Far field tsunami simulations of the 1755 Lisbon earthquake: Implications for tsunami hazard to the U.S. East Coast and the Caribbean

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A R T I C L E   I N F O
Article history:
Received 1 February 2008
Received in revised form 16 July 2008
Accepted 21 October 2008

Keywords:
tsunami modeling
1755 Lisbon earthquake
Azores–Gibraltar plate boundary
U.S. East Coast
Caribbean tsunami

A B S T R A C T
The great Lisbon earthquake of November 1st, 1755 with an estimated moment magnitude of 8.5–9.0 was the most destructive earthquake in European history. The associated tsunami run-up was reported to have reached 5–15 m along the Portuguese and Moroccan coasts and the run-up was significant at the Azores and Madeira Island. Run-up reports from a trans-oceanic tsunami were documented in the Caribbean, Brazil and Newfoundland (Canada). No reports were documented along the U.S. East Coast. Many attempts have been made to characterize the 1755 Lisbon earthquake source using geophysical surveys and modeling the near-field earthquake intensity and tsunami effects. Studying far field effects, as presented in this paper, is advantageous in establishing constraints on source location and strike orientation because trans-oceanic tsunamis are less influenced by near source bathymetry and are unaffected by triggered submarine landslides at the source. Source location, fault orientation and bathymetry are the main elements governing transatlantic tsunami propagation to sites along the U.S. East Coast, much more than distance from the source and continental shelf width. Results of our far and near-field tsunami simulations based on relative amplitude comparison limit the earthquake source area to a region located south of the Gorringe Bank in the center of the Horseshoe Plain. This is in contrast with previously suggested sources such as Marquês de Pombal Fault, and Gulf of Cádiz Fault, which are farther east of the Horseshoe Plain. The earthquake was likely to be a thrust event on a fault striking ~345° and dipping to the ENE as opposed to the suggested earthquake source of the Gorringe Bank Fault, which trends NE–SW. Gorringe Bank, the Madeira-Tore Rise (MTR), and the Azores appear to have acted as topographic scatterers for tsunami energy, shielding most of the U.S. East Coast from possible future earthquakes along the Azores–Iberia plate boundary. Additional simulations to assess tsunami hazard to the U.S. East Coast from the 1755 Lisbon tsunami. Additional simulations to assess tsunami hazard to the U.S. East Coast from possible future earthquakes along the Azores–Iberia plate boundary indicate that sources west of the MTR and in the Gulf of Cadiz may affect the southeastern coast of the U.S. The Azores–Iberia plate boundary west of the MTR is characterized by strike-slip faults, not thrusts, but the Gulf of Cadiz may have thrust faults. Southern Florida seems to be at risk from sources located east of MTR and South of the Gorringe Bank, but it is mostly shielded by the Bahamas. Higher resolution near-shore bathymetry along the U.S. East Coast and the Caribbean as well as a detailed study of potential tsunami sources in the central west part of the Horseshoe Plain are necessary to verify our simulation results.

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1. Introduction

The Azores–Gibraltar plate boundary is the source of the largest earthquakes and tsunamis in the north Atlantic basin. These include the 1941 M8.4 and 1975 M1979 strike-slip earthquakes west of the Madeira-Tore Rise (MTR) and the 1969, Ms 8.0 earthquake in the Horseshoe Plain south-east of the Gorringe Bank (Fukao, 1973; Buforn et al., 1988, 2004) (Fig. 1). This plate boundary is also believed to have been the source region of the 1761 tsunamigenic earthquake (Baptista et al., 2006) and of the great November 1st, 1755 Lisbon earthquake (Machado, 1966; Moreira, 1985; Johnston, 1996). The 1755 earthquake, which was estimated to be of magnitude Mw 8.5–9.0 (e.g., Gutscher et al., 2006), had the largest documented felt area of any shallow earthquake in Europe (Martinez-Solares et al., 1979; Johnston, 1996) and was the largest natural disaster in Europe in the past 500 years. It inflicted up to 100,000 deaths (Chester, 2001) through destruction by ground shaking, ensuing fires and tsunami waves of 5–15 m that devastated the coasts of Southwest Iberia and Northwest Morocco and were even reported as far north as Cornwall, England (Baptista et al., 1998a). Additionally, Grácia et al. (2003a,b) showed clear evidence of...
submarine landslide deposits from acoustic-backscattering, suggesting that the slope failure process could have contributed to tsunami generation and reports of tsunami waves along the European and Moroccan coasts.

The large tsunami-wave generated by the earthquake also caused damage in the eastern Lesser Antilles, as far north as Newfoundland, Canada and as far south as Brazil (Kozak et al., 2005; Ruffman, 2006). However, no reports were documented from cities along the U.S. East Coast (Reid, 1914; Lockridge et al., 2002; Ruffman, 2006). Table 1 summarizes the tsunami run-up reports from around the Atlantic Ocean (Frishman, 1755; Reid, 1914; Baptista et al., 1998a; O'Loughlin and Lander, 2003; Kozak et al., 2005; Ruffman, 2006). Fig. 2 shows relevant locations on the map as well as cities along the U.S. East Coast, which existed in 1755.

Although many attempts have been made to characterize the 1755 Lisbon earthquake and tsunami (Johnston, 1996; Baptista et al., 1998a; Gutscher et al., 2006; Grandin et al., 2007) only one study (Mader, 2001) had considered the far field effects of the tsunami. Studying far-field effects is advantageous in determining a possible source location and fault orientation because such effects are less influenced by near-source bathymetry and are unaffected by components of the tsunami wavefield generated by submarine landslides which are significant in the near-field (Gisler et al., 2006), but attenuate rapidly. Mader (2001) generated a numerical model for a source centered at the location of the Mw 7.8, 1969 earthquake (Fig. 1), which provided estimates of the deep water wave amplitudes along the U.S. East Coast and the Caribbean. However, the study did not attempt to characterize the earthquake's source parameters, using instead a 30-m vertical drop of a 300-km radius area as a source; nor did it endeavor to compare tsunami hazard along the U.S. East Coast and the Caribbean from different sources in the region.

In this study we first investigate constraints on the epicenter of the 1755 Lisbon earthquake from far field numerical tsunami simulations. Second, features such as fault orientation, distance from source, and near-source and regional bathymetry are tested in order to determine what governs tsunami propagation in the Atlantic Ocean. We then assess the tsunami hazard to the U.S. East Coast and the Caribbean from possible future earthquake sources located in the east Atlantic region.

2. Tectonic setting and the 1755 Lisbon earthquake

The eastern end of the Azores–Gibraltar plate boundary, which separates the Eurasian and African plates, is a region of complex bathymetry. Plate kinematic models together with focal mechanisms show that the motion between the two plates is slow (0.7–5 mm/yr) (Argus et al., 1989; Nocquet and Calais, 2004; Fernandes et al., 2007), changing along the boundary from extension in the Azores to compression towards the east that includes the Gorringe Bank and the Gibraltar arc (Fig. 1, inset). The precise location of the plate boundary

![Fig. 1. Plate tectonic setting (inset) and bathymetric map of the Iberian–African plate boundary.](image-url)
close to Iberia is uncertain and the plate boundary deformation there might be diffuse over a 200–330 km wide zone (Griffin and Chen, 1986; Hayward et al., 1999). The dominant active structures in this region are the Gorringe Bank Fault (GBF), the Marqués de Pombal Fault (MPF), the St. Vincente Fault (SVF) and the Horseshoe Fault (HSF), which have been studied by several authors (Sartori et al., 1994; Baptista et al., 2003; Grácia et al., 2003a; Terrinha et al., 2003). These structures and most of the faults in this area trend NE–SW (Borges et al., 2001; Zitellini et al., 2004; Buflorn et al., 2004) (Fig. 1).

Thus far the source of the great Lisbon earthquake remains unknown (Gutscber, 2004). A consensus attributed the origin of the earthquake to a structure located between the Gorringe Bank and the Coral Patch Ridge (Machado, 1966; Moreira, 1985; Johnston, 1996) (Fig. 1). Yet the relatively modest surface area of this fault region makes it difficult to explain the high seismic moment of $\sim 2 \times 10^{18}$ Nm, for a reasonable set of fault parameters (e.g., co-seismic displacement, rigidity, and recurrence) (Gutschber et al., 2006). Three major solutions were proposed based on seismic reflection and multibeam echosounder data, estimates of shaking intensity, and backward ray tracing of tsunami propagation. These fault solutions are shown in Fig. 1 and will be referred later in this paper as:


Marqués de Pombal Fault (MPF) — Zitellini et al. (2001) and Grácia et al. (2003a) suggested active thrusting along the MPF, located 80 km west of Cape Sao Vicente (strike 02°).


3. Methodology

3.1. Tsunami model simulations

All simulations presented in this study were generated using COMCOT (Cornell Multi-grid Coupled Tsunami Model) developed by P.-F. Liu, X. Wang, S.-B. Woo, Y.-S. Cho, and S.B. Yoon, at the School of Civil and Environmental Engineering, Cornell University (Liu et al., 1998). All calculations were performed on the Arctic Region Supercomputing Center in Alaska, using the Tsunami Computational Portal at: http://tsunamiportal.naoc.go.jp/wizard.php. COMCOT solves both linear shallow water (LSW) and non-linear shallow water (NLSW) equations in spherical coordinates. Two simplifying assumptions were made to create the initial sea surface deformation, which serve as the initial boundary conditions for the numerical simulations. First, the sea surface responds instantaneously to seafloor earthquake deformation. Second, the initial sea surface displacement is identical to that of the seafloor (Ruff, 2003). The initial sea surface deformation, computed based upon user-provided fault parameters, is identical to the seafloor displacement generated by Coulomb 3.0 (Lin and Stein, 2004; Toda et al., 2005; http://coulombs-tress.org). Aside from the governing equations, the difference in using linear vs. non-linear hydrodynamic models lies in the boundary conditions. The linear model uses reflective boundary conditions and is therefore unable to perform explicit run-up calculations at the shallow water areas along the coast. On the other hand, the non-linear model uses moving boundary conditions and is capable of explicit run-up calculations. The linear model was used in this study, because no attempt was made to calculate run-up. The output files used for all interpretations are depth and maximum wave amplitude files. The depth file contains the bathymetry of the region where the simulation took place. An ETOP02, 2551 $\times$ 1457 bathymetry grid with 2 resolution was used for all simulations. The maximum wave amplitude file contains the calculated maximum sea level amplitude for a selected region, throughout an entire simulation run (tsunami propagation time of 10–11.25 h).

3.2. Tsunami theory and numerical model limitations

Tsunami theory has been studied by many authors. The following section sums up tsunami theory based upon Liu et al. (1998) and Ward (2002). The leading wave of a tsunami has a wavelength proportional to the longitudinal dimensions of the earthquake source region, which could be several hundreds to a thousand kilometers for a major earthquake. It is considered to be a shallow water gravity wave, where the ocean depth is negligible compared to the wavelength. Its phase speed is proportional to $\sqrt{gh}$, where $g$ is the acceleration of gravity and $h$ is the water depth in meters. The wave period ranges between several hundreds to several thousand seconds. During propagation in deep water, tsunami wave slope is small, resulting in insignificant convective inertia forces, which can be ignored. As tsunamis propagate into the shallower water region, the wave amplitude increases and the wavelength decreases due to shoaling. The nonlinear convective inertia force becomes increasingly important. In the very shallow water, the bottom frictional effects become significant as well. Therefore, the nonlinear shallow water equations including bottom frictional terms should be used in the description of the tsunami inundation. In principle, numerical computation of wave heights based on linear shallow water equations is sufficient and...
accurate as long as the modeled tsunami wavelength is much greater than water depth and the wave amplitude is much smaller than water depth. This principle holds up until the deep part of the continental shelf. Consequently, this study is unable to provide definite run-up results and only relative amplitudes can be taken into consideration.

The time step chosen for each simulation must meet the Courant–Friedrichs–Lewy (CFL) condition (Courant et al., 1928) in order to assure numerical stability. The CFL condition for explicit numerical methods assures that the algorithm used for solving partial differential equations is convergent. For the COMCOT modified explicit scheme, the largest allowable Courant number is 0.8660 (Liu et al., 1998). Therefore, in order to assure stability the time step used in this study never exceeded 3 s.

3.3. Tsunami amplitude

Two methods were used to reliably calculate wave amplitude. First, the amplitude was calculated at depths of 250 m (see ‘shelf point’ in Fig. 3), similar to ten Brink et al. (Chapter 7, 2007), in selected sites along the U.S. East Coast, the Caribbean Islands, Europe, and Africa (Fig. 2). This depth falls within the minimal wavelength to grid size ratio (see Section 3.2 for detail), allowing for accurate propagation and amplitude calculations. Second, a rectangular patch of different sizes (Fig. 3) was chosen seaward of each location along the Atlantic, Caribbean, African and European coasts (Fig. 2). The average amplitude was calculated for all points within the depth range of 150 to 50 m in each patch. The size of the patches varied depending on the geographical locations where the amplitudes are measured. Along the U.S. East Coast for instance, where the shelf is wide, larger patches were selected to account for as many points as possible within the 150 to 50 m depth range. In the Caribbean, where the shelf is narrower, smaller patches were sufficient to incorporate a representative number of points in the same depth range. Although amplitudes calculated at such shallow depths may be inaccurate in terms of their geographical locations, averaging them out over a large area gives a good indication of the wave amplitude in that particular region. This method also verifies that the amplitude calculated at a nearby shelf edge point of 250 m depth is not anomalous. Figs. 4a and b shows a comparison between amplitudes calculated using the two methods, from an earthquake source located in location 8 (Fig. 3). Indeed, the average amplitudes calculated in the patches in the shallower water show similar or higher amplitudes in comparison to the ones calculated in the slightly deeper shelf edge points, as one would expect from the amplification effects of shallow waters.

3.4. A method to overcome unreliable historical reports of run-up observations

Caution must be exercised when using historical reports in order to compare between possible epicenter locations. Table 1 shows the variability of run-up amplitudes in historical reports, particularly in the Azores, Madeira, Lisbon and Tangier. It is therefore impossible to compare our model results to individual run-up reports. Moreover, run-up amplitudes are highly sensitive to the near shore bathymetry and onshore topography whereas, because of the model limitations discussed in Sections 3.1 and 3.2, amplitudes were calculated at a water depth of 250 m. We therefore grouped together places in the Caribbean, along the Portuguese and Moroccan coast, in Madeira and the Azores, as locations representing consistent reports of high amplitudes. Earthquake sources generating high tsunami amplitudes in those locations are therefore assigned as a good fit to the 1755 Lisbon earthquake epicenter. Similarly, we joined together places along the U.S.
East Coast and in Vigo and La Coruña in the northern Spanish coast, under a category of places where no historical reports were documented (i.e., negative evidence). Blanc (2008) quotes a French consular letter from 1755 about a tsunami striking La Coruña, with a crest to trough amplitude of 1.95 m, which frighten people but caused no damage (Frishman, 1755). The particular locations along the U.S. East Coast (with the exception of Virginia Key in Florida), and Vigo in Spain, were chosen because they were already populated at the time of the earthquake yet there were still no tsunami reports found in the literature. In places along the U.S. East Coast, the tsunami should have struck during daylight hours. The semi-diurnal tidal ranges along the U.S. East Coast are 3 m and the difference between the times that high-tide reaches different locations along the East Coast is as large as 5 h. Therefore, had a significant tsunami impacted the U.S. East Coast, some sites there would have experienced flooding during high tide. In NW Spain, both the time the tsunami should have struck and the tide conditions are similar to the other locations further south along the coast. Therefore, neither tidal variations nor time of the day are likely to explain the absence of reports in these locations. Table 2 summarizes the criteria used to group the historical reports.

In order to quantify the results we compared and normalized the amplitudes of all sources relative to source 5 (shown in Fig. 3). For each location $j$ out of a total of $n$ along the coasts (shown in Fig. 2 and Table 1) where no amplitudes were reported, we calculated the amplitudes of different model sources relative to that of source 5 using:

$$Amp_{ij}^\text{min} = \sum_{k=1}^{m} (Amp_k - Amp_j) / Amp_5$$

where $i$ represents the 16 model epicenter locations shown in Fig. 3. A better fitting epicenter location for any one of the examined model locations along the coasts would generate wave amplitudes lower than that of source 5 and, thus, receive a positive rating relative to source 5. Similarly, for each location $k$ out of a total of $m$ where high amplitudes were reported (shown in Fig. 2 and Table 1), we calculated the amplitudes of the sources relative to that of source 5 using

$$Amp_{ij}^\text{max} = \sum_{k=1}^{m} (Amp_k - Amp_j) / Amp_5$$

where $i$ represents the 16 epicenter locations shown in Fig. 3. A better fitting epicenter location for any one of the locations along the coasts would generate wave amplitudes higher than source 5 and, consequently, receive a positive rating relative to source 5. As a result, the best fitting source $i$ should maximize:

$$[Amp_{ij}^\text{min} + Amp_{ij}^\text{max}].$$
Figs. 5–7 and 17 were created using Eqs. (1)–(3). Similar results were also obtained when we excluded the Azores, Madeira and Lisbon, where there was a large variation in the reported run-up amplitude, from the calculations.

4. Results

Fig. 3 and Table 3 show all the earthquake sources that were modeled. To facilitate a meaningful comparison among the models, and for lack of detailed geologic constraints for any of the sources, all the models used the same fault dip, dimensions, slip and rigidity (Table 4) as those proposed for GBF (Johnston, 1996). Gorringe Bank is the most prominent morphological feature in the area and was suggested to be capable of generating an earthquake with a moment magnitude of $1.26 \times 10^{22}$ Nm, similar to the one calculated for the 1755 Lisbon earthquake (Johnston, 1996). The rigidity value used for the moment magnitude calculation was very high ($6.5 \times 10^{10}$ Pa), to account for a fault that is almost entirely within oceanic mantle lithosphere (Johnston, 1996). Furthermore, the use of a pure thrust fault with rake 90°, would result in the highest possible transoceanic tsunami amplitudes (see Geist, 1999), enabling us to test each individual feature that govern tsunami propagation, separately.

4.1. The effect of fault orientation on tsunami propagation and amplitudes

The first set of simulations was designed to examine the effect of strike orientation on tsunami propagation. Source 3 was chosen for this set because it is the one least susceptible to near-source bathymetric effects in the fault region. The fault strike was rotated 360° at 15° interval. Fig. 8 shows the variations of maximum wave amplitude as a function of fault orientation, for sites along the U.S. East Coast and the Caribbean. A pattern of two maxima at fault strikes of 165°–180° and 345° yields the highest amplitudes in the Caribbean. A fault strike of 345° is the equivalent to a thrust fault dipping to the ENE (see dashed fault over source 3 in Fig. 3) and was chosen as a reference configuration. In this configuration, the leading westward propagating wave is a depression phase (ocean withdrawal), followed by an elevation phase (flooding), in agreement with observations from Madeira (Reid, 1914), Brazil (Kozak et al., 2005; Ruffman, 2006), Newfoundland (Ruffman, 1990), and the Caribbean (O'Loughlin and Lander, 2003). The minima are for fault strikes of 75°–90° and 270°–285°. Note that GBF, which was suggested as a possible source for the 1755 Lisbon earthquake (Johnston, 1996), has the worst fit.
4.2. The effect of different source locations on tsunami propagation and amplitudes

A fault strike of 345° yields the highest amplitudes in the Caribbean in accordance with historical reports and was therefore used when searching for fault location of the 1755 Lisbon earthquake (see Section 4.3). Sixteen fault locations were modeled as tsunami sources in the region of study (Fig. 3) and tsunami amplitudes were calculated in locations along the U.S. East Coast and the Caribbean as well as along the European and African coasts.

Fault orientation for all locations was assumed to be 345° following the analysis in Section 4.1. Fig. 5 shows a comparison between the different source locations relative to source 5. Based on the method outlined in Section 3.4, only source 8 fits better than source 5 and source 2 fits slightly worse. Note that source locations 8, 5, and 2 are all located within the Horseshoe Plain. Fig. 7 shows a comparison between source 5, source 8 and the three previously suggested source locations GBF, MPF, and GCF. It is clear that these three source locations are the ones that are better fit to the observations than sources 5 and 8. Figs. 9–11 show maximum wave amplitude plots from earthquake sources located in GBF, GCF and MBF respectively. Figs. 9 and 10 highlight the same conclusion that is portrayed graphically in Fig. 7. The maximum wave amplitude generated from GBF (060°) is seen in a direction that is almost perpendicular to that observed by the historical reports. As a result, the Caribbean Islands are unaffected. Contrary to historical reports the wave amplitudes along the U.S. East Coast, generated from GCF (349°) are high (~0.5 m) and spread over a relatively wide area (as far north as Charleston). MPF from Fig. 11 cannot be discounted, because it shows that the U.S. East Coast remains relatively untouched and high wave amplitudes are seen in the direction of the Caribbean, thus in agreement with historical reports. Nevertheless, the results shown in Fig. 7 as well as comparing between MPF and sources 5 and 8 (Figs. 13 and 14), indicate that MPF is less likely to be the 1755 Lisbon earthquake source.

4.3. The 1755 Lisbon earthquake epicenter and fault strike

Figs. 5 and 7 indicate that the most likely epicenter of the 1755 Lisbon earthquake according to our model simulations is in the Horseshoe Plain area of sources 5 and 8 and not in the previously suggested locations: GBF, MPF and GCF. The Horseshoe Plain area is characterized by high seismicity and is cut by NE–SW trending thrust faults which reach the seafloor (e.g., Sartori et al., 1994; Zitellini et al., 2004). Figs. 6 and 8, however, illustrate that the fault was most likely trending NW–SE as opposed to the previously interpreted NE–SW

Table 4
Fault parameters used for all simulations.

<table>
<thead>
<tr>
<th>Source depth (km)</th>
<th>Fault length (km)</th>
<th>Fault width (km)</th>
<th>Average slip (m)</th>
<th>Dip (deg)</th>
<th>Rake (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>200</td>
<td>80</td>
<td>13.1</td>
<td>40</td>
<td>90</td>
</tr>
</tbody>
</table>

Source depth corresponds to the top of the fault plane.
strike orientation. The only known tectonic feature with a NW–SE trend in this area is the inferred Paleo Iberia–Africa Boundary (PIAB), the equivalent structure to the Newfoundland transform fault on the North American plate, which was formed during the opening of the central Atlantic ocean in the Late-Jurassic–Early Cretaceous (Rovere et al., 2004) (Fig. 1). However, further seismic and multibeam investigations of the west Horseshoe Plain are necessary to test if the PIAB is currently active.

4.4. Near field tsunami travel times

Constraining source location based on tsunami travel time is problematic (Gutscher et al., 2006) due to the inaccuracy of historical reports (e.g., a 30 min difference in arrival time between Porto Santo and Madeira Islands which are only 50 km apart), due to the possibility of landslide-generated tsunamis, and due to the difficulties in simulating tsunami propagation at shallow depths (see Section 3.2)
Nevertheless, we computed travel times to locations of historical reports assuming simple aerial distance, tsunami phase speed of \( \sqrt{gh} \) with water depths ranging from 2500 m to 4500 m for sources 5 and 8 and 1000 m to 4000 m from source 2 (Table 5). Travel times from historical reports were listed by Baptista et al. (1998a) and Gutscher et al. (2006). Although source location 2 (near MPF) seems to be the best with respect to some of the historical reports, the overall time differences between source location 2 and sources 5 and 8 are minor, implying that an epicenter located farther west than source 2 is possible.

5. Discussion

5.1. The effects of regional and near-source bathymetry on tsunami propagation and amplitude

Regional and near-source bathymetry have a significant effect on tsunami propagation in the Atlantic. In a hypothetical case lacking bathymetric features, a tsunami is expected to propagate uniformly in all directions along great circle paths. Fig. 12 shows a plot of maximum wave amplitude across the Atlantic ocean from source 5. The black lines indicate great circles from earthquake source 5 to different locations along the U.S. East Coast and the Caribbean. The trace of relatively high wave amplitudes in the direction of Virginia Key in southern Florida represents the only wave packet closely following a great circle. All other wave amplitude traces relevant to the locations along the U.S. East Coast and the Caribbean suggest that the corresponding wave packets were either dispersed or deflected by various bathymetric features. Figs. 13 and 14 show a maximum wave amplitude plot from sources 8 and 5 focusing on far-field and near-source effects, respectively. Fig. 14 suggests that the wave propagating eastward toward the Portuguese coast is unaffected by deep ocean bathymetry, whereas Fig. 13 implies that propagation westward has a fingering pattern due to wave scattering by bathymetry. The near-source bathymetric elements causing such scattering are the Gorringe Bank, the Ampere and Coral Patch seamounts as well as Madeira Island and the MTR. These bathymetric elements are shallower than 1500 m, which is the minimal depth required to scatter a tsunami wave according to the analytical analysis of Mofjeld et al. (2000). The

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Table 5

Comparison of historically observed tsunami arrival times with calculated arrival times from sources 5, 8 and 2 (S5, S8 and S2) in Fig. 3 and compared to calculated arrival times from two sources (1 and 2) at the Marques de Pombal "source B" (N 160) and "source C" (N 160 N 115) (Baptista et al., 1998b) and a source in the Gulf of Cadiz (Gutscher et al., 2006).

<table>
<thead>
<tr>
<th>Location</th>
<th>Historical time</th>
<th>Travel time S5</th>
<th>Travel time S8</th>
<th>Depth (m)</th>
<th>Travel time S2</th>
<th>Depth (m)</th>
<th>Travel time MPF1</th>
<th>Travel time MPF2</th>
<th>Travel time GCF</th>
</tr>
</thead>
<tbody>
<tr>
<td>St. Vincente</td>
<td>16±7</td>
<td>16–17</td>
<td>21–22</td>
<td>4000–3500</td>
<td>16–19</td>
<td>1500–1000</td>
<td>23</td>
<td>21</td>
<td>22</td>
</tr>
<tr>
<td>Huelva</td>
<td>50±10</td>
<td>39–44</td>
<td>45–51</td>
<td>2500–2000</td>
<td>39–47</td>
<td>1500–1000</td>
<td>80</td>
<td>74</td>
<td>52</td>
</tr>
<tr>
<td>Cadiz</td>
<td>78±15</td>
<td>43–48</td>
<td>50–56</td>
<td>2500–2000</td>
<td>44–54</td>
<td>1500–1000</td>
<td>70</td>
<td>70</td>
<td>36</td>
</tr>
<tr>
<td>Madeira</td>
<td>90±15</td>
<td>54–57</td>
<td>49–52</td>
<td>4500–4000</td>
<td>64–68</td>
<td>4000–3500</td>
<td>78</td>
<td>78</td>
<td>72</td>
</tr>
<tr>
<td>Figueira</td>
<td>45±10</td>
<td>52–58</td>
<td>54–61</td>
<td>2500–2000</td>
<td>61–75</td>
<td>1500–1000</td>
<td>53</td>
<td>50</td>
<td>83</td>
</tr>
<tr>
<td>Porto</td>
<td>63±71</td>
<td>63–71</td>
<td>66–74</td>
<td>2500–2000</td>
<td>76–94</td>
<td>1500–1000</td>
<td>90</td>
<td>87.5</td>
<td>96</td>
</tr>
</tbody>
</table>

All times are in minutes.
Fig. 12. Maximum wave amplitude projected on a sphere from an earthquake source located in source 5. The scale ranges from 0–1 m. Warm colors indicate high amplitudes and cold colors low amplitudes. Black lines indicate great circle paths between source 5 and locations along the U.S. East Coast and the Caribbean. The wave energy heading toward Virginia Key in southern Florida is the only one following a great circle path. All other wave energies are scattered by topography.

Fig. 13. Maximum wave amplitude from the best fit earthquake source located in source 8. Scale — same as in Fig. 9. Wave scattering is mainly caused by the Madeira Island, Madeira Tore-Rise (MTR), the Azores, the Great Meteor (CM) and Cruiser (Cr) seamounts. The ray passing in between the Azores and the Great Meteor seamount reaches southern Florida. The rest of the U.S. East Coast is relatively unaffected by the tsunami. NBB—northern Bahamas Banks.
energy is first highly influenced by the Ampere and Coral Patch seamounts as well as the MTR and Madeira Island. Farther to the west, wave propagation seems to be influenced by the Mid-Atlantic ridge; in particular the Azores and the Great Meteor and Cruiser seamounts. Higher amplitudes are shown in the vicinity of these bathymetric elements. However, the wave amplitudes decay quickly behind these bathymetric features because these features tend to attenuate the low frequency waves. On the other hand, tsunami wave energy is inferred to be traversing through the deep part of the MTR (arrow in Fig. 14) and later in between the Azores and Great Meteor seamounts, following a great circle toward southern Florida; this wave phase maintains its low frequency content and reaches its trans-Atlantic destination with much higher amplitude. We believe the reason why there are no reports from the 1755 tsunami in southern Florida could be attributed to the northern Bahamas Banks (NBB) which may have acted as a barrier to that area. The rest of the U.S. East Coast remains relatively protected. The northern part of the MTR may have played an important role in shielding the United States, scattering wave energy in that direction. Similarly, the Coral Patch and Ampere seamounts as well as Madeira Island seem to partially scatter the energy in the direction of the Caribbean. The same energy is later scattered a bit more by the Great Meteor and Cruiser seamounts. It is possible that the trace of relatively high amplitudes southward of the Great Meteor seamount may correspond to refracted tsunami energy, responsible for run-up reports in Brazil (Kozak et al., 2005; Ruffman, 2006). Scattering energy by seamounts, however, is relatively ineffective (Mofjeld et al., 2000), allowing enough energy to reach the Caribbean, thus explaining the historical reports. Additional simulations using high-resolution near-shore bathymetry could verify the historical reports claiming that some islands in the Caribbean have experienced greater run-ups than others. Historical run-up reports
exist for the entire Antilles arc beginning in Santiago de Cuba and ending in Barbados with the exception of San Juan, Puerto Rico. A possible explanation for the absence of a tsunami report from San Juan is the presence of the ultra-deep Puerto Rico trench (−8350 m) north of San Juan, which may have deflected the energy of the ray path that arrived in a sub-critical angle (Mei, 1989; Mofjeld et al., 2000). The waves propagating northward (as indicated from the high wave amplitudes), amid the Gorringe Bank and the Josephine seamount and then passing north of the Azores, may have eventually reached Newfoundland, Canada, explaining the historical reports there. Finally, the wave energy that passed southward east of the Coral Patch seamount may explain the historical reports in the Canary Islands (Reid, 1914).

5.2. Implications to tsunami hazard to the U.S. East Coast

The effect of near-source bathymetry on tsunami propagation was tested in order to assess tsunami hazard to the U.S. East Coast from possible future earthquakes in the study area. Two sources were compared: one east and one west of the MTR because both regions have the potential to generate sufficiently strong earthquakes (Buforn et al., 1988). For both sources the maximum wave amplitude was calculated for fault strike orientations varying from 0°–360° at 15° interval as described in Section 4.1. The wave amplitudes were then averaged out over 360° and measured at deep water locations 3500 and 4000 km (shown by asterisks in Fig. 2) from sources 16 and 3, respectively. These deep water locations lie along the azimuths of the U.S. East Coast and the Caribbean coastal sites. A 10% amplitude reduction was factored in to compensate for the difference in distance between 3500 and 4000 km (Ward, 2002) in order to properly compare between the two sources (Fig. 15). If bathymetry had no effect on wave propagation one would expect wave amplitudes to be identical. The fact that amplitudes vary further demonstrates the significant effect of the bathymetry on transatlantic tsunami propagation. The calculations from source 3 illustrate an amplitude distribution pattern very similar to that depicted in Fig. 10 with a maximum in the direction of Virginia Key. Wave amplitudes from an earthquake source west of the MTR (source 16) show an entirely different amplitude distribution pattern, revealing higher amplitudes in the direction of Baltimore and southward down to Cape Hatteras (Azimuth 292 from source), signifying possible tsunami hazard to these regions. All other places calculated from source 16 show a decrease in amplitudes, except for the waves heading towards Charleston, while the amplitude for Dominica remains relatively unchanged. Fig. 16 shows a maximum wave amplitude plot from source 16, for a fault with a strike of 30°, west of and adjacent to the MTR. This plot may suggest a possible greater hazard to the U.S. East Coast from earthquakes located in the region west of MTR. We should note, however, that the region west of MTR has so far generated only

![Fig. 16. Maximum wave amplitude from an earthquake source located in source 16 and oriented 30°. Scale — same as Fig. 9. High amplitudes are seen in a wider area along the U.S. East Coast relative to Fig. 13, highlighting the greater hazard from earthquake sources located west of MTR.](image)

![Fig. 17. Comparison between all of the modeled sources relative to source 5, for sites along the U.S. East Coast (see Section 3.4 for explanation). See Fig. 3 and Table 3 for source locations. Positive bars represent sources that may have a lower impact than source 5 on the U.S. East Coast. Negative bars represent sources that are calculated to have greater impact than source 5 to the U.S. East Coast (see Section 5.2). Sources 1, 3, 12, 16 and 10 are calculated to have the greatest impacts to the U.S. East Coast.](image)
strike–slip earthquakes (Grimson and Chen, 1986; Bufrin et al., 1988) and relative motion there is predicted by plate kinematic models to be strike–slip (Argus et al., 1989; Nocquet and Calais, 2004). Fig. 17 compares all the different earthquake sources relative to source 5 with respect to the U.S. East Coast only (excluding the Virginia Key), in the same way described in Section 3.4. In all cases the fault strike was 345°, because it yields the highest amplitudes in the direction of the United States, as shown in Fig. 8. Source locations 3 and 1 in the Gulf of Cádiz and locations west and north of the Gorringe Bank are calculated to generate the highest amplitude tsunamis along the U.S. East Coast, highlighting the potential hazard from these sources. Fig. 10 further demonstrates the potential tsunami hazard to the U.S. East Coast from earthquake sources located in the Gulf of Cádiz. Fig. 11, on the other hand, shows low tsunami risk from an earthquake source located in the MTF. We can therefore conclude that the Gorringe Bank and the north MTR may protect the U.S. East Coast from earthquakes in the Horseshoe Plain, the MTF, the SVF and their surrounding area, but not from the Gulf of Cádiz. Finally, it is important to note that only thrust earthquakes, roughly striking northward may pose tsunami hazard to the U.S. East Coast.

5.3. Other considerations — shelf width

The continental shelf along the U.S. East Coast is much wider than along the Caribbean Islands. The large shelf width may have contributed to the dissipation of tsunami amplitude along the U.S. East Coast and is perhaps one reason for the lack of historical reports from the 1755 Lisbon tsunami. Due to the limitations imposed by the low-resolution bathymetry (Section 3.2), we were unable to quantitatively calculate the shelf width effect on wave amplitudes. Nevertheless, Figs. 12 and 13 illustrate that amplitudes in southern Florida are higher than in other areas along the East Coast although the continental shelf in Florida is wider. This suggests that shelf width affects tsunami propagation and amplitudes less than the source fault strike orientation and the seafloor bathymetry along the wave paths.

6. Conclusions

Methodological tsunami simulations based upon historical reports of both far field and near field effects of the November 1st, 1755 Lisbon tsunami suggest three important conclusions: First, the earthquake seems to have been generated by a NW–SE trending fault located in the center of the Horseshoe Plain, south of the Gorringe Bank. This orientation is almost perpendicular to previously suggested NE–SW trending faults such as GFB and structures south of the Gorringe Bank (Zitellini et al., 2001). The only known tectonic structure with a NW–SE orientation in this area is the PIAB, although its potential for reactivation remains ambiguous. Moreover, the modeling results allow us to discount the GCF and to a lesser extent the MPF, because both are located too far to the east of the Horseshoe Plain. The GCF can be discounted as a tsunami source because it is predicted to generate relatively high wave amplitudes along the U.S. East Coast, and relatively low ones along the Caribbean. The orientation and location of the MPF are slightly less favorable than our preferred sources in the Horseshoe Plain, even when considering historical reports of tsunami arrival times.

Second, seafloor bathymetry is a significant factor in dictating transatlantic tsunami propagation. In particular, the bathymetry of the Gorringe Bank, the MTR (Josephine Seamount) and the Azores allows waves to reach Newfoundland, but blocks them from reaching most of the U.S. East Coast, with the exception of southern Florida. The Ampere and Coral Patch seamounts, Madeira Island, and the Great Meteor and Cruiser seamounts reduce wave propagation toward the Caribbean. The latter two features partially refract wave energy toward Brazil. Furthermore, high run-up reports in the Caribbean are most likely due to the steep rise in the bathymetry near to shore.

The third conclusion concerns tsunami hazards to the U.S. East Coast from sources located along the eastern Iberian–African plate boundary, which generate sufficiently strong thrust earthquakes. The Gorringe Bank and the north MTR act as near source barriers, protecting most of the U.S. East Coast. For sources located east of MTR and south of the Gorringe Bank, Florida might be at risk if sufficient wave energy manages to pass through the Bahamas. Sources in the Gulf of Cádiz may present a wider hazard to the U.S. East Coast, because they are sufficiently south as to not be affected by the Gorringe Bank, north MTR, and the Azores. For sources located west of the MTR, the risk is shifted northward in the direction of Baltimore. It is important to note that the interpretations in this report considered relative amplitudes only. High resolution near-shore bathymetry is crucial for more accurate run-up calculations and tsunami hazard assessments.

Acknowledgments

Propagation models were obtained using the Cornell Multi-grid Coupled Tsunami Model (developed by Liu et al., 1998) in the Tsunami Computational Portal; the Portal is a joint project of the Northwest Alliance for Computational Science & Engineering at Oregon State University (www.nacse.org) and the Artic Region Supercomputing Center (www.arsc.edu) at the University of Alaska-Fairbanks. We thank Dylan Keon and Harry Yeh (Oregon State University) and Tom Logan and Elena Suleimani (Arctic Region Supercomputing Center) for assistance with running the models. We would further like to thank Xiaoming Wang (Cornell University) for technical consultations and assistance, Eric Geist (U.S. Geological Survey, Menlo Park) for fruitful discussions, and Hyun-Sook Kim, Jason Chaytor, and Brian Andrews (U.S.Geological Survey, Woods Hole) for technical assistance. Paul-Louis Blanc, Institut de Radioprotection et de Sûreté Nucléaire, France, kindly provided a copy of Frishman letter from 1755. Helpful reviews by Eric Geist and Alberto Lopez, two anonymous reviewers, and Editor David Piper are gratefully acknowledged. Roy Barkan thanks the Woods Hole Oceanographic Institution Summer Student Fellowship Program for their financial support and hospitality. Roy Barkan would further like to thank Shmulik Marco (Tel Aviv University) for providing the technical means to complete this study. This study was funded by the U.S. Nuclear Regulatory Commission under NRC Job Number N6408 Physical study of tsunami sources.

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