



HAL
open science

Neotectonics of the Owen Fracture Zone (NW Indian Ocean): structural evolution of an oceanic strike-slip plate boundary

Mathieu Rodriguez, Marc Fournier, Nicolas Chamot-Rooke, Philippe Huchon, Julien Bourget, M. Sorbier, Sébastien Zaragosi, Alain Rabaute

► To cite this version:

Mathieu Rodriguez, Marc Fournier, Nicolas Chamot-Rooke, Philippe Huchon, Julien Bourget, et al.. Neotectonics of the Owen Fracture Zone (NW Indian Ocean): structural evolution of an oceanic strike-slip plate boundary. *Geochemistry, Geophysics, Geosystems*, 2011, 12 (1), pp.1-25. 10.1029/2011GC003731 . hal-00643186

HAL Id: hal-00643186

<https://hal.sorbonne-universite.fr/hal-00643186v1>

Submitted on 21 Nov 2011

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **Neotectonics of the Owen Fracture Zone (NW Indian Ocean): structural evolution of an**
2 **oceanic strike-slip plate boundary**

3 *M. RODRIGUEZ^{1,2,3*}, M. FOURNIER^{1,2}, N. CHAMOT-ROOKE³, P. HUCHON^{1,2}, J.*
4 *BOURGET⁴, M. SORBIER³, S. ZARAGOSI⁴, A. RABAUTE^{1,2,3}*

5
6 *(1) Institut des Sciences de la Terre de Paris, UMR 7193, Université Pierre et Marie Curie,*
7 *case 129, 4 place Jussieu, 75252 Paris cedex 05, France*

8 *(2) iSTeP, UMR 7193, CNRS, F-75005 Paris, France*

9 *(3) Laboratoire de Géologie, Ecole normale supérieure, 24 rue Lhomond, 75231 Paris cedex*
10 *05, France*

11 *(4) Université Bordeaux 1, UMR 5805 EPOC, avenue des facultés, 33405 Talence Cedex,*
12 *France*

13
14 **Corresponding author: rodriguez@geologie.ens.fr*

15
16 **Abstract**

17 **The Owen Fracture Zone is a 800 km-long fault system that accommodates the**
18 **dextral strike-slip motion between India and Arabia plates. Because of slow pelagic**
19 **sedimentation rates that preserve the seafloor expression of the fault since the Early**
20 **Pliocene, the fault is clearly observed on bathymetric data. It is made up of a series of**
21 **fault segments separated by releasing and restraining bends, including a major pull-**
22 **apart basin at latitude 20°N. Some distal turbiditic channels from the Indus deep-sea fan**
23 **overlap the fault system and are disturbed by its activity, thus providing landmarks to**
24 **date successive stages of fault activity and structural evolution of the Owen Fracture**
25 **Zone from Pliocene to Present. We determine the durability of relay structures and the**
26 **timing of their evolution along the principal displacement zone, from their inception to**
27 **their extinction. We observe subsidence migration in the 20°N basin, and alternate**
28 **activation of fault splays in the vicinity of the Qalhat seamount. The present-day Owen**
29 **Fracture Zone is the latest stage of structural evolution of the 20-Myr-old strike-slip**
30 **fault system buried under Indus turbiditic deposits whose activity started at the eastern**
31 **foot of the Owen Ridge when the Gulf of Aden opened. The evolution of the Owen**
32 **Fracture Zone since 3-6 Myr reflects a steady state plate motion between Arabia and**
33 **India, such as inferred by kinematics for the last 20 Myr period. The structural**
34 **evolution of the Owen Fracture Zone since 20 Myr— including fault segments**

35 **propagation and migration, pull-apart basin opening and extinction - seems to be**
36 **characterized by a progressive reorganisation of the fault system, and does not require**
37 **any major kinematics change.**

38

39 **1. Introduction**

40 Large strike-slip faults in continental or oceanic domains display a variety of
41 geological features that received considerable attention from researchers [*see Mann, 2007 for*
42 *a synthesis*]. Of particular interest are structures transverse to the main strike-slip fault that
43 take place in areas where the layout of the principal displacement zone is discontinuous or
44 curved. These relay structures accommodate the transfer of slip between adjacent fault
45 segments on both sides of the stepover region. Depending on fault geometry and local stress
46 field, these discontinuities may undergo compression or extension, leading to the formation of
47 a positive relief or a basin, forming a restraining bend or a releasing bend, respectively
48 [*Sylvester, 1988; Cunningham and Mann, 2007*]. Strike-slip stepover regions are often
49 considered as progressively increasing in structural relief (subsidence or uplift) with
50 increasing slip along the principal displacement zone [*Aydin and Nur, 1982; Mann et al.,*
51 *1983*].

52 In contrast, recent studies of the San Andreas fault system [*Wakabayashi et al. (2004)*
53 *and Wakabayashi (2007)*] showed that some stepover regions may migrate along the main
54 strike-slip fault. Further evidence of migration is found in the narrow, elongated Dead Sea
55 pull-apart basin along the Levant fault [*Kashai and Croker, 1987; Ten Brink and Ben*
56 *Avraham, 1989; Garfunkel and Ben Avraham, 1996; Ten Brink and Rybakov., 1999; Smit et*
57 *al., 2008*]. In both cases, little structural relieves are generated as compared to what would be
58 expected from the progressive growth model. Basin migration is in contrast to prior
59 observations that, within stepover basins, mature strike-slip systems tend to develop simple
60 through-going faults that accommodate the relative motion [*Zhang et al, 1989; Le Pichon et*
61 *al., 2001; Rangin and Le Pichon, 2004; Ben-Zion and Sammis, 2003; Wesnousky, 2005;*
62 *Schattner and Weinberger, 2008; Wu et al., 2010; Schattner, 2010; Garcia-Moreno et al.,*
63 *2010*]. These works thus raise the question of the durability of relay structures and the timing
64 of their evolution along the principal displacement zone, from their inception to their
65 extinction.

66 We recently surveyed a major strike-slip feature in the Northwest Indian Ocean, which
67 forms the present day India-Arabia plate boundary: the Owen Fracture Zone (OFZ hereafter)
68 [*Matthews, 1966; Whitmarsh et al., 1979; Gordon and DeMets, 1989; Fournier et al., 2008b,*

69 2011]. The OFZ is a 800 km-long dextral strike-slip fault which connects the Makran
70 subduction zone to the north and the Aden-Owen-Carlsberg triple junction to the south
71 (Figure 1Figure 1). New bathymetry data collected during the Owen and FanIndien 2009
72 cruises reveal that fault scarps are well preserved on the seafloor and can be followed
73 continuously over hundreds of kilometers, pointing to very active tectonics [Fournier et al.,
74 2011]. Although the overall geometry of the OFZ follows a small circle about the India-
75 Arabia pole of rotation (i.e. is a pure transform fault), the fault system displays a succession
76 of releasing and restraining bends, with related tectonic features such as the 20°N pull-apart
77 basin (Figure 1Figure 1). South and North, the OFZ terminates into the Beautemps-Beaupré
78 rhomboidal pull-apart basin and the Dalrymple Trough, respectively (Figure 1Figure 1)
79 [Fournier et al., 2008a, 2011; Edwards et al., 2008]. The objective of this paper is to
80 constrain the timing of the structural evolution of the OFZ over the Plio-Pleistocene period.
81 Since the OFZ is located at the western end of the Indus turbiditic system, distal turbiditic
82 channels are strongly disturbed by neotectonics of this fault system and provide good
83 landmarks to date the successive stages of fault activation and structural evolution of the
84 OFZ. In some areas, transition from mass transport to pelagic deposition mode can also be
85 related to tectonic events and constrain their dating.

86

87 2. Tectonic and kinematic setting

88

89 2.1. Main structural features of the Owen Fracture Zone and present-day kinematics

90 The overall shape of the OFZ is curved (Figure 1Figure 1): the fracture zone azimuth
91 increases progressively from N10°E at latitude 15°30'N (Beautemps-Beaupré Basin) to
92 N31°E at the entrance of the Dalrymple Trough at 22°N. The OFZ closely follows a small
93 circle about the rotation pole determined with GPS and seismicity data, which is consistent
94 with a pure strike-slip motion along the entire fracture zone [Fournier et al., 2011], in contrast
95 with the increasing transtension north of 18°N predicted by the MORVEL closure-enforced
96 pole [DeMets et al., 2010]. The OFZ is made up of six apparently uninterrupted fault
97 segments, ranging in length from 60 to 180 km [Fournier et al., 2011]. The fault system is
98 associated with a major morphological feature, the Owen Ridge, which rises up to 2000 m
99 with respect to the surrounding seafloor. The Owen Ridge is disrupted by two morphological
100 thresholds at 18°10'N and 20°N. The entire plate boundary can be divided into five
101 geographic segments, starting from the Beautemps-Beaupré Basin (Figure 1Figure 1): the
102 southern ridge (300 km long), the central ridge (220 km long), the 20° N pull-apart basin

103 (100 km long), the Qalhat seamount, and the Murray Ridge (which is not covered by our
104 dataset). As indicated by dextral strike-slip focal mechanisms of earthquakes along the OFZ
105 (Figure 1) [Quittmeyer and Kafka, 1984; Gordon and DeMets, 1989; Fournier et al., 2001],
106 the Arabian plate moves northwards slightly faster than the Indian plate with a relative motion
107 of 3 ± 1 mm yr⁻¹ estimated independently from geodetic [Reilinger et al., 2006; Fournier et
108 al., 2008b] and geological [DeMets et al., 1990, 1994, 2010] data.

109

110 2.2. History and origin of the Owen Ridge and the Owen Fracture Zone

111 Offsets on the seafloor imply a finite dextral displacement of 10-12 km along the OFZ
112 (Figures 2, 3) [Fournier et al., 2008b, 2011]. Considering a steady motion of 3 ± 1 mm yr⁻¹,
113 this indicates that the present-day trace of the OFZ has been active since 3-6 Myr. The
114 reconstruction of the India-Arabia motion based on magnetic anomalies suggests that the rate
115 of dextral strike-slip motion remained stable along the OFZ since the first stages of seafloor
116 spreading in the Gulf of Aden [Merkouriev and DeMets, 2006; Chamot-Rooke et al., 2009;
117 Fournier et al., 2010]. This would imply the existence of an at least ~20-Myr-old strike-slip
118 fault system accommodating the Arabia-India motion at the Eastern foot of the Owen Ridge.
119 The trace of the OFZ observed on the seafloor is the latest stage of this strike-slip activity.
120 The southern and central segments of the Owen Ridge were uplifted along this ~20-Myr-old
121 OFZ 19 Myr ago, as attested by the rapid transition from turbiditic to pelagic deposits in
122 DSDP and ODP cores [Whitmarsh et al, 1974; Shipboard Scientific Party, 1989; Weissel et
123 al., 1992]. [Patriat et al., 2008] identify a kinematic change in seafloor spreading on the
124 Southwest Indian Ridge at 24 Ma, and hypothesizes that the collision between Arabia and
125 Eurasia induced a major plate reorganization phase at the scale of the Western Indian Ocean.
126 The uplift and the initiation of seafloor spreading in the Gulf of Aden (dated at 19.7 Myr
127 (Chron 6) [Whitmarsh, 1979; Fournier et al., 2010]) most probably occurred in response to
128 this plate reorganization event. The Dalrymple Trough at the northern end of the OFZ is of
129 Early Miocene age too, with an increase in subsidence observed since the Late Miocene
130 [Edwards et al., 2000; Gaedicke et al., 2002]. OBS data suggest that the trough developed
131 along a small piece of continental crust, probably inherited from the fragmentation of the
132 Gondwanaland [Edwards et al., 2000, 2008; Gaedicke et al., 2002]. At the northern end of the
133 OFZ, the history of the Qalhat Seamount is not clearly established. The nature of the
134 underlying basement remains unknown, as it has never been directly sampled, but the
135 presence of the Little Murray Ridge volcanic seamounts buried under the Oman basin
136 [Gaedicke et al., 2002; Mouchot, 2009], coupled with the existence of a strong magnetic

137 anomaly in the vicinity of the seamount and a typical flat top morphology, strongly suggest
138 that the Qalhat Seamount is a volcanic guyot [Edwards et al., 2000; Fournier et al., 2011].
139 Onlap of Paleocene sediments onto the Qalhat seamount [Edwards et al., 2000, 2008;
140 Gaedicke et al., 2002] demonstrates that the seamount is older than Paleocene.

141

142 2.3. The Owen Ridge and the Indus Fan

143 The Owen Ridge acts as a topographic barrier for the Indus turbiditic system and
144 isolates the Owen Basin, located west of the Owen Ridge, from any sediment supply from the
145 east [Whitmarsh, 1979; Mountain and Prell, 1989]. The Indus turbiditic system covers
146 $1.1 \times 10^6 \text{ km}^2$, stretching 1500 km into the Indian Ocean from the present delta front [Clift et
147 al., 2001]. The oldest sediments drilled suggest a Paleogene age for the emplacement of this
148 sedimentary system [Whitmarsh et al., 1974; Shipboard scientific party, 1989; Weedon and
149 McCave, 1991; Qayrum et al., 1997, Ellouz-Zimmermann et al., 2007]. At its thickest part the
150 fan is more than 9 km thick, but its thickness decreases westward when approaching the
151 Owen and Murray Ridges [DSDP site 222, Shipboard scientific party, 1972; Clift et al., 2001;
152 Calvès, 2008]. Seismic stratigraphy revealed that channel and levee complexes are most
153 pronounced after the Early Miocene, coincident with a sharp increase in sedimentation rates
154 related to the uplift of the Himalaya [Clift et al., 2001]. ODP site 720 located at the southern
155 extremity of the fan documents an alternation of pelagic and short turbiditic episodes over the
156 entire Pleistocene sequence, in response to shifts in the loci of Indus Fan deposition
157 [Shipboard scientific party, 1989; Govil and Naidu, 2008].

158

159 3. Methods and data set

160 3.1. Identification of the structure of the Owen Fracture Zone

161 Bathymetry and acoustic imagery were collected using a Kongsberg-Simrad EM 120
162 echosounder on board the R/V Beautemps-Beaupré operated by the French navy during the
163 Owen and FanIndien 2009 surveys. A DEM at 80 m grid interval was produced as well as
164 mosaics of the acoustic imagery (bottom reflectivity). In addition to bathymetry and
165 reflectivity, the SBP120 sub-bottom profiler coupled with the EM120 provided a set of high
166 frequency (3.5 kHz), high resolution profiles with penetration down to 100 m in fine grained
167 sediments and about 25 m in sand rich floor. Only a AGC has been applied. Because of slow
168 pelagic sedimentation rates along the OFZ since the Early Pliocene [Shipboard scientific
169 party, 1974], the trace of the fault is well preserved on the seafloor. The fault traces observed
170 on both bathymetric data and SBP120 profiles were mapped. Active faults are expected to be

171 identified by a scarp at the seafloor. However, for some of these faults, it is difficult to assess
172 whether they are active or not. Indeed, in areas dominated by pelagic deposition, the pelagic
173 cover takes the exact shape of the underlying morphology and preserves fault offsets through
174 times. Thus the detection of offsets at the seafloor is not by itself an indicator of the present-
175 day activity of the fault. On the other hand, active faults are well observed in areas dominated
176 by turbiditic processes, since their frequency allows a good record of the progressive offset of
177 growth-faults and associated sedimentary tilting.

178

179 3.2. Interpretation of sedimentary deposits on SBP120 (3.5 kHz) profiles

180 We recognize pelagic and turbiditic deposits in the sub-bottom high resolution profiles
181 based on their seismic characters. Where SBP120 profiles display well stratified, continuous
182 and conformable horizons, the deposits are interpreted as pelagic in origin. This interpretation
183 is supported by correlation with DSDP site 222 drill hole (see [Figure 1](#) ~~Figure 1~~ for location),
184 and by the analysis of a Küllenberg core [*Bourget et al., submitted*]. Where SBP120 profiles
185 display well-stratified horizons with successions of high and low amplitude reflections, or
186 thick transparent horizons in pull-apart areas, the deposits are interpreted as turbiditic in
187 origin. This interpretation is also supported by the analysis of a Küllenberg core (Figure 4 for
188 location) [*Bourget et al., submitted*].

189

190 3.3. Age estimates

191 Several DSDP and ODP drillings are available in our studied area. Estimated
192 sedimentation rates for the Pliocene pelagic interval are used to date the tectonic and
193 sedimentary events we observe on SBP profiles. All pelagic sedimentation rates estimated at
194 different drilling sites range between ~30 and ~50 m.My⁻¹ [*Shipboard scientific party, 1974,*
195 *1989*]. There is little spatial variation of Plio-Pleistocene pelagic sedimentation rates along the
196 Owen Ridge, allowing large scale interpolation of these values. In this study, we use
197 sedimentation rates estimated at site DSDP 222, because of its proximity to the OFZ.
198 Sedimentation rates at site DSDP 222 quoted between 43 and 53 m.My⁻¹ for the last 2 Myr
199 interval [*Shipboard scientific party, 1974*]. Two-way travel time to seismic reflectors was
200 converted to depth using a P-wave velocity of 1500 m.s⁻¹.

201 A different approach based on the analysis of Küllenberg cores is used to estimate the
202 age of turbiditic deposits in the 20°N pull-apart basin. Küllenberg cores were sampled in areas
203 where turbiditic deposits are the thinnest in order to sample the highest number of turbiditic
204 events. The high resolution of SBP 120 profiles allows correlation of turbiditic reflectors with

205 turbiditic deposits identified on Küllenberg cores. Turbiditic deposits are dated using the ^{14}C
206 method on planktonic foraminifera sampled in undisturbed layers of the core. ^{14}C ages
207 obtained for the first turbiditic events are extrapolated to deeper horizons using the constant
208 tilting rate of turbiditic horizons in the 20°N Basin. The method and the role of relative sea
209 level variations in the turbiditic infill of the 20°N Basin are detailed in [Bourget *et al.*,
210 *submitted*].

211

212 **4. Results**

213 In this section, we describe the different segments of the OFZ fault system. We first
214 focus on the dominantly strike-slip portions, and then describe stepover areas. Each
215 description focuses on the geomorphic expression of the OFZ, in relation with the turbiditic
216 channels from the Indus Fan where present. We describe channels whose activity can be used
217 as a time indicator for the relative and/or absolute age of the deformation in stepover areas.
218 The depth of pull-apart basins is given with respect to the surrounding seafloor.

219

220 *4.1. Dominantly strike-slip segments: the south and central Owen Ridge*

221 The southern Owen Ridge consists in a 300 km-long, 50 km-wide, and 2000 m-high,
222 asymmetric relief, with a steep east-facing scarp associated with the OFZ and a gentle western
223 slope dissected by numerous landslides [Rodriguez *et al.*, *submitted*] (Figure 2). To
224 the south, the OFZ offsets the southern ridge dextrally over 12 km (Figure 2) [Fournier *et al.*,
225 2008b]. The area shows no evidence of vertical motion on the bathymetric data, suggesting
226 dominantly pure strike-slip motion along this segment.

227 Northwards, a minor restraining bend is observed on the Indian plate side at the
228 northern end of the southern ridge (Figure 2). The restraining bend corresponds to a
229 series of gentle folds related to the slight deviation of the OFZ trend between 17°N and 18°N.
230 At the same latitude, the vertical throw of the OFZ is maximum. The folds generate a relief up
231 to 120-m-high (Error! Reference source not found. Figure-2). A similar gentle fold has
232 been observed on ODP seismic profiles [Shipboard scientific party, 1989], which appears to
233 have been active during the entire Pleistocene, accommodating about 150 m of vertical
234 motion. Thus a small component of fault-normal compressional motion can be inferred in this
235 area.

236 The central Owen Ridge is a 220-km-long, 50-km-wide, and up to 1700-m-high relief
237 (Figure 3). The central ridge displays an irregular morphology, with a 1300 m-high plateau
238 spreading over 485 km² and seamounts dissected by complex networks of gullies and

239 landslides [*Rodriguez et al., submitted*]. The OFZ crosscuts the central ridge between 18°50'N
240 and 19°35'N and offsets it dextrally over 10 km (Figure 3). Southeast of the central ridge (at
241 latitude 18°40'N), a 7-km-wide stepover between two segments of the OFZ gives birth to a
242 releasing bend associated with a small 25-km-long and 90-m-deep pull-apart basin.

243

244 4.2. The 20°N pull-apart basin releasing bend

245 The 20°N Basin is 90-km-long and up to 35-km-wide pull-apart basin (**Figure**
246 **4Figure-4**). The basin developed in a 12-km-wide stepover between two major strike-slip
247 faults trending N25°E south of the basin and N30°E north of it. The overall structure of the
248 basin is asymmetric, with the OFZ as a steep master fault on the western side, and a single
249 normal fault dividing northwards into several arcuate splays on the eastern side. The throw of
250 arcuate normal faults clusters between 100 m and 300 m, and the associated topographic
251 uplift decreases northwards as the number of faults increases, suggesting partitioning of local
252 extensional deformation. To the southeast of the 20°N basin, a system of en-échelon faults,
253 oriented 25° clockwise with respect to the OFZ is observed between 19°18'N and 19°45'N
254 (Figure 4).

255 The bathymetric map shows three distinct sub-basins (labelled SB1, SB2, and SB3 in
256 **Figure 4Figure-4**). SB1 and SB2 extend over 70 km² and 340 km², respectively. SB3 extends
257 over 590 km². E-W-trending transverse normal faults separate the sub-basins. The 20°N basin
258 deepens abruptly northwards, as sub-basins 1, 2, 3 form terraces respectively 60, 100, and
259 360 m deep with respect to the surrounding seafloor (Figure 5). Pelagic deposits blanket SB1
260 and SB2, as well as the half grabens of the eastern hanging wall, whereas recent turbiditic
261 deposits filled in SB3 [*Bourget et al., submitted*]. In SB3, Indus turbiditic deposits adopt
262 vertically a fanning configuration on the longitudinal SBP120 profile (**Figure 5Figure-5**).
263 The turbiditic infill forms a subtle and asymmetric syncline on transverse SBP120 profiles
264 (**FigureFigure-6**). A change of sense of asymmetry occurs along strike within SB3, where
265 horizons within the southern part of the basin dip toward the west (profile P07b), whereas in
266 the northern section, horizons dip to the east (profile P07a).

267 Although SBP120 profiles do not reach the substratum of the 20°N basin, they give
268 some insights about its deep structure. Turbiditic horizons onlap tilted blocks at the southern
269 end of the 20°N basin (**Figure 5Figure-5**). Turbiditic reflectors are slightly disturbed in the
270 northern part of the 20°N basin, as a series of small buried benches, probably related to the
271 activity of a minor normal fault (Figure 5).

272

273 4.3. Age of the Indus turbiditic channels affected by the opening of the 20°N basin

274 The opening of the 20°N basin affected the evolution of seven turbiditic channels
275 belonging to three distinct generations labelled A, B, and C in younging order (Figures 5, 7).
276 These channels are located 600 km away from the Indus canyon and are part of the distal
277 Indus fan.

278

279 *Channel A-generation before the onset of the Owen fracture zone (>3-6 Myr old)*

280 The oldest channel (A1) is located in the Owen Basin and is now disconnected from
281 the Indus fan. This highly meandering channel is now abandoned and covered by a thick
282 (>100 m) pelagic drape (**Figure 4**~~Figures~~–4, 7). No trace of activity of this channel is
283 observed on SBP120 profiles. Thus, the channel is older than the oldest sub-bottom reflector,
284 the age of which is estimated at 3 Myr based on sedimentation rates at DSDP site 222. It is the
285 minimum age for this channel, which could be older. The presence of channel A1 in the Owen
286 Basin indicates that no significant relief acting as a topographic barrier for sediments coming
287 from the Indus existed during its activity. This channel thus pre-dates the onset of the present-
288 day trace of the OFZ.

289 The meandering channel A2 is supposed to be of the same generation, since no trace
290 of channel activity is observed on SBP120 data (**Figure 7**). Arcuate normal faults of the 20°N
291 basin crosscut channel A2, the channel being now located on an uplifted area. We thus infer
292 that the uplift post-dates the channel activity.

293 Other channels (group A3) that do not display any trace of activity at the depth of
294 SBP120 penetration are located in the Indus abyssal plain (**Figure 4**~~Figure~~–4). This group of
295 channels was subjected to avulsion. As normal faults dissected the bifurcation point, it is
296 difficult to discriminate which channel predates the other. Channel avulsion is known to be an
297 autocyclic process that spans thousands of years [*Babonneau et al., 2002*], but it can be driven
298 by tectonics processes related to the tectonic evolution of the 20°N basin which spans millions
299 years. For that reason, we cannot use the occurrence of avulsion as a tectonic indicator. The
300 channels run parallel to the OFZ and do not seem to have been “attracted” by the presence of
301 any topographic depression during their activity. A normal fault of the 20°N Basin dissects
302 the eastern channel A3. Thus the channel activity seems to pre-date the opening of the basin,
303 and is probably coeval with the channel A1. We infer that the group of channels A3 could be
304 the southward prolongation of channel A2.

305

306 *Channel B-generation and the opening of the southern sub-basins of the 20°N basin*

307 The generation B is composed of three channels. The paleo-activity of these channels
308 is marked by a lens-shaped, rough and chaotic facies on SBP120 profiles (**Figure 7**). Two of
309 them (B1 and B2) are located at the southernmost part of the 20°N basin (**Figure 4****Figures 4,**
310 5). The normal fault bounding the southernmost sub-basins 1 and 2 dissects channel B1,
311 which is also incised by the system of en-échelon faults to the east (**Figure 4****Figures 4, 8**).
312 The complex arcuate normal fault system that borders the eastern side of the 20°N basin
313 strongly dissects channel B3. Small relics of channel B3 are also observed close to the
314 currently active channel on bathymetry. A thick pelagic drape covers lens-shaped features
315 since 1.32 to 0.95 Myr according to sedimentation rates at DSDP site 222.

316

317 *Channel C-generation and the filling of the present-day 20°N basin*

318 The currently active channel of the 20°N basin (generation C) is recognisable on the
319 seafloor morphology because of its deep incision through the complex normal fault system
320 and its highly meandering sinuosity (**Figure 5****Figures 4, 7**). It displays a U-shape in cross
321 section. The incision becomes wider at the mouth of the channel. A transparent body is
322 observed on both sides of the channel, and is interpreted as a HARP (high amplitude
323 reflections package) incised by the channel (**Figure 8**). The path of channel C was probably
324 imposed by the growing topography formed by normal faults footwalls.

325

326 *4.4. Present-day deformation east of the Qalhat Seamount*

327 The northern Owen Ridge, or Qalhat Seamount, is a volcanic guyot that extends
328 between 20°30'N and 22°10'N (**Figure 9**). It is a 210-km-long, more than 55-km-wide,
329 and up to 2700-m-high relief, characterized by a flat and 400-m-deep top. It is eroded by
330 complex networks of mass wasting features [Rodriguez *et al.*, *submitted*]. The OFZ runs on
331 the eastern side of the Qalhat Seamount, where it is divided into three major splays. The
332 principal splay (labelled 1 in **Figure 9**) trends N 30E° and bends to N20E° at latitude
333 21°30'N. A second splay (labelled 2 in **Figure 9**), located west of the first one, also
334 trends N30E°. Those two splays (1 and 2) delimitate a 160-m-deep, 45-km-long and up to 20-
335 km-wide pull-apart basin. We propose the name of "Qalhat Basin" for this still-unnamed
336 feature. The third splay (labelled 3 in **Figure 9**) is composed of an en-échelon fault
337 system, which trends 14° to the east with respect to the main OFZ (splay 1). At latitude
338 21°35'N, a 700-m-high and arcuate relief interrupts the en-échelon fault system, which merges
339 northwards with the Dalrymple Trough.

340 The deep structure of this portion of the OFZ has been partly imaged by seismic lines
341 collected by the RSS Charles Darwin (**Figure 3Figure-10**) [*Edwards et al., 2000, 2008*]. As
342 suggested by strong magnetic anomalies, seismic lines confirm that an eastward prolongation
343 of the Qalhat Seamount buried under Indus turbiditic deposits underlies the OFZ. Thus, the
344 arcuate relief that disrupts the en-échelon fault system (labelled 3 in **FigureFigure-9**) is
345 volcanic in origin. The OFZ forms a flower-structure in this area, with several sub-parallel
346 vertical splays imaged on seismic profile (Figure 10). The currently active fracture zone is one
347 of the several splays of the flower structure.

348 The Qalhat Basin is the most conspicuous tectonic feature in this area. It displays a
349 needle geometry, with a very elongated depocentre. Arcuate normal faults with vertical
350 throws of 20 to 70 m delimitate two half grabens on the eastern side of the basin. The active
351 OFZ crosses the basin and has accommodated 30 m of vertical slip since its onset, suggesting
352 a negligible extensional component to the strike-slip motion in this area. Complex slope
353 failures are observed on both sides of the Qalhat Basin, which is a catchment for mass
354 transport deposits. Two distinct seismic facies observed on the SBP120 profile that crosses
355 the half graben allow to distinguish two modes of sedimentary deposition in this basin
356 (**Figure 4Figure-11**). The transparent and thick body at the bottom is interpreted as the result
357 of mass transport, whereas the covering well-stratified facies is interpreted as pelagic
358 deposits. The currently active depocentre displays a low reflectivity facies, whereas half-
359 grabens display a highly reflective facies (**Figure 5Figure-12**). This indicates that the active
360 depocentre is filled in by mass transports deposits from the Qalhat Seamount, whereas the
361 elevated graben is now isolated from such supply and mainly filled in by pelagic deposits,
362 which imprint their signature on seafloor reflectivity.

363 Two turbiditic channels belonging to the Indus system are of particular interest with
364 regards to the tectonic activity of the OFZ. The first channel (labelled A on **FigureFigure-9**)
365 is dissected by an en-échelon fault system. No trace of its activity is observed on SBP120
366 profiles, suggesting an age older than 3 Myr. The second channel (labelled B on
367 **FigureFigure-9**) is dissected by the two splays of the OFZ (labelled 1 and 2 in Figure 10) that
368 border the current depocentre of the Qalhat Basin. The western splay of the OFZ (labelled 2
369 in **FigureFigure-9**) crosscuts some of the meanders of channel B. It indicates that the channel
370 used to pass west of the splay 2. Thus, the channel pre-dates the onset of the OFZ in this area.
371 These turbiditic channels could possibly be coeval with turbiditic channels identified West of
372 the Dalrymple Trough and at the top of the Murray Ridge [*Ellouz et al, 2007; Mouchot,*
373 *2009*].

374 Since Indus turbiditic channels on the north-western side of the Dalrymple Trough all
375 pre-date the onset of the trough [Ellouz *et al.*, 2007; Mouchot, 2009] and since the turbiditic
376 system of the Makran accretionary wedge does not form any channel [Bourget, 2009; Bourget
377 *et al.*, 2010], they cannot be used as time indicators for the propagation of the deformation
378 towards the Dalrymple horsetail structure.

379

380 **5. Summary of the constraints on the age of the deformation along the Owen Fracture** 381 **Zone**

382 In this section we discuss the age and relative chronology of deformation along the
383 different segments of the OFZ. In areas where different scenarii are possible, we decipher
384 which one is the most likely in order to propose a detailed history of the structural evolution
385 of the OFZ since the Early Pliocene.

386

387 *5.1. Age of deformation in the restraining bend (southern Owen Ridge)*

388 The fold observed on ODP seismic line [Shipboard scientific party, 1989] along the
389 main restraining bend of the southern Owen Ridge offers a good opportunity to follow the
390 deformation rate through time. The fold has been active during the entire Pleistocene. The
391 most recent episode of folding is observed on SBP profile that crosses the restraining bend
392 located between 17°N and 18°N (Figure 3). The restraining bend consists in two folds. A
393 quantitative analysis of the folding rate using sedimentation rates estimated at site ODP 720
394 (32 m.Myr⁻¹) suggests an increase around 0.8 Myr followed by a smooth and constant
395 deformation.

396

397 *5.2. Age of deformation in the vicinity of the 20°N basin*

398 The best chronological record comes from the 20°N pull-apart basin. The onset of the
399 OFZ and the opening of the 20°N basin clearly post-date the activity of channels of the
400 generation A (Figure 7, Figures 4, 7). The first stage of deformation is an uplift associated
401 with normal fault activity. The uplift is recorded at 1.8-1.5 Myr at DSDP site 222, and marks
402 the transition from turbiditic to pelagic deposits [Shipboard scientific party, 1974]. This could
403 be a local uplift associated with the local formation of a new fault. However, no trace of
404 channel activity of the B generation is observed either on seafloor bathymetry or on SBP120
405 profiles in the Owen Basin, west to the 20°N Basin. These could be proofs of an incipient
406 20°N basin where channels of the B generation used to discharge turbiditic material. The
407 second stage of extensional deformation corresponds to the development of a series of half

408 grabens bounded by the system of en-échelon normal faults between 19°18'N and 19°45'N
409 (Figure 4Figures-4, 8). The onset of normal faults seems to be responsible of the channel
410 deactivation, as all the channels of the B generation are faulted. Thus, dating the deactivation
411 of channels, i.e. the base of the pelagic blanket that covers lens-shaped features, gives the age
412 of the deformation. If faulting is not responsible of channel deactivation, then faults emplaced
413 latter than the age of the base of the pelagic blanket that covers lens-shaped features, which is
414 therefore the maximal age of faults. The age of the base of the pelagic blanket that seals traces
415 of channel activity is estimated at 1.32 to 0.95 Myr using sedimentation rates calculated at
416 DSDP site 222 (Figure Figures-7, 8). Arcuate normal faults on the hanging wall of the 20°N
417 Basin and the en-échelon fault system must be the latter stage of the normal fault system that
418 started to develop at 1.8-1.5 Myr. With regards to the sub-basin SB3, the extrapolation of the
419 ages of turbiditic deposits measured on Küllenberg core samples gives an age of 420 kyr for
420 the deepest sub-bottom reflector imaged (Figure 13) [Bourget *et al.*, *submitted*]. Thus, the
421 sub-basin SB3 is at least 400 kyr old.

422

423 *5.3. Age of deformation in the vicinity of the Qalhat Seamount*

424 All the turbiditic channels in this area pre-date the onset of the present OFZ and the
425 opening of the Qalhat basin (FigureFigures-9, 14). We could not assess with field evidences
426 the age of the en-échelon fault system labelled “3” in FigureFigure-9. Field evidences are
427 compatible with two different scenarii. On one hand, the en-échelon fault system could be the
428 very early stage of the present-day OFZ. The alternative is that it is the location of the future
429 OFZ.

430 In the first case (Figure 14), the horsetail termination develops in the continuity of the
431 present-day OFZ and not along the en échelon-fault system, suggesting that the formation of
432 the en-échelon fault system predates the formation of the horsetail structure. This scenario is
433 consistent with several analog modelling works [Aydin and Nur, 1982; Rahe *et al.*, 1998,
434 Basile and Brun, 1999; Schlische *et al.*, 2002; Wu *et al.*, 2010] which show that en-échelon
435 fault systems are usually emplaced at the very early stage of strike-slip deformation. In the
436 second case, the direction of the en-échelon fault system is more consistent with the small-
437 circle predicted for a pure-strike-slip motion than the direction of the present-day OFZ
438 [Fournier *et al.*, 2011]. In this latter case, the en-échelon fault system is considered as the
439 incipient stage of the future OFZ. If confirmed, this scenario would lead to a reorganization of
440 the horsetail termination.

474 The study of the sedimentary infill of the Qalhat Basin can be used to decipher the
475 relative chronology of the deformation in this area. Indeed, the presence of turbiditic or mass
476 transport deposits sealed by a pelagic cover on the graben of the Qalhat Basin (Figure 12)
477 indicates that the graben used to be a part of the Qalhat Basin depocenter, and has been
478 subsequently slightly uplifted (by about 30 m). Thus, the Qalhat Basin used to be wider, and
479 has been restrained by the onset of the currently main arm of the fracture zone which cross-
480 cuts diagonally the basin (**Figure Figures-9**, 11). We assume that the onset of the diagonal
481 fault is responsible of the graben isolation from mass transport deposits. The age of the base
482 of the isolated graben pelagic cover - and thus the age of the Qalhat Basin extinction- ocured
483 0.7 to 0.6 Myr ago assuming pelagic sedimentation rates close to the ones estimated at DSDP
484 site 222 [*Shipboard scientific party, 1974*] (Figures 11, 14).

485 The uplift observed at the southern end of the Qalhat Basin is likely to post-date the opening
486 of the Qalhat Basin, as the related normal faults are uplifted too. The age of the extinction of
487 the Qalhat Basin is nearly coeval with the onset of the increase in folding observed at the
488 restraining bend along the southern ridge. The uplift could be due to local compression along
489 the OFZ, which could be responsible for the basin extinction too.

490

491 5.4. Summary of the history of the Owen Fracture Zone for the last 3-6 Myr

492 Although the present trace of the OFZ on the seafloor is only 3-6 Myr old, kinematic
493 studies suggest that the relative motion between India and Arabia plates is located along the
494 Owen Ridge since the opening of the Gulf of Aden 20 Myr ago [*Chamot-Rooke et al., 2009*;
495 *Fournier et al., 2010, 2011*]. Thus the present-day OFZ is the exposed part of a strike-slip
496 corridor which is partly buried under the Indus deep-sea fan. The following chronology
497 focuses on the structural evolution of the exposed part of the OFZ whose activity spans the
498 Plio-Pleistocene period.

499 1) The Present-day trace of the OFZ emplaced about 3 to 6 Myr ago, as deduced from
500 finite offsets on the seafloor [*Fournier et al., 2011*]. It runs at the foot of the eastern flanks
501 of the southern and the central ridges since this time (**Figure 2Figure 2, Figure Figure**
502 **3**). A horsetail termination (**FigureFigure-9**) and a rhomboidal pull-apart basin (**Figure**
503 **2Figure-2**) began to develop at the northern and southern ends of the fracture zone,
504 respectively [*Edwards et al., 2000, Fournier et al., 2008a*]. The onset of the horsetail
505 termination could be related to the increase in subsidence of the Dalrymple Trough
506 observed at ~5 Myr by *Gaedicke et al. [2002]*.

Formatted: Font
or grammar

- 507 2) 1.8 to 1.5 Myr ago, a tectonic activity possibly related to the first stages of formation of
508 the 20°N Basin is observed on DSDP drilling [*Shipboard scientific party, 1974*].
- 509 3) An en-échelon fault system and asymmetric transform basins were emplaced 1.3 to
510 0.95 Myr ago at 20°N (**Figure 4** ~~Figures~~ 4, 8, 15).
- 511 4) A small shortening phase is recorded at the northern extremity of the southern ridge
512 0.8 Myr ago. It could be associated to a slight deviation of the OFZ from its previous
513 trace. It is the last step of a more regional shortening, which started in the Early
514 Pleistocene [*Shipboard scientific party, 1989*].
- 515 5) It is still unclear when the Qalhat Basin opened, and when the uplift to the south of it
516 occurred. However, it was intersected by the master OFZ fault 0.7 to 0.6 Myr ago
517 (**Figure** ~~Figure~~ 9). The uplift south of the Qalhat Basin could be coeval with the
518 shortening phase identified on the southern ridge.
- 519 6) Since its opening, the subsidence of the 20°N basin has been relocated in different sub-
520 basins. The currently active sub-basin is at least ~400 kyr old (**Figure 4** ~~Figures~~ 4-5, 13).

521 Uncertainties remain in the dating exposed above. The age of the en-échelon fault
522 system to the east of the Qalhat Basin, and the age of the uplift south of it, are not assessed by
523 clear field evidences, and the timing of 20°N sub-basins migration needs to be further
524 constrained. This summary must be considered as the most likely relative ages with regards to
525 the general tectonics of the OFZ.

526

527 **6. Discussion**

528 6.1. *Does the structure of the Owen Fracture Zone reflect a steady-state India-Arabia*
529 *relative plate motion?*

530 The structures that we describe above are those formed along a dextral strike-slip
531 boundary since at least 3 Ma. The relative motion between India and Arabia is small (3 ± 1
532 mm/yr), and consequently very sensitive to any change of plate velocities. Indeed, a slight
533 decrease of the northward motion of Arabia would have inverted the motion along the OFZ
534 from dextral to sinistral, which is not observed. We thus conclude that the relative motion has
535 been stable during the Pliocene, which is further supported by the agreement between
536 geodetic and geologic kinematic modelling [*Fournier et al., 2008b; DeMets et al., 2010*]. The
537 finite offset itself (10 to 12 km) is compatible with a constant shear rate during the same
538 period of time (about 3 mm/yr). Taken all together, this confirms that the major structures that
539 we observe along the OFZ, and in particular the pull-apart basins, are younger than 3-6 Myrs.

540 On the other hand, several observations suggest that the dextral strike-slip motion
541 started much earlier, probably when oceanic accretion initiated in the Gulf of Aden. This is
542 supported by kinematics reconstructions [*Chamot-Rooke et al., 2009*] that suggest an
543 initiation of the dextral motion around 17 Ma or slightly earlier, and a finite differential
544 motion between Arabia and India of the order of 80 km. Furthermore, the amount of
545 horizontal extension at both ends of the OFZ seems to be larger than the 10-12 km offsets.
546 Indeed, 10-12 km of offset are not sufficient to create the subsidence calculated at the
547 Beautemps-Beaupré Basin [*Fournier et al., 2008a*]. The Dalrymple Trough itself may have
548 been active as an oblique rifting as early as the Early Miocene, based on stratigraphic studies
549 [*Gaedicke et al., 2002*]. There is thus an apparent mismatch between the present-day active
550 trace of the fault and the inferred age of initiation of strike-slip motion along the Owen Ridge.
551 This consequently raises the question of the location of the OFZ prior to the Pliocene. One
552 possibility is that the old fault system is located in the same area as the present-day fault
553 system, but is buried under Indus turbiditic deposits, as shown by one seismic profile across
554 the northern OFZ (**Figure 10**). This profile clearly shows more splays than active fault traces
555 at the surface. Acceleration of the turbiditic sedimentation rate between 15 Ma and 5 Ma
556 [*Shipboard scientific party, 1974, 1989; Clift et al., 2001*], in relation with the uplift of the
557 Tibet and the correlative increase in the erosional rates, may have favored such a burial.
558 Considering that the trace of the fault system on seafloor morphology is the result of the
559 competition between the fault activity and sedimentation rates, the presence of the turbiditic
560 channel A1 in the Owen Basin can be explained by a period where sedimentation rates were
561 high enough to entirely cover the active trace of the OFZ fault system.

562 The initiation of the present-day fault trace is synchronous with an increase of the
563 subsidence rate in the Dalrymple Trough and the uplift of the Murray Ridge around 4-5 Myrs
564 [*Gaedicke et al., 2002*]. However, this Pliocene change does not correlate with any known
565 plate-scale regional kinematic changes or major regional tectonic events [*Lepvrier et al.,*
566 *2002; Delescluse and Chamot-Rooke, 2007; Fournier et al., 2004, 2006, 2011*]. Therefore,
567 the structural evolution of the OFZ since 20 Myr and the onset of the present-day trace of the
568 OFZ 3-6 Myr ago both fit into the scheme of a continuum of deformation and a gradual
569 reorganization of the OFZ unrelated to any major kinematics change.

570

571 6.2. Main characteristics of the structural evolution of the Owen Fracture Zone

572 In this section, we emphasize the main characteristics of the last structural
573 reorganization of the OFZ since 3-6 Myr through a comparison with other transform plate

574 boundaries, mainly the Levant Fault [Aydin and Nur, 1982; TenBrink and Ben Avraham,
575 1989; Lazar et al., 2006], the San Andreas Fault [Wakabayashi et al., 2004; Wakabayashi,
576 2007], and a large continental strike-slip fault: the Haiyuan fault in China [Zhang et al., 1989].
577

578 *6.2.1. Role of structural inheritance in the structural evolution of the Owen Fracture* 579 *Zone*

580 Several observations suggest a control of structural inheritance in the reorganization of
581 the OFZ fault system. First, the topographic relief of the Owen Ridge does not seem to affect
582 the location of the OFZ, which runs alternatively across the slope or at the eastern foot of the
583 Owen Ridge (Figures 2, 3). This conspicuous configuration could be the result of the
584 reactivation of a buried splay of the OFZ. Secondly, the activation of the 0.7-0.6 Myr old
585 segment of the OFZ in the area of the Qalhat Basin could be the result of the reactivation of a
586 buried splay of the flower structure identified in this area (FigureFigures-9, 10, 14). Basin
587 inception and extinction thus seem to be related to the alternate activation of fault segments in
588 the fracture zone. This situation is very similar to the one observed at the Dayinshui Basin
589 located along the Haiyuan Fault in China [Zhang et al., 1989]. Even if the OFZ
590 accommodated only 10-12 km of relative motion [Fournier et al., 2011], the structural
591 evolution of the OFZ in the area of the Qalhat Basin could possibly show a transition from a
592 diffuse to a more localized deformation pattern with increasing maturity of strike-slip motion,
593 as observed elsewhere [Le Pichon et al., 2001; Ben-Zion and Sammis, 2003; Wesnousky,
594 2005; Schattner and Weinberger, 2007; Wu et al., 2010; Garcia-Moreno et al., 2010]. The
595 deformation tends to localize on pre-existing major fault segments with time.

596

597 *6.2.2. Asymmetry and inception of pull-apart basins along the Owen Fracture Zone*

598 All pull-apart basins developed along the OFZ are asymmetric (18°40'N, 20°N,
599 Qalhat Basin). Asymmetric basins are commonly observed along strike-slip faults [Ben
600 Avraham and Zoback, 1992; Brothers et al., 2009; Seeber, 2010] and interpreted as the result
601 of transform normal extension. Asymmetry exists even at the scale of the sub-basin SB3 of
602 the 20°N Basin (FigureFigure-6), which compares closely with the Zofar Basin along the
603 Levant fault because of the reversal of its asymmetry [Frieslander, 2000].

604 The activity of pull-apart basins along the OFZ spans over different time-scales. The
605 opening of pull-apart basins at 20°N 1.8 to 1.5 Myr ago is not coeval with the activation of the
606 present-day OFZ 3-6 Myr ago. (Figure 7Figure-15). When the main releasing bend formed at
607 20°N 1.8 to 1.5 Myr ago, the OFZ was already emplaced since 20 Myr. This is similar to the

608 timing of emplacement of the Dead Sea Basin 3 to 5 Myr ago along the 17 to 11.5 Myr old
609 Levant Fault [*Le Pichon and Gaulier, 1988; Garkunkel and Ben Avraham, 2001*] in response
610 to a small gradual change in relative plate kinematics [*Garfunkel, 1981*]. It may also be
611 compared to the emplacement of the Olema Creek formation 110 to 185 kyr ago along the 18
612 Myr old San Andreas Fault [*Wakabayashi et al., 2007*].

613

614 6.2.3. Migrating subsidence along the Owen Fracture Zone

615 Several observations suggest that subsidence migration is one of the main
616 characteristic of the OFZ fault system reorganisation since 3-6 Myr. The 20°N basin displays
617 a series of three sub-basins (~~Figure 4~~Figure-4). Sub-basin 2 is 40 m deeper than sub-basin 1,
618 and sub-basin 3 is 260 m deeper than sub-basin 2. Sub-basins 1 and 2 were topographically
619 isolated from the turbiditic supply of the currently active channel since its inception in this
620 area and are sealed by a pelagic blanket. The topography of sub-basins 1 and 2 is flat, whereas
621 the topography of sub-basin 3 is gently tilted, so as the corresponding turbiditic layers
622 (~~Figure 5~~Figure-5). These observations indicate that subsidence is currently inactive in sub-
623 basin 1 and 2, and is localized in sub-basin 3. Subsidence initiated to the south of the 20°N
624 Basin, and migrated northwards, creating deeper and younger sub-basins bounded by
625 transverse faults.

626 This situation is very close to what is observed in the case of the Dead Sea pull-apart
627 basin [*Aydin and Nur, 1982; TenBrink and Ben Avraham, 1989; Lazar et al., 2006*] and the
628 San Andreas Fault system [*Wakabayashi et al., 2004; Wakabayashi, 2007*] where episodes of
629 migration of subsidence are documented [*Kashai and Croker, 1987; Garfunkel and Ben*
630 *Avraham, 1996*]. The active Dead Sea Basin is indeed divided in two sub-basins separated by
631 a transverse fault, the northern one being 350 m deeper than the southern one. However,
632 subsidence migration in the Dead Sea is still a matter of debate, and alternate interpretations
633 have been proposed [*Ben Avraham and Schubert, 2006, Lazar et al., 2006 and Ben Avraham*
634 *et al., 2008*].

635 Sandbox laboratory experiments performed by [*Smit et al., 2008*] show that migration
636 of sub-basins occurs where the ratio between the strike-slip faults spacing and the thickness of
637 the deforming layer is <1 . [*Smit et al., 2008*] conclude that this ratio determines not only the
638 basin width, but also its geometry and the migration of subsidence. In these models simulating
639 the migration of subsidence, the intrabasinal transverse faults appear during basin migration,
640 and do not result from the reactivation of previous faults dividing the basin, consistently with
641 observations of [*Kashai and Croker, 1987*] on the Dead Sea Basin.

642 In the case of the 20°N Basin, the subsidence migrated much faster than the
643 displacement accommodated along the OFZ during the last million years (more than 60 km of
644 displacement for the depocenter, versus less than 10 km of displacement along the master
645 fault). Following the nomenclature proposed by [Wakabayashi *et al.*, 2007], sub-basins SB1
646 and SB2 are interpreted as the “wake” of the active depocenter (**Figure 4**Figure-4). In the
647 light of the model of [Smit *et al.*, 2008], the active transverse syn-subsidence faults observed
648 in the active sub-basin 3 (**Figure 5**Figure-5) could be incipient stages of future transverse
649 faults that isolate and concentrate the subsidence. If verified, this would confirm the tendency
650 of subsidence to localize in a smaller but deeper sub-basin, without increasing the overall size
651 of the releasing bend with increasing offset along the master fault.

652

653 7. Conclusions and perspectives

654 The present day India-Arabia plate boundary emplaced at the location of the OFZ 3 to
655 6 Myr ago. The structural evolution of the OFZ was principally marked by the development
656 of a major releasing bend at 20°N ~1.8-1.5 Myr ago and subsequent subsidence migration,
657 local compression at 17°30'N and 20°30'N ~0.8 Myr ago, and the extinction of the Qalhat
658 Basin 0.7 Myr ago. The structural evolution of the OFZ since the Pliocene suggests a
659 continuous adjustment to a steady-state relative plate motion between the Arabian and Indian
660 plates. The present day OFZ strike-slip system is the latest stage of evolution the India-Arabia
661 plate boundary. Older stages of evolution need to be further constrained by deeper seismic
662 lines. Indeed, the structure of the inferred 20-Myr-old buried strike-slip system that
663 accommodated the motion between India and Arabia since the opening of the Gulf of Aden
664 has not been clearly observed. The location of the India-Arabia plate boundary for the
665 Paleogene-Early Miocene interval is far less known. Kinematics modelling [Royer *et al.*,
666 2002] and geological studies [Edwards *et al.*, 2000] postulate that the OFZ was located in the
667 Owen Basin at those times, but there is no structural evidence for this hypothesis.

668

669 **Acknowledgements:** We are indebted to the Captain Geoffroy de Kersauson, officers, and
670 crew members of the *BHO Beautemps-Beaupré*, and to the French Navy hydrographers
671 Vincent Lamarre and Yves-Marie Tanguy, and the hydrographic team of the ‘Groupe
672 Océanographique de l’Atlantique’, for their assistance in data acquisition. We thank Thierry
673 Garland for the supply of the Fanindien cruise data and the support of ARTEMIS for the use
674 of ¹⁴C ages measured on the cores collected during the Fanindien cruise. We greatly
675 acknowledge Tim Minshall, an anonymous reviewer and G3 editors for their constructive

676 comments. We thank Baptiste Mulot, Pierpaolo Dubernet and Matthias Delescluse for their
677 technical assistance. We acknowledge the support of SHOM, IFREMER, CEA (LRC Yves-
678 Rocard) and INSU-CNRS for the Owen and the Fanindien 2009 cruises and their support of
679 M. Rodriguez thesis.

680

681 **Figure captions**

682

683 Figure 1. Multibeam bathymetric map of the Owen Fracture Zone acquired during the Owen
684 and Fanindien cruises, with location of figures 2 to 5. Shallow seismicity since 1973 (focal
685 depth < 50 km, magnitude > 2), from USGS/NEIC database (yellow dots), Engdhal et al.
686 (1998, white dots), CMT Harvard database (red dots), and Quittmeyer and Kafka (1984; green
687 dots), and earthquake focal mechanisms for the Owen fracture zone. The seismicity along the
688 OFZ is moderate, the maximum magnitude recorded to date being a Mw 5.8 earthquake.
689 However infrequent but large earthquake may be expected as at other slow boundaries. Inset
690 shows the regional tectonic setting of the India-Arabia plate boundary. AOC: Aden-Owen-
691 Carlsberg triple junction, B³: Beautemps-Beaupré basin, ITS: Indus turbiditic system, OFZ:
692 Owen fracture zone, Sh: Sheba Ridge.

693

694 Figure 2. a) Slope gradient map of the OFZ along the southern Owen Ridge. b) Inset shows
695 the restraining bend which initiates where the trend of the linear segment of the OFZ slightly
696 deviates and divides into two splays. The inactive splay is buried under mass transport
697 deposits from the southern ridge. The maximal vertical throw of the OFZ is associated to the
698 maximal vertical throw of the restraining bend. c) 3.5 kHz profile running through the
699 restraining bend (see inset b) for location). A slight and abrupt increase in folding is observed
700 at ~0.8 Myr (see text).

701

702 Figure 3. Slope gradient map of the OFZ along the central Owen Ridge. A pull-apart basin is
703 observed at 18°40'N. The OFZ offset dextrally the Owen Ridge over 10 km.

704

705 Figure 4. Slope gradient map of the asymmetric 20°N pull-apart basin, and interpretative
706 structural scheme. SB: Sub-basin. A, B, C: channels. Inset shows the en-échelon fault system
707 located to the south east.

708

709

710 Figure 5. a) topography of the 20°N Basin. SB1 and SB2 are flat, whereas SB3 is gently
711 tilted. 3.5 kHz profiles oriented longitudinally along the 20°N basin b) across the inactive sub-
712 basin SB1 and SB2; and c) across the active sub-basin SB3 (see [Figure 4](#) for
713 location). Major transverse faults are observed at the seafloor ([Figure 4](#)) whereas
714 minor transverse faults show only on SBP120 profiles since vertical deformation is erased by
715 rapid deposition. The pelagic blanket of SB1 and SB2 was gently and progressively folded by
716 flexural deformation.

717

718 Figure 6. 3.5 kHz profiles oriented transverse to the 20°N basin (see [Figure 4](#) for
719 location). The profile a) runs across the sub-basin SB3 and the DSDP 222 site, allowing
720 correlations of the ages deduced from sedimentation rates. The profile b) runs across sub-
721 basin SB3. c) zoom of profile a) showing the location of DSDP site 222 and the picking of the
722 pelagic reflector that seals the activity of turbiditic channel B3. The pelagic drape is imaged
723 by conformable reflectors and is composed of detrital clay nanno-ooze to nanno-rich detrital
724 carbonate silty clay [DSDP site 222, *Shipboard scientific party, 1974*]. The trace of activity of
725 the turbiditic channel is imaged by a chaotic and rough facies. The age of 0.95 - 1.32 Myr is
726 deduced from pelagic sedimentation rates calculated at site DSDP 222.

727

728 Figure 7. 3.5 kHz profiles across the turbiditic channels affected by the opening of the 20°N
729 basin (see [Figure 4](#) for location). Channel activity is imaged by a rough and chaotic
730 facies on SBP profile.

731

732 Figure 8. 3.5 kHz profile across the "en-échelon" fault system located at the southern end of
733 the 20°N basin (see [Figure 4](#) for location). The sedimentary filling that onlaps tilted
734 blocks must be the result of fast deposition related to channel B2, as it lies between two
735 concordant layers of pelagic deposits on the southern part of the SBP profile. It is therefore
736 interpreted as mixed turbiditic and pelagic deposits.

737

738 Figure 9. Slope gradient map of the OFZ along the Qalhat Seamount, and interpretative
739 structural scheme. 1, 2, 3: splays of the OFZ; A, B: channels. Inset shows a turbiditic channel
740 dissected by the splays of the OFZ. A part of the slope gradient map is derived from data
741 published in [*Bourget et al., 2009*].

742

743 | Figure 340. Seismic profile modified from Edwards et al. (2000), showing a negative flower
744 | structure buried under turbiditic deposits at the eastern foot of the Qalhat Seamount (see
745 | [FigureFigure-9](#) for location).

746

747 | Figure 441. 3.5 kHz profile across the elevated graben of the Qalhat Basin (see [FigureFigure](#)
748 | [9](#) and [Figure 5Figure-12](#) for location). Mass Transport Deposits (MTD) are observed beneath
749 | the pelagic cover. The graben used to be a catchment for MTD but it is now isolated because
750 | of the onset of the main OFZ that crosscuts the Qalhat Basin.

751

752 | Figure 542. Reflectivity map of the Qalhat Basin (see [FigureFigure-9](#) for location). The low
753 | reflectivity area indicates that the present-day depocenter is still a catchment for Mass
754 | Transport Deposits (MTD), whereas the high reflectivity facies of the graben indicates that it
755 | is not supplied by MTDs anymore.

756

757 | Figure 13. A) Blow-up of 3.5 kHz profile 2 (see figure 5) showing the location of the site of
758 | tilt measurement. The first ten meters have been sampled by Küllenberg cores. B) Relation
759 | between the tilt of turbiditic deposits and their age. The tilt of turbiditic horizons in sub-basin
760 | 3 is measured on SBP profiles (see [Figure 5Figure-5](#) for location) and corrected from vertical
761 | exaggeration ¹⁴C ages are available for turbiditic events 1 to 8 [*Bourget et al., submitted*]. The
762 | relation between the tilt of turbiditic horizons and ¹⁴C ages is fairly linear, suggesting a
763 | constant rate of tilting through times. C) The measurement of the tilt of turbiditic horizons can
764 | be used as a time indicator for deeper horizons unsampled by küllenberg cores according to
765 | the relation obtained in Figure B. The deepest turbiditic deposit we could date is 420 kyr old.

766

767 | Figure 644. Chronology of the evolution of the Owen fracture zone (OFZ) in the vicinity of
768 | the Qalhat Seamount. Stages are younger from a) to d), although they cannot be precisely
769 | dated. Figure 16 shows the structural evolution of the OFZ in the case where the en-échelon
770 | fault system pre-dates the onset of the horsetail termination. The identification of mass
771 | transport deposits (MTD) on the 3.5 kHz profile of figure 12 indicates that the Qalhat Basin
772 | used to be wider (stage c) before being crosscut by the present-day main OFZ 0.8 to 0.7 Myr
773 | ago (stage d) An alternative interpretation is that the en-échelon fault system developed
774 | recently and could be the incipient future OFZ.

775

776 | Figure 745. Chronology of the opening and the evolution of the 20°N basin.

777 a) Schematic reconstruction of the OFZ at 20°N before 3 Myr. Channels of the generation A
778 are active. The location of channel A1 west of the OFZ could indicate that the active trace of
779 the fault system was probably buried under turbiditic deposits at those times. b) Schematic
780 reconstruction of the OFZ at 20°N at 1.8-1.5 Myr. Transition of sedimentary facies recorded
781 at DSDP site 222 indicates fault activity at those times, which could be related to incipient
782 stages of formation of the 20°N pull-apart basin. Turbiditic channels of generation B were
783 probably active and could have canalized turbiditic deposits to the incipient 20°N Basin. c)
784 Schematic reconstruction of the OFZ at 20°N at 1.32-0.95 Myr. Turbiditic channels of the
785 generation B are inactive, probably because of the opening of the 20°N Basin. The active
786 depocenter was wider than it is today. d) Schematic reconstruction of the OFZ at 20°N since
787 at least 400 kyr. Between 1.32-0.95 Myr and 400 kyr, the 20 °N Basin has undergone
788 subsidence delocalization in the smaller but deeper sub-basin 3. Turbiditic channel C
789 activated and filled in the sub-basin 3.

790

791

792

793 **References**

- 794 Aydin, A. and A. Nur (1982), Evolution of pull-apart basins and their scale independence,
795 *Tectonics*, *1*, 91-105, doi:1029/TC001i001p00091.
- 796 Babonneau, N., B. Savoye, M. Cremer, and B. Klein (2002), Morphology and architecture of
797 the present canyon and channel system of the Zaire deep-sea fan, *Mar. Pet. Geol.*, *19*, 445-
798 467.
- 799 Basile, C. and J.P. Brun (1999), Transtensional faulting patterns ranging from pull-apart
800 basins to transform continental margins: an experimental investigation, *J. Struct. Geol.*, *21*,
801 23–37.
- 802 Ben-Avraham, Z. and M.D. Zoback (1992), Transform normal extension and Asymmetric
803 basins: an alternative to pull-apart models, *Geology*, *20*, 423–426.
- 804 Ben Avraham Z and G. Schubert (2006), Deep "drop down "basin in the southern Dead Sea,
805 *Earth Planet. Sci. Lett.*, *251*, 254-63.
- 806 Ben Avraham Z., Z. Garfunkel, M. Lazar (2008), Geology and evolution of the southern Dead
807 Sea Fault with emphasis on subsurface structure, *Annu. Rev. Earth Planet. Sci.*, *36*, 357-87
- 808 Bourget, J. (2009) Les systèmes turbiditiques du golfe d'Oman et de la marge Est-africaine:
809 architecture, évolution des apports au quaternaire terminal, et impact de la distribution
810 sédimentaire sur les propriétés géoacoustiques des fonds, PhD thesis, 404 pp., Univ.
811 Bordeaux 1, France.
- 812 Bourget, J., S. Zaragosi, T. Mulder, J.-L. Schneider, T. Garlan, A. Van Toer, V. Mas, N.
813 Ellouz-Zimmermann (2010), Hyperpycnal-fed turbidite lobe architecture and recent
814 sedimentary processes: A case study, *Sediment. Geol.* *229*, 144-159
815 doi:10.1016/j.sedgeo.2009.03.009
- 816 Bourget, J., S. Zaragosi, N. Ellouz-Zimmermann, N. Mouchot, T. Garlan, J-L Schneider, V.
817 Lanfumey, S. Lallemand (2010), Turbidite system architecture and sedimentary processes
818 along topographically complex slopes: the Makran convergent margin, *Sedimentology*, *58*,
819 376-406, doi: 10.1111/j.1365-3091.2010.01168.
- 820 Bourget, J., S. Zaragosi, M. Rodriguez, M. Fournier, T. Garlan, N. Chamot-Rooke, Late
821 Quaternary megaturbidites of the Indus Fan: origin and stratigraphic significance.
822 *Submitted to Marine Geology*.
- 823
- 824

825 Brothers, D.S., N.W. Driscoll, G.M. Kent, A.J. Harding, J.M. Babcock, R.L. Baskin (2009),
826 Tectonic evolution of the Salton Sea inferred from seismic reflection data, *Nature*
827 *Geoscience*, 2, 581-584, doi: 10.1038/ngeo590.

828 Calvès, G. (2008), Tectonostratigraphic and climatic record of the NE Arabian Sea, PhD
829 thesis, 305 pp., Univ. Aberdeen, U.K.

830 Chamot-Rooke, N., M. Fournier, Scientific Team of AOC and OWEN cruises (2009),
831 Tracking Arabia-India motion from Miocene to Present, American Geophysical Union,
832 Fall Meeting 2009.

833 Clift, P. D., N. Shimizu, G. D. Layne, J. S. Blusztain, C. Gaedicke, H. U. Schluter, M. K.
834 Clark, and S. Amjad (2001), Development of the Indus Fan and its significance for the
835 erosional history of the Western Himalaya and Karakoram, *Geol. Soc. Am. Bull.*, 113,
836 1039-1051.

837 Cunningham, W. D. and P. Mann, (2007), Tectonics of strike-slip restraining and releasing
838 bends, in *Tectonics of Strike-Slip Restraining and Releasing Bends*, edited by W.D.
839 Cunningham and P. Mann, *Geol. Soc. Spec. Publ.*, 290, 1–12.

840 Delescluse, M. and N. Chamot-Rooke (2007), Instantaneous deformation and kinematics of
841 the India-Australia Plate, *Geophys. J. Int.*, 168, 818-842, doi: 10.1111/j.1365-
842 246X.2006.03181.x

843 DeMets, C., R.G. Gordon, D.F. Argus and S. Stein (1990), Current plate motions, *Geophys. J.*
844 *Int.*, 101, 425-478.

845 DeMets, C., R.G. Gordon, D.F. Argus and S. Stein (1994), Effect of recent revisions of the
846 geomagnetic reversal time scale on estimates of current plate motions, *Geophys. Res. Lett.*
847 21, 2191-2194

848 DeMets C., R. G. Gordon and D.F. Argus (2010), Geologically current plate motions,
849 *Geophys. J. Int.*, 181, 1-80, doi: 10.1111/j.1365-246X.2009.04491.x,.

850 Edwards R.A., T. A. Minshull and R. S. White (2000), Extension across the Indian–Arabian
851 plate boundary: the Murray Ridge, *Geophys. J. Int.*, 142, 461-477.

852 Edwards, R. A., T. A. Minshull, E. R. Flueh and C. Kopp (2008), Dalrymple Trough: An
853 active oblique-slip ocean-continent boundary in the northwest Indian Ocean, *Earth Planet.*
854 *Sci. Lett.*, 272, 437-445.

855 Ellouz Zimmermann, N. et al. (2007b), Offshore frontal part of the Makran accretionary
856 prism (Pakistan) the Chamak Survey, in *Thrust Belts and Foreland Basins: From Fold*
857 *Kinematics to Hydrocarbon Systems*, edited by O. L. Lacombe et al., pp. 349–364,
858 Springer, Berlin.

859 Engdahl, E.R., R. van der Hilst and R. Buland (1998), Global teleseismic earthquake
860 relocation with improved travel times and procedures for depth determination, *Bull. Seism.*
861 *Soc.Am.*, *88*, 722-743.

862 Fournier, M., N. Bellahsen, O. Fabbri, and Y. Gunnell (2004), Oblique rifting and
863 segmentation of the NE Gulf of Aden passive margin, *Geochem. Geophys. Geosyst.*, *5*,
864 Q11005, doi:10.1029/2004GC000731.

865 Fournier, M., C. Lévrier, P. Razin and L. Jolivet (2006), Late Cretaceous to Paleogene post-
866 obduction extension and subsequent Neogene compression in the Oman Mountains,
867 *GeoArabia*, *11*, 17-40.

868 Fournier, M., C. Petit, N. Chamot-Rooke, O. Fabbri, P. Huchon, B. Maillot, and C. Lévrier
869 (2008a), Do ridge-ridge-fault triple junctions exist on Earth? Evidence from the Aden-
870 Owen-Carlsberg junction in the NW Indian Ocean, *Basin Research*, *20*, 575-590, doi:
871 10.1111/j.1365-2117.2008.00356.x

872 Fournier, M., N. Chamot-Rooke, C. Petit, O. Fabbri, P. Huchon, B. Maillot, and C. Lévrier
873 (2008b), In-situ evidence for dextral active motion at the Arabia-India plate boundary,
874 *Nature Geoscience*, *1*, 54-58, doi:10.1038/ngeo.2007.24.

875 Fournier, M., N. Chamot-Rooke, C. Petit, P. Huchon, A. Al-Kathiri, L. Audin, M.-O. Beslier,
876 E. d'Acromont, O. Fabbri, J.-M. Fleury, K. Khanbari, C. Lévrier, S. Leroy, B. Maillot,
877 and S. Merkouriev (2010), Arabia-Somalia plate kinematics, evolution of the Aden-Owen-
878 Carlsberg triple junction, and opening of the Gulf of Aden, *J. Geophys. Res.*, *115*, B04102,
879 doi:10.1029/2008JB006257.

880 Fournier, M., N. Chamot-Rooke, M. Rodriguez, P. Huchon, C. Petit, M.-O. Beslier, and S.
881 Zaragosi (2011), Owen Fracture Zone: the Arabia-India plate boundary unveiled, *Earth*
882 *Planet. Sci. Lett.*, *302*, 247-252, doi:10.1016/j.epsl.2010.12.027.

883 Fournier, M., P. Patriat and S. Leroy (2001), Reappraisal of the Arabia-India-Somalia triple
884 junction kinematics, *Earth Planet. Sci. Lett.*, *189*, 103-114.

885 Frieslander, U. (2000), The structure of the Dead Sea transform emphasizing the Arava using
886 new geophysical data, Ph.D. thesis, 204 pp., Hebrew University of Jerusalem, Israel.

887 Gaedicke, C., A. Prexl, H.U. Schlüter, H. Roeser and P. Clift (2002), Seismic stratigraphy and
888 correlation of major regional unconformities in the northern Arabia Sea, in *The Tectonic*
889 *and Climatic Evolution of the Arabian Sea Region*, edited by P. Clift, D. Kroon, C.
890 Gaedicke and J. Craig, *Geol. Soc. Spec. Publ.*, *195*, 25-36.

891 Garcia Moreno, D., A. Hubert Ferrari, J. Moernaut, J.G. Fraser, X. Boes, M. Van Daele, U.
892 Avsar, N. Cagatay and M. De Batist (2010), Structure and recent evolution of the Hazar

893 Basin: a strike slip basin on the East Anatolian Fault, Eastern Turkey, *Basin Research*, 23,
894 191-207, doi: 10.1111/j.1365-2117.2010.00476.

895 Garfunkel, Z. (1981), Internal structure of the Dead Sea leaky transform (Rift) in relation to
896 plate kinematics, *Tectonophysics*, 80, 81-108.

897 Garfunkel, Z. and Z. Ben-Avraham (1996), The structure of the Dead Sea basin,
898 *Tectonophysics*, 266, 155–176.

899 Garfunkel, Z. and Z. Ben-Avraham (2001), Basins along the Dead Sea transform, in *Peri-*
900 *Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins*, edited by P.A.
901 Ziegler, W. Cavazza, A.H.F. Robertson, and S. Crasquin-Soleau, Mémoires Museum
902 National d'Histoire Naturelle de Paris, 186, 607–627.

903 Gordon, R.G. and C. DeMets (1989), Present-day motion along the Owen fracture zone and
904 Dalrymple trough in the Arabian Sea. *J. Geophys. Res.* 94, 5560-5570.

905 Govil, P. and P. D. Naidu (2008), Late Quaternary changes in depositional processes along
906 the western margin of the Indus Fan, *Geo-Marine Letters*, 28, 1-6.

907 Kashai, EL. and P. F. Croker (1987), Structural geometry and evolution of the Dead Sea-
908 Jordan rift system as deduced from new subsurface data, *Tectonophysics*, 141, 33-60.

909 Lazar, M., Z. Ben-Avraham and U. Schattner (2006), Formation of sequential basins along a
910 strike-slip fault - geophysical observations from the Dead Sea basin, *Tectonophysics*, 421,
911 53–69.

912 Le Pichon X., and J.M. Gaulier (1988), The rotation of Arabia and the Levant fault system,
913 *Tectonophysics*, 153, 271-94

914 Le Pichon X., A.M.C. Sengor, E. Demirbag, C. Rangin, C. Imren, R. Armijo, N. Gorue, N.
915 Cagatay, B. Mercier de Lepinay, B. Meyer, R. Saatcilar and B. Tok (2001), The active
916 main Marmara fault, *Earth Planet. Sci. Lett.*, 192, 595 – 616.

917 Lepvrier, C., M. Fournier, T. Bérard, and J. Roger (2002), Cenozoic extension in coastal
918 Dhofar (southern Oman): Implications on the oblique rifting of the Gulf of Aden,
919 *Tectonophysics*, 357, 279-293.

920 Mann, P., M. R. Hempton, D.C. Badley, and K. Burke (1983), Development of pull-apart
921 basins, *Journal of Geology*, 91, 529–554.

922 Mann, P. (2007), Global catalogue, classification and tectonic origins of restraining- and
923 releasing bends on active and ancient strike-slip fault systems, *Geol. Soc. Spec. Publ.*, 290,
924 13-142.

925 Matthews D. H. (1966), The Owen Fracture Zone and the northern end of the Carlsberg
926 Ridge, *Phil. Trans. Royal Soc., A*, 259, 172-186.

927 Menesguen C. (2010), Etude du remplissage et de l'évolution d'un bassin pull-apart le long de
928 la Zone de Fracture d'Owen. Master thesis, 37 pp., Univ. Bordeaux 1, France.

929 Merkuriev, S. and C. DeMets (2006), Constraints on Indian plate motion since 20 Ma from
930 dense Russian magnetic data: Implications for Indian plate dynamics, *Geochem. Geophys.*
931 *Geosyst.*, 7, Q02002, doi:10.1029/2005GC001079.

932 Mouchot, N. (2009), Tectonique et sédimentation sur le complexe de subduction du Makran
933 pakistanais. PhD thesis, 364 pp., Univ. Cergy-Pontoise, France.

934 Mountain, G.S. and W. L. Prell (1989), Geophysical Reconnaissance Survey for ODP Leg
935 117 in the Northwest Indian Ocean, in *Proc. ODP, Init. Repts., Leg 117*, edited by W.L.
936 Prell and N. Niitsuma, 117, pp. 51-64, College Station, TX.

937 Patriat, P., H. Sloan and D. Sauter (2008), From slow to ultraslow : a previously undetected
938 event at the Southwest Indian Ridge at ca. 24 Ma, *Geology*, 36;207-210, doi:
939 10.1130/G24270A.1

940 Qayyum M, R. A. Lawrence and A. R. Niem (1997), Molasse-Delta-Flysch continuum of the
941 Himalayan orogeny and closure of the Paleogene Katawaz remnant ocean, Pakistan, *Int.*
942 *Geol. Rev.*, 39, 861–875.

943 Quittmeyer, R. C. and A. L. Kafka (1984), Constraints on plate motions in southern Pakistan
944 and the northern Arabian Sea from the focal mechanisms of small earthquakes, *J. Geophys.*
945 *Res.*, 89, 2444-2458.

946 Rahe, B., D. Ferrill and A. Morris (1998), Physical analog modeling of pull-apart basin
947 evolution, *Tectonophysics*, 285, 21–40.

948 Rangin C. and X. Le Pichon, Strain localization in the Sea of Marmara: Propagation of the
949 North Anatolian Fault in a now inactive pull-apart, *Tectonics*, 23, TC2014,
950 doi:10.1029/2002TC001437, 2004

951 Reilinger R., et al. (2006), GPS constraints on continental deformation in the Africa – Arabia
952 – Eurasia continental collision zone and implications for the dynamics of plate interactions,
953 *J. Geophys. Res.*, 111, B05411, doi:10.1029/2005JB004051.

954 Royer J.Y., A.K. Chaubey, J. Dymant, G.C. Bhattacharya, K. Srinivas, V. Yateesh and T.
955 Ramprasad (2002), Paleogene plate tectonic evolution of the Arabian and Eastern Somali
956 basins, in *The Tectonic and Climatic Evolution of the Arabian Sea Region*, edited by P.
957 Clift, D. Kroon, C. Gaedicke and J. Craig, *Geol. Soc. Spec. Publ.*, 195, 7-23.

958 Schattner, U. and R. Weinberger (2008), A mid-Pleistocene deformation transition in the Hula
959 basin, northern Israel: Implications for the tectonic evolution of the Dead Sea Fault,
960 *Geochem. Geophys. Geosyst.*, 9, Q07009, doi:10.1029/2007GC001937.

961 Schattner, U. (2010), What triggered the early-to-mid Pleistocene tectonic transition across
962 the entire eastern Mediterranean? *Earth Planet. Sci. Lett.*, 289, 539-548.

963 Schlische, R. W., M. O. Withjack, and G. Eisenstadt (2002), An experimental study of the
964 secondary deformation produced by oblique slip normal faulting, *AAPG Bull.*, 86(5), 885-
965 906.

966 Seeber, L., C. Sorlien, M. Steckler, and M.-H. Cormier (2010), Continental transform basins:
967 Why are they asymmetric?, *Eos Trans. AGU*, 91(4), 29–30.

968 Shipboard Scientific Party, Site 222 (1974), in *DSDP Init. Repts*, leg 23, edited by R.B.
969 Whitmarsh, O.E. Weser and D.A. Ross, doi:10.2973/dsdp.proc.23.106.

970 Shipboard Scientific Party (1989), site 731 , in *Proc. ODP, Init. Repts., leg 117*, edited by
971 Prell, W.L., N. Niitsuma et al., College Station, TX (Ocean Drilling Program), 585-652.

972 Smit J., J. P Brun, S. Cloetingh and Z. Ben-Avraham (2008), Pull-apart basin formation and
973 development in narrow transform zones with application to the Dead Sea Basin, *Tectonics*,
974 27, TC6018, doi:10.1029/2007TC002119.

975 Sylvester, A. G. (1988), Strike-slip faults, *Geol. Soc. Am. Bull.*, 100, 1666–1703.

976 Ten Brink, U. S. and Z. Ben-Avraham (1989), The anatomy of a pull-apart basin: seismic
977 reflection observations of the Dead Sea, *Tectonics*, 8, 333-350.

978 Ten Brink, U. S. and M. Rybakov (1999), Anatomy of the Dead Sea transform; does it reflect
979 continuous changes in plate motion? *Geology*, 27, 887-890.

980 Wakabayashi, J., J. V. Hengesh and T. L. Sawyer (2004), Four-dimensional transform fault
981 processes: progressive evolution of step-overs and bends, *Tectonophysics*, 392, 279–301.

982 Wakabayashi, J. (2007), Stepovers that migrate with respect to affected deposits: field
983 characteristics and speculation on some details of their evolution, in *Tectonics of Strike-
984 Slip Restraining and Releasing Bends*, edited by W.D. Cunningham and Paul Mann, *Geol.
985 Soc. Spec. Publ*, 290, 169–188.

986 Weedon, G.P. and I.N. McCave (1991), Mud turbidites from the Oligocene and Miocene
987 Indus Fan at sites 722 and 731 on the Owen Ridge, in *ODP Proc. Scientific results, leg
988 116*, edited by Prell, W.L., N. Niitsuma, K. Emeis et al., College Station, Texas, p. 215–
989 220.

990 Weissel, J.K., V. A. Childers and G. D. Karner (1992), Extensional and Compressional
991 Deformation of the Lithosphere in the Light of ODP Drilling in the Indian Ocean.
992 Synthesis of Results from Scientific Drilling in the Indian Ocean, Geophysical
993 Monography 70, American Geophysical Union.

- 994 Wesnousky, S. G. (2005), The San Andreas and Walker Lane fault systems, western North
995 America: transpression, transtension, cumulative slip and the structural evolution of a
996 major transform plate boundary, *J. Struct. Geol.*, 27, 1505–1512
- 997 Whitmarsh, R.B., O. E. Weser and D. A. Ross (1974), Initial report DSDP, U.S. Government
998 Printing Office, Washington, D.C., v. 23, p. 1180.
- 999 Whitmarsh, R.B. (1979), The Owen Basin off the south-east margin of Arabia and the
1000 evolution of the Owen Fracture Zone, *Geophys. J. Royal Astron. Soc.*, 58, 441-470.
- 1001 Wu, J. E., K. McClay, P. Whitehouse and T. Dooley (2010), 4D analogue modelling of
1002 transtensional pull-apart basins, *Mar. Petrol. Geol.*, 26, 1608-1623.
- 1003 Zhang, P.-Z., B. C. Burchfiel, S. Chen and Q. Deng (1989), Extinction of pull-apart basins.
1004 *Geology*, 17, 814–817.

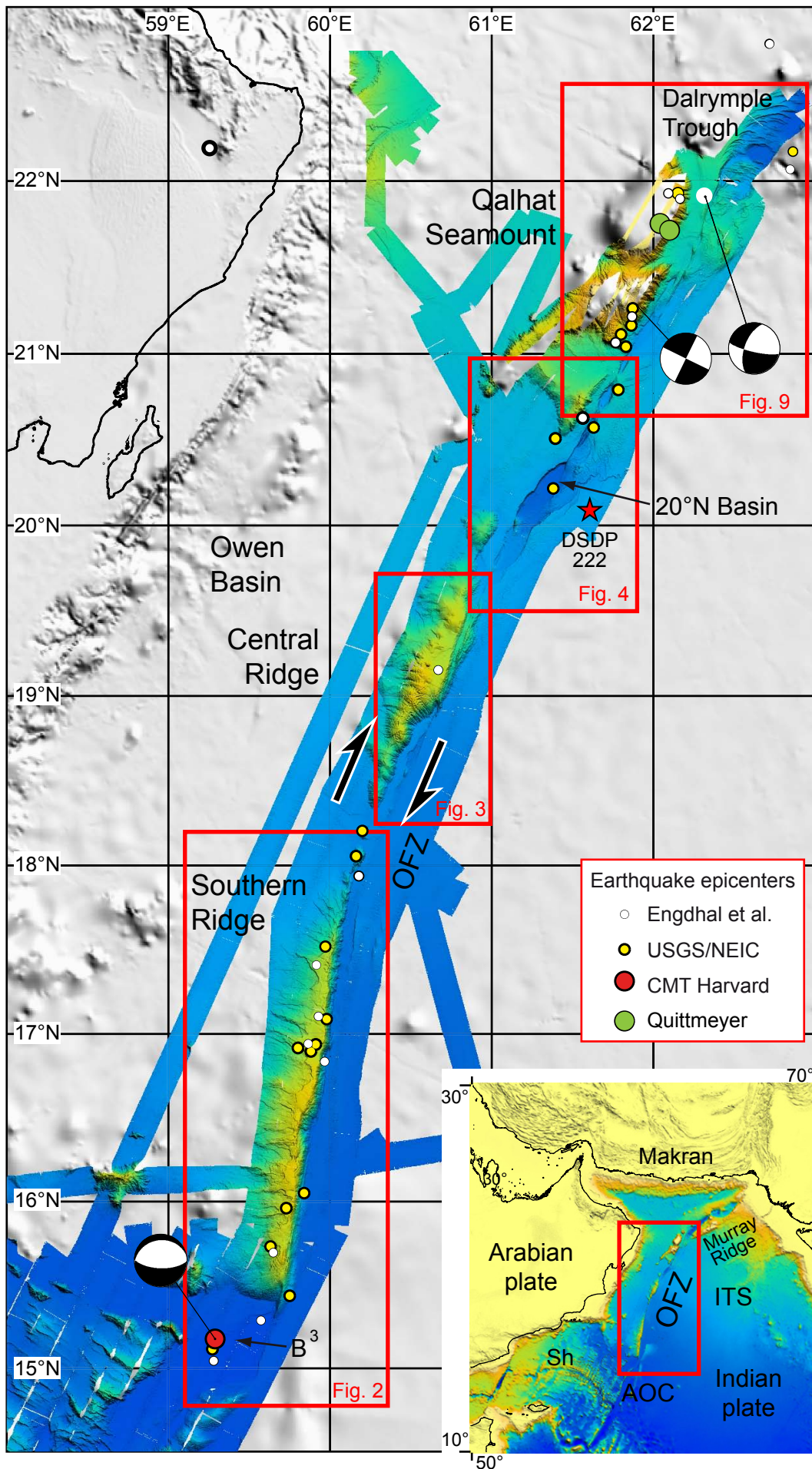


Figure 1

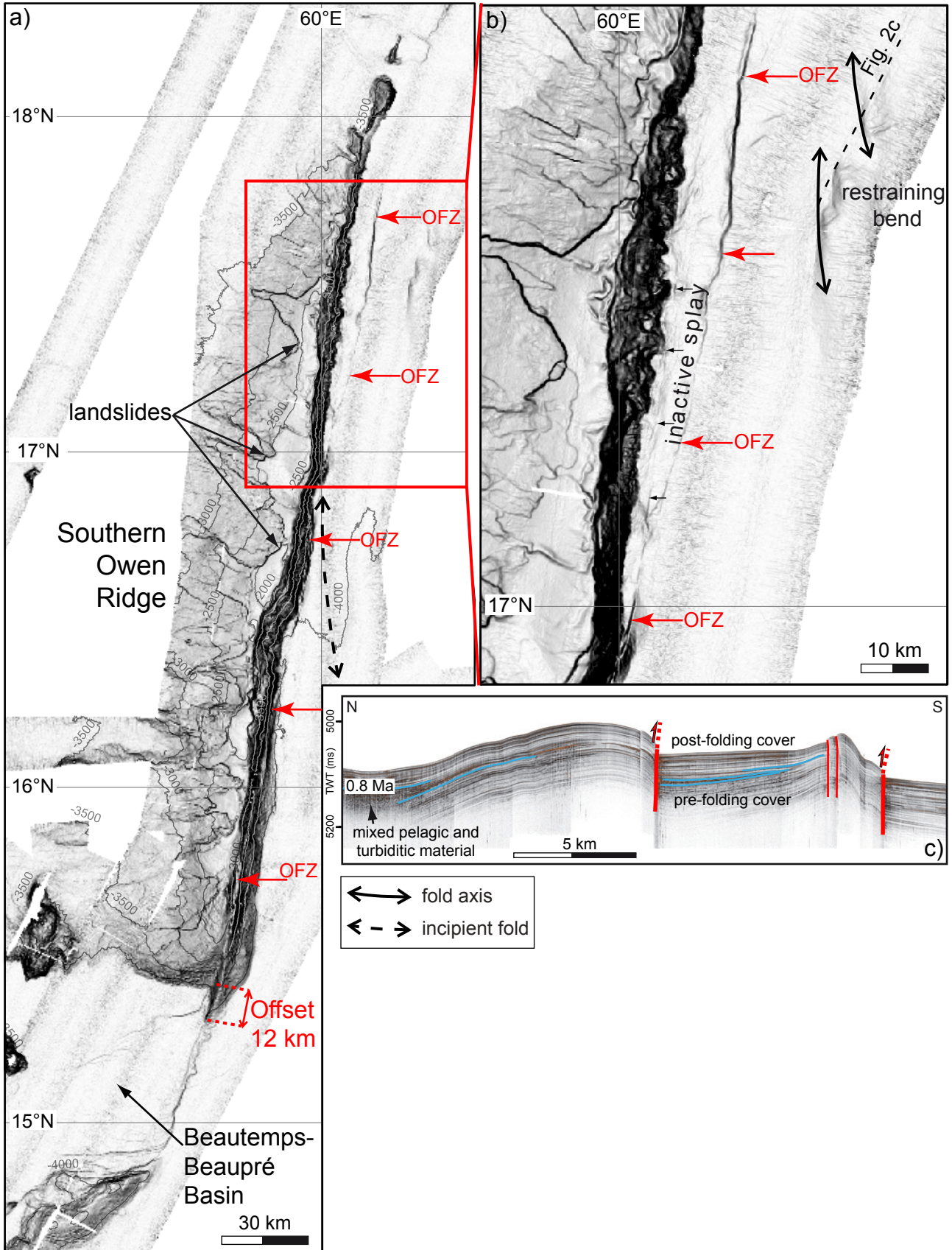


Figure 2

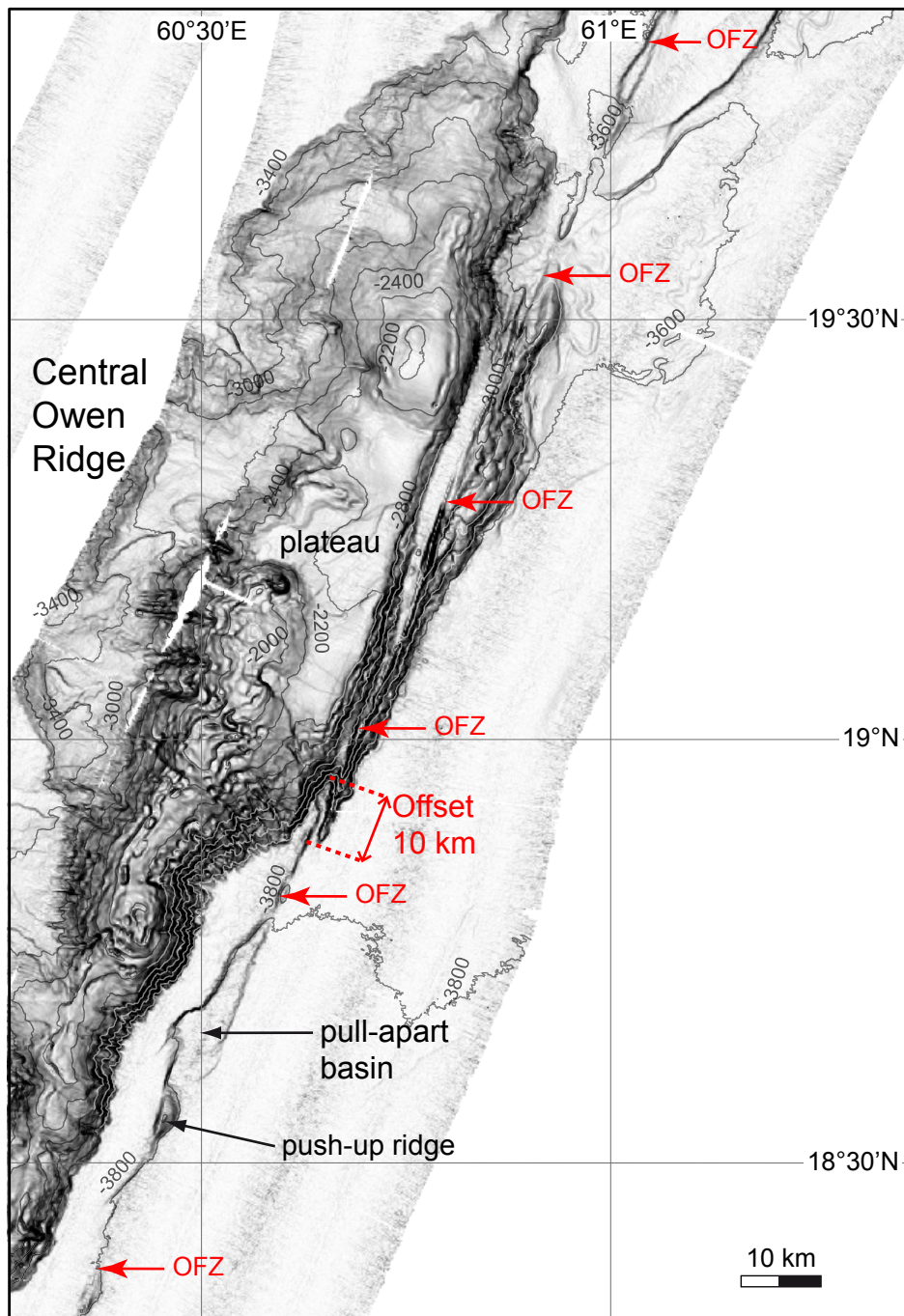


Figure 3

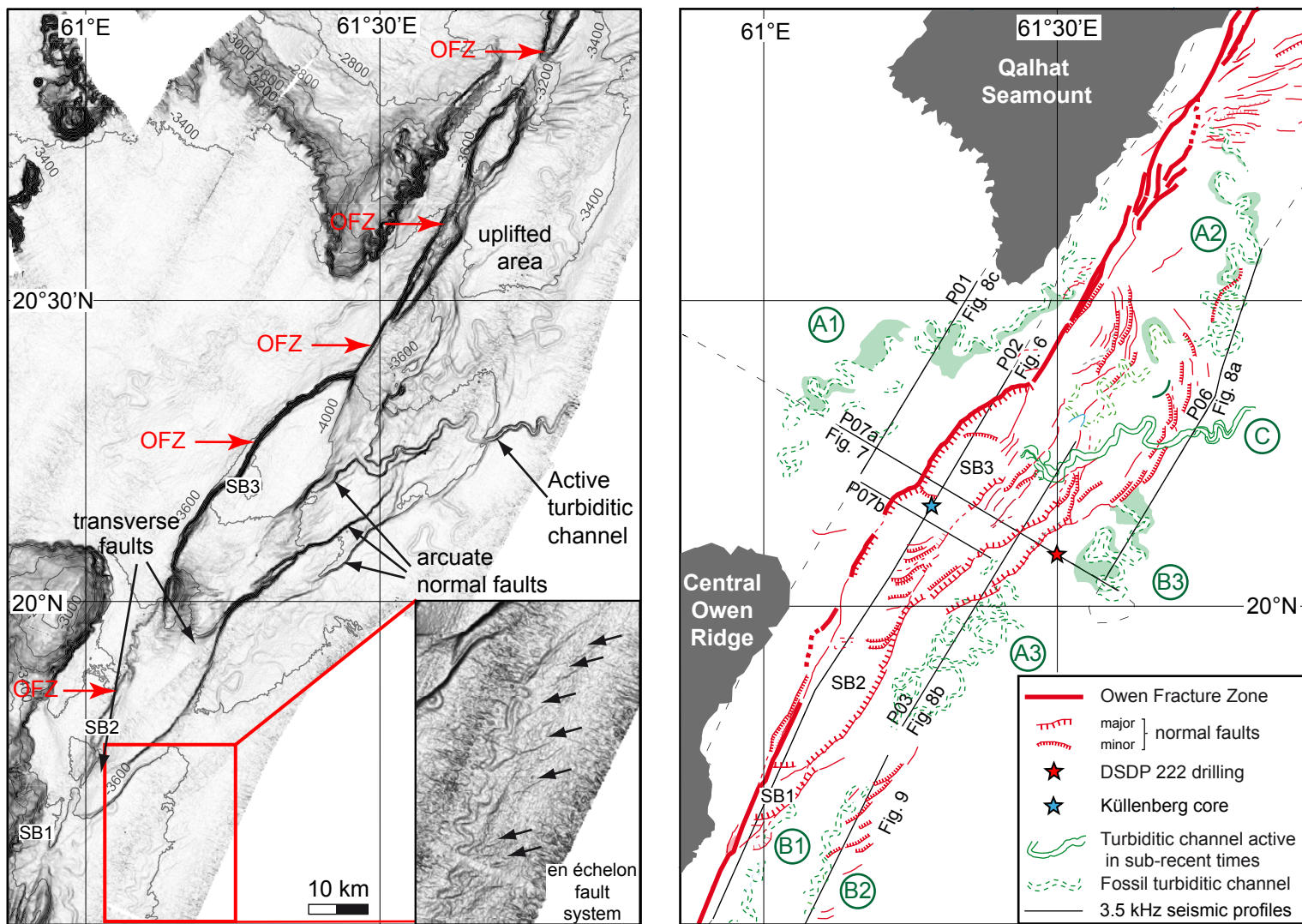


Figure 4

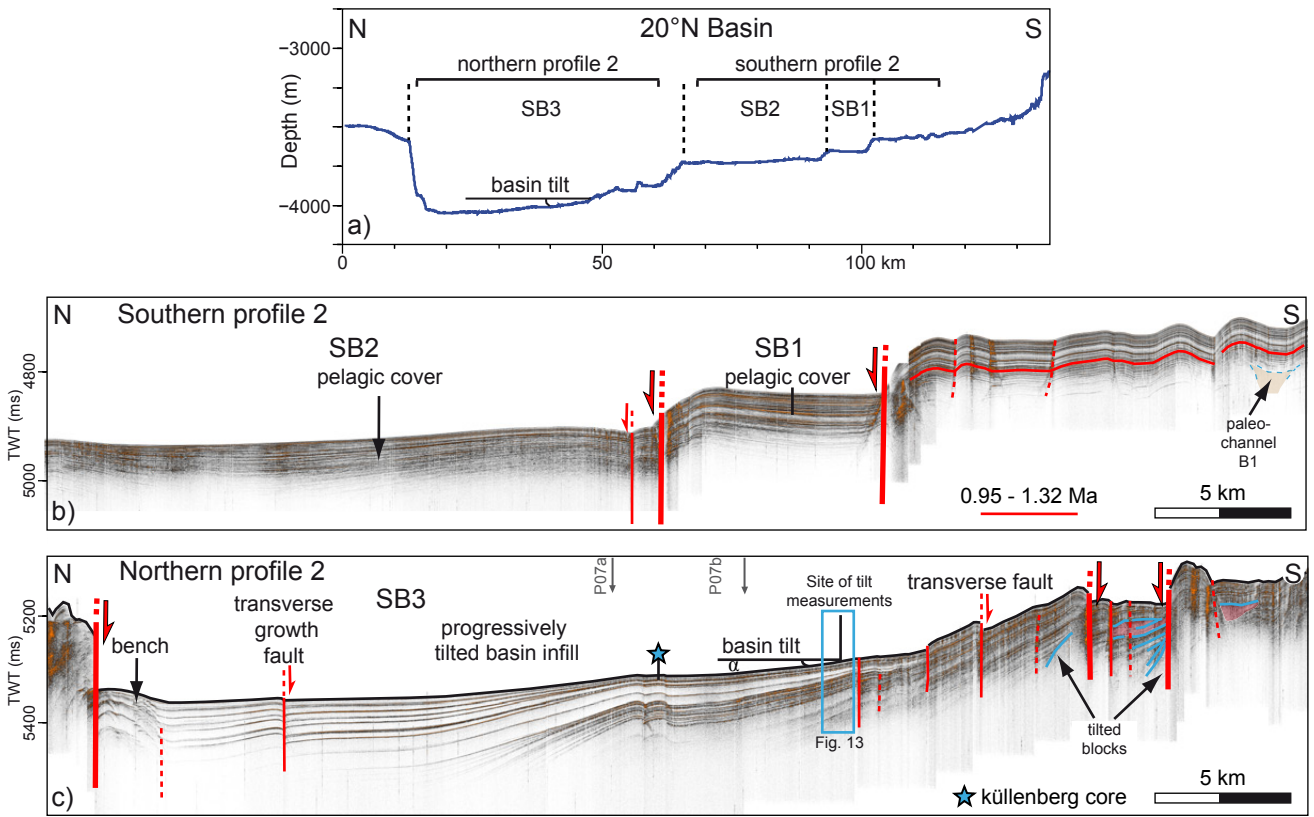


Figure 5

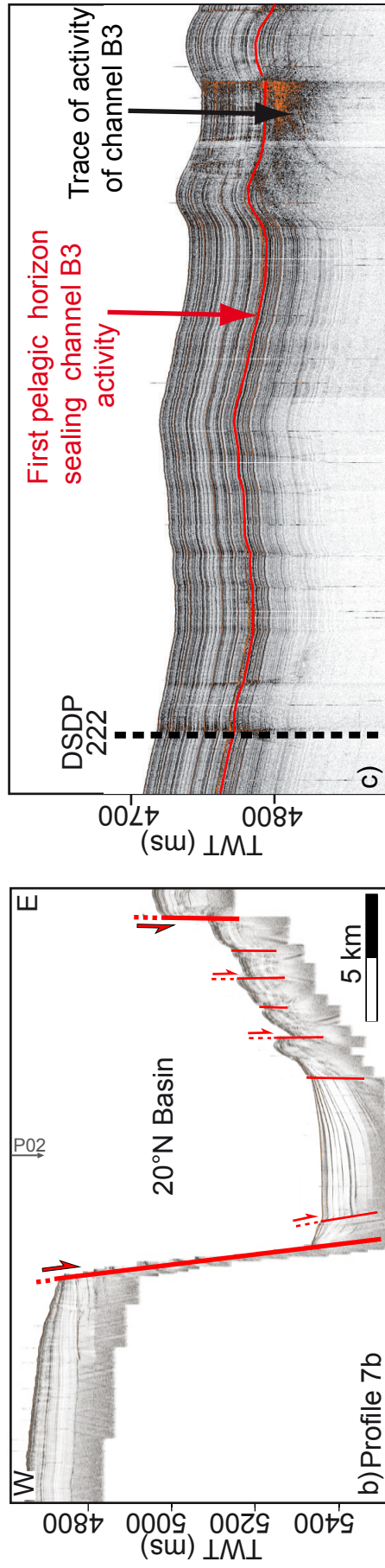
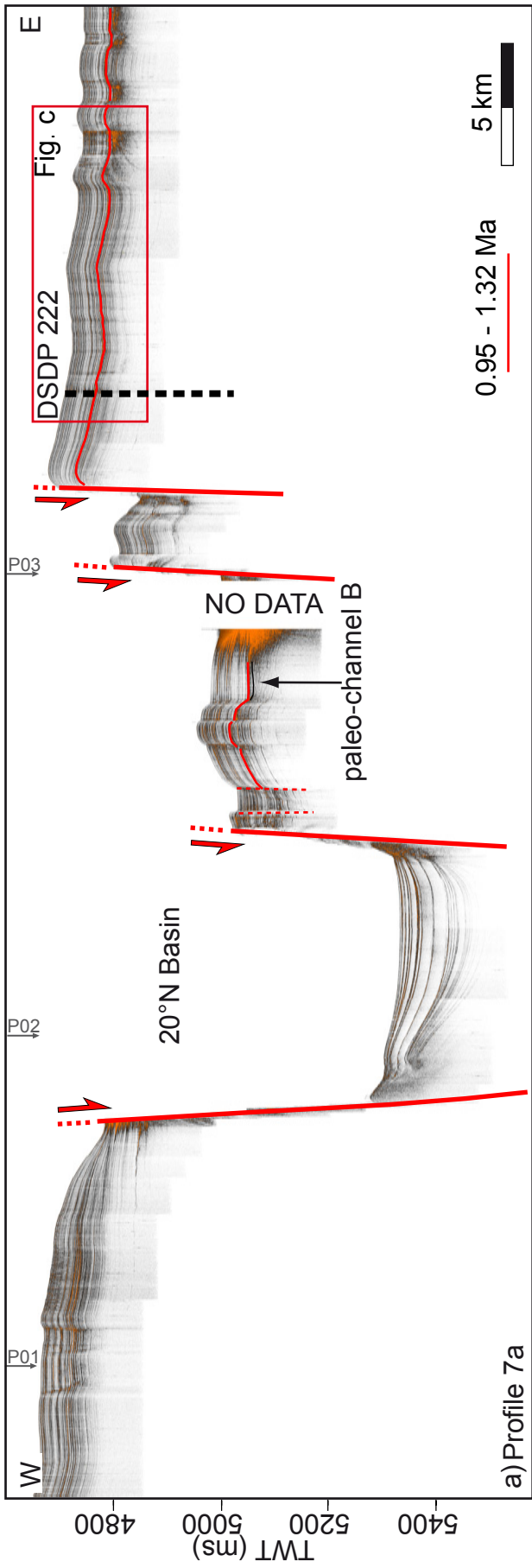


Figure 6

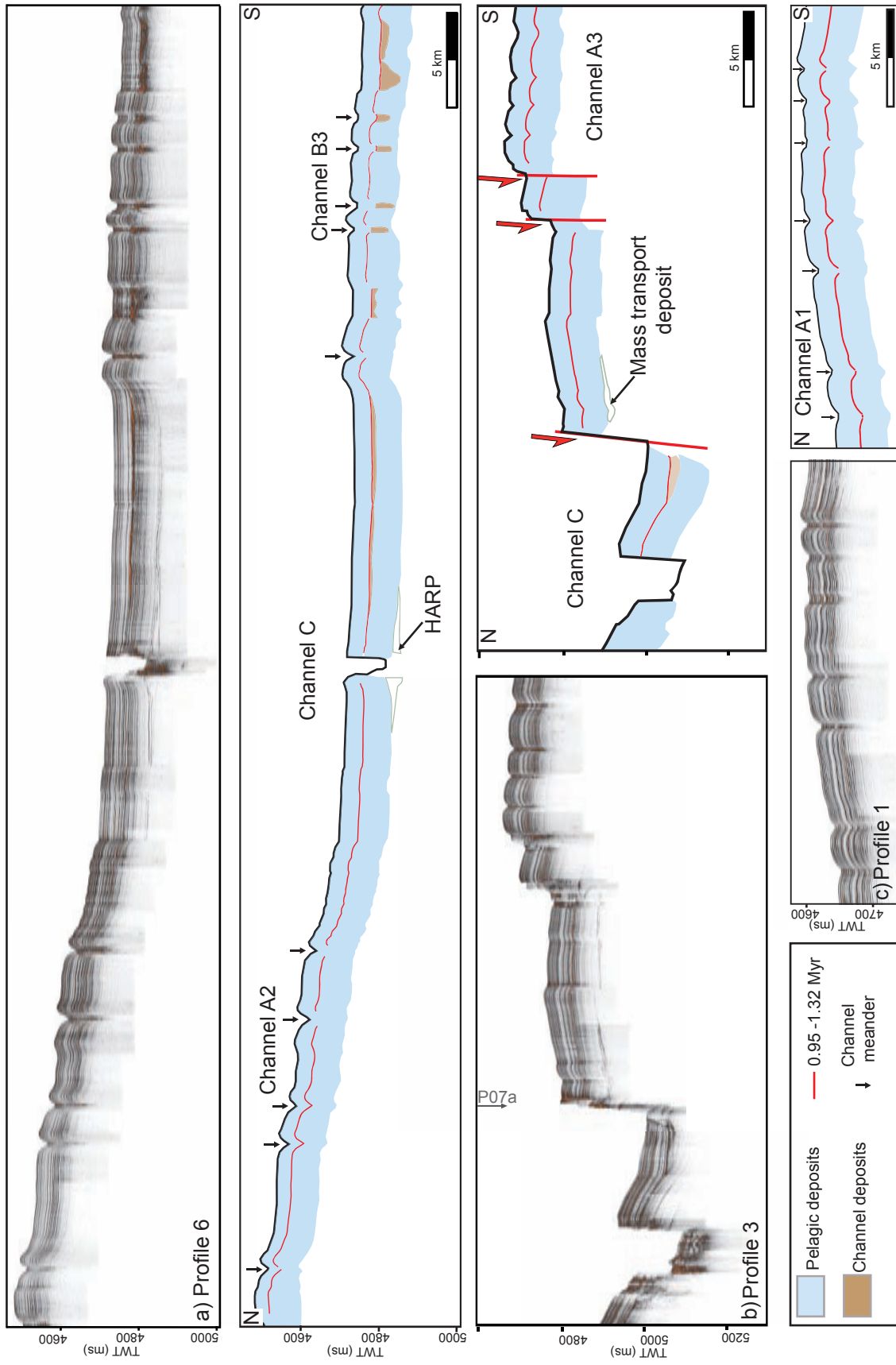


Figure 7

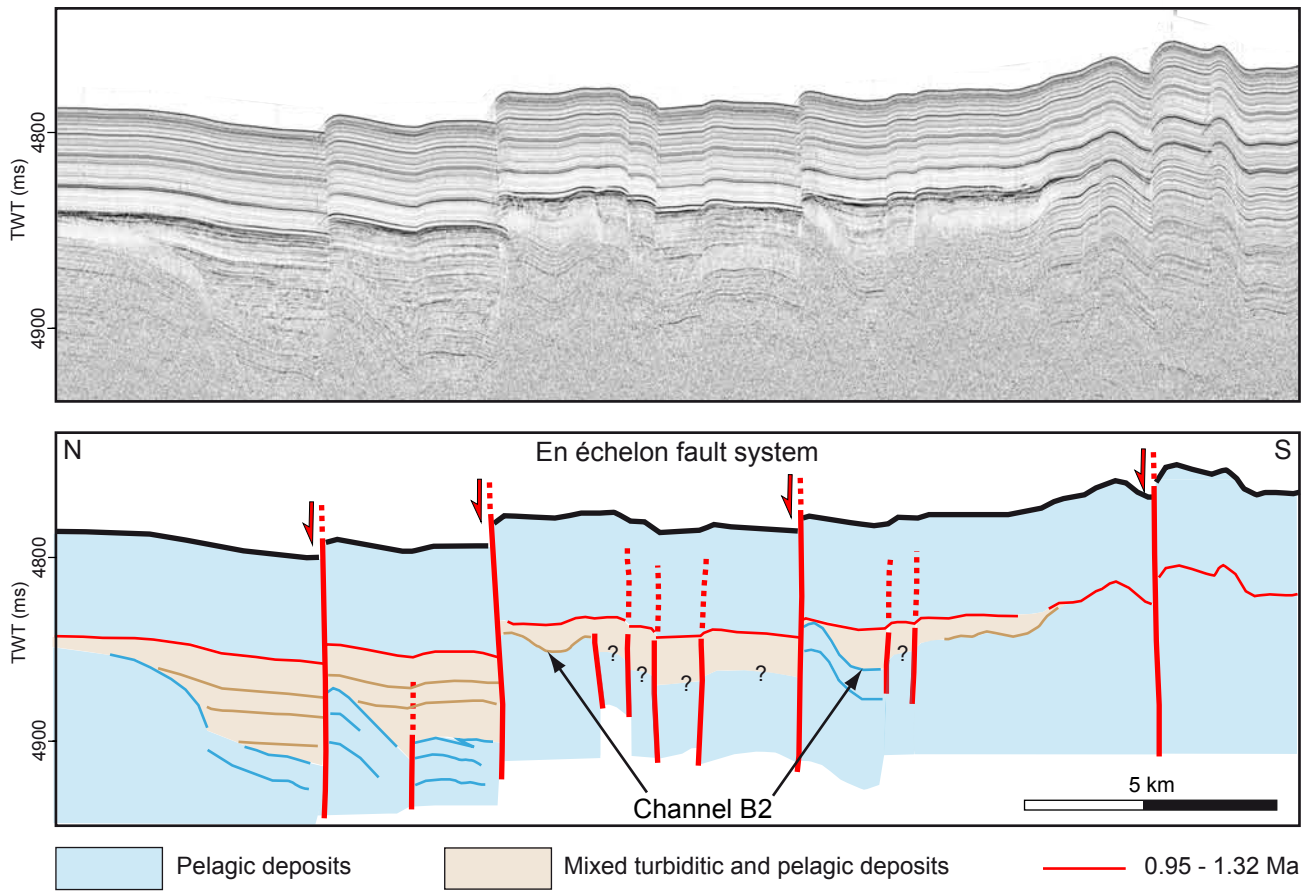


Figure 8

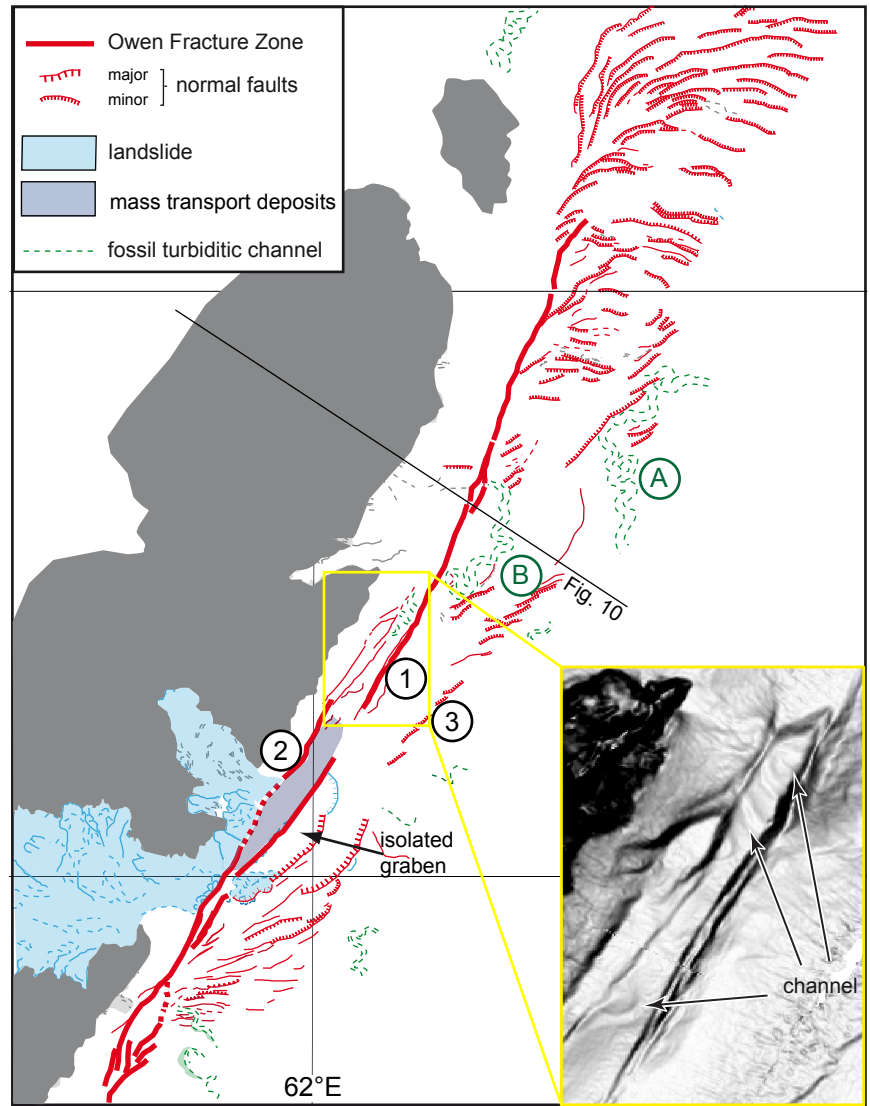
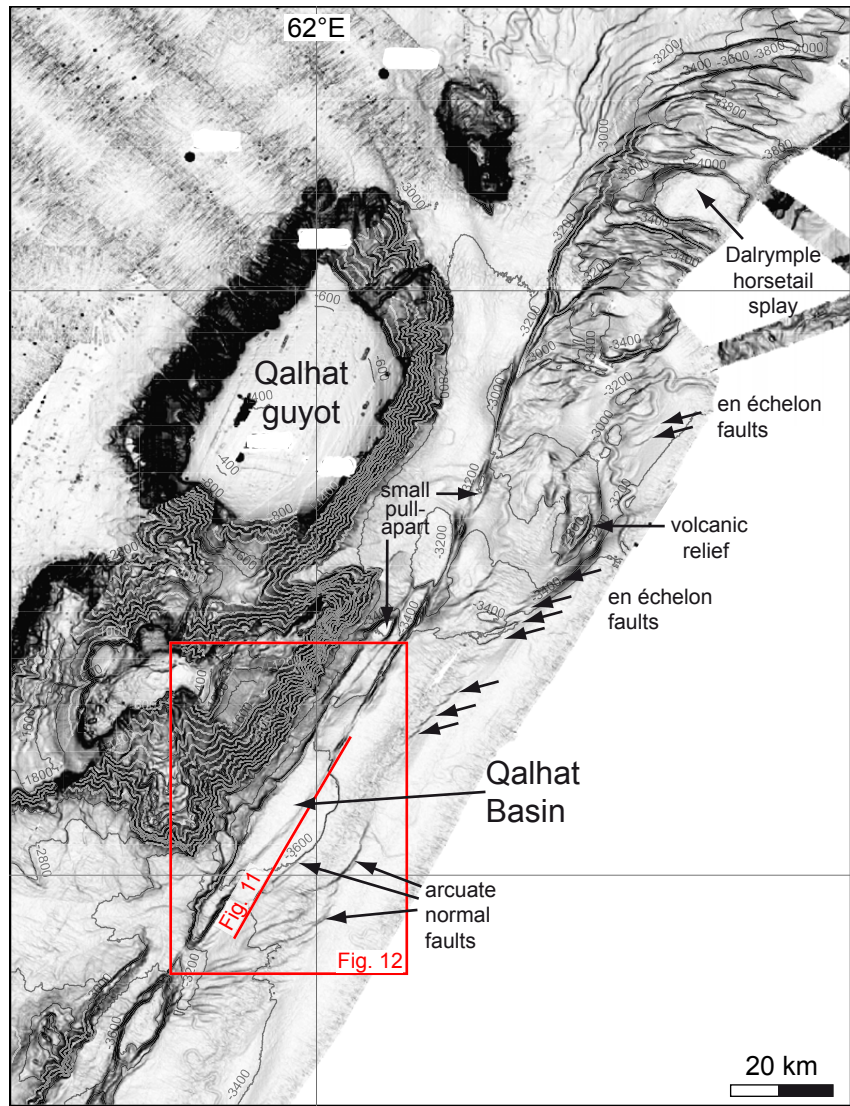


Figure 9

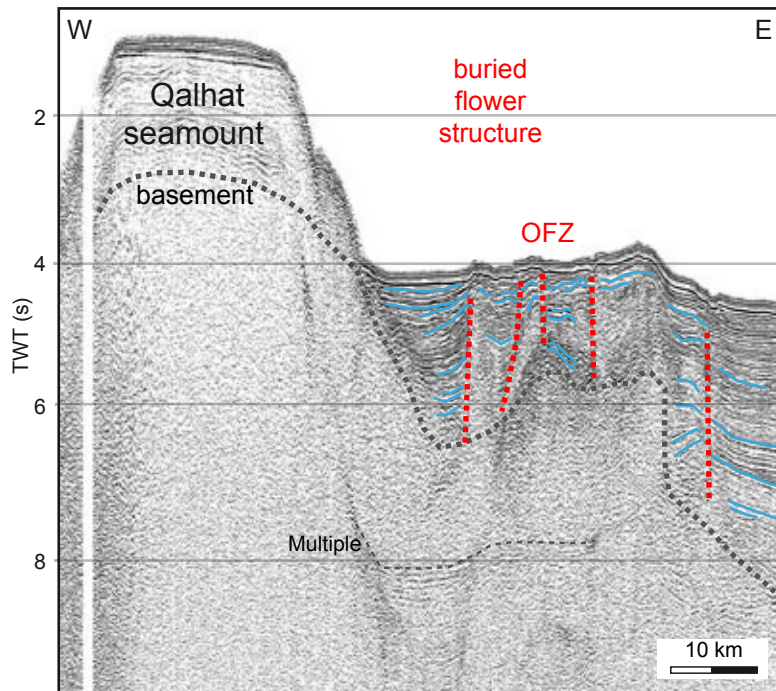


Figure 10

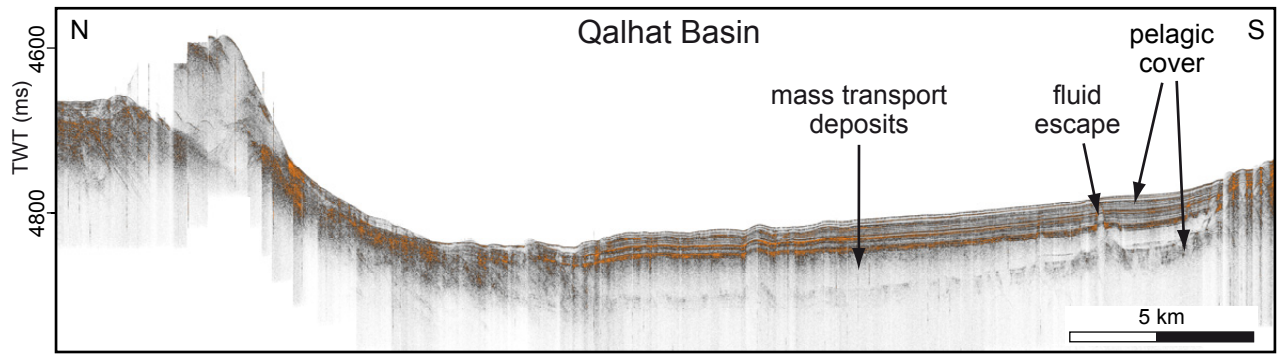


Figure 11

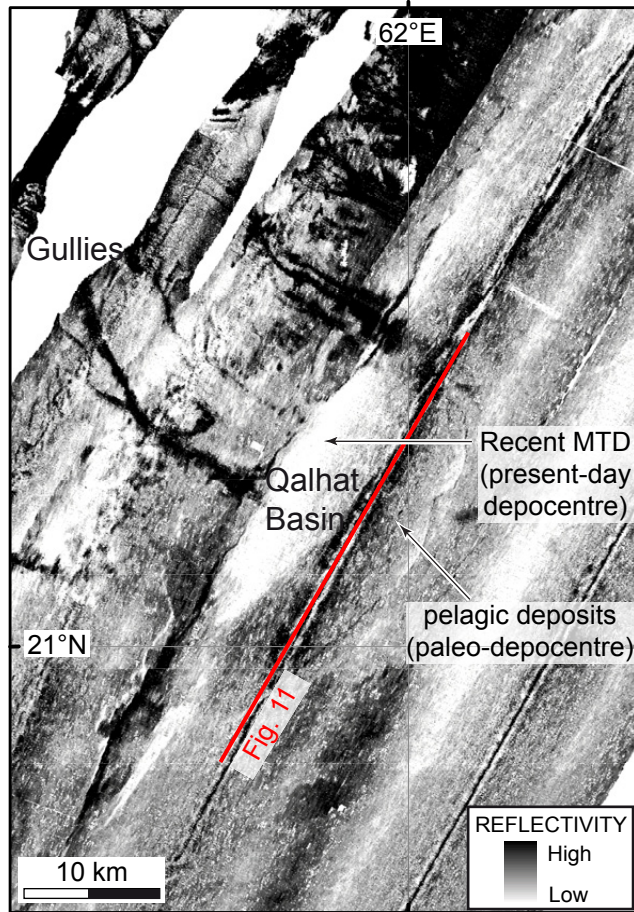


Figure 12

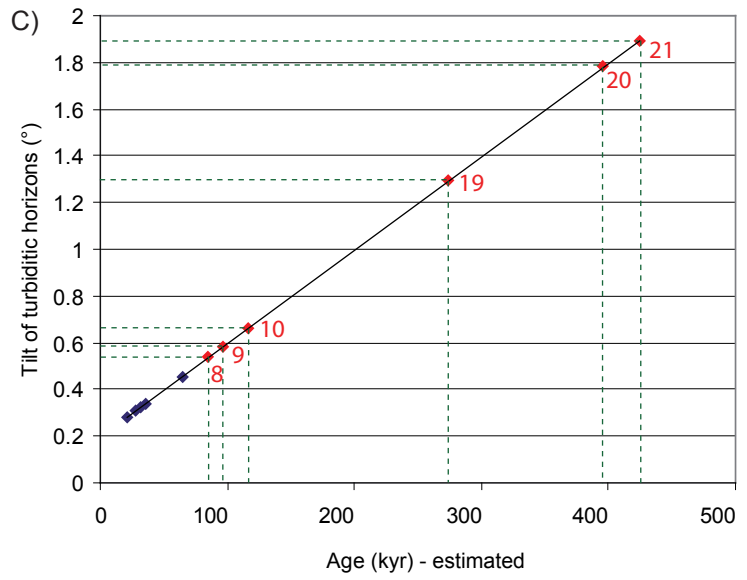
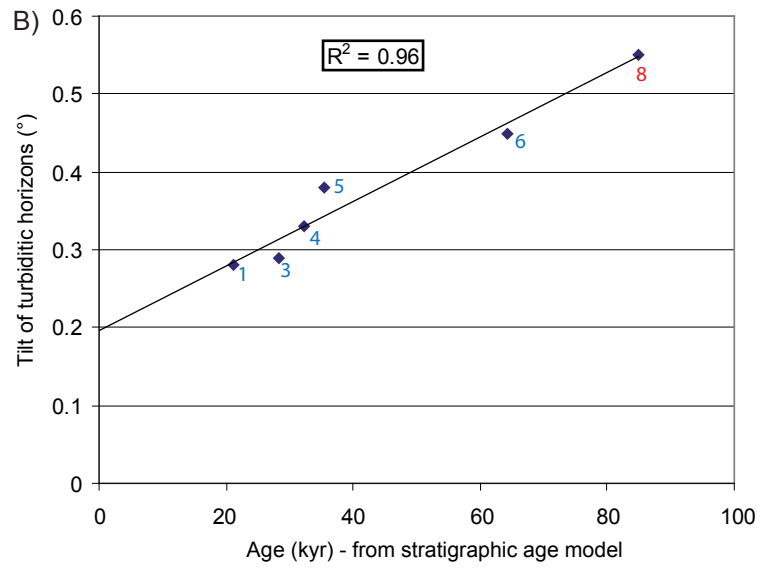
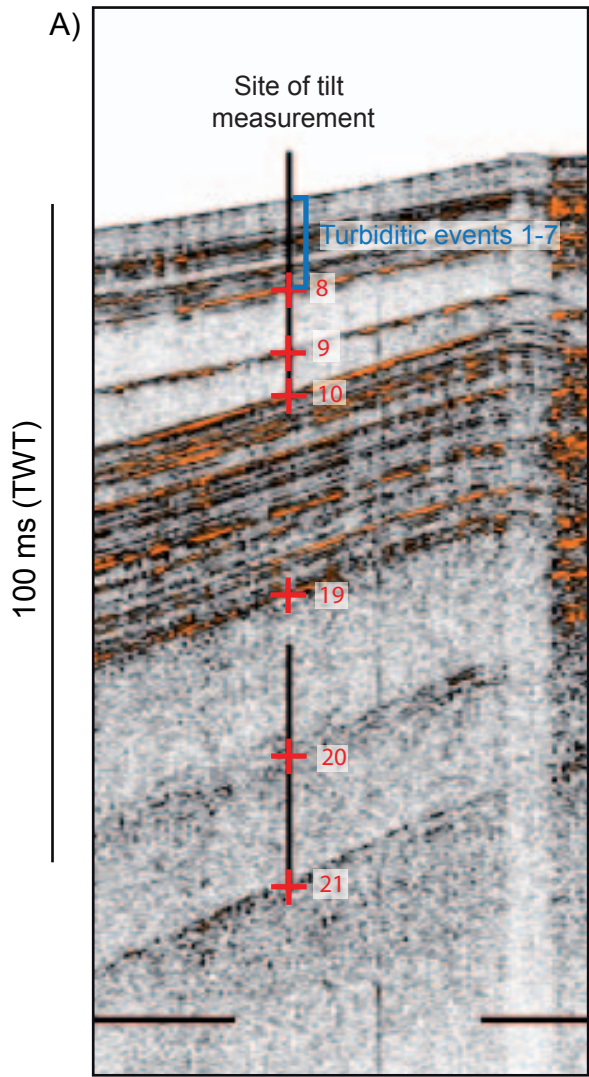


Figure 13

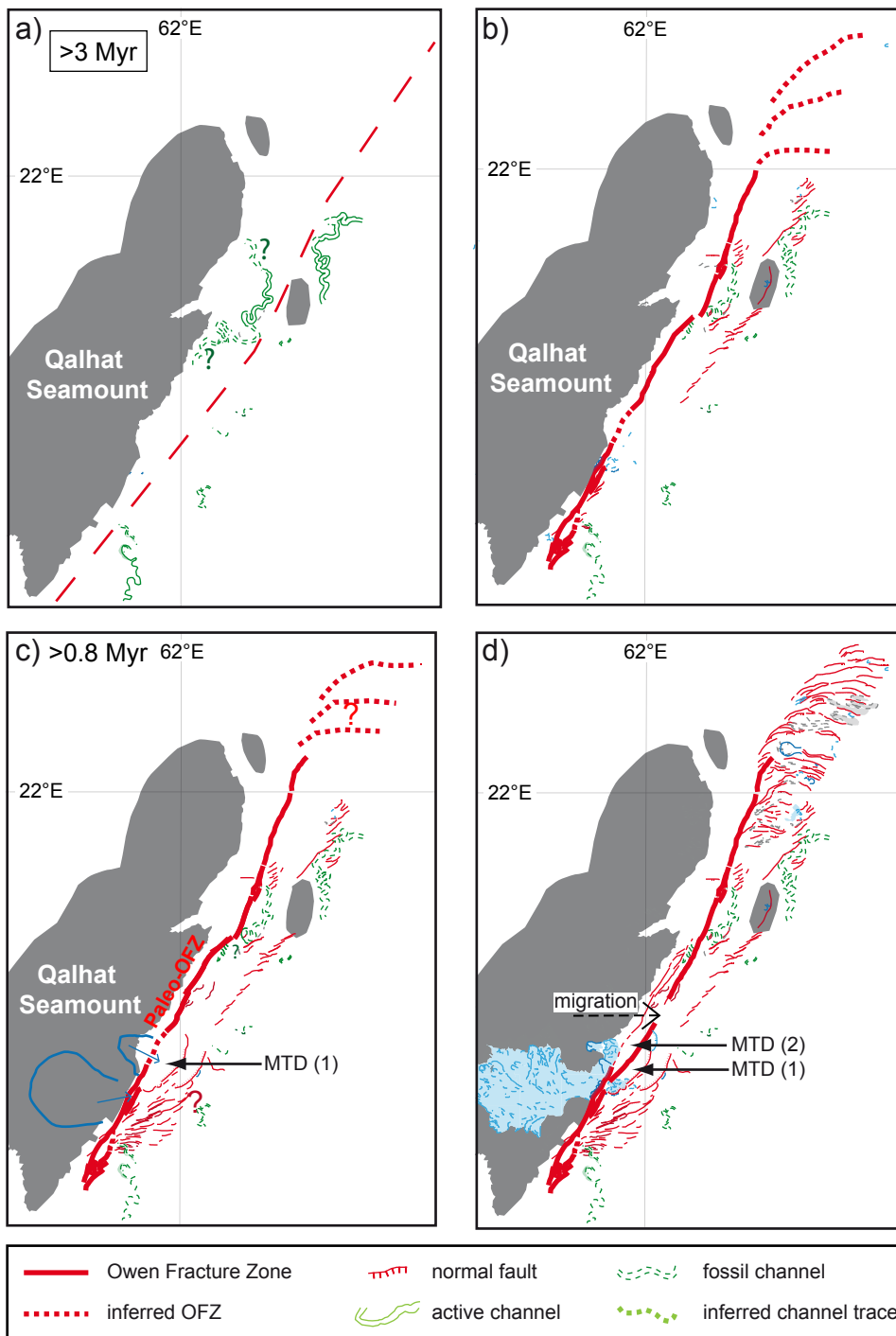


Figure 14

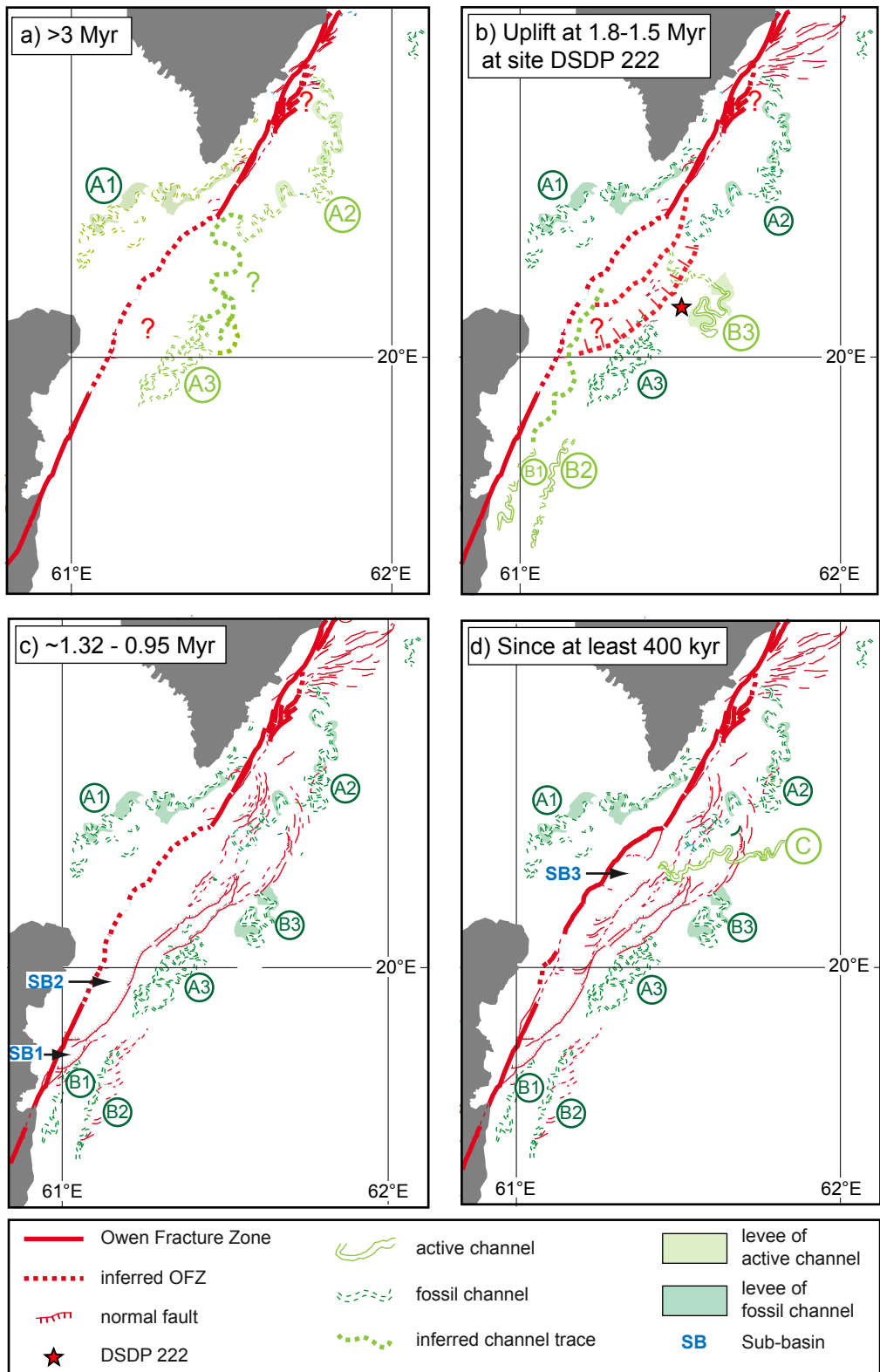


Figure 15