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Short Notes

In Search of the 31 March 1761 Earthquake and Tsunami Source

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Abstract Earthquake catalogs for the Iberian Peninsula report three strong-magnitude events in the eighteenth century: 27 December 1722, 1 November 1755, and 31 March 1761. These events have magnitudes greater than 7 and generated tsunamis that damaged the Portuguese coasts. However, their source areas are controversial because of the lack of detailed and coherent historical descriptions.

The 31 March 1761 earthquake was felt in Lisbon at noon, alarming the inhabitants and throwing down ruins of the past 1 November 1755 earthquake. According to several sources the earthquake was followed by a tsunami that was observed as far as Cornwall (United Kingdom), Cork (Ireland), and Barbados (Caribbean). The Portuguese catalogs locate this event on the Horseshoe Abyssal Plain, south of Gorringe Bank, and attribute a magnitude of 7.5. The Caribbean tsunami catalog (Lander *et al.*, 2002) locates the event further north 37° N 10° W and estimates of its epicenter intensity as IX.

In this study we present a reappraisal of the available historical reports concerning the 1761 event, a revision of the macroseismic intensities along Iberia, and the tsunami observations along the western Portuguese and Galicia coasts, England, Ireland, and the West Indies. With this dataset we use backward raytracing techniques to discuss the location of the event and its integration with one of the major tsunami generation areas in the western Portuguese margin. We conclude that the 31 March 1761 earthquake took place at 12:01 a.m. (Lisbon time). Its epicenter was located about 34.5° N 13° W and had a tsunami magnitude close to 8.5.

Online material: Felt reports from the 31 March 1761 earthquake.

Introduction

Tsunamis are among the world's most destructive coastal hazards. The increase in coastal population and the development of large leisure areas set up a scenario with a large potential of disaster. Portugal has an important record of historical tsunamis, which several times damaged the city of Lisbon and the coasts of Algarve (south Portugal). Instrumental data, on the contrary, are scarce and so historical events must be accurately studied to provide some information for a quantitative evaluation of tsunami risk.

The 1761 earthquake and tsunami are poorly known. Existing catalogs give contradictory information on the location: 37.00° N 10.00° W (Mezcua and Solares, 1983) or 36.00° N 10.50 W (Oliveira, 1986); the epicenter intensity is evaluated as IX (Mezcua and Solares, 1983), 8.5 (Oliveira, 1986), or VIII (Munuera, 1963); the time of occurrence is 12:05 a.m. Lisbon time (Mezcua and Solares, 1983) or 12:10 a.m. (Oliveira, 1986; Moreira *et al.*, 1993).

Most of the original information available for the 1761 earthquake and tsunami comes from newspapers published in Lisbon (*Gazeta de Lisboa*) and London (the *London Chronicle*). A series of letters published in the *Philosophical Transactions of the Royal Society* reproduce and complete the former descriptions. In the Annual Register for 1761 (Annual Register, 1761), most of this information is reorganized and the earthquake constitutes one of the most remarkable events of the whole year.

The first study of the 1761 earthquake and tsunami was published by Borlase (1762) in the *Philosophical Transactions*. It summarizes most of the information published in the transactions and addresses the origin of the earthquake and its propagation. Eighty-five years later Perrey (1847) and ninety years later Mallet (1852) recompiled all the available data. These compilations were used in most of the modern studies (e.g., Moreira, 1984; De La Torre, 1997).

In this work we have two main objectives: first, to synthesize the available historical information concerning both the earthquake and the tsunami, separating original records from later interpretations and, second, to focus on a few parameters, the most useful in our view to model the tsunami propagation. Using this data set we compute the most probable location of the 1761 source area. Finally, we discuss the macroseismic data to assess the likelihood of the tsunami source location.

Data

Earthquake Observations

In the last day of March 1761, close to noon, an earthquake was felt in Lisbon, lasting for more than 3 min and alarming the inhabitants of the city. Several walls collapsed in the downtown but 3 hr later everything was in perfect tranquility. Close to Lisbon, greaver damage is reported in Setubal, 30 km south of Lisbon, and Vila Franca, 25 km north of Lisbon. Oporto, 300 km north of Lisbon, reported damage greaver than observed in the 1755 event, although contradictory descriptions exist in the 1761 case.

In Spain, a systematic inquiry was made at that time by the president of the Real y Supremo Consejo de Castilla (Royal and Supreme Council of Castilla) also bishop of Cartagena, Diego de Rojas y Contreras, who sent a letter to all local authorities to ask for information about the 31 March 1761 earthquake, including reports of the earthquake occurrence, number of victims, and damage (De La Torre, 1997). The answers to the bishop's queries, which correspond to the largest dataset for this event, were reviewed recently by De la Torre (1997).

Later reports, mainly published in newspapers and reproduced and discussed in the *Philosophical Transactions* describe earthquake effects in Funchal (Madeira island), Terceira (Azores archipelago), South Barbary (Agadir), southern United Kingdom and Ireland, Spain, and the Netherlands. (Ⓔ Details are available in the electronic edition of BSSA.)

There are numerous descriptions reporting the earthquake effects that can be used to assess MSK macroseismic intensities. Most values were already compiled by De La Torre (1997); the values were re-evaluated to ensure that the original information was clear enough. The final data set is presented in Table 1, where data concerning geographic location for all macroseismic data are also included.

Tsunami Observations

About 1¼ hr after the earthquake a tsunami was observed in Lisbon. Its amplitude is estimated as eight feet, affecting several ships that were left dry at some intervals (Molloy, 1761). Seawater changes were also observed along the Spanish coasts, namely Ayamonte, Puerto de Santa Maria, Cadiz, and Barcelona, but we did not find quantitative details on arrival time or run-up. The tsunami was also observed in Funchal and in Terceira.

Good descriptions of the tsunami came from southern Ireland and the United Kingdom, in particular, Kinsale, where it was observed at about six o'clock in the evening, Mount's-bay (Cornwall), Penzance, and the Scilly Islands. There is also a reference for Barbados, where the tide ebbed and flowed, in about 8 min, between 18 inches and 2 feet, which was attributed to the 1761 tsunami. This information is included, for example, in tsunami catalogs for the Caribbean (Zahibo and Pelinovski, 2001). (Ⓔ Details are available in the electronic edition of BSSA.) In a few exceptional cases both the arrival time of the tsunami wave, the wave height, and the duration of the phenomenon are registered. Table 2 summarizes tsunami information. (Ⓔ Direct quotes of observations can be found in the electronic edition of BSSA.)

T Phases

A few ships navigating offshore Portugal reported violent shocks related to the 1761 earthquake. The descriptions include "felt off the spindle of a magnetic needle," submarine noise (Van Hoff, 1841, in De La Torre, 1997), shaking similar to "striking on a sunken rock" (Annual Register, 1761), up to the description of a shock so large that "the crew threw out the boat in order to leave her" (Moreira, 1993). (Ⓔ Complete accounts of the *T* waves are available in the electronic edition of BSSA.)

T phases have been known for a long time, as elastic compressive waves generated in the ocean (also known as *T* phases) due to the earthquake-induced vertical motion of the seafloor and the slight compressibility of the water mass. They propagate over great distances in the ocean sound channels but their amplitudes decrease rapidly, so that their manifestation must be noticeable close to the tsunami source up to a distance not exceeding several sizes of the source (Nosov, 2000).

We summarize the presumable seaquake observations in Table 3. To characterize each of the observations we used the Rudolph Scale (NCDC, 2003). Observations run between 4 ("Slight Shock felt as if a heavy anchor was dropped rapidly.") to 6 ("Rather Strong, Cups, glasses, etc. are vibrated"). The Rudolph scale has a maximum value of 10.

Data Processing

Magnitude

Instrumental earthquakes that generated significant tsunamis observed in Lisbon were the events of 28 February 1969 ($-10.57°$ E, 35.01° N) and 26 May 1975 ($-17.60°$ E, 35.09° N). Both had M_s magnitudes estimated at 7.9 (Fukao, 1973, Lynnes and Ruff, 1985). The greater tsunami observed in Lisbon was generated by the 1 November 1755 earthquake of which the magnitude is considered to be larger than 8.75 (Abe, 1979).

Tsunami run-up heights can be used to infer the earthquake magnitude. Abe (1981) determined the empirical relationship:

Table 1
Macroseismic Intensities

Location	Earthquake Time	Latitude (° N)	Longitude	Duration (min)	MSK Int.	Source
A Coruña	13h	43.33	8.42° W	6	IV	RT
Alcantara	12h 40m	39.72	6.88° W		IV	RT
Algeciras		36.13	5.45° W		II	RT
Alicante		38.34	0.48° W		II	RT
Almeria		36.83	2.43° W		I	RT
Ayamonte		37.02	7.04° W	7–8	IV	RT
Baeza	12h 30m	37.99	3.47° W		IV	RT
Barcelona		41.40	2.17° E		I	RT
Beja		38.02	7.86° W		VI	
Cadiz		36.53	6.29° W	2–3	II	RT
Carmona		37.48	5.63° W	4	II	RT
Cartagena		37.61	0.98° W		I	RT
Castro Marim		37.22	7.44° W		IV	RT
Ceuta		35.91	5.30° W		II	RT
Coria		39.99	6.54° W		III–IV	RT
Cork Harbor	12h 15m (Lisbon time)	51.85	8.25° W		II	M
Cuenca	11h 30m	40.08	2.14° W	2	III	RT
Écija	12h 15m	37.54	5.09° W	7–8	II	RT
Évora		38.57	7.91° W		V	
Funchal. Port	11h 35m	32.60	16.91° W		VII	
Granada		37.17	3.59° W		IV–V	RT
Guadix		37.26	3.14° W	4	II	RT
Horta		32.78	16.85° W		VI	
Jaca		42.58	0.55° W	1	I	RT
Léon		42.59	5.57° W	4	IV	RT
Lisboa		38.72	9.14° W		VII	S, MO
Logroño	12h 15m	42.47	2.44° W		II	RT
Malaga	12h 30m	36.72	4.42° W		II–III	RT
Orihuela	13h 00m	38.09	0.95° W		II–III	RT
Oviedo	12h 30m	43.35	5.83° W	5	II	RT
Plasencia	12h 30m	43.17	2.42° W	6–7	II	RT
Oporto		41.15	8.62° W		VI	
Pto Sta Maria	12h 15m	36.61	6.23° W	6–8	III	RT
Ronda		36.74	5.16° W		II	RT
San Sebastien	12h 20m	38.16	1.48° W	5–6	II–III	RT
Santa Cruz in Barbary		30.40	9.60° W	0.25	III	M
Santander		43.47	3.80° W	3	III	RT
Segovia	12h 20m	40.94	4.11° W		II–III	RT
Sevilla	12h 30m	37.40	5.98° W		IV	RT
Tarragona		41.12	1.24° E		II	RT
Toledo		39.86	4.03° W		II	RT
Valencia		39.48	0.39° W		II–III	RT
Vélez		36.79	4.10° W	2	II	RT

Sources are flagged as: M (Mallet, 1852), S (unknown, 1761), MO (Molloy, 1761), RT (De La Torre, 1997). All other correspond to this work.

$$M_t = \log H_2 + \log R + 5.55,$$

where M_t represents the "tsunami magnitude," which we can consider close to M_w, H_2 is the maximum crest-to-trough amplitude on tide gauge record in meters, and R is the distance from epicenter to station along the shortest oceanic path in kilometers. If we also consider that run-up heights described in the historical records correspond roughly to the crest-to-trough amplitude on tide gage records (Kajiura, 1983, in Abe, 1995), and we consider a candidate epicenter close to 34.5° N 13° W, we get an average M_t 8.8. For two locations, however, Lisbon and Penzance, we have run-up estimations for 1755 (Baptista *et al.*, 2003) and 1761 events. If we consider that the magnitude of 1 November 1755 is well evaluated as 8.75 (Abe, 1979), then a small corrective term must be added to the preceding expression, and we get a final estimation M_t 8.5 for the 1761 earthquake. This corresponds to an epicenter intensity of XI.

Isoseismal Analysis

In Figure 1 we plot the available MSK intensities already listed in Table 1. Most of the isoseismal contours point to a

Table 2
Tsunami Travel Times and Wave Heights

Location	Latitude (° N)	Longitude (° E)	Travel Time	Wave Height (m)	Source
Barbados	13.04	−59.57	8:30	1.2	Zahibo and Pelinovski. (2001)
Antígua	17.05	−61.80	N/A	Observed	Zahibo and Pelinovski. (2001)
Barcelona	41.40	+02.17	N/A	Observed	De La Torre (1997)
Penzance	50.10	−05.50	N/A	1.2	Borlase (1762)
Lisbon	38.72	−09.13	1:15	2.4	unknown (1761); Molloy (1761)
Mount Bay	50.08	−05.48	5:00	1.2	Borlase (1762)
Scilly Islands	49.92	−06.33	5:00	0.6	Van Hoff (in De La Torre, 1997); Borlase (1762)
Kinsale Harbor	51.67	−08.51	6:00	0.6	Borlase (1762)
Terceira	38.65	−27.22	N/A	Large	Fearns (1761)

Table 3
T-Wave Observations

Latitude (° N)	Longitude (° W)	Time	Rudolph Scale	Description and Source
43.00	10.00	10 min. am (pm?)	5	*London Chronicle* 2 May 1761
44.48	11.32	N/A	4	Robert Muirwood (personal comm.) in Moreira *et al.* (1993)
44.80	14.24	11:45	6	1761 Annual Register (page 93)

Figure 1. Isoseismal curves for 1761 earthquake. Locations plotted correspond to macroseismic intensities from Table 2.

source west-southwest of Lisbon. We can extract more information from intensity data if we fix a specific attenuation law and consider that a probable location will correspond to the minimum of the averaged intensity errors (observed minus predicted) employing the least-squares approach where each position in space is treated as a possible point source for the earthquake and the averaged squared error is calculated for all locations where macroseismic intensities are available.

Given the average quality of macroseismic intensity data and the poor azimuthal coverage we decided to fix *a priori* the Medvedev–Sponheuer–Karnik (MSK) attenuation law, deduced for the best constrained earthquakes in the southwest Iberia margin. To do so, we use the law fitted by Casado *et al.* (2000) to a large set of isoseismal maps within the Iberian area, considering that 1761 source region corresponds to the "low attenuation" behavior verified in most of the large earthquakes in the southwest Iberian margin (Casado *et al.*, 2000).

$$I^P = I_0 - 1.762 * \log(r) - 0.00207 * r$$
$$I_0 = 5.557 + 0.902 * I_0 + 0.014 * I_0^2,$$

where I^P is the predicted intensity value at each point, I_0 is the epicentral intensity, r is the distance from each site to the edge of the fault plane. The average prediction error was defined as:

$$\varepsilon = \frac{1}{n} \sqrt{\sum_{k=1}^{n} (I_k - I_k^P)^2},$$

where n is the number of locations where intensity I is known. For the 1761 earthquake we used an epicentral intensity of XI and the same set of parameters as earlier. Results are shown in Figure 2.

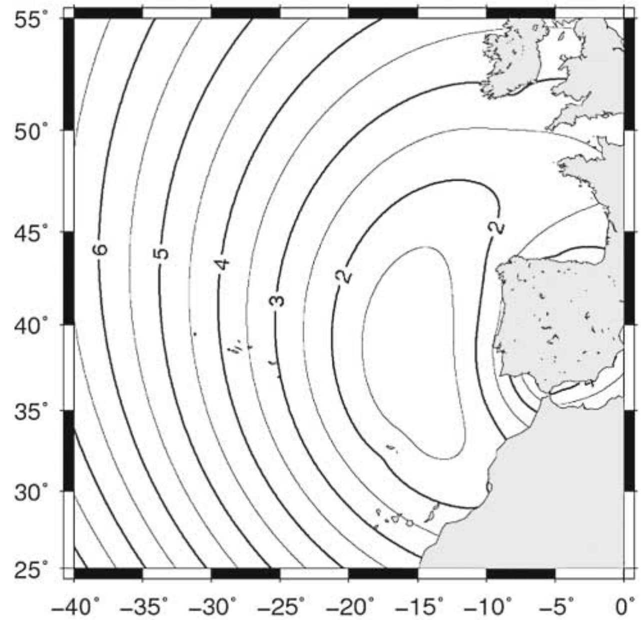

Figure 2. Location of the best-fit location for the 1761 source using Baptista *et al.* (1998) attenuation law.

Tsunami Backward Ray-Tracing

From the tsunami descriptions we conclude that there are only a few quantitative observations of the tsunami parameters (see Table 2), including tsunami travel times for five different places (Lisbon, Scilly Islands, Barbados, Mount Bay, and Kinsale Harbor) and four observations of wave heights (Barbados, Penzance, Lisbon, and Kinsale Harbor). With this data set we can compute an approximate tsunami source location, using backward raytracing techniques.

Raytracing techniques simulate the horizontal displacement of a sea-surface perturbation using a linear approximation This is valid only when depth relative to the wavelength is small and when the pressure distribution is assumed to be hydrostatic; the first condition is valid for propagation in the deep ocean, leading to a simple computation of the wave speed as an exclusive function of depth: $c = \sqrt{gh}$. Although in most cases raytracing is used to produce travel-time charts for tsunami warning purposes (Choi *et al.*, 2003), it can be reversed to compute, instead, the location of the tsunami source (Miyabe, 1934; Hatori, 1969; Gjevik *et al.*, 1997; Baptista *et al.*, 1998; Ortiz and Bilham, 2003).

For the 1761 tsunami we have only five different arrival times and so five backward raytracing simulations were computed using a point source located in each of the locations (virtual tide gauge coordinates are listed in Table 2). In Figures 3 and 4 we present the results of the backward raytracing simulations for all locations.

If we compare (even visually) the five determinations, we can easily conclude that the Barbados observation is incompatible with the other four. In particular, it is incompatible with the travel time observed at Lisbon, where both earthquake and tsunami were observed and so the time difference is better established. The incompatibility between Barbados and the other four observations also reinforces the observation made by Shepherd (2001) that found no local sources to support the observation of the 1761 tsunami in Barbados. Another explanation could be that the observed agitation of the water did not correspond to the leading wave of the tsunami.

We can determine a preliminary source area location as the minimum of the averaged time square errors employing the least-squares approach where each position in space is treated as a possible point source for the tsunami and the averaged travel-time-squared errors are calculated for all station data from the corresponding travel time (Baptista *et al.*, 1998). This is presented in Figure 5, where the stripped area corresponds to an average error less than 0.5 hr. This is a reasonable value, because the accuracy of the earthquake

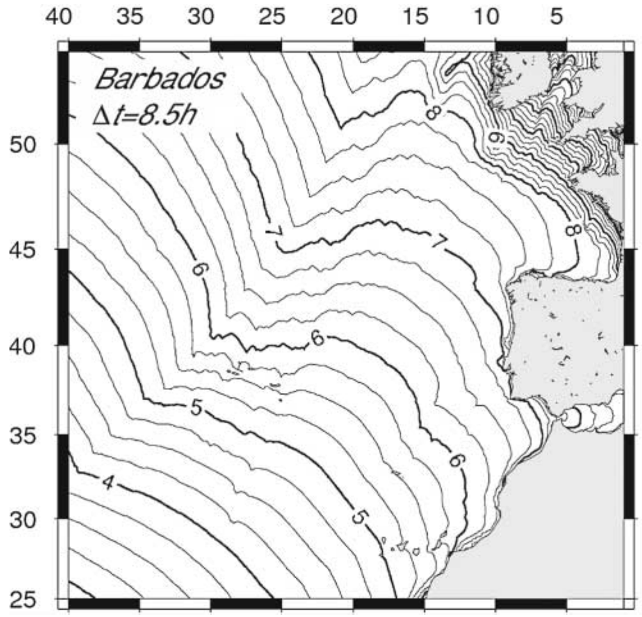

Figure 3. Backward raytracing for Barbados travel time.

time is evaluated as close to 16 min (see below) and the tsunami arrival time is a much more complex phenomenon, where the observation can refer to its first observation, the inundation (or retreat) of the waters, or the maximum run-up.

There are also some limitations in raytracing techniques because they are based on a linear approximation. This implies that the speed of the wave propagation, in shallow waters, is slightly higher than reality; this effect may be important close to harbors and coastal areas. Also, there is a geometric constraint given by the poor azimuth coverage of observations: they are all located north of the presumed source areas, and so errors do not cancel in the root-mean-square (rms) determination.

Discussion and Conclusions

1. Time of the earthquake. If we consider only the observation of the earthquake along Iberia, we get significant differences for its time of occurrence. In Figure 6 we plot all available observations compiled by De La Torre (1997) for Spain, where a simple correction was applied to local times to make them comparable to Lisbon time. We get a final value of 12:01 a.m. (Lisbon time), with 16 min rms error. This is reasonable given the means available for time determinations. Also, descriptions were made several days or weeks after the events.

2. Location of the earthquake source. Available data do not allow a clear identification of the source area. De la Torre (1997) considers three possible locations for the earthquake source (36° N 10.5° W, 37° N 10.0° W, and 41° N 14.0° W). The choice made by each catalog is a function of the relative weight given to the descriptions for Canarias, Madeira, Azores, and Barbados (which would favor a "southern solution") or to the effects described at Oporto, Ireland and Southern England (which favor a "northern" solution). We consider three candidate locations, A (43° N 12° W), B (34.5° N 13° W), and C (36° N 10.5° W), that correspond, respectively, to the "*T*-phase" solution, the "tsunami solution," and the "Gorringe solution." Coordinates for solution A were derived from the location of the vessels where "seaquakes" were described; solution B from the intersection of the better location from the intensity and tsunami point of view; solution C corresponds to Gorringe Bank, considered in most catalogs as the most probable source of several large magnitude earthquakes felt in southwest Portugal (Oliveira, 1986).

3. Magnitude. Historical descriptions of wave heights were used to estimate the earthquake tsunami magnitude using the Abe (1981) approach. However, to do so, we had to fix a location for the earthquake epicenter and some relationship between the observed run-up and the tsunami crest-to-trough amplitude that would have been observed in tide gauges if they existed in 1761. The value obtained is similar to the other instrumental events that generated noticeable tsunamis in the Portuguese coasts and it is not sensitive to small changes in the location of the epicenter.

4. *T* Phases. The descriptions of the *T*-phase effects (Ⓔ available in the electronic edition of BSSA) are very vivid but are reported on a somewhat wide area (compare Fig. 5). Positions range from "some ships, at sea, at a certain distance from Lisbon," to "the latitude of 43°, not many leagues off shore," or "Off Lisbon; 44°29′ N; 11°19′ W." Ambraseys (1985) refers to the occurrence of thousands of "seaquake reports" in the insurance companies and, in particular, the well documented case of the 28 February 1969 (M_s 7.9) earthquake in the Horseshoe Abyssal Plain, where a 32,000-ton vessel suffered serious structural damage and was obliged to return to Lisbon, where it was dry-docked and surveyed (Ambraseys, 1985). Large effects were felt up to 190 km away from the epicenter. If we compare the historical descriptions here with the ones given by Ambraseys (1985) we can conclude that vessels that reported the effects summarized in Table 3 were not over the source area. *T*-phase effects seem to favor solution A but the source area can also be located some hundreds of kilometers to the south.

5. Tsunami data. The tsunami travel times, with the exception of Barbados, allow the computation of a mean solution within a relatively restricted area. The exclusion of the Barbados travel time is based on its total inconsistency with respect to the other four travel times. Though the Lisbon travel time, which can be considered the most reliable of all, is compatible with solutions A and B and less compatible with solution C (see Fig. 4); the travel times observed in southern England and Ireland are compatible only with solution B.

6. Intensity reports. Reports on the level of destruction in Oporto are highly variable: according to some, lots of

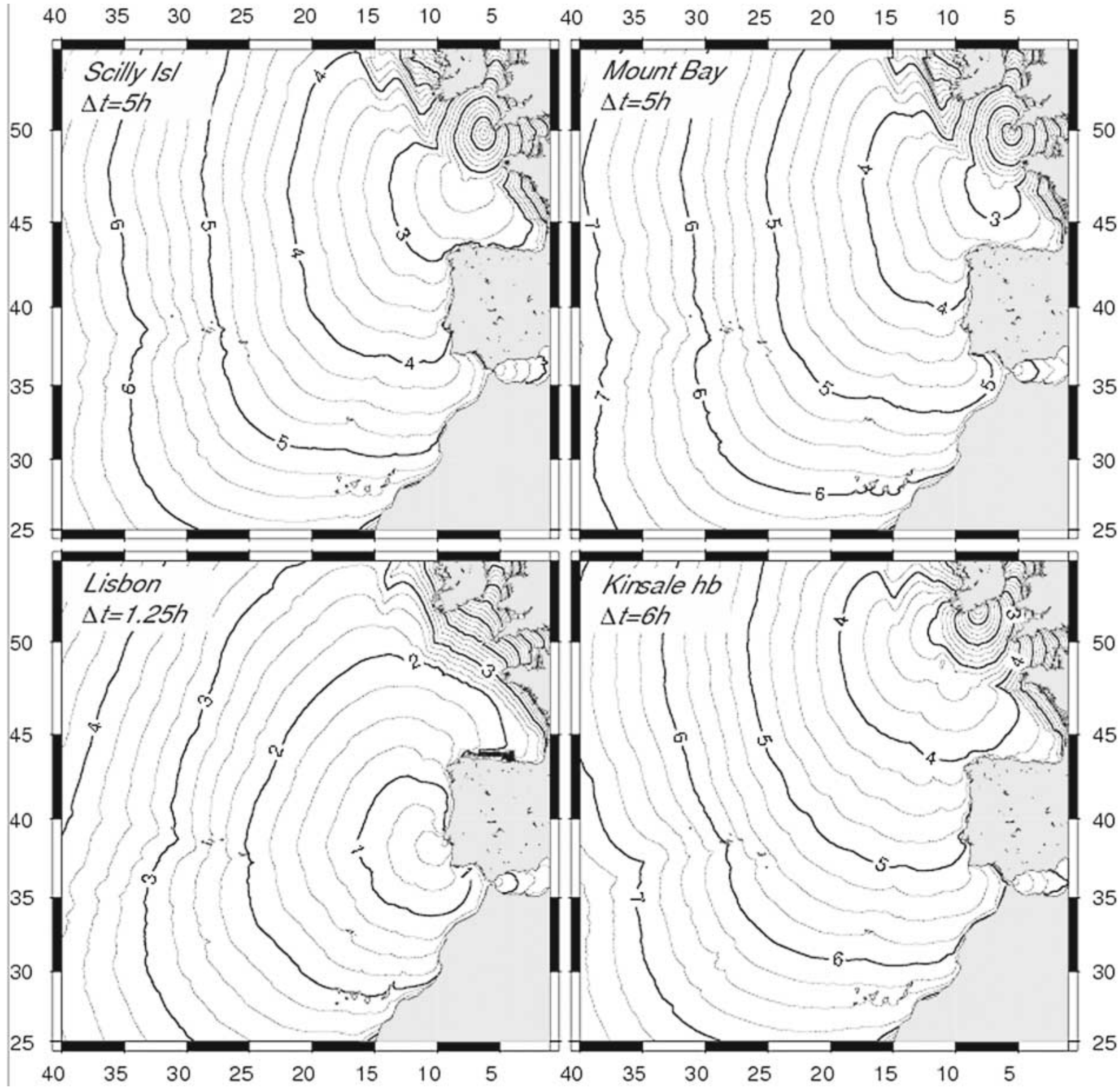

Figure 4. Backward raytracing for four locations where travel times are known (Lisbon, Scilly Islands, Mount Bay, and Kinsale Harbor).

buildings were thrown down, but according to others little damage was observed (Ⓔ see details in the electronic edition of BSSA). In the Lisbon area, the situation is similar: although the level of destruction in Lisbon does not seem to have been very severe, the cities of Vila Franca and Setúbal at a distance of 20 and 90 km, respectively, are reported to have suffered heavily. On the other hand, in La Coruña (north of Oporto) no houses fell down, but in Madeira the shock was felt strongly enough to make rocks fall off the cliffs and to destroy a church and kill four people. Overall, the intensity data (Figs. 1 and 2) favor location B with respect to solutions A or C.

7. How to reconcile historical data? The observation of important *T*-phase effects on ships was considered by previous works (Moreira, 1993; De La Torre, 1997) to imply that the source area should be located close to the ships that reported its occurrence. This implies rejecting nearly all other reports ranging from Ireland to Madeira Island. A magnitude 8 earthquake, as discussed earlier, if located in the region where *T*-phase effects were reported, would have been highly destructive to the northwest region of the Iberian Peninsula. Historical descriptions, and, in particular, the systematic Spanish inquiry (De La Torre, 1997), rule out the implied degree of destruction.

It is known that *T* waves propagate very far in the oceans through the SOFAR channel that acts as a wave guide where *T* waves lose very little energy. This could justify the observation of effects 800 km north of the source area. It is

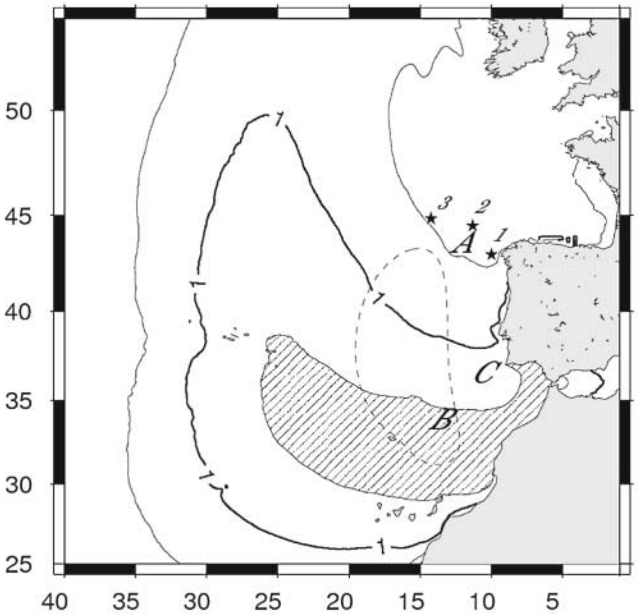

Figure 5. Contour lines represent the average misfit of backward raytracing (hours). Stripped area corresponds to an average error smaller than 0.5 hr. Seaquake observations (stars) are numbered according to the order of Table 3. Dashed line encloses the area where the MSK average misfit is less than 1.5.

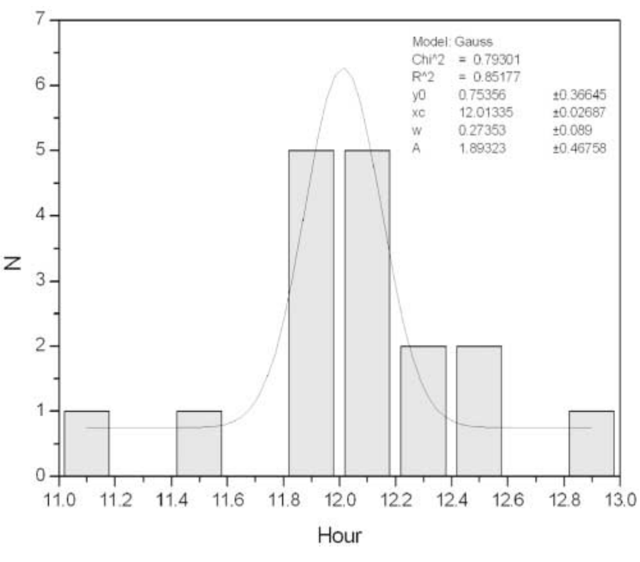

Figure 6. Histogram of time of occurrence for the 31 March 1761 earthquake (Lisbon time). Observations correspond only to Iberia.

also known that when T waves interact with smooth and steep physical barriers (islands for example) they are effectively converted back into P and S waves. The summit of the Galicia Bank has a depth of about 700 m which is enough to completely cut the SOFAR channel (that spans between 700 and 1000 m). It has also very steep flanks, so it assembles the conditions to an efficient conversion from T to P waves that would radiate from the Bank. This same mechanism could also work efficiently on the continental slopes that meet the necessary geometric conditions.

The historical information that we were able to gather for this study has strong limitations. We tried to reconcile tsunami and intensity information from several independent sources to extract the most probable location for a significant earthquake that is included in a few seismic catalogs and so used for seismic-hazard assessment. Within the limitations of historical data we conclude that the 31 March 1761 earthquake took place at 12:01 a.m. (Lisbon time). Its epicenter was located about 34.5° N 13° W and had a magnitude of about 8.5.

Acknowledgments

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