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1 **Interconnection salt diapir–allochthonous salt sheet in Northern Tunisia: the Lansarine–**
2 **Baoula case study**

3

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11

12 **Abstract**

13 Surface and subsurface data are used to illustrate the halokinetic style, structural evolution and
14 kinematics of the Lansarine–Baouala salt structure in the Alpine domain of Tunisia. It corresponds to a
15 salt structure showing a large salt canopy overhanging buried diapir. The salt structure is located along
16 the main fault systems, inherited from the southern Tethyan passive margin periods. The Northern
17 domain of Tunisia is undergoing diapirism, which has been recognized as active during the Upper Aptian
18 and Lower Albian periods. Above the diapir apex, salt flows outward and spreads by gravity downslope.
19 A minibasin has formed during this period of salt outward evacuation. The minibasin has a thickness
20 practically correlated to the budget of salt evacuation toward the diapiric structure. During Cenozoic
21 times, compressional deformation stages do not reflect active diapirism. The Lansarine–Baouala salt
22 structure, consisting of connected salt sheets, was passively transported on the Southeast-verging major
23 thrusts that are deeply rooted in the salt beneath the Mateur basin. The Northern Atlas of Tunisia
24 reveals, as many others salt provinces, an interconnection of salt diapirs and significant laterally
25 spreading allochthonous salt sheets (canopies?). In this domain, the inherited allochthonous Triassic salt
26 sheets interlayered within recent series and connected to buried diapir have a significant role in the
27 shallow structural complexity observed in the sedimentary cover during the subsequent contractional
28 events.

29 **Keys-words:** North Tunisian Atlas; diapir; allochthonous salt sheets; mini-basin; thrust

30 **1. Introduction**

31 Rifts and continental margins are usually characterized by salt deposition followed by salt tectonics,
32 which strongly influence their structural styles (e.g. Talbot et al., 1991; Weijermars et al., 1993; Jackson
33 and Vendeville, 1994; Hudec and Jackson, 2007; Callot et al., 2012). Salt tectonics is currently recognized
34 to have a crucial role in the deformation of several geologic systems around the world in both active and
35 passive margins (e.g. Rowan et al., 2003; Hudec and Jackson, 2007; Alsop et al., 2015; Moragas et al.,
36 2017). Rift and passive continental margins tectonic style and/or evolution were the subject of new
37 tectonic concepts developed during the last 40 years (e.g. Bally, 1983; Allen and Allen, 1990; Brune et
38 al., 2014). These geologic systems are commonly described as dominated by intra-basin growth listric
39 normal faults in control of an asymmetrical seafloor that is characterized by repetitive depocenters. The
40 intra-basin faults system is responsible for the basin's geometry that is characterized by hanging-wall
41 rollovers, rafts, salt-related features and important thickness and facies variations of sedimentary
42 sequences. These systems are frequently inverted and deformed by subsequent compressions hence
43 the area is subject to tectonic inversion. The presence of salt is usually involved in the deformation of
44 several inverted systems such as the Alpine orogens of Europe, the Betics, the Pyrenees, the North
45 African alpine belts, Zagros (e.g. Canerot et al., 2005; Masrouhi and Koyi., 2012; Callot et al., 2014; Saura
46 et al., 2015). The early interpretations of salt tectonics that predominated during the first half of the
47 20th century were also updated by the upsurge of the thrust tectonics models. This evolutionary trend
48 has modernized salt tectonic concepts. Recent improvement in drilling and seismic technologies led to a
49 surge of large allochthonous salt on deep water passive margin. Although salt structures are being
50 recognized as having large overhangs (e.g. Jackson and Talbot, 1989; Talbot et al., 1991; Talbot, 1993),
51 several hundreds of salt overhangs are known to exist now (e.g. Wu et al., 1990; Talbot, 1993; Fort and
52 Brun, 2004; Hudec and Jackson, 2004, 2006, 2009) and are confined to only a few of the many
53 sedimentary basins containing salt (Gulf Coast, Campos basin, Angola margin, Red Sea). Past salt
54 overhangs are neglected in literature because they are usually comprised within orogens.

55 After a while, the interaction between salt diapir and salt sheets has been studied by several works in a
56 large proportion of salt provinces: Australia (e.g. Hearon et al., 2015; Rowan et al., 2019), Iran (e.g.
57 Heydarzadeh et al., 2020), Turkey (e.g. Kergaravat et al., 2016), Central Europe (e.g. Warsitzka et al.,
58 2019), the Algarve Portuguese basin (e.g. Ramos et al., 2017), Mexico (e.g. Giles and Rowan, 2012), USA
59 (e.g. Gradmann and Beaumont, 2017) and among others. Processes of mobilization of salt usually assists
60 gravitationally driven deformation of rock mass associated with this growth. In passive margin setting,

61 during sedimentation a wide range of gravity driven structures in successive depositional events are
62 generated along the slope. Salt is itself associated to this evolution in which the overburden moves with
63 gravity spreading driven by the differential load of tectonic blocks (Alsop and Marco, 2014). The half-
64 graben, which usually characterizes this growth, develops salt diapir in case of sufficient buried
65 underlying salt budget. When the salt reaches the sea floor, the interaction of several parameters
66 guides the evolution of the structure such as the rate of salt evacuation versus the sedimentation rate
67 (Fernandez et al., 2020), the impact of base-salt relief on salt flow (Dooly et al., 2018), Extension versus
68 sedimentation rate (Allen et al., 2020)...

69 The Maghreb exposes numerous salt structures and is described as one of the largest worldwide salt
70 provinces. Salt structures are more abundant in the eastern part of this province which corresponds to
71 the Northern Tunisian Atlas. This 10000 km² area exposes more than 40 structures, most of them are
72 hundred square kilometers structures (Fig. 1). Salt structures in this region have been the subject of
73 controversial geologic interpretations (e.g. Perthuisot et al., 1988; Vila et al., 1994; Masrouhi et al.,
74 2014a). In this province, two different types have been proposed to explain the salt growth and
75 configuration: (1) the salt pierces the overburden and the sequences' thickness is reduced in the limbs
76 of the structure which define the diapir or dome model (Bolze 1950; Perthuisot 1978; Perthuisot et al.
77 1998; Jallouli et al. 2005; Benassi et al. 2006). This model corresponds to the means of 1970s ideas
78 based on buoyancy -driven convection. Halokinetic movement was attributed to autonomous, isotactic
79 rise of salt and vertical salt movement versus its overburden due to Rayleigh-Taylor instabilities. (2)
80 Allochthonous salt interpretation (Ghanmi et al. 2001; Vila et al., 2002; Slama et al., 2009; Masrouhi and
81 Koyi, 2012; Zouaghi et al., 2013) which suggests a setting with salt flowing at the sediment-water
82 interface through a feeder frequently acknowledged as the normal faulting affecting the sedimentary
83 cover over the Triassic salt. The normal synsedimentary faults, formed during the thin-skinned
84 Cretaceous extension of the southern Tethyan passive margin, create a space for salt to reach the
85 surface during their motion and then flows on the surface, thus defining a salt glacier structures.
86 Although the interconnection between salt domes-salt glaciers is well documented in many passive
87 margins, the interconnection between those two types of structures has never been identified in the
88 Northeastern Maghreb salt province.

89 This study presents such an interconnection for the first time in this province and demonstrates a
90 reappraisal of the structure and the diapirism growth style with significant allochthonous salt sheet
91 (canopy?) spreading laterally from the diapir stem during the lower Cretaceous period. We present new

92 surface and subsurface data for both diapir growth and the salt sheet-related structures, which are now
93 partially obliterated by subsequent Tertiary compressions that need to be considered in the study of salt
94 tectonics in the Northern African area.

95 **2. Geological setting**

96 **2.1. Structural framework**

97 The Northern Tunisia Atlas is the eastern segment of the Maghreb folds-thrust belt known as the
98 Maghrebides chain (Fig. 1). The Maghrebides is an Alpine belt; occupying the Northwestern African
99 plate, which is related to the Africa–Eurasia convergence during Tertiary–present-day period. After a
100 long Mesozoic extensional Tethyan period, the positive inversion of the basins in Tunisia is recognized to
101 operate during Late Cretaceous (Guiraud and Bosworth, 1997; Masrouhi et al., 2008). During the
102 Cenozoic compressional history, the first significant contractional event is dated as Middle-Late Eocene
103 and accepted as the Atlassic compression (e.g. Vila et al., 1975; Frizon de Lamotte et al., 2000). The
104 Atlassic event is well traced by regional unconformity on the entire Tunisian Atlas (Masrouhi et al., 2007;
105 Gharbi et al., 2015). During the Oligocene–Lower Miocene period, the Atlas domain shows a post
106 compressional tectonic quiescence stage. The Oligocene–Miocene sequences are deposited in
107 distinguished sub-basins such as the Numidian Basin and the Medjedra basin. The tectonic setting seems
108 to be well correlated to the West-Mediterranean geodynamic setting (Corsica–Sardinia and the
109 Algerian–Provencal Miocene back-arc basin opening (e.g. Vergés and Sàbat, 1999; Frizon de Lamotte et
110 al., 2009; Roure et al., 2012; Van Hinsbergen et al., 2014). The Oligocene–lower Miocene extension is
111 caused at the scale of the margin by the retreat of the slab (e.g. Jolivet and Faccenna, 2000; Deverchere
112 et al., 2005). In the Tunisian Northern Atlas, the paroxysmal compressional event is considered to have
113 occurred during Middle Miocene to present-day times by several authors (e.g. Dlala, 1992; Chihi, 1992;
114 El Ghali et al., 2003; Gharbi et al., 2014). All authors who worked in this region agree that this event is
115 the major compressional tectonic event deforming the Atlassic domain of Tunisia (e.g. Gharbi et al.,
116 2014; Bahrouni et al., 2014; Soumaya et al., 2015; Masrouhi et al., 2019). The main regional
117 compression direction was identified by several studies as NW-trending during Miocene-Pliocene and N-
118 Trending during the Late Pliocene-Quaternary (e.g. Dlala and Hfaiedh, 1993; Billi et al., 2011; Gharbi et
119 al., 2014; Soumaya et al., 2015). The northern Atlas of Tunisia shows major NE-trending thrust systems,
120 with three most important traced fronts from North to South being (Fig. 1): the tell thrust range front,
121 the Teboursouk Thrust Unit Front (TTUF) and the Zoughan-Ressas Thrust Unit Front (ZRTUF). The
122 structural style of this belt has recently been identified as the result of the inversion of Mesozoic

123 inherited structures of the earlier passive margin basins and/or sub-basins along the south Tethyan
124 margin (e.g. Jaillard et al., 2017; Naji et al., 2018a; Masrouhi et al., 2019).

125 The inherited system is well linked to the Jurassic and Cretaceous periods with the southern margin of
126 the Tethys (Boughdiri et al., 2007; Frizon de Lamotte et al., 2013). The long Tethyan evolution generated
127 major basins from Morocco to Tunisia during the Late Permian–Late Cretaceous (Roure et al., 2012;
128 Moragas et al., 2016, 2017; Naji et al., 2018a). The long Mesozoic evolution began with Triassic basins
129 from an epicontinental environment and characterized by evaporitic facies covering the entire Maghreb
130 sub-basins (e.g. Soto et al., 2017). The Triassic sub-basins frequently display subordinate basalts
131 preserved in several localities in Tunisia (Laaridhi-Ouazâa, 1994) that are interpreted as the result of rift-
132 related magmatism (e.g. Mattoussi Kort et al., 2009). During the Jurassic period, a well-documented
133 main N-trending regional extension was associated to the development of main basins (e.g. Boughdiri et
134 al., 2007; Saura et al., 2014; Martin-Martin, 2017; Soussi et al., 2017; Moragas et al., 2018). These basins
135 are subject to significant thickness and facies variations (Peybernes et al., 1994; Boughdiri et al., 2007).
136 During Early Cretaceous a NE-SW regional tectonic extensional setting became dominant (e.g. Jaillard et
137 al., 2017; Naji et al., 2018b). This evolution from E-W (present-coordinates) to NW-SE shape is well-
138 matched to E-W transform faults of the Atlantic that change to NW-SE in the Alpine Tethys (e.g.
139 Rosenbaum et al., 2002; Masrouhi et al., 2019). During the lower Cretaceous, a paleogeographic
140 differentiation was established with a continental platform in the south and center and a deep domain
141 in the north. This differentiation is associated with volcanic and salt structures, all of them characterized
142 by a passive margin context (e.g. Mattoussi Kort et al., 2009; Masrouhi et al., 2013; Jaillard et al., 2017;
143 Dhahri and Boukadi, 2017).

144 Halokinetic signature in the Maghreb domain has long been recognized and mostly interpreted as
145 Cretaceous–related structures (Bolze, 1954; Perthuisot, 1988; Vila et al., 1994; Masrouhi et al., 2014b).
146 This belt corresponds to one of the largest worldwide salt provinces in which a debate has been active
147 for decades between two main interpretations i.e. “vertical” diapir structures [autochthonous
148 emplacement] and salt sheets [allochthonous mobile evaporites emplaced at younger stratigraphic
149 levels]. This paper presents new data of the LBSS to highlight a new model of relationship between the
150 allochthonous salt canopy and the autochthonous diapir feeder, in which the relationship is partially
151 obliterated by Tertiary compressions.

152 **2.2. Lithostratigraphic successions**

153 Mesozoic and Cenozoic series outcrop along the LBSS region and comprise sequences from Triassic to
154 quaternary deposits.

155 The mainly evaporitic Triassic series shows a thickness that varies from 150 to 1000 m. The Jurassic
156 series consist of classic limestones Oust, Zaghouan, Staa, Zaress and Beni klab Formations (Peybernes et
157 al., 1996; Boughdiri et al., 2007). These series are reduced in the southern part of the LBSS compared to
158 their southern equivalent outcropping in the Tunisian Dorsale (Soussi et al., 1999). They vary from 40 to
159 450 m in outcropping series near the LBSS and include stratigraphic gaps linked to the tilted block
160 geometry characterizing this margin during Mesozoic times (Naji et al., 2018b). In the northern domain,
161 the Jurassic reach a 1000m thickness in the Mateur – Ichkeul area (Alouani et al., 2006). In this area, the
162 Jurassic sequences usually indicate the presence of siliceous radiolarites revealing an important oceanic
163 depth (Cordey et al, 2005).

164 The Tithonian-Berriasian displays a facies made of debris-flow and mud-flow sandstones and siltstones
165 sequences. The Valanginian–Hauterivian is made of siliciclastic carbonate usually covered by mixed shelf
166 deposits that are dated as Hauterivian and defined as Seroula Formation (Peybernes et al., 1996). The
167 Barremian sequences are described in this region to contain a lower tempestites unit followed by a thick
168 carbonate unit and topped by storm deposits and black-shales parasequences (Khessibi, 1967, Souquet
169 et al., 1997). In northern Tunisia, the Valanginian–Aptian series are defined as M'Cherga Formation
170 (Buroillet, 1956; Ben Ferjani et al., 1990). The Aptian–Early Albian sequences are typically of deep-water
171 environment made of gray to black marls. They also habitually contain abundant slumps and nodules
172 features usually interpreted as related deep and irregular sea floor. This period (Aptian–Albian are the
173 two longest epochs of the lower Cretaceous) is interpreted as displaying the maximum extensional rate
174 of the Tethyan basin in northern Tunisia (Masrouhi et al., 2014b, Jaillard et al., 2017; Naji et al., 2018a).
175 Moreover, more than 80% of the salt structures appeared to be active during Aptian–Albian (e.g.
176 Perthuisot, 1988; Vila et al., 1994; Ghanmi et al., 2001; Masrouhi et al., 2014a).

177 After the lower Cretaceous deposition variation, the northern Tunisian domain shows a late Cretaceous
178 homogenous facies cover the entire basin with no significant facies variation. The late Cretaceous basin
179 is mainly consisting of pelagic to hemipelagic facies. The basin reflects a transgression that invaded and
180 sealed the aforementioned lower Cretaceous configuration. This deposition of the top of former
181 dislocation is attested by significant thickness variation along the entire basin. The thicknesses vary
182 considerably from some meters to hundreds of meters for the same interval from an area to another.

183 This period is characterized by homogenous facies with shales, marls and limestones characteristic of
184 post rift sedimentation (Fig. 2). The post rift sedimentation started in late Cretaceous times and
185 smoothly sealed tilted blocks geometries, shaped and accentuated during Early Cretaceous by major
186 synsedimentary normal fault systems. This second cycle shows limestone and marl alternations at the
187 base dated as Albian–Cenomanian and defined as Fahdene Formation. These sequences are topped by
188 organic-rich limestones thinly laminated defined as Bahloul member. On the Fahdene Formation, the
189 Turonian–Early Campanian Aleg Formation is formed by marl and shale with fine-grained limestones
190 (Burolet, 1956). Upper Cretaceous sequences, defined as Abiod Formation, display Campanian–
191 Maastrichtian chalky limestones deposits covering the entire basin of Tunisia (Burolet, 1956).

192 The Tertiary consists of the following sequences: marine black shale series dated as Paleocene and
193 defined as the El Haria Formation. El Haria Formation is frequently overlain by the lower Eocene
194 Formation made of limestones. Then, middle to upper Eocene thick series of marls and shales
195 corresponding to the Souar Formation. Due to the tectonic history (large thrust systems), the upper
196 Eocene does not outcrop in the Lansarine region. Oligocene sequences involve deltaic siliceous
197 sandstones. The Neogene sediments are made of thick marine Aquitanian–Burdigalian formations. The
198 upper Miocene–Pliocene sequences are continental strata recognized as the result of the large erosion
199 during Cenozoic uplifting and cover unconformably the Neogene and Cretaceous Formations.

200 **3. Field data and results**

201 The structural analysis based on a study of detailed geologic maps, controlled by field cross-sections.
202 Interpretations are based on the detailed geological map (1:25 000 scale) from existing data and data
203 prepared for this study. Work consists also of the use of the existing detailed micropalaeontological data
204 from the formations, in order to date the lithologies and illustrate the position of the Triassic salt
205 relative to the surrounding Mesozoic and Cenozoic sequences. the characteristics and behavior of the
206 contact between salt and other strata is detailed to draw back the nature of salt (vertical diapir and/or
207 horizontal salt sheet). Fault kinematics analysis is conducted in some areas, to quantify the brittle
208 deformation and the tectonic regime contemporaneous to the salt body evolution.

209 The ~ 360 km² LBSS corresponds to the second largest salt body in the north Tunisian salt province. In
210 order to highlight the significant variations of the salt geometry in different areas, we will first discuss
211 the salt contacts in key-zones and then present the structural style by segment.

212 In the northern part of the LBSS, all the southeast part of the Triassic salt body is in contact with
213 Cretaceous strata. The Triassic salt overlies Albian sequences in the south (Wadi Maiou). In this region,
214 lower Albian dips 45° to the NW under the Triassic units that are parallel to the Cretaceous layers and do
215 not show an extrusive character (Fig. 4). The parallelism between salt contact and Albian layers was the
216 subject of the most curious interpretations by former study of Bolze (1954) which attribute an Albian
217 age to the Triassic salt. This parallel contact can be mapped along more than 2 km from Wadi Maiou to
218 Wadi Tell and, regarding the ravinant character, was interpreted as a “conformable contact” between
219 Triassic salt and Albian Sequences due to an extrusive flow of salt (Masrouhi et al., 2013). This
220 terminology was used to qualify the bottom or the top contact of the salt sheet that flowed over the sea
221 water-sediment interface or was covered by later deposition. This contact is genetically different from a
222 sharp vertical diapiric contact. In the northwestern part, salt is joined in an anticline structure without
223 the classical complication of salt (Fig. 4). The Jebel Bou Djebbah anticline structure shows once more a
224 salt with conformable contact and a cretaceous series that plunges under the Triassic sequences.
225 Northward, in the Wadi El Ahmer locality, an Aptian sequence crops out in the main salt structure. This
226 outcropping Cretaceous series was initially described and dated by Masrouhi et al. (2013) as an upper
227 Aptian sequence. For the first time in this region, a work highlighted a fold structure in which Triassic
228 salt is incorporated in recent series. The Wadi El Ahmer locality shows a NNE-trending anticline structure
229 including upper Aptian Triassic salt cover sequences (for dating see Masrouhi et al., 2013) without the
230 typical salt folding. In the middle part of the large salt outcrop, an Aptian sequence was also identified as
231 the bedrock of the Triassic evaporites (Fig. 4). This anticline shows a “conformable” Triassic contact
232 (ravinant, glauconitic and reworked Triassic insoluble elements) at a dipping angle of 30° subparallel to
233 the above Aptian sequences. Far northwest, the Lella Selilia offers an outcrop of salt sheet with a salt
234 contact subparallel to its cover that is constituted by a normal stratigraphic succession from upper
235 Aptian to Campanian sequences. In the northeastern part (the Sebaa Regoud) a “lower conformable
236 contact” (with respect to the salt position) crops out in this entire region.

237 The central part of this structure shows the main salt body of Lansarine in contact with two cretaceous
238 outcrops i.e. the Ain El Karma Locality and the Jebel Djedaria (Fig. 5): At Ain el Karma, the Triassic salt
239 covers dark greenish clays with many nodules associated with slumps (a) themselves covering
240 limestones (b). For 2 km from the south to the north, the base of the Triassic salt is in contact only on
241 the term (a). This series has long been regarded as overturned middle Albian series by all authors
242 (Jauzein, 1962; Zargouni, 1975; Perthuisot, 1978). Masrouhi et al. (2013) presents a detailed

243 micropaleontological association in which a distinction has been established with lower Albian in the
244 base and middle Albian in the upper part, confirming a normal stratigraphic succession. The Jebel
245 Djedaria shows the only southern outcrop of Cretaceous sequences under the tertiary complex
246 configuration of the Jebel Lansarine. This locality offers an outcrop showing a reduced Cretaceous
247 sequence (compared to their equivalents in northern Tunisia) covering Triassic salt. This section is
248 formed by Upper Albian to Santonian sequences with normal lithostratigraphic polarity. Thickness and
249 contact nature (glauconitic, transgressive and phosphatic) testify to a transgressive Cretaceous
250 sequence on an existing submarine salt sheet that was extruded on a water–sediments interface. The
251 interbedded position confirms a rapid salt spreading (1 to 2 Ma). Moreover, the Lansarine central part
252 shows a 500m thick Triassic salt sequence in the East (Ain Karma) and 150 m thick sequences in the
253 Jebel Kef Debba (Fig. 5), possibly indicating a northwestward thinning of salt and thus a possible
254 northwestward spreading.

255 The LBSS corresponds to the second largest Triassic salt body in contact with only a few Mesozoic
256 outcrops, making the interpretation of this structure difficult. The geologic maps and field data (Figs. 3-
257 6) display evidences that the LBSS is a curved deformed zone bounded mainly by NE- to E-trending faults
258 systems. Its northern side is affected by the 8-km long N40°E-trending Baouala fault. In the main salt
259 body this fault is well identified by microfolding along the fault trace. Along the southeastern side of the
260 Jebel Malouia this fault is confirmed as the main thrust in this region, delimiting a hanging-wall with
261 Triassic salt and a footwall formed by Campanian to Miocene sequences. The offset is estimated to be
262 between 1 and 2 km, regarding the salt position versus the Cretaceous sequences and Miocene
263 sequences. This thrust bounds two varied domains in relation to the salt position: to the northwest the
264 Triassic salt overlies upper Aptian (described afore in Wadi el Marir, Wadi Ahmer, Jebel Bou Djebbah
265 and Sebbaa Regoud) and to the southwest salt overlies Albian sequences (Wadi Maiou). The folding
266 curves and the related structures attest a post Miocene shortening. Three fault sites in this region were
267 analyzed and display a major NW-SE Miocene to quaternary compression state. Using the fault diagram
268 (Fig. 4), the first site (sf_1) indicates a compressional tectonic regime locally characterized by N337°-
269 trending maximum stress axis (σ_1). The second site (sf_2) measured in the Campanian limestone
270 forming the footwall of the Baouala thrust indicates a similar compressional tectonic regime locally
271 characterized by N145°-trending maximum stress axis (σ_1). The third site (sf_3) measured in the
272 Miocene limestones also indicates a compressional tectonic regime locally characterized by N343°-
273 trending maximum stress axis (σ_1). These three sites are compatible with the majority of the tectonic
274 studies in northern Tunisia illustrating a NW-trending Miocene to present day main compression (e.g.

275 Melki et al, 2011; Soumaya et al., 2015; Booth-Rea et al., 2018). The fault kinematics analysis is
276 compatible with field data illustrating a southeast verging of the Baouala thrust.

277 The Lansarine region of northern Tunisia shows a plateau with horizontal Eocene sequences resting on a
278 well-outcropping marine Miocene series on its southern side and overlapping a detritic series of the
279 Fortuna formation commonly attributed to the Oligocene (Fig. 6). The Fig.6 shows the configuration and
280 reproduces the geological history of this region. The Oligocene–lower Miocene sequences cover a
281 Cretaceous unit of the Jebel Djedaria and show a moderate angular unconformity of about 15°. The
282 Paleogene (Paleocene and Eocene) subhorizontal sequences of Lansarine consist of two units forming
283 the so-called “Lansarine nappe” (Masrouhi et al., 2007). The previous mapping estimates a
284 southeastward advance of about 10 km from the peel thrust of Mateur, themselves are displaced, which
285 therefore constitutes a minimum estimation based on erosion limits. The thrust overlaps are well
286 fossilized in the entire Lansarine plateau. The top of the Miocene sequences under the thrust was dated
287 as Serrevalian. This implies that this configuration attests at least two tertiary compressional events. A
288 first Eocene episode with a moderate folding which corresponds to the well-known Atlassic
289 compressional event (Masrouhi et al., 2008) recognized on the overall scale of the Maghreb (Frizon de
290 lamotte et al., 2000; Dhahri and Boukadi, 2010). A second paroxysmal Tortonian compression (the
291 Alpine phase) is responsible for the present structure and deformations observed along this region
292 (Chihi, 1992; Bouaziz et al., 2002; Bahrouni et al., 2014, Gharbi et al., 2014; Ahmadi et al., 2019).

293 The second characteristic, well identified by maps and geometric relationships between entities, is the
294 E-trending fault systems. This Atlassic domain shows active N110-120° E-trending faults systems
295 involved in the tertiary deformation affecting all sediments including the Mio-Pliocene sequences, which
296 confirms their recent activity. This system frequently displays a multiphase striation for which the first
297 one is classically related to the Mesozoic extensional period. This system is usually associated to
298 thickness variations, slump folds and conglomeratic horizons indicative of their early Mesozoic
299 evolution. Using the fault diagram, we rotated the collected fault data (Fig. 7) to restore the bedding
300 plane to its horizontal orientation. We also used the back-tilting fault kinematics analysis of Naji et al.,
301 (2018b) carried out in the northern Tunisian domain. The back-tilting result shows that these faults were
302 normal faults predating the Cenozoic tilting (Fig. 7). Back-tilting of the measured lower Cretaceous fault
303 population indicates a normal faulting that result from an extensional tectonic regime locally
304 characterized by NE-SW minimum stress axis (Fig. 7). The back-tilting of the Aptian–Albian measured
305 fault population indicates a normal faulting that result from an extensional tectonic regime locally

306 characterized by E-W to NE-SW minimum stress axis (Fig. 7). This phase is mainly characterized by NW-
307 trending fault systems and associated sub-basins. This change from lower Cretaceous to Aptian-Albian
308 periods, is interpreted in Tunisia as the translation from the Jurassic E-W (present-coordinates)
309 dominant shape to Early Cretaceous NW-SE shape, which is concordant with the general E-W transform
310 faults of the Atlantic that translate to NW-SE during the Alpine Tethys faults (Rosenbaum et al, 2002;
311 Frizon de Lamotte et al., 2013; Masrouhi et al., 2019). The tectonic regime changes during the late
312 Cretaceous post-Rift period. Locally, the Lansarine domain seems to be controlled by normal faults
313 characterizing a NW-SE to N-S tectonic extension (Fig. 7).

314 **4. Seismic interpretation**

315 Three seismic profiles are used and interpreted, for which some of them are composite ones. The
316 seismic sections were provided by ETAP Petroleum National Company and the Carthago Oil Company
317 Tunisia. For some section the acquisition was carried out by Shell-Tunirex Company in 1983. Seismic
318 profile data are used to correlate the fault systems and/or tectonic blocks with the above-described
319 sedimentary sequences. These Seismic profiles were correlated by petroleum wells as well as the
320 stratigraphic sequences and outcrops established by the present-study fieldwork.

321 In this section, seismic profile interpretation is used to document the presence of allochthonous salt
322 sheets and a buried salt diapir in the study area. Three seismic profiles are used to highlight the
323 geometry at depth, halokinetic type and entities relationship. The seismic section L18 is a composite
324 section derived along the Lansarine–Baouala region (Fig. 8). The ~ 35 km long NW-SE L18 profile crosses
325 the LBSS near the Baouala structure (Fig. 3). Furthermore, two seismic profiles are used to highlight the
326 distribution and the relationship between entities that are respectively the NE-SW L05 profile in the
327 northern part and the NE-SW L11 profile in the southern part (with respect to the salt structure). These
328 seismic profiles were analyzed and interpreted using calibration from the W-1, W-2 and W-3 petroleum
329 wells (see Fig. 3 for location). The internal geometry of the shallow portion of the profiles in the LBSS is
330 calibrated by the surface observations of this study.

331 The seismic section L18 highlights the Teboursouk and the Baouala fault zone that is characterized by a
332 deep north-dipping basement fault and shallow listric normal faults within inferred Triassic evaporites,
333 which confirms previous results (Rigo et al., 1996; Masrouhi et al., 2013). This basement fault zone was
334 mostly active as a normal fault during Mesozoic times, allowing the deposition of a thick sedimentary
335 pile in this northern Tunisian domain, reaching up to a thickness of 6 km. This major basement normal

336 faulting controls Mesozoic basin with associated changes in thickness well observed for the Jurassic and
337 lower Cretaceous sequences. Structurally, it is clear that the present tectonic style of the area is the
338 result of a post Cretaceous structural inversion. In this zone, the thick sedimentary cover is deformed by
339 thrusts and large-scale folding above the southeastward major thrusts connected in the interface
340 basement-cover (Triassic salt?). Thrusts are superimposed on deep seated basement faults which were
341 active during Mesozoic times as major normal deep faults. The diapir, detected at depth for the first
342 time, seems to be a minor complicating factor in Tertiary shortening deformations. Geometry, thickness
343 variations, together with Triassic evaporites associated to a major thrust, suggest that the basement-
344 cover interface geometry of this domain is poorly deformed (Fig. 8). The profile shows major thrust
345 systems developed beneath the Chougui basin at the base of the basement–cover interface. The far
346 northern part shows an imbrication of southeast-verging thrusts and gently dipping asymmetric folds
347 filled by Oligocene-Miocene series (Fig. 8). The absence of the lower Cretaceous series in the hanging-
348 wall and the base of these minor thrusts suggest a Cenozoic development. They probably do not
349 correspond to inverted faults. In between, major southeast verging thrust systems are developed and
350 major normal fault between Baouala block and Mateur plain are preserved. This seismic section
351 underlines a buried diapir at depth. This stem of the LBSS is revealed as a salt body now buried under
352 the complex thrust configuration of the LBSS. The main information highlighted by this seismic section is
353 the compatibility between surface and subsurface data showing the presence of allochthonous salt
354 sheets detected on the surface and their thinning towards the north. The salt sheet is detached from its
355 autochthonous original position. A salt sheet (canopy?) is interlayered between lower Cretaceous beds.
356 This 8 to-10 km-long salt sheet is actually located in the hanging-wall of the main southeastward thrust.
357 The seismic section also highlights that salt tectonics seems to be controlled by Mesozoic normal
358 faulting associated with well-developed rim synclines (salt mini-basins?) in lower Cretaceous thick
359 sequences around the main salt body. Salt mini-basins are actually well recognized as being associated
360 to salt movement in many worldwide salt provinces (Hudec and Jackson, 2007; Callot et al., 2012; callot
361 et al., 2016). Mini-basins were likely formed as the result of the synsedimentary upward movement of
362 Triassic salt conducted by the differential sedimentation loading itself guided by the major normal
363 faulting. At present, the absence (or thinning?) of salt under Mateur plain may be explained by the
364 lateral flow of salt from the main stem during Mesozoic times. This phenomenon, called salt evacuation,
365 is well known in many other salt provinces (e.g. Gulf of Mexico, Santos basin ...).

366 The seismic section L05 crosses the northern domain and mainly shows an extensional geologic setting
367 in this area (Fig. 9). This section was interpreted by Booth-Rea et al., (2018) as one of the low angle fault

368 systems named the Ghezala fault characterizing the extensional late Miocene collapse of northern
369 Tunisia. This work interprets the major low angle normal faults as associated to high-angle faults
370 producing a geometry characterized by half-graben and hanging-wall syncline basins during the late
371 Miocene times. This interpretation of extensional collapse in the Atlas system during the late Cenozoic
372 African-Eurasian convergence setting was related to deep mantle tectonic mechanisms in the
373 concomitant extension of Tyrrhenian basin to East. While we agree with the core interpretation of
374 Booth-Rea et al., (2018), we enhance and add some details in this section. The Miocene extensional
375 system appears to be connected down on a major normal fault system (affecting both basement and
376 cover, Fig. 9) such faults are probably inherited from the former Mesozoic Tethyan extensional period
377 later reactivated during late Miocene times. This section highlights a lower Miocene extensional period.
378 This extension, which occurred after the short Eocene compression, is mainly accommodated by large
379 extensional faults connected down on basement faults systems that seem to be inherited from the
380 former Tethyan basin.

381 The NE-SW seismic line L11 is located in the front of the thrust systems of LBSS and shows a narrow
382 basin filled by Mio-Plio-Quaternary sediments reaching a 3 km of thickness (Fig. 10). In opposition to the
383 northern domain, this section mainly exhibits high-angle normal fault systems. This narrow basin is
384 mainly recorded by Miocene syntectonic strata. Regarding the position (between the Baouala thrust and
385 the major Teboursouk thrust), the connected high angle normal faults and the sequences in this narrow
386 basin are all interpreted as an orthogonal extension in which the thrust strongly influences the
387 development of normal faults. This section exhibits a strong variation in thickness associated to faults. In
388 addition, the thickness variations are remarkable for Jurassic and lower Cretaceous sequences, which
389 testifies that this fault is inherited from Mesozoic times.

390 The seismic data seems to be well correlated with gravity results recently presented by Sarsar et al.
391 (2017). According to the superimposition of the Euler solutions to Bouguer Anomaly map used to
392 present the distribution of anomalies and curved gravity lineaments about positive anomalies and also
393 estimate the depth of those anomalies, Sarsar et al. (2017) suggest that the west of the LBSS
394 corresponds to a fault concentric zone and provides the evidence of a pre-existing diapiric geometry
395 body located at the depth between 1 and 2 Km and later buried by Mesozoic and Cenozoic sediments.
396 The NW-trending of the main gravity contours reach a depth of 2 km and confirm the existence of the
397 Teboursouk fault as a major thrust fault system. This system is probably inherited from Mesozoic times
398 and acted as the main control in favor of diapir upward (for more details see Sarsar et al., 2017). The

399 concentric behavior in the west of the LBSS is probably connected downward to the Teboursouk thrust,
400 which explains the presence of sub-Triassic structures in this belt.

401 5. Discussion

402 *i. Previous interpretations:* Several interpretations were made separately by several authors to
403 explain the diapiric structure that may be summarized as follows: 1- Bolze (1954) suggests a salt move-
404 up because of space creation via normal faulting, 2- Zargouni (1975) suggests a scenario in which salt
405 intrudes forcefully into its overburden during Mesozoic and Eocene times, and 3- Perthuisot (1978)
406 suggests a way in which salt intruded forcefully into its overburden during late Cretaceous-Eocene
407 compression. This structure, called by all previous studies "the Lansarine diapir », was presented as a
408 cylindrical, 45 km long and 15 km-wide vertical diapir. Our field mapping does not support this
409 hypothesis of a single 15 km-wide vertical diapir. The structure and the salt budget do not permit to
410 interpret it as one rooted salt structure. 4- Truillet and Deltail (1982) propose a large thrusting system.
411 Based on the observation that the Triassic salt is interbedded and subparallel to the Cretaceous layers,
412 this work propose that all this region corresponds to a large tectonic nappe and must be integrated to
413 the front of the Numidian allochthonous zone. 5- Finally Masrouhi et al. (2013) propose a salt canopy
414 model. This model proposes a scenario in which extension induces normal faults that provide routes to
415 the salt to reach the surface and flow downslope.

416 *ii. Field data discussion and tectono-sedimentary features:* Field work, together with the
417 interpretation of seismic profiles, shows that the present LBSS may result from a long-time polyphasic
418 geological history. This 45 km-long and 15 to 5 km-wide salt structure was likely passively transported on
419 major thrusts during the Atlasic and Alpine shortening. These main tectonics events are recognized to
420 occur during the Late Eocene (Atlassic) and Miocene–to present-day for the alpine event (e.g. El Ghali et
421 al., 2003; Masrouhi et al., 2008; Dhahri and Boukadi, 2010). The present structure is also the result of
422 considerable erosion and probably dissolution during late Cretaceous and Cenozoic times. It likely
423 covered a wider area during Mesozoic times. In this study we confirm by seismic examination the
424 presence of a salt diapir body deeply buried under a complex thrust system which was absent in the
425 previous interpretation. We also confirm by surface data that the outlines of the Triassic salt are
426 commonly subparallel to the bedding of the adjoining rocks succession of the northern and southern
427 parts. The contacts of the Triassic rocks are not discordant with them. Geologic map and field cross
428 sections are consistent with an allochthonous sheet of Triassic salt emplaced over the Early Cretaceous
429 successions. For the LBSS, field data emphasizes the absence of radial faulting that is usually associated

430 with diapirism growth well observed and documented in others salt province (e.g. Rowan et al., 2003;
431 Yin et al., 2009). Based on field investigations, we therefore suggest (for the central and northern part) a
432 composite salt sheet covering Early Cretaceous sediments. Established by seismic interpretation, a
433 diapiric stem is detected under the LBSS large salt sheet. In the LBSS area, Early Cretaceous sediments
434 are characterized by considerable variation in thickness and exhibit numerous sedimentary features
435 (nodules reworking, slumps and conglomerates lenses) within an active extensional tectonic regime
436 (Naji et al., 2018b). This created an irregular seafloor resulting from normal faulting. In the LBSS area,
437 several preserved extensional structures as small-scale sealed normal faults constitute a useful marker
438 for this tectonic activity. Structurally, the LBSS structure occupies a position in the intersection of the
439 major faults systems i.e., the E-trending strike-slip and the NE-trending thrusts fault systems. Surface
440 data together with the seismic profile interpretation highlight a general shallow north-dipping high-
441 angle normal fault above a major deep listric one. Mechanically, this can explain the weakened
442 overburden of the Triassic salt. The LBSS corresponds at present to a complex structure involving Triassic
443 salt and bounded by north dipping fault systems (Figs. 7-10). It seems to be the stem of the LBSS diapir
444 during Mesozoic time before the Atlasic shortening. This emplacement was likely driven by normal
445 faulting, which affected both the cover and the basement, along with differential loading. This context
446 fits with the southern Tethyan passive margin setting well documented by previous studies (e.g.
447 Souquet et al., 1997; Jaillard et al., 2013; Soua, 2014; El Amari et al., 2016; Naji et al., 2018a).

448 **iii. Diapirism style:** Taking in account the overall structure, the LBSS shows strong similarities to
449 those recently described by authors around many salt provinces (e.g. Saura et al., 2014; Alsop et al.,
450 2015; Martin-Martin et al., 2017; Jaillard et al., 2017; Zucker et al., 2019). The halokinesis process was
451 particularly active during the rifting stage, in which tilted block geometry was responsible for differential
452 loading. During the Alpine compression, diapiric salt corresponds to weak surfaces used by the Alpine
453 thrust systems. Structural analysis indicates that the LBSS emplacement was guided by a basement
454 normal fault system related to the south Tethyan rifting. The general framework of this margin, together
455 with similarities between the LBSS and recently published diapirs from other salt provinces, led to the
456 proposal of polyphase diapirism history. Within the Maghreb framework (e.g. Soussi et al., 2017; Dhahri
457 and Boukadi, 2017; Ahmadi et al., 2019), the Jurassic–Early Cretaceous extensional context of the south
458 Tethyan margin in Tunisia allows a reactive mobilization of salt layers. This tectonic setting offers two
459 conditions allowing salt to extrude onto the sea-floor. (1) The extension weakens the overburden by
460 normal faulting creating space for salt to rise forming the diapir structure and (2) induces thickness
461 variations leading to differential loading which initiate the salt movement (Fig. 11). During the Tethyan

462 rifting period, lower Cretaceous extensional tectonic regime induced thick depositional basin sediments
463 controlled by major basement normal faults. Within this structural setting, well recorded in North
464 Tunisia, the salt's overburden was weakened by faulting creating space, along with salt (acknowledged
465 to be Newtonian fluid at depth) withdrawing from the weakened zone by faulting. Accumulation of thick
466 salt overburden in subsiding area acts as a rapid differential loading that encourages salt flow from the
467 diapir stem. This evolution corresponds to a reactive diapir (sensu Vendeville and Jackson, 1992b), which
468 extruded along the main fault system (Fig. 11). Active diapirism would correspond to a short lower
469 Cretaceous event, currently not observable because it has been overprinted by tectonic compressional
470 deformations. The basement fault movement, salt withdrawal and growth of diapir itself; all of them
471 may have induced a significant asymmetry creating salt mini-basins well recognized as associated to salt
472 movement in many worldwide salt provinces. Above-mentioned observations indicate a structural
473 configuration that led to lateral spreading onto the depositional surface over the LBSS diapiric stem (Fig.
474 11). It corresponds to a well-known salt process that could be active both in submarine (Fletcher et al.,
475 1995) and sub-aerial environments (Talbot, 1998).

476 **iv. Regional consideration:** The structural style of the LBSS seems to be very similar to other
477 structures recognized as being the result of basement faults movement, on which extensional thin-
478 skinned deformation is connected. The trigger mechanism of salt migration of the LBSS can be compared
479 to other similar salt provinces: e.g., Pyrenean orogen (Canérot et al., 2005; Saura et al., 2016), the
480 Eastern Betics (Roca et al., 2013; Rubinat et al., 2013), Nordkapp Basin (Koyi et al., 1995), or North Africa
481 (Bouchouata et al., 1995; Saura et al., 2013, Masrouhi et al., 2014a, Moragas et al., 2017, 2019). The
482 original criteria of the Salt structure is preserved likely because it was passively transported on major
483 thrusts, superimposed on deep seated basement faults, during the Atlassic and Alpine shortening
484 periods. The Northeastern Maghreb domain is one of the world's least-known areas in terms of salt
485 tectonics (Hudec and Jackson, 2007). It appears similar to many others inverted salt provinces, for which
486 it is difficult to interpret the original geometry and evolution of salt diapirs. In this domain, the
487 buoyancy-driven halokinetic model has long been considered as the unique way of salt growth. This
488 model postulated an autonomous isostatic rise of salt and piercement of the overburden due to the
489 Rayleigh–Taylor instabilities. Based on this model, the diapirs of North Tunisia have been considered to
490 expose only “vertical” diapir structures (Perthuisot, 1978; Perthuisot et al., 1998; Rouvier et al., 1998;
491 Jallouli et al., 2005). Recently, an allochthonous salt model has been documented by Masrouhi et al.
492 (2013) in the Alpine thrust front of Teboursouk (Teboursouk thrust unit, Fig. 1). The authors suggest a

493 scenario where salt broke out at the sediment–water interface or below a thin layer of marine
494 sediments forming an allochthonous extrusion, similar to the present-day salt tectonics described in the
495 deep–water Gulf of Mexico (Fletcher et al., 1995) or Angola margin (Fort et al., 2004; Ge et al., 2019).
496 Later, the pre-existing allochthonous salt was deformed during Miocene–present day Alpine
497 compressions that used the weakness of the salt sheet as thrust décollement. The interconnection
498 between these two types (i.e. vertical diapir and large connected salt sheet) has never been described
499 yet.

500 **v. *The LBSS style:*** The LBSS clearly shows asymmetric overhangs flowing northward (the subsiding
501 area). The first part of the LBSS evolution for the sheet connected to the Lansarine salt diapir can be
502 easily distinguished as an extrusive advance (*sensu* Hudec and Jackson, 2006), in which an exposed salt
503 sheet spreads at the water–sediment interface without any roof. Above the apex of the diapir, that
504 corresponds to the feeder, salt flows outward and spreads by gravity downslope (Fig. 11). The
505 overburden on the salt may have formed a synclinal minibasin on top of the salt sheet (Fig. 11). In this
506 case, the minibasin had a thickness usually determined by the salt evacuation budget forming in some
507 cases a salt weld (e.g. Diegel et al., 1995; Schuster, 1995; Callot et al., 2012). In this domain, the low-
508 amplitude syncline of Chouigui (Fig. 9) is likely the minibasin developed by evacuation of allochthonous
509 Triassic salt during Mesozoic times (see Fig. 11). From late Cretaceous to Eocene time, the onset of the
510 compressional Atlassic stage (e.g. Guiraud and Bosworth, 1997; Frizon de Lamotte et al., 2000) resulted
511 in widespread shortening associated with the inversion of the normal faults and probably without
512 significantly compressing the LBSS. The diapir was passively transported above the southeast-verging
513 major thrust which is deeply rooted in the salt beneath the Mateur plain (Fig. 11). During this period, the
514 LBSS was eroded due to fold-related uplifting.

515

516 **6. Conclusion**

517 Based on the results of detailed field work and seismic profiles' interpretation of the Lansarine-Baouala
518 Salt Structure in the Northern Atlassic domain of Tunisia, the following conclusions can be made:

- 519 • Salt diapirism in Northern Tunisia appears to be contemporaneous with tectonic extension. The
520 Jurassic–Early Cretaceous extensional tectonic context of the southern Tethyan margin induced
521 thick depositional basin sediments controlled by major basement normal faults. The extension

522 weakens the overburden by normal faulting and the thickness variations leading to differential
523 loading which allow a reactive mobilization of Triassic evaporites,

524 • During Mesozoic diapirism phase, a significant allochthonous salt spreads laterally from the diapir
525 stem. Salt flows outward and spreads by gravity downslope highlighting an extrusive sheet
526 advance. In this domain, the low-amplitude syncline of Chouigui seems to be a minibasin
527 developed in conjunction with the evacuation of allochthonous Triassic salt during Mesozoic
528 times,

529 • Compressional deformation stages do not reflect active diapirism. The Lansarine diapir and its
530 connected salt sheet were passively transported on the southeast-verging major thrust deeply
531 rooted in the salt beneath the Mateur plain during the Atlassic and Alpine shortening. In the north
532 Tunisian domain, the identification of allochthonous Triassic salt sheets connected to buried
533 diapir have a significant role in the shallow structural complexity observed in the sedimentary
534 cover during the subsequent contractional events.

535

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899 **Figure captions**

900 Figure 1: (a) Geodynamic setting of the Maghreb fold-thrust-belt and location of north Tunisian domain.
901 (b) Simplified sketch of the Tunisian Atlas showing the major tectonic units, major thrust systems, main
902 outcropping Triassic salt and the location of the LBSS.

903 Figure 2: Stratigraphy of the north Tunisian domain with indication of average thickness variation,
904 lithostratigraphic short description, the average density of main units and the regional tectonic regime.

905 Figure 3: Detailed geological map of the LBSS and its surrounding areas showing the main units, their
906 relationships, the Triassic salt shape, the location of the detailed maps of figs 4 and 5 and the location of
907 seismic profiles of figs 8-10.

908 Figure 4: Detailed geological map of the northern part of the LBSS showing the salt contact, the main
909 outcrops of Cretaceous that constitutes the bedrock of the large salt canopy, and fault kinematics
910 analysis of few fault sites highlighting a major NW-SE Cenozoic compression.

911 Figure 5: Detailed geological map of the central part of the LBSS showing the salt contact, the main
912 outcrops of Cretaceous that constitutes the bedrock of the large salt canopy (Ain el Karma) and the
913 cover of the Triassic salt (Jebel Djedaria), under the thrust systems of the Lansarine structure.

914 Figure 6: Panoramic view looking northeastward of the Jebel Djedaria structure shows the configuration
915 and reproduces the geological history of this region with (a) uninterpreted and (b) interpreted section.
916 In the south, the Triassic is covered by normal polarity Cretaceous sequences, themselves
917 unconformably covered by Oligocene–Lower Miocene (15-20° angular Atlassic unconformity). The
918 Oligocene–Lower Miocene sequences are overthrust by southeast-verging the Lansarine nappe
919 formed by Paleocene shale and Eocene limestone.

920 Figure 7: fault kinematics analysis of abundant meso-scale synsedimentary faults within the Cretaceous
921 sequences. The Stereonets are back-tilted faults that indicate a NE-SW Aptian–Albian tectonic extension;
922 E-W to NE-SW late Cretaceous tectonic extension; and Santonian NW-SE to N-S tectonic extension.

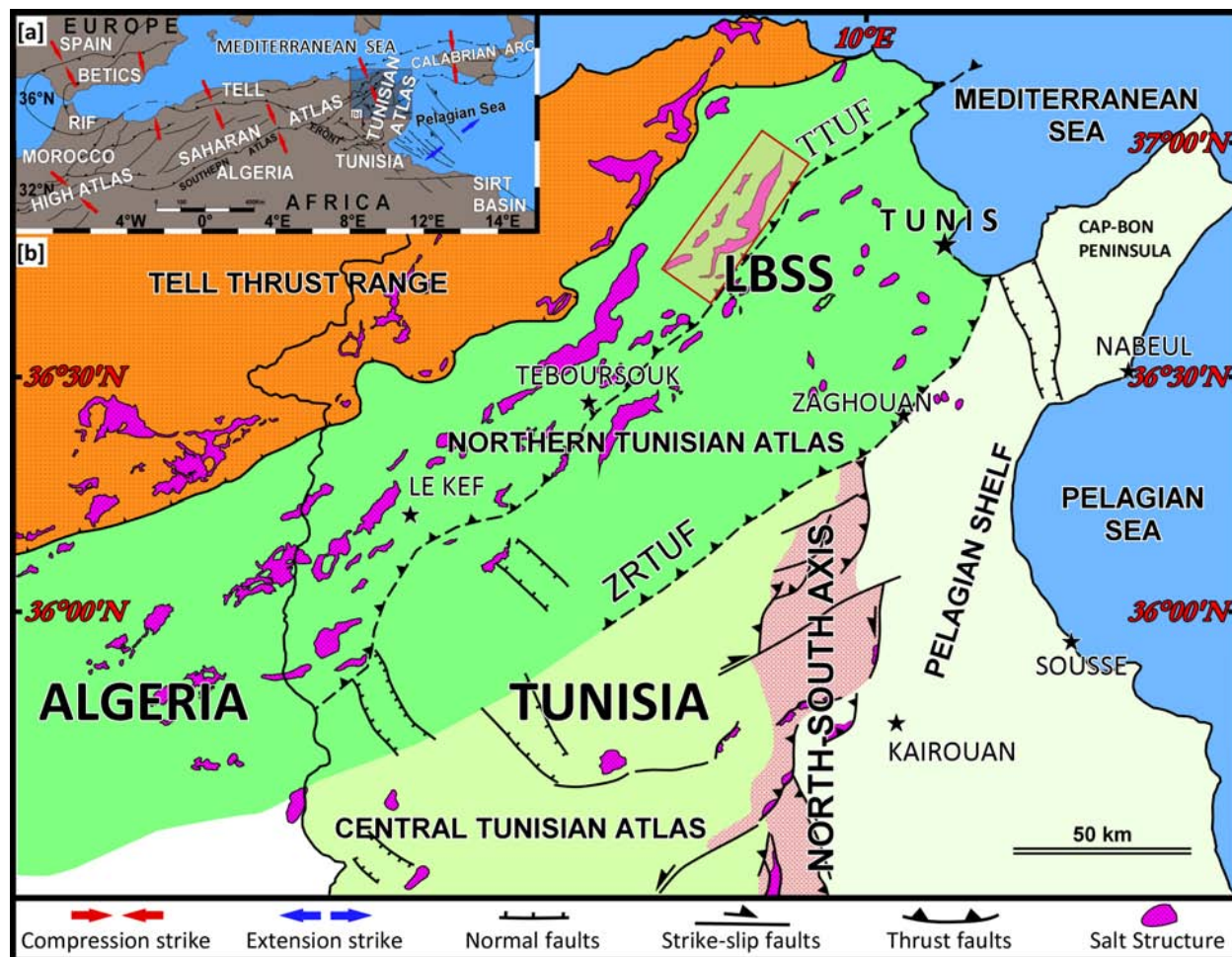
923 Figure 8: (a) Uninterpreted, migrated, depth-converted seismic reflection data from the LBSS area. (b)
924 Interpreted seismic line of L18A and L18B showing the preserved extensional Chouigui mini-basin to the
925 south associated with the major Teboursouk thrust fault. Note that the thrust fault systems are
926 connected to intra-basement detachment normal fault. The salt sheet covers a Cretaceous sequences at

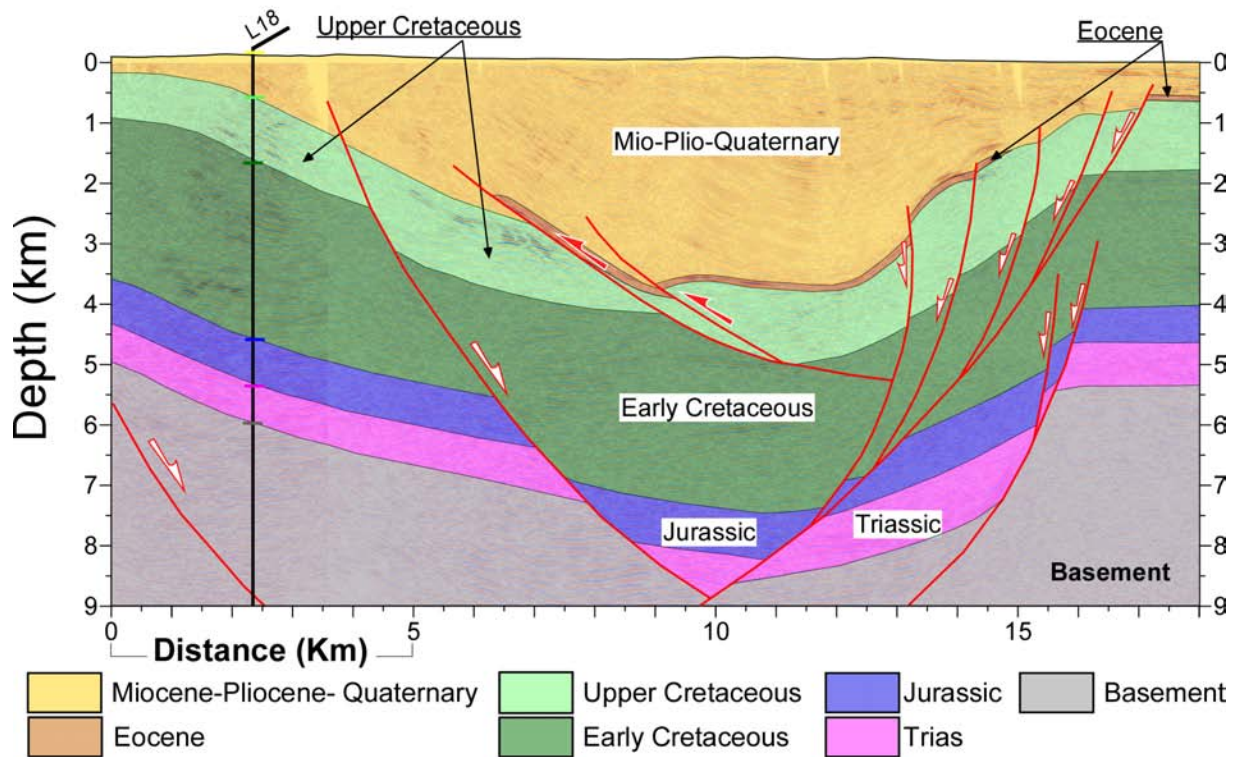
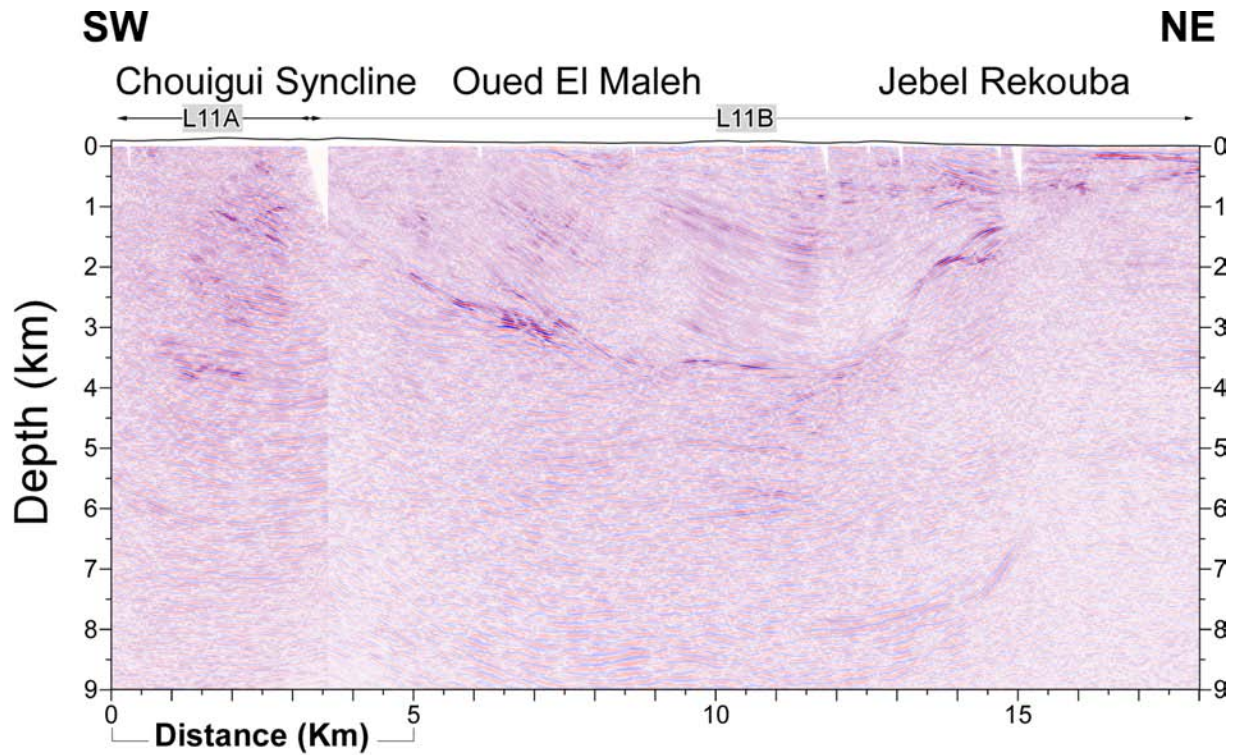
927 depth. Under Lansarine structure the velocity and the shape of reflectors seem to highlight a salt body
928 (Lansarine diapir). The thrust fault is connected at depth to rollover anticline limited by the listric normal
929 fault connected to the intra-basement detachment all of them are inverted structures.

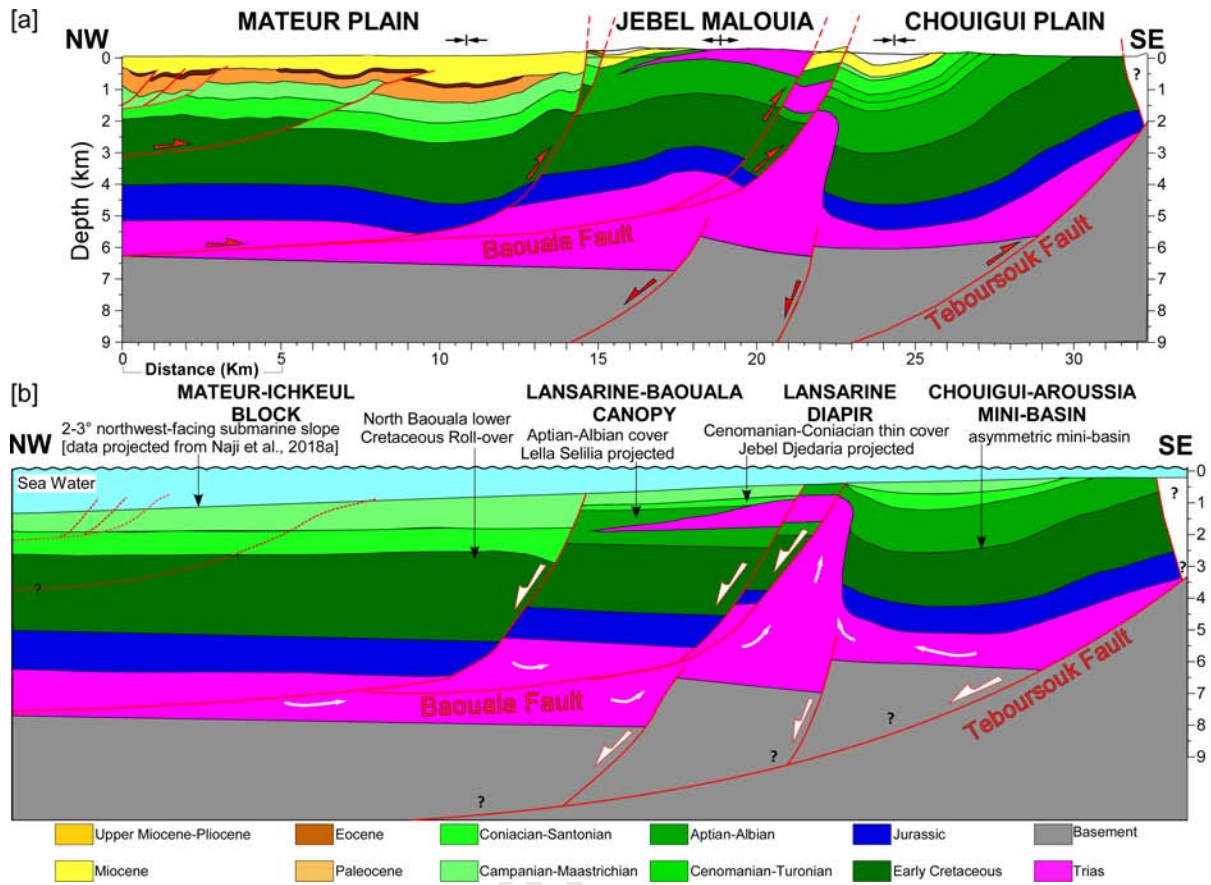
930 Figure 9: seismic section L05 crosses the northern domain. It mainly shows an extensional geologic
931 setting in this area with the low angle fault systems characterizing the extensional late Miocene collapse
932 associated to high-angle faults producing a geometry characterized by half-graben and hanging-wall
933 syncline basins

934 Figure 10: NE-SW seismic line L11 located in the front of the thrust systems of LBSS showing a narrow
935 basin filled by Mio-Plio-Quaternary sediments reached a 3 km and mainly recorded by Miocene
936 syntectonic strata. The high angle connected normal faults and the sequences in this narrow basin are
937 interpreted as an orthogonal extension in which the thrust strongly influences the development of
938 normal faults.

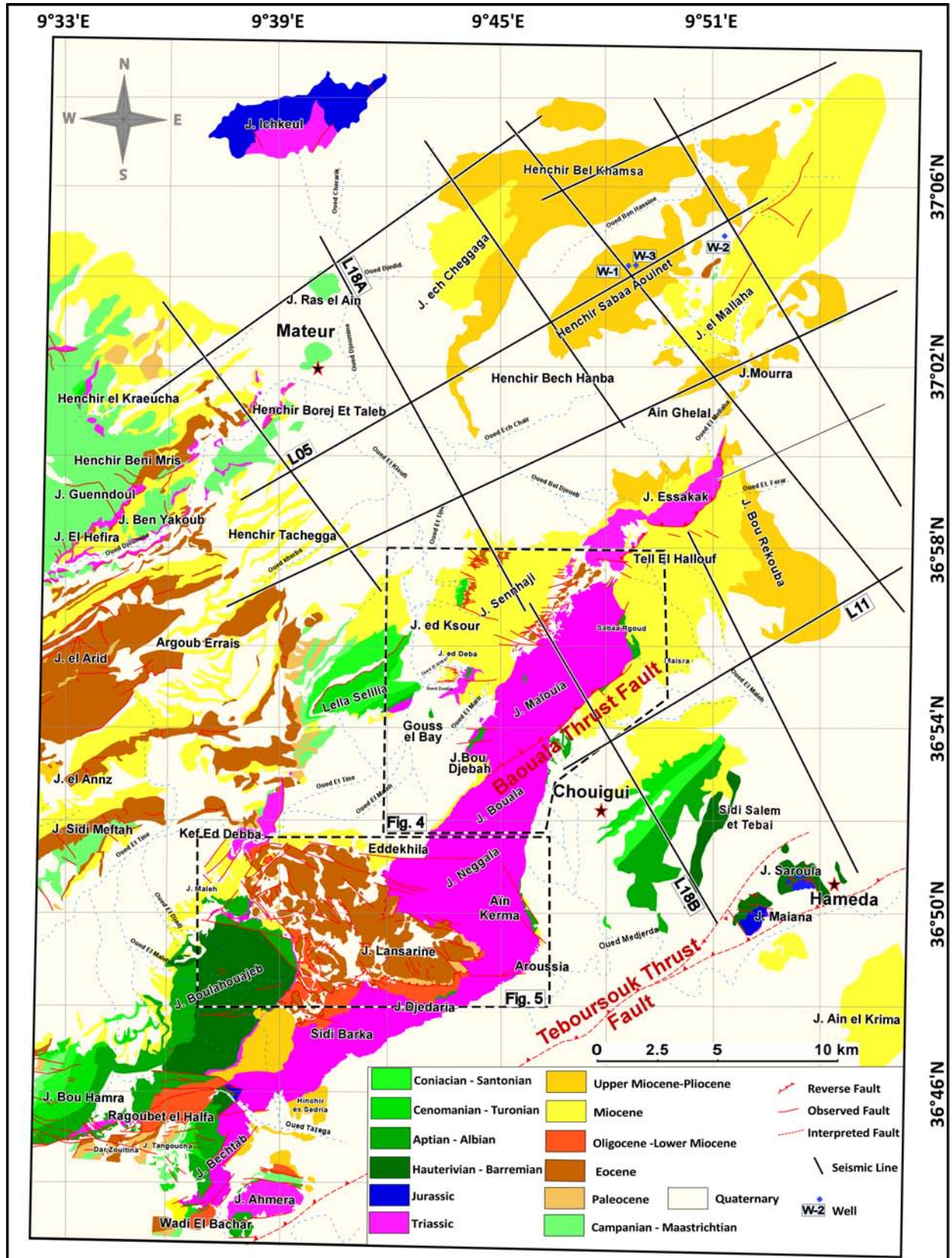
939 Figure 11: simplified Palinspastic stages reconstruction cross-section. (a) Present-day. (b) Late
940 cretaceous state. Shortening amount is hypothesized to be ± 6.5 km ($\sim 17-18\%$). See text for detail on
941 halotectonics stages and styles.

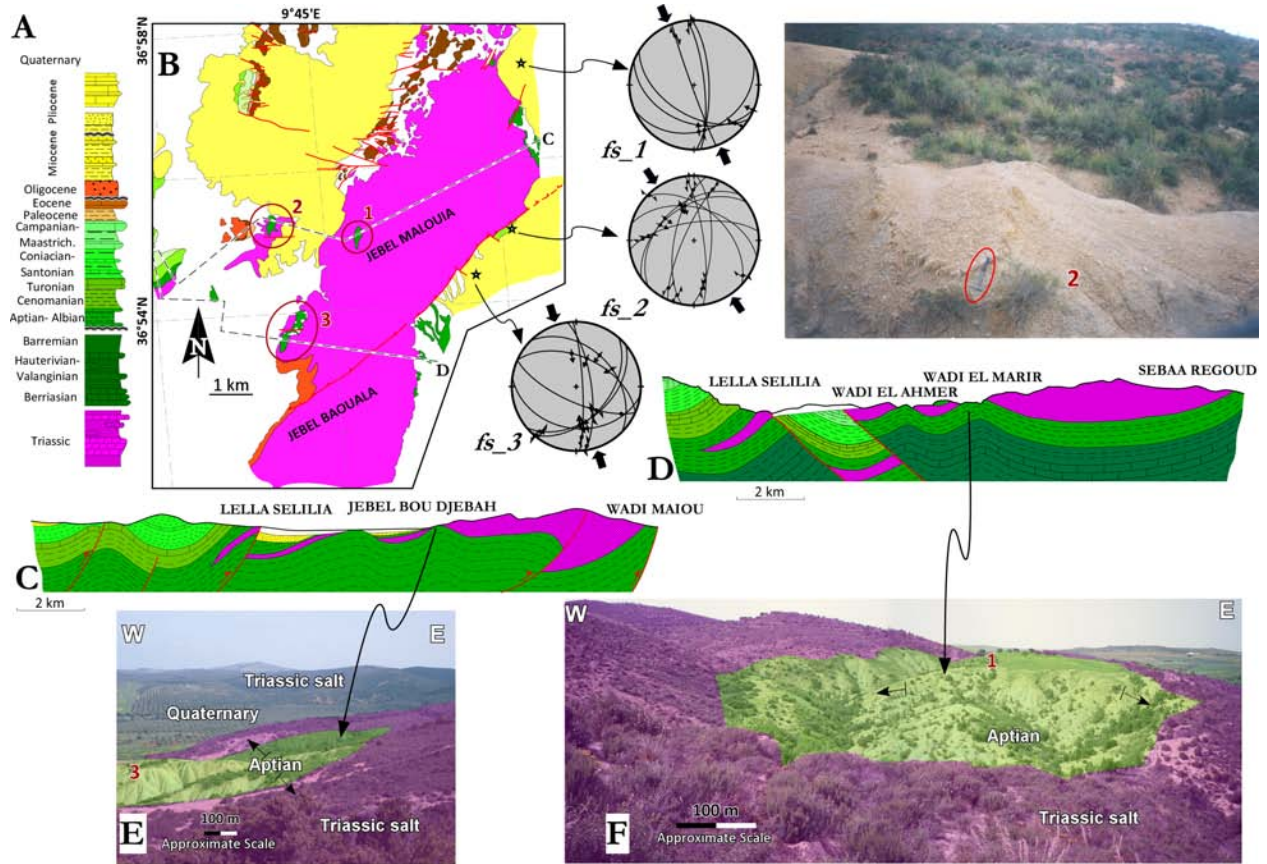


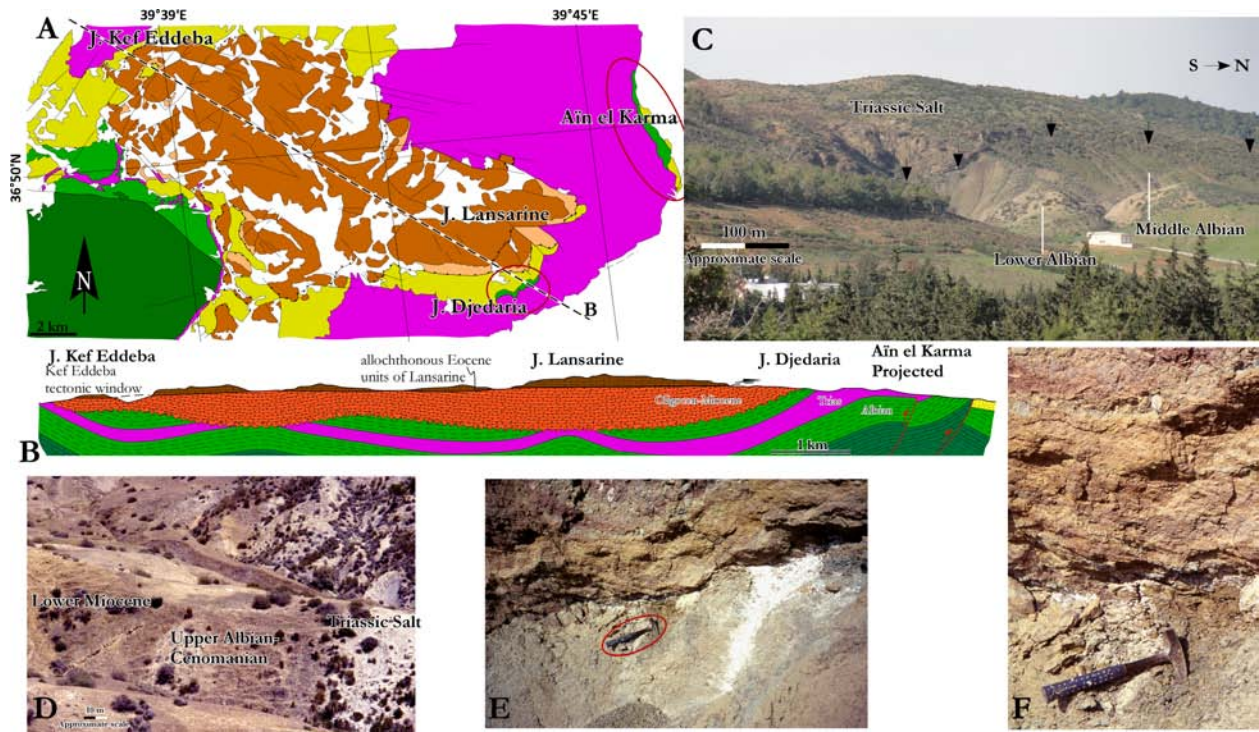


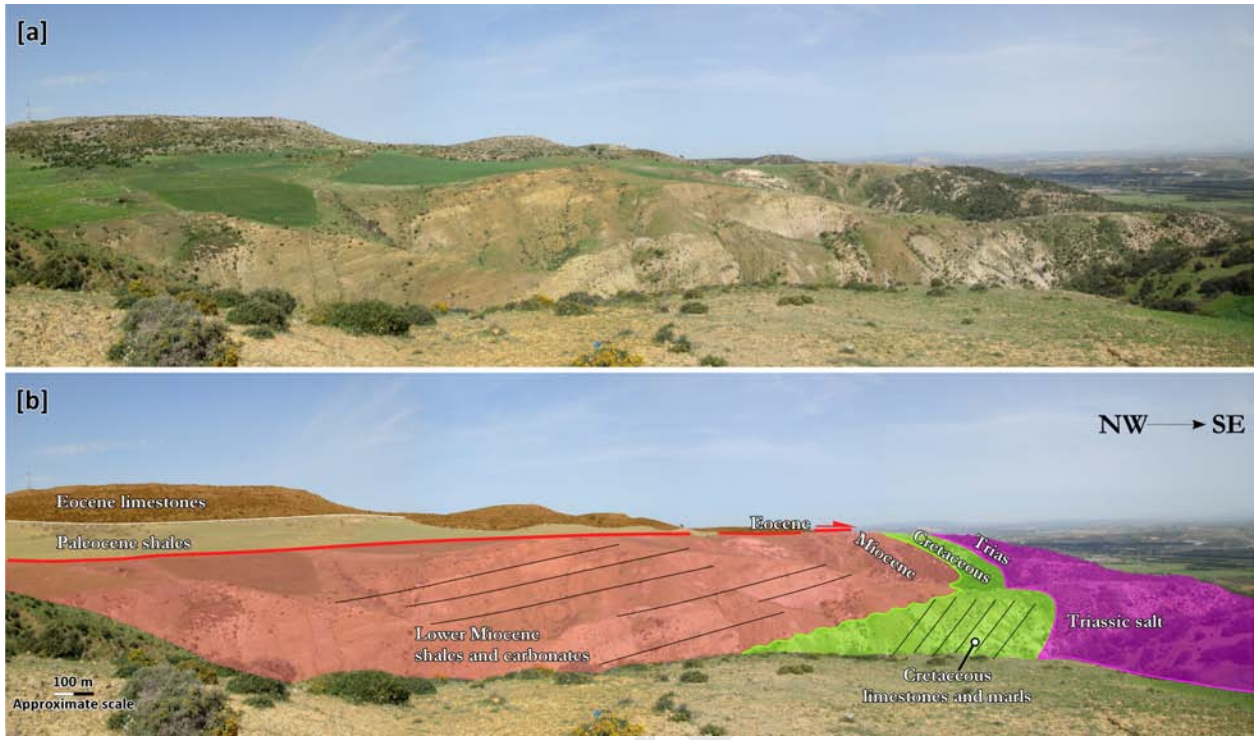


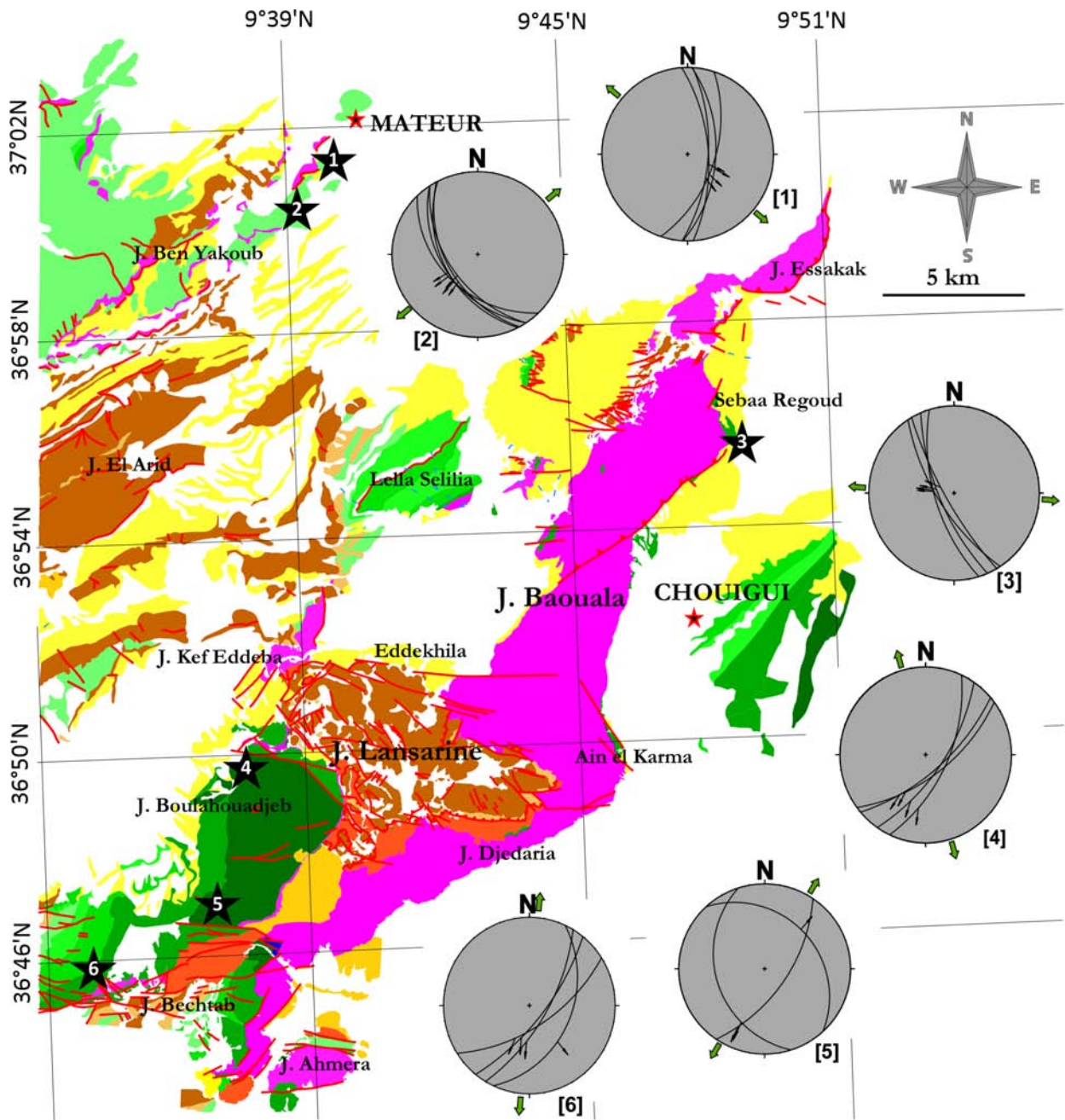
Age		Lithology and Density g/cm ³	Thickness variation	Lithostratigraphical description	Regional Tectonic	
Quaternary			2.05	0–100 m	Soil	
Pliocene			2.60	250–370 m	Sands and sandstones with rare clays intercalations	
Miocene	Messinian		2.60	150–300 m	Black clays with lacustrine marls and gypsum	
	Serrevalian-Tortonian		2.60	500–2700 m	Gray clays and marls with rare sandy intercalation	
Paleogene	Oligocene		2.58	0–50 m	Siliceous grainstone	
	Upper Eocene		2.62	50–100 m	Siliceous yellowish limestones	
	Paleocene		2.62	800–1000 m	Dark marls	
Cretaceous	Campanian-Maastrichtian		2.53	60–160 m	Marine limestones and claystones presented by two carbonate members	
	Coniacian-Santonian		2.52	350–900 m	Thick sequences of marls-limestone with marls dominance	
	Turonian		2.54	60–200 m	Marls and limestones alternation	
	Cenomanian		2.63	100–1000 m	Black limestones and green black marls	
	Aptian- Albian		2.63	100–1000 m	Black to green marls with calcareous layers	
	Barremian		2.63	450–700 m	Marls and dark nodular limestones alternations	
	Hauterivian-Valanginian		2.62	130–800 m	Marls with limestones and frequents sandstones alternations	
	Berriasian		2.62	60 m	Greenish muds and thin bedded sandstones and siltstones	
Jurassic			2.58	40–440 m	Carbonates composed of calcites and dolomites	
Triassic			2.17	150–1000 m	Thick continental evaporitic salt (halite-gypsum), clays and dolomites	

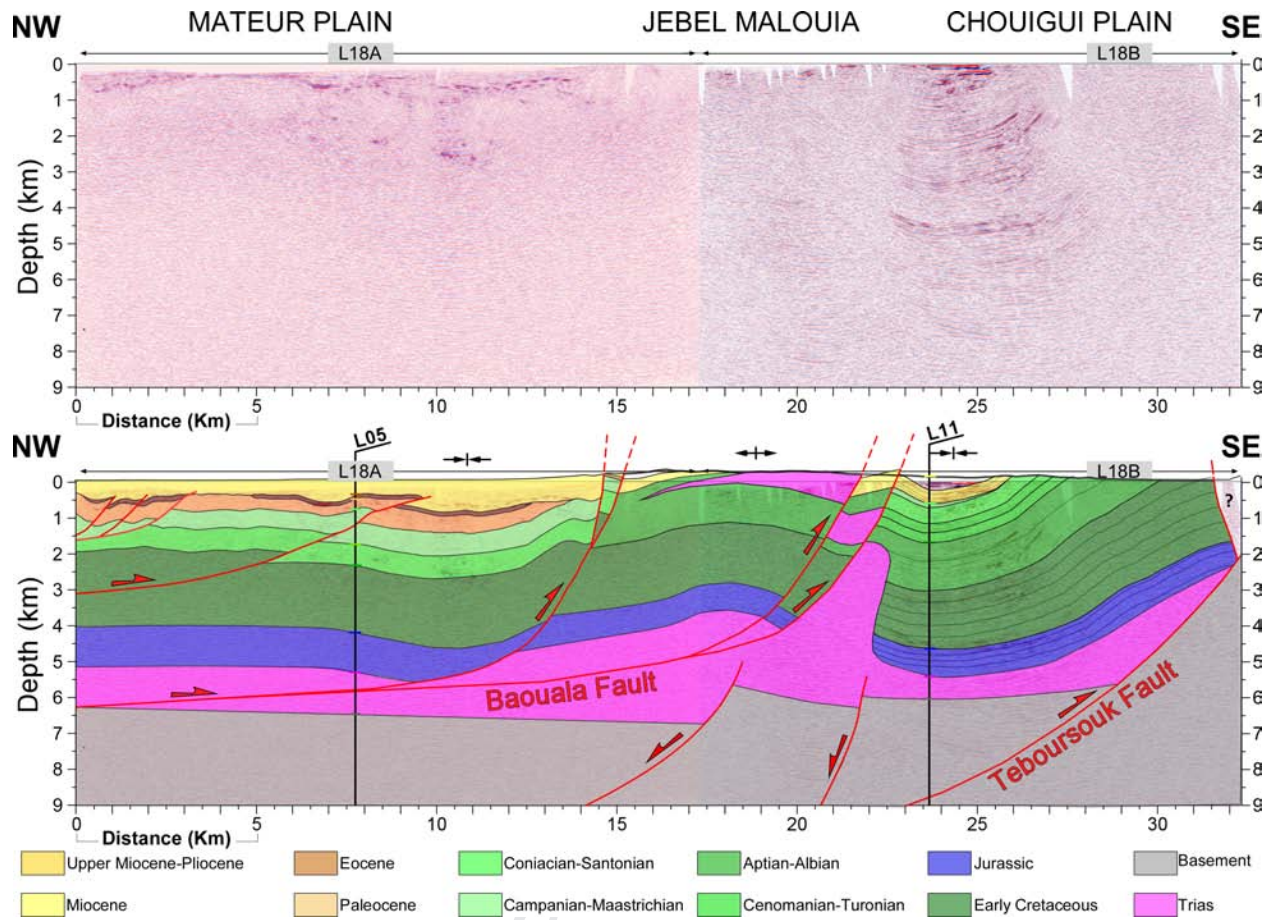


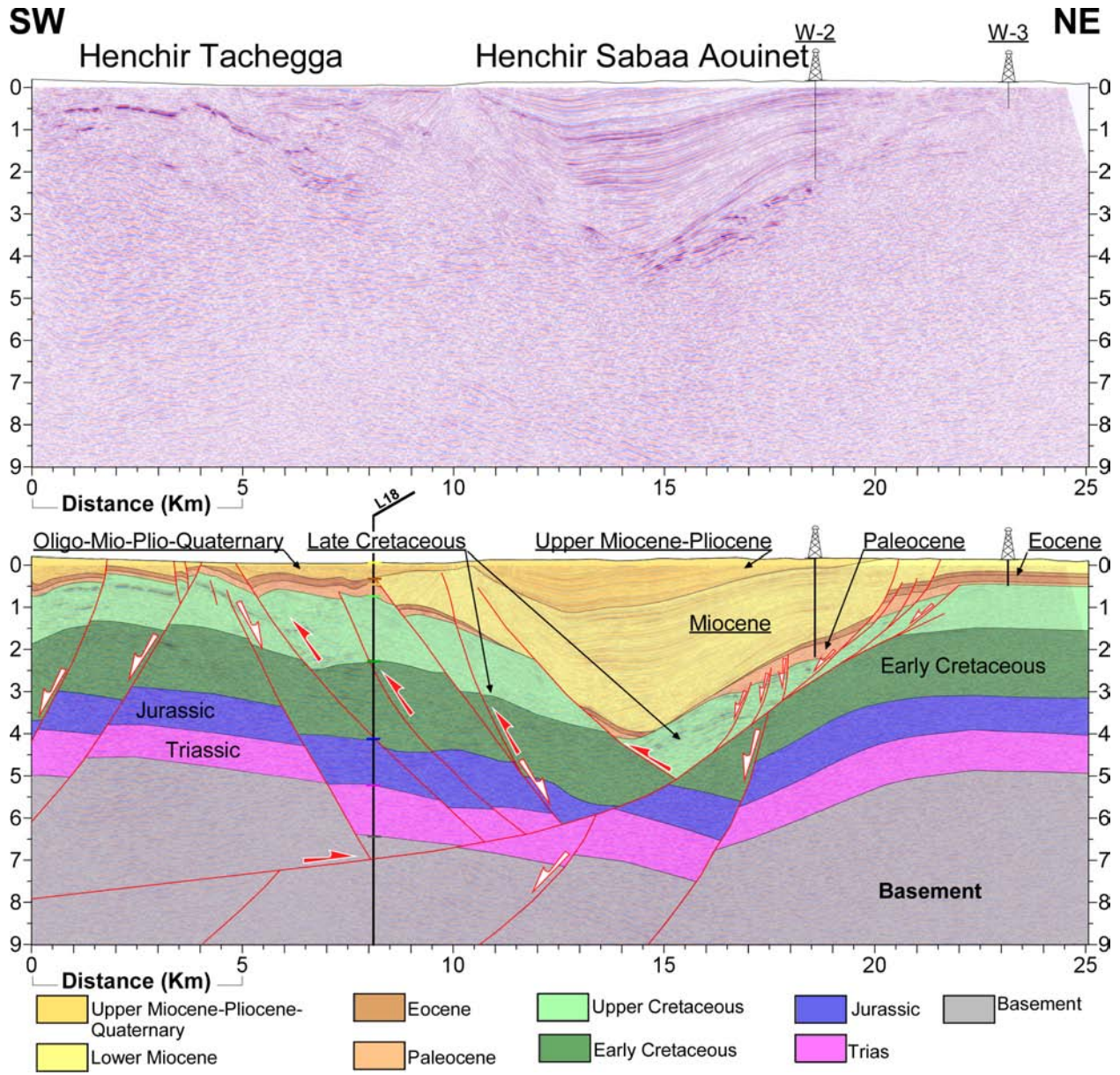












Highlights

Active Salt diapir developed during Tethyan Mesozoic extension in Northern Tunisia

Significance of allochthonous salt sheets spreading laterally from the diapir stem

During subsequent compressional stage diapir is squeezed and passively transported above major thrust

Journal Pre-proof