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Structural styles of Albian rafts in the Espírito Santo Basin (SE Brazil):

Evidence for late raft compartmentalisation on a 'passive' continental margin

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Abstract

In recent years, hydrocarbon exploration offshore SE Brazil has been focused on Early Cretaceous units that were deformed due to Albian-Cenomanian gravity gliding above Aptian salt. A three-dimensional (3D) seismic volume from the Espírito Santo Basin, SE Brazil is here used to: a) test the parameters considered to control raft tectonics on a margin tectonically reactivated in the Cenozoic, and b) investigate the impact of prolonged halokinesis on raft deformation. Offshore Espírito Santo, the combined effects of halokinesis and multiple (Andean) tectonic phases are expressed by local collapse, fault reactivation and late segmentation of Albian rafts. As a result of this deformation we observe four main raft geometries: a) rolled-over rafts, b) tabular rafts, c) collapsed rafts, and d) folded and tilted rafts on the flanks of salt rollers. This work shows that salt rollers formed buttresses to moving Albian-Cenomanian rafts, with withdrawal of salt from underneath some of the rafts leading to their collapse and welding onto pre-salt strata. This process occurred in the studied part of the Espírito Santo Basin with minimum control of post-raft overburden thickness on raft compartmentalisation. Salt withdrawal from underneath the rafts is an important phenomenon as it enhanced connectivity between pre-salt and post-salt units, potentially promoting the migration of hydrocarbons from syn-rift source units into post-salt reservoirs.

Keywords: South Atlantic, SE Brazil, raft tectonics, compartmentalisation, halokinesis, overburden thickness.

28 1. Introduction

29 Raft tectonics comprises one of the most extreme deformation styles on salt-influenced
30 continental margins (Duval et al., 1992; Gaullier et al., 1993; Mauduit et al., 1997; Penge et al.,
31 1999; Alves, 2012; Pilcher et al., 2014). It is characterised by downslope translation of large blocks
32 of strata above a ductile detachment layer (Gaullier et al., 1993). A key characteristic of raft
33 tectonics is that thin-skinned stretching in overburden strata reaches beta (β) values of 2-3, with
34 associated gravitational gliding contributing to the fragmentation of post-salt units (Duval et al.,
35 1992; Gaullier et al., 1993; Mauduit et al., 1997; Vendeville, 2005). The majority of published work
36 suggests this fragmentation results from the interaction between gliding blocks (rafts), faulting and
37 a thickening overburden. Based largely on the interpretation of regional 2D seismic data and the
38 analysis of physical models, published results consider the thickness of the post-raft overburden and
39 the slope gradient as the main controlling parameters on the degree and style of raft segmentation
40 and downslope movement (Brun and Mauduit, 2009; Duval et al., 1992; Gaullier et al., 1993;
41 Mauduit et al., 1997; Vendeville, 2005). According to these authors, differences in post-raft
42 overburden thickness can maintain downslope gliding of rafts even if slope gradient is close to zero,
43 as long as an efficient basal décollement is present at depth. The proposed models essentially
44 suggest that increasing rates of syn-kinematic sedimentation increase downslope displacement of
45 rafts and make listric normal faulting more likely (Mauduit et al., 1997). However, the role of salt
46 thickness and tectonic reactivation in raft evolution is still poorly understood in basins such as the
47 Espírito Santo Basin, in which significant tectonic and igneous events are known to have controlled
48 its structural evolution (e.g. Fiduk et al., 2004). In fact, distinct tectonic episodes controlled the Late
49 Cretaceous-Lower Cenozoic evolution of the basin, inducing local shortening, diapir growth and
50 fault reactivation (Fiduk et al., 2004; Baudon and Cartwright, 2008; Alves, 2012).

51 This work is based on 3D TWT (two-way time) seismic data from Espírito Santo Basin, offshore
52 SE Brazil, to describe and discuss the effect of tectonic reactivation and halokinesis on the structure

of Albian rafts and overlying strata. It focuses on a region of offshore SE Brazil where a direct relationship between post-raft overburden thickness and raft internal deformation is not observed, and concludes on the factors that may have controlled raft evolution in the Espírito Santo Basin (Figs. 1a to 1c). Importantly, the study area records multiple episodes of tectonic reactivation related to the Andean tectonic phases and Paleogene emplacement of the Abrolhos Volcanic Plateau (Fiduk et al., 2004) (Fig. 1a). The first of these episodes, the Late Cretaceous Peruvian phase (Scheuber et al., 1994), had a deep control on fault reactivation and local erosion in the study area. The main advantage of this work, when compared with most published data, is that it uses a high-quality 3D seismic data volume to describe in great detail the fault families associated with salt structures and adjacent rafts. In such a context, we will map and describe main faults and types of rafts in a sedimentary basin known for its hydrocarbon potential.

The paper starts with a description of the data and methods used. It is followed by a section introducing the geological setting of the Espírito Santo Basin. The results section describes the main raft geometries, quantifies overburden thickness, and documents the main fault families observed in the study area. It also relates the styles of halokinesis imaged on seismic data with the styles of deformation observed within the rafts. We conclude the paper by answering important questions related to raft evolution, including:

a) Is the thickness of post-raft overburden the key control on raft deformation offshore Espírito Santo?

b) Are growing salt structures capable of imposing renewed compartmentalisation in otherwise welded (and stable) rafts?

c) What is the importance of halokinesis to hydrocarbon migration and structure charging of Albian rafts in the study area of the Espírito Santo Basin?

2. Data and methods

80 A high-quality 3D TWT seismic volume from CGG was used in this paper to interpret the
81 structural evolution of Albian-Cenomanian rafts in the Espírito Santo Basin, SE Brazil. The
82 interpreted 3D seismic volume covers 2400 km² of the continental slope area immediately south of
83 Abrolhos Plateau, in Block BES-100 (Fig. 1a). The seismic volume was acquired using a 6 x 5700-
84 6000 m array of streamers. It has a bin spacing of 12.5m x 25 m and is zero-phased migrated. The
85 seismic volume uses the European SEG standard for polarity, in which the a change of acoustic
86 impedance from low to high has positive amplitude and is visualised on-screen as a red seismic
87 reflection. Data were vertically sampled every 2 ms. Data processing included resampling, spherical
88 divergence corrections and zero-phase conversions undertaken prior to stacking (Fiduk et al., 2004;
89 Alves, 2012).

90 Four N-trending rafts were investigated to constrain their spatial distribution and deformation
91 (Figs. 1b, 1c and 2). The top and base of the interpreted rafts coincide with a reflection of strong
92 amplitude that was mapped every two lines (25 m). Detailed structural maps were generated to
93 highlight the rafts' external and internal structure, and the orientation and distribution of their faults.
94 In detail, prominent stratigraphic unconformities were mapped across the entire seismic volume to
95 compute: (i) isochron maps for post-raft overburden units, and (ii) Root-Mean Square (RMS)
96 amplitude maps, which are useful to highlight faults and chasms inside and between the interpreted
97 rafts. RMS amplitude maps average the squared amplitudes of seismic reflections mapped within a
98 pre-defined interval (Brown, 2004).

99 Post-raft overburden thickness was measured every 20 inlines and crosslines (i.e. every 250 m).
100 Raft thickness, length and width were also measured to document changes in rafts geometry.
101 Seismic stratigraphic interpretations were based on França et al. (2007) and Alves (2012). In our
102 calculations we used velocity data from the DSDP Site 516 (located on the Abyssal Plain to the
103 southeast of the study area), which estimated V_p velocity varying from 1700 m/s TWT for Late
104 Cenozoic strata, to 2000 m/s for Paleogene and Late Cretaceous strata and 3500 m/s for the

105 interpreted rafts (Barker et al., 1993). A seismic velocity of 1560 m/s TWT was used for the water
106 column (Gamboa et al., 2012) (Figs. 2 and 3a).

107

108 **3. Geological framework of the Espírito Santo Basin**

109

110 *3.1 Tectono-stratigraphic evolution*

111

112 The Espírito Santo Basin comprises a series of Late Jurassic-Cretaceous rift basins, trending N-S
113 to NNE-SSW, located between the Vitória-Trindade Chain and Abrolhos Plateau (Fig. 1a). Its
114 tectonic evolution records four distinct stages: rift onset, syn-rift, transitional and drift (Alves, 2012;
115 Chang et al., 1992; Fiduk et al., 2004; Gamboa et al., 2012). The initial rift-onset stage occurred
116 during the Late Jurassic to earliest Cretaceous and chiefly comprises continental deposits (Figs. 3a
117 and 3b). The syn-rift stage, dated from the late Berriasian/Valanginian to the early Aptian, is
118 marked by significant tectonic activity that led to the formation of rift basins (Demercian et al.,
119 1993; França et al., 2007; Gamboa et al., 2011; Mohriak, 2005; Ojeda, 1982). During this time,
120 lacustrine sediments accumulated in a series of fault-controlled basins before carbonate deposition
121 commenced at the start of the Aptian.

122 The transitional stage occurred between the Aptian and the early Albian, and records widespread
123 tectonic quiescence with the cessation of basement fault activity (Gamboa et al., 2011). Thermally,
124 the basin records a sudden increase in heat flow accompanying continental breakup in its more
125 distal parts, whereas proximal regions of the margins should have recorded a relative cooling
126 (Lentini et al., 2010). In addition, thick salt in parts of the basin was potentially able to cool the
127 basin relatively to regions with no salt, keeping some of the pre-salt source rocks in the oil and gas
128 windows. The effect of thick accumulations of salt is even more marked when considering discrete
129 episodes of rifting offshore SE Brazil, each one capable of recording increases in local heat flow to
130 the margin (Lentini et al., 2010). Stratigraphically, the transitional stage marks a shift from

131 continental syn-rift strata to marine drift units. These units mark the first marine incursion into the
132 central graben of the southeast Brazilian rift basins (e.g. Dias, 2005). The transitional stage in SE
133 Brazil records the deposition of >3000 m of evaporites, mainly halite and anhydrite, resulting from
134 extreme marine evaporation in arid climatic conditions (França et al., 2007; Mohriak, 2003;
135 Mohriak et al., 2008).

136 The drift stage reflects the onset and the spreading of ocean crust between the South American
137 and African tectonic plates and is dominated by open marine deposition. The deposition of marine
138 shales, turbidite sands and marked episodes of mass-wasting define this stage in the Espírito Santo
139 Basin (Fiduk et al., 2004). The drift stage can be sub-divided into a transgressive early-drift
140 megasequence (Albian-Early Eocene) and a regressive late-drift Megasequence, which together
141 span from Eocene to Holocene (Ojeda, 1982). In the Espírito Santo Basin, the drift stage is
142 dominated by continental-slope embankment, incision of submarine channel systems, and
143 widespread slope mass-wasting, with salt tectonics playing a significant role in post-evaporitic
144 sequence deformation. Post-salt deformation in the Espírito Santo Basin has been driven by a
145 combination of gravity gliding and gravity spreading (Fiduk, et al., 2004). Salt structures and
146 associated overburden units can be divided into three structural domains: a) proximal extensional,
147 b) mid-slope translational, and c) distal compressional (Fig. 3b). The extensional domain is located
148 in proximal, upper slope areas and is characterised by salt rollers, salt walls, normal faults, turtle
149 anticlines and rafts (Fiduk et al., 2004; Gamboa et al., 2011; Mohriak et al., 2008; Omosanya and
150 Alves, 2013). The transitional, mid-slope domain is dominated by salt diapirs, whereas the
151 compressional domain is developed on the distal parts of the slope and is dominated by
152 allochthonous salt canopies and tongues that deform the seafloor (Fiduk et al., 2004) (Fig. 3b).

153

154 *3.2 Andean tectonic phases and their effect on SE Brazil*

155

156 The study area was affected by several tectonic events (França et al., 2007), some of which can
157 be correlated to deformation episodes in the Central Andes (Fiduk et al., 2004; Gamboa et al., 2011;
158 Mohriak et al., 2008; Omosanya and Alves, 2013). These deformation episodes are expressed in the
159 Espírito Santo Basin by stratigraphic unconformities of regional expression (França et al., 2007) as
160 shown in detail in Fig. 3a. The direction of shortening was roughly orientated E-W following the
161 trend of the Andean Ranges, and acted together with gravitational tectonics to form complex
162 structures at post-salt level. At the scale of the South American Plate, the Andean Orogeny was
163 triggered in the Late Albian, during the Mochica phase (Mégard, 1984; Mégard et al., 1984),
164 leading the formation of the Pre-Urucutuca unconformity (Horizon 3) in the study area (Fig. 3a).
165 The following tectonic event, the Peruvian phase (80-90 Ma; Scheuber et al., 1994) resulted in the
166 deposition of extensive turbidite-filled submarine channels, which are particularly prevalent in the
167 study area (Fig. 3a).

168 A major reconfiguration of oceanic plates occurred at 49 Ma in the SE Pacific, during the Eocene
169 Incaic phase (Isacks, 1988; Mégard, 1984) (Fig. 3a). The last Andean compressive events are
170 divided into three discrete phases: the Quechua 1, 2 and 3 (Mckee and Noble, 1982; Mégard et al.,
171 1984), as shown in Figure 3a. The Quechua 1 phase occurred between ~20 and 12.5 Ma (Early to
172 Middle Miocene; Mégard, 1984), whereas the Quechua 2 phase occurred between 9.5-8.5 Ma
173 (Pliocene) and was marked by strike-slip movements (Fig. 3a). Broadly east-west orientated
174 shortening occurred during Quechua 3 (at ~6 Ma), which contrasts with the N-S shortening
175 recorded at present in the eastern part of South America (Lima, 2003).

176

177 **4. Controls on raft movement and segmentation**

178

179 Raft tectonics is the most significant style of deformation accompanying thin-skinned extension
180 on continental margins. Raft tectonics can generate regions in the sedimentary sequence where the
181 overburden stretches by two or three times its original length (Duval et al., 1992; Gaullier et al.,

182 1993; Mauduit et al., 1997). Where fault blocks at the base of stretched overburden units are
183 disconnected, they are termed rafts. If they are still partly in contact, they are termed pre-rafts
184 (Duval et al., 1992).

185 Most published studies used experimental or numerical models to understand the mechanisms of
186 raft tectonics (Duval et al., 1992; Gaullier et al., 1993; Mauduit et al., 1997). These models were
187 often supported by 2D seismic data, and they were based on various assumptions concerning the
188 physical processes involved in rafting. Such seismic data were crucial to understand the mechanical
189 behavior of raft systems and test the applicability of physical and numerical models (Brun and Fort,
190 2011).

191 One of the key parameters mentioned in published models as capable of controlling raft
192 displacement is overburden thickness (Mauduit et al., 1997). In their physical analogue models,
193 Mauduit et al. (1997) tested how the rafting structures are controlled by overburden sedimentation.
194 The experiment resulted in the formation of a wide deformation zone in the lab, with tilted blocks
195 delimited by extensional normal faults and rafts (Figs. 2 and 4). The first structures to develop are
196 symmetric grabens and, as sedimentation rate increases, the number of rafts or blocks increase
197 proportionally (Fig. 4). The models of Mauduit et al. (1997) indicate that an increase of the
198 sedimentation rate enhances the displacement rate of rafts as a response to increasing vertical
199 loading. Vendeville (2005) later showed that regional sediment deposition can trigger gravity
200 spreading, even without an oceanward dipping basal slope. As a key example, rafts in the Gulf of
201 Mexico record pure spreading driven by sedimentary loading. This setting requires a thick
202 sedimentary overburden, high sediment density and low frictional angles of the sediments (Brun
203 and Fort, 2011; Rowan et al., 2012). It will also imply the creation of large amounts of lateral space
204 into which overburden units can accumulate during stretching, as recorded in the Kwanza Basin
205 (Angola) by Duval et al. (1992). Here, two different types of gravity related movements have been
206 identified; gravity gliding and gravity spreading.

207 Salt as a viscous evaporitic layer has been described as an important factor not only in raft
208 formation, but also in raft gliding and subsequent deformation (Brun and Mauduit, 2009,
209 Vendeville, 2005,). Salt acts as a lubricant layer and forms rollers, pillows and diapirs adjacent to
210 individual rafts (Alves, 2012; Brun and Mauduit, 2009; Gaullier et al., 1993). Brun and Mauduit
211 (2009) performed laboratory experiments to study the development of growth faults during rafting.
212 They suggested that the concave shape of rollover faults is not the only valid argument for the
213 generation of large-scale listric faults in areas of raft tectonics. Instead, their concave shape results
214 from the connection between a steeply dipping normal fault and a flat-lying or gently dipping
215 décollement, a geometry prone to cause important tectonic reactivation in adjacent rafts due to the
216 mechanical instability of rollover faults (Brun and Mauduit, 2008). In other words, changes in the
217 dip of roller faults at depth results in the transfer of horizontal displacement towards the surface
218 through the rolling over of strata in the rafts, and in post-raft overburden strata every time roller
219 fault sole out into the detachment salt layer and significant lateral movement is recorded in rafts. In
220 support of this, Alves (2012) documented significant Late Cretaceous-Early Cenozoic reactivation
221 in raft-related faults in the Espírito Santo Basin, a phenomenon triggered by regional (Andean)
222 tectonics and related slope oversteepening.

223 This paper develops the ideas of Alves (2012), recognizing that the thickness of post-raft
224 overburden units does not vary significantly in the study area, a character suggesting that the salt
225 thickness and the evolution of salt rollers are the main controlling factor in their
226 compartmentalization and ramping up on the flanks of growing salt rollers. For that reason, we
227 name this latter stage of thin-skinned deformation ‘late rafting’, as it occurred in the late Cretaceous
228 after the main stage of raft movement in the Espírito Santo Basin.

229

230 **5. Seismic stratigraphy and structural features of the Espírito Santo Basin**

231

232 The seismic stratigraphy of the Espírito Santo Basin follows França et al. (2007), Alves and
233 Cartwright (2009) and Alves (2012). Figures 2 and 3a show seismic sections illustrating the entire
234 seismic sequence and the horizons interpreted in this work: a) base raft (Horizon 3a), b) top raft
235 (Horizon 3), c) base Santonian (Horizon 4) and d) the seafloor. In the next few paragraphs the
236 seismic imaging and a lithological description of the complete sedimentary sequence are described.

237

238 *5.1 K20 to K40 sequence (Earliest Cretaceous to Early/Mid Aptian)*

239

240 The K20-K40 sequences comprise sub-salt, syn-rift and early post-rift strata. The lower
241 boundary of K20 is marked by a moderate-amplitude, locally diffractive reflection marking the top
242 of crystalline basement rocks. The contact between K20 and K30 is irregular, and the imaged
243 sequences comprise moderate to high amplitude, low frequency reflections. K40 has a higher
244 amplitude and is more continuous than K20 and K30 (França, et al., 2007). The boundaries between
245 K20, K30 and K40 are difficult to distinguish on the interpreted seismic volume, partly because the
246 sequences comprise moderate to high amplitude, low frequency reflections (Fig. 6).

247 K20 comprises the oldest unit in the study area (Valanginian), deposited at the base of the Nativo
248 Group (Cricaré Formation). Heterolithic conglomerates and coarse sandstones observed in proximal
249 regions of Espírito Santo grade into fine-grained mudstones in more distal areas (França et al.,
250 2007). The K30 sequence comprises volcanic and volcanoclastic rocks intercalated with sandstones
251 and conglomerates (Jaguaré Member), which change into shales, marls and carbonate units in more
252 distal regions (França et al., 2007). The basal post-rift Sequence K40 comprises conglomerates and
253 sandstones that grade into fine sandstones and shales deposited in lacustrine and sabkha
254 environments (Membro Mucuri) (Figs. 3a and 6).

255

256 *5.2 K50 sequence (Aptian)*

257

258 The K50 sequence represents the main salt interval in the Espírito Santo Basin. This unit is
259 particularly well imaged forming the core of salt pillows and diapirs, where it is characterized by
260 chaotic, low amplitude reflections (Fig. 6). Its lower boundary consists of an irregular, moderate to
261 high amplitude reflection, below which high amplitude strata are observed. Its upper boundary
262 coincides with the first continuous strata above the low amplitude salt structures. In the study area,
263 the K50 sequence is only preserved within triangular salt anticlines (rollers) formed between rafts
264 (Fig. 6).

265 K50 was deposited in a series of confined basins in conditions of high evaporation. Carbonate
266 and anhydrite intervals predominate in shallow marginal areas of Espírito Santo, whereas halite is
267 more abundant in the central and distal parts of the basin (França et al., 2007).

268

269 5.3 K62 to K70 sequences (Albian)

270

271 The K62 to K70 sequences comprise a package of high-amplitude internal reflections overlying
272 the K50 and the basal K20 to K40 sequences. The lower boundary of K62-K70 is marked by
273 Horizon 2, whereas the top (i.e. Horizon 3) comprises an angular unconformity in the study area
274 (França et al., 2007).

275 K62-K70 is up to 600 ms two-way travel time thick and comprises marine strata, mainly sands,
276 silt, shales and oolitic limestones and marls, which are partly time equivalent to Unit 7 at DSDP
277 Site 356 (Kumar et al., 1977) and to the onshore Regência Formation (Bruhn and Walker, 1997;
278 Fiduk et al., 2004; França et al., 2007) (Fig. 3a). The top of K62-K70 is an angular unconformity in
279 the proximal regions of the Espírito Santo Basin, changing into a paraconformity in more distal
280 regions (França et al., 2007).

281 The two-way time structure of the top raft Horizon 3 is shown in Figure 7a. The map reveals the
282 presence of six (6) rafts in the study area, which are separated by local chasms (Fig. 7a). These

283 isolated depocentres were filled by post-Albian strata as explained in the following section (Figs. 3a
284 and 8).

285

286 *5.4 K82 to K88 sequences (Late Albian to Santonian)*

287

288 The K82 to K88 sequences comprise continuous, low-amplitude reflections deposited above the
289 Albian-Aptian rafts and associated salt structures. The lower boundary of K82-K88 is sharp and
290 marked by growth onto major listric faults above Horizons 2 and 3. The upper boundary of Late
291 Albian-Santonian strata is an irregular high-amplitude reflection representing an erosional
292 unconformity of Santonian age (Horizon 4) (França et al., 2007).

293 The K82-K88 sequences comprise shales and turbidite sands (França et al., 2007), belonging to
294 the lower Urucutuca Formation (Fig. 3b). Locally, the lower boundary of K82 contains carbonate
295 breccias derived from eroded Albian carbonate platforms.

296 In Figs. 7b and 7c are highlighted the two-way time structure and thickness of Late Albian-
297 Santonian sequence. Deposits of this latter age fill local inter-raft basins and cover older rafts to a
298 maximum thickness of 3.5 s (~2700 m) over raft 4 (Fig. 8). Isochron maps show a minimum
299 thickness over structural highs (i.e., rafts) and structures reactivated at the end of the Cretaceous
300 (Figs. 7b and 7c). Main sub-basins and associated salt rollers strike N-S to NNE-SSW (Figs. 7b and
301 7c).

302

303 *5.5 K90-K130 sequence (Late Santonian to Maastrichtian)*

304

305 A major unconformity related to the incision of a Late Santonian to Maastrichtian channel
306 system is observed above low-amplitude strata in K82-K88 (Golfinho Field, Vieira et al., 2007).
307 This boundary (Horizon 4, Fig. 3a) is overlaid by high-amplitude seismic reflections. Growth of

308 strata into listric, salt-detached faults is observed below the unconformity (Fig. 9). A regional
309 unconformity of Late Maastrichtian age marks the upper boundary of K90-K130.

310 K90-K130 comprises the middle part of the Urucutuca Formation (Fiduk et al., 2004; França et
311 al., 2007). The main lithologies in the sequence are turbidites and shales, changing into marly
312 successions towards more distal parts of the Espírito Santo Basin (França et al., 2007).

313 The two-way time isopach of K90-K130 is shown in Fig. 7c. The figure highlights the complex
314 set of faults that developed in this sequence, partly in response to basin reactivation at the end of the
315 Cretaceous. Reactivated structures at K90-K130 level are shown with thickness minima in the study
316 area, a character further investigated in Section 6 (Figs. 7c and 9).

317

318 *5.6 E10-E30 sequence (Paleocene to Early Eocene)*

319

320 The E10-E30 sequence comprises high-amplitude reflections affected by closely-spaced normal
321 faults. Its lower boundary coincides with the Maastrichtian unconformity (Fig. 3a), whereas the
322 upper boundary is marked by Horizon 6, a regional unconformity above which moderately faulted
323 reflections are observed. This unconformity has been interpreted to result from tectonic uplift of the
324 basin-shoulder during the early Cenozoic (França et al., 2007). A key characteristic of this interval
325 is the high degree of faulting observed to extend from Upper Cretaceous strata (Fig. 10).

326

327 *5.7 E40 to N10 sequences (Eocene to Early Miocene)*

328

329 The E40 to N10 sequences are composed of east-dipping clinoforms deformed by closely-spaced
330 normal faults (Figs. 3a and 9). The age of E40-N10 ranges from the Eocene to the Early Miocene,
331 i.e. it comprises equivalent strata to the upper part of the Urucutuca Formation (França et al., 2007).
332 The E40-N10 sequences are composed of turbidite sands intercalated with volcanoclastic deposits

333 (França et al., 2007). There is evidence on seismic data that syn-sedimentary fault activity
334 continued at this level. (Figs. 9 and 11)

335

336 *5.8 N20 to N60 (Early Miocene to Holocene)*

337

338 The N20-N60 sequences comprise chaotic to continuous reflections. The sequences are, in places,
339 eroded by submarine channels and comprise sandstones (Rio Doce Formation), calcarenites
340 (Caravelas member) and turbidite sands and marls (Urucutuca Formation). Mass-transport
341 complexes and channel-fill deposits are abundant throughout the basin after the Early Miocene
342 (França et al., 2007) (Fig. 3a).

343 The mid-Miocene TWT structure in Figure 10 shows the complex faulting observed at this
344 level, and the formation of a gentle slope dipping towards the east.

345

346 **6. Results**

347

348 *6.1 Evidence for tectonic reactivation and late halokinesis*

349

350 The Andean tectonic phases were key events controlling the structural evolution of the Espírito
351 Santo Basin (Lima, 2003). Tectonic reactivation was chiefly recorded at the end of the Cretaceous
352 and in the Eocene, as revealed on seismic data. Figure 7 shows a series of isochron maps between
353 the Maastrichtian and Santonian unconformities i.e., between the stratigraphic unconformities that
354 mark Late Cretaceous and Eocene tectonic episodes in the Espírito Santo Basin (Alves, 2012). In
355 these maps is important to highlight the thickness variations recorded at Santonian-top
356 Maastrichtian and Maastrichtian-Eocene levels. Areas recording reactivation and local erosion
357 present the lower thickness values in Figure 7. On seismic data, tectonic reactivation on seismic
358 data is marked by low-amplitude folding and reactivation of extensional structures (Figs. 2 and 5).

359 Erosion of Horizons 5 and 6 accompanied Late Cretaceous and continued during Eocene tectonism,
360 and resulted in the deposition of less than 400 ms TWT between the two mapped unconformities
361 (Fig. 5).

362 Figure 8 illustrates one of the regions on the Espírito Santo continental slope where tectonic
363 reactivation is more pronounced. Reactivated faults occur in the imaged seismic line between rafts 1
364 and 2, towards the upper part of the continental slope. In other regions, pop-up structures intersect
365 Late Cretaceous and early Cenozoic strata (Figs. 2 and 5).

366 In summary, Late Cretaceous tectonic reactivation on the Espírito Santo continental slope area is
367 marked by: a) localised inversion of raft-bounding normal faults, forming local pup-up structures b)
368 shortening of Meso-Cenozoic strata to form local pop-up structures (Fig. 2).

369 370 *6.2 Rafts geometry and thickness variations in the post-raft overburden*

371
372 At present, Aptian salt forms isolated accumulations, some of which are observed beneath the
373 interpreted rafts in the form of rollers. Above the Aptian salt are observed symmetric and
374 asymmetric rafts with distinct structural styles and inferred evolutions (Figs. 5 and 6).

375 A structural map of Horizon 3 (top rafts) illustrates the plan-view geometry of rafts 1 to 6 (Fig.
376 7a). In the northwest part of the study area, the rafts are intensely faulted. In east-west profiles, i.e.
377 perpendicular to the strike of rafts, raft 1 is irregular and discontinuous, showing important
378 segmentation (see Section 7 and Fig. 7a). In the map in Fig. 7a, this raft is at least 36 km long. For
379 raft 2, two-way time (TWT) raft thickness ranges from 34 ms to 815 ms along the north-south
380 profile in Fig. 11. This corresponds to a thickness of 45 m to 1107 m, using velocity data from
381 Barker et al. (1983) (Fig. 3a).

382 In contrast to raft 1, the north-south profile in Fig. 11 shows raft 2 to be continuous with a well-
383 defined branch in its northeast portion. The gap between the main body of the raft 2 and this latter
384 branch is occupied by a chasm with a small salt roller (Fig. 6). The TWT thickness of raft 2 varies

385 between 52 ms and 991 ms, i.e. between 72 m and 1340 m. Rafts 3 and 4 are geometrically similar
386 without any visible branches developed along their long axes (Figs. 1b and 7a). Raft 4 comprises a
387 north-trending raft with a tabular shape showing distinct degrees of bucking and faulting at
388 Cretaceous level (Figs. 1c and 6)

389 Rafts 5 and 6 are the structures less visible on seismic data, with their base outside the available
390 seismic data in most of the study area (Fig. 1c). Their tops are irregular, with several segments
391 visible on structural data (Fig. 7a). Aptian salt is present as isolated accumulations (rollers) beneath
392 of rafts 1 to 6.

393 Thickness plots were calculated from seismic data (Fig. 12). The plots are separated into two
394 main packages comprising post-raft sediments: (i) top rafts (Albian to Early Cretaceous) to base
395 Santonian (Late Cretaceous) and (ii) top rafts (Albian to Early Cretaceous) to seabed (Figs. 3a, 5
396 and 6). Over the northern part of raft 1, overburden thickness is 700 ms (875 m) from top raft to the
397 base Santonian (K82 to K88 sequences); and 1280 ms (~1600 m) for the Santonian to the seafloor
398 (K90 to N60 sequences) (Fig. 12). Trend curves for overburden thickness are similar for the two
399 intervals considered: top raft to base Santonian and base Santonian to seafloor, and when plotting
400 the curves for the total post-raft overburden (Fig. 12).

401 Overburden strata draping rafts 2 and 3 show a similar thickness trend to equivalent strata above
402 raft 1, recording ~1250 ms (1562 m) and 2600 ms (2860 m) for the top raft to base Santonian, and
403 base Santonian to seafloor intervals (Fig. 6). Strikingly, rafts 4 and 5 show marked thickness
404 variations in north-south profiles, but with the thicker overburden strata occurring to the north and
405 central parts of the rafts, where structural compartmentalisation is greater (Fig. 12). Raft 6 shows
406 larger thickness in its central part (Fig. 12). In essence, the thickness of post-raft overburden
407 increases towards the south when considering the sequences between Horizon 4 and 5, and
408 decreases for the Santonian-Seafloor sequence (Fig. 12).

409

410 *6.3 Deformation styles and fault families*

411

412 Figure 7c shows an isochron map between Horizons 4 and 5 (Santonian to top Maastrichtian). In
413 addition, the seismic profiles in Figures 5, 6 and 8 highlight the main fault families developed above
414 the Aptian salt and in Albian rafts. Figure 7c is complemented by the TWT structural maps in
415 Figure 10. The maps show the complex sets of faults affecting post-salt overburden units near the
416 base of the Santonian and above. The seismic profiles in Figure 8 show that overburden faults
417 propagated vertically until they reached horizon 4 (base Santonian) and overlying strata. Fault
418 families in rafts 1 to 6 include: a) roller, b) rollover, c) keystone; d) reactivated, and f) concentric
419 faults. A schematic map of these types of faults is shown in Figure 13.

420

421 6.3.1. *Roller Faults*

422

423 Roller faults accommodated bulk downslope displacement in rafts. Roller faults dip both
424 oceanwards (east) and landwards (west), offsetting strata in rafts 1 to 4 and overlying strata above
425 them (Figs. 5, 6 and 8). Roller faults sole out into the Aptian salt. Triangular salt rollers are
426 observed in the footwalls of roller faults (Fig. 14). Some of the roller faults propagated upward into
427 lower Cenozoic strata, tipping out at the base of a mass-transport deposit that contains large
428 remnant blocks (Alves, 2012).

429

430 6.3.2 *Rollover Faults*

431

432 Rollover faults comprise closely-spaced antithetic and synthetic faults generated on top of
433 monoclinal rollovers and turtle anticlines, themselves formed due to movement on adjacent roller
434 faults. Together with keystone faults, rollover faults accommodate some of the bending strain
435 related to the downslope displacement of collapse of underlying rafts (Fig. 15). They are formed
436 due to progressive bending of rollover structures above the Albian rafts (Fig. 5, 6 and 8).

6.3.3 *Keystone faults*

Keystone faults are pairs of conjugate normal faults that dip in the opposite direction to, and accommodate displacement occurring on rollers faults (Figs. 5 and 6). Keystone faults can also manifest as planar growth faults rooted into the crests of triangular salt rollers. Throws on keystone faults are small in the regions where they intersect collapsed salt rollers (Alves, 2012) (Fig. 16).

6.3.4 *Reactivated faults*

Reactivated faults comprise fault sets initially formed by the arching of overburden units above the Albian rafts. They were later reactivated in late Cretaceous anticlines, as shown in Fig. 8. The geometry of reactivated roller faults resemble that of crestal (or keystone) faults, but they form anticlinal structures towards their top (Figs. 9d and 13). They are interpreted as rollover, keystone and crestal faults that were reverse-reactivated.

6.2.5 *Concentric faults*

Concentric faults are observed above the depocentres formed by raft tectonics (Alves, 2012) (Fig. 9c and 9b). They are developed on the margins of extensional sub-basins, dying out downwards the main Cretaceous depocentres, accommodating local strain at the tips of the oval-shaped sub-basins formed on the hanging-wall blocks of roller faults.

7. Structural styles documenting raft deformation over Aptian salt structures

7.1. Rolling-over and internal strata growth

463

464 Rolled-over rafts are those showing important growth of strata adjacently to salt rollers and roller
465 faults (Fig. 10). These rafts are not cross-cut by major faults, and are mostly bounded landwards
466 and oceanwards by large roller faults. Rolled-over rafts formed during Cretaceous gravitational
467 collapse of the margin, but with most of extension concentrated on the larger roller faults. Turtle-
468 back structures are not developed above these rafts (Figs. 5 and 8).

469 The raft imaged in Figure 14 show important growth of strata in the areas where salt was
470 withdrawn from the base of the raft towards adjacent salt pillows. It is also noted the increasing
471 angle of basal strata in the raft as the rolling-over of the raft continues in time (Fig. 8).

472

473 *7.2 'Passive' fragmentation in the form of tabular rafts*

474

475 Tabular rafts are structures displaced over salt without significant control of roller faults on the
476 growth of strata inside the rafts (Figs. 8 and 15). Instead, these rafts are interpreted to have evolved
477 with large salt rollers separating them from adjacent rafts, and hindering any rolling over of strata
478 on their flanks (Figs. 8 and 15). Faults are scarce in their interior and, when present, show
479 predominant normal offsets resulting from salt withdrawal at flanks of the rafts. Turtle anticlines
480 can form in younger overburden units in response to folding of latest Cretaceous-early Tertiary
481 strata (Fig. 11).

482 Figure 15 shows tabular raft 2 in the study area of Espírito Santo. Tabular rafts show no
483 significant growth of strata in their interior, suggesting they were 'passively' translated and
484 fragmented on the continental slope. In the study area, most of the tabular rafts seem to be partly
485 welded on pre-salt successions and are bounded by small to moderate size salt rollers that did not
486 deformed their flanks (Fig. 15).

487

488 *7.3 Collapse and folding of rafts' flanks due to salt withdrawal*

489

490 Collapsed blocks are observed in flanking strata to sub-tabular rafts, always in association with
491 withdrawal of salt from evolving salt rollers (Fig. 16). Resulting fault styles include normal faults
492 showing no growth at the level of the Albian rafts, but showing growth and erosional truncation at
493 Late Cretaceous level (Fig. 16). This character suggests the faults post-date the deposition of
494 Albian-Cenomanian strata drapping the interpreted rafts. As a result of the withdrawal of salt from
495 salt rollers, most of these collapsed blocks are, at present, welded onto pre-salt successions (Fig.
496 16).

497 Rafts are usually folded in the immediate footwall of the larger roller faults (Fig. 16). They
498 reflect later growth of roller faults and associated growth of salt rollers on the flanks of relatively
499 stable rafts. Roller faults show predominant normal offsets and, in some parts of the study area,
500 were reactivated to form pop-up structures expressed in Upper Cretaceous strata.

501 Figure 16 depicts collapse structures in raft 2. Here, we interpret the withdrawal of salt from
502 underneath the raft, and subsequent growth of the salt pillow to the west, resulted in the collapse of
503 the flank of the imaged rafts. This structural style is more obvious to the north of the study area,
504 where rafts 1 to 6 are close together and segmented in smaller rafts. Local collapse structures
505 accompany the tilting and fragmentation of rafts on the flanks of salt rollers that grew, or where
506 shortened, during the Late Cretaceous and Cenozoic.

507

508 *7.4 Tilting and fragmentation of rafts on the flanks of growing salt rollers*

509

510 The most striking example of late reactivation in rafts comes from the array of faults and rafts
511 formed and tilted on the flanks of growing salt rollers. An example of one of such structures is
512 shown in Fig. 12, in which the oceanwards half of raft 2 is fragmented, collapsed and tilted on the
513 flank of the salt roller separating raft 2 into two parts. Normal faults related to the collapse of rafts
514 over withdrawn Aptian salt are observed in Figs. 6 and 16). Faults show predominant normal offsets

515 resulting from extension and salt withdrawal, but do not extend up into Paleogene strata i.e., they
516 were chiefly generated by short-lived collapse of rafts during the Late Cretaceous. As a result of
517 collapse, complex sets of conjugate normal faults are often observed in Upper Cretaceous rollovers,
518 as structures formed to accommodate the collapse of underlying rafts (Fig. 16).

519 Figure 16 shows an example of such structural style. The imaged raft was tilted and fragmented
520 on the flank of a growing salt roller, which shows evidence for Late Cretaceous reactivation. Part of
521 this fragmentation results from withdrawal of salt from the base of the rafts to inflate the adjacent
522 salt pillow, thus resulting in complex structural compartmentalisation of intra-raft strata.

523

524 **8. Discussion**

525

526 Fiduk et al. (2004) assumed tectonic contraction in post-salt units began early in the Albian and
527 continued until the present day. Rafting ceased at different times depending on the initial thickness
528 of the salt available and overburden thickness. Based on these two principles, we discuss in this
529 section: (i) the different styles of deformation observed on the rafts and (ii) reactivation of faults as
530 a function of salt roller growth.

531

532 *8.1 Why is there a poor correlation between overburden thickness and the degree of raft*
533 *deformation?*

534

535 The key question posed by this work is why there is a poor correlation between overburden
536 thickness and the degree of raft deformation in the Espírito Santo Basin? Based on the evidence of
537 moderate, but widespread tectonic reactivation of the continental slope during the Late Cretaceous
538 and Eocene, a plausible explanation should consider important raft movement in Espírito Santo
539 after the Santonian. An example of late-stage raft tectonics, in which the reactivation of salt rollers
540 is a key control on rafts' structural deformation, is provided by raft 2 (Figs. 10 to 12). Ramped-up

541 strata on the flanks of a salt roller, with associated uplift and erosion of the Late Cretaceous Horizon
542 4 demonstrates a later stage of deformation in the study area (Figs. 5 and 6). We interpret this
543 geometry as reflecting late evacuation of evaporites from upper-slope regions of the Espírito Santo
544 Basin to the base of the continental slope. Downslope salt flow resulted in the collapse of minor salt
545 pillows below individual rafts, in the growth of the larger salt rollers, and in the progressive welding
546 of rafts 1 to 6 onto pre-salt units. Most of this collapse occurred in the Late Cretaceous, as shown by
547 the collapse faults developed above Horizon 4.

548 The history of gravity-gliding extension of the Albian rafts and the relationship with Aptian salt
549 layer are summarized in Figure 17. At the scale of the interpreted 3D seismic volume, we observe
550 that syn-kinematic sediment thickness is relatively constant, a character suggesting that vertical
551 loading imposed by overburden strata was not the key factor controlling raft movement and
552 deformation in the study area. Instead, lateral spreading and downslope gliding of the rafts was
553 likely controlled by the presence of intra-raft salt structures – which closely controlled the degree of
554 downslope movement and faulting experienced by rafts 1 to 6 (Figs. 14 to 17).

555 Based on the interpreted data, we suggest that ‘late’ compartmentalisation of rafts offshore
556 Espírito Santo is an important phenomenon, with overburden thickness playing only a minor role.
557 In this work we demonstrate that salt rollers and the relative thickness of salt underneath individual
558 rafts are the main factors controlling raft movement. In regions where raft movement was
559 overprinted by the growth of salt rollers, rafts are highly segmented by reactivated faults (Figs. 5, 6,
560 9 and 17). In regions where smaller volumes of salt were available below the rafts and ramping-up
561 over growing salt rollers was hindered, we suggest rafts were static throughout most of their late
562 evolution and structural compartmentalisation was accordingly moderate.

563 A second question that arises when interpreting the seismic data in this paper is why are
564 structural collapse, tilting and local deformation so prominent in raft 2? One possible answer to this
565 question assumes that extension-related faulting was predominant in the study area, and that no
566 major reactivation occurred in response to the Andean tectonic stages. A second potential

567 explanation is that tectonic reactivation was significant, and that a later stage of roller growth and
568 salt withdrawal may have occurred, even if in a predominantly extensional regime.

569 Locally reactivated faults and associated pop-up structures indicate that a late stage of horizontal
570 shortening affected the study area, particularly during the latest Cretaceous and Palaeogene (e.g.
571 Alves, 2012) (Figs. 8 and 17). These structures were previously interpreted as partly
572 accommodating strain across the hinge of extensional rollovers. We interpreted them as reflecting a
573 later stage of gravitational gliding in the study area, in which Late Cretaceous strata (K82-K88)
574 were compressed against Aptian rafts (and overburden strata) due to the change from vertical
575 subsidence, recorded on the upper part of roller faults, to sub-horizontal strain in the regions where
576 roller faults sole into the Aptian salt (Alves, 2012).

577 We interpret deformation in raft 2 to result from the combination of factors described above, but
578 the surprising result in our analysis is that the thickness of the sediments overburden is not a key
579 factor in the onset of late raft deformation. Instead, we suggest that deformation in raft 2 resulted
580 from a combination of factors, including slope oversteepening and resulting stresses imposed by a
581 gravitationally unstable, downslope-moving overburden sequence against rafts 1 to 6. In this model,
582 the evacuation of salt from beneath the rafts, and their eventual grounding, was an important
583 process controlling the degree of deformation in Albian-Cenomanian rafts. Rafts overlying thin salt
584 successions were quickly grounded, and faults mostly occur within Late Cretaceous-Early Cenozoic
585 overburden strata (Figs. 5 and 6). Rafts with significant thickness of salt underneath record
586 important collapse, with salt withdrawal contributing to the growth of adjacent salt rollers. The
587 combined effect of salt roller growth and horizontal shortening of these same salt structures
588 (contributing to an increase in the angle of rollers' flanks) acted together to further tilt and deform
589 Albian-Cenomanian strata (Fig. 17). As a result, we observe in the study area styles of raft
590 compartmentalisation distinct to those published in the literature, with the thickness of overburden
591 units and slope oversteepening being locally replaced, as primary factors in raft
592 compartmentalisation, by the thickness of available salt below and adjacent to fully developed rafts.

593

594 *8.2 Importance of collapse features to the generation of salt welds*

595

596 Salt welds are formed at the base of post-salt strata by the complete evacuation of salt from
597 below these strata (Jackson and Cramez, 1989; Rowan et al., 1999). A consequence of welding of
598 post-salt strata onto pre-salt units, in relationship with the timing of potential hydrocarbon
599 generation and migration, is the establishment of fluid conduits between stratigraphic intervals that,
600 otherwise, would be hydrodynamically separated (Rowan, 2004). A key observation from the
601 interpreted seismic data is the generation of salt welds in regions recording collapse and tilting of
602 strata on the flanks of salt rollers (Fig. 17). In these cases, the timing in which the salt was
603 withdrawn from the base of the rafts, and a full weld was formed, is an important piece of
604 information when assessing the degree and timing of connectivity between pre-salt and post-salt
605 units. An example of these salt welds is shown in Figures 11 and 15, in which only a small part of
606 the raft is in contact with pre-salt strata (salt weld). We postulate that most of this welding occurred
607 relatively late in the Espírito Santo Basin, allowing the migration of fluids from pre-salt source
608 intervals into rafts and Cretaceous reservoirs only after welds were present below individual rafts.
609 This is an important observation, and one that confirms that palaeoreconstructions of raft
610 movement, and fluid migration, are key to explaining discrepancies in the charging of post-salt
611 reservoirs on continental margins dominated by gravitational collapse. The reasons for such
612 discrepancies are highlighted in the following paragraphs.

613

614 *8.3. Enhanced petroleum potential of post-salt strata due to raft grounding and salt welding*

615

616 After the deposition of Aptian salt, an Albian carbonate platform was developed all over SE
617 Brazil and later fragmented over a progressively thinner salt layer (e.g. Demercian et al., 1993).
618 According to Mauduit et al. (1993), sediment prograding from the continental shelf may have

619 helped this process in a first stage. In a second stage, tectonic pulses led to further oversteepening
620 and remobilisation of the Aptian salt, resulting in enhanced halokinesis, folding and faulting of the
621 Meso-Cenozoic successions capping the rafts (Fig. 17).

622 An immediate result of salt withdrawal from below the Aptian rafts is the generation of
623 important fluid flow paths from pre-salt source units (Lagoa Feia equivalent) towards reservoir
624 successions in and above rafts (Fig. 18). Complete withdrawal of salt from below the rafts leads to
625 the formation of salt welds, with subsequent transmission of fluids from below the salt into post-salt
626 successions across areas where salt thickness is below a certain threshold, or where salt is impure to
627 allow fluid flow through it. This phenomenon has important implications to the petroleum potential
628 of the Espírito Santo Basin; broad salt welds will favour the migration of fluids to the interior of
629 rafts, and through the large listric faults that bound them (Figs. 17 and 18). In turn, the presence of
630 relative thick salt below the rafts will isolate the post-salt reservoirs from pre-salt sources, with the
631 added cooling effect of salt to the overall thermal evolution of the basin (Lentini et al., 2010).

632 In the study area, evidence of fluid flow through listric faults is ubiquitous in the southern part of
633 the 3D seismic volume, where dim zones likely associated with the presence of gas are observed
634 over raft 3 (Fig. 18). Upper Cretaceous channel-fill deposits overlie the dim zones and hint at fluid
635 charging from pre-salt source rocks into the Golfinho Field and other associated oil and gas
636 accumulations (Fig. 18). This setting is, nonetheless, complicated by the presence of post-
637 Cretaceous faults that were associated with progressive halokinesis and younger tectonic phases
638 affecting the Espírito Santo Basin. Structural and stratigraphic hydrocarbon traps subsequently
639 formed at Late Cretaceous level could have been breached by fault families developing in Cenozoic
640 seal units (Fig. 18). To understand the sealing capacity of such faults is paramount to assessing the
641 petroleum potential of Late Cretaceous reservoirs.

642 Closer to the Andes, tectonics has been an important factor controlling the development of
643 structural traps, and promoting fault-related paths for hydrocarbon migration from source to
644 reservoir rocks. Examples of such control are well expressed in the Colombian Caribbean margin

645 (Sarmiento and Rangel, 2004; Ortiz-Karpf et al., 2015), onshore Colombia (Dengo and Covey,
646 1993; Mora et al., 2006), in Venezuela (Roure et al., 1997) and in multiple locations along the
647 Argentinian Andes (e.g. Belotti et al., 1995). A similar tectonic evolution is observed in SE Brazil,
648 where the combined effect of Andean tectonic phases and gravitational tectonics was capable of
649 controlling trap formation and, on a larger scale, source rock maturation (e.g. Beglinger et al.,
650 2012). The maturation history of the proximal margin of SE Brazil indicates that fault-controlled
651 fluid migration and overburden exhumation occurred in association with the Andean tectonic phases
652 (Mello and Maxwell, 1991). However, differences between distinct basins (and individual
653 depocentres) are observed in relation to the degree and timing of trap formation, and thermal
654 maturation of the basin. Assuming a homogeneous distribution of pre-salt source units, thermal
655 maturation and fluid migration is highly dependent on the thickness of Aptian salt (Lentini et al.,
656 2010).

657 In the study area of Espírito Santo, tectonic inversion is relatively moderate, but may have
658 increased raft segmentation and overburden folding to favour (1) the reactivation of salt minibasins
659 and detachment of overburden rocks over areas of thick evaporites; (2) forced folding of the post-
660 Aptian overburden, particularly against grounded rafts; (3) the reactivation of slope-bounding
661 normal faults and associated tilting of the slope area, promoting the formation of extensional and
662 compressional forced faults over reactivated structures (Figs. 17 and 18). In addition, some of the
663 structural traps observed at depth, particularly those related to individual rafts, might not be related
664 uniquely to Andean tectonism. This latter aspect broadens the range of potential traps on the
665 Espírito Santo Basin, increasing the range of drilling targets towards older Albian-Cenomanian
666 successions capping the extensional rafts (Fig. 17).

667 As a result of such setting, grounding of and enhanced segmentation of rafts will favour the
668 transmission of pre-salt fluid into the post-salt reservoirs, either by diffusion of oil through
669 permeable strata or by directed flow through faults and fractures. Late Cretaceous–Early Cenozoic
670 tectonics was therefore important for the relative development of traps and, in turn, for the loss of

671 fluids through the development of faults in post-Cretaceous units. The degree of segmentation
672 observed in rafts, and faults developed above them, denote that overburden thickness did not
673 exclusively control raft tectonics, and Late Cretaceous-Cenozoic tectonism is proposed here to be
674 an important control on salt withdrawal and raft segmentation well after the Albian-Cenomanian
675 onset of halokinesis offshore Espírito Santo.

676

677 **9. Conclusions**

678

679 This paper shows that the most developed styles of faulting and raft compartmentalisation occur
680 where salt was withdrawn from the base of rafts in the Late Cretaceous/Early Cenozoic. This
681 withdrawal likely resulted from tectonic imbalance between overburden loading and slope gradient
682 imposed by the Andean tectonic phases affecting SE Brazil. As a result, we observed the following
683 types of structures in rafts from the Espírito Santo Basin:

684

- 685 a) Rolling-over and internal strata growth in rafts that were displaced in the Albian-Cenomanian;
- 686 b) Fragmentation in the form of sub-tabular rafts whenever they were 'passively' translated on the
687 continental slope;
- 688 c) Collapse of rafts' flanks due to salt withdrawal from beneath them;
- 689 e) Tilting and fragmentation of raft on the flanks of growing salt rollers.

690

691 In the study area, rafts with significant thickness of salt underneath record important collapse,
692 with salt withdrawal contributing to the growth of salt rollers. The combined effect of salt roller
693 growth and horizontal shortening of these same salt structures acted together to further tilt and
694 deform Albian-Cenomanian strata. As a result, we observe in the study area styles of raft
695 compartmentalisation distinct to those published in the literature, with the thickness of overburden
696 units and slope oversteepening being as primary factors in raft compartmentalisation.

697 A result of this setting is the relative enhancement of fluid flow through listric faults. Upper
698 Cretaceous channel-fill deposits overlie the dim zones and hint at fluid charging from pre-salt
699 source rocks into younger strata. This setting is complicated by the presence of post-Cretaceous
700 faults. To understand the sealing capacity of such faults is paramount to assessing the petroleum
701 potential of Late Cretaceous reservoirs.

702

703 It is suggested that similar settings to the one documented in this paper occur in other Atlantic-
704 type margins subject to raft tectonics, with ramping-up of reactivated rafts contributing to a larger
705 degree of faulting and deformation in Albian reservoirs. The data in this paper will add to future
706 palaeoreconstructions of raft movement, and associated fluid migration from pre-salt units in SE
707 Brazil, and will help to tackle any discrepancies in the interpretation of gravitational collapse on
708 continental margins.

709

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716

717 **References**

718 Alves, T.M., 2012. Scale-relationships and geometry of normal faults reactivated during
719 gravitational gliding of Albian rafts (Espírito Santo Basin, SE Brazil). *Earth and Planetary*
720 *Science Letters*, 80-86.

721 Alves, T.M., Cartwright, J.A., 2009. Volume balance of a submarine landslide in the Espírito Santo
 722 Basin, offshore Brazil: Quantifying seafloor erosion, sediment accumulation and depletion.
 723 Earth and Planetary Science Letters 288, 572-580.

724 Barker, P.F., Buffer, R.T., Gombôa, L.A., 1993. A seismic reflection study of the Rio Grande Rise.
 725 Barker, P. F., Carlson, R. L., Hohnson, D. A. (Eds.), Initial Reports of the Deep Sea Drilling
 726 Program. Government Printing Office, Washington, D. C., 953 - 976.

727 Baudon, C., Cartwright, J., 2008. The kinematics of reactivation of normal faults using high
 728 resolution throw mapping. Journal of Structural Geology 30, 1072-1084.

729 Beglinger, S.E., van Wees, J.-D., Cloething, S., Doust, H., 2012. Tectonic subsidence history and
 730 source-rock maturation in the campos Basin, Brazil. Petroleum Geoscience 18, 153-172.

731 Belotti, H.J., Saccavino, L.L., Schachner, G.A., 1995. Structural Styles and Petroleum Occurrence
 732 in the Sub-Andean Fold and Thrust Belt of Northern Argentina. AAPG Memoir 62, 545–555.

733 Brown, A.R., 2004. Interpretation of three-dimensional seismic data, Sixth ed. American
 734 Association of Petroleum Geologists, Tulsa.

735 Bruhn, C.H.L., Walker, R.G., 1997. Internal architecture and sedimentary evolution of coarse-
 736 grained, turbidite channel-levee complexes, Early Eocene Regência Canyon, Espírito Santo
 737 Basin, Brazil. Sedimentology 44, 17-46.

738 Brun, J.-P., Fort, X., 2011. Salt tectonics at passive margins: Geology versus models. Marine and
 739 Petroleum Geology 28, 1123-1145.

740 Brun, J.-P., Mauduit, T.P.O., 2008. Rollovers in salt tectonics: The inadequacy of the listric fault
 741 model. Tectonophysics 457, 1-11.

742 Brun, J.-P., Mauduit, T.P.O., 2009. Salt rollers: Structure and kinematics from analogue modelling.
 743 Marine and Petroleum Geology 26, 249-258.

744 Chang, H.K., Kowsmann, R. O., Figueiredo, A. M. F., Bender, A. A., 1992. Tectonics and
 745 stratigraphy of the East Brazil Rift system: an overview. Tectonophysics 213, 97 - 138.

746 Demercian, S., Szatmari, P., Cobbold, P.R., 1993. Style and pattern of salt diapirs due to thin-
747 skinned gravitational gliding, Campos and Santos basins, offshore Brazil. *Tectonophysics* 228,
748 393-433.

749 Dengo, D.A., Covey, M.C., 1993. Structure of the Eastern Cordillera of Colombia: Implications for
750 Trap Styles and Regional Tectonics. *AAPG Bulletin* 77, 1315-1337.

751 Dias, J.L., 2005. Tectônica, estratigrafia e sedimentação no Andar Aptiano da margem leste
752 brasileira. *B. Geoci. Petrobras*, Rio de Janeiro v. 13, n. 1, 27 - 25.

753 Duval, B., Cramez, C., Jackson, M.P.A., 1992. Raft tectonics in the Kwanza Basin, Angola. *Marine*
754 *and Petroleum Geology* 9, 389-404.

755 Fiduk, J.C., Brush, E.R., Anderson, L.E., Gibbs, P.B., Rowan, M.G., 2004. Salt-Sediment
756 Interactions and Hydrocarbon Prospectivity: Concepts, Applications and Case Studies for the
757 21st Century. Gulf Coast Society of Economic Paleontologists and Mineralogists Foundation,
758 24th Bob F. Perkins Research Conference Proceedings (CD-ROM). Salt deformation,
759 magmatism, and hydrocarbon prospectivity in the Espírito Santo Basin, offshore Brazil, 370-
760 392.

761 França, R.L., del Rey, A.C., Tagliari, C.V., Brandão, J.R., Fontanelli, P.R., 2007. Bacia do Espírito
762 Santo. *Bol. Geocienc. Petrobras* 15, 501 - 509.

763 Gamboa, D., Alves, T., Cartwright, J., 2011. Distribution and characterization of failed
764 (mega)blocks along salt ridges, southeast Brazil: Implications for vertical fluid flow on
765 continental margins. *Journal of Geophysical Research: Solid Earth* 116, B08103.

766 Gamboa, D., Alves, T.M., Cartwright, J., 2012. A submarine channel confluence classification for
767 topographically confined slopes. *Marine and Petroleum Geology* 35, 176-189.

768 Gaullier, V., Brun, J.P., Gue'rin, G., Lecanu, H., 1993. Raft tectonics: the effects of residual
769 topography below a salt de'collement. *Tectonophysics* 228, 363-381.

770 Isacks, B.L., 1988. Uplift of the Central Andean Plateau and bending of the Bolivian Orocline.
771 *Journal of Geophysical Research: Solid Earth* 93, 3211-3231.

772 Jackson, M., Cramez, C., 1989. Seismic recognition of salt welds in salt tectonics regimes, Gulf of
 773 Mexico salt tectonics, associated processes and exploration potential: Gulf Coast Section
 774 SEPM Foundation 10th Annual Research Conference, pp. 66-71.

775 Kumar, N., Gamboa, L., Schreiber, B., Mascle, J., 1977. Geologic history and origin of Sao Paulo
 776 Plateau (Southeastern Brazilian Margin), comparison with the Angolan margin and the early
 777 evolution of the Northern South Atlantic. Supko, PR, Perch-Nielsen, K. (Eds.), 927e945.

778 Lentini, M.R., Fraser, S.I., Sumner, H.S., Davies, R.J., 2010. Geodynamics of the central South
 779 Atlantic conjugate margins: implications for hydrocarbon potential. *Petroleum Geoscience* 16,
 780 217-229.

781 Lima, C., 2003. Ongoing compression across South American plate: observations, numerical
 782 modelling and some implications for petroleum geology. Geological Society, London, Special
 783 Publications 209, 87-100.

784 Mauduit, T., Guerin, G., Brun, J.P., Lecanu, H., 1997. Raft tectonics: the effects of basal slope
 785 angle and sedimentation rate on progressive extension. *Journal of Structural Geology* 19, 1219-
 786 1230.

787 Mckee, E.H., Noble, D.C., 1982. Miocene volcanism and deformation in the western Cordillera and
 788 high plateaus of south-central Peru. *Geological Society of America Bulletin* 93, 657-662.

789 Mégard, F., 1984. The Andean orogenic period and its major structures in central and northern Peru.
 790 *Journal of the Geological Society* 141, 893-900.

791 Mégard, F., Noble, D.C., McKee, E.H., Bellon, H., 1984. Multiple pulses of Neogene compressive
 792 deformation in the Ayacucho intermontane basin, Andes of central Peru. *Geological Society of*
 793 *America Bulletin* 95, 1108-1117.

794 Mello, M. R., Maxwell, J. R., 1991. Organic geochemical and biological marker characterization of
 795 source rocks and oils derived from lacustrine environments in the Brazilian continental margin.
 796 In: Katz, B. J. (ed.) *Lacustrine Basin Exploration – Case Studies and Modern Analogs*. AAPG,
 797 *Memoir*, 50, 77–98.

798 Mohriak, W.U., 2003. Bacias Sedimentares da Margem Continental Brasileira. *Geologia, Tectônica*
799 e Recursos Minerais do Brasil, CPRM, São Paulo, Capítulo III In: Bizzi, L. A.,
800 Schobbednhaus, C., Vidotti, R. M. & Gonçalves, J. H. (eds), 87 - 165.

801 Mohriak, W.U., 2005. Interpretação geológica e geofísica da Bacia do Espírito Santo e da região de
802 Ambrolhos: Petrografia, datação radiométrica e visualização sísmica das rochas vulcânicas. .
803 Bol. Geocien. Petrobras 14, 73 - 87.

804 Mohriak, W.U., Nemcok, M., Enciso, G., 2008. South Atlantic divergen margin evolution:
805 rift□bored uplift and salt tectonics in the basins of Southeastern Brazil. In: R.J. Pankhurst,
806 R.A.J. Trouw, B.B. Brito Neves and M.J. de Wit (Eds.), *West Gondwana pre□Cenozoic*
807 correlations across the South Atlantic region. Geological Society London, Special Publications
808 294, 365 - 398.

809 Mora, A., Parra, M., Strecker, M.R., Kammer, A., Dimaté, C., Rodríguez, F., 2006. Cenozoic
810 contractional reactivation of Mesozoic extensional structures in the Eastern Cordillera of
811 Colombia. *Tectonics* 25, TC2010, doi: 10.1029/2005TC001854.

812 Ojeda, H.A.O., 1982. Structural framework, stratigraphy, and evolution of Brazilian marginal
813 basins. *AAPG Bulletin* 66, 732-749.

814 Omosanya, K.d.O., Alves, T.M., 2013. Ramps and flats of mass-transport deposits (MTDs) as
815 markers of seafloor strain on the flanks of rising diapirs (Espírito Santo Basin, SE Brazil).
816 *Marine Geology* 340, 82-97.

817 Ortiz-Karpf, A., Hodgson, D.M., McCaffrey, W.D., 2015. The role of mass-transport complexes in
818 controlling channel avulsion and the subsequent sediment dispersal patterns on an active
819 margin: The Magdalena Fan, offshore Colombia. *Marine and Petroleum Geology* 64, 58-75.

820 Penge, J., Munns, J.W., Taylor, B., Windle, T.M.F., 1999. Rift–raft tectonics: examples of
821 gravitational tectonics from the Zechstein basins of northwest Europe. Geological Society,
822 London, *Petroleum Geology Conference series* 5, 201-213.

- 823 Pilcher, R.S., Murphy, R.T., Ciosek, J.M., 2014. Jurassic raft tectonics in the northeastern Gulf of
824 Mexico. Interpretation 2, SM39-SM55.
- 825 Roure, F., Colletta, B., De Toni, B., Loureiro, D., Passalacqua, H., Gou, Y., 1997. Within-plate
826 deformations in the Maracaibo and East Zulia basins, western Venezuela. Marine and
827 Petroleum Geology 14, 139-163.
- 828 Rowan, M.G., Jackson, M.P.A., Trudgill, B.D., 1999. Salt-related fault families and fault welds in
829 the northern Gulf of Mexico. AAPG Bulletin 83, 1454-1484.
- 830 Rowan, M.G., Lawton, T.F., Giles, K.A., 2012. Anatomy of an exposed vertical salt weld and
831 flanking strata, La Popa Basin, Mexico. Geological Society, London, Special Publications 363,
832 33-57.
- 833 Sarmiento, L.F., Rangel, A., 2004. Petroleum systems of the Upper Magdalena Valley, Colombia.
834 Marine and Petroleum Geology 21, 373-391.
- 835 Scheuber, E., Bogdanic, T., Jensen, A., Reutter, K.-J., 1994. Tectonic Development of the North
836 Chilean Andes in Relation to Plate Convergence and Magmatism Since the Jurassic, in:
837 Reutter, K.-J., Scheuber, E., Wigger, P. (Eds.), Tectonics of the Southern Central Andes.
838 Springer Berlin Heidelberg, pp. 121-139.
- 839 Vendeville, B.C., 2005. Salt tectonics driven by sediment progradation: Part I—Mechanics and
840 kinematics. AAPG Bulletin 89, 1071-1079.
- 841 Vieira, P.E., Bruhn, H.L.C., Santos, C.F., Del Rey, A.C., Alves, R.G., 2007. Golfinho field-
842 discovery, development, and future prospects. Offshore Technology Conference, Houston,
843 Texas.

844

845 **Figure Captions**

846

847 Fig. 1 – a) Map of southeast Brazilian margin highlighting the location of the study area (Block
848 BES 100). The map shows the main structural elements that separate Espírito Santo from the
849 Campos and Santos basins. Note the prominent bathymetric high (Abrolhos Plateau) that dominates
850 the northern half of the Espírito Santo Basin, and the presence of an East-West seamount chain
851 (Vitória-Trindade Chain) to the east of the study area. Raft movement in most of the study area was
852 from the northwest and west towards the southeast and east. b) Structural map with interpreted rafts
853 summarising the relative position of rafts in the study area, and highlighting the geometry of North-
854 South Albian rafts on the continental slope of Espírito Santo. Numbers 1 to 4 denote the distinct
855 rafts referred to in the text. c) Interpreted West to East seismic profile highlighting the style of raft
856 tectonics, and geometry of surrounding units, for general context. The top and base of Rafts 1 to 3
857 are observed in the seismic section. Only the top horizon is observed in Rafts 4 to 6.

858

859 Fig. 2 - Seismic profile highlighting the presence of reactivated structures (including local pop-up
860 structures) in the study area. Highlighted are also roller faults (RF), rollover faults (RoF), keystone
861 faults (KF) and reactivated Faults (RvF).

862

863 Fig. 3 – a) Correlation panel between the interpreted seismic units and stratigraphic information
864 from the Espírito Santo Basin based on França et al. (2007). Velocity data for ODP Site 516 was
865 taken from Barker et al. (1983). Maximum thickness and depositional environments of the
866 interpreted units are also shown in the figure. b) Schematic representation highlighting the study
867 area on the continental slope of Espírito Santo Basin as modified from Fiduk et al., (2004); Gamboa
868 et al., (2010) and Omosanya and Alves (2013). SR – Syn-Rift sequence, T – Transitional sequence,

869 ED – Early Drift sequence, LD – Late Drift sequence. The raft tectonics area is located in the
870 proximal extensional domain (dashed square).

871

872 Fig. 4 – Simplified schematic evolution of raft tectonics during the Albian-Santonian period in the
873 study area (modified from Duval et al., 1992 and Pilcher et al., 2014). In a first stage (1), early rafts
874 are formed together with extensional faults. In the second stage (2), the post-raft overburden fills
875 the gaps between the rafts. In the last stage (3), the tabular rafts remained isolated become
876 progressively welded on the pre-salt strata. Note the erosion at the end of this stage (Santonian).
877 The salt accumulated into salt rollers, pillows and the rafts growth internally. In the last two stages
878 are observed extensional faults into the post-raft overburden that laterally confined the raft and/or
879 the salt accumulations. The arrows indicate the slope direction in the Espírito Santo Basin (not to
880 scale).

881

882 Fig. 5 - Seismic profile highlighting a phase of widespread movement and erosion of rafts at the end
883 of the Cretaceous (Horizons 4 and 5). As with other figures, the seismic profile shows roller faults
884 (RF), rollover faults (RoF), keystone faults (KF) and reactivated faults (RvF). The rafts reactivation
885 is observed on the base Santonian unconformity, showing local pop-up and tight anticlinal
886 structures (square dashed line).

887

888 Figure 6 - Seismic profile showing the geometry of collapsed rafts (see Horizon 3 and 4 for
889 reference). As with other figures, the seismic profile shows roller faults (RF), rollover faults (RoF),
890 keystone faults (KF) and reactivated faults (RvF). In this profile, raft 2 collapsed by probable
891 withdrawal of salt from underneath.

892

893 Fig. 7 – TWT structure and isochron maps of key horizons in the study area. a) TWT of the top rafts
894 horizon 3, showing the relative location of rafts 1 to 6. b) Isochron map for strata between top rafts

895 (horizon 3) and base Santonian (horizon 4). c) Isochron map for strata between horizons 4 and 5
896 (Santonian to top Maastrichtian). Note the marked variations in thickness in these last two maps.
897

898 Fig. 8 – Seismic profile highlighting the principal fault families related to raft movements. The
899 figure shows roller faults (RF), rollover faults (RoF), keystone faults (KF) and reactivated faults
900 (RvF). The rafts reactivation is observed on the base Santonian unconformity, showing local pop-up
901 and tight anticlinal structures (square dashed line). The main horizons considered for thickness plots
902 in Figure 12 are also pointed out: base and top rafts, base Santonian and seafloor. The figure include
903 a line (in grey) for horizon reference.

904
905 Fig. 9 – Seismic profiles highlighting the major fault types triggered by the movement of rafts and
906 post-raft overburden: a) Roller faults; b) Keystone faults; c) Crystal faults; b) reactivated faults. See
907 Figure 1b for location seismic profiles.

908
909 Fig. 10 – Structural maps for key horizons mapped in the study area: a) seafloor, b) Eocene
910 unconformity (horizon 6), c) top Maastrichtian unconformity (horizon 5), d) intra-Santonian
911 unconformity (horizon 4). Note the marked faulting of the mapped horizons.

912
913 Fig. 11 – Profile North-South above raft 2. It is showing the elongated body of raft 2. The figure
914 highlights any interpreted horizons together with main sedimentary and structural bodies in the
915 study area. Dashed line (grey) included for reference.

916
917 Fig. 12 – Thickness plots for overburden strata above Albion rafts in the study area, acquired in a N-
918 S direction. Data and trend lines refer to the intervals top raft to base Santonian (in grey), base
919 Santonian to seafloor (dashed line, black), and total overburden thickness above Albion rafts
920 (black). The graphs highlight the existence of thicker overburden units towards the central region in

921 rafts 1, 5 and 6. Conversely, rafts 2, 3 and 4 show the thickest overburden units to the north of the
922 study area. This character contrasts with highest degree of internal deformation recorded in the
923 northern and central parts of the interpreted rafts, as explained in more detail in the text.

924

925 Fig. 13 – a) Amplitude map from 25 ms-thick window below the base Santonian showing the main
926 fault families that intersect Horizon 4 (base Santonian); b) interpretation based on the amplitude
927 map (Fig. 13a) highlighting the faults families; c) block diagram through segment A-B (Fig. 13a),
928 with ~5x vertical exaggeration. It shows rafts 2, 4 and 5, and the main roller faults adjacent to the
929 rafts.

930

931 Fig. 14 – a) Uninterpreted and b) interpreted West to East seismic profile showing gentle internal
932 strata growth in raft 2. Note the presence of growth raft strata above the salt roller to the east, and
933 the initiation of a triangular-shaped structure above raft 2. The raft is lateral confined by salt
934 structures, salt roller to the west and salt pillow to the east.

935

936 Fig. 15 – a) Uninterpreted and b) interpreted West to East seismic profile showing collapsed lateral
937 part of raft 2, listric normal faults and raft welded on the pre-salt units.

938

939 Fig. 16 - a) Uninterpreted and b) interpreted West to East seismic profile showing the structural
940 deformation in raft 2. Deformation styles include the tilting of flanking strata, ramping up on the
941 salt structures and collapse of the central part of raft 2, lateral constrained by extensional faults and
942 welded on the pre-salt units. In the figure is highlighted the thickness (m) between the i) top raft to
943 base Santonian and ii) base Santonian to seafloor, for reference.

944

945 Fig. 17 - Conceptual schematic evolution of rafts in the study area, highlighting the effect of salt
946 pillow growth on the structural compartmentalisation of Albian (and younger) strata in the Espírito

947 Santo Basin. Fault systems in the figure are associated with different styles of raft deformation, as
948 described in this paper. Modified from Alves (2012).

949

950 Figure 18 - Uninterpreted and b) interpreted West to East seismic profile highlighting the presence
951 of fluid-flow features above listric faults in raft 3. Fluid putatively migrates from pre-salt and intra-
952 raft units to accumulate above the listric faults in Late Cretaceous strata. Note the presence of
953 propagated Rollover faults into Cenozoic units to the East of raft 3.

a)

b)

Stage 1

Stage 2

Stage 3

- Post-raft overburden
- Raft
- Salt
- Pre-salt strata

Structure map (TWT) - Top Rafts (Horizon 3)

Isochron map - Top Rafts (Horizon 3) to Santonian (Horizon 4)

Isochron map - Santonian (Horizon 4) to Top Maastrichtian (Horizon 5)

Seafloor TWT map

a)

Mid Eocene (Horizon 6) TWT map

b)

Top Maastrichtian (Horizon 5) TWT map

c)

Base Santonian (Horizon 4) TWT map

d)

a)

b)

c)

d)

West
East

Present day configuration of the Espirito Santo Basin

STAGE 4: Middle/Late Eocene

STAGE 3: Early Santonian

STAGE 2: Post-Albian to Coniacian

STAGE 1: Albian