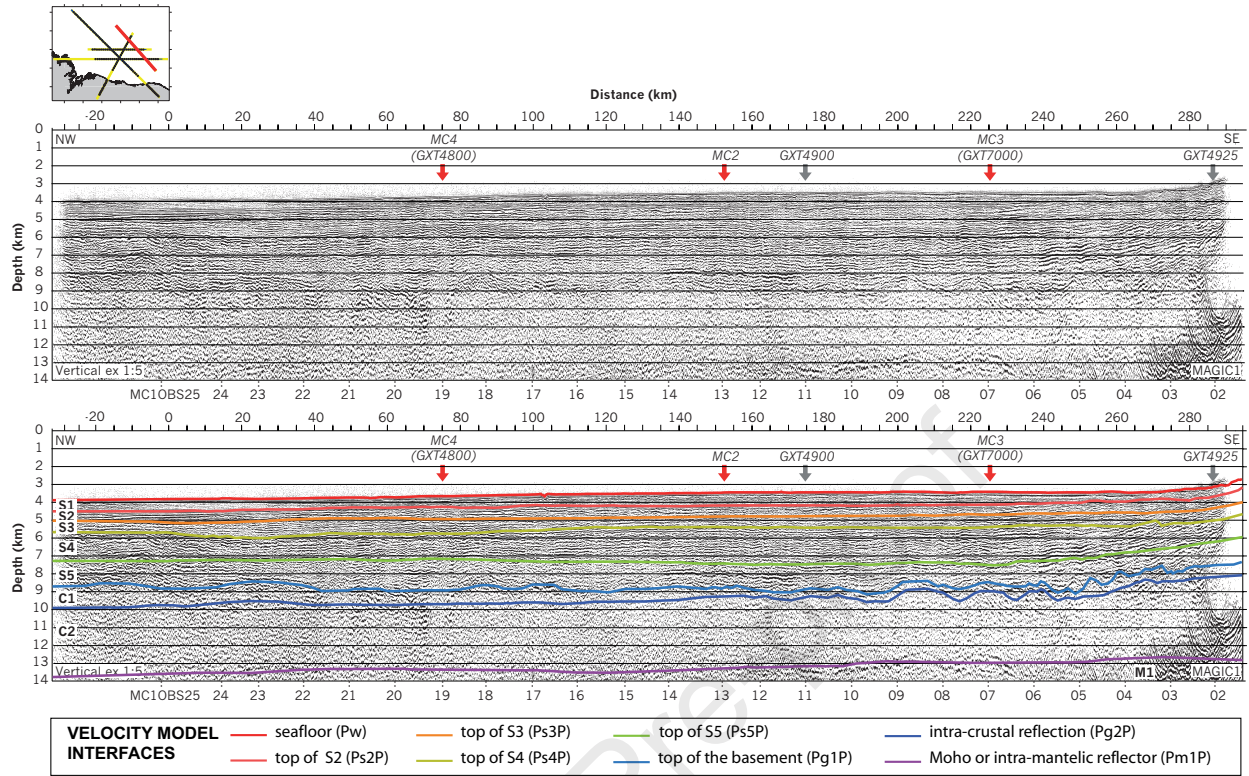


Figure 12 - Moulin et al.

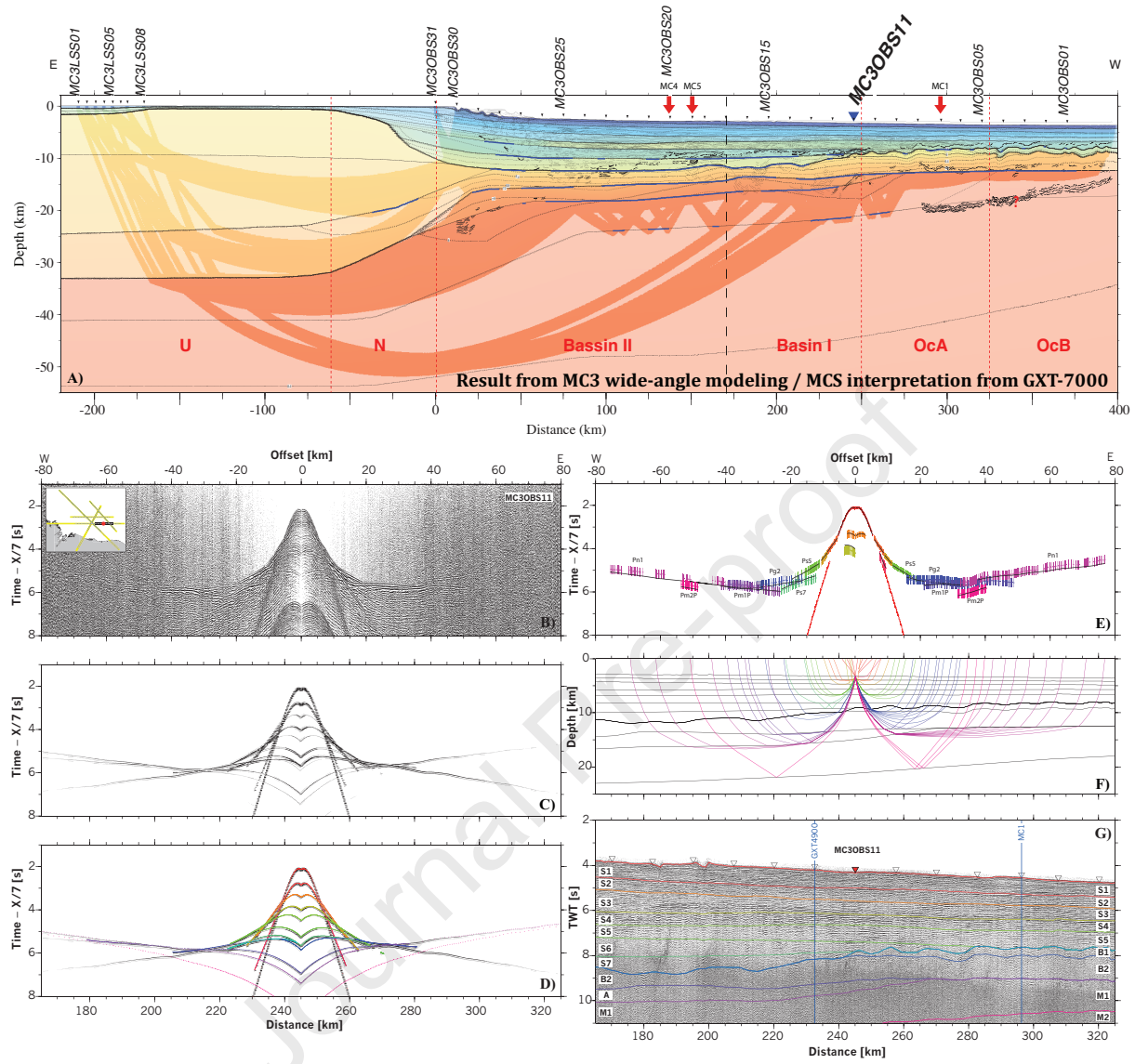


Figure 13 - Moulin et al.

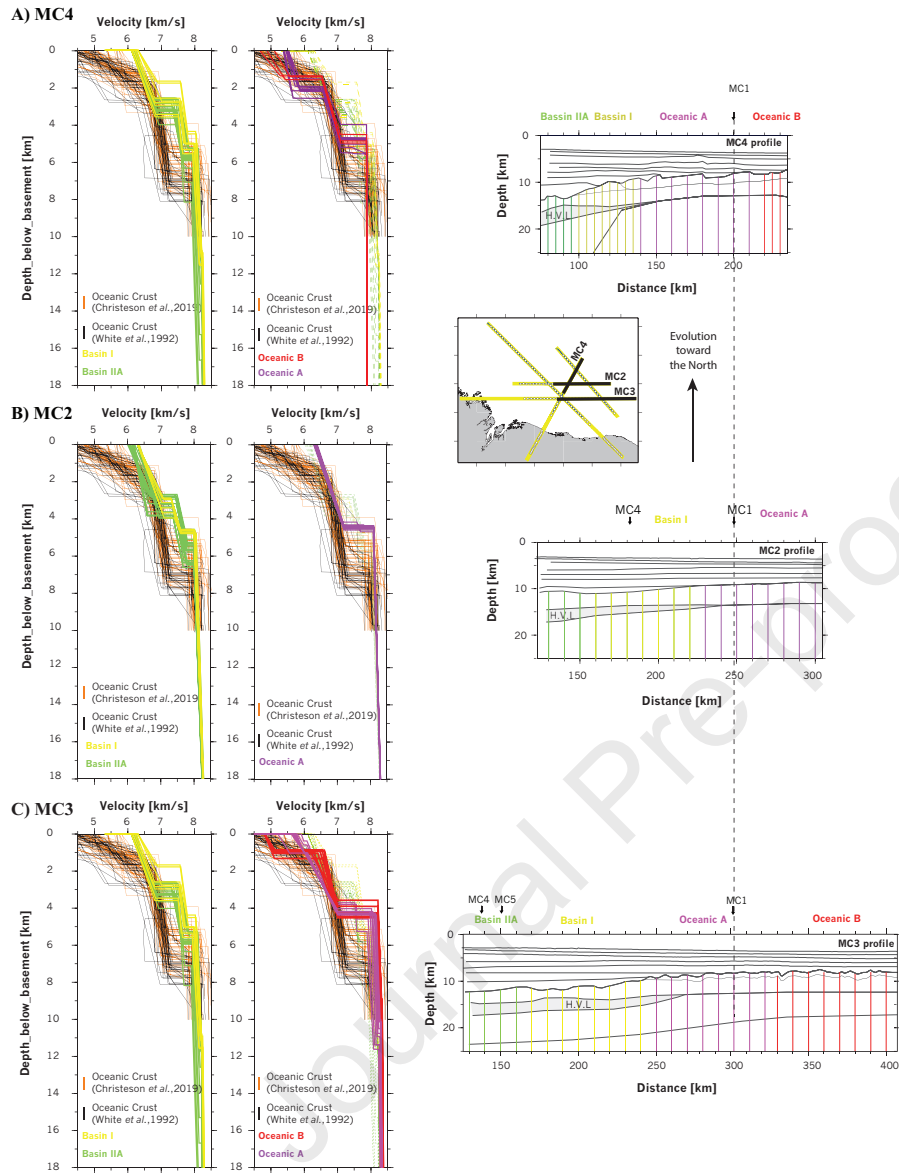


Figure 14 - Moulin et al.

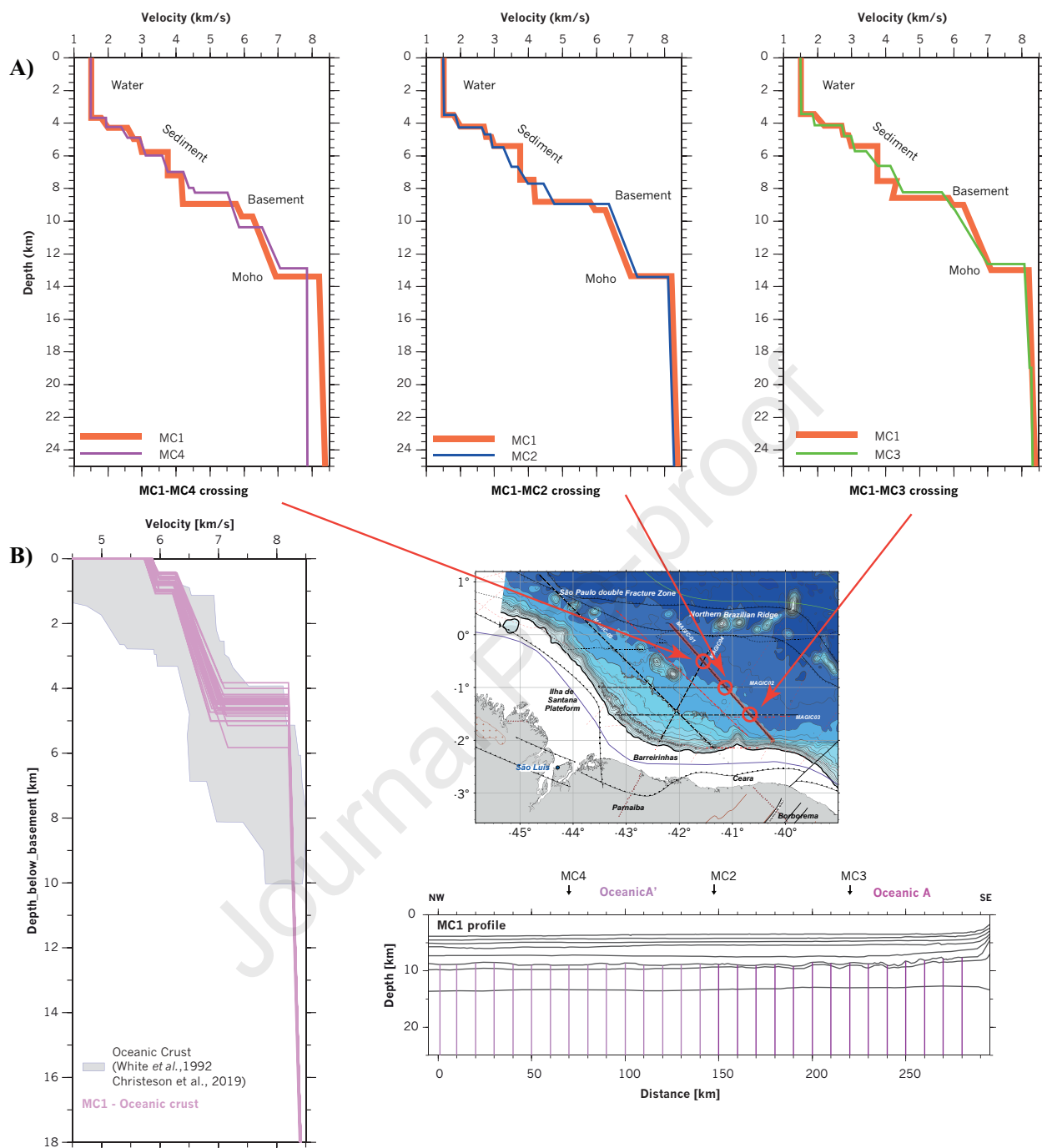
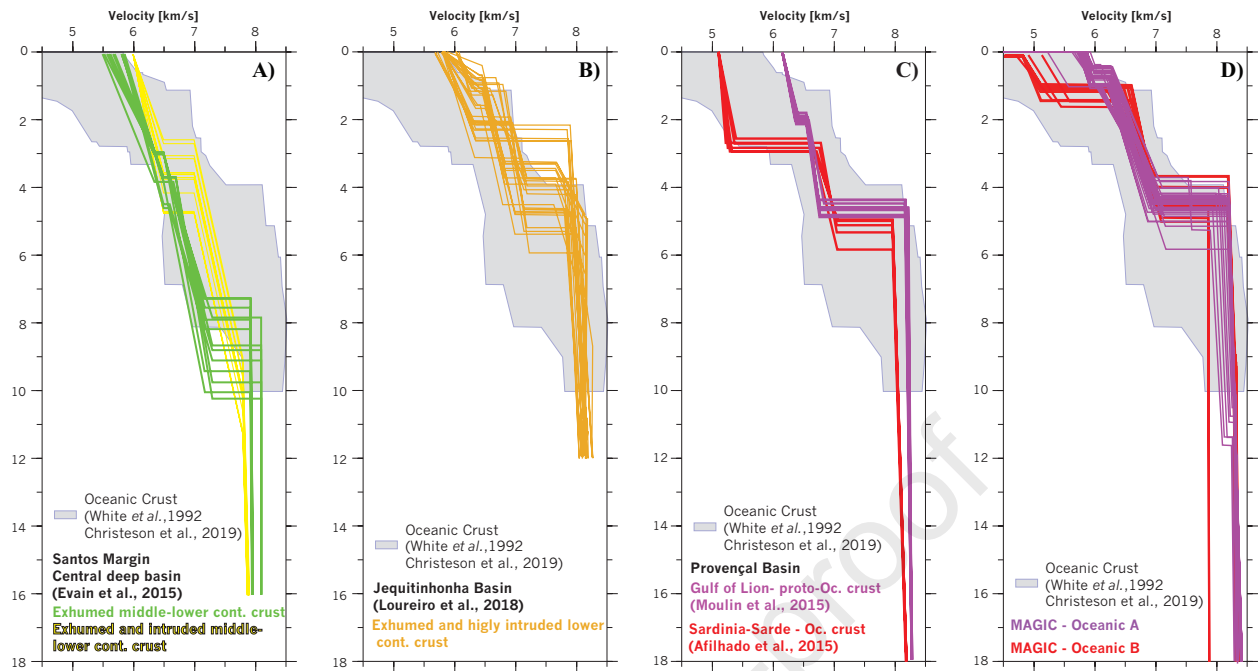


Figure 15 - Moulin et al.

Figure 16 - Moulin *et al.*

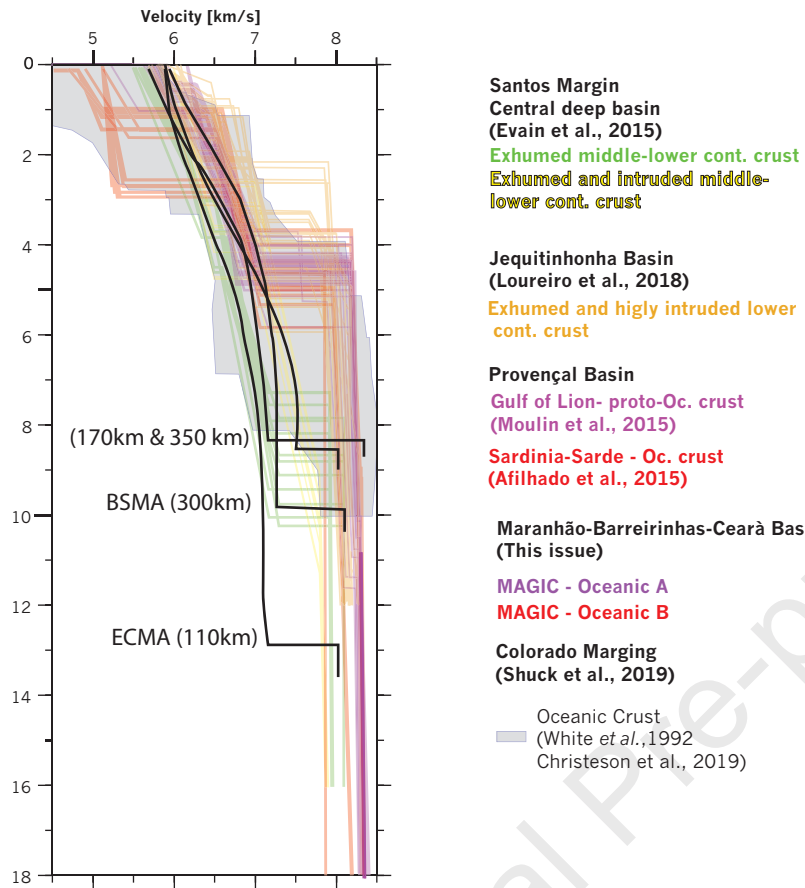


Figure 17 - Moulin et al.

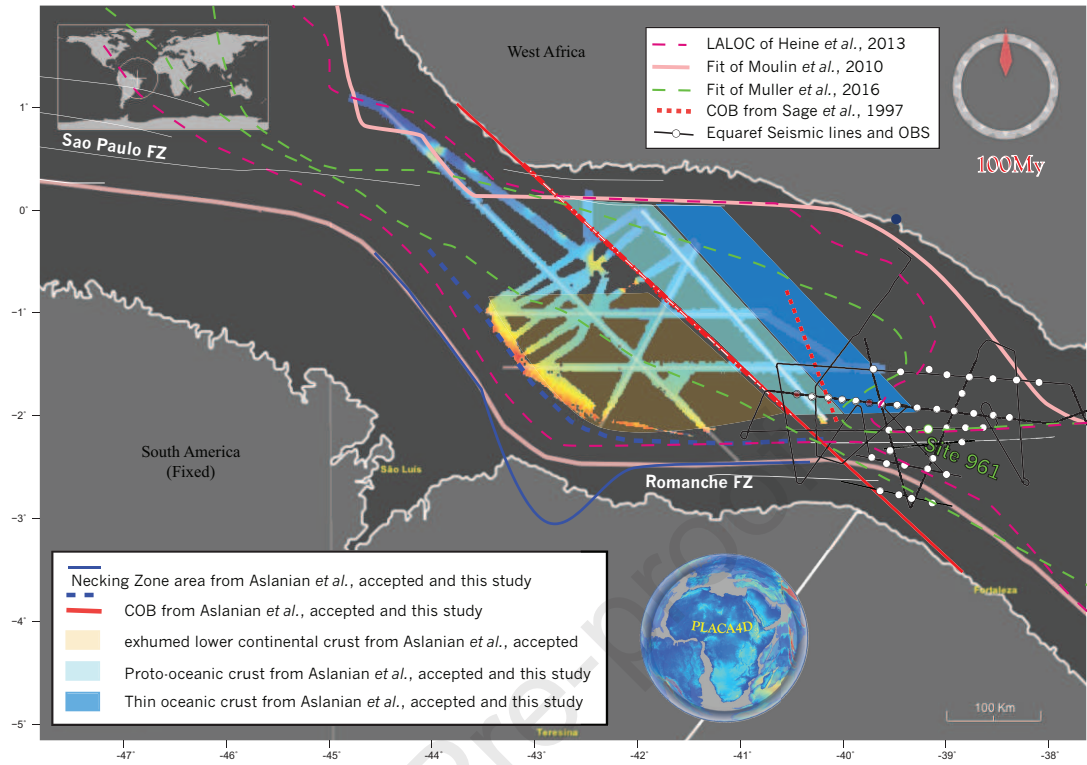


Figure 18 - Moulin *et al.*

The Pará-Maranhão-Barreirinhas-Ceará passive margin in the Equatorial Atlantic segment presents a strong E-W-segmentation with:

- A transition to oceanic crust very complex, with the presence of an about 60 km-wide band of proto-oceanic crust before a “more typical” but still thin oceanic crust eastward.
- The passage between an intermediate domain and a typical oceanic crust is not abrupt in terms of composition but takes place in stages
- This eastward evolution suggests the involvement of lower continental crust and /or continental mantle lithosphere erosion over time until the inception of « typical » oceanic crust.

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We confirm that the manuscript has been read and approved by all named authors and that there are no other persons who satisfied the criteria for authorship but are not listed.

We further confirm that the order of authors listed in the manuscript has been approved by all of us.

We understand that the Corresponding Author is the sole contact for the Editorial process.

Maryline Moulin is responsible for communicating with the other authors about progress, submissions of revisions and final approval of proofs

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Declaration of interests

- ✓ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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