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Salt morphologies and crustal segmentation relationship: new insights from Western Mediterranean Sea and other salt passive margins

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Abstract

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

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

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

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

1. Introduction

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

gliding of the brittle-ductile series during gravity spreading (Obone Zue Obame et al., 2011), (ii) initial conditions and morpho-structure evolution of the basin (Gaullier et al., 2008), (iii) local geodynamic setting that could induce thick-skinned tectonics (Maillard et al., 2003; Déverchère et al., 2005). The application of the two salt tectonic end-member models highlights problems when we take into account the differences in salt morphologies in the Western Mediterranean Sea. In the Western Mediterranean Sea, salt was deposited in a closed basin, already formed, above a thick marine sequence deposited during the Late Oligocene and Miocene intervals. Moreover, the purely vertical subsidence in the deep basin implies small difference in initial salt thickness that could have controlled salt supply variations.

We present here a compilation of the relationships between salt morphologies and crustal nature in well-studied salt-bearing basins throughout the world. We first focus on the Western Mediterranean Sea, for which we present a new detailed map of salt morphologies. We categorized regional salt morphologies in order to superimpose them onto the crustal segmentation, constrained by wide-angle seismic studies, magnetic and gravity maps. We then discussed the possible causes for this relationship, taking into account the Western Mediterranean Sea peculiarities. We then applied the same method to South-Atlantic margins (Angolan and Santos Basins), Gulf of Mexico, Morocco and Nova Scotia margins in order to observe if this coincidence is also confirmed in different salt tectonics systems.

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

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

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

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

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

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

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

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

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

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

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

3. Dataset and method

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

3.1 Salt morphologies interpretation

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

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

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

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

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

3.2 Crust segmentation

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

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

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

4. Western Mediterranean salt deformation

4.1 Salt tectonics mechanisms

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

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

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

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

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

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

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

4.2. Salt morphologies description and distribution

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

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

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

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

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

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

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

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

5. Crustal segmentation and nature

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

domain, which follows the onshore unthinned continental crust domain, is the "crustal thinning" domain, area characterized by the main thinning of the crust, mainly focused on the lower crust where, in the Gulf of Lion, thickness decreases from 15 to 7 km (Fig. 3). Basinward, the transitional domain is characterized by a relatively high crustal velocity, 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 lower layer, in the Gulf of Lion and West Sardinian, respectively. The lower layer is interpreted as exhumed lower continental crust (Moulin et al. 2015; Afilhado et al., 2015) or exhumed lower crust and upper mantle (Jolivet et al., 2015). The central domain is characterized by a decrease in velocity and velocity gradient in the lower crust (Moulin et al., 2015) and presents an "atypical" relatively thin (5 km) oceanic crust (e.g. Le Douaran et al., 1984; Gailler et al., 2009; Moulin et al., 2015). The oceanic crust is thinner than the Atlantic type but the velocity and gradients confirm the oceanic nature (Gailler et al., 2009) and its close relationship with the transitional domain, regarding the velocities and the Moho discontinuity depth (Moulin et al., 2015; Afilhado et al., 2015). The segmentation is confirmed by gravimetric (Sandwell and Smith, 1997) and aeromagnetic (Galdeano and Rossignol, 1977) data (Leroux et al., 2019).

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

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

6. Discussion

6.1 Geometrical correlation between salt morphologies and crustal segmentation

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

6.2 Salt morphologies segmentation hypotheses

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

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

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

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

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

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

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

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

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

6.3 The potential role of thermicity on salt deformation

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

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

bring episodically heat to the surface by magmatic intrusions (King and Anderson, 1998, Lucazeau et al., 2008). Analysis from heat flow in the transitional crust domains estimates that these processes can lead to an increased mantle heat flow input of 20-30 mW/m² (Goutorbe et al., 2008b) and a doubling of the surface heat flow (Lucazeau et al., 2010). This equally means higher temperatures at the top of basement crust that will be transferred to the salt deposits. Moreover, also regional hydrothermal circulations are known to occur in margins and can redistribute heat in specific margin segments (Lucazeau et al., 2010; Poort et al., 2020).

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

7. Other salt passive margins

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

7.1 Salt morphologies

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

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

7.2 Passive margins crustal segmentation

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

9. Conclusion

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

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

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

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

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Contributions

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

Figure Captions

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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