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ESTIMATION OF EXPECTED MAXIMUM WATER LEVEL DUE TO TIDE AND TSUNAMI INTERACTION ALONG THE COASTAL BELTS OF PENANG ISLAND IN PENINSULAR MALAYSIA

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ABSTRACT

In this paper, an estimate of the expected maximum water levels associated with tide and tsunami interaction is computed along the coastal belts of Penang Island in Peninsular Malaysia. For this purpose, a nonlinear Polar coordinate shallow water model of the Indonesian tsunami of 2004 by Roy et al. (2007b) is used. Appropriate tidal condition is generated in the domain by applying tidal forcing through the western open sea boundary. For studying tide and tsunami interaction, the 2004 Indonesian tsunami is introduced in the previously generated tidal oscillation. The expected maximum possible water level along the coastal belts of Penang Island is estimated based on the interaction of tide and tsunami for different tidal conditions (high and low tidal periods). It is seen that the surge level is very sensitive towards the coastal belts due to interaction during high tide. The west coast of Penang Island is found to be vulnerable for stronger surge due to interaction. The influence of tidal wave on the tsunami wave height towards Penang Island is also investigated. The tide was found to have a significant effect in tsunami enhancement in the coastal regions.

***Key words**: Tide, Surge, Penang Island, Indonesian Tsunami 2004.*

1. INTRODUCTION

The coastal environment is diverse, with astronomical tides, wind, waves and currents; all contributing to the forces experienced by coastal features. An example of a devastating natural calamity is the event that occurred in the Indian Ocean on 26 December 2004. An important phenomenon in applied coastal oceanography is the interaction of high astronomical tides and tsunami. Tides are the rising and falling of Earth's ocean surface caused by the tidal forces of the Moon, the Sun and other planets acting on the oceans. The interaction of tides and tsunami leads to significant high water levels, thus increasing the risk of coastal flooding, shoreline erosion and damage of urban drainage systems. Furthermore, an increase in the mean water level together with large tsunamis may produce severe damages to coastal structures, which are usually designed without taking into account an abnormal rise of in the sea level.

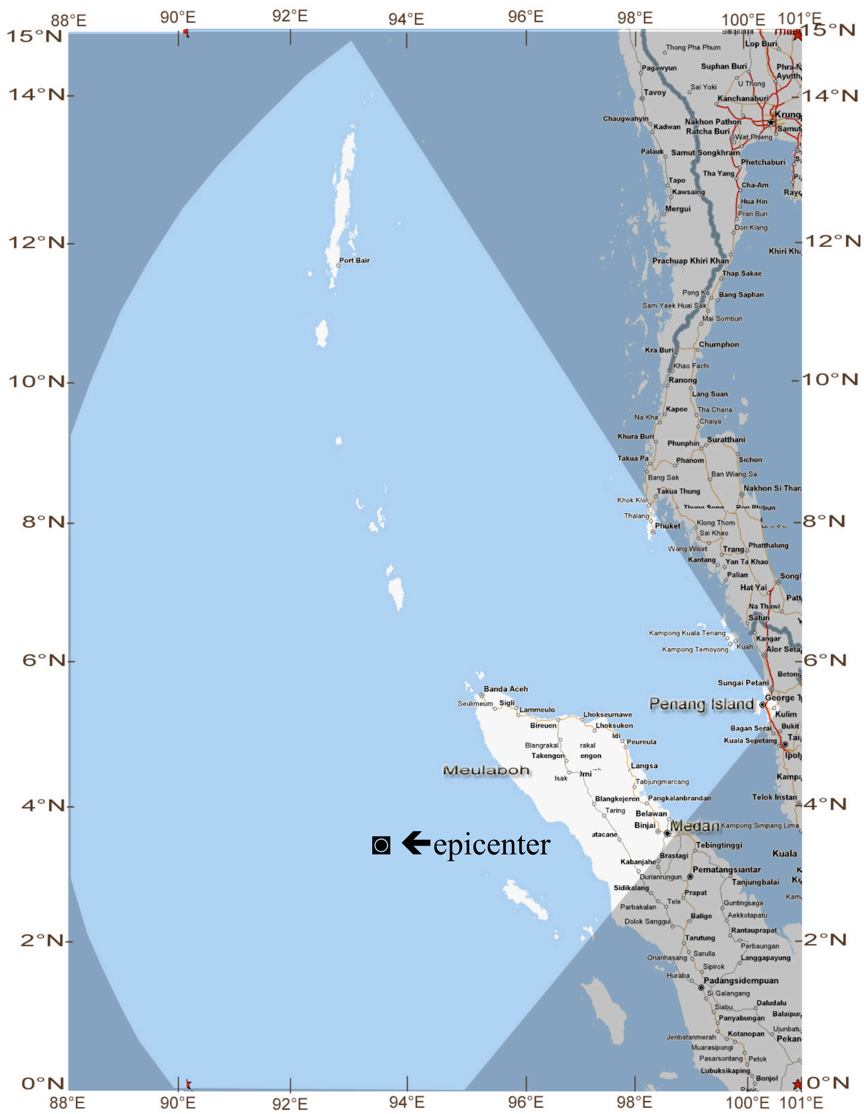

Figure 1: Map of the model domain including North Sumatra and Penang Island

The Malacca strait (Singapore up to Penang Island) is a large tidal range (difference between high and low tide) area. Among the tidal constituents *M*2 and *S*2, due to attractions of the moon and the sun respectively, are predominant in the region. Since the astronomical tidal phenomenon is a continuous process in the sea, tsunamis always interact with the astronomical tide and so the pure tidal oscillation in the whole basin should be considered as the initial dynamical condition for the tide-tsunami interaction phenomenon. The interaction with tide is generally nonlinear and the nonlinear effect is prominent in the very shallow regions; for example near a island or coastal belt. So in order to compute accurately the tsunami in a particular coastal region it is necessary that the interaction of tide and tsunami should be carried out.

Many studies concerning tide, surge due to tropical storms and their interaction can be found in the literature (e.g. Flather 1976, 1981; Flather and Davies, 1976; Heaps and Jones, 1981; Nihoul, 1977, 1982; Johns et al. 1985, Roy 1995). Considerable studies have also been done for the computation of Indonesian tsunami of 2004. Examples include Kowalik et al. (2005), Kowalik and Proshutinsky (2006), Roy and Izani (2006), Roy et al. (2007 a, b) and Karim et al. (2007). Kowalik and Proshutinsky (2006) investigated the dynamics defining tsunami enhancement in the coastal regions and related to interaction with tides. In their study, two simple cases of tide/tsunami interactions along a narrow and wide shelf have been investigated to define importance the nonlinear interactions. However, the tidal effect on tsunami propagation has not been fully investigated.

The present paper addresses this important aspect in applied coastal oceanography: the nonlinear interaction of high astronomical tides and tsunami which lead to higher water levels. The initial tsunami wave is generated in the deep ocean with the strength that of Indonesian tsunami 2004 which occurred at approximately 160 km west of North Sumatra (Fig 1). The purpose of the paper is to estimate the expected maximum possible water level due to tide and tsunami interaction along the coastal belts of Penang Island in Peninsular Malaysia. This information might be needed, for example, to build the base of a tsunami and a storm surge shelter at a particular region along a coastal belt.

2. THE MATHEMATICAL MODEL

2.1. *Shallow water equations*

We introduce the vertically integrated shallow-water equations in Polar form, which are, in standard notation (Roy et al. 2007b):

$$\frac{\partial \zeta}{\partial t} + \frac{1}{r}\frac{\partial}{\partial r}\left[r(\zeta + h)v_r\right] + \frac{1}{r}\frac{\partial}{\partial \theta}\left[(\zeta + h)v_\theta\right] = 0 \qquad (1)$$

$$\frac{\partial v_r}{\partial t} + v_r \frac{\partial v_r}{\partial r} + \frac{v_\theta}{r}\frac{\partial v_r}{\partial \theta} - f v_\theta = -g\frac{\partial \zeta}{\partial r} - \frac{C_f\, v_r\, (v_r^2 + v_\theta^2)^{1/2}}{\zeta + h} \qquad (2)$$

$$\frac{\partial v_\theta}{\partial t} + v_r \frac{\partial v_\theta}{\partial r} + \frac{v_\theta}{r}\frac{\partial v_\theta}{\partial \theta} + f v_r = -\frac{g}{r}\frac{\partial \zeta}{\partial \theta} - \frac{C_f\, v_\theta\, (v_r^2 + v_\theta^2)^{1/2}}{\zeta + h} \qquad (3)$$

where v_r = radial component of velocity of the sea water, v_θ = tangential component of velocity of the sea water, f = Coriolis parameter = $2\, \Omega \sin\varphi$, Ω = angular speed of the earth , φ = latitude of the location, g = acceleration due to gravity, ρ = the sea water density, C_f = the coefficient of friction.

2.2. *Boundary conditions*

For a closed boundary the normal component of velocity is considered as zero. The radiation type of boundary conditions are used for open boundaries which allow the disturbance created within the analysis area to go out of the area. The analysis area is bounded by the radial lines $\theta = 0°$, $\theta = \Theta = 110°$ through O and the circular arc $r = R$ (Fig. 1). Following Roy et al. (1999) the northern and southern open sea boundary conditions are respectively given by

$$v_\theta + \sqrt{(g/h)}\ \zeta = 0 \quad \text{along } \theta = 0 \tag{4}$$

$$v_\theta - \sqrt{(g/h)}\ \zeta = 0 \quad \text{along } \theta = \Theta \tag{5}$$

For generating tide in the basin at the circular open sea west boundary, the condition is taken as

$$v_r - \sqrt{(g/h)}\ \zeta = -2\left(\frac{g}{h}\right)^{\frac{1}{2}} a \sin\{(2\pi t)/T + \phi\} \quad \text{along } r = R \tag{6}$$

where a is the amplitude, T is the tidal period, φ is the initial phase.

2.3. *Transformation for uneven resolution along radial direction*

The polar coordinate system automatically ensures finer resolution along tangential direction near the region of the pole. By setting the Pole suitably at the location where fine resolution is required, a uniform grid of size $\Delta\theta$ is generated in the tangential direction by a set of radial lines through the Pole. The arc distance between any two successive radial lines decreases towards the Pole and increases away from the pole. Thus uneven resolution is achieved in the tangential direction although uniform grid size $\Delta\theta$ is used.

To achieve uneven resolution along radial direction, fine to coarse in the positive radial direction, according to Haque et al. (2005), the following transformation is used:

$$\eta = c \ln\left(1 + \frac{r}{r_0}\right) \tag{8}$$

where r_0 is a constant of the order of total radial distance and c is a scale factor. From this transformation we obtain a relationship between Δr and $\Delta\eta$ which is as follows:

$$\Delta r = \frac{r + r_0}{c} \Delta\eta \tag{9}$$

This relation shows that, keeping the value of $\Delta\eta$ as constant, we can generate a variable Δr. For a constant value of $\Delta\eta$, Δr will increase with increase of r, so that we obtain uneven resolution (fine to coarse) in the radial direction in the physical domain while in computational domain the resolution remains uniform. The boundary conditions will remain unchanged under the transformation.

3. GRID GENERATION AND NUMERICAL SCHEME

3.1. *Grid generation*

In the physical domain the grid system is generated through the intersection of a set of straight lines given by θ = *constant* through the Pole O (5° 22.5′ N, 100° 30′ E) and concentric circles, with centre at O, given by r = *constant*, where the total number of grids are 778×307. The angle, $\Delta\theta$, between any two consecutive straight grid lines through O is constant; whereas the distance between any two consecutive circular grid lines, Δr, increases in the positive radial direction. After the transformation (8), both $\Delta\theta$ and $\Delta\eta$ become uniform where $\Delta\theta$ is set to be (110/306)° and $\Delta\eta$ is set to be 1/777.

3.2. *Numerical scheme and implementation of the tsunami source*

The governing equations and the boundary conditions are discretized by finite-difference (forward in time and central in space) and are solved by a conditionally stable semi-implicit method using a staggered grid system, similar to the Arakawa C, as described in Roy et al. (1999). The time step is taken as 10 seconds that ensures the CFL stability criterion of the numerical scheme. The values of the friction coefficient is taken as uniform (C_f = 0.0033) throughout the physical domain. The depth data used in this study are collected from the Admiralty bathymetric charts.

The generation mechanism of the 26 December 2004 tsunami was mainly due a static sea bed deformation caused by an abrupt slip at the India/Burma plate interface. The estimation of the extent of the earthquake rapture as well as the maximum uplift and subsidence of the seabed is given in Kowalik et al. (2005). From the deformation contour, it is seen that the estimated uplift and subsidence zone is between 92° E to 97°E and 2°N to 15°N with a maximum uplift of 5.07 m at the west and maximum subsidence of 4.74 m at the east (Fig. 1). In the source zone of Indonesian tsunami 2004, an appropriate value of sea surface rise/fall has been assigned as initial condition to each ζ-point of the staggered grid system. Since the pole is considered as on land, where no computation is done, there is no problem of instability during numerical computation.

4. TIDE AND TSUNAMI INTERACTION PRINCIPLE

In order to simulate the tide and tsunami interaction, the initial step is to generate the exact tidal oscillation during a tsunami period in the model basin by prescribing the appropriate sea surface elevation along the open boundary. This provides the initial sea-state condition for the interaction process. It is required that at the time of introducing the tsunami source in the model simulation process the tidal response, already generated, must match with the tidal oscillation in the actual basin. To incorporate the effect of tsunami, it is required to activate the tsunami source and to allow the propagating tsunami wave across the analysis area. In the Malacca Straits *M*2 and *S*2 constituents are predominant, so that in the spring-neap cycle there is significant variation of the tidal range. The diurnal inequalities (difference in heights between two successive high/low waters) with respect to high/low waters are also significant (Fig. 2). Moreover, due to the superposition of tidal constituents of different periods, the tidal oscillation is not exactly periodic. Thus the complexity associated with tidal phenomena leads to the difficult task of generating the exact tidal oscillation in the model basin.

Instead of generating such a complex oscillatory system, a relatively simple procedure is adopted. This leads to an oscillatory tidal solution with period of *M*2 constituent where the computed high water at each location is in agreement with the average of high waters predicted in the tide table for the tsunami period. For interaction purpose, the tsunami source would have to be activated at a particular time, say $t = T_0$, into the initially generated tidal oscillation. Let the tsunami wave approach the coast at time, say $t = T$. Now the procedural requirement is to adjust the time $t = T_0$ with such a phase of the tidal oscillation that at the time $t = T$ the phase of pure tidal oscillation at each location is in agreement with that in the actual basin.

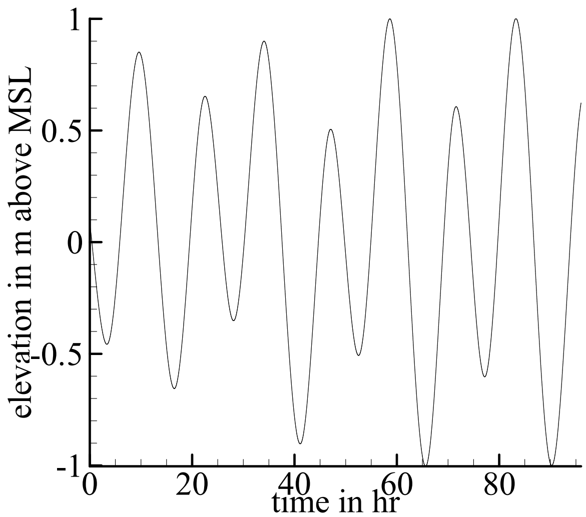

Figure 2. Computed water level due to tide at George Town (North-East coast of Penang Island).

5. INTERACTION METHODOLOGY FOR INDONESIAN TSUNAMI 2004

To investigate the influence of tidal phenomenon on tsunami wave height along the coastal belts of Penang Island associated with the Indonesian tsunami 2004, three model runs were set up:

- only tsunami wave
- tsunami wave input during high tide
- tsunami wave input during low tide

For the purpose of analysis, the results are presented in various forms at two coastal locations; one at north-west coast and another at George Town (north-east coast) of Penang Island, where tidal information are available in Malaysian tide table. The Indonesian tsunami 2004 arrived at north-west coast at the time of high tide and arrived at north-east coast at the time of low tide. For computing water level due to superposition or interaction at a particular location, tidal information of that location must be available. For the purpose of interaction/superposition of tide and tsunami at a coastal location, the time series of tide at that location is necessary. But in general, the tidal information is available, as high and low values, four times a day in the tide table of Malaysia. The corresponding time series is generated through a cubic spline interpolation method. Figure 2 represents the tidal information at George Town situated at the north-east Coast of Penang Island.

The tide is generated in the model domain through the west open sea boundary condition (6) with appropriate values of a, T and φ in absence of tsunami. It is observed from the tide table that though there is variation in the tidal period at the head of Penang Sea, the average period is approximately of $M2$ tide and so we chose T = 12.4 hours and φ = 0. The information of the tidal amplitude along the west boundary is not available. We have chosen a = 0.8 m to test the response of the model along the coastal belt. The response is found to be sinusoidal with the same period (12.4 hours). But the amplitude at every location may not be exactly same in reality. By providing appropriate values of the amplitude of a, and phase φ along the western open sea boundary, the model is expected to generate representative tidal oscillation in the whole basin.

6. RESULTS AND DISCUSSIONS

Figure 3 depicts the time series of water level due to tsunami (only) at a coastal location (north-west coast) of Penang Island. The arrow (in the horizontal axis) indicates the time of initiating the tsunami source. The maximum elevation is approximately 3.1 m. The time series of water level at the same coastal location due to the interaction of high tide and tsunami is presented in Fig. 4. The tsunami source is activated in such a phase that tsunami wave arrives at the north-west shore of Penang when the tidal amplitude achieves maximum (during high tide). It is seen that major oscillation occurred during high tide period. Due to the interaction of tsunami with high tide the water level is found to be more (3.4 m) than that due to tsunami only (Fig. 3). Thus tide has significant effect on water height along the coastal belt.

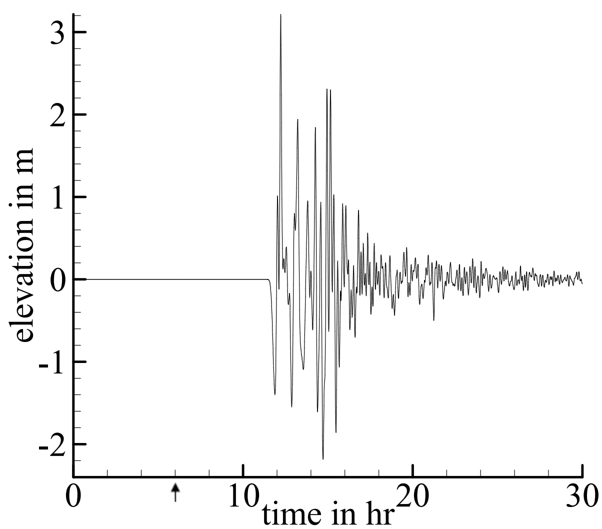

Figure 3. Time series of water level due to tsunami only at north-west coast. The arrow indicates the time of initiating the tsunami source.

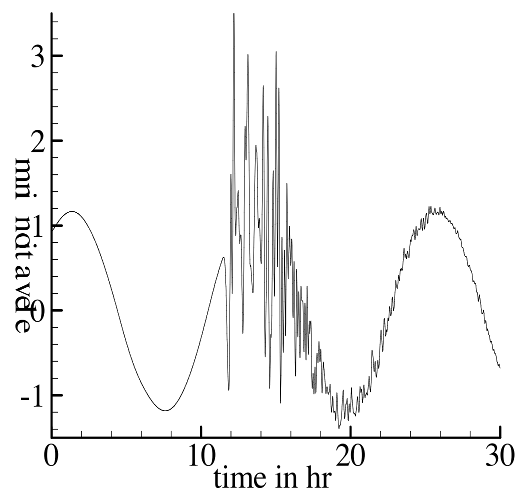

Figure 4. Water level due to interaction of tsunami and high tide at the same coastal location.

In order to investigate the significance of non-linear interaction of tide and tsunami, comparisons is done between superposition of tide and tsunami and nonlinear interaction of tide and tsunami. Figure 5a shows the time series of water level obtained by superposition of tide and tsunami (tide + tsunami) at the North- west coast of Penang Island. This is obtained by superimposing linearly the time series of tsunami response obtained through model simulation and that of tidal oscillation obtained from tide table (Fig. 2). Figure 5b shows the time series of water level due to the nonlinear interaction of tide and tsunami at the same coastal location. Comparison shows that oscillation is identical but with different amplitude. It is seen that superposition gives the over estimation of water level. So, it is important to compute non-linear interaction near the beach.

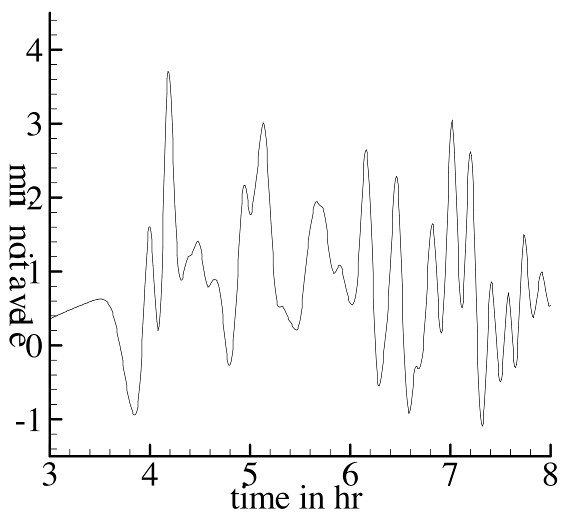

Figure 5a. Water level due to superposition of tide and tsunami (tide + tsunami) at north-west coast of Penang

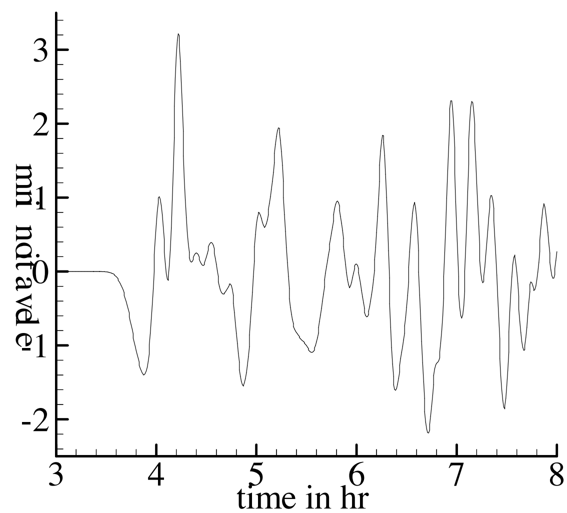

Figure 5b. Water level due to nonlinear interaction of tide and tsunami at the same location.

The influence of tide on tsunami is also investigated. Figure 6a show the water level at the same coastal location of Penang Island due to tsunami only. On the other hand, the water level at the same location is obtained by subtracting tide from tide and tsunami interaction (tide and tsunami interaction – tide) and presented in Fig. 6b. Since both the curves are not identical, it follows that the tide has a significant effect on ultimate tsunami response along the coastal regions.

Investigation on the influence of low tide on tsunami has been carried out at George Town. In this computation tsunami source is activated in such a phase that tsunami wave arrives at the north-east shore when the tidal amplitude reaches to minimum (during low tide). Figure 7 show the interaction of tide and tsunami at George Town. The surge response due to interaction of tide and tsunami during low tide period is found to be less (1 m) than that due to tsunami surge only.

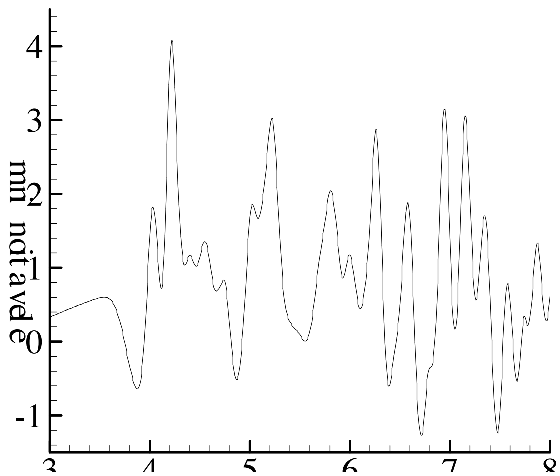

Figure 6a. Time series of water level due to tsunami at a coastal location of Penang Island.

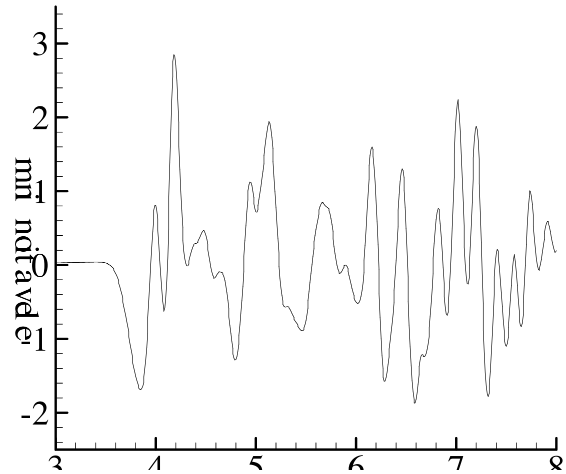

Figure 6b. Water level due to tide, tsunami interaction – tide at the same coastal location.

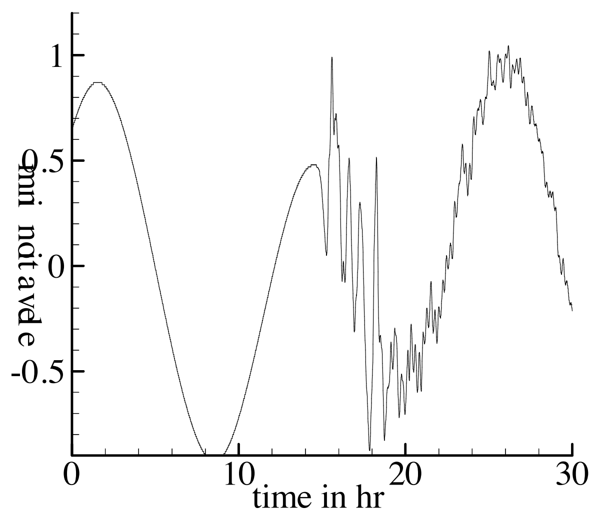

Figure 7: Water level due to the interaction of tide and tsunami at north-east coast.

A post tsunami survey report for some coastal locations of Penang Island, done by the authors, is available in Roy et al. (2007b). Computed results at two coastal locations are found to be in good agreement with the observations. Figure 8 depicts the computed maximum water levels due to tide and tsunami interaction during the high tide period along the west coast of Penang Island, where the maximum elevation is found to be 5 m. The surge intensity is found to be higher at the north-west coast. On the other hand, the maximum water levels due to tide and tsunami interaction during low tide period along the west coast are found to be 3.6 m (Fig. 9). Thus, tide has a significant effect on tsunami enhancement along the coastal regions.

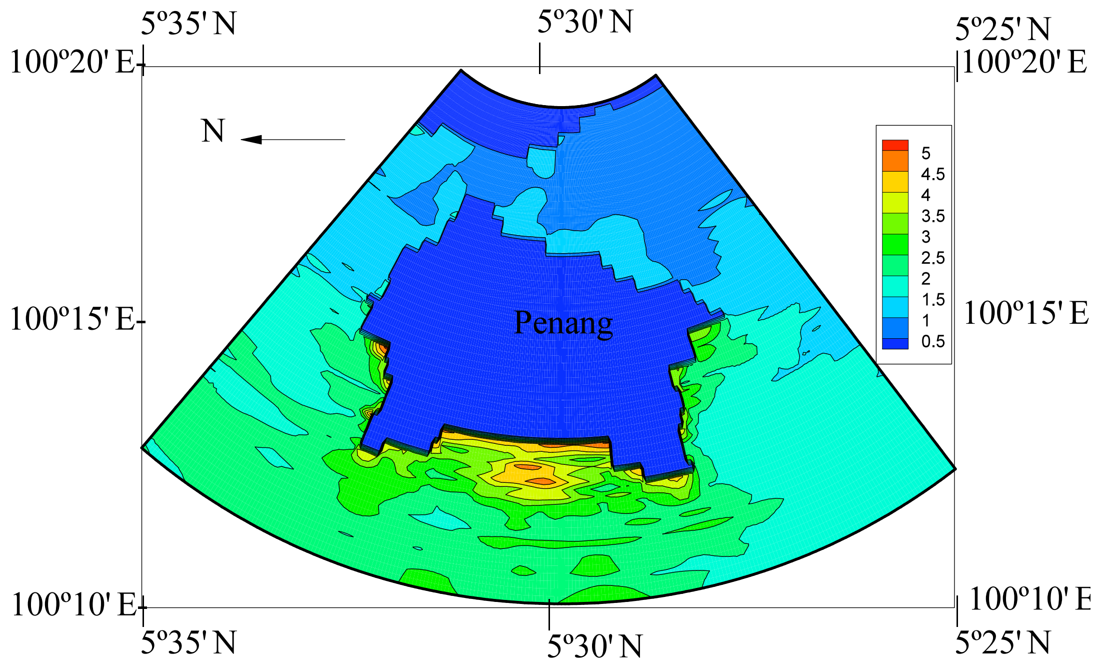

Figure 8. Maximum elevation along the coastal belts of Penang Island during high tidal period

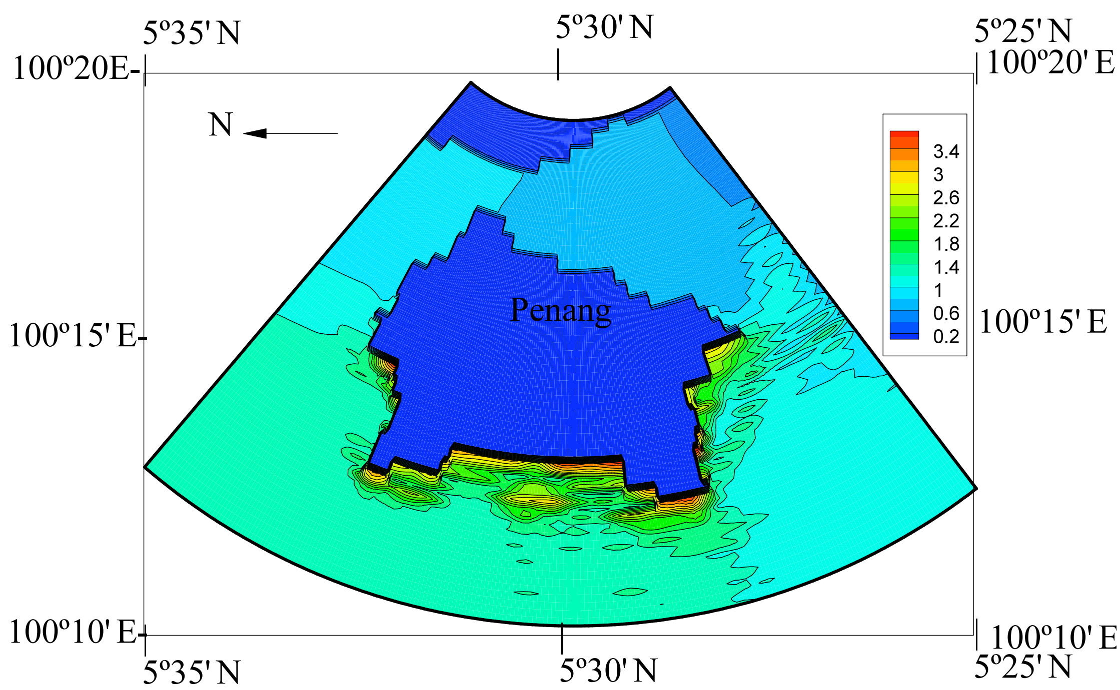

Figure 9. Maximum elevation along the coastal belts of Penang Island during low tidal period

7. CONCLUSION

It has been shown that the astronomical tides have significant effect in tsunami enhancement in the coastal regions. The nonlinear interaction of the tide with tsunami is important, as it generates stronger sea level changes along a coast. So for accurate prediction of tsunami along a coastal belt, it is necessary to consider the tidal effect on tsunami.

Acknowledgements

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REFERENCES

Flather, R. A., Davies, A. M. Note on a preliminary scheme for storm surge prediction using numerical models, Quart. J. Roy. Metrol. Soc. 102, 1976, 123-132.

Flather, R. A. A tidal model of northwest European continental shelf, Mem. Soc. Roy. Sci. Leg. 10, 1976, 141-164.

Flather, R. A. Storm surge prediction using numerical models for surge prediction, In: proc. on monsoon dynamics. Cambridge: Cambridge University Press. 1981, 659-687.

Haque, M.N., Miah, S., Roy, G.D. *A Cylindrical Polar Coordinate Storm Surge Model with Radial Stretching for the coast of Bangladesh*, SUST Studies, 6(1), 2005, 1-13.

Heaps, N. S., Jones, J. E. Three dimensional model for tide and surges with vertical eddy viscosity prescribed in two layers, II Irish sea with bed friction layer, Geophys. J. Roy. Astron. Soc. 64, 1981, 303-320.

Nihoul, J. C. J. Three-dimensional model of tides and storm surges in a shallow well-mixed continental sea, Dyn. Atmos. Oceans, 2, 1977, 29-47.

Nihoul, J. C. J. Hydrodynamic models of shallow continental seas (application to the North Sea), Riga, E., ed. La Salle, Belgium, Neupre, 1982.

Johns, B., Rao, A. D., Dube, S. K., Sinha, P. C. Numerical modeling of tide-surge interaction in the Bay of Bengal, Phil. Trans. Roy. Soc. London A 313, 1985, 507 – 535.

Karim, M. F., Roy, G. D., Ismail, A. I. M. An Investigation on the Effect of Different Orientation of a Tsunami Source along the Coastal Belt of Penang Island: A Case Study of the Indonesian Tsunami 2004**,** *Far East J. of Ocean Research*, 1 (1), 2007, 33-47.

Kowalik. Z. and Proshutinsky. T. Tide – Tsunami interactions, Sc. Tsunami Hazards. 24(4), 2006, 242 – 256.

Kowalik, Z., Knight, W., Whitmore, P. M. Numerical Modeling of the Global Tsunami: Indonesian Tsunami of 26 December 2004. Sc. Tsunami Hazards. 23(1), 2005, 40 – 56.

Roy, G.D. Estimation of expected maximum possible water level along the Meghna estuary using a tide and surge interaction model, *Environment International*, 21(5), 1995, 671-677.

Roy, G.D., Humayun Kabir, A.B.;Mandal, M.M.; Haque, M.Z.1999. Polar Coordinate Shallow Water storm surge model for the coast of Bangladesh, Dyn. Atms. Ocean, 29, 397-413.

Roy, G.D. and Izani, A. M. I. Numerical modeling of Tsunami along the coastal belt of Penang using a Polar coordinate shallow water model, Far East J. App. Math., 23 (3),2006, 241-264.

Roy, G. D., Karim, M. F., Ismail, A. M. A Linear Polar Coordinate Shallow Water Model for Tsunami Computation along North Sumatra and Penang Island, *Oriental J. Mathematical Sciences*, 1 (1), 2007a, 1-16.

Roy, G. D., Karim, M. F., Ismail, A. M. A Non-Linear Polar Coordinate Shallow Water Model for Tsunami Computation along North Sumatra and Penang Island, *Continental Shelf Research*, 27, 2007b, 245–257.

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POTENTIAL DEFICIENCIES IN EDUCATION, INSTRUMENTATION, AND WARNINGS FOR LOCALLY GENERATED TSUNAMIS

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ABSTRACT

A review of historical data for Hawaii reveals that significant tsunamis have been reported for only four of twenty-six potentially tsunamigenic earthquakes from 1868 through 2009 with magnitudes of 6.0 or greater. During the same time period, three significant tsunamis have been reported for substantially smaller earthquakes. This historical perspective, the fact that the last significant local tsunami occurred in 1975, and an understandable preoccupation with tsunamis generated around the margins of the Pacific, all combine to suggest apparent deficiencies in: (1) personal awareness of what to do in the event of a possible local tsunami; (2) the distribution of instrumentation capable of providing rapid confirmation that a local tsunami has been generated; and (3) the subsequent issuance of timely warnings for local tsunamis. With these deficiencies, far more lives may be lost in Hawaii due to local tsunamis than will result from tsunamis that have originated along the margins of the Pacific. Similar deficiencies may exist in other areas of the world threatened by local tsunamis.

Key words: Tsunami preparedness; tsunami education; tsunami instrumentation; warning systems; local tsunamis, Hawaii tsunamis

1. INTRODUCTION

Public officials, educators, residents, and visitor industry personnel need to know that the greatest threat to loss of life from natural hazards in the Hawaiian Islands may be from locally generated tsunamis rather than from Pacific-wide tsunamis. Although the discussions that follow are based on Hawaii's history and current state of preparedness for future tsunamis, the resulting conclusions of this report may be applicable to agencies and individuals in other coastal areas of the world threatened by local tsunamis.

2. HISTORICAL DATA

Of the thirty-one earthquakes with magnitudes of 6.0 or greater reported in the Hawaiian Islands from 1868 through 2009 (U.S. Geological Survey on-line data), five of these were in the interior of the Big Island and the other twenty-six were in or near the Big Island's coastal areas. Most, for which depths are available, were shallow with the greatest reported value being 50 km. Of the twenty-six earthquakes, only four of these generated significant local tsunamis (Table 1; Figs. 1 through 3). The other reported local tsunamis were generated by much smaller earthquakes or by possible submarine landslides (Table 1; Fig. 2). In addition to these reported tsunamis, unexplained flooding with characteristics consistent with a tsunami generated by a submarine landslide was observed along the eastern shoreline in the Pohoiki and Opihikao area of the Big Island. The vertical height of flooding on land (i.e., run-up) was estimated to be about 27 feet and the inland extent of flooding (i.e., inundation) was estimated to be about 800 feet. No earthquake was felt prior to the flooding and no other shoreline areas reported flooding. Although the flooding was initially considered to possibly be the result of a storm (Cox and Morgan 1977), later unpublished investigations indicated that a storm could not have been the cause of this flooding (personal communication).

In summarizing the available historic data, we find that most earthquakes with magnitudes of 6.0 or greater have not generated significant local tsunamis, while much smaller earthquakes or submarine landslides that may not even be felt could generate highly localized but potentially deadly tsunamis. As will be revealed in discussions that follow, this fact poses a daunting challenge for our public officials, educators, and visitor industry in terms of education, instrumentation, and warnings.

3. EDUCATIONAL DEFICIENCIES

Some very simple, life saving concepts related to locally generated tsunamis are not well known in our State. This deficiency is, in part, a result of our understandable preoccupation with past tsunamis originating along the margins of the Pacific that have struck Hawaii. A common singular recommendation for people in inundation areas is: "If you feel the ground shake, move as quickly and safely as possible to higher elevations"**.** However, as the historical data indicates, this advice may not be sufficient to save some lives.

An additional recommendation should be the following. “If you feel the ground shake, do not assume that a siren or other warning will be given to you in time to save your life. Also, do not assume that the earthquake is far enough away so that you can take your time in getting to higher elevations.” The tsunami generating the earthquake that is felt may be very large and hundreds of miles away, or very small but very close. There may be a few minutes or a few tens of seconds to avoid death or injury.

Knowledge of the following could also save lives. “Tsunamis are a series of waves and the first wave is not necessarily the largest. You may not feel a small earthquake or submarine landslide that could send a highly localized but potentially deadly tsunami towards your coastal area. Even small tsunamis can flood far inland for several minutes. Subsequent tsunami waves can interact with earlier waves draining off the land back into the ocean producing even higher waves with powerful, debris-laden currents. Always pay attention to any early signs of a tsunami. These signs could include the unexplained exposure of reefs, inland flooding, unusual currents, or unusual changes in the locations of breaking waves.”

Table 1
Locally Generated Tsunamis in the Hawaiian Islands *

Year	Month	Day	Ms	Run-ups on Tide Gauge Readings in Feet
1868	04	03	7.5	45
1901	08	09	---	4
1908	09	21	6.8	4
1919	10	02	---	14
1951	08	21	6.9	4
1952	03	17	4.5	10
1975	11	29	7.2	47

** Data are taken from Lander and Lockridge (1989). Run-ups are maximum vertical measures of a tsunami's wave height on land relative to mean sea level. The run-ups or tide gauge readings given here are the largest reported for the Big Island. Significant run-ups (i.e., 3 feet or more) were not reported on any island other than the Big Island. Additional run-up values for the Big Island are given in Figures 1 through 3. See Walker (2000) for a discussion of the 1901 event. Newspaper articles were examined for indications of felt earthquakes possibly associated with the 1901 and 1919 events (Cox and Morgan, 1977; and Walker, 2000). None were found. Also, in searching Hawaiian Volcano Observatory (HVO) earthquake reports based on seismic data, no events could be found that might be responsible for the 1919 tsunami. There is no evidence to suggest that the HVO seismic network was not operational on 2 October 1919.*

Public officials, educators, and visitor industry personnel should be responsible for ensuring that residents and visitors respect and understand the unique characteristics of locally generated tsunamis, requiring quick evacuations and the avoidance of false assumptions. Many of these educational deficiencies were apparent during the 6.7 Kiholo earthquake of 15 October 2006 off the west coast of the Island of Hawaii as hundreds of visitors and residents stood along the shorelines immediately after the earthquake wondering what to do. People in inundation zones on other islands also felt the shaking but did not move quickly to higher elevations. Fortunately, in this instance, no significant local tsunami was generated. Should such educational deficiencies persist in the event of an 1868 or 1975 type tsunami along the west coast of the Big Island rather than along the east coast of the Big Island, fatalities would be especially large on the Big Island and Maui, with additional fatalities possible on Oahu and Kauai

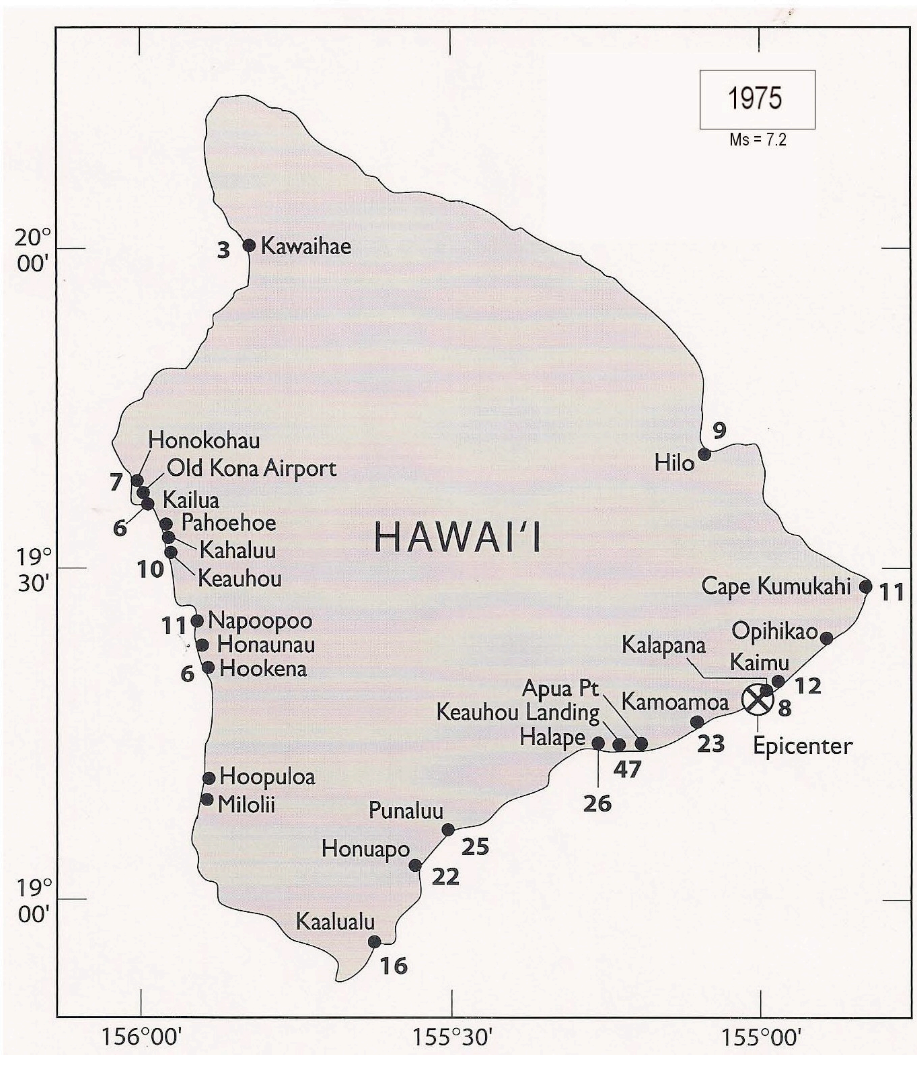

Figure 1. Earthquake epicenter, magnitude, and run-up values in feet for the 1975 tsunami (Lander and Lockridge, 1989).

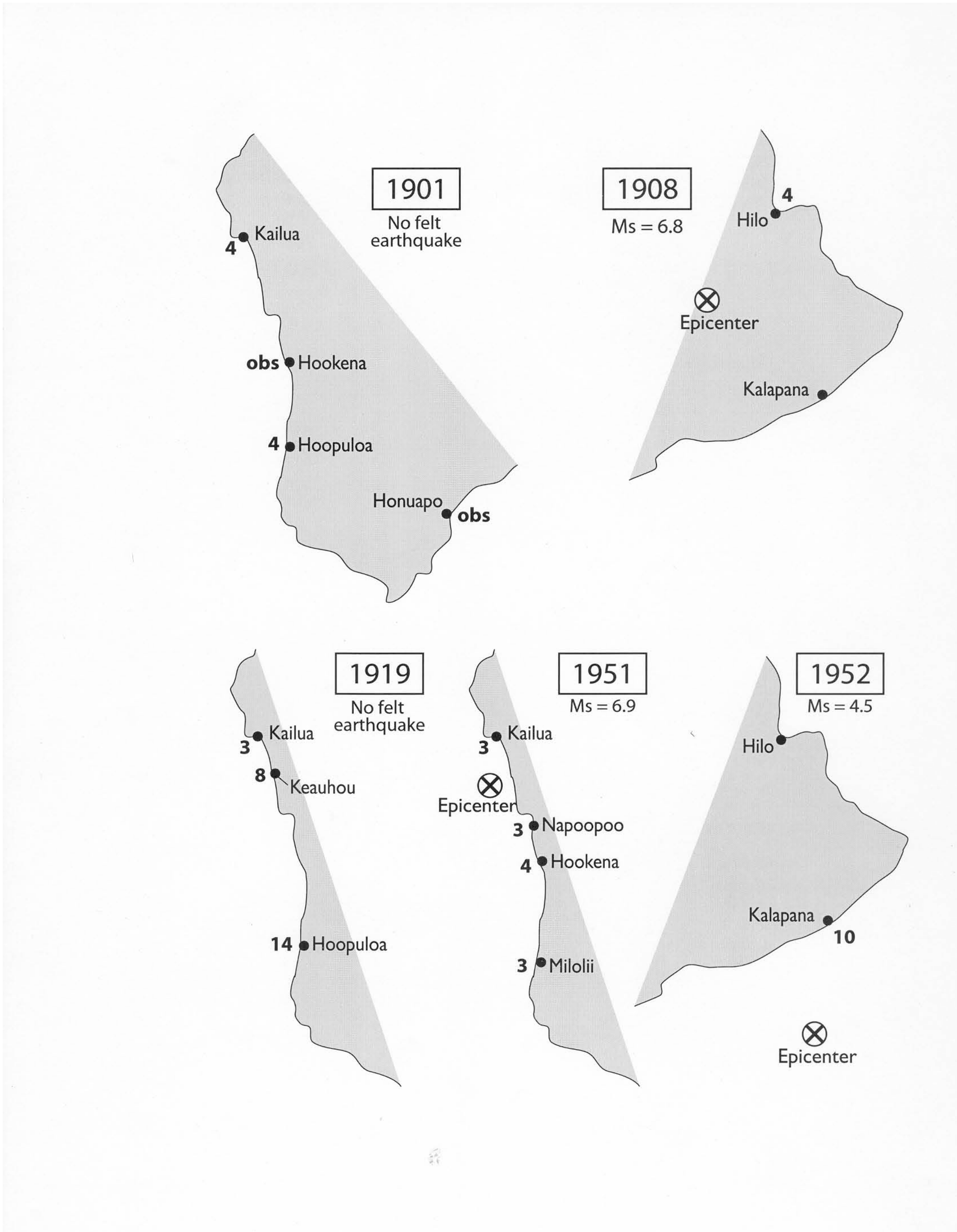

Figure 2. Earthquake epicenters, magnitudes, and run-up values in feet for other significant local tsunamis in the 20th century (Lander and Lockridge, 1989).

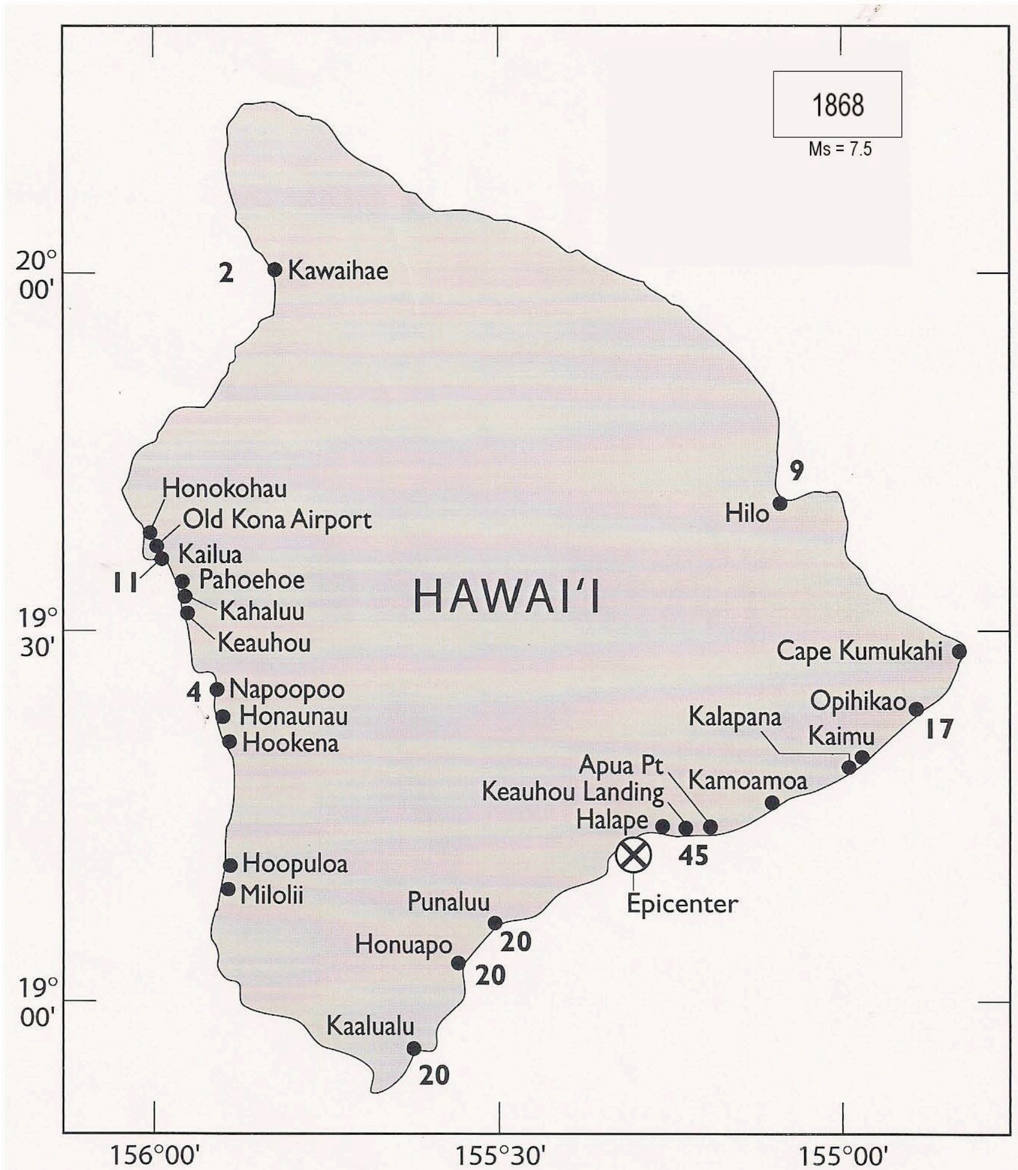

Figure 3. Earthquake epicenter, magnitude, and run-up values in feet for the 1868 tsunami (Lander and Lockridge, 1989).

In addition, there are reports of the tsunami being observed at other Big Island location. A value of 7 feet indicated for Apua may not be consistent with reports that all of the houses in the area were washed away. It should be noted that the waves could have been much larger as they swept across this extensive low-lying point of land. There may have been no topographic features in the area capable of providing reliable evidence of higher run-ups.

4. INSTRUMENTAL DEFICIENCIES

Deep ocean instrumentation can provide the data necessary for warnings of ocean-wide tsunamis. However, by the time signals are recorded on these instruments, destruction in the near field of the tsunami source area will already have occurred. As the historical data indicates, warnings for local earthquakes cannot be based only on earthquake magnitudes. Water level data is also required. In this regard Cellular Run-up Detectors (CRD's) were installed at various locations on the Big Island (Figure 4) to complement existing sea level gauges (Walker 2002). Mounted in inundation areas at elevations of about 10 feet above sea level on Civil Defense siren poles, these solar powered water sensors send cell phone transmitted signals of flooding to the Pacific Tsunami Warning Center (PTWC) in about 30 seconds. Signals received from these sensors subsequent to a nearby earthquake would confirm that a tsunami has in fact been generated. Although official warnings may not be possible for shorelines first struck by the tsunami, nearby and more distant coastal areas may receive timely official warnings. These land-based low cost instruments are relatively easy to install and maintain. In areas with little or poor cell phone coverage, sensors using satellite transmitters (SRD's) have been developed, tested, and installed (Walker 2010). These units are hidden in artificial rocks and are powered by eight primary lithium ion D-cells. Thus far these units have operated in development testing and in the field without maintenance for more than one year along the seismically active eastern shoreline of the Big Island (Figure 4). Support is needed to maintain these instruments and to install some additional units to provide warnings for beach parks and campsites along the Big Island's eastern shore and to provide for more advanced warnings for other coastal areas of the State – especially for other areas of the Big Island (e.g., Kona and Hilo) and for Maui. Also, siren systems triggered by satellite transmissions are needed for remote campgrounds in Hawaii Volcanoes National Park and possibly for other remote campsites throughout the State.

5. WARNING DEFICIENCIES

At present there is no difference between the siren tones for local tsunamis and distant tsunamis. People are advised to turn on their radios or televisions for further information when they hear a warning siren. Question: "Should people in inundation zones who hear a siren take time to find a radio or television to determine what they should do, or should they evacuate first and then find a radio or television?" Answer: "They should first get out of the danger zone." Without a difference in siren tones, they have no way of knowing whether the siren is for a local tsunami or for an ocean-wide tsunami. Evacuating first is the safest assumption.

Another potential deficiency in warning procedures may be unnecessary and timely human interventions. The recording of a local earthquake followed by the detection of flooding transmitted by ocean or land based instrumentation should be sufficient to automatically trigger tsunami warnings that automatically expand or terminate as more sensors are monitored. The speed of warnings is critically important for local tsunamis – especially for the Big Island and Maui. Again, a repeat of the 1868 and 1975 tsunamis (powerful 45 and 47 foot high water levels on land, respectively) occurring on the Kona (west) Coast rather than on the Puna and Kau (east) coasts, would be devastating along the western shores of the Big Island, the western and southern shores of Maui, and the southern coastlines of Oahu, and possibly Kauai. Even highly localized tsunamis for which no warning is given can undermine confidence in warning agencies. This loss of credibility could result in substantial additional fatalities even if a warning is issued for a subsequent larger tsunami.

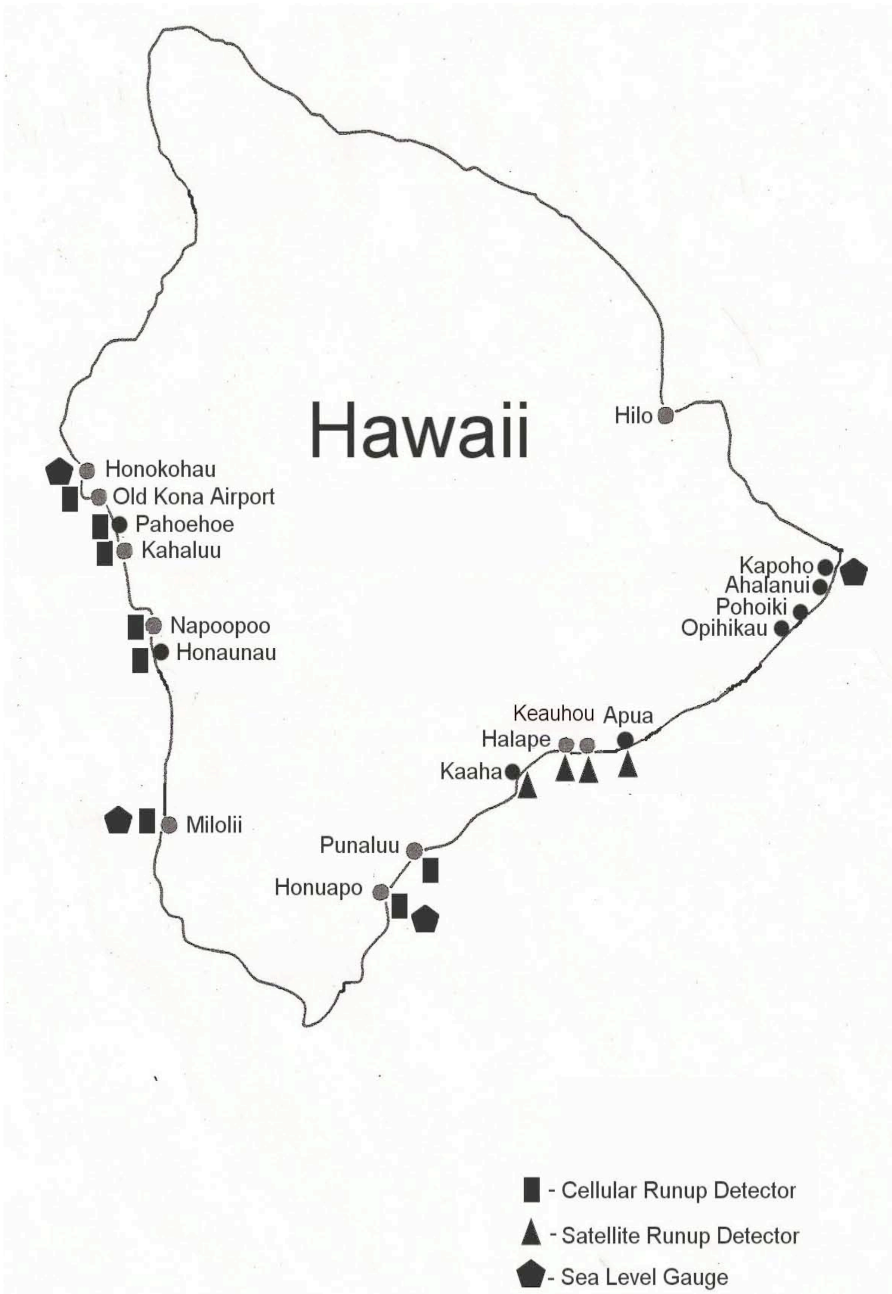

Figure 4. Site locations for Cellular Run-up Detectors (CRD's), Satellite Run-up Detectors (SRD's), and nearby Pacific Tsunami Warning Center sea level gauges.

6. CONCLUSIONS

Because of existing potential deficiencies in education, instrumentation, and warnings, locally generated tsunamis may be a far greater risk to loss of life in the Hawaiian Islands than Pacific-wide tsunamis. These deficiencies should be addressed and corrected by public officials, educators, residents, and visitor industry personnel prior to, rather than after, the next locally generated tsunami.

Acknowledgements

The Cellular Run-up Detectors were developed, tested, and installed several years ago with funding given to the State of Hawaii's Civil Defense Agency by the National Tsunami Hazards Mitigation Program. The Pacific Tsunami Warning Center (PTWC) as part of their monitoring network now maintains these instruments. The development, testing, and installation of the Satellite Run-up Detectors were similarly funded. However, at present, support is unavailable for the maintenance of these critical units and for their integration into PTWC's monitoring network. Hawaii Volcanoes National Park and the Hawaiian Volcano Observatory provided assistance with the installation of these units on the Big Island.

REFERENCES

Cox, D.C., and Morgan, J. (1977). Local Tsunamis and Possible Local Tsunamis in Hawaii, HIG-77-14, Hawaii Inst. of Geophysics, 118pp.

Lander, J.F., and Lockridge, P.A. (1989). United States Tsunamis (including United States possessions) 1690-1988, National Geophysical Data Center, Publication 41-2, Boulder, Colorado, 265 pp.

Walker, D.A. (2000). Twentieth Century Ms and Mw Values as Tsunamigenic Indicators for Hawaii, Sci. Tsunami Haz., 1&2, 69 - 76.

Walker, D.A., and Cessaro, R.K. (2002). Locally Generated Tsunamis in Hawaii: A Low Cost, Real Time Warning System with World Wide Applications, Sci. Tsunami Haz. 20, 4, 177 – 182.

Walker, D.A. (2010). Satellite Transmitted Flood Alerts to Reduce Fatalities and Injuries on the Island of Hawaii Associated with Locally Generated Tsunamis, Sci. Tsunami Haz. 29, 1, 1 – 10.

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TSUNAMI CATALOG AND VULNERABILITY OF MARTINIQUE (LESSER ANTILLES, FRANCE)

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ABSTRACT

In addition to meteorological hazards (hurricanes, heavy rainfalls, long-period swells, etc.), the Caribbean Islands are vulnerable to geological hazards such as earthquakes, landslides and volcanic eruptions caused by the complex tectonic activity and interactions in the region. Such events have generated frequently local or regional tsunamis, which often have affected the island of Martinique in the French West Indies. Over the past centuries, the island has been struck by destructive waves associated with local or regional events - such as those associated with the eruption of the Saint-Vincent volcano in 1902 and by tsunamis of distant origin as that generated by the 1755 Lisbon earthquake.

The present study includes a classification of tsunamis that have affected Martinique since its discovery in 1502. It is based on international tsunami catalogs, historical accounts, and previous scientific studies and identifies tsunamigenic areas that could potentially generate destructive waves that could impact specific coastal areas of Martinique Island. The potential threat from tsunamis has been greatly increasing because of rapid urban expansion of coastal areas and development of tourism on the island.

Key- words: Tsunami, earthquakes, landslides, volcanic eruptions, Martinique, Caribbean, risk, hazards, vulnerability.

1. INTRODUCTION

1.1 Generalities

The Caribbean region, with its complex geodynamic and climatic context, is particularly prone to tsunami generation (O'Loughlin and Lander, 2003). The Caribbean tectonic plate is bordered to the north and south by numerous strike-slip faults where major earthquakes can occur. The magnitude Mw 7.0 Haiti earthquake of 12 January 2010 (Leroy et al, 2010) is a recent example of a destructive seismic event - similar to others that has occurred in the past along the Northern Caribbean margin (Pararas-Carayannis, 2010). Also, destructive earthquakes occur along the eastern boundary of the Caribbean plate (Germa, 2008) where there is active subduction with the North American plate at a rate of 2 cm/year, as well as along the western boundary – particularly along the Pacific coast - which is characterized by a higher rate of subduction of 9 cm/year (Grindlay et al, 2005).

Martinique is located within the Lesser Antilles islands group along the Atlantic subduction zone. This group is distinguished from the Greater Antilles (to the North), which is separated by the Anegada Strait between Puerto-Rico and the Virgin Islands (Fig. 1). The volcanic arc, which constitutes the Lesser Antilles, is over 850 km long and 450 km wide (Zahibo and Pelinovsky, 2001). Its formation results from tectonic processes associated to the subduction of the Atlantic plate beneath the Caribbean plate in two distinct phases: the first from the Eocene to the Oligocene (50 to 20 Myr BP), and the second during the Miocene (10 Myr BP) (OVSG-IPGP, 2005). Martinique and the neighbouring island of Sainte Lucie are located near the center of the volcanic arc and were created by the volcanic activity of these two phases (MacDonald et al, 2000). Numerous volcanoes are still active in the Lesser Antilles and some of them had eruptions and associated collateral events which generated tsunamis: Montserrat (Herd et al, 2005), Guadeloupe (Feuillard et al, 1983), Saint-Vincent (Le Friant et al, 2009), Mount Pelée on Martinique (Pararas-Carayannis, 2006; Leone and Lesales, 2008), Kitts in Saba, Liamiuga in St Kitts and Nevis and the Kick'em Jenny submarine volcano (Smith and Shepherd, 1993, Pararas-Carayannis, 2006) (Fig. 1).

Associated with such geological activity, the tropical climate of the archipelago results in violent storms, frequently causing coastal landslides that can generate local tsunamis. Also, along with the geological hazards associated with earthquakes, volcanic eruptions represent a real tsunamigenic threat. The mechanisms of tsunami generation from volcanic eruptions, debris avalanches, pyroclastic flows and collateral flank failure mechanisms in the Lesser Antilles - and Martinique in particular - have been examined and evaluated (Pararas-Carayannis, 2006). Also several historical catalogs have been compiled for tsunamis and tsunami-like events that have occurred in the region (Lander, 1997; Lander et al, 2002; O'Loughlin and Lander, 2003; Lander et al, 2003), with a special focus on the Lesser Antilles (Zahibo and Pelinovsky, 2001, Saffache, 2005b). However, there seems to have been no specific study of the tsunami hazard in Martinique itself. Thus, the primary objective of the present study is to review the pre-existing databases on events that impacted Martinique and to propose a new catalog, composed only of the well-known and documented tsunamis. Furthermore, the present study reviews briefly the island's vulnerability, with a special emphasis on La Trinité Bay. However, in order to understand the historical accounts of tsunamis that have affected Martinique in the past, we need to review its geographical and geo-tectonic setting.

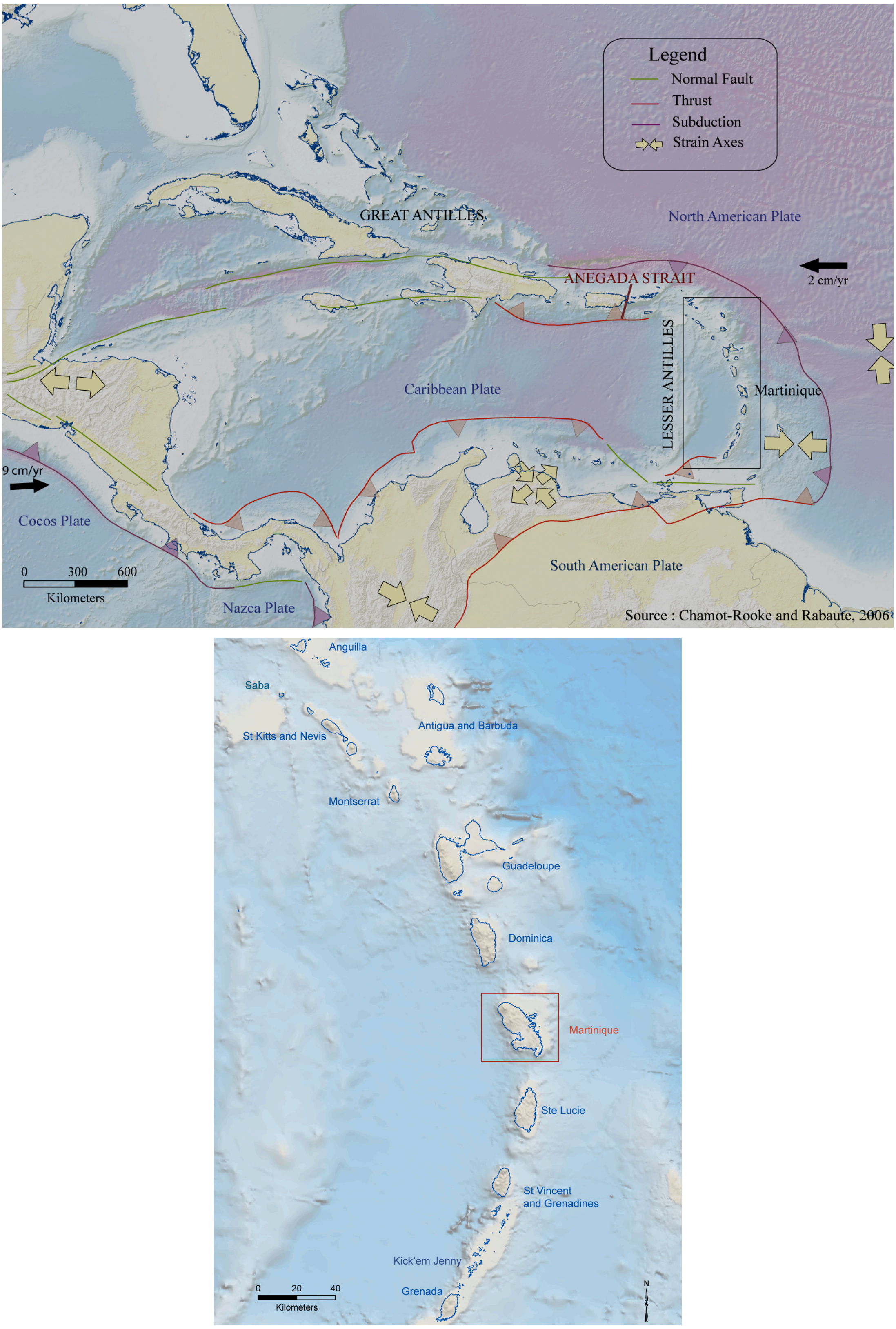

Fig. 1: Geodynamical context of the Antilles (tectonic scheme from Chamot-Rooke and Rabaute, 2006; bathymetric data from GEBCO (IOC, IHO and BODC, 2003) and geographic location of Martinique Island within the Lesser Antilles.

1.2 Martinique

1.2.1 Geography

Martinique is a volcanic island located upon a subduction zone and which has Mount Pelée, an active volcano (Westercamp and Tazzieff, 1980; MacDonald et al, 2000, Pararas-Carayannis, 2006). The island exhibits various reliefs ranging from swamplands near the sea to volcanoes. Mount Pelée reaches 1398 m in height and three other peaks exceed 1000 m. If we draw a line between Fort-de-France (west coast) and La Trinité (east coast) The highest reliefs are found in the northern part of the island (Fig. 2). The southern part of the island is mainly constituted of low hills and swamplands. The littoral exhibits alternation between steep cliffs in the north and sand beaches (Les Anses d'Arlet, la Caravelle, St Anne, etc.) and mangrove forests (Le Lamentin for example) in the south (Saffache, 2005a).

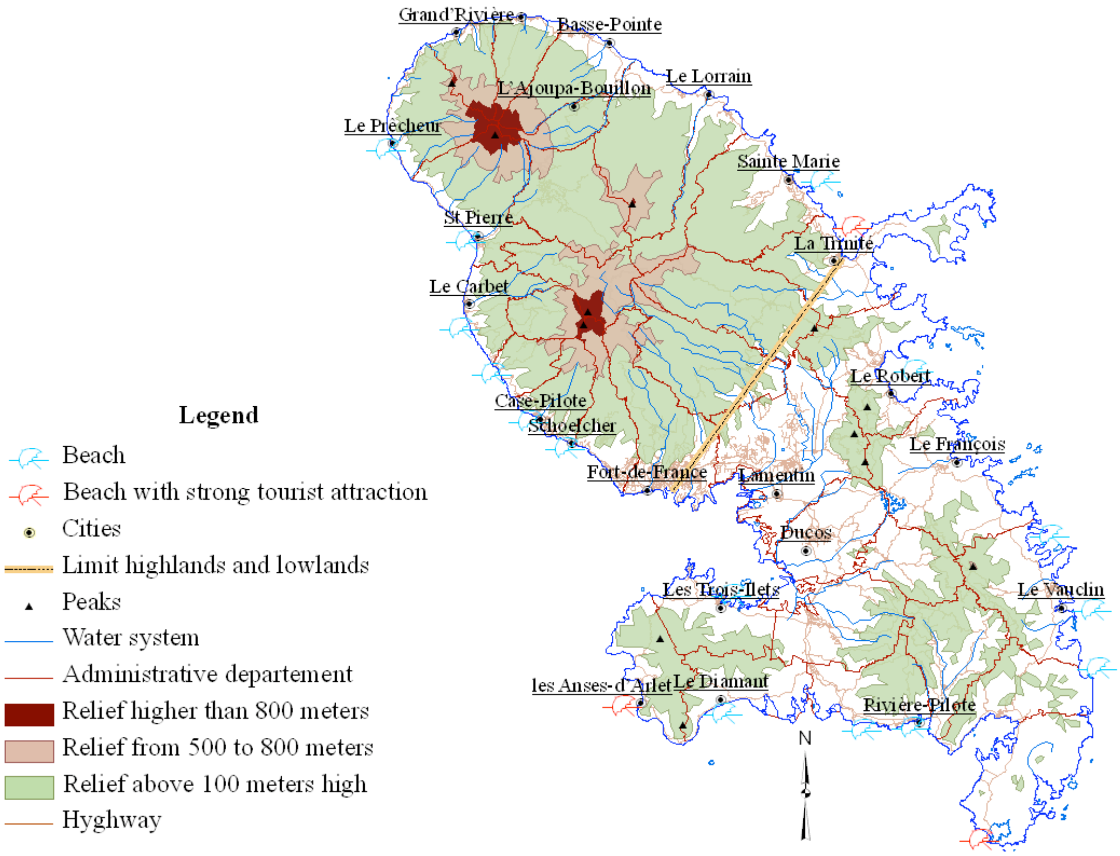

Fig. 2: Geography of Martinique Island. The yellow line indicates the limit between the hilly landscape (northern part) and the plain (southern part).

The great 2004 Sumatra event demonstrated that a steep continental slope and coastal features such as coral reefs, mangrove forests and lagoons help reduce the degree of a tsunami's impact by slowing down its propagation and absorbing or reflecting part of its energy (Kathiresan and Rajendran, 2005, Kunkel et al, 2006). However other coastal features can increase a tsunami's adverse impact by resonance amplification or by inducing strong currents generation usually in bays or harbours (Sahal et al, 2009, Roger et al, 2010a). Beaches with gentler slopes also tend to amplify the wave heights due to a shoaling effect.

1.2.2 Historical Development of Martinique

In pre-colonial times, the island of Martinique was subjected to two successive waves of human settlement, first in the 4th century BC (with the arrival of the Arrawak people) and then in the 13th century AD (with the Caribbean people) (Lalung, 1948). Following the arrival of the Europeans in 1502 and the settlement of French colonists in 1635 (Lambolez, 1905; Chauleau, 1993; Charbit, 2006), the island was integrated into the French colonial empire and its population increased tenfold thanks to triangular trade (Charbit, 2006; Clément, 2009). Nowadays, Martinique has a population of over 400.000 (INSEE, 2009), which is mostly concentrated in the conurbation of Fort-de-France, Schoelcher and Le Lamentin (Calmont and Vassoigne, 1999), as well as in a few other towns deemed attractive, such as Sainte Marie and Le Robert. These five agglomerations concentrate nearly fifty percent of the population of the island (INSEE, 2009).

The bay of Fort-de-France concentrates poles of economic activity, political and decision-making centres and luxury commuter towns (Schoelcher) (Calmont and Vassoigne, 1999). Some southern cities benefit from the attractiveness of heliotropical tourism, which attracts about 700.000 visitors every year (Schleupner, 2007). The increase in tourist influx along the coastal zones and the popularity of seaside activities require that a study of the tsunami hazard has a high priority and must be undertaken, thus we begun by documenting historical tsunami data into the following catalogue.

2. TSUNAMI CATALOG

2.1 Data

The conclusions of the present study rely mainly on the historical database of tsunamis that have been recorded in Martinique and the neighbouring islands. Material used for this database includes original historical documents (testimonies, letters, etc.), which are considered as primary sources, whereas secondary sources are derived from recent studies, tsunami catalogues, simulation results, etc.

2.1.1 The Catalog

Existing historical tsunami catalogs provide a general idea of the tsunami hazard in the Antilles. These catalogs are archived and updated by different government organizations such as NOAA (U.S. National Oceanic and Atmospheric Administration) and the Tsunami Laboratory at Novosibirsk (Russia), or compiled by research of historical records (Lander et al, 2002; O'Loughlin and Lander, 2003; Saffache et al, 2003). Although derived from the same original historical data – which includes dates, source areas, wave heights or other miscellaneous information - these catalogs produce different results because of differences in the methodology of compilation. In our study, all of these existing catalogs have been analysed and inter-correlated in order to produce a single, better-documented tsunami catalog.

The worldwide tsunami catalog from NOAA provided at the following internet site (http://www.ngdc.noaa.gov/hazard/tsu_db.shtml) and that of the Tsunami Laboratory of Novosibirsk (http://tsun.sscc.ru/proj.htm) do not index all the tsunamis that have occurred but only the most significant.

Catalogs by O'Loughlin and Lander (2003) and Lander et al. (1997, 2002, 2003) list tsunamis for the entire Caribbean. The catalog by Saffache et al. (2003) includes earthquakes and abnormal sea level rises that have occurred in the French Antilles (Guadeloupe and Martinique Islands). Although helpful, these catalogs have a broader regional focus and do not provide sufficient information on specific events to be adequate for a risk assessment study of Martinique. To properly assess the risk, it is necessary to review carefully local historical archives as indicated subsequently.

2.1.2 Historical Documents

Analysis of historical documents (Du Tertre, 1668, Boyer-Peyreleau, 1823; Hess, 1902; Lacroix, 1904; Lambolez, 1905), and particularly of various testimonies recorded in Martinique archives (http://www.manioc.org; http://gallica.bnf.fr), highlights details that were omitted from earlier catalogs and scientific papers. Most of these documents report on the 1902 catastrophe on the island (Hess, 1902, Lacroix, 1904) and some provide useful information on its historical development since 1635 (Royer-Peybeleau, 1826; Du Tertre, 1668, Lambolez, 1905). Review of these accounts is helpful in documented the destructive impact of tsunamis on the local population, on the resulting losses and on reactions (Hess, 1902; Lambolez, 1905). Moreover, these accounts provide an important amount of information for specific locations, as for example the impact of the May 5, 1902 tsunami in the vicinity of the Guérin factory (Hess, 1902), as well as on records of false alarms and the lack of proper warning by public authorities during the eruption of the Mount Pelée volcano (Lambolez, 1905). However, in spite of the wealth of details that can be found in these historical sources, many more historical documents pertaining to the Martinique's maritime trade were lost due to the destruction of archives (by the 1902 eruption of Mount Pelée). Hopefully, research studies may help compensate for these gaps in knowledge.

2.1.3 Scientific Studies

Some tsunamis have been investigated in detail through scientific research. Such research provides details, which can complement the data already gathered through historical analysis. Events that have been researched include the 1755 Lisbon earthquake and tsunami (Baptista et al, 1998; Chester, 2001; Baptista et al, 2003; Barkan et al, 2009; Roger and Baptista, 2009; Roger et al., 2010a, 2010b), the 1761 earthquake (Baptista et al, 2006), and those associated with recurring eruptions of the submarine volcano Kick'em Jenny (Smith and Shepherd, 1993, Pararas-Carayannis, 2006). These studies help visualise what could happen in an areas where information is lacking and help assess the generating sources and impacts of potential tsunamis.

2.2 Data Selection

2.2.1 Storm Surges

In reviewing the data in existing catalogs attention was paid in identifying and excluding storm surges that may have been wrongly listed as tsunamis. It appeared that numerous phenomena recorded as tsunamis in Martinique were in fact storm surges generated by storms and hurricanes, which are frequent in the Antilles region (Royer-Peybeleau, 1826; Lambolez, 1905; Saffache et al, 2003; O'Loughlin and Lander, 2003). Thus, seventeen events (in 1642, 1694, 09/12/1756, 08/24/1757, 08/14/1766, 09/05/1776, 10/12/1780, 08/14/1788, 09/06/1816, 10/21/1817, 09/21/1818, 07/26/1825, 09/20/1834, 09/09/1872, 09/04/1883, 08/18/1891 and 08/08/1903) had to be taken away from our catalogue because they had been generated by hurricanes. For example, Revert's account (1949) lists a total of 34 hurricanes between 1633 and 1903. Among hurricane generated storm surges, seven events (in 1642, 1694, 1756, 1757, 1766, 1780 and 1883) were particularly destructive to crops and ships. The worse of these storm surges appears to have been that of August 18, 1891, which wrecked the city of Le Lamentin (Lambolez, 1905) and caused severe damages (Revert, 1949).

2.2.2 Uncertain Tsunamis

After cross-matching data gathered from different scientific studies, some tsunamis had to be left out of the catalogue, since no evidence indicates that they actually reached Martinique. Among these, many are of volcanic origin, whether at the global scale associated with the 1883 eruption of the Kratatau volcano in Indonesian (Choi et al, 2003; Pararas-Carayannis, 2003), or at the scale of the Lesser Antilles with the low intensity eruptions of the Kick'em Jenny (Smith and Shepherd, 1993) and of the Souffrière Hills in Montserrat (1824, 1897, 1997), 320 kms away from Martinique. Finally, this selective process also excludes the formation of an 'ephemeral mud island' as the one off the coasts of Trinidad at various points in time, specifically in 1853, 1874, 1911, 1928, 1934) (O'Loughlin and Lander, 2003).

2.3 Criterion of Validity

The criterion of validity used in the present study is different from that which was used in the catalog compiled by the Laboratory of Tsunamis of Novosibirsk (www.tsun.sscc.ru/) and that used by O'Loughlin and Lander (2003). The scale of validity used for the Martinique catalog adhered to factual information as specifically as possible (Table 1). Thus, events reported as 'abnormal oscillations' by some observers (with degree 1 on the tsunami scale), were separated from tsunamis that appear more frequently in different sources (with degree 4), or from those to which all sources systematically refer (degree 5). Degree 2 corresponds both to known tsunamis, but about which it is not certain whether they impacted Martinique or not (degree 2a) and to tsunamis that may have occurred but were not recorded in Martinique, in spite of a violent earthquake or a landslide recorded by observers (2b). In other words, such events may be known to have occurred, but none of the sources available either suggests that the event in question caused a tsunami or that it affected Martinique. However, it was decided to include them in the study because it is possible that such events were not recorded on the island simply because there were no observers on site, or for other reasons (occurring at night, micro tsunami, etc.).

3. RESULTS AND DISCUSSION

3.1 Results

Overall, the present study included at least thirty-three tsunamis, of which only five did not reach Martinique according to original references. These included the 1831 Granada tsunami, the 1874 and 1880 Dominica tsunamis and the 1985 and 2004 Guadeloupe tsunamis (Table 2). In addition to date and localisation, the catalog indexes the origin, the degree of validity associated to some tsunamis, as well as some parameters such as amplitude and run up. Also, the catalog lists various other notes and information, regarding contradictions that were found in the researched sources (as for example, the 1751 tsunami). The catalog stands out due to the heterogeneity of the level of information available for each event. Thus, some tsunamis, such as those of 1755 or 1902, benefit from the abundance of details in well-documented sources and additionally conducted research. However, other events such as the 1657 and 1874 tsunamis are very poorly documented, since they occurred a long time ago (1657 was the year of the first earthquake ever perceived by the colonists, who had only just settled on the island in 1635), or because of the sparseness of damage caused by the reported 'abnormal oscillations' (as for the 1874 event).

3.1.1 Detailed Analysis

A more detailed review of the various tsunamis reported in the database has allowed the present study to group some events in accordance to their origin characteristics (seismic, volcanic), or in accordance to the propagation characteristics (local, regional, or far-field tsunamis). Of all the events indexed in the catalog, only four are absent from Figure 3 (the Hispaniola earthquake of 1751; the Lisbon earthquake of 1755; the Surinam earthquake of 1767 and the Costa Rica earthquake of 1991).

Figure 3 shows the origin of twenty-six tsunamis generated near the Lesser Antilles, of which twenty-one were observed on Martinique. These include the tsunamis which occurred on the following dates: 6 April 1690; September 1702; 6 March 1718; 30 November 1823 and 1824; 30 November 1827; 26 October 1829; 26 July 1837; 11 January 1839; 8 February 1843; 18 November 1867; 8 November 1876; 5 May 1902; 7 May 1902; 8 May 1902; 9 May 1902; 20 May 1902; 30 August 1902; 16 February 1906; 24 July 1939; 25 December 1969; 22 April 1991 and 21 November 2004.

Of the thirty-three tsunamis listed in the catalog, half are of seismic origin generated mainly around the Virgin Islands (as the 18 November 1867 event) and Guadaloupe (as the 1843, 1969, 1985 and 2004 events). Tsunamis in the region are also generated from volcanic sources and more than ten of those recorded were associated with the volcanoes of the Caribbean archipelago (i.e. Kick'em Jenny, Souffrière Saint-Vincent, Mount Pelée and the 'Souffrière' of Montserrat). Tsunamis generated by landslides in the region seem to be under-represented in the diagram – with only one event being included. However, it should be noted that landslides could be triggered by both seismic and volcanic events. It is therefore necessary to take into account that 13% of all tsunamis are of multiple origins (Fig. 4).

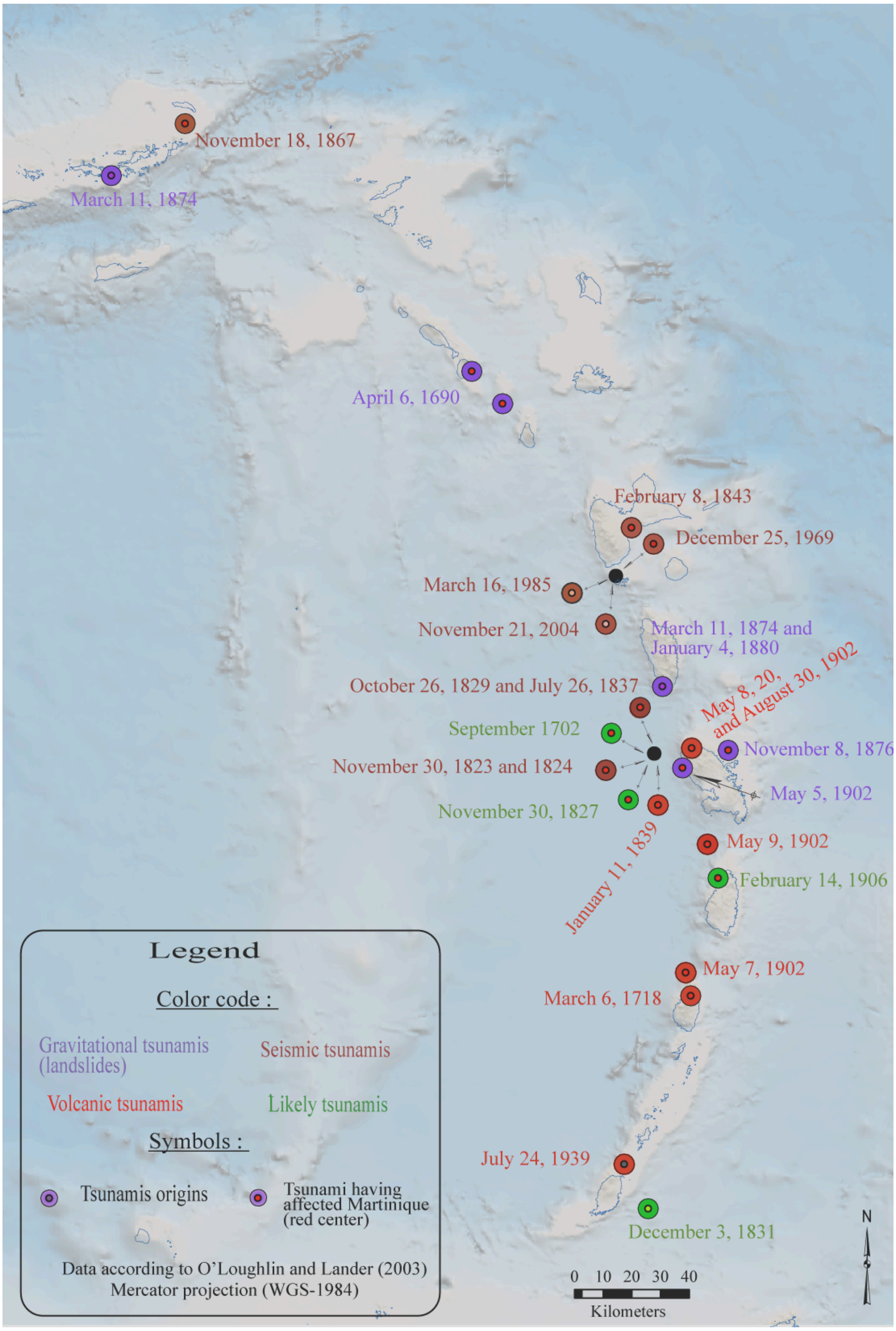

Fig. 3: Sources of tsunamis in the arc of the Lesser Antilles.

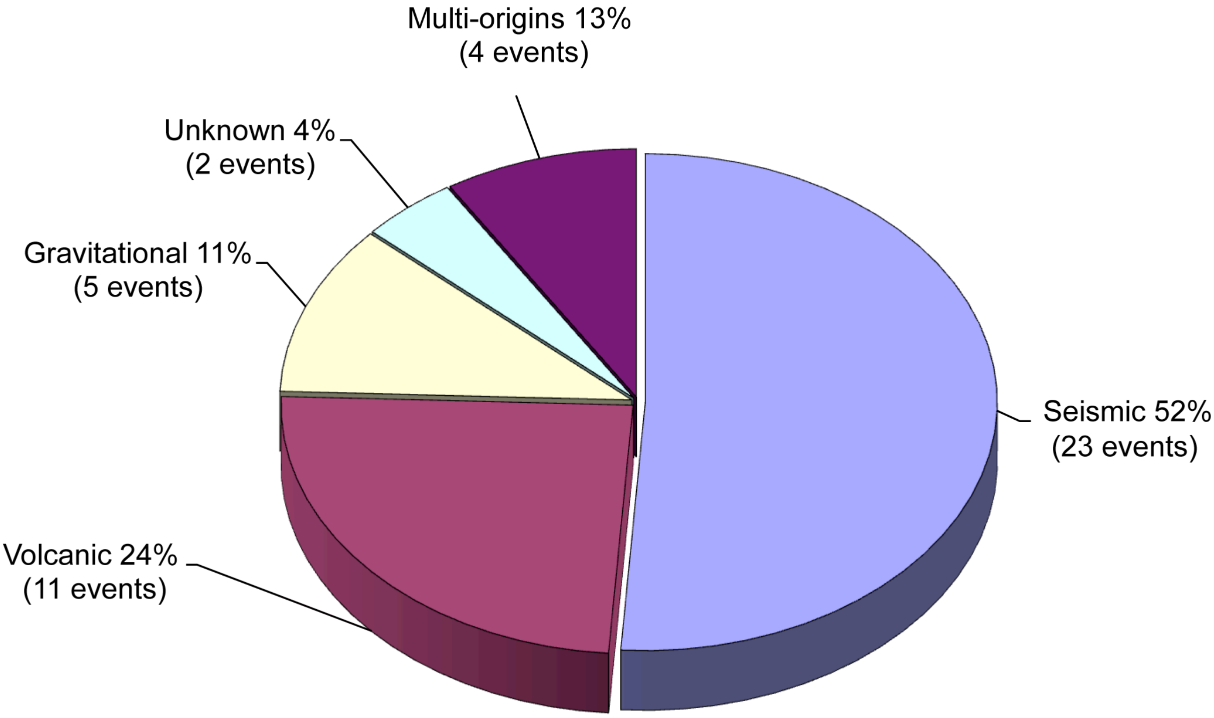

Fig. 4. Distribution of tsunamis according to origin.

Lastly, two events (1829 and 1837) are of unknown origin. This lack of information is attributed to the inadequacy of historical accounts and the inability to determine the tsunami source mechanism. Thus, these two events may have been caused either by a passing hurricane or by an earthquake. It should be noted that in those times, the violence of hurricanes could foster 'jolts' that were thought to be earthquakes. However, this confusion regarding the origin of tsunamis only affects 6% of the total number (i.e. the two events).

The second diagram (Fig. 5) illustrates the prevalence of local tsunamis (comprising of 49% of the total) over regional tsunamis (comprising of 36%). Martinique is therefore mostly subjected to tsunamis originating near neighbouring islands such as Saint Vincent, or Guadaloupe. The only tsunami of distant origin is represented by the single Lisbon tsunami of 1755.

Although many tsunamis have been observed on the coasts of Martinique, few have been located with geographic accuracy. Thus, in the majority of cases, the original references mention Martinique without attempting to classify the coasts on the island that were affected by tsunami wave action.

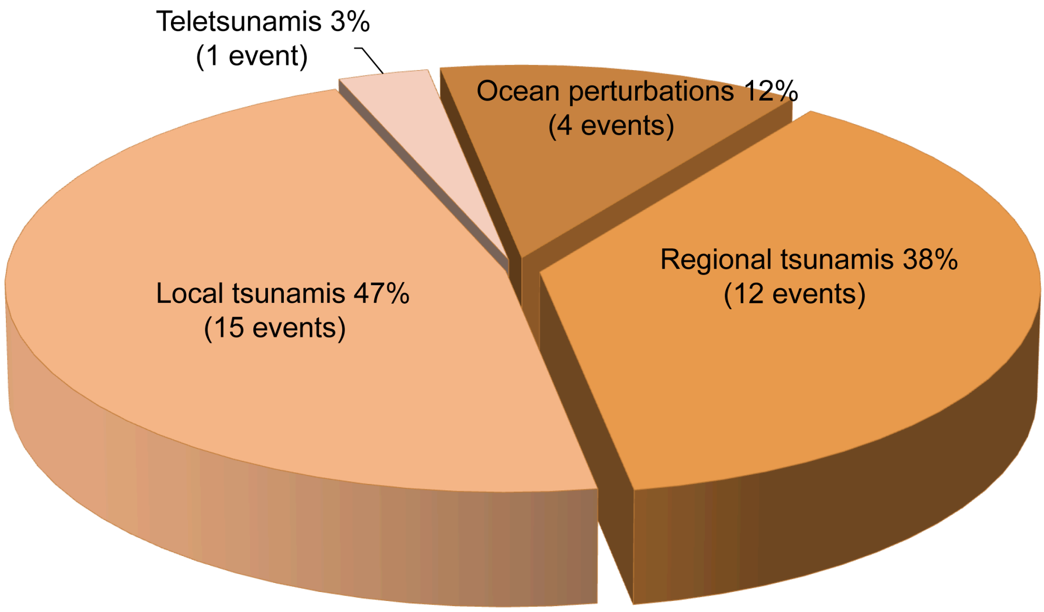

Fig. 5. Classification of tsunamis according to impacted zone.

The map (Fig. 6) shows towns in Martinique that were affected by tsunamis, according to observations found in different reference sources. These indicate which of the coastal areas are particularly vulnerable to tsunami impact. Among them, is the coastal area along the flanks of the volcano stretching from Macouba to The Carbet via Le Précheur (Northwest), but also the coast stretching from Case-Pilote to Trois-Ilets, along the bay of Fort-de-France and, finally, the tourist area stretching from Sainte Marie to the Robert, near the presqu'isle of the Caravelle (La Trinité). Other sites could have been impacted but lack of educated population to document what happened limited most of the observations in major towns.

Figure 7 helps visualise the cities that are most vulnerable to the tsunami hazard in Martinique. Saint Pierre (27%) and La Trinité (26%) are the towns that would be most affected. The 27% mentioned for St Pierre indicate that 27% of the listed tsunamis in the catalog affected St Pierre. The destruction of the capital city of Saint Pierre during the cataclysmic eruption of 1902 was a definitive blow since it resulted in the deaths of the city's population (28.000 people), but also its status as capital at the benefit of Fort-de-France. La Trinité presents an original profile in so far as it is affected to locally produced tsunamis (such as those of 1876, 1902, etc.), but also to tsunamis of distant origin (such as that of 1755), which makes this location as the most sensitive on the island in terms of exposure to the tsunami hazard. Some towns in the North of the island (Basse-Pointe, Macouba, le Précheur) are also exposed to 'oscillations', as it was shown in 1902, when Mount Pelée was particularly active.

However, it should be noted that an overwhelming majority of the population on Martinique is concentrated in coastal towns and areas, which are particularly vulnerable to tsunamis (Goiffon, 2003).

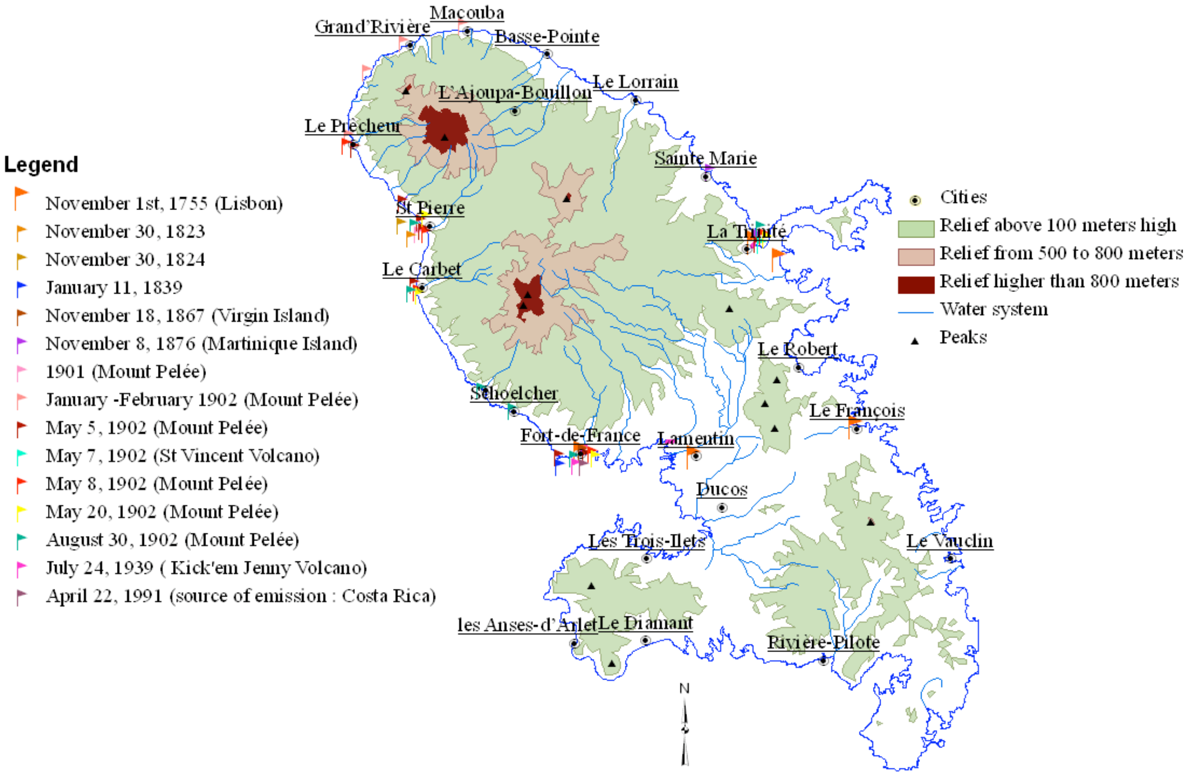

Fig. 6. Cities in Martinique impacted by past tsunamis.

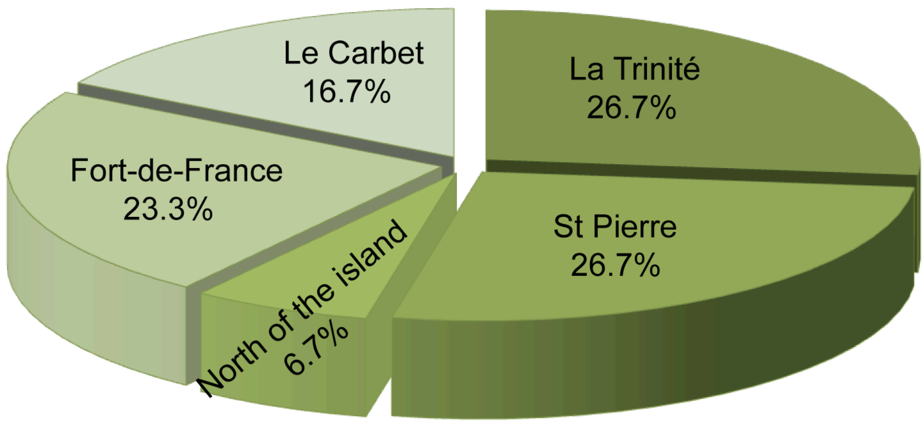

Fig. 7. Towns in Martinique affected by tsunami hazard.

3.1.2 Major Events

1755 November 1

The tsunami of 1 November 1755 is the only event of distant origin ever recorded in Martinique. The 1755 earthquake and tsunami were extremely destructive in Lisbon. The tsunami was recorded in Madera and in Morocco (Barkan et al, 2009) and reached the American continent and the Caribbean region (Roger et al, 2010a, 2010b, 2010c). The first of its waves reached the Island of Martinique in the afternoon and its subsequent effects lasted for about four hours (from 14:00 hrs to 18:00 hrs). This gap is consistent with the results obtained by numerical simulations, since the tsunami's origin time was at about 09:30 on November 1 and took over 7 hours to reach the coasts of Martinique (Roger et al, 2010). The Presqu'isle of Caravelle and the small town of La Trinité (west coast) were the first ones affected. Subsequent tsunami wave refraction around the island was observed as in the Baleares during the tsunami of Zemmouri in 2003 (Alasset et al, 2006) - and the bay of Fort-de-France was flooded. This phenomenon of wrap around is well described by Yeh et al. (1994). At the same time there was a rise in sea level along the coasts, the level of some rivers. There was sharp increase in sea level at Lamentin (Fort Royal) and Epinette (La Trinité), leading to further inland inundation (Lambolez, 1905; Baptista et al, 1998; Baptista et al, 2003; Lander et al, 2003; Roger et al, 2010b).

The Tsunamis of 1902

Various tsunamis that occurred in 1902 (on May 5th, 7th, 8th, 20th and 30th) affected extensive areas stretching from Grand-Rivière, in the North of the island, to La Trinité, via the area of Le Prêcheur, le Carbet–Case Pilote and the Trois Ilets. Fig 6 shows the various areas that were flooded. Inspite of the diversity of tsunami origins, the waves exhibited high concentrations at the same localities. Thus, the tsunami generated by the May 5th lahar flow impacted St-Pierre, the Carbet, Fort-de-France and La Trinité, as was the tsunami generated by the eruption of La Souffrière on the Island of St-Vincent, although with lesser intensity.

3.2 Discussion

Documenting through a regional study - within the framework of the Caribbean - the tsunamis that affected the island of Martinique allows to identify as well coastlines vulnerable to the tsunami hazard. However, the list of events described above, just like the catalog and the maps, show that tsunamis are potentially destructive in this area of the Caribbean, where high-magnitude earthquakes (in 1690, 1751, 1831, 1843, 1867, 1969 and 1991) and volcanic eruptions (in 1718, 1880, 1902 and 1939) are frequent and can also result in landslides (as in 1690, 1718, 1876, 1880, 1902 and 1985). Tsunamis that can affect Martinique can be generated from diverse local as well as distant sources. Mainly tsunamis that can affect Martinique tend to be of local origin, thus there is not sufficient time to issue a tsunami warning. Earthquakes generate the majority of tsunamis, although many are caused by gravitational flank collapses as well as lahars and landslides. Although the sources of many tsunamis have been identified as located in many of the islands of the Caribbean arc, Martinique on its own, principally because of the re-awakening of the Mount Pelée volcano has generated over eight local tsunamis within a year (1902), of which two were registered as 'abnormal oscillations'.

From a spatial point of view, according to Figure 8, we can say that three towns are particularly vulnerable: Fort-de-France (the political and economical capital of the island), La Trinité and Saint-Pierre. Presently, a tsunami generated by a new eruption of Mount Pelée would potentially affect the towns of Le Précheur, Grand-Rivière, Macouba, and Saint-Pierre, which have a total population of 8,360 inhabitants (plus the tourists), i.e. 2% of the population of the island (INSEE, 2009). On the contrary, a tsunami like that of 1755 would affect the towns of Fort-de-France, Le Lamentin, La Trinité and Le François, i.e., where over 40% of the population (162,131 inhabitants) are located (INSEE, 2009). Relatively, only the southern and the west coasts of the island seems to be protected from tsunamis because of the presence of natural and residual coral reef barriers.

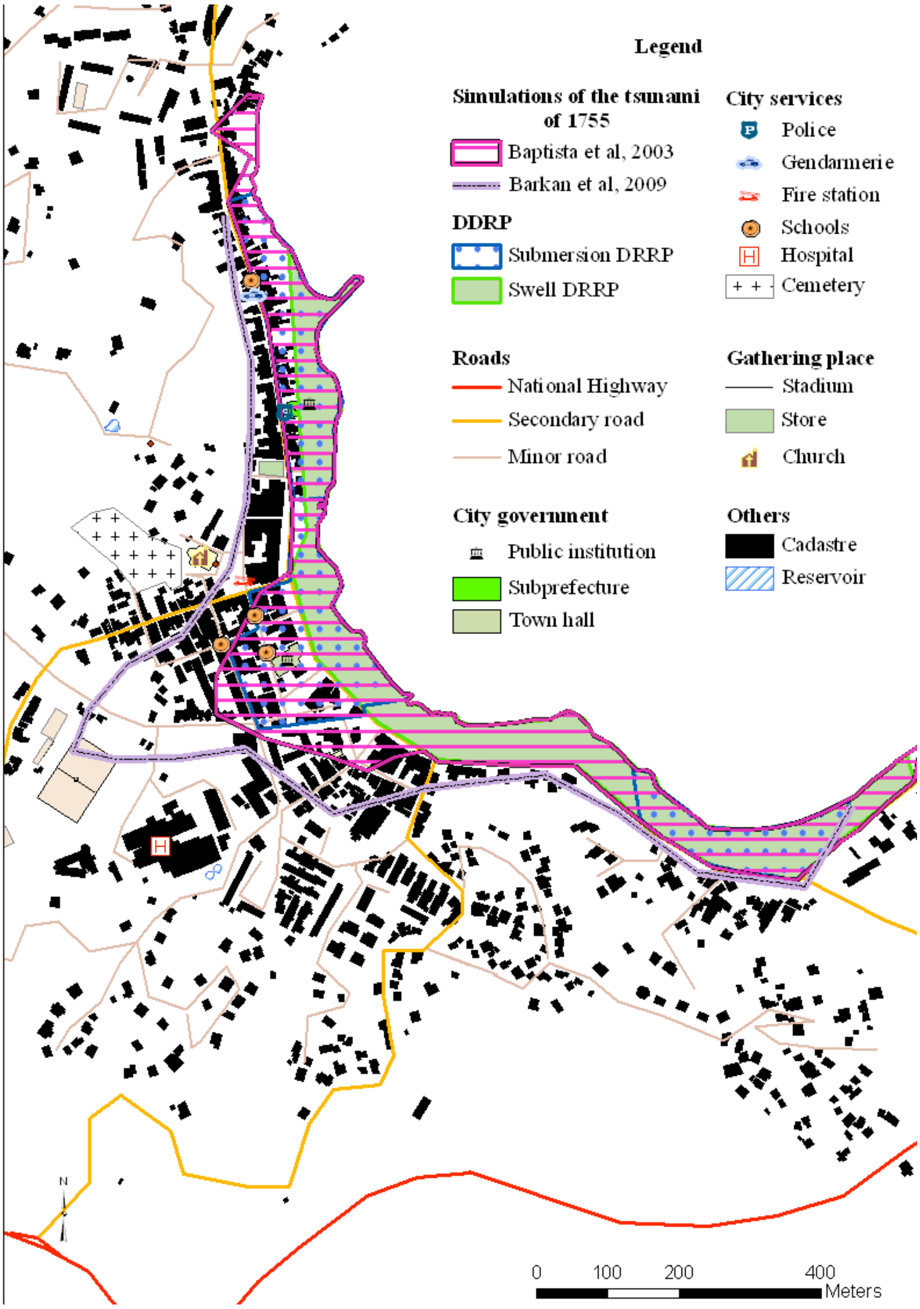

Fig. 8: Vulnerability of La Trinité: local submersion hazard limit and proposition of extension according to tsunami simulation. All the strategic sites and roads are indicated.

3.2.1 Uncertain Tsunamis

The detailed catalog (chart) also lists tsunamis, which cannot be documented with certainty. The following is a discussion of seven such events.

The first is a set of two tsunamis that appear in all written testimonies with only date, time and origin. Archived exactly one year apart, respectively on 30 November 1823 and 30 November1824, both of these events purportedly caused 'damages in the harbour' on those two dates - seemingly the Saint-Pierre harbour. Moreover, both events appear to have occurred following periods of high temperatures that were accompanied by torrential rains (Mallet, 1852, 1853, 1854). These similarities lead to two distinct hypotheses. It is entirely possible that these two purported tsunamis were in fact one single event, wrongly reported as separate events due to transcription of calendar mistakes. Alternatively, it is somewhat possible that the two events resulted from earthquakes that occurred exactly one-year apart. However it is highly unlikely that such coincidence in the day and month actually occurred. Finally, it is also entirely possible that the reported tsunamis were storm surges generated by hurricanes.

The tsunami of 8 November 1876 is reported in a letter that is referred in Lambolez (1905). This letter recollects 'waves' and 'bellowing swell' between the Presqu'isle of the Caravelle and Sainte Marie, stating that 'often to occur here'. This purported tsunami seems to have originated from a submarine landslide if we take its description into account. However, there is a significant margin of error when dealing with historical sources that contain vague descriptions. Significantly, this event is not listed in any of the other reference sources.

Finally, insufficient information is provided regarding the remaining uncertain tsunamis as to their geographic origin location, so that these cannot be really considered as real events by the present study. In fact, many of these events are referred to in an unspecific way and may have been induced by landslides, earthquakes, and storms - as was the case for the 3 June 1718 or 2 August 1837 events. Other tsunami events listed as 'abnormal oscillations' (i.e. the 1657 event) make it even harder to assess the validity of their tsunamigenic nature. All of these tsunami events, because of vagueness of the information, are associated with a degree of validity (see Table 1) that allows us to visualise them quickly and dissociate them from those that are better documented as certain.

Overall, only the tsunamis that caused serious and significant damages, both in material and in human lives, are verifiable with any kind of factual certainty in assessing Martinique's potential tsunami hazard. These major events have become the subject of numerous studies, which further permit the assessment of tsunami risk for other coastal areas of the island. Unfortunately some tsunamis, although destructive, are harder to document adequately because of lack of interest. For example the 1761 tele-tsunami, associated through its origin with the Lisbon earthquake of 1761, illustrates the difficulties in the compiling of historical catalogs, since good registration of events depends on the validity of historical reports and of various documents. In the case of 1761, only a few specialists (Zahibo and Pelinovski, 2001; Baptista et al, 2006) have studied this tele-tsunami, and the results do not allow us to integrate this event in the study of tsunamis in Martinique, due to the lack of proper historical accounts, regarding the impact on the island.

3.2.2 Uncertainty of Sources

In view of the above listed uncertainties, the results need to be evaluated critically, since they rely partly on sources that may depend on author's subjectivity.

Historical sources must be used with caution as they may introduce erroneous data, due to inadequate understanding by the reporting past scientists and observers. Thus, of the many recorded 'raz-de-marées' (common French for tsunami), only a few are actually tsunamis. Let us note that the observation and localisation of tsunamis depends mainly on human presence along impacted areas and that at coastal areas with low population density significant tsunami impact may not have been properly reported. Furthermore many of the names of localities on the island have changed through centuries, so it is not always easy to locate precisely the extent of the flooded zones (i.e. 'as far as the stone bridge on river Roxelane', (Lambolez, 1905)). The use of secondary sources is also a factor of uncertainty, as the problem raised by O'Loughlin and Lander (2003) shows, when they explain the existence of different dates for a single event due to calendar mistakes. Finally, the absence of information in the archives does not necessarily mean an absence of an actual tsunami, since literate men – who were scarce in this colonial island, only transcribed these events.

Additionally, the results of tsunami modelling depend on the data gathered by the research specialists; therefore, as with any other scientific endeavour, they come with a margin of error or uncertainty. These models can overestimate or underestimate tsunami amplitudes. Only a study of sedimentary deposits (as in Morton et al, 2006) can confirm whether the listed tsunamis actually occurred. However, subsequent development of the heavily populated coastal areas makes such investigations almost impossible, unless construction works expose accidentally tsunami deposits (Nicolae-Lerma, pers. comm., 2010).

Lastly, the data gathered by the mareographs of SHOM may have been useful, but no access was possible. Moreover, most of the mareographic equipment has only been installed recently (October 2005) on Martinique (Créach, pers. Comm., 2010).

Up to now the present study concentrated on the tsunami hazard in Martinique but without including a social-economic dimension: the vulnerability. Only the coastal town of La Trinité was chosen for further assessment of the tsunami hazard.

3.2.3 Vulnerability of La Trinité

The town of La Trinité is located on the Atlantic coast of the island. It has a population of 13,582, distributed over an area of 45.8 km². Although the town concentrates only 3.5% of the Martinique's total population, it is the designated local administrative centre ('sous-préfecture' - district level) and is host to the only hospital on this part of the coast. Moreover, La Trinité has been affected in the past by eight tsunamis of various origins (seismic, volcanic, landslides) and generating sources (Lisbon, Saint-Vincent, Martinique). Of those recorded in La Trinité, which seem to be the largest, only the 1755 tsunami provides sufficient data to conduct a vulnerability study of the town nowadays. Moreover, La Trinité is located at very low elevation (< 5 m) and concentrates its public offices in a narrow area very close to the sea, which increases the town's tsunami vulnerability. Thus, Figure 8 shows the strategic sites of the town (schools, administrative centres, firehouses, hospitals, etc.) as well as the simulation results that were produced using the source of Barkan et al. (2009), maximising a scenario for the Caribbean, located in the Iberian peninsula but not based on existing submarine or geological structures (Roger et al, 2010a). The results of modelling, however, have not been yet published for Martinique.

As a matter of fact, no Disaster Risk Reduction Plan (DRRP) seems to be interested in indexing the vulnerability of the coastal areas with regards to the tsunami risk, whereas the Swell and Submersion DRRP are already taken into account in urbanization planning. The use of maximising the scenarios with tsunamis generated either by earthquakes affecting the Caribbean and/or based on already-existing geological structures (Baptista et al, 1998; Baptista et al, 2003; Barkan et al., 2009; Roger et al, 2010a,b) on the locality of La Trinité, allowed the present study to compare the modelled inundation of the 1755 event (the models are consistent with the historical accounts and observations, Roger et al., 2010a,b) with those modelled by the two DRRP. The studies indicate that the tsunami inundation similar to that of 1755 would flood the town and all its strategic sites (district offices, mayor's house, schools, police stations, firehouse and emergency centres, etc.), with a maximum flow depth much more significant than that indicated by DRPP, and that serious material and human damage is possible, if prevention measures are not properly implemented (education, works, buildings displacements, etc.) (Leone et al., 2010). Moreover, the present study indicates that the road network will be directly impacted by a potential sea-level rise associated with a similar tsunami. For example, the main coastal road, which separates the town from the sea, would be inundated, thus causing serious problems with the evacuation of people and the transportation of supplies. Additionally, all the buildings within the first 200 m from the coastline will be inundated according to the simulation of the maximum potential scenario for the Caribbean (Barkan et al., 2009). Of all the buildings in La Trinité, only the hospital s is located outside this inundation zone. It should also be noted that the 1755 tele-tsunami flooded the bay of Fort-de-France and of Lamentin, which nowadays is the center of the island's political and economic activity and of the island's main airport and harbor (Lamentin) as mentioned by Roger et al. (2010b).

The DEM (Digital Elevation model) used for tsunami simulations (propagation and inundation) in La Trinité Bay has been constructed using high resolution bathymetry (obtained from high resolution multi-beam and re-sampled bathymetric data of the French Hydrographic Service, SHOM; Roger et al. 2010b) and from topographic data (IGN, 2006); The model reproduces submarine features, coastal configuration and the aerial landscape.

4. CONCLUSION

The compilation of the present tsunami catalog for Martinique Island leads to the following conclusions:

After an accurate comparison of the different available historical catalogs, only twenty-eight events have been classified as tsunamis that reached Martinique since its discovery. Among them, twenty-three tsunamis – corresponding to 50% of the total - have been generated by earthquakes, directly or indirectly (induced landslides). For the majority of cases it was not possible to discriminate between a direct or indirect cause due to the lack of adequate historical data. In the same way, 28% of those tsunami events have occurred following volcanic eruptions but again, it is very difficult to make the distinction between a direct origin associated with a lahar entering the sea, or a volcano's flank instability, or else a combination of an earthquake and an eruption. 49% of the events (16/33) have had a local impact.

The present study allowed the conduct of a preliminary vulnerability study of the tsunami hazard. In fact, tsunamis have affected several localities along the coast of Martinique, but some of them, due mainly to the geographical location, coastal configuration and mainly because of lack of proper observations, may show a lack of impact. Thus, the town of La Trinité was the perfect example for this study of tsunami risk and vulnerability, since this is the area of increased tourist activity and coastal urbanization, indicative of similarly exponential vulnerability of other coastal towns. The choice of the 1755 Lisbon tsunami as the worse possible scenario for tsunami impact on Martinique highlights the need to improve the current anti-flood DRR plan for the island.

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REFERENCES

Alasset P.J, H. Hébert, S. Maouche, V. Calbini, M. Meghraoui (2006). The tsunami induced by the 2003 Zemmouri earthquake (MW = 6.9, Algeria): modelling and results, Geophys. J. Int, 166**,** 213–226.

Baptista M. A., S. Heitor and L. Mendes Victor (1998). The 1755 Lisbon tsunami; Evaluation of the tsunami parameters, J. of Geodynamics, 25(2), 143–157.

Baptista M.A., J. M. Miranda, F. Chierici and N. Zitellini (2003). New study of the 1755 earthquake source based on multi-channel seismic survey data and tsunami modeling, Natural Hazards and Earth System Sciences, 3, 333–340.

Baptista M.A., J.M. Miranda and J.F. Luis (2006). In Search of the 31 March 1761 Earthquake and Tsunami Source, Bulletin of the Seismological Society of America, 96(2), 713–721.

Barkan R, U. Brink, J. Lin (2009). Far field tsunami simulations of the 1755 Lisbon earthquake: Implications for tsunami hazard to the U.S. East Coast and the Caribbean, Marine Geology, n°264, 109–122.

Boyer-Peyreleau, E.E. (1823). Les Antilles Françaises, particulièrement la Guadeloupe, depuis leur découverte jusqu'au 1er Novembre 1823, 3 volumes, Brissot-Thivars, Paris.

Calmont A. and C. De Vassoigne (1999). Guadeloupe, Martinique, Guyane: Des espaces tropicaux entre insularité et continentalité, Mappemonde n°54 (1999.2).

Charbit Y. (2006). Les colonies françaises au xviie siècle : mercantilisme et enjeux impérialistes européens, Revue européenne des migrations internationales.

Chauleau Y. (1993). Dans les iles du vent, la Martinique XVIIe-XIXe siècle, l'Harmattan.

Chester D-K. (2001). The 1755 Lisbon earthquake, Progress in Physical Geography, 25, 363–383.

Choi B. H., E. Pelinovski, K.O. Kim and K.O. Lee (2003). Simulation of the trans-oceanic tsunami propagation due to the 1883 Krakatau volcanic eruption, Natural Hazards and Earth System Sciences,3, 321–332.

Clément A. (2009). "Du bon et du mauvais usage des colonies" : politique coloniale et pensée économique française au XVIIIe siècle, Cahiers d'économie politique, N°56, pp. 101-127.

Du Tertre J.-B. (1668-1671). Histoire générale des Antilles habitées par les Français.

Feuillard M., C.J. Allegre, G. Brandeis, R.Gaulon, J.L. Le Mouel, J.C. Mercier, J.P .Pozzi and M.P. Semet (1983). The 1975–1977 crisis of la Soufriere de Guadeloupe (F.W.I): A still-born magmatic eruption, Journal of Volcanology and Geothermal Research**,** 16(3-4), 317-334.

Germa A. (2008). Evolution volcano-tectonique de l'île de la Martinique (Arc insulaire des petites Antilles): Nouvelles contraintes géochronologiques et géomorphologiques, Thèse de Doctorat, Université d'Orsay.

Goiffon M. (2003). Pression foncière et littoralisation à la Martinique: Pression anthropique et environnement en Amérique latine, Les Cahiers d'Outre-Mer, N° 223.

Grindlay N. R., M. Hearne and P. Mann (2005). High Risk of Tsunami in the Northern Caribbean, Eos, 86,(12).

Herd R.A., M. Edmonds, V-A Bass (2005). Catastrophic lava dome failure at Soufriere Hills Volcano, Montserrat, 12–13 July 2003, Journal of Volcanology and Geothermal Research, 148, 234– 252.

Hess J. (1902). La Catastrophe de la Martinique, Note d'un reporter, Ed. Eugène Fasquelle, Paris.

Institut Géographique National (2006). Le Lamentin, presqu'île de la Caravelle, PNR de la Martinique. Carte de Randonnée, 4502 MT, édition 2, 1 : 25000.

INSEE (2009). Enquête de recensement sur la population en Martinique, le 1er Janvier 2006, INSEE Antilles-Guyane, N°39, Janvier 2009.

Kathiresan K. and N Rajendran (2005). Coastal mangrove forests mitigated tsunami, Estuarine, Coastal and Shelf Science, 65, 601-606.

Kunkel C.M., R.W. Hallberg and M. Oppenheimer (2006). Coral reefs reduce tsunami impact in model simulations, Geophysical Research letters, 33, L23612, doi:10.1029/2006GL027892.

Lacroix A. (1904). La Montagne Pelée et ses éruptions, Editions Masson. 663 pp.

Lambolez C.L. (1905). Saint-Pierre – Martinique, Annales des Antilles françaises, Journal et album de la Martinique : Naissance, vie et mort de la cité créole (1635 – 1902), Livre d'or de la Charité, Ed : Berger-Levrault et Cie, Paris.

Lander F. (1997). Caribbean tsunami, an Initial History, Naturals Hazards and Hazards Management in the Greater Caribbean and Latin America, 3, 1–18.

Lander F., L.S. Whiteside and P.A. Lockridge (2002). A Brief history of tsunamis in the Caribbean Sea, The International Journal of The Tsunami Society, 20(2), 57–94.

Lander F., L.S. Whiteside and P.A. Lockridge (2003). Two Decades of global tsunamis (1982-2002), Science of Tsunami Hazards, 21(1), 3-88.

Lalung H. (1948). Les Caraïbes, peuple étrange, aujourd'hui disparu, Bourrelier et Cie, Paris.

Le Friant A., G. Boudon, A. Arnulf, R. Robertson (2009). Debris avalanche deposits offshore St. Vincent (West Indies): Impact of flank-collapse events on the morphological evolution of the island, Journal of Volcanology and Geothermal Research, 179, 1–10.

Leone F. and T. Lesales (2008). L'évaluation intégrée du risque volcanique. Application à la montagne Pelée (Martinique), in Pagney Benito-Espinal (Ed) - Les interfaces ruptures, transitions et mutations, coll. Espaces tropicaux. Presses Universitaires de Bordeaux.

Leone F., F. Lavigne, R. Paris, J-C Denain and F. Vinet (2010). A spatial analysis of the December 26th, 2004 tsunami-induced damages: Lessons learned for a better risk assessment integrating buildings vulnerability, Applied Geography, in press.

Macdonald R., C.J. Hawkesworth, E. Heath (2000). The Lesser Antilles volcanic chain: a study in arc magmatism, Earth-Science Reviews,49, 1–76.

Mallet, R. (1853). Catalogue of Recorded Earthquakes from 1606 B.C. to A.D. 1850, Part I, 1606 B.C. to 1755 A.D. Report of the 22nd Meeting of the British Association for the Advancement of Science, held in Belfast, Sept. 1852, John Murray, London, 177 pp.

Mallet R. (1854). Catalogue of Recorded Earthquakes from 1606 B.C. to A.D. 1850, Part II, 1755 A.D. to 1784 A.D., Report of the 23Td meeting of the British Association for the Advancement of Science, held in Hull, Sept. 1853, John Murray, London, 118-212.

Mallet R. (1855). Catalogue of Recorded Earthquakes from 1606 B.C. to A.D. 1850, Part III, 1784 A.D. to 1842 A.D., Report of the 24" Meeting of the British Association for the Advancement of Science, John Murray, London, 326 pp.

Morton R., B. Richmond, B. Jaffe, G. Gelfenbaum (2006). Reconnaissance Investigation of Caribbean extreme wave deposits – preliminary observations, interpretations, and research directions, Open-File Report 2006-1293, USGS.

O'Loughlin K. and F. Lander (2003). Caribbean Tsunamis, A 500 year History from 1498 – 1998, Kluwer Academic Publishers.

Pararas-Carayannis, G., 2003. Near and Far-Field Effects of Tsunamis Generated by the Paroxysmal Eruptions, Explosions, Caldera Collapses and Slope Failures of the Krakatau Volcano in Indonesia, on August 26-27, 1883, Presentation for the International Seminar/Workshop on Tsunami "In Memoriam 120 Years Of Krakatau Eruption Tsunami And Lesson Learned From Large Tsunami" August 26th 29th 2003, Jakarta and Anyer, Indonesia. Also published in the Journal of Tsunami Hazards, Volume 21, Number 4. 2003

Pararas-Carayannis, G., 2006. Risk Assessment of Tsunami Generation from Active Volcanic Sources in the Eastern Caribbean Region - In "CARIBBEAN TSUNAMI HAZARDS"- Proceedings of National Science Foundation Caribbean Tsunami Workshop, March 30-31, 2004, Puerto Rico, Aurelio Mercado-Irizarry - Philip Liu, Editors. ISBN 981-256-545-3, 341 pp. 2006, Hard Cover Book. World Scientific Publishing Co. Pte.Ltd., Singapore.

Pararas-Carayannis G., (2010). Assessment of Tsunamigenic Potential along the Northern Caribbean Margin – Case Study: Earthquake and Tsunamis of 12 January 2010 in Haiti. Science of Tsunami Hazards, Vol 29, No. 3 (Review of draft report).

Revert E. (1949). La Martinique : Etude géographique et humaine, Paris : Nouvelles Éditions latines, 559 pp. Collection : Bibliothèque de l'Union française.

Roger J. and M.A. Baptista (2009). Synthèse des informations historiques relatives au tsunami associé au séisme de Lisbonne de 1755 dans les Antilles, Programme ANR RISKNAT, projet MAREMOTI**,** Rapport d'activité n°2.1

Roger J., M-A Baptista, A. Sahal, F. Accary, S. Allgeyer, H. Hébert (2010b). The transoceanic 1755 Lisbon tsunami in Martinique, Pure and Applied Geophysics, proceedings of the International Tsunami Symposium, Novosibirsk, Russia. (in press).

Roger, J., S. Allgeyer, H. Hébert, M.A. Baptista, A. Loevenbruck, F. Schindelé (2010a). The 1755 Lisbon tsunami in Guadeloupe Archipelago : sourve sensitivity and investigation of resonance effects, The Open Oceanography Journal, 4, 58-70.

Roger, J., M.A. Baptista, D. Mosher, H. Hébert, A. Sahal (2010c). Tsunami impact on Newfoundland, Canada, due to far-field generated tsunamis. Implications on hazard assessment. Proceedings of the 9th U.S. National and 10th Canadian Conference on Earthquake Engineering, July 25-29, 2010, Toronto, Canada, n°1837.

Saffache P. (2005a). Petit manuel de géographie de la mer et des littoraux : essai de compréhension du milieu littoral martiniquais, Matoury : Ibis Rouge Editions, Presses Universitaires Créoles, Collection Géographie & Aménagement des Espaces Insulaires, 53 pp.

Saffache P. (2005b). The vulnerability of the island of Martinique to the risk of a tsunami, Études caribéennes, n°3. http://etudescaribeennes.revues.org/653

Saffache P., J.V Marc, J. Mavoungo, V. Huyghues-Belrose , O. Cospar (2003), Tremblements de terre et raz de marée dans les Départements Français d'Amérique (1643-2002) : éléments pour un aménagement raisonné et une prise de conscience de la vulnérabilité du milieu. Éditions Publibook Université, Collection Sciences Humaines et Sociales, Série Géographie, Paris.

Sahal A., J. Roger, S. Allgeyer, B. Lemaire, H. Hébert, F. Schindelée, F. Lavigne (2009). The tsunami triggered by the 21 May 2003 Boumerdès-Zemmouri (Algeria) earthquake: field investigations on the French Mediterranean coast and tsunami modeling, Natural Hazards and Earth System Sciences,9, 1823–1834.

Schleupner C. (2007). Spatial assessment of sea level rise on Martinique's coastal zone and analysis of planning frameworks for adaptation, J Coast Conserv, 11, 91–103.

Smith M. and J. Shepherd (1993). Preliminary Investigations of the Tsunami Hazard of Kick'em Jenny Submarine Volcano, Natural Hazards, 7, 257-277.

Yeh, H., Liu, P., Briggs, M., Synolakis, C. (1994). Propagation and amplification of tsunamis at coastal boundaries. Nature, 372, 353-355.

Westercamp D. and H. Tazieff (1980). Martinique, Guadeloupe Saint-Martin, La Désirade, Guides Géologiques Régionaux, Masson, Paris.

Zahibo N. and E. Pelinovsky (2001). Evaluation of tsunami risk in the Lesser Antilles, Natural Hazards and Earth System Sciences, 1, 221–231.

Zahibo N., E. Pelinovsky, A. Yalciner, A. Kurkin, A. Koselkov, A. Zaitsev (2003). Le tsunami de 1867 aux ÎlesVierges : observations et modélisation, Oceanologica Acta, 26, 609–621.

Zahibo N., E. Pelinovski, E. Okal, A.Yalçiner, Ch.Kharif, T. Talipova, A. Kozelkov (2005). The earthquake and tsunami of november 21, 2004 at Les Saintes, Guadeloupe, Lesser Antilles. Science of Tsunami Hazards, 23(1), 25-39.

Table 1: Criterion of validity

Degree of validity	Meaning
1	‘Abnormal oscillations’
2 a	Tsunamis generated but not observed in Martinique
2 b	Known earthquake or landslides which did not lead to a tsunami, according to the available sources
3	Events sometimes recorded as tsunamis, but about which sources disagree and/or give contradictory accounts
4	Tsunamis regularly referred to by a significant proportion of sources
5	Tsunamis systematically and unanimously recorded as such

DATE				Localisation of tsunamis			Parameters							Validity coefficient	Consequences	sources	Additional information
Year	**Month**	**Day**	**Hours**	**Origin of the tsunami**		**Affected towns and rivers in Martinique**	**Type**	**Cause (Intensity, Magnitude)**	**Period**	**Range**	**run-up**	**Nb waves**	**Withdrawal**		**Damages**		
1657				Martinique	Martinique		Abnormal sea oscillations	S						1	Houses (seism)	Lambolez, 1905	First earthquake since 1635
1690	April	6	16h LT	Antigua and Guadeloupe (S) + Redonda (L)	Charlestown, Charlotte-Amalie, Guadeloupe, Martinique, Barbade/Barbuda (?), Ste Lucie, Montserrat, Antigua, St Christopher		Regional tsunami	S (XI, 8,1), L					201 m (Charlestown) , 16,5 to 18,5 m à St Thomas	4		Lander, 1997 ; Lander et al, 2002 ; O'Loughlin and Lander, 2003	Similar event in 1843, 1985, 2004
1702	Septem ber			Martinique, Guadeloupe	Martinique, Guadeloupe, Antigua		Regional tsunami	S (VIII, 6,5)						2 a	-	Saffache et al, 2003 ; O'Loughlin and Lander, 2003	Excited animals
1718	March	6	night	Martinique, St Vincent	Martinique, St Vincent		Local tsunami	S + L (Mart) + V (St-Vincent)						4		Lambolez, 1905 ; O'Loughlin and Lander, 2003	
1751	Novem ber	21	7h50 LT	Hispaniola	Antilles (including Martinique)		Regional tsunami	S (XI, 8)						3		O'Loughlin and Lander, 2003 ; [www.tsun.sscc.ru]	
1755	Novem ber	1st	2h à 6 h	Lisbonne	Martinique	Trinité, Galion, Fort-Royal, Lamentin, Cul-de-sac François	Télétsunami	S (Ms : XII, 8,75 - 9)	15 min (Trinité)		12 pieds (Trinit é)	3	200 "pas" (= 124 m)	5	Damaged houses, docks, shops, and boats	Lambolez, 1905 ; Saffache et al, 2003	No tsunami referenced in Ste Marie and Le Robert. No awareness of tsunami event by slaves (slaves collected fishes on the beach during the sea withdrawal caused by the tsunami).
				Lisbonne	Martinique	Epinette River (Trinité), Lamentin River and Fort-Royal River	Télétsunami	S (Ms : XII, 8,75 - 9)			3 pieds (Lam)	3		5	Damaged houses, docks, shops, and boats	Lambolez, 1905 ; Saffache et al, 2003	
				Lisbonne	Martinique		Télétsunami	S (Ms : XII, 8,75 - 9)		4,6 (VII) to 1,8 m		3	1,6 km	5	Damaged houses, docks, shops, and boats	Zahibo and Pelinovski, 2001 ; O'Loughlin and Lander, 2003	
1767	April	24	6h30 LT	Surinam	Martinique, Barbade		Regional tsunami	S						4		O'Loughlin and Lander, 2003	
1823	Novem ber	30	3h10 LT	Martinique	Martinique	St Pierre (Harbour)	Local tsunami	S (4,8)						4	Damaged boats	O'Loughlin and Lander, 2003 ; Saffache et al, 2003	Similar events on a 1 year interval

1824	November	30	3h30 LT	Martinique	Martinique	St Pierre (Harbour)	Local tsunami	S (4,8)						4	Damaged boats	O'Loughlin and Lander, 2003 ; Saffache et al, 2003	
1827	November	30	3h LT	Martinique, Guadeloupe, Antigua			Regional tsunami	S (VIII, 6,5)						2 a		O'Loughlin and Lander, 2003	
1829	October	26		Martinique	Martinique		Local tsunami	S ou Storm						4		O'Loughlin and Lander, 2003	
1831	December	3	19h40LT	Grenada	St Kitts, St Vincent, Guyana, Trinidad		Regional tsunami	S (IX, 7)						2ab		O'Loughlin and Lander, 2003	Effect of the tsunami reported at St Kitts and Trinidad
1837	July	26	12h51 UT	Martinique	Martinique		Local tsunami	S ou Storm						2ab	Numerous casualties	Lander, 1997 ; Lander et al, 2002 ; O'Loughlin and Lander, 2003	
1839	January	11	6h LT	Martinique	Martinique	harbour	Local tsunami	S (IX, 6,9) : id Seaquakes						4	Boats damaged by the tsunami, 400 houses destroyed in Fort Royal (S), 400 deads	Lambolez, 1905 ; O'Loughlin and Lander, 2003	
1843	February	8	10h35 LT	Guadeloupe	Antilles (including Martinique)		Regional tsunami	S (XII, 8,3)		1,2 m				2 b		Lander, 1997 ; Lander et al, 2002 ; O'Loughlin and Lander, 2003	Similar event in 1690, 1985, 2004
1867	November	18	14h50 LT	Iles Vierges	Antilles (including Martinique)	Fort-de-France	Regional tsunami	S (X, 7,3)		0,7 m				5		Zahibo and Pelinovski, 2001; Zahibo et al, 2003 ;; O'Loughlin and Lander, 2003	
1874	March	11	4H30 LT	Dominique and St Thomas	Dominique and St Thomas		Regional tsunami	L						1		O'Loughlin and Lander, 2003	
1876	November	8	15h	Martinique	Martinique	between Ste Marie and the Presqu'île de la Caravelle (Trinité)	Local tsunami	L ?						4	No damages	Lambolez, 1905	
1880	January	4	11 h LT	Dominique	Dominique		Local tsunami	V + L						2 b		O'Loughlin and Lander, 2003	River level rised by 3,7 m (Roseau)
1901				Martinique	Martinique	Rade de St Pierre	Abnormal sea oscillations, violent currents	V				3 or 4		4	No damages	Saffache et al, 2003	
1902	February-March			Martinique	Martinique	North of the island : Macouba - Le Prêcheur	Abnormal sea oscillations, violent currents	V						4	No damages	Saffache et al, 2003	

1902	May	5	13h LT	Martinique (Factory Guerin)	Martinique	La Guérite - Bellevue (between Fort de France and the Pointe des nègres)	Local tsunami	V (3rd lahar)					100 m	5	Flooded houses and roads (Fonds Core). Flooded shops, boats moved towards coast, destroyed docks	Hess, 1902 ; Lambolez, 1905 ; Saffache et al, 2003	
			13h LT	Martinique (Factory Guerin)	Martinique	Blanche River, Roxelane River(St Pierre)	Rising of rivers level	V (3rd lahar)	2 min (Blanche)	8 mètres (Rox)	"Pont de Pierre" (Rox)	1	10 m to 300 feet (Blanche)	5		Hess, 1902 ; Lambolez, 1905 ; Saffache et al, 2003	
			13h LT	Martinique (Factory Guerin)	Martinique	St Pierre (Port :Company Girard, Square Bertin, Fonds-Coré, le Mouillage) + Carbet	Local tsunami	V (3rd lahar)	1 to 2 min	3 to 4 m for the first wave to 20 m	20 m (with the fontain of the square Bertin)	2 to 15 vagues	From 60 to 70 m (Mr Sully) and to 1m20 (person resqued). About 20 to 30 m	5		Hess, 1902 ; Lambolez, 1905 ; Saffache et al, 2003	
			13h LT	Martinique (Factory Guerin)	Martinique	Trinité	Local tsunami	V (3rd lahar)		80 cm		3		5		Hess, 1902 ; Lambolez, 1905 ; Saffache et al, 2003	
1902	May	7	19h LT	St Vincent	Martinique	Trinité	Local tsunami	V				3	80 cm	5		Lander, 1997 ; Lander et al, 2002 ; O'Loughlin and Lander, 2003	
1902	May	7	14 - 15h LT	St Vincent	Martinique	Madame River (Fort de France), Des Pères River (St Pierre)	Local tsunami	V			25 cm			5		Saffache et al, 2003	
1902	May	8	19H - 20h LT	Martinique (Mount Pelée)	Martinique	St Pierre, le Précheur, Carbet, Trinité, Fort-de-France	Local tsunami	V (Nuées ardentes)		3 m (St Pierre), 2m (Carbet)	40 cm (Fort). 200 m (Carbet)	3	1,50 to 2 m (Fort de France)	5	Destruction of all boats in the harbour, except Roddam. 52 km2 destroyed, 38,000 deads, 1 survivor (prisonnier)	O'Loughlin and Lander, 2003	
1902	May	9		Martinique, Ste Lucie	Martinique		Anormal perturbations	V (vent volcanique)						5		O'Loughlin and Lander, 2003	
1902	May	20		Martinique (Mount Pelée, Souffrière St Vincent)	Martinique	Saint Pierre, Carbet, Petite Anse to St Pierre, Fort de France, Trinité	Local tsunami	V		3,50 m (wave hight) : Carbet	50 m at the Petite Anse, 40 cm to St Pierre Fortification			5	Destroyed houses, boats, docks	Saffache et al, 2003	

1902	August	30	21h25 LT	Martinique (Mount Pelée)	Martinique	Saint Pierre, Carbet, Fort de France, Schoelcher, Case Pilote, Trinité	Local tsunami	V		3 m (St Pierre), 1 m at Fort-de-France and Trinité	100 m at Case Pilote, 30 m at Schloecher			5	Flooded docks in Fort de France, including La Savanne square	Saffache et al, 2003 ; O'Loughlin and Lander, 2003	
1906	February	16	1h25 LT	Ste Lucie	Martinique, St Vincent, Guadeloupe, Grenade, Dominique, Barbade		Local tsunami	S (VIII)						2 a		Lander et al, 2002	Tsunamigenic seism reported the same year, Dec. 31, in Venezuela.
1939	July	24	12h LT	Kick'em Jenny	Antilles (including Martinique)	Fort-de-France, Le Vauclin	Regional tsunami	V (VEI: 1)						4		Smith and Shepherd, 1993 ; O'Loughlin and Lander, 2003	
1969	December	25		Guadeloupe	Antilles (including Martinique)		Regional tsunami	S (X - XI, 7,7)						2 b		O'Loughlin and Lander, 2003 ; Zahibo et al, 2005	Barbade, Antigua, Dominique
1985	March	16	14h54 UT	Guadeloupe	Guadeloupe		Local tsunami	S (VI, 6,3) + L						2 b		O'Loughlin and Lander, 2003	Similar event in 1690, 1843, 2004
1991	April	22	21h56 UT	Costa Rica	Antilles (including Martinique)	Fort-de-France Bay	Regional tsunami	S (X - XI; 7,6)						4		Lander et al, 2002 ; O'Loughlin and Lander, 2003	
2004	November	21	11h40 UT	Guadeloupe (Les Saintes)	Guadeloupe		Local tsunami	S (6,3)						2 b		Zahibo et al, 2005	Similar events in 1690, 1843, 1985

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ASSESSMENT OF THE TSUNAMIGENIC POTENTIAL ALONG THE NORTHERN CARIBBEAN MARGIN
Case Study: Earthquake and Tsunamis of 12 January 2010 in Haiti.

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ABSTRACT

The potential tsunami risk for Hispaniola, as well as for the other Greater Antilles Islands is assessed by reviewing the complex geotectonic processes and regimes along the Northern Caribbean margin, including the convergent, compressional and collisional tectonic activity of subduction, transition, shearing, lateral movements, accretion and crustal deformation caused by the eastward movement of the Caribbean plate in relation to the North American plate. These complex tectonic interactions have created a broad, diffuse tectonic boundary that has resulted in an extensive, internal deformational sliver slab - the Gonâve microplate – as well as further segmentation into two other microplates with similarly diffused boundary characteristics where tsunamigenic earthquakes have and will again occur. The Gonâve microplate is the most prominent along the Northern Caribbean margin and extends from the Cayman Spreading Center to Mona Pass, between Puerto Rico and the island of Hispaniola, where the 1918 destructive tsunami was generated. The northern boundary of this sliver microplate is defined by the Oriente strike-slip fault south of Cuba, which appears to be an extension of the fault system traversing the northern part of Hispaniola, while the southern boundary is defined by another major strike-slip fault zone where the Haiti earthquake of 12 January 2010 occurred. Potentially tsunamigenic regions along the Northern Caribbean margin are located not only along the boundaries of the Gonâve microplate's dominant western transform zone but particularly within the eastern tectonic regimes of the margin where subduction is dominant - particularly along the Puerto Rico trench. The Haiti earthquake of 12 January 2010 and its focal mechanism are examined, as they provide additional clues of potential tsunami generation that can occur along transform zones and, more specifically, from interplate and intraplate seismic events and subsequently induced collateral hazards, such as aerial or submarine landslides triggered by strong surface seismic waves.

Key Words: Tsunami, Haiti, Hispaniola, Caribbean northern margin, Seismotectonics.

1. INTRODUCTION

A major earthquake struck Haiti in the early evening of January 12, 2010. It resulted in extensive destruction and thousands of deaths and injuries in the capital city of Port-au-Prince and surrounding areas. It was the worst earthquake to strike the country in the last 200 years. The quake generated a local tsunami in the Gulf of Gonâve and triggered landslides along the southern coast of the island of Hispaniola - which also generated local tsunamis. To understand the12 January 2010 earthquake as well as the specific geotectonic processes that affect Hispaniola, the adjacent Greater Antilles Islands and the potential for tsunami generation in the region, we examine the seismotectonics of the Eastern and Northern Caribbean margin. The present study provides background on the seismotectonics of Hispaniola and of the rest of the Greater Antilles Islands, reviews past destructive earthquakes and tsunamis, documents both the earthquake and tsunami in Haiti and assesses the potential risk of future tsunami generation from seismic events along convergent, divergent and transform tectonic boundaries of the diffuse Northern Caribbean margin.

2. SEISMOTECTONICS OF THE CARIBBEAN REGION - Brief Overview

The Caribbean is a region of considerable tectonic complexity. The Caribbean tectonic plate is surrounded on three sides by the much larger North and South American plates, both of which are moving approximately westward at an estimated rate of about 20 to 30 millimeters per year. There is a moderate level of inter-plate seismicity and interplate and intraplate seismic and volcanic activity. The region is characterized by three main types of plate boundaries: convergent, divergent and transform (Figure 1).

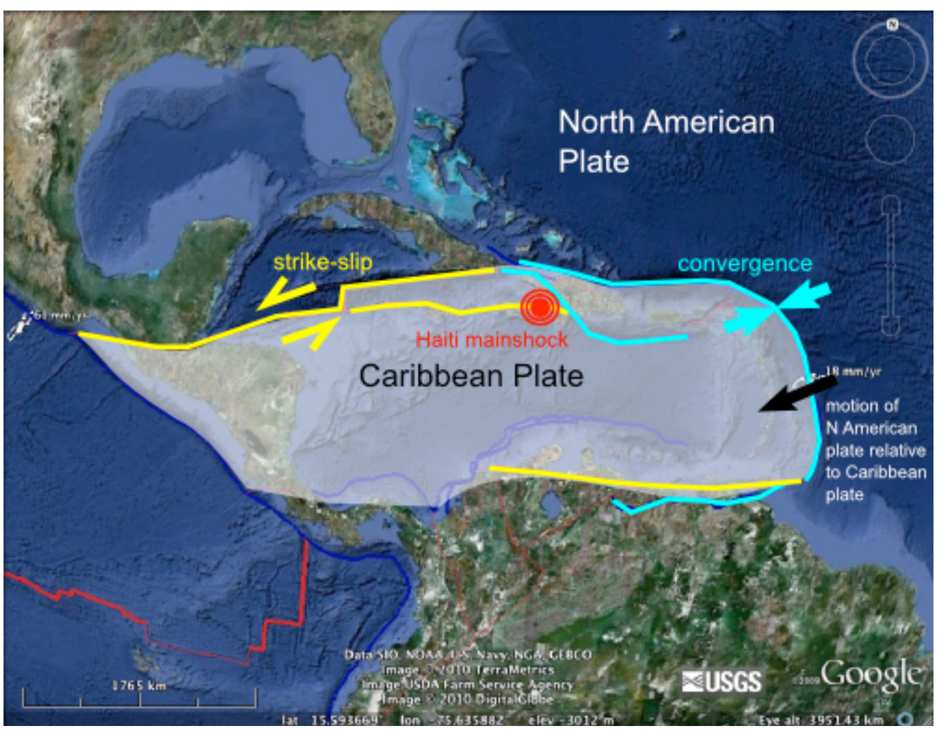

Figure 1. The Caribbean tectonic plate (USGS/Google graphic)

This tectonic activity results in frequent earthquakes, volcanic eruptions, landslides and mass edifice failures of volcanic island flanks. Most of the destructive events occur near or along the geo-tectonically active plate boundaries and are associated with complex mechanisms characteristic of each source. Generally, active seismic source mechanisms in regions of subduction in the Caribbean involve relatively small crustal blocks (Pararas-Carayannis, 2006). The present study examines the tsunamigenic potential along the segment of the Northern Caribbean margin extending from the Anegada Trough in the east to the Cayman Trough Spreading Center in the west (Fig. 2).

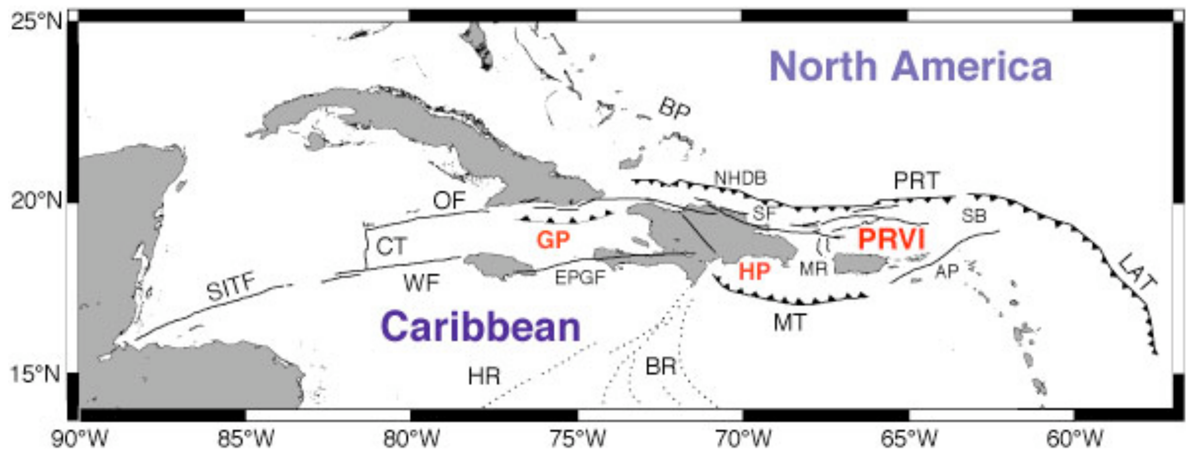

Figure 2. Northern Caribbean Margin (AP: Anegada Trough; BP: Bahamas Platform; BR: Beata Ridge; CT: Cayman Trough Spreading Center; EPGF: Enriquillo-Plantain Fault zone; GP: Gonave Microplate; HP: Hispaniola Microplate; HR: Hess Rise; LAT Lesser Antilles Trench; MR: Mona Pass Rift; MT: Muertos Trench; PRVI: Puerto Rico-Virgin Islands block; SITF: Swan Islands Transform Fault; SF: Septentrional Fault Zone; WF: Walton Fault zone (modified internet graphic).

2.1 Geotectonics of the Eastern Caribbean Margin -- Brief Overview

In the Eastern Caribbean, the interactions between the larger tectonic plates and the Caribbean plate are responsible for zones of subduction and the formation of the West Indies Volcanic Island Arc on the overlying plate. Seismic events are principally associated with a subduction zone along a north-south line, just east of the main island arc where the North American Plate dips from east to west beneath the Caribbean Plate. Additionally, the down-dip compression on the North American plate has created a tensional volcanic back-arc (Fig. 3), which is characterized by spreading and shallower seismic activity (Pararas-Carayannis, 2006).

As the fore-arc is driven by the mantle drag toward a trench – the zone of subduction - the resulting compression is balanced with the slab pull. This flow in the mantle causes back-arc spreading (Seno and Yamanaka, 1998). Arc stresses and such back-arc spreading result in increased volcanic activity in the Lesser Antilles region. Also, the inter-plate tectonic interaction near or along the marginal boundaries often results in moderate to large earthquakes. Superimposed on the inter-plate tectonic interaction is a pattern of intra-plate activity - which is more pronounced in the Leeward Islands region where subduction by the Barracuda Rise adds additional strain on both the "subducted" North American Plate and the overlying Caribbean Plate. In this region earthquakes are generally shallow and potentially tsunamigenic (Pararas-Carayannis, 2006).

Figure 3. The West Indies Volcanic Island Arc (after Pararas-Carayannis, 2006)

In the region northwest of Trinidad there is another concentration of earthquake activity where the strike of the plate boundary changes direction. The earthquakes in this region are of intermediate depth and thus not tsunamigenic. However, the potential for local tsunami generation from individual, active volcanic sources and active shallow seismic regions is relatively high (Pararas-Carayannis, 2006a). Examples of tsunamigenic volcanic activity include recent events related to the eruptions of Soufriere Hills volcano on Montserrat island (Pararas-Carayannis, 2006).

2.2. Geotectonics of the Northern Caribbean Margin

A review of geotectonic processes along the northern Caribbean margin can help assess the tsunami risk for Haiti and the Dominican Republic on Hispaniola, as well as for the other Greater Antilles Islands. Along the Northern Caribbean margin, including areas in the vicinities of Jamaica, Hispaniola and the Virgin Islands, convergent, compressional and collisional tectonic activity - caused primarily from the eastward movement of the Caribbean plate in relation to the North American plate - is also responsible for the creation of zones of subduction and the formation of volcanic island arcs on the overlying plate. The island of Hispaniola where Haiti is located, lies along the eastern Northern Caribbean margin, within the transition zone between different tectonic regimes, with subduction dominant to the east and a transform zone dominant to the west (Mann et al., 1998).

The two tectonic plates meet mainly through a combination of strike slip and interplate thrust faulting (Calais et al., 2002). As the Caribbean plate moves eastward with respect to the North American plate along the Northeastern margin at about 20 mm/yr, the oblique convergence is distributed between the subduction interface and major strike-slip faults within the overriding plate (Syed et al., 2008). In the last 250 years, this northeastern region has experienced eleven large ($M \geq 7.0$) earthquakes. Most of the significant historical tsunamis were generated along the dominant, eastern subduction regime.

The inter-plate tectonic interaction near or along these northern marginal boundaries results in moderate seismic and volcanic activity. Most of the earthquakes are of shallow depth. However, near the plate boundaries moderate, deeper, intra-plate earthquakes can also occur. The intra-plate earthquakes are caused primarily by internal deformation in a slab of the North American Plate. Concentrations of these earthquakes can occur to focal depths of up to 200 kilometers. The deformation within the North American-Caribbean plate boundary zone has resulted in what appears to be a segmentation of the Caribbean plate into three major microplates with diffused boundaries, each requiring an individual pole of rotation to describe its motion relative to the North American plate (Heubeck & Mann, 1991). The three microplates are the Gonâve (GP) and Hispaniola (HP) microplates and the Puerto Rico-Virgin Islands block (PRVI) (Fig. 2).

2.2.1 The Gonâve Microplate

The existence of the postulated Gonâve sliver microplate on the Northern Caribbean margin has been confirmed from GPS measurements of lateral movements, seismic data and patterns of folding, faulting, and tectonically-induced, Quaternary uplift of coral reefs in Western and Central Hispaniola (Dixon et al., 1998; DeMets & Wiggins-Grandison, 2007). Further west, the northern boundary of the Gonâve microplate is defined by the Oriente strike-slip fault south of Cuba (Calais and Mercier de Lepinaya, 2003), which appears to be an extension of the Septentrional fault system along northern Hispaniola. On the south the Gonâve is bounded by the Walton fault, the Jamaica restraining bend and the Enriquillo-Plantain Garden fault zone. The east boundary of the microplate is defined by faults in Hispaniola and the west boundary by the Cayman Spreading Center (CSC) (Fig. 4). The motion relative to the North American Plate is to the east. The elongated Gonâve microplate occupies an area of about 190,000 square kilometers of the northeastern Caribbean plate, which is in the process of shearing off and accreting to the North American plate (DeMets & Wiggins-Grandison, 2007; Mann, et al., 2008). Earthquakes mainly occur along the boundaries of the Gonâve microplate and some, along the subduction zone north of the Septentrional fit, have the potential to generate local tsunamis, particularly in the region along Northeast Hispaniola and closer to Puerto Rico.

Along the southern margin of the Gonâve microplate, earthquakes near the Muertos Trench also have the potential to generate local tsunamis on Hispaniola and Puerto Rico, with lesser effects on the Virgin Islands.

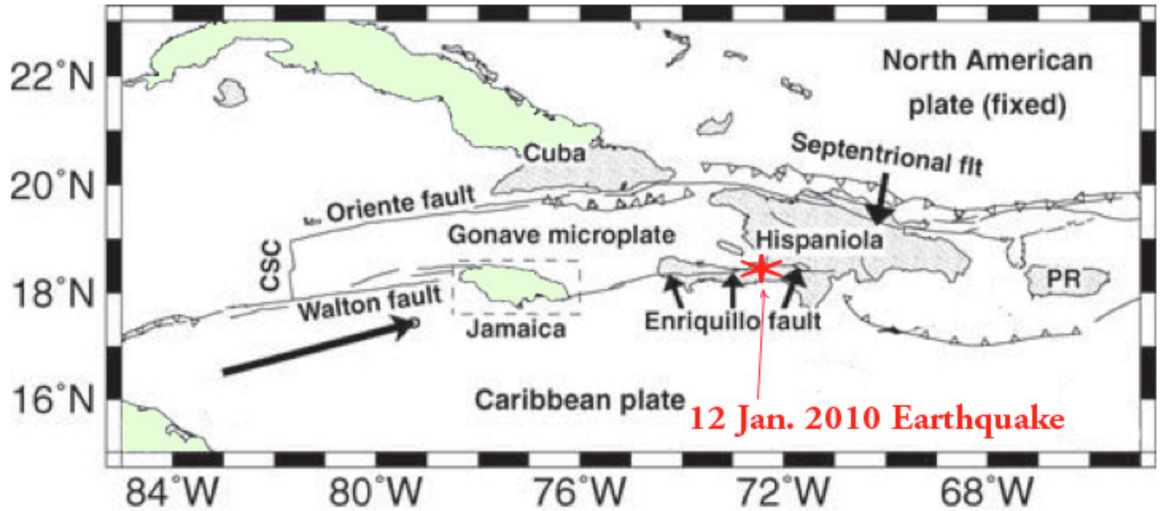

Figure 4. Location map and tectonic setting of the Gonâve microplate on the North-Central Caribbean margin. The arrows indicate the relative motion of the Caribbean plate in relation to the North American as predicted by a model (DeMets et al., 2006). CSC: Cayman spreading center and PR: Puerto Rico (Modified after DeMets and Wiggins-Grandison, 2007)

2.2.2 Hispaniola's Major Fault Zones

Part of the island of Hispaniola is within the Gonâve microplate. Active major plate-boundary structures affecting Haiti and the rest of Hispaniola, include two major left-lateral strike-slip faults and an offshore fold and thrust belt known as the North Hispaniola Deformation Belt, which is part of the subduction zone that extends beneath the northeastern Dominican Republic (Dolan and Wald, 1998; Tuttle et al., 2003). The two major, left lateral, strike-slip fault systems, which traverse Hispaniola in an approximate east-west direction, are the Enriquillo-Plaintain Garden (EPGFZE) along the southern part and the Septentrional (SFZ) along the north (Mann et al., 1995). Between the major fault zones there are a series of interconnected fault lines. Strike-slip faults are clearly visible on satellite imageries as lineaments (Thielen, 2010).

The two fault systems are part of the broader, diffuse boundary that has resulted in an extensive, internal deformational slab of the North American Plate by the aforementioned interaction, collision and subduction with the Caribbean tectonic plate. Tectonic motions are accommodated on the two faults and their offsets, which are located a little bit away from the actual plate boundary, further inside the plate's interior. However, most of the seismic activity on the Gonâve microplate is concentrated in Jamaica rather than Haiti - as shown by the epicenters of earthquakes shown in Figure 5.

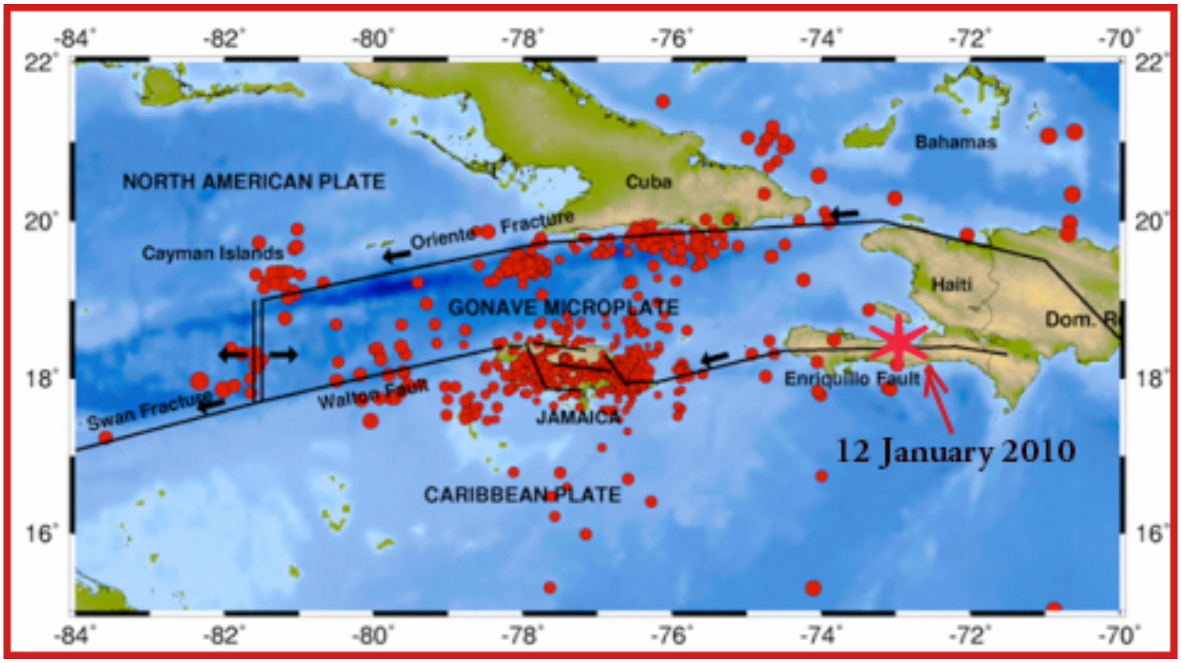

***Figure 5. Seismicity of the Northern Caribbean Margin** (modified University of West Indies graphic)*

Between the EPGFZE and SFZ strike-slip fault systems, the central part of Haiti shows a diffused fractured deformational zone. The most recent phase of tectonism in this zone involves strong uplifts and broad open folding along NW-striking axes.

The Enriquillo-Plaintain Garden Fault System (EPGFZE) - The 1200-km-long EPGFZE is the southern moving edge of the Gonâve microplate (Mann et al., 1995). It provides a "bypass", strike-slip fault system which has allows continued, unimpeded eastward motion of the smaller Caribbean plate in southern Hispaniola - past zones of blocking convergence from the north. Satellite imagery and field observations indicate that the fault zone extends onshore from central Hispaniola east of Lake Enriquillo in the Dominican Republic to the westernmost end of Hispaniola, where it has a total offset of 30-50 km along Haiti's southern peninsula – considered to be a fragment of the Caribbean oceanic crust of the "Late Cretaceous Caribbean plateau basalts" (Mann et al., 1998)(Fig. 6).

The EPGFZE is similar in structure and character to the San Andreas fault of California and the Northern Anatolian fault system in Asia Minor, in that it involves strike-slip displacements, although along portions of its offsets there is evidence of transpression (compression and shear) and seismic events can include small vertical displacements that can generate local tsunamis. To the west, the fault zone crosses beneath the 50-km wide Jamaica Passage and continues west to eastern and central Jamaica where there is an apparent increase in seismicity. To the east, it extends into the Muertos Trench, a zone of subduction where earthquakes can potentially generate local tsunamis on Hispaniola and the adjacent Caribbean islands.

The earthquake of January 12, 2010 - as most of past destructive earthquakes in Haiti- occurred on an offset segment of EPGFZE, associated mostly with shallow earthquakes having left lateral strike slips (Fig. 6). However there is evidence that both vertical and lateral movement occurred - as when the 1701 earthquake struck. This segment of the EPGFZE had been locked for almost for 250 years. The last major earthquake in this particular segment had occurred in 1770. However, in 2008 there was some precursory seismic activity in the area.

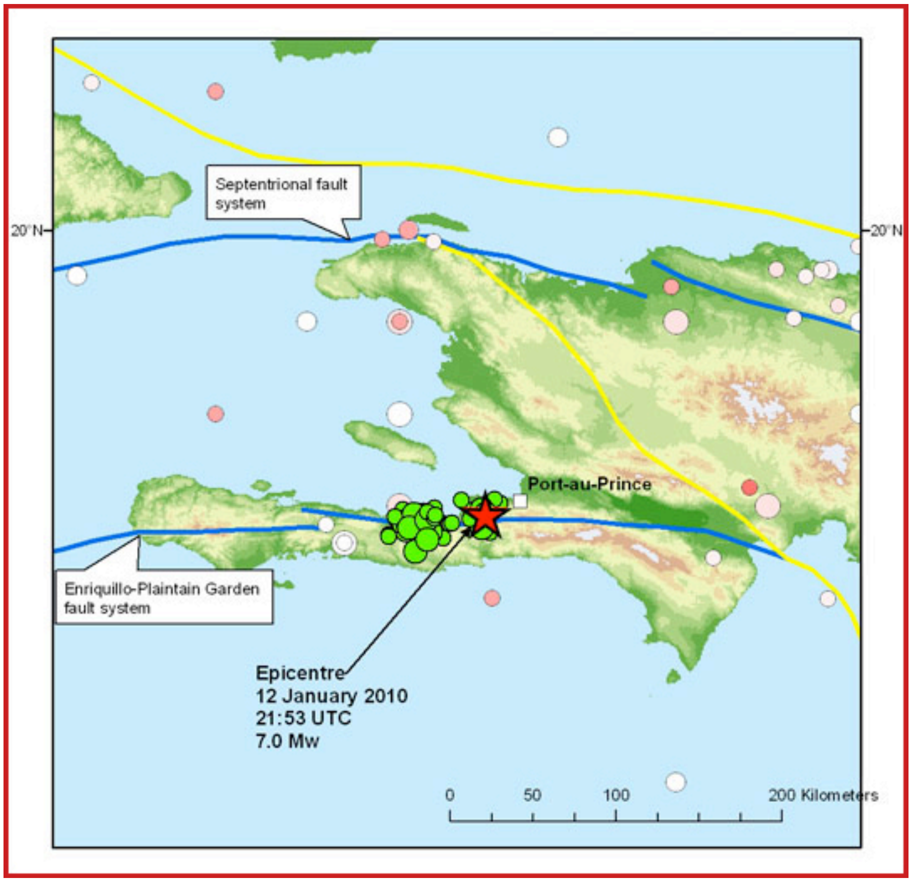

Figure 6. EPGFZE and SFZ in Haiti. Aftershock Distribution of 12 January 2010 Earthquake (modified graphic of the British Geological Survey)

The Septentrional Fault Zone (SFZ) - The Septentrional Fault Zone (SFZ) is a left-lateral fault zone located north of Hispaniola between the Caribbean and the North American plates (Fig. 6). This is a highly seismic zone. It forms part of the diffused north boundary on the Gonâve microplate and the Northern Caribbean margin. It runs through the Cibao valley in the northern Dominican Republic. To the east, it extends into the Puerto Rico subduction zone. To the west, its offset crosses northern Haiti and extends to the southern part of Cuba. There is a high probability that a strong, destructive earthquake on the SFZ could occur in the near future and impact both Haiti and the Dominican Republic.

Haiti's Regional Deformation Patterns - Between the two major strike-slip fault systems (the EPGFZE and the SFZ) the central part of Haiti shows a diffuse fracture zone that trends at N130. The most recent phase of tectonism in Haiti involves strong uplifts and broad, open folding along NW-striking axes, which is consistent with the regional maximum deformation pattern predicted for E-W left-lateral shear along the North America—Caribbean plate boundary (Pubellier et al, 2003).

3. CASE STUDY: EARTHQUAKE AND TSUNAMIS OF 12 JANUARY 2010 IN HAITI.

The traditional belief is that tsunamis are mainly generated from earthquakes along regions of subduction and not from seismic events in regions of lateral fault zones. However, as the 1999 tsunami in the Sea of Marmara indicated, transitional deformation associated with lateral faulting - such as that of the North Anatolian Fault (NAF), can result in earthquakes that can generate destructive tsunamis either by inducing sea floor uplift and subsidence by compression or by strong ground motions triggering sub aerial, submarine landslides or large scale coastal ground liquefaction. The following is a review of the earthquake of 12 January 2010 in Haiti as an interesting case of how local tsunamis can be generated along transitional segments of the Northern Caribbean margin from convergent, compressional and collisional tectonic activity.

3.1 The Earthquake of 12 January 2010 in Haiti

Haiti is on the western side of the Island of Hispaniola, one of the Greater Antilles Islands situated between Puerto Rico and Cuba. The 12 January 2010 earthquake in Haiti was an exceptional episode of sudden strain release over a wide area along a lateral offset segment of the strike-slip EPGFZE zone, which had been locked for 250 years and had accumulated great stress. There had been several minor precursor tremors in 2008 at Petionville, Delmas, Croix des Bouquets and La Plaine, which were indicative of stress increase along this segment of EPGFZ and that a large, long overdue, earthquake was highly probable. Unfortunately, not much attention was given to the precursor events, as the last major, destructive earthquake in the Port-au-Prince region had occurred in 1770. In the past major earthquakes had occurred in 1554, 1701, 1751, 1770, 1783, 1842, 1860 and 1887 along both the southern (EPGFZ) and northern zone (SFZ) (Fig. 6).

The 12 January 2010 earthquake was an extremely destructive shallow event with a focal depth of 13 km and a Moment Magnitude Mw = 7.0 (Fig. 7). It occurred at 21:53:10 UTC (04:53:10 PM local time) and its epicenter was at 18.457 N, 72.533 W (USGS), about 25 km (15 miles) WSW of PORT-AU-PRINCE; 130 km (80 miles) E of Les Cayes, 150 km (95 miles) S of Cap-Haitien, Haiti 1125 km (700 miles) SE of Miami, Florida (USGS data). Its strong ground motions lasting for 35-40 seconds were felt as far away as Jamaica. As in 1751, there was extensive ground cracking and liquefaction in the wide Cul-de-Sac plain - a rift valley that extends eastwards into the Dominican Republic.

Following the main shock up to January 25, 2010, there were at least 24 aftershocks ranging in magnitude from 5.0 to 5.9. Most of the aftershocks occurred to the west of the main earthquake in the region known as the Mirogoane Lakes, which is a basin formed by a 5 km pull-apart displacement from EPGFZ (Mann et al., 1995).

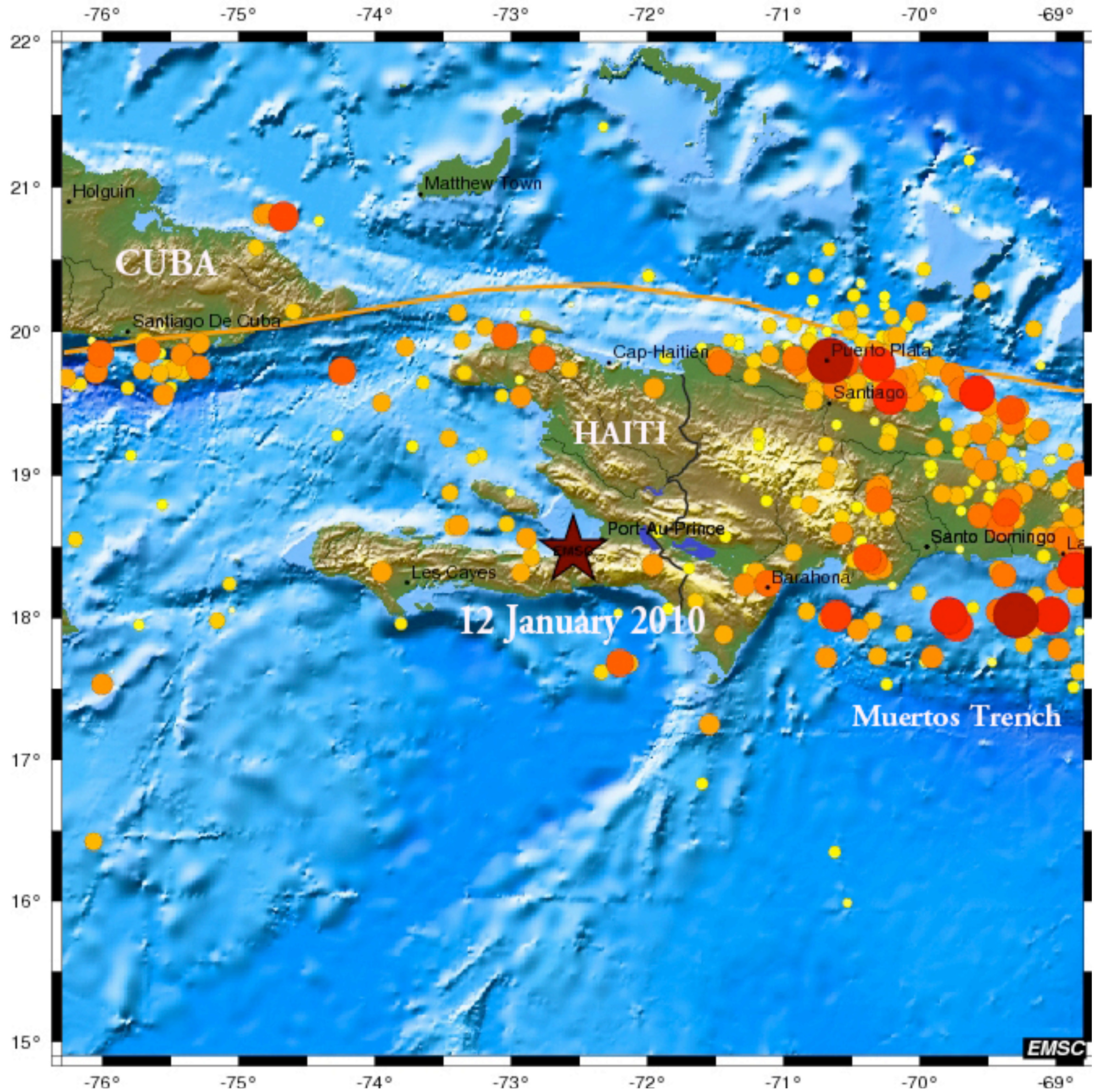

Figure 7. Seismicity from 1964 to 12 January 2010. Earthquakes ranging from M4 to M7.5, including the 12 January 2010, M7.1 earthquake. (Modified graphic of EMSC)

Based on the distribution of the aftershocks it appears that the earthquake's rupture was relatively short and estimated at about 75 km (Fig. 8). However, field investigations around the epicentral area found no clear evidence of rupture on the surface above the plate-boundary fault (Bilham, 2010), which is consistent with satellite radar images taken before and after the earthquake which indicate that the rupture started more than 8 kms in depth but ended at least 2 k below the earth's surface. The significance of such absence of clear rupture is discussed in a subsequent section as it relates to estimates of earthquake recurrence frequencies and potential remaining stress in the region that could result in significant earthquakes and the generation of local tsunamis.

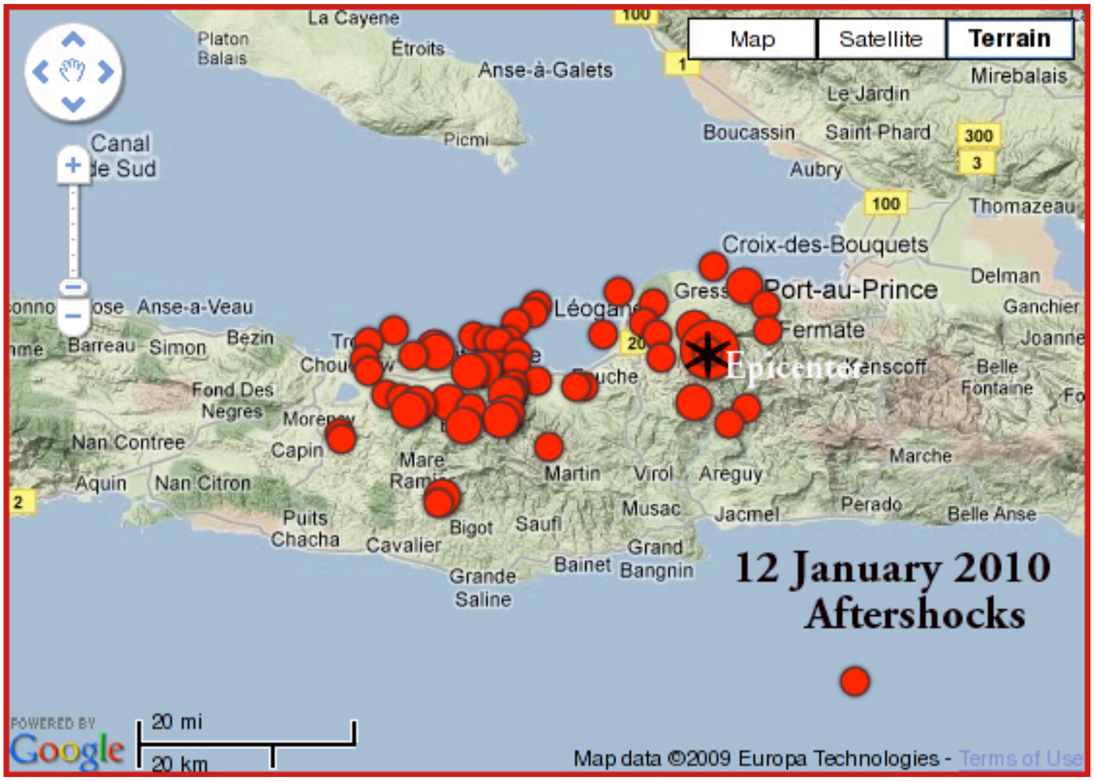

Figure 8. Aftershocks up to January 25, 2010 (modified after Europa Technologies)

3.1.1 Focal Mechanism

The earthquake's focal solution indicates that crustal movements resulted mainly from strike-slip shear. However, the focal mechanism solution (Fig. 9) shows evidence of transtension (extention & shear) as small vertical displacements – both uplift and subsidence - occurred along a portion of the coastline of the region known as the Mirogoane Lakes in the Gulf of Gonave, which as mentioned is a basin formed by a 5 km pull-apart displacement from EPGFZ (Mann et al., 1995).

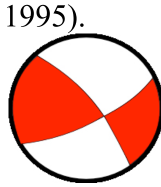

Figure 9. Focal Mechanism of the 12 January 2010 Earthquake.

3.2.2 Remaining Stress

It appears that not all the stress has been released on the EPGFZ. Coulomb stress is evident along both the east and west segments adjacent to that affected by the 12 January 2010 earthquake (Fig. 10). The segment to the east appears to have undergone the greater change in Coulomb stress, thus an earthquake with epicenter closer to Port Au Prince is very possible. Stress transference may also result in strong earthquakes or aftershocks in southern Haiti (east and west) along segments of EPGFZ – and as far west as Jamaica. The significance of the remaining stress for future earthquakes and tsunamis is evaluated in a subsequent section.

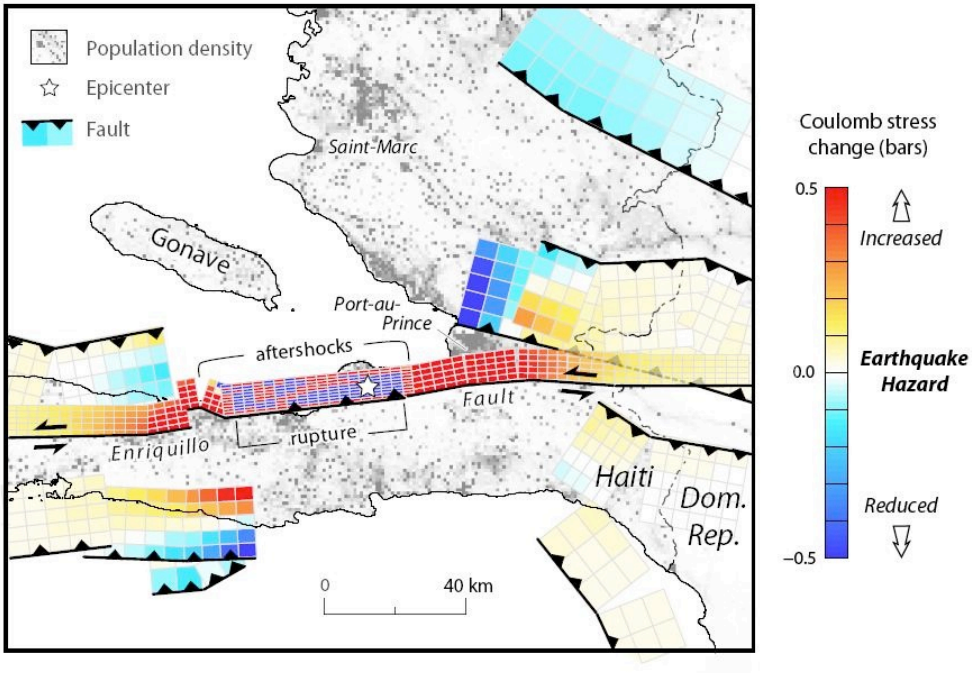

Figure 10. Coulomb stress change after the 12 January 2010 Earthquake (web graphic)

3.3 The Tsunamis of 12 January 2010 in Haiti

An overview of tsunami events in the region was provided by NOAA's Satellite and Information Service (NESDIS). Accordingly, the earthquake generated a local tsunami in the Gulf of Gonâve with a height of 3.3 meters at Petit Paradis - about 62 miles (100 kilometers) away from the earthquake's epicenter. The waves run 70 meters inland causing damage (Lovett, 2010). Seven people were swept out to sea. Trees in the water provided evidence of ground subsidence, as in 1770.

3.3.1 Tsunami Generation Mechanisms

As a general rule, strike-slip earthquakes do not generate tsunamis of any significance as motions involve primarily lateral crustal movements. However, the January 12, 2010 earthquake involved pull-apart motions, as well as small offshore uplift and coastal subsidence, similar to those caused by the 1770 quake. The tsunami was generated mainly by vertical changes of up to 50 cms. Figure 11, below shows the tsunami generating area. However, there is also evidence that the observed tsunami waves were caused by landslides in the Gulf of Gonâve. Both satellite imagery and on-the-ground photos show drastic changes in inland as well as coastal configuration. For example, a field survey found that a palm tree had slid from the shore to a water depth of 7 meters (Fritz, 2010). Waves with 15-second period arrived almost immediately after the quake, which also supports generation by landslides, and localized uplift and subsidence along the coast.

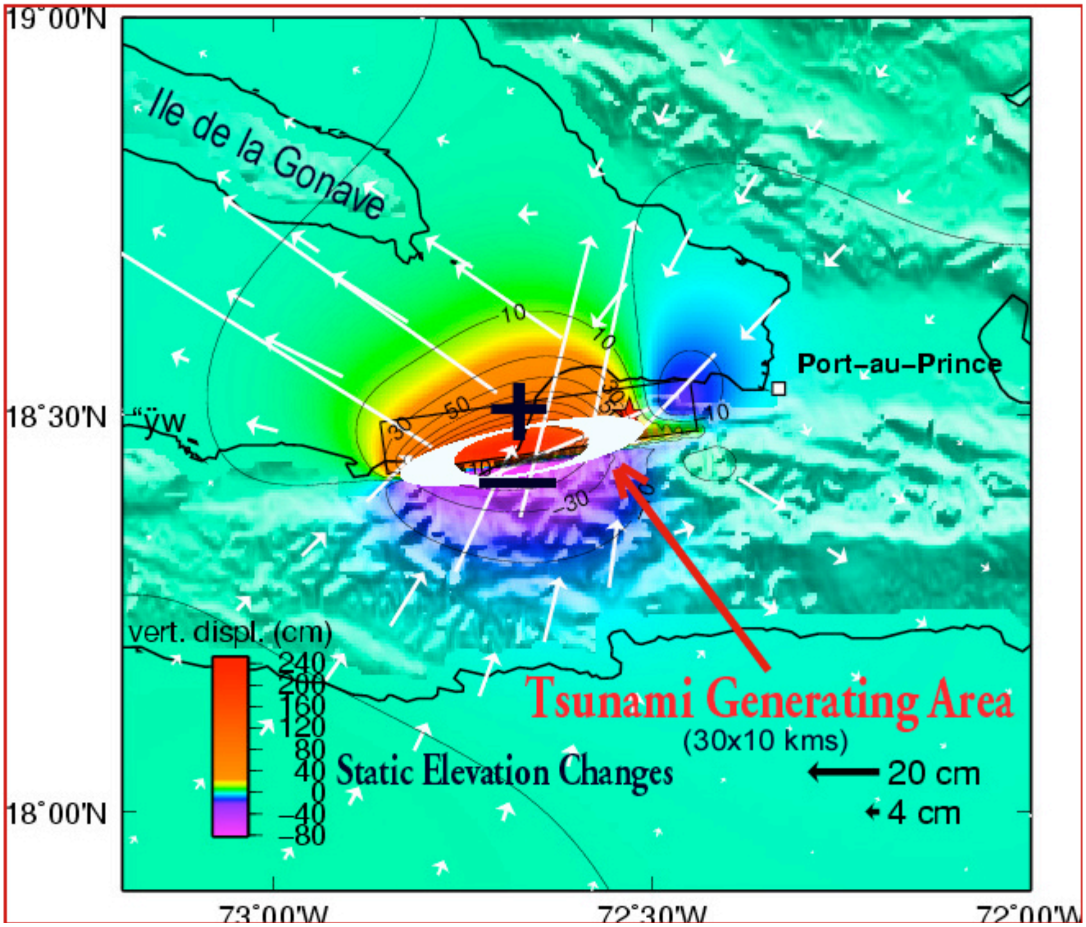

Figure 11. Tsunami Generating Area as inferred from Distribution of Static Elevation Changes and visual field observations (Modified web map of reported static elevation changes with superimposed tsunami generation area estimated at 300sq. Kms.)

Additional field investigations found that sizeable tsunami waves with periods of about 5 minutes and as high as three meters struck along a 100-kilometer stretch of the shoreline on the southern coast of Haiti, all the way into the neighboring Dominican Republic. This led to the conclusion that the tsunami generation in this region was caused by multiple coastal and underwater landslides or perhaps from a larger offshore region. A major landslide is evident on satellite imagery (Fig. 12) at the Bay of Jacmel, on the southern coast of Haiti where a field survey documented strong wave activity. Tsunami wave action washed boats and boulders ashore and knocked down walls at the town of Jacmel (Fritz, 2010). There is also evidence that strong seismic waves from past earthquakes triggered numerous landslides along the southern coast (Bilham, 2010).

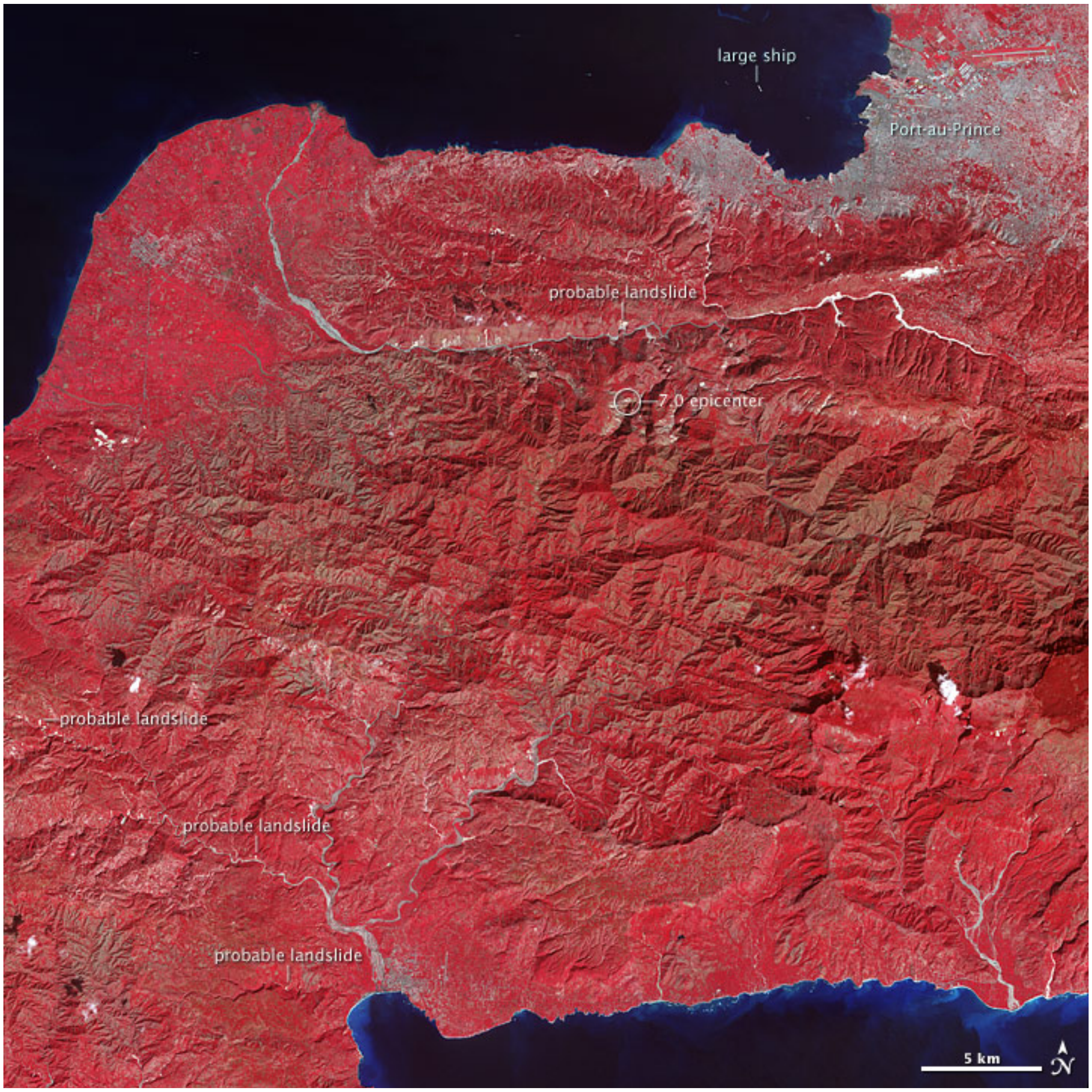

Figure 12. Satellite image (false color) of Haiti's southern peninsula taken on 21 January 2010 with ASTER instrumentation (resolution 15 m) from NASA's Earth Satellite shows (in lighter color) landslides generated by the strong ground motions of the 12 January 2010 earthquake and the location of the ones that generated a local tsunami at the Bay of Jacmel, and elsewhere on the southern coast (Modified NASA graphic haiti_ast_2010021_lrg-1.jpg)

4. HISTORICAL EARTHQUAKES AND TSUNAMIS ALONG THE NORTHERN CARIBBEAN MARGIN

Assessment of the tsunami risk along the Northern Caribbean margin requires review of past events, of their source mechanisms and of their specific impacts. There is a long record of earthquakes throughout the Caribbean region since the arrival of Christopher Columbus in 1492. Some of the larger quakes generated destructive local tsunamis. However not all of the tsunamis were sufficiently documented and could be confirmed.

The historic record shows that at least six of the more significant tsunamis were generated along the Northern Caribbean margin. Also, that a total of twenty-three tsunami-like events occurred along the Lesser Antilles region on the Eastern Caribbean margin. Sixteen of those in the Lesser Antilles were identified as having seismic origin, four of volcanic origin and three of unknown source (Zahibo et al., 2003).

Jamaica has a long history of earthquakes and was severely struck by two destructive earthquakes in 1692 and 1907. As previously stated, most of the larger destructive earthquakes in Haiti and the Dominican Republic have occurred along the EPGFZ in the southern part of Hispaniola and along the SFZ in the northern part of the island.

Figure 13 below shows the focal mechanisms of earthquakes that have occurred along the eastern segment of the Northern Caribbean Margin (Ali et al., 2008). As described subsequently, some of these quakes generated destructive tsunamis.

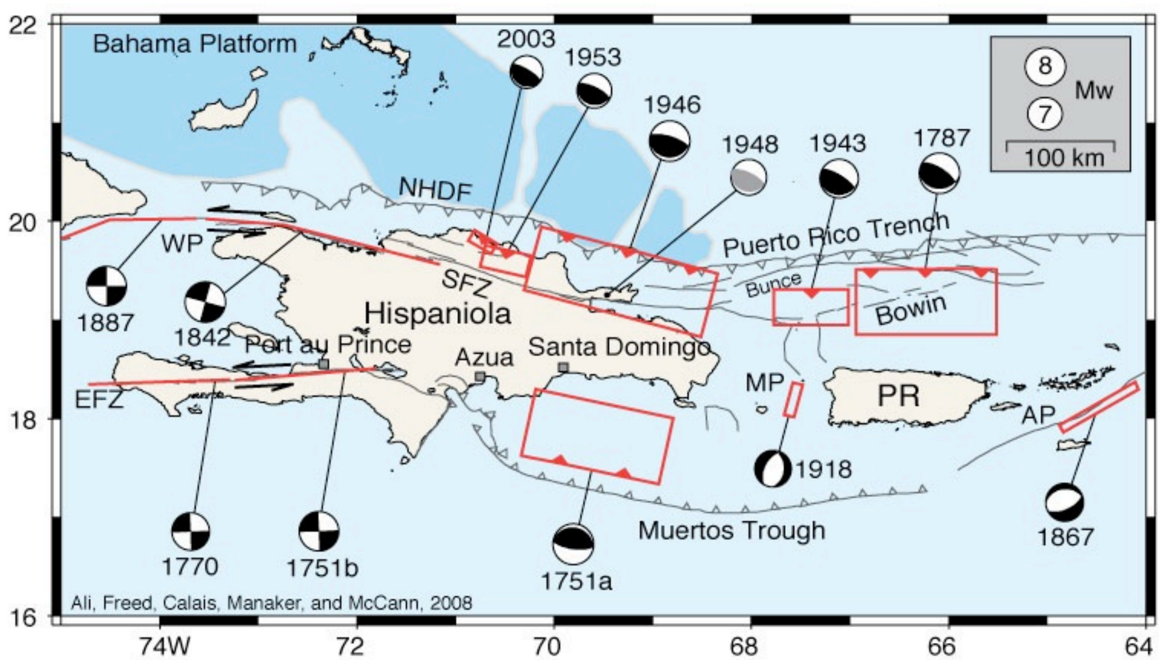

Figure 13. Focal mechanisms and surface projections of estimated rupture planes/geometries for large (M>7.0) historic and a recent M6.5 earthquakes in the Gonave microplate boundaries region since 1751. (Acronyms - NHDF: North Hispaniola Deformation Front; NHF: North Hispaniola Fault; WP: Windward Passage; EFZ: Enriquillo Fault Zone; SFZ: Septentrional Fault Zone; MP: Mona Passage; PR: Puerto Rico) (From Ali et al., 2008).

Along the EPGFZ on Hispaniola, large, destructive earthquakes occurred in 1701, 1751, 1770, 1842, 1860 and 1887 (Dixon et al., 1998; Mann, 2005). Along the SFZ, large destructive earthquakes occurred at Cape Haitien in 1842 and at Pointe-a-Pitre in 1843 (Tomblin, 1981). The following is a listing of the most destructive earthquakes and tsunamis in the northern Caribbean region that impacted Hispaniola and some of the other Greater Antilles Islands up to 1946.

1554 - The first recorded earthquake occurred in 1554 when Haiti was still the Spanish colony of Española. This earthquake destroyed Concepción de la Vega and Santiago de los Caballeros.

1692 June 7 - An earthquake (Epicenter 17.80 N - 76.70 W) that destroyed Port Royal, Jamaica, caused a tsunami that inundated the south coast of the island. About 2,000 people died.

1701 November 9 – The earthquake was one of the most destructive and believed to have occurred between Ile de la Gonave and Haiti's southern peninsula. Port-au-Prince had not yet established. There was severe damage and subsidence around Léogâne. Reportedly, the coastal road from Léogâne to Petit Goâve sank into the sea (Taber, 1922). A tsunami was generated but there are no details of its height of extent of destruction.

1751 October 18 – A major shallow quake with an estimated magnitude up to M8 and epicenter at 18.50 N -70.70 W, destroyed Port-au-Prince and adjacent region of southern Haiti (Taber, 1922).

1751 November 21, 22 - Two more severe quakes occurred on the segment of Enriquillo Fault Zone between Petionville and Tiburon and destroyed the newly established town of Port-au-Prince. There was extensive ground cracking and liquefaction on the Cul-de-Sac plain, which caused the collapse of many buildings.

1770 June 3 – A major, shallow quake occurred at 7:15 pm. Its epicenter was estimated to have been at 18.30 N -72.20 W along the segment of Enriquillo Fault Zone extending from Petionville to Tiburon. There was a subsequent shock. Reportedly the ground motions lasted for about four minutes. Strong ground motions felt as far as in Cap-Haïtien, about 160 kilometers (99 mi) away. Reports that even chimneys in Jamaica collapsed. There were landslides in the mountains, and rivers were dammed. The quakes caused ground cracking and liquefaction, which resulted in extensive destruction from Croix de Bouquets in the east through the plain of the Cul-de-Sac, to Port-au-Prince, as well as along the north coast of the Tiburon Peninsula as far as Miragoâne to the west (Taber, 1922). All buildings between Lake Miragoâne and Petit-Goâve, to the west of Port-au-Prince were leveled. At Grand Goâve the foot of the mountain of La Saline was partly submerged. The village of Croix des Bouquets sank below sea level.

Also, the earthquake generated a tsunami in the Gulf of Gonâve that inundated the coastline inland by as much as 7.2 kilometers (4.5 mi) in the Cul-de-Sac depression (O'Loughlin & Lander, 2003). However this extensive inundation may have been partially due to the effects of subsidence and ground liquefaction. - as in the Gulf of Izmit, in Turkey when the 1999 struck. The death toll from the tsunami is not known.

The death toll of the earthquake was surprisingly low. The reason may have been a rumbling sound that preceded the earthquake, which served as a warning for people to evacuate structures that subsequently collapsed. Only 200 people died in Port-au-Prince from the collapse of buildings, including 79 of 80 people in the hospital. Fifty more people died in Léogâne. However, following the earthquake, 15,000 more people died from famine and diseases.

1781 October 22 - A 10-foot high sea wave swept away houses and killed 10 people on the south coast of Jamaica after an earthquake (epicenter 18.20 N -78.10 W) that occurred during a hurricane.

1783 - A strong quake caused partial collapse of a principal church in Santigo.

1842 May 7 - A disastrous quake (M7.7) near Cap Haitien (epicenter 18.50 N -72.50 W) killed half the population of the town of ten thousand. There were two distinct shocks - the second lasting longer. The earthquake did serious damage to Henri Christophe's palace at Sans Souci image and to the Citadelle La Ferrière near Milot. Milot was Haïti's former capital under the self-proclaimed King Henri Christophe, who ascended to power in 1807, three years after Haïti had gained independence from France.

The earthquake generated a tsunami with an approximate run-up of 4 or 5 meters (13 to 16 feet), which struck the nearby city of Port de Paix (O'Loughlin & Lander, 2003). The tsunami was preceded by a 60-meter (about 197 ft) withdrawal of the sea. The tsunami struck along the northern coasts of the Dominican Republic, Haiti and the Virgin Islands. Several hundred fatalities were recorded at the time, but it is not clear how many were tsunami-related.

1843 February 8 - An earthquake (M8.3) at Pointe-a-Pitre (epicenter 16.50 N - 62.20 W). No information available on a tsunami

1867 November 18 – A 7.5 earthquake along the north scarp of the Anegada Trough, followed by another strong quake ten minutes later (epicenter 18.40 N - 64.30 W about 15 to 20 km southwest of St. Thomas), generated a tsunami that struck the U.S. Virgin Islands. Only 17 lives were lost. A maximum tsunami height of 10m was reported for two coastal locations (Deshaies and Sainte-Rose) in Guadeloupe (Zahibo et al 2003). The US Navy ship *Monongahela* was cast ashore at St. Croix by the tsunami (refloated six months later).

1907 - Major earthquakes occurred in Jamaica. No information available on a tsunami.

1918 October 11 - A 7.5 magnitude earthquake in the Mona Passage, between Puerto Rico and Hispaniola (Epicenter 18.50N - 67.50 W), produced a tsunami waves that struck the western coast of Puerto Rico with run-up of up to 6 meters. There were 116 fatalities, 40 of them directly from the tsunami. Two-inch waves were recorded at Atlantic City, N.J. A second strong tsunamigenic earthquake in the same epicentral origin occurred on 25 October 1918.

1946 August 4 – A magnitude M8.1 earthquake off the northeast coast of the Dominican Republic (epicenter 19.30 N - 69.00 W) triggered a tsunami that killed about 1,800 and partially impacted Haiti. 20,000 people rendered homeless (O'Loughlin & Lander, 2003). The wave was also recorded as far as Daytona Beach in Florida and Atlantic City in New Jersey. A second large earthquake (M7.9) occurred on August 8, 1846.

A more recent tsunami list for the Caribbean published in 2002 shows that in the last 35 years prior to 2002, four local tsunamis occurred along the Eastern

Caribbean margin. Subsequently, however, there have been additional local tsunamis in the Lesser Antilles generated by volcanic flank collapses and pyroclastic flows – specifically from recent eruptions of the Soufriere Hills volcano on Montserrat.

1969 December 25 - An earthquake with magnitude M7.6 in the Lesser Antilles (epicenter 15.79 N - 59.64 W) generated a local tsunami. Maximum tsunami height in the Barbados was 46 cm.

1985 March 16 – A moderate earthquake with magnitude 6.3 in Guadeloupe, generated a local tsunami of several centimeters in height which was recorded at Basse Terre, on Guadeloupe.

1997 July 9 - An earthquake of magnitude 6.8 off the coast of Venezuela generated a small local tsunami on Tobago.

1997 December 26 - A volcanic eruption of the Soufriere Hills volcano in Montserrat generated a wave with a height of 3 m at Old Road Bay (Hooper and Mattioli, 2001).

5. ASSESSMENT OF TSUNAMIGENIC POTENTIAL ALONG THE NORTHERN CARIBBEAN MARGIN

A total of 88 tsunamis - most of them moderate - have been reported in the earthquake-prone, volcano-ringed Caribbean area since 1489. Several of these were generated by volcanic eruptions, collateral volcanic flank failures, debris avalanches and landslides (Pararas-Carayannis, 2006). Based on the historic record, there is a high probability that a destructive tsunami will be generated again in the Northern Caribbean region that will affect coastal areas of the Virgin Islands, Puerto Rico, the Dominican Republic and Haiti. Tsunami waves with a height of up to 10-12 meters could strike coastal areas. The effects of past tsunamis generated along the eastern segment of the Northern Caribbean margin have extended up to 1,320 miles. Thus, there is a high probability that future events may have some far-field impact. Additionally, there is a possibility that an outer ridge earthquake on the Northern Caribbean margin (Fig. 14) but further away from the zone of subduction, could trigger a significant tsunami.

Because of the oblique subduction, there is a great deal of crustal deformation, not only along the southern side of the tectonic boundaries, but also along the outer ridge of the North American plate to the north.

5.1 Potential Tsunamigenic Earthquakes along the Eastern Segment of the Northern Caribbean Margin.

The strain that has been accumulating along the eastern segment of the Northern Caribbean margin was not fully released by the 1867 Anegada Trough or the 1918 Mona Pass events. In addition to past earthquakes, many small underwater landslides and cracks, 20 miles or more long, exists on the sea floor off the coast of Puerto Rico, near where the 1918 tsunami originated. Cracking indicates that these areas are close to failure. Besides the 1918 event occurred in Mona Pass, along the eastern inter-plate convergent boundary of the Gonâve microplate and not exactly on the northern edge of the Caribbean margin. As previously mentioned, three main types of plate boundaries characterize the northeastern Caribbean region: convergent, compressional and collisional. There is also an abrupt change in the obliquity of tectonic convergence. Thus, there is a high probability that a major or even a great tsunamigenic earthquake ranging in magnitude from 7 to 8 - or even greater - could occur somewhere along the northeastern segment of the Northern Caribbean margin. Also a repeat of the May 7, 1842 quake (M7.7) near Cap Haitien on Hispaniola could generate a destructive tsunami.

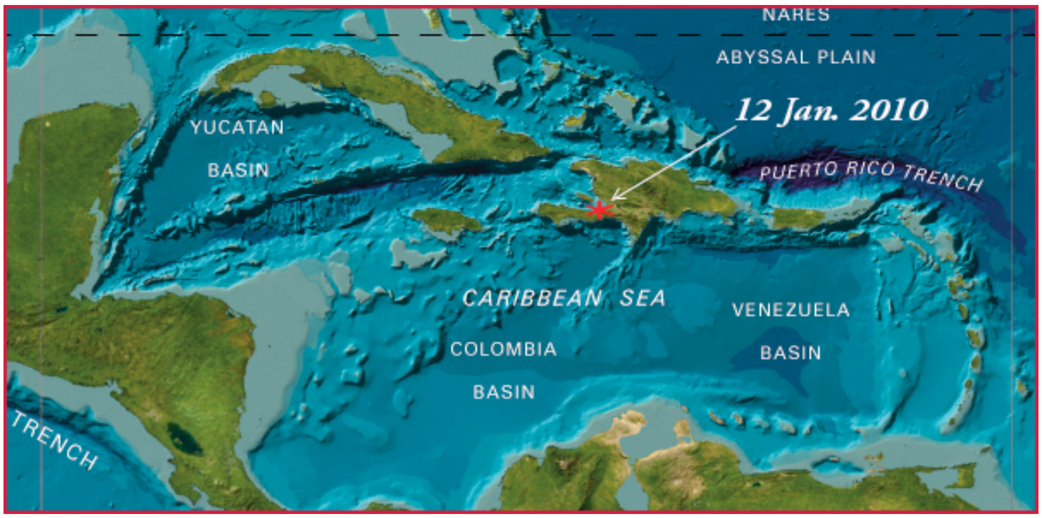

Figure 14. The Northern Caribbean Margin.

5.1.1 Potential Tsunami Impact in the Virgin Islands, Puerto Rico and Hispaniola.

As indicated, the northeastern region of the Northern Caribbean margin is tectonically complex. A great amount of stress has been building up along the segment closer to the Virgin Islands and Puerto Rico caused by about 20 to 30 millimeters of westward movement of the North American plate. Based on localized crustal movements, as well as the oblique convergence between the North American and Caribbean plates, it would be expected that stress is distributed, not only between the inter phases of subduction but also along the major strike-slip faults within the overriding Gonave microplate and the adjacent segment of the Antilles Arc where subduction is more oblique. The next tsunami generated in this northeastern region of the Northern Caribbean margin can be expected to be very destructive since a great deal of development has taken place.

Virgin Islands and Puerto Rico

Continuous underthrusting by the North American plate has also increased the possibility of a major earthquake to occur along the Puerto Rico Trench as well as the likelihood of an outer ridge event to the north of the tectonic boundary - where there is a lot of tectonic deformation and tilting of large fractured, limestone slabs. A major or even a great earthquake along this northeastern boundary has the potential of generating a destructive tsunami in the Virgin Islands and Puerto Rico and could also have significant far-field impact in Florida, the Bahamas and elsewhere. Large-scale landslides could also be triggered and there is evidence that such events have occurred in the past (Fig. 15) that perhaps generated tsunamis.

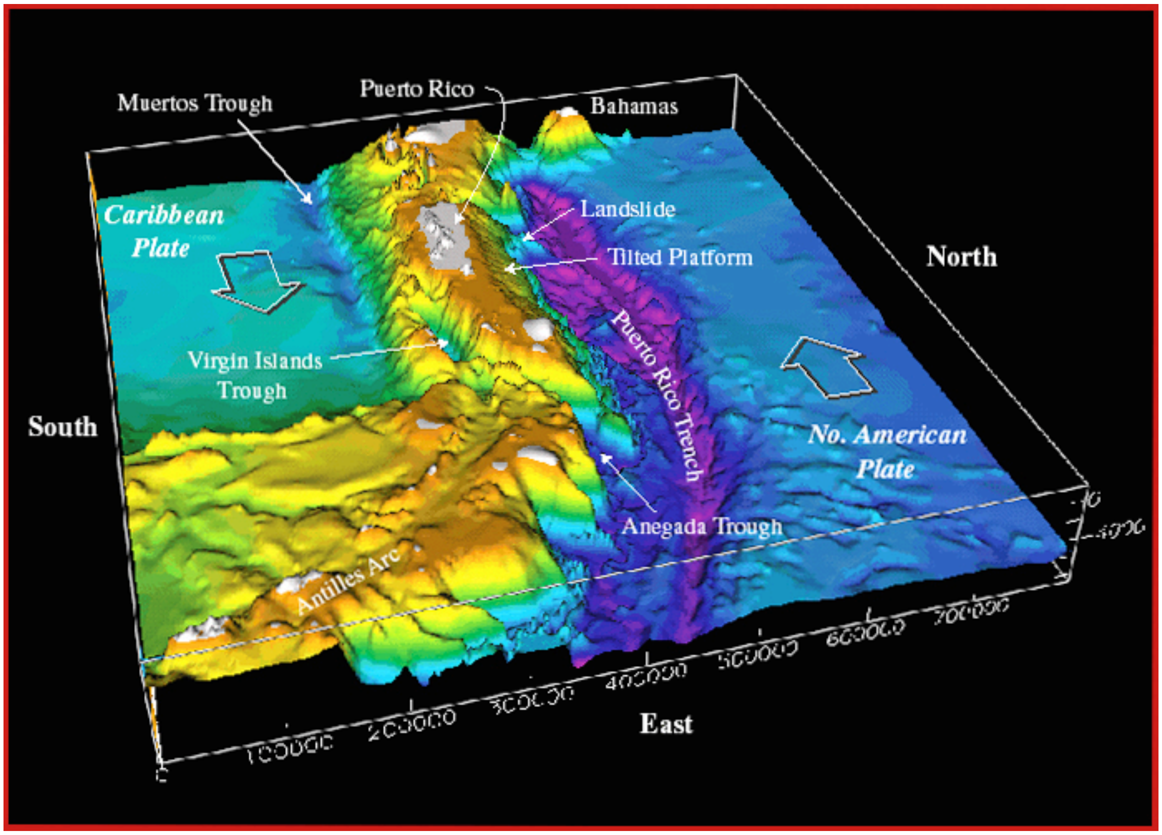

Figure 15. Morphology and bathymetry of the Northeastern segment of the Northern Caribbean Margin (Woods Hole Oceanographic Institute and NGDC graphic)

Additionally, tectonic irregularities in both the Anegada Trough and the Mona Pass regions could also result in major earthquakes. Thus, there is great potential for tsunami generation along the bend of the Antilles Arc closer to the Anegada Trough where the 1867 earthquake occurred (Fig. 15). Such an event is long overdue. This time, the tsunami that will be generated will have a more severe impact in the Virgin Islands because of population growth and development along coastal areas. It could also have far-field impacts in Florida, Bahamas and as far north as the eastern U.S. coast.

Also, as previously indicated, the Mona Pass marks the boundary of the western boundary of the Puerto Rico microplate east of Hispaniola (Figs. 13, and 15). Major earthquakes along this microplate boundary have the potential to generate local tsunamis, particularly in the region closer to Puerto Rico. Another major earthquake similar to the two which occurred in 1918 (M7.5, Epicenter 18.50N - 67.50 W) can be expected to occur in the future. Such earthquake will generate destructive tsunamis that will strike mainly the western coast of Puerto Rico and to a lesser extent the eastern coast of Hispaniola and perhaps the Virgin Islands. Waves with run-up of up to 6 meters can be expected on the western coasts of Puerto Rico, with lesser heights elsewhere in the region.

Hispaniola (Haiti and Dominican Republic)

As previously indicated, the 12 January 2010 earthquake in Haiti did not release totally the strain that has build up along this segment of the Enriquillo-Plaintain Garden fault (EPGFZ) on Hispaniola. Figure 16 (JAXA image, Univ. of Miami, 2010) depicts the 2010 earthquake's rupture and the potential future rupture to the east that may result from stress transference.

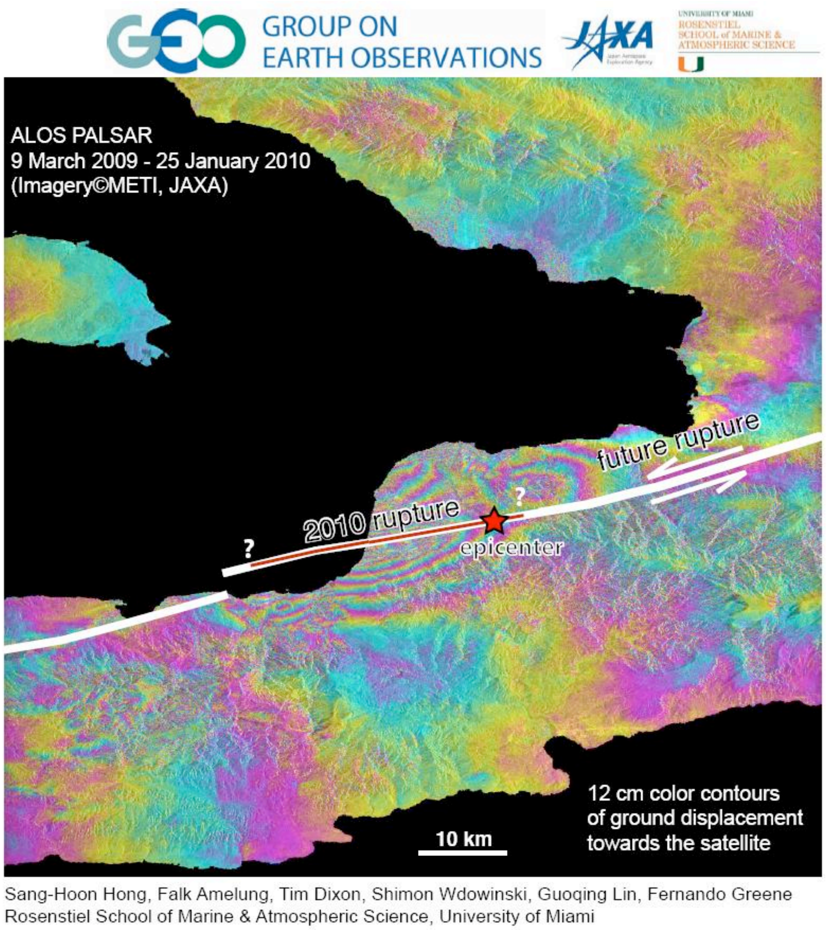

Figure 16. JAXA image illustrating the segment to the east of the 2010 earthquake on the EPGFZ that is likely to have a destructive earthquake in the future.

Coulomb stress is evident to the segments east, west and south of the stricken area. The adjacent segments of the subsurface rupture on the EPGFZ are now near breaking point because of stress transferred to them (Bilham, 2010). Such progression of Coulomb stress occurs along similar strike slip faults, as for example the Northern Anatolian fault in Turkey (Pararas-Carayannis, 1999). Thus a major earthquake similar to the 2010 event is likely to occur again – more likely along the eastern segment where the stress is greater (Ali et al. 2008).

Such an earthquake could result in further transtension (extension and shear) and pull-apart displacement on the fault or its offsets, as in the Mirogoane Lakes in the Gulf of Gonâve. The earthquake epicenter will be closer to Port au Prince, so it could be very destructive, given the population density and poor construction of buildings. Also, crustal extention and shear along offsets will result in strong ground motions that would trigger again landslides that could generate local tsunamis in the Gulf of Gonâve and along the southern coasts of Haiti and the Dominican Republic.

Additional study of Coulomb stress changes in the areas affected by the February 12, 2010 earthquake in Haiti, based on coseismic slip distribution (Fig. 17), is also indicative of stress to the south of the stricken area. An earthquake further south could result in landslides that could generate local tsunamis at the Bay of Jacmel and elsewhere along the southern coast of Hispaniola.

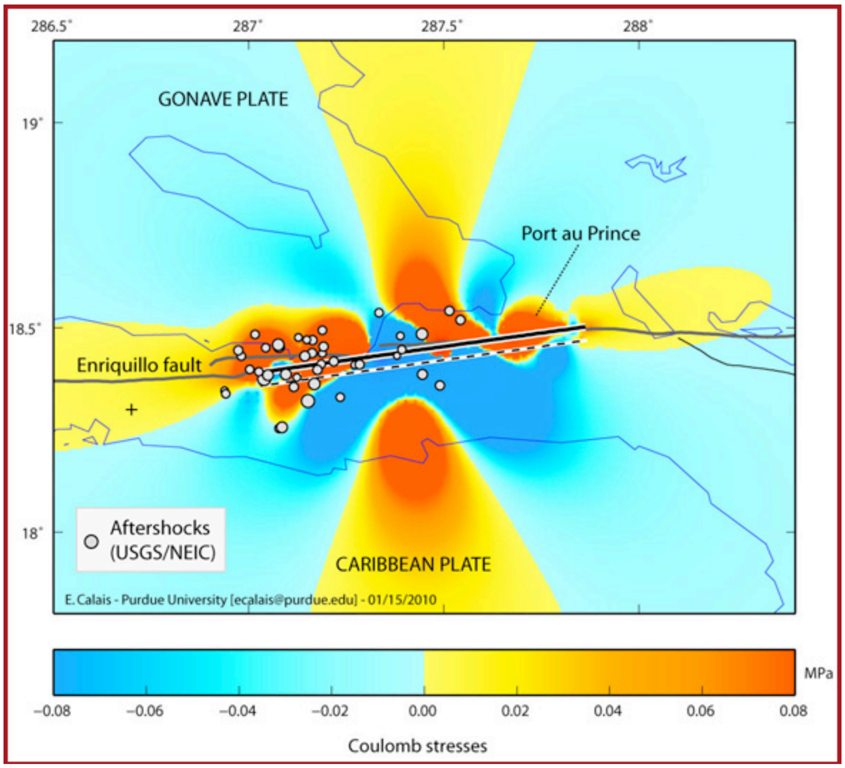

Figure 17. Changes in Coulomb failure stresses caused by the February 12, 2010 Haiti Earthquake, with red indicating regions that have been brought closer to rupture. Calculations assume friction of 0.2 on receiver faults with strike = 90°, dip = 90°, and rake = 0° (i.e., assuming pure left-lateral strike-slip). (Modified from: http://web.ics.purdue.edu/~ecalais/haiti/ based on coseismic slip distribution from G. Hayes (USGS/NEIC)

Along the southern margin of the Gonâve microplate, earthquakes near the subduction region characterized by the Muertos Trench (Figs. 13 &15) have the potential to generate local tsunamis that can impact mainly Hispaniola and Puerto Rico, with lesser effects on the Virgin Islands. Another earthquake similar to the1751 on the Muertos Trench (Fig. 13) is very possible. Also, further west off the southern peninsula of Haiti, there are short subduction zones where earthquakes could generate destructive local tsunamis along the coasts of southern Hispaniola.

Additionally, major or even great earthquakes can occur off the northeast coast of the Dominican Republic along the North Hispaniola Deformation Front (NHDF), or along the North Hispaniola Fault (NHF) or along the Puerto Rico subduction margin (Fig. 13). Earthquakes along the NHDF, or along the NHF, have the potential to generate very destructive tsunamis similar to those of August 4, 1946, that will impact greatly the northern coasts of the Dominican Republic and Haiti but may also have significant far-field impact elsewhere. Based on the historical record and rates of crustal movements, large tsunamigenic earthquakes on the NHDF can be expected on the average every 100 years, but may occur at more frequent intervals. Thus tsunamigenic earthquake on the northeast coast of the Dominican Republic is highly probable.

Finally, and as previously mentioned, earthquakes along the Puerto Rico subduction margin can also generate tsunamis, which can affect the northern coasts of Hispaniola. Also, another earthquake along the Mona Pass similar to the 1918 is very likely to occur again that will affect mainly the eastern coast Hispaniola.

5.2 Potential Tsunamigenic Earthquakes Along the Western Segment of the Northern Caribbean Margin.

5.2.1 Potential Tsunami Impact in Jamaica, Cuba and the Cayman Islands.

Jamaica

The historic record indicates that only two confirmed tsunamis have occurred in Jamaica over the past 300 years from local earthquakes. Accordingly, a destructive earthquake in 1692 that killed 2,000 people at Port Royal resulted in ground subsidence and the generation of a local tsunami in the harbor which reached a maximum height of 1.8 meters (6 feet). Another earthquake along the north coast of Jamaica in 1907 generated a tsunami with waves ranging from 1.8 - 2.4 meters between Portland and St. Ann and was accompanied by 70-90 meter withdrawal of the sea. Waves of 2.5 meters from the same event were observed in Kingston Harbor.

Cuba

Of a total of six purported tsunamis in Cuba (in 1766, 1775, 1852, 1831, 1932 and 1939) only the 1755 event in Santiago De Cuba appears to be an actual tsunami event. All others in Santa Clara, Las Villas and Santiago de Cuba represent questionable tsunamis

Cayman Islands

The western boundary of the Northern Caribbean margin extends for about 1500 km from the Sierra Maestra Mountains of Cuba to the Misteriosa Bank near Belize and the Gulf of Honduras. Movement of the North American tectonic plate along the Caribbean boundary at a velocity of about 20 mm/year has formed the Cayman Ridge and Trough, also known as the Cayman Trench or Bartlett Deep. West of Jamaica and south of Grand Cayman Island group lies a pull-apart basin, a divergent feature known as the Cayman Spreading Center, which represents the broader western boundary of the Gonâve microplate, consisting of two parallel branches separated by approximately 125 km. This west boundary zone of the Northern Caribbean margin is characterized by a relatively simple set of transform faults. The historic record does not document tsunamis of any significance generated by earthquakes in this western segment of the northern margin – mainly because seismic events in the region are of lesser magnitude and involve primarily lateral moments along regional faults. Thus, there is little danger of significant tsunami generation in this western region, or of the Cayman Islands being struck by a significant tsunami. The most recent significant earthquake near the Grand Cayman Island occurred on 14 December 2004 and measured 6.7. Its epicenter was about 30 kms south of George Town and its focal depth was 10 kms. It was reported to be the strongest earthquake to hit Grand Cayman since 1900 and was also felt in Cancun, Mexico and Kingston, Jamaica. No tsunami was generated.

5.3 Recurrence Frequency of Tsunamigenic Earthquakes Along the Northern Caribbean Margin

The overall tsunami recurrence period in the Caribbean has been estimated to average about 19 ± 22 years between deadly events (Proenza & Maul, 2010) - which include tsunamis from volcanic sources such as those caused by debris avalanches and pyroclastic flows of the Soufriere Hills Volcano in Montserrat (Hooper and Mattioli, 2001; Pararas-Carayannis, 2006). A rough evaluation of the cumulative frequency of tsunamis was done for Barbados and Antigua (Zahibo et al, 2003).

Along the Northern Caribbean margin boundary, the historic record of the past 500 years indicates that at least ten destructive tsunamis were generated by earthquakes. Thus, the recurrence frequency for the Northern Caribbean margin averages to one significant tsunami every 50 years. Of the ten known tsunamis in the region, those, which occurred in 1692, 1781, 1842, 1867, and 1918 and in 1946, resulted in deaths and destruction. The most significant of the six were the two tsunamis of 1946. Since 1946, sixty-four (64) years have passed without another significant tsunami in the region. Thus, based on statistical probability alone, another destructive tsunami is long overdue to occur somewhere along the Northern Caribbean margin.

5.4 Potential Tsunami Generation from an Outer Rise Earthquake near the Eastern Segment of the Northern Caribbean Margin

There is geological evidence that large tsunamis were generated in prehistoric times (before 1400 AD) along the northern margin of Puerto Rico that were much larger than any of those known from 500 years of historical records. There is evidence that a major earthquake along the Northern Caribbean plate boundary occurred about 800 years ago. This may have been an outer ridge event. Thus, a repeat of such an event is possible – and perhaps overdue - given the estimated rate of about 20 to 30 millimeters that the Caribbean plate is moving in relation to the North American plate.

REFERENCES

Ali, S. T., Freed, A. M., Calais, E., Manaker, D. M. & McCann, W. R., 2008. *Geophys. J. Int.* **174**, 904–918 (2008). | Article

Bilham R., 2010. Lessons from the Haiti earthquake. *Nature* **463**, 878-879 (18 Feb. 2010)

Calais, E., Mazabraud, Y., de Le'pinay, M. B., Paul Mann, P., Mattioli, G., and Jansma, P., 2002: Strain partitioning and fault slip rates in the northeastern Caribbean from GPS measurements, Geophys. Res. Lett., 29(18), 1856, doi:10.1029/2002GL015397, 2002.

Calais E. and Mercier de Lepinaya, B., 2003. From transtension to transpression along the northern Caribbean plate boundary off Cuba: implications for the Recent motion of the Caribbean plate. Report of Institut de Géodynamique, Université de Nice-Sophia Antipolis, Valbonne, France.

Dixon, T.H., Farina, F., DeMets, C., Jansma, P., Mann, P. and Calais, E., 1998. Relative motion between the Caribbean and North American plates and related boundary zone deformation from a decade of GPS observations, J. Geophys. Res., vol 103, pp 15157-15182.

DeMets, C, and Wiggins-Grandison, M., 2007. Deformation of Jamaica and motion of the Gonave microplate from GPS and seismic data. Geophys. J. Int. (2007) 168, 362–378.

Dolan, J. and Wald, D., 1998: The 1943–1953 northcentral Caribbean earthquakes: Active tectonic setting, seismic hazards, and implications for Caribbean-North America plate motions, in: Active Strike-slip and Collisional Tectonics of the Northern Caribbean Plate Boundary Zone, edited by: Dolan, J. and Mann, P., Geol. S. Am. S., Boulder, 326, 143–169, 1998.

Fritz, H., 2010 (unpublished survey)

Heubeck, C. and Mann, P., 1991, Geologic evaluation of plate kinematic models for the North American-Caribbean plate boundary zone. Tectonophysics, Vol. 191, Issues 1-2, 20 May 1991, Pages 1-26

Hooper, D. M. and Mattioli, G. S., 2001. Kinematic modelling of pyroclastic flows produced by gravitational dome collapse at Soufriere Hills. Natural Hazards, 23, 65–86, 2001.

Lovett, R. A., 2010: Haiti earthquake produced deadly tsunami – waves up to three metres high hit sections of the nation's coastline, Nature News, doi:10.1038/news.2010.93, published online, available at: http://www.nature.com/news/2010/100225/full/news.2010.93.html (last access: 28 May 2010), 25 February 2010.

Mann, P., Prentice, C. S., Burr, G., Pena, L. R., and Taylor, F. W., 1998: Tectonic geomorphology and paleo-seismology of the Septentrional fault system, Dominican Republic, Geol. S. Am. S., 326, 63–123, doi:10.1130/0-8137-2326-4.63, available at: http://www.jsg.utexas.edu/news/pdfs/011310/Mann SFZ 1998.pdf (last access: 28 May 2010), 1998.

Mann, P., Taylor, F. W., Edwards R.L. and Ku, T.-L., 1995. Actively evolving microplate formation by oblique collision and sideways motion along strike-slip faults: An example from the northeastern Caribbean plate margin, Tectonophysics, vol 246, pp 1-69.

Mann, P.; Taylor, F. W. R.; Edwards, L. and Teh- Lung Ku, 2000. Actively evolving microplate formation by oblique collision and sideways motion along strike-slip faults: An example from the northeastern Caribbean plate margin

Mann, P.; Prentice, C.; King, W.; Demets, C.; Wiggins-Grandison, M.; Benford, B., 2008. Late Quaternary Activity and Seismogenic Potential of the Gonave Microplate: Plantain Garden Strike-Slip Fault Zone of Eastern Jamaica. American Geophysical Union, Fall Meeting 2008, abstract #T11B-1869

Mann, P. 2005. Active Tectonics and Seismic Hazards of Puerto Rico, the Virgin Islands, and Offshore Areas (book).

O'Loughlin, K. F and Lander J. F. (2003). Caribbean Tsunamis: A 500-Year History from 1498–1998. Springer. p. 85-86. ISBN 978-1402017179.

Pararas-Carayannis, G. 1999. The Earthquake and Tsunami of August 17, 1999 in the Sea of Marmara, Turkey. http://www.drgeorgepc.com/Tsunami1999Turkey.html

Pararas-Carayannis, G. 2002: Evaluation of the threat of mega tsunami generation from postulated massive slope failures of island stratovolcanoes on La Palma, Canary Islands, and on the Island of Hawaii, *Science of Tsunami Hazards, 20, 5, 251–277, 2002*

Pararas-Carayannis, G., 2006. Risk Assessment of Tsunami Generation from Active Volcanic Sources in the Eastern Caribbean Region - In "CARIBBEAN TSUNAMI HAZARDS"- Proceedings of National Science Foundation Caribbean Tsunami Workshop, March 30-31, 2004, Puerto Rico, Aurelio Mercado-Irizarry - Philip Liu, Editors. ISBN 981-256-545-3, 341 pp. 2006, Hard Cover Book. World Scientific Publishing Co. Pte.Ltd., Singapore.

Pararas-Carayannis, G., 2006a. Assessment of Seismic Generating Source Mechanisms in the Caribbean Region (unpublished report).

Proenza W. X. and Maul G. A., 2010, Tsunami Hazard and Total Risk in the Caribbean Basin, *Science of Tsunami Hazards, Vol. 29, No. 2, pp. 70- 77 (2010)*

Pubellier M., Vilab J. M. and D. Boissonc, 2003. North Caribbean neotectonic events: The Trans-Haitian fault system. Tertiary record of an oblique transcurrent shear zone uplifted in Hispaniola. Report of Département de Géotectonique, Laboratoire de Pétrophysique et de Tectonique, Université Paul Sabatier, Toulouse, France and Bureau des Mines et de l'Energie, B.P. 2174, Port-au-Prince, Haiti.

Seno T. and Yamanaka Y., 1998. Arc stresses determined by slabs: Implications for back-arc spreading. Earthquake Research Institute, University of Tokyo, Geopys. Res. Lett.,Å@3227-3230, 1998

Syed Tabrez Ali, Andrew M. Freed, Eric Calais, David M. Manaker, and William R. McCann, 2008. Coulomb stress evolution in Northeastern Caribbean over the past 250 years due to coseismic, postseismic and interseismic deformation. Geophys. J. Int. (2008) 174, 904–918.

Sykes, L.R., McCann, W.R. and Kafka, A.L., 1982. Motion of Caribbean plate during the last 7 million years and implications for earlier Cenozoic movements, J. Geophys. Res., vol 87, pp 10656-10676.

Taber, S., 1922. The seismic belt in the Greater Antilles, Bulletin of the Seismological Society of America, vol 12, 199-219.

Theilen-Willige, B., 2010. Detection of local site conditions influencing earthquake shaking and secondary effects in Southwest-Haiti using remote sensing and GIS-methods, Nat. Hazards Earth Syst. Sci., 10, 1183–1196, 2010 www.nat-hazards-earth-syst-sci.net/10/1183/2010/

Tomblin, J. (1981). "Earthquakes, volcanoes and hurricanes A review of natural hazards and vulnerability in the West Indies." Ambio 10(6): 340-345.

Tuttle, M. P., Prentice, C. S., Dyer-Williams, K., Pefla, L. R., and Burr, G., 2003: Late Holocene Liquefaction Features in the Dominican Republic: A Powerful Tool for Earthquake Hazard Assessment in the Northeastern Caribbean, B. Seismol. Soc. Am., 93(I), 27–46, available at: http://www.jsg.utexas.edu/news/ pdfs/011310/Tuttle et al 2003 DR.pdf (last access: 28.05.2010), February 2003.

Scherer, J., 1912. "Great earthquakes in the island of Haiti". Bulletin of the Seismological Society of America 2: 174–179.

Zahibo N., Pelinovsky E., Yalciner A. C., Kurkin A., Koselkov A., and Zaitsev A., 2003. The 1867 Virgin Island Tsunami. Natural Hazards and Earth System Sciences (2003) 3: 367–376

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