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► **To cite this version:**

A. Loureiro, P. Schnurle, F. Klingelhöfer, A. Afilhado, J. Pinheiro, et al.. Imaging exhumed lower continental crust in the distal Jequitinhonha basin, Brazil. *Journal of South American Earth Sciences*, 2018, 84, pp.351-372. 10.1016/j.jsames.2018.01.009 . hal-02328861

HAL Id: hal-02328861

<https://hal.science/hal-02328861>

Submitted on 12 Apr 2021

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Imaging exhumed lower continental crust in the distal Jequitinhonha basin, Brazil

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Abstract :

Twelve combined wide-angle refraction and coincident multi-channel seismic profiles were acquired in the Jequitinhonha-Camamu-Almada, Jacupe, and Sergipe-Alagoas basins, NE Brazil, during the SALSA experiment in 2014. Profiles SL11 and SL12 image the Jequitinhonha basin, perpendicularly to the coast, with 15 and 11 four-channel ocean-bottom seismometers, respectively. Profile SL10 runs parallel to the coast, crossing profiles SL11 and SL12, imaging the proximal Jequitinhonha and Almada

basins with 17 ocean-bottom seismometers. Forward modelling, combined with pre-stack depth migration to increase the horizontal resolution of the velocity models, indicates that sediment thickness varies between 3.3 km and 6.2 km in the distal basin. Crustal thickness at the western edge of the profiles is of around 20 km, with velocity gradients indicating a continental origin. It decreases to less than 5 km in the distal basin, with high seismic velocities and gradients, not compatible with normal oceanic crust nor exhumed upper mantle. Typical oceanic crust is never imaged along these about 200 km-long profiles and we propose that the transitional crust in the Jequitinhonha basin is a made of exhumed lower continental crust.

Highlights

► The transitional domain in the Jequitinhonha basin is, at least, 150 km wide. ► The transitional crust is composed of exhumed lower continental crust. ► Necking occurs within less than 100 km. ► An anomalous velocity zone is imaged at the base of the crust.

Keywords : NE Brazil, South Atlantic Ocean, Passive margins, Wide-angle refraction seismic, PSDM, Crustal structure, Cretaceous breakup, Lower continental crust

42 **1 Introduction**

43 The processes that led to the breakup of West Gondwana and the opening of the South Atlantic
44 Ocean are still not fully understood. One of the main hindrances for an accurate reconstruction of
45 West Gondwana is the lack of magnetic anomalies to establish a time-line for the oceanic crust-
46 spreading rate, as the breakup occurred during the Cretaceous Normal Superchron, chiefly in the
47 Central Segment of the South Atlantic Ocean (Moulin et al., 2010). The lack of magnetic anomalies
48 is counterbalanced by the presence of well-marked fracture zones and lineaments that, with the
49 knowledge of the intra-plate deformation on both Africa and South America, tightly constrain the
50 plate movements (Moulin et al., 2010; Aslanian & Moulin, 2012).

51

52 The SALSA experiment is aimed at constraining the crustal structure, the segmentation and the

53 geodynamical setting of the Camamu triple junction (Fig. 1), where the aborted Recôncavo –
54 Tucano – Jatobá rift system connects with the Jequitinhonha – Camamu-Almada and Jacuípe –
55 Sergipe-Alagoas rift systems. Here, the basins are set in extremely narrow margins, from less than
56 ~40 km to less than 100 km wide, and with very narrow continental shelves, which is quite rare in
57 passive margin settings (Dominguez et al., 2013). In this paper, we present wide-angle refraction
58 and coincident reflection data along two parallel profiles located on the Jequitinhonha basin (SL11
59 and SL12), extending approximately 170 km and 180 km from the continental shelf to the distal
60 Jequitinhonha basin, and a third profile (SL10), running approximately 270 km parallel to the coast
61 and crossing the Jequitinhonha and Camamu-Almada proximal basins (Fig. 1b). The conjugates of
62 these basins are the South Gabon and Congo basins (Fig. 1c). The remaining basins studied during
63 the SALSA project (Camamu, Jacuípe, Tucano and Sergipe-Alagoas basins) will be discussed in
64 companion papers.

66 2 Geological setting

67 The Jequitinhonha – Camamu-Almada rift system extends from the Royal Charlotte bank and
68 Cumuruxatiba basin, on the south, to the Barra and Itapuã faults, and the Recôncavo and Jacuípe
69 basins, on the north (Fig. 1). The Jequitinhonha basin borders the south-eastern margin of the São
70 Francisco craton, but the sampled basement is the Rio Pardo Group, low metamorphic
71 Neoproterozoic rocks linked to the Araçuaí Orogen that make up the Rio Pardo-Nyanga aulacogen
72 (Ledru et al., 1989; Gordon et al., 2012). This contrasts with the basement of the northern Camamu-
73 Almada basin, which is the Archean crust of the São Francisco craton (Schobbenhaus et al., 2003),
74 but this part of the craton is itself underlain by the Itabuna branch of the Paleoproterozoic Itabuna-
75 Salvador-Curaçá belt (Delgado et al., 2003)

76
77 During the disaggregation of Rodinia, the São Francisco and Congo cratons were never completely
78 detached and formed a cratonic bridge (Porada, 1989, Dias et al., 2016). Neoproterozoic rifting
79 reached the present-day Araçuaí orogen (Trompette, 1997), creating the Macaúbas-Jequitinhonha
80 basin, a gulf-like branch of the Adamastor Ocean with an undetermined extension of oceanic crust.
81 The main Cretaceous rift trends are strongly controlled by basement inheritance, with reactivation
82 of previous rift structures and fold belts from the Paleoproterozoic Itabuna – Salvador – Curaçá and
83 Neoproterozoic Araçuaí – West Congo orogens (Ferreira et al., 2013).

84
85 Rifting most likely started at the Camamu triple junction, failing at the Recôncavo – Tucano –
86 Jatobá rift system, and propagating southwards to the Camamu-Almada and Jequitinhonha basins.
87 Rifting started in the Sergipe-Alagoas basin at a later phase (Moulin et al., 2012; Chaboureau et al.,
88 2013). Intense fault activity occurred during the Early Aptian in the Jequitinhonha and Almada
89 basins, south of the Taipus-Mirim Accommodation Zone (TMAZ), but also on the extreme north of
90 the Camamu basin, forming isolated large grabens strongly bounded by basement lineaments. In the
91 Middle Aptian, the rift architecture changed with the formation of conspicuous N-S and NE-SW
92 hinge faults mostly concentrated in Camamu basin, north of the TMAZ. In the rest of the rift
93 system, fault activity decreased and thermal subsidence started (Ferreira et al., 2013).

94
95 The syn-rift sedimentary sequences in the Jequitinhonha and Camamu-Almada basins are bound by
96 two major regional discordances that, according to biostratigraphic data, indicate the rift phase
97 lasted about 30 Ma and ended in the Aptian/Albian transition (Küchle et al., 2005). This chronology
98 is confirmed in the Jequitinhonha basin (Rangel et al., 2007; Chaboureau et al., 2013), but disputed
99 in the Camamu-Almada basin, where the syn-rift sediments deposit from the Late Berriasian to the
100 Barremian/Aptian transition (Scotchman & Chiassi, 2009) or even to as early as the Middle Aptian
101 (Caixeta et al., 2007; Gontijo et al., 2007). In the Early to Middle Aptian there is a major hiatus in
102 the sedimentation with the formation of a Central Elevated Block (Chaboureau et al., 2013).

103
104 Salt composition, morphology and repartition in the northeastern Brazilian basins and their African

105 conjugates is quite varied (Chaboureau et al., 2013). Salt is absent in the Jacuípe basin, but there are
106 small anhydrite and halite deposits in the Camamu-Almada basin and larger deposits in the
107 Jequitinhonha basin. Salt deposits in their conjugate basins are larger and more homogeneous, but
108 with a different composition from the south of the Congo basin northwards. Here, they are
109 potassium-rich evaporites with a probable hydrothermal origin that would suggest a magmatically
110 active extensional environment (Hardie, 1990; Chaboureau et al., 2013). The same composition is
111 also found in the Sergipe-Alagoas basin, further north on the Brazilian coast, but in almost
112 inexpressive deposits.

113

114 **3 Data and Method**

115

116 **3.1 Seismic data**

117 The SALSA (Sergipe-Alagoas Seismic Acquisition) is a joint project of the Department of Marine
118 Geosciences (IFREMER: Institut Français de Recherche pour l'Exploitation de la MER, France)
119 and Petrobras, in collaboration with the Laboratory of "Oceanic Geosciences" (IUEM: Institut
120 Universitaire et Européen de la Mer, France), the Faculdade de Ciências da Universidade de Lisboa
121 (IDL, Portugal), and the Laboratório de Estudos da Litosfera, Universidade de Brasília (Brazil). The
122 mission was conducted on the French R/V L'Atalante between April and May 2014, with the
123 acquisition of twelve combined wide-angle refraction and high-resolution multi-channel seismics in
124 the northeastern coast of Brazil. In this study we present the results from the three southernmost
125 profiles, with emphasis on the two profiles that run perpendicular to the coast.

126

127 Profile SL11 was the southernmost profile of the SALSA mission (Fig. 1, Tab. 1), extending from
128 water depths of 629 m to 4136 m. Profile SL12 was parallel to SL11 (Fig. 1, Tab. 1), at a distance of
129 approximately 40 km, extending from water depths of 1118 m to 3992 m, with the loss of
130 instrument SL12OBS11. Both profiles cross profile SL10 (Fig. 1, Tab. 1), which runs parallel to the
131 coast, between 50 km and 100 km from the coastline. Profile SL12 was extended onshore with 21
132 seismometers (Fig. 1), but no in-line shots were recorded as the batteries of the acquisition systems
133 were already depleted at the time of shooting due to numerous marine mammal sightings that
134 severely delayed the operations.

135

136 The OBS were deployed at 7 nautical miles intervals, and were capable of recording on four
137 channels (1 hydrophone and a 3 component geophone). A 4.5 km long digital seismic streamer
138 ensured near-offset multi-channel seismic (MCS) recording of the shots with 360 hydrophones
139 (Figs. 2a, 2b and 2c). The seismic source was a tuned array of 16 air guns with a combined volume
140 of 6544 in³.

141

142 The reflection seismic data were pre-processed with the Geocluster (CGG-Veritas) software to
143 include geometry corrections, wave equation multiple attenuation, shot-gather predictive
144 deconvolution, time variant band-pass filter, and radon transform multiple attenuation. The OBS
145 data were pre-processed to include clock drift corrections, and location corrections due to drift from
146 the deployment position during their descent to the seafloor using the direct water wave arrival.
147 Data quality was generally very good on all instruments and channels, with clear arrivals to offsets
148 over 80 km on most instruments. Times of first and secondary arrivals were picked at their onset,
149 without filtering whenever possible, and with band-pass Butterworth filters in all other cases.

150

151 **3.2 Forward modelling**

152 The wide-angle OBS data were modelled using the RAYINVR (Zelt & Smith 1992; Zelt 1999)
153 software package using a layer stripping-approach and iterative damped least-squares travel-time
154 inversion at later stages. The starting parametrization for each layer was defined by screening all
155 instruments for the most important features, either strong reflections or clear turning waves and
156 critical refractions (Figs. 3 to 7). To avoid over-parametrization issues, only the interfaces

157 discernible in the OBS data were included in the models. Arrival times of near-offset reflections
158 were picked from the MCS data for the main sedimentary interfaces (Figs. 2a, 2b and 2c) that were
159 also identifiable on the OBS data. These arrival times were converted to depths using the
160 propagation velocities obtained from the OBS data. Depth and velocities of the crustal layers and
161 the upper mantle were modelled using exclusively the OBS data. Velocity gradients, relative
162 amplitudes, and cut-off points were constrained by comparison of synthetic seismograms with the
163 record sections (Figs. 3a and 3b, or 6a and 6b, for example). The final models (Fig. 8) show the
164 velocity field and interface geometries of all the main sedimentary layers and basement. On SL11
165 (Fig. 8c), eight sedimentary layers were modelled, reaching a total thickness of 3.3 km at the eastern
166 profile end and 4.1 km at the top of the continental slope. The propagation velocities increase
167 gradually inside the sedimentary column (Tab. 2), although with a slight decrease towards the distal
168 basin and only a small velocity inversion is identified, between Ps4 and Ps3. The velocities at the
169 top of the layers are very well-constrained, with the vertical velocity gradient established from cut-
170 off distances and relative amplitudes.

171

172 Five basement layers were modelled, based on the different refracted and reflected arrivals: upper
173 crust, middle crust, lower crust, anomalous velocity zone, and lithospheric mantle. The upper crust
174 seems to be present only on the continental slope, with a thickness of about 3.7 km, but the base of
175 the evaporites and transition to basement is not discernible in the data, which could mean that the
176 upper part of this layer may be a mixture of evaporites and sediments. The middle crust thins from
177 6.5 km on the continental slope to 1.5 km on the distal basin. Thinning is almost symmetrical with
178 respect to the top and base of the layer. The lower crust thins from 8.5 km on the continental slope
179 to 2 km on the distal basin. The anomalous velocity zone has a relatively constant thickness,
180 varying from 2.3 km to 1.5 km at the edges of the profile and a maximum thickness of almost 3 km
181 at the necking zone. The lithospheric mantle has a propagation velocity of 8.3 km/s 20 km below
182 the Moho to provide a gradient capable of explaining the observations. The thickness of the
183 "unthinned" continental crust at the western edge of the profile was set to approximately 21 km,
184 based on reflected arrivals at 20 km offset and on gravity modelling.

185

186 The model for SL11 is able to justify around 95% of the picked travel-times, and the uncertainty of
187 each pick was estimated according to its signal-to-noise ratio (SNR), as proposed by Zelt & Forsyth
188 (1994). The model is very well adjusted, with a normalized χ^2 value of 1.002 (Tab. 3).

189

190 On SL12 (Fig. 8b), eight sedimentary layers were modelled, reaching a total thickness of 4.1 km at
191 the eastern extremity of the profile and 6.2 km at the sub-basin at the western end of the profile. The
192 maximum sediments thickness is 9.1 km, at 0 km model distance, but the model has a greater
193 uncertainty in this region. The propagation velocities increase gradually inside the sedimentary
194 column (Tab. 4), although with a slight decrease towards the distal basin on the shallowest layers.
195 The deeper layers show a slight increase towards the distal basin.

196

197 The basement structure of SL12 has only four modelled layers, as no arrivals from an anomalous
198 velocity zone similar to that of profile SL11 were identified: upper crust, middle crust, lower crust
199 and lithospheric mantle. The upper crust has a thickness between 1.9 km and 2.0 km in the distal
200 basin and increases to 2.9 km at the base of the continental slope,. The middle crust thins from 6.0
201 km in the continental slope to 1.5 km in the distal basin. Thinning is mostly achieved on the base of
202 the layer, with the top of the layer almost horizontal. The lower crust is 5.5 km thick at the western
203 end of the model and completely thins out at 45 km model distance. The lithospheric mantle has a
204 propagation velocity of 8.3 km/s 10 km below the Moho to provide a gradient capable of explaining
205 the observations. The thickness of the lower crustal unit at the western edge of the profile is based
206 on the reflected arrivals identified in instruments SL12OBS10 and SL12OBS12 and on gravity
207 modelling.

208

209 The model for SL12 is able to justify around 95% of the picked travel-times, and the uncertainty of
210 each pick was estimated according to its SNR. The model is slightly over-adjusted, with a
211 normalized χ^2 value of 0.794 (Tab. 5).

212

213 The model for SL10 is able to justify about 93% of the picked travel-times, and is well adjusted,
214 with a normalized χ^2 value of 1.742 (Tab. 7). On SL10 (Fig. 8a), five sedimentary layers were
215 modelled, with an average total thickness of about 3 km, and reaching a maximum total thickness of
216 5.4 km at 125 km model distance. Ps5, that overlays the basement, is probably a mixture of
217 sediments and evaporites.

218

219 The basement of SL10 has four main layers: upper crust, middle crust, lower crust and lithospheric
220 mantle. The upper basement layer has a thickness of around 2.5 km. The middle unit has a thickness
221 that varies from 7.3 km at the model edges to around 3.2 km at 150 km model distance, with
222 thinning occurring at the top of the layer. The thickness of the lower crustal unit is approximately
223 constant at 5.3 km. The propagation velocity in the upper mantle increases to 8.20 km/s 10 km
224 below the Moho to provide a gradient capable of explaining the observations. Although the
225 anomalous velocity zone imaged on SL11 reaches the crossing with SL10, the poor ray coverage of
226 the upper mantle means that there are not enough observations to justify the inclusion of this feature
227 on SL10, keeping in line with the minimum structure approach, i.e., the final velocity model should
228 only contain the structure required by the data (Zelt & Smith, 1992).

229

3.3 PSDM

230 To verify the accuracy of the final velocity model, the MCS data of each profile were pre-stack
231 depth migrated (PSDM), and a residual move-out analysis was performed, using the Seismic Unix
232 package (Stockwell 1999; Cohen and Stockwell, 2015). A script is available in the supplementary
233 material. The final velocity layered model was converted to a 50 x 25 m spaced grid, and used to
234 compute travel-time tables regularly spaced at 150 m (the same spacing as the recorded shots) along
235 the profile by paraxial ray tracing. Solving the eikonal equation compensates travel-times in shadow
236 zones. The travel-time tables are used to calculate a common offset Kirchhoff depth migration.
237 Migrated traces are output as common image gathers (CIG) binned at 25 m, with 30 offset-classes
238 between 249 and 4596 m, spaced at 150 m. SALSA11 (Fig. 9a), SALSA12 (Fig. 9e), and SALSA10
239 (Fig. 9c) were migrated up to a depth of 18 km, showing very good resolution in the sedimentary
240 layers, and a good resolution in the crust, with good agreement between strong reflectors and their
241 wide-angle estimated depths.

242

243 Calculating the residual move-out (RMO) allows a dip-independent velocity analysis on the
244 migrated CIG. This implies that, if the velocity model is close to the true medium velocity, all
245 common offset migrated panels map the recorded seismic events to the same reflector depth. If the
246 velocity model significantly deviates from the true medium velocity, the move-out from near to far
247 offset translates into an interval velocity correction (Liu and Bleistein, 1995). Additionally, depth
248 migrated gathers are excellent records of amplitude variations with offset, and therefore are
249 indicators of in-situ rheological changes. The residual move-out behaviour coupled with the seismic
250 character from PSDM images are key elements to locate accurately major geological contacts,
251 moreover with higher horizontal resolution when compared to the OBS records.

252

253 The RMO analysis of the migrated SALSA11 (Fig. 9b), SALSA12 (Fig. 9f), and SALSA10 (Fig.
254 9d) sections shows mostly sub-horizontal arrivals, indicating a good agreement between the
255 modelled and true medium velocities in the entire sedimentary basin and upper crust. Coherence is
256 lost immediately below the basement, due to the arrival of the free surface multiples.

257

3.4 Gravity modelling

258 During the cruise, filtered gravity data were acquired every 10 s using a Lockheed Martin BGM-5

260 dynamic gravimeter installed as close as possible to the centre of gravity of the vessel. This
261 gravimeter additionally calculates the Eötvös correction, and the Free Air and Bouguer gravity
262 anomalies. The data were merged with the navigation data and corrected for instrumental drift using
263 the Caraïbes software, developed by Ifremer. Gravimeter drift was corrected using three
264 measurements at absolute gravity points in Maceió, São Paulo and Salvador. Outliers were
265 manually cleaned. The precision of the measurements is evaluated at 1 mGal.

266
267 The observed gravity in the Jequitinhonha basin is relatively smooth, with a positive anomaly on
268 the edge of the continental shelf, and a negative anomaly that starts at the location of profile SL10
269 towards the distal part of the basin (Fig. 1a). On the global satellite-derived gravity (Sandwell &
270 Smith, 2009; Sandwell et al., 2013, 2014), this negative anomaly is connected to a negative
271 anomaly concentric with the pronounced gravimetric high at the Royal Charlotte bank (Fig. 1a).

272
273 An estimate of crustal density can be obtained by correlating the acoustic wave propagation
274 velocities with rock densities (Christensen & Mooney, 1995). Although there is no absolute
275 relationship between acoustic propagation velocity and density, gravity modelling shows that the
276 seismic model is compatible with the measured gravity anomaly.. Areas of the model unconstrained
277 or poorly constrained by seismic data can be further constrained as a first order interpolation by
278 gravity modelling.

279
280 For profile SL11, we built a 2-D model consisting of 171 homogeneous density blocks, by
281 conversion of seismic velocity to density according to Ludwig et al. (1970). The density conversion
282 of our velocity model (Tab. 8) is able to predict the main trend of the gravity anomaly (Fig. 8c). The
283 regional trend was removed by varying the densities in the deep lithospheric mantle in the range
284 between 3330 kg/m^3 and 3345 kg/m^3 .

285
286 The densities for the upper mantle must be consistent with the geological setting, but in this
287 complex region, with Archean and Proterozoic units and at the northern limit of the Neoproterozoic
288 rifting, several hypotheses are possible. The modelled mantle densities are consistent with: a
289 moderately depleted Archean Sub-continental lithospheric mantle (SCLM), that can have mean
290 densities as low as 3310 kg/m^3 ; a depleted Proterozoic mantle, with mean densities as low as 3330
291 kg/m^3 ; (Poudjom Djomani et al., 2001); or with a very reworked and possibly enriched
292 Paleoproterozoic serpentized forearc mantle (Chaves et al., 2016). Present-day lightly
293 serpentized forearc mantles have mean densities starting at 3200 kg/m^3 (Hyndman & Peacock,
294 2003).

295
296 The largest difference between observed and calculated gravity anomaly, 13.0 mGal (Fig. 8c),
297 occurs close to the continental slope where thinning is more pronounced and seismic coverage is
298 poorer. The calculated gravity anomaly is also consistent with the satellite-derived gravity values
299 observed on parallel profiles extracted north and south of the profile .

300
301 For profile SL12, we built a 2-D model consisting of 106 homogeneous density blocks. The density
302 conversion of our velocity model (Tab. 8) is able to predict the main trend of the gravity anomaly
303 (Fig. 8b). The density of the mantle is consistent with the geological setting.

304
305 The largest difference between observed and calculated gravity anomaly occurs close to the western
306 edge of the model where the maximum difference reaches 9.6 mGal (Fig. 8b). The calculated
307 gravity anomaly is well within the values observed on parallel profiles extracted north and south of
308 the profile, derived from satellite gravity measurements.

309
310 Gravity modelling for profile SL12 is also consistent with the existence of the lower crustal unit in
311 the proximal basin. Without it, the calculated gravity anomaly does not match the data.

312
313 For profile SL10, we built a 2-D model consisting of 92 homogeneous density blocks, but precise
314 gravity modelling is hindered by the presence of salt, sometimes mixed with sediments, chiefly in
315 the Camamu-Almada basin, with differences between modelled and observed values up to 50 mGal.
316 In the region between profiles SL11 and SL12, the largest difference is under 20 mGal (Fig. 8a). ..
317 On the southern end of the profile, ray coverage is very poor in the upper mantle, and we cannot
318 exclude the contribution of the anomalous velocity zone imaged on SL11. To avoid boundary
319 effects the velocity model is extended north and south before calculating the gravity anomaly, but
320 without taking into account the change of topography, bathymetry and crustal thickness. The mass
321 deficit that is apparent in the gravity anomaly of SL10 is due to the regional influence of the thick
322 continental crust of the São Francisco craton in the north and the Royal Charlotte bank in the south.
323 Load anomaly in all profiles is within acceptable ranges for a crust in isostatic equilibrium
324 (Whitmarsh et al. 1996).

325 326 **3.5 Crossings with SL10**

327 To avoid operator and interpretation bias, profile SL10 was modelled independently of profiles
328 SL11 and SL12, by another operator, using different tools for data picking and filtering. SL10 runs
329 parallel to the coast and images the crust perpendicularly to the major structural features. Some
330 differences are expected, chiefly for the deepest interfaces and foremost on those constrained by
331 wide-angle reflections. However, when comparing the 1D velocity-depth profiles extracted at the
332 crossing points, the most noticeable discrepancy is the total crustal thickness of SL11 and SL10
333 (Figs. 10b and 10c). SL11 has a 8 km thick crust, while the crust on SL10 reaches a thickness of 9
334 km. This is explained by the inclusion of the anomalous velocity layer seen in SL11 in the lower
335 crust of profile SL10. The direction of shooting of SL10 and the presence of salt makes it very
336 difficult to model this feature.

337 338 **4 Model evaluation**

339 340 **4.1 Indirect model evaluation**

341 Models are evaluated not only by the number of justified observations and global data fit, but also
342 by the uniformity and density of their ray coverage, smearing, resolution and the number of rays
343 that constrain each node (hitcounts).

344
345 All three models are globally well covered with rays, and well-constrained from arrival times.
346 Localized defocusing effects caused by the salt layer are apparent in the lower crustal layers of
347 profile SL11. The resolution is generally very good (Fig. 11). Some sedimentary layers show lower
348 resolution, but they were constrained using the near-offset reflection data.

349
350 All layers show hit-counts larger than 2000, on profiles SL10 and SL11, and larger than 1000, on
351 profile SL12, indicating that the velocities are well constrained. Hit-counts for interface depths are
352 larger than 1000, on profiles SL11 and SL12, and larger than 2000, on profile SL10, indicating a
353 good constraint also on interface depths and topographies. Smearing is low on all three models,
354 suggesting that they are not over-parameterized.

355 356 **4.2 Uncertainty estimation using VMONTECARLO**

357 VMONTECARLO (Loureiro et al., 2016) was applied on the crustal layers of the final models of
358 SL10, SL11 and SL12, using the Metropolis algorithm and adaptive variance to increase the
359 convergence. The first quality threshold used to establish the model ensemble (ME) was set to twice
360 the preferred model's quality of fit (in terms of RMS, number of traced rays and χ^2). Two million
361 random models were generated for each profile.

362
363 For SL11, 16,289 models were at least as good as the preferred model, in terms of number of

364 justified observations and data fit, and used to build a global uncertainty map. . The global
365 uncertainty map for profile SL11 (Figs. 12c and 12d) shows that the model is generally well
366 constrained, apart from very localized defocusing effects caused by the salt layers. Only the deeper
367 crustal layers admit a broader range of velocity values while still satisfying the global data fit, but
368 the ray coverage in these regions is poor. Interface depths are also well determined.

369
370 A direct result of the VMONTECARLO uncertainty estimation is the output of a model that fits the
371 dataset better than the preferred model. In this case, the best model is capable of tracing 8097 rays
372 (justifying 97.0% of the data), with an RMS of 0.069 s and a χ^2 of 1.930. The original model had
373 justified 7910 observations, had a χ^2 of 3.193 and an RMS of 0.089 s. Even if the best model has a
374 better data fit and justifies more observations than our preferred model, the best model (dashed lines
375 in Figs. 12a and 12b) does not deviate significantly from the preferred solution (solid lines in Figs.
376 12a and 12b). This means that the preferred model is already a good solution.

377
378 For SL12, we used VMONTECARLO with the same options as for SL11. The first quality
379 threshold used to establish the ME was set to twice the preferred model's quality of fit (in terms of
380 RMS, number of traced rays and χ^2). SL12 has less data points (6490 observations) than SL11
381 (8348 observations), but with a better quality of fit, measured in both terms of RMS and χ^2 . Given
382 the current data interpretation, the model solution also seems to be more unique, as some arrivals
383 are constrained to very specific features.

384
385 For SL12, only 6 models were at least as good as the preferred model, in terms of number of
386 justified observations, RMS and χ^2 value, indicating that the preferred model is a very good
387 solution. If parameters are relaxed to match the quality of fit of model SL11 (the same RMS and χ^2
388 values) but requiring at least the same number of traced rays as the preferred model, this number
389 increases to 26,043 models (1.3% of the ME). However, this subset of the ME does not include the
390 best random model.

391
392 To generate an uncertainty map comparable to that of SL11 (Figs. 12c and 12d), the ME of SL12
393 was filtered to show only models with a χ^2 of 3.193 and an RMS of 0.089 s (the same values used
394 for SL11). The ME was further filtered to show only models capable of tracing at least 5,500 rays to
395 use a similar number of models on both profiles. The final ME had 15,324 models.

396
397 The global uncertainty map for profile SL12 (Figs. 12e and 12f) shows that the central part of a
398 model is generally well constrained. The lower crustal units in the necking zone, sparsely-covered
399 by rays, admit very different values while keeping the global fit at acceptable levels. Interface
400 depths are well constrained.

401
402 The best model found was capable of tracing 5852 rays (justifying 90.2% of observations), with an
403 RMS of 0.067 s and a χ^2 of 1.808. The original model traced 5886 rays, had a χ^2 of 2.035 and an
404 RMS of 0.071 s. It too, did not deviate significantly from the preferred model (dashed and solid
405 lines in Figs. 12c and 12d), indicating that the preferred model is already a very good solution.

406
407 For SL10, 2,528 models were capable of fitting the data with a χ^2 value better than 3.193 (the same
408 value used for SL11), an RMS up to 0.179 s, and capable of tracing at least 85% of the preferred
409 model's traced rays. The global uncertainty map for profile SL10 (Figs. 12a and 12b) shows that the
410 crustal layers are generally well constrained, with some noticeable defocusing effects caused by the
411 presence of salt. The velocities of the upper mantle have higher uncertainties. The best random
412 model does not deviate significantly from the preferred solution and is capable of tracing 11,961
413 rays, with an RMS of 0.129 s and a χ^2 of 1.673. It has a better fit than the preferred model for SL10,
414 but it justifies less observations.

415

416 1D slices of these global uncertainty maps, coloured according to model score, were used to study
417 the profile crossings. We find that each of the preferred models is compatible with the uncertainty
418 bounds of the profile it crosses (Fig. 10), and generally close to the random models with higher
419 scores.

420

421 **4.3 Comparison with GXT**

422 SALSA11 was shot along the track of a still not published seismic acquisition line, profile ION
423 GXT-1900. This dataset provides additional constraints to the sedimentary basin and crustal
424 structure. On the line drawing of this profile there are several deep crustal reflectors and what
425 appears to be a highly reflective, undulating Moho. Due to the spacing of the instruments, refraction
426 data are typically unable to retrieve small reflector wavelengths, but the final velocity model of
427 SL11 is capable of bounding the major units identified in the ION GXT-1900 profile with a very
428 good agreement (Fig. 13).

429

430 **5 Discussion**

431 Profiles SL11 and SL12 were aimed at imaging the necking zone and the deep Jequitinhonha basin.
432 Profile SL10, running parallel to the coast, was aimed at imaging the necking zone and proximal
433 Jequitinhonha and Camamu-Almada basins.

434

435 The interpretation of the SL11 and SL12 wide-angle models allowed us to identify two main
436 regions (Figs. 15 and 16). The first region is the necking zone, between 0 km and ~50 km model
437 distance, where the continental crust thins from 19-23 km to about 13 km. The second region, from
438 50 km model distance to the end of profiles SL11 and SL12, is the deep basin.

439

440 The southernmost region of profile SL10 shows a crustal structure of a thinned continental crust
441 nature (Fig. 17) that is consistent with what is observed on profiles SL11 and SL12, at the crossing
442 with this profile. North of the crossing with SL12, already in the Camamu-Almada basin, it neither
443 shares features with typical continental crust nor typical oceanic crust, but those of a transitional
444 domain that is discussed next.

445

446 In profile SL12, a third crustal unit is present, modelled with lower propagation velocities than the
447 anomalous velocity zone of profile SL11. This layer is constrained by reflections at both the top and
448 base (Figs. 6 and 7), and is also required to improve the fit in the gravity anomaly model.

449

450 The unthinned continental crust is not imaged by any of the presented profiles, however, the width
451 of the first region, the necking zone, can be estimated less than 100 km, taking into account a total
452 crustal thickness of between 37 km and 41 km (Assumpção et al., 2013) taken at approximately 50
453 km inland.

454

455 The second region, the deep basin, has a crust that cannot be classified as having an oceanic nature,
456 as the propagation velocities, gradients and layer thicknesses are inconsistent with typical Atlantic
457 oceanic crust of comparable age (Figs. 15 and 16). Two crustal layers are present, with an upper
458 layer that reaches a maximum thickness of 3 km and has high seismic propagation velocity (above
459 5,5 km/s) and a strong gradient. The Moho is also marked by a sharp increase in velocity and clear
460 PmP reflections. Two hypotheses can be made for the nature of the crust in this region: exhumed
461 mantle exhumed lower continental crust.

462

463 **5.1 Exhumed mantle**

464 Continental mantle exhumation and subsequent serpentization has been proposed as an
465 explanation for the formation of the ocean-continent transitional crust at the later phases of the
466 rifting process in the Atlantic Ocean, mainly based on the very specific Galician margin studies
467 (Boillot et al., 1987; Brun & Beslier, 1996; Manatschal, 2001; Lavier & Manatschal, 2006).

468

469 To test the hypothesis of mantle exhumation, the 1D velocity-depth profiles extracted along the
470 profiles were compared to similar profiles extracted from wide-angle seismic models (Figs. 18a,
471 18b and 18c) in the Newfoundland Grand Banks margin, Iberia Abyssal Plain, and southern
472 Galician margin, where the ocean-continent transition zones have been interpreted as exhumed
473 mantle, serpentized upper mantle, or a mixture of both. In the southern Galician margin
474 serpentized upper mantle and lower continental crust materials were drilled during Legs 173 and
475 149 of the Ocean Drilling Program (Chian et al., 1999), with the lower continental crust materials
476 being interpreted as rafts that float in an overall upper mantle exhumation regime that occurs along
477 a deep detachment (Boillot et al., 1987; Manatschal, 2001). The 1D velocity-depth profiles are also
478 compared a compilation of mean velocity-depth variations in the ocean-continent transition (Fig.
479 18d - Minshull, 2009) of profiles in the previously compared regions and the UK western
480 approaches margin, where an anomalously high V_p/V_s ratio indicates the presence of strongly
481 serpentized rocks (Bullock & Minshull, 2005).

482

483 In the southern Iberian Abyssal Plain (IAM-9 profile, Fig. 18a), a zone inferred to be exhumed
484 upper mantle has a thickness of 5 km divided into two layers (Dean et al., 2000). An upper, high
485 velocity-gradient layer, 2 km to 4 km thick, interpreted to be upper mantle serpentized peridotite
486 (with a mean bulk serpentization from 25% to 100%) along faults, which may have been
487 subsequently locally intruded by the products of decompression melting in the mantle, and a lower
488 layer up to 4 km thick with a P wave propagation velocity between 7.3 km/s and 7.9 km/s,
489 representing mantle peridotite with a mean bulk serpentization of < 25%, possibly concentrated
490 along fewer, but steep cutting faults. On the IAM-9 profile, reflections from the Moho, representing
491 the serpentization front, are weak in the wide-angle seismic data and absent in the normal incident
492 data.

493

494 On the eastern Grand Banks of Newfoundland, profile SCREECH-2 (Fig. 18b) shows a ~25 km-
495 wide zone with a smooth basement, characterized by velocities higher than 6.3 km/s, increasing to
496 7.7 km/s at a depth of 5 km. This region is interpreted, from analogies with the Iberian Margin, as
497 being exhumed upper mantle material (van Avendonk et al., 2006). Similar zones are much wider
498 (60 – 160 km) on SCREECH-3 (Lau et al., 2006) and also on the Iberia Abyssal Plain (Dean et al.,
499 2000).

500

501 However, in the southern Galician margin, where upper mantle materials were drilled, the crustal
502 structure was imaged by the CAM wide-angle profile (Minshull et al., 1998; Chian et al., 1999),
503 showing a continuous increase of velocity with depth and absence of intra-basement interfaces (Fig.
504 18c).

505

506 The velocity structure of the Jequitinhonha basin is noticeably different from that of these profiles,
507 where the basement is presumably formed of exhumed mantle. When compared to the other three
508 areas, the Jequitinhonha basin has an upper layer with much higher velocities and lower gradients,
509 and there are clear reflections from the Moho seen in both wide-angle and near offset seismic
510 sections that are associated with strong velocity contrasts. These reflections at the base of the lower
511 layer are absent in the other three profiles. These differences clearly demonstrate that the basement
512 of the Jequitinhonha basin cannot be formed of exhumed mantle.

513

514 **5.2 Exhumed lower continental crust**

515 Between thinned continental crust and pure oceanic crust, the presence of exhumed lower
516 continental crust has been proposed for other segments of the South Atlantic Ocean (Moulin et al.;
517 2005; Aslanian et al., 2009; Klingelhöfer et al., 2015; Evain et al., 2015; Moulin et al., 2016) and
518 the Mediterranean Sea (Afilhado et al., 2015; Moulin et al., 2015).

519

520 To test the exhumed lower continental crust hypothesis, we compared the 1D velocity-depth profiles
521 from the Jequitinhonha basin with those extracted from wide-angle seismic models in the Santos
522 basin (SANBA experiment) and Gulf of Lion and Provençal basin (Sardinia experiment), where the
523 ocean-continent transition zones have been interpreted to be exhumed lower crust. Bott (1971),
524 Aslanian et al. (2009) have proposed that this lower continental crust may feed the first oceanic
525 accretion process

526
527 The 1D velocity-depth profiles extracted along SL11 and SL12 are comparable to those from
528 regions where exhumed lower continental crust is proposed, such as the Provençal basin (Fig. 18e),
529 where the gradients and thicknesses are similar; and the Santos basin (Fig. 18f), where the
530 propagation velocity gradients are also very similar, but the crust is thicker. There are also only two
531 crustal layers and a marked velocity step at the top of the mantle (Figs. 18e and 18f) in both regions.
532 These observations lead us to infer that this domain is most likely comprised of exhumed lower
533 continental crust.

534
535 The Almada section of profile SL10 shows similarities with the proximal part of the exhumed lower
536 continental crust domain of profiles SL11 and SL12, most notably in terms of upper crust thickness
537 and velocity gradient in the lower crust. The closeness to the necking zone may explain most of the
538 differences found between profile SL10 and the profiles that run perpendicularly to the coast. Layer
539 thicknesses and propagation velocities at the top of the crust are comparable to those of the thinner
540 parts of the exhumed continental crust domain of the Santos basin (Fig. 18f), as well as the
541 propagation velocities at the top of the basement.

542 543 **5.3 Boudin-like feature**

544 A specific feature of the southern-most profile (SL11) is an anomalous velocity zone that underlays
545 the lower crust. The P-wave propagation velocities in this relatively thin layer are over 7.6 km/s,
546 making the reflection at the top of this layer a candidate seismic Moho (Prodehl et al., 2013), but it
547 is much more reflective than the mantle, indicating a crustal nature (Cook et al., 2010; Hammer &
548 Clowes, 1997). The internal reflectivity is apparent in the GXT-1900 line drawing (Fig. 13), but also
549 in the PSDM (Fig. 2b) and MCS sections (Fig. 9a).

550
551 The anomalous velocity zone bounds an high-amplitude undulated reflector identified in cyan in the
552 line-drawing of profile ION GXT-1900 (Fig. 13). This reflector shares similarities with the “M-
553 reflector”, proposed by Blaich et al. (2010) for the Camamu-Almada basin. The wavy character of
554 deep reflectors is also found in the conjugate margin, for example in profile PROBE23 (Rosendahl
555 et al., 1991). In the Middle South Gabon Basin, a similar undulating reflector is also identified, the
556 “G-reflector” (Dupré et al., 2011). The “G-reflector” appears also at approximately 10 s twt, and is
557 interpreted as the transition between an upper crust and the result of mafic underplating of the
558 continental crust that has been thinned during extension. A similar anomalous velocity zone, with P-
559 wave seismic propagation velocity exceeding 7.7 km/s, is also present in the Upper South Gabon
560 basin, on profile SPOG2 (Wannesson et al. 1991).

561
562 The same feature is not modelled on profile SL12, but as the maximum reciprocal offsets are
563 smaller in this profile, due to the smaller number of deployed instruments, it may not be possible to
564 accurately image this feature if it exists. Additionally, profile SL12 is located on the transition
565 between the Jequitinhonha and Almada basins, near the Olivença basement high, which could mark
566 a different basement fabric. However, taking into account the estimated uncertainties for this model
567 (Figs. 12a and 12b), an anomalous velocity zone similar to the one modelled on profile SL11 cannot
568 be excluded.

569
570 A possible interpretation for the anomalous velocity layer is that the lower continental crust, as it is
571 thinned and/or exhumed, begins to flow (Bott, 1971; Buck et al., 1999; Aslanian et al., 2009) and

572 form lithospheric-sized boudins (Fig. 14). A similar mechanism has also been proposed for the
573 Uruguayan margin (Clerc et al., 2015). In this process, part of this crust is enriched with mafic and
574 ultramafic mantle materials, which explain the increase of the seismic propagation velocity. Such
575 intrusions of mantle-derived melts into the lower crust have been proposed in the Paleoproterozoic
576 block of the Ukrainian Shield, where a layer with seismic propagation velocity of 7.6 km/s is
577 observed above the Moho (Thybo et al., 2003; Thybo & Artemieva, 2013). As the anomalous
578 velocity zone is not apparent on profile SL12, other possibilities are that this feature is linked to the
579 Miocene Abrolhos volcanism (intruded upper mantle), connected to upper mantle foliation due to
580 plate movement along the Bode Verde Fracture Zone, or even indicative of boudinage of the deepest
581 parts of the continental crust induced by a transfer zone that is expressed by the Royal Charlotte
582 Bank.

583

584 **6 Conclusions**

585 The Jequitinhonha basin is set on a narrow margin, with the complete necking of the continental
586 crust occurring within less than 100 km.

587

588 The typical oceanic crust was not reached in profiles SL11 and SL12. Taking this into account and
589 the limit of the necking zone, the transitional domain in the Jequitinhonha basin is, at least, 150 km
590 wide, which is in agreement with the transitional domain proposed for the same area by Blaich et al.
591 (2008). Seismic velocity gradients seem to rule out a purely continental origin and Moho reflections
592 discard a serpentinized upper mantle provenance. The transitional domain is instead most probably
593 composed of exhumed continental lower crust. This is in accordance with the hypothesis of
594 Aslanian et al. (2009), who proposed a transitional crust in this region.

595

596 On profile SL11 there is an anomalous velocity zone that underlays the lower crust and bounds
597 high-amplitude undulating reflectors. The seismic propagation velocities and the geometry of the
598 reflectors suggest the existence of a boudin-like feature probably linked to the crustal thinning, with
599 oceanward creep of the lower crust.

600

601 In the Jequitinhonha basin, the upper continental crust completely thins out at the limit of the salt
602 province (Fig. 19). The middle and lower crusts are exhumed in the distal basin, with shearing and
603 oceanward creep of the deepest parts of the lower crust. Ultramafic intrusions in the lower crust are
604 also possible.

605

606 **Acknowledgements**

607 We thank two anonymous reviewers for their excellent suggestions, and their constructive and
608 exhaustive reviews that significantly improved this manuscript. We are indebted to the captain,
609 crew, and MCS technical team of the R/V L'Atalante. We also thank J. Crozon, P. Fernagu, P.
610 Pelleau and M. Roudaut, the OBS technical team, and J. Soares, R. Fuck, M. Lima, J. Pinheiro, F.
611 Lima, M. Sobrinho, P. Resende, R. Oliveira, N. Dias, C. Corela, J. Duarte, D. Alves and L. Matias,
612 the land stations deployment team. The GMT (Wessel & Smith 1998), Seismic Unix (Stockwell
613 1999; Cohen and Stockwell, 2015), and Geocluster (CGG-Veritas) software packages were used
614 extensively in the preparation of this paper. Processing of the high resolution seismic data was done
615 by A. Baltzer, M. Benabdellouahed, and M. Rabineau. The geodynamic interpretations were done
616 by D. Aslanian and M. Moulin. R. Fuck acknowledges CNPq, Brazil research fellowship and INCT
617 Estudos Tectônicos research grant.

618

619 The dataset collected during the SALSA experiment is protected under a partnership with Petrobras.
620 Any request has to be addressed to Daniel Aslanian (aslanian@ifremer.fr) and Adriano Viana
621 (aviana@petrobras.com.br).

622

623 **Contributions**

624 The SALSA Project was led by D. Aslanian and M. Moulin, from Ifremer, and A. Viana, from
 625 Petrobras. Processing of the high resolution seismic data was done by A. Baltzer, M.
 626 Benabdellouahed, and M. Rabineau. The geodynamic interpretations were done by D. Aslanian and
 627 M. Moulin. Modelling of profiles SL11 and SL12 was done by A. Loureiro. Modelling of profile
 628 SL10 was done by F. Klingelhöfer. Profile SL09 was modelled by A. Afilhado. F. Gallais modelled
 629 profiles SL07 and SL08. M. Evain modelled profiles SL05 and SL06. Profiles SL01 and SL02 were
 630 modelled by J. M. Pinheiro. Modelling of profiles SL03 and SL04 and processing of the deep-
 631 sounding reflection seismic data was done by P. Schnürle.

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Figure 1: Location of profiles of the SALSA mission. a) satellite-derived gravimetric anomaly in the sea (Sandwell & Smith, 2009; Sandwell et al., 2013, 2014) and topography on land (IHO-IOC GEBCO, 2014). Circles denote OBS deployments. Triangles denote land stations. Black solid lines indicate MCS sections. Dashed black line denotes faults. Dashed white lines indicate marine basin limits (Bizzi et al., 2003). Orange area indicates the Recôncavo-Tucano-Jatobá (RTJ) basins (Bizzi et al., 2003). Light blue area indicates the limits of the São Francisco Craton, adapted from (Hasui, 2012). TMAZ: Taipus-Mirim Accommodation Zone. Main river indicated by solid blue lines (Wessel & Smith, 1996). b) Bathymetry (IHO-IOC GEBCO, 2014) around the profiles discussed in the present paper. c) General location map of studied area and conjugate margin.

Figure 2: MCS record sections for: a) SALSA10, b) SALSA12, and c) SALSA11. Solid coloured lines indicate the modelled interfaces. White triangles denote OBS locations. Red vertical line indicates crossings with other profiles.

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Figure 3: Instrument SL11OBS05. a) Record section. b) Synthetic record section. c) Synthetic record section with arrivals coloured according to their identified phases. d) Picked arrivals coloured according to phase, their error bars, and solid black lines indicating the predicted arrival times. e) Ray tracing over the model. f) Comparison with the MCS section, coloured lines match the reflected arrivals colours from d) and colour scale from Fig. 2. All time sections reduced to 7 km/s.

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Figure 4: Instrument SL11OBS09. a) Record section. b) Synthetic record section. c) Synthetic record section with arrivals coloured according to their identified phases. d) Picked arrivals coloured according to phase, their error bars, and solid black lines indicating the predicted arrival times. e) Ray tracing over the model. f) Comparison with the MCS section, coloured lines match the reflected arrivals colours from d) and colour scale from Fig. 2. All time sections reduced to 7 km/s.

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Figure 5: Instrument SL11OBS13. a) Record section. b) Synthetic record section. c) Synthetic record section with arrivals coloured according to their identified phases. d) Picked arrivals coloured according to phase, their error bars, and solid black lines indicating the predicted arrival times. e) Ray tracing over the model. f) Comparison with the MCS section, coloured lines match the reflected arrivals colours from d) and colour scale from Fig. 2. All time sections reduced to 7 km/s.

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Figure 6: Instrument SL12OBS12. a) Record section. b) Synthetic record section. c) Synthetic record section with arrivals coloured according to their identified phases. d) Picked arrivals coloured according to phase, their error bars, and solid black lines indicating the predicted arrival times. e) Ray tracing over the model. f) Comparison with the MCS section, coloured lines match the reflected arrivals colours from d) and colour scale from Fig. 2. All time sections reduced to 7 km/s.

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Figure 7: Instrument SL12OBS09. a) Record section. b) Synthetic record section. c) Synthetic record section with arrivals coloured according to their identified phases. d) Picked arrivals coloured according to phase, their error bars, and solid black lines indicating the predicted arrival times. e) Ray tracing over the model. f) Comparison with the MCS section, coloured lines match the reflected arrivals colours from d) and colour scale from Fig. 2. All time sections reduced to 7 km/s.

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Figure 8: Final velocity models for profiles: a) SL10, b) SL12, and c) SL11. Thick blue lines indicate interfaces constrained by wide-angle reflections. Shaded areas indicate ray coverage. Observed (dashed black lines) and calculated (solid red lines) gravity anomalies under each profile.

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Figure 9: Pre-stack depth migrated MCS record sections for: a) SALSA11, c) SALSA10, and e) SALSA12. RMO analysis of the pre-stack depth migrated MCS record sections of: b) SALSA11, d) SALSA10, and f) SALSA12. Solid coloured lines indicate the modelled interfaces. White triangles denote OBS locations. Red vertical line indicates crossings with other profiles.

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Figure 10: Comparison of 1D velocity depth profiles at the crossings of SL11 (black solid lines) and SL12 (purple solid lines) with SL10 (blue dashed line). Coloured regions show uncertainties estimated with VMONTECARLO for a) SL12; b) SL11; c) SL10 at the crossing with SL11; d) SL10 at the crossing with SL12. Different colours indicate normalized random model scores.

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Figure 11: Model resolution for all velocity nodes of: a) SL11, b) SL12, and c) SL10. Gray and yellow areas are considered well resolved in terms of velocities. Squares indicated well-resolved depth nodes in the basement.

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Figure 12: a) and b): Model uncertainties for profile SL10, using 5116 models that were capable of tracing at least 85% of the rays of the preferred model, with a maximum residual of 210 ms and a χ^2 value of 4.4. c) and d): Model uncertainties for profile SL11, using 16 289 models that were at least as good as the preferred model. e) and f) Model uncertainties for profile SL12, using 15 324 models that were at least as good as the preferred model. Red shades [a), c) and d)] denote maximum admissible velocity increases from the preferred solution. Blue shades [b), d) and f)] denote maximum admissible velocity decreases from the preferred solution. Yellow bands indicate interface depth uncertainties resulting from the increase of propagation velocity [a), c) and e)], and from the decrease of propagation velocity [b), d) and f)].

Figure 13: Overlay of SL11 model (dashed lines) to line-drawing of ION GXT-1900 (solid lines). Blue arrow marks crossing with profile SL10. Inset shows the coincident tracks of SL11 and ION GXT-1900 profiles.

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Figure 14: Cartoon for the boudinage of the lower continental crust as it is stretched and enriched during the rifting process.

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Figure 15: 1D velocity-depth profiles below the basement of SL11, taken at 10 km intervals. Interfaces identified by thick coloured lines have a velocity contrast over 0.1 km/s. Interfaces identified by thin grey lines have velocity contrast under 0.1 km/s. Blue regions indicate a compilation of 1D velocity-depth profiles from (White et al., 1992) for oceanic crusts in the Atlantic Ocean aged 59 Ma to 127 Ma and 142 Ma to 170 Ma. Pink regions indicate a compilation of continental crusts (Christensen & Mooney, 1995). Green 1D profiles mark thinned continental crust, orange profiles mark exhumed lower crust.

Figure 16: 1D velocity-depth profiles below the basement of SL12, taken at 10 km intervals. Interfaces identified by thick coloured lines have a velocity contrast over 0.1 km/s. Interfaces identified by thin grey lines have velocity contrast under 0.1 km/s. Blue regions indicate a compilation of 1D velocity-depth profiles from (White et al., 1992) for oceanic crusts in the Atlantic Ocean aged 59 Ma to 127 Ma and 142 Ma to 170 Ma. Pink regions indicate a compilation of continental crusts (Christensen & Mooney, 1995). Green 1D profiles mark thinned continental crust, orange profiles mark exhumed lower crust.

Figure 17: 1D velocity-depth profiles below the basement of SL10, taken at 10 km intervals. Blue regions indicate a compilation of 1D velocity-depth profiles from White et al. (1992) for oceanic crusts in the Atlantic Ocean aged 59 Ma to 127 Ma and 142 Ma to 170 Ma. Pink regions indicate a compilation of continental crusts (Christensen & Mooney, 1995). Red and green profiles are from the thinned continental crust domain, purple profiles are from the transitional domain. Black profiles are from both domains.

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Figure 18: Comparison between 1D basement velocity-depth profiles of SL11 and SL12, extracted every 10 km after the necking zone (orange lines), of SL10, in the Camamu-Almada basin (purple lines), and from wide-angle seismic models where ocean-transition zones were interpreted as exhumed and/or serpentized upper mantle (shaded areas): a) IAM-9 in Iberia Abyssal Plain (Dean et al., 2000); b) SCREECH-2 in the Grand Banks margin, offshore Newfoundland (van Avendonk et al., 2006); c) CAM in the Southern Galician margin (Chian et al., 1999); d) Mean velocity-depth variations with one standard deviation above and below the mean of SCREECH-1, SCREECH-2, SCREECH-3, WAM, ISE-1, CAM-144 and IAM-9 models from Minshull (2009). Same comparison, but where ocean-transition zones were interpreted as exhumed lower continental crust (shaded areas) e) Sardinia profiles in the Provençal Basin (Moulin et al., 2015); f) SanBa Experiment in Santos Basin (Evain et al., 2015).

Table 1: Details of profiles SL10, SL11 and SL12.

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Table 2: Vp Propagation velocities of SL11, in km/s. Velocities are specified at the top and bottom of each layer for the proximal and distal parts of the basin.

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Table 3: Statistics for model SL11.

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Table 4: Vp Propagation velocities of SL12, in km/s. Velocities are specified at the top and bottom of each layer for the proximal and distal parts of the basin.

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Table 5: Statistics for model SL12.

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Table 6: Vp Propagation velocities of SL10, in km/s. Velocities are specified at the top and bottom of each layer.

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Table 7: Statistics for model SL10.

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Table 8: Densities converted from seismic propagation velocities for SL10, SL11 and SL12, according to Ludwig et al. (1970).

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Profile	Maximum offset (km)	OBS	No. of shots
SL10	230	15	1686
SL11	180	15	1161
SL12	215	12 (1 lost)	1399

Table 1: Details of profiles SL10, SL11 and SL12.

Layer	Proximal	Distal
Ps1	1.8	
	1.9	
Ps2	2.2	2
Ps3	2.6	
	2.7	
Ps4	2.8	2.6
	2.9	2.7
Ps5	3	2.7
	3	2.8
Ps6	4.1	3.1
	4.2	3.2
Ps7	4.4	4.3
	4.5	4.7
Ps8	4.9	4.7
	5	
C1	5.2	-
	5.3	-
C2	5.6	5.8
	6.1	6.4
C3	6.4	6.3
	6.8	6.8
AVZ	7.6	
	8.1	
M1	8.1	8.2
	8.3	

Table 2: Vp Propagation velocities of SL11, in km/s. Velocities are specified at the top and bottom of each layer for the proximal and distal parts of the basin.

Phase	No. of Points	RMS (s)	χ^2
Pw	5538	0.042	0.739
Ps2P	513	0.043	0.113
Ps2	633	0.077	1.357
Ps3P	417	0.026	0.045
Ps3	209	0.038	0.922
Ps4P	248	0.014	0.034
Ps4	72	0.048	0.695
Ps5P	365	0.038	0.08
Ps6P	76	0.019	0.011
Ps6	226	0.046	1.753
Ps7P	257	0.036	0.126
Ps7	288	0.037	0.942
Ps8P	444	0.05	0.188
Ps8	916	0.048	0.859
Pg1P	165	0.075	0.91
Pg1	469	0.044	0.496
Pg2P	530	0.052	0.372
Pg2	446	0.098	2.872
Pg3P	258	0.06	0.631
Pg3	1739	0.106	2.616
Pu1P	994	0.103	2.209
Pu1	808	0.113	1.776
PmP	902	0.086	0.679
Pn	2288	0.072	0.511
Global	18801	0.068	1.002

Table 3: Statistics for model SL11.

Layer	Proximal	Distal
Ps1	1.8	
	1.9	
Ps2	2.3	2
	2.4	2.1
Ps3	2.6	
	2.7	
Ps4	3.8	3.6
	3.9	3.7
Ps5	3.8	
	3.9	
Ps6	4.1	4.3
	4.3	4.4
Ps7	4.4	4.5
	4.4	4.7
Ps8	5.1	
	5.2	
C1	5.2	
	5.3	
C2	6.7	7
	7.2	7.2
C3	7.2	-
	7.3	-
M1	8	
	8.3	

Table 4: Vp Propagation velocities of SL12, in km/s. Velocities are specified at the top and bottom of each layer for the proximal and distal parts of the basin.

Phase	No. of Points	RMS (s)	χ^2
Pw	3454	0.041	0.734
Ps2P	194	0.032	0.04
Ps2	288	0.031	0.391
Ps3P	414	0.042	0.304
Ps3	163	0.076	1.223
Ps4P	267	0.083	0.505
Ps4	331	0.072	1.561
Ps5P	162	0.047	0.168
Ps6P	27	0.009	0.004
Ps6	226	0.06	1.367
Ps7P	274	0.065	0.307
Ps7	70	0.04	1.358
Ps8P	153	0.083	1.456
Ps8	540	0.051	0.98
Pg1P	240	0.056	0.457
Pg1	767	0.058	0.972
Pg2P	407	0.085	1.106
Pg2	2826	0.073	0.937
Pg3P	990	0.08	1.102
PmP	37	0.066	0.251
Pn	1304	0.057	0.371
Global	13134	0.061	0.794
PmP	902	0.086	0.679
Pn	2288	0.072	0.511
Global	18801	0.068	1.002

Table 5: Statistics for model SL12.

Layer	Almada-Camamu	Jequitinhonha
Ps1	1.9	
	2	
Ps2	2.3	
	2.4	
Ps3	2.6	2.4
	2.8	2.7
Ps4	3.8-4.2	4.2
	4.3	
Ps5	4.3	
	4.8	4.4
C1	5	5.3
	5.7	5.6
C2	5.8	
	6	
C3	6.6	
	7	
M1	8.1	
	8.2	

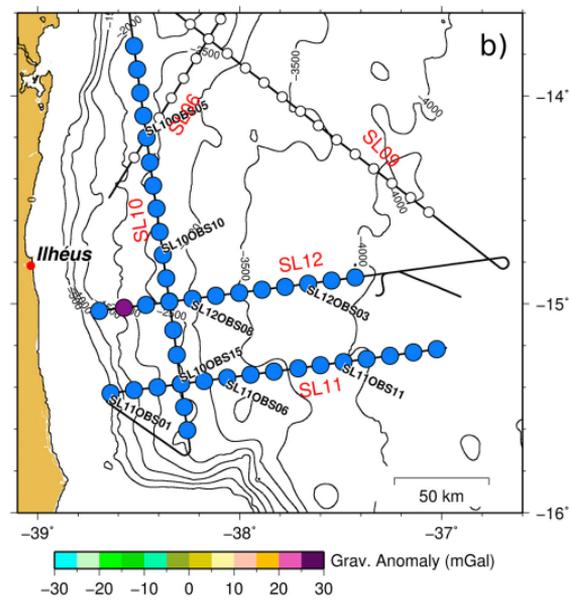
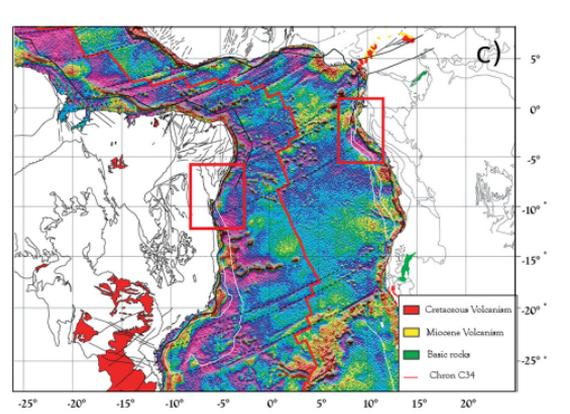
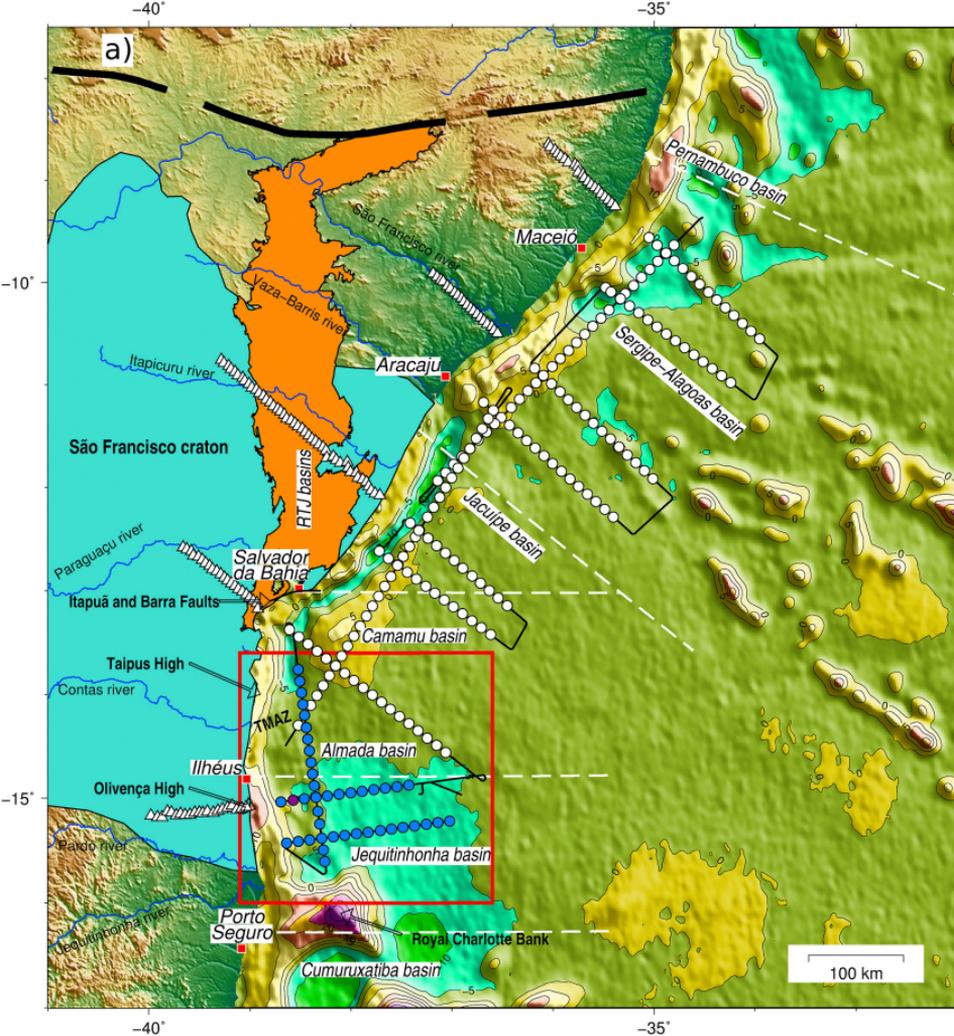
Table 6: Vp Propagation velocities of SL10, in km/s. Velocities are specified at the top and bottom of each layer.

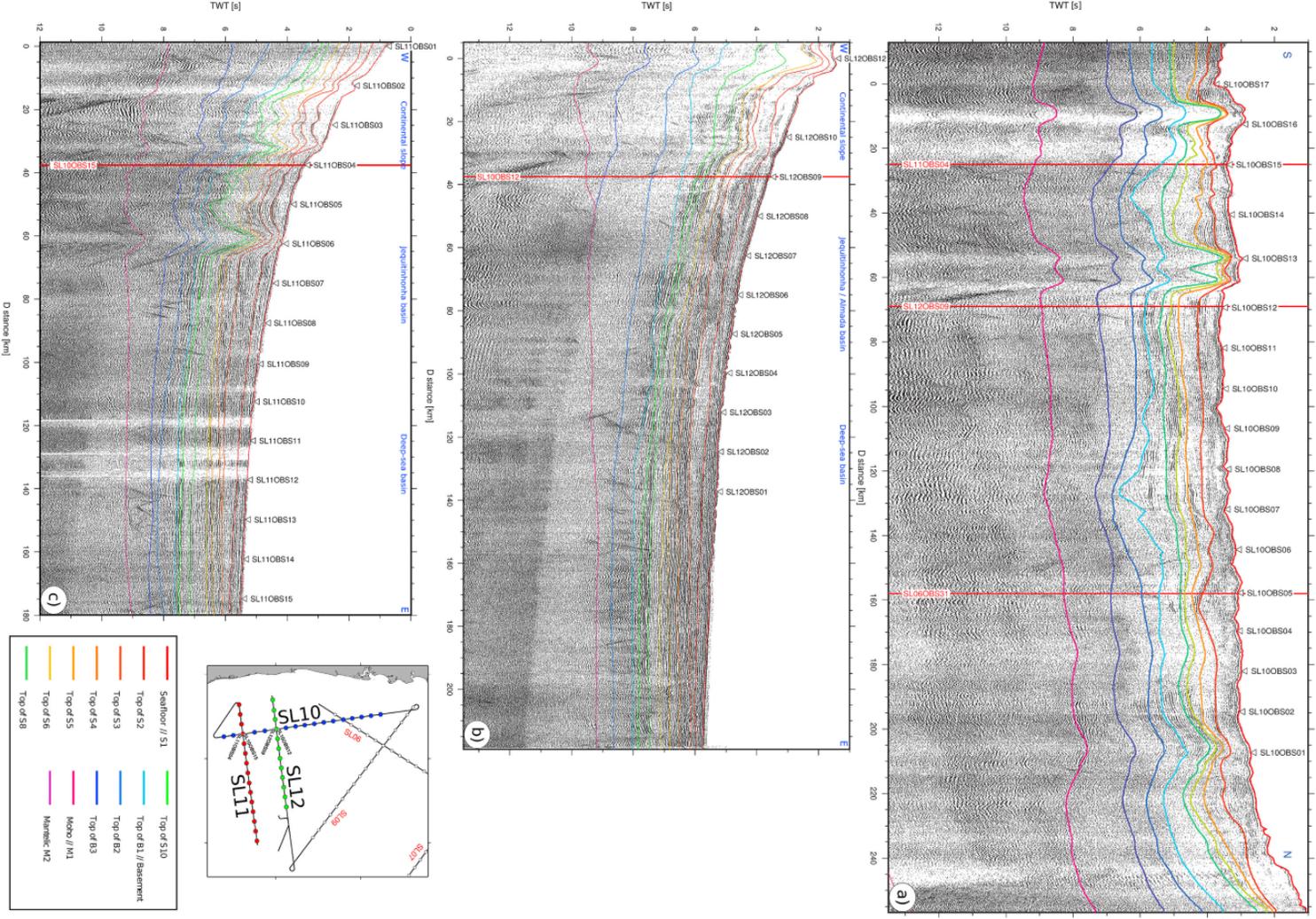
Phase	No. of Points	RMS (s)	χ^2
Pw	1032	0.012	0.015
Ps1P	236	0.085	0.726
Ps1	399	0.141	1.981
Ps2P	517	0.087	0.751
Ps2	51	0.244	6.064
Ps3P	341	0.143	2.048
Ps3	816	0.178	3.183
Ps4P	87	0.13	1.699
Ps4	232	0.048	0.236
Pg1P	1299	0.138	1.916
Pg2	2577	0.157	2.457
PmP	1803	0.136	1.857
Pn	3579	0.122	1.484
Global	12969	0.132	1.742

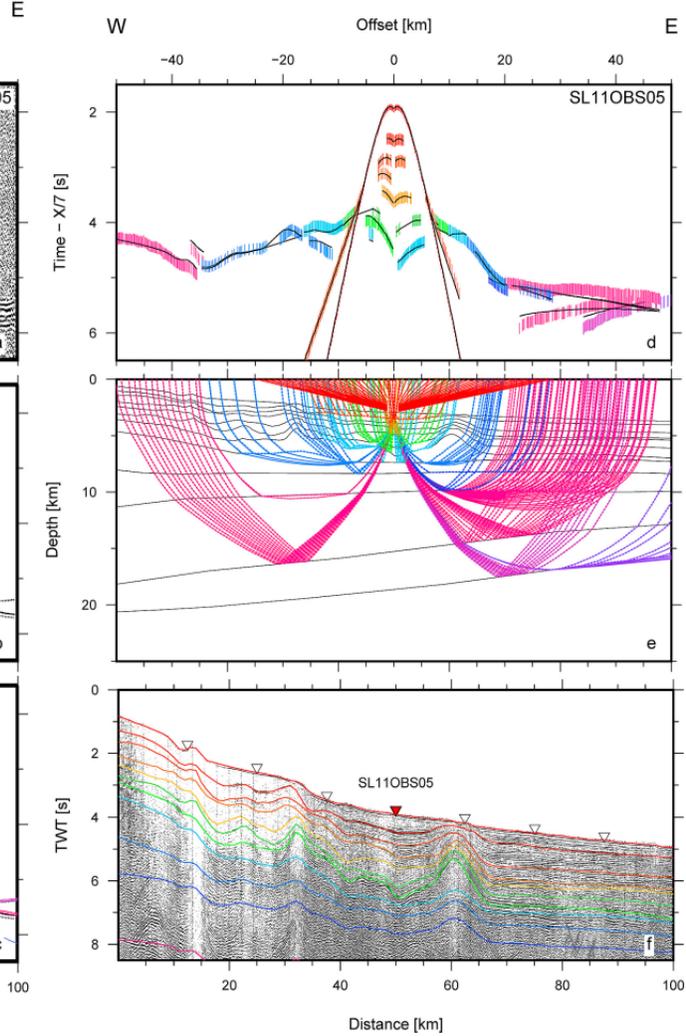
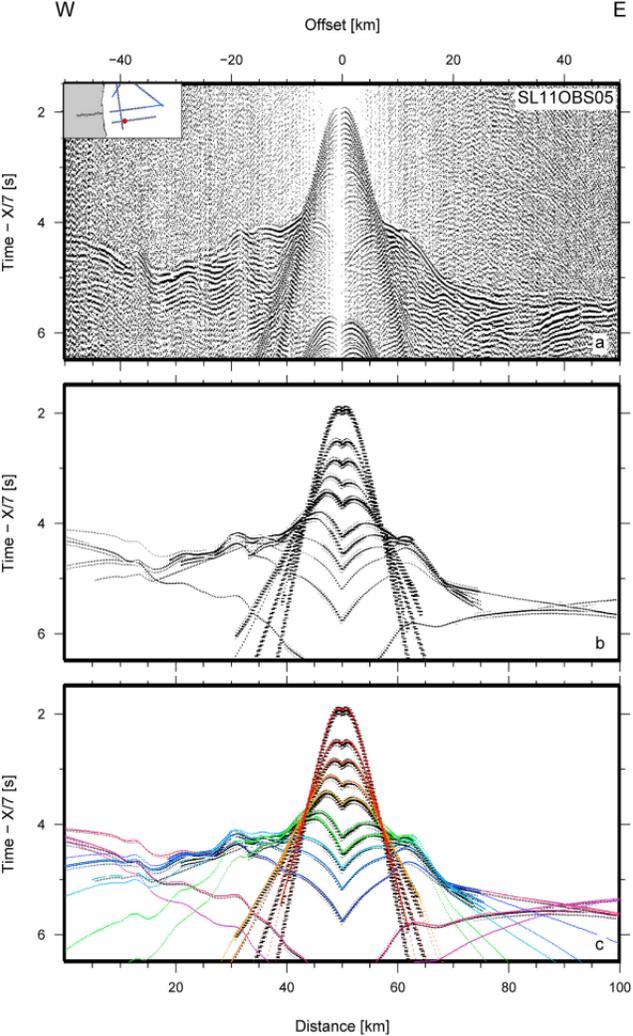
Table 7: Statistics for model SL10.

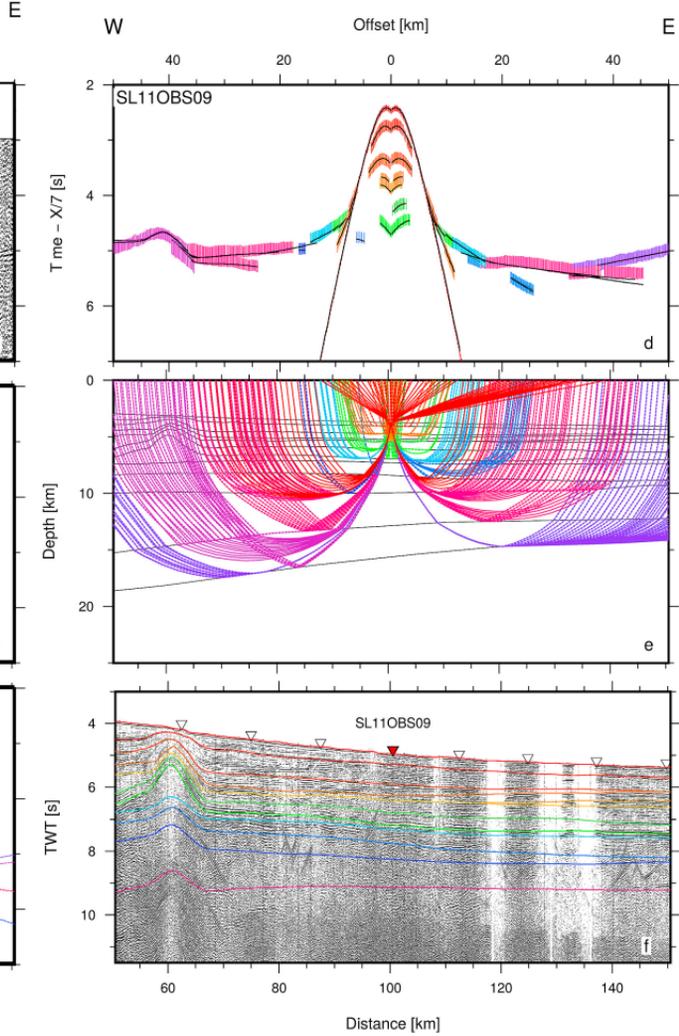
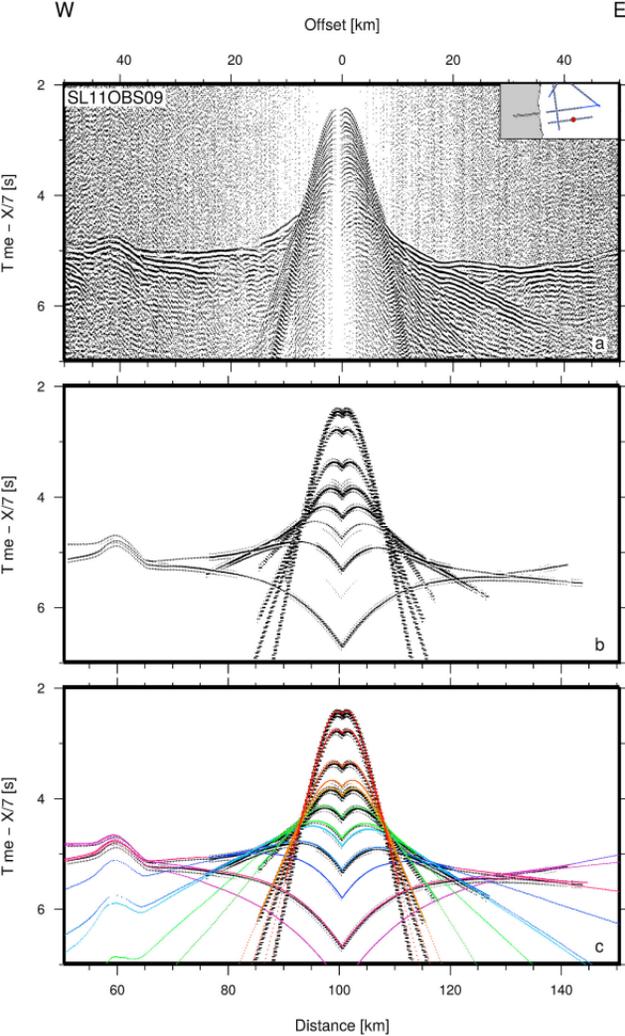
Domain	SL10	SL11	SL12
Sediments	2200 – 2580	2200 – 2520	2200 – 2540
Crust	2660 – 2880	2530 – 3250	2680 – 3030
Upper Mantle	3330	3320 – 3330	3320 – 3330

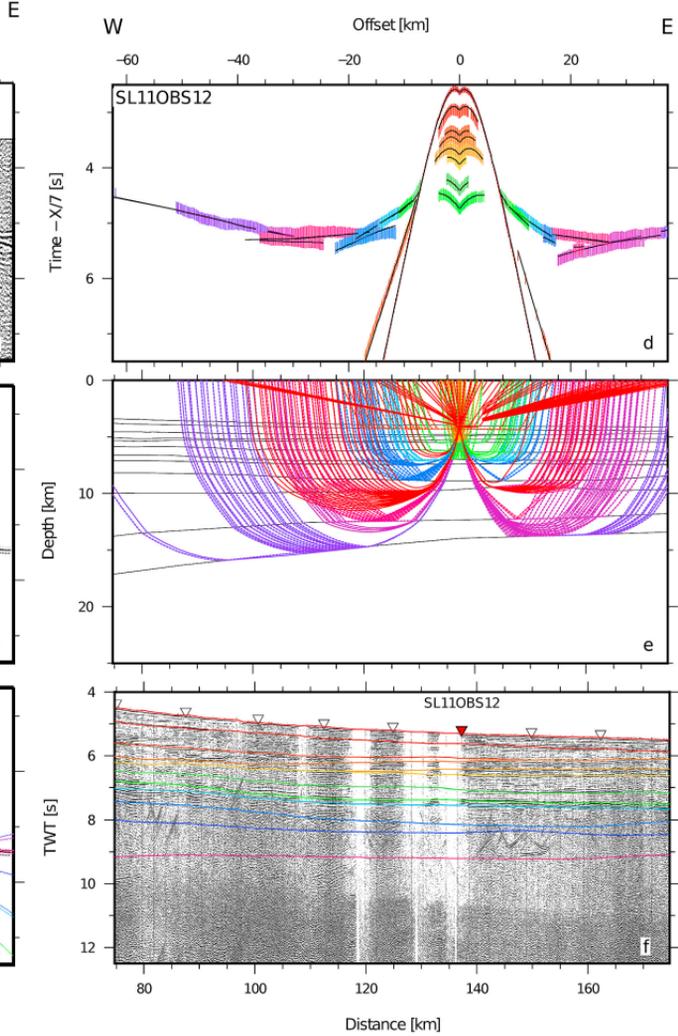
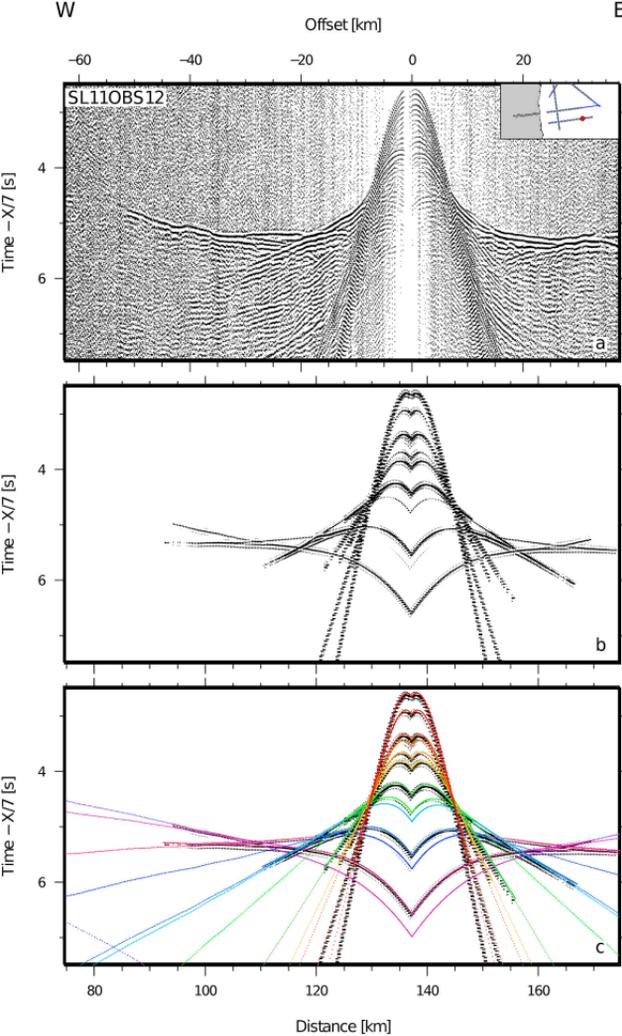
Table 8: Densities converted from seismic propagation velocities for SL10, SL11 and SL12, according to Ludwig et al. (1970).

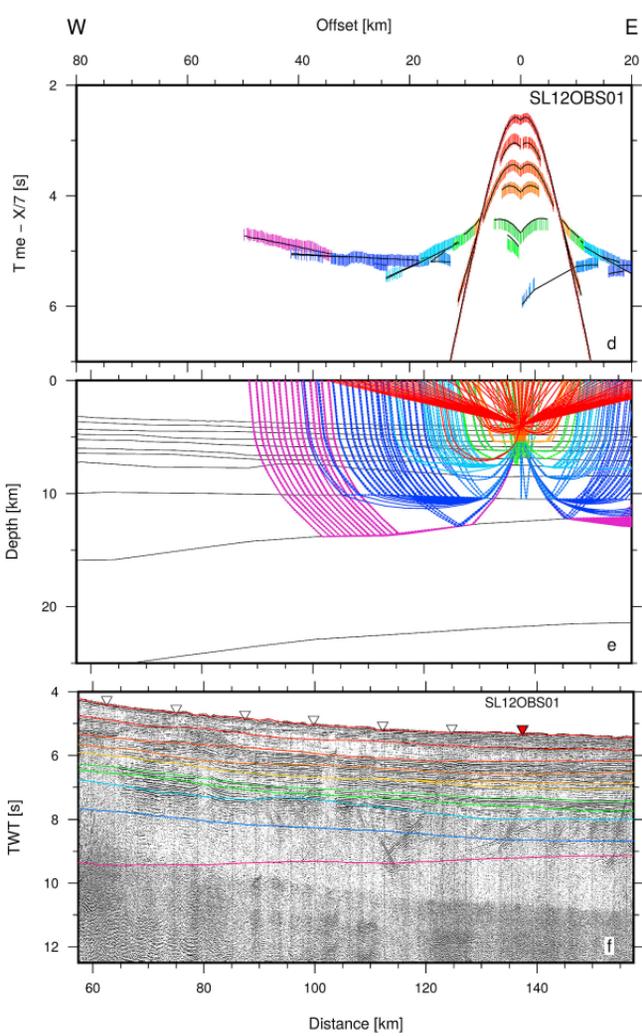
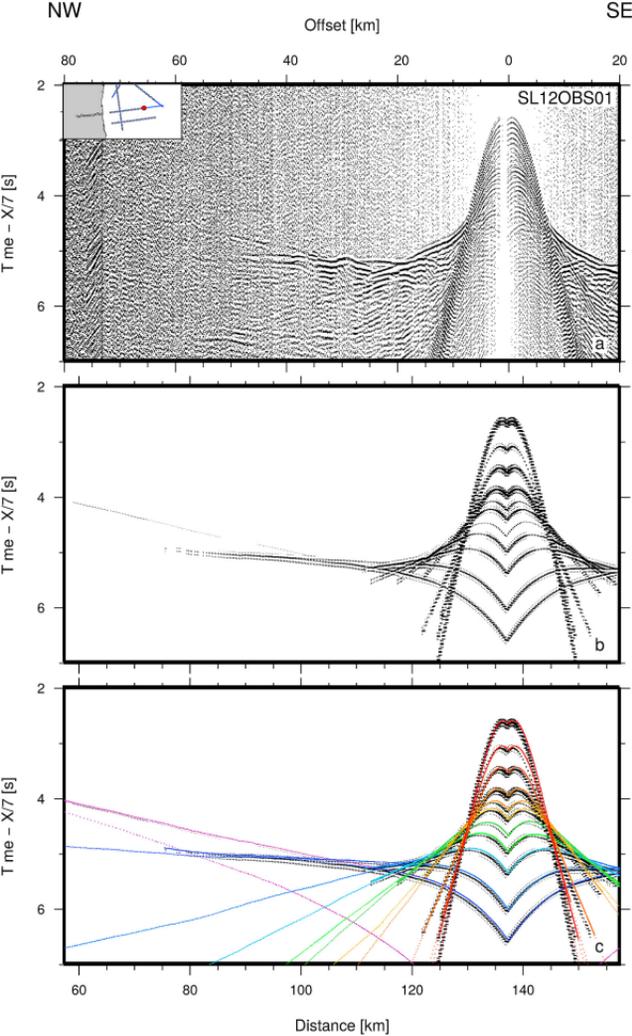


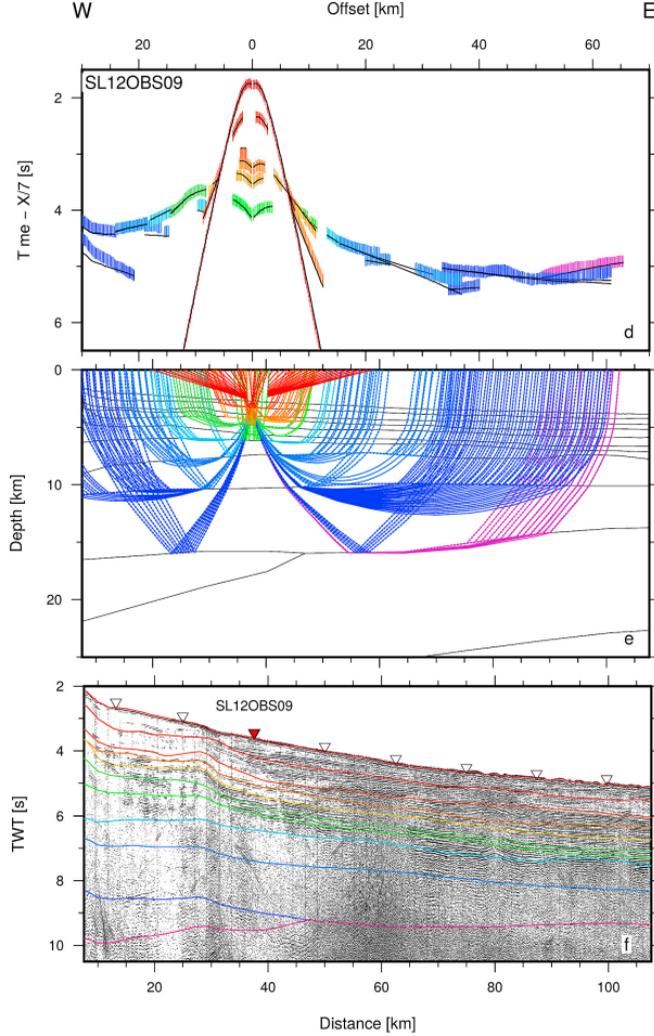
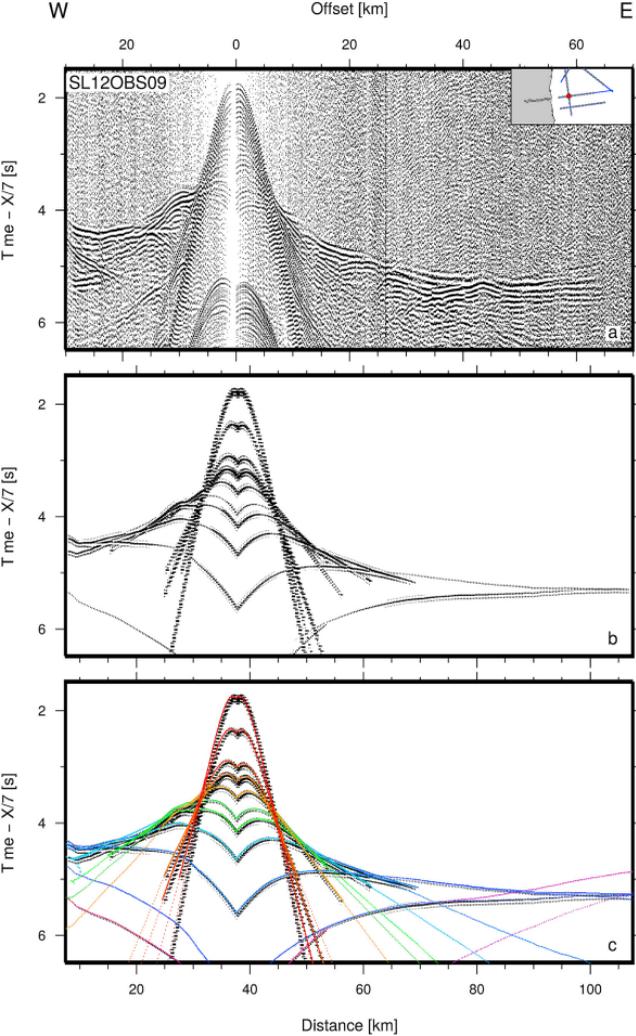


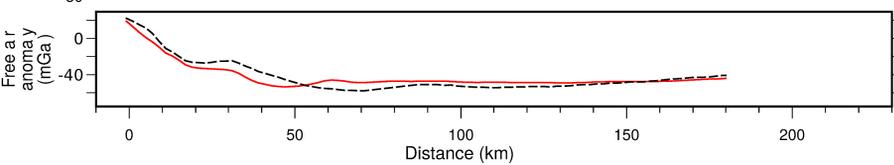
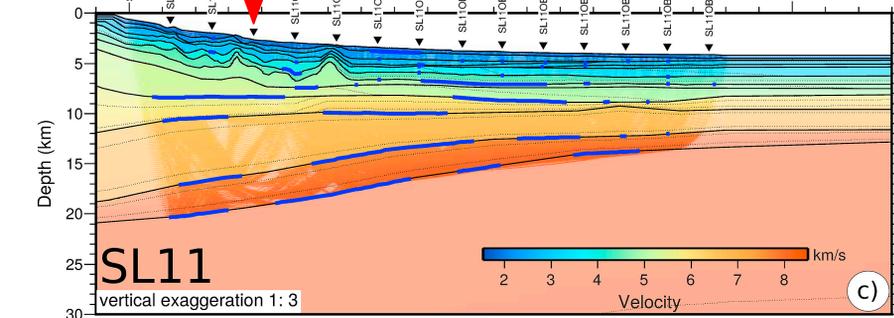
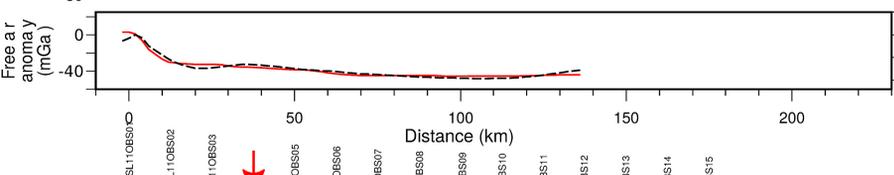
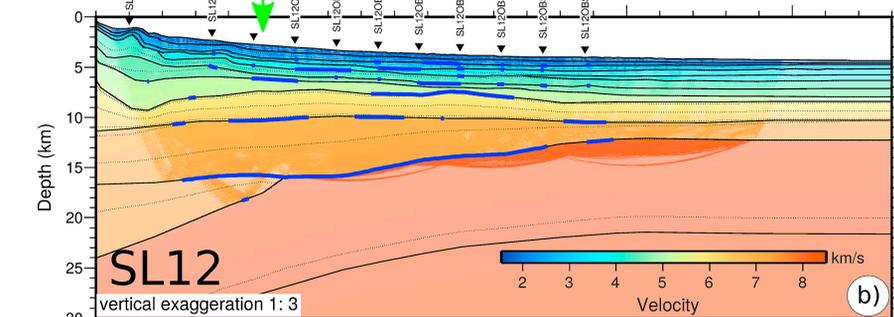
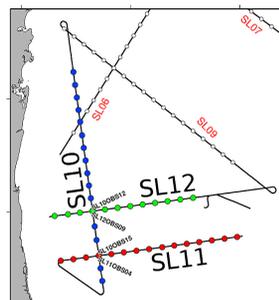
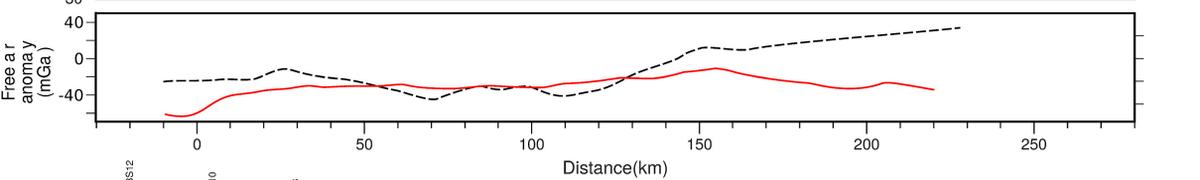
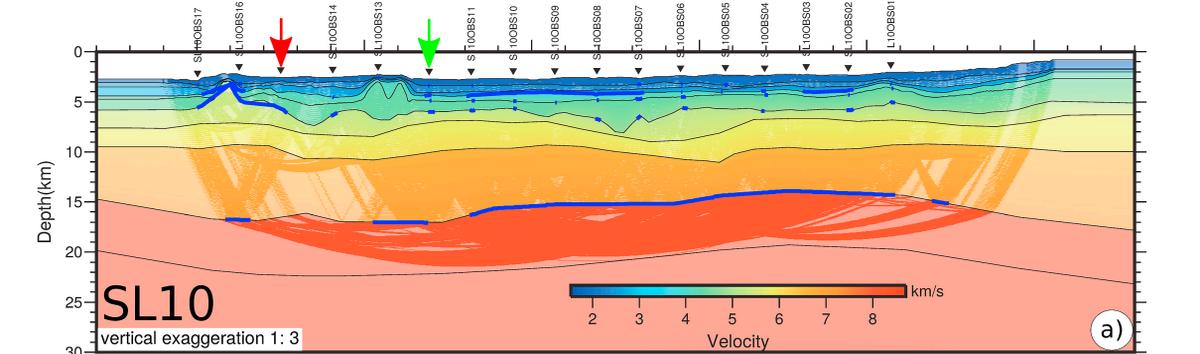


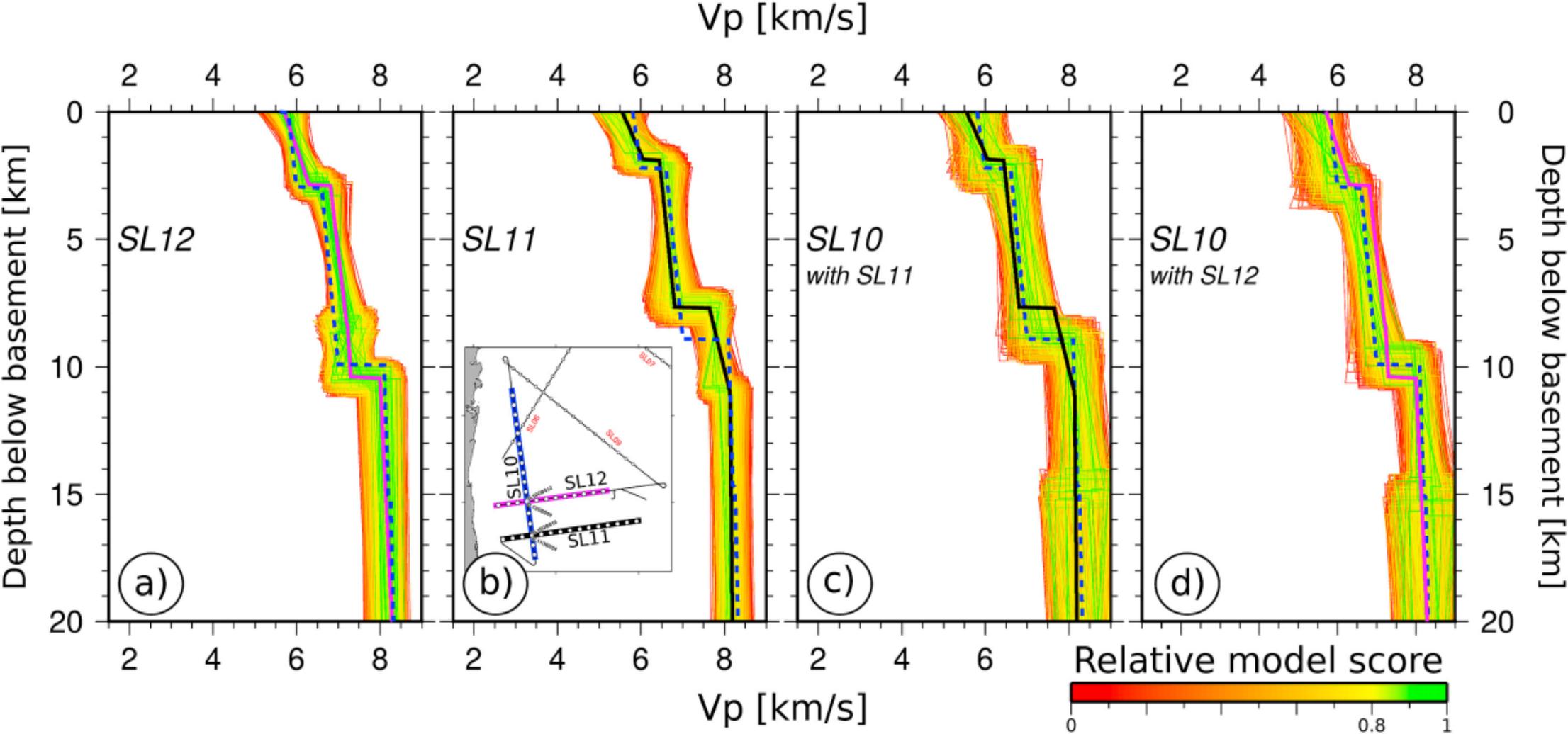


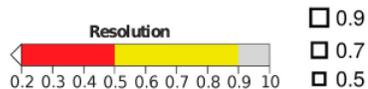
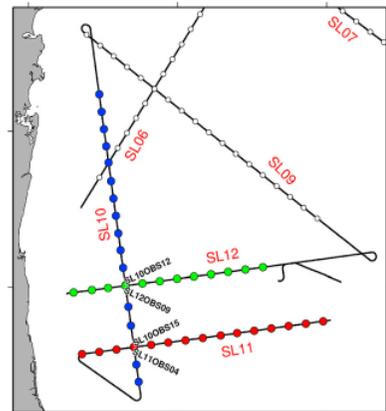
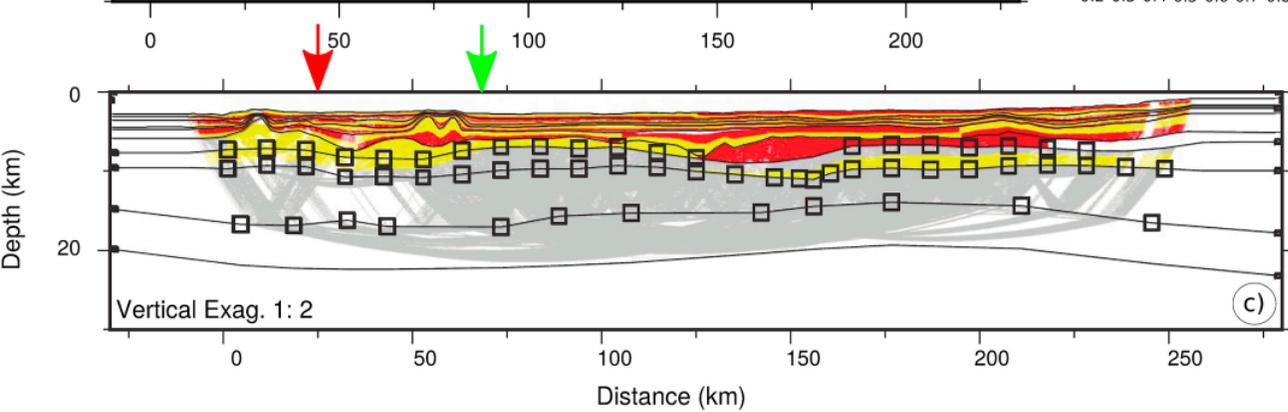
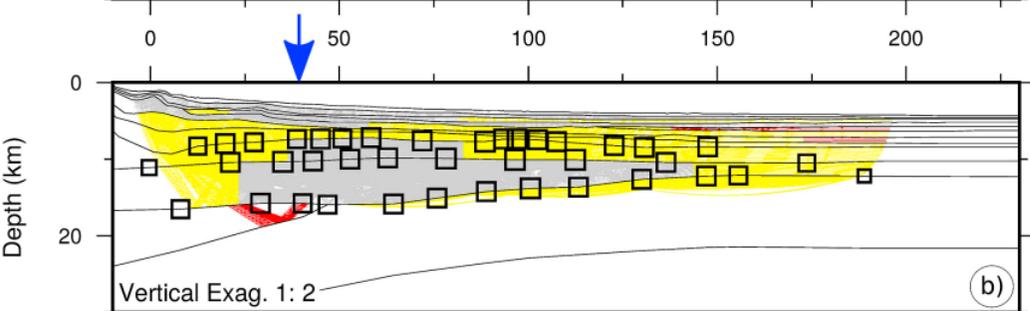
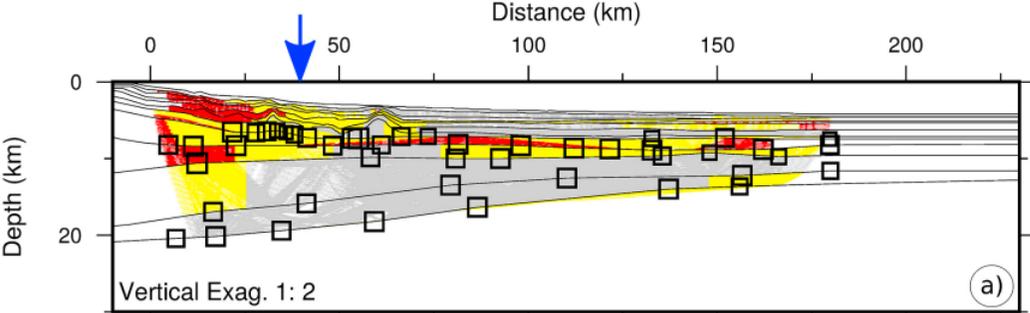


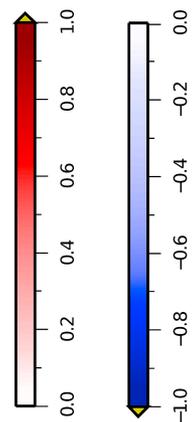
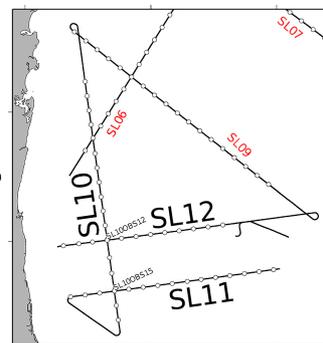
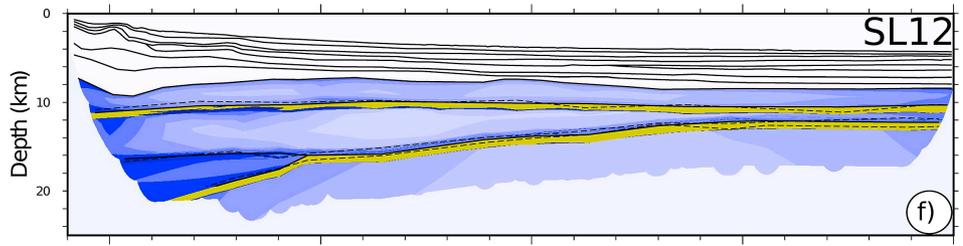
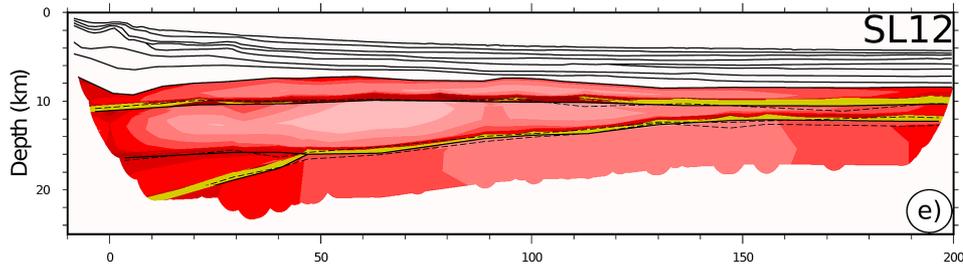
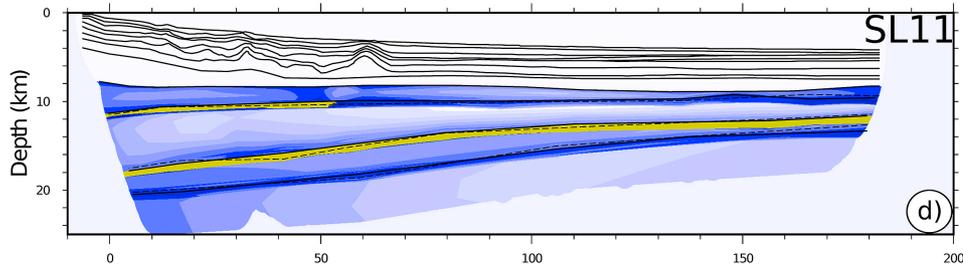
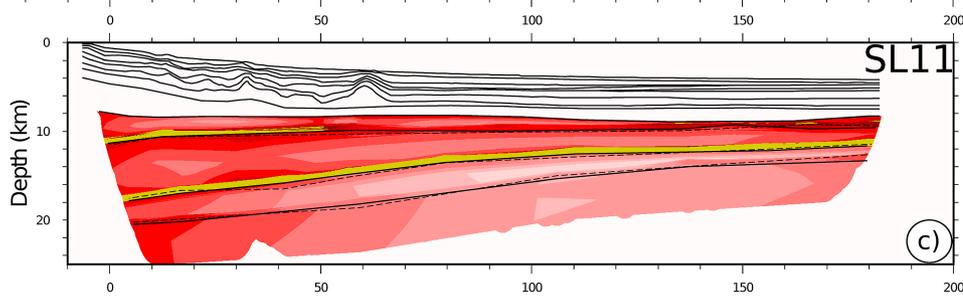
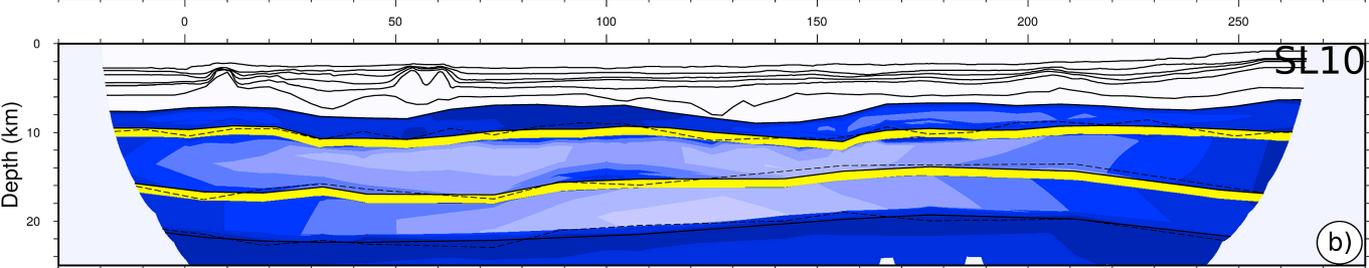
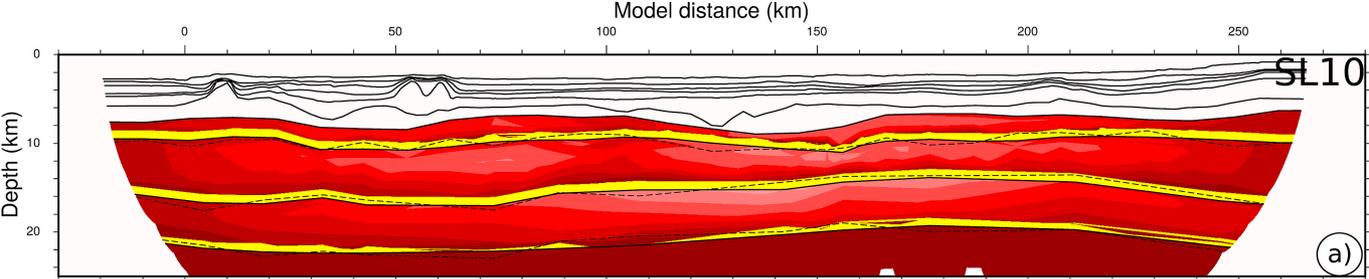




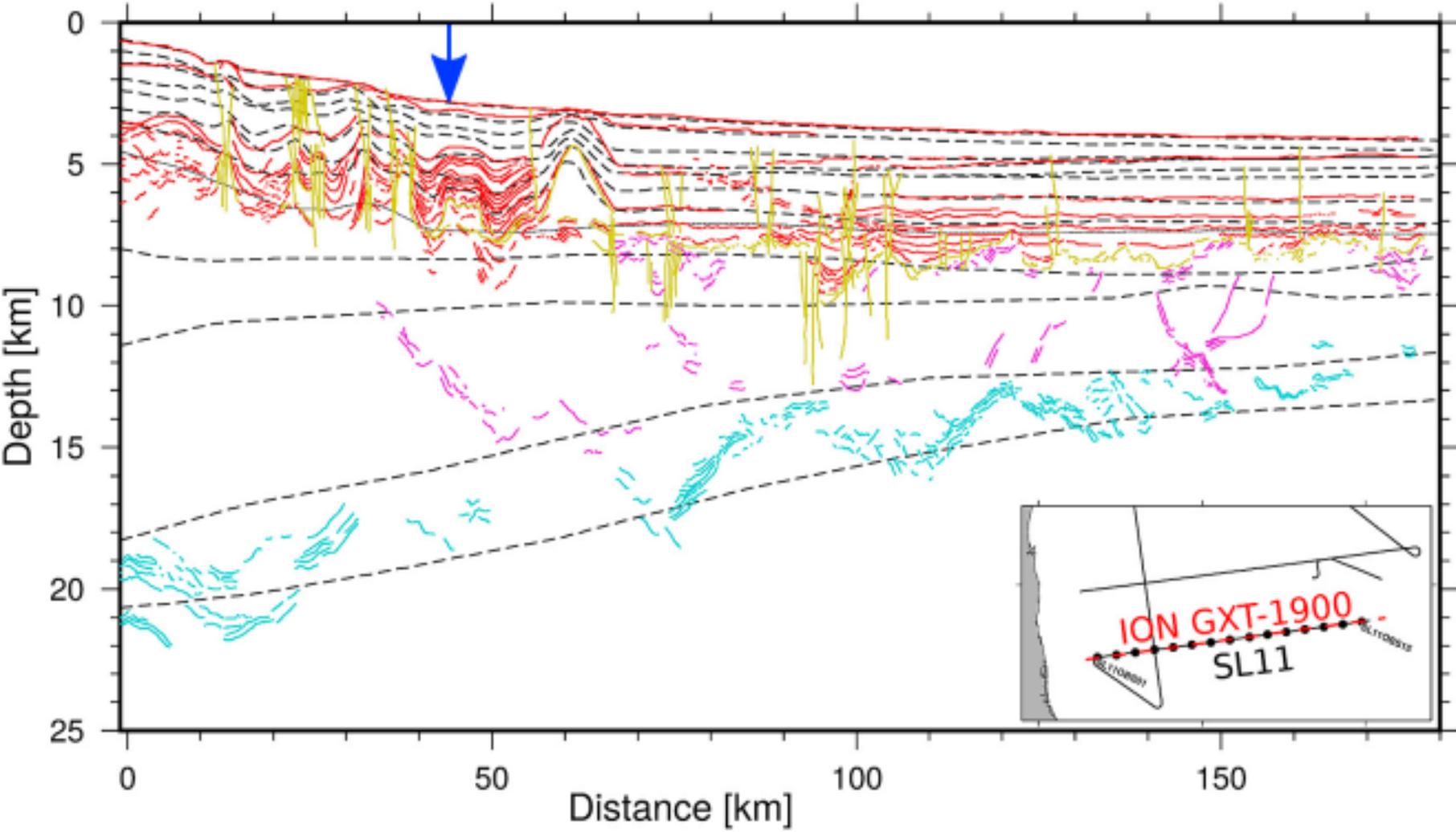


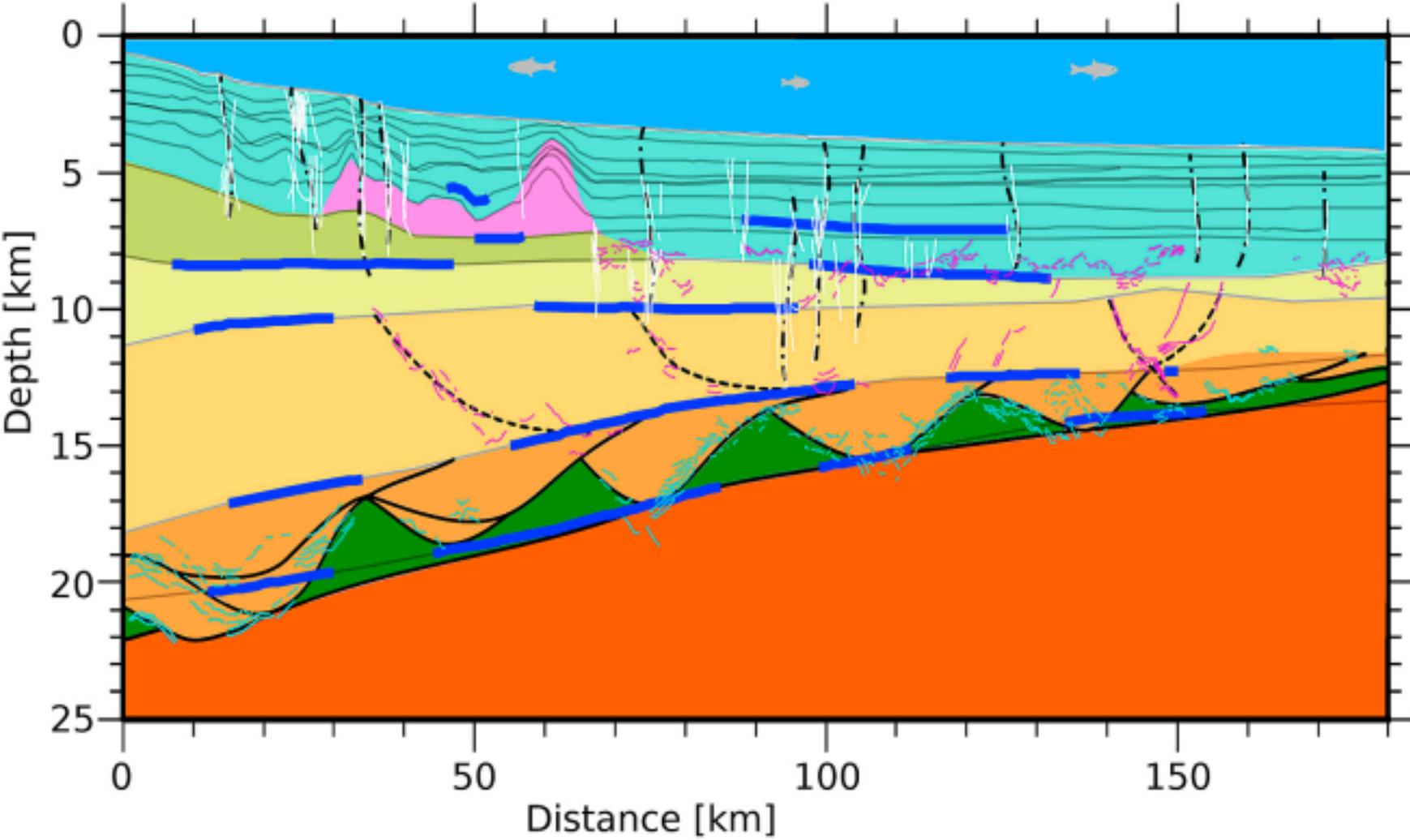




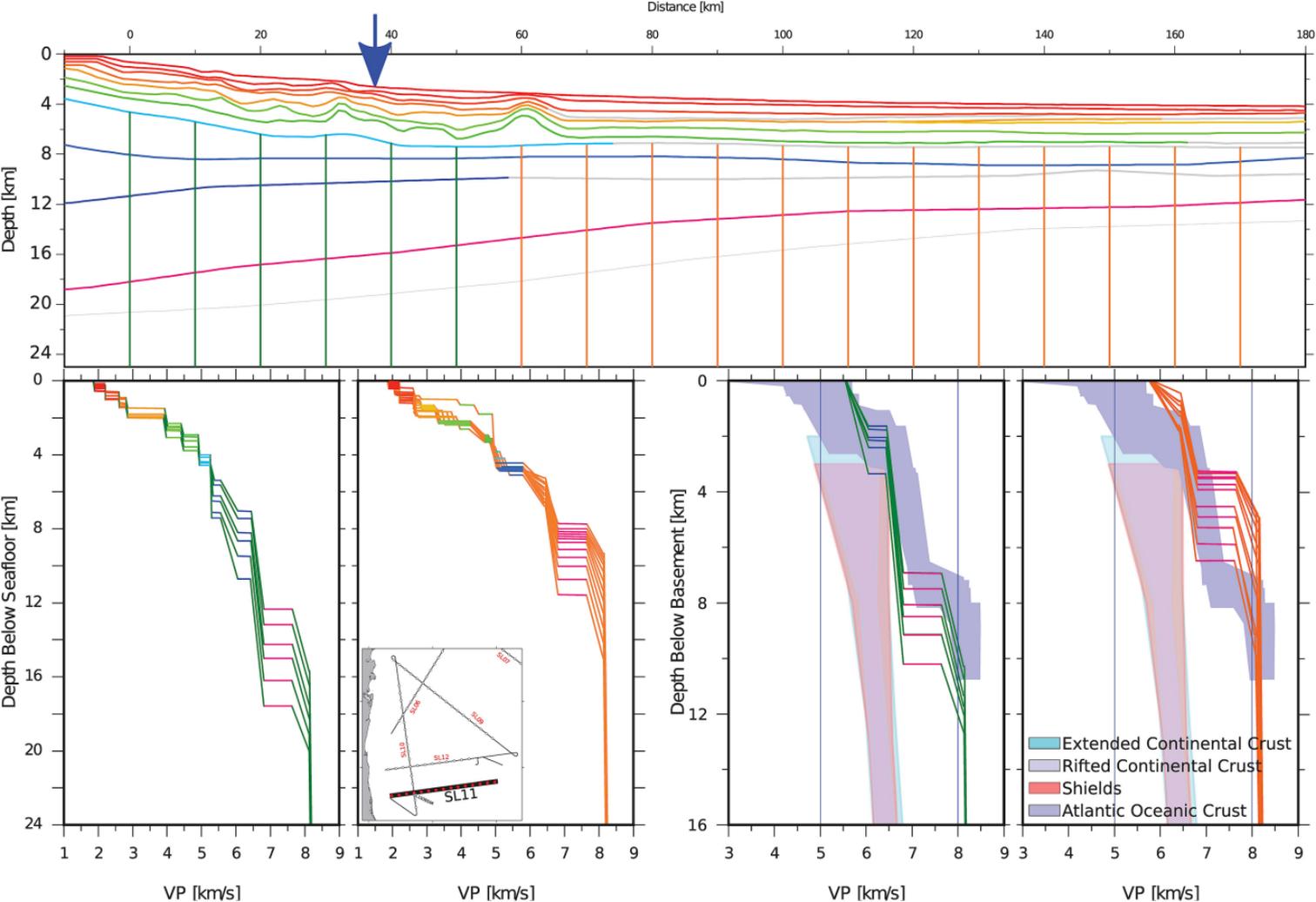


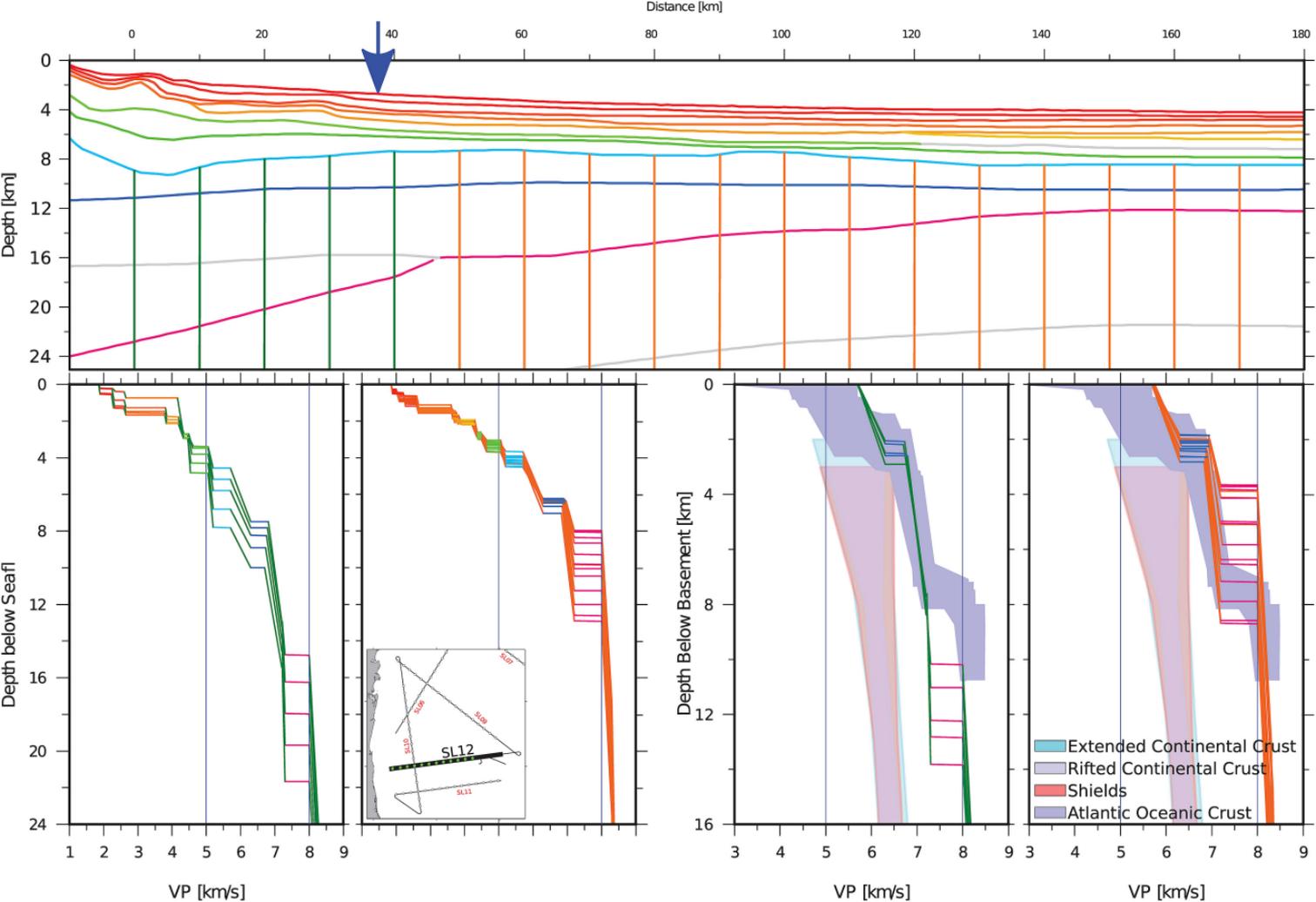
Velocity uncertainty (km/s)

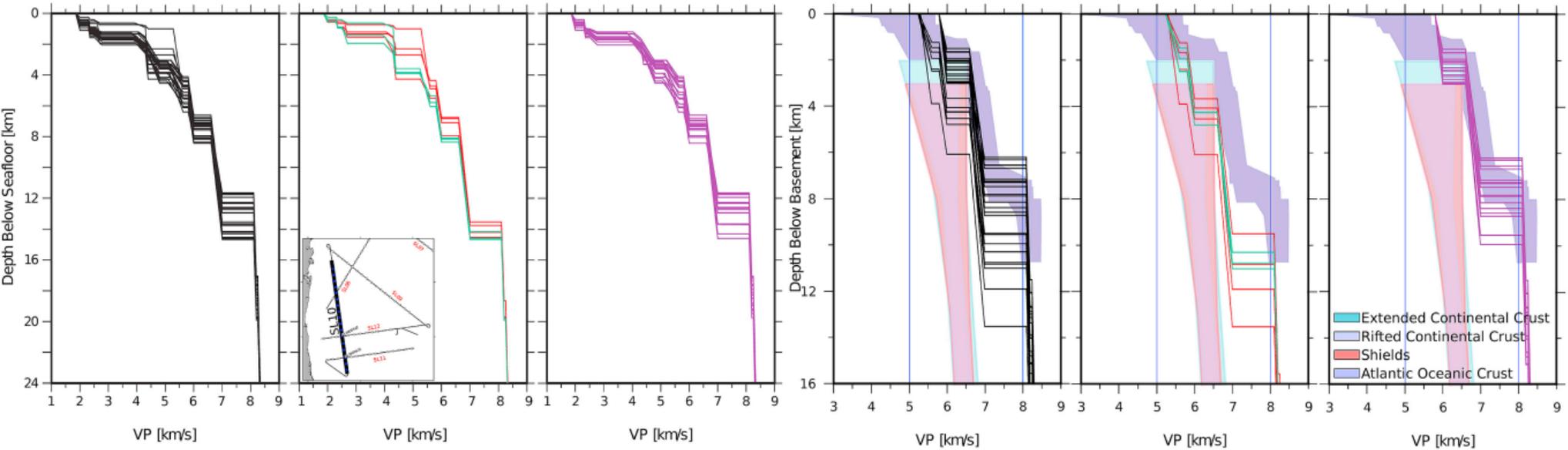
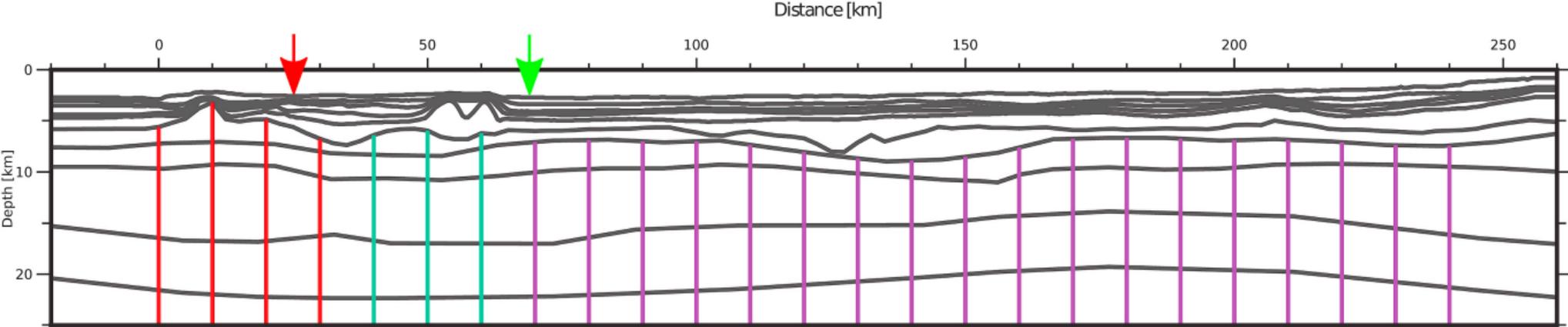




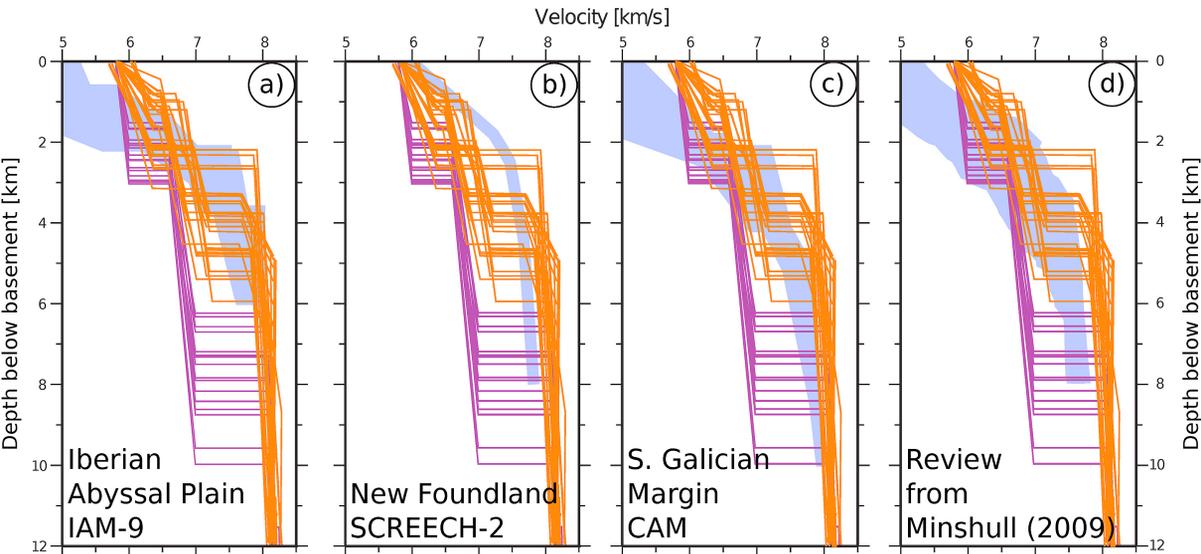
- Sediments
- Salt
- Upper Crust
- Middle Crust
- Lower Crust
- Thinned and/or exhumed Lower Crust
- Enriched Lower Crust
- Upper Mantle
- Wide-angle reflection
- Listric fault
- Detachment fault
- Fluids vent



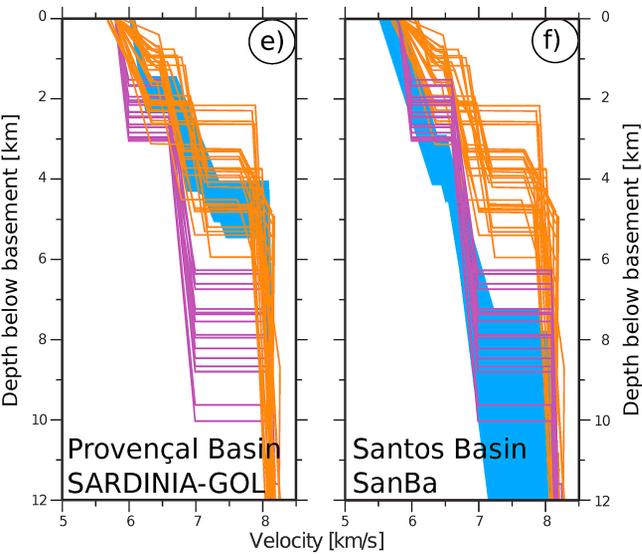




Exhumed / serpentinised Mantle



Exhumed Cont. Crust



JEQUITINHONHA BASIN

