Forearc crustal faults as tsunami sources in the upper plate of the Lesser Antilles subduction zone. The Case study of the Morne Piton fault system.

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Abstract

In this study, alternatively to the megathrust, we identify upper plate normal faults orthogonal to the trench as a possible tsunami source along the Lesser Antilles subduction zone. The Morne Piton fault system, is such a trench-perpendicular upper crustal fault at the latitude of Guadeloupe. By the means of seismic reflection, high resolution bathymetry, remotely operated vehicle (ROV) imaging and dating, we reassess the slip rate of the Morne Piton fault since 7Ma, *i.e.* its inception, and quantify an average rate of 0.25 $mm \cdot yr^{-1}$ since ca. 1.2Ma dividing by two previous estimations and thus increasing the earthquake time recurrence and lowering the associated hazard. ROV dive revealed a metric scarp with striae at the toe of the Morne Piton fault system suggesting a recent fault rupture. We estimate a fault rupture area of ~ 450-675 km^2 and then a magnitude range for a maximum seismic event around M_w 6.5 ± 0.5 making this fault potentially tsunamigenic as the nearby Les Saintes Fault responsible for a tsunami following the 2004, M_w 6.3 earthquake. Consequently, we simulate a multi-segment tsunami model representative for a worst-case scenario as if the whole identified Morne Piton fault segments ruptured together. Our model provides clues for the potential impact of local tsunamis on the surrounding coastal area as well as for local bathymetric controls on tsunami propagation. We illustrate that (i) shallow water plateaus act as secondary sources and are responsible for a wrapping of the tsunami waves around the island of Marie-Galante, (ii) canyons indenting the shallow water plateau slope break are focusing and enhancing the wave height in front of the most touristic and populated town of the island, (iii) resonance phenomenon is observed within Les Saintes archipelago showing that the waves' frequency content is able to perturbate the sea-level during many hours after the seismic rupture.

Keywords: subduction zone, forearc, crustal fault, slip rate, tsunami hazard, Lesser Antilles, Guadeloupe Island

1. Introduction

Regions at the vicinity of active subduction zones are prone to seismic and related hazards, including tsunamis, exposing their inhabitants to multiple threats. Megathrust earthquakes represent the greatest threat with the highest seismic moments and consequently huge tsunamigenic potential (Satake and Tanioka, 1999). Earthquakes triggered on crustal faults in the overriding plate represent an additional hazard that needs to be quantified (Bilek, 2010). In order to assess the hazards and mitigate the risks associated with these crustal faults, it is essential to estimate their slip rates.

On land, slip rates on active faults are determined from paleo-seismic trenches (McCalpin, 1996), high resolution geophysical investigation (Wallace, 1981; Zhang et al., 2014), satellite imagery (Tronin, 2009), InSAR (Biggs and Wright, 2020), geodetic measurement (GNSS: Symithe et al., 2013) as well as seismicity which account for the present-day strain accumulation of the crust. Offshore, slip rate estimates are provided by the means of underwater geodesy (*i.e.* acoustic geodesy: Kido et al., 2006; Petersen et al., 2019; Fujita et al., 2006) or fiber optic monitoring (Hirata et al., 2002; Gutscher et al., 2019). The recurrence time of events may be estimated by the study of turbidite deposits cores (e.g., Cascades: Goldfinger et al., 2012; Taiwan: Lehu et al., 2016; Antilles: Seibert et al., 2016; New Zealand: Lewis et al., 1980), high resolution marine seismic and multibeam echo-sounder data (e.g., Escartin et al., 2016, 2018), and submarine dives survey (e.g., Geli et al., 2011). However, constraining hazard models in areas undergoing slow strain rates remains challenging as the earthquakes recurrence time overcomes the historical period.

Indeed, geodetic measurements require decades-long time series as the resolution of the method is not accurate enough and erosion or high sedimentation rates may have erased or covered, respectively, the active fault scarps making it difficult to identify active faults segments. Therefore, datasets based on the last ten to hundred years of record along tectonic systems undergoing slow strain rates may not be representative of the bulk strain and may be at the origin of biased estimations of slip rate along these faults.

The Lesser Antilles (Eastern Caribbean) records slow deformation rates as the north and south American tectonic plates slowly subduct under the Caribbean plate ($20 mm \cdot yr^{-1}$ - Figure 1). Extensional tectonics and normal faulting affect the forearc (Feuillet et al., 2002, De Min et al., 2015, Boucard et al., 2021) but available historical data do not report tsunami events related to forearc fault rupture. However, the Les Saintes M_w 6.3 earthquake of December 2004 ruptured the Roseau normal fault (Feuillet et al., 2011a, Bazin et al., 2010). The earthquake reached an intensity up to VIII in the Guadeloupe Archipelago (Figure 1), being felt by most of its ~400,000 inhabitants, and was responsible for one casualty. This earthquake triggered a tsunami with up to 2 m high waves at the coast and a maximum measured run-up distance of 42 m in Les Saintes (Zahibo et al. 2005; Le Friant et al. 2008; Cordrie et al. 2020). Prior to this event, this fault was unmapped and therefore not identified as an active fault (Feuillet et al., 2002; Terrier and Combes, 2002). Forearc normal faults, similar to Les Saintes fault system, may pose a threat to the 4 million inhabitants of the Lesser Antilles that are living on volcanic arc islands facing the subduction trench to the east and literally sitting over the subduction interface.

The present study focuses on the Morne Piton fault system, perpendicular to the subduction trench, which is one of the most prominent onshore-offshore fault systems which cuts the Guadeloupe Archipelago arc and forearc islands (Figure 1). Regarding the seismic and tsunami hazards related to this fault system and the vulnerability of the coastal population and infrastructures of the archipelago, the objectives are to (1) estimate the fault slip rate (2) determine the geometry of the fault segments, and (3) model the associated tsunami hazard, since such a joint approach has been lacking so far. In the following study, the fault geometry is refined in order to provide an up-to-date map of the fault segments thanks to high-resolution (HR) bathymetric data. Then, we integrate its long-term slip rate over the last *ca.* 7 My, *i.e.* from fault initiation to present-day, by the mean of HR seismic reflection lines and available biostratigraphic and isotopic dates.

Secondly, Remotely Operated Vehicles (ROV) explorations of seafloor rupture allowed us to measure the height of the fault scarp and to determine the fault kinematics from striations observed along a recent co-seismic scarp. Because the overall geometry of this fault system is comparable to the Les Saintes fault system in terms of length, seafloor scarp and dip, we postulate that a rupture along the Morne Piton fault may trigger a local tsunami close to the coasts of the Guadeloupe Archipelago. Therefore we study the seismogenic and tsunamigenic potential of the Morne Piton fault system providing an overview of what could happen in terms of tsunami generation if all the segments of the Morne Piton fault ruptured simultaneously, *i.e.* a plausible worst-case scenario. This scenario allows to identify and discuss the local bathymetric controls on the propagation of the resulting tsunami wave and the consequences (*e.g.,* amplifications and interferences) in near-shore areas of the neighboring populated islands. We do not assess coastal inundation scenarios, as our scenario can't be refined by observational rupture data accurate enough to realize such a specific hazard study. Finally, we conclude on the importance of forearc crustal faults as potential major tsunami sources in subduction zones.

2. Geological settings

Oceanic lithosphere of the North and South American plates is slowly subducting beneath the Caribbean plate at a convergence rate of ~20mm·yr^{-1} (Figure 1, DeMets et al., 2000; Philippon and Corti, 2016). The convex trench geometry results in along strike variations of obliquity, increasing northward from Guadeloupe. Along the arc, oblique subduction is accommodated by trench-parallel left-lateral strike slip faults such as the Harvers-Montserrat-Bouillante / Les Saintes corridor (located within the volcanic arc), the Bunce fault (located along the crustal buttress), and a series of trench-perpendicular grabens forming a sinistral horsetail (Feuillet et al., 2002; Feuillet et al., 2010; Ten Brink et al., 2004; Laurencin et al., 2019; Boucard et al., 2021) (Figure 1, Figure 2A).

In the central Lesser Antilles, the Marie-Galante Basin (Guadeloupe Archipelago), is located at the southern end of the aforementioned regional horsetail system and is described as a conjugated normal fault system defining a trench perpendicular graben (Figure 2 A; Feuillet et al., 2001, 2011). This graben affects sediment deposits comprising three regional mega-sequences: an Eocene(?) - Early Miocene MG-MS1 sequence, a mid-

Miocene – late Tortonian / early Messinian MG-MS2 sequence, and a Messinan to present MG-MS3 sequence (Bouysse et Mascle, 1994; De Min, 2014; De Min et al., 2015; Cornée et al., 2023). It shapes the Marie-Galante Basin (up to 1200 m water depth) and surroundings, Grande-Terre and Marie-Galante Islands, respectively (Figure 1). The northern boundary of the Marie-Galante Basin is the east trending, south dipping Gosier fault that runs primarily onshore along the southern coast of Grande-Terre (Garrabé and Andreieff, 1988; Figure 2A). The southern boundary of the basin consists of the N100° trending, ~50 km-long, north dipping Morne Piton fault, which crosscuts the northern edge of Marie-Galante Island (Bouysse et al., 1993) and extends offshore on both sides of the island (Feuillet et al., 2002, 2004).

The Morne Piton fault system consists of five main 5-15 km-long segments trending N90°E ± 30° separated by N140°E shorter right-lateral relays (Figure 2). The fault scarp is exposed at Anse Piton, eastern coast of Marie-Galante, and shows dip-slip striations (Feuillet et al., 2002). Onshore Marie-Galante, the fault offsets the Pliocene to middle Pleistocene platform by ~60 m. It also crosscuts a series of 3 uplifted late-mid to late Pleistocene terraces along the eastern side of the island. Feuillet et al. (2004) calculated a 5 km dislocation depth and a 70 to 80°N fault dip to model the observed flexure of the footwall. Considering that the Marie-Galante Plateau is a flat abandoned 330 Ka old marine terrace, these authors estimate the average slip rate of the Morne Piton at about 0.5±0.2 $mm \cdot yr^{-1}$ since 330 Ka. Regarding the uplifted terraces they estimated a maximum earthquake moment magnitude (M_w) ranging from 5.8 up to 6.5 with a 400-1000 to 1400-3300 years of recurrence time, respectively (250 km^2 of estimated ruptured area). Moreover, it was later demonstrated that this plateau emerged between 1.77 and 1.07 Ma (magnetostratigraphic Chron 1R2r: Cornée et al., 2012; Münch et al., 2014; De Min et al., 2015; Léticée et al., 2019; Cornée et al., 2021). Note that considering an older age for the Plateau emergence would drastically lower the slip rate estimate and increase the recurrence time calculated by Feuillet et al. (2001).

3. Historical seismicity

Upper plate seismicity in the Marie-Galante Basin provided by (i) CDSA Seismic database (Antilles Seismological Data Center - Bengoubou-Valerius et al. (2008); Massin et al.(2021)), (ii) IRIS database (IRIS https://www.isc.ac.uk (Figure 2B) and (iii) the deployment of Ocean Bottom Seismometers (OBS) (Ruiz et al., 2013; Bie et al., 2019) shows a widely distributed pattern of moderate magnitude earthquakes (M_w≤5.3), with the exception of the 2004 seismic cluster in Les Saintes. Wide Angle Seismic (WAS) profiles together with earthquakes data indicate a seismogenic crustal thickness limited to the first 15-20 km west of Marie-Galante suggesting a brittle-ductile transition at this depth (Kopp et al., 2011; Ruiz et al., 2013; Gonzalez et al., 2017; Padron et al., 2021). Among the very few focal mechanisms available in the Marie-Galante Basin (Gonzalez et al., 2017), the 25 February 2014 M_w 3.8 earthquake occurred beneath the southern Grande-Terre platform and shows pure normal motion along sub-E-W trending nodal planes (Gonzalez et al. 2017, Event n°9, hypocentral location accuracy of ca. 5 km; Figure 2C). The location, depth and nodal plane characteristics (57° dip and N102°E) of the earthquake indicate that the event may correspond to a rupture along the Gosier fault system, which is the only major fault system in the vicinity of the hypocentral location able to trigger such magnitude earthquake. Feuillet (2000) provided more than 20 focal mechanisms for earthquakes showing local magnitude 2<Ml<3.7 and one Ms=5.6 earthquake, located in and around the Marie-Galante Graben. All focal mechanisms show nearly pure normal motion along sub-E-W trending nodal planes, consistent with kinematics indicators observed along the Gosier and Morne Piton faults. This tectonic pattern is confirmed by GNSS velocities which indicate that a small trench-parallel extension is accommodated in the upper plate forearc (van Rijsingen et al., 2021).

Two historical earthquakes are reported along these two fault systems: (i) the 16 May 1851 earthquake with a maximum intensity of VII recorded in the southeastern part of Basse-Terre, is attributed to the Morne Piton fault with an estimated moment magnitude M_w=6.0 (Feuillet et al., 2011; and (ii) the 29 April 1897 earthquake with a maximum intensity of VIII recorded in the Pointe-à-Pitre area being either attributed to the Gosier Fault system with an estimated moment magnitude M_w=5.5, or to the Montserrat fault zone, with an estimated moment magnitude M_w=6.5 (Bernard and Lambert, 1988; Feuillet et al., 2011b). Overall, at the latitude of Guadeloupe, regional earthquake data suggests that normal fault systems are active with an ability to generate earthquakes of moment magnitude M_w 6 and above. This magnitude range is potentially able to trigger tsunami according to tsunami catalogues (as explained, for example, in Roger et al., 2019).

4. Historical tsunami

Southwest of the Marie-Galante Basin, the 2004 M_w 6.3 earthquake (Bazin et al., 2010; Feuillet et al., 2011) showed that upper plate crustal faults can generate strong earthquakes and tsunamis. The main shock occurred along the NNW-SSE trending, *ca.* 40 km-long arc-parallel Les Saintes Fault System (Feuillet et al., 2011; Leclerc et al., 2016). The recurrence of such a rupture is estimated to be a few hundred years or more (Escartin et al., 2016; Escartin et al., 2018; Feuillet et al., 2011; Le Friant et al., 2008). Focal mechanisms of the main shock as well as five aftershocks provided an overall pure normal motion along NNW-SSE nodal planes (Figure 2D). Source models from Salichon et al. (2009), Bazin et al. (2010), and Feuillet et al. (2011), well constrained by the long duration of the aftershock sequence, proposed a main source localized along the N135°E trending, 50°E dipping Roseau Fault (westernmost fault of Les Saintes fault system) with a 30 km-long and 21 km-downdip width fault plane. Aftershock seismicity reactivated several nearby conjugate faults with a maximum seismic depth at *ca.* 15 km. The main rupture occurred at two asperities located 8 km below the surface with a maximum slip of 1.8 m, and propagated to the surface triggering a coseismic offset of the seafloor of 0.3-0.6 m along a *ca*. 10 km-long segment. Escartin et al. (2016) investigated the fault scarp by the mean of HR bathymetry highlighting a 3 km-long, up to 0.9 m-high scarp, but concluded that part of the observed slip may be post-seismic. The Les Saintes earthquake generated up to 2 m-high tsunami waves at the coast and a maximum horizontal run-up of 42 m in some bays of Les Saintes (Zahibo et al., 2005; Le Friant et al., 2008; Cordrie et al., 2020). However, tsunami models using fault parameters based on seismological data resulted in an underestimation of the tsunami wave amplitude and run-up (Le Friant et al., 2008). Cordrie et al. (2020) consider that their best fit models require greater slip on the fault plane and a greater magnitude for the earthquake than those given by the seismological data in order to accurately reproduce the observed tsunami, suggesting that the observed scarp is the surface expression of co-seismic slip (source parameters: M_w =6.4-6.5 – fault plane 15x15 km – Strike N325°E – Dip 55°E – rake ca.90° - slip=2.5-3.5 m).

Over the last ~500 years of historical written archives in the Lesser Antilles, a few dozen confirmed tsunamis from different origins (local, regional or far-field sources including earthquakes, landslides, volcanic eruptions or combinations of them) have been reported. Starting with the 16 April 1690 Ms~8.0 Barbados earthquake (which presumably triggered the first reported tsunami in the Lesser Antilles), it includes the widely studied 1 November 1755 Lisbon transoceanic tsunami (*e.g.,* Gutscher et al., 2006, Accary and Roger, 2010; Roger et al., 2010, 2011, Martinez-Loriente et al., 2021) and the 18 November 1867 Virgin Islands tsunami (*e.g*., Zahibo et al., 2003, 2005; Barkan and ten Brink, 2004). Landslide sources and/or pyroclastic flows, are also known for their tsunamigenic potential. There are more and more studies led to assess the hazard associated to these "silent tsunamis" (e.g., Roger et al., 2024). In the Caribbean Region, a few tsunamis triggered by landslides and/or pyroclastic flows have been reported in catalogues of events (e.g., O'loughlin and Lander, 2003; Accary and Roger, 2010), bathymetric surveys helped to identified large submarine landslide scars and deposits (e.g., Deplus et al., 2001; Le Friant et al., 2009, 2019) and a few studies have highlighted the capacity of these landslides to trigger large tsunamis (e.g., Smith and Shepherd, 1996; Teeuw et al., 2009; Leslie and Mann, 2016).

On the basis of an extensive literature review, including cross-checking of information, we conclude that only four tsunamis reported in Guadeloupe are likely of upper crustal seismic origin (Mallet, 1853, 1854, 1855; Lander, 1997; Zahibo and Pelinovsky, 2001; Lander et al., 2003; O'Loughlin and Lander, 2003; Zahibo et al., 2003; Accary and Roger, 2010; Nikolkina et al., 2010; Roger et al., 2013; online databases: NGDC/WDS, 2023; TL/ICMMG, 2023). These tsunamis have been observed or recorded following earthquakes occurring on regional faults (indicated magnitude and epicenter coordinates are from the USGS online earthquakes catalogue: https://www.usgs.gov/programs/earthquake-hazards): the M_w~8.0-8.5 earthquake on 8 February 1843 (NE of Guadeloupe, 16.73°N, 61.17°W) and the M_w 7.2 earthquake on 25 December 1969 (SE of Guadeloupe, 15.648°N, 59.694°W) are arguably attributed either to a rupture along the megathrust or to upper plate faulting ; the M_w 6.5 earthquake on 16 March 1985 (along the Harvers-Montserrat-Bouillante fault system between Montserrat and Nevis, north of Basse-Terre, 17.013°N, 62.448°W); and the M_w 6.3 earthquake on 21 November 2004 (Along the Les Saintes fault system, south of Basse-Terre, 15.679°N, 61.706°W)(Figure1). The largest earthquakes and tsunamis produced at subduction zones are expected to originate from rupture at the plate interface megathrust. However, historical records in the Lesser Antilles reveal that neither the

1843 M_w 7.5-8.5 nor the 1839 M_w 7.5-8.5 largest known earthquakes, although destructive, have been followed by large tsunamis. However, Roger et al. (2013) showed that the simulation of a Mw 8.5 1843-like megathrust earthquake would have produced wave amplitudes of 5 m and more along the exposed coasts of Guadeloupe, which was not reported in coeval documents. Feuillet et al. (2011) explain these two major earthquakes by the great depth of the rupture along the megathrust that led to little seafloor deformation. However, numerical simulation of worst-case scenarios for these two ruptures along the megathrust evidence the possibilities of tsunami amplitudes up to 10 meters and above in some embayment (Roger et al., 2013; Colon-Useche et al., 2023. As magnitude of crustal earthquakes is constrained by fault length, events occurring along such crustal fault show a much smaller magnitude than megathrust earthquakes consequently may form smaller seafloor offsets. Thus, most crustal earthquakes able to trigger a tsunami do not produce significant sea surface deformation (only a few centimeters amplitude in most cases) compared to subduction interface earthquakes. Associated tsunamis are typically only visible on pressure gauge records (coastal gauges or DART systems) after processing the data (*e.g.*, de-tiding, high-frequencies filtering, etc.).

3. Material and method

3.1 Seismic lines

We present eight multichannel seismic (MCS) lines acquired during five oceanographic campaigns (location on Figure 3A). These include high-resolution sparker source seismic data from KaShallow 1 (Lebrun et al., 2009) and GEOBERYX03 oceanographic campaigns (Thinon and Bitri, 2003; Thinon et al., 2004 and 2010), mid resolution GI airgun arrays seismic data from KaShallow 2 (Lebrun et al., 2009) and Aguadomar (Deplus et al., 1999) cruises, and deep penetrating MCS data from the Sismantilles 1 seismic experiment (Hirn, 2001) (Table 1).

Sismantilles 1 seismic data have been processed using CGG-Veritas Geovecteur® software on board the R/V Nadir (Hirn et al., 2001). Processing includes band pass filtering, internal and external mute, one step velocity analysis, NMO correction, stack, predictive deconvolution and post-stack constant water-velocity time migration. The

KaShallow 1 and 2, Aguadomar and Geoberyx have been processed with Seismic Unix software (Cohen and Stokwell, Center for Wave Phenomena, Colorado School of Mines). The seismic processing includes band pass filtering, sea waves and spherical divergence corrections, constant velocity or simple velocity gradient NMO correction and stack, and constant water-velocity time migration. The reflection seismic lines are in millisecond two-way-travel-time (mstwt). The velocities of the Wide Angle Seismic refraction (WAS) profiles are in second two-way-travel-time (stwt).

3.2 Bathymetry

High-resolution bathymetric data were acquired during the KaShallow2 oceanographic campaign (Lebrun, 2009) using a Simrad EM300 multibeam echosounder. We merged this data with Aguadomar (Deplus, 1999) and Sismantilles 2 (Laigle et al., 2007) cruises Simrad EM12 Dual multibeam echosounder data available for the Marie-Galante Basin. Vertical accuracy for these echosounders is plurimetric for typical water depth found in the Marie-Galante Basin (<2000 m below mean sea-level, noted bsl hereafter). Near-shore (0-200 m bsl) and onshore, very high-resolution bathymetric and topographic data comes from the Litto3D database - (https://www.geoportail.gouv.fr/donnees/litto3d last accessed on September 2024- that includes airborne lidar survey and KaShallow-3 multibeam data acquired with a RESON Seabat 8101 multibeam echosounder). The vertical accuracy for this second dataset is better than one meter. We used the Caraibes software (ifremer) to process the data and to produce a 25 m grid spacing Digital Elevation Model of the Marie-Galante Basin and surrounding islands. Maps are produced using the open-access QGIS software (https://www.qgis.org)

3.3 Depth and time calibration of main geological boundaries.

In order to measure offsets of unconformities on time-migrated seismic lines we need to constrain the seismic velocities within the sediments. We used velocities from the WAS profile (Kopp et al., 2011) in the south of Marie-Galante, that trends parallel to the MCS line Agua116 (Cornée et al., 2023). The WAS velocities in the *ca*. 0.4 stwt (second two way time) thick upper unit (MG-MS3 – Cornée et al., 2023) ranges between 2 and 2.5 km/s. The 3.25 km/s isochrones mimic the base of unit MG-MS2 and the 4.5 km/s isochrones follow the acoustic basement below MG-MS1. Moreover, KaShallow 2 cruise

MCS data (Table 1) acquired with a 600 m long streamer allows us to determine the Normal Move Out velocities down to a depth of *ca*. 0.75 stwt in well-layered units such as shown on the seismic lines (Figure 3). Once converted into interval velocities using the "Hewitt Dix formula", we determine velocities in the upper unit from 1500 to 2750 m/s (Dix, 1955). Therefore, we use 1500 m/s in the water and 2000 m/s and 2500 m/s in the sediments to estimate (and bound) the depth of unconformities observed on time-migrated seismic lines (Table 2).

Offshore, several first order unconformities and sedimentary units were accurately dated using bio-stratigraphy analysis or radiochronology (Bouysse and Mascle, 1994; KaShallow Reasearch Program results: Münch et al., 2013; De Min, 2014; De Min et al., 2015; Cornée et al., 2023). The deepest dated unconformity, MG-SB2, which corresponds to the top of the MG-MS1 sequence, occurs on seismic lines east of Marie-Galante (thick orange line on Figure 5, lines AGUA97 – K09-09 – K09_45 - Sis7C). Along the seismic line Agua 97, the F8 fault scarp has been sampled at 514 m bsl just beneath the unconformity (KaShallow Cruise ROV dive, Figure 6). The samples, BMG2 and 4, yielded a Late Burdigalian/earliest Langhian age (Cornée et al., 2023). Thus, we propose 16 Ma ± 1Ma for the age of MG-SB2**.** Above, another regional unconformity MG-SB3 (Cornée et al., 2023) is identified east of Marie-Galante. It corresponds to the top of MG-MS2 sequence (thick purple line on the Figure 5 lines Agua97 – K09-09-08 – K09-45-44 – Sis7C). The age for this surface is bracketed between the overlaying Late Messinian GT carbonate platform (zones N18, 5.8-5.33 Ma –Cornée et al., 2023) and the underlying sedimentary unit dated Late Tortonian 8.57 ± 0.43 Ma (Ar-Ar, Münch et al., 2014). We thus consider 7 ± 1.5 Ma for the age of MG-SB3. West of Marie-Galante, the angular unconformity on line K09-90 North-West of Marie-Galante may corresponds to MG-SB3 (Figure 4). However, this reflector is too deep to be followed across the whole fault system. Within the uppermost sequence, MG-MS3, a remarkable unit boundary corresponding to a second order unconformity, can be easily correlated throughout the basin and onshore (red thick line on Figure 4). This unit boundary is Middle-Late Piazencian offshore and correlates onshore with the 3–2.9 Ma tectonically-induced erosional unconformity SB1 (see above; Cornée et al., 2023). Along the seismic line Ber03-30-31, the fault scarp immediately north of F3 and F4 has been sampled at 283 m bsl (KaShallow Cruise ROV dive, Figure 4). Samples, BC1 and BC2, yielded ages of 1.33 ± 0.23 Ma and 1.15. ± 0.12 Ma, respectively (Ar^{40}/Ar^{39} ages on plagioclases, Münch et al., 2013; 2014). These samples correspond to a prominent seismic reflector within the upper unit of MG-MS3 sequence that can also be easily correlated through all the seismic lines west of Marie-Galante. We thus retain an average age of 1.29 ± 0.26 Ma for this seismic reflector (green line on Figure 4).

3.4 Tsunami modeling

In order to test the tsunamigenic potential of the fault system proposed herein, a rupture scenarios has been elaborated and is presented hereafter.

Numerical simulations of tsunami generation and propagation were carried out using COMCOT software (Cornell Multi-grid Coupled Tsunami: Liu et al., 1998; Wang, 2008; Wang and Power, 2011). COMCOT is widely used by the research community and constantly tested notably through various real tsunami cases (e.g. Prasetya et al., 2011; Gusman et al., 2019; Paris et al., 2021; Gusman et al., 2022; Roger et al., 2023). COMCOT uses a modified staggered finite-difference scheme to solve linear and non-linear shallow water equations in either spherical or Cartesian coordinate systems throughout a set of nested grids allowing refinement of the bathymetric resolution in coastal areas. A two-way nested grid configuration is implemented in the model to balance computational efficiency and numerical accuracy (Wang 2008; Wang and Power 2011).

For this study, nesting has been used with two grid levels: the first grid is a 0.5 arcmin (~900 m) resolution grid of the Lesser Antilles (extent: 295°E, 302°E, 12°S, 18°S) built from the global dataset GEBCO 2021 (GEBCO Compilation Group, 2021); the second grid is a 3.75 arcsec (~115 m) spatial resolution grid focusing on the Guadeloupe Archipelago and Dominica Island, including the location of the investigated Morne Piton fault system as shown on Figure 1 (extent: 297.92°E, 300.22°E, 14.94°N, 16.717°N). This second grid has been built from different datasets including the aforementioned bathymetric data (§ 3.2). The highest resolution and the most recent data have been kept first. Data gaps have been filled in with data from GEBCO (GEBCO Compilation Group, 2021) for offshore regions, and SRTM version 3.0 Global 1 arcsec data (NASA SRTM, 2013) for onshore regions. Continuity of the different datasets has been ensured using kriging interpolation, which has proven to be one of the best methods to produce a well-defined DEM, especially for smooth transitions between different resolution areas (e.g. Bernardes et al., 2006; Arun, 2013; Ajvazi and Czimber, 2019). Note that Dominica was included in the second grid in order to look at potential effects which could occur between the different islands and also to assess the potential tsunami threat resulting from the Morne Piton scenario on this neighboring island.

The initial sea-bottom displacement is calculated by COMCOT considering an instantaneous rupture of the fault using the surface deformation model of Okada (1985), and transmission of the deformation to the water column above is considered instantaneous. Calculations of wave propagation have been done at mean sea level (MSL) assuming a constant Manning's roughness coefficient of 0.011 for the seabed friction (Wang et al., 2017). A higher friction coefficient leads to more energy dissipation of tsunami waves, especially in shallow waters, slowing down their speed and reducing their amplitude and impact (e.g. Dao and Tkalich, 2007). Considering the limited extent of the interest zone (~250km x 200km), the rupture parameters (leading to a small coseismic rupture) and the objective to look at potential localized effect as inter-islands resonance, tsunami waves propagation time was set to 10 hours.

4. RESULTS

4.1 The Morne Piton fault system

The HR bathymetric data presented here above (section 3.2) allows to refine the structural pattern of the Morne Piton fault system, especially offshore (Figures 3A - B). The fault system splays eastward from the N120-N135°E trending Eastern Les Saintes fault system located east-south-east of Basse-Terre to the N110-115°E trending Petite-Terre fault system south-south-east of Grande-Terre (Figures 2 and 3). Thus, the fault zone spreads over a 5-8 km wide and 50 km long zone with an average N100°E trend.

Morpho-bathymetric analysis allows us to identify surficial segments of the faults that are reaching out the seafloor. The main fault scarp of the Morne Piton fault system is the southernmost one, along which 11 fault segments of 1-10 km length can be identified (Figure 3B). From west to east, the F1 segment trends N110° and then the fault steps left along the N75°E trending F2 segment. A little farther east, the fault cuts the northern Colombie Bank and the eastern Marie-Galante platform along closely spaced N90E trending left or right stepping segments F3, F3', F4 and F5. Across the island, the F6 segment is a N130°N trending, 6 km long right step relay linking the F5 segment to the F7 N90°E trending one. Further east offshore Marie-Galante, two N80°E trending fault segments, F8, 9, 10 and 11, arrange as overlap right steps. There, the fault scarp vanishes in just a few kilometers. To the east of the line Sis7C, neither the sediments nor the basement are affected by the north-dipping Morne Piton fault system (Figure 5). In contrast, the seismic line Sis7C shows that the basement is southwardly downthrown by the Petite-Terre fault system, along south dipping active and sealed faults to the north and south of the Morne Piton fault system, respectively (Figure 5 location on Figure 3A) . West of Marie-Galante, the Morne Piton fault system widens as closely spaced fault splays trending N95°E to N100°E link the main fault scarp (F2, F3 and F4) to the antithetic Goyave Fault system or die westward (Figure 4 location on Figure 3A). Eastward of the F6 segment, some synthetic and a few antithetic faults splay northeastward and link with the N110-115°E Petite-Terre Fault system.

The mean fault-scarp height west of Marie-Galante Island is ca. 100 m (Figure 3C). Across Marie-Galante Island, the mean fault-scarp height reaches 200 m and controls the staircase morphology of the island. East of the island, the Marie-Galante Canyon carved the sedimentary units, clearing some of the fault planes increasing their apparent scarps heights up to 400 m. To the east, the canyon meanders and cuts through the eastern tip of the Morne Piton fault system (Figure 3 B and C). West of the island, at the vicinity of the volcanic island of Basse-Terre, either recent deposits or the uppermost sedimentary units seal most of the faults. These observations seem to indicate that the sedimentary rate west of Marie-Galante and the erosional rate east of Marie-Galante (in the canyon) exceed the vertical slip rate of the fault.

4.2 Vertical slip rate estimates along the Morne Piton fault system

To assess the average vertical slip rate along the Morne Piton fault system, we estimated the fault offset of key dated reflectors across the entire length of the fault system (Figures 4 and 5 – Table 2). West of Marie-Galante, the main offset of the 1.29 ± 0.26 Ma seismic reflector (green on Figure 4) increases from west to east (i.e., from the westernmost extremity of the fault toward its center). Close to the eastern shore of Basse-Terre Island (Figure 4, profile K09-96, and Table 1), the 1.29 Ma reflector is downthrown by 110-115 m. Eastward, the reflector offset increases up to 230-257 m (Figure 4, line K09-90, K08_24 and Table 2) and reaches a maximum of 300-322 m (Figure 4, lines ber03_30-31, K08-59). Accordingly, the number of sealed structures across the fault system decreases eastward (Figures 3 and 4). Thus, West of Marie-Galante, the Morne

Piton fault system accommodates a vertical slip rate increasing eastward up to 0.25 ± 0.08 mm·yr^{-1} over the last 1.29 ± 0.26 Ma (Table 2).

The 2.95 ± 0.05 Ma Unit Boundary (red line on figure 4) can only be correlated across the fault system along the K08-59 seismic line. Growth strata are observed in the deposits above the 2.95 Ma unit boundary (Figure 4, gray shadow on seismic lines), attesting for syn-sedimentary fault activity. This unit boundary offset reaches 550-620 m, leading to 0.20 ± 0.02 mm·yr^{-1} average vertical slip rate over the last 2.95 ± 0.05 Ma, i.e ca. 0,16 mm·yr^{-1} for the period 2,95 - 1,29Ma (Table 2). West of Marie-Galante Island, deeper reflectors cannot be identified and correlated across the fault system because of the limited seismic penetration.

However, east of Marie-Galante, the MG-SB2 sequence boundary dated to 16 ± 1Ma (orange in Figure 5 – see also Figure 5 in Cornée et al., 2023) is the only reflector that can be correlated on both sides of the fault system. In the hanging wall of the fault, the younger MG-SB3 7 ± 1.5Ma boundary (Purple in Figure 5) as well as a large part of the fault scarp are eroded by the Marie-Galante Canyon. The seismic line K09-09 (Figure 5) shows that the MG-SB3 unconformity records the Morne Piton fault inception: in the footwall of the fault, the stratigraphy of MG-MS2 sequence (comprised between MG-SB2 and MG-SB3) shows conformal deposits flexed upward while approaching the fault, whereas MG-MS3 deposits onlap onto MG-SB3 and present clear growth strata. We thus propose that the 16 ± 1 Ma sequence boundary is pre-tectonic and is tilted by the fault since its inception 7 ± 1.5 Ma ago (the age of MG-SB3; Figure 5, profile K09-45-45). Along the fault system East of Marie-Galante we calculated the strain rate from MG-SB2 offset since 7±1Ma. From east to west the slip rate ranges from 0.067±0.03 mm·yr^{-1} at the K09_44-45 seismic line, to 0.071±0.02 mm·yr^{-1} along the K09_08-09 line. Seismic line Agua97 (Figure 5) presents the greatest offset of MG-SB2. However, this seismic line does not cross the southernmost F7 segment (the water depth is too shallow for ship navigation in the footwall compartment). We estimated the depth of MG-SB2 in the footwall compartment from the closest seismic line available that crosses the fault located 1 km east of the Agua97 line. We obtain an offset of 830-860 m leading to a maximum average vertical slip rate of 0.12 ± 0.03 mm·yr^{-1} since 7 ± 1.5 Ma, i.e ca. 0,07 mm·yr^{-1} for the period 7 - 2,95Ma (Table 2). Consequently, we propose that i) the vertical slip rate accommodated by the Morne Piton fault system increases progressively from each extremity of the fault toward its center, and ii) that the Morne Piton fault system is characterized by an increasing slip rate from *ca.* 0.07 from its inception (*i.e.* 7 Ma ago) to *ca.* 0.25 mm·yr^{-1} since 1.29 Ma (Table 2).

4.3 Earthquakes parameters of Tsunami modeling

ROV dive along the ber03_30-32 seismic line allowed observation of one of the main morphologic scarps of the Morne Piton fault system across the F2 and F3' segments (Figures 3 and 6). Across the upper plateau, between F2 and F3', we observed several N90°E trending fractures parallel to the fault segments (Figure 6A). While descending across the F3' scarp, the slope progressively steepens up from 45° at 157 m bsl to more than 80° at 280 m bsl just a few meters above the toe of the scarp (Figure 6A, B and C). This morphology suggests a ca. 128 m-high cumulative scarp for the F3' segment at that location. The very last meter of the fault scarp above the toe of the slope presents a 100 cm-high polished vertical surface, partly altered, showing dip-slip striations indicating pure normal motion along this fault segment (Figure 5D E and F). This exposed and partly altered fault slip plane breaches the sea floor at high angle. Such a polished striated fault scarp morphology is similar, although more altered, to the co-seismic fault scarp observed at the toe of the Roseau Fault plane, after the Les Saintes earthquake (Escartin et al., 2016). We conclude that this observation of the Morne Piton polished striated scarp may correspond to one of the last co-seismic scarps formed during a major earthquake (including possible post-seismic slip motion) along this fault. Alteration of the fault slip plane suggests that the slip event occurred tens to several hundred years ago, *i.e.*, this fault slip plane may correspond to a pre-instrumental earthquake (see discussion). From this observed scarp we obtained a ratio of last event scarp over total scarp height (proxy for the cumulative slip as determined on Figure 3) of ~2,6%. With this ratio we calculated an average scarp of ~75 cm along the whole length of the fault and a maximum scarp of 2 m. Such an average scarp value corresponds to the surface expression of a magnitude M_w ~6.7 earthquake using the criteria of Wells and Coppersmith (1994), or even 7 according to Thingbaijam et al. (2017). The same studies also provide a calculated maximum displacement of ~2m, consistent with the maximum observation along the scarp. Moreover, the 45 km total length of the Morne Piton fault system measured from HR bathymetry, together with the width of the fault given by the 10 to 15 km thick seismogenic crust, lead to a rupture area ranging between ~ 450 to 675 km^2 that would generate a magnitude M_w ~6.7 ± 0.1 earthquake corroborating the afore range of magnitude deduced with other observations (e.g., M_w ranging between 6.6 and 6.8 according Wells and Coppersmith, (1994) and Leonard (2010) and around 7 according to Thingbaijam et al., 2017). The rupture parameters for the different identified segments of the fault shown on figure 3B are provided in Table 3. Geographic location of the center of top of the fault plane and azimuth are provided based on our structural analysis (section 4.1 and 4.3). Neither seismic lines which illustrate only a few hundred of meters nor in-depth earthquake distributions (which is not enough resolved) allow to estimate the dip of the Morne Piton fault system. Thus, after considering the influence of dip on surface deformation which turns out to be negligible, we choose a mean dip of 75° for the fault segments after Feuillet et al. (2004). The shape of the rupture area (along-strike length x downdip width) and a slip of 1.89 m (maximum displacement estimated from scarp heigh measurement) is implemented for each segment in order to fit with a total fault surface of ca. 500 km^2, corresponding to a magnitude M_w 6.5 earthquake. Finally, conformably to pure dip slip striations observed along the F3' segment (section 4.3) and the F7 segment (Feuillet et al. 2002), we apply a rake of 90°, corresponding to a pure normal faulting mechanism as observed by Feuillet et al. (2004) at Anse Piton and by the ROV picture of the present study. Fault segments F8 and F9 are not straight, for the purpose of modeling they have been divided into F8-F9 and F10-F11 (Figure 7). Hereafter, we use these parameters to evaluate the potential tsunami hazard from the Morne Piton fault for a worst-case plausible scenario rupturing all the segments simultaneously.

4.4 Plausible worst-case tsunami scenario and resulting hazard

We present a worst-case plausible scenario, related to a rupture along all the identified segments of the Morne Piton fault system as these 1 to 10 km-long segments most probably root in-depth along a single fault zone (Feuillet et al., 2004). We rule out the eventuality of testing a single 10 km-long segment of the Morne Piton fault system rupture as it would generate a $M_w<6.0$ earthquake and would thus unlikely consist in a tsunami source (e.g., Roger et al., 2019). Here, we use a plausible M_w 6.5 scenario, i.e a magnitude slightly lower that the maximum magnitude M_w 6.7 deduced from the morphological analysis, but close enough to the Les Saintes earthquakes magnitude as both Les Saintes and Morne Piton fault systems share close morphological characteristics. Our model would generate a tsunami with a significant energy/amplitude to accurately highlight the potential consequences of tsunami waves' propagation and interaction with the peculiar shallow reliefs and major embayment located in and around the Marie-Galante Basin. Quantifying horizontal run-up at the coast and assessing tsunami risk following a rupture along this fault is out of the scope of the present study as such quantifications necessitate a better knowledge of the fault dynamic itself (return period of large events, etc.).

The initial surface elevation directly resulting from the Okada (1985)'s formulation is presented in Figure 7. Due to the high inclination of the fault planes (dip = 75°) dipping globally northward, a profile cut of the initial displacement is represented from the north to the south by a crest (positive elevation) and a trough (negative elevation). At t=0, the shallow water equations take over from this initial deformation and the propagation of the tsunami waves is calculated over the nested grids at adequate time steps. Figure 8 presents the state of the virtual water surface at six different times of the tsunami waves propagation from 1 to 16 minutes. The wave front initially parallel to the fault axes ($t \leq 1$ min) is progressively influenced by the bathymetry within the very first minutes following the rupture, leading to an anisotropic propagation of the waves showing variability in space and time. In addition, the fact that the fault literally crosses Marie-Galante leads to the tsunami source being divided in two independent sources located on the west and east of this island: two tsunamis are therefore generated and called TsuW (on the west) and TsuE (on the east) hereafter. These two tsunamis propagate from their origin and wrap around Marie-Galante as shown at t=4 min of propagation. Then, between 4 and 9 min, the two tsunami fronts meet on the north and south of Marie-Galante. Meanwhile, the propagation of the TsuW waves meet the shallow waters of the Banc Colombie shoal (approx. coordinates: 15.98°N/-61.43°W; minimum water depth is about 35m), west of Marie-Galante: the waves' amplitude increases as they slow down and their interaction leads to a constructive interference resulting in a "new" tsunami source at the Banc Colombie shoal, mainly showing a negative wave propagating southward with some extensions toward Marie-Galante on the east and Les Saintes on the west.

Approaching the coasts, wave shoaling takes over, the reduction in water depth slowing down the waves and simultaneously increasing their amplitude. It leads to wave amplification as particularly shown along Marie-Galante north shore, the southeast coast of Basse-Terre and the south of Petite-Terre and eastern Grande-Terre (Figure 9).

After 10 hours of tsunami propagation, the maximum values of wave amplitude reached on each point of the simulation domain are shown in Figure 9. The overall impact of such an event is that the maximum wave amplitude is not going over 1.2 m, carefully considering the 100 m resolution of the simulation domain: in fact, grid refinement at the coast showing higher resolution would probably highlight higher wave amplitude in very localized areas because of interaction with small underwater structures not represented at this resolution, as well as non-linear effects. The patterns of those amplitudes indicate that not only the fault region but also some coastal regions are exposed to tsunami waves of 50 cm or more, which is above the usual beach and marine threat 30-cm threshold. It is the case for the neighboring coasts of Marie-Galante, southeast Basse-Terre, south Grande-Terre and the natural reserve of Petite-Terre. The southeast coast of Basse-Terre is particularly exposed with wave amplitudes of more than 1 m (Figure 9a). A focus on Les Saintes Archipelago highlights also wave amplitudes of more than 1 m, even between the islands (Figure 9b). The northeast coast of Dominica is also affected but to a lesser extent (maximum wave amplitudes of ~50 cm). Further high-resolution simulation, including flow speed calculations, would help to correctly assess the related hazard on this island. Virtual sea-level gauges have been added at different locations of grid 2 (Figure 9) in order to check if the model is stable and to look for possible resonance (especially in Les Saintes Archipelago). The raw signal of seven VG (top figure 10) highlights a clear decrease of the amplitude over the time on all stations. However, for stations VG_1, VG_2 and VG_3, a low-frequency oscillation is clearly visible and lasts for at least 10 hours. The amplitude spectrums on these three stations show that two peaks with a period of ~8.5 min and 15 min respectively are present on the three signals at Les Saintes, which is not the case for the stations out of the archipelago. Moreover, the high amplitude negative wave southwardly propagating generated by the “new” tsunami source at the Banc Colombie shoal, is shown by VG_1 on figure 10 with a peak to through value of ~0.6 m. VG_2 and VG_3 also show it to a lesser extent a bit later.

5. Discussion

5.1. Upper plate fault tsunamigenic potential.

Les Saintes earthquake demonstrated that upper plate normal faults may generate $M_w > 6$ tsunamigenic earthquakes in the Lesser Antilles. Les Saintes tsunami produced up to 2 m high waves at the coast and 42 m distance run-up in a peculiar embayment (Zahibo et al. 2005). Such normal faults are prone to be tsunamigenic because their rupture is relatively shallow (compared to the megathrust), and their slip motion is favorable to large seafloor displacement. Together with their proximity to the islands, they are able to produce metric-high tsunami waves at the coast and tens to hundreds of meters of run up distances (depending on the topography). Therefore, upper plate crustal faults may represent a major potential tsunami hazard in the Lesser Antilles islands and particularly in the Guadeloupe Archipelago as pointed out by the Intergovernmental Oceanographic Commission held in Fort-de-France in 2019 (IOC-UNESCO, 2020). Similarly to Les Saintes Fault, we assume that the 50 km long Morne Piton fault poses a potential earthquake and tsunami hazard. The large scarp we observed at the toe of the Morne Piton fault suggests recent seismogenic rupture(s) along this structure, potentially tsunamigenic. However, this scarp might not be related to the 1851 historical event as the estimated magnitude M_w 5-5.5 for this earthquake appears too low to explain the observed scarp. Thus, a rupture of the fault along its whole length must not be excluded. Several other prominent onshore-offshore faults affect the seafloor and the topography of the archipelago and may represent both an earthquake and a tsunami hazard. However, the relationships between faults, earthquakes and tsunami is not clearly established as shown in the following examples.

Along the arc, the Harvers-Montserrat-Bouillante and Les Saintes fault systems are the most prominent tectonic features (Feuillet et al., 2010). To the south, Les Saintes Fault system dips east and defines a half-graben (Leclerc et al., 2016). The westernmost fault, the Roseau fault, ruptured during the 2004 M_w 6.3 earthquake and is most likely reloading stress and therefore quiet. Recurrence time for this earthquake has been estimated to be more than 1,000 years given the regional slow strain rate. However, the eastern normal faults of the system offset the seafloor over more than 30 km and present tilted blocks filled by fan shaped late Pleistocene deposits attesting for recent deformation. In the light of these observations, the eventuality of a tsunamigenic earthquake along these faults should be considered. To the North-West along the Harvers segment, a rupture occurred in 1985 with a M_w 6.5 earthquake showing strike-slip mechanisms. Beck et al., (2012) estimated a recurrence time of 6,500-7,000 years for such a M_w 6.5 event based on the vertical offset of coseismic deposits in hemipelagites imaged by very high-resolution seismic lines across the fault. In between these two segments, the Montserrat-Bouillante segment is seismically quiet except if the 1897 (estimated M_w 7.0) earthquake occurred along this fault (Feuillet et al., 2011b). However, no tsunami related to such a rupture has been reported. Seismic lines across the Montserrat-Bouillante fault (Feuillet et al., 2010; Legendre, 2018) reveal that the fault offsets the most recent units including the oldest reflector drilled during the IODP1395, that dates upper Gelasian *ca*. 1.8-2Ma (Le Friant et al. 2013). Given an offset of 0.3m stwt and a 2000-2500 m/s sediments velocity this provides a 0.15-0.2 mm·yr^{-1} slip rate. Thus, the Montserrrat-Bouillante segment should also be considered tsunamigenic (Figure 2).

South of Grande-Terre of Guadeloupe, the N90°E trending Gosier Fault system bounds 45 km of coastal area (Figure 2). The fault system offset the Mid-Pleistocene Grande-Terre plateau that culminates at +150 m from the offshore plateau that rests 15-20 mbsl (Münch et al., 2013). This suggests a long-term vertical slip rate ca. 0.10 mm·yr^{-1}. To the east of Grande-Terre, the fault crosscuts the MIS5e terrace attesting for Late Pleistocene activity of the fault. However, evaluation of paleo-seismicity along one eastern segment of the fault system by the means of trenches allowed the identification of recent surface ruptures, although superficial deposits remain undated (Terrier and Combes, 2002).

East of Guadeloupe, offshore, the Marie-Galante Basin is bounded to the east by the Karukera spur (Figure 2D), a 75 km long N-S trending submerged plateau that culminates 30 mbsl to the north offshore La Désirade Island, and gently dips southward down to *ca*. 1500 mbsl (De Min et al., 2015). The spur is bounded to the west by N150°E to N0°E trending, west dipping, and normal faults. These faults offset the middle Miocene sequence boundary (SB2) by up to *ca*. 2,700-2,900 m, leading to a long-term vertical slip rate of 0.16-0.18 mm·yr^{-1}. Recent deposits are clearly affected by tectonic activity (Siebert et al., 2020). Located far from the islands, the earthquake intensity felt onshore would be relatively low in the island, but a tsunami could propagate across the Marie-Galante Basin directly toward the coasts of the Lesser Antilles Arc islands.

5.2. Slip rate reassessment along the Morne Piton fault system

With this study, we evidence that the slip rate along the Morne Piton fault system increases through time with a maximum slip rate of 0.25 ±0,08 mm·yr^{-1} since the last 1.29 Ma. This slip rate is up to four times slower than previous estimates. Over the last 330 Ka, Feuillet et al. (2004) estimate a bulk slip rate along the Morne Piton fault as high as 1

$mm \cdot yr^{-1}$ over 330 – 125 Ka then decreasing to 0.3 $mm \cdot yr^{-1}$ since the last 125 Ka. This last value is close to the long-term slip rate obtained offshore in this study (Figure 11). These results suggest that the fault may present a fast slip rate during short periods of time (few 100 ka.) separated by long periods (million years) of low slip rate. The fast 1 $mm \cdot yr^{-1}$ rate was obtained considering that the terrace T2MG is offset by the fault by 159 m and dates MIS7e (249 Ka) and the upper-plateau of Marie-Galante corresponds to an abrasion surface from the MIS9e high stand (330 Ka) (Feuillet et al., 2004).

This latter statement can be reconsidered. The same *Agaricia sp.* limestone unit, and *Acropora sp.* limestone unit top the three islands, Grande-Terre – La Désirade – Marie Galante, suggesting they emerged synchronously (Feuillet et al., 2002; Cornée et al., 2012; Munch et al., 2013). In the 3 islands the youngest formation, *Acropora* unit, is not younger than 1.07 Ma and not older than 1.54Ma (Cornée et al., 2012; Münch et al., 2014). From the geological map of Bouysse et al. (1993), this unit rims Marie-Galante Island. In La Désirade, the Upper Plateau culminates at 276 m asl (above sea level), whereas the 330 Ka terrace is at 35 m asl (Lardeaux et al., 2014; Leticée et al., 2019). Consequently, the hypothesis stating that the plateau emerged 330 Ka ago can be ruled out in Marie-Galante. Based on the age of the latest deposit of the Marie-Galante Plateau that range between 1.54 and 1.07 Ma, a vertical slip rate of 0.15-0.22 $mm \cdot yr^{-1}$ can be calculated. This value is close to the ca. 0.25 $mm \cdot yr^{-1}$ obtained offshore for the slip rate along the Morne Piton fault system over the same period of time (Figure 11). As a consequence, it is not possible to conclude that the Morne Piton fault system has short periods of fast slip rate, but instead it probably increases through time, a reaching a maximum slip rate of 0.25 $mm \cdot yr^{-1}$ over the last million years. As a consequence, dividing by four the slip rate along the Morne Piton fault system is increasing the earthquake time recurrence along this fault system and thus the time recurrence of potential associated tsunami. Constraining the fault slip rate at the time scale of one or few seismic cycles may allow better estimates of seismic and tsunamigenic hazards of the Morne Piton fault system. This would require a better knowledge of in-depth fault geometry and identification of its active segments that could be obtained by the means of a microseismic survey (using Ocean Bottom Seismometers acquisition over 1-2 years or more). At present-day, BOTDR laser reflectometry is used to perform long-term monitoring of the Morne Piton fault using the network of submarine telecom fiber optic cables connecting Marie-Galante to the larger islands of Basse-Terre and Grande-Terre (Gutscher et al., 2023). It is to note that given the slow strain rate and in the absence of rupture occurring along the fault during the survey period, identifying slip rate across this fault system may require hundreds to thousands of years. Moreover, very high-resolution seismic data across the fault in areas of high sedimentation rates (*i.e.*, along the eastern coast of Basse-Terre Island) may constrain the Holocene fault activity. Slip rate estimates can be obtained by coring and dating of (i) the most recent deformed sediments as well as (ii) tsunami deposits in salt marshes. Finally, such regional monitoring would also contribute to a survey of past and potential landslides that may also be induced by earthquakes and which may locally generate destructive waves.

5.3. Bathymetric features control on tsunami wave propagation

The aim of the tsunami simulation associated to the present study is not to produce a precise hazard assessment for the islands of Guadeloupe, but rather to give an overview of what could happen in terms of tsunami generation if all the identified segments of the Morne Piton fault system ruptured together, and to identify a few gaps in terms of scientific knowledge and operational activities. An accurate hazard assessment study would require many rupture scenarii including combinations of the segments used in this study, with variations of their parameters and sensitivity tests.

The main outcome of the simulation presented above lies in the fact that submarine features play an important role on the tsunami waves behavior and amplitude. Submarine canyons are known to focus the waves (e.g.: along the continental slope of the middle American trench: Álvarez-Gómez et al., 2012; or at Nazarè Portugal: Martins et al., 2010; do Carmo et al., 2022; Delpey et al., 2021). This behavior also occurs along the rime of the island submarine plateau rising the wave amplitude as exemplified in front of the most populated cities of south-eastern Grande-Terre (Figure 9). Shallow water plateaus located around or between islands slow down the waves which leads to particular propagation patterns like the wrapping around the relief (Figure 8). There, the wrapping effect of the waves around Marie-Galante and the Colombie Bank results in two distinct tsunami sources, i.e. a primary source at fault and a secondary one at the Colombie Bank (Figure 8). Such a behavior, already shown in other regions, is able to considerably amplify the impact of the tsunami on the coast opposite to the fault rupture (Chadha et al., 2005; Chen et al., 2010). It is important to notice that the low resolution of the grid used for the present simulations is a limiting factor in quantifying correctly the wave amplitude along the shoreline. A higher resolution simulation grid would better reproduce the bathymetric features, especially in shallow waters, having a non-negligible impact on the waves' behavior and amplitude.

Our simulation also highlights interesting phenomena that would require further consideration in the framework of further tsunami hazard studies: wave oscillation, which could be attributed to a resonance effect, is clearly visible within the Les Saintes Archipelago, and potential wave trapping is also visible around those islands. If the second case is purely observation, the resonance between Les Saintes islands is clearly revealed by the single-sided Fourier amplitude spectrum (Figure 10) and the peak at ~15 min seems to be associated with the negative wave coming time after the initial wave front and related to the tsunami interaction with the Colombie Bank shoal. The records provided by the virtual gauges located beforehand within the archipelago (VG_1, VG_2 & VG_3) clearly shows a long-period oscillation of the signal which is not present on the gauges located outside of Les Saintes (VG_4, VG_5, VG_6 & VG_11). It shows how the frequency content of the incoming signal can affect the sea-level during many hours after the seismic rupture.

The numerical simulations performed by Cordrie et al. (2020) of the tsunami having followed the M_w 6.3 Les Saintes earthquake were able to match the witnesses' observations in Les Saintes (Zahibo et al., 2005). Despite the low resolution (100m) of the present simulation on Les Saintes, there are some similarities between the two studies of potential impacted zones, for example in Marigot or Grande Anse Bays. It also shows that other bays, like the ones located between Terre-de-Haut and Ilet à Cabrit, appear to be quite well protected and not exposed to relatively strong tsunami waves.

Finally, this study also highlights the exposed coastline of Dominica: on the north-northeast coast of the island (Figure 9), 50+ cm waves are simulated, showing that this island should integrate such a scenario of crustal fault rupture within its tsunami hazard assessment plan.

Conclusions

Thanks to HR bathymetry, reflection seismic data and rock/sediment samples, the analyses of the morphology and tectonic structures of the Marie Galante Basin located in the middle of the Guadeloupe Archipelago, allow to detail the structural pattern of this region and to estimate a slip rate of *ca*. 0.1 mm·yr^{-1} increasing over the last million year to 0.25 mm·yr^{-1}, along the Morne Piton fault system, , cross-cutting the basin. This estimate divides the previously published estimations of the slip rate by four, and thus increases the earthquake recurrence time associated to the Morne Piton fault system from 1Ka to 4-5 Ka. We show that a seismic rupture associated to an earthquake showing a moment magnitude M_w ~6.5 can occur along the Morne Piton fault system in case of the rupture of the full length of the fault (all segments being considered connected at depth in the present demonstration). Such an event would be tsunamigenic according to numerical simulation results. The multi-segment tsunami modeling illustrates how submarine morphological and structural features influences the propagation pattern of the tsunami leading to constructing interferences and resonances, thus increasing the tsunami threat on nearby islands, especially highlighting a resonance effect within the Les Saintes Islands, not discussed so far (and potentially the explanation of the so-far unreproduced run-up values of the 2004 tsunami). At a regional scale we evidenced that several other regional faults such as Montserrat-Harvers-Bouillante Fault, Gosier Fault, Karukera Spur Border Fault, may also be tsunamigenic. Indeed, although they have the potential to produce relatively low magnitude (<7) earthquakes, their rupture could occur at shallow depth and close to a highly populated coast. Therefore, scenarii with arc and forearc crustal fault ruptures must be integrated within their tsunami hazard assessment plan. For that, it is necessary to have a better knowledge of onshore-offshore structural and seismogenic patterns of each individual major faults system as the regional low strain rate leads to large recurrence time of tsunamigenic earthquakes (> 1,000 years), *i.e.* much greater than the historical record. In addition, these earthquakes could also have the capacity to destabilize the sedimentary layers at the edge of plateaus and canyons, triggering submarine mass failures, capable of triggering large but more localized tsunamis.

Competing interests

The contact author has declared that none of the authors has any competing interests.

Acknowledgements and data

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Data availability and request

Litto3D data can be found at https://diffusion.shom.fr/loisir/litto3d-guad2016.html
French oceanographic fleet data can be obtain on demand via the Sismer online interface following the cruise doi – Aguadomar: https://doi.org/10.17600/98010120; Sismantilles: https://doi.org/10.17600/1080060 and https://doi.org/10.17600/7010020; Kashallow: https://doi.org/10.17600/9020010;
Requests for the Geoberyx03 data must be addressed to the BRGM.
Some figures have been prepared using the Generic Mapping Tools Version 6 software (Wessel et al., 2019).

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FIGURES

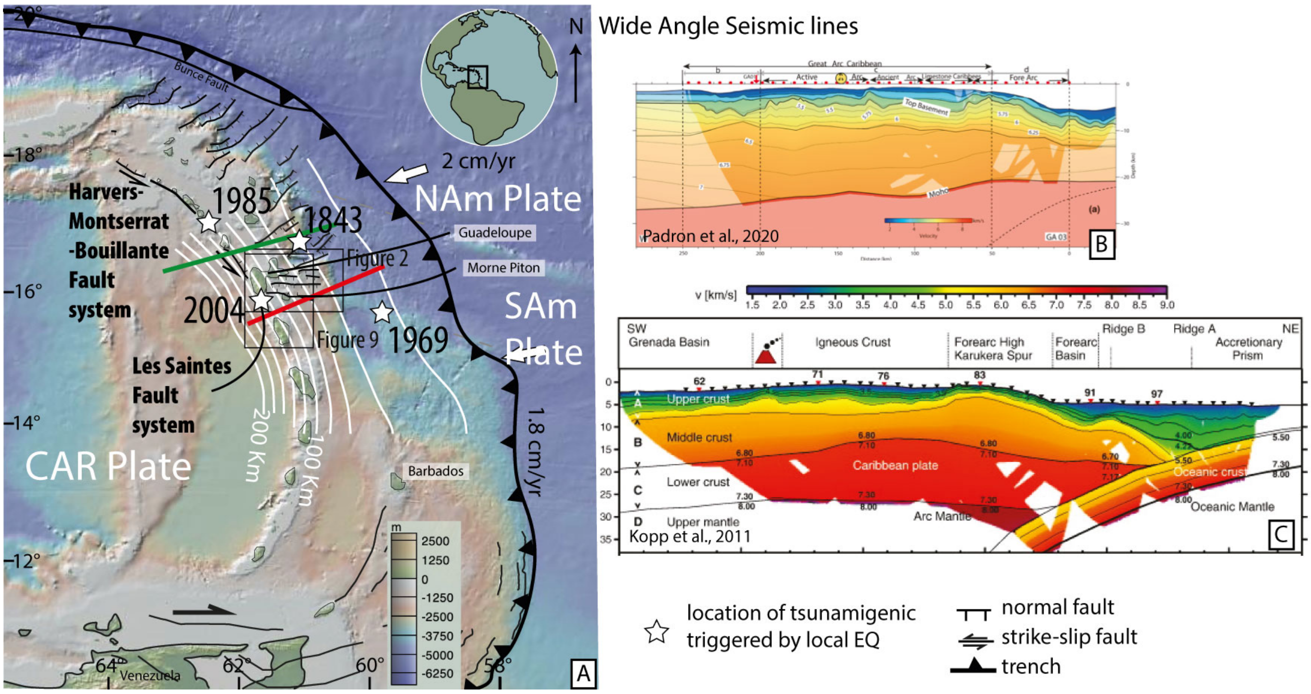

Figure 1: A: Synthetic tectonic map of the Lesser Antilles forearc. Structures after Feuillet et al. (2002), De Min (2014), Laurencin et al. (2019), Legendre (2018), Boucard et al. (2021). Red and green thick lines indicate location of the Wide Angle Seismic lines from B. Kopp et al. (2011) and C. Padron et al. (2021) respectively. White Star: location of tsunamigenic earthquakes. Thick white contour lines: Slab depth isocontour from Bie et al. (2020).

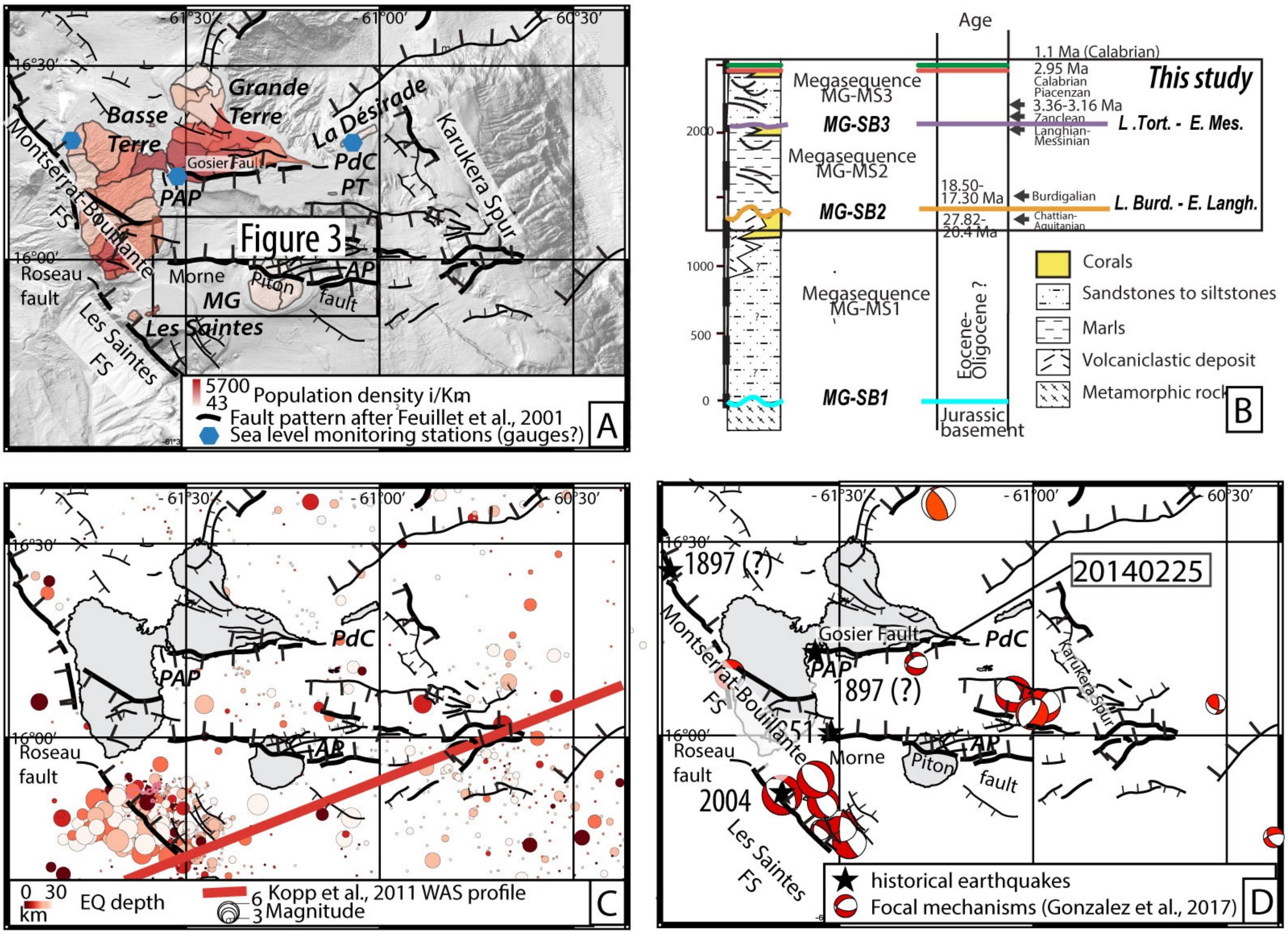

Figure 2. The Marie-Galante Basin A) structural pattern after Feuillet et al. (2002) and De Min et al. (2015) on the shaded-relief bathymetric map. blue hexagons : tide gauges http://refmar.shom.fr/fr/liste-maregraphes-data.shom.fr. Red colors scale: Guadeloupe population density per km2 after GEOFLA (https://www.data.gouv.fr/fr/datasets/geofla-r/). B) Sismostratigraphic scheme of the Marie-Galante basin modified after Cornée et al., 2023. C) Colored dots: Crustal seismicity (from IRIS seismic database 2023) for magnitude earthquakes (EQ) $3>M_w>6.5$ and located from 0 to 30 km depth), red line locates the WAS line (Kopp et al., 2011). D) Focal mechanisms solutions are indicated by red beachballs after Gonzalez et al. (2017). The location of historical earthquakes is indicated by black stars (after: Feuillet et al., 2011b). AP, PdC, PAP, PT stand for Anse Piton, Pointe des Chateaux, Pointe-à-Pitre and Petite Terre.

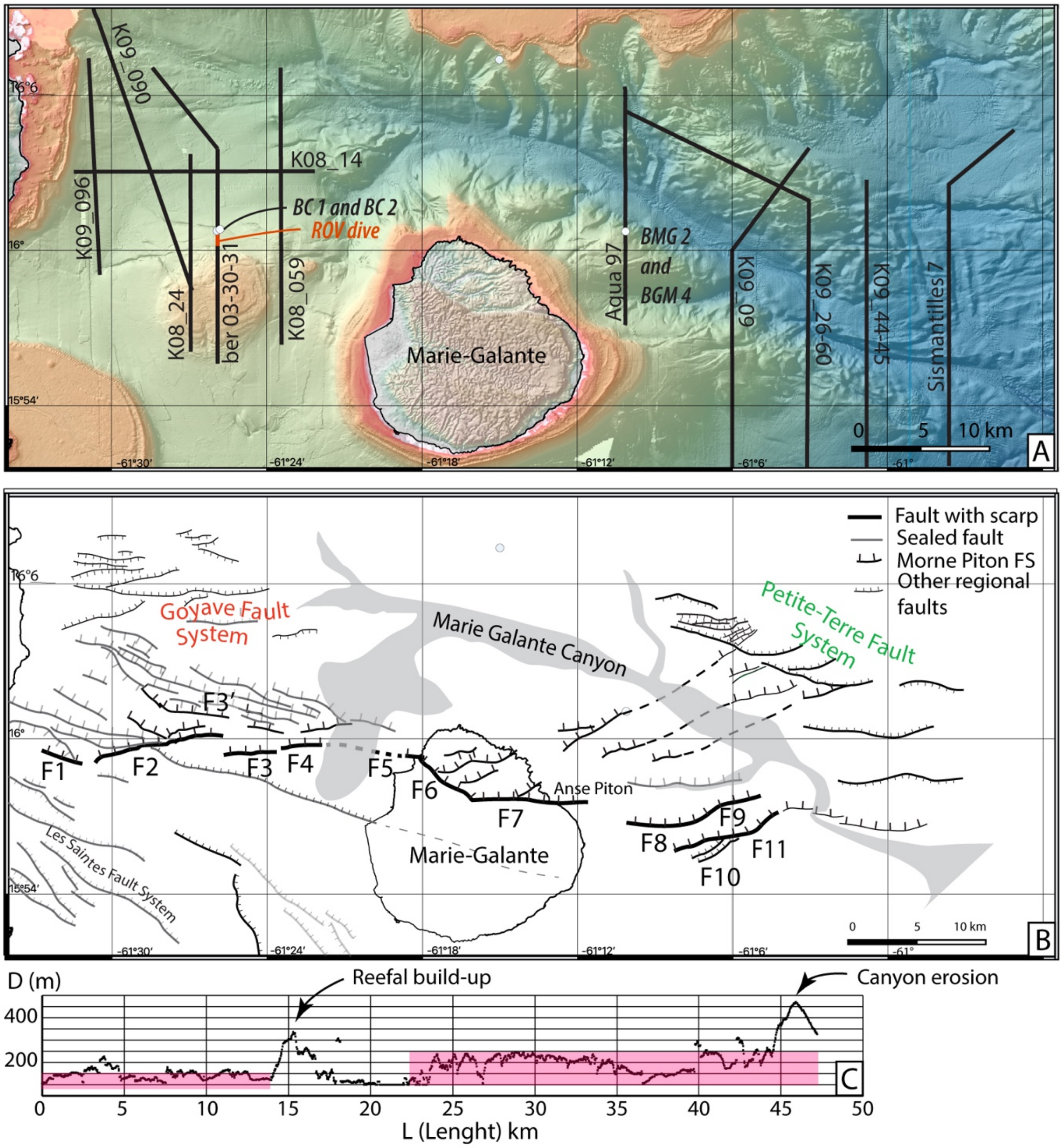

Figure 3. A) High resolution (25m grid spacing) bathymetric map [UMO16] *of the Marie-Galante Basin, offshore Guadeloupe and location of the seismic profiles shown on figures 4 and 5, and location of dredge samples used for the seismic units age calibration (Münch et al. 2013). B) Structural interpretation of the E-W trending Morne Piton fault system. C) A proxy for cumulative strain given by the graphic displaying the D (fault surface displacement) taken as the difference between the top and the toe of the fault scarp versus L (fault length) along the whole system.*

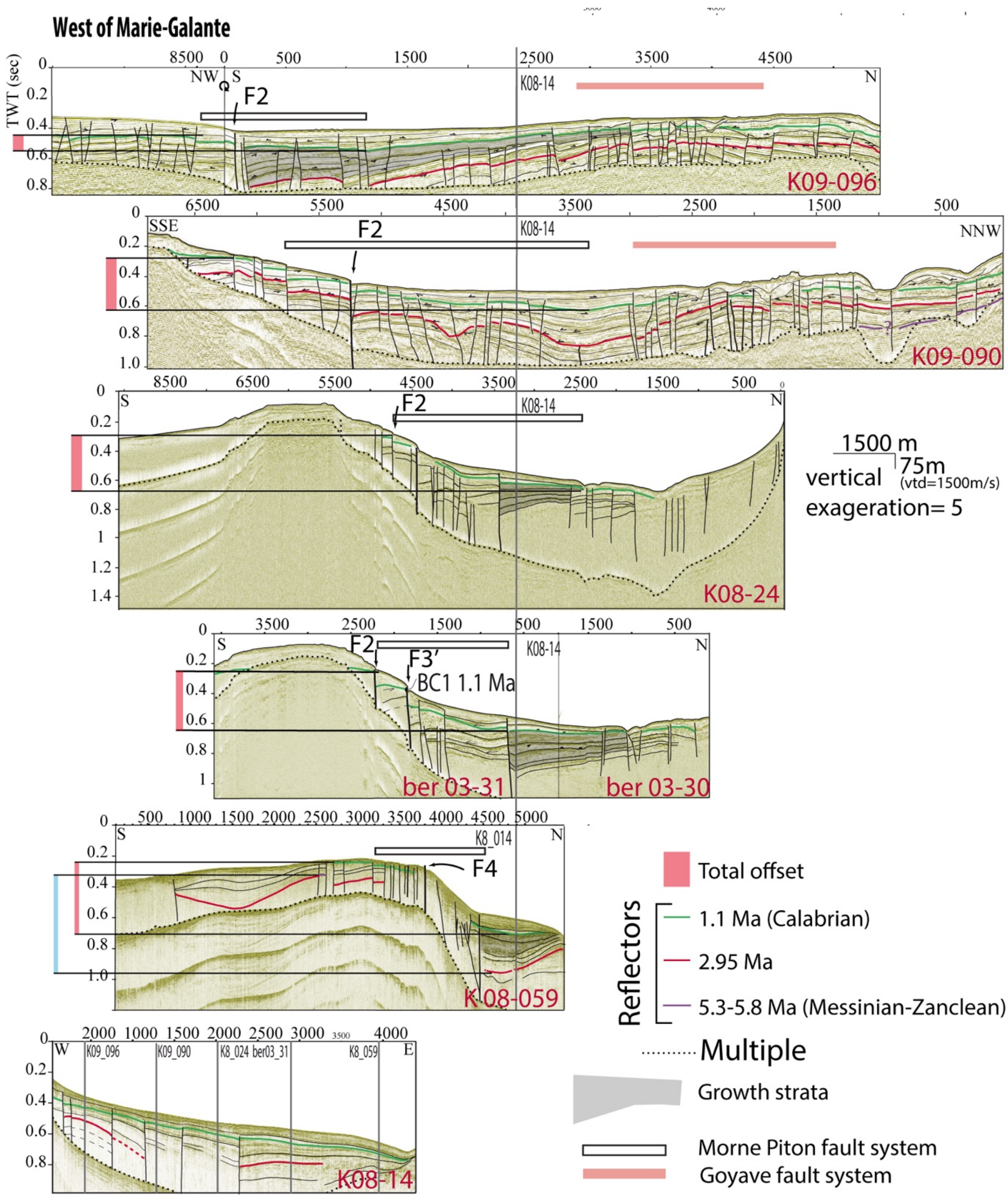

Figure 4: Seismic lines West of Marie-Galante (location on Figure 3) showing the correlation across the Morne Piton fault system of the 1.29Ma unit reflector (Green) that correspond to the reflector dredged at BC1 and BC2 location along the Ber03_30-31 seismic line. The 2.95Ma Unit Boundary (Red) is correlated from seismic lines south of the Colombie Bank and Eastern Marie-Galante Basin (line K09_90 and K09_26-62 location on figure 2). Notice that the basin sedimentary slope is in the

1286 *E-W direction (parallel to line K08-14). Therefore, the initial topography of the reflectors in the N-*
1287 *S direction, across the fault system, can be neglected when measuring the offset along the seismic*
1288 *lines.*

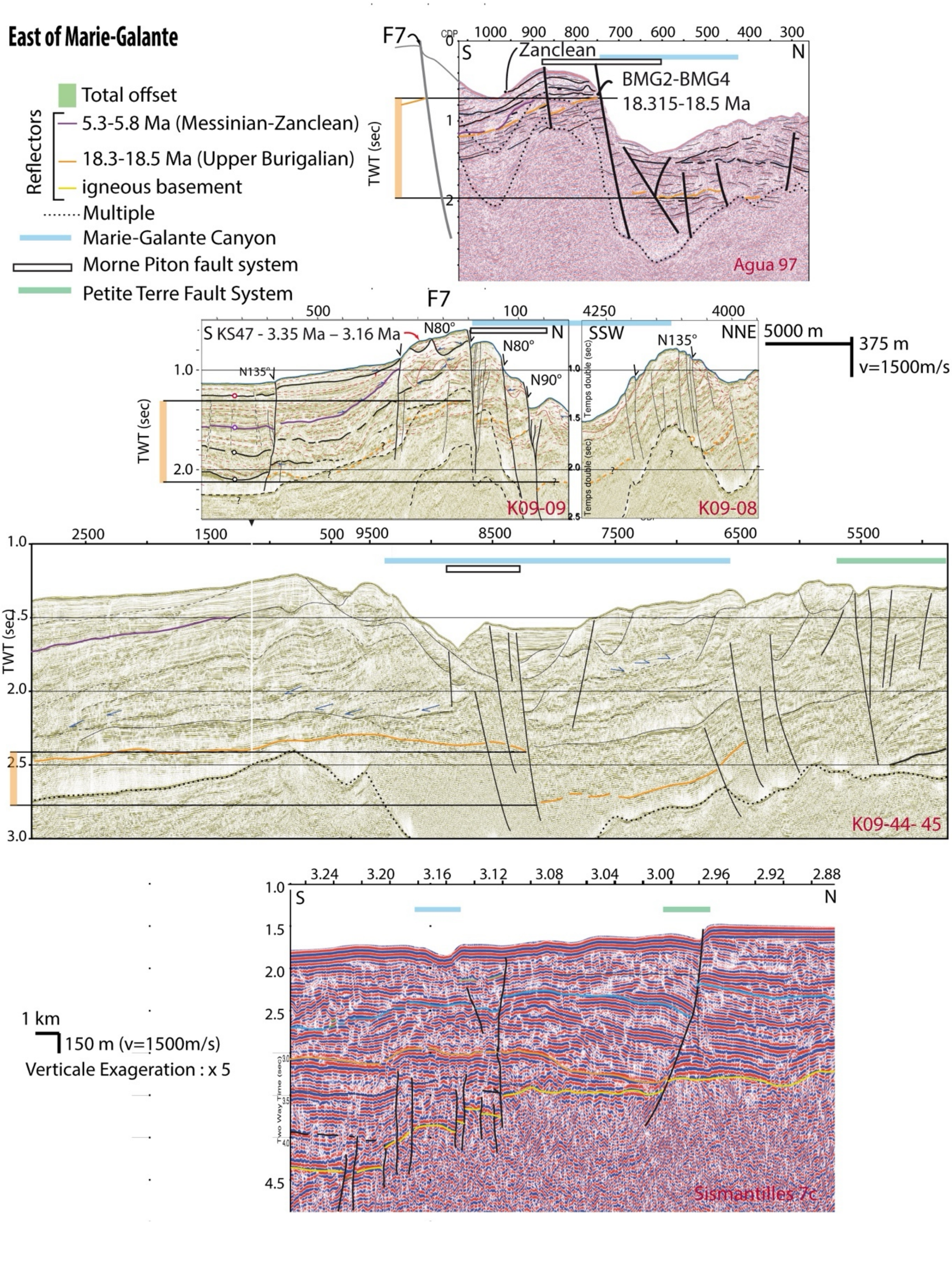

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Figure 5: Seismic lines east of Marie-Galante illustrate the correlation across the Morne Piton fault system of the 7 Ma (Tortonian/Messinian) MG-SB3 sequence boundary (Purple) and the 16 Ma, (Burdigalian) MG-SB2 Sequence Boundary (Orange). Seismic line location on Figure 3).

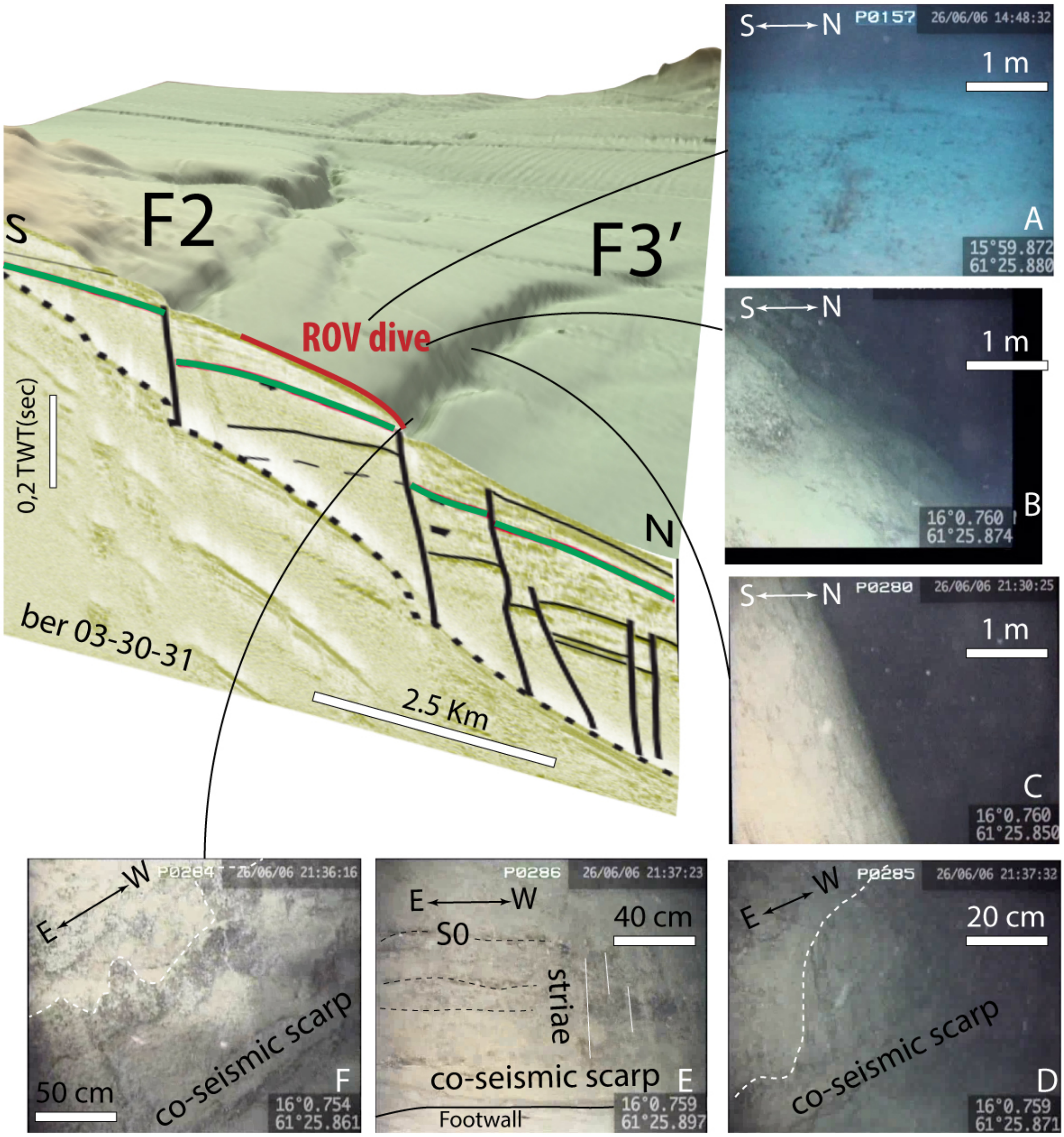

Figure 6: ROV photographs of fault identified on the seafloor along the BER03-30-31 seismic line across the F3' Morne Piton fault segment (location on Figure 3) *. (A) photography of the hangingwall of F3'. The eroded F3' fault plane presents a progressive downward slope steepening (B and C) until the toe of the fault, which is marked by a characteristic co-seismic scarp with a dip slip striae (D E F). On each photograph, white numbers starting with a P is the water depth in meters Latitude is North and Longitude is West (WGS84). A B and C views show several tens square meters wide areas,*

D E F are close up showing ca 1m high escarpment just above the foot of the fault scarp (visible at the bottom of each photos).

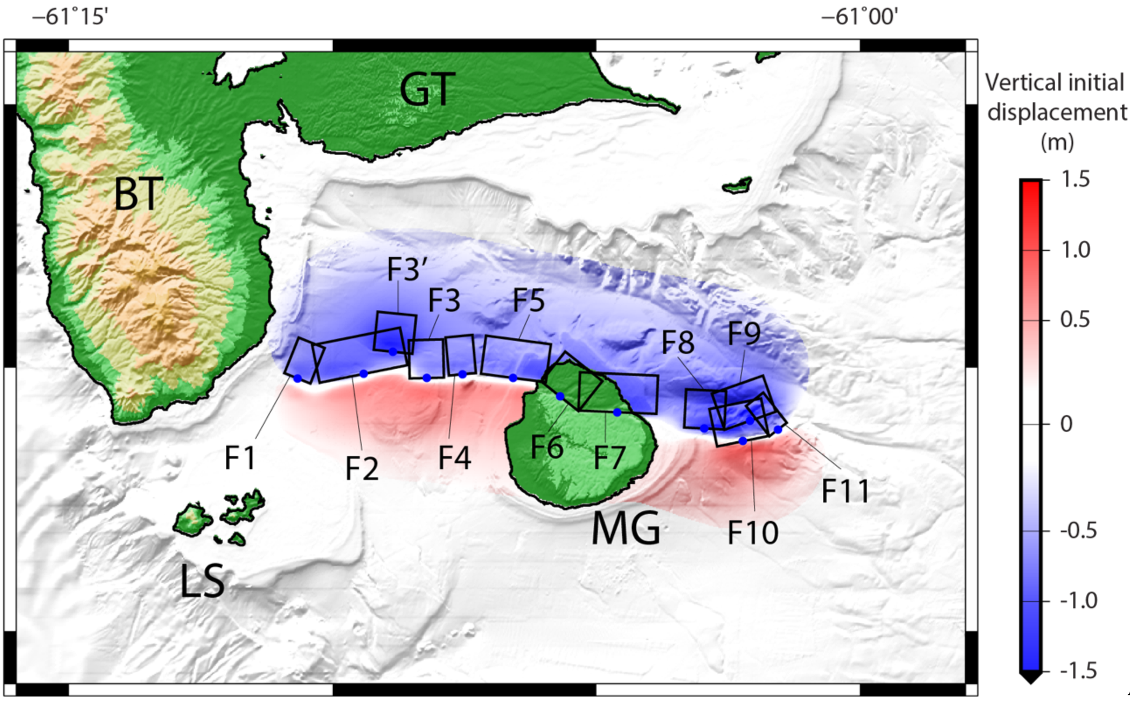

Figure 7: Initial surface elevation for a maximum credible scenario built with the 11 fault segments detailed in Table 3. Blue dots indicate the top fault center. Acronyms stand for Grande-Terre (GT), Basse-Terre (BT), Les Saintes (LS) and Marie-Galante (MG).

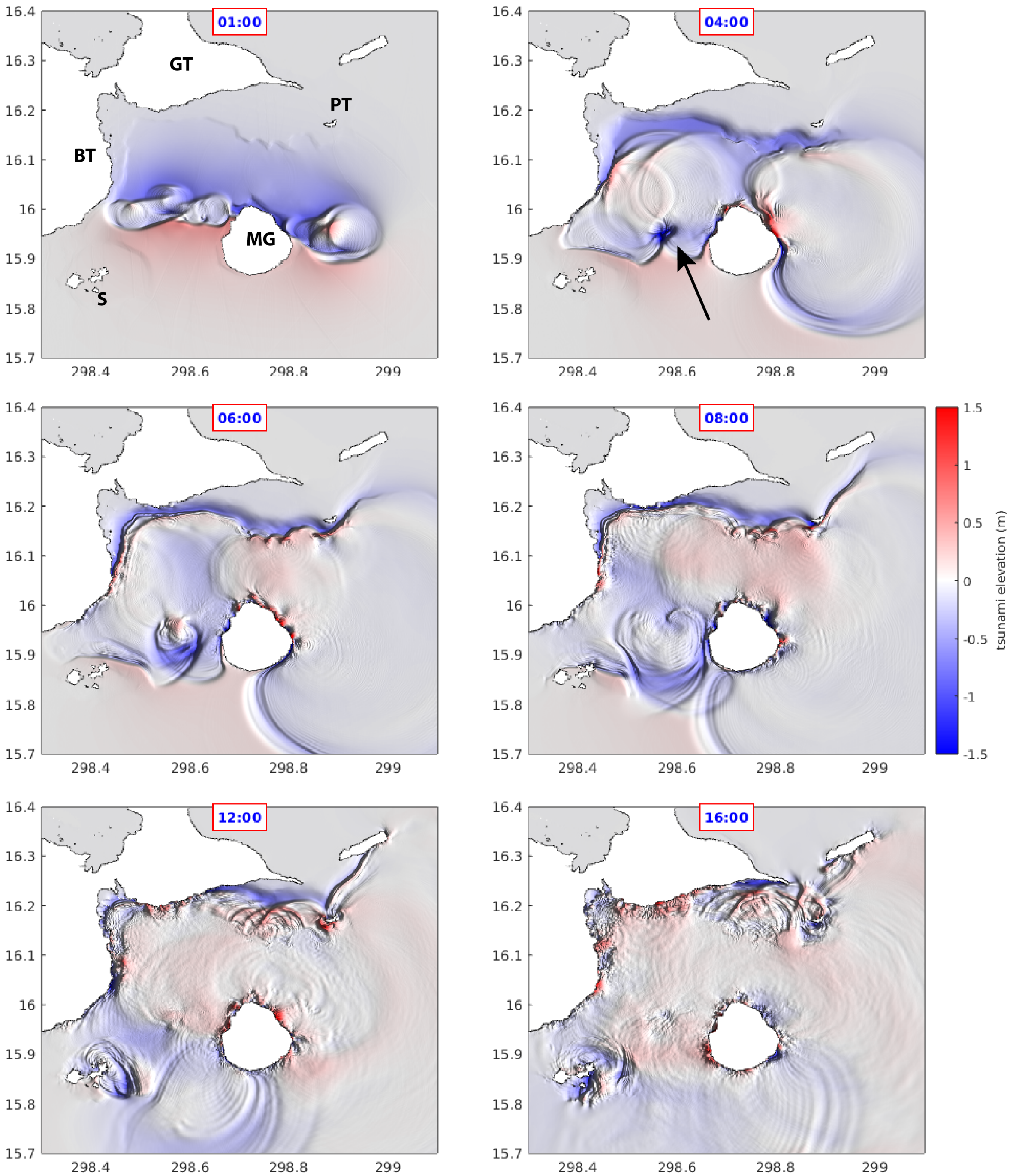

Figure 8: Snapshots of tsunami elevation within the Guadeloupe Archipelago at 1, 4, 8, 6, 12 and 16 minutes of waves propagation. Red and blue colors correspond to wave crests and troughs respectively. The black arrow shows the Banc Colombie shoal. BT: Basse-Terre; GT: Grande-Terre; S: Les Saintes; PT : Petite-Terre ; MG : Marie-Galante.

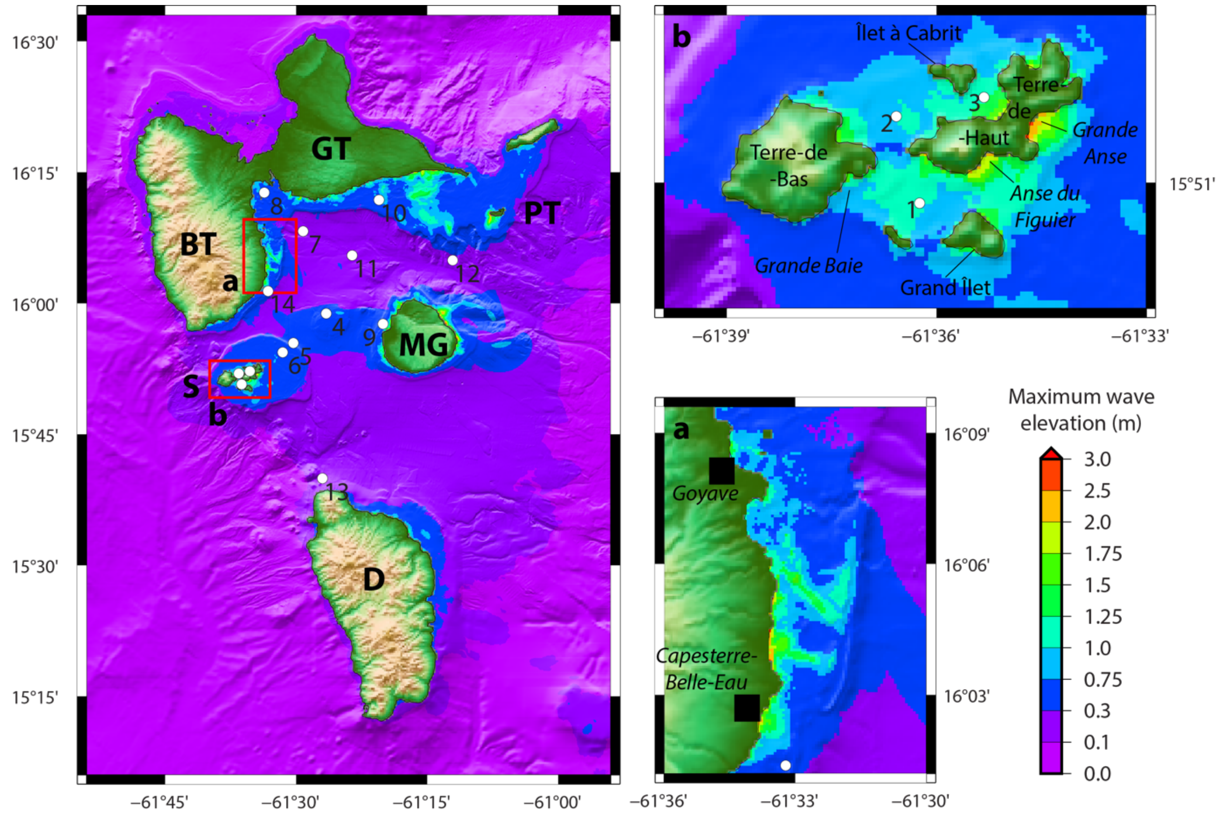

Figure 9: Shadowed bathymetric map with tsunami maximum wave elevation. Numbered white dots: fourteen virtual sea-level gauges (VG) : . BT: Basse-Terre; GT: Grande-Terre; S: Les Saintes; PT : Petite-Terre ; MG : Marie-Galante ; D : Dominica. VG_4 & VG_11 are located near the fault rupture region, the VG_5 & VG_6 are near the Les Saintes Fault system. VG_1, VG_2 & VG_3 are within the Les Saintes archipelago.

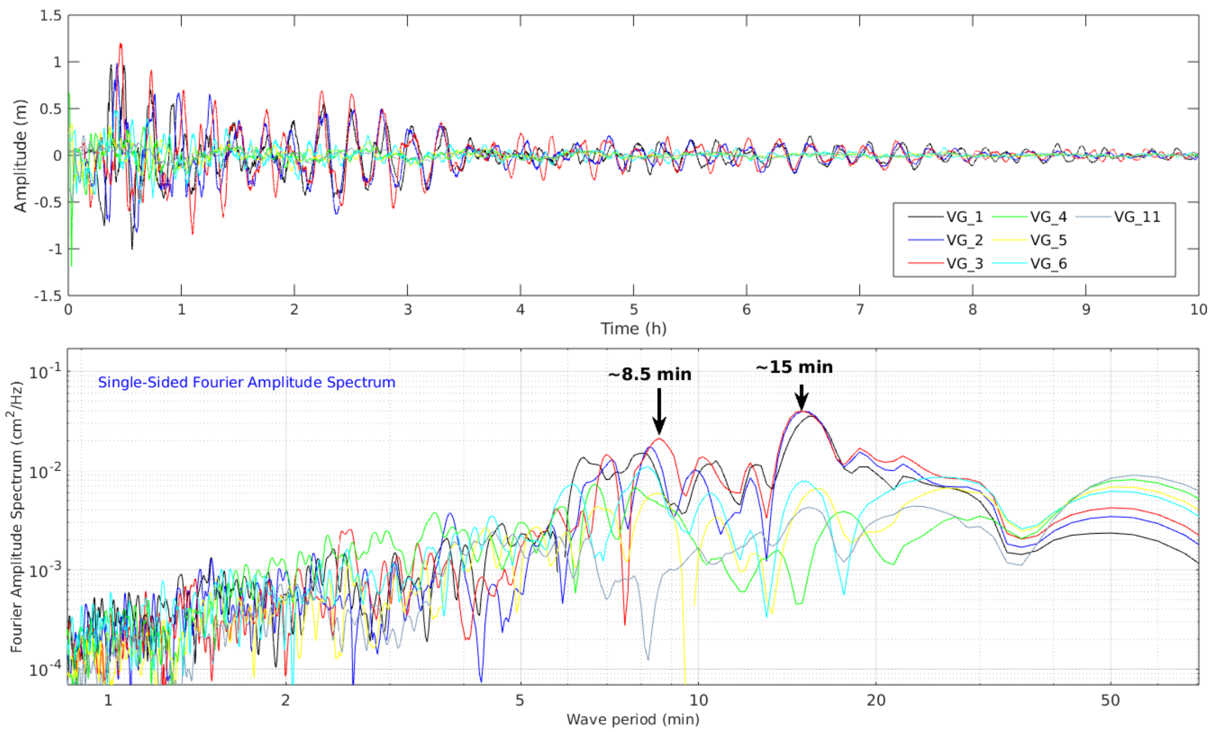

Figure 10: Post-processing of virtual gauge records. Top: sea-level records at 7 different locations (VG 1 to 6 and VG11 – location on figure 9); bottom: single-sided Fourier amplitude spectrum. The blue arrows symbolize the location of the 2 peaks of period ~6.5 and 17 min.

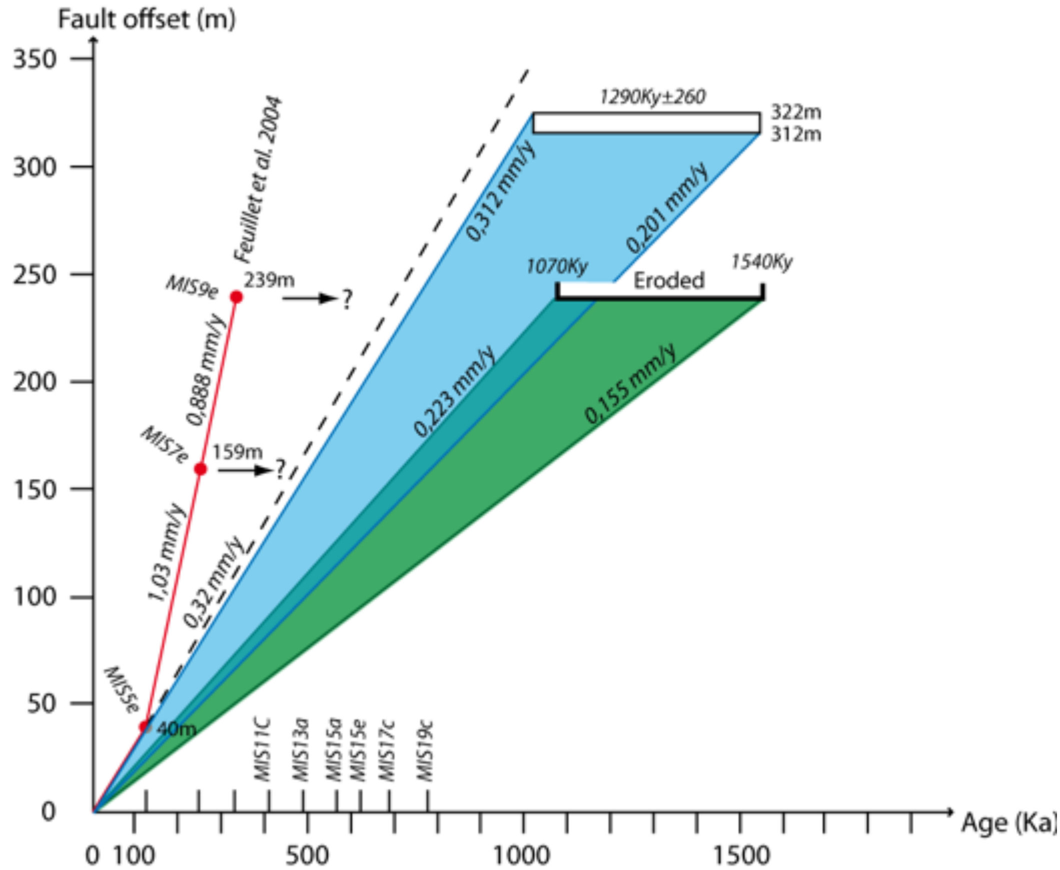

Figure 11: Fault offset along the Morne Piton fault against the age of the strain marker. Red: data from Feuillet et al. (2004) based on absolute age of terrace T4 (MIS5e), the estimated age of terrace T2 (MIS7e) and the suggested age of Marie-Galante upper plateau (MIS9e). Green: strain range calculated using upper plateau unit age from Münch et al. (2014) (note that erosion may lower this estimation strain rate). Blue: strain range calculated from the fault offset of the seismic unit dated ca 1,2Ma along the seismic line K08-59 (green reflector on Figure 4). Dotted line indicates the 0.32 $mm \cdot yr^{-1}$ *strain rate from the estimated offset of the MIS5e Terrace in Marie-Galante (Feuillet et al. 2004).*

Cruise	KaShallow 1	KaShallow 2	Aguadomar	SismAntilles	GEOBERYX03 SISM BGM
Seismic source	1000 J sparker	35-45 in3 GI Airguns array	45-105 in3 Two GI Airguns	4400 in3 Airguns array.	1000 J sparker
Peak frequency (far field)	250-400Hz	40-70Hz	30-50Hz	15-20Hz	250-400Hz
Number of traces	6 traces	72 traces	6 traces	360 traces	6 traces
Fold coverage	3/6 fold	9/18 fold	3 fold	30 fold	3/6 fold
Inter CDP distance	4 m	3.125 m	4 m	6,25 m	4 m

Table 1: Main acquisition parameters of the seismic data shown in this study (Figures 4 and 5).

Profile Name Sorted from East to West	Age	Uncertainty (Ma)	horizon depth in the footwall (stwt)		horizon depth in the hangingwall (stwt)		horizon depth in the footwall (m) V en m/s 2500	horizon depth in the hanging wall (m) V en m/s 2500	horizon depth in the footwall (m) V en m/s 2000	horizon depth in the hanging wall (m) V en m/s 2000	**TOTAL OFFSET (m) V en m/s 2500**	**TOTAL OFFSET (m) V en m/s 2000**	Strain rate (mm/years)				Interval strain rate (mm/yrs)
			water	rock	water	rock	max	max	min	min	**max**	**min**	min	max	**Average**	**nty**	
K09-44-45 – MG-SB2	**16**	**1**	1,575	0,83	1,575	1,2	2 212,5	2 700,0	2 006,3	2 396,3	**487,5**	**390,0**					
since inception	**7**	**1,5**											0,046	0,089	**0,067**	**0,030**	
K09_08-09 – MG-SB2	**16**	**1**	0,595	0,91	1,27	0,9	1 577,5	2 052,5	1 351,3	1 832,5	**475,0**	**481,3**					
since inception	**7**	**1,5**											0,057	0,086	**0,071**	**0,021**	
Agua 97 – MG-SB2	**16**	**1**	0,05	0,72	1,36	0,6	937,5	1 770,0	757,5	1 620,0	**832,5**	**862,5**					
since inception	**7**	**1,5**											0,101	0,151	**0,126**	**0,035**	0,07
UB4	**2,95**	**0,05**	0,23	0,09	0,59	0,4	285,0	905,0	262,5	812,5	**620,0**	**550,0**	0,183	0,214	**0,199**	**0,022**	0,16
K08-059	**1,29**	**0,26**	0,22	0,02	0,5	0,1	190,0	512,5	185,0	485,0	**322,5**	**300,0**	0,194	0,313	**0,253**	**0,085**	0,25
Ber 03-31	**1,29**	**0,26**	0,21	0,04	0,53	0,1	207,5	522,5	197,5	497,5	**315,0**	**300,0**	0,194	0,306	**0,250**	**0,079**	
K08-24	**1,29**	**0,26**	0,25	0,04	0,6	0	237,5	487,5	227,5	480,0	**250,0**	**252,5**	0,163	0,243	**0,203**	**0,056**	
K09-090	**1,29**	**0,26**	0,33	0,02	0,49	0,1	272,5	530,0	267,5	497,5	**257,5**	**230,0**	0,148	0,250	**0,199**	**0,072**	
K09-096	**1,29**	**0,26**	0,32	0,12	0,44	0,1	390,0	505,0	360,0	470,0	**115,0**	**110,0**	0,071	0,112	**0,091**	**0,029**	

Table 2: Measured offset of seismic reflectors across the Morne Piton fault system and calculated total vertical strain rates. See text for the ages estimate. Seismic reflectors depth on each side of the fault system is measured in time (stwt – second two way time) and converted in depth in using water velocity (1500m/s) and two end-member velocities for the sediment (see text for explanation), providing a minimum and a maximum offset value. The minimum strain rate is obtained from the ratio between the min offset and the max age bound of the reflector and vice versa.

Fault segment	Lon (°)	Lat (°)	Length (m)	Width (m)	Top of the fault plane depth (m)	Strike (°)	Dip (°)	Rake (°)	Slip (m)
F1	-61.5335	15.98987	2838	12500	500	111	75	-90	1.89
F2	-61.4708	15.994	9070	12500	500	78	75	-90	1.89
F3'	-61.4428	16.01503	3918	12500	500	95	75	-90	1.89
F3	-61.4108	15.99009	3474	12500	500	88	75	-90	1.89
F4	-61.377	15.99367	2735	12500	500	84	75	-90	1.89
F5	-61.3287	15.99021	6623	12500	500	97	75	-90	1.89
F6	-61.2846	15.97275	4813	12500	500	128	75	-90	1.89
F7	-61.2301	15.95767	7904	12500	500	92	75	-90	1.89
F8	-61.1477	15.94221	4052	12500	500	92	75	-90	1.89
F9	-61.1042	15.94958	5495	12500	500	70	75	-90	1.89
F10	-61.111	15.93014	5336	12500	500	78	75	-90	1.89
F11	-61.0777	15.94126	2228	12500	500	52	75	-90	1.89

Table 3: Parameters used for the tsunami source simulation of a rupture along the multisegment fault presented in figure 7.