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# 1 **Salt morphologies and crustal segmentation relationship: new insights from**

## 2 **Western Mediterranean Sea and other salt passive margins**

3  
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### 15 **Abstract**

16 Salt tectonics at salt-bearing margins is often interpreted as the combination of gravity  
17 spreading and gravity gliding, mainly driven by differential sedimentary loading and margin  
18 tilting, respectively. Nevertheless, in the Western Mediterranean Sea, the classical salt  
19 tectonics models are incoherent with its morpho-structural setting: the Messinian salt was  
20 deposited in a closed system, formed several Ma before the deposition, horizontally in the  
21 entire deep basins, above a homogenous multi-kilometre pre-Messinian thickness. The  
22 subsidence is purely vertical in the deep basin, implying a regional constant initial salt  
23 thickness, the post-salt overburden is homogenous and the distal salt deformation occurred  
24 before the mid-lower slope normal faults activation.

25 Instead, the compilation of MCS and wide-angle seismic data highlighted a clear coincidence  
26 between crustal segmentation and salt morphology domains. The salt structures change  
27 morphology at the boundary between different crustal natures. Regional thermal anomalies  
28 and/or fluid escapes, associated with the exhumation phase, or the mantle heat segmentation,  
29 seem therefore play a fundamental role in adding a further component on the already known  
30 salt tectonics mechanisms. The compilation of crustal segmentation and salt morphologies in  
31 different salt-bearing margins, such as the Santos, Angolan, Gulf of Mexico and Morocco-  
32 Nova Scotia margins, seems to depict the same coincidence. In view of what observed in

33 Western Mediterranean Sea, the heat segmentation influence in the passive margins should  
34 not overlooked and deserves further investigation.

35

36 **keywords:** salt tectonics; passive margin; Western Mediterranean Sea; thermicity; Messinian  
37 Salinity Crisis

38

39

## 40 **1. Introduction**

41 The study of salt tectonics is a widely discussed topic, especially since 1990 and mainly due  
42 to its importance in oil exploration. Under geological strain rate and due to its rheological  
43 properties, salt or halite, flows like a fluid causing spectacular geometries. Triggering can be  
44 induced by several mechanisms, including gravity, tectonic force, salt buoyancy or  
45 differential loading (Ge et al., 1997; Hudec and Jackson, 2007). The latter is dominant in  
46 sedimentary salt-bearing basins. On passive margins, salt tectonics is distributed into three  
47 domains (Jackson et al., 1994; Letouzey et al., 1995; Ings and Shimeld, 2006; Hudec and  
48 Jackson, 2007): an extensional zone on the slope dominated by basement-involved or  
49 basement-detached reactive diapirism (Duval et al., 1992; Jackson et al., 1994; Vendeville et  
50 al., 1995), a mid-slope more or less deformed transitional domain (Cramez and Jackson, 2000;  
51 Dooley et al., 2013, Ge et al., 2019) and an area of contraction characterized by salt  
52 thickening, thrust faults and folding (Hudec and Jackson, 2006). Regional differential loading  
53 mechanism induced by detachment surface slope and/or sedimentary loading are often  
54 summarized as the cause of gravity gliding and gravity spreading end-member models. The  
55 discussion of which of the two models prevails is still debated (Brun and Fort, 2011; 2012;  
56 Rowan et al., 2012; Schultz-Ela, 2001; Peel, 2014). During gravity spreading, salt moves by  
57 sedimentary differential loading (Ge et al., 1997; Jackson and Talbot, 1986; Vendeville, 2005;  
58 Hudec and Jackson, 2007) whilst during gravity gliding, the salt and its overburden slide  
59 towards the deep basin (Cobbold et al., 1989; Demercian et al., 1993; Fort et al., 2004)  
60 triggered by the inclination of the margin occurring during thermal subsidence. During the  
61 well-known Messinian Salinity Crisis (MSC) in the Mediterranean Sea, evaporite deposition  
62 (up to 3 km thick, Haq et al., 2020) was triggered by a combination of global tectonic events,  
63 which led to the closure of the Strait of Gibraltar (e.g. Leroux et al., 2019) and climatic  
64 evolution (e.g. CIESM, 2008). Salt tectonics in the Liguro-Provençal basin has been studied  
65 since the works of Le Cann, (1987), Gaullier, (1993) and Gorini,(1993) and, along with recent  
66 studies, is outlined as the combination of several parameters: (i) differential loading and

67 gliding of the brittle-ductile series during gravity spreading (Obone Zue Obame et al., 2011),  
68 (ii) initial conditions and morpho-structure evolution of the basin (Gaullier et al., 2008), (iii)  
69 local geodynamic setting that could induce thick-skinned tectonics (Maillard et al., 2003;  
70 Déverchère et al., 2005). The application of the two salt tectonic end-member models  
71 highlights problems when we take into account the differences in salt morphologies in the  
72 Western Mediterranean Sea. In the Western Mediterranean Sea, salt was deposited in a closed  
73 basin, already formed, above a thick marine sequence deposited during the Late Oligocene  
74 and Miocene intervals. Moreover, the purely vertical subsidence in the deep basin implies  
75 small difference in initial salt thickness that could have controlled salt supply variations.  
76 We present here a compilation of the relationships between salt morphologies and crustal  
77 nature in well-studied salt-bearing basins throughout the world. We first focus on the Western  
78 Mediterranean Sea, for which we present a new detailed map of salt morphologies. We  
79 categorized regional salt morphologies in order to superimpose them onto the crustal  
80 segmentation, constrained by wide-angle seismic studies, magnetic and gravity maps. We  
81 then discussed the possible causes for this relationship, taking into account the Western  
82 Mediterranean Sea peculiarities. We then applied the same method to South-Atlantic margins  
83 (Angolan and Santos Basins), Gulf of Mexico, Morocco and Nova Scotia margins in order to  
84 observe if this coincidence is also confirmed in different salt tectonics systems.

85

## 86 **2. Particularity of Western Mediterranean salt-bearing basin: a divergent closed and** 87 **fossil system**

88 The Cenozoic opening of the Western Mediterranean in an overall convergence context  
89 between Africa and Eurasia is closely linked to the creation of the Alpine chains, which  
90 surround it (e.g. Carminati et al., 2012). The Algero-Provençal basin opened during the Late  
91 Oligocene-Early Miocene times in the Tethyan subduction zone context. Since Late  
92 Oligocene, the slab retreat is predominant in the formation of back-arc basins, such as the  
93 Alboran, the Liguro-Provençal, the Tyrrhenian, the Aegean basins (Jolivet and Faccenna,  
94 2000). The Liguro-Provençal basin was created by the counter-clockwise rotation of the  
95 Corso-Sardinian block (e.g. Auzende et al., 1973; Olivet, 1996), dated from Late Eocene  
96 (~Priabonian: 33.7 Ma) ranging between 23 to 19 Ma according to authors (Leroux et al.,  
97 2019). The reconstruction of initial conditions of the Liguro-Provençal rifting (Bache et al.,  
98 2010) shows similarities with those of Central (Labails et al., 2009) and South Atlantic  
99 margins (Moulin et al., 2005; Aslanian et al., 2009; Aslanian and Moulin, 2012), such as the  
100 high structural position before the break-up and the strong subsidence that followed.

101 While there is a general consensus on the geodynamics of the Liguro-Provençal basin, the  
102 opening direction of the Algerian basin is still a matter of debate, as well as the initial position  
103 of the AlKaPeCa (Alboran, Kabylies, Peloritan and Calabria) blocks. There are several  
104 hypotheses regarding the opening direction and the nature of the margins (Leroux et al.,  
105 2019). The first group of authors claim a dominant N-S opening (e.g. Gueguen et al., 1998;  
106 Frizon de Lamotte et al., 2000; Rosenbaum et al., 2002; Schettino & Turco, 2006) with no  
107 significant displacement of the Alboran block (Faccenna et al., 2004). The second hypothesis  
108 claims an E-W opening with an important westward migration of the Alboran block (e.g.  
109 Malinverno & Ryan, 1986; Royden, 1993). The third hypothesis proposes a two-step opening  
110 scenario, with a first N-S movement which led to the opening of the East Algerian basin and a  
111 second E-W, with the opening of the West Algerian basin (Etheve et al., 2016; Lepretre et al.,  
112 2018). Another model claimed by Vergés and Fernández, (2012) proposes an initial SE-  
113 dipping subduction that shifted to a faster E-dipping subduction which could explain some  
114 observations made in the Betic-Rif orogenic system, including the Alboran back-arc basin  
115 opening.

116 Other authors do not agree with the rollback process and propose a collapse of the thickened  
117 continental crust (Dewey, 1989) or a lithospheric delamination (Roure et al., 2012).

118 In the Algerian basin, the Plio-Quaternary period is affected by tectonic inversion caused by  
119 the NW-SE convergence between the African and European plates with a slow rate of  
120 shortening (0.5 to ~1 cm/yr) (Kherroubi et al., 2009). This still present-day shortening causes  
121 deformation and tectonic inversion offshore Algeria and even in the SE Iberian margin  
122 (Maillard and Mauffret, 2013). The Algerian tectonic inversion is the only remarkable post-  
123 salt deposition tectonic activity which has an impact on the present-day salt structures.

124 About 6 million years ago (~5.96- 5.32 Ma, Krijgsman et al., 1999a), when the Algerian and  
125 Liguro-Provençal basins were formed, the Mediterranean underwent a rapid and extreme  
126 paleo-environmental event known as Messinian Salinity Crisis. The closure of the Strait of  
127 Gibraltar (Hsu et al., 1973), or deeper lithospheric processes (Garcia-Castellanos and  
128 Villasenor, 2011), caused a significant sea-level fall and the deposition of evaporites  
129 associated to considerable erosion mostly located in the upper parts of the margin (e.g.  
130 Clauzon et al., 1996; Rouchy and Caruso, 2006). In the Western Mediterranean Sea, the  
131 Messinian deposits have been described as a trilogy, composed of lower evaporites, salt and  
132 upper evaporites (Montadert et al., 1970; Réhault et al., 1985). The salt layer is considered to  
133 be composed of pure halite, while the evaporites can be interpreted as intercalations of  
134 gypsum/anhydrite and clastics deposits. The uppermost part of Upper Evaporites (also called

135 Upper Unit, Lofi et al., 2011, 2018) has been sampled during DSDP Leg XIII (Hsü et al.,  
136 1973) revealing dolomitic marls and anhydrite layers. Nevertheless, the vertical and lateral  
137 facies variability suggest that the lithology of the Upper Unit can vary at the basin scale (Lofi  
138 et al., 2011). The Upper and Lower evaporites are considered less ductile than the pure halite,  
139 as anhydrite creeps slowly than gypsum which in turn creeps slowly than halite (Jackson and  
140 Hudec, 2017). In most cases, they brittly accommodated the underlying salt deformation.

141 The Messinian units are deposited above a thick Miocene sedimentary sequence characterized  
142 by thickness variations between the Western Mediterranean sub-basins. The Oligocene syn-  
143 rift to Miocene post-rift sedimentary units are composed, by analogy with onshore drill (e.g.  
144 onshore Kabyle, Algeria, Géry et al., 1981; Onshore Camargue, France, Triat and Truc,  
145 1983), by silty marl, conglomerates and sandstones, showing facies from shallow water to  
146 deep marine. In seismic, these deposits onlaps the acoustic basement reflector and fill the  
147 hollows in the relief. At present day, they are at basin scale sub-horizontal, parallel to the base  
148 of salt and mostly unfaulted.

149 Salt is ductile and, in the Western Mediterranean Sea, is always deposited as flat horizontal  
150 layer above the pre-Messinian units. The present-day slope of this surface thus gives the post-  
151 deposition movement, mainly caused by differential subsidence. In the Gulf of Lion, first  
152 Rabineau et al., (2014) with a 2D dataset and then Leroux et al., (2015) with a 3D grid  
153 calculated using sedimentary geometries the subsidence up to present day. They define three  
154 subsidence domains, from the shelf to the deep basin, which fit with the hinge lines defined  
155 by refraction studies by Moulin et al., (2015) (see chapter below) and by interpretation by  
156 Bache et al., (2010). On the platform and on the slope, they observed a seaward tilting of  
157 different amplitudes, varying from the Miocene ( $0.06^\circ \text{ Ma}^{-1} - 0.12^\circ \text{ Ma}^{-1}$ ) to the Plio-  
158 Quaternary age ( $0.16^\circ \text{ Ma}^{-1}$ ) (Leroux, 2012; Leroux et al., 2014). Instead, in the central deep  
159 basin, they calculated a purely vertical post-rift subsidence, with a mean rate of 500 m/Ma  
160 since 20 Ma, supposedly to be the end of rifting age (Séranne, 1999). The Gulf of Lion  
161 therefore acts as a sag basin, with features similar to those of the Angolan margin (Moulin et  
162 al., 2005; Aslanian et al., 2009) and Santos basin (Klingelhoefer et al., 2015; Evain et al.,  
163 2015). In this work, the term “sag basin” refers to the sedimentary sequence below the salt,  
164 characterized by flatly layered sedimentation (Moulin et al., 2005; Karner and Gamboa, 2007;  
165 Bache et al., 2010) and purely vertical subsidence.

166 The post-Messinian series (Pliocene and Quaternary deposits) in the Algerian basin has a  
167 lower thickness than in the Liguro-Provençal (0.8 km and 2 km, respectively; from Mauffret

168 et al., 2004), whereas it is roughly similar between the eastern and western Algerian basins  
169 and western Sardinian margins (Dal Cin et al., 2016).

170

### 171 **3. Dataset and method**

172 We compared salt morphologies and crustal domains in Western Mediterranean Sea margins,  
173 Brazilian margin (Santos and Campos basins), Angolan margin, northern Gulf of Mexico  
174 margin, north-west African margin and Nova Scotia margin.

175

#### 176 **3.1 Salt morphologies interpretation**

177 In the Western Mediterranean Sea, we interpreted a large dataset composed of different  
178 resolution seismic lines, collected through the *French GDR Marges* and *Action Marges*  
179 programs, with additional recent seismic data from *OGS* and *Ifremer* (Fig. 1). Reflection  
180 seismic lines have different resolution and quality allowing a good interpretation of deep and  
181 shallow reflectors.

182 In the Santos and Campos basins, we interpreted salt structures from published seismic  
183 profiles (Fig. 9) from Modica et al., (2004); Mohriak et al., (2009; 2012); Torsvik et al.,  
184 (2009); Lentini et al. (2010) ; Scotchman et al., (2010) ; Unternehr et al., (2010) ; Zalan et al.,  
185 (2011) ; Davison et al. (2007 ; 2012) ; Kumar et al., (2013) ; Moulin et al., (2012) ; Peron-  
186 Pinvidic et al., (2013) ; Jackson et al. (2015) ; Strozyk et al., (2017) ; Kukla et al., (2018).

187 For the Angola margin, we interpreted profiles (Fig. 10) from Marton et al., (2000); Tari et  
188 al., (2003) ; Hudec and Jackson, (2004) ; Moulin et al., (2005) ; Unternehr et al., (2010) ; Fort  
189 et al., (2004) ; Cowie et al., (2016).

190 For the Gulf of Mexico, we compiled salt morphologies from several published seismic  
191 profiles (Fig. 11) (Wu et al., 1990; Radovich et al., 2007; Pilcher et al., 2011; Rodriguez,  
192 2011; Fort and Brun, 2012; Hudec and Jackson, 2013a; Ismael, 2014; Pindell et al., 2014;  
193 Rowan, 2014; Weimer et al., 2017; Curry et al., 2018; Liu et al., 2019) and from already  
194 published salt-distribution maps (Seni, 1992; Diegel et al., 1995; Peel et al., 1995)

195 In the Morocco and Nova Scotia conjugated margins (Fig. 12), the salt structures originate  
196 from works by Hinz et al., (1982); Keen and Potter, (1995); Sahabi, (2004); Shimeld, (2004);  
197 Neumaier et al., (2016); Tari et al., (2000, 2003, 2013)

198

#### 199 **3.2 Crust segmentation**

200 The crustal segmentation is constrained by wide-angle profiles from ECORS and SARDINIA  
201 Experiments (Pascal et al., 1993; Gailler et al., 2009; Moulin et al., 2015; Afilhado et al.,

202 2015) in the Liguro-Provençal basin and by Spiral Project (Leprêtre et al., 2013; Medaouri et  
203 al., 2014; Mihoubi et al., 2014; Badji et al., 2015; Bouyahiaoui et al., 2015; Hamai et al.,  
204 2015, Arab et al., 2016a, 2016b; Aïdi et al., 2018) in the Algerian Basin. The lack of wide-  
205 angle surveys in the South Balearic basin compels us to use the segmentation of this margin  
206 derived from gravity and magnetic data interpretations (Leroux et al., 2019).

207 In the Santos and Campos basins, we used the crustal interpretation from the SanBa Project  
208 (Evain et al., 2015) and from the Zaiango Project (Contrucci et al., 2004; Moulin et al., 2005;  
209 Aslanian et al., 2009) in the Angolan margin. In Gulf of Mexico, the segmentation is from the  
210 refraction seismic project GUMBO (Eddy et al., 2014, 2018; Christeson et al., 2014; Van  
211 Avendonk et al., 2015) and interpretation from (Hudec et al., 2013a; Sawyer et al., 1991 and  
212 Curry et al., 2018). In Morocco-Nova Scotia conjugated margins the crustal segmentation is  
213 taken from Biari et al., (2015); Contrucci et al., (2004); Funck et al., (2004); Wu et al., (2006).

214

#### 215 **4. Western Mediterranean salt deformation**

##### 216 **4.1 Salt tectonics mechanisms**

217 Given its rheological properties, salt can acts as a level of *décollement* and can flow driven by  
218 sedimentary loading and margin tilting (e.g. Ge et al., 1997; Vendeville, 2005; Hudec and  
219 Jackson, 2007; Gaullier et al., 2006; Brun and Fort, 2011; Rowan et al., 2012).

220 In the Western Mediterranean, salt tectonics was previously locally described by some  
221 authors, as in the Gulf of Lion (Le Cann, 1987; Pautot et al., 1984; Gaullier, 1993; Gorini,  
222 1993; Dos Reis, 2001; Dos Reis, 2005, 2008; Maillard et al., 2003; Granado et al., 2016), in  
223 the Provence margin (Obone-Zue-Obame et al., 2011), in west Sardinian margin (Geletti et  
224 al., 2014; Dal Cin et al., 2016), in South Balearic basin (Camerlenghi et al., 2009; Dal Cin et  
225 al., 2016; Blondel et al., 2020) or in the north Algerian margin (e.g. Déverchère et al., 2005;  
226 Domzig et al., 2006; Arab et al., 2016a). Following these works, the salt structures in the  
227 Western Mediterranean Sea are mainly the result of: i) multi-directional thin-skinned tectonics  
228 driven by the seaward dipping base of salt and the sedimentary loading given by thickness  
229 differences in the overburden, ii) the local geodynamic context which, via differences in sub-  
230 salt setting, could influence the salt deformation, as shown by several authors in several other  
231 passive margins (e.g. Haddou and Tari, 2007; Evans and Jackson, 2020; Pichel et al., 2020).  
232 The flowing of salt along a detachment surface in passive margin is a well-known process  
233 (Brun & Fort, 2011; Hudec and Jackson, 2004; Jackson et al., 1994, Schultz- Ela, 2001) and  
234 causes three kinematically linked zone: i) an updip area of extension, ii) a mid-slope transition  
235 and a iii) downdip contraction areas, as also observed in the Gulf of Lion (Le Cann, 1987;



236 Gorini, 1993; Gaullier, 1993; Dos Reis et al., 2005). The updip extension is characterized by  
237 normal basinward-dipped faults, salt rollers and rollover strata and is considered to be coeval  
238 with the thrusting and buckling in the distal contraction domain (Cobbold and Szatmari, 1991;  
239 Marton et al., 2000, Cramez and Johnson, 2000; Tari et al., 2001; Brun and Fort, 2004, 2011;  
240 Rowan et al., 2004). Nevertheless, in the Gulf of Lion, the listric faults were activated only in  
241 the Early Pliocene, as evidenced by growth strata and rollover structures clearly visible above  
242 the Messinian deposits (Fig. 2a): the basinward-dipping normal faults in that proximal domain  
243 were activated only after the complete deposition of the UU, as highlighted by the constant  
244 thickness of this unit involved in the displacement. Conversely, in the deep basin, the  
245 difference of thicknesses and the onlaps in the Upper Unit (UU) (Fig. 2b) suggest a salt  
246 deformation that began with the deposition of this unit, still belonging to the Messinian  
247 Salinity Crisis (Gaullier et al., 2012): the sliding postdates the deep deformation. It seems that  
248 early deformation in the deep basin has induced a space that allowed the salt to glide from the  
249 lower slope, leading to the listric fault formation. The Early Pliocene thin-skinned extension  
250 tectonics therefore appears here to be a consequence of the space created by the distal diapir  
251 formation.

252 Several authors (Le Cann, 1987; Pautot et al., 1984) observed a striking correlation between  
253 the OCT (Ocean-Continent Transition) limit and a variation in salt structures in Gulf of Lion,  
254 as shown in figure 3. Pautot et al., 1984 and Le Cann, (1987) claimed for the reactivation of  
255 rifting-involved faults which could influence the salt deformation and Maillard et al., (2003)  
256 for an influence of differential compaction of subsalt sediments given by basement steps filled  
257 by pre-Messinian sediments.

258 The Messinian salt system peculiarity and the coincidence observed in Gulf of Lion (Fig. 3)  
259 led us to extend the salt structures-crustal segmentation analysis on the global Western  
260 Mediterranean (Fig. 4).

261 We interpreted salt structures in Liguro-Provençal and Algerian basins, leaving aside the  
262 Alboran Sea due to the absence of Messinian halite (e.g. Do Couto et al., 2014) and the  
263 Ligurian basin due to lack of data (Fig. 1). We defined a nomenclature for regional salt  
264 morphologies (Fig. 4) in order to classify them in 5 groups sharing the same geometrical  
265 features. The classification allowed to directly compare the salt morphologies to the crustal  
266 segmentation and nature and then to regionally discuss the salt structures genesis and  
267 evolution. The classification (Fig. 4) is mainly based on geometric criteria, such as: shape,  
268 base diapir width, the geometrical relationship with the overlying layers and associated  
269 dominant features, such as faults, mini-basins (Jackson and Talbot, 1992) or welds (Jackson

270 and Cramez, 1989). Although every salt structure is the result of processes interaction, we  
271 tried to discriminate them only by their geometrical characteristics, without a detailed  
272 discussion on genesis and timing involved. The salt morphologies groups are then plotted on  
273 map, showing wide variations between the proximal and distal domains and between the sub-  
274 basins composing the Western Mediterranean Sea (Fig. 5).

275

## 276 **4.2. Salt morphologies description and distribution**

277 In this chapter we will describe the main present-day salt morphologies characteristics (Fig. 4)  
278 and their spatially distribution along the Western Mediterranean Sea (Fig. 5). Figure 6 (A, B,  
279 C, D, E) and Figure 7 (F, G, H, I) present nine seismic transects across the different sub-  
280 basins.. The profiles provide a regional view of salt morphologies following the nomenclature  
281 described in fig. 4 and, respectively, highlight: (1) the Liguro-Provençal basin (LPB): the Gulf  
282 of Lion margin and the western Sardinia margin (Figs. 6A; 7F,G); (2) the east Algerian Basin  
283 (EAB) and the southern part of Liguro-Provençal Basin by crossing the North Balearic  
284 Fracture Zone (NBFZ, e.g. Olivet, 1996) (Figs. 6B,C; 7H ); (3) the western Algerian Basin  
285 (WAB) and the EAB by crossing the Hannibal Area (Fig. 6C); (4) the eastern and western  
286 part of the WAB (Figs. 6D,E; 7I). These long transects show the relationship between salt  
287 morphologies and crustal segmentation all along the Western Mediterranean Sea and the  
288 regional salt structures variations, here summarized:

289 • The Listric domain is characterized by salt rollers and rollover anticlines and basinward-  
290 dipping normal faults involving the displacement of the Upper Unit and Plio-Quaternary  
291 deposits (e.g. Figs. 6A; 7I). The faults can be still active or buried (e.g. Gulf of Lion, Dos  
292 Reis, 2001) and are located at the mid-lower slope of the margins. The sliding of salt along its  
293 base detachment induces a sediments subsidence and thinning of salt, which is proportional to  
294 the density and thickness of the overburden and the initial salt layer thickness (Vendeville,  
295 2005), and does not affect the pre-salt sequence (thin-skinned tectonics). The structures  
296 formed by the salt movement are the results of extensional process, which involves salt and  
297 overburden basinward movements. This domain is located at mid-lower continental slope of  
298 the margins (Figs. 6; 7) and is always present albeit with some differences in amplitude,  
299 mainly given by base of salt steepness and wideness of the margin (Fig. 5): in the Gulf of  
300 Lion it occupies an area that can extend up to 50 km basinward from the upslope limit of salt  
301 (Fig. 6A) while in the Algerian margin and in the east Alboran margin is narrower, of 10-20  
302 km (Figs. 6B,D; 7I). In the south Balearic and in the west Sardinian margins is between 10  
303 and 30 km (Figs. 6C,E; 7F,G,H). As mentioned above (Fig. 2a), the listric faults activated

304 after the deposition of the Upper Unit and are mainly active in the Pliocene but they can reach  
305 the seafloor (Badhani et al., 2020).

306 • The Anticlinal salt domain (Fig. 4) is characterized by the absence of normal faults rooted at  
307 the base of salt and it consists of anticlinal salt-cored and sparsely distributed normal faults at  
308 the top. The outer-arc extension faults are mainly located at the top of Upper Unit affecting  
309 the Lower Pliocene sequence. The base of salt can be sub-horizontal (Gulf of Lion; Fig. 3) or  
310 even dipping landward, as in the north Algerian margin (Fig. 6B,D). The base of anticlines is  
311 large and the ratio between length and height is greater than five. The Anticlinal domain is  
312 located in the proximal deep basins (Fig. 5), characterized by a purely vertical subsidence (as  
313 shown in Gulf of Lion by Rabineau et al., 2014; Leroux et al., 2015) as also attested by sub-  
314 horizontal pre-salt deposits. It is not present in the western Sardinia, south Balearic and north-  
315 western Algerian margins, where basinward the Listric domain the deep basins are dominated  
316 by diapirs structures characterized by discordant overburden (Figs. 6C,D,E; 7F,G,H). As  
317 shown in Gulf of Lion (Fig. 2b), the anticlinal began to grow early, just after their deposition,  
318 during the deposition of the Upper Unit still belonging to the Messinian Salinity Crisis time.  
319 The evolution is then characterized by period of quiescence during the Pliocene and strong  
320 deformation in the Quaternary.

321 • The Large Isolated Diapir (Fig. 4) and the Large Diapirs domains exhibit piercing diapirism  
322 affecting the seafloor. In Gulf of Lion (Fig. 5), diapir crests can rise to 250 m above the  
323 seafloor (Pautot et al., 1984) and the structures are isolated, forming wide mini-basins and  
324 welding together (Fig. 3). In the western Algerian deep basin the structures are less spaced-  
325 apart and generally smaller in size (Figs. 5; 7I). In both basins the overburden is discordant  
326 except for the Upper Quaternary deposits, which are concordant to the top of diapirs. The  
327 overburden deposits could form narrow and steep or, less frequently, gentle drape-folds  
328 geometries on diapirs flanks, typical of hook and wedge halokinetic sequences described by  
329 Giles and Rowan, (2012) (see also Mianaekere and Adam, 2020). Both salt stocks and walls  
330 are present: in the Gulf of Lion the salt walls have a dominantly N-S direction while in the  
331 north Algerian margin they are parallel to the margin, as also interpreted by Badji, (2014) and  
332 Leprêtre, (2012). In the Gulf of Lion they grew far from the upslope salt limit (at least 100 km  
333 basinward; figs. 5; 6A) while in the west Algerian basin they appear closer to the margin  
334 (about 10 km, just basinward the Listric domain; fig. 5) and dominate the deep basin. The  
335 structures show an initial active growth history early after the deposition followed by a  
336 passive down-building which characterized mostly all the Plio-Quaternary diapirs growing.

337 • The Small, Regular and the Narrow, Irregular domains (Fig. 4) are interpreted in the deep  
338 basins (Fig. 5) and are characterized by different-sized connected diapirs (Figs. 6, 7). The  
339 Connected, Regular domain is characterized by closely-spaced diapirs that deform the  
340 Pliocene but very rarely the Quaternary sequences (Fig. 6B,C). The overburden reflectors are  
341 mainly concordant and the width to height ratio is close to one. This domain is located in the  
342 EAB, where a maximum thickness of 0.5 s twt and a regional sub-horizontal base of salt  
343 characterized the entire basin (Figs 5; 6B,C). The Plio-Quaternary sedimentary thickness is  
344 more or less constant in all basin, except in the north eastern Algerian margin, with a mean  
345 thickness of 0.8 s twt. The Connected, Irregular domain is characterized by close and frequent  
346 diapirs (Figs. 6A; 7F). The overburden is mostly discordant and the structures are smaller and  
347 narrower than in the Large Diapirs domain previously described: they have developed more in  
348 height than in width (Fig. 4) and can deform the seafloor.

349 In addition to the salt morphologies domains described above, we interpreted two areas where  
350 the base of salt is around 500 m shallower than in the deep basin (Fig. 5). In the western  
351 Sardinian and south Balearic margins (Figs. 6B,C,D,E; 7F,G,H) the base of salt extends for 50  
352 km (western Sardinian) to 20-100 km (south Balearic) from the Listric domain, located in the  
353 lower slope, and is low basinward tilted. The diapirs structures are close and connected. In the  
354 south Balearic margin (Fig. 6E) the sudden deepening of the salt base could be provoked by  
355 southward basement tilting associated with the compressive reactivation of the Algerian  
356 margin (Leprêtre, 2012).

357 We also observed an area of thin or absent salt (hard to decipher with our seismic data  
358 resolution), concomitant with the Hannibal Area between the east and west Algerian basins  
359 (Figs. 5; 6C). Here the salt is not deformed and, where present, has a maximum thickness of  
360 0.2 s twt. The Hannibal Area is characterized by positive magnetic anomalies (Aidi et al.,  
361 2018) and basement highs (Mauffret et al., 2004), which suggest the presence of magmatic  
362 structures (Fig. 6C).

363

## 364 **5. Crustal segmentation and nature**

365 During the SARDINIA Experiment, three wide-angle profiles in the Gulf of Lion (Moulin et  
366 al., 2015) and three in the Western Sardinian margin (Afilhado et al., 2015) were acquired  
367 (Fig. 1). The authors interpreted each margin in three domains, with possible subdomains, as  
368 also described by Bache et al., 2010 (Figs. 3; 5). These domains are similar between the  
369 conjugate margins, showing a symmetrical distribution (Gailler et al., 2009), except for the  
370 dimensions of the transitional domain, which is wider in the Gulf of Lion (Fig. 5). The first

371 domain, which follows the onshore unthinned continental crust domain, is the “crustal  
372 thinning” domain, area characterized by the main thinning of the crust, mainly focused on the  
373 lower crust where, in the Gulf of Lion, thickness decreases from 15 to 7 km (Fig. 3).  
374 Basinward, the transitional domain is characterized by a relatively high crustal velocity,  
375 varying between 6.0 and 6.4 km/s for the upper layer, and 6.6-7.5 km/s, 6.8-7.25 km/s for the  
376 lower layer, in the Gulf of Lion and West Sardinian, respectively. The lower layer is  
377 interpreted as exhumed lower continental crust (Moulin et al. 2015; Afilhado et al., 2015) or  
378 exhumed lower crust and upper mantle (Jolivet et al., 2015). The central domain is  
379 characterized by a decrease in velocity and velocity gradient in the lower crust (Moulin et al.,  
380 2015) and presents an “atypical” relatively thin (5 km) oceanic crust (e.g. Le Douaran et al.,  
381 1984; Gailler et al., 2009; Moulin et al., 2015). The oceanic crust is thinner than the Atlantic  
382 type but the velocity and gradients confirm the oceanic nature (Gailler et al., 2009) and its  
383 close relationship with the transitional domain, regarding the velocities and the Moho  
384 discontinuity depth (Moulin et al., 2015; Afilhado et al., 2015). The segmentation is  
385 confirmed by gravimetric (Sandwell and Smith, 1997) and aeromagnetic (Galdeano and  
386 Rossignol, 1977) data (Leroux et al., 2019).

387 Whilst in the EAB a well-organized set of N-S to NW-SE magnetic anomalies suggests an  
388 oceanic nature for this basin (e.g. Leroux et al., 2019), in the WAB the absence of clear  
389 aligned magnetic anomalies testifies for a clear oceanic crust, leaving open all questions  
390 concerning the crustal nature and consequently the geodynamic models of the area. The north  
391 Algerian margin was surveyed during the SPIRAL survey (Groupe SPIRAL, 2014) and five  
392 wide-angle profiles were acquired with a more or less N-S direction, from the west to the east  
393 ends of the Algerian basin (Fig. 1). The results show a narrow thinning continental crust  
394 domain (from 20 to 60 km wide) (Fig. 5) along the entire Algerian margin with possible  
395 heterogeneous thinning between upper and lower crust in some transects (Lepretre et al.,  
396 2013; Arab et al., 2016a; Aidi et al., 2018). Likewise, the transitional domain is narrow (Fig.  
397 5), often associated with graben structures at the surface and interpreted by some authors as  
398 the surface expression of the subduction-transform edge propagator (STEP) fault (Govers and  
399 Wortel, 2005) with several accretion axes (Badji et al., 2015; Aidi et al., 2018). The SPIRAL  
400 refraction results show a central part of the Algerian basin characterized by a 5.5 km thick  
401 crust with velocity differences between the sub-basins that lead the authors to interpret their  
402 nature as follow: (i) the EAB is composed of magmatic thin crust (Mihoubi et al., 2014;  
403 Bouyahiaoui et al., 2015); (ii) the Hannibal High could have undergone a magmatic  
404 reworking during the second phase of basin opening (Aidi et al., 2018); (iii) in the WAB the

405 presence of some mantle serpentinization is possible (Aidi et al., 2018). Recently, heat flows  
406 suggested different formation ages (Poort et al., 2020), confirming the differences in  
407 geodynamic sub-basin history.

408

## 409 **6. Discussion**

### 410 **6.1 Geometrical correlation between salt morphologies and crustal segmentation**

411 A correlation exists between salt morphology and the structural segmentation and nature of  
412 each sub-domain of the Western Mediterranean area (Figs. 5; 6; 7). In the Liguro-Provençal  
413 basin the Listric domain is located on the edges of the margins, where the authors interpret a  
414 transitional and/or a thinned crust (Moulin et al., 2015; Afilhado et al., 2015), as well as all  
415 along the north Algerian margin (Lepretre et al., 2013; Mihoubi et al., 2014; Badji et al.,  
416 2015; Bouyahiaoui et al., 2015; Aidi et al., 2018) and at the transition between the WAB and  
417 the Alboran basin (Medaouri et al., 2014). In the western Sardinian margin, the boundaries of  
418 the high salt base domain match well with the transitional crust interpreted by Afilhado et al.,  
419 (2015). In the deep Liguro-Provençal basin the Connected, Irregular salt morphology  
420 corresponds to a proto-oceanic and/or oceanic crust following the interpretation of refraction  
421 data by Moulin et al., (2015) and Afilhado et al., (2015). Instead, a Connected, Regular  
422 domain is interpreted in the EAB, which seems to be constituted of thin oceanic crust, as  
423 evidenced by refraction data (Mihoubi et al., 2014; Bouyahiaoui et al., 2015) and clearly  
424 aligned magmatic anomalies (Leroux et al., 2019). The WAB, whose crustal structure is still a  
425 matter of debate, is characterized by generally more deformed salt structures, with differences  
426 in the eastern and north-western sectors: in the southwest of the South Menorca Block, the  
427 salt structures are connected and the salt base is shallower while the deep basin is  
428 characterized by the deepening of salt and larger morphologies. The transition between  
429 continental and oceanic domains in the Alboran-WAB limit corresponds to a change in salt  
430 morphologies, between an Anticlinal to a Large/Connected domain. The sharp and quick  
431 transition from a thinned continental crust to an oceanic crust along the north Algerian margin  
432 is associated with a quick transition in salt morphologies.

433

### 434 **6.2 Salt morphologies segmentation hypotheses**

435 We have mentioned some peculiar features of the Western Mediterranean which making it a  
436 particular area for the salt tectonics study in passive margin context. Here, the salt was  
437 deposited during the Messinian Salinity Crisis (about 6 million years ago) in a closed system  
438 where all sub-basins considered in this paper were already formed, well after the rifting

439 phases and oceanic spreading. The salt is therefore not involved in any rifting-extension  
440 movement, as some authors claimed for instance in Gulf of Mexico (Hudec et al. 2013b).  
441 Therefore, in the Mediterranean Sea, the salt was deposited everywhere in the deep, already  
442 formed oceanic basins and lower slopes, onlapping the pre-salt sedimentary sequences. The  
443 lower slope (concomitant with the Llistric salt domain) is characterized by a margin basinward  
444 tilting while the deep basin by a purely vertical subsidence, started just after the oceanic  
445 break-up (Rabineau et al., 2014; Leroux et al., 2015; Bache et al., 2010). The vertical  
446 subsidence suggests an approximately constant initial salt thickness in all the deep basins,  
447 without regional differences given by variations in accommodation space during the  
448 deposition. Salt initial thickness is thus envisioned constant in all the deep basin, above a  
449 thick pre-salt sedimentary layer that cover all initial topographic variations, except for the  
450 margin edges, in the lower slope, where probably was thinner. The tilting in the slopes  
451 actually induced a well-developed thin-skinned tectonics, as previously described by different  
452 authors (Gaullier, 1993; Dos Reis et al., 2005; Brun and Fort, 2011). Nevertheless, we have  
453 shown as the salt has started to deform during the Messinian while the listric faults in the  
454 margin were activated only in the Lower Pliocene, as a direct consequence of the deep basin  
455 deformation. The infra-Messinian deformation was active in all the deep basin, as also  
456 observed in the north Algerian (Bouyahiaoui et al., 2015) and western Sardinian margins and  
457 was already observed in the Levant basin (Gvirtzman et al., 2013). If, as we said, we  
458 envisioned a constant salt initial thickness and no salt translation at time, an imbalance of salt  
459 flux (Pichel et al., 2020; Evans and Jackson, 2020) which could have created accommodation  
460 for growth sedimentation and so early differential loading is to exclude. Another possible  
461 interpretation is an erosional phase infra-UU which could have created differential load in the  
462 early overburden, inducing an early salt deformation. An infra-UU unconformity named IES  
463 (Intermediate Erosion Surface) is observed in western Sardinian and south Balearic margin  
464 (Lofi et al., 2018; Geletti et al., 2014) and in the intermediate-depth basins, such as Valencia  
465 and east Corsica basins (Lofi et al., 2011). The IES has an erosional truncation character and  
466 divides the UU into two sub-units. It is tricky to understand if the erosional surface could be  
467 the cause or if it marks the end of the infra-Messinian deformation. In figure 2 the IES is not  
468 observed (probably due to the low resolution of the line) and the deeper sub-unit of the UU  
469 seems to go in onlap on the salt. The relation between the IES surface with respect the infra-  
470 UU salt deformation is therefore still doubtful.

471 Following Bache et al., (2012) the evaporites in Mediterranean deep basin deposited during a  
472 relatively moderate and slow sea-level rise. The salt deposition was probably brief (<50 000

473 years, Bache et al., 2012) and the following dilution of waters led the progressive deposition  
474 of the Upper Unit. The pure halite seismic facies and the rapidity of salt deposition in a  
475 relatively calm and constant sea level rise inclines us to exclude any erosive surfaces inside  
476 the salt that could have triggered differential spatial loading. Moreover, the seismic facies  
477 seems typical of *pure* halite and no regional lithological levels are interpreted, as for instance  
478 depicted in eastern Mediterranean salt deposits (Gorini et al., 2015).

479 The main issue still unresolved is the geometrical coincidence between the salt morphologies  
480 and the crustal segmentation (Figs. 3, 5, 6, 7). As mentioned, the salt is deposited above a  
481 thick (up to 2.5 km in Liguro-Provençal basin), regionally unfaulted, pre-salt sequence. The  
482 base of salt is also unfaulted and do not shows regionally spatially depth variations (Figs. 6;  
483 7). The only margin affected by thick-skinned tectonics is the north Algerian margin, which  
484 undergo to tectonic inversion caused by the NW-SE convergence between the African and  
485 European plates (e.g. Kherroubi et al., 2009). The still present-day shortening causes the base  
486 of salt offset by low angle south-dipping reverse faults (e.g. Domzig et al., 2006; Leprêtre,  
487 2012; Deverchère et al., 2005) and the salt involving in the play of active ramps and flats (Fig.  
488 6B,D,E). The north Algerian margin is however a particular case in the Western  
489 Mediterranean Sea, not observed in other sub-basins. An active tectonic influence is thus to  
490 discard for a possible cause of crustal segmentation-salt morphologies coincidence.

491 As mentioned above, Le Cann, (1987) and Pautot et al., (1984) claimed for the reactivation of  
492 rifting-involved faults which could influence the salt structures distribution and Maillard et  
493 al., (2003) for a differential pre-salt sediments compaction influence. However, there are no  
494 main faults system that could have affect the base of salt, since that the pre-Messinian  
495 sediments are thick and mostly unfaulted in the deep basin. The majority of inherited rifting  
496 faults are no longer actives and there is no correlation between salt deformation and faults  
497 activation timing. We then agree on basement steps relevance, even below a thick  
498 sedimentary sequence which filled any rugosity but, unless probably for the North Balearic  
499 Transfer Zone (Fig. 5), there are no significant step in the basement that could explain such a  
500 difference. Moreover, the physical experiments of Maillard et al., (2003) are unrealistic  
501 scaled, with a basement steps three times the silicone thickness and a simulated base salt tilt  
502 of 3°, not observed in the deep basin. As highlighted in figure 3, at the transitional-oceanic  
503 crust boundary, the basement has small depth variations, the pre-salt is mostly unfaulted and  
504 the base of salt is sub-horizontal. Moreover, the observations of Pautot et al., (1984) and  
505 Maillard et al., (2003) are mainly localized in the Gulf of Lion deep basin, while the crustal



506 segmentation-salt morphologies coincidence seems to be present in all the Western  
507 Mediterranean sea.

508 It is well-known that the sedimentary loading may be a dominant mechanism on passive  
509 margin salt tectonics (e.g. Vendeville, 2005; Rowan et al., 2012; Warsitzka et al., 2013;  
510 Gaullier et al., 2014). The progradation of a brittle sedimentary wedge over a buoyant salt  
511 layer induces pressure gradient differences caused by differential loading (e.g. Vendeville,  
512 2005). In the Gulf of Lion, the Quaternary Rhône Fan represents a large turbidite system  
513 which comprises a 1500-thick accumulation of turbidites and mass-transport deposits (e.g.  
514 Leroux et al., 2017; Droz et al., 2020; Badhani et al., 2020). The strong salt deformation in the  
515 deep Gulf of Lion basin occurred in the Quaternary is often put in relation with this  
516 sedimentary fan, which would be pushed and squeezed the salt toward the distal basin,  
517 forming salt stocks and walls (Large Diapirs domain, Fig. 5). Nevertheless, once again, this  
518 phenomenon cannot explain the coincidence between crustal segmentation and salt  
519 morphologies observed everywhere in Western Mediterranean Sea. Moreover, the Plio-  
520 Quaternary sedimentation is variable in the deep sub-basins. It can reaches 2 km in the Gulf of  
521 Lion basin but is, with smaller thicknesses, almost equal in the EAB and in the WAB (less  
522 than 1 km). As described (Figs. 5; 6; 7), the salt morphologies in the EAB and WAB differ  
523 considerably, questioning the role of sedimentary loading in both deep basins.

524 We propose that, despite being deposited above a thick sedimentary column, the salt layer  
525 could be affected by a crustal and mantle thermal anomaly. The main causes of salt tectonics,  
526 which have been discussed here and which have been known until now, seem to be unable to  
527 explain the coincidence between crustal nature and salt morphology. The thermal  
528 segmentation seems to be the only hypothesis that could explain the observed correlation.  
529 Below, we will try to give some elements to discuss our hypothesis which has still to be  
530 proved and quantified, but which should not overlooked.

531

### 532 **6.3 The potential role of thermicity on salt deformation**

533 Since the hypothesis of thermal convection argued by Talbot, (1978) the heat influence on salt  
534 deformation has intrigued the salt tectonics scientific community. Following this author, the  
535 geothermal gradients can add an extra component to the classical salt tectonics mechanisms.  
536 Salt rheology is mainly described by creep equations, strongly affected by grain salt  
537 properties, water content and temperature (e.g. Li and Urai, 2016). Temperature plays an  
538 important role, causing a decrease in viscosity (Urai et al., 1986, 2008; Carter et al., 1993;  
539 Van Keken et al., 1993; Li and Urai, 2016; Marketos et al., 2016; Peach et al., 2001; Jackson

540 and Hudec, 2017). For instance, Carter et al. (1993) found a strong decrease of salt viscosity  
541 (up to 4 orders of magnitude) from a temperature of 0°C to 100°C. A high temperature can be  
542 localized adjacent to a rise of magma which may be the main cause of its movement  
543 (Schofield et al., 2014). In the southern North Sea, Underhill (2009) suggested a heating  
544 induced by Paleogene dikes as main trigger for the salt movement. Recently, Magee et al.  
545 (2020) suggested a salt flow enhanced by the hot magma emplacement in the Santos basin.  
546 Moreover, the salt morphologies seem to change near the sill-complex: where the sills are  
547 located, salt rollers are formed whilst far from the sill, the area is characterized by diapiric  
548 deformation (Magee et al., 2020). The post-rift volcanic activity is observed in several  
549 margins, as in the oceanic crust of Gulf of Aden (Leroy et al., 2010) and in Newfoundland  
550 and Labrador margins (Karner and Shillington, 2005). In the Gulf of Aden, Lucazeau et al.  
551 (2009) measured very high heat flow concomitant to a volcanic structure, whose activity  
552 lasted well beyond the rifting phase. The magma transfer is close related to fault-zones  
553 (Georgen and Lin, 2003) which could be inherited from the rifting period and can channelize  
554 the heat distribution. Volcanic activity and magmatic intrusions can somehow influence salt  
555 morphologies, but they are local phenomena and do not fully explain the coincidence between  
556 crust and salt, observed at a regional level. More regional thermal anomalies can be produced  
557 by a pronounced crustal heat segmentation. An extremely thinned crust with exhumed mantle  
558 or lower crust can, with the help of fluids, affect the thermal regime in a large transition zone  
559 and lead to regional variations in temperature at the base of the salt. The passive margin  
560 genesis is a complex process, where the lower continental crust and the upper mantle play a  
561 crucial role. The discovery of serpentized mantle in a few transitional domains (Boillot et  
562 al., 1980), the role of lower continental crust exhumation (Moulin et al., 2005; Aslanian et al.,  
563 2009; Afilhado et al., 2015; Moulin et al., 2015; Evain et al., 2015; Loureiro et al., 2018), the  
564 anomalous heat flow in the transitional domain (Dupré, 2003, Lucazeau et al., 2008, Goutorbe  
565 et al., 2008a, Goutorbe et al., 2011), the role of mafic intrusion in the lower continental crust  
566 (Thybo & Nielsen, 2009, Thybo and Artemieva, 2013; Tozer et al., 2017; Moulin et al.,  
567 2020), the fact that the passive margins remain in a high position, close to the sea-level, until  
568 the break-up and even much later (Moulin et al., 2005; Aslanian et al., 2009; Labails et al.,  
569 2009; Péron-Pinvidic & Manatschal, 2009; Bache et al., 2010) provide insights into crustal  
570 and mantle thermal segmentation and its role on subsidence. The replacement of the  
571 continental mantle by hot asthenosphere produces an important thermal anomaly that affects  
572 the subsidence of the transitional domain (Huisman and Beaumont, 2011, 2014). These  
573 thermal anomalies can be persistent in time due to edge-driven convection in the mantle and

574 bring episodically heat to the surface by magmatic intrusions (King and Anderson, 1998,  
575 Lucazeau et al., 2008). Analysis from heat flow in the transitional crust domains estimates  
576 that these processes can lead to an increased mantle heat flow input of 20-30 mW/m<sup>2</sup>  
577 (Goutorbe et al., 2008b) and a doubling of the surface heat flow (Lucazeau et al., 2010). This  
578 equally means higher temperatures at the top of basement crust that will be transferred to the  
579 salt deposits. Moreover, also regional hydrothermal circulations are known to occur in  
580 margins and can redistribute heat in specific margin segments (Lucazeau et al., 2010; Poort et  
581 al., 2020).

582 Although the direct link between salt morphology segmentation and the thermal regime  
583 remains speculative, we argue that the hypothesis of a relationship between temperature and  
584 salt deformation at regional level should not be overlooked. Only more in-depth work on the  
585 thermal regime in passive margins can help in quantifying the temperature variations needed  
586 to affect the salt flow.

587

## 588 **7. Other salt passive margins**

589 In the previous section we have described the coincidence between crustal segmentation and  
590 salt morphologies in Western Mediterranean Sea. Although we have listed some of the  
591 potential hypothesis that could explain this coincidence, the discussion is still open and a clear  
592 answer is still unclear. In the next section, we will explore others, well-known, passive  
593 margins in the world (Fig. 8) in order to understand if this correlation can be observed in other  
594 salt-bearing margins context. We will not go into detail about the discussion of salt tectonics  
595 in every margin as is not the objective of this paper. Nevertheless, we will try to summarize  
596 the key information and the analogies with the Western Mediterranean Sea, in order to discuss  
597 the salt morphologies-crustal segmentation relationship. We compared regional salt  
598 morphologies and crustal segmentation in the South-Atlantic margins (Angolan and Santos  
599 basins), Gulf of Mexico, Morocco and Nova Scotia margins.

600

### 601 **7.1 Salt morphologies**

602 The definition of salt morphology domains is not a simple task as each passive margin differs  
603 in main salt tectonics mechanisms and resulting structures. We have grouped the salt  
604 morphologies into 3 main domains, so that they can be regionally compared with the crustal  
605 domains constrain by refraction seismic data. We have referred the salt structures  
606 interpretation to the wide salt tectonics literature present in the different passive margins,  
607 using published profiles (References in chapter 3.1) and salt isopachs maps (e.g. Kukla et al.,

608 2018; Mohriak et al., 2008). The 3 domains are so-defined based to their main geometric  
609 characteristics and associated structures, as done in the Western Mediterranean Sea, in order  
610 to give a global and homogenous view, and are named: –Raft domain, –Diapirs domain and -  
611 Massive-Canopy domain. The Raft domain is so-called for the typical extensional thin-  
612 skinned structures as salt rollers, turtlebacks and base salt-rooted normal faults. The Diapirs  
613 domain includes salt stocks and walls characterized by discordant overburden, autochthonous  
614 connected or disconnected structures. To note that in Gulf of Mexico the Diapirs domain can  
615 exhibits coalescence sheets and allochthonous salt as well as basinward listric faults (Diegel et  
616 al., 1995; Fort and Brun, 2012). The Massive-Canopy domain is located in the downdip part  
617 of the salt basins and is characterized by thickening of salt deposits, thrust, folds and  
618 compressional diapirs. This domain can be characterized by massive thicken salt and/or  
619 canopy structures. The term canopy describes salt that is emplaced at stratigraphic levels  
620 above the autochthonous evaporites and comprises sub-horizontal or moderately dipping  
621 structures (Hudec and Jackson, 2006). The salt moved vertically from a single or multiple  
622 autochthonous feeder forming a salt sheet which, after coalescence of numerous structures,  
623 can form a canopy. The spatial distribution of the three domains just described shares  
624 characteristics with passive margins kinematically linked zones already described by different  
625 authors (e.g. Brun and Fort, 2011; Hudec and Jackson, 2004; Jackson et al., 1994; Peel, 2014;  
626 Vendeville, 2005; Rowan et al., 2012), characterized by and updip extension, a mid-slope  
627 translation and a downdip contraction domain. We have not used the terminology already  
628 proposed in the literature because we consider it interpretative with regard to the processes of  
629 salt tectonics involved. The Raft domain in the passive margins present similarities with the  
630 Listric salt domain interpreted in the Western Mediterranean Sea. The Diapirs domain has the  
631 main similarities with the Large Diapirs domain in the Algerian and Gulf of Lion deep basin  
632 (Figs. 6A and 10). Instead, the Canopy-Massive domain is not interpreted in Western  
633 Mediterranean salt basins, where the salt is mainly autochthonous.

634

## 635 **7.2 Passive margins crustal segmentation**

636 The crustal segmentation and nature of each margin is described in this small paragraph,  
637 listing the interpretation proposed in literature.. As in Western Mediterranean sea, the  
638 segmentations is constrain by wide-angle seismic data with the complementation of gravity  
639 and magnetic data.

640 In **SSPS (Santos Sao Paulo plateau System)** we have considered the SanBa Experiment, a  
641 joint project between Ifremer, IUEM, Lisbon's faculty of Science, the University of Brasilia

642 and Petrobras which collected six multichannel (MCS) and coincident wide-angle seismic  
643 profiles in the in the Brazilian margin: profiles SB01 and SB02 run in a NW-SE direction,  
644 separated from each other by less than a hundred kilometres (Fig. 9a); profiles SB03, SB04,  
645 SB06 and SB07 cross these two principal lines. The velocity models enabled the authors  
646 (Evain et al., 2015) to present a new segmentation composed of seven distinct domains (Figs.  
647 9), also coherent with the observations of gravity and magnetic data:

- 648 - Domain CC: unthinned continental crust domain with a 35-40 km thick crust and  
649 velocities between 5.6 and 7.0 km/s.
- 650 - Domain N: the necking domain, characterized by thinning, is wide less than 100 km.  
651 The Moho rises and the basement deepens.
- 652 - Domain A: thin crust with velocities between 6.0 and 7.0 km/s. Towards the deep  
653 basin (SE), the Moho and the top of the basement rise slowly. This domain  
654 corresponds to a low gravity anomaly and a high magnetic anomaly and is interpreted  
655 as exhumed lower continental crust (Evain et al., 2015).
- 656 - Domain B: very thin crust that shows the most heterogeneous crustal structure. The  
657 basement deepens by 1-2 km and the area is characterized by a large and mostly  
658 negative magnetic anomaly and a relatively high gravity anomaly. Large lateral  
659 variations in crustal velocities are observed between the two profiles. This domain is  
660 interpreted either as very thin exhumed lower continental crust intruded by  
661 magmatism or as atypical oceanic crust, overlying altered mantle (Evain et al., 2015).  
662 Whatever the nature of domain B in the central part of the SSPS, it is characterized by  
663 a Moho rise, an extremely thin crust (7 to 11 km thick from Evain et al., 2015) and a  
664 lower layer characterized by high velocity (7.0-7.8 km/s) (Fig. 9b).
- 665 - Domain C: 11 to 17 km thick crust with a pattern similar to Domain A except for the  
666 lower upper crust velocity (5.2 km/s).
- 667 - Domain D: 5 km thick atypical crust with a triangular shape characterized by a step in  
668 bathymetry, low gravity and magnetic anomalies. Klingelhoefer et al., (2014) describe  
669 this domain as characterized by anomaly velocities as proto-oceanic crust.
- 670 - Domain OC: Oceanic crust

671 In the **Angolan margin** we have considered the crustal nature interpreted after the ZaiAngo  
672 wide-angle project, led by Ifremer and TotalFinaElf. Seven MCS and refraction profiles were  
673 acquired between the Lower Congo and the Angolan basins (Fig. 10). Contrucci et al.,

674 (2004a) and Moulin et al., (2005) interpreted the Angolan margin formed by four domains  
675 (Fig. 10), also coherent with gravity and magnetic data:

- 676 - CC: corresponds to the unthinned continental crust, 30-35 km thick (based on gravity  
677 data)
- 678 - N: refers to the 50 km-wide necking zone, with an abrupt thinning of crust of about 20  
679 km.
- 680 - T1: refers to the landward region of transitional domain characterized by a thin crust  
681 of about 16-8 km thick, and an anomalous velocity zone below (7.2 to 7.5 km/s). This  
682 layer is interpreted partly in the necking zone and it reaches a maximum thickness  
683 when the basin displays the maximum depth.
- 684 - T2: refers to the basinward region of transitional domain, characterized by a slightly  
685 thinning crust and the absence of the high velocity lower layer seen in T1. The  
686 transitional domain is bounded to the west by a basement high. Crust thickness does  
687 not exceed 5 km.
- 688 - OC: the oceanic crust is 6-7 km thick.

689 The velocity model in the upper crustal layer of segment T2 does not seem to be coherent  
690 with an oceanic nature and, at the same time, shows velocity and thickness that appear to be  
691 different to those of a “typical” continental crust (Contrucci et al., 2004a). The anomalous  
692 velocity layer in domain T1 can be interpreted as lower continental crust because of its  
693 velocities (Holbrook et al., 1992) or as serpentinized mantle (Contrucci et al., 2004a). Based  
694 on kinematic studies, wide-angle and multichannel seismic analysis, Aslanian et al., (2009)  
695 propose a three-phase model of evolution where domain T2 is allochthonous and could  
696 represent the exhumation of the lower continental crust from the autochthonous T1 domain.

697 In **Gulf of Mexico**, the crustal segmentation interpreted as part of the GUMBO Marine  
698 Seismic Experiment consists of four wide-angle profiles in the northern GOM margin (Fig.  
699 11). The continental crust is between 36 km (Offshore Florida, Eddy et al., 2018) and 40 km  
700 (Huerta and Harry, 2012) thick and it gets thinner along about 200 km (Gumbo3, Eddy et al.,  
701 2014). In the transitional domain, the crust stretched to a thickness of less than 10 km  
702 (GUMBO1 offshore Texas, Van Avendonk et al, 2015; Fig. 7b), and 6-7 km offshore Florida  
703 (Christeson et al., 2014). In the north-western GOM, concomitant to GUMBO3 and  
704 GUMBO4 profiles, the authors interpreted high-velocity lower crust as syn-rift magmatic  
705 intrusion into the lower crust (Eddy et al., 2014) while offshore Texas, the low velocity is not  
706 consistent with magmatic underplating (Van Avendonk et al., 2015). Here, a 45 km-wide zone

707 of low-velocity and thin crust might be associated with exhumed mantle, locally  
708 serpentized. The oceanic crust limit (OCT) is still debated in literature (Fig. 11). The  
709 oceanic crust has a thickness ranging from 6 km (offshore Florida, GUMBO4) to 8 km. Many  
710 interpretations of the nature of the transitional crust domain in the north GOM are proposed:  
711 thinned continental crust (Roberts et al., 2005); oceanic and proto-oceanic crust (Imbert et al.,  
712 2005; Mickus et al., 2009); serpentized mantle or thin oceanic crust (Kneller and Johnson,  
713 2011); hyperextended continental crust with exhumed mantle in central and north-western  
714 sectors and proto-oceanic crust in the east (Rowan et al., 2012). Whatever the interpretation,  
715 the velocity models show a wide transitional domain characterized by a Moho rise (Fig. 11)  
716 (up to 16 km depth, GUMBO3) and crustal thinning (up to 7 km, GUMBO4), associated with  
717 high velocity in the lower layer (Fig. 11b).

718 In **Morocco-Nova Scotia** conjugated margins, we considered crustal segmentation using  
719 MIRROR (Biari et al., 2015), SISMAR (Contrucci et al., 2004b) and SMART (Funck et al.,  
720 2004; Wu et al., 2006) reflection-refraction experiments for, respectively, the Morocco (the  
721 former two) and Nova-Scotia margins (Fig. 12). The Moroccan margin segmentation is  
722 divided into three crustal domains (Fig. 12, from Biari et al., 2015): unthinned continental  
723 crust of 36 km thick, a thinning continental crust domain and an oceanic domain. The thinning  
724 domain is less than 90 km wide and, at the limit with the oceanic domain, the crust is  
725 approximately 13 km thick. The oceanic domain is characterized by an area with high lower  
726 crust velocities (atypical oceanic crust from Biari et al., 2015) and an oceanic crust thickness  
727 of 8 km in the most distal part. In the Nova Scotia margin (Fig. 12), Funck et al., (2004)  
728 (SMART1) and Wu et al., (2006) (SMART2) interpreted a 36 km-thick continental crust,  
729 which thins seawards to 5-6 km, mostly over the slope region. To the north (SMART1 and  
730 SMART2), the authors propose a wide area of partially serpentized upper mantle rocks,  
731 characterized by high lower-crustal velocities (up to 7.6 km/s). Further south (SMART3)  
732 Louden et al., (2013) interpreted a thicker oceanic crust (7 km) and an area of magmatic  
733 underplated bodies. The comparison between the two conjugated profiles (SMART1 and  
734 MIRROR1), following the cinematic reconstitution of Sahabi et al., (2004), shows structural  
735 and crustal similarities in the area of the necking zone and a thinning of continental crust.  
736 However, the presence of a transitional domain in the Canadian margin characterized by an  
737 exhumed serpentized mantle and a West African domain formed by intruded oceanic crust  
738 led the authors to favour the hypothesis of a symmetrical rifting followed by an asymmetrical  
739 oceanic accretion (Klingelhoefer et al., 2016; Biari et al., 2017), although there are other  
740 hypothesis, including an age difference between the two oceanic crusts imaged by refraction

741 profiles or a greater thickness of the oceanic crust on the African side due to the influence of  
742 the Canary hot spot (Biari, 2015).

743

## 744 **8. Geometrical correlation between salt morphologies and crustal segmentation in other** 745 **passive margins and potential explanations**

746 The spatial salt morphologies distribution variation (Figs. 9; 10; 11; 12) suggests a  
747 relationship with crustal segmentation of each margin.

748 In Santos margin, the salt base is evidently interpretable except in the Massive-Canopy  
749 domain, where seismic imaging is hampered by absorption and scattering of salt deposits. In  
750 the Large Diapir domain, the salt base shows a landward vergence (Fig. 9b; Davison, 2007;  
751 Guerra and Underhill, 2012; Strozyk et al., 2017). Moreover, it is largely unfaulted in the  
752 proximal to distal domains and underlain by similarly unfaulted sag sequence (Moulin et al.,  
753 2005; Karner and Gamboa, 2007; Unternher et al., 2010; Rowan, 2014), composed of syn-rift  
754 sediments (Kumar et al., 2013) of about 2 km thick (Evain et al., 2015). Salt deformation is  
755 entirely thin-skinned, without involvement or reactivation of pre-salt structures (Guerra and  
756 Underhill, 2012). The transition between the continental domain (Domain A and A') and  
757 Domain B coincides with a change in salt morphology, from a Diapir Salt Domain to a  
758 Canopy-Massive Domain (Fig. 9). In domain B the wide-angle profile shows an irregular top  
759 of the crust (Fig. 9; dashed black line) but do not seems to have a structural influence on the  
760 salt. The SSSP represents a buffer zone of the South Atlantic Ocean (Moulin et al., 2012) with  
761 its two conjugate margins and a failed ridge in its centre (Evain et al., 2015). Therefore, in the  
762 southern part of the basin, the salt onlaps the Sao Paulo Ridge (Fig. 9), interpreted as of  
763 continental nature (Evain et al., 2015).

764 In the Angolan margin (Fig. 10), the slope exhibits a Raft extensional domain, described by  
765 different authors (e.g. Brun and Fort, 2011; Marton et al., 2000; Duval and Cramez, 1992),  
766 basinward followed by the Diapir domain in the proximal sag basin, characterized by salt  
767 stocks, weld structures and wide pillows. In the Canopy-Massive domain, as in Santos basin,  
768 the salt is thick, stands above the sag basin and the base of salt is hampered. It can extends  
769 over the oceanic crust, forming nappe geometries (Fig. 10). Whatever the crustal nature of the  
770 different segments in the transitional domain, there is a strong coincidence between: 1) crustal  
771 domain T1 and the overlaying Large Diapir Domain, 2) crustal domain T2 and the Canopy-  
772 Massive Salt domain (Fig. 10b). In the Santos and Angolan margin, the limit between the



773 Diapirs domain to the Massive-Canopy domain corresponds to a shallower Moho depth (from  
774 ~20 km to ~15 km) and a thinning of crust (from ~12-15 to ~5-8 km).

775 In the Gulf of Mexico, the crustal segmentation uncertainty (Fig. 11) and the wide spatial  
776 distribution of salt structures make difficult the comparison between crustal nature and salt  
777 morphologies. Nevertheless, the canopy structures in the Sigsbee Escarpment lays on  
778 extremely thinned transitional crust, probably exhumed (Rowan, 2014, 2018), and on oceanic  
779 crust. The crust below the wide canopy domain thins (up to 8 km) and the Moho rises and  
780 could be characterized by mantle exhumation (Rowan, 2014).

781 The salt tectonics system of Morocco-Nova Scotia conjugated margins exhibits some  
782 differences with respect the Gulf of Mexico and South Atlantic margins. The Diapir salt  
783 domain is mainly centred on the Necking Zone domain, where the crust is of continental  
784 nature and thinned (Fig. 12; Biari et al., 2015) and the deformation is affected by the initial  
785 salt thickness and the underlying basement structures (Ings and Shimeld, 2006; Krézsek et al.,  
786 2007). The Massive-Canopy salt domain occupies a vast area only in the Nova Scotia margin  
787 and is characterized by salt derived from the Diapir Domain, which glided towards the deep  
788 basin after its deposition (Sahabi et al., 2004) (Fig. 12). In the Moroccan margin, the seaward  
789 movement is limited (Holik et al., 1991), except in the Tafelney Plateau (Fig. 12), where some  
790 salt canopies may have locally mobilized over a distance up to 20 km.

791 In contrast to the Western Mediterranean Sea, several mechanisms may be at the origin of this  
792 concomitance. The salt passive margin segmentation caused by interaction between gliding-  
793 spreading mechanisms (e.g. Ge et al., 1997; Vendeville, 2005; Hudec and Jackson, 2007;  
794 Gaullier et al., 2006; Brun and Fort, 2011; Rowan et al., 2012) likely explain the downdip  
795 succession of salt morphologies domains but does not fully demonstrate their spatially  
796 coincidence with the crustal segmentation. The present-day salt structures distribution could  
797 be related to the salt deposition timing with respect the rifting phases (Jackson and  
798 Vendeville, 1994; Tari et al., 2003; Rowan, 2014, 2018; Jackson and Hudec, 2017; Hudec and  
799 Peel, 2019). The crustal structure controls the topography at the time of salt deposition and  
800 thus exerts a primary control on the initial salt thickness. The salt thickness can govern the  
801 evolution of the salt morphologies, which derived, among other causes, by the salt supply  
802 (e.g. Hudec and Jackson, 2007). A good examples are the Morocco-Nova Scotia conjugated  
803 margins, considered as a syn-rift (Tari et al., 2003, 2013) or syn-stretching (Rowan, 2014) salt  
804 basins. Here, the salt was deposited during the last stage of rifting, prior to the break-up, in the  
805 Upper Triassic/Lower Jurassic (Hafid, 2000; Tari et al., 2003). The salt structures are thus

806 attributed to the underlying basement structure, which determined the initial configuration of  
807 the basins and influenced the salt initial thickness (Ings and Shimeld, 2006; Krézsek et al.,  
808 2007). In Gulf of Mexico and South Atlantic salt basins, several authors claimed for  
809 significant variations in initial salt thickness (Gulf of Mexico: Rowan, 2014; 2018; Hudec et  
810 al., 2013a) while others (Angolan margin: Moulin et al., 2005) for a roughly constant  
811 thickness and no post-deposition horizontal movements. As a matter of fact, the salt  
812 deposition context is still debated as well as its impact on present-day salt structures.

813

## 814 **9. Conclusion**

815 We have showed that due to its peculiar characteristics, in the Western Mediterranean the  
816 classical salt tectonics models fail to explain the regional differences in salt morphologies  
817 distribution, the early deformation in the deep basin and the spatial coincidence with the  
818 crustal segmentation.

819 A different temperature distribution at the base of salt related to fluids and/or heat  
820 segmentation resulting from different crustal natures could play a role, affecting the salt  
821 rheology and thus adding a further component on the well-known mechanisms.

822 The same coincidence is observed in some world salt-bearing passive margins. Although the  
823 discussion on the salt tectonics main mechanisms in the passive margins is still open, the heat  
824 segmentation hypothesis should not be overlooked and deserves further investigation,  
825 especially in view of what has been observed in Western Mediterranean Sea.

826 Further heat flow analysis are needed to correlate the salt deformation timing with the thermal  
827 regime in passive margin. Moreover, analogical and/or numerical modelling might be useful  
828 to isolate the temperature parameter impact and understand how the salt may changes the  
829 morphology and timing with different temperature regime at its base.

830

831

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843 L’Atalante 2018, <https://doi.org/10.17600/18000402>).

844

#### 845 **Contributions**

846 The study was conceived by DA. MB has re-interpreted and compiled the different salt  
847 morphologies with the help of RP, DA, MM, MR, EL, CG, ADB. The compilation and  
848 analysis of the published wide-angle data were done by MB, MM, DA. The thermal data were  
849 provided by JP. AC provided new seismic data. All authors provided critical feedback and  
850 helped shape the research, analysis and manuscript.

851

#### 852 **Figure Captions**

853

854 **Figure 1:** A) Location of the study area marked by the black outlined rectangle. Note the  
855 structural threshold separating the Western Mediterranean Sea to the Atlantic Ocean and the  
856 Tyrrhenian basin, making it a present-day closed system. B) Dataset and bathymetry of the  
857 Western Mediterranean Sea. Bathymetric contours every 200 m. MCS and wide-angle seismic  
858 lines are represented. Red circles represent the OBS and the land stations deployed for the  
859 wide-angle acquisition. *References* profiles are from work of Leprêtre et al., (2013); Badji,  
860 (2014); Arab et al., (2016). Labels: Gulf of Lion (GoL), LPB (Liguro-Provençal basin), WS  
861 (west Sardinian margin), MB (Minorca basin), VB (Valencia basin), EAB (east Algerian  
862 basin), HA (Hannibal area), WAB (western Algerian basin), EA (east Alboran margin), SMB  
863 (south Minorca Block), SB (south Balearic margin), BP (Balearic promontory).

864

865 **Figure 2:** (A) Zoom on high-resolution seismic profile of Gulf of Lion lower slope (location  
866 Fig. 1) showing salt rollers and rollover structures from the Early Pliocene deposits. (B) Zoom  
867 on low-resolution seismic profile (Fig. 3; location Fig. 2) showing early salt deformation syn-  
868 deposition of Upper Messinian Unit (green). The Lower Unit limit is from Lofi et al., (2011).

869

870 **Figure 3:** From top to bottom: NW-SE Sardinia 01 time-migrated seismic profile,  
871 uninterpreted and interpreted (location in Fig. 1). At the bottom, Sardinia 01 wide-angle

872 refraction profile showing crust segmentation along Gulf of Lion margin (modified from  
873 Moulin et al., 2015). Dashed black lines represent the lower and upper limit of the crust.

874

875 **Figure 4:** Salt morphologies nomenclature and description used in the Western Mediterranean  
876 Sea. Seismic zooms are all presented at the same scale to enable comparison. Vertical  
877 exaggeration 1:4.

878

879 **Figure 5:** Map showing the distribution of salt morphologies and crust segmentation in  
880 Western Mediterranean Sea. The contour lines represent the base of MUM unit (Gorini et al.,  
881 2015) in the deep basin and the BES (Basal erosional surface) and MES (Messinian erosional  
882 surface) (Lofi et al., 2011; Gorini et al., 2015) in the slope and shelf. In the deep basin, within  
883 the red dashed lines, the base of MUM unit represents the base of salt. Contour every 500 m.  
884 The categories associated with the different colours are described in Fig. 4. Solid black lines  
885 correspond to the crustal segmentation from Moulin et al., (2015) (Gulf of Lion), Afhilado et  
886 al., (2015) (West Sardinia) and Leprêtre et al., 2013; Medaouri et al., 2014; Mihoubi et al.,  
887 2014; Badji et al., 2015; Bouyahiaoui et al., 2015; Aïdi et al., 2018 (north Algerian margin).  
888 Dashed lines represent segmentation interpretation from gravity and magnetic data (Leroux et  
889 al., 2019). Transfer zones are from Rehault et al., (1985) and Pellen et al., (2016). Dashed red  
890 line represents salt upslope limit before this work, taken from Bache et al. (2009), Obone Zue  
891 Obame et al., (2011), Bellucci, (2017), Driussi et al., (2015).

892

893 **Figure 6:** Regional line drawings of seismic sections from the Gulf of Lion to the West  
894 Algerian margin (location in Fig. 5). Are represented the main salt morphologies domains in  
895 Western Mediterranean Sea. The crustal segmentation is from literature (references are shown  
896 below the segments). Position of the Transfer Zone is from Pellen et al., (2016). Note the  
897 horizontally pre-salt and base salt reflectors and the salt morphology differences between east  
898 (B-C profiles) and west (C-D-E) Algerian sub-basins. Vertical exaggeration 1:6.

899

900 **Figure 7:** Additional regional line drawings in Western Sardinian and West Algerian margins  
901 (location in Fig. 5), (same colour-code as previous Figures). Vertical exaggeration 1:6.

902

903 **Figure 8:** Regional bathymetric map showing location of the world passive margins study  
904 areas.

905

906 **Figure 9:** (A) Map showing the salt morphologies and crust segmentation in the Santos and  
907 Campos basins. On the left it is shown only the crustal segmentation while on the right also  
908 the salt morphologies domains. Blue area represents the Raft salt domain distribution; pink  
909 area the Large and Narrow Diapirs domains; yellow area the Canopy-Massive salt domain.  
910 Wide-angle profile location and crust segmentation are from Evain et al., (2015). Volcanic  
911 structures are from Mohriak, (2001). The hinge line and the Capricornio and Cruzeiro do Sul  
912 Lineaments are from Moulin et al., (2012). Black lines represent seismic profile interpreted  
913 (references in the text). (B) From top to bottom, Sanba01 time-migrated reflection seismic  
914 line uninterpreted and interpreted (location in Fig. 9a). The interpreted salt morphologies are  
915 represented, showing their evolution from the lower slope to the deep basin. Sanba01 wide-  
916 angle profile, coincident with the reflection seismic line showing crust segmentation along  
917 SSPS (from Evain et al., 2015). Dashed black lines represent the lower and upper limit of the  
918 crust.

919

920 **Figure 10:** (A) Map showing the salt morphologies and crust segmentation in Angolan  
921 margin. On the left it is shown only the crustal segmentation while on the right we have  
922 superposed the salt morphologies. Blue area represents the Raft salt domain distribution; pink  
923 area the Large Diapir domain; yellow area the Canopy-Massive salt domain. Wide-angle  
924 profile location and crust segmentation are from Contrucci et al. (2004) and Moulin et al.,  
925 (2005). Black lines represent seismic profile interpreted in this study (references in the text).  
926 (B) From top to bottom, time migrated reflection profile and line-drawing interpretation  
927 (modified after Moulin et al., 2005) showing salt morphologies domains. Z-03 wide-angle  
928 coincident profile showing crust segmentation by Contrucci et al., (2004). Location in Fig.  
929 10a. Dashed black lines represent the lower and upper limit of the crust.

930

931 **Figure 11:** (A) Map showing crustal segmentation and salt morphologies in Gulf of Mexico.  
932 On the left it is shown only the crustal segmentation while on the right the superposed salt  
933 morphologies. Outlined in red the salt structures from Diegel et al., (1995). Salt morphologies  
934 boundary are redrawn from Seni, (1992). The oceanic domain boundaries from Hudec et al.,  
935 2013a; Sawyer et al., 1991, Christeson et al., 2014 and Curry et al., 2018 are represented.  
936 Black lines represent Gumbo wide-angle survey profiles (references in the text). Fault trends  
937 are from Diegel et al., (1995) and Stephens, (2010). Black lines represent seismic profile,  
938 interpreted in this study (references in the text). (B) PSDM ION GulfSPAN 2450 profile from  
939 Van Avendonk et al., (2015) and Dickinson, (2017) approximately in the location of GUMBO

940 1 profile (location in Fig. 11a). The salt morphologies are from this study. Below, the  
941 GUMBO1 wide-angle profile. Crustal segmentation in black is from this study and discussion  
942 in Van Avendonk et al., (2015). Dashed black lines represent the lower and upper limit of the  
943 crust.

944

945 **Figure 12:** Map showing salt morphologies and crustal segmentation in Nova Scotia and  
946 Morocco conjugated margins. Crust segmentation is taken from SMART, SISMAR and  
947 MIRROR refraction studies (references in the text). Salt structures are from Shimeld, (2004);  
948 Sahabi, (2004) and Tari et al. (2000). Black lines represent the seismic profile, interpreted in  
949 this study (references in the text). Canopy (allochthonous) salt in Nova Scotia is from Sahabi,  
950 (2004) interpretation. (B) Mirror01 reflection and wide-angle associated profiles (location in  
951 Fig. 12a). Modified after Biari et al., (2015). Dashed black lines represent the lower and upper  
952 limit of the crust.

953

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