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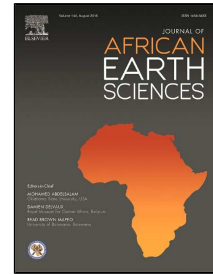


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1 **Distinguishing rift-related from inversion-related anticlines:**
2 **Observations from the Abu Gharadig and Gindi Basins,**
3 **Western Desert, Egypt.**

4
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6
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11
12 **Abstract**

13 Distinguishing the tectonic origin of anticlinal structures is problematic in regions with a
14 complex history of rifting and inversion. We present the results of seismic mapping, in the
15 form of time-depth (isochron) and time-thickness maps to characterize how sedimentary
16 thickness differentials evolved in response to normal faulting and to inversion events on faults
17 within the Abu Gharadig and Gindi Basins in the Western Desert of Egypt. Late Cretaceous
18 rift-related faults in the Abu Gharadig Basin strike NW-SE, W-E and SW-NE. In the eastern
19 part of the basin, a prominent SW-NE trending interbasinal saddle formed in response to
20 preferential subsidence forming half-grabens to its north-west and southeast, during the Mid-
21 Turonian to Santonian interval. Santonian to Palaeogene inversion in the Abu Gharadig Basin
22 developed on its northern basin margin, the absence of SW-NE striking faults in the eastern
23 central basin resulting in any inversion effects being minor. In the central Gindi Basin, Upper
24 Cenomanian to Lower Turonian SW-NE striking rift faults underwent inversion as early as
25 the Mid-Turonian. The orientation of existing rift faults and modification of the local stress
26 fields control the extent to which inversion was taken up in each basin through time. The Abu
27 Gharadig and Gindi Basins are two of the rift basins developed in West and Central Africa
28 that underwent rifting, inversion and dextral shearing during the Late Cretaceous. We
29 emphasize the value of high-resolution stratigraphic mapping to characterize short-lived and
30 subtle pop-up events that may have gone unnoticed.

31
32 **Keywords:** Anticlines - Rifting – Inversion tectonics - Abu Gharadig Basin - Gindi Basin.

33

34 1. Introduction

35 A common problem in the structural interpretation of rift basins with long and complex
36 tectonic histories is the distinction of structural and stratigraphic geometries that are the
37 product of extensional rifting from those that result from compression and structural
38 inversion. A range of fold types may form in response to normal fault growth, either a) as a
39 response to growing fault hanging wall and footwall displacement fields (e.g. Ellis and
40 McClay, 1988; Schlische 1995; Contreras et al. 2000) or b) at a shorter wavelength, to the
41 propagation of fault-tip monoclines (Wilson et al., 2009), or c) due to the interaction of
42 displacement fields associated with adjacent normal faults (Morley et al., 1990; Cartwright et
43 al., 1995; Dawers and Underhill 2000; Rotevatn and Jackson, 2014). Inversion of pre-existing
44 normal faults and concurrent folding or tightening of rift-related folds may be a record of a
45 change in the far-field stress regime from extension to compression (Cooper and Williams
46 1989; Coward 1991; Tavernelli 1996; Turner and Williams 2004; Scisciani 2009; Withjack et
47 al., 2010; Bonini et al., 2012), or to the rotation of principal stress directions (e.g. Paton and
48 Underhill 2004). The latter authors showed that subtle shifts in the orientation of the
49 maximum deviatoric stress direction during rifting can create structures that in many cross-
50 sections appear to be inversion features. There is a possibility that rift-related structures may
51 be misinterpreted as due to structural inversion resulting from a switch to a regional
52 compressive regime. Examples of inverted rift basins worldwide are many and include the
53 Australian Dampier Sub-Basin (Cathro and Karner, 2006); the Cameros Basin in Spain (Salè
54 et al., 2014) and the Brazilian Rio do Peixe Basin (Nogueira et al., 2015). Detection and
55 interpretation of inversion features within rift basins, and their distinction from rift-related
56 structures, is important for accurate deductions of orientations of principal stress axes and
57 appreciation of the region's plate tectonic history through time.

58
59 Several hypotheses regarding the tectonic history of the Western Desert province in Egypt
60 arose due to uncertainties in the origin and timing of anticlinal growth in various rift basins in
61 the region. For example, Bayoumi and Lofty (1989) invoked a broad phase of Late
62 Cretaceous to Early Cainozoic compression and inversion in the Western Desert. Whereas,
63 Moustafa (2008) referred to continuous normal fault activity in the Late Cretaceous to Early
64 Cainozoic along WNW-ESE striking faults, which are thought to be parallel to the
65 contemporaneous shortening direction related to convergence of Afro-Arabia and Eurasia.
66 Moustafa (2008) proposed that multiple pulses of inversion took place across the Western
67 Desert in the Late Cretaceous to Oligocene, inverting early rift faults of NE-SW, ENE-WSW

68 and NNE-SSW orientation and developing similarly oriented structural traps. Bosworth et al.
69 (2008) noted that arc collisions on the north-east African margin from the Santonian onwards
70 yielded varied structural expressions across the Western Desert, depending on whether the
71 affected structure lay within the stress shadow of the Cyrenaican inversion belt. Shortening
72 was most severe in a belt reaching from northern Libya across to the Levant and diminished
73 to the south and east of this deformation belt. In this context, it is the northern margin of the
74 Abu Gharadig Basin (Fig. 1) which has accommodated the major part of the shortening of this
75 basin, across the Mubarak Inversion Complex (as illustrated in Bosworth et al. 2008).
76 Moustafa (2013) proposed that the NE-SW orientation of the Abu Gharadig Field anticlinal
77 trap was due to rift-related subsidence to both the NW and SE of the fold axis.

78
79 This study focuses on the Abu Gharadig and Gindi Basins in the Western Desert, Egypt
80 (Figure 1), and illustrates the use of stratigraphic thickness changes in time-thickness maps as
81 the primary tool to distinguish rift-related from inversion-related structural geometries. As
82 outlined above, the different anticline geometries and orientations present across the adjacent
83 basins have been explained by differing tectonic interpretations in the recent past. In this
84 study, we aim to show that with detailed mapping, even where only a really restricted 2D
85 seismic surveys are available, coherent tectonic interpretations can be derived from these data.
86 We present time-depth maps of key stratigraphic surfaces within each basin, and then produce
87 time-thickness maps for key intervals. These time thickness maps together with cross-sections
88 are critical for distinguishing the timing of extensional activity on normal faults of varied
89 orientations, and for isolating which fault orientations may have undergone any
90 compressional inversion effects and when this occurred.

91

92 **2. Geologic setting**

93 The Triassic-Early Cretaceous rift basins in the northern Western Desert of Egypt resulted
94 from the opening of the Tethys Ocean (e.g. Ziegler, 1987; Frizon de Lemotte et al., 2011) and
95 subsequent inversion (Bayoumi and Lofty, 1989; Guiraud, 1998; Bosworth, 1994; Bosworth
96 et al., 2008; Wescott et al., 2011). The Abu Gharadig Basin and the Gindi Basin are
97 productive hydrocarbon basins in the northern part of the Egyptian Western Desert. Both
98 basins originated as N- and NE-tilted half-grabens in the Jurassic-Early Cretaceous concurrent
99 with the opening of the Neo-Tethys and Atlantic oceans (Moustafa, 2008).

100

101 The stratigraphy of the northern Western Desert, including both the Abu Gharadig and Gindi
102 Basins is summarized in Figure 2. This study is focused on the Upper Cretaceous to
103 Oligocene strata. The Late Cretaceous rock units are divided into three formations from base
104 to top, the Bahariya, Abu Roash and Khoman Formations. The Bahariya Formation
105 (Cenomanian age) conformably overlies the Early Cretaceous Kharita Formation (EGPC,
106 1992). It consists of fine to medium-grained sandstone with thin shale and carbonate interbeds
107 (Soliman and El Badry, 1980; Schlumberger, 1984). The Bahariya Formation is thought to
108 have been deposited by fluvial systems passing into inner shelf, shallow marine environments
109 (Abd El Kireem et al., 1996).

110
111 The Abu Roash Formation (Turonian-Coniacian age) conformably overlies the Bahariya
112 Formation and unconformably underlies the Khoman Formation. It is composed of
113 alternations of carbonate and clastic rocks and is subdivided into seven members named A to
114 G from top to base. It was deposited under shallow marine shelf conditions (Schlumberger,
115 1984; EGPC, 1992; Abd El Kireem et al., 1996) as successive oscillating transgressive-
116 regressive marine cycles (Sarhan, 2017a). The Khoman Formation (Santonian-Maastrichtian
117 age) overlies the Abu Roash Formation unconformably (Hantar, 1990; Abd El Kireem and
118 Ibrahim, 1987; Schrank and Ibrahim, 1995) and comprises a thick chalky limestone section
119 deposited under outer shelf conditions (Sultan and Halim, 1988; EGPC, 1992; Abd El Kireem
120 et al. 1996; Tantawy et al., 2001; Mahsoub et al., 2012) during widespread sea-level rise
121 covered most of the northern parts of Egyptian lands associated with basin subsidence
122 (Sarhan, 2017a). It is subdivided into two members by an unconformity surface; the lower
123 member B is composed of shale and argillaceous limestone whilst the upper A member is
124 composed of chalky limestone (Fawzy and Dahi, 1992). These unconformities tend not to be
125 obvious within depocentres, where continuous sedimentation resulted in their correlative
126 conformities.

127
128 At the end of Cretaceous time, a marked depositional gap and erosional truncations developed
129 over much of the Western Desert, during the Palaeocene (Schlumberger, 1984; EGPC, 1992).
130 The Eocene Apollonia Formation unconformably overlies the Late Cretaceous Khoman
131 Formation and consists of hard limestone and rare chalky limestones, with some shale
132 intercalations deposited in an open marine environment (Fawzy and Dahi, 1992; EGPC, 1992;
133 Bakry, 1993). The Oligocene Dabaa Formation is composed of a thick shale succession and
134 overlies the Apollonia Formation unconformably (Fawzy and Dahi, 1992). It was deposited

135 under inner-shelf to littoral environments, shallower than the depositional environment of the
136 underlying Apollonia Formation (EGPC, 1992).

137

138 Most of the hydrocarbons in the Western Desert, including in the Abu Gharadig and Gindi
139 Basins were discovered by targeting structural prospects in sandstone and carbonate
140 reservoirs (e.g. Sarhan et al., 2017a & b). These structural traps appear as three-way fault-
141 bounded or elongate, dome-like four-way closures, within the Cretaceous sedimentary
142 package (e.g. Sarhan, 2017b; Sarhan et al., 2017b). These traps formed when Upper
143 Cretaceous tectonic inversion affected the northern part of the Western Desert (Sultan and
144 Abd El Halim 1988; EGPC, 1992, and David et al., 2003). Structural geometries in the
145 inverted basins display great variation from one basin to another, depending on the pre-
146 inversion structure of each basin.

147

148 The present work aims to characterize the orientation of inversion structures in the Abu
149 Gharadig Basin and the Gindi Basin, to understand how Upper Cretaceous tectonic inversion
150 affected the two sedimentary basins in the northern Western Desert of Egypt.

151

152 3. Data and methodology

153 The first study area is the central part of the Abu Gharadig Basin in the northern Western
154 Desert that lies between latitudes 29.5° and 30° N and between longitudes 28.3° and 28.8° E
155 (Figure 1). The seismic data available for this study were 30 (thirty) 2D migrated seismic
156 reflection profiles at 2 km spacing extracted from a 3D seismic cube. Eleven (11) seismic
157 lines-oriented ENE-WSW are up to 55.5 km long, and nineteen lines are oriented NNW-SSE
158 trend and 25.5 km (Figure 3a). These seismic sections are tied to five wells (AG-2, AG-5,
159 AG-6, AG-15 and SAWAG-1).

160

161 The second study area lies between latitudes 29.7° and 29.8° N and longitudes 30.4° and 30.5°
162 E within the central part of the Gindi Basin (Figure 1). The available data comprised 20
163 (twenty) 2D migrated seismic reflection sections, comprising eleven of N-S orientation and
164 nine trending E-W (Figure 3b). These seismic data sets are tied to three wells, SWQ-4, SWQ-
165 21 and SWQ-25.

166

167 The procedures used in the present work to achieve the aim of the paper are a) the picking of
168 faults in each basin, and b) the picking of distinct and laterally extensive seismic horizons,

169 which are tied to the stratigraphic data from the wells in the basins. The interpreted faults and
170 key horizons were mapped over the grid of 2D seismic lines to construct time-depth
171 (isochron) maps, which allow the preparation of time-thickness (time-isochore) maps for each
172 stratigraphic interval.

173

174 **4. Results**

175 **4.1. Abu Gharadig Basin**

176 Within this dataset, five seismic reflectors represent the clearest surfaces along all the
177 examined seismic profiles. These surfaces display strong, relatively continuous and high
178 amplitude reflectors across the seismic profiles (Figures 4a and b). According to the well-
179 seismic ties, these horizons represent: (a) the top of the Bahariya Formation (green), varying
180 in time-depth from 1400 to 2900 ms (Figure 2); (b) the top of the Abu Roash D Member (pale
181 blue) that ranges from 1300 to 2500 ms in time-depth; (c) a high amplitude reflector
182 designated near the top of Khoman B Member (yellow), picked higher than the real top of
183 Khoman B surface by about 50 ms and varying in time-depth between 800 and 1800 ms; (d)
184 near the top of Khoman A Member (pink), picked about 150 ms higher than the real top of
185 Khoman A surface, ranging in time-depth from 500 to 1300 ms; and (e) the uppermost
186 horizon being a near-top Dabaa Formation surface (dark blue) at about 100 ms below the true
187 top Dabaa Formation, and which varies in time-depth between 300 and 625 ms.

188

189 Fault picks reveal that all the interpreted faults in this study area have net normal
190 (extensional) offsets. The time-depth maps for the interpreted surfaces show that the faults
191 form two groups, faults of NW-SE trend with lengths between 5 km and at least 16 km, and
192 faults of approximately E-W trend, which range in length from 5 km to at least 20 km (Figure
193 5). These intrabasinal fault trends compare with the Abu Gharadig Basin border faults, which
194 have NE-SW to E-W to ESE-WNW strikes (Figure 1). Note that the map areas presented in
195 Figure 5 and other figures have been trimmed to avoid edge effects in the contouring.

196

197 The largest fault in this area forms one of the Abu Gharadig basin-bounding faults with throw
198 ranging from 900 ms up to about 1100 ms at top-Bahariya level (F1, coloured red in Figure 4b
199 and in Figure 5). Sedimentary strata in the hanging wall of this main fault have a component
200 of dip toward its plane and form a significant dip-fan into the fault across all Upper
201 Cretaceous to Oligocene mapped intervals, suggesting persistent reactivation and progressive
202 rotation of the fault block during basin infilling episodes on its hanging wall side.

203
204 Mapped surfaces (Figure 5) are tilted to the north and north-west, towards the basin-bounding
205 fault set that lies outside this study area, but which controls the northern margin of the Abu
206 Gharadig Basin (Figure 1). The northern basin margin is delimited by E-W trending faults
207 (Bosworth et al., 2008). An anticline in the central part of the study area trends NE-SW and
208 includes the structural closure of the Abu Gharadig Field (Moustafa, 2008). This anticline is
209 prominent at Top Bahariya and Top Abu Roash D levels (Figures 5a and 5b).

210
211 Line 5240 (Figure 4b) shows a major E-W fault (F2, coloured green) which downthrows to
212 the north with significant offsets at Top Bahariya Formation and Top Abu Roash D levels. In
213 other lines, it is clear that this fault cuts through all the interpreted horizons from the Top
214 Bahariya horizon up to the Near-top Dabaa Formation horizon (Figure 5). Growth on this
215 fault (and associated secondary faults of similar orientation) during the Top Abu Roash D
216 Member to near the top of Khoman B Member interval contributed to the formation of the
217 anticlinal closure, as did growth in the hanging wall to the F1 fault of opposite dip. The
218 anticline may be considered to lie in the hanging wall of F1, the more major fault of the two,
219 because all picked stratigraphic surfaces of the anticline are at a lower structural level than the
220 equivalent surfaces in the footwall to F1, as seen in the WSW-ENE oriented line of section in
221 Figure 4b.

222
223 The time-thickness maps for the four intervals bounded by the five interpreted horizons are
224 presented in Figure 6. The Top Bahariya Formation to Top Abu Roash D interval (Figure 6a)
225 displays thickening over the area of study to the north-west, into the northern basin margin
226 fault set, and thickening to the north-east into the F1 NW-SE trending normal fault. Extension
227 was thus being taken up on these faults and secondary faults of E-W to NW-SE strike during
228 this period.

229
230 The second time-thickness map displays the interval between Top Abu Roash D and near the
231 top of Khoman B Member (Fig. 6b) and shows a significant thickening to the NE again,
232 towards fault F1 of NW-SE strike, with time-thickness values reaching 750 ms on the
233 downthrown side of the fault. Also prominent within this time interval is thickening on either
234 side of the Abu Gharadig central antiform. Within our study area, thicknesses can be seen to
235 increase northwestwards to the north-west of this structure and to increase to the east
236 southeast to the southeast of this antiform. Faults active within the study area are oriented E-

237 W to NW-SE, but the thickening to the NW and ESE of the antiform implies that faults at the
238 basin margins to the north (dipping S or SE) and to the south-east (dipping NW) were
239 controlling sedimentary thickening, such that the antiform formed as a saddle between two
240 half-grabens of opposing dip during this period of active rifting. This saddle thus formed an
241 interbasinal high between two fault-controlled depocentres of opposing polarity.

242

243 The time-thickness map of the stratigraphic interval between the near the top of Khoman B
244 and near the top of Khoman A surfaces (Fig. 6c) displays less dramatic thickness differentials
245 of up to 200 ms TWT in the study area. This suggests either less variable sedimentation rates
246 with lesser tectonic activity during this period of deposition or a period of basin under filling.
247 In the south and east of the study area, thicknesses increase toward the basin margin that lay
248 to the southeast. In the north of the study area, thicknesses increase on the south-west,
249 downthrown side of F1, indicating continued normal displacement on this structure.
250 Thicknesses also increase to the north-west of the central antiform axis. Within the limited
251 area mapped in this study, it is not clear whether the thickness differentials toward the
252 northern and southeastern basin margins were due to continued extensional growth on basin
253 margin faults or due to thickening into inversion-related growth synclines to NW and SE. The
254 marker horizons remain below their regional levels, in net extension. Either way, the central
255 antiform saddle is accentuated by differential subsidence to the north or north-west and to the
256 south of its axis during this interval.

257

258 The time-thickness map of Figure 6d represents the interval from near the top of Khoman A
259 to near the top of Dabaa Formation, which post-dates the Cretaceous-Palaeogene
260 unconformity. This interval again shows a significant increase in thickness on the
261 downthrown side of the F1 NW-SE trending normal fault. However, across the study area,
262 thicknesses increase toward the NNW to a maximum time-thickness of 680 ms and higher
263 values toward the northern basin margin faults. There is no expression of the central high or
264 saddle. The south-dipping half-graben to the south of the Abu Gharadig anticline was
265 therefore inactive. Reactivation of rifting therefore appears to have been focused onto F1 and
266 the northern basin margin during this episode, with only minor faulting on secondary faults
267 within the study area. It is worth noting that compared to the frequency of syn-depositionally
268 active NW-SE faults during the Top Abu Roash D to near the top of Khoman B interval
269 (Figure 6b), in the southern part of the study area, the near the top of Khoman A to near the

270 top of Dabaa Formation interval (Figure 6d) is characterised by activity across E-W to WNW-
271 ESE faults in the study area, except for reactivation of F1.

272

273 **4.2. El-Gindi Basin**

274 Seismic horizons within the grid of 2D lines from the central Gindi Basin (Figures 1 and 3)
275 have been picked corresponding to the Top Bahariya Formation (yellow in Figure 7), Top
276 Abu Roash F Member (blue), Top Abu Roash D Member (dark green) and Top Abu Roash B
277 Member (light green), as constrained by ties to the wells SWQ-4, SWQ-21 and SWQ-25.

278

279 All picked horizons on the seismic sections display a clear anticline bounded by two faults,
280 which show modest net reverse fault offsets of about 10-100 ms TWT at all the picked
281 stratigraphic levels (Figure 7a, b and c). There is, however, some along-strike variation with
282 the fault to the southeast of the anticline retaining a net extensional offset (e.g. Figure 7c).
283 The time-depth maps for the picked horizons reveal a distinctive anticline, the axis of which
284 plunges toward the NE and which is bounded by the two, 4-5 km long, NE-SW striking
285 reverse faults (Figure 8).

286

287 The first time-thickness map for the Top Bahariya to Top Abu Roash F interval (Figure 9a)
288 displays thinning to the south-west and south of the mapped faults. The interval shows slight
289 thickening within the area bounded by the two faults, in the northeast of the mapped area.
290 This suggests that these faults were active normal faults during this episode, the upper
291 Cenomanian to lower Turonian, and controlled the formation of a minor graben in the central
292 Gindi Basin.

293

294 The second time-thickness map (Figure 9b), of the interval between Top Abu Roash F and
295 Top Abu Roash D (middle Turonian), shows a clear thinning in the area between the two
296 interpreted NE-SW faults and thickening to the north-west and south-east of these (see also
297 Figure 7). This suggests that these faults were active in the middle Turonian time interval as
298 reverse faults. The time-thickness differential is up to ca. 40 ms TWT, from 190 ms TWT
299 within the faults to 230 ms TWT outside the faults.

300

301 The third time-thickness map (Figure 9c) represents the interval between the Top Abu Roash
302 D Member and the Top Abu Roash B Member (middle to upper Turonian). There is subtle
303 thinning in the area between the two interpreted NE-SW faults and thickening to the north-

304 west and southeast of the faults. This suggests that the faults continued to behave as reverse
305 faults into the middle to upper Turonian, although with a thickness differential of only ca. 10
306 ms TWT.

307

308 **5. Discussion and Conclusions**

309 In many petroleum provinces, positive basin inversion structures represent important
310 structural hydrocarbons traps (e.g. Bally, 1983; Fraser and Gawthorpe, 1990). The timing of
311 hydrocarbon migration relative to the development of inversion structures is also critical in
312 establishing a viable petroleum play, i.e. whether the migration may have occurred before or
313 during or after the tectonic inversion (Sibson, 1995). In this study, we have constrained the
314 structural timing element regarding Late Cretaceous rifting and tectonic inversion events in
315 hydrocarbon-bearing sedimentary basins in the northern Western Desert, in the Abu Gharadig
316 and Gindi Basins.

317

318 In lower Turonian times, both the Abu Gharadig Basin and the Gindi Basin show evidence of
319 extensional faulting on faults with strikes ranging from NE-SW to E-W to NW-SE, implying
320 a maximum deviatoric stress (extension) direction of approximately N-S. Through the mid-
321 Turonian to Santonian history of the eastern Abu Gharadig Basin, a prominent saddle
322 developed as a high between two opposing half-graben depocentres that were controlled by
323 basin margin faults. This forms the prominent NE-SW striking faulted antiform of the Abu
324 Gharadig Field (Moustafa, 2008). In contrast, detailed mapping has shown that after
325 Cenomanian-lower Turonian rift faulting a subtle inversion event took place in the central
326 Gindi Basin during the mid-Turonian, continuing into the upper Turonian. This was
327 accommodated by the reverse offset of NE-SW striking faults. This event may be an early
328 expression of the dextral shear environment across the West and Central African rift system
329 and the rifts of the Western Desert and Sudan that existed throughout the Santonian to
330 Maastrichtian (Fairhead et al., 2013). Pulses of rifting and inversion occurred as responses to
331 changes in the intraplate stress regime that were driven by changes in spreading rates and
332 azimuthal directions in the Central and South Atlantic oceans. In this context, minor mid-
333 Turonian inversion of NE-SW faults in the Gindi Basin could be considered as a “pop-up” in
334 a local dextral shear phase, with NE-SW maximum extension and NW-SE compression
335 directions. The clear absence of this mid-Turonian inversion in the nearby Abu Gharadig
336 Basin could be explained as due to either a) it not being resolved seismically and
337 stratigraphically, or b) the absence of NE-SW striking faults within the mapped area, i.e.

338 faults of an orientation suitable for pop-up development, or c) the local stress field in the Abu
339 Gharadig Basin not having a significant deviatoric compressional component.

340

341 The Top Abu Roash D Member to near the top of Khoman B interval, mid-Turonian to
342 Santonian, in the Abu Gharadig Basin shows net extensional displacements on faults within
343 the basin and on the NW-SE oriented basin margin fault (F1 in Figure 4b). This does not
344 contradict the initiation of inversion on the northern margin of the basin during the Santonian,
345 as described and illustrated by Moustafa (2008) and Bosworth et al. (2008). Nevertheless, any
346 reverse reactivation of faults within the eastern Abu Gharadig Basin study area within this
347 time interval was minor, such that faults retain net extensional offsets. Structural shortening
348 thus appears to have been focused onto faults of appropriate strike on the main northern basin
349 margin (the Mubarak Inversion trend of Bosworth et al., 2008).

350

351 The near the top of Khoman B to near the top of Khoman A interval in the Abu Gharadig
352 Basin study area shows further thickness differentials associated with the central saddle,
353 implying extensional faulting continued into this stratigraphic phase. However, across the
354 basin more dramatic unconformities are clear at the top of the Khoman B Member and within
355 the Khoman A Member (Figure 10, after Moustafa, 2008), and these converge toward the
356 northern basin margin (Moustafa, 2008; Bosworth et al., 2008). It is therefore difficult to be
357 precise with respect to when preferential subsidence and sediment accumulation to the north-
358 west and southeast of the central saddle switched from being due to normal faulting to due to
359 inversion of basin margin structures. Compressive strain was localised onto the northern basin
360 margin (see Figure 2D of Bosworth et al., 2008), but extension may have continued on
361 structures of suitable orientation, such as the NW-SE trending basin margin fault (F1 in
362 Figure 4b). It is not clear whether the central saddle antiform underwent any tightening
363 associated with these inversion events, but net normal offsets remain on all the intrabasinal
364 faults in the study area.

365

366 Above the Maastrichtian-Palaeogene unconformity, the interval from near the top of Khoman
367 A to near the top of Dabaa Formation thickens to the NNW, consistent with extensional
368 reactivation of the northern basin-bounding fault set and to the northeast, on the downthrown
369 side of the F1 basin-bounding fault. There is no sign of thickening southeastwards. E-W to
370 WNW-ESE oriented intrabasinal faults show normal fault reactivation (Figure 6d). So a pulse
371 of rifting involving tilting down to the north can be invoked. The lack of expression of the

372 central saddle suggests that any normal fault-related growth or inversion-related tightening of
373 this structure had ceased by Apollonia Formation times. This however, does not preclude
374 further inversion on the northern basin margin continuing after burial of the base-Palaeogene
375 unconformity in the eastern Abu Gharadig Basin.

376

377 This study highlights the value of mapping at a high stratigraphic resolution to identify subtle
378 changes in the local stress regime of sedimentary basins, which have undergone pulses of
379 rifting and inversion. Basins across the West and Central African rift system and the rifts of
380 North-East Africa were deforming in response to Late Cretaceous rifting, collision-related
381 inversion, and regional dextral shear (Fairhead et al., 2013). In such shear-influenced settings,
382 the Gindi Basin provides an example of how certain basins may show short-lived inversion or
383 “pop-up” events in response to local changes in principal stress directions or their relative
384 magnitudes. However, these events may only be characterized when isochron and isochore
385 mapping is carried out at the scale of age (e.g. Turonian) or shorter time intervals. The
386 recognition of a Mid- to Upper Turonian inversion event in the Gindi Basin suggests that
387 other subtle pop-up phases on structures of suitable orientation might have occurred and
388 should be searched for in other West, Central and Northeast African rift basins earlier than the
389 main Santonian-Palaeogene inversions described.

390

391 In conclusion, how Late Cretaceous tectonic inversion processes affected the northern
392 Western Desert varies from one sedimentary basin to another. The Abu Gharadig Basin and
393 the Gindi Basin are two examples, which display different styles and timing of inversion
394 events. The inversion in the Abu Gharadig Basin was focused on its northern basin margin. In
395 contrast, in the Gindi Basin, the NE-SW striking, intrabasinal rift faults allow a Mid- to Upper
396 Turonian inversion to be identified.

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Figures Captions

- Fig. (1) Regional map highlighting the study areas within the Abu Gharadig Basin and the Gindi Basin in the northern Western Desert (after Bosworth et al, 2008). See Figure 3 for data distributions in the boxed study areas.
- Fig. (2) Regional stratigraphic column of the northern part of the Egyptian Western Desert including the Abu Gharadig and Gindi Basins (stratigraphic nomenclature after Moustafa, 2008). Lithostratigraphic column is based on the well data used in this study.
- Fig. (3) Available seismic lines and well locations of the study areas. (A) Abu Gharadig Basin and (B) Gindi Basin.
- Fig. (4) Interpreted seismic lines within the Abu Gharadig Basin. (A) Dip seismic line no. 1390 and (B) strike line no. 5240 (see Fig. 3 for relative locations), which shows long-lived growth into fault F1, coloured red, and an oblique section through the central Abu Gharadig anticline prominent at Top Abu Roash D level (pale blue).
- Fig. (5) Time-depth contour maps (ms TWT) of the picked seismic horizons within the Abu Gharadig Basin. (A) Top Bahariya Formation, (B) Top Abu Roash D Member, (C) Near the top of Khoman B Member, (D) Near the top of Khoman A Member, (E) Near the top of Dabaa Formation. Note the NE-SW trending, NE-plunging central anticline in (A) and (B).
- Fig. (6) Time-thickness contour maps (ms TWT) of the picked seismic intervals within the Abu Gharadig Basin. (A) From Top Bahariya Formation to Top Abu Roash D Member, (B) from Top Abu Roash D Member to near the top of Khoman B Member, (C) from near the top of Khoman B Member to near the top of Khoman A Member, (D) from near the top of Khoman A Member to near the top of Dabaa Formation. Note the thickening to the NW and to the ESE of the central NE-SW trending anticline or saddle, most prominent during the (B) Top Abu Roash D to near the top of Khoman B interval.

637 Fig. (7) Interpreted seismic lines within the Gindi Basin, showing its central anticline. (A)
638 Dip line no. 4766 and (B) strike line no. 10476 (see Fig. 3 for relative locations).

639

640 Fig. (8) Time-depth contour maps (ms TWT) of the picked seismic horizons within the Gindi
641 Basin, characterizing the prominent NE-SW anticline. (A) Top Bahariya Formation,
642 (B) Top Abu Roash F Member, (C) Top Abu Roash D Member, (D) Top Abu Roash
643 B Member.

644

645 Fig. (9) Time-thickness contour maps (ms TWT) of the picked seismic intervals within the
646 Gindi Basin. (A) From Top Bahariya Formation to Top Abu Roash F Member, (B)
647 from Top Abu Roash F Member to Top Abu Roash D Member, (C) from Top Abu
648 Roash D Member to Top Abu Roash B Member. Note the switch from minor graben
649 development (interval A) to thinning onto the crest of the structure and inferred
650 inversion during intervals B and C.

651

652 Fig. (10) Sketch modified from an interpreted E-W seismic profile in Moustafa (2008) across
653 the Abu Gharadig Basin showing the positive structural inversion of a NE-SW
654 oriented basin margin fault during the Santonian to Palaeogene. Orange arrows
655 denote directions of reflector (strata) terminations.

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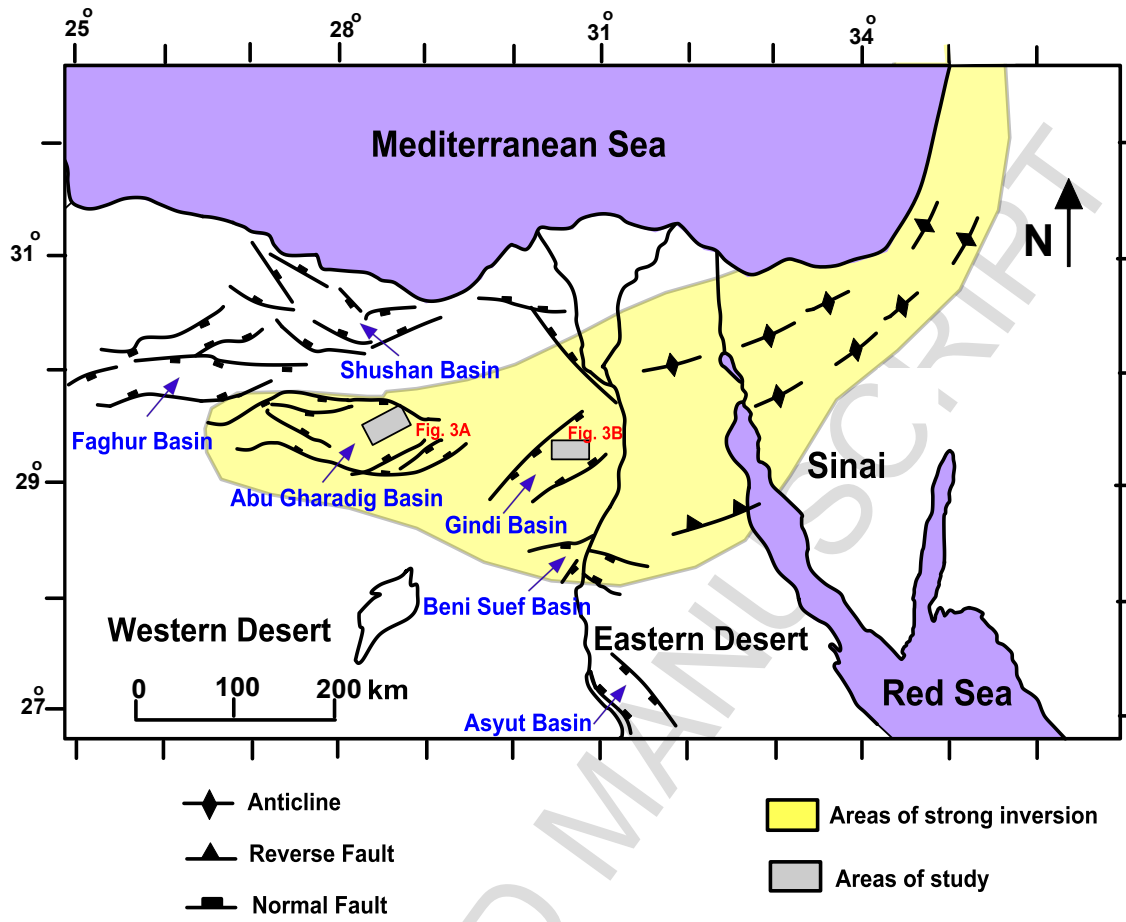


Figure 1

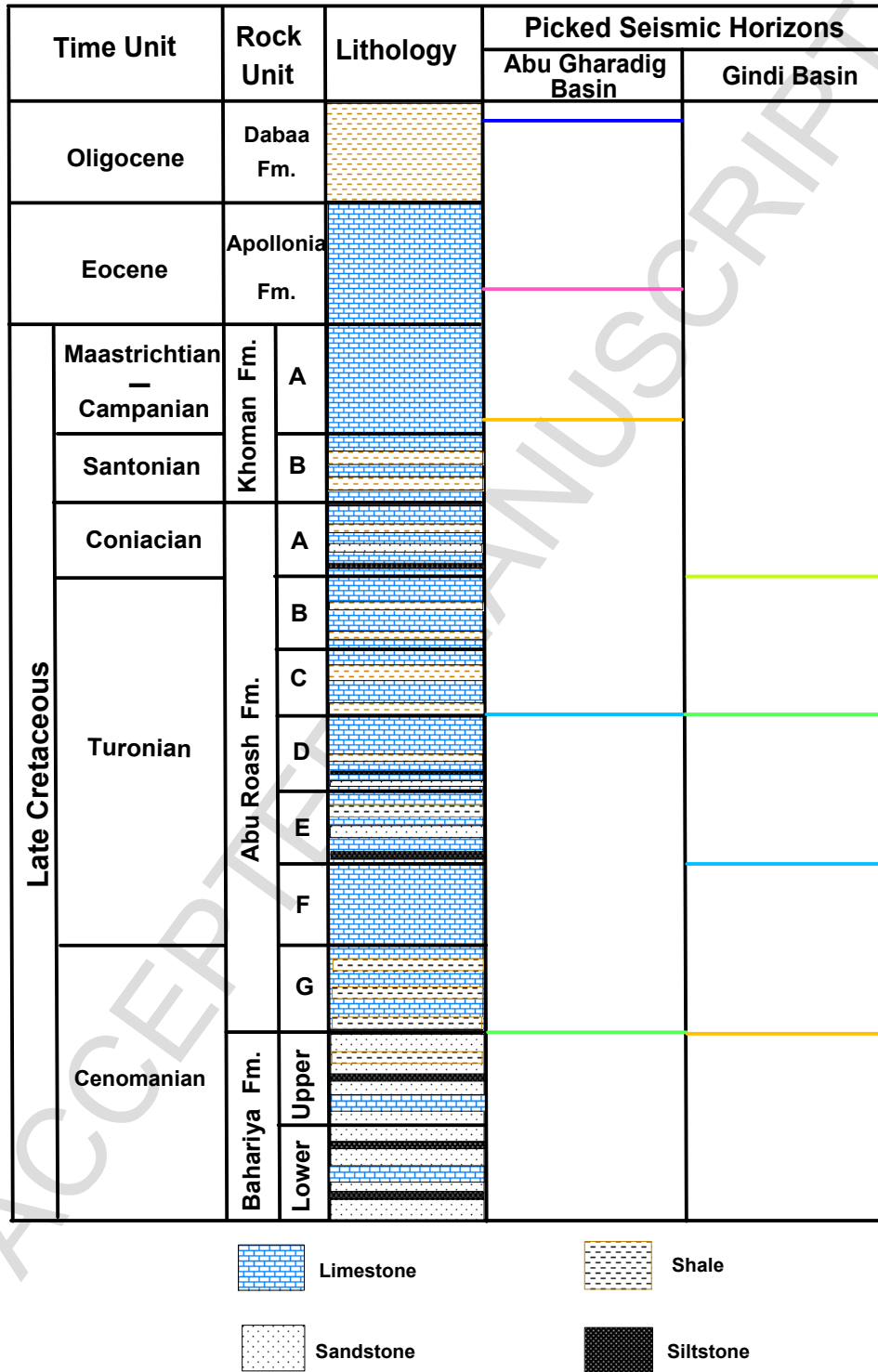


Figure 2

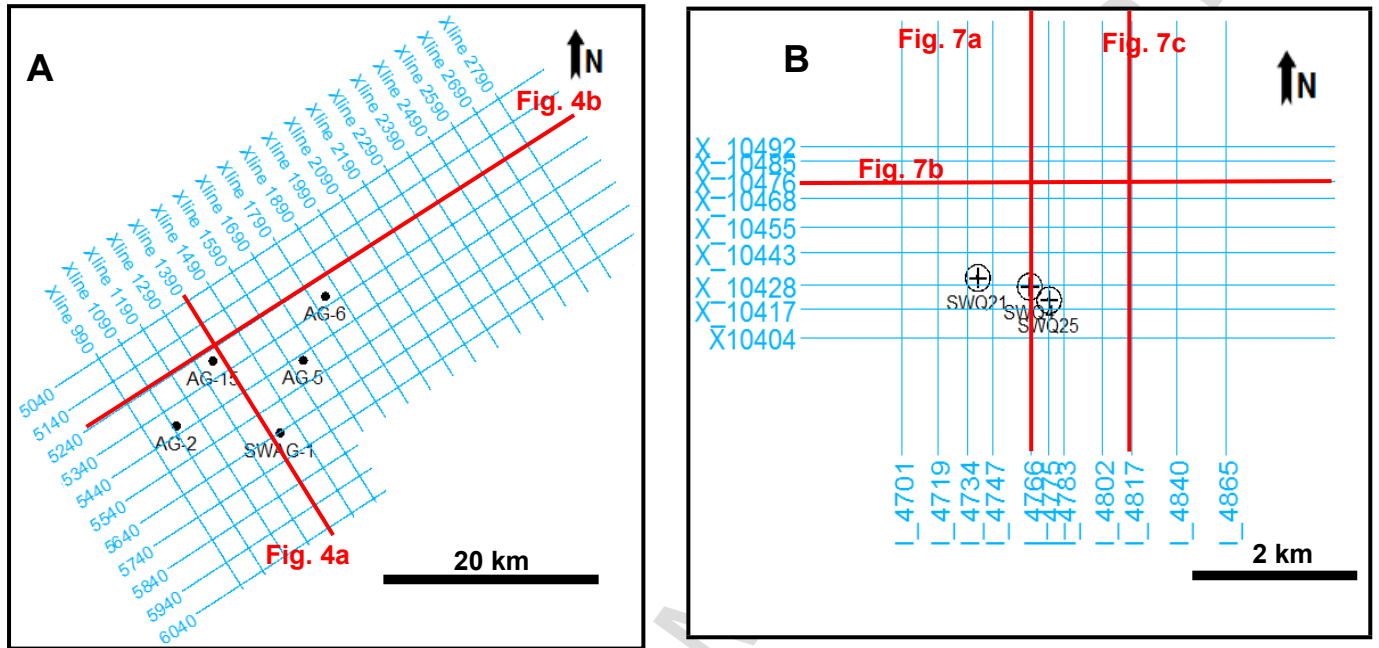


Figure 3

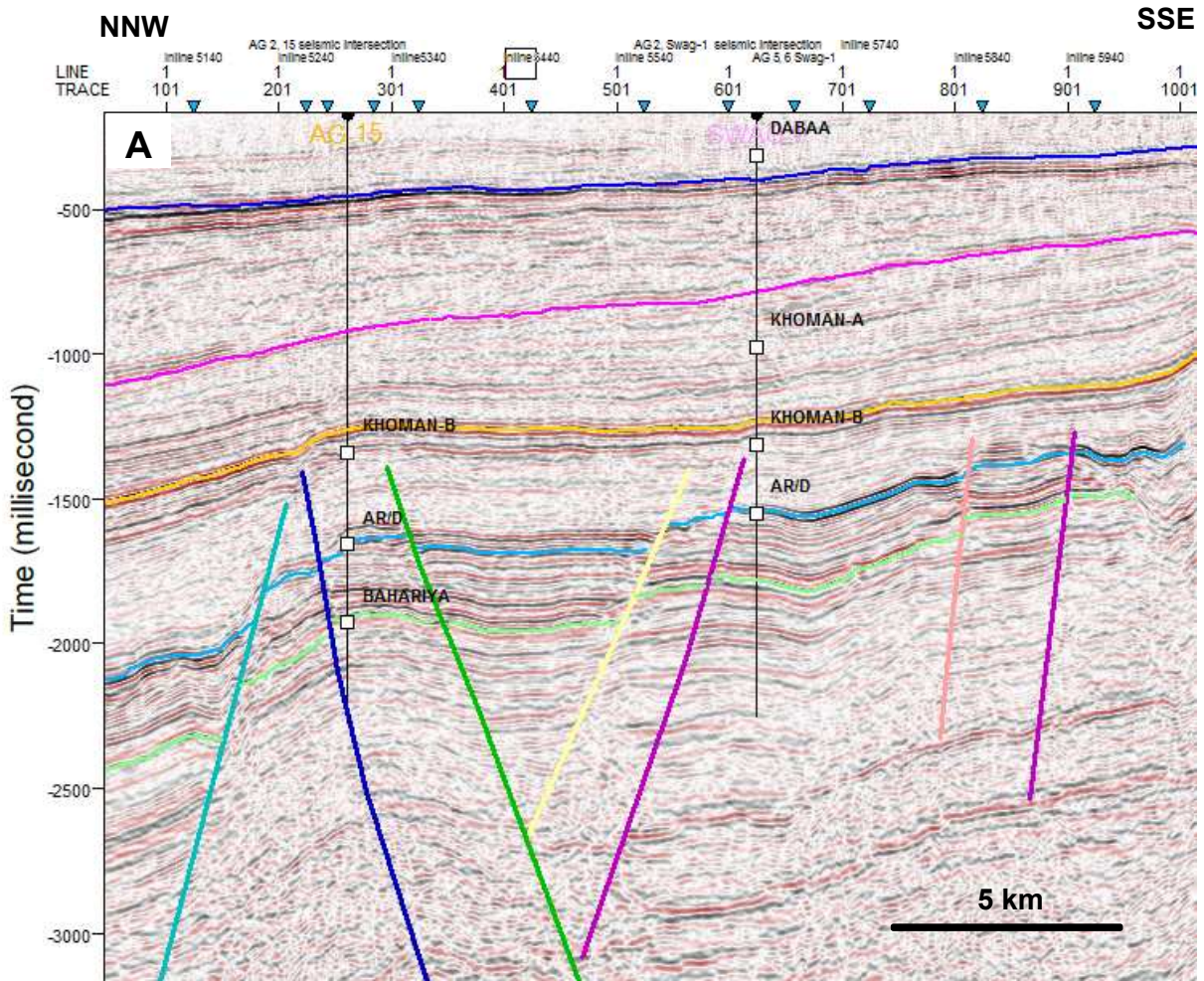


Figure 4a

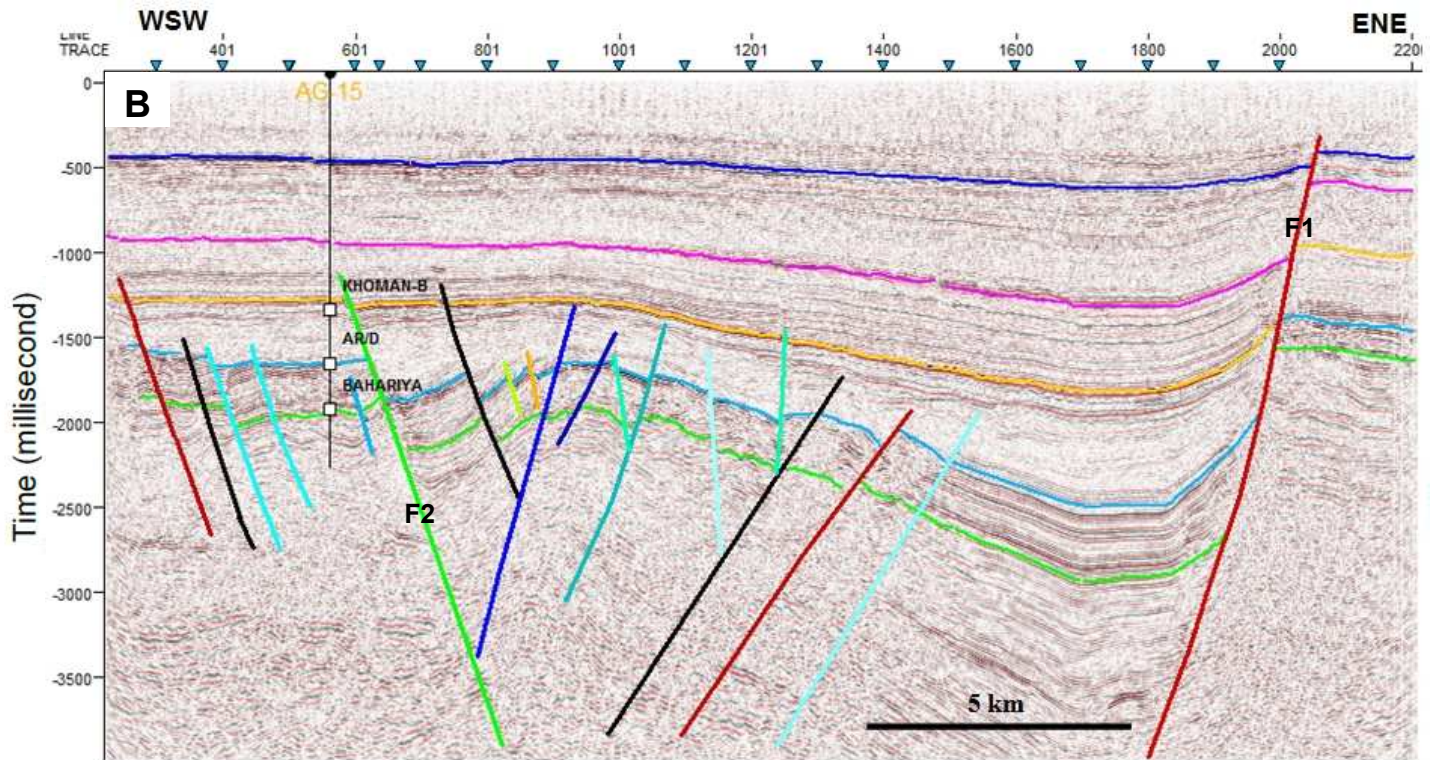


Figure 4b

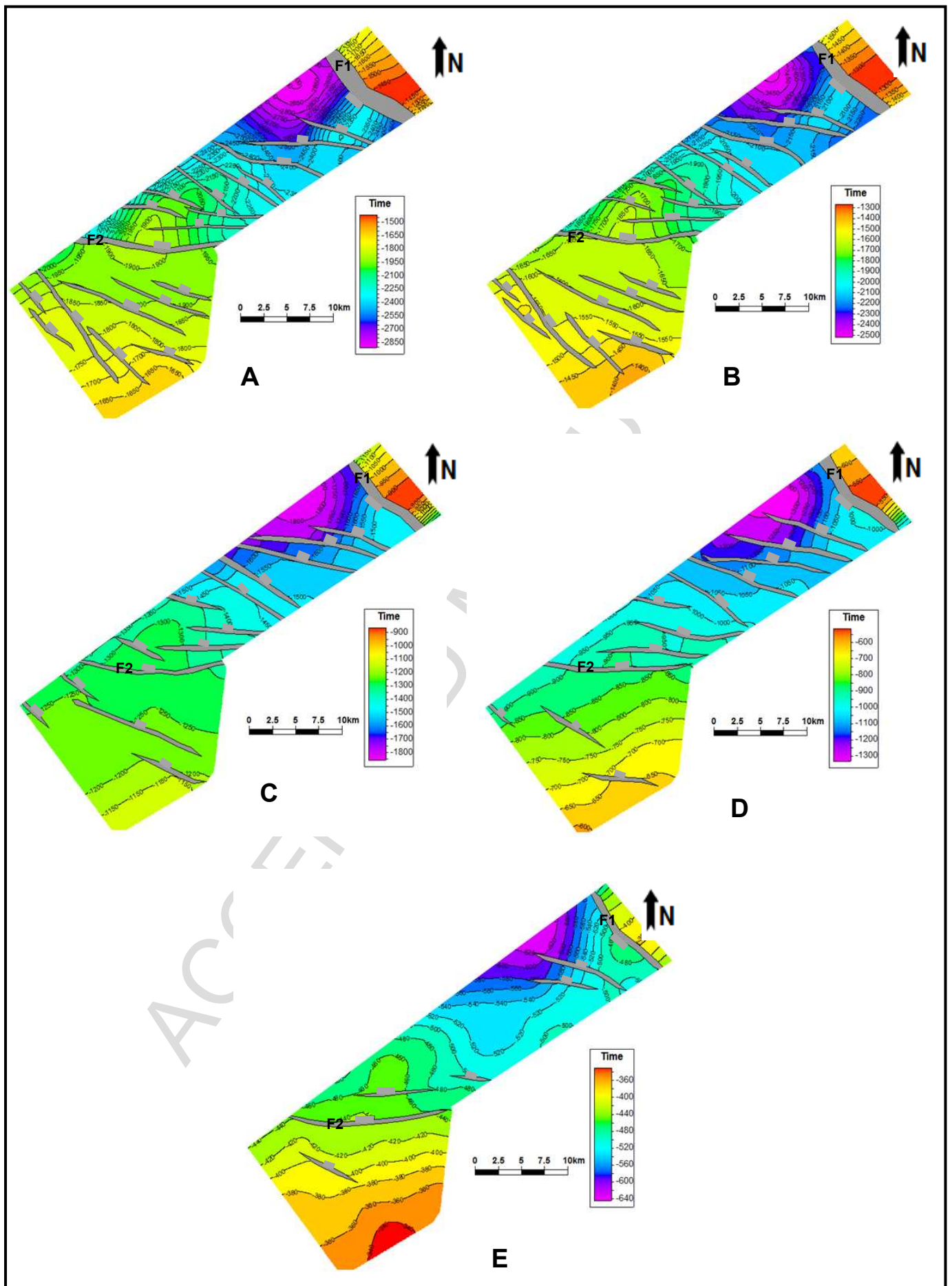


Figure 5

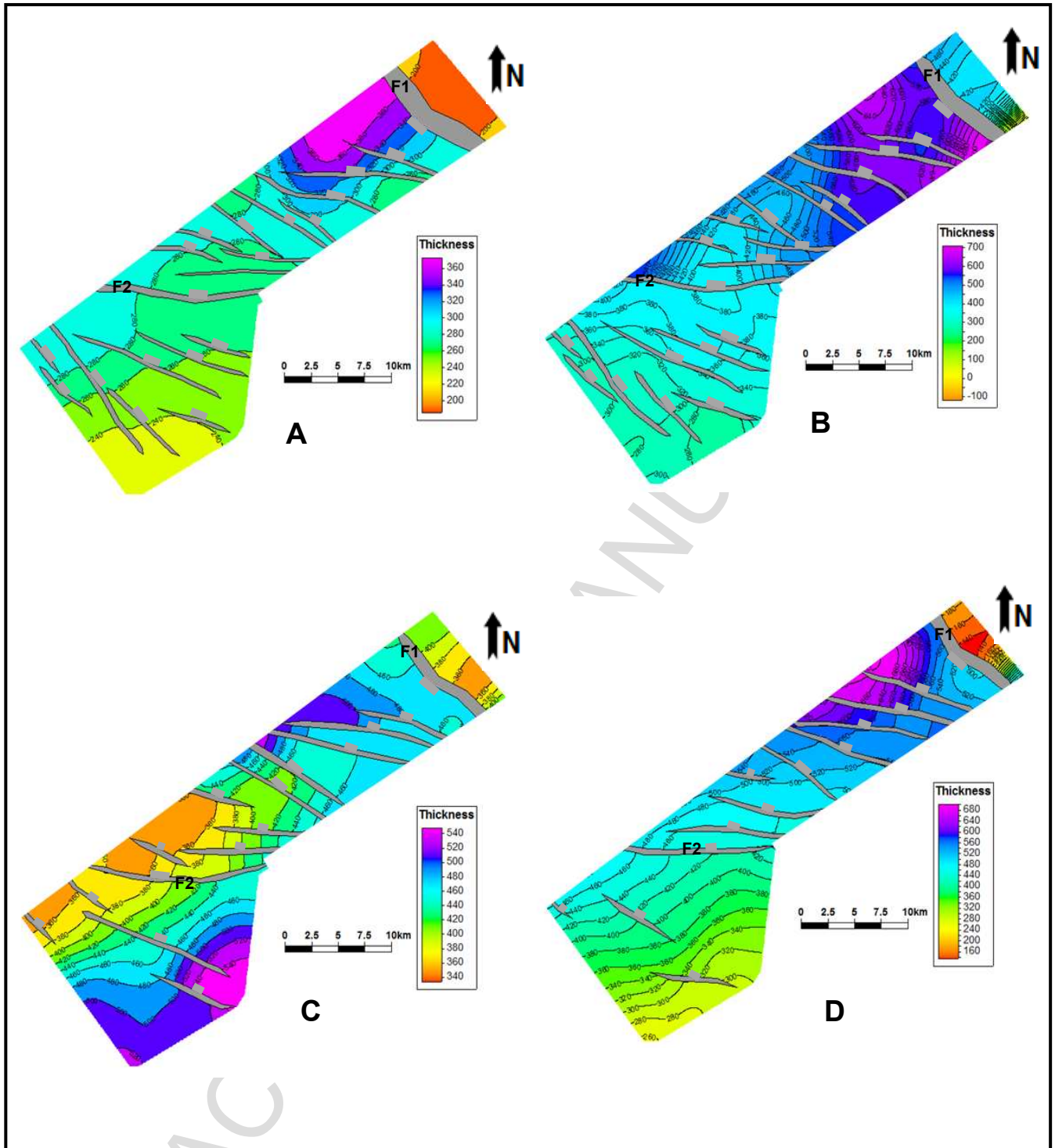


Figure 6

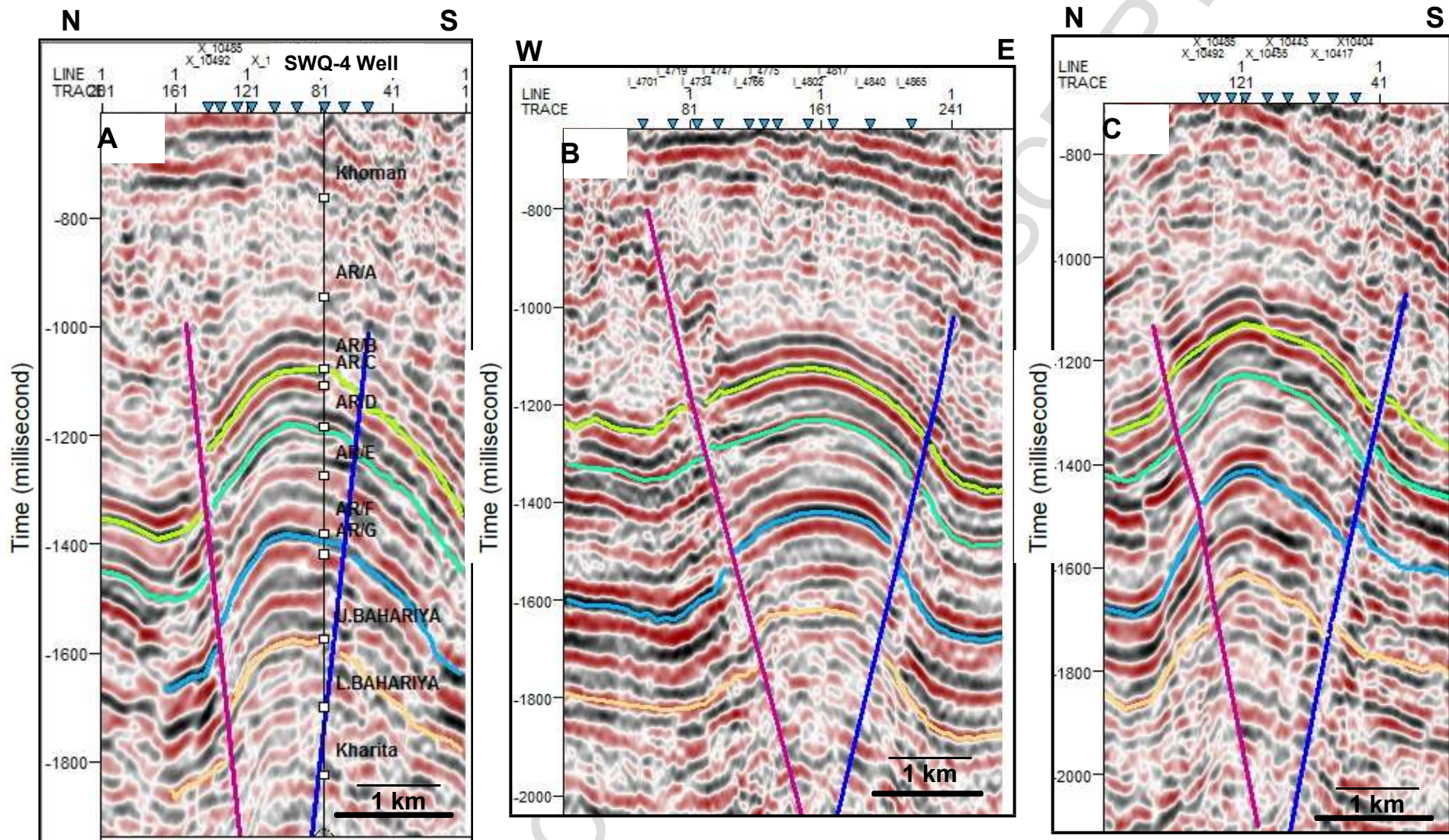


Figure 7

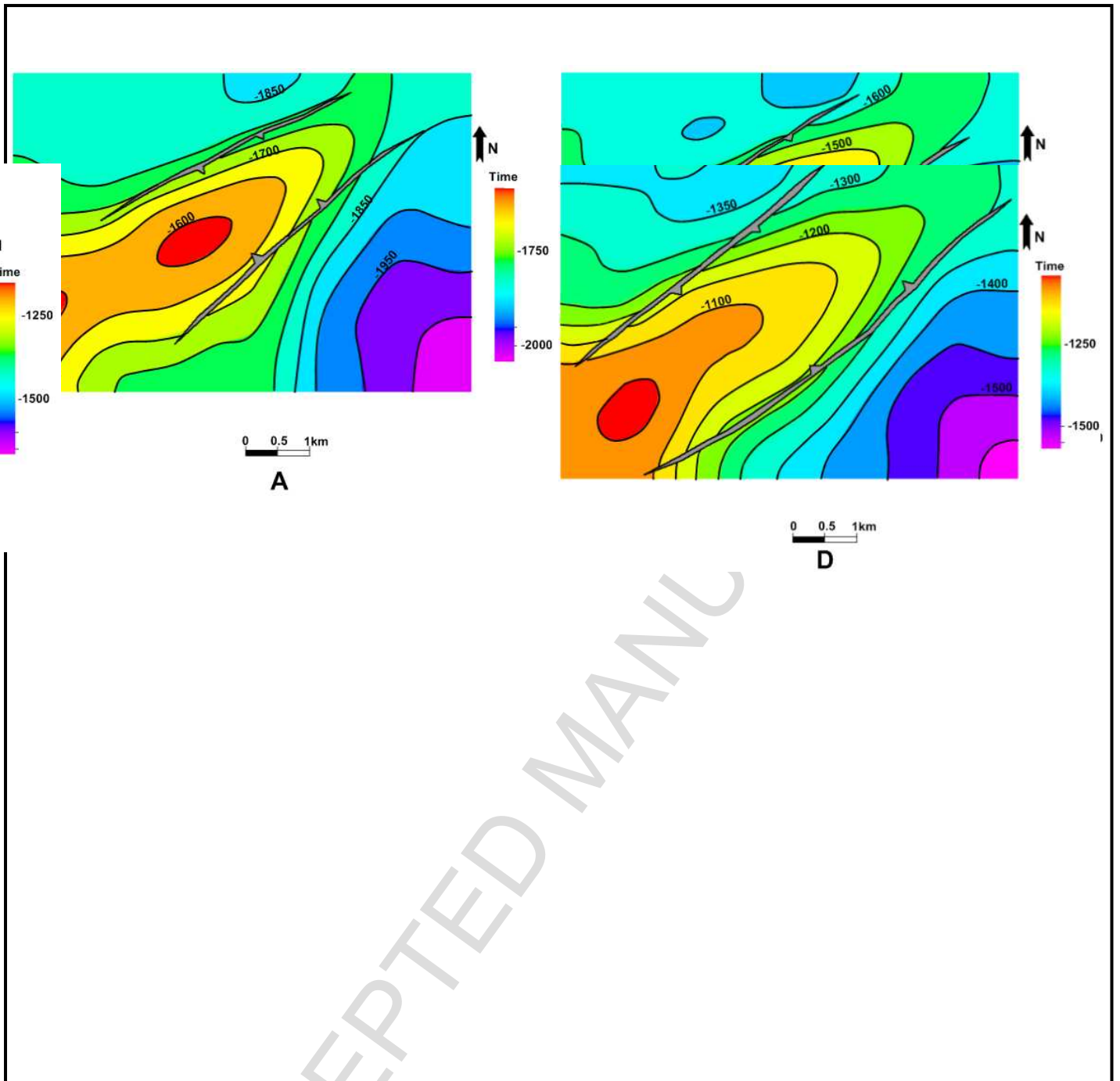


Figure 8

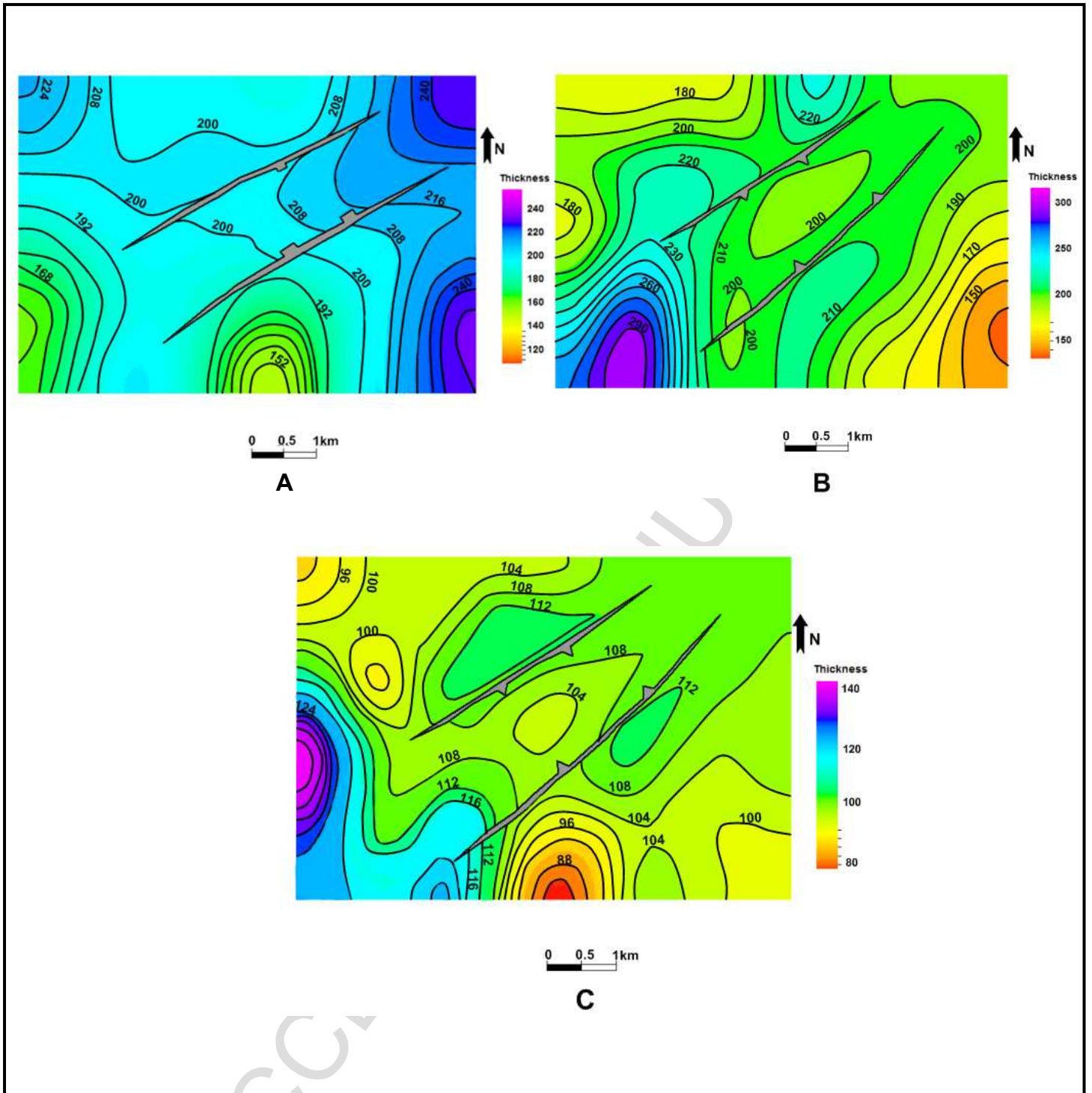


Figure 9

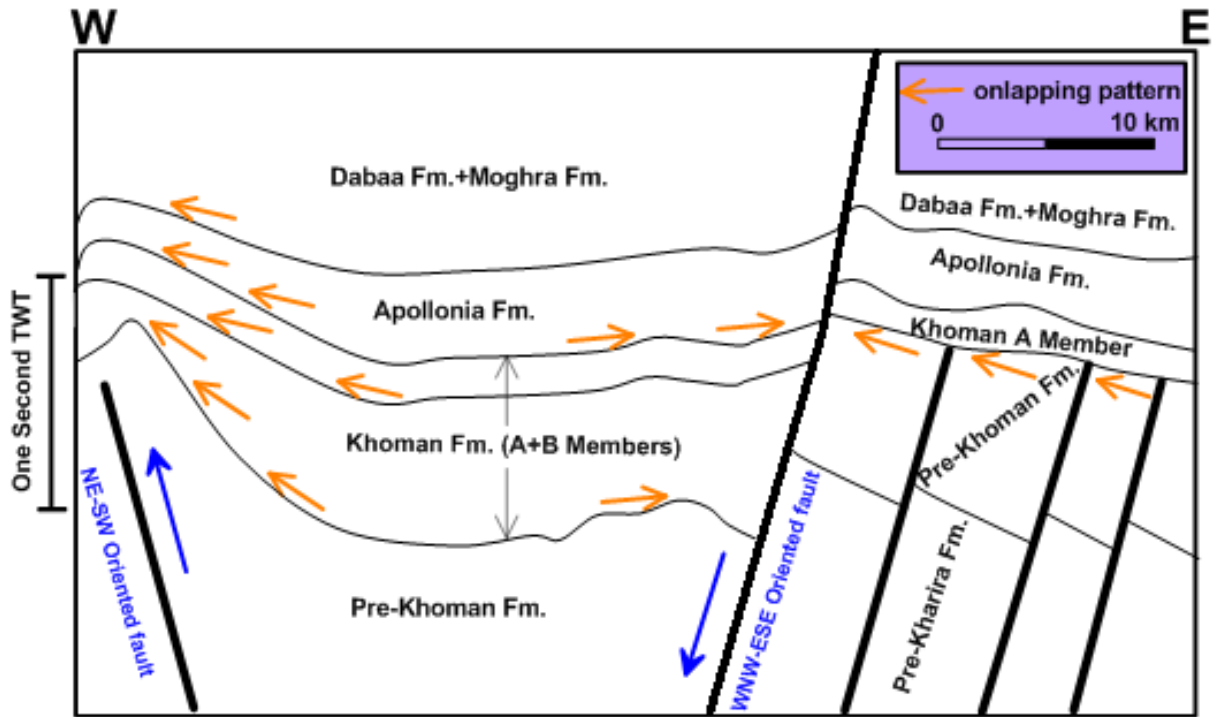


Figure 10

Highlights

- 1- Rift-related and inversion-related anticlines.
- 2- Tectonic inversion in the Abu Gharadig and Gindi Basins.
- 3- Value of mapping at high stratigraphic resolution.