Historical perspective on seismic hazard to Hispaniola and the northeast Caribbean region

Uri S. ten Brink[1*], William H. Bakun[2], and Claudia H. Flores[1]
utenbrink@usgs.gov
[1]U.S. Geological Survey, Woods Hole, Massachusetts
[2]U.S. Geological Survey, Menlo Park, California
*Corresponding author: utenbrink@usgs.gov

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Abstract

We evaluate the long-term seismic activity of the North-American/Caribbean plate boundary from 500 years of historical earthquake damage reports. The 2010 Haiti earthquakes and other earthquakes were used to derive regional attenuation relationships between earthquake intensity, magnitude, and distance from the reported damage to the epicenter for Hispaniola and for Puerto Rico and the Virgin Islands. The attenuation relationship for Hispaniola earthquakes and northern Lesser Antilles earthquakes is similar to that for California earthquakes, indicating a relatively rapid attenuation of damage intensity with distance. Intensities in Puerto Rico and the Virgin Islands decrease less rapidly with distance.

We use the intensity-magnitude relationships to systematically search for the location and intensity magnitude M_I which best fit all the reported damage for historical earthquakes. Many events occurred in the 20th-century along the plate-boundary segment from central Hispaniola to the NW tip of Puerto Rico, but earlier events from this segment were not identified. The remaining plate boundary to the east to Guadeloupe is probably not associated with M>8 historical subduction-zone earthquakes. The May 2, 1787 earthquake, previously assigned an M 8-8.25, is probably only M_I 6.9 and could be located north, west or SW of Puerto Rico. An M_I 6.9 earthquake on July 11, 1785 was

probably located north or east of the Virgin Islands. We located $M_I < 8$ historical earthquakes on April 5, 1690, February 8, 1843, and October 8, 1974 in the northern Lesser Antilles within the arc. We speculate that the December 2, 1562 (M_I 7.7) and May 7, 1842 (M_I 7.6) earthquakes ruptured the Septentrional Fault in northern Hispaniola. If so, the recurrence interval on the central Septentrional Fault is ~300 years, and only 170 years has elapsed since the last event. The recurrence interval of large earthquakes along the Hispaniola subduction segment is likely longer than the historical record. Intra-arc M ≥ 7.0 earthquakes may occur every 75-100 years in the 410-km-long segment between the Virgin Islands and Guadeloupe.

1. Introduction

Hydrodynamic models show that large earthquakes on the North America/Caribbean subduction zone have the potential to cause trans-oceanic tsunamis that will impact the U.S. East Coast, western Europe, and the nearby Caribbean islands (e.g., *Geist and Parsons*, 2009]. Although the potential magnitude of intra-arc Caribbean earthquakes may be smaller than subduction-zone events, the seismic risk may be greater because of their shallow depth and proximity to Caribbean-region population centers. The 500-year written history of the Caribbean includes accounts of devastating earthquakes, volcanic eruptions, and tsunamis [*Perrey*, 1857] that can be used to quantify the seismic hazard of the region.

The Caribbean islands were discovered and named by Columbus in 1492-1498 [*Morison*, 1942] and were quickly populated, first by Spanish and then by other nationalities. European migration to these islands and the international struggle over the control of the islands and the sea routes during the 16th -18th centuries, have resulted in rich written records in the form of bureaucratic reports and letters from the islands to the mother countries. These reports and letters sometimes include descriptions of damages from hurricanes and earthquakes, often accompanied by requests for money to rebuild damaged property (e.g., *de Utrera*, [1995] quoting Archivo General de las Indias, Santo Domingo, AGI-IG 95). The record is most complete in Hispaniola (present-day Haiti and the Dominican Republic) and Puerto Rico, which were settled at the end of the 15th century; records are sparser and start later in the smaller islands.

In this paper we compile reports of damage by significant earthquakes in Hispaniola, Puerto Rico, the Virgin Islands, and the northern Lesser Antilles in the past 500 years to understand the long-term seismic activity of the trench and arc regions. The reader is referred to *Bakun et al.* [2012] for a discussion of historical earthquakes on the Enriquillo Fault in southern Hispaniola. We focus on several key issues: 1) Variations in earthquake intensity between different islands, which may indicate different hazard for a given earthquake magnitude; 2) The implications to earthquake and tsunami hazards of the general absence of subduction earthquakes along most of the Puerto Rico trench; 3) The recurrence interval of major earthquakes along the Septentrional fault, the principal strike-slip fault across northern Hispaniola.

2. Tectonic setting

Cretaceous-age North American lithosphere presently subducts under the Puerto Rico trench, a 1300-km-long section of the Caribbean island arc from central Hispaniola to Guadeloupe (Fig. 1). Because of the direction of motion (250°) relative to the plate boundary, subduction is nearly arc-parallel north of the Dominican Republic, Puerto Rico, and the Virgin Islands, and is more arc-perpendicular along the Lesser Antilles, where it is accompanied by arc volcanism. The convergence rate is 19-20 mm/y [*Mann et al.*, 2002]. *Geist and Parsons* [2009] estimated a recurrence interval of ~1000 years for M 8.5 earthquakes and ~4000 years for M 9 earthquakes. The 500+ years of historical record can be supplemented using tsunami deposits and liquefaction features as evidence for earlier earthquakes. To date, evidence of a large tsunami overwash has been found at only one location along the north shores of the Antilles Islands. This overwash event, found on the island of Anegada, British Virgin Islands (A in Fig. 1), dates between AD1650-1800 [*Atwater et al.,* 2010].

Intra-arc fault ruptures can cause great damage in the Greater and Lesser Antilles islands, as was demonstrated by the January 12, 2010 M 7.0 Haiti earthquake. The November 18, 1867 Virgin Islands earthquake [*Reid and Taber*, 1920] is another example of a moderately large earthquake (M ~7.2, *Barkan and ten Brink*, 2010) on an intra-arc fault that generated a devastating tsunami and delayed the purchase of the U.S. Virgin Islands (USVI) from Denmark by 50 years [*Dookan*, 1994]. Although the potential magnitude of intra-arc earthquakes may be smaller than along the subduction zone, their seismic risk may be more severe because of their shallow depth and proximity to population centers.

One of the largest intra-arc faults is the Septentrional fault system, which accommodates a significant portion of the left-lateral component of the oblique convergence between the North American and Caribbean plates. The Septentrional fault system underlies Santiago (Santiago de los Caballeros), the second largest city in the Dominican Republic. *Prentice et al.* [2003] suggested from analyzing trenches across the fault that the last large earthquake on the Septentrional fault near Santiago occurred 800-1000 years ago. GPS-based kinematic models suggest strike-slip accumulation at a rate of 12.3 mm/y along the fault system [*Calais et al.*, 2010]. Geomorphic features both offshore and onshore indicate that the fault system is primarily strike-slip with some local vertical scarps with alternate directions [*Mann et al.*, 1998; *ten Brink and Lin*, 2004] that may indicate local components of normal faulting [*Mann et al.*, 1998]. The fault systems split into two branches in west-central Hispaniola (Fig. 1). The southern branch crosses the Cibao valley where its trace is largely obscured by fluvial sedimentation and erosion [*Mann et al.*, 1998] and continues along the northern coast of Haiti to the Windward Passage [*Calais and Mercier de Lepinay*, 1995]. From there it connects to the Oriente fault system along the southern coast of Cuba to the Cayman spreading center. The northern branch extends for 70 km along the southern edge of the Cordillera Septentrional mountain-front [*Mann et al.*, 1998] and probably terminates offshore [*Dillon et al.*, 1996]. The northern branch does not show evidence for late Quaternary activity and may no longer be active [*Mann et al.*, 1998].

3. Methodology

We have compiled a database of damage descriptions from primary sources and 19th century catalogs [*Flores et al.*, 2011], and assigned our own intensities to the descriptions of the historical earthquakes following the criteria listed in Table 2 of *Bakun et al.* [2012], and did not use the intensities listed in previous catalogs. Previous attempts at locating and estimating the magnitudes of historical earthquakes in the region were based on the location and intensity assignment of the most severe damage [*McCann*, 2006]. Here we follow the method of *Bakun and Wentworth* [1997], which uses a training set of instrumentally-recorded earthquakes with damage reports to derive regional relationships between earthquake intensity, magnitude, and distance from the reported damage to the intensity center. These relationships are then used to estimate the magnitude and intensity center of historical earthquakes by searching a grid of trial epicenters. Errors are objectively evaluated. Because the method uses damage reports, an intensity center is determined, rather than the epicenter.

Bakun et al. [2012] used 93 intensity assignments from the 2010 Haiti earthquakes (M7.0) and two aftershocks (M5.9, and M4.7) to estimate the intensity attenuation for Haiti. A regression on these assignments yielded the relation:

$$MMI = -(1.69 \pm 0.81) + (1.70 \pm 0.19) * \mathbf{M} - (0.00165 \pm 0.00054) * \Delta_h - (2.13 \pm 0.34) * \log_{10}(\Delta_h), \quad (1)$$

where **M** is moment magnitude and Δ_h is the distance in kilometers of the MMI site from a point source at h = 10 km depth. The intensity attenuation relation (1) is approximately similar to that obtained for southern California [*Bakun*, 2006](Fig. 2b) and can be used to provide unbiased estimates of location and magnitude for crustal and subduction zone earthquakes throughout Hispaniola.

We use (1) to estimate M from individual intensity observations for a trial epicenter [*Bakun and Wentworth*, 1997]. That is,

$$M_I = \text{mean}(M_i), \quad (2)$$

where

$$M_i = \{(MMI_i + 1.69 + 0.00165\Delta_{h,i} + 2.13 \log_{10}(\Delta_{h,i})\}/1.7, \quad (3)$$

MMI_i, and $\Delta_{h,i}$ are the MMI value, and the hypocentral distance, respectively, at site i. We find the misfit for each trial epicenter from

$$rms[M_I] = [rms(M_I - M_i) - rms_0(M_I - M_i)], \quad (4)$$

where rms $(M_I - M_i) = \{\Sigma_i[W_i(M_I - M_i]^2/\Sigma_i W_i^2\}^{1/2}$, $rms_0(M_I-M_i)$ is the minimum rms (M_I-M_i) over the grid of trial epicenters, and W_i is the distance-weighting function [Bakun and Wentworth, 1997]

$$W_i = \begin{cases} 0.1 + \cos[(\Delta_i/150)(\pi/2)] & \text{for } \Delta_i < 150 \text{ km} \\ 0.1 & \text{for } \Delta_i > 150 \text{ km} \end{cases} \quad (5)$$

The preferred intensity center (green triangle in Figures 3-6) is the trial source location for which rms $[M_I]$ is minimum [*Bakun*, 1999] and corresponds more to the moment centroid than to the epicenter. The rms $[M_I]$ contours (red contours in Figures 3-6) bound the intensity center region and are associated with confidence levels that the epicenter is located within the contour (heavy and dashed green lines)[*Bakun and Wentworth*, 1997]. The M_I at the intensity center is the best estimate of moment magnitude M for that earthquake. Uncertainties in M appropriate for the number of intensity assignments are also estimated at the 68% and 95% confidence levels (Table 2)[*Bakun and Wentworth*, 1999].

The bootstrap data resampling strategy has been developed to provide estimates of the uncertainty of model parameters estimated from a given finite data set. In the bootstrap re-sampling strategy, n random samples are drawn with replacement from a set of n observations. For example, consider a data set with three observations: A, B, and C. For the data set, there are 9 possible bootstrap resampling sets: AAA; BBB; CCC; AAB; AAC; ABB; ACC; BBC; and BCC. The bootstrap resampling approach is particularly useful because it has been shown [*Efron*, 1982] that the statistical properties of the family of bootstrap resampled sets is identical to the statistical properties of the original data set. For a data set of 4 points, such as the1562 earthquake (Fig. 3a), the variance may be larger than the expected value, so the bootstrap method may not be very reliable. The bootstrap resampling distributions for location and magnitude are presented in Figures 3-6. We interpret divergence in the locations and magnitudes between the bootstrap analysis and the original grid search to indicate solutions that are not well constrained.

4. Results: variations in intensity attenuation

We used 19 modern earthquakes in the NE Caribbean with intensity assignments (Table 1) to test the Haiti attenuation model (equation 1) across the region. Intensity assignments for events after the year 2000 are taken from the USGS Earthquake Hazards Program (Did you Feel It (DYFI), http://earthquake.usgs.gov/dyfi) as real numbers, not rounded integers. DYFI questionnaires were designed to match the descriptions of MMI intensities and are assumed to be equivalent (D. Wald, written comm., 2011). Damage reports from events before 2000 were compiled from different sources [*Bodle and Murphy*, 1984; *Coffman and von Hake*, 1984a; *Coffman and Stover*, 1984; *Coffman and von Hake*, 1984b; *Stover and Brewer*, 1991]. The number of locations of intensity reports for these calibration events varies from 3 to 55. Earthquakes in Hispaniola and the northern Lesser Antilles span a wide magnitude range (M_w 4.3-7.6; M_s where M_w is not available). Calibration events from Puerto Rico and the Virgin Islands span a more limited magnitude range (M 3.4-5.8).

Intensity magnitudes, M_I, obtained with equation (1) for earthquakes in Hispaniola and the northern Lesser Antilles are consistent with M_w. (Fig. 2a). That is, the Haiti intensity attenuation relationship (equation 1) is appropriate for earthquakes throughout Hispaniola and for earthquakes in the northern Lesser Antilles. The M_I calculated using the Haiti intensity attenuation relation are larger than the instrumental magnitudes of earthquakes near Puerto Rico and the Virgin Islands by about 1.0 magnitude units (Fig. 2a). The consistency of the magnitude mismatch suggests that the Haiti intensity attenuation is not appropriate for earthquakes near Puerto Rico and the Virgin Islands.

We therefore used regression analysis for the Puerto Rico and Virgin Islands (PR-VI) events to obtain an intensity attenuation model that appears to be appropriate for Puerto Rico and the Virgin Islands

$$MMI = -1.06 + 1.45*\mathbf{M} - 0.00136*\Delta_h - 1.3*\log_{10}(\Delta_h), \qquad (6)$$

The calculated M_I using (6) are consistent with the instrumental magnitudes for M ≥ 5.0 earthquakes there (Table 1). For M<5.0, M_I tends to over-estimate the earthquake magnitude, and this may be due to either an inaccurate intensity model or to the inaccuracy in the instrumental magnitudes for the small earthquakes. We assume that equation (6) is also appropriate for larger (M >6) earthquakes in Puerto Rico and the Virgin Islands. The individual MMI residuals (observed MMI – calculated MMI) on average do not depend on epicentral distance, providing an independent support for the relationship in equation 6.

The intensity attenuation relation for PRVI earthquakes, equation (6), is intermediate between the intensity attenuation relation for the eastern United States (ENA-SCR) and California (Fig. 2b). For a given earthquake magnitude, the intensity drops off with increasing distance more slowly in Puerto Rico and the Virgin Islands than for Hispaniola and the Lesser Antilles (Fig. 2b). The PRVI model is consistent with the Quality Factor (Q) for Puerto Rico being intermediate between that for California and eastern North America [*Motazedian and Atkinson*, 2005]. Their Q was derived from modern M 3.0-5.5 earthquakes.

5. Analysis of historical earthquakes

5.1. Hispaniola and Mona Passage

The first reported severe earthquake in Hispaniola took place in what is now the northern Dominican Republic in 1562 [*de Utrera,* 1995, citing primary sources]. The date of this earthquake is controversial. A date of November 2, 1564, mentioned in an undated letter by Echagoian to King Philip II of Spain, was later adopted by others [*Charlevoix*, 1731; *del Monte y Tejada*, 1890; *García*, 1900; *Moreau de Saint-Méry*, 1796; *Poey*, 1857; *Scherer*, 1912; *Southey*, 1827; *McCann*, 2006]. *Garcia* [1900] also mentioned April 20, 1564. *De Utrera* [1927] noticed the discrepant 1562 and 1564 dates and argued that Echagioan wrote his letter in early 1568 after leaving Hispaniola. *De Utrera* [1927] consulted the Archivo General de Indias (AGI) in Seville and wrote "*In the Indies Archives can be found the following papers: Letter from the honorable Herrera to His Majesty, in his Real Consejo de Indias, over various matters and among them news of an earthquake that occurred December 2 of the year before, which is dated February 16, 1563. Another letter dated February 13, 1563 co-written by the Honorable Herrera and the Honorable Echagoian and by the doctor Caceres to His Majesty, in their Real Consejo de Indias, over the earthquake that occurred on December 2 of the year before,* *between eight and nine at night, which resulted in the fall of the Catheral de la Vega. Another letter from the clergyman Cabildo from Conception to His Majesty in his Real Consejo de Indias, over the destruction caused by the earthquake of December 2, 1562, and is a letter that contains the date October 6, 1563*". We follow *de Utrera* [1927] in assigning the date of December 2, 1562 to the earthquake.

The 1562 earthquake completely destroyed Santiago de Los Caballeros [*del Monte y Tejada*, 1890]. That town was located on the Septentrional fault trace [*Mann et al.,* 1998] and was subsequently abandoned and rebuilt 10 km to the SE. Most of the city of Concepcion de La Vega (also known as La Vega), 35 km SE of present-day Santiago was destroyed [*Moreau de Saint Mery*, 1796], and the city was abandoned. The church and Franciscan monastery, which were built of masonry, partly reinforced by iron bars, were almost completely destroyed [*Scherer,* 1912]. The convent and dormitory in Puerto Plata, the only brick and stone buildings in that town, were severely damaged [*de Utrera*, 1927], as were several buildings in Santo Domingo that were built of weak masonry [*de Utrera*, 1927].

We could not find damage reports from eight additional towns that are shown on a map from that era [*Ortelius*, 1579] as having churches, monasteries, or other types of large buildings. Only one of these towns (Puerto Real, 12 km SE of Cap Haitien) was located along the northern part of the island, and there is no archeological evidence for severe damage during the 1562 earthquake. However, the town was forcibly burned and abandoned in 1578 by the Spanish authorities because of the failure of the population to restrict contraband trade, so it is difficult to identify earthquake damage occurring 16 years earlier (K. Deagan, written comm., 2011). An eyewitness on a ship at Monte Cristi 100 km west of Santiago saw the earth shake ashore in 1562 (AGI -IG 1002, 13 May 1563).

Our intensity location for the 1562 earthquake is between Santiago and La Vega about 15 km south of the Septentrional fault near Moca, and the intensity magnitude M_I is 7.7 (Fig.

3a). Bootstrap analysis gives similar preferred location and magnitude estimates, but also allows for an alternative location 20 km to the southeast.

Damage from several earthquakes during the 17th century was reported in Santo Domingo and old Azua (Table 2). Reports for these events are generally available only from these towns, probably because of the abandonment of the northern and western coasts of Hispaniola in 1606 by a decree of the king of Spain [*Charlevoix*, 1731; *Southey,* 1827; *García*, 1900], and the general decline in Spanish population during the 17th century. On the other hand, there were still a total of 13 towns in central, southern, and eastern Hispaniola during that time [*Charlevoix*, 1731], and a map of Hispaniola [*Mercator*, 1628] shows 4 towns with stone churches or other buildings in south and southeast Hispaniola, 4 in north-central Hispaniola, and 4 in the Cordillera Central. French settlement, in what is now Haiti, started around 1670 [*Charlevoix*, 1731]. In any case, the locations and magnitudes of the M_I 6.5-7.5 earthquakes in 1615, 1665, 1673, 1684, and 1691 (Table 2) cannot be confidently determined with the available data, but they appear to have occurred near the south coast of the Dominican Republic.

The May 7, 1842 earthquake caused extensive damage all along northern Hispaniola. "In a minute at most, the cities of Cap Haitien, Port-de-Paix, Mole St. Nicolas, Fort Liberté and Saint-Yague (Santiago) became a heap of rubble. In Cap Haitien, it lost about 5,000 souls half of the population; in Port-de-Paix about 200; at Saint-Yague (Santiago) 200; in other places a little less" [*Ardouin*, 1860]. Based on newspaper reports [L'Ami de la religion, June 28, 1842; The Public Ledger, July 15, 1842; Journal de la Drome, June 22, 1842; el Constitusional, July 9, 1842] and a compilation of damage by *Scherer* [1912], we estimate the highest intensity (IX) to have been in Fort Liberté, Cap Haitien, Port de-Paix, and Mole Saint Nicolas along the northern coast of Haiti and in Santiago and Hato del Yaque (8 km west of Santiago) in north-central Dominican Republic. The intensity was also high (VIII) in Puerto Plata, and in La Vega, 30 km north and 30 km southeast of Santiago, respectively, and in Monte Cristi (VII-IIX). The earthquake intensity diminished to the west to V in Santiago de Cuba and III in Jamaica. To the east, the intensity continued to be high (VII) within the area 50 km east of Santiago in the

Dominican Republic, and in easternmost Dominican Republic (VI-VII in Higuey and Seibo), and decreased to V in western Puerto Rico. The intensity decreased more rapidly south and north of the fault zone. To the south, we estimate intensity VII at Cotui, Dominican Republic, VI 1/2 in Santo Domingo, VII at Gonaives, Haiti, VI at St. Marc, Haiti, V at Port Au Prince, and III – IV in the southern Peninsula of Haiti. North of the fault zone, we estimate IV at Cockburn Town, Turks and Caicos. Aftershocks were felt from Mayaguez, Puerto Rico to Port-au-Prince, Haiti [*Flores et al*., 2011].

The intensity center of the 1842 earthquake is 10 km west of Santiago and its intensity magnitude, M_I is 7.6 (Fig. 3b). Bootstrap analysis gives similar magnitude and location, but also permits alternate locations (Fig. 3b).

The last moderately large earthquake located on or near the Septentrional fault took place on September 23, 1887. Damage was most severe at Mole St. Nicolas at the northwestern tip of Haiti, where liquefaction and perhaps a tsunami occurred [*Tippenhauer*, 1893; *Scherer*, 1912]. Damage was less severe at Port de Paix, Cap Haitien, and Gonaives to the east, and Santiago de Cuba to the west. The earthquake was felt in Manzanillo, Cuba, and Kingston, Jamaica to the west [*Anonymous*, 1887a; *Anonymous*, 1887b; *Hall*, 1922]. Damage was mentioned but not specified in Santo Domingo [*Scherer*, 1912]. We have assigned intensities only to damage levels from ground shaking, not ground failure and tsunami. Damage at Mole St. Nicholas appears to include liquefaction (ground failure) and/or tsunami damage, and has therefore been discarded from our analysis. In a solution that excludes the reports from Mole St. Nicholas, the intensity center is on or near the Septentrional fault between Port de Paix and Cap Haitien, with M_I 6.7 (Fig. 3c), but a solution that includes Mole St. Nicholas (intensity VIII) gives an intensity center in the Dominican Republic and M_I 7.2. There are no damage reports related to this earthquake from the Dominican Republic.

Several significant earthquakes occurred along the north coast of Hispaniola and Mona Passage starting in 1897. The December 29, 1897 earthquake caused serious damage in north-central Hispaniola, including “irreparable damage” to the governor’s residence, a cathedral, and a chapel in Santiago [*Agamennone*, 1898], and to the cathedral in Altamira [*Anonymous*, 1898]. It destroyed buildings in Puerto Plata and damaged the railroad there (*Tomblin and Robson* [1977], quoting Jamaica Post, 1898). The submarine cable tore in Puerto Plata [*Agamennone*, 1898]. The 1897 earthquake was felt in south and SW Hispaniola (Jacmel, Port Au Prince, Santo Domingo). The intensity center is near Puerto Plata, 10 km south of the northern Dominican Republic coast, and M_I is 6.5 (Fig. 4a). *Abe* [1994] proposed an instrumental magnitude 6.8 and an epicenter located 250 km to the southwest, but instrumental locations from that time can be a few hundreds of kilometers in error (W.H.K. Lee, personal communication, 2011). The 1897 event is similar in size and location to the September 12, 2003 Mw 6.4 Puerto Plata shallow thrust earthquake [*Dolan and Bowman*, 2004].

The April 23, 1916 23:36L (April 24, 1916 04:26:42 UT) earthquake occurred in a sparsely-populated area at the northeast tip of Hispaniola, resulting in relatively little damage. However, the earthquake was felt from St. Thomas, USVI, in the east to Grand Turk Island in the northwest, Port-de-Paix, in northwestern Haiti, and to the southern coast of the southern Peninsula of Haiti, roughly 350-450 km from the earthquake location. Our analysis places the earthquake under Samana Peninsula with M_I 6.8 (Fig. 4b). *Gutenberg and Richter* [1954] located the earthquake 160 km to the SE in the middle of Mona Passage at a depth of 80 km (Fig. 4b). They determined a magnitude of m_B7.0. *Doser et al.* [2005] estimated a magnitude M_w 6.8 and a reverse fault mechanism with location at the SE tip of Hispaniola (Fig. 4b) at a depth of 16±7 km. Instrumental locations of earthquakes from that time can be 100-200 kilometers in error because of the small number of recording stations, their poor frequency range, and their large distance from the epicenter (W.H.K. Lee, personal communication, 2011). The intensity analysis, on the other hand, is based on 24 nearby reported locations (Fig. 4b).

M >6.5 earthquakes continued in northern Hispaniola and Mona Passage in the 20th century (Table 3). We analyzed the August 4, 1946 earthquake, the largest of a series of six earthquakes (M_s 7.0 - 8.1) that occurred between 1943 and 1953 along the subduction zone from central Hispaniola to the northwest corner of Puerto Rico [*Dolan and Wald*, 1998; *Kelleher et al.*, 1973]. Five of these events had predominantly reverse mechanisms and are thought to represent stress release on the subduction interface [*Dolan and Wald,* 1998]. The epicenter of the M 7.8-8.1 August 4, 1946 was located on land 22 km south of the Septentrional strike-slip fault (Fig. 4c; [*Kelleher et al.*, 1973]). The earthquake was accompanied by a tsunami that drowned nearly 100 people is the village of Mantanzas [*Lynch and Bodle*, 1948] 100 km to the west-northwest of the epicenter. The intensity center for this event is located at the tsunami location, ~100 km WNW of the instrumental epicenter (Fig. 4c). The intensity magnitude is M_I 7.8±0.2.

5.2. Puerto Rico and the Virgin Islands

In contrast to the large earthquakes along northern Hispaniola and Mona Passage, only moderate-size earthquakes have occurred in the 20th century north of Puerto Rico and the Virgin Islands (http://www.globalcmt.org/CMTsearch.html; [*Doser et al.*, 2005; *Engdahl and Villaseñor*, 2002). The only two known large earthquakes in this area occurred in 1785 and 1787. *McCann* [1985] estimated an M 8-8.25 for the May 2, 1787 earthquake and assumed a location in the Puerto Rico trench. His magnitude estimate was based on reports of serious damage to masonry throughout Puerto Rico, particularly along the north shore [*McCann et al.,* 2011]. It is difficult to locate this earthquake, because it was felt only in Puerto Rico. The intensity center is near the Muertos Trough, southwest of the island (Fig. 5a), although the most severe damage was reported along the northern coast of the island [*McCann et al.,* 2011]. The solution was pushed away from the north coast of Puerto Rico because of the mixture of lower and higher intensities there, which perhaps reflects varying soil conditions. The bootstrap analysis for this event locates the earthquake along the north coast, because it draws many subsets without the small intensities. Regardless, the solution in Fig. 5a should be interpreted as having almost equal probability to be located SW or north of the island. From geological considerations, one possible intensity center is north-northeast of Puerto Rico under Main Ridge, an aseismic ridge that appears to have subducted starting ~3.3 m.y. ago [*ten Brink*, 2005]. Using equation (6), the intensity attenuation relationship for Puerto Rico and the Virgin Islands, M_I is 6.9 for both a Muertos Trough and a Main Ridge location (Fig. 5a). Because the location is not well constrained, the magnitude could range from 6.4, if located under the north coast of Puerto Rico, to 7.3, if located under the outer rise north of the trench. Our analysis indicates that the magnitude of the 1787 earthquake was less than 7 1/2. This conclusion is qualitatively supported by the lack of felt or damage reports from the surrounding islands [*McCann et al.,* 2011].

The July 11, 1785 earthquake was felt most strongly (VI 1/2) in Virgin Gorda at the eastern end of the British Virgin Islands (BVI), less strongly in Tortola, BVI, and Antigua, (V) and least strongly in St. Kitts and St Eustasia (IV). An earthquake was felt that day in northern Haiti [*Moreau de Saint-Méry*, 1796], but it is not clear whether it was the same earthquake. The earthquake might have been accompanied by a tsunami, although the descriptions are equivocal. In Spanish Town, Virgin Gorda, “there was uncommon agitation of the sea” (*McCann et al.,* 2011, quoting the Times of London), and “The island of Tortola, which was swept over during this convulsion by an earthquake wave” [*Perrey*, 1847; *Shaler*, 1869]. It is likely that the source was located north or northeast of the BVI because the earthquake was not reported felt in St. Croix, and was felt less strongly in the western volcanic chain of the Lesser Antilles islands (St. Kitts and St. Eustasia) than in the eastern (Antigua) and northern islands (BVI). Our preferred location is north of Tortola with M_I 6.9 (Fig. 5b). Sea floor maps show large outer-rise normal faults northeast of the Virgin Islands [*ten Brink et al.,* 2004]. If the earthquake occurred on one of these faults, its magnitude could have been as high as 7.2.

5.3. The Lesser Antilles

Three large events have occurred in the northern Lesser Antilles islands since Europeans settled these islands in the first half of the 17th century. The April 5, 1690 earthquake was felt from St. Thomas, USVI to Barbados. It was felt most strongly (IX) in Nevis and almost as strongly in the adjacent islands of St. Kitts and Montserrat, all along the western chain of the Lesser Antilles and in Antigua on the eastern chain. We located the earthquake near Nevis with a magnitude M_I 7.5 (Fig. 6a). The intensity center is 280 km from the trench, where the slab interface is about 200 km deep [*Feuillet et al.*, 2002], suggesting that it may have been an intra-arc event and not a subduction zone event. Comparison of the locations of tsunami reports to locations of predicted flooding also suggests that the 1690 earthquake was not a subduction event. Tsunami modeling from a subduction zone event predicts significant flooding along the trench-facing islands of Antigua and Barbuda (Fig. 6d), where no flooding was reported (despite these islands being populated at the time). In contrast, tsunami was reported from Nevis and Charlotte Amelie on the south side of St. Thomas and in Nevis [*O'Loughlin and Lander*, 2003], where negligible flooding is expected from a subduction zone event (Fig. 6d). Hence, the observations support intra-arc origin for the 1690 earthquake.

The February 8, 1843 event was the largest historical earthquake in the northern Lesser Antilles with an estimated magnitude of M 7.5 - 8 [*Bernard and Lambert*, 1988]. We assigned MMI IIX-IX to damage reports from various locations in Guadeloupe and Antigua, and MMI VII-IIX to damage reports from various locations in St. Kitts, Nevis, and Dominica [*Flores et al.*, submitted]. Although there was one tsunami report from Barbados, the earthquake did not produce a tsunami or a noticeable vertical deformation in Guadeloupe or Antigua [*Bernard and Lambert*, 1988]. Using our intensity assignments we determined an M_I 7.8 with the intensity center located beneath Guadeloupe (Fig. 6b).

The October 8, 1974 Ms 7.1-7.6 earthquake was located near Barbuda on a southeast dipping normal fault above the subduction zone [*McCann et al.*, 1982]. The intensity center is located within 10 km of *McCann et al.*'s [1982] epicenter with M_I 7.0 (Fig. 6c).

6. Discussion

6.1. Subduction zone events

Our review of historic earthquakes in the northeast Caribbean suggests that the only part of the plate boundary where M 7-8 shallow subduction events are known to have occurred is the 415-km-long segment from central Hispaniola to the NW tip of Puerto Rico (Fig. 7). The significant earthquakes on this segment all took place in the 20th century, and there is no clear evidence for earlier M >7 1/2 events. Adding the seismic moment for the 20th century earthquakes, listed in Table 3, yields a total seismic moment of 2.13×10^{28} dyne-cm. An uncertainty of ±0.1 in the magnitude of the two largest events, the 1943 and August 4 1946 earthquakes outweighs other uncertainties, because of their large magnitude relative to the other earthquakes. (Those uncertainties include whether the 1918 and other smaller earthquakes occurred on the subduction interface or whether older magnitude scales for the 1916 earthquakes and for the 1948, and 1953 earthquakes can be equated with Mw.) Assuming an 100-km-wide seismogenic subduction interface (i.e., from the epicenter of the October 4, 1946 earthquake to the trench), a 415-km-long segment, and elastic rigidity of $3x10^{10}$ dynes/cm^2, a total seismic moment release of $2.13x10^{28}$ dyne-cm is equivalent to an average of 17 m slip on the subduction interface, if the slip components for the earthquakes are aligned. The slip azimuths of the 1943 and 1946a earthquakes were probably ~45° clockwise to the direction of plate convergence *[Dolan and Wald*, 1998, *Doser et al.,* 2005]. If a 45° slip azimuth is representative of earthquakes on the subduction interface, the expected slip accumulation is ~14 mm/y. A slip of 17 m would accumulate in 1200 years, provided the subduction zone is fully coupled. If some of the slip were released aseismically, then the recurrence interval would be longer. This simple estimate is consistent with the conclusion that the historical record of subduction interface earthquakes in Hispaniola is complete; there have been no other M >7 1/2 subduction earthquakes since the 15th century (Fig. 7).

With the possible exception of the 1785 earthquake, there is no evidence for large subduction events north of Puerto Rico or the Virgin Islands. *McCann* [2006] suggested that the 1787 earthquake occurred north of Puerto Rico because damage was most extensive along the island's northern coast. If so, the 1787 earthquake could have been generated by the subduction of Main Ridge (Fig. 1 and 7), which locally elevates the forearc bathymetry by up to 2000 m [*ten Brink et al.*, 2004]. An earthquake in such a tectonic setting could be analogous to the M_w 6.9-7.1 March and May 1947 earthquakes near the toe of the Hikurangi subduction zone east of New Zealand [*Bell et al.*, 2010]. These events were interpreted to have been caused by subducting seamounts in an otherwise aseismic section of the subduction zone [*Bell et al.*, 2010].

The 1785 earthquake could have taken place in one of three locations based on known tectonic features on the seafloor north and east of the Virgin Islands. It could have been an intra-arc M_I 6.8-6.9 earthquake in the 6000 m deep Sombrero basin east of the British

Virgin Islands (Fig. 7), or an M_I 7.0-7.1 in the trench, or an M_I 7.1-7.2 outer-rise event north of the trench, where fault scarps up to 1500 m high are mapped (Fig. 1; [*ten Brink et al.,* 2004]). *Atwater et al.* [2010] have documented an overwash event on the island of Anegada (15 km north of Virgin Gorda) during the period between 1650-1800. *Wei et al.* (in prep.) have modeled several hypothetical tsunami sources and compared the predicted inundation from these models to the observed overwash. These sources included an M 8.4 earthquake along the segment of the Puerto Rico trench west of the BVI, an M 8.7 earthquake along the segment of the Puerto Rico trench east of the BVI, an M 8 outer-rise normal fault event north of the BVI, and the 1755 M 9(?) Lisbon transatlantic tsunami. The Lisbon tsunami is known to have caused damaging tsunami in the Lesser Antilles and Brazil, [e.g., *Barkan et al.*, 2009]. A tsunami from an M 8 outer rise event and the 1755 Lisbon tsunami are the only earthquake sources capable of overtopping the offshore reef and the onshore sand ridge and flooding the island. The 1785 earthquake appears too small to generate a tsunami, unless it was a slow “tsunami earthquake” [*Kanamori*, 1972], that generated relatively little shaking at frequencies that affect people and man-made structures. There is no evidence to support this speculation so the 1755 Lisbon earthquake appears to be the likely source of the Anegada island overwash event.

In the Lesser Antilles, the three significant historical events appear to be intra-arc earthquakes, although the 1843 earthquake could have been a subduction event. The depth to the slab interface beneath our preferred location in Guadeloupe is 125 km [*Feuillet et al.*, 2002]. However, if the earthquake was located east of Barbuda (see Bootstrap inset in Fig. 6b) where the depth to the slab is only 30 km [*Feuillet et al.*, 2002], then the earthquake could have been an M_I 8.2 subduction earthquake. Instrumentally-recorded earthquakes in the Lesser Antilles between 1950-1978 [*Stein et al.*, 1982; *Stein et al.*, 1983] and later earthquakes in 1985, 1992, 2001 [*Feuillet et al.*, 2002] had mainly strike-slip or normal fault mechanisms and were located within the arc. *Stein et al.* [1982, 1983] suggested that the plate boundary is largely decoupled and that the downgoing slab is in extension.

Geist and Parsons [2009] estimated the recurrence interval for M 7.5 events along the entire Hispaniola-Puerto Rico-Lesser Antilles subduction zone to be 67-125 years, provided the subduction zone is fully coupled. The low frequency of M >7.5 earthquakes in the historical record suggests, that with the exception of the segment north of Hispaniola, seismic coupling is low.

7.2. Recurrence intervals on the Septentrional fault

Inferences about rupture modes and recurrence intervals of the Septentrional fault system depend not just on the interpretation of the 1562 and 1842 earthquakes but also on the earthquake histories inferred from trenches along the Septentrional fault (Fig. 1). At one trench site, the most recent identified rupture dates to AD 1040-1230 and involved a minimum of 4 m of left-lateral motion and 2.3 m of normal slip [*Prentice et al.*, 2003]. Horizon folding at a second site occurred sometimes after 3900 BP and before AD 1440-1640 and was ascribed by *Prentice et al.* [2003] to that same event. Their other two sites did not show evidence for AD 1040-1230 faulting. *Prentice et al.* [2003] inferred a penultimate event occurring before AD 30-240 at one site, although samples as young as AD 1680-1940 were dated adjacent to the sample of older date within the faulted surface deposits (their Figure 10). *Prentice et al.* [2003] concluded from these inferred histories a recurrence interval in the range 800-1200 years on that segment of the Septentrional fault. Here we propose that the 1562 and 1842 earthquakes both ruptured the central Dominican Republic section of the Septentrional fault, and that the 1842 rupture extended to the west along the northern Haiti coast. Rather than 800-1200 years, the recurrence interval on the central Dominican Republic section of the Septentrional fault is about 300 years (Fig. 7).

Prentice et al [2003] did not interpret evidence of the nearby 1562 earthquake (Fig. 3a) in any of their trenches. Perhaps it occurred on the subduction interface, or on a secondary thrust feature, as did the 2010 Haiti earthquakes [e.g., *Calais et al.,* 2010]. *Hengesh et al.* [2000], however, found stratigraphic evidence of the 1562 earthquake in their trench across the Septentrional fault at old Santiago de los Caballeros, the town destroyed by that earthquake. We assign the 1562 earthquake to the Septentrional fault because of the extensive damage and large intensity magnitude. The earthquake is not constrained well enough to exclude the possibility that it occurred on a thrust fault in Cordillera Septentrional, the mountain range separating the fault from the northern coast.

The apparent absence of the 1842 earthquake in the trench records could be explained by an earthquake source farther west along the northern Haiti coast [e.g., *Calais et al.,* 2010]. However, given the broad extent of severe damage from the 1842 earthquake (intensity IX in Mole St. Nicolas and in Santiago which are 290 km apart), the extent of the felt aftershocks from Mayaguez, Puerto Rico to Port-au-Prince, Haiti [*Flores et al.,* submitted], and the size of the earthquake (M_I 7.6-7.7), the earthquake is most simply explained by strike-slip motion on a fault plane with relatively short down-dip width. Global empirical relationships [*Wesnousky*, 2008] show an average rupture length of 220-290 km for an M_w 7.6-7.7. The only known strike-slip fault in northern Hispaniola long enough to support an M_w 7.6-7.7 earthquake is the Septentrional fault.

Given the broad geographical extent of severe damage (Fig. 3b), the magnitude of the 1842 earthquake would have had to be much larger than M_I 7.6, if it were a thrust event on the subduction interface. It is possible that the earthquake ruptured the segment of northern Haiti, propagated smoothly through western Cibao valley, and ruptured the central Dominican Republic. This rupture scenario would give rise to two centers of maximum moment release, as seen in our analysis (Fig. 3b). An eyewitness in Puerto Plata described “A second shock followed, yet stronger than the former, accompanied by the same appearances, effects and terrors” [The Public Ledger, July 15, 1842, printed letter dated May 20, 1842].

Locating the 1562 and 1842 earthquakes on the Septentrional fault simplifies the interpretation of the fault slip rates. The estimated Holocene rate is between 6-12 mm/y [*Prentice et al.,* 2003], and the estimated present slip accumulation rate from kinematic models that fit GPS measurements in Hispaniola is 12.3 mm/y [*Calais et al.,* 2010]. If the last earthquake occurred between 1040-1230 AD [*Prentice et al.*, 2003], then the average slip accumulation on the Septentrional fault to date is between 4.7 - 11.9 m (780 yr x 6 mm/y and 970 yr x 12.3 mm/y), an unusually large amount of accumulated slip for a moderately long strike-slip fault. For comparison, a global empirical relationship for strike-slip faults show an average slip for 3.0 m for a 221 km long rupture (the equivalent of M_w 7.7), 3.4 m for a M_w 7.7 source [*Wesnousky*, 2008]. The largest documented average slip for a strike-slip source is 4.7 m for the 1857 San Andreas earthquake [*Wesnousky*, 2008]. If the 1562 and 1842 earthquakes were strike-slip earthquakes on the Septentrional fault, then the recurrence interval is ~300 yr and the average slip accumulation between earthquakes would be 1.8 – 3.7 m (6 - 12.3 mm/y), more consistent with the slip reported for other M 7.5-8 strike-slip earthquakes [*Wesnousky*, 2008].

Static stress models predict that the 1943-1953 subduction earthquakes increased the Coulomb stress on parts of the Septentrional fault [*Dolan and Bowman*, 2004; *ten Brink and Lin*, 2004] bringing it closer to failure. The absence of subsequent rupture on the Septentrional fault is more understandable if the last earthquake on the fault occurred in 1842 rather than 800 years ago. That is, if the 1842 event occurred on the fault, then the 1943-1953 earthquakes took place early in a ~300-year-long loading cycle of the fault. The absence of rupture on the Septentrional fault is difficult to understand if the last earthquake on the fault occurred 780-970 years ago, because of the 4.7-11.9 m of accumulated slip *(Dolan and Bowman*, 2004).

7.3 Muertos Trough

Southward thrusting of eastern Hispaniola and western Puerto Rico over the Caribbean plate has produced the Muertos thrust belt and Muertos Trough [*Granja et al.*, 2009; *Ladd and Watkins*, 1978] The compression direction appears to be perpendicular to Muertos Trough [*ten Brink et al.*, 2009]. *Byrne et al.* [1985] attributed the thrusting to northward subduction of the Caribbean plate beneath the eastern Greater Antilles, based on the analysis of an M_s 6.7 1984 earthquake. That earthquake took place at a depth of 32 km on a gently northward-dipping fault south of the Dominican Republic. *Byrne et al.* [1985] further suggested that the October 18, 1751 event was an M~8 subduction earthquake, but the weight of the evidence suggests an on land location near Azua [*Bakun et al.,* 2012].

Other recent earthquakes beneath the Muertos thrust belt, the mb 5.6-5.8 May 2, 1968, the mb 6.1 June 11, 1971, the M_s 6.7 March 23, 1979, and the M_s 6.1 November 5, 1979 [*Engdahl and Villaseñor,* 2002], were deep (59-106 km), and were perhaps located on the downgoing North American slab. The M_I 6.5-7.5 earthquakes in 1615, 1665, 1673, 1684, and 1691 (Table 2) appear to have occurred near the south coast of the Dominican Republic, although the number of damage reports is too small to locate them. It is possible that these events were either intermediate-depth subduction earthquakes under the south coast of the Dominican Republic, or onshore blind thrust-fault earthquakes, or thrust events from the Muertos trough.

Ten Brink et al. [2009] have argued that the Muertos thrust belt is being formed by the transfer of compressive stresses from the Hispaniola-Puerto Rico subduction zone across the rigid arc to the backarc. They highlighted similar tectonic regimes, where compressive stresses are transferred from the subduction to the back arc region, such as north of Flores and Wetar arc in Indonesia, east of Vanuatu, and north of Panama and western Costa Rica. All 3 regions have experienced M 7.5-7.9 earthquakes within the past 20 years and in two of the regions these earthquakes were accompanied by devastating tsunamis. Thus, the occurrence of a large relatively shallow thrust earthquake in the Muertos Trough should not be discounted in hazard assessments.

8. Conclusions

The northeast Caribbean islands offer a unique opportunity to study the long-term seismic activity of a plate boundary, because of the availability of more than 500 years of written records that include damage reports from earthquakes. We assigned intensities to these damage reports, and estimated the location and magnitude of large historical earthquakes. Our analysis leads to the following conclusions:

1) The North American subduction zone in the NE Caribbean extends from central Hispaniola in the west to Guadeloupe, Lesser Antilles, in the east, a distance of 1300 km.

However, historical earthquakes activity on the subduction interface can be assigned only to the 415 km–long segment from central Hispaniola to the northwest tip of Puerto Rico. A series of large earthquakes in the 20th century appear to have released perhaps 1200 years of oblique slip accumulation in this segment. Earlier historical earthquakes could not be attributed to this subduction segment.

2) The historical record is inconclusive as to the possibility of large deep (>70 km) earthquakes under Hispaniola in the 17th century, and instrumental magnitudes of deep earthquakes under Hispaniola are less than M_I 6.7.

3) We did not identify any large subduction zone earthquakes along the Puerto Rico - Virgin Islands segment. The May 2, 1787 earthquake is smaller than previously suggested and was perhaps generated by the subduction of the "aseismic" Main ridge.

4) The July 11, 1785 earthquake is poorly located and had a magnitude range of M_I 6.9-7.2, depending on its location, whether within the arc east or north of the British Virgin Islands, or near the trench, or in the outer rise.

5) We did not identify large subduction earthquakes along the Lesser Antilles. Therefore, either a large slip deficit has accumulated in the past 500 years along the subduction zone from Puerto Rico to Guadeloupe, or the subduction interface, with the exception of subducting seamounts, is largely decoupled and aseismic. We favor the latter interpretation.

6) Intra-arc earthquakes constitute the primary earthquake and tsunami hazards to the NE Caribbean. We estimate a ~300 year recurrence interval and magnitudes ≤ 7.7 for earthquakes on the Septentrional fault in northern Hispaniola, comparable to the ~300 year recurrence interval for the Enriquillo fault in southern Hispaniola proposed by *Bakun et al.* [2012]. If the Septentrional fault last ruptured in 1842, a large earthquake may not be as imminent, as *Prentice et al.* [2003] have concluded.

7) Four or five M_I 7.1-7.8 intra-arc earthquakes (in 1690, 1843, 1867, and 1974, and perhaps the July 11, 1785) have occurred in the past ~380 years along the 420-km long arc segment between the Virgin Islands and Guadeloupe.

8) Intra-arc seismic activity in Puerto Rico appears minor. The only large strike-slip fault system, the Bunce fault, is located 15-20 km south of the trench and far from the island.

9) Earthquake intensity appears to decay more slowly with distance in Puerto Rico and the Virgin Islands than in Hispaniola and the Lesser Antilles.

10) Although large earthquakes are not documented in the Muertos thrust belt south of the Dominican Republic and Puerto Rico, the tectonic setting of this thrust belt is similar to other backarc thrust belts where large destructive earthquakes and tsunamis have occurred in the past 20 years.

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Figure 1 –Tectonic elements and place names in the northeast Caribbean. Inset – place names in Hispaniola. Barbed lines – mapped thrust faults. Black and white lines – normal or mixed normal and strike-slip faults. Continuous lines – strike-slip faults. St.Th – St. Thomas; T – Tortola; VG- Virgin Gorda; A – Anegada.

Figure 2 – (a) M_I vs. M. for calibration events listed in Table 1. M_I for the Dominican Republic (DR), Haiti, and Lesser Antilles events (red dots) were calculated using equation (1). M_I for earthquakes in Puerto Rico and the Virgin Islands (PRVI) were calculated using the PR-VI model, equation (6). DR – Dominican Republic. (b) MMI

attenuation for an M 6.0 source at 10 km depth in Haiti (eqn. 1) is shown in green and in Puerto Rico and the Virgin Islands (eqn. 6) in black relative to the same source in California [*Bakun*, 2006] and in the stable continental region of eastern North America [*Bakun and Hopper*, 2004] shown in green and red, respectively. Note that intensity data for Puerto Rico and the Virgin Islands is within a distance < 240 km from the epicenters, hence the intensity in larger offsets is poorly constrained.

Figure 3 –Earthquakes near the Septentrional fault system. Left: Location of reported intensity (blue, diameter proportional to intensity), intensity center (green triangle), contours of magnitude (red lines), and contours of the 68% (solid green line) and 95% (dashed green line) confidence level of location. Note that the 95% confidence level in (A) is outside the dashed contour and covers most of the map. Grid search was carried out every 1 km within the area covered by red contours. Top right: Bootstrap locations (green circles). The contour (black line) encloses 68% of the bootstrap locations. Red triangle – grid location with maximum density of bootstrap locations. Blue star - location of minimum RMS value of the mean magnitudes from the corresponding plot on the left. Bottom right: Distribution of bootstrap magnitudes (Dashed line). Lines show the magnitude range of 68% and 95% of the solutions. Black triangle – median magnitude. Star - preferred magnitude at the minimum RMS value from the corresponding plot on the left column.

Figure 4 – Same as Figure 3 for possible earthquakes along the subduction zone in Hispaniola analyzed in this paper. Stars on map for 1916 earthquake –epicenter from *Gutenberg and Richter*, 1954 (GR) and *Doser et al., 2005* (D). Star on map for 1946 earthquake – epicenter from *Kelleher et al.*, 1973.

Figure 5 – Same as Figure 3 for earthquakes in Puerto Rico and the Virgin Islands.

Figure 6 – (a-c) - Same as Figure 3 for earthquakes in the Lesser Antilles. Star on map for 1974 earthquake – epicenter from *McCann* [1982]. (d) Simulation by Yong Wei, NOAA Pacific Marine Environmental Laboratory, of maximum tsunami wave height from a

hypothetical M 8.7 subduction earthquake along the northeast corner of the Caribbean subduction zone. The red line is wave amplitude above the shallow edge of the subduction at 5 km depth below the sea floor.

Figure 7 – Locations of intensity centers from our analysis in this paper and in *Bakun et al.* [2012] and modern epicenters of moderate and large earthquakes between Hispaniola and Guadeloupe. Our evaluation of recurrence intervals for different tectonic regions, discussed in the text, is marked by red lines. Those parts of the subduction zone not aligned by red are not expected from this study to generate large earthquakes.

Table 1. Calibration events

Event	Long	Lat	Depth	M_W,m_b,M_S	MI (Haiti)	#MMI	M - M_I	M_I (PRVI)	M - M_I	Damage references
DR19460804	-68.94	18.92	10	-, -, 7.8-8.1	7.8	32	0.0-0.3			L&B
DR19710611	-69.80	18.00	57	-, -, 6.5	7.0	26	-0.5			C&vH, 1984b
HA20100112*	-72.54	18.45	13	7.0, -, -	6.8	65	0.2			DYFI
HA20100120*	-72.80	18.43	10	5.9, -, -	5.9	14	0.0			DYFI
HA20100222*	-72.55	18.52	10	-, 4.7, -	5.1	8	-0.4			DYFI
LI19741008	-62.00	17.35	47	-, 6.6, 7.1-7.6	7.0	9	0.1-0.6			C&S, 1984a
LI19850316	-62.45	17.01	13	-, 6.3, 6.8	6.2	4	0.6			S&B, 1991
LI20041121	-61.71	15.68	14	6.3, -, -	6.2	16	0.1			DYFI
LI20091005	-62.85	18.02	43	-, 4.7, -	5.4	14	-0.7			DYFI
PR19720202	-66.90	18.50	92	-, 4.7, -	6.0	18	-1.3	5.0	-0.3	C&vH, 1984c
PR19720522	-67.00	18.50	35	-, 4.6, -	6.0	15	-1.4	5.4	-0.8	C&vH, 1984c
PR19720903	-65.70	17.80	56	-, 4.6, -	6.0	7	-1.4	5.3	-0.7	C&vH, 1984c
PR19750617**	-66.34	18.50	111	-, 5.1, -	6.4	38	-1.3	5.3	-0.2	C&S, 1984b
PR19850721**	-67.15	18.96	23	-, 5.7, 5.3	5.9	35	-1.7	5.1	0.2	S&B, 1991
PR19851029	-67.15	18.96	33	-, 4.2, 4.0	5.7	7	-1.3	5.0	-0.6	S&B, 1991
PR20090312**	-66.37	19.07	15	5.0 (M_d)†	6.5	105	-1.5	5.0	0.0	DYFI
VI19700708**	-64.60	18.00	150	-, 5.8, -	6.8	19	-1.0	5.8	0.0	C&vH, 1984a
VI20090721	-64.12	19.12	30	-, 4.7, -	5.8	6	-1.1	4.8	-0.1	DYFI
VI20091024	-64.86	18.80	34	3.7 (Md)†	5.4	8	-1.7	4.2	-0.5	DYFI

*Events used to derive equation (1).
** Events used to derive magnitude in equation (6).
† (M_d) Magnitudes calculated by using the duration of shaking as measured by the time decay of the amplitude of the seismogram. These were provided by the Puerto Rico Seismic Network.
L&B – Lynch and Bodle; C&vH – Coffman and von Hake; C&S – Coffman and Stover; S&B – Stover and Brewer; DYFI – Did You Feel It (earthquake.usgs.gov/dyfi/)

Table 2. Historical earthquakes analyzed in this paper
(a) Hispaniola

yr mo day	Long Lat (optimal location)	M_I	# of damage locations	± 68% ** confid-ence level	± 95% ** confid-ence level	M instrum-ental	Instrumental coordinates
1562 12 02	-70.68 19.37	7.7	4	±0.3	-0.6, +0.5	-	-
1615 09 07	So. Coast of DR	7.5	2	~0.7	~±1.0	-	-
1665 01 ?	So. Coast of DR	6.8	2	~0.7	~±1.0	-	-
1673 05 09	So. Coast of DR	7.3	2	~0.7	~±1.0	-	-
1684 ? ?	So. Coast of DR	7.0	2	~0.7	~±1.0	-	-
1691 ? ?	So. Coast of DR	7.5	2	~0.7	~±1.0	-	-
1842 05 07	-70.80 19.42	7.6	44	+0.2, -0.1	-0.3, +0.2	-	-
1887 09 23	-72.65 19.86	6.7	13	±0.2	-0.4, +0.3	-	-
1897 12 29	-70.76 19.70	6.5	11	±0.2	-0.4, +0.3	-	-
1916 04 24	-69.38 19.20	6.8	24	±0.2	-0.4, +0.3	6.8-7.0	-68 18.5
1946 08 04	-69.80 19.35	7.8	32	+0.2, -0.1	-0.3, +0.2	7.8-8.1	-68.94 18.92

(b) Puerto Rico, Virgin Islands, and the Lesser Antilles

yr mo day	Long Lat (optimal location)	M_I	# of damage locations	± 68% confid-ence level	± 95% confid-ence level	M instrum-ental	Instrumental coordinates
1690 04 05	-62.51 17.08	7.5	11	±0.2	-0.4, +0.3	-	-
1785 07 11	-64.60 19.21	6.9	5	-0.3, +0.2	-0.6, +0.5	-	-
1787 05 02	-67.54 17.33	6.9*	11	±0.2	-0.4, +0.3	-	-
1843 02 08	-61.49 16.34	7.8	29	+0.2, -0.1	-0.3, +0.2	-	-
1974 10 08	-61.76 17.36	7.0	9	±0.2	-0.4, +0.3	7.1-7.6	-61.976 17.349

*Using attenuation relationship for Puerto Rico and the Virgin Islands. M_I7.2 if using relationship for Hispaniola

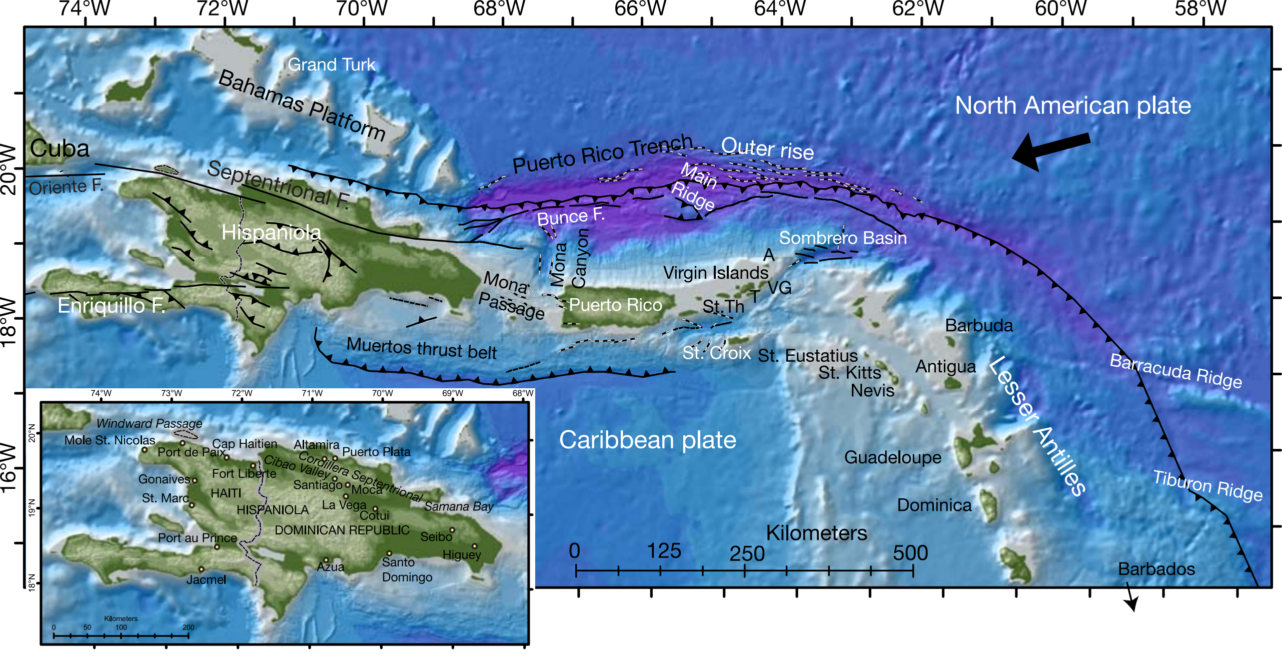
** M confidence levels as a function of the number of intensity assignments were calculated in *Bakun and Wentworth* [1997] for the location at the epicenter. If the true location is unknown and M_I varies with different potential locations, the uncertainty is likely larger than given in the columns.

Table 3. Location and magnitude of significant (M ≥6.5) 20th century earthquakes in northern Hispaniola and Mona Passage

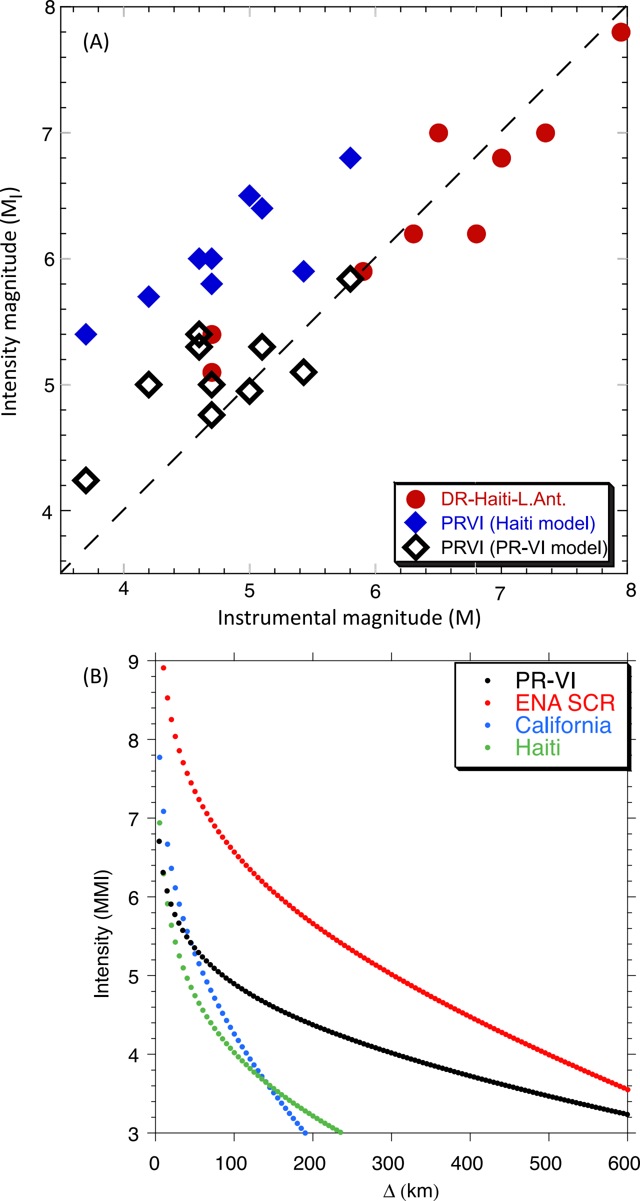
yr mo day	Long Lat	Long Lat #	mB	Ms	Mw	Mw #	M_I	Other scales
1897 12 29	-70.76 19.70*						6.5	6.8&
1915 10 11	-67 19	-67.10 19.04				6.4		6.6-6.8
1916 04 24	-69.38 19.20*	-68.53 18.26	7.0			6.8	6.8	7.2
1916 11 30	-70 19							6.6-6.8
1917 07 27	-67.5 19	-67.66 18.28	7.0	7.0		6.9		7.0
1918 10 11	-67.631 18.473	-67.62 18.28	7.2	7.3-7.5	7.3	7.2		7.5
1920 10 20		-67.28 19.07				6.5		
1943 07 29	-66.983 19.09	-66.97 18.99	7.5	7.7-7.8	7.6	7.8-7.9		7.8-7.9
1946 08 04	-69 18.959		7.6	7.8-8.1	7.9		7.9	8.1
1946 08 08	-69.631 19.63		7.6	7.6	7.5			7.6-7.9
1946 10 04	-68.77 18.68**			7.0				
1948 04 21	-69.703 19.319		7.3	7.1,7.3**				7.3
1953 05 31	-70.4 19.4			6.7,7.0**				6.8

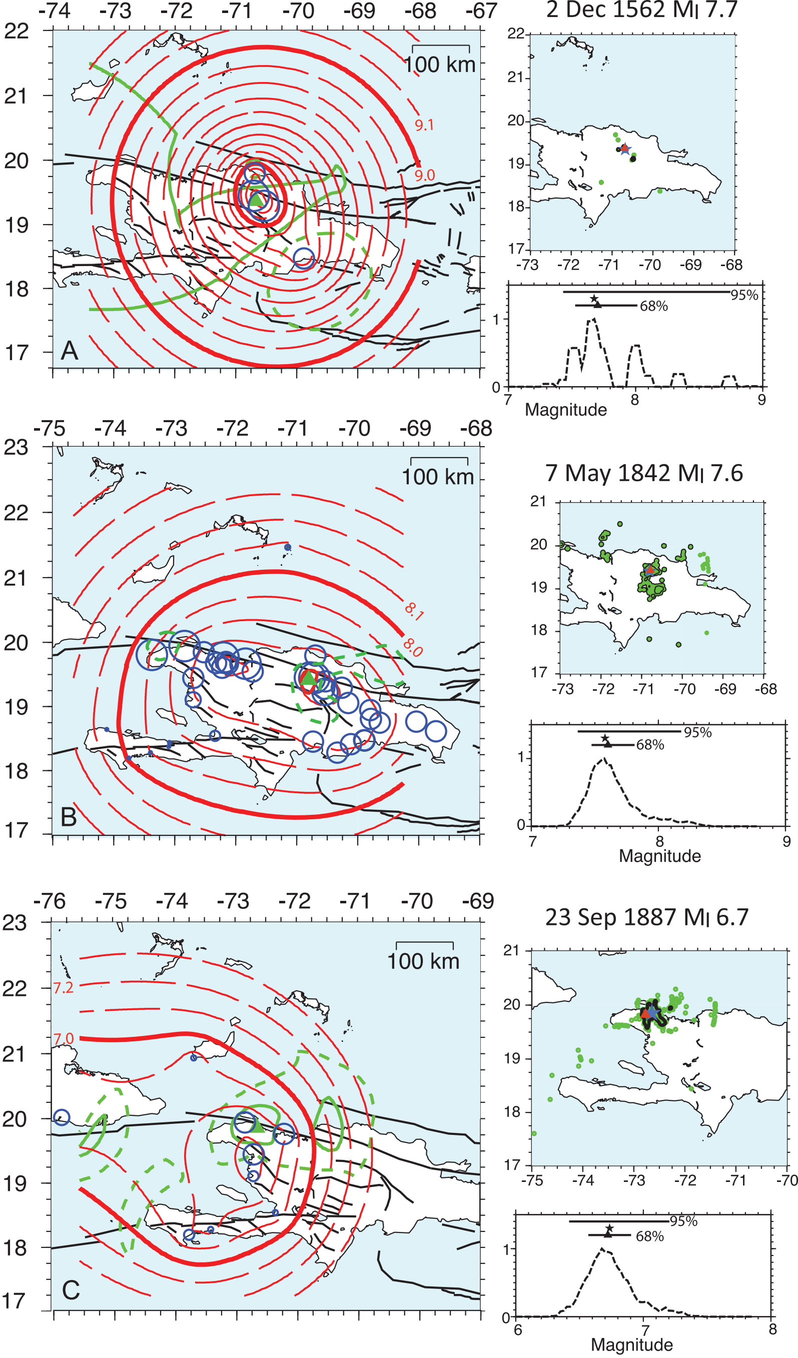
-All epicenters are from Engdahl and Villaseñor [2002], except *Intensity location from this paper; ** from Kelleher [1973]; #Listed in Doser et al. [2005].

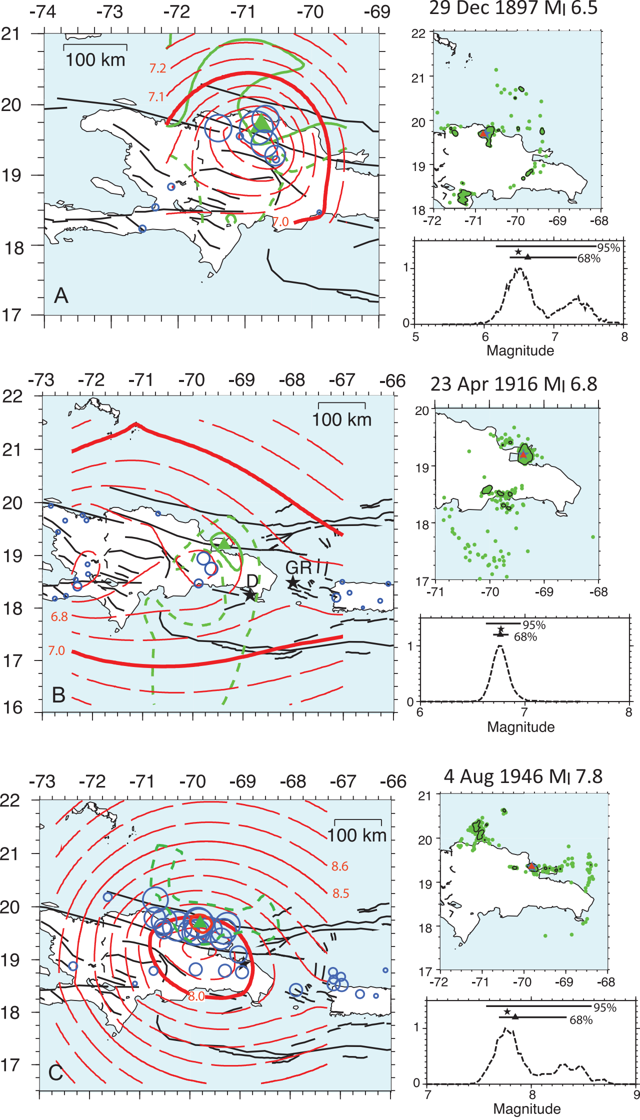
-All magnitudes are from Engdahl and Villaseñor [2002], except M_I - Intensity magnitude from this paper; **from Kelleher [1973]; #Mw calculated by Doser et al. [2005], & from Abe [1994].

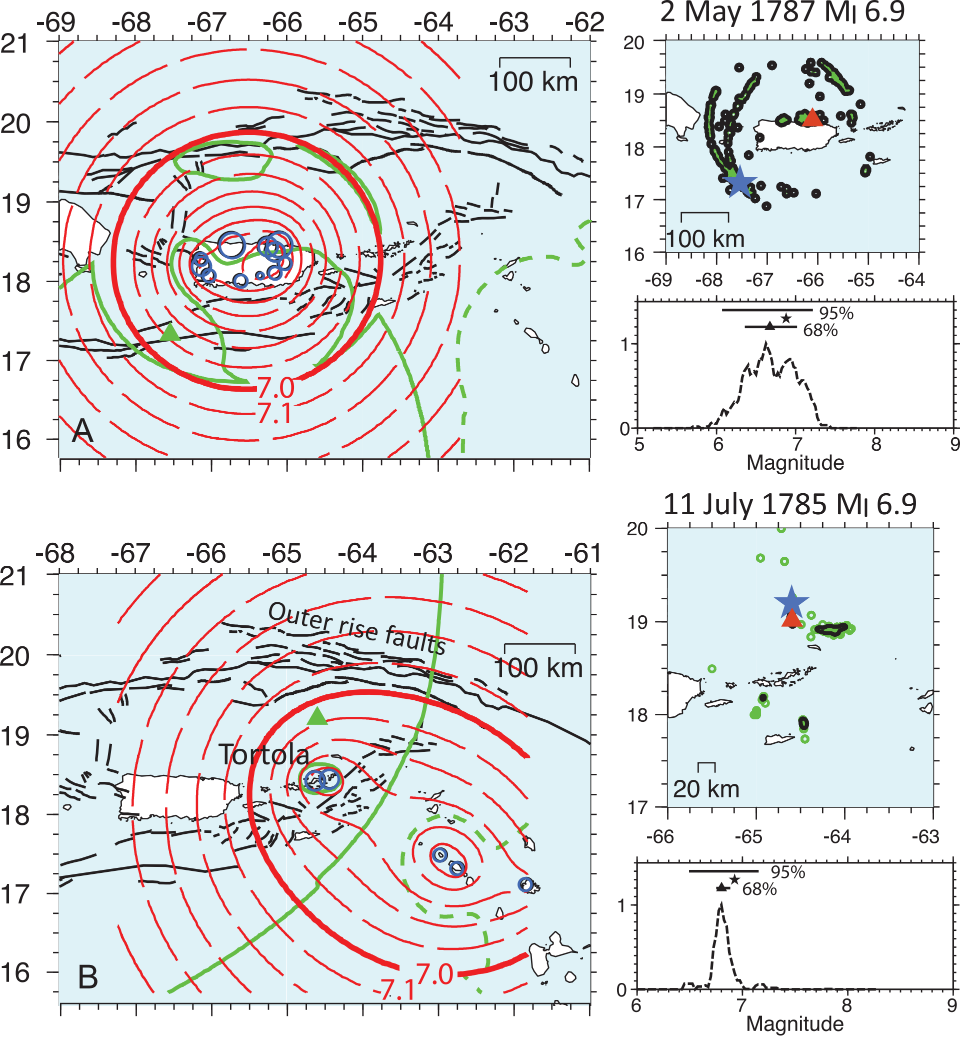


74°W
72°W
70°W
68°W
66°W
64°W
62°W
60°W
58°W
20°W
18°W
16°W
Grand Turk
Bahamas Platform
North American plate
Cuba
Oriente F.
Septentrional F.
Puerto Rico Trench
Outer rise
Main Ridge
Bunce F.
Hispaniola
Mona Canyon
Sombrero Basin
A
Virgin Islands
VG
T
St.Th
Mona Passage
Puerto Rico
Enriquillo F.
Muertos thrust belt
St. Croix
St. Eustatius
St. Kitts
Nevis
Barbuda
Antigua
Lesser Antilles
Barracuda Ridge
Caribbean plate
Guadeloupe
Tiburon Ridge
Dominica
Kilometers
0
125
250
500
Barbados
74°W
73°W
72°W
71°W
70°W
69°W
68°W
20°N
19°N
18°N
Windward Passage
Mole St. Nicolas
Port de Paix
Cap Haitien
Altamira
Puerto Plata
Cibao Valley
Cordillera Septentrional
Fort Liberte
Gonaives
Santiago
Moca
St. Marc
HAITI
La Vega
Cotui
Samana Bay
HISPANIOLA
DOMINICAN REPUBLIC
Seibo
Port au Prince
Higuey
Azua
Santo Domingo
Jacmel
Kilometers
0
50
100
200

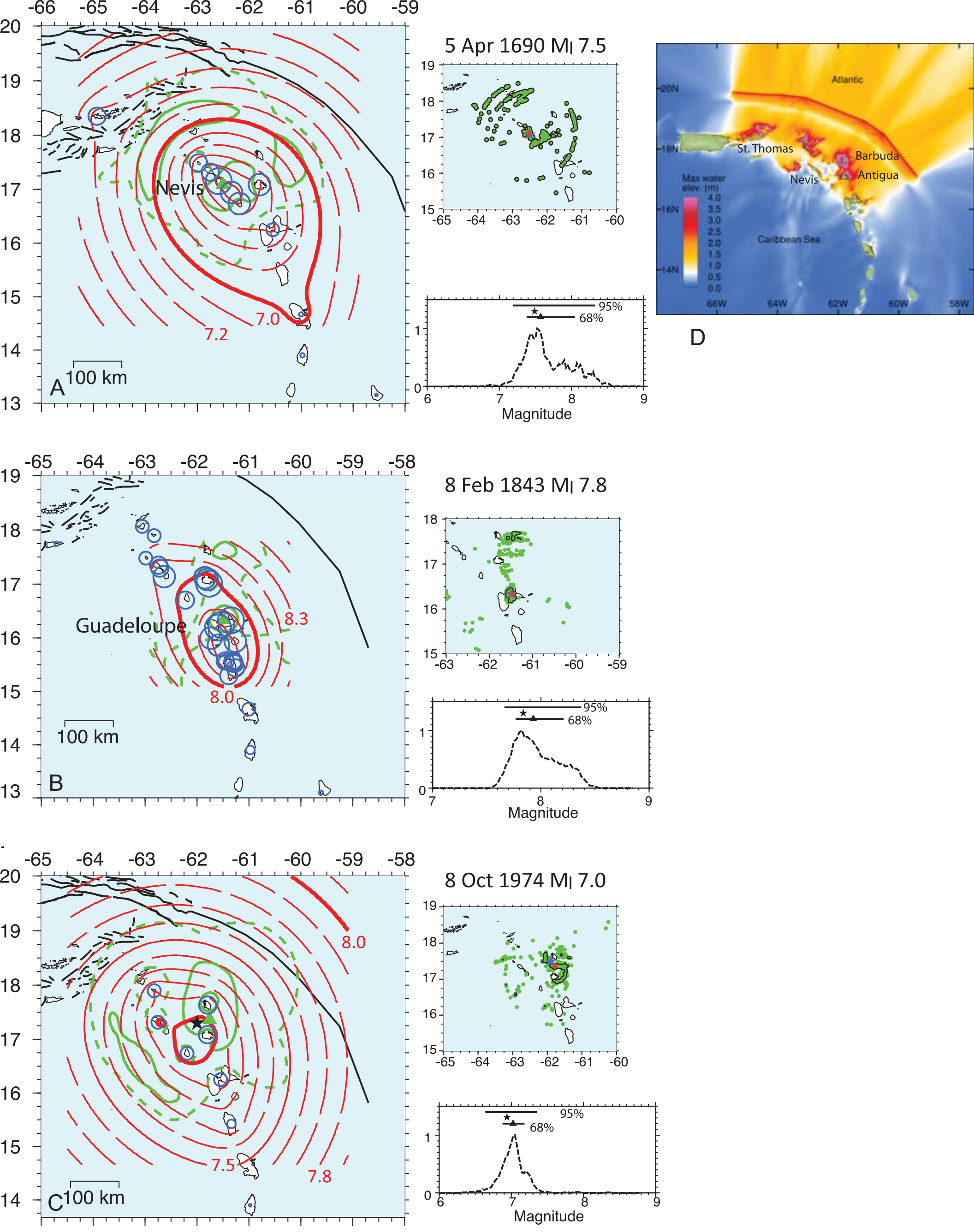

(A)
Intensity magnitude (M_I)
Instrumental magnitude (M)
DR-Haiti-L.Ant.
PRVI (Haiti model)
PRVI (PR-VI model)
(B)
Intensity (MMI)
Δ (km)
PR-VI
ENA SCR
California
Haiti


2 Dec 1562 Ml 7.7
100 km
9.1
9.0
A
95%
68%
Magnitude
7 May 1842 Ml 7.6
100 km
8.1
8.0
B
95%
68%
Magnitude
23 Sep 1887 Ml 6.7
100 km
7.2
7.0
C
95%
68%
Magnitude

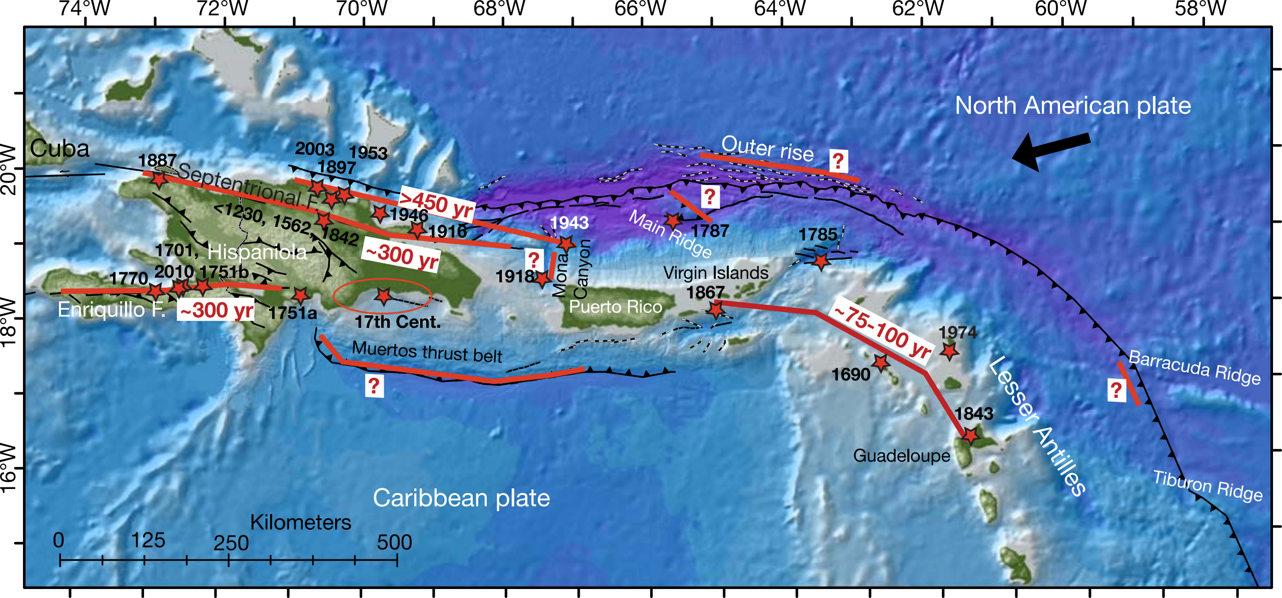

29 Dec 1897 Ml 6.5
23 Apr 1916 Ml 6.8
4 Aug 1946 Ml 7.8
100 km
A
B
C
7.2
7.1
7.0
6.8
8.6
8.5
8.0
D
GR
95%
68%
Magnitude



2 May 1787 Ml 6.9
11 July 1785 Ml 6.9
100 km
20 km
7.0
7.1
A
B
Tortola
Outer rise faults
95%
68%
Magnitude



5 Apr 1690 Ml 7.5
Nevis
7.0
7.2
100 km
A
95%
68%
Magnitude
8 Feb 1843 Ml 7.8
Guadeloupe
8.3
8.0
100 km
B
95%
68%
Magnitude
8 Oct 1974 Ml 7.0
8.0
7.5
7.8
100 km
C
95%
68%
Magnitude
Atlantic
St. Thomas
Nevis
Barbuda
Antigua
Max water elev. (m)
Caribbean Sea
D


74°W
72°W
70°W
68°W
66°W
64°W
62°W
60°W
58°W
20°N
18°N
16°N
North American plate
Outer rise
?
Cuba
1887
Septentrional F.
2003
1953
1897
>450 yr
<1230, 1562,
1842
1946
1916
~300 yr
1943
Main Ridge
1787
1785
1701,
Hispaniola
1770
2010
1751b
?
1918
Mona Canyon
Virgin Islands
1867
Enriquillo F.
~300 yr
1751a
17th Cent.
Puerto Rico
~75-100 yr
1974
Muertos thrust belt
?
1690
Barracuda Ridge
Lesser Antilles
1843
Guadeloupe
Tiburon Ridge
Caribbean plate
Kilometers
0
125
250
500