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Tsunami Hazard Evaluation of the Eastern Mediterranean: Historical Analysis and Selected Modeling

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Abstract Seismic sea waves in the eastern Mediterranean have been reported since written history first emerged several thousand years ago. We collected and investigated these ancient and modern reports to understand and model the typical tsunamigenic sources, with the ultimate purpose of characterizing tsunami hazard along the Levant coasts. Surprisingly, only 35% of the tsunami reports could be traced back to primary sources, with the balance remaining questionable. The tsunamis varied in size, from barely noticeable to greatly damaging, and their effects ranged from local to regional. Overall, we list 21 reliably reported tsunamis that occurred since the mid second century B.C. along the Levant coast, along with 57 significant historical earthquakes that originated from the "local" continental Dead Sea Transform (DST) system. An in-depth evaluation shows that 10 tsunamis are clearly associated with on-land DST earthquakes, and therefore, as formerly suggested, they probably originated from offshore, seismogenically induced slumps. Eight tsunamis arrived from the "remote" Hellenic and Cypriot Arcs, one from Italy, and two are left with as yet unrecognized sources. A major conclusion from this work is that onshore earthquakes commonly produce tsunamis along the Levant coastline, and that analogous situations are present elsewhere in the Mediterranean, as well as along the California coast and in another regions with active faults near the coast.

We modeled three typical scenarios, and in light of the Sumatra experience, we examined the more likely severe magnitudes. This of course leads us toward the upper range of expected run-ups. The models show that sooner than five minutes after a strong earthquake produces an offshore slump, which occurs after close to a third of the large DST earthquakes, a 4- to 6-m run-up may flood part of the Syrian, Lebanese, and Israeli coasts. Tsunamis from remote earthquakes, however, arrive later and produce only 1- to 3-m run-ups, but are more regional in extent.

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Introduction

Tsunamis in the eastern Mediterranean have attracted much attention for their occurrence in an area with a long and significant history, and because many of them appear to have occurred after onshore earthquakes along the Dead Sea Transform (DST, also referred to as the Levant fault), making it likely that they result from seismogenically induced submarine landslides (Shalem, 1956; Almagor and Garfunkel, 1979; Ambraseys and Melville, 1988; Ambraseys and Barazangi, 1989; Arieh, 1989). Considering the proposed mechanism and the evidence for more than 800 years of seismic quiescence along several segments of the DST (Meghraoui *et al.*, 2003; Daëron *et al.*, 2005; Zilberman *et al.*, 2005), as well as the lack of a reliable list of events and a comprehensive understanding of the tsunamigenic environment of the eastern Mediterranean, the need for tsunami hazard evaluation is obvious.

Tsunamis in the Middle East were first described in written history a few thousand years ago with early cuneiform texts reporting the flooding of Ugarit, a city along the Syrian coast (near Latakia), by a sea wave (Dussaud, 1896; Virolleaud, 1935; Ambraseys, 1962; Ambraseys *et al.*, 2002). Ancient religious scripts described how the sea fled while the Holy Land trembled (*Amos* 9: 5–9), and later chronicles reported observations of destructive sea waves associated with violent earthquakes (e.g., Guidoboni *et al.*, 1994; Guidoboni and Comastri, 2005, and references therein).

The original reports are rare, hard to find, and in many cases could not be unequivocally interpreted. Many difficulties appear in interpreting the primary chronicles, and catalog editors inevitably introduce their personal understanding. Consequently, "storm sea waves" may be regarded as seismic sea waves (Ambraseys, 1962), a single earthquake may be separated into several events, and several events may have been merged into one (Karcz, 2004).

Although these and more are genuine complexities, some catalogs unfortunately include entries with no referencing and unverified accounts, many of which have been shown to be borrowed from elsewhere, erroneously listed because of poor translation, or made from inaccurate calendar determinations. This may result in amalgamation, duplication, or omission of true events, exaggeration or diminution of the real size of the events, assignment of a wide range of source parameters (e.g., origin time, epicenter, magnitude) to the same event, and other problems. With time, subsequent generations of seismological compilations appear to compound the problem by recompiling dubious events, unavoidably increasing the bias in the long-lasting effort to construct a complete list of true events.

These real difficulties are extensively and thoroughly addressed by several researchers (e.g., Guidoboni *et al.* 1994; Karcz, 2004; Ambraseys, 2005a). with an emphasis on their implications for the evaluation of earthquake hazards. For example, Ambraseys *et al.* (2002) refer to several case studies where about half the entries were found to be false and emphasize the need for caution in editing such historical data. For this study, we present a detailed compilation of historical reports of tsunamis and earthquakes for the eastern Mediterranean, along with an analysis as to the validity of these reports through cross-referencing with primary sources. We then characterize the typical tsunamigenic sources for this area and model three of them to assess their potential impact in this turbulent region. Finally, we draw parallels with similar settings of near-shore active structures that may have generated tsunamis in the past in other areas of the world.

Assessment of the Historical Data

We focused our attention on the easternmost coast of the Mediterranean, because this is the area affected most by the DST fault system. It includes, from south to north, the coasts of Egypt along the Nile Delta and Sinai, Israel, Lebanon, Syria, and as far north as the Bay of Iskendrun (Alexandretta) in southern Turkey (Fig. 1). Because tsunamis may arrive from afar, we also assessed the potentially remote tsunamigenic sources relevant to the eastern Mediterranean, namely the Cypriot and the Hellenic arcs, the Aegean Sea, and as far west as Italy.

First, we searched the available literature and collected all of the records for tsunamis along these coasts, as well as all of the accounts of earthquakes that are attributed to rupture along the DST system. We then examined the authenticity of each of the events and found that several detailed studies in the existing literature allow us to validate or question many of these events. The most credible studies were those that analyzed the primary sources and extracted an accurate description of the event with minimal necessary interpretation (translation, historical context, etc.). We focused on constructing a reliable list specifically suitable for our study; but, of course, as new data are discovered and further in-depth analysis of already known chronicles appears, our list will need to be updated.

We screened the data and distinguished between reliable and doubtful reports by cross-referencing each to their original sources and by tracing back to where modern cataloguers compiled their data. We were thus able to resolve many of the ambiguities encountered, merge different entries originating from the same event, exclude duplicated and questionable events, and minimize uncertainties. This resulted in a condensed and more reliable (in our opinion) list of tsunamis and earthquakes than if we had listed every possible event.

Once the list was compiled, it was possible to systematically correlate tsunamis with earthquakes. It then became clear that only about 50% of the tsunamis on the reliable list were parented by local DST earthquakes. For those that had no local source, we expanded our search for causative earthquakes to the entire eastern Mediterranean region, as far west as Italy, and found large earthquakes elsewhere that provided the likely trigger for all but three tsunamis.

All the reliable tsunamis and earthquakes are summarized in Table 1 and presented in Figure 1. The criteria and reasons for how we constructed this table are explained in the following sections.

Sources of Data

Shalem (1956) and Ambraseys (1962) were the first to compile a specific list of tsunamis for the eastern Mediterranean. This was followed by regional compilations by Antonopoulos (e.g., 1979, 1980a–f, 1990), Soloviev *et al.* (2000), and Papadopoulos (2001), and areal investigations for the coasts of Greece (e.g., Galanopoulos, 1960; Papadopoulos and Chalkis, 1984; Papazachos *et al.*, 1986), Turkey (e.g., Altinok and Ersoy, 2000), the Marmara Sea (Ambraseys, 2002), Italy (e.g., Tinti *et al.*, 2004), and Cyprus (Fokaefs and Papadopoulos, 2006). Many examinations targeted specific events, the most notable being the tsunami triggered by the Late Minoan Thera (Santorini) eruption and collapse (e.g., Yokoyama, 1978; McCoy and Heiken, 2000; Minoura *et al.*, 2000, and references therein) and the 9 July 1956, southern Aegean tsunami (e.g., Ambraseys, 1960; Papazachos *et al.*, 1985; Goldsmith and Gilboa, 1986; Van Dorn, 1987; Perissoratis and Papadopoulos, 1999). Far fewer direct field investigations have been conducted for tsunamis

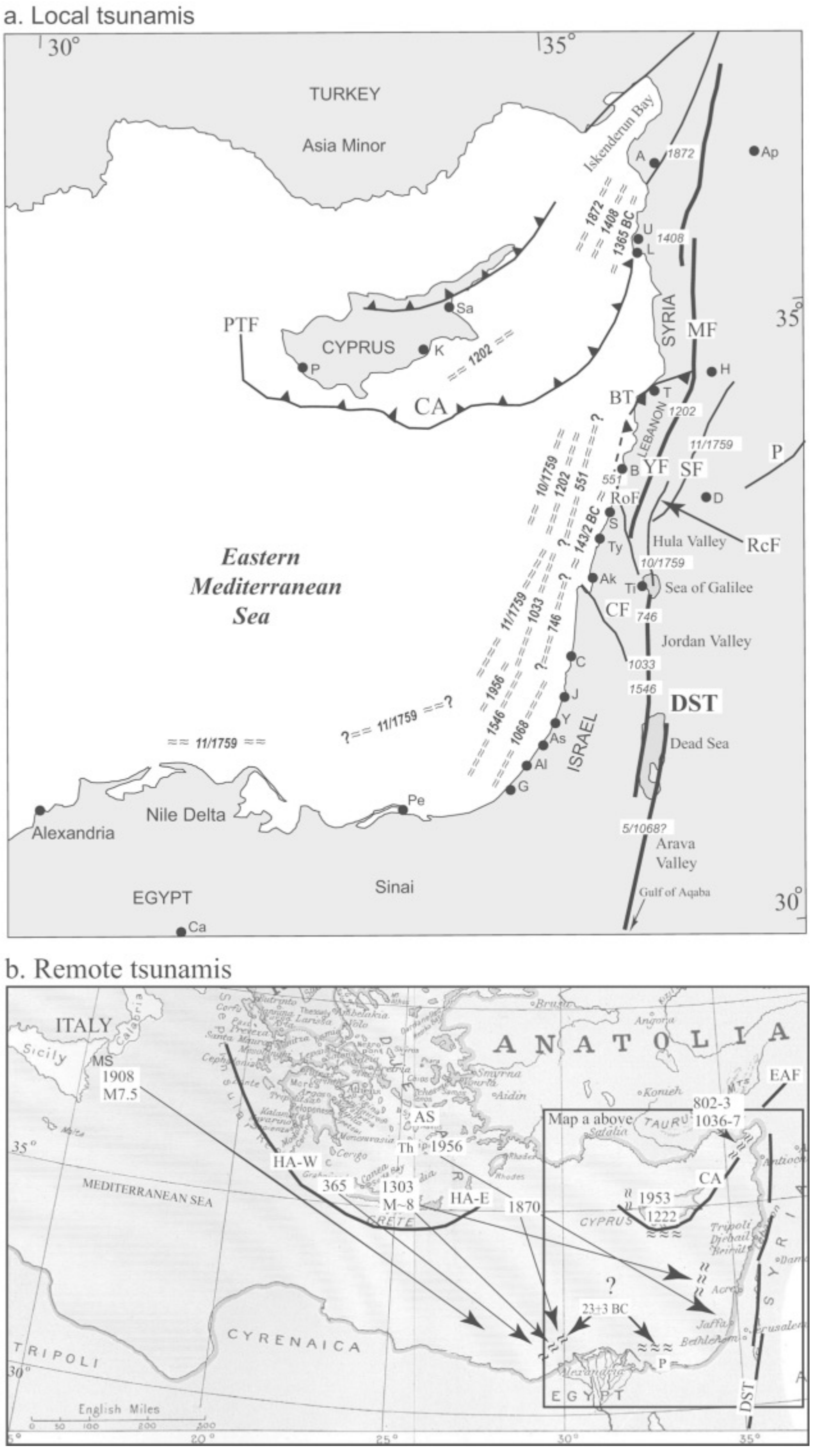

Figure 1. Location of study area and occurrence of local and remote tsunamis that arrived to the eastern Mediterranean (Levant) coast, from Alexandria, Egypt, to Iskenderun Bay, Turkey, since the mid second century B.C. Note the probable location of the tsunamigenic earthquakes. Tectonic elements from Daëron *et al.* (2005). (a) Local tsunamis originated from earthquakes along the DST system. (b) Remote tsunamis originated from sources off the DST system, including the Cypriot Arc which is not far from the DST. Map shows Ottoman area during the nineteenth century (modified after Miller, 1913). Plate borders (bold lines) and regions: AS, Aegean Sea; CA, Cypriot Arc; DST, Dead Sea Transform; HA-E, Hellenic Arc east; HA-W, Hellenic Arc west; EAF, East Anatolian fault, Faults: BT, Beirut thrust; CF, Carmel fault; MF, Missyaf fault; P, Palmerides; PTF, Paphos Transform fault; RcF, Rachaiya fault; RoF, Roum fault; SF, Serghaya fault; YF, Yammaouneh fault. Localities: A, Antioch (Antakya); Ak, Akko; Al, Ashkelon; Ap, Allepo; As, Ashdod; B, Beirut; C, Caesarea; Ca, Cairo; D, Damascus; G, Gaza; H, Homs (Hims); J, Jaffa; K, Kition (near Larnaca); L, Latakia; MS, Messina Straits; P, Paphos; Pe, Pellusium; S, Sidon; Sa, Salamis (near Famagusta); T, Tripoili; Th, Thera (Santorini) Island; Ti, Tiberias; Ty, Tyre; U, Ugarit; Y, Yavneh.

in the Mediterranean, but their potential contribution for risk assessment cannot be underestimated (Dominey-Howes, 2002; Whelan and Kelletat, 2002). Databases of worldwide tsunamis, including those that have occurred in the Mediterranean, which are based on existing catalogs, are available via the internet (at www.ngdc.noaa.gov/ by the National Geophysical Data Center (NGDC) and omzg.sscc.ru/tsulab/intc522.html by Gusiakov, 1997).

Complimentary information is also found in studies of historical seismicity that describe natural seismogenic effects. The most important and dependable accounts are those that report a tsunami directly from the primary historical sources. Included here, for example, are the catalogs of Guidoboni *et al.* (1994) and Guidoboni and Comastri (2005); historical reviews of Poirier and Taher (1980), Ambraseys (1989, 2004), Ambraseys *et al.* (1994), and Ambraseys and

Table 1
Integrated List of Tsunamis and Significant Earthquakes of the Eastern Mediterranean

Earthquakes				Tsunamis		Affected Coasts[#,**]													
							Levant												
							Israel							Lebanon					
Date*	Size[†]	Remote[‡]	Local DST[§]	Date*	Size[¶]	Egypt	Gaza	Ashkelon	Ashdod	Yavneh	Jaffa	Caesarea	Akko	Tyre	Said/Beir/Tri	Syria	Cyprus	Asia Minor	Greece/ Crete
?				− 1365 ± 5	S											Ugarit ∼∼∼			
760–750 B.C.	?		C	speculated															
199–198 B.C.	M		N	–															
Mid second century B.C.	?		N	?															
?				Mid second century B.C. (143/2 B.C.)	S								∼∼∼ ↑	∼∼∼ ↑					
148 02 21 B.C. or 130 B.C.	M		N	–															
69–65 B.C.	M		N	–															
31 B.C.	M		C	–															
?				23 ± 3 B.C.	S	Alx, Pel ∼∼∼ ↑													
17 B.C.	M	CA		?															
37 03 23 A.D.	M		N	–															
115 12 13	L		N	?															
127–130	M		C	–															
303/304	M		N	–															
363 05 18–19	L		C	–															
365 07 21	V	HA-W		365 07 21	L	Alx ∼∼∼													G, C ∼∼∼
419	M		C	–															
502 08 19	M		N?	–															
526 05 20/29	M		N	–															
528 11 29	M		N	–															
551 07 09	L		N	551 07 09	M							?	?	?	∼∼∼ ↓↑	?			
ca. 570	M		N	–															
601–602	M		N	–															
659/660	M		C	–															
713 02 28/03 10	M		N	–															
746 01 18	M		C	746 01 18	?						?	?	?	?					
802 12 30 – 803 12 19	?	CA-EAF		802 12 30 – 803 12 19	S													Isk ∼∼∼	
847 11 24	M		N	–															
859 12 30 – 860 01 29	L		N	–															
991 04 05	M		N	–															
1002 11 10 – 1003 10 29	M		N	–															

1033 12 05	M		C	1033 12 05	S-M					?	~~~	~~	~~~ ↓					
1036 03 12 – 1037 03 11	M	CA-EAF		1036 03 12 – 1037 03 11	S												Isk ~~~	
1063 07 30 – 08 27	M		N	–														
1068 03 18 08 30	L		S	–														
1068 05 29	M		S-C	1068 05 29	S		~~ ↓↑	~~ ↓↑	~~~ ↓↑	~~~ ↓↑								
1138 10 11	M		N	–														
ca. 1150	M		C	–														
1156 12 09	M		N	–														
1157 04 02	M		N	–														
1157 07 05	M		N	–														
1157 08 12	L		N	–														
1163 08	M		N	–														
1170 06 29 03 45	L		N	–														
1202 05 20 02 40	L		C-N	1202 05 20	M													
1212 05 01	L		S	–														
1222 05 11 06 15	M	CA		1222 05 11	S													
1287 02?–03 22	M		N	–														
1293 01 11–02 08	M		S-C	–														
1303 08 08 03 30	V	HA-E		1303 08 08	M-L	Alx ~~ ↓↑					~	~	~~~ ↓↑	~				C ~~~ ↑↓
1404 02 20	M		N	–														
1408 12 29	M		N	1408 12 29	S											~~~		
1458 11 08/16	M		S	–														
1546 01 14 afternoon	M		C	1546 01 14	M		~~~ ↓↑	~~ ↓↑	~~ ↓↑	~~ ↓↑	~~~ ↓↑							
1588 01 04 13 00	M		S	–														
1705 11 24	M		N	–														
1738 09 25	M		N	–														
1759 10 30 03 45 LT	M		C-N	1759 10 30	S–M								~~~ ↑	~	~~~			
1759 11 25 19 23 LT	L		N	1759 11 25	S–M	Nile Delta ~~~	?	?	?	?	?	?	~~~ ↑					
1796 04 26 09 05	M		N	–														
1822 08 13 20 40	L		N	–														
1834 05 26 04 00	M		C	–														
1837 01 01 14 34	L		C-N	–														
1870 06 24 17 00	L	HA-E		1870 06 24	M	Alx ↑~~~												~~~

(*continued*)

Table 1
(Continued)

Earthquakes				Tsunamis		Affected Coasts[#,**]													
							Levant												
							Israel							Lebanon					
Date*	Size[†]	Remote[‡]	Local DST[§]	Date*	Size[¶]	Egypt	Gaza	Ashkelon	Ashdod	Yavneh	Jaffa	Caesarea	Akko	Tyre	Said/Beir/Tri	Syria	Cyprus	Asia Minor	Greece/ Crete
1872 04 03 07 40	L		N	1872 04 03	S											Ant $\sim\sim\sim$ ↑			
1908 12 28 05 20	L	IT		1908 12 28	L	Alx $\sim\sim\sim$													
1918 09 29 12 07	M		N	–															
1927 07 11 13 04	M		C	–															
1953 09 10 04 06	M	CA		1953 09 10	S												$\sim\sim$	?	
1956 07 09 03 11	M	AS		1956 07 09	M						$\sim\sim\sim$								
1995 11 22 04 15	L		S	–															G, C $\sim\sim\sim$

Number of Tsunamis	Egypt	Gaza	Ashkelon	Ashdod	Yavneh	Jaffa	Caesarea	Akko	Tyre	Said/Brt/Tri	Syria	Cyprus	Iskendrun	Greece/Crete
Since mid second century B.C., for this coast	NC[#] 6	3	3	3	4	6	5	8	6	3	3	NC[#] 3	NC[#] 2	NC[#]
Since mid second century B.C., for this country					Israel: 11				Lebanon: 6		3			
Since mid second century B.C., Levant, Gaza to Syria							Levant: 13							

(E) See complementary information, sources of information and references regarding the lists of tsunamis and earthquakes in the electronic edition of BSSA.)

*Events are marked by time of occurrence, as detailed as known: year, month, day, hour, minute. Timing of earthquakes and tsunamis does not always match and it may reflect ambiguities in the historical data. We therefore associated a tsunami with its most likely causative earthquake. 3/M, The "reliable period" used for our hazard evaluation starts at about mid second century B.C.

†Size of the earthquakes. We follow the broad categories suggested by Ambraseys and Jackson (1998): V, very large event ($M_s \geq 7.8$); L, large ($7.8 > M_s \geq 7.0$); M, moderate ($7.0 > M_s \geq 6.0$); and S, small ($M_s < 6.0$). Estimations were taken from historical, geological, and paleoseismological studies. Where not available, we estimated the size according to our best judgment.

‡Regions of remote earthquakes: AS, Aegean Sea; CA, Cypriot Arc; CA-EAF, the transition zone from the Cypriot Arc to the East Anatolian fault, near the triple junction with the DST; HA-E, Eastern Hellenic Arc; HA-W, Western Hellenic Arc; IT, Italy.

§Local earthquakes along the DST system, not necessarily on the main transform: N, northern part, in Syria and Lebanon; C, central part, in Israel, from the Hula Valley to the Dead Sea; S, southern part, in Israel, Arava Valley, and southward.

¶Areal extent of the tsunami: L, occurred over a wide range or in several distant coasts along the eastern Mediterranean; M, spread along several nearby coasts; S, limited to a few coasts only.

#Affected coasts along the eastern Mediterranean are from south to north. Locality abbreviations and historical names: Akko, Acre, Ptolemais, in Israel; Alx, Alexandria, in Egypt; Ant, Antakya (Antioch), in Turkey near the Syrian border; Brt, Beirut, in Lebanon; C, Crete; G, Greece; Isk, Iskenderun, Alexandretta Bay, in Turkey; Pel, Pellusium in Sinai in Egypt; Said, Saida, Sidon, in Lebanon; Tri, Tripoli, in Lebanon; Tyre, Sur, Tsur, in Lebanon.

**Tsunami symbols: $\sim\sim\sim$, reported for a specific site; $\sim\sim$, reported at coasts of this country; $\sim$, reported at a general region; "?" tsunami inferred or questionable. Notable phase of the tsunami (if explicitly mentioned in the literature), not necessarily the first one: ↓, down; ↑, up. Order of phases in time is from left to right.

Finkel (1995); critical reviews and reappraisals (Ambraseys and White, 1997; Ambraseys 2002, 2005a,b; Karcz, 2004); and focused investigations of specific events (Ambraseys and Melville, 1988; Ambraseys and Barazangi, 1989; Ambraseys and Karcz, 1992; Darawcheh *et al.*, 2000; Guidoboni *et al.*, 2004a,b). Many other lists draw from both primary and secondary sources, and summarize and update past catalogs (e.g., Plassard and Kogoj, 1968; Ben-Menahem, 1991; Amiran *et al.*, 1994; Khair *et al.*, 2000; Sbeinati *et al.*, 2004). We refer to these works after cross-correlating their data with the primary sources.

Indirect information was also useful for our compilation. Field evidence for historical earthquakes provides a better estimate of the source parameters that might have triggered tsunamis. Among these are paleoseismic investigations that may associate (although with some uncertainties) a specific surface rupture with a given historical event (e.g., Reches and Hoexter, 1981; Marco *et al.*, 1997, 2003, 2005; Ellenblum *et al.*, 1998; Amit *et al.*, 1999; Klinger *et al.*, 2000; Gomez *et al.*, 2001; Neimi *et al.*, 2001; Meghraoui *et al.*, 2003; Zilberman *et al.*, 2004, 2005; Daëron *et al.*, 2005) surface faulting associated with earthquakes (Ambraseys and Jackson, 1998), and reports on lacustrine seismites and deformed layers from the Dead Sea Basin (Enzel *et al.*, 2000; Ken-Tor *et al.*, 2001; Migowski *et al.*, 2004) that may attest to the strength of shaking or distance to many of these events.

List of Tsunamis

The earliest known account of a tsunami reported the flooding of Ugarit, a coastal city along the Syrian coast, in about 1365 ± 5 B.C. (see Ambraseys *et al.*, 2002). The second reported tsunami occurred in the mid second century B.C., probably in about 143 or 142 B.C. and was investigated by Karcz (2004), although its tsunamigenic source has not been fully clarified. As we later examined the seismicity and realized that reports prior to the mid second century B.C. are very likely incomplete, it is reasonable to assume that so is the millennial gap between these two tsunamis. We therefore list the tsunami at Ugarit but disregard it for the statistics.

For each of the selected tsunamis, we examined the time of occurrence and areal spread along the Levant coast (Table 1 and Ⓔ supplemental material available in the electronic edition of BSSA). In most cases, the sources cite the specific port or coastal city where the tsunami hit, or just mention the occurrence of the tsunami. In secondary catalogs, however, the region or the country affected by the tsunami is also mentioned but it is not always clear whether this was in the original report or simply the interpretation given by the cataloguer. For example, the chronicles of the 18 January 746 earthquake report that "There was also an extraordinary storm in the sea, such that the waves rose up to the sky" (from Guidoboni *et al.*, 1994; Ambraseys, 2005b), but they do not mention the name of the sea. Later cataloguers put this tsunami "on Mediterranean coast" (e.g., Amiran *et al.*, 1994) or "in Lebanon and Egypt" (Soloviev *et al.*, 2000) which is a reasonable interpretation, but not the only one.

The tsunamis varied in size from barely noticeable to greatly damaging, and ranged from affecting local ports to flooding several coasts from the same event. The historical descriptions do not contain quantitative parameters such as run-up or extent of inundation, but the need for assigning magnitude or intensity to each of the tsunamis is essential. Even the semiquantitative description provided for the 1759 tsunami in Acre (Akko) should be questioned in terms of its interpretation. In that account, "The water rose to 8′ (~2.5 meters) . . ." is a vague record because the height above sea level and the tide at the time of the tsunami are not mentioned. In fact, this figure describes the inundation height at an unknown location in Akko, and in the narrow alleys of the old city, the flowing water has no way to go but to pile up. Thus, the actual run-up cannot be determined without more data.

Nevertheless, these reports provide valuable insights into the tsunamigenic process along the Levant coast. The association of a tsunami with a specific earthquake, the distance they traveled and the extent of their affect along the coast, the reported presence of a receding phase and the damage it caused are all firm data that allow us to qualitatively understand what happened. We estimated the extent of each tsunami by determining whether it was local and limited to only a short part of the coast (S in Table 1), was spread along an extended part of the coast (M), or affected a large part of the eastern mediterranean coastline (L). We also mention if the tsunami was associated with the receding of the sea, although in most cases it was not possible to determine whether this was also the first phase of the tsunami. At the end of Table 1, we summarize the number of tsunamis listed for each of the coasts and countries included in our study. In total, we count 22 reliably reported tsunamis, of which 21 are from the period between mid second century B.C., when historical accounts are better and more frequent, and the present.

Doubtful Tsunamis

Tsunami reports that could not be substantiated by primary sources were considered doubtful and were therefore excluded from our primary list in Table 1. (Ⓔ Supplemental material is available in the electronic edition of BSSA.) Most are doubly reported events or those that were listed erroneously, and some are those with highly questionable dates (e.g., likely due to misinterpretations in calendar date corrections). However, it is possible that some tsunamis on this list actually did occur but are very poorly recorded; future new discoveries may elucidate these events.

In contrast, many of the tsunamis in this group have already been investigated and found questionable. For example, regarding the doubtful tsunami in A.D. 76, Ambraseys (1962) states: "Correct translation of Greek text indicates storm sea-waves, not uncommon in Cyprus."

Interestingly enough, about 65% of all tsunami entries we found in the literature were categorized here as doubtful, and this is not a unique example (Ambraseys *et al.*, 2002). Tinti *et al.* (2004) listed 45 false tsunamis for the Italian tsunami catalog (ITC), which overall contains 67 reliable entries.

Also, we list at least one real tsunami in the doubtful category because it apparently did not affect the Levantine coast, although it is also possible that there are simply no accounts of it for that region. Specifically, the eruption and collapse of Thera Island around 1627–1600 B.C. (dating by Friedrich *et al.*, 2006) caused a tsunami that struck the Aegean region and affected the Late Minoan culture. The pumice from the eruption found in Cyprus and Israel was critically questioned as a true tsunamite deposit of this event (Dominy-Howes, 2002), although models (that are based on the pumice findings) potentially show that the Late Minoan tsunami could have reached to the Levant (Yokoyama, 1978). We list this event as doubtful, not because it may not have occurred, but because it is doubtful that it produced a tsunami along the Levant coast.

List of Earthquakes

Constructing the list of earthquakes (also in Table 1) helped us to better understand the origin of past tsunamis, define their typical sources, and produce realistic models of possible future scenarios. The relevant "local" earthquakes are those that occurred within the tectonic framework of our study, and this is mainly the DST and its associated structures. Inclusion of all the significant earthquakes allowed us to describe the type of events that do, or do not, generate tsunamis, and to estimate the probability of a future tsunami if a large earthquake occurs in the region. We define "significant" earthquakes as those that caused damage or loss of life in at least two localities, and earthquakes that were associated with a tsunami. Earthquakes that were "only" reported as "felt" or where damage was only reported to have occurred at one site were not included because they were probably of smaller magnitude and the historical record is incomplete for such events. Thus, inclusion of these events may bias the interpretation for the relationship between tsunami generation and large earthquakes. Furthermore, Khair *et al.* (2000) have already presented a general catalog of DST earthquakes for the past four millennia, but for our purposes, we excluded the smaller-magnitude events and screened out the dubious events by crosscorrelation to primary sources.

For each of the events, we examined the source of information, and if found to be reliable, we cited its time of occurrence to the extent that it is known (year, month, day, hour, and minute), and its estimated size (magnitude) and area of occurrence (Ⓔ See supplemental material in the electronic edition of BSSA).

We distinguish moderate from large earthquakes to examine the possible relationship between earthquake size and the potential for tsunami generation. With the present understanding of the magnitude of historical seismicity, which is mainly based on macroseismic damage and the area where the shock was felt, rather than on quantitative measurable parameters, it is impossible to assign a clear threshold magnitude for the large, moderate, and "excluded" events. We follow the broad categories suggested by Ambraseys and Jackson (1998) where V is a very large event ($M_s \geq 7.8$), L is large ($7.8 > M_s \geq 7.0$), M is moderate ($7.0 > M_s \geq 6.0$), and S is small ($6.0 \geq M_s$), this last category being the likely threshold for our "excluded" events. Where available, our estimates are taken from previous studies that were based on primary sources, on "M_e" determinations ("equivalent magnitude value calculated using the method of Gasperini *et al.* [1999] and Gasperini and Ferrari [2000]," as given by Guidoboni and Comastri [2005]), and paleoseismology. If not available, we estimated the size according to the degree of macroseismic damage and extent of affected area. Obviously, this is a subjective procedure because some large events may be underreported and listed as moderate. Similarly, inflated reports of moderate earthquakes, some closely timed moderate events, a mainshock followed by an intensive aftershock sequence (e.g., "earthquakes lasted 40 days"), an earthquake swarm with several strong events, and a sequence of strong events, may all seem like one large earthquake to the people of the time. Altogether we note 57 significant DST earthquakes since the mid second century B.C. and another one before that.

The location of historical events is no less subjective and tricky than determination of magnitude. For example, given the long and narrow pattern of the populated area of the Levant (constrained by the Mediterranean Sea on the west and the Syrian and Arabian deserts in the east), which is more or less parallel to the north–south trend of the DST, the isoseismals of strong earthquakes may tend to stretch along the inhabited regions and always "coincide" with the strike of the DST. Moreover, the directivity effect may cause the maximal damage in an area that is north or south of the earthquake epicenter and the rupture zone, and thus shift the high isoseismals away from the epicenter. For these reasons here we give only a rough estimate of the location of the historical events, whether they occurred in the northern part of the DST (N, in Syria and Lebanon), the central part (C, Israel, from the Hula Valley to the Dead Sea), or the southern part (S, south of the Dead Sea, in the Arava Valley and Gulf of Aqaba [Elat]).

We also referred to the occurrence of seismites (mixed or deformed layers) in the Holocene deposits of the Dead Sea Basin as indicators for the strength of the shaking there (Enzel *et al.*, 2000; Ken-Tor *et al.*, 2001; Migowski *et al.*, 2004). Those observations suggest the occurrence of several historical earthquakes that have not (yet?) been recognized or reported in the historical literature (e.g., A.D. 175, A.D. 90, 700 B.C.), although that record apparently is also not complete because it lacks evidence of some significant events, either by being masked by subsequent events, by having occurred during a sedimentary hiatus, or for some other as yet unknown reason. For example, the events of

A.D. 306, 349, and 363 are all obscured, although the latest is known to have generated a sea wave in the Dead Sea. This effect also does not allow for discrimination between remote strong events and close moderate ones. Nevertheless, it supports the occurrence of questionable earthquakes, such as the one of 1656 that had been reported to affect Tripoli, either in Libya, or Lebanon, or both.

Paleoseismology seems to be most helpful in resolving earthquake sources, although in most cases it provides a time window for the occurrence of an event and the correlation with a given historical earthquake within that time frame is only an association. Currently there are a limited number of observations on the DST, but they still support the contention that some of the tsunamigenic events resulted from on-land surface ruptures, and thus establish the proposed hypothesis of this unique but typical scenario that DST on-land earthquakes trigger tsunamis via seismogenically induced submarine slumping (Ⓔ See supplemental material in the electronic edition of BSSA).

Several tsunamis are reported to have arrived from distant sources in the Mediterranean and the Aegean seas, and From as far as Italy. We systematically searched the literature for parenting sources for these tsunamis and listed them in Table 1, together with all other "local" events (Ⓔ See supplemental material in the electronic edition of BSSA).

Uncertainties and Completeness of the List

There are large uncertainties that relate to all of the reported historical earthquakes and tsunamis, as well as to the cumulative lists of those events. This may include the location, time, magnitude, areal extent, effects, and sometimes even the very occurrence of the reported events. Two more issues are relevant to constructing such a list if one intends to statistically analyze the data. The first is completeness. Are all major events recognized and documented? Historical catalogs show a dramatic increase in reporting after about the mid second century B.C. (probably in part reflecting the growing influence of the Roman Empire in the Middle East). Prior to about this time, the reports are very sparse and we consider the record to be largely incomplete. After this time, the record is probably also incomplete but very likely captures the largest and most important events. Khair *et al.* (2000) suggested that continuous accumulation of records of significant earthquakes started at about 184 B.C., and that later variations in the seismicity reflect low- and high-seismic-activity periods. Consequently, we discuss only the past ~2150 years of this catalog that we consider "reliable" and take this era as the start of the "reliable period."

The second issue relates to the threshold for when a tsunami is sufficiently noticeable to have been reported. It is highly likely that many small tsunamis occurred, but for which the rise in sea level was too small to notice, there was no damage, it happened at night, or it may have occurred in an area where no one was literate to report it. A small tsunami at low tide may go unreported, whereas the same-sized tsunami at a critically high tide may cause damage. Thus, the list of reliably reported tsunamis is probably a minimum accounting of the actual occurrences.

Relationship between Tsunamis and Earthquakes: Tsunamigenic Sources

The list of reliable accounts (Table 1) allows for a complete cross-correlation of earthquakes and tsunamis. This is an unconventional presentation in the sense that it brings together tsunamis that occurred in a given region together with their local and remote sources (if identified), the significant nontsunamigenic earthquakes of that region, and the orphan tsunamis. Nevertheless, it allows for the generalization of the tsunamigenic environment of the region and sets up typical future scenarios. All in all, we identified ten tsunamis that originated from "local" sources, four tsunamis that arrived from the Hellenic Arc and the Aegean Sea, and four from the Cypriot Arc, one from Italy, and an additional two tsunamis that are left with as-yet no identified trigger. We discuss these sources next.

Not included in the summarizing list, but no less important, are local earthquakes not related to the DST system that did not produce a tsunami, either because they were not large enough or because they were too far from the coast. Such is the large A.D. 1042 or 1043 event from the Palmyra region of northeastern Syria (Guidoboni and Comastri, 2005), located about 200 km east of the Mediterranean coast. The same could be inferred from modern recorded earthquakes for which accurate location and magnitude determination are available. For example, the southern Suez Rift (m_b 7.0, 1969) and the Gulf of Aqaba (M_w 7.1, 1995) did not generate a tsunami in the Mediterranean (Ⓔ See supplemental material in the electronic edition of BSSA). These are worth noting because they represent seismic sources that can produce large earthquakes in the future but are unlikely to produce tsunamis (unless they generate a considerably stronger shock much closer to the Mediterranean).

We did not identify any historical tsunamis that reached the Levant coast from volcanic eruptions or remote landslides. Nevertheless, future discovery of field evidence or of as-yet-unknown chronicles for past events, or occurrence of such events in the future, should not be ruled out.

Local Earthquake Source: The Dead Sea Transform System

The 57 moderate to large earthquakes we list since about the mid second century B.C. range in location from southernmost Israel to northern Syria and southern Turkey, so they essentially encompass most of the DST system from the Gulf of Aqaba to its junction with the East Anatolian fault, near the Bay of Iskenderun. Nearly a sixth of these, ten events, produced tsunamis somewhere along the eastern Mediterranean coast that were large enough to have been noticed and documented (Fig. 1a). Under closer inspection, however, it is also clear that the size of the earthquake, among many other factors, affects whether a tsunami is likely to be generated. Of the 43 moderate earthquakes interpreted in Table 1, only six (about 14%) produced a tsunami, whereas four of the 14 larger events, or about 29%, produced a tsunami. Thus, between a quarter to a third of the largest and a seventh of the moderate DST earthquakes were tsunamigenic. Clearly, larger events exhibit a much higher (factor of 2) likelihood to produce a tsunami than smaller (moderate) earthquakes.

Note that the 1546 earthquake, which is considered a moderate event by Ambraseys and Karcz (1992), generated a significant tsunami that affected the southern coast of Israel. They state that the magnitude estimate for this event is "Ms about 6.0, in many respects similar to that of the earthquake of 1927," which was M_L 6.2 but did not trigger a tsunami. The other low-magnitude events that produced tsunamis were the May 1068 and 1408 events, both estimated as M_e 6.0 (Guidoboni and Comastri, 2005). These are probably the smallest tsunamigenic DST earthquakes, although it is also possible that their magnitudes have been underestimated because of the lack of recorded reports. Because we have no records for $\sim M < 6$ tsunamigenic DST earthquakes, the threshold for tsunami generation may be in the range of M 6.0–6.5. That is, it apparently requires large enough earthquakes to trigger offshore landslides that in turn produce tsunamis, although no more than a seventh of these were demonstrated to be "successful."

Most of the earthquakes north of the Dead Sea, for which a surface rupture was suggested by paleoseismic observations, produced a tsunami. These are the 551 (Elias *et al.*, 2001; Daëron *et al.*, 2004), 746 (Marco *et al.*, 2003), 1202 (Marco *et al.*, 1997, 2005; Ellenblum *et al.*, 1998; Daëron *et al.*, 2005), October 1759 (Marco *et al.*, 1997, 2005; Ellenblum *et al.*, 1998), and November 1759 (Gomez *et al.*, 2001; Daëron *et al.*, 2005) earthquakes, with the exception of 1170 (Meghraoui *et al.*, 2003). Some findings (Reinhardt *et al.*, 2006) may also suggest the occurrence of a tsunami after the 115 rupture (Meghraoui *et al.*, 2003). These were also among the largest DST earthquakes (551, 1202, November 1759), but the data we have are insufficient to conclude that they also produced the largest "local" tsunamis. In contrast to the north, no tsunamis have been reported for the events of 1212 and 1458, for which possible surface ruptures were suggested in the Arava Valley (Klinger *et al.*, 2000), south of the Dead Sea. Regarding the 1068 events, a tsunami was related to the second shock that occurred on 29 May, north of the former one of 18 March (Guidoboni and Comastri, 2005). It is therefore reasonable to assume that the surface rupture found in the southernmost Arava Valley by paleoseismology (Amit *et al.*, 1999; Zilberman *et al.*, 2005) belongs to the first event, 18 March, which was nontsunamigenic. The large 1995 M 7.1 earthquake in the northern Gulf of Aqaba did trigger a tsunami in the Gulf (Wust, 1997), but not in the Mediterranean.

Examination of the location and extent of the tsunamis in relation to the location of the earthquakes that generated them indicates that most of the tsunamis struck the coast adjacent to the onshore area damaged by the earthquake. The relationship between the occurrence of a tsunami and the distance that the seismogenic source is from the coast or from the continental margin is not clear. No tsunamis were reported for the strong earthquakes that occurred in northern Syria during the twelfth century A.D. (Ambraseys, 2004; Guidoboni *et al.*, 2004a,b), although the moderate 1408 earthquake, located the same distance from the coast, did generate a tsunami. Furthermore, the May 1068 and 1546 earthquakes that originated farther away from the coast both produced tsunamis. It appears that the largest distance between the DST and the coast for which we have information of an earthquake-generated tsunami is about 80–100 km, but of course, it should also be magnitude dependent. In any case, this appears to be the upper limit.

Another contributing complication is that the DST system distributes deformation along the neighboring plates, with normal faults along its rift in Israel and thrusts along its restraining bend in Lebanon. Paleoseismic findings suggest that these secondary structures have generated earthquakes that produced tsunamis as well. Such was the case in the 551 earthquake that has been attributed to the Beirut thrust (Daëron *et al.*, 2004) and the October and November 1759 events that apparently ruptured the Rachaya and Serghaya faults, respectively (Gomez *et al.*, 2001; Daëron *et al.*, 2005).

Overall, it appears that all of the tsunamigenic DST earthquakes (with the exception of May 1068?) have been located north of the Dead Sea with magnitudes stronger than M 6–6.5, and most were probably associated with surface rupture. In a predictive sense, future tsunamis will most likely occur opposite the rupture and damage zone of future large earthquakes, and there is a known significant seismic hiatus along at least the Jordan Valley, one of the closest sections of the DST to the coast.

Onshore versus Offshore Sourcing

The notion that tsunamis in the eastern Mediterranean are generated by seismogenic submarine landslides was suggested to explain the occurrence of tsunamis right after onshore earthquakes. Supporting evidence comes from historical descriptions of tsunamis that start with or include a remarkable receding phase of the sea (e.g., 1068, 1546), in accordance with a scenario of a tsunami generated by a seaward-moving submarine landslide. Also, the bathymetry of the continental margins of the eastern Mediterranean is spotted with numerous typical slump scars (Almagor and Hall, 1984), either on its steep southern slopes or deep northern canyons. Moreover, geotechnical studies by Frydman and Talesnik (1988) show that these margins become unstable under seismic shaking of about 0.1g, which is the expected value given by the Israeli building code (based on the 500-year recurrent event, acceleration of up to 0.175g

should be expected for the northern coast of Israel, with an expected lower value for the continental shelf and slope, some 20–30 km westward).

Thus, most of the tsunamis that followed DST earthquakes should have resulted from submarine slumps and paleoseismology suggests that such were the earthquakes of 746, 1202, October 1759, and November 1759. Of special interest is the 9 July 551 event. That earthquake presumably ruptured the Beirut thrust (Elias *et al.*, 2001; Daëron *et al.*, 2004) offshore of Lebanon, including its continuation onland toward the Roum fault that branches off of the DST. A portion of this fault was also suspected as the source of the $M \sim 7$ 1837 earthquake (Ambraseys, 1997). We suggest that the 551 tsunami could have originated either by a direct rupture in the sea or as a submarine landslide or as a result of both.

Distant Tsunamigenic Earthquakes

Eight tsunamis originated from outside the DST system (Fig. 1b). Four arrived from the Hellenic Arc and their sources are believed to be close to or greater than M 7.5–8 (365, 1303, 1870, and 1956). However, only the 1303 and the 1956 tsunamis were reported to have hit the easternmost Mediterranean coast. Nonetheless, the impact of the other two, which were reported "only" on Alexandria, especially the 365, should not be underestimated. The tsunami traveling the farthest arrived from the 1908 Messina Straight event (estimated as M 7.5 by International Seismological Centre [ISC] [2001] and M 7.1 by Boschi *et al.*, 2000), barely reaching Alexandria and the Nile Delta.

Four tsunamis are listed for the Cypriot Arc, which generated tsunamis in 1222 and 1953, the last of which also arrived to Asia Minor. The 802–803 and 1036–1037 tsunamis in Iskendrun Bay may have originated from the northeasternmost tip of that arc, already in Anatolia, where it approaches the East Anatolian fault. No damage was reported for the 1036–1037 tsunamigenic earthquake (Guidoboni and Comastri, 2005). The only tsunami that originated from an earthquake along the DST that is reported to have struck Cyprus was produced by the 1202 earthquake. We did not investigate tsunamis arriving at Cyprus from the Hellenic Arc or from Asia Minor; therefore, our list for Cyprus may not be complete.

The preceding observations suggest that tsunamis along the easternmost Mediterranean do arrive from distant sources and from near non-DST sources, but they are not as destructive as the local ones, with perhaps the significant exception only of the 1303 event. The 365 catastrophe in Alexandria, however, suggests that the Nile Delta is more vulnerable and needs further attention.

Unidentified Tsunamigenic Sources

The last category we report is for tsunamis that have no known earthquake or volcanic trigger, and they all date to the third and fourth millennium before the present. It is reasonable to presume that these tsunamis could be related to either nearby earthquakes or distant sources for which there are currently no direct documentation or reports. For instance, the flooding of Ugarit, ca. 1365 B.C., could have resulted from some of the many earthquakes that destroyed this coastal city during the Bronze Age, but there are no direct indications for that (Ambraseys *et al.*, 2002). We know that in the past 2200 years, the Ugarit region was damaged repeatedly by DST earthquakes, at least one of which was tsunamigenic (A.D. 1408). Deformed layers in the Dead Sea are associated with more than half of the earthquakes and tsunamis introduced in our list, and there is also a deformed layer related to the date of the Ugarit flooding (1365 B.C.) (Migowski *et al.*, 2004). From this, it is possible that the Ugarit tsunami was associated with a DST earthquake but there are no direct accounts of this so we leave this as an orphan event.

Similarly, the mid second century B.C. flooding of the shore in southern Lebanon could be related with some of the notable shocks at that time. Karcz (2004) suggests that the flooding may be related to an earthquake in Sidon (not dated) but excludes the 198 B.C. earthquakes as its potential source. Migowski *et al.* (2004) document a deformed layer at 140 B.C., which is also close in time to this tsunami. Nevertheless, we cannot say with certainty that this tsunami directly resulted or followed an earthquake.

The last orphan tsunami on our list occurred in about 23 ± 3 B.C. in eastern Egypt. It is possible that it is associated with the 17 B.C. earthquake in Cyprus, which is the nearest in time and place to it. However, no tsunami was reported for this quake, and this would require an additional error in the reported date of the tsunami.

Although one can claim that these orphans may have been triggered by undocumented earthquakes, we cannot exclude a scenario that a tsunami was triggered by a spontaneous submarine slide or a small offshore earthquake that was not felt on land.

Other Possible Local and Distant Sources

There could possibly be other tsunamigenic sources in the eastern Mediterranean that have not produced tsunamis in the past two millennia. Although we could not verify their effect within the historical time frame and the given resolution of our data, their potential to generate future tsunamis should not be excluded. Among these we list the seismogenic offshore area north of Egypt and Sinai, where M 6 events have been recorded in modern times (potential source of the 23 ± 3 B.C. tsunami?), the Nile Delta slopes, the great submarine slide near the Anaximander Seamount (Ten Veen *et al.*, 2004), and the volcanoes of the Aegean Sea that were already shown to be tsunamigenic (the LM tsunami and another one, on A.D. September 1650, which was local and probably originated with the collapse of Mt. Colombo, near Thera, that erupted at that time [Dominey-Howes *et al.*, 2000; Dominey-Howes, 2002]).

The only hint of an asteroid-generated tsunami on the eastern coastal area of the Mediterranean Sea was given by Soloviev *et al.* (2000) for the 198 B.C. event: "Shortly after the appearance of a big comet an earthquake occurred that was accompanied by an overflow of the sea water." However, this event was probably duplicated from the event of 373 or 372 B.C. that occurred in the Gulf of Corinth (Greece). We therefore list it with doubtful events.

Last but not least, the passive margins along the eastern Mediterranean are still being loaded with sediments primarily derived from the Nile Delta, although at a decreased rate after construction of Aswan Dam. This implies that sediments are still accumulating on the continental slopes and increasing the potential for slope failure. It is reasonable to assume that most of this potential is cleared (out) by seismically induced ground shaking before it naturally matures to a spontaneous failure, but this should certainly be verified.

Threshold Magnitude for Tsunamigenic Earthquakes

Because no "local" tsunamis have been observed along the Levant coast during the instrumental period, we can only approximate the threshold magnitude of nontsunamigenic DST earthquakes. Specifically, neither the M_L 6.2 1918 earthquake in Syria nor the M 6.2 1927 Dead Sea earthquake in Israel produced a tsunami (Ⓔ See supplemental material in the electronic edition of BSSA). On the other hand, it appears that many of the strongest or most damaging historical earthquakes (e.g., 363 and 1837), including those that ruptured the surface, have not produced tsunamis either (e.g., 1170). Therefore, there is no 100% "successful" tsunamigenic earthquake and no clear threshold magnitude. Obviously, this should be questioned since we cannot be sure that all past tsunamis were indeed observed and, if observed, whether they were reported.

Two similar-magnitude earthquakes may in one case trigger a tsunami, whereas in a different configuration, they may not. Clearly the magnitude is not the only factor and the distance from the coast, the focal mechanism or depth, and the effects of directivity, probably all play a role. Therefore, past experience gives only a rough impression of that threshold and detailed modeling is needed to determine the typical properties of a tsunamigenic earthquake in the region. We can only estimate the threshold magnitude for nontsunamigenic earthquakes, and this seems to be about $M \sim 6$. Above this threshold, historical earthquakes in the range of $M \sim 6.0$–7.0, were only 14% successful, and for higher magnitudes 29% successful.

Mechanical considerations may suggest that landslide-generated tsunamis could be related to a time-dependent process. The continuous accumulation of sediment loading on the continental slope may result in decreased stability with time, and progressively smaller accelerations would be needed to generate a slope failure. Therefore, the threshold magnitude for tsunamigenic earthquakes may not be constant through time but may gradually decrease after the previous tsunami.

Tsunami Modeling

Several studies modeled tsunamis at the Levant coast. Striem and Miloh (1976) examined the occurrence of an offshore submarine slump in southern Israel with an area of 6×2 km and rock mass as thick as 50 m. They concluded that this will result in a tsunami associated with lowering of the sea level of up to 10 m, and a retreat of the sea to a distance of 0.5–1.5 km for about 0.5–1.5 hr, and which may occur once or twice per millennium. Miloh and Striem (1978) studied offshore surface faulting along the southern Israeli coast and found that this may cause a tsunami of up to 5 m high with a wave frequency of 20 min. In some tectonic configurations it could also be associated with recession of the sea level. Yokoyama (1978) simulated the arrival of the Thera tsunami to the coasts of Israel, constraining its run-up to 7 m in height based on the presence of pumice from this event on the coastal terraces near Tel Aviv (near Jaffa). Recently, El-Sayed *et al.* (2000) modeled the propagation of the 1303 tsunami from near Crete to Alexandria and Acre (Akko), and found that the first arrivals of the calculated tsunami are strongly regressive, in agreement with the historical reports. Hamouda (2006) also computed the propagation of that tsunami to Alexandria and concluded an arrival time of 35–45 min after the earthquake, with run-ups of 2–9 m along the Nile Delta.

A tsunami in the Dead Sea was simulated by Begin and Ichinose (2004) to understand the deposition of gypsum above brecciated beds in the Late Pleistocene sediments. They concluded that gypsum deposition was the result of mixing of the water column in the lake due to seismic sea waves in this lake, associated with strong earthquakes.

Ben-Menahem and Vered (1982) examined mareograms (tide gauge) at some Israeli ports and found that seiches have apparently been excited by wind in Haifa Bay. They showed that "the shape and size of the resonating water body is dependent to some degree on the direction and intensity of the exciting meteorological front."

The tsunami simulations here derive from linear dispersive water wave theory. It is explained in brief with a reference to many examples in the Appendix.

Models of Tsunami Sources

We recognized two principal tsunamigenic mechanisms: submarine landslides that follow local DST earthquakes and remote earthquakes originating in Italy and the Hellenic and Cypriot arcs. Global seismicity and seismotectonics of modern times show that earthquakes in the eastern mediterranean tend to concentrate mainly along plate boundaries (Salamon *et al.*, 1996, 2003). It is therefore reasonable to assume that earthquakes along these elements were in the past, and will probably be in the future, the main tsunamigenic sources, either directly if located in the sea or via submarine landslides. Potentially more sources could exist, but their rate of seismicity is lower, and therefore they are not considered here. Based on these conclusions we simulated three scenarios and with the great concern arising after the Sumatra tsunami of December 2004, we modeled the upper range of the probable magnitudes of these sources. A similar approach of simulating tsunamigenic earthquakes of magnitude equal to or greater than the highest ones registered in historical times can also be found in Tinti *et al.* (2005).

The 2004 Sumatra $M > 9$ great earthquake resulted from rupture of a very long fault, which in the past broke in only shorter segments. Such is also the case in the Cascadia region, where it was thought that the subduction interface was incapable of generating earthquakes larger than M 7.5 until it was recognized as the seismogenic source for an orphan tsunami in Japan in January 1700 (Satake *et al.*, 1996; Atwater *et al.*, 2005). These events show that accounting only for the instrumental data and historical information, without taking into account the tectonic framework as an indicator of the scale of the magnitude of the tsunamigenic source, may result in underestimation of the actual hazard. Future work, of course, is needed to evaluate the potential contribution of the extreme events to the overall hazard. Nevertheless, in our present state of knowledge, we believe it is necessary to outline the possible range of the hazard.

The Akhziv Landslide Scenario. The "on land earthquake–submarine landslide" is apparently the most common mechanism for the Levant coast, and the tsunami that followed the 1202 earthquake with a rupture proposed to be on land in Lebanon and northern Israel (Marco *et al.*, 1997, 2005; Ellenblum *et al.*, 1998; Daëron *et al.*, 2004) is a good example of this. For the occurrence of significant submarine slides along the continental margins of the Levant, opposite the proposed 1202 rupture, we followed Almagor and Hall (1984) and Almagor (1993): "At the point of Akhziv Canyon, great slabs of detached sediment blocked the thalweg which has excavated detours around them." Also Almagor and Garfunkel (1979) specifically pointed to: "Chunks of continental loess . . . were sampled down to depths of 900 m."

This scenario finds an origin in the 5×10 km Akhziv canyon headwall region (blue box in Fig. 2a), where a 25-m-thick sediment slice breaks loose and slides 30 km downslope at 40 m/sec. The material comes to rest in a 7×15 km run-out zone that is 18 m thick (red box, Fig. 2a). The drop height of the material is about 1200 m. A 30-km run-out with 1200 m fall height gives an effective coefficient of basal friction of $\mu = 1200/30000 = 0.04$. Low coefficients of friction of this size are typical of submarine landslides.

As with most landslide tsunamis, the initial sequence starts with a sea level draw-down over the excavation at the canyon headwall, and a sea level elevation at the toe of the slide mass farther offshore. The picture changes quickly, however, as the initial waves propagate out at speeds that exceed the advancing slide and the initial draw-down rebounds into a dome. Waves reach shore in about 5 min. The wave period is about 5 min. Predicted average run-up exceeds 4 m for 100 km of the coastline to the north and south. Being somewhat closer to the slide direction, locations toward the north experience a bit larger run-up. Figure 2b shows the waveforms arriving at different locations along the coast and the average run-up there. (Ⓔ See the movie in the supplemental material in the electronic edition of BSSA.)

Cypriot Arc Earthquake. Here we modeled a remote source of which the most notable tsunami followed the $\sim M$ 8, 1303 earthquake in Crete. The recent 1953 M_L 6.2 Cyprus earthquake, possibly along the Paphos Transform fault (Papazachos and Papaioannou, 1999) southwest of the island, has shown that tsunamis can be generated by that arc as well (Ambraseys and Adams, 1992), even with relatively small-magnitude events. Activity southeast of Cyprus is not much less intensive. The earthquake of September 1961 was an M_L 6.0 thrust event (Papazachos and Papaioannou, 1999; Salamon *et al.*, 2003), and the overall pattern of seismicity dips north-northwest to a depth of 80–100 km (Rotstein and Kafka, 1982) and extends about 300 km in length. Thus, from tectonic considerations, it is reasonable to assume that the subduction part of the Cypriot Arc may potentially generate a considerably strong earthquake. Our scenario originates as a strong earthquake along the Cypriot Arc, just opposite the northern Levant coast (Fig. 3a). The rupture occurs on two fault segments dipping 15 degrees with a down-dip width of 40 km. The total length of the two segments is 120 km. Five meters of pure thrusting gives the earthquake a moment magnitude of M 7.8. The large size given to this modeled event also illustrates the expected effect of a $\sim M$ 8 event coming from a longer distance, like that of 1303.

Like tsunamis from all long earthquake faults, most of the wave energy is radiated in directions perpendicular to the strike, southeast in this case, directly toward the Israeli and Lebanese coasts. Most of the Mediterranean coast from Port Said to Beirut should experience average tsunami run-ups of about 1.5 m, but keep in mind, as mentioned previously, factors of two deviations from the average should always be anticipated. The first waves arrive to Lebanon in about 20 min, and wave period is about 15 min. Figure 3b shows the waveforms and average run-up at different locations along the coast. (Ⓔ See the movie in the supplemental material in the electronic edition of BSSA.)

Beirut Thrust Earthquake. This is an example that is based on a possible historical event sequence. The mechanism and size of the modeled event follow Darawcheh *et al.* (2000), Elias *et al.* (2001), and Daëron *et al.* (2004), who suggested rupture of the Beirut thrust as the origin of the 551 M_S 7.2 earthquake and tsunami. Inland to the south, the pattern of this thrust follows the trace of the Roum fault that was also

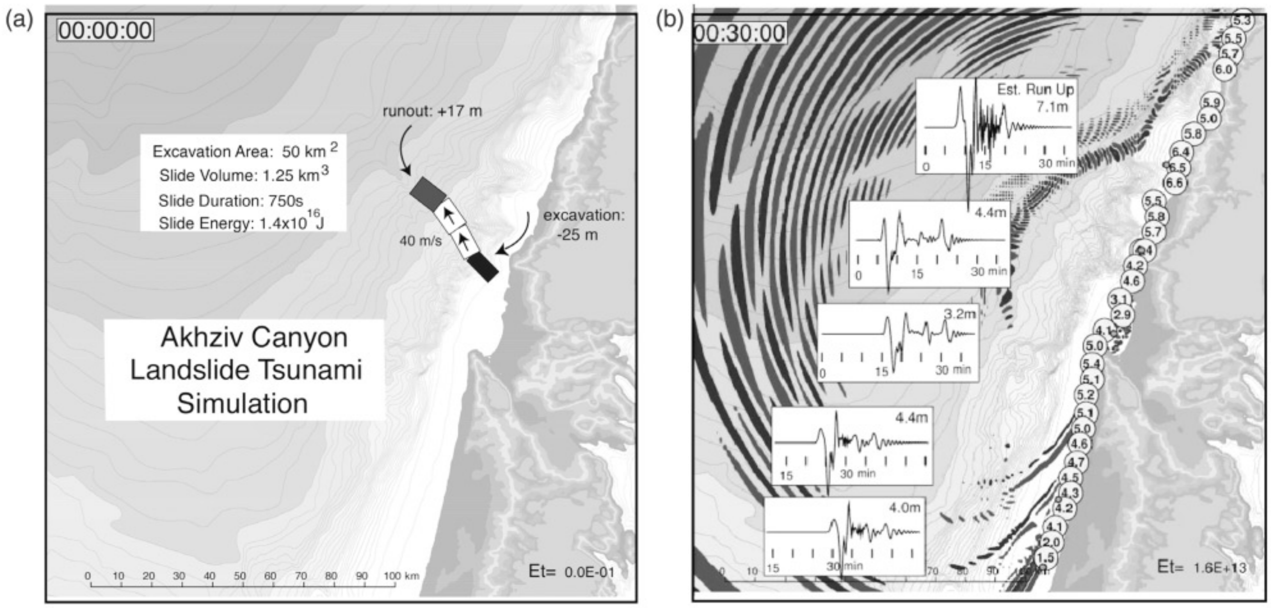

Figure 2. (a) Assumed parameters of the Akhziv Canyon landslide scenario. Quarter-minute bathymetry is from Hall (1981). (b) Estimated average run-up heights (numbers in circles) and waveforms at selected sites along the coast.

suspected as the origin of the 1837 *M* 7 earthquake (Ambraseys, 1997).

Here we modeled one fault segment dipping 30 degrees, with a down-dip width of 30 km and along-strike length of about 80 km (Fig. 4a). Pure thrusting of 4 m gives the source a moment magnitude of *M* 7.4. The faulting in this example was largely under land with only its northwestern edge in the sea, so the tsunami generation efficiency was low. At Beirut itself, the wave might run up to 3 m. Most of the Mediterranean coast north and south of Beirut experiences average tsunami run-ups of less than one meter and a wave period of 6 min. Figure 4b shows the waveforms and average run-up at different locations along the coast. (Ⓔ See the movie in the supplemental material in the electronic edition of BSSA.) A submarine landslide also could have been triggered by the 551 earthquake, thus resulting in a composite (earthquake- and landslide-driven) tsunami.

Model Results

Tsunamis from the two earthquake scenarios produced run-ups ranging from 1 to 3 m at the shore, in the range of the slip on the source fault ($<$5 m). As a rule of thumb, a tsunami run-up even proximal to an earthquake cannot be much more than the slip on the parent fault, and with distance, run-up values get even smaller because of geometrical spreading and frequency dispersion of the water waves. All of the earthquake tsunami simulations are nearly linearly dependent on fault slip, with all other fault parameters fixed. So, to produce larger run-ups, we would need to simulate extreme magnitudes, much higher than what we did for this study. Working against a strong tsunami at distance for the Beirut Thrust scenarios was the fact that only a small portion of the fault extended out into the sea. The result was a small effective tsunami source area. Tsunamis from small sources tend to spread at faster rates than tsunamis from large sources.

The tsunami from the landslide scenario produced an average run-up of 4–6 m on the nearby coast, and this is larger than any of the earthquake simulations. As illustrated in Figure 5, submarine landslides basically act like a moving uplift source, 17–25 m high in this case. These values far exceed the uplift associated with the earthquakes and hence account for larger initial waves. As in the earthquake tsunami simulations, tsunamis from landslides are nearly linearly dependent on landslide thickness, with all other slide parameters fixed. So if one wishes to increase or decrease the slide thickness from 25 m to construct other scenarios, the tsunami would scale proportionately. Working against a strong tsunami at distance for the landslide scenario was the fact that the uplift source area is fairly narrow, compared with that experienced in most earthquakes. Tsunamis from smaller landslide sources tend to have shorter periods and spread faster than tsunamis from larger earthquake sources.

Summary and Conclusions

In this article, we constructed a list of 21 reliably reported tsunamis that have struck the Levant coast, along with 57 moderate to large earthquakes that have occurred along the DST system, since about the mid second century B.C. (Table 1). Ten of the tsunamis were triggered by earthquakes that originated along the DST system (Fig. 1a), six of which followed moderate earthquakes and four followed large

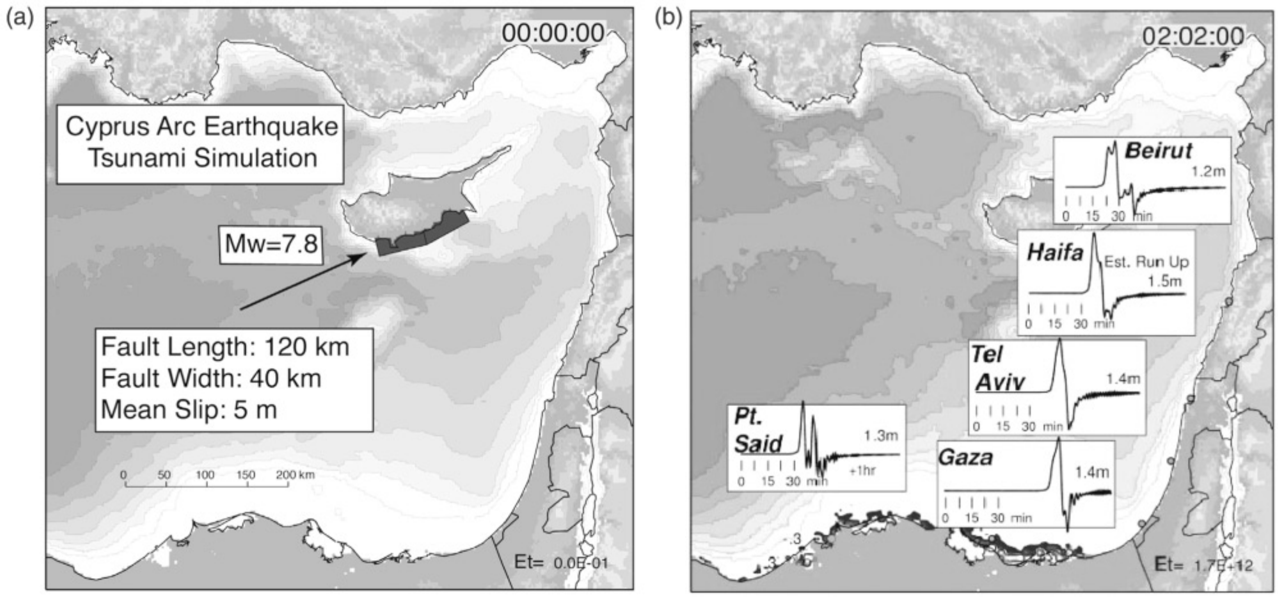

Figure 3. (a) Assumed parameters of the Cyprus Arc earthquake scenario. Two minutes bathymetry is from ETOPO2. (b) Waveforms and estimated average run-ups at selected sites along the coast.

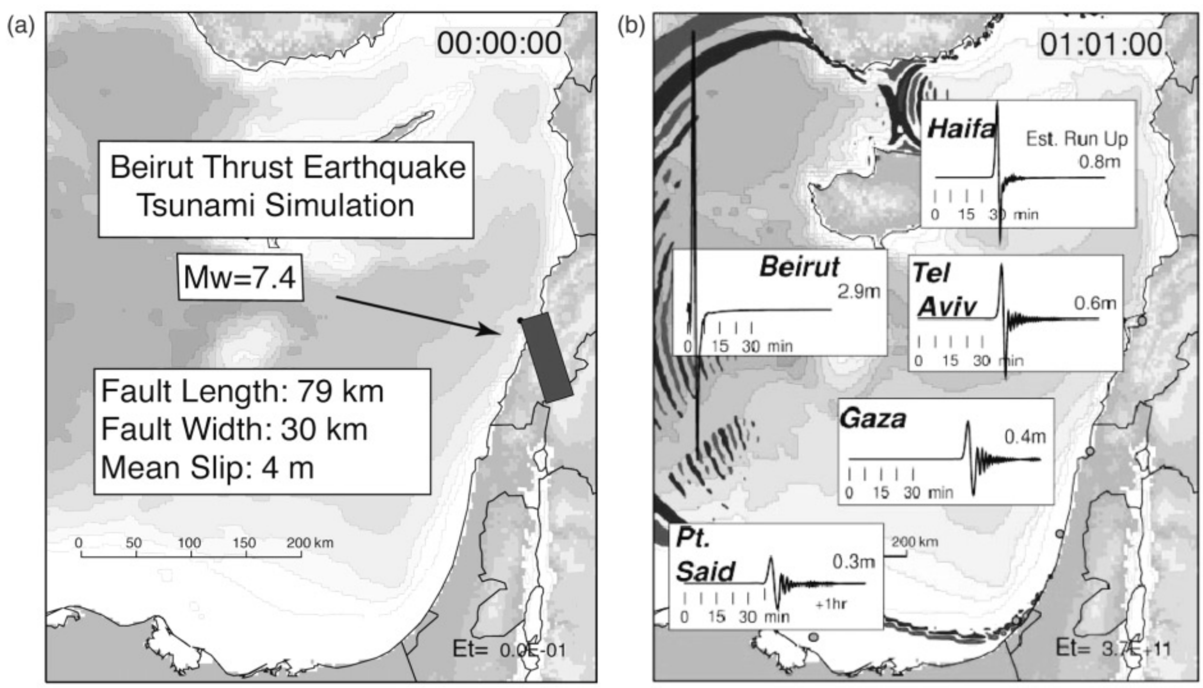

Figure 4. (a) Assumed parameters of the Beirut Thrust earthquake scenario. Two minutes bathymetry is from ETOPO2. (b) Waveforms and estimated average run-ups at selected sites along the coast.

earthquakes. These observations indicate that about a seventh (14%) of the moderate and from a quarter to a third (29%) of the large DST earthquakes were tsunamigenic. We estimate that the threshold of tsunamigenic DST earthquakes is likely to be in the range of *M* 6–6.5. Of the other 11 tsunamis, nine were associated with non-DST sources (Fig. 1b) in the Cypriot and the Hellenic arcs, and Italy, and two have no known trigger.

Nearly two-thirds of the tsunamis mentioned in the literature were found here to be doubtful, and if not ignored, would increase the estimated or perceived hazard. For example, considering all the tsunamis reported for the eastern

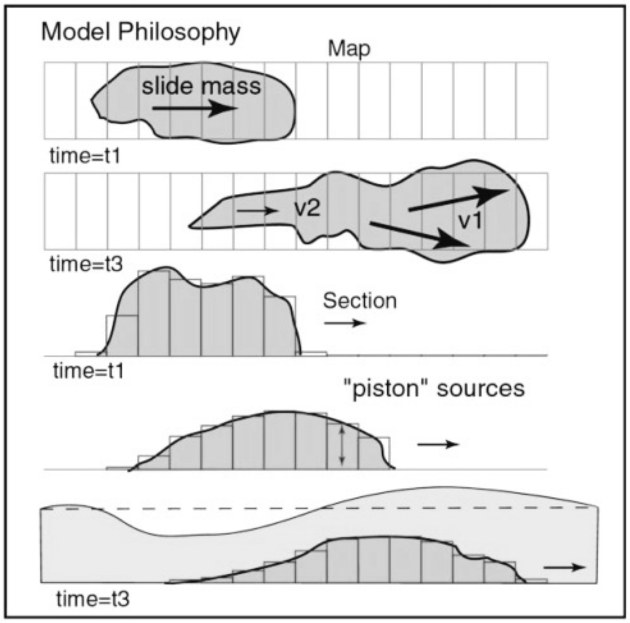

Figure 5. Philosophy in modeling tsunami from landslides. Real landslides deform and spread during sliding and velocities of particles can be in many directions (top two rows). We replace the deforming mass by a series of "piston sources" that move up and down according the volume of slide material over the piston (bottom three rows).

Mediterranean (Nile to Iskendrun Bay and Cyprus) during the past 2150 years results in an averaged repeat time of about 40 years, whereas accounting only for the reliable reports, the repeat time is extended to about 100 years. For the easternmost Mediterranean coast, from Gaza to Syria, the shortest period between tsunamis is 26 days (between the 30 October and 25 November 1759 events), whereas the longest period without a tsunami was apparently about 700 years, from the mid second century B.C. to A.D. 551. Hence the value of $\sim$100 years for the average repeat time of tsunamis is merely the product of 2150 years divided by 21 tsunami occurrences and cannot be used in a predictive sense.

These observations allowed us to recognize and model two principal mechanisms for the generation of tsunamis that will likely affect this region in the future. The most frequent events are those that originate from submarine landslides produced by on-land rupture of active faults (DST system), and from earthquakes originating beneath the sea on nearby subduction zones (Hellenic Arc, Cypriot Arc). A third mechanism is a combination of the two and it relates to a tsunami generated by movement on faults with at least some submarine ground rupture (the Beirut thrust). The landslide-related tsunamis produced the largest average run-ups of 4–6 m, whereas tsunamis produced by direct offshore fault motion are smaller, 1–3 m, a result of smaller seafloor motions.

Because nearly a third of the large historical earthquakes along the DST have produced tsunamigenic landslides, it is important to assess the likelihood of a future earthquake from that source. From examination of the seismic history of the DST, it appears that several of its significant segments have not ruptured in more than 800 years. These include the Missyaf segment in Syria that last ruptured on 1170 (Meghraoui *et al.*, 2003), the Yammaouneh segment in Lebanon that last produced an earthquake and a tsunami in 1202 (Daëron *et al.*, 2005), and potentially all of the Jordan Valley segment if the 1546 earthquake is as small as purported to be by Ambraseys and Karcz (1992). Considering that the geologic and geodetic strain accumulation rate is on the order of 4–6 mm/yr (see up-to-date detailed analysis for different timescales, fault segments, and measuring methods in Klinger *et al.*, 2000; Daëron *et al.*, 2004; Marco *et al.*, 2005; Ambraseys, 2006), these portions of the DST have accumulated at least 3–5 m of potential slip that is most likely to be released in a large earthquake (M 7). Thus there should be great concern in Syria, Lebanon, and Israel, not only for the possible occurrence of a near-future large earthquake, but also for the significant likelihood of a tsunami produced by an offshore slump within minutes of the mainshock.

Worldwide Application. The eastern Mediterranean and the DST system stand out only for their rather long and detailed earthquake and tsunami historiography, and not as a unique active structure near a continental margin. Similar settings also exist along other anciently settled coasts of the Mediterranean, as well as in areas that have only short periods of seismic reporting. This is exemplified by the 1373 earthquake in the Central Pyrenees-Catalonia region, along with its associated tsunami in Barcelona. Similarly, the 1169 and 1456 earthquakes and tsunamis in Italy are also good examples (Guidoboni and Comastri, 2005). Farther away, along the California coast, past records are only three centuries long at the most, yet several of the recent near-field tsunamis such as 1812, 1865, and 1868 (Chowdhury *et al.*, 2005) followed nearby onshore earthquakes. From this, it is clear that generation of tsunamis by offshore slumps produced by onshore earthquakes is not limited to the eastern Mediterranean, but rather, appears to be a potential trigger for all regions with active faults near the coast.

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Appendix

The tsunami simulations here derive from linear dispersive water wave theory. Taking the origin of coordinates at a representative location in the source region, vertical tsunami motions at the sea surface at $\mathbf{r} = (x,y)$ and time t in an ocean of uniform depth h, is:

$$u_z(\mathbf{r},t) = \sum_{m=-\infty}^{m=+\infty} \int_0^\infty \frac{kdk}{2\pi \cosh(kh)} J_m(kr)e^{im\theta} \times \int_{A(t)} d\mathbf{r}_0 J_m(kr_0)\, e^{-im\theta_0} \int_0^t dt_0 \dot{u}_z^{bot}(\mathbf{r}_0,t_0) \cos(\omega(k)(t-t_0)) \quad (1)$$

In (1), the $J_m(x)$ are cylindrical Bessel functions, θ is the azimuth angle measured from x toward y, $r = |\mathbf{r}|$, k is wavenumber, $\omega(k)$ is the frequency associated with tsunami waves of wavenumber k in water of depth h, and $\omega^2(k) = gk \tanh(kh)$. Assumed to be known is $\dot{u}_z^{bot}(\mathbf{r}_0, t_0)$, the time derivative of the vertical displacement of the seafloor. $\dot{u}_z^{bot}(\mathbf{r}_0, t_0)$ integrated over the source area $A(t)$ and source duration drives the tsunami. If the time history of seafloor uplift is everywhere "ramp like", starting at $t = t_0$ and lasting for time T_R, then (1) becomes

$$u_z(x,y,t) = \sum_{m=-\infty}^{m=+\infty} \int_0^\infty \frac{kdk}{2\pi \cosh(kh)} J_m(kr)e^{im\theta} \cos(\omega(k)(t-t_0-\tau/2)) \times \frac{\sin(\omega(k)\tau/2)}{-\omega(k)T_R/2} \Bigg|_{\tau=0}^{\tau=\min(t-t_0,T_R)} \int_{\text{Area}} d\mathbf{r}_0 U(\mathbf{r}_0) J_m(kr_0) e^{-im\theta_0} \quad (2)$$

It is in the prescription of $U(\mathbf{r}_0)$ where local information comes into play. For an earthquake source, (2) is evaluated once with $U(\mathbf{r}_0)$ being the vertical component of the fault's static deformation field. Fault information like strike, dip, rake, length, width, location, and predicted slip amount are needed here. For a landslide source, (2) is evaluated several times with $U(\mathbf{r}_0)$ being "piston-like" uplifts or draw-downs over rectangular regions shifted in time and space to mimic the passing of landslide material (Fig. 5). Local details of slide volume, area, speed, and path are needed in this case. Equation (2) also can be written

$$u_z(\mathbf{r},\ t) = \sum_{m=-\infty}^{m=+\infty} \int_0^\infty \frac{k(\omega)d\omega}{2\pi u(\omega)\ \cosh(k(\omega)h)} \tag{3}$$

$$J_m\ (\omega T(\omega,\ \mathbf{r}))e^{im\theta}\ \cos(\omega(t - t_0 - \tau/2))$$

$$\times\ \frac{\sin(\omega\tau\ /\ 2)}{-\omega T_R\ /\ 2}\ \Big|_{\tau=0}^{\tau=\min(t-t_0,T_R)} \int_{\text{Area}} d\mathbf{r}_0 U(\mathbf{r}_0) J_m(k(\omega)r_0 e^{-im\theta_0}$$

with travel time $T(\omega,\mathbf{r}) = r/c(\omega)$. The $c(\omega) = \omega/k(\omega)$ and $u(\omega)$ are the tsunami phase and group velocities. In moving to a variable depth ocean, (3) becomes

$$u_z(\mathbf{r},t) = \sum_{m=-\infty}^{m=+\infty} \int_0^\infty \frac{k_c(\omega)d\omega}{2\pi u_c(\omega)\ \cosh(k_c(\omega)h_c)}$$

$$J_m\ (\omega T_p(\omega,\mathbf{r}))e^{im\theta}\ \cos(\omega(t - t_0 - \tau/2))$$

$$\left[\frac{u_c(\omega)}{u(\omega)}\right]^{1/2} G(\mathbf{r})\ \times\ \frac{\sin(\omega\tau\ /\ 2)}{-\omega T_R\ /\ 2}\ \Big|_{\tau=0}^{\tau=\min(t-t_0,T_R)} \tag{4}$$

$$\int_{\text{Area}} d\mathbf{r}_0 U(\mathbf{r}_0) J_m(k_c(\omega)r_0)e^{-im\theta_0}$$

where variables with subscript c are calculated using the water depth h_c at the origin, e.g., k_c is found from $\omega^2 = gk_c \tanh(k_c h_c)$. The principal differences between (4) and (3) are:

1. Travel time $T(\omega,\mathbf{r}) = P(r)/\bar{c}(\omega)$ is now calculated over a curved "ray path" of length P and mean phase speed $\bar{c}(\omega)$ over that path.
2. A new shoaling factor $[u_c(\omega)/u(\omega)]^{1/2}$ accounts for wave height changes due to water depth.
3. A new geometrical spreading factor $G(r) \leq 1$ takes into consideration the reduction of wave amplitudes into shadow zones.

Examples of tsunami modeling of various scenarios and sources from around the world can be found in Ward and Asphaug (2000, 2002), Ward (2001, 2002), Ward and Day (2001, 2002, 2003, 2005, 2006), and Schnellmann *et al.* (2002).

Run-up Estimates

Simple linear wave theory cannot follow the waves all the way on to land. Instead we follow Chesley and Ward (2006) and take the wave amplitude A (one-half peak to peak) in shallow water depth D ($D > A$) some distance offshore and estimate run-up height R as

$$R = A^{0.8}\ D^{0.2} \tag{5}$$

Formula (5) fits well with experiments of two-dimensional breaking and nonbreaking solitary waves over a range of conditions on a smooth planar beach. Run-up, however, being an extreme measure, is not a particularly stable quantity in three-dimensional real-world situations. Within very short distances, field run-ups can vary easily by factors of 2 or 3. Operationally, we consider run-ups from formula (5) to represent a mean value of a statistical distribution that has a standard deviation roughly equal to the mean. That is, for a quoted run-up of 2 m say, perhaps 15% of nearby locations would experience run-ups >4 m and a few percent of nearby locations would experience run-ups >6 m.

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