The 1867 Virgin Island Tsunami

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The 1867 Virgin Island tsunami: observations and modeling

Le tsunami de 1867 aux Îles Vierges : observations et modélisation

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Abstract

The 1867 Virgin Island tsunami was of a great effect for the Caribbean Islands. A maximal tsunami height of 10 m was recorded for two coastal locations (Deshayes and St. Rose) in Guadeloupe. The historical data of this event for the Caribbean Sea are discussed. The modeling of the 1867 tsunami is performed in the framework of the nonlinear shallow-water theory. The four different orientations of the tsunami source in the Anegada Passage are examined. The directivity of the tsunami wave in the Caribbean is investigated. The time histories of water surface fluctuations are calculated for several coastal locations on the coasts of the Caribbean Sea. Results of the numerical simulations are in reasonable agreement with data of observations.

Résumé

Le tsunami de 1867 des Îles Vierges a eu un grand effet sur les îles de la Caraïbe. Deux vagues de tsunami d'une hauteur maximale de 10 m ont été enregistrées en deux points des côtes de la Guadeloupe (Deshaies et Sainte-Rose). Les données historiques de cet événement pour la mer des Caraïbes sont discutées. La modélisation du tsunami de 1867 est effectuée dans le cadre de la théorie non linéaire en eau peu profonde. Les quatre orientations possibles de la source du tsunami dans le passage d'Anegada sont étudiées. La directivité de la vague de tsunami dans la Caraïbe est examinée. Les fluctuations temporelles de la surface de la mer sont calculées pour plusieurs points de la côte de la mer des Caraïbes. Les résultats de la simulation numérique sont raisonnablement en accord avec les données d'observation.

Keywords: Tsunami; Shallow-water theory; Numerical model; Caribbean Sea

Mots clés : Tsunami ; Théorie en eau peu profonde ; Modèle numérique ; Mer des Caraïbes

1. Introduction

Recently, Lander et al. (2002) published a brief history of tsunamis in the Caribbean Sea. They collected 91 reported waves that might have been tsunamis within the Caribbean region. Of these, 27 are judged by authors to be true, verified tsunamis and an additional nine are considered to be very likely true tsunamis. Only for last 35 years, there were six true and almost true tsunamis: 25 December 1969 (earthquake with magnitude 7.6 in Lesser Antilles, maximal tsunami amplitude of 46 cm at Barbados); 16 March 1985 (moderate earthquake with magnitude 6.3 in Guadeloupe, several-centimeter tsunami was recorded at Basse-Terre, Guadeloupe); 1 November 1989 (weak earthquake with magnitude 4.4 off the north coast of Puerto Rico generating a small wave in Cabo Rojo); 22 April 1991 (earthquake with magnitude 7.6 created a tsunami affected the coast of Central America from Costa Rica to Panama; wave height is 2 m in Cahuito Perto Viejo, Costa Rica); 9 July 1997 (earthquake of

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magnitude 6.8 occurred off the coast of Venezuela induced a weak tsunami). Our goal is to study the tsunami damage for Lesser Antilles. Totally, 23 tsunami-like waves were recorded in this area for last 400 years: 16 events have the seismic origin, four events have volcanic origin and three events have unknown source (Zahibo and Pelinovsky, 2001). Many of these tsunamis are not well documented and cannot be qualified as the true tsunamis. On the basis of these data, the rough evaluation of the cumulative frequency of tsunami was performed for Barbados and Antigua. Of course, the accuracy of such estimates is very low, and the numerical simulation of the historical and prognostic tsunamis should be done to create the reliable tsunami database. In particular, the volcano can effectively generate tsunami waves. For instance, the Soufriere Hills Volcano on Montserrat erupted several times in 1990s (Hooper and Mattioli, 2001) and generated tsunami waves on 26 December 1997, with height 3 m. Heinrich et al. (1998, 1999a, b, 2001), studying the danger of the volcanic eruption in the Soufriere Hills Volcano, Montserrat, has shown that the potential debris avalanche can induce the tsunami waves of 1–2 m in nearest zone and 50 cm at Guadeloupe and Antigua. Friant (2001) simulated the tsunami waves from the potential eruption of the St. Pierre Volcano, Martinique. Recently, the submarine evidence for large-scale debris avalanches for many islands in the Lesser Antilles Arc was found (Deplus et al., 2001).

The Lesser Antilles has also experienced transoceanic tsunamis. In 1755, during the Lisbon earthquake, the tsunami crossed the Atlantic Ocean in 7 h and attacked Saba (7 m runup), St. Martin (4.5 m runup), Dominica (3.7 m runup), Antigua (3.7 m runup) and Barbados (1.5 m runup). The propagation of this tsunami has recently been modeled by Mader (2001a). According to his calculations, the wave amplitude east of Saba at 4747 m depth is 2.5 m, and at depth 825 m is 5 m close to observed value (7 m). We would also like to mention the possible tsunami expected from a lateral collapse of the Cumbre Vieja Volcano on La Palma (Canary Islands); according to Mader (2001b), its height may be 3 m high on the Caribbean Islands (Saba Island). Pararas-Carayannis (2002) discusses 40 m waves during this event in the Caribbean.

Tsunami waves, generated at the Virgin Islands on 18 November 1867, were significant for many islands in the Lesser Antilles: wave height exceeded 10 m at Guadeloupe and 3 m at Grenada. Historical material of this event is collected (Reid and Taber, 1920; Lander et al., 2002; Zahibo and Pelinovsky, 2001). The main goal of this paper is to simulate 1867 event with different orientations of the tsunami sources and to compare the numerical results with the available data observations. Historical data of the 1867 tsunami are summarized in Section 2. Numerical model based on the nonlinear shallow-water theory is briefly presented in Section 3. Results of numerical simulations (the sea state at different time steps and the time history of water surface fluctuations at selected coastal stations) with respect to four different orientations of the tsunami source in the Anegada Passage are given in Section 4. The possible amplification of the tsunami amplitude due to wave interference is discussed in Section 5. Computed results are compared with available data of observations.

2. Historical data of the 1867 tsunami

Tsunami of 1867 in Caribbean is a well-documented event. On 18 November 1876, at approximately 2:45 p.m. (18:45 UT) a violent earthquake occurred at Virgin Islands. Its surface magnitude is 7.5 and focal depth is 33 m. The earthquake was strong; it had intensity nine on British Virgin Islands (Tortola, St. John), US Virgin Islands (St. Thomas, St. Croix) and Puerto Rico (Viequez and Culebra); these locations are shown in Fig. 1. Lander et al. (2002) indicate that an earthquake occurred in the Anegada Passage between St. Croix and St. Thomas, US Virgin Islands. The same location,

Fig. 1. The distribution of tsunami runup (meters) in the Virgin Islands (the epicentre is shown by star).

18.0° N 65.0° W is indicated in web of Noaa/Nesdis/National Geophysical Data Center. In ETDB/ATL (2002), epicenter coordinates are 18.4° N 64.3° W, and this location corresponds to the British Virgin Islands, near Virgin Gorda, to east from Tortola. Actually, both coordinates are in the axis of the Anegada Passage inclined to the latitude on 30°. This deepest passage (maximal depth is 4.5 km) is minimum 56 km in width. Reid and Taber (1920) conclude that the length of the source is some 10 km, while the vertical displacement of the sea floor is less than 10 m; the strike of the fault must have been approximately east–west, following the general direction of the scarp. They also mention that according to the observations there were two severe shocks separated by an interval of about 10 min, and each of these shocks were followed by a great sea wave.

Original information of the 1867 event was collected by Reid and Taber (1920), 50 years after earthquake and tsunami. Lander et al. (2002) in their catalogue of tsunamis in the Caribbean give the summarized information about this event. Zahibo and Pelinovsky (2001) examined the wave heights for Guadeloupe and revised some values. Some descriptions of the 1867 tsunami mainly from the paper (Reid and Taber, 1920) with additional information from other sources are given in the following.

At St. Thomas (US Virgin Islands), "directly after the second shock (10 min after the first shock) the ocean, which shortly before the first shock had receded from the land several 100 feet was seen to rise like one huge wave and come in toward the harbor. It stood up like a straight white wall, about from 15 to 20 feet high, and advanced very fast into the harbor, sweeping or upsetting small vessels before it, and raising the large men-of-war and steamers to its top. The appearance of this wave was like a white masonry wall, erect and straight as if built with the aid of rule; it did not have the appearance of ordinary waves. It broke in over the lower parts of the town to the height of a couple of feet and to extent of about 10 min, and the second appeared to be even a little larger than the first, and went a little further inland. After these two waves had passed away, the ocean remained quite calm again, just as it was before the first shock of the earthquake". Lander et al. (2002) added that the tsunami killed 12 people. The same impressions have been reported by Navy officers in this harbor after the earthquake and tsunami. "The extraordinary spectacle of a heavy wall of sea, some 20 feet in height, apparently distant about 3 miles, was coming towards the harbor with terrible power. The damage on shore has been far more ruinous to the merchants than that occasioned by the hurricane. The first heavy roller went up into the town swamping the stores which were mostly on the bay front, throwing out and finally stranding their goods in unheard-of directions". "In a few minutes a great wave was seen approaching from the southeast between the Islands of St. Thomas and St. Croix reaching soundings it began to break, assuming the appearance of a great bore, not less than 23 feet in height. A second wave soon came in a much greater volume than the first, of any of them—I think the water rose 20 feet above where it had fallen. The third wave was not so large as the two first, and the fourth did not appear to rise more than 12 of 15 in.". Lander et al. (2002) pointed that the tsunami waves induced damage at Hassel Island (small island near St. Thomas, opposite Charlotte Amalie). We are able to find the description of tsunami in Virgin Island in local Guadeloupe newspaper (Le Commercial de la Guadeloupe, 1867), "15 min after the great jolt of 18 November, the sea formed, at the entry of the bays, a bar of more than 100 m height which precipitated on the city like an avalanche; but the floods, broken on its passage on the white rocks which are in the middle of the passage, have decreased their violence". There must be exaggeration in this description.

In St. Croix (US Virgin Islands), "the waves broke upon the northern and western coasts of the island with great violence, washing many vessels and boats ashore, sweeping away some smaller houses and doing great injury to others". In particular, the US ship "Monongahela" moved on the beach in front of the town of Fredericksted (west part of island). Lander et al. pointed that tsunami was a wall of water 7.6 m high. Five people were killed. Totally, the waves at St. Croix were 7–9 m. At Christensted (north part) waves swept inland 90 m.

In Road Town, Tortola (British Virgin Islands), "the sea sank and then rose 4 or 5 feet above its usual level, submerging the lowest part of the town and sweeping away most of the smaller houses". Lander et al. (2002) added that "at Peter Island, British Virgin Islands, waves 1.2–1.5 m were reported".

The distribution of tsunami runup heights in the Virgin Islands (near the origin) is presented in Fig. 1.

Tsunami waves were recorded in the most of the islands of the Lesser Antilles; see Fig. 2. "A high wave is said to have invaded Saba Island; and the sea rose pretty high at St.

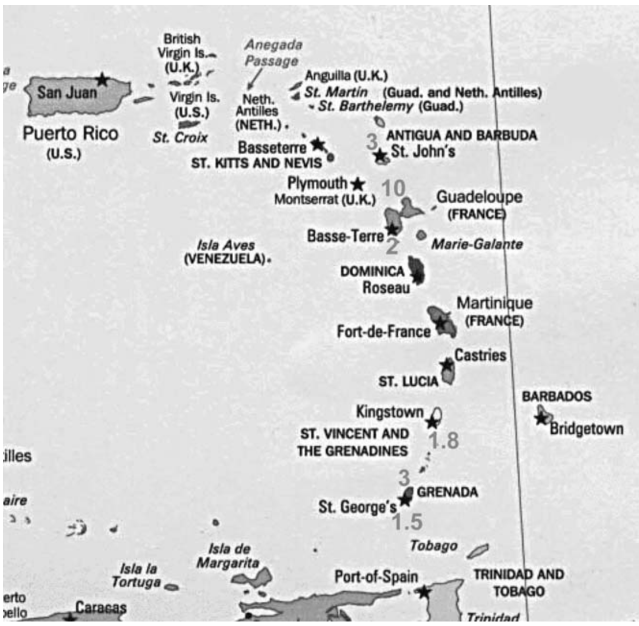

Fig. 2. The distribution of tsunami runup (meters) at Lesser Antilles during the 1867 event.

Christopher" (now St. Kitts). In Antigua, the sea rose 8 or 10 feet in the harbor of St. John's.

At Basse-Terre, Guadeloupe, "the sea suddenly retired a long distance and then advanced, this phenomenon was repeated once, and then all was quiet. The total range in the height of the sea from its lowest to its highest level was about 2 m". In the northern locations, Deshayes and St. Rose on the Basse-Terre Island (Guadeloupe), "sea has suddenly withdrawn to more than 100 m from the littoral and then returned in a wave at least 60 feet high, which broke over the shore and carried off all floatable objects". This big value for tsunami height at Deshayes is repeated in the catalogue by Lander et al. (2002); they pointed that this the largest height recorded in the Caribbean. The description of this event is based on the original letter by Devill (1867). We found this letter written in French. At Deshayes, "the sea devastated and trusted almost all the houses of the village. The inhabitants took refuge in the church". The inspection of Deshayes shows that the church is located 10 m above sea level (Zahibo and Pelinovsky, 2001), and, therefore, tsunami wave height did not exceed 10 m. According to Devill (1867) at St. Rose, after withdrawn on 100 m "a first blade, at least 60 feet high, rising about 3 miles to the north in the open sea has rolled violently towards the ground, immersing all the littoral and flooding the houses. A second and third of these enormous blades, rolling from the north to south, followed, with short intervals, reversing all in their passage". The possibility of determining the 18 m wave height from coast, 3 miles in the open sea, seems to be unrealistic, and we believe looking on descriptions of tsunami in both locations that tsunami wave height at St. Rose did not exceed 10 m also. As it is pointed by Reid and Taber (1920), Pointe-a-Pitre (southern side of Guadeloupe) is "so protected, at the head of the Bay, Petit Cul de Sac, that the waves were barely, if at all, noticeable there." Lander et al. (2002) characterize waves at Pointe-a-Pitre as a slight swell. They also mentioned that at Isles des Saintes, there was a slight swell, and at Fond-du-Cure houses were inundated to a depth of 1 m. The wave was observed at Martinique.

At St. Vincent, "the water was observed to be unusually high; but nothing occurred to attract attention". At Bequia Island (The Grenadines), "were the three great slow waves, the water rising about 6 feet above its usual level; the whole event lasted above 40 min, and the water was not in the great agitated".

In Grenada, Saint George's, "the sea suddenly sank 4 or 5 feet, leaving the reef, in front of the lagoon, bare; it then rose as much". This was repeated six times and then all was quiet. At Gouyave (former Charlotte Town), "the sea began to ebb and flow with a range of about 20 feet, doing some damage to the town". Reid and Taber (1920) pointed that 20 feet seem to be an exaggeration. Lander et al. (2002) give 3 m for Gouyave.

A tsunami is also mentioned at Isle de Margarita, Venezuela.

The 1867 earthquake was strong all over Puerto Rico to west from the origin (Fig. 3). In Yabucoa Harbor (southeastern part), "the sea retired about 150 yards, and then advanced an equal distance over the land, which in this neighborhood is low." In Fajardo (eastern part), tsunami "was very small". Lander et al. (2002) pointed that tsunami waves were 1–6 m at Puerto Rico. At San Juan (north) and Arroyo (south, near Guayama), water rose 0.9–1.5 m, and high waves were observed at Vieques (east of Puerto Rico).

Quantitative information of the observed tsunami wave runup height is summarized in Table 1. In fact, this table can be found in electronic ETDB/ATL (2002), but we modified it analyzing all geographical locations and eliminating duplicated and incorrect data.

It is important to give here the general description of the tsunami propagation during the 1867 event (Reid and Taber, 1920). "A great sea wave was started by the first shock, and a second larger one by the second shock some 10 min later; but other waves followed were relatively unimportant. They trav-

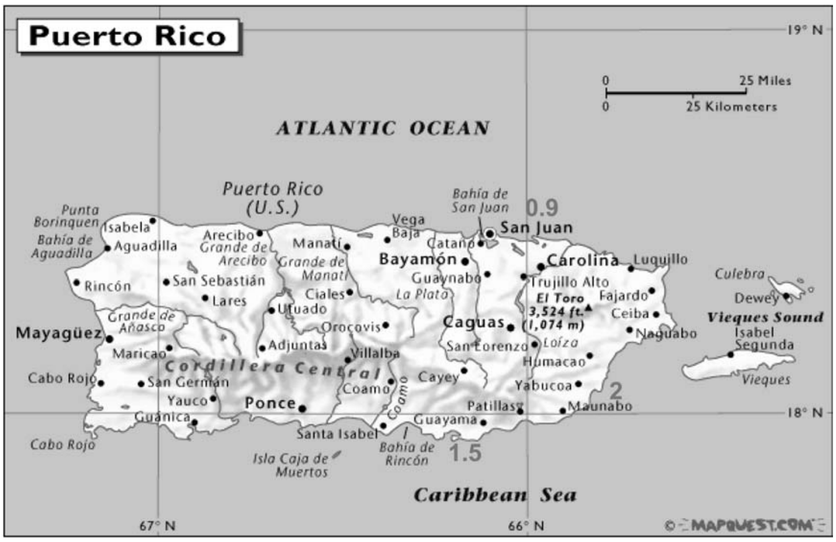

Fig. 3. The distribution of tsunami runup (meters) along Puerto Rico coast during the 1867 event.

Table 1
The distribution of tsunami runup at some locations for the 1867 event

Island	Town	Height, m
British Virgin Islands, Tortola	Road Town	1.5
British Virgin Islands, Peter Island		1.5
US Virgin Islands, St. Thomas	Charlotte Amalie	6.0
US Virgin Islands, Hassel Island		4.9
US Virgin Islands, St. Croix	Fredericksted	7.6
Antigua	St. John's	3.0
Guadeloupe	Deshayes	10.0
Guadeloupe	St. Rose	10.0
Guadeloupe	Basse-Terre	2.0
Guadeloupe, Isles des Saintes		1.0
Grenadines, Bequia Island	Port Elizabeth	1.8
Grenada	Gouyave	3.0
Grenada	Saint George's	1.5
Puerto Rico	San Juan	0.9
Puerto Rico	Yabucoa Harbor	2.0
Puerto Rico	Arroyo (near Guayama)	1.5

Historical information of the 1867 tsunami will be used for comparison with the results of the numerical simulation.

elled in all directions: rapidly toward St. Croix and Vieques and the south coast of Puerto Rico on account of the deep water, losing energy slowly, and becoming small toward the western part of the south coast; rapidly towards to east, where they fell upon the northern Leeward Islands; rapidly also around the western end of St. Croix and across the deep water of the eastern Caribbean Sea to Guadeloupe and the islands south of it, as far as Grenada, nearly 800 km from the origin. The shallow-water north of the origin rapidly reduced the speed and the energy of the waves moving in that direction; however, they were still strong in the nearby islands, but were much weaker when they reached the north-eastern coast of Puerto Rico; they were not noticed at all on the northern coast".

3. Numerical model and tsunami source

To describe the tsunami wave propagation, the shallow-water theory in the Cartesian coordinates is used. Due to lowest latitude of the Caribbean, the Coriolis effect is neglected. These equations are

$$\frac{\partial M}{\partial t} + \frac{\partial}{\partial x}\left(\frac{M^2}{D}\right) + \frac{\partial}{\partial y}\left(\frac{MN}{D}\right) + g\,D\frac{\partial \eta}{\partial x} + \frac{k}{2gD^2}M\sqrt{M^2+N^2} = 0 \quad (1)$$

$$\frac{\partial N}{\partial t} + \frac{\partial}{\partial x}\left(\frac{MN}{D}\right) + \frac{\partial}{\partial y}\left(\frac{N^2}{D}\right) + g\,D\frac{\partial \eta}{\partial y} + \frac{k}{2gD^2}N\sqrt{M^2+N^2} = 0 \quad (2)$$

$$\frac{\partial \eta}{\partial t} + \frac{\partial M}{\partial x} + \frac{\partial N}{\partial y} = 0 \quad (3)$$

where η is the water surface elevation, t is time, x and y are horizontal coordinates in zonal and meridional directions, M and N are discharge fluxes in horizontal plane along x and y coordinates, $D = h(x, y) + \eta$ is the total water depth, $h(x, y)$ is unperturbed basin depth, g is the gravity acceleration and $k = 0.025$ is the bottom friction coefficient. Numerical simulations used the tsunami propagation model Tunami-N2 developed by Prof. Imamura in Tohoku University (Japan) and provided through the Tsunami Inundation Modeling Exchange (Time) program, see Goto et al. (1997). It has been applied to several case studies in the Sea of Marmara (Yalciner et al., 2002) and in the Aegean Sea (Yalciner et al., 2001), and also for Puerto Rico (Mercado and McCann, 1998). The model solves the governing equations by the finite difference technique with leap-frog scheme (Goto et al., 1997). The bathymetry of the Caribbean Sea was obtained from the Smith and Sandwell global seafloor topography (Etopo2) with a 3 km grid size. The time step is selected as 6 s to satisfy the stability condition. The total number of grid points in the study area is 433,580 (815 × 532). Along the depth of 10 m contour line the vertical wall boundary condition is assumed. Free outward passage of the wave is permitted at the open sea boundaries. In fact, nonlinear and dissipative terms are not important for simulation of the tsunami waves in the open sea, and we may consider all solutions as linear and dissipativeless.

The earthquake epicenter is assumed to be located at the site with coordinates: 18.0°N 65.0°W according to the data of Noaa/Nesdis/National Geophysical Data Center (see Fig. 1). The surface magnitude of tsunamigenic earthquake is chosen as 7.5 and focal depth less than 30 km according to ETDB/ATL (2002). The length of the fault is 120 km and the width is 30 km. We consider several orientations of the fault in the Anegada Passage (Fig. 4). Reid and Taber (1920) suggested that the fault is oriented from west to east. By using this information, the first assumption (S1) for the major axis of the source is selected as parallel to latitude. The major axes of other sources (S2, S3 and S4) are selected as inclined to the latitude with the angles of 15°, 20° and 25°, respectively, close to the axis of the Anegada Passage and the slope of the trench.

The computed tsunami source (initial positions of sea level displacements) is in elliptical shape and similar in all of these seismic models. Fig. 5 shows the initial water displacement according to the source assumption S3 (its geographical location will be shown in Fig. 7). The depression of the water surface is assumed to be at south (on deepest part of the trench) with the negative amplitude 1.8 m. The elevation source is assumed to be at north (on shallowest part of the trench) with the positive amplitude 3.9 m.

4. Computed wave distribution in Caribbean

In the application, we have used the propagation model by assuming each tsunami sources separately and computed the sea state at different time steps, time histories of water surface oscillations and the maximum positive amplitudes at every grid points. In analysis the directivity of tsunami is discussed by using the distribution of positive amplitudes (crest amplitudes) in the domain computed throughout 300–

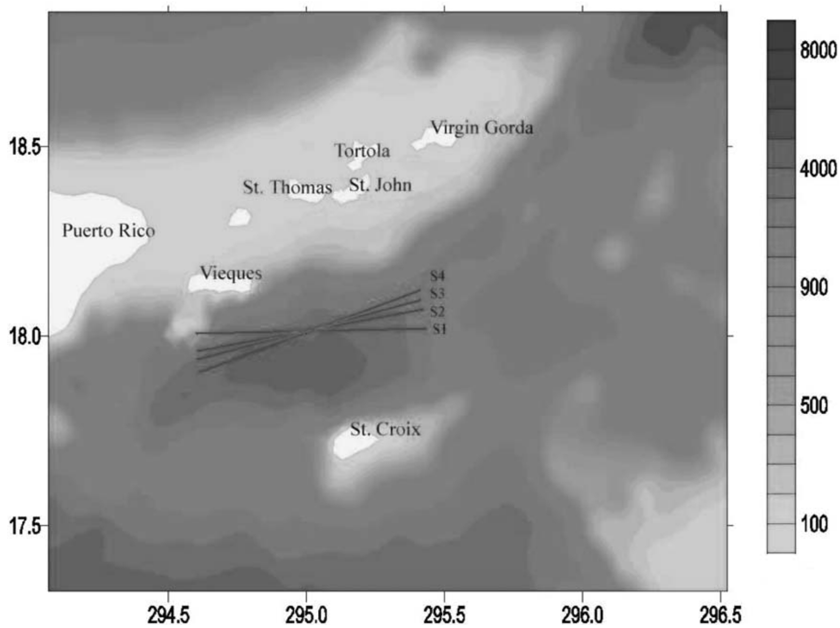

Fig. 4. The location of the fault and its selected orientations of the major axis in the numerical domain.

min simulation. Fig. 6 shows the distributions maximal elevation of the sea level (maximum positive tsunami amplitudes). In all cases the tsunami is significant near the epicentral area: Virgin Islands and Puerto Rico. If the fault is oriented from west to east (source S1), as suggested by Reid and Taber (1920), the tsunami propagates mainly in south direction (Grenada, Trinidad and Isla de Margarita) and in east direction (Saba, St. Kitts, Antigua and, particularly, northern Guadeloupe). If the major axis of the tsunami source is selected as inclined 15–25° with the latitude (along the main axis of the Anegada Passage, sources S2–S4), the directivity diagram has also two peaks: the northern peak spreads from Saba to Guadeloupe and southern peak (Grenadines, Grenada) become weaker. The central part of Lesser Antilles (Dominica, Martinique and St. Lucia) is weakly affected by tsunami waves. According to the observations (Table 1), tsunami was significant in many islands of the Lesser Antilles (except its central part) and, of course, in Puerto Rico and Virgin Islands. Therefore, the theoretical

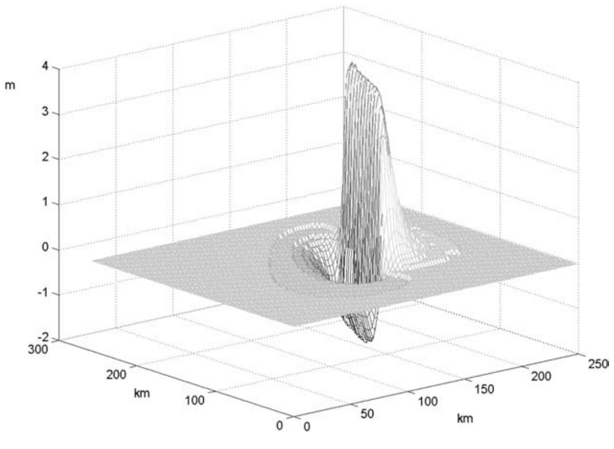

Fig. 5. Computed tsunami source (assumption S3) of the 1867 event.

model predicts correctly main ways of tsunami propagation in the vicinity of the Lesser Antilles. More correct selection of the preferable tsunami source can be done after the analysis of the tsunami height distribution along the coast and the comparison of the computed results with observed data.

The snapshots of the tsunami wave propagation for various orientations of the sources are almost similar and here they are given in Fig. 7 for the source S4 (25° angle of major axis with the latitude). Tsunami waves affected all islands of the Lesser Antilles for about 1 h. Previous calculations of the tsunami travel time for the tsunami generated at Charlotte Amalie (US Virgin Islands) give about 1.5 h (Weissert, 1990), but in his case the source centre is farther than tsunami source considered here (Fig. 5). After 2 h, there is the complicated picture of the tsunami waves in Lesser Antilles after reflection and diffraction on islands and shelf zone.

The computed time histories of water surface fluctuations for the source S4 at several locations in the epicentral area (Virgin Islands, Puerto Rica) are shown in Fig. 8. The crest amplitudes here are high, up to 6 m, and the trough depth is up to 6 m. Of course, the used bathymetry has no good resolution in the coastal zone (3 km), and the locations of the computed records cannot exactly correspond to the “real” locations of the observations and eyewitness reports at the coastal locations. But roughly they should describe observed features of tsunami. For instance, according to the observations, there are two giant waves generated by two shocks in the Virgin Islands and Puerto Rico with the time interval of 10 min. It was also observed that the initial motion of the sea was as the receding sea in most of locations. Our simulation includes the one shock only, so the theory should explain the one large wave in the epicentral zone. The computed water surface fluctuations at Frederiksted, St. Croix, Virgin Islands (Fig. 8) show definitely the huge wave up to 6 m coming after

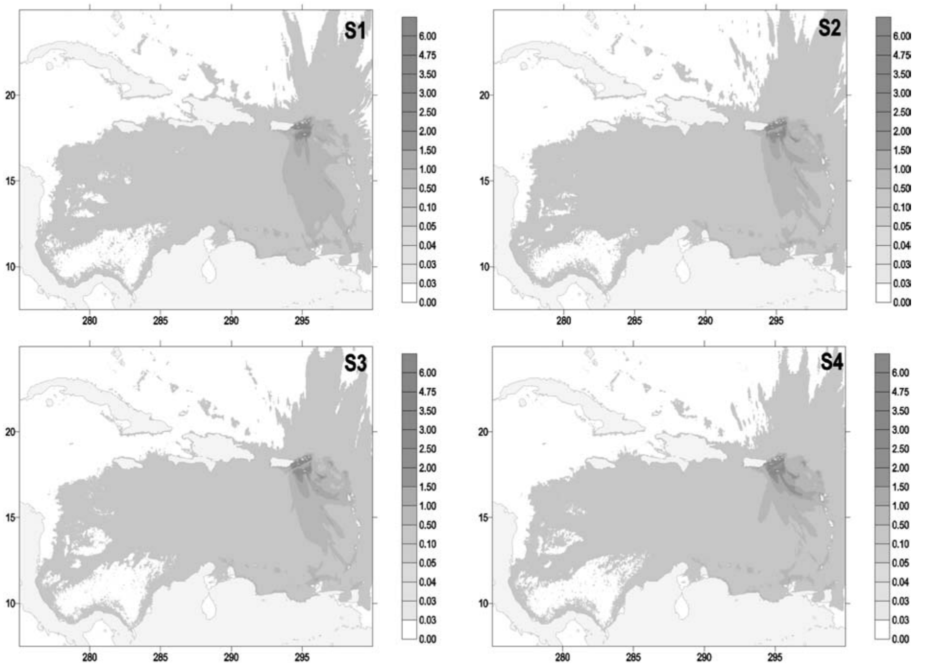

Fig. 6. The distribution of the maximal elevation of sea level (maximum positive amplitudes) for all sources (S1–S4).

the depression of 1–2 m in good agreement with observations. For Christiansted, St. Croix, the simulations predict three large waves of 3 m coming after the depression of 1 m. But the same features are not obtained for other places. In Cruz Bay, St. John tsunami began from the rise up to 3 m and the second wave arrived after deep depression on 6 m. The rising of sea to north from the source in our calculations corresponds to the seismic source model (elevation in northern part). Perhaps, the earthquake initiated the landslide; in this case, the depression of the wave is at shore side, and this can explain the observed sea receding on St. Thomas, St. John and Tortola. We would like to mention that the results (mainly, the number of large waves and their amplitudes) are very sensitive to the location of tide-gauges and the reproduction of the coastal line of these relative small islands (about of 10 grid points).

Fig. 9 describes the computed time histories of water surface fluctuations at several locations in the Lesser Antilles (source S4). The computed amplitudes here are less than 2 m. Tsunami approaches to the Lesser Antilles in 40–60 min after the earthquake, and this corresponds to the observations. In all cases tsunami come from the receding sea, and then large wave is arrived; it also corresponds to the observations. Tadepalli and Synolakis (1994, 1996) suggested the physical model of tsunami source with the leading depression wave and our source model is in agreement with the physical model.

In most of the locations, the first wave has maximal amplitude. Such a wave is evident on the computed record for Dehayes, Guadeloupe. Its crest amplitude exceeds 2 m, but significantly less than the value pointed by the witness report, 18 m. Results of our calculations show the appearance of the group of the tsunami waves, and this may be related as with the resonance effects between various islands, as well as with the propagation in the form of the edge waves along the Lesser Antilles. Significant oscillations of the sea level can continue 1 h or more and this corresponds to the observations. As it is pointed in Section 2, “the whole event lasted above 40 min” at the Grenadines, and there were six waves at Grenada. So, the computed results are qualitatively in reasonable agreement with data of observations.

The influence of the source orientation on the computed time series is shown in Fig. 10 for Dehayes, Guadeloupe. The results are very sensitive with the orientation of the source. The wave amplitude and the form of the wave train can vary significantly. Mainly, this variability is related with the wave propagation, diffraction and reflection in basin with complicated bathymetry and irregular coastal line, forming the directivity diagram (see, Fig. 6). The maximal tsunami height in the northern part of Guadeloupe can be expected for the source inclined to the latitude under large angle.

More detail information about the computed waves are summarized in Table 2. Each computed “tide-gauge” is provided by geographical coordinates; and they are given in table with high accuracy. The name of the coastal location near the computed tide-gauge is indicated only for navigation; as we have mentioned, the numerical model does not take into account the coastal zone and uses the reflected “vertical wall” as the boundary condition. We also give the depth in the place of the computed tide-gauge calculated from the used bathymetry Etopo2; it does not coincide with real depth at the location. Calculated maximum positive (+) and negative (–) elevations in meters are given in Table 2 for all tsunami sources. As we pointed, the wave height depends from the source orientation. According to the calculations, the amplitude of the 1867 tsunami exceeds 10 cm in most of

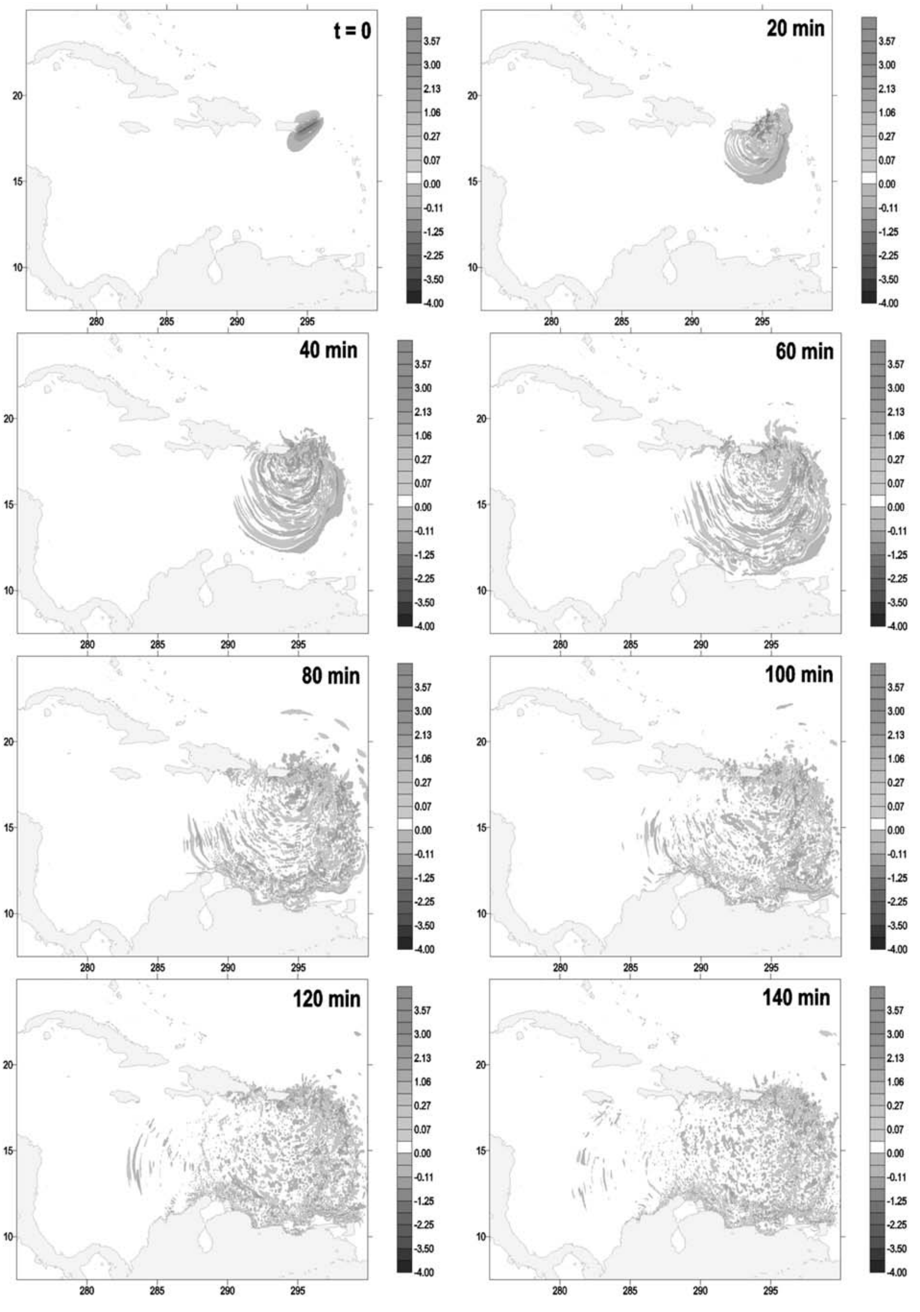

Fig. 7. Snapshots of the tsunami wave propagation at various times with interval 20 min (source S4).

the countries of the Caribbean (Puerto Rico, Dominican Republic, Haiti, Cuba, Nicaragua, Panama, Colombia, Venezuela, Lesser Antilles and Virgin Islands). Of course, to detect the waves with amplitudes about 10–50 cm in 1867 by eyes was very difficult, and most of them were invisible or not reported. The computed amplitudes exceeded 1 m are obtained for the Virgin and Lesser Antilles Islands and Puerto Rico, where tsunami waves are observed significantly. Therefore, the results of computing in average are in satisfied agreement with data of observations.

Comparison between the observed data and the numerical results for different sources is shown in Fig. 11. The points of observations are located along *x* axis starting from the point at west near the source (Puerto Rico) to the east to southern Lesser Antilles through the Virgin Islands (the exact locations of these points are presented in Figs. 1 and 3). The observed 10 m. runup in Dehayes and St. Rose (Guadeloupe) are in evident contrast with computed maximum positive amplitudes. Early, the height of 18 m was cited for these locations in Guadeloupe (Reid and Taber, 1920; Lander et al., 2002). After an inspection these places and investigation of historical materials, Zahibo and Pelinovsky (2001) suggested that the tsunami waves could not exceed 10 m. Perhaps, the positive amplitude of the wave near the coast significantly less, about 5 m; such waves can induce the damage described in literature (“wave broke over the shore and carried off all

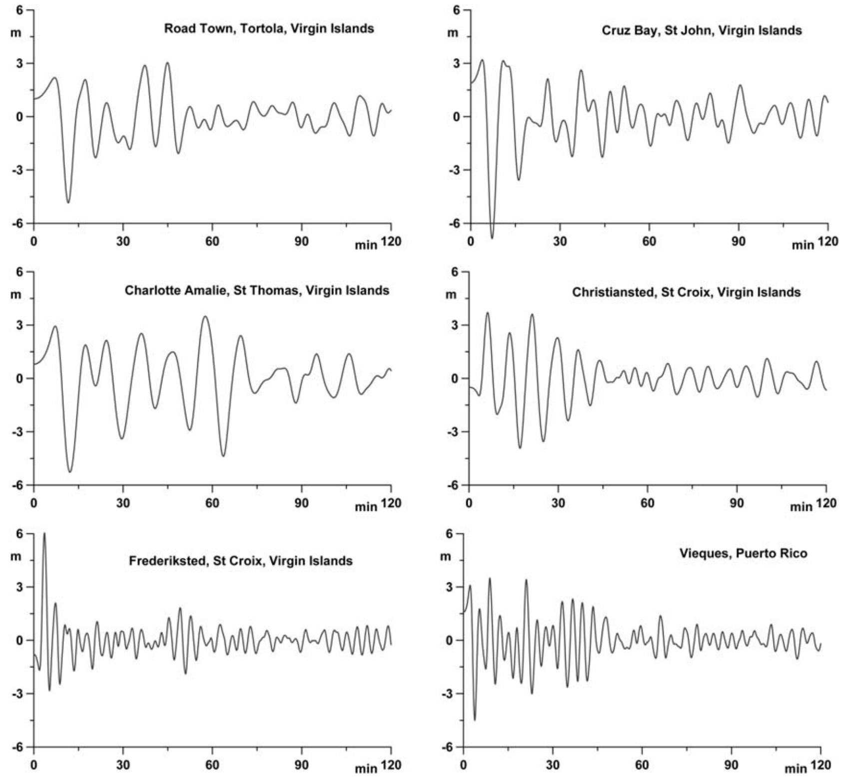

Fig. 8. The computed time histories of water surface fluctuations at several locations in the epicentral zone (source S4).

floatable objects"), but it is not confirmed. In this case, the correlation between observation and computation will be more evident. We may not expect that there is local bathymetry effect which is due to coarse grid used. Computed wave heights are maximal in the epicentral zone and also in the southern Lesser Antilles, in reasonable agreement with observations.

The formal definition of the correlation coefficient ($\Sigma H_{\mathrm{i\ obs}} H_{\mathrm{i\ com}} / \langle H_{\mathrm{obs}} \rangle \langle H_{\mathrm{com}} \rangle$) between observed amplitudes and computed amplitudes (source S4) gives the value 0.16, meanwhile the same with no observed data for Dehayes and St. Rose is 0.63. It gives the additional arguments in necessity to continue to examine the wave characteristics in the northern locations of Guadeloupe.

It is important to mention that tsunami waves are localized mainly in the Caribbean Sea, and the penetration of tsunami waves into the Atlantic through the Lesser Antilles straits and passages as well through the Virgin Island passages is relatively weak. The explanation is evident: tsunami waves effectively reflect and refract from the deepest Puerto Rico Trench behind the Caribbean Islands. As a result, the tsunami energy will mainly disperse in the Caribbean Sea.

5. Double wave generation during the 1867 event

As we mentioned in the previous section, there were two strong shocks with a time interval of 10 min. Reid and Taber (1920) emphasized that "a great sea wave was started by the first shock, and a second larger one by the second shock some 10 min later". We have no information about the characteristics of each shock in the epicentral zone that to simulate tsunami waves generated by two shocks. We may assume the second shock was on the same place with the same polarity, or it could be on opposite side of trench in the Anegada Passage, close to St. Croix (Fig. 4), in this case the polarity of water displacement should be opposite. In our numerical method, we can simulate the propagation of tsunami waves generated by two different shocks. But, in fact, the contribution of nonlinear terms is weak, and we may analyze the effect of the interference of two wave systems generated by each shock using the principle of the linear superposition of the obtained solutions. Let us assume that the second shock occurred 10 min later, and it has the same geometrical characteristics as the first shock, but its polarity may be as the same as well as opposite. The resulting wave at Dehayes, Guadeloupe (according to the orientation of the source S4) is presented in Fig. 12. If the single shock generated the wave train with the large first crest (see, Fig. 10), two shocks of the same polarity generate the wave train with the two large waves (the first and the third) of the almost the same amplitude; meanwhile, two shocks of the opposite polarity induce the wave train with the largest first and fourth waves. The maximal crest amplitude in all cases has almost the same magnitude of 2 m. The amplitude of the largest negative wave is increased significantly (up to 3 m), when the shocks induced the tsunami of opposite polarities. Due to linearity of considered problem, we may analyze the wave field formed by the first positive and second negative displacement in the

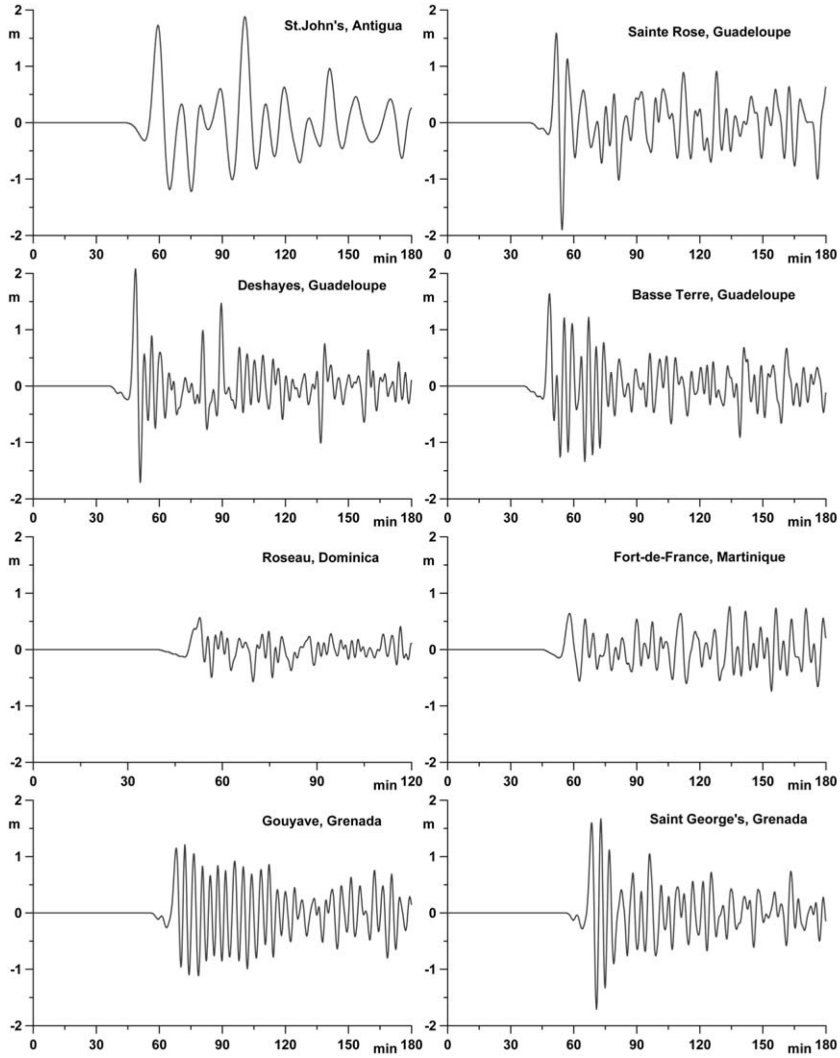

Fig. 9. The computed time histories of water surface fluctuations at several locations in the Lesser Antilles (source S4).

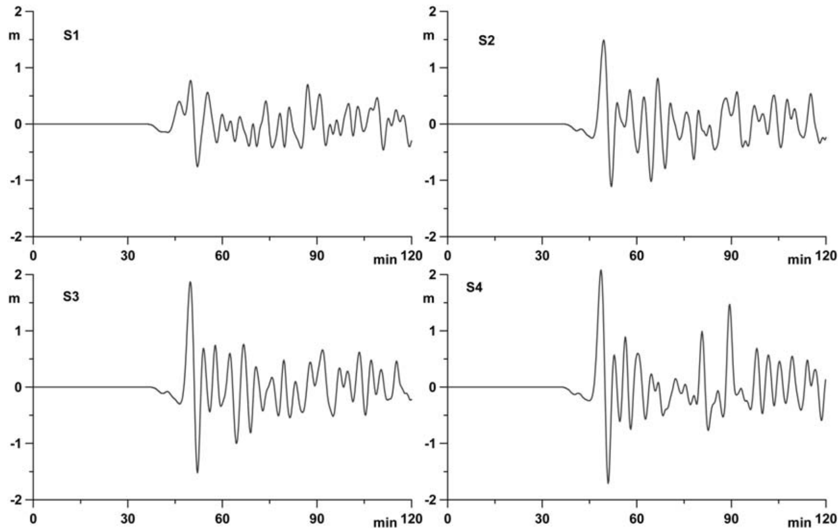

Fig. 10. The computed time histories of water surface fluctuations at Dehayes, Guadeloupe for various source orientations.

Table 2
Computed tsunami amplitudes (meter) in the Caribbean during the 1867 event

Location	Coordinates		*H* (m)	S1		S2		S3		S4	
	W	N		+	–	+	–	+	–	+	–
Guayama, Puerto Rico	66.028	17.970	23.5	3.9	5.1	4.1	3.9	3.3	3.3	3.3	3.6
Yabucoa, Puerto Rico	65.844	18.036	65.1	2.5	3.0	3.2	3.0	3.4	3.4	2.6	2.0
San Juan, Puerto Rico	65.906	18.398	26.8	0.5	0.5	0.9	0.9	1.0	0.9	0.8	0.7
Vieques, Puerto Rico	65.17	18.135	18.1	3.0	4.5	3.3	4.3	3.3	4.3	3.5	4.5
Charlotte Amalie, St. Thomas, Virgin Islands	64.924	18.365	59.8	2.9	5.1	5.9	7.7	6.1	7.9	3.5	0.3
Road Town, Tortola, Virgin Islands	64.587	18.497	77.4	2.5	4.4	2.3	4.9	2.7	5.2	3.0	4.8
Cruz Bay, St. John, Virgin Islands	64.702	18.332	22.8	5.7	7.5	3.7	7.0	4.1	6.4	3.2	6.9
Christensted, St. Croix, Virgin Islands	64.71	17.739	27.2	3.8	3.0	3.6	3.0	3.5	3.0	3.7	3.9
Fredericksted, St. Croix, Virgin Islands	64.894	17.74	74.9	6.0	2.8	5.3	2.8	5.1	2.4	6.0	2.8
St. John's, Antigua	61.857	17.081	9.2	0.8	0.8	1.3	0.9	1.7	1.0	1.9	1.2
St. Rose, Guadeloupe	61.735	16.39	26.9	0.9	0.8	1.2	1.1	1.4	1.6	1.6	1.9
Deshayes, Guadeloupe	61.827	16.291	21.3	0.8	0.8	1.5	1.1	1.9	1.5	2.0	1.7
Basse-Terre, Guadeloupe	61.765	16.028	42.9	0.6	0.8	1.5	0.9	1.5	1.1	1.6	1.3
Pointe-a-Pitre, Guadeloupe	61.489	16.225	67.1	0.4	0.6	0.7	0.7	0.6	0.6	0.7	1.0
Roseau, Dominica	61.428	15.237	48.2	0.4	0.4	0.6	0.4	0.4	0.5	0.6	0.6
Fort-de-France, Martinique	61.09	14.579	34.3	0.7	0.7	0.8	0.7	0.8	0.8	0.8	0.7
Kingstown, St. Vincent	61.275	13.229	28.8	0.7	1.0	0.9	1.1	1.0	1.1	0.8	1.0
Gouyave, Grenada	61.704	12.208	28.5	1.4	1.5	1.9	1.6	1.5	1.9	1.2	1.1
Saint George's, Grenada	61.765	12.109	15.9	2.3	2.8	2.2	2.4	2.4	2.4	1.7	1.7
Port-of-Spain, Trinidad	61.459	10.792	9.3	1.2	0.8	1.1	0.6	0.5	1.1	0.6	0.9
Isla de Margarita	63.943	11.221	8.5	0.9	1	1.1	1.1	0.9	0.9	0.7	0.6
Curmana, Venezuela	64.188	10.430	18.8	0.9	0.6	0.9	0.4	1.0	0.5	0.5	0.3
Barcelona, Venezuela	64.863	10.101	8.8	0.8	0.7	0.3	0.3	0.5	0.3	0.5	0.3
Caracas, Venezuela	67.47	10.529	8.4	0.4	0.5	0.6	0.6	0.6	0.6	0.6	0.6
Puerto Cabello, Venezuela	68.083	10.529	25.4	0.6	0.6	0.8	0.6	0.7	0.7	0.7	0.6
Bonaire, Netherlands Antilles	68.85	12.11	26.5	0.1	0.1	–	–	0.8	0.8	0.1	0.1
Cartagena, Colombia	75.475	10.661	48.9	0.5	0.4	0.1	0.1	0.2	0.1	0.2	0.1
Colon, Panama	80.168	9.212	9.7	–	–	–	–	0.2	0.2	–	–
Puerto Cabezas, Nicaragua	83.296	14.052	8.4	0.2	0.2	0.1	0.1	–	–	0.1	0.1
Santiago de Cuba, Cuba	75.843	19.979	7.8	–	–	0.1	0.1	–	–	0.2	0.2
Port-au-Prince, Haiti	72.408	18.596	6.0	0.3	0.3	0.4	0.3	–	–	0.2	0.2
Cayes, Haiti	73.758	18.168	27	0.3	0.3	0.4	0.4	0.3	0.4	0.3	0.3
Barahona, Dominican Republic	71.058	18.168	77.1	0.4	0.3	0.2	0.4	0.2	0.2	0.5	0.2
Santa Domingo, Dominican Republic	69.77	18.461	44.5	0.4	0.3	0.4	0.4	0.4	0.4	0.3	0.4
La Romana, Dominican Republic	68.912	18.398	24.2	0.9	0.8	1	0.9	0.2	0.2	0.5	0.3

southern part of the tsunami source. The resulting wave will have the polarity opposite to those shown in Fig. 12 (right). Inthis case, the maximal crest amplitude will be higher as a result of linear superposition; such a wave will be the third in the train. The intermediate waves (between the first wave and a wave of maximal amplitude) are not such visible, and the witness can describe only the two large waves.

Effects of the wave interference may induce also the increasing of the runup amplitude, as Tadepalli and Synolakis (1994, 1996) and Soloviev and Mazova (1994) shown theoretically. Due to small wave period, in the coastal zone the dispersion effects become important, and they may lead to the spatial-temporal focusing of the tsunami waves; these mechanisms are now studied actively (Pelinovsky et al., 2000; Mirchina and Pelinovsky, 2001). Perhaps, this mechanism is responsible for anomalous amplification of the tsunami waves in the northern part of Guadeloupe, but just now we have no enough information to check this hypothesis. We should also remind here the examples of tsunami events with the single initial wave by the occurrences of two main shocks with approximately 12 min intervals. They are 1956 Southern Aegean tsunami (Perrisoratis and Papadopoulos, 1999) and 1998 Papua New Guinea tsunami (Synolakis et al., 2002).

6. Conclusions

The 1867 tsunami in the Virgin Islands was recorded in many islands of the Caribbean Basin, in particular in Puerto Rico, Virgin Islands (St. Thomas, St. Croix, Tortola, Peter Is) and the Lesser Antilles (Saba, St. Kitts, Antigua, Guadeloupe, Grenadines, Grenada). Historical data of this tsunami are compared with the results of the numerical simulation. The mathematical model applied for the tsunami analysis is based on nonlinear long wave theory in the Cartesian coordinates. The bathymetry used is obtained from Etopo2 with spatial resolution 3 km. Numerical simulation of the tsunami propagation in the Caribbean has been performed for several

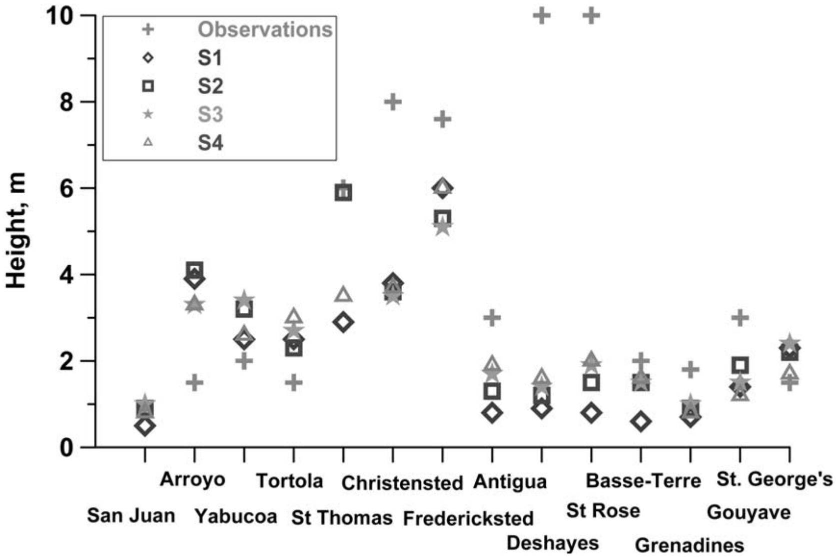

Fig. 11. Distribution of the crest amplitudes, comparison of computed and observed data.

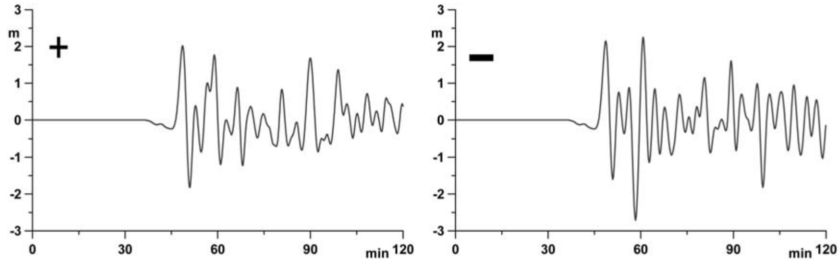

Fig. 12. Tsunami waves at Dehayes generated by two shocks of the same (+) or opposite (–) polarities.

orientations of the fault in the Anegada Passage. It is shown that the results are very sensitive to the source orientation and the location of the computed tide-gauges. But the directivity and the character of the wave distribution are varied relatively weakly. In the average, numerical results confirm the observed directivity of the tsunami propagation having two peaks in the northern and southern Lesser Antilles. Particularly, the observed form of the tsunami wave trains in different locations in the Caribbean is explained in the computed experiments. The distribution function of tsunami crest amplitude along the coast is in reasonable agreement with data of observations if the recorded 10 m heights of tsunami waves in the northern part of Guadeloupe (Dehayes and St. Rose) are ignored. This huge value of tsunami waves recorded in the Caribbean for whole history may come from the exaggeration or very local amplification of the wave. The tsunami wave interference due to two shocks with an interval in 10 min is also discussed.

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