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► To cite this version:

Newdeskarl Saint Fleur, Yann Klinger, Nathalie Feuillet. Detailed map, displacement, paleoseismology, and segmentation of the Enriquillo-Plantain Garden Fault in Haiti. *Tectonophysics*, 2020, 778, pp.228368. 10.1016/j.tecto.2020.228368 . hal-03063915

HAL Id: hal-03063915

<https://hal.science/hal-03063915>

Submitted on 14 Dec 2020

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1 **Detailed map, displacement, paleoseismology, and segmentation of the Enriquillo-Plantain Garden Fault**
2 **in Haiti**

3

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8

9 **Abstract.-** The Enriquillo-Plantain Garden Fault (EPGF) ruptured several times in Haiti, producing large
10 historical earthquakes (e.g., 18 October 1751, 21 November 1751, and 3 June 1770). Their location and lateral
11 extents are poorly known. The most devastating one, the Mw7, 12 January 2010 earthquake should have been
12 the occasion to constrain the fault kinematics and geometry, unfortunately no significant surface rupture was
13 observed. Here, combining LiDAR data, high-resolution aerial photographs, geological map and observations in
14 the field, we revisited the fault map at a large variety of scales to improve the assessment of seismic hazard
15 associated with the EPGF. This detailed mapping allowed us to target several sites for paleoseismic trenches. We
16 carried out paleoseismological investigations in the Clonard pull-apart basin during which one trench was
17 opened perpendicular to the fault. That trench showed stratigraphic evidence for one paleoearthquake.
18 Radiocarbon dating in a dry and buried stream led to the conclusion that the paleoearthquake might be that of
19 1770. Then, based on kilometric-scale geometric complexities, we showed that the EPGF was composed of more
20 than four segments, each capable to generate Mw>7 earthquakes. Finally, we identified several geological and
21 geomorphic markers of long-term deformation along the fault. The geological offsets of ~40 km, ~15 km and ~8
22 km from west to east support the idea of an eastward propagation of the EPGF.

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24 **1. Introduction**

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26 The left-lateral strike-slip Enriquillo-Plantain Garden Fault (EPGF) extends over a length of about
27 1100 km, with a mean direction of N85°E, from the Cayman Trough to the west, to the Enriquillo Valley

28 (Dominican Republic) to the east. . It is part of a complex transpressive plate boundary between the North
29 American and Caribbean plates which converge at a rate of about 2 cm/year in a NE-SW direction. This main
30 fault runs on the southern peninsula of Haiti where it is broadly well expressed (Fig. 1) and thus has long been
31 identified [e.g., Duplan, 1975; Mann, 1983; Vila et al., 1985]. Mann et al. (1995) proposed a summary map of
32 the fault, using mainly SEASAT radar imagery and aerial photographs (scale : 1/40000). That map pointed out
33 the main typical geomorphic markers of a strike-slip fault zone such as linear valleys, pull-apart basins and
34 pressure ridges. The fault affects Quaternary deposits, attesting to its activity [Mann et al., 1995]. Several
35 historical earthquakes, in 1701, 1751 and 1770 were attributed to the EPGF, using macroseismic data [Ali et al.,
36 2008; Bakun et al., 2012; McCann, 2006]. Comparing the intensity of the 2010 earthquake to those of the
37 historical ones, Bakun et al. (2012) assigned a M_l 6.6 to the 9 November 1701 event. In 1751 two large
38 earthquakes occurred in southern Hispaniola within one month interval. The first one on 18 October of M_l 7.4-
39 7.5 to the east of the EPGF system may be due to an offshore rupture of the Muertos Trench (Fig. 1, inset). The
40 second one on 21 November of M_l 6.6 occurred near Port-au-Prince is, with less ambiguity, attributed to the
41 EPGF [Bakun et al., 2012]. The 3 June 1770 event, of M_l 7.5, is located broadly to the west of the Lake
42 Miragoâne [Bakun et al., 2012].

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56 Based on the date of OCCURRENCE of the last event on the fault and the strain rate indicated by the GPS data,
57 Manaker et al. (2008) estimated that the slip deficit along the EPGF fault was large enough in 2005 that it could
58 break in a near future to produce a M7 class earthquake earthquake. Although this was probably verified few

59 years later when the 2010 earthquake occurred (by considering that the rupture broke the EPGF, as inferred by
60 Saint Fleur et al. (2015)), this study was preliminary because 1) the segmentation of the fault had not been
61 considered; 2) the lateral extents of the historical events were not known ; and 3) the GPS data did not have
62 enough resolution that would permit to estimate slip rate deficits on separated sections of the fault. To better
63 characterize the seismic hazard associated with the EPGF, paleoseismic studies were needed. Gold *et al.* (2012)
64 carried out paleoseismic investigations along the eastern part of the fault, especially, in the Momance Valley and
65 eastern Pétion-Ville (Fig. 2a). The choice of this part of the fault was only based on its geomorphic expression in
66 this area. They excavated three trenches in which they identified recent ground ruptures that might be associated
67 with the 1751. Unfortunately, the radiocarbon and luminescence samples yielded ages that could not be
68 interpreted with confidence (Prentice, pers. com., 2013).

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70 In this paper, we mapped in detail the Enriquillo-Plantain Garden Fault, mainly by using high-
71 resolution aerial photographs (pixel 30 cm) and Light Detection and Ranging (LIDAR) data (pixel 1 m)
72 complemented with field observations. We first worked at the scale of the southern-peninsula to map the main
73 traces of the fault by using an ASTER Digital Elevation Model (DEM) with a horizontal resolution of 30 m. On
74 the basis of this first order mapping, we then divided the southern peninsula into five parts to map the fault in
75 much more detail from east to west (Fig. 1). For each part, we gradually decreased the scale of the observations
76 to reach a resolution of few tens of meters depending on the dataset and we combined them to observation in the
77 field when possible. . Then, based on criteria of morphological analysis, fault geometry, sedimentary record, and
78 accessibility [e.g., *Klinger, 2007; McCalpin, 1996*], we selected one site for a paleoseismological
79 investigation. An exploratory trench was excavated and has shown stratigraphic evidence for one
80 paleoearthquake. Indirect radiocarbon dating suggests that this paleoearthquake likely corresponds to a event
81 during the historical period. In addition, we collected information for several other favorable sites for future
82 paleoseismic investigations. Our mapping at all scale eventually allows us for discussing the
83 geometry of the fault and for proposing a first-order segmentation model of the fault. The seismic
84 hazard associated with the segments and the geological and morphological offsets along them are also
85 discussed.

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87 **2. Tectonic background**

88 The formation of the island of Hispaniola is marked by plutonism, volcanism and island-arc
89 metamorphism during the pre-Aptian time. The island-arc (the northern and central parts) has undergone several
90 deformation phases, mainly folding, which was active until the Lower Miocene [*Mann et al.*, 1991a]. A major
91 episode in the formation of Hispaniola was the suture, in the Lower Miocene, of the island-arc with the southern
92 peninsula [*Mann et al.*, 1991a]. Besides this major episode occurred in Lower Miocene, the sharp escarpments
93 bordering the structures of the island-arc made *Mann et al.* (1991a,b) suggest transpressive tectonics from the
94 Late Miocene to the present. As the island-arc, NW-SE Lower Miocene folds (e.g., Massif de la Selle, Massif de
95 la Hotte, Fig. 1) have been documented along the southern peninsula of Haiti [*Momplaisir and Boisson*, 1988;
96 *Mann et al.*, 1991a; *Mann et al.*, 1995]. During the Middle and Upper Miocene time, sandstone, marls and
97 chalky limestone unconformably deposited on the borders of the later folds.

98 Using magnetic data, *Pindell et al.* (1988) suggested that the EPGF initiated at the Middle Eocene,
99 propagating eastward from the Mid-Cayman Spreading Center (Fig. 1, inset). The fault seems to have reached
100 the southern peninsula of Haiti at the Pliocene time, disrupting Late Miocene terranes [*Wessels et al.*, 2019;
101 *Calais and Mercier de Lépinay*, 1995; *Calmus*, 1983; *Calmus and Vila*, 1988; *Cooper*, 1983; *Heubeck and*
102 *Mann*, 1991a; b; *Mann et al.*, 1984; *Van Den Berghe*, 1983; *van den Bold*, 1975]. However, based on
103 stratigraphic and structural analyses, *Mann et al.* (1991b) suggested an earlier episode of E-W strike-slip
104 deformation in the peninsula from the Eocene to the Lower Miocene. This episode would be followed, since the
105 Miocene, by the present-day transpressional regime. Thus, in the Pliocene time, the EPGF would have
106 reactivated old zones of weakness and crosscut en echelon folds along the southern peninsula of Haiti.

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110 About fault segmentation, *Leroy et al.* (2015) had identified several segments of the EPGF offshore
111 between Haiti and Jamaica using high-resolution multi-beam echo-sounder and shallow seismic reflection data.
112 The segments crosscut pre-existing structures inherited from former tectonic regime [*Leroy et al.*, 2015]. *Cowgill*
113 *et al.* (2012) summarily identified two main segments along the southern peninsula: an eastern and a western
114 segments separated by a huge step-over extended from Fayette to Grand-Goâve bay. They proposed a conceptual
115 model of earthquake behavior along the EPGF. According to that model the 1751b earthquake would have
116 broken the eastern segment from Dumay to Fayette, and the 1770 event would have broken the western segment

117 from Grand-Goâve bay to Clonard pull-apart [Cowgill et al., 2012; Bakun et al., 2012]. The 2010 earthquake
118 would have broken the step-over from Fayette to Grand-Goâve bay, between those two major segments [Cowgill
119 et al., 2012]. After the 2010 Haiti earthquake, several studies have focused on the area of Léogâne and the Cul-
120 de-Sac plain. Using high-resolution topographic data and aerial photograph Saint Fleur et al. (2015) mapped in
121 detail fault around the epicentral area of the 2010 earthquake. We found that the strike-slip deformation induced
122 by the EPGF stops near Pétion-Ville, southern Port-au-Prince and then a WNW-ESE fold and thrust network
123 prevails eastward. That pattern was confirmed by Calais et al. (2016) using GPS data. They highlighted a strong
124 component of shortening around the Cul-de-Sac basin, while the GPS vectors are nearly parallel to EPGF fault
125 trace western Léogâne. Saint Fleur et al. (2019) analyzed the drainage network and found that the paleofan of
126 Port-au-Prince was offset by ~8 km across the EPGF. In addition to the strike-slip deformation, we used high-
127 resolution topographic data, aerial photographs, geological information and field observations to map in detail
128 the WNW-ESE folds and thrusts in Port-au-Prince and the Cul-de-Sac – Enriquillo plain. A model of nearly pure
129 strike-like from Tiburon to Pétion-Ville, then transpression and compression on the Dumay Fault and on the
130 Bahoruco Fault, respectively, was inferred [Saint Fleur et al., 2019]. Seismicity, focal mechanism and crustal
131 structure analysis in southern Haiti revealed a transition in the EPGF system from a near vertical strike-slip fault
132 along the Southern Peninsula to a southward-dipping oblique-slip fault along the southern border of the Cul-de-
133 Sac-Enriquillo basin (Possee et al., 2019; Rodriguez et al., 2018). In the lake Azuéli Cormier et al. (2019)
134 acquired 220 km of multichannel seismic reflection profiles across its surface. They found two major structures:
135 a NW-SE fold bounds the western side of the lake and another one to the southern side. The latter is poorly
136 imaged below a shallow gas front [Cormier et al., 2019]. However, the study of Wang et al. (2018) proposed
137 that the folds are produced by shear along the EPGF that continues all along the Cul-de-Sac – Enriquillo basin
138 including in the lake Azuéli.

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198 3. Detailed mapping of the EPGF

199 3.1. The Momance Valley

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201 The Froide and Momance rivers are the most prominent drainages controlled by the EPGF in the
202 southern peninsula. The valleys of the latter rivers are linear and mark the morphology (Fig. 2-5) [*Mann et al.*,

203 1995]. Figure 2a shows that the fault is morphologically well expressed along the Pétion-Ville - Léogâne section
204 where the main rivers (Momance and Froide) are left-laterally offset by the fault to form two dog-legs (Fig. 2a,
205 inset). In the area where the valleys are running parallel to the fault, the latter is ~35 km long, up to 700 m high,
206 linear, and has a steep uphill-facing (i.e. south-facing) scarp (Fig. 2, and 3). The scarp portrays numerous
207 geomorphic features of ongoing uplift north of the EPGF surface trace. The sharp uphill-facing
208 Enriquillo Plantain Garden Fault scarp is incised by numerous small, narrow, intermittent
209 gullies, some flowing in perched wineglass valleys, implying that transverse stream incision has
210 not kept pace with relative tectonic uplift (Fig. 4). In addition, numerous wind gaps, ponded sediments, upstream
211 of deep gorges incised by streams flowing towards the bay of Port-au-Prince may indicate a drainage reversal
212 due to tectonic uplift. Overall, the Enriquillo-Plantain Garden Fault and the Lamentin thrust bound an uplifting
213 block which is limited to the south by the salient EPGF scarp [Saint Fleur et al., 2015].

214 The two drainages (Momance and Froide) are separated by a 500-m-wide, elongate ridge (Chafa). West
215 of the Chafa ridge, on two meanders of Momance river, we can observe a sharp uphill-facing scarp through the
216 limestone (Fig. 5b, c, Fig. S1a, b) mapped as Eocene [*Bien-Aimé Momplaisir*, 1986].

217 On the eastern meander and on the southern wall of Momance river, a ~N75°E fault has been identified
218 [*Lacassin et al.*, 2013; *Prentice et al.*, 2010]. This fault exhibits a ~5-m high uphill-facing scarp evidenced by an
219 orthogonal topographic profile on the LIDAR DEM (Fig. 5b, inset). The fault blocks and deflects the tributaries
220 of the main river. In the center-east of figure 5b, we note the clearest offset tributary that exhibits a nice dog-leg.
221 As we consider that the stream has been offset because of slips on the fault during several earthquakes, we
222 tentatively reconstruct, by back-slipping, its initial geometry. Since we mapped the boundaries of the terraces of
223 the stream, they can be used as piercing lines. The terraces are the abandoned beds of the stream. Thus, we can
224 infer that the stream used to flow along those terraces before being offset. We first consider that the stream used
225 to flow along the western terrace of the downstream; we match the upstream to the border of this western terrace.
226 We obtain a maximum offset of about 60 m (Fig. S1c). A minimum offset of about 48 m is also obtained by
227 matching the upstream to the border of the eastern terrace (Fig. S1d). But, the most probable reconstruction is
228 done by corresponding directly the upstream to the downstream. This reconstruction yields an offset of about 55
229 m (Fig. 5b, Fig. S1e). This reconstruction appears to put the western and the eastern terraces of the downstream
230 in front of those of the upstream. Our most probable offset measurement is comparable to the 52 +18/-13 m of
231 *Cowgill et al.* (2012), using LIDAR data. In this area, minor lateral offsets (1.8 - 2 m) have also been measured

232 in the field [Prentice *et al.*, 2010] and may be associated with one historical earthquake. This morphology is well
233 preserved in spite of the high erosion rate in the region. At that scale the geometry of the fault is quite
234 ambiguous. Indeed, at the westernmost border of this eastern meander, our field observations confirm an uphill -
235 south-facing scarp (Fig. 5d) suggesting a north-dipping fault, assuming that the fault has a small reverse
236 component in this area. However, in less than 100 m to the west, at eastern Jameau, the fault exposes a \sim N100°E,
237 60°S plane on the northern wall of Momance river (Fig. 5e) [Prentice *et al.*, 2010; This study].

238

239 At the Conette site (the western meander), a N84°E fault is observed. Topographic profile on the
240 LIDAR data shows that the fault scarp is about 13.5-m high (Fig. 5c, inset). This fault clearly offsets left-
241 laterally a channel. We tentatively propose to reconstruct this channel. We first, by back-slipping, put the eastern
242 terrace of the upstream in front of the western terrace of the downstream. We obtain the maximum possible
243 offset of about 203 m (Fig. S2c). This maximum offset may be overestimated, as it is simpler to match the
244 western terrace of the upstream to the western terrace of the downstream. In the same manner, we propose a
245 minimum possible offset of about 158 m (Fig. S2d). Indeed, this minimum offset is obtained by putting the
246 western terrace of the upstream in front of the eastern terrace of the downstream. It is thus the least offset that we
247 can get. More realistically, we try to match the southern part of the downstream and the northern part of the
248 upstream, following their general direction. This reconstruction yields an offset of about 169 m (Fig. S2e). It is
249 our best reconstruction and is compatible with the 163 m found by Prentice *et al.* (2010) using field
250 measurement. We can follow this channel northward where it abuts against another less significant E-W fault
251 scarp. This scarp is \sim 6.5-m high (Fig. 5c, inset). The channel appears to be also offset by this second fault.

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293 3.1.1. The site of Fayette (Jean-Jean)

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295 The site of Fayette is located at the western outlet of the linear Momance Valley. This site is at the
296 junction of at least three geologic formations (Fig. 6a). At this place, the fault passes through the basement
297 (Cretaceous basalt) of the southern peninsula and continues eastward through the Eocene limestone. North of the
298 fault trace, the village of Fayette is built on a terrace of Momance river. This northern part is less eroded than the
299 southern one. Several landslides, which are probably induced by the 2010 earthquake [*Gorum et al.*, 2013], are
300 observed at the base of the less eroded northern part. On a 3D view of the area of Fayette (Fig. 6b) we can see
301 that mounds on the southern wall of Momance river are clearly truncated by a linear fault. The drainage network
302 that is coming from the southern block undergoes strike-slip deformation. Indeed, detailed mapping of this area
303 shows offset, deflected and beheaded channels (Fig. 6c). Some of the drainages stop abruptly at the piedmont of
304 the mountain, near the fault scarp. They abut against fans, probably their own fans, without re-incising them.

305 Few tens of meters toward the north, we find other incisions with no clear catchments. Such a pattern may
306 indicate that the fans are abandoned. The upstreams, instead of re-incising the fans, would infiltrate at the fault
307 scarp and reappear further north.

308

309 In this site of Fayette, Prentice *et al.* (2010) measured in the field 7 minor channel offsets. They are
310 ranging from 1.3 to 3.3 m. Those offsets may be associated with one of the eighteenth century earthquakes
311 [Prentice *et al.*, 2010].

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313 3.2. Léogâne-Miragoâne Complex Zone

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315 From Léogâne to Miragoâne, the Enriquillo-Plantain Garden Fault consists of a ~45-km long zone of
316 geometric complexities (Fig. 7a). The fault is well expressed along the Momance Valley. In contrast, from
317 Fayette to Dufort the fault is not well expressed, except the sharp topographic limit between the basalt of the
318 Massif de la Selle and the Quaternary deposits of Léogâne Delta Fan. From Dufort, no strike-slip deformation is
319 clear over a distance of ~17 km until the town of Grand-Goâve where evidence for reverse faulting is rather
320 seen. To the south of the town of Grand-Goâve, a linear hill front is observed, separating Pliocene rocks to the
321 south and Quaternary deposits to the north (Fig. 7b). On the hill, we identified several hanging valleys that attest
322 to active reverse faulting. Figure 7c shows details of an old tributary of Grand-Goâve river. This tributary left its
323 ancient trace on the hill characterized by alluvial material in a sinuous depression. This environment is quite
324 favorable for dense low vegetation. Following this vegetation, we reconstituted the trace of the old stream until
325 its paleofan. This valley is perched at ~100 m.

326

327 West of Grand-Goâve, the strike-slip fault reappears along the ~N110°E Tapion pressure ridge (Fig.
328 7d). This typical spindle-shaped restraining bend is bending about ~17° of the fault in this area. It may have first
329 formed the pressure ridge and offset it later or simultaneously. As a restraining bend, the Tapion pressure ridge
330 may be bordered by secondary thrusts. Hence, an offshore thrust may explain the abrupt northern coast of Tapion
331 where numerous landslides and uplifted coral reefs were observed (Fig. 7e).

332

333 From Tapion pressure ridge, the EPGF likely continues offshore for only about 4 km before being quite
334 visible onland along the southern flank of Fort-Royal hill. At this place, the fault marks the morphology by a
335 linear slope break. Similarly to Tapion pressure ridge, the northern coast of Fort-Royal hill is abrupt and exhibits
336 uplifted beaches (Fig. 7f). The beaches are uplifted by up to ~4 m. These observations may be compatible with
337 an offshore thrust, likely the Trois-Baies Fault, bounding the hill (Fig. 7a).

338

339 After running along southern Fort-Royal, the EPGF steps down to form an up-to-5-km wide and 20-km
340 long depression: the Miragoâne pull-apart. With about 43 m depth, the lake Miragoane, the main feature of the
341 pull-apart, is the deepest lake of the caribbean [*Brenner and Binford*, 1988; *Higuera-Gundy et al.*, 1999; *Wang et*
342 *al.*, 2018]. Sonar survey in the lake revealed that the sediments are highly deformed mainly by normal faulting,
343 typical of a pull-apart basin [*Wang et al.*, 2018]. Core and pollen data from the lake Miragoâne revealed that the
344 formation of the pull-apart would be at least 43kyr [*Wang et al.*, 2018].

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399 3.3. Fond-des-Nègres area

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401 West of the Léogâne-Miragoâne Complex Zone (LMCZ), in the middle of the southern
402 peninsula, the EPGF is very well expressed and has a clear single trace in the morphology.
403 A hillshade image of an ASTER DEM superimposed by the 1/250000 geologic map of Haiti
404 [Momplaisir and Boisson, 1988] enables us to combine the information given both by the
405 topography and the geology (Fig. 8). The fault cuts across the southern flank of Rochelois
406 plateau and juxtaposes the Cretaceous and Paleogene units composing the plateau and
407 Neogene units and Quaternary deposits which pond in Fond-des-Nègres basin. North of the
408 town of Fond-des-Nègres, the fault exhibits a ~24-m-high uphill-facing (i.e. north-facing)
409 scarp (section AA', Fig. 8b, inset). The fault continues westwards where it offsets a part of

410 the drainage network: many small gullies that flow southwards, on the northern compartment
411 of the fault, are disrupted by the fault and remain confined
412 on its northern side instead of flowing into the main Serpente river (Fig. 8b)
413 We observe the same feature eastward in the Jacquot area as highlighted by
414 very high-resolution aerial photograph (pixel 30 cm, Fig. 9. At this scale, the fault trace is long and
415 linear and made of several few kilometer-long small segments . We can observe that the drainage network
416 flowing from the Rochelois plateau is also disrupted by the fault . As a result, several ponds and small basins
417 have developed along the fault scarp. East of this area , we can observe that some gullies are even deflected in a
418 sinistral sense before flowing in the ponds. ... The Fond-des-Nègres area, especially the three basins shown in
419 figure 9a, has been investigated in the field for further paleoseismological studies. In the following we described
420 in more detail on of this site (Ka' Jacquot pond, Fig. 9b). (.

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422 3.3.1. Ka' Jacquot pond area

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424 In the Ka' Jacquot pond area, the EPGF cuts across Paleogene limestones, highly eroded due
425 to intense tropical rainfalls. Its scarp is high enough(>10 m high) to . . block the drainage network, forcing the
426 water to flow along its base , before to fill one of the small basins. The southward view of the landscape
427 (Fig.9b) strikingly illustrates this relationship. We can clearly distinguish a well-developed alluvial fan of a
428 stream that abuts against the scarp. We inferred that the stream used to flow downward towards the lower
429 topographic points (beyond the scarp to the south) before having been blocked by the progressive growth (by
430 cumulative slips) of the EPGF fault scarp. This scarp now acts as a shutter ridge forcing the water to flow more
431 eastward into the Etang Ka' Jacquot basin. This process is confirmed by the extent of the alluvial fan eastward.
432 The intensive erosion by the water flow contributes to increase the height of the scarp enhancing the effect of the
433 tectonics. West of the alluvial fan the fault scarp gradually decreases in height (Fig. 9a).
434 . In any case, the fault is active enough to defeat the whole drainage network in this area.

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451 3.4. L'Asile, Clonard pull-apart and Camp-Perrin basins

452 L'Asile basin is located between the Rochelois plateau to the east and the Clonard basin to the west
453 (Fig. 10a). It is filled with continental Miocene units which lie unconformably upon the older Cretaceous
454 folded units [Bien-Aimé Momplaisir, 1986]. In this area, the scarp of the EPGF is less clear than in other sites.
455 However, it can be evidenced by analyzing slight variations of slopes in the basin (Fig. 10b). The basin can be
456 clearly distinguished in the slope map because it forms a rather flat area (with slopes ranging between 0 and
457 36%) compared to the surrounding higher reliefs. It is crosscut by discontinuous few kilometers-long linear E-
458 W-striking slope breaks (between 0 and 10°), which highlight the fault trace. West of the basin, a linear E-W-
459 striking slope break is continuous over 5km. It corresponds in part to the Citronnier river bed which course is
460 deflected and controlled by faulting (Figure 10 c). Along this 5km long portion, the direction of the fault veers
461 progressively by 4° clockwise before cutting in the middle of Clonard basin westwards. (Fig. 10c).
462 We infer that this basin is a ~24-km² spindle-shaped pull-apart along the EPGF. In the middle of the Clonard
463 basin the main fault is well-expressed and exhibits a well-preserved ~2-m high scarp in the Quaternary alluvium
464 (Fig. 13). In this area, the fault controls the course of the main river (Cavaillon) flowing from the basin of Camp-

465 Perrin (Fig. 11a). aThe Camp-Perrin and Clonard basins are connected by the wide Cavaillon river bed, overall
466 forming a ~30-km long and up-to-5-km wide depression. This depression is filled with quaternary sediments
467 brought by numerous active rivers (e.g., Bonne Fin, Gros Marin) which all converge toward this area.f (rThe
468 basin of Clonard, in particular, is completely filled with late quaternary sediment which are offset by faulting
469 attesting for Late Quaternary faulting. West of the Clonard basin, several meanders and terraces of Cavaillon
470 river are crosct by the fault.In such areas, the terraces of the rivers are probably flooded regularly but the scarp
471 is still well preserved attesting for recent activity along the fault. . Near one of the westernmost meanders of the
472 river on figure 11c, a minor pull-apart basin (Laplace) is formed by a 140-m left step-over of the main fault. This
473 little basin is not bounded by two fault subsegments but it expands widely to the north and slightly to the south
474 of the fault traces. Because the basin consists of a good sediment trap, it is preferentially used for agriculture.

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492 3.5. Macaya pressure ridge and Tiburon River

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494 West of Camp-Perrin, the direction of the EPGF veers counterclockwise by 10° where it bounds the
495 southern flanks of Macaya Mountain. The highest point of this mountain is the Macaya peak (2347 m), which is
496 the second highest peak of Haiti. (Fig. 12a). Another secondary fault (southern ravine fault, SRF), is identified
497 to the north of the main EPGF and controls the course of the southern Ravine river [Calmus, 1983; This study].
498 The latter river flows in an ~E-W direction before abruptly veering to flow southwards along the western
499 border of the Camp-Perrin basin. A 20km-long, 3km wide, ~2.2-km high, ~N95°E elongate ridge is caught
500 in between the SRF to the north and the main EPGF to the south. Although, as most of the Macaya mountain,
501 this relief is partly inherited from earlier Laramide folding [Calmus and Vila, 1988], the faults may have
502 contributed to shape and enhance the relief of the ridge which we inferred to be mainly a pressure ridge of the
503 main EPGF. The shape and direction of the ridge is consistent with the overall geometry of the fault zone.

504

505 To the west of the Macaya pressure ridge, the EPGF continues linearly and controls the course of the
506 main rivers (e.g., Tiburon River). Figure 12b shows an aerial photograph of the area of Nan Cosse River. The
507 fault clearly encroaches the southern terraces of that east-flowing river before stepping down along the west-
508 flowing Tiburon River. From there, the Tiburon River runs along the EPGF, forming a linear valley. Then, the
509 fault continues offshore.

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519 4. Paleoseismic investigations – Identified potential sites

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521 During our fault mapping, we identified several potential sites for paleoseismological studies (Table 1).

522 In the areas of Fond-des-Nègres and L'Asile basin the sites were overall inappropriate because of limestones

523 outcrop along the fault scarp. However, the eastern Jacquot area and the Clonard basin particularly show better

524 stratigraphic criteria for paleoearthquake records because the different soil units are relatively thin and clear.

525 (Fig. S3). In particular, the basin of Clonard is likely a closed basin where the sedimentary record should be well

526 preserved. (Fig. 11). In addition, the fault scarp is well expressed in the Quaternary alluvium.

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528 4.1. Site selection

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530 At the western edge of the Clonard basin, we identified two favorable sites especially in the area of the

531 Cavaillon River: 1) the Laplace small pull-apart described earlier (Fig. 11c); and 2) The Siline area, where an

532 alluvial terrace is crosscut by the fault, which scarp is ~2-m high at this place (Fig. 13), consistent with the

533 observations made further east by Mann *et al.* (1995). We selected the Siline site because: 1) the fault has a

534 simple geometry; 2) the area is well fed in sediment; 3) despite unfavorable conditions such as erosion, flooding

535 and agricultural activities, the fault marks the morphology, attesting to Late Quaternary activity; and 4) the site is

536 quite accessible.

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Table 1: Accurate location of the potential trench sites identified through our detailed mapping.

Sites	Longitude	Latitude	Description
Laplace	-73.714646	18.350145	Well-defined site characterized by a minor pull-apart in the huge Camp-Perrin – Clonard valley.
Siline	-73.692107	18.348457	Site of our exploratory trench. It is located in the Clonard pull-apart, on an alluvial terrace. See text for details.
Easter Clonard	-73.635782	18.360795	Well-defined site in the Clonard pull-apart basin [<i>Mann et al.</i> , 1995; This stu
L’Asile 1	-73.503962	18.373775	Site located at the western edge of the basin of L’Asile. This site is chosen with uncertainties on the fault trace and/or sedimentary record.
L’Asile 2	-73.322251	18.372547	Site located at the eastern edge of the L’Asile basin. This site is chosen with uncertainties on the fault trace and/or sedimentary record.
Eastern Jacquot	-73.156105	18.379147	Well-defined site with stratigraphic evidence.
Fayette	-72.540099	18.471470	Site investigated by Gold <i>et al.</i> (2012)
Dumay 1	-72.190605	18.512608	Site investigated by Gold <i>et al.</i> (2012) and Saint Fleur <i>et al.</i> (2015, 2019)
Dumay 2	-72.174967	18.512568	Site investigated by Gold <i>et al.</i> (2012)

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577 4.2. Excavating an exploratory trench

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579 Despite the fact that we were satisfied by the morphology and the stratigraphy exposed along riverbeds
580 in order to excavate a trench, we faced several major problems: 1) it was nearly impossible to find an excavator
581 in the area; 2) the whole area, especially the fault scarp was used for agriculture, we had to negotiate and refund

582 the bean crops; and 3) the whole local population feared that we had come for mining reasons and challenged our
583 work. In that hostile situation, we could only excavate manually an exploratory trench (Fig. 13c). This trench is
584 ~14-m long, 1.5-m wide and 1.5-m depth. Figure 14 shows the eastern wall of the trench that exhibits fine
585 alluvial materials. The stratigraphic units were not very sharp but we could distinguish at least 5 different units, a
586 sixth one being the arable soil. From the bottom to the top, the first unit is a sandy light greyish unit. This unit is
587 topped by yellowish, blackish, dark greyish, and brownish silts and clays. The uppermost unit is the organic soil
588 made of humus, clay and silt. The three first units step down to the south (Fig. 14b, c) whereas the overall
589 morphology of the fault scarp at the surface shows a down-thrown to the north. This suggests that -only a
590 secondary branch of the fault was exposed in our trench, the main fault being likely further south Unfortunately,
591 because of the problems related earlier, we could not continue the trench to the south. The units 1, 2 and 3 are
592 vertically offset by ~20 cm by that branch..

593

594 Above the offset units, we identified a dark greyish unit (unit number 4 in Figure 14c), which is much
595 thicker on the northern compartment of the fault . This unit was likely deposited after an earthquake to fill the
596 accommodation space created by the vertical coseismic deformation (i.e, the depressed area on the downthrown
597 side of the fault) . This suggests the existence of paleoseismic horizon between units 3 and 4. The unit 5 is
598 uniformly thick (up to 1m-thick), horizontal and not deformed. It was deposited after the complete levelling of
599 the coseismic scarp by the unit 4. Overall the architecture of the fault growth sequence strongly supports the
600 record of the last paleoearthquake at this place. The earthquake offset by the same quantity (20 cm) the three
601 lower “pre-seismic” units. The unit 4 and 5 are post-seismic sedimentary units. The coseismic offset of 20 cm
602 recorded by the lower units is on an individual secondary branch of the fault. The 2m-high scarp in the landscape
603 likely corresponds to the last coseismic vertical offset along the main branch of the fault. The unit 5 is much
604 thicker (up to 1m-thick) than the other units and was likely deposited by a recent spate of Cavillon River, as
605 confirmed by eyewitnesses. Through the sedimentation dynamics at this place is ideal for preserving the entire
606 fault sequence during a complete seismic cycle, we could not unfortunately find any organic material in our
607 trench for direct radiocarbon dating. This is a classical difficulty in tropical environment where the organic
608 matter oxidizes rapidly [e.g., Zech *et al.*, 1997]. However, by chance we found a paleochannel exposure on one
609 of the main riverbanks, at the immediate northern side of the fault (Fig. 15a). This riverbank is a natural cross-
610 section that we cleaned in order to expose well the paleochannel. We found that the paleochannel was

611 approximately at the same corresponding depth as the bottom of the trench (Fig. 15b). Assuming that the fault
612 scarp observed on the terrace was formed by the last event we can infer that the paleochannel used to flow on
613 that terrace toward the main Cavaillon River prior to that event. Since the paleochannel was rich in charcoal, we
614 collected samples to have a maximum age for the paleoearthquake. The samples are HA_12-01, HA_12-02, and
615 HA_12-03.

616

617 Figure 16 shows the calibrated radiocarbon age for the three samples. We consider that the samples
618 were part of the same unit. In details, regarding the probability density, the most probable dates for HA_12-01
619 are ranging from 1717 to 1784 AD. For HA_12-02 and HA_12-03, we have 1728-1810 AD and 1726-1814 AD,
620 respectively (in green on Fig. 16). According to the coherence of the two latter date ranges, they are the most
621 representative of the real age of the paleochannel. Thus, the date of deposition of the paleochannel is 1726-1814
622 or 1770 ± 44 AD.

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643 5. Discussion

644 5.1. Paleoseismology

645 Considering that no event was attributed to the EPGF in the early 1800s [*Bakun et al.*, 2012], the event
646 found in our exploratory trench may be one of the eighteenth century's events (1701, 1751 or 1770). The
647 maximum intensities of the 1701 and the 1751 events were reported in Léogâne and Port-au-Prince. In addition,
648 the 1701 event caused severe damage from the Cul-de-Sac plain to Petit-Goâve, the 1751 (21 November) event
649 also caused severe damage in Port-au-Prince and in the Cul-de-Sac plain [*Moreau de Saint Méry*, 1803]. The
650 latter reports rather suggest that the 1701 and 1751 events ruptured an eastern segment of the EPGF. However,
651 for the 1770 event, the historical documents rather reported that the earthquake was felt in the entire island of
652 Hispaniola and in Jamaica. In Haiti, this event severely damaged the cities of Port-au-Prince, Léogâne, Petit-
653 Goâve and particularly Les Cayes (Fig. 1), suggesting that it ruptured a western segment of the EPGF [*Bakun et*
654 *al.*, 2012; *McCann*, 2006]. Thus, the ground rupture found in our exploratory trench may be associated with the
655 1770 event. The vertical offset of ~20 cm found in our trench is quite large for a secondary fault branch and the
656 2m-high scarp visible in the landscape likely corresponds to the vertical offset on the main fault. Such offsets
657 are compatible with those observed for large strike slip earthquakes worldwide (e.g., *Shi et al.*, 2019). and
658 support a large magnitude for the 1770 event which was one of the greatest earthquakes in Haiti. Such an
659 event had the potential to cause severe damage in Port-au-Prince.

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670 **5.2. Lateral Displacement along the EPGF**

671 Our detailed fault mapping enabled us to evidence several sinistral offsets of several morphological and
672 geological markers. They are of different scales, ranging from few meters to several tens of meters and even
673 hundreds of meters and they attest for quaternary activity of the fault. Although systematic dating of fault-offset
674 markers would be require to establish precise model for seismic cycle along the different segments of the fault
675 (i.e. to evidence Holocene characteristic coseismic slips for example), which is beyond the scope of this study,
676 their analysis at different spatio-temporal scales is however still useful to get clues on the longer term activity
677 and slip rate of the fault. n. As examples, by , using morphometric parameters of the drainage network, Saint
678 Fleur *et al.* (2019), have evidenced that the paleofan of Port-au-Prince and the Froide river are offset
679 horizontally by ~8 km since Pleistocene times . The offset is larger westwards as the Momance River is offset
680 by ~15 km (Fig. 2), probably over the Plio-Pleistocene, considering an eastward propagation criterion. This
681 implies a decrease of the slip rate from 4-6 mm/yr to 3 mm/yr eastward . In the following we described evidence
682 of lateral offsets of much older geological markers by the EPGF, which helps to reconstruct the history of the
683 faulting since the Pliocene.

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687 **5.2.1. Offset of the Massif de la Selle, the maximum offset of the EPGF along the southern peninsula of**
688 **Haiti**

689 The La Selle Mountain anticline is a ~N100°E, Lower Miocene anticline [*Momplaisir and Boisson,*
690 1988; *Mann et al., 1995*] that is assumed to have been offset by the EPGF (Fig. 17) [*Calais and Mercier de*
691 *Lépinay, 1995; Calmus, 1983; Calmus and Vila, 1988; Heubeck and Mann, 1991a; b; Mann et al., 1984; Van*
692 *Den Berghe, 1983*]. On the basis of our detailed fault map superimposed to the geological map and the
693 topographic information, it is possible to describe in more detail the offset of the Massif de la Selle, as well as
694 that of other anticlines further west, to get constraints on slip rates and the timing of faulting at the scale of the
695 southern peninsula (Figures 17 and 18). The morphology of the present-day northern edge of the anticline to the
696 south of the Léogâne Delta Fan suggests that this anticline is sharply crosscut and offset by the fault.

697 A striking observation is that the Cretaceous tholeiitic complex (dark blue) of the core of the anticline is sharply
698 disrupted by the fault and outcrops against the Quaternary unit of the Léogâne Delta Fan. The same geological
699 units were mapped north of the fault. They form the core of another anticline ~40 km further west, which is
700 disrupted and in sharp contact with cretaceous volcanic units to the north and quaternary units to the south. This
701 anticline compose the Fort-Royal hill.

702 The La Hotte Mountain is another anticline, having the same orientation than the Massif de la Selle. It is also
703 sharply cut to the north by the EPGF.

704

705 North of Lake Miragoâne (Fig. 8 and 17) the Fort-Royal hill is likely the offset northern closure of the Massif de
706 la Selle fold . It can not be the offset counterpart of the La Hotte anticline because in the latter case, the offset is
707 smaller (8km) than that of the Momance River (15km) which is located at the same place and which is likely
708 younger. . By considering that the fault extends eastwards to branch the contractionnal horsetail splays
709 composed of the Dumay thrust and other reverse faults on the southern edge of the Enriquillo lake, we
710 attempted a first reconstruction of the La Selle anticline. We obtain offsets slightly different when the trace of
711 the EPGF changes in the Miragoâne lake region, which is a 5km large pull-apart. It is 37 ± 10 km when the fault
712 runs south of the pull-apart and it is 40 ± 10 km when we consider that the fault cuts across the center of the pull-
713 apart We also tried a reconstruction by considering that the EPGF prolongates until the northern edge of the
714 Cabritos Island (Lake Enriquillo, Dominican Republic), which morphology was previously assumed to be

715 controlled by a section of the EPGF [Mann *et al.*, 1995]. We obtained an offset of 41 ± 10 km (Fig. S4).] Thus,
716 we retained an average offset of about 40 ± 10 km (Fig. 17b). With such a value, it is also possible to
717 reconstruct the La Hotte Anticline. The Cretaceous tholeiitic core of the southern part of the anticline strikingly
718 corresponds to the same geological units which outcrop in the L'Asile basin. However uncertainties are quite
719 large because the fold axes are not well located. Nevertheless, our estimate is comparable to those found by
720 Calmus (1984) and Van Den Berghe (1983) by using other markers in the field only. Regarding the river offset
721 (Momance and Froide) to the east over the Pleistocene and the Plio-Pleistocene, respectively, the offset of the
722 Massif de la Selle and the Massif de la Hotte to the west should have happened earlier, likely during the Early-
723 Middle Pliocene. The 40 km would imply slip rates of 7 to 11 mm/yr. Considering the uncertainties, the offset
724 could be as low as 30 km implying slip rates of 6-8 mm/yr. For the maximum offset border, that is 50 km, we get
725 slip rates ranging from 9 to 14 mm/an.

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727 **MES CALCULS (Ref: International Chronostraphic Chart)**

728 **40km/5.33Ma → 7.5 mm/an; 40km/3.6Ma → 11.11 mm/an; 30km/5.33Ma → 5.62mm/an;**

729 **30km/3.6Ma → 8.33 mm/an; 50km/5.33 Ma → 9.33 mm/an; 50km/3.6Ma → 13.88 mm/an.**

730 The decrease of the geological offsets and related slip rates from west to east is consistent with the magnetic data
731 of Pindell *et al.* (1988). Figure 18 mimics the eastward propagation of the EPGF. The NW-SE folds forming the
732 highlands of the southern peninsula are inferred to be of Lower Miocene age as their borders are made of Lower
733 Miocene rocks (Fig. 18a). Mid-Upper Miocene rocks unconformably deposited on the folds. At the same time,
734 the EPGF, having an eastward propagation, arrived in the area of the Jamaica passage, west of the southern
735 peninsula of Haiti (Fig. 18b). The EPGF continued to propagate, entering the peninsula and beheading the Lower
736 Miocene folds. Since the propagation was progressing, the amount of the western offsets increased while new
737 offsets were setting up to the east (Fig. 18c). Figure 18d shows the present-day situation where the Massif de la
738 Hotte and Massif de la Selle are offset by 40 ± 10 km. To the east of the Massif de la Selle, the fault zone mainly
739 accommodates compressive deformation that is distributed on several thrust faults. This is a fold-propagation-
740 thrust due to the suture between the southern peninsula and the island-arc, and to the present-day Hispaniola-
741 Bahamas platform $\sim 45^\circ$ -shortening direction in the area [Saint Fleur *et al.*, 2015; Wang *et al.*, 2018]. In addition
742 to the regional shortening, such a pattern is consistent with the propagation direction [Perrin *et al.*, 2016] of the
743 EPGF and the fact that the main shear zone stops in the area [Saint Fleur *et al.*, 2019].

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784 5.3. First-order segmentation of the EPGF and seismic hazard

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786 Fault segmentation represents a critical piece of information to constrain seismic cycle models and
787 various scaling laws [*Manighetti et al.*, 2007; *Schwartz and Coppersmith*, 1984; *Sieh*, 1996; *Weldon et al.*,
788 2004]. It can allow for estimating where future ruptures could initiate and stop, thus rupture length, which is very
789 important for seismic hazard assessment. A segment is assumed to have a relatively uniform behavior along its
790 length during earthquake ruptures. Different factors may control rupture length, going from mechanisms related
791 to the earthquake nucleation to fault maturity, fault geometry, or variations of the rupture velocity due to
792 properties such as heat generation, lithology, fluid content, and pore fluid pressure in the fault zone [e.g.,
793 *Michael and Eberhart-Phillips*, 1991; *Schwartz and Coppersmith*, 1984; *Segall and Pollard*, 1980; *Sibson*, 1986;
794 *Wesnousky*, 2006; *Wesnousky and Jones*, 1994]. These works seem to show that variations in the length of
795 earthquake rupture are not necessarily controlled by the relative size of initial slip pulses or the stress drops, but
796 rather by the geometric complexity of fault trace and variations in accumulated stress levels due to past
797 earthquakes along faults [*Etchebes*, 2011; *King et al.*, 1994].

798

799 Here, as no surface rupture map is available, our segmentation is based on structural observations such
800 as the superficial geology or the fault geometry [Etchebes, 2011]. The segments are separated by geometric
801 complexities or discontinuities [Knuepfer, 1989]. These geometric complexities are mainly fault azimuth change,
802 jog or relay zone, jog-bend zone and fault branching (Fig. 19).

803 It has been observed that large earthquake ruptures are generally bounded by a fault azimuth change of
804 greater than 30° [Barka and Kadinsky & Cade, 1988]. But bends do not often exceed a few degrees.

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818 The jogs can be extensive or compressive, depending on the sense of the steps with respect to the sense
819 of fault motion. The jogs may initiate, slow and stop rupture propagation [Barka and Kadinsky & Cade, 1988;
820 Duan and Oglesby, 2006; Harris et al., 1991; King and Nabelek, 1985; Klinger et al., 2006; Klinger, 2010;
821 Knuepfer, 1989; Sibson, 1986; Wesnousky, 1988; 2006]. Again, the size of a jog plays an important role.
822 Analysing 22 historical surface rupture maps, Wesnousky (2006) showed that earthquake ruptures cannot
823 propagate through jogs that exceed ~5-km large.

824

825 Along the southern peninsula of Haiti, the strike-slip deformation is relatively obvious. However, to the
826 east of Pétion-Ville the trace of the EPGF disappears where diffuse compressional deformation is observed
827 [Saint Fleur et al., 2015; 2019]. This may partly correspond to an off-fault tip splay network. Thus, and it

828 is cautious to discuss the geometry and the first-order segmentation of the EPGF, starting from Pétion-Ville to
829 Tiburon (Fig. 20).

830 During our detailed mapping, we identified both small-scale geometric asperities (tens of meters) as
831 well as large-scale ones (few kilometers). The large-scale features, except azimuth change, do not necessarily
832 stop a dynamic rupture [Harris *et al.*, 1991; King and Nabelek, 1985; Sibson, 1986; Wesnousky, 2006]. But, they
833 can initiate, accelerate or slow down rupture propagation, as it appears to be the case during the 11 August 1931
834 Fuyun earthquake, in Northern Xinjiang, China [Etchebes, 2011]. Consequently, various segmentations can be
835 obtained, depending on the scale considered, as it has already been evidenced along several strike-slip faults [Du
836 *et al.*, 2019; Allen *et al.*, 1991; Etchebes, 2011; Klinger, 2010]. To be sure of the persistence of the geometry and
837 the segmentation of the EPGF, we propose a first-order segmentation, considering only the kilometeric-scale
838 discontinuities as segment endpoints. In other words, we only take into account fault steps and bends that build
839 kilometeric-scale features (e.g., Macaya, Clonard). These discontinuities are the most susceptible to affect the
840 dynamics of earthquake rupture [Duan and Oglesby, 2006; Harris and Day, 1993; 1999; Oglesby, 2008;
841 Wesnousky, 2006; 2008]. As they are persistent over several seismic cycles, the segment lengths they delimit can
842 be used as key parameters to estimate probable earthquake size [Albee and Smith, 1966; Bonilla *et al.*, 1984;
843 Stemmmons, 1977; Wells and Coppersmith, 1994; Wesnousky *et al.*, 1984].

844

845 From Pétion-Ville to Fayette (at the outlet of Momance River), the EPGF portrays a well-defined strike
846 of $\sim N80^{\circ}E$. But, detailed mapping and field investigation along the Momance valley show several subfaults of
847 different geometries. Indeed, the eastern meander of figure 5b is crosscut by a $\sim N75^{\circ}E$ south-facing fault scarp,
848 supported by field observation (Fig. 5e). This description alone may indicate that the fault is either north- or
849 south-dipping. A south-dip may imply that the fault could even be normal or has a normal component. But,
850 neither available seismic data nor geomorphic observation allows confirming this. A north-dip would suggest a
851 reverse component in the area. However, at only ~ 100 m to the west, we observed a fault exposure that is
852 $\sim N100^{\circ}E$ -striking and dips 60° to the south. Since the latter faults are too different in strike and dip over
853 relatively short distance, they cannot be representative of the overall geometry of the EPGF. Those minor faults
854 may represent surface complexities of the main fault or secondary fractures, which are common in major strike-
855 slip fault zones [Hobbs *et al.*, 1976]. In addition, the minor faults mentioned above have very local actions. They
856 portray scarps of a few meters and displace geomorphic features by a few meters to several tens of meters. Those

857 minor faults do not appear to affect large-scale geomorphic features. At larger scale (Fig. 2), the main scarp is up
858 to ~0.7-km high, albeit it may contain a part of apparent vertical, and portrays numerous related-uplift
859 geomorphic features (Fig. 3 and 4) [Saint Fleur et al., 2015]. As the Massif de la Selle anticline is obliquely
860 offset by the EPGF (Fig. 17), this topographic high may not be explained by an eventual dip toward the south of
861 the EPGF as previously suggested [e.g., *Calais et al.*, 2010]. Figure 21b shows a stack of 64 topographic profiles
862 across the Massif de la Selle anticline. Its two flanks are well defined. Its crest is intensively eroded, consistent
863 with the stretchiness and the sparseness of the profiles in the area. However, all the profiles are strikingly tight
864 on the northern flank and particularly delineate a slope break corresponding to the EPGF. The topographic
865 profiles systematically exhibit a regular slope to the south of- and a “bulge” north of the EPGF surface trace.
866 This “bulge” unambiguously constitutes the main scarp of the EPGF that also offsets two long rivers: the
867 Momance that is offset by ~15 km, and Froide by ~8 km. Such a geomorphic expression indicates that the main
868 EPGF is constant in strike, probably steeply dips to the north, and constitutes a single segment from Pétion-Ville
869 to Léogâne (Fayette in particular). We call this segment the Pétion-Ville – Léogâne segment (Fig. 20) [Saint
870 Fleur et al., 2015].

871

872 From Dufort to Grand-Goâve, the EPGF trace is not seen in the morphology, but we only see evidence
873 for reverse faulting (Fig. 7). Thus, in terms of geomorphology, it is difficult to see the continuation of the Pétion-
874 Ville – Léogâne segment until Grand-Goâve. If we assumed that the EPGF continued, it would bend to the south
875 by about 10°. This configuration would imply a pull-apart extending from Dufort to Grand-Goâve, which would
876 contradict the proposition of Cowgill *et al.* (2012) on the existence of a compressive stepover in the area. Their
877 proposition was based on evidence for possible active folding (e.g., wind gap on Morne Diable, Fig. 7a) along
878 the southern margin of the Léogâne Delta Fan.

879 If the Dufort - Grand-Goâve section cannot be neither a compressive stepover nor a restraining bend,
880 the observed evidence for reverse faulting should be independent of the geometry of the EPGF in the area. Thus,
881 we propose that the reverse faults may be explained by a possible onland continuation of the ~N120°E Trois-
882 Baies Fault (Fig. 7, inset). At the connection of the fault systems, the dip-slip deformation may be dominant,
883 supporting the absence of surface strike-slip deformation in the area. Although it does not appear at the surface,
884 the EPGF should be present in the ~17-km-long Dufort – Grand-Goâve section, as one needs this section to form
885 the eastern bordering fault of the Tapion pressure ridge. In addition, as fault bends have been proven to be a

886 robust criterion to distinguish discrete segments [Bilham and Williams, 1985; Klinger, 2010], the $\sim 17^\circ$ of
887 azimuth change of the EPGF along the Tapion pressure ridge must separate two distinct segments. The northern
888 bordering fault of the Tapion pressure ridge is to the south of Fort-Royal hill. The latter fault, adding to its
889 possible extension through the bay of Petit-Goâve, is about 15-km long. Thus, the Fort-Royal and Dufort –
890 Grand-Goâve sections appear to have lengths for typical individual seismic segments [Klinger, 2010]. One can
891 also consider the area extending from Dufort to western Lake Miragoâne as a huge geometric complexity: the
892 ~ 45 -km-long Léogâne – Miragoâne Complex Zone (LMCZ).

893

894 Then, from western Lake Miragoâne to western basin of L'Asile, the EPGF lies along a well-defined
895 single trace (Fig. 8 and 10). The fault is $\sim N85^\circ E$ -striking. However, the dip of the fault is difficult to constrain.
896 For example, in the Fond-des-Nègres area, the fault exhibits a general north-facing scarp of ~ 10 - 20 -m high that
897 affects the Paleogene limestone (Fig. 8 and 9). Since no geophysical data are available to constrain the fault
898 structure at depth in this area, the north-facing scarp may be related to an either north- or south-dip. A dip to the
899 north would indicate a normal (extensional) component in the area. A normal component appears to be in
900 agreement with the ponds on figure 9. However, the hydrogeological context may be favorable as well to the
901 formation of the ponds. A dip to the south would indicate a reverse (compressional) component. But, during our
902 field investigations we did not see any tectonic feature that would be associated with a reverse component.
903 Hence, in the absence of any geophysical data at depth, we assume that the fault is vertical in the area. In this
904 case, the north-facing scarp is likely related to an apparent vertical or shutter ridge. The simple geometry, albeit
905 in variable scarp heights, appears to continue westward through the basin of L'Asile. Thus, from western Lake
906 Miragoâne to western L'Asile (or eastern Clonard), the geometry of the EPGF is stable enough at the southern-
907 peninsula scale to consider this stretch as a unique segment (Fig. 20). This is the Miragoâne-Clonard segment.

908

909 The EPGF bends along the middle of the Clonard pull-apart and continues along the basin of Camp-
910 Perrin (Fig. 10). In this area, the fault does not exhibit a prominent scarp, except for the ~ 2 -m-high scarp
911 observed in the field [Mann *et al.*, 1995; This study]. This morphology and geometry is relatively constant until
912 eastern Macaya pressure ridge. In addition, the $\sim 4^\circ$ of azimuth change of the fault in the Clonard pull-apart is
913 sufficient to separate two distinct segments, as a fault bend of only 2° to 3° can significantly affect rupture
914 propagation [Duan and Oglesby, 2006; King *et al.*, 2005; Klinger *et al.*, 2005; Klinger, 2010; Nielsen and

915 *Knopoff, 1998*]. Thus, with $\sim 4^\circ$ of azimuth change, we can consider that we have a new segment extending from
916 western Clonard to eastern Macaya: the Clonard-Macaya segment.

917

918

919 Then, the fault runs along southern Macaya pressure ridge (Fig. 12a). This pressure ridge is explained
920 by an inferred $\sim 10^\circ$ bending of the EPGF at the western edge of the ridge. . The geometry of the ridge suggests
921 that it may also be controlled by the Ravine du Sud Fault. To the west, the EPGF is well expressed along the
922 terraces of Nan Cosse River before stepping down along Tiburon River. The relay zone of ~ 250 -m wide between
923 the two sub-faults should be a pull-apart while it is in fact a mound (Fig. 12b). We suggest that this mound be
924 part of the overall topography both to the north and south of the fault. We then consider, at larger scale (Fig.
925 12a), that the two sub-faults are part of one segment: the Macaya-Tiburon segment.

926

927 In short, the detailed mapping of the EPGF has enabled us to distinguish at least four clear segments
928 (Pétion-Ville – Léogâne; Miragoâne – Clonard; Clonard – Macaya; and Macaya – Tiburon) along the southern
929 peninsula of Haiti (Fig. 20). We have also seen that the segments contain hundred-of-meter-scale asperities (e.g.,
930 Laplace minor pull-apart, Fig. 11) that would divide them into several sub-segments. These sub-segments may
931 play a significant role in the initiation of seismic rupture but hardly stop it [*Harris et al.*, 1991; *King and*
932 *Nabelek*, 1985; *Sibson*, 1986; *Wesnousky*, 2006]. Using high-resolution satellite imagery, Etchebes (2011)
933 mapped the 160-km long surface rupture of the 1931 Fuyun earthquake. From this mapping, she identified more
934 than 80 geometric complexities along the rupture trace. However, she proposed a first-order segmentation
935 containing only 7 segments of 10 to 40 km long. Her segments were characterized by the greater geometric
936 complexities, but mainly by their faulting style supported by their related long-term geomorphic features.

937 Similarly, we rather consider the southern-peninsula-scale segments separated by kilometric-scale,
938 geometric complexities. From east to west, the Macaya – Tiburon segment starts western Macaya pressure ridge
939 and continues several kilometers offshore of the town of Tiburon. Beside the Clonard – Macaya segment, which
940 is 32 km long, the others are ~ 50 -km long in average. However, the Clonard – Macaya segment is separated
941 from the Miragoâne – Clonard one by a relatively small azimuth change ($\sim 4^\circ$), so a large earthquake from one of
942 those segments may easily go through that geometric complexity and break both segments at the same time. For
943 example, since Port-au-Prince was severely damaged and our paleoseismic investigation highlighted the 1770

944 event-related surface rupture along the Clonard-Macaya segment, the 1770 event might simultaneously break the
945 two segments (Clonard – Macaya and Miragoâne – Clonard). Even if a rupture can go through the azimuth
946 change, the segment change may be marked by change in faulting style probably normal, as in the case of the
947 Mw 7.8, 2001 Kokoxili earthquake [King *et al.*, 2005; Klinger *et al.*, 2005; van der Woerd *et al.*, 2002].

948 On the other hand, it appears very unlikely that a rupture starting along the Miragoâne – Clonard
949 segment would reach the Pétion-Ville – Léogâne segment. This rupture should stop at the 5-km wide Miragoâne
950 pull-apart, the western complexity of the LMCZ. Indeed, it has already been shown that stepovers wider than 3-5
951 km often arrest seismic ruptures [Harris *et al.*, 1991; Wesnousky, 2006]. Alike, a rupture propagating westward
952 along the Pétion-Ville – Léogâne segment would probably stop at the Tapion pressure ridge or the Miragoâne
953 pull-apart. At the eastern edge of Tapion pressure ridge, the rupture may already encounter a first barrier, which
954 is the branching of the Trois-Baies Fault. If possible, the rupture may preferentially continue along the Troies-
955 Baies Fault. On the contrary, a rupture initiating along the Trois-Baies Fault may jump along the EPGF, as it was
956 the case with the Lamentin Fault during the 2010 Haiti earthquake [Saint Fleur *et al.*, 2015]. This has already
957 been observed for several earthquakes where a secondary branch activated the main fault, and vice versa. This is
958 related to the dynamic stress conditions at the branch in addition to the regional stress field and geometry [Saint
959 Fleur *et al.*, 2015; Etchebes, 2011; Bhat *et al.*, 2004; Kame *et al.*, 2003; Sowers *et al.*, 1994].

960

961 Overall, the geometry of the EPGF is simpler to the west where fault-branching is rare. One may
962 consider that the fault is more mature western Miragoâne. To the east, compressional deformation and fault-
963 branching begin to be salient with the Trois-Baies Fault. Then, the pattern continues with the Lamentin Fault
964 and the compression is more and more remarkable eastward. This pattern is strikingly highlighted by the GPS
965 vectors that are slower and more parallel to the EPGF to the west and oblique to the east with increasing rates
966 (Fig. 21a) [Calais *et al.*, 2016]. As the eastern part is immature, it is more complex and active than the western
967 part. More historical and instrumental earthquakes are recorded to the east [Bakun *et al.*, 2012; Rodriguez *et al.*,
968 2018]. The 2010 earthquake occurred in that complex, active zone. The active Lamentin thrust, as a branching
969 fault, broke first and the rupture jumped to the EPGF (Fig. 21b) [Saint Fleur *et al.*, 2015]. From a long-term
970 kinematics point of view, that complex rupture occurred near a triple junction (Fig. 21b inset), as in the case of
971 the Kaikoura, New Zealand 2016 earthquake [Shi *et al.*, 2019]. Regarding the eastward propagation of the EPGF,
972 the triple junction was first formed by the Troies Baies Fault, the Dumay Fault and the EPGF (trench-trench-

973 transform geometry). Then, the Lamentin Fault replaces the Trois Fault in the present-day triple junction. In this
974 regard, like other western secondary thrusts, the Trois Baies Fault is as a remnant of the triple junction
975 propagation but still active because its geometry is compatible with the regional stress field.

976

977

978 **5.3.1. Comparison with other strike-slip faults**

979 Comparing our first-order segmentation to the one of Etchebes (2011), we remark that we obtained less
980 segments in spite of the longer stretch of fault studied here (>250 km). Our segments are thus longer. This may
981 indicate that the EPGF is smoother and has a simpler geometry than the Fuyun Fault. This is consistent with the
982 maximum offset of the EPGF (40 ± 10 km) that is greater than that of the Fuyun Fault (20 km). For numerous
983 strike-slip faults, clear relationships have already been shown between their total offset and their trace
984 complexity [Stirling *et al.*, 1996; Wesnousky, 1988]. The relatively continuous trace of the EPGF (western part)
985 is due to its maturity [Ben-Zion and Sammis, 2003; de Joussineau and Aydin, 2010; Manighetti *et al.*, 2007]. In
986 contrast, the Fuyun Fault is roughly active since the Early Paleocene-Eocene [Besse *et al.*, 1984; Molnar and
987 Tapponnier, 1975; Tapponnier and Molnar, 1979], whereas the EPGF is thought to be of Pliocene age in Haiti
988 [Calais and Mercier de Lépinay, 1995; Calmus, 1983; Calmus and Vila, 1988; Heubeck and Mann, 1991a; b;
989 Mann *et al.*, 1984; Van Den Berghe, 1983]. Thus, fault maturity may be mostly related to the evolution and the
990 growth of the fault instead of its age. It is widely accepted that brittle faults start from a series of short en echelon
991 arrays and grow by linkage of the neighbouring segments [Cartwright *et al.*, 1995; de Joussineau and Aydin,
992 2010; Kim *et al.*, 2004; Myers and Aydin, 2004; Segall and Pollard, 1980]. The small discontinuities are thus
993 destructed by through-going planar surfaces to form more continuous and smoother fault segment geometry at
994 larger scale while new segments achieve a greater capability to interact with each other at a greater distance
995 [Aydin and Schultz, 1990; Scholz, 2002].

996

997 Analyzing the data sets of 22 strike-slip faults from Stirling *et al.* (1996), and the 10 strike-slip faults
998 from Klinger (2010), the EPGF seems to bear lots of structural resemblance with the North Anatolian Fault
999 (NAF). The latter has 12 asperities over 980 km, that is 0.12 asperities/km against ~0.16 for the EPGF. The NAF
1000 has a total minimum offset of 35 km in average [Barka, 1992; Barka and Gülen, 1989] against ~40 km for the
1001 EPGF along the southern peninsula of Haiti. The NAF initiated as a wide right-lateral shear zone in the Late

1002 Miocene time, whereas the main fault trace was formed during the Late Pliocene time [Barka, 1992; Sengör,
1003 1979; Sengör et al., 2005]. From these values a Neogene slip rate ranging from 5 to 10 mm/yr has been inferred
1004 [Barka, 1992; Hubert-Ferrari et al., 2002]. The Holocene slip rate seems to be higher ranging from 10 to 20
1005 mm/yr [Hubert-Ferrari et al., 2002; Kozaci et al., 2007; Polonia et al., 2004]. All these parameters suggest that
1006 the trace of the NAF is as smooth as that of the EPGF.

1007

1008 6. Conclusions

1009

1010 This work is a significant step for future studies of the Enriquillo-Plantain Garden Fault as it addresses
1011 several aspects. We have mapped the EPGF in detail using high-resolution imagery, geological data and field
1012 observations. This precise map can be used in the reconstruction perspective of Haiti.

1013 Analyzing several criteria like fault geometry, sedimentary record and accessibility, several favorable
1014 sites for paleoseismic studies have been identified. An exploratory trench has been excavated in Siline area, in
1015 the basin of Clonard. That trench showed evidence for one paleoearthquake, which is likely the 1770 event.
1016 However, this preliminary result needs to be strengthened by more direct radiocarbon dating and excavation
1017 across the main fault.

1018 The segmentation of the fault and seismic hazard implications have been discussed. We highlighted at
1019 least four clear segments by considering only kilometer-scale geometric complexities. More extensive
1020 paleoseismic studies may bring evidence for more segments as well as rupture scenarios along the Enriquillo-
1021 Plantain Garden Fault in Haiti.

1022 Finally, the geological offsets of ~40 km, ~15 km and ~8 km from west to east support the idea of an
1023 eastward propagation of the EPGF. In the easternmost part of the fault, east of Port-au-Prince, the fault zone
1024 rather portrays diffuse and compressional deformation, consistent with fault immaturity, the oblique convergence
1025 between Hispaniola and the Bahamas platform, and the indentation of the Beata Ridge in the area.

1026

1027 Acknowledgments

1028

1029 We thank INSU CT3 for funding and the Haitian Bureau of Mines and Energy, especially D. Anglade and O.
1030 Desliens for their assistance in the field. We also thank the National Center for Geospatial Information and B. E.
1031 Piard for providing the high-resolution aerial photographs. E. Calais kindly shared the GPS data with us and
1032 provided useful advice. We thank R. Lacassin, Y. Gaudemer, R. Gold, and C. Prentice for fruitful discussions.
1033 We are thankful to Paul Mann, and an anonymous reviewer for constructive comments to improve the quality of
1034 the paper.

1035

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