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1 **The Southern Atlas Front in Tunisia and its foreland basin: structural style and**

2 **regional-scale deformation**

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8 **Abstract:**

9 Surface and subsurface data together with balanced cross-sections are used to emphasize the structural
10 style and regional-scale deformation of the Southern Atlas Front in Tunisia and its foreland basin. This E-
11 trending shaped belt reflects successive tectonic events. Mixed thick- and thin-skinned structural style
12 characterizing this belt is defined by shallow thrusts rooted down into deep E-trending basement thrusts
13 with major decollement level located within Triassic salt. A NW-trending system is also recognized in the
14 Front and particularly responsible for the belt's curved shape, which probably acts as transfer areas
15 between the wide-ranging E-trending major structures. The intermediate areas are characterized by salt
16 tectonics probably active during Jurassic and Lower Cretaceous, in response to active synsedimentary
17 normal faults and differential sedimentation loading. The shortening amount varies and decreases from
18 north to south and from west to east: ~7.45% in the West, ~3.3% in west-central part, ~2% in central
19 area to insignificant (~1.4%) in the east. From north to south, the Northern Chotts Range narrow zone
20 accommodates overall the Cenozoic shortening with ~6.95% and less than 1% in the Chotts region. This
21 variation coincides with the pre-Cenozoic basin's architecture acquired during Mesozoic Tethyan
22 evolution, which generates the main Jurassic E-trending Chotts basin. Inside this basin, distributary sub-
23 basins formed in relation with NW-trending faults active during Cretaceous period. The Chotts basin
24 appears, therefore, as Mesozoic rift rollover structure well-preserved in the front of the belt. The
25 compilation of data suggests that the Front of the southeasternmost end of the Atlassic domain is
26 distinguished by the reactivation of inherited fault systems acting during alpine orogeny as major
27 southward thrusts. The foreland shows a structural singularity to be formed by a preserved almost
28 unmodified and well exposed large Mesozoic rollover structure.

29 **Keywords:** Southern Atlas Front; Tunisia; balanced cross-sections; tectonic inversion; structural
30 inheritance

31 **1. Introduction**

32 The evolution of sedimentary basins typically involves extensional deformation of both basement and
33 sedimentary cover. The initial rifting stage usually progress to continental passive margin, which is
34 frequently characterized by an irregular sea floor dominated by repetitive depocenters associated to
35 intra-basin active listric normal growth faulting. This is well observed in the present-day Atlantic-type
36 passive margin and is well preserved in several inverted margins such as the Tethyan margins around the
37 Mediterranean (present coordinates). These margins are characterized by associated sub-basins
38 produced by major fault systems themselves made-up by regional tectonic extension (Allen and Allen,
39 1990). This system is in control of basin geometry (graben and half graben) strongly attested by
40 important thickness variations of sedimentary sequences. The tectonic inversion is defined once the
41 aforementioned basins are deformed by subsequent compressional tectonic (Letouzey, 1990). During the
42 inversion of the margin, normal faults are partly or entirely inverted (Bally, 1983). The reactivation of
43 normal faults is commonly well observed for those with an opposite strike to the subsequent
44 compression direction. At the scale of structures, the reverse movement of the pre-existent extensional
45 systems, is typically derived by preexistent normal faults that bound graben or half graben (Bally, 1983).
46 In addition, the existence of salt layers in the sedimentary sequences characteristically made additional
47 complication to the inversion process. The presence of salt is also involved in the mode of inversion, with
48 classically occurrence of the preferential major decoupling along the salt layers (Roure et al., 1992). The
49 Atlas Mountains of North Africa are recently documented by many authors as one of worldwide belt
50 resulting from the basin inversion processes (e.g. Vially et al., 1994; Ziegler et al., 1995; Roure et al.,
51 2012; Frizon de Lamotte et al., 2013; Lepretre et al., 2018a, b).

52 The Southern Atlas of Tunisia forms the front of the North African Alpine orogenic belt, between the
53 deformed Atlassic domain to the North and the relatively undeformed Saharan platform area to the
54 South (Fig. 1). This region also bounds the easternmost segment of the Cenozoic deformation front in the
55 Maghreb region, which extends over 2000 km from High Atlas in Morocco to the West, the Algerian
56 Saharan Atlas in the center and the Tunisian Atlas to the East. From West to East, the southern
57 deformational front of the Maghrebides belt drastically changes with typical intra-continental chain in
58 the High Atlas of Morocco and a fold-and-thrust belt in the Southern Atlas of Tunisia (Teixell et al., 2003;
59 Benaouali-Mebarek et al., 2006; Missenard et al., 2007; Gharbi et al., 2015). The physiographic
60 characteristic of the Southern Atlassic domain in Tunisia is a series of narrow elongated massifs
61 separated by large basins (Figs. 2, 3). It shows a distinguished distribution of fold's geometry which led to

62 a varied structures and basin's interpretations by previous studies during the past 65 years. 1) 1950's to
63 early 1990's: strike-slip faulting was proposed, which produced the fold distribution and the style of the
64 Tunisian Southern Atlas (e.g. [Castany, 1954](#); [Zargouni and Ruhland, 1981](#); [Abdeljaouad and Zargouni](#)
65 [1981](#); [Abbes and Zargouni 1986](#); [Fakraoui 1990](#); [Ben Ayed, 1993](#); [Bedir et al., 2001](#)). Two major NW-
66 trending right lateral strike-slip fault systems (Gafsa and Negrine-Tozeur, [Fig. 2](#)) were documented as the
67 major ones, which are associated to left lateral NE-trending one recognized as less active faults systems.
68 The structures are identified, in this model, as curved, right-stepping *en echelon* shaped folds related to
69 the strike-slip fault systems. 2) 1990's late: few works acquaint with salt tectonics model ([Hlaiem et al.,](#)
70 [1997](#); [Hlaiem, 1999](#)). These works introduce an evolution in which a significant diapirs took place during
71 Early Jurassic period. This structural model locates the subsequent Tertiary folding as preferentially
72 nucleated on the preexistent diapir apex. 3) Fault-related folding 2000's model ([Outtani et al., 1995](#);
73 [Ahmadi et al., 2006](#)) was dominated by south verging thrust system on which folds are located. This
74 model interprets the anticline structures in this zone as developed by the result of ramp-related folding,
75 disturbing only the sedimentary cover (thin-skinned model) without basement involving. 4) 2010's:
76 structural reconsideration of this belt was based on the role of the pre-existent faults systems in the
77 subsequent tertiary deformation. This model highlights a structural evolution controlled by deep-seated
78 basement faults ([Riley et al., 2011](#); [Said et al.,2011b](#); [Gharbi et al., 2015](#); [El Amari et al., 2016](#)).

79 This paper focuses on the structural style and regional-scale deformation of the Southern Atlas Front in
80 Tunisia and its foreland basin as an inverted basin. To document these goals, surface data (geologic
81 mapping, cross-sections, markers of tectonics deformation) are analyzed. New seismic reflection profile's
82 interpretation, well data and long-transect sequences correlations are as well used to highlight the main
83 structure and/or borders of different domains. Three long balanced cross-sections across the southern
84 Atlasic belt are constructed to estimate the amount and the rate of shortening, as well as the variations
85 in term of structural architecture and tectonic style using methods described in [Dalstrom, 1969](#); [Coward,](#)
86 [1996](#); [Decelles and Giles, 1996](#).

87 **2. Geological setting and lithostratigraphic overview**

88 The Southern part of the Atlas Mountains of Tunisia is located in the southeastern part of the Maghreb
89 orogenic belt called the Maghrebides. The Maghrebides orogen is qualified as the North African Alpine
90 chain, which is shaped during Tertiary–present-day compressional events related to Africa–Eurasia
91 convergence setting. The timing of the deformation, as well as the main compressive events are now
92 well established after sound arguments advanced during the 1990^s–2000^s (e.g. [Beauchamp et al., 1999](#);

93 [Bosworth et al., 1999](#); [Frizon de Lamotte et al., 2000](#); [Guiraud et al., 2005](#); [Masrouhi et al., 2007](#); [Khomsi](#)
94 [et al., 2009](#); [Frizon de Lamotte et al., 2011](#); [Gharbi et al., 2014](#)). The convergence between African and
95 Eurasian plates is expressed firstly in Tunisia by the positive inversion of major Tethyan basin and now it
96 is mostly accepted as operated during Late Cretaceous ([Guiraud and Bosworth, 1997](#); [Benaouali-](#)
97 [Mebarek et al., 2006](#)). Even if in the southern Atlassic domain of Tunisia, the Late Cretaceous inversion
98 markers are not yet well-identified; this issue is well highlighted in central and northern Tunisia
99 ([Masrouhi et al., 2008](#); [Dhahri and Boukadi, 2010](#)). The first significant compressional event causative of
100 the Atlas building is the middle–late Eocene Atlassic compression ([Vila et al., 1993](#); [Frizon de Lamotte et](#)
101 [al., 2000](#)). This later is well underlined by local as well as regional unconformities on the overall Tunisian
102 Atlas domain ([Vila et al., 1975](#); [Khomsi et al., 2006](#); [Gharbi et al., 2015](#)). In southern Tunisian domain,
103 previous field data and regional geological mapping are consistent with this unconformity dating an
104 Eocene compressional event (e.g. [Slimane et al., 1991](#); [Regaya et al., 1991](#); [Riley et al., 2011](#)). This
105 compressional stage is followed by a period of tectonic quiescence during Oligocene and Lower Miocene
106 times. The thick Numidian siliciclastic series deposited in the Tell-Rif sub-basins in addition to the general
107 geodynamic setting (Miocene back-arc opening of the Algerian–Provencal basins and the rotation of
108 Corsica and Sardinia) led to a model in which the northern edge of this chain is interpreted as an
109 accretionary prism developed in the western Mediterranean as back-arc oceanic basin ([Vergés and](#)
110 [Sàbat, 1999](#); [Frizon de Lamotte et al., 2009](#); [Roure et al., 2012](#); [Van Hinsbergen et al., 2014](#), [Hamai et al.,](#)
111 [2015](#)). The Oligo-Miocene extensional setting is interpreted by several authors to be caused by the slab
112 retreat ([Jolivet and Faccenna, 2000](#); [Deverchere et al., 2005](#)). As summarized by several authors, events
113 of widespread deformation in the Southern Atlassic Front of Tunisia are thought to have occurred during
114 middle-Miocene to present-day times ([Zargouni, 1985](#); [Chihi, 1992](#); [Swezey, 1996](#); [Bouaziz et al., 2002](#);
115 [Gharbi et al., 2014](#)). All authors, having worked in this region, agree that this event is the major
116 compressional tectonic event building the southern Atlassic domain in Tunisia (e.g. [Ben Ayed, 1980](#);
117 [Zargouni, 1985](#); [Boukadi, 1989](#); [Dlala, 1992](#); [Chihi, 1992](#); [Zouari et al., 1990](#); [Gharbi et al., 2014](#); [Bahrouni](#)
118 [et al., 2014](#); [Soumaya et al., 2015](#)). This period is characterized in southern Atlassic domain by NW-
119 trending main regional compression direction during Miocene-Pliocene and N-Trending tectonic during
120 the late Pliocene-Quaternary ([Dlala and Hfaiedh, 1993](#); [Billi et al., 2011](#); [Gharbi et al., 2014](#); [Soumaya et](#)
121 [al., 2018](#)). The present structure of the Southern Atlassic Front in Tunisia comprises a series of elongated
122 structural edifices widely accepted today as deriving from the inversion of Mesozoic depocenters
123 recording the former passive margin basin and/or mini-basins along the south Tethyan margin (e.g. [Chihi](#)
124 [et al., 1992](#); [Dhahri and Boukadi, 2010](#); [Briki et al., 2018](#); [Frifita et al., 2019](#)).

125 The structural style of the southern Atlassic domain of Tunisia is dominated by E-W–trending structures.
126 In major cases, these structures are shaped by deep E-W–trending fault systems. These systems are
127 widely documented as inherited systems from the former south Tethyan margin (Gharbi et al., 2013;
128 Gabtni et al., 2013; Tanfous-Amri et al., 2017). From a geodynamic point of view, this system is well
129 correlated during Jurassic and Cretaceous periods with the Alpine Tethys evolution within the overall
130 context of the relative left movement of Africa with respect to Eurasia (Boughdiri et al., 2007; Frizon de
131 Lamotte et al., 2013). The breakup of Pangea has been expressed by successive extensional episodes that
132 generate some major basins from Morocco to Tunisia (Moragas et al., 2016, 2017; Naji et al., 2018a;
133 Jaillard et al., 2019). This period is started during the Late Permian and continued until Late Cretaceous.
134 The Initial Mesozoic basin geometry was characterized by Triassic basin subject to prevalent evaporitic
135 facies across overall Tunisia (Burolet, 1956; Kamoun et al., 2001). Triassic evaporitic sequences are
136 indicative of an epicontinental environment which was recognized in entirely Maghreb sub-basins
137 (e.g. Masrouhi and Koyi, 2012; Masrouhi et al., 2014a; Moragas et al., 2017). Triassic sequences are
138 usually associated to subordinate basalts, locally preserved in Tunisia (Burolet, 1973; Laaridhi-Ouazâa,
139 1994) and commonly viewed as products of rift-related magmatism (Masrouhi et al., 2013; Masrouhi et
140 al., 2014b). The Jurassic period is well documented as active rifting period overall the Maghreb area
141 (Bracene et al., 2002; Boughdiri et al., 2007; Saura et al., 2014; Martin-Martin, 2017; Moragas et al.,
142 2018). This stage is well preserved in south Tunisia by E-W–striking basins (such as the Chotts
143 basin; Patriat et al., 2003) produced by major E-trending fault systems themselves made-up by regional
144 N-trending tectonic extension. This system is locally in control of basin geometry (graben and half
145 graben) strongly attested by important thickness variations of sedimentary sequences (Ben Ferjani et al.,
146 1990; Bahrouni et al., 2016; Tanfous-Amri et al., 2017; this work). The N-S–extensional context prevailed
147 during Jurassic become a NE-SW regional tectonic extensional setting during Early Cretaceous (Louhaichi
148 and Tlig, 1993; Jaillard et al., 2017; Naji et al., 2018b). This second phase is mainly characterized by NW-
149 trending faults systems and associated sub-basins. This translation from Jurassic E-W (present-
150 coordinates) prevailing configuration to Early Cretaceous NW-SE shape is compatible with general E-W
151 transform faults of Atlantic which change to NW-SE ones in the Alpine Tethys (Rosenbum et al, 2002;
152 Frizon de Lamotte et al., 2011). In Southern Tunisia, the recent works establish a paleogeographic
153 partition of the continental platform, which is scattered in some localities by related volcanic and
154 halokinetic movement that are recognized overall Tunisian domain (Burolet, 1991; Mattoussi Kort et al.,
155 2009; Jaillard et al., 2017; Dhahri and Boukadi, 2017). The main synsedimentary normal faults form a
156 general tilted blocks basin’s geometry, for which the Aptian–Albian ages are documented to have the

157 maximum extensional related-structures period in Tunisia (Gharbi et al., 2013; Masrouhi et al., 2014a;
158 Naji et al., 2018a). The rapid active extension characterizing Early Cretaceous period seems to decrease
159 during late Cretaceous times. Basin shows a thick deposition of open marine shale and shallow
160 carbonates sealing the above-mentioned tilted basin's geometry.

161 The compilation of lithostratigraphic sequences data in the Southern Atlassic mountains of Tunisia,
162 presented in this work, is derived from petroleum wells and outcrops around the studied structures (Ben
163 Ismail, 1982; Ben Youssef and Peybernes, 1986). The Triassic series are thick of about 400-800 m showing
164 important thickness variations. In the Northern Chotts range, the Triassic series are mainly evaporitic
165 showing carbonate and siliciclastic beds. The Jurassic strata consist of limestones of the Nara Formation
166 (Soussi and Ben Ismail, 2000). Three members were recognized: lower, Middle and Upper Nara showing
167 similarly important thickness variations (250 m to 2500 m). On the Jurassic limestones, Lower Cretaceous
168 series show a siliciclastic, clay, dolomites and sandstones layers. The Lower Cretaceous shows in its
169 bottom 300 m thick clay and dolomites of the Meloussi Formation. This later is overlain by the Boudinar
170 Formation which is thick of about 200 m (Fig. 2).

171 The Bouhedma Formation consists of a repetitive mixed carbonate-clay-evaporite sequences and
172 considered to be late Hauterivian age. This Formation shows also an important thickness variation (from
173 400 m to 2000 m). The Bouhedma Formation is overlain by Sidi Aïch Formation composed mainly of shaly
174 sands and ferruginous sandstone. This Formation, thick of about 200 m, corresponds to fluvial
175 environment and dated as Barremian age deduced from its stratigraphic position. In the northern Chotts
176 range, the Sidi Aïch Formation is capped by the Orbata Formation dated as Aptian age (Ben Youssef et
177 al., 1985). This later is composed of three members i.e. the lower Orbata corresponds to a dolomitic bar
178 named in this range Barrani bar dated as Bedoulian age, the middle Orbata is a marly and sandy
179 carbonates and the upper member consists of a sandy unit. In the Chotts basin, the upper Aptian to the
180 lower and Middle Albian is characterized by hiatus. This discontinuity is due to non-deposition or to
181 subsequent erosion. The Zebbag Formation was subdivided into four members, the upper Albian is
182 formed by an alternation of marls and carbonates with *Knemiceras* named M'Razig member that is
183 overlain by the massive dolomitic bar of the Rhadouan member. On this member rest the marls,
184 carbonates and evaporites of Kerker member which is dated as Cenomanian age. The top of the Zebbag
185 Formation is composed of massive bar of dolomite and dolomitic limestones named Guettar. This bar is
186 dated as Turonian. The Aleg Formation, which is dated Turonian–Early Campanian, is characterized by
187 marls and limestones layers. The Berda Formation, dated as Late Campanian–Early Maastrichtian, is

188 essentially formed by two carbonate members, which are separated by a middle member consisting of
189 an alternation of marl and limestones (Fig. 2). Upper Cretaceous limestone sequences in the Chotts Basin
190 are overlain by Lower Paleocene, lay the marine green shale of El Haria Formation that is overlain by the
191 Eocene deposits which consist of the Metlaoui Formation limestones and the continental unit of
192 Bouloufa Formation. This later is formed by conglomerates, red clay, and lacustrine limestones. In the
193 southern Atlassic domain structures, the sequences are generally overlain unconformably by the Segui
194 Formation, which is interpreted as syntectonic growth strata acknowledged to be the product of the
195 erosion of the structures uplifted during Cenozoic compressional events.

196 **3. Surface data**

197 Recently the knowledge about the west Mediterranean geodynamic evolution is upgraded after several
198 structural and geodynamics wide-ranging studies (e.g. [Tricart et al., 1994](#); [Faccenna et al., 2004](#); [Lacombe
199 and Jolivet, 2005](#); [Vergés and Fernández, 2012](#), among others). The southern Tunisian Domain is rather
200 understudied, compared to Algeria and Morocco, in term of regional-scale deformation. During the last
201 decade, many studies carried out in the southern Atlassic front of Tunisia (e.g. [Ahmadi et al., 2006](#); [Riley
202 et al., 2011](#); [Said et al., 2011a](#); [Gharbi et al., 2013, 2015](#)). However, all these mentioned works were
203 focused along the northern part of this belt (Metlaoui, Gafsa, Sehib and Orbata structures, Fig. 2). The
204 southernmost segment (Northern Chotts range) and its relationships with the Saharan platform, together
205 with taking into consideration the preexistent basin structure are yet to be appraised. The Northern
206 Chotts Range (NCR) is a 165 Km-long and narrow 5 km-wide chain (Fig. 3). It constitutes the
207 southernmost domain showing folded structures well highlighted by their physiographic characteristic as
208 well as their remarkable and well-preserved closures. From West to East, this series of anticlines, with
209 distinguished curved axes, are closely associated with E- to NW- to WNW-trending fault systems. The
210 following sections address the structural analysis of the NCR, which is based on field observations across
211 the eastern and the central parts of the Range as well as the northern isolated anticlines (Berda and
212 Sehib), and the southern regions that is the Chotts basin together with the Southern Chotts Range (SCR).
213 Data used are derived from this study, the preexisting 1/100000 geologic map published by the national
214 official of Mines, Gharbi, (2013) and El Amari et al. (2016). This field work is controlled by systematic
215 transects (with dips reading, faulting and folding analysis, entities relationships...), surface geologic cross-
216 sections together with detailed mapping of the key-zones.

217 The Jebel Sehib structure, which extends in the South of Metlaoui Range, corresponds to an isolated
218 anticline structure in the wide plain named the Segui plain (Figs. 3, 4UU'). The surface analysis shows a

219 dissymmetric folding (Ahmadi et al., 2006) for which the forelimb shows 70°–dipping strata and the
220 backlimb displays gentle dipping strata of about 20° (Fig. 4TT'). This anticline illustrates a curved axis
221 which changes from N120 in the West (near the NW-trending fault cutting the structure), to N80 in the
222 middle and to N30 in the East (Fig. 3). This structure was for long time interpreted as the result of the
223 master strike-slip movement and genetically related to strike-slip tectonics (Zargouni, 1985; Bedir et al.,
224 2001; Zouari et al., 1991; Zouaghi et al., 2005, 2011). The elongated central part shows a constant and
225 long (compared to East and West closure) geometry which mean that the formation of the fold was not
226 complicated by the faults bordering the structure. The geometry of the fold shows a long gentle dipping
227 backlimb and a short high dipping forelimb which seems to be not controlled by the adjacent strike-slip
228 fault. The Sehib anticline structure is particularly well outlined by the Campanian limestones (Ahamdi et
229 al., 2006). At the front, this anticline shows local-scale bend folding, which is probably the surface
230 expression of deep thrust on which the fold is shaped (Fig. 4TT'). This anticline is characterized by south-
231 verging bending style of folding showing a clear kink in the southern limb with Cretaceous and Eocene
232 strata which suggests a north-dipping axial surface. The anticline is limited by two North and South flat
233 domain plains filled by thick Mio-Plio-Quaternary Segui Formation displaying characteristics of
234 syntectonic alluvial deposits exhibiting growth-strata pattern.

235 The Jebel Berda shows particularly unusual geometry showing an inverted northern back-limb and a long
236 normal southern forelimb (Fig. 4UU'). The second singularity of this fold is the notable entangled form
237 between an eastern part with an inverted northern limb which corresponds to the Ain El Beida, and a
238 classically south verging anticline in the western part which corresponds to the Jebel Atra. This structure
239 shows series extending from Cenomanian–Turonian sequences in the core to Mio-Plio-Quaternary
240 deposits in the limbs. The interlacing between the two parts of the structure (Ain El Beida and Atra) is
241 located within the Coniacian–Santonian marls sequences. The distinctive entangled form within
242 Coniacian–Santonian marls is associated to thickness change, which vary from 140 m in the West to 50 m
243 in the East. In the eastern part, the cross-section (Fig. 4UU') cuts this structure and illustrates the local
244 Berda style with north-verging folding. This anticline is associated to E-trending fault, which corresponds
245 to north-verging thrust. The hanging wall of Berda thrust is formed by north-verging anticline structure
246 showing a north limb with 65-70°–north-dipping Cretaceous strata and a southern limb with 45°south-
247 dipping strata. The footwall of the Berda thrust is formed by an overturned 75-80°–south-dipping layers.
248 In the western part, i.e. Jebel Atra, the structure corresponds to N095/10°W anticline corresponding to a
249 typical E–trending south–verging Atlassic structure, regarding the overall belt structure.

250 The central and the western part of the Northern Chotts Range is an E-trending narrow belt (Mahjoub
251 and Fakraoui, 1990). This belt is generally characterized in the Jebel Askar, Sif El Laham and Teferna-
252 Zitouna structures by two differentiated types of folding and/or style of structures between the northern
253 and the southern limbs (Figs. 3, 4VV'). All these above-mentioned structures show high-dipping northern
254 limbs with upper Cretaceous strata unconformably covered by Eocene and Miocene sequences. The
255 southern limbs are formed by gentle-dipping to horizontal Lower Cretaceous sequences (Fig. 5). The
256 southern side of all these structures are truncated by E-trending kilometric faults systems parallels to the
257 axis of major anticlines (Fig. 3). These E-trending faults show usually inverse movement within Aptian-
258 Albian sequences, in which the inverse throw rarely exceeds some tens of meters. This structural
259 singularity of long faults with restricted throw is usually associated to hectometric folding well-arranged
260 along faults, and which is observable over several kilometers (Fig. 4). In addition, along this E-trending
261 fault systems a disharmonic folding is frequently observed particularly in Aptian marls and carbonates
262 alternations. This fault systems (with restricted throw) vs folding is likely incompatible with the Tertiary
263 compressions, which is more developed in the nearby structures. Regarding the folding style, with south
264 gentle-dipping Early Cretaceous and North high-dipping Late Cretaceous, this south dipping fault system
265 (with unimportant displacement, Figs. 4VV', 5) seems to be an inherited system deeply rooted and
266 passively transported on north-dipping major thrust, which accommodate the Tertiary compressions.

267 Eastward, the Bir Oum Ali-Hachichina structure (BOAHS) comprises two major anticlines separated by
268 the Bir Oum Ali-Hachichina fault systems (BOAHFS). This area exposes, in term of deformation, two
269 different styles with a northern highly deformed anticline and a southern gentle deformed one (Fig.
270 4WW'). The BOAHS shows a curved-axis which changes from West to East (El Amari et al., 2016). As
271 shown on the geologic map (Fig. 3), the BOAHS have a NE-trending axis in the Jebel Bir Oum Ali, a NW-
272 trending one in the Jebel Hachichina and an ESE-trending axis in the Jebel Naimia. The field investigation
273 (Figs. 5), along the BOAS anticline, highlight that the southern part is formed by the Early Cretaceous
274 series showing a long southern limb and a short northern one characterized by ~10-15°S-dipping
275 southern forelimb and ~45-65° N-dipping northern backlimb (El Amari et al., 2016). While the southern
276 side is simple and shows a little deformed side of the BOAHS (Fig. 5), the northern part shows a complex
277 pattern. This part is characterized by the anticline's succession with curved hinge line and showing a right
278 stepping "en echelon" arrangement. These two different sides are separated by the BOAHFS and does
279 not show a syncline in-between (Figs. 4UU'). The curved secondary (in respect to general structure)
280 anticline succession is well observed along the central part of the NCR. Furthermore, from West to East
281 geometric complexity increase with usually south overturned limbs (Figs. 4, 5). This area offers a well

282 traced ~15-20°-angular unconformity between the aforementioned Cretaceous series and Miocene–
283 Pliocene sequences (Fig. 5). Eastward, this angular unconformity is localized between Paleocene–Eocene
284 and Miocene–Pliocene series along 20 km northern limbs of this area (Figs. 3).

285 Eastward of the BOAHS, the Jebel Beidha structure, is the second remarkable fold in South Tunisia
286 showing a northern inverted limb. The Jebel Beidha structure is an uncharacteristic sub-circular anticline
287 cored by Lower Cretaceous sequences (Fig. 3). This anticline shows a 25°-south-dipping overturned
288 northern backlimb and a 31°-south-dipping southern forelimb (Fig. 4XX'). The northern forelimb and
289 Beidha anticline's core are abruptly affected by the 70°-south-dipping thrust, well observed in the Oum
290 Aguel region. The Beidha structure exhibits along the core numerous ENE-trending normal faults,
291 probably related to the Tethyan period regarding their tendency versus the fold deformation (Figs. 6, 7).

292 Eastward, the Beidha structure is bordered by rhomboidal area named the Bou Loufa basin typical of
293 pull-apart basin shape (Zargouni, 1985; Abbes and Zargouni, 1986; Abbes et al., 1986; Fakraoui et al.,
294 1991). This structure is shaped by northwest-trending dextral faults comprising two en echelon
295 segments, which are the Jebel Stah – Chebket Bou Loufa and the Jebel El Haira segments (Fig. 3). These
296 segments delimit right-stepover fault zone defining releasing stepover. The northwestern corner of this
297 pull-apart basin developed at the end stage the formation of circular to elliptic salt intrusion;
298 documented in previous literature as the Hadifa diapir (Abbes and Zargouni, 1986).

299 The eastern part of the NCR is an E-trending structure (Haidoudi structure, Fig. 3), which change at the
300 easternmost segment (Zemlet el Beidha, Fig. 3) to NE-trending anticline. All this part is called the Zemlet
301 el Beidha-Jebel Haidoudi structure (ZBHS). This structure, of about 24 km long, have a curve-shaped
302 geometry with two main anticlines separated by the NW-trending Fejej fault (Fig. 3). The Zemlet el
303 Beidha is a south-to-south-east verging asymmetric anticline with only an east closure and shows, from
304 West to East, a curved axis that changes from E-strike to NE-strike (Fig. 3). This 16 Km long and 3-to -5 km
305 wide structure, is located between the Menzel Habib plain to the North and the Chotts Fejej and Draa
306 Oudhref respectively Southwest and Southeast. The Zemlet el Beidha anticline is respectively cored by
307 the Hauterivian–Barremian in the central part, the Albian in the Jebel Meida and the Coniacian–
308 Santonian in the Jebel Romana and Gouada plain (Fig. 3). The north-eastern segment of the ZBHS is
309 formed by gentle deformed Cretaceous sequences covered unconformably by Late Miocene–Pliocene to
310 Quaternary series. The surface geometry of the ZBHS shows a ~20° N-dipping backlimb made of
311 Coniacian–Santonian green marls alternated with bioclastic limestones unconformably overlain by
312 Miocene continental sandstones (Fig. 4ZZ'). The forelimb is formed by the Aleg 45° S-dipping Formation,

313 itself unconformably overlain by the Segui late Plio-Quaternary Formation. Eastward, of the Jebel Tebaga
314 Fatnassa a disturbance of dips is recorded in the limestones of the Aleg Formation possibly associated
315 with N-dipping reverse faulting (Fig. 4XX'). Westward at the central part of the ZBHS, numerous N100–
316 110°E apparent trending strike-slip faults, along Khanguet Aïcha and Khanguet Amor, regions affect the
317 45° S-dipping limb. Along the 9 km-long NNW–SSE surface cross section, across Jebel Es Smaïa at the
318 North and the Jebel Haidoudi anticline at the South (Figs. 4YY'), the structure shows an asymmetric fold
319 with a core occupied by Coniacian dolomites showing a ~15 to ~20° N-dipping backlimb and ~40° S-
320 dipping southern forelimb. In this southern limb, the Late Cretaceous strata are unconformably overlain
321 by Plio-Quaternary continental deposits. The Northern limb is affected by ENE-trending reverse fault that
322 separates the Jebel Haidoudi anticline from the Oued Krenafess syncline in the south. This later syncline
323 forms the hanging-wall of E-trending fault systems which likely correspond to the inversion of E-striking
324 inherited system that delimits the Jebel Es Smaïa monocline at the north. In the southern limb of Jebel
325 Haidoudi anticline, a main ~45 to 70°–NE-dipping normal faults are observed, most of them are a
326 preserved synsedimentary sealed normal faults. This normal faulting is usually sealed by the Santonian
327 marls (Figs. 6, 7) attesting an ante-Santonian synsedimentary activity. The Coniacian dolomitic sequence
328 thickness increases toward the NE hanging-wall and are characteristic of synsedimentary faults activity.
329 In previous works, the numerical analysis of data (Gharbi et al., 2013), consisting of bedding rotation data
330 to restore the horizontal bedding planes, highlights a ~N30 to 65°–E-trending tectonic extension that
331 prevailed during the Coniacian–Santonian times.

332 The southern Chotts Range is an E-W striking physiographic chain, which extends over than 100 Km from
333 Kebili in the West into Jebel Aziza in the El Hamma in the East (Ghanmi and Potfaj, 1987; Ghanmi et al.,
334 1988). Two distinguished crests are formed by the lower Albian and Turonian dolomitic bars which are
335 separated by marls intercalation. These two crests are well parted in the western part by the
336 Cenomanian sequences. Both crests get closer in eastern part and becoming one on another
337 corresponding to complete disappearance of the Cenomanian intercalations (Ghanmi and Potfaj, 1991).
338 This chain was interpreted previously as the southern limb of a mega fold structure called the Chotts
339 mega-anticline. Along this chain, the layer's dips collected during field work are usually between 5° and
340 10°, and never exceeds 15°, which cannot be taken necessarily as related to tectonic compressions.

341 In the end, the main fault systems controlling and associated to the folding, are of two types: 1) E-
342 trending north-dipping faults act as major southward thrust systems and 2) E-trending south-dipping
343 faults act as back-thrust likely connected down on the major north-dipping ones. These facts are

344 demonstrated in some structures by their preserved character as former normal faults deeply rooted
345 (will be developed in the next seismic section) and passively transported on the north-dipping major
346 ones.

347 **4. Subsurface data**

348 Eleven oil wells and nine seismic profiles are used in this works to describe and define the deep
349 configurations of key-zones in the objective to connect and extend geological data in depth.

350 **4.1. Correlation**

351 Three correlations (C1, C2 and C3) between the sedimentary sequences of the study area and the nearby
352 structures (Southern Central Tunisia in the North, and the Saharan Platform in the South) is used to
353 shows lateral variations of thickness, facies and gaps (Figs. 8-10). The correlation is given here without
354 major faults control with a primary objective to describe them in absence of structural adjustment, which
355 will be integrated in the next section. For confidentiality reasons, in this work, the wells are numbered
356 independently from their original names.

357 In the Saharan platform, the Triassic series are of typical Germanic facies composed of the three series
358 (Soussi et al., 2017); (i) the basal shale-sandy unit (TAGI: Trias argilo-gréseux inférieur). The TAGI is
359 interpreted as having been deposited in a fluvial deltaic environment, the TAGI shows habitually lateral
360 injected volcanic sills (Ben Ismail, 1991), (ii) the intermediate marine carbonate unit is characterized by
361 dolomites of the Messaoudi and Rehache Formations and (iii) the evaporitic unit of the Adjadj
362 (Messaoudi and Mhira Evaporites) which show lagoonal deposits (Busson, 1970; Ben Ferjani et al., 1990;
363 Bouaziz et al., 1999; Kammoun et al., 2001). The Triassic series thickness varies from 0 m (absent) in the
364 P8 well to 500 m in the P8 and P9 wells. Towards the North, the subsurface data show a thick Triassic
365 series, exceeding the 1500 m (Tanfous Amri et al., 2005, 2017). However, there are no wells that crossed
366 the total Triassic sequences at the Southern Tunisian Atlas. At the P6 well (Fig. 8), the 1700 m-thick
367 crossed sequences show a chaotic mixture facies of gypsum, clay, silt and anhydrite deposit with a few
368 thin intercalations of dolomite (Buroillet, 1973; Ben Ferjani et al., 1990).

369 In the study area, many petroleum wells cross the totality of the Jurassic series. Its maximum thickness is
370 found in the Chott Fejej where they reach a total thickness of 2500 m at the P2 well and passes even to
371 2650 m in nearby wells (Fig. 9). The thickness of Jurassic progressively decreases, and the facies becomes
372 evaporitic and detrital, and reach 1100 m in the P12 and P10 wells in the far south and south-west part
373 of the Chotts basin. Toward the North, the thickness of the Jurassic sediment does not exceed 500 m in

374 the P6 well and passes to 600 m at the P14 (Fig. 10) well where the Jurassic series represent the typical
375 Nara Formation with three carbonate members indicating deposition in moderate deep marine
376 environments of an external platform in southern and central Tunisian Atlas (Tlig, and Steinberg, 1980;
377 Soussi, 2002).

378 In the Chott Fejej, the Melloussi, Boudinar, Bouhedma and Sidi Aich Formations, are thick of about 3000
379 m. Towards the North, the thickness decreases to reach 1100 m in the P6 and P14 wells (Figs. 9, 10).
380 Laterally, in the Saharan Platform, the Tithonian-Barremian deposit passes to sandy facies with
381 intercalations of clays, carbonates and evaporites named the "Wealdian facies" or Asfer Formation (Tlig,
382 2015). This Formation shows also important thickness variation ranging from 50 m in the P10 well to 600
383 m in the P11 well. From Aptian to Maastrichtian, the deposit is mainly made up of carbonates, clays and
384 rare evaporitic series, outcropping widely in the two sides of the Fejej structure (Marie et al., 1984;
385 Fakraoui, 1990; Fakraoui et al., 1991; Abdallah et al., 2000). The Aptian-Maastrichtian deposits typifies a
386 transgressive cycle, showing several variations of thickness, facies and gaps. The Aptian–Late Turonian
387 series are exposed along all the Southern Chotts range. The thickness changes from East to West
388 showing a clear reduction towards the East (Ghanmi and Potfaj, 1991) and important thickness variations
389 from South to North (Gharbi et al., 2015; Amri Tanfous et al., 2017; this work). These sequences have a
390 170 m–thickness in the eastern part of the SCR, in which the Turonian Guettar bar covers, via
391 Cenomanian conglomeratic horizon, the Albian alternation. Towards the West, the maximum thickness
392 of the Orbata and Zebbag Formations is recognized in the Fom El Argoub section (550 m). This E-W
393 variation, is similarly observed along the NCR. In the eastern part (Zemlet el Beida), the Coniacian–
394 Turonian Aleg Formation unconformably overlies the Aptian Barrani bar. Westward, at the Jebel Naimia,
395 the thickness of the Aptian–Turonian series is about 95 m for the Orbata Formation and 215 m for the
396 Zebbag Formation. Westward, At the khanguet Besbessa, the Orbata Formation shows a thickness of 120
397 m and the Zebbag Formation is about 390 m (Abdallah et al., 1995; Gharbi et al., 2013; El Amari et al.,
398 2016). Toward the western part of the Range, the maximum thickness measured in the Khanguet El
399 Asker is about 130 m for the Aptian–Orbata Formation and of 650 m for the Conomanian–Turonian
400 Zebbag Formation. In subsurface, the P13 well crosses 1000 m of the Aptian-late Turonian series.
401 Towards the South and North, the thickness of the Aptian-late Turonian series changes. It is about 550 m
402 in the Saharan Platform (500 m in the P10, 550 m in P12 and 630 m in P3 wells) and about 870 m in the
403 southern Central Tunisia (855 m in P6 and 885 m in the P14 wells).

404 In southern Atlassic domain of Tunisia, the Coniacian–Santonian Aleg Formation shows an important
405 thickness and facies variations in both N-S and E-W striking. Along the SCR, this Formation is a 100 m–
406 thick well exposed on the southern part of the Tebaga of Kebili mountains. Far South, the petroleum
407 wells crossed 600 m of the Aleg formation. It is formed by two units, evaporitic with a lagoonal character
408 at the base and carbonate at the top. Toward the North of the Chotts basin, the Aleg Formation is
409 developed on the northern limbs of the NCR structures and even in two limbs of Jebel Berda. From East
410 to West the variations are clear with 25 m at the Jebel Naimia, 130 m in Khanguet Besbessa (Abdallah,
411 1989; El Amari et al., 2016). Far west, the Aleg Formation have a thickness of 260 m at the Khanguet El
412 Asker, 300 m at the Jebel Morra and reduced to 135 m at the Jebel Sidi Bou Hellel. At the outcrop, the
413 thickness of the Aleg formation reached its maximum at Jebel Berda structure (450 m). Northward along
414 the northern structures, the same variation is further accentuated with 450 m Jebel Berda and 1000 m in
415 the P13 well (Fig. 10). The upper part of Cretaceous Formations shows the same thickness variations,
416 although its rate of difference has been considerably reduced. The Campanian – Maastrichtian Berda
417 Formation shows respectively 200 m in the Oued Batoum, 330 m in the Jebel Kherfane and 420 m at the
418 P13 well.

419 **4.2. Seismic data**

420 Seismic interpretation is carried out in a structural goal. Nine seismic profiles are used and interpreted,
421 for which some of them are composite ones. Seismic profile data are used to join the fault systems
422 and/or tectonics block with the above-described lateral variations of thickness and facies of the
423 sedimentary sequences. These Seismic profiles were correlated by many oil wells (Figs. 8-10 for more
424 details) as well as the stratigraphic sequences established by field work. In the study area, three main
425 parts are studied (northern, central and southern) based on the seismic data availability and structural
426 style. We chose individual nine main stratigraphic limits (LS) that we considered to be important for the
427 comprehension of the sedimentary history in addition to tectonics configuration. These three-selected
428 areas well covered by the nine selected seismic profiles and well calibrated in north by P5, P6, P13 and
429 P14; while in the central part by P1, P2, P4 and P7 and in the south P3, P8, P9, P10, P11 and P12 (Fig. 3).
430 For confidentiality reasons, the seismic profiles are labelled as LSx independently from their original
431 names.

432 The seismic section LS1 crossing the Chott Jerid area from SW to NE illustrates well the southwestern
433 part of the Chotts basin (Fig. 11). This section proves a thickness increase of Jurassic and Lower
434 Cretaceous sequences toward the middle part of the Chotts basin. Straight fault-plane reflections and

435 lithostratigraphic cutoffs constrains the location of high angle north–dipping normal fault systems. The
436 hanging-wall of this faults is characterized by thick Jurassic (1000 m) and Lower Cretaceous (2000 m in P7
437 well) sequences. This thick block is probably deposited in deep area with high subsidence in response to
438 the E–trending faults activity. In addition, this major E–trending fault seems to be sealed by the
439 Cenomanian–Turonian Zebbag Formation. The northern side of this section shows also some normal
440 faults that seem to be intra-basin growth normal faults. This section reproduces a good image of Lower
441 Cretaceous beds highlighted by interference of irregular reflectors, which is probably the expression of
442 growth strata pattern. A second section LS2, crossed by LS1 seismic profile, confirms this geometry of
443 steep north-dipping normal faults associated to the growth of the Lower Cretaceous sequences (Fig. 12).
444 These two seismic profiles (calibrated by nearby wells) argues for a basin showing in its southern side
445 north-dipping normal faults responsible of the creation of deep zone during Lower Cretaceous in what is
446 now the Chotts basin.

447 The lines LS3a and LS3b describe the eastern central part of basin (Figs. 13, 14). The central (north south
448 speaking) part of the Chotts basin, is constantly recording the initial Mesozoic extension. Along the LS3a
449 section, the top Jurassic is highlighted by well-identified reflectors, which become discontinuous toward
450 the southeastern part possibly related to the change of facies (Hlaim, 1999). These reflectors are
451 topped by irregular reflector of the Lower Cretaceous. This section shows an important normal faults
452 system connecting the north dipping Chotts basin as revealed by the above-mentioned seismic profiles
453 (Fig. 13). The key information given by this section is the increase to the Jurassic series thickness toward
454 the East. This give evidences that the Chotts basin, with major E-W striking governed by E–trending
455 normal fault systems, shows also a lateral thickness variation from West to East parallel to the basin
456 strike and confirm an eastward deepening.

457 This section (Fig. 13) shows that eastern part of NCR is marked by active salt tectonics. The northern side
458 on this section exposes a remarkable thinning of Jurassic and Lower Cretaceous toward an area with
459 internal confuse reflection patterns. This configuration is characteristically of structure related salt rise as
460 seen in the flanks of diapiric structures in many Atlassic domains (Hlaim, 1999; Masrouhi et al., 2014a;
461 Moragas et al., 2016, 2017). On this section, salt tectonics is probably controlled by Mesozoic normal
462 faulting that is associated with rim synclines (salt mini-basins?) well developed in Jurassic thick
463 sequences around the main salt body (Fig. 13).

464 Northward, the deep structure of the NCR is well illustrated by the NNW–SSE trending seismic reflection
465 line LS3b (Fig. 14). Reflectors have been calibrated using projected data from P2 and P5 wells. To the

466 South, the Chott Fejej basin is formed above a normal movement of the fault system which delimited a
467 deep graben beneath and associated with active subsidence during Jurassic and Early Cretaceous. To the
468 North, a similar extensional structure created by listric normal faulting delimits a North-facing tilted
469 blocks geometry beneath the Jebel Es Smaïa (Fig. 14).

470 Two long composite seismic profiles are derived along the Southern Atlas of Tunisia, i.e. the LS4a,
471 approximately 100 km and the LS4b about 65 km-long (Figs. 15, 16). The seismic section LS4a crossing
472 the SCR preserved subsurface structures of the Chott Fejej and the northern side of the Telemzane arch
473 or the Telemzane High (Gabtni et al., 2012; Gabtni et al., 2013; Dhaoui et al., 2014). The southwestern
474 part of the seismic section shows a considerable thinning of the Mesozoic series above the faulted
475 Paleozoic basement? (Fig. 15). The faulted and probably folded Lower Paleozoic series form the anticline
476 structure of Telemzane E-trending arch. These series are largely eroded and covered with an angular
477 unconformity by the series of Permian-Carboniferous. The stratigraphic limit sited between the
478 Telemzane high lower Paleozoic series and the Permian-Carboniferous series corresponds to the
479 discontinuity known as Hercynian as suggested by the presence of Onlap figure in the seismic line LS4a.
480 In the southwestern part of the study area, the P12 well shows that the upper part of the Paleozoic
481 series (Carboniferous and Permian?) is absent due to the major uplift of Telemzane Arch during the
482 Hercynian orogeny (Busson, 1967; Underdown and Redfern, 2007). The Permian-Carboniferous series
483 showing chaotic seismic facies become increasingly thick towards the Chotts basin (Fig. 15). Eastward,
484 the LS4b seismic section shows the remarkable geometry of the Hercynian unconformity between the
485 Paleozoic and the Mesozoic sequences (Fig. 16). This unconformity correlates well with which was
486 described far West in the Libyan domain (Boote et al., 1998). This section displays a remarkable angular
487 unconformity configuration with successive toplap figures on the Paleozoic basement showing a
488 northward slope varying from 2° to 4°. The toplap figure, the zones of truncated reflections and the
489 correlation with nearby wells indicate successive unconformities. The succeeding unconformities are
490 successively observed at the base of Triassic series, at the base of Jurassic and then at the base of Lower
491 Cretaceous series filling all the basin (Fig. 16). The Early Triassic series are represented by continuous
492 reflections and high amplitude can be interpreted as the continental sandstones of the Ouled Chebbi and
493 Kirchaou Formations (Ben Ferjani et al., 1990). This Formation is overlain by chaotic seismic facies that
494 corresponds to the Late Triassic evaporites. The Triassic series are affected by a set of an E-trending
495 parallel normal faults dipping to the North defining a general system of half-graben containing smaller
496 grabens. The major fault located in southern side of the Chott Fejej developing a structure of a tilted
497 block.

498 The seismic reflection profile LS5, with Approximately E–W strike, is located in the Metlaoui plain
499 between Sehib and Metlaoui mountains (Fig. 17). This line is presented here to discuss the NW–trending
500 fault systems previously considered as the major faults controlled the formation and the growth of folds
501 in South Tunisia (Zargouni and Tremoliere, 1981; Boukadi and Zargouni, 1991; Hlaïem, 1999; Bedir et al.,
502 2001). On the basis of well continuous sub-horizontal reflectors on the both sides of the section and
503 slightly deformed in the central, together with calibration by the P14 well, this section shows a series of
504 NE-dipping steep normal faults characterizing different tectonics blocks with inconstant thick series.
505 These faults are especially active during lower cretaceous and Coniacian–Santonian showing a clear
506 variation of thickness. These data are consistent with an earlier hypothesis of inherited Mesozoic fault
507 systems demonstrated by field data (see section 3). The reactivation of this fault system is clear in the
508 fault located in the western part of this section, which show reverse movement. The LS6 seismic profile is
509 also used to highlight at depth the Sehib structure. This profile shows in the forelimb a well identified
510 reflector’s truncated zone stating a thrust fault on which the Sehib fold is generated (Fig. 18). This
511 section is well calibrated by the P13 well drilled in northern back-limb. This section is an individual one in
512 the southern Tunisia that does not show any inherited closely controlling the subsequent compressional
513 structure. The Sehib anticline seems to be a fault-related-fold developed during tertiary compressional
514 events. However, the parallel near seismic line LS7 crossing the structure in the eastern part of the Sehib
515 anticline shows that the northern part of the basin is formed above an inverted normal fault which
516 delimits a deep graben beneath Sehib and associated with rapid subsidence during Early Cretaceous (Fig.
517 19). This unusual interconnection can be interpreted within two different scenarios: the eastern part of
518 the Sehib anticline is separated from the main structure by NW–trending fault and forms in this case an
519 isolated part like the Berda structures (Fig. 19). The second hypothesis is that the fault is formed above
520 inherited systems, but in this case the LS6 section does not permit to highlight it. This last option can be
521 sound if we know that the LS6 is not perpendicular to the structure and then to the deep inherited
522 system if it exists.

523 As a summary of this section, the compilation of data coming from surface as well as the seismic and well
524 data highlight a strong variation from North to South and from West to East, which coincide with the pre-
525 Cenozoic shortening basin’s architecture acquired during successive Mesozoic Tethyan extensional
526 periods (Fig. 20). Data highlight a former Permian? to Early Cretaceous basin with multipart geometries
527 that varies from East to West and from North to South. The Fig. 20 summarize a basin essentially
528 controlled by earlier Mesozoic E- and NW–trending fault systems themselves related to the extensional
529 period of the southern Tethyan margin. The E–trending faults are expressively active during Jurassic

530 controlling thick sequences especially in the eastern part of the basin (see [Fig. 20](#) for more details). The
531 NW-trending faults are expressively active during Early Cretaceous showing a variation in the rate of
532 sedimentation which become thicker in the central and western part of the basin. The E-trending fault
533 systems, which generate a dominant E-W-striking basin (Chotts basin), which is bounded by two high
534 Paleozoic zones: the Telemzame arch and Northern Chotts Range. Inside this main basin, distributary
535 sub-basins established probably in conjunction with NW-trending faults. These results seem to be well
536 integrated in the general Tethyan context ([Frizon de Lamotte et al., 2013](#)), which have demonstrate that
537 the transform faults are E-W in the Atlantic and change to the NW-SE in the Alpine Tethys ([Rosenbaum,](#)
538 [2002](#)).

539 **5. Balanced cross-sections and structural style**

540 To illustrate the structural style and the regional-scale deformation as well as the evolution of the
541 Southern Atlasic Front of Tunisia and its foreland, three long balanced cross-sections are constructed.
542 Balanced cross-sections, further detailed below, are based on field extensive work, mapping of
543 concerned areas, seismic reflection profiles across South Tunisia calibrated by detailed lithostratigraphic
544 data from outcrops and petroleum wells. During processing, surface data are digitized from ASTER and
545 DGEM, in which sections are built basically perpendicular to structures' axes to minimize the out of plane
546 transport. Restoration and balancing of cross-sections are constructed referring the usual thrust
547 tectonics concepts ([Suppe, 1983](#); [Suppe and Medwedeff, 1990](#); [Mitra, 2002](#); [Shaw et al., 2005](#); among
548 others). Balancing and restoration were performed using the "Move 2017" structural modeling software
549 (Midland Valley Inc.). For the cover, the thrust tectonics concepts fundamentally used in modeling are
550 bed-length, thickness conservation and flexural slip algorithm ([Ramsay and Huber, 1987](#)). For the deep
551 thrust tectonics' concepts and basement modeling, the restoration was based on the regional
552 lithostratigraphic data ([Butler, 1989](#); [Butler et al., 2006](#)), well calibrated by petroleum wells.

553 As in the orogens worldwide, it was difficult to acknowledge the role of the basement tectonics control in
554 the far Southern Atlas front of Tunisia in previous works. Currently, the outers part of orogens is
555 recognized also to be controlled by thick-skinned deformation (e.g. [Mouthereau et al., 2007](#)). The
556 basement-involved shortening as well as the tectonic style in the subsequent cross-sections, are
557 constructed based on methods and/or natural examples recently published ([Mouthereau et al., 2012](#),
558 [Lacombe and Bellahsen, 2016](#)).

559 The NNW-SSE constructed [S1] section was made from the Sehib structure in the North to the Saharan
560 platform in the South ([Fig. 21](#)). This 110 km-long section is pinned in the Chebket el Mtarkeb plain in the

561 North and the southwestern Kebili plain in the South. This section highlights at least two different zones
562 with northern deformed side and a southern part not as much of deformation compared to northern
563 one. The northern portion shows tightly juxtaposed folds with two observable south-verging anticlines
564 separated by syncline hidden under thin quaternary cover. The Jebel Sehib and the Jebel Taferna Zitouna
565 anticlines connect down on large north-dipping faults themselves connected on the Triassic evaporites
566 (transition basement-sedimentary cover?). Between these anticlines, the syncline is connected to depth
567 on short-cut, possibly in sedimentary cover, itself connected to the north-dipping fault of the Jebel Sehib
568 (Fig. 21). Considering all the above-presented data and taken in account the regional depth of
569 dispositional accommodation zone and rate (Marshak and Wilkerson, 1992), the southern thrust -named
570 in this work the Taferna thrust- involves possibly a high inherited basement structure beneath the
571 Taferna–Zitouna. This thrust appears to be connected down on a major basement normal fault active
572 during Mesozoic Tethyan rifting and controlled deep northern depositional area (regarding Taferna-
573 Zitouna series) at least until lower Cretaceous times. Approximately the southern two-thirds of the
574 section correspond to the Chott El Jerid basin and the northern part of the Saharan platform (Fig. 21).
575 The Chott El Jerid basin, the Jebel Tebaga Kebili and the Saharan Platform structures are closely
576 connected down on major north-dipping faults. From dense and abundant data (seismic reflection lines
577 and petroleum wells), these structures are intended to connect three major normal faults controlling a
578 thick Mesozoic sedimentary basin beneath this zone. The calculated data show that basement is at about
579 a depth of 2 km (below sea level) in the Saharan platform, 5 km beneath the Chott El Jerid basin and
580 about 3 km at Taferna–Zitouna (Fig. 21). Data are consistent with the hypothesis that the Taferna–
581 Zitouna could be located on inherited basement high and forms the boundary between two different
582 sedimentary depositional area. The present construction shows a model in which the major faults are
583 preserved as normal and connected down onto a major sub-horizontal detachment at a depth of ~5 km
584 beneath the southern part. The present model is also adequate when further combined with large
585 geometrical sequences attitude. The hanging-wall, of the Chotts basins faults, reflects a pleasant
586 reproduction of gently kinked rollover anticlines widely known in passive margin context (McClay, 1989;
587 Brun and Nalpas, 1996). The hanging-wall rollover geometries were described in literature as closely
588 controlled by the dips of the bounding faults (condition fulfilled in Chotts region) and amount of
589 extension. In the southern Atlassic Front, a recent study of Gharbi et al. (2015), was estimated a rate of
590 extension of about 6.5 km in Chotts region during rift stage from Permian? to Early Cretaceous. The
591 conservation of layers length -based on the absence of observation of internal layers deformation-
592 excepting the evaporitic Triassic series, reveals an estimation of shortening of about 7.6 km (~7.45%)

593 with 2.8 km (~2.7%) as deep shortening and 4.8 km (~4.7%) as shallow shortening. The shortening is
594 widely dissimilar from a deformed northerly zone with 7.1 km shortening amount (~6.95%) and southerly
595 almost undeformed zone with 0.5 km shortening amount (~0.5%). The Paleozoic high, defined in this
596 study as delimited by major basement faults, had the role of an accommodation zone between the
597 northern region with thick and thin-skinned tectonic style and the southern region showing preserved
598 inherited extensional structures of the southern Tethyan margin (Fig. 21).

599 A second 107-km-long balanced cross-section [S2] is constructed from northwest to southeast crossing
600 the following structures: Jebel Berda, Jebel Sif El Laham, Jebel Barrani, the Chotts basin, Jebel Tebaga
601 Fatnassa and the Saharan Platform (Fig. 22). This NW-SE section is pinned in the Berda plain in the North
602 and the southwestern El Hamma plain in the South. Similarly, this section shows two structurally distinct
603 styles from North to South. The northern part, which we considered all the region located in the north of
604 the NCR, corresponds to narrow anticline structures forming an E-W-striking major curved narrow range
605 (Fig. 3). This part shows that shortening in the basement and cover is accommodated by different
606 structural organization. Two distinguished anticlines, regarding the general style of NCR folds, bounded
607 by south-dipping north-verging back-thrusts with moderate offset. The Jebel Berda as well as the Jebel
608 Sif El Laham are plausibly up-lifted by these north-verging back-thrust systems. Back-thrust connects
609 down on large thrust ones located between Basement and sedimentary cover and linked in the Triassic
610 evaporitic layers at a depth of about 6 km (below sea level). This section records, from West to East, a
611 net variation in depth of basement, which was computed in the last section at 3 km (below sea level).
612 This geometry of shallow back-thrust connected down on major thrust systems shows a down-branches
613 creating a basement short-cut beneath the Jebel Sif El Laham. We interpret this zone as Paleozoic high
614 accentuated during Mesozoic extension which creating a horst structure beneath the NCR. During the
615 subsequent Tertiary compression, the in-sequences forward propagation on the south-verging major
616 deep thrusts attach on their footwall a north verging back-thrusts in favor of the formation of a deeper
617 down triangle zone (Boyer and Elliot; 1982; Huyghe and Mugnier, 1992). These structures are very
618 frequent at the front of many major thrust systems. They are described as the hinterland-dipping of
619 normal faults that are inverted as foreland-directed thrusts, and early forelandward-dipping normal
620 faults that are inverted as back-thrusts (McClay and Buchanan, 1992; Cooper et al., 1989; Bonini, et al.,
621 2012). The restoration of this section suggests at least two phases of compressive deformation featured
622 by the structures growth. The synclines, limited the two above-mentioned anticlines, shows a thin
623 Oligocene-Miocene series unconformably resting onto the Eocene-Paleocene and Late Cretaceous
624 sequences. Oligocene-Miocene series are themselves well deformed later such as the inverted ones

625 under the Berda back-thrust. This scenario reflects a first Middle to Late Eocene moderate tectonic
626 compression and a post Lower Miocene important second event. The first one, is recognized as the
627 Atlassic event and confirmed by recent studies in the southern Tunisia (Riley et al., 2011; Gharbi et al.,
628 2015, El Amari et al., 2016, this work). This cross section suggests that during the Miocene–present-day
629 successive compressions, the shortening within these structures is mainly accommodated by the
630 inversion of north-dipping faults acting now as major thrusts deeply rooted in the basement with
631 shallower thin-skinned south-dipping back-thrusts. Far South, the structure is approximately undeformed
632 by Tertiary successive compressions and show a thick Jurassic and Lower Cretaceous sequences. The
633 southern two-thirds of the section reproduce a large rollover anticline associated with the major north-
634 dipping listric faults (Fig. 22). Mesozoic series filled a deep basin (basement is at about 7 km below sea
635 level in the Chotts basin), broadly downthrown relative to northern and southern areas. This basin is also
636 getting deeper from West to East compared to the western deep computed from the last section S1. This
637 variation, from West to East, is perceived in term of deformation by decrease of shortening. The section
638 reveals an estimation of shortening of about 3.65 km (~3.3%) with 1.5 km (~1.3%) as deep shortening
639 and 2.15 km (~2%) as shallow shortening. The shallow shortening is widely different from the northern
640 zone showing 3.4 km shortening amount (~3.1%) and southern zone practically undeformed with 0.25
641 km shortening amount (~0.2%). The Paleozoic High, beneath the NCR demonstrated by the present study
642 which bounded by major basement faults, performs the role of an accommodation zone between north
643 side region deformed by shallow thrust systems connected on deep basement ones and the southern
644 region which is preserved as inherited extensional large rollover anticline formed during Mesozoic times
645 (Fig. 22).

646 A short section [S3], compared to the western ones, is constructed eastward in the Chott Fejej zone
647 based on well information and seismic reflection data (Fig. 23). This section is presented to confirm the
648 same arrangement developed before and validates that the Cenozoic compressional deformation
649 strongly decreases toward the East. This 43-km–long section illustrates the Chott Fejej basin as formed
650 above a normal movement of the fault systems. These systems, which delimit a deep graben beneath the
651 Chott Fejej, is associated with rapid subsidence during Jurassic and Early Cretaceous. To the North, a
652 similar extensional structure created by listric normal faulting delimits a tilted blocks geometry with
653 north-verging beneath the Jebel Es Smaïa. The constructed section shows a negligible shallow shortening
654 of about 0.65 km (~1.4%).

655 6. Discussion

656 The Southern Atlas Front in Tunisia, which is the easternmost end of the Atlas Mountains across
657 northwestern Africa, was presented for long time as developed in a strike-slip zone. The fold dispersal
658 and the style of the Tunisian Southern Atlas were interpreted as linked to Gafsa and Negrine-Tozeur
659 major NW-trending right lateral strike-slip fault systems (e.g. [Castany and Lapparent, 1952](#); [Abdeljaouad
660 and Zargouni 1981](#); [Zargouni and Ruhland, 1981](#); [Zargouni et al., 1985](#); [Abbes and Zargouni 1986](#);
661 [Fakraoui 1990](#)). The fold's distribution, in this model, was qualified of all of them as curved, right-
662 stepping *en echelon* folds strictly related to the strike-slip fault systems. The second interpretation of
663 structures is based on salt tectonics model, in which the Tertiary folding is preferentially nucleated on
664 the preexistent diapir apex originally growing during Early Jurassic period ([Hlaiem et al., 1997](#), [Hlaiem,
665 1999](#); [Zouaghi et al., 2011](#)). The third thin-skinned interpretation with south verging thrust system
666 ([Ahmadi et al. 2006](#); [Ahmadi et al., 2013](#); [Trigui et al., 2016](#)) interprets the anticline structures in this
667 zone as developed and the result of ramp-related folding, deformed the sedimentary cover during a
668 single tectonic phase (Miocene to recent). Recently new ideas have come to the fore, with introducing
669 the role playing by structural heritage in the Cenozoic deformation of the belt ([Riley et al., 2011](#); [Gharbi
670 et al., 2015](#); [El Amari et al., 2016](#); [Tanfous-Amri et al., 2017](#)). This study is integrated in the extensive
671 present way for renewing the longstanding geological conceptions as it is the case in many West
672 Mediterranean Alpine areas (e.g. [Vergés and Sàbat, 1999](#); [Faccenna et al., 2004](#); [Lacombe and Jolivet,
673 2005](#); [Deverchere et al., 2005](#); [Frizon de Lamotte et al., 2011](#); [Roure et al., 2012](#); [Bestani et al., 2015](#);
674 [Lepretre et al., 2018a](#)).

675 The mixed thick- and thin-skinned structural style that typifies the southernmost deformed structures of
676 the NCR in southern Tunisia is defined by shallower thrusts connected down on deep basement thrusts
677 having a net variation throughout the NCR and the Chotts region. This study highlights a shortening
678 amount, calculated from the above-exposed balanced sections, tending to decrease from North to South
679 and from West to East. The total shortening (deep + shallow) amount varies from 7.6 km (this work,
680 section S1) in the West to 3.65 km in the west-central part (this work, section S2), to 2 km ([Gharbi et al.,
681 2015](#)) in central part and 0.65 km in the eastern part (this work, section S3). This variation coincides with
682 variation of the top of basement which varies from 3.5 km below sea level in the S1 section, to 5.5 km in
683 the S2 section and 6 km in S3 section. From North to South, the Cenozoic deformations are concentrated
684 in the narrow zone of the NCR, which accommodate ~6.95 % of shortening and ~0.5% in the Chotts
685 region. This variation coincides with net variation of total sequences thickness which is almost 4 km in
686 the NCR and can reach 7.6 km of thickness in the Chotts. Additionally, this remarkable variation from
687 North to South and from West to East, coincide with the pre-Cenozoic shortening basin architecture

688 acquired during successive Mesozoic Tethyan extensional periods. Surface and subsurface data analyzed
689 before highlights a former Mesozoic basin with multipart geometries that varies from East to West and
690 from North to South. To simplify the earlier Mesozoic basin's style, the general geometry can be
691 described as follow: on a North-South strike, the central part consistent of the Chotts basin corresponds
692 to a large rollover anticline structure developed during Jurassic and Early Cretaceous times. This large
693 Mesozoic half-graben is developed between two high structures i.e. the NCR to the North and the
694 Telemzane arch or the Telemzane High in the South (Gabtni et al., 2006; Gabtni et al., 2011; Dhaoui et
695 al., 2014). The uplift of the Telemzane Arch occurred probably during the Hercynian orogeny (Busson,
696 1967; Underdown and Redfern, 2007) attested in southern Tunisia and Libya by remarkable large-scale
697 unconformity of the Mesozoic series on an E-trending preexistent Paleozoic arch (Boote et al., 1998). On
698 the northern side of the Chotts basin mega-rollover, an uplifted zone is defined by the present study as
699 the horst structure in the basement along the NCR. This high zone is uplifted during Mesozoic times
700 probable by south-dipping normal faults connecting down on the major listric large north-dipping fault
701 systems. On an East-West strike evolution, this basin is deeper during Jurassic period in its eastern part
702 showing Jurassic sequences with more than 2 km of thickness with southward dipping. The western side
703 is not as much of deep and shows a north-dipping 1 km-thick Jurassic sequences. During Early Cretaceous
704 an inversion of subsidence is started, and the western and central parts display deeper zone than the
705 eastern part showing 2 km of sequences thickness near Kébili to the West and almost 1 km in the East
706 near El Hamma region (see compilation on Fig. 20).

707 The central part of the NCR shows the outcrop of the only outcropping salt diapir called the Hadifa diapir,
708 which is located in the western side of Hadifa graben (Fig. 3). Although the present work is not focused
709 on this central part, the seismic section LS3a (Fig. 13) close to this area shows clearly the development of
710 salt structures. The development of this diapir seems to be localized at the side of Mesozoic half graben.
711 Triassic salt activity is probably initiated during Jurassic times and accentuated during Lower Cretaceous
712 times. Salt structures in this central part look to be closely associated and triggered during continuous
713 sedimentary loading. In addition, Hlaim (1999), in the far west of the NCR (Bouhlel structure, Fig. 2),
714 shows the emplacement of salt structure, in which salt migration is associated to synsedimentary normal
715 faults during Jurassic and Lower Cretaceous times. Recently, Gharbi et al. (2015) were similarly highlight
716 a salt activity beneath the Beidha structure (close to Hadifa salt diapir). Authors establish a scenario, in
717 which northward salt migration is controlled by inherited normal faults responsible for the emplacement
718 of the Beidha salt pillow.

719 Finally, the Southern Tunisian Fold-and-Thrust belt, which was interpreted for long time without
720 incorporation of its structural inheritance, is comparable to many others famous Cenozoic belts governed
721 by basin inversion and reactivation of inherited systems (Bonini et al., 2012; Espurt et al., 2014): e.g. the
722 Provençal orogenic system (e.g. Roure and Colleta, 1996; Lacombe and Mouthereau, 2002; Espurt et al.,
723 2012; Bestani et al., 2016), the Western Alps (e.g. Tricart and Lemoine, 1986; Bellahsen et al., 2012,
724 2014; Jourdon et al., 2014; Boutoux et al., 2014), the Apennines (e.g. Ghisetti and Vezzani, 1997; Scisciani
725 et al., 2014) and the Pyrenees (e.g. Guimera, 1984; McClay et al., 2004; Mouthereau et al., 2014; Diaz et
726 al., 2018).

727 The southern Atlassic domain of Tunisia appears to have strong influence of inherited former Mesozoic
728 E–and NW–trending fault systems, as well as the sedimentation rate which reveals a dominant E–W–
729 striking basin developed during successive rifting phases from Permian? to Early Cretaceous periods. The
730 NCR’s folded structures are interpreted to be formed above the E–trending inherited faults. These
731 systems are established as inherited deep basement fault systems shaped an E–W high and deep zones
732 during Mesozoic times. This E–trending systems is particularly active during Jurassic period. This period is
733 characterized by the development of the Chotts basin limited by two major high Paleozoic zones
734 (Telemzane arch in the South and the NCR high in the North). During Cenozoic compressions, the north-
735 dipping major faults are inverted and subsequently acting as major deep thrust systems. The south-
736 dipping ones are recognized as north–verging back-thrusts with moderate offset (i.e. Berda, Sif El Laham,
737 Beidha) connecting down on large thrust ones with a strong coupling between Basement and
738 sedimentary cover located in the Triassic evaporitic layers. During Cenozoic compressions, this system is
739 responsible for the chain building and is expressly active during Miocene and Pliocene N-trending
740 compression attested by thick Mio-Pliocene syntectonic strata. The NW–trending system is particularly
741 active during Early Cretaceous time controlling a strong thickness variation of Early Cretaceous
742 sequences. During Cenozoic compressions, this system is particularly responsible for the curved nature of
743 this chain, which probably acts as transfer areas between the wide-ranging E–W strike major structures.
744 The Chotts basin, which, until today, was considered as a large anticline related to the Cenozoic
745 compressional events, is presumably a large rollover anticline preserved since the earlier rifting stage.
746 The tertiary compressional front will be, therefore, repositioned in the front of NCR Front, and not in the
747 SCR as it was previously accepted.

748 **7. Conclusions**

- 749 - The Southern Atlas Front in Tunisia is characterized by mixed thick- and thin-skinned structural style
750 defined by shallow thrusts rooted down into deep basement thrusts with a preferential decollement
751 level located within the Triassic evaporites. The E-trending N-dipping thrusts are recognized as major
752 deep thrust systems. The S-dipping thrusts are recognized as N-verging back-thrusts with moderate
753 offset connecting down on large north-dipping ones. A NW-trending system is also recognized,
754 responsible for the curved nature of this chain, acting as transfer areas between the wide-ranging E-
755 trending major structures. The intermediate areas are characterized by salt tectonics predominantly
756 active during Jurassic and Lower Cretaceous, in response to active synsedimentary normal faults
757 associated with differential sedimentation loading.
- 758 - Along the belt, the total shortening amount tends to decrease from North to South and from West to
759 East. The calculated shortening varies from 7.6 km (~7.45%) in the West to 3.65 km in West-central
760 part, about 2 km in central area to insignificant (0.65 km) in the eastern part. From north to south,
761 deformations are concentrated in the narrow zone of the Northern Chotts Range, which
762 accommodate almost 7% of deformation and less than 1% in the Chotts region.
- 763 - The strong variation from North to South and from West to East, coincide with the pre-Cenozoic
764 shortening basin's architecture acquired during Mesozoic Tethyan evolution. This former evolution
765 generates a main Jurassic E-trending basin (Chotts basin) bounded by two high Paleozoic zones.
766 Inside the main basin, distributary sub-basins are recognized in relation to NW-trending faults mainly
767 active during Cretaceous period. The Chotts basin appears, therefore, as well-preserved Mesozoic rift
768 rollover structures taking place on the southern border of Atlasic orogen.
- 769 - The compilation of data, in this paper, suggests that the Front of the southeasternmost end of the
770 Atlasic domain is distinguished by the reactivation of inherited fault systems acting during alpine
771 orogeny as southward low-dip thrusts. The foreland shows a structural singularity to be formed by a
772 preserved almost unmodified and well exposed large Mesozoic rollover structure.

773

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786

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1220 **Figure captions**

1221 **Figure 1.** Simplified structural sketch of west Mediterranean region with location of the Southern Atlas
1222 Front of Tunisia with a base map is produced using ASTER and DGEM data.

1223 **Figure 2.** (a) Simplified lithostratigraphic column of the Southern Atlassic area, derived from outcrop and
1224 petroleum wells data with thickness variation of Mesozoic sequences (minimum and maximum). (b)
1225 Simplified geologic map of the southern Tunisia showing the major E-trending structural ranges with
1226 location of the study area.

1227 **Figure 3.** (a) Location of the studied seismic profiles (some of them are composite ones) of [Figs. 11-19](#)
1228 and the petroleum wells used in calibration, the base map is produced using ASTER and DGEM data. (b)
1229 Geologic map of the study area, data are derived from present work, [Abbes et al., 1986](#); [Ghanmi and](#)
1230 [Potfaj, 1987](#); [Ghanmi et al., 1988](#); [Mahjoub and Fakraoui, 1990](#); [Zouari et al., 1991](#); [Fakraoui et al., 1991](#);
1231 [Gharbi et al., 2015](#); [El Amari et al., 2016](#); revealing the detailed structures of southern Atlassic domain
1232 with location of surface geologic cross-sections of [Fig.4](#) and the long balanced cross-sections [S1-S2-S3]
1233 across the chain ([Figs. 21-23](#)).

1234 **Figure 4.** Geologic cross sections prepared from surface data across the main structures of the study
1235 area, see [Fig. 3](#) for location.

1236 **Figure 5.** (a) 3D Google earth WSW-looking view showing the central part of the Northern Chotts Range
1237 highlighting the interfering between E- and NW-trending fault systems with the location of the E-
1238 trending major thrust (Bir Oum Ali Structure) and Back-thrust (J. Sif Laham); the NW-trending acts as
1239 transfer areas between the wide-ranging E–W strike major structures. (b) West looking panoramic view
1240 of J. Sif Laham pointing the emplacement of the north-verging back-thrust between southern sub-
1241 horizontal Early Cretaceous sequences and northern folded Late Cretaceous sequences. West looking
1242 panoramic view of J. Bir Oum Ali showing the position of south-verging major thrust (incomplete reverse
1243 displacement from former normal throw). (c) Panoramic view, looking ENE, of reactivated normal fault
1244 (thickness variation from north to south) to thrust forming the fold above a triangular geometry
1245 established by north-verging back-thrust connected down on south-verging thrust. (d) East looking photo
1246 of the angular unconformity of Miocene–Pliocene series on Paleocene–Eocene sequences, which
1247 outcrops along 20 km of the Bir Oum Ali’s northern limb.

1248 **Figure 6.** Common preserved extensive deformation sealed within Cretaceous sequences in south
1249 Tunisia, (a) S-looking photo illustrating sealed normal faults within the Coniacian–Santonian sequences in

1250 the J. Haidoudi structure, (b) NW-looking photo of preserved normal faults within Hauterivian–Barremian
1251 sequences in the J. Beidha.

1252 **Figure 7.** E-looking panorama of an incompletely inversion of earlier normal fault as reverse observed
1253 within Late Cretaceous (Campanian) sequences in the northern side of the Bir Oum Ali structure.

1254 **Figure 8.** Lithostratigraphic correlation [C1] of regional extent showing the thickness variation from
1255 southwest to northeast. The correlation is given here without the major faults control with a primary
1256 objective to describe them in absence of structural adjustment (for this see [Fig. 20](#)).

1257 **Figure 9.** Lithostratigraphic correlation [C2] of regional extent showing the thickness variation across the
1258 central part of the study area. The correlation is given here without the major faults control with a
1259 primary objective to describe them in absence of structural adjustment (for this see [Fig. 20](#)).

1260 **Figure 10.** Lithostratigraphic correlation [C3] of regional extent showing the thickness variation across
1261 the eastern part of the study area. The correlation is given here without the major faults control with a
1262 primary objective to describe them in absence of structural adjustment (for this see [Fig. 20](#)).

1263 **Figure 11.** seismic profile LS1 (see [Fig. 3](#) for location), crossing the Chotts Jerid area from SW to NE,
1264 illustrates the southwestern part of the Chotts basin showing increase toward the middle part of the
1265 Chotts basin of Jurassic and Lower Cretaceous thickness associated to deep E–trending faults.

1266 **Figure 12.** seismic profile LS2, located in the central part of Chotts basin (see [Fig. 3](#) for location), shows a
1267 geometry of steep north-dipping normal faults associated to the growth strata pattern of the Lower
1268 Cretaceous sequences.

1269 **Figure 13.** seismic profile LS3a, located in the Eastern part of the Chotts basins (see [Fig. 3](#) for location),
1270 shows important normal fault systems relating the north dipping Chotts. The northern side exposes a
1271 remarkable thinning of Jurassic-Lower Cretaceous toward an area with internal confuse reflection
1272 patterns. This configuration is characteristically of structure related salt rise. salt tectonics is probably
1273 controlled by Mesozoic normal faulting that is associated with rim synclines (salt mini-basins?) well
1274 developed in Jurassic thick sequences around the salt body.

1275 **Figure 14.** seismic profile LS3b (see [Fig. 3](#) for location), Reflectors have been calibrated using projected
1276 data of P2 well onto the section. To the South, the Chott Fejej basin is formed above a normal movement
1277 of the fault systems which delimited a deep graben beneath Chott Fejej and associated with rapid
1278 subsidence during Jurassic and Early Cretaceous.

1279 **Figure 15.** 100 km-long composite seismic profile LS4a (see Fig. 3 for location), from the Saharan
1280 platform to the Chotts basin crossing the Southern Chotts range, the southwestern part of the seismic
1281 section shows a considerable thinning of the Mesozoic series above the faulted Paleozoic series. the
1282 deep central part highlights intra-basin growth faulting associated to Jurassic and Lower Cretaceous
1283 growth strata pattern.

1284 **Figure 16.** 65 km-long composite seismic profile LS4b (see Fig. 3 for location); southeastward, the seismic
1285 section shows a remarkable geometry indicating successive unconformities. The Hercynian unconformity
1286 between the Paleozoic and the Mesozoic sequences is well highlighted by successive toplap figures on
1287 the Paleozoic basement showing a northward slope varying from 2° to 4°.

1288 **Figure 17.** Seismic profile LS5 crossing the Metaloui plain (see Fig. 3 for location), shows continuous sub-
1289 horizontal reflectors on the both side of the section and a little deformed area in the central part. This
1290 section shows a series of NE-dipping steep normal faults characterizing different tectonics blocks with
1291 inconstant thick series of Early Cretaceous and Coniacian–Santonian.

1292 **Figure 18.** Seismic profile LS6 (see Fig. 3 for location), crossing the Sehib structure. The profile shows in
1293 the forelimb a well identified reflector's truncated zone stating a thrust fault above which the Sehib fold
1294 is formed.

1295 **Figure 19.** Seismic profile LS7 (see Fig. 3 for location), crossing the structure in the eastern part of the
1296 Sehib anticline, shows that the north part basin is formed above inverted normal faults which bounds a
1297 deep graben beneath Sehib and associated with important subsidence during Early Cretaceous.

1298 **Figure 20.** schematic sketch collecting all data coming from seismic and wells together with surface data.
1299 Data highlight a dominant former Permian? to Early Cretaceous E-W–striking basin (Chotts basin)
1300 bounded by two high zones: the Telemzame arch and Northern Chotts Range, with multipart geometries
1301 that varies from east to west and from north to south (see text for more details).

1302 **Figure 21.** balanced cross-section [S1] across the western part of the southern Atlas Front of Tunisia, see
1303 Fig. 3 for location.

1304 **Figure 22.** balanced cross-section [S2] across the central part of the southern Atlas Front of Tunisia, see
1305 Fig. 3 for location.

1306 **Figure 23.** balanced cross-section [S3] across the eastern part of the southern Atlas Front of Tunisia, see
1307 Fig. 3 for location.