Bulletin of the Seismological Society of America, 92, 1, pp. 1–18, February 2002

The Seismic Activity of the Marmara Sea Region over the Last 2000 Years

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Abstract We examined the seismicity of the region of the Sea of Marmara in northwest Turkey over the last 2000 yr using the historical record of the region. We found no evidence for truly large earthquakes of a size comparable to those in the North Anatolian Fault zone in the Marmara Basin. Events are smaller in keeping with the known fault segmentation of the Basin. Over the last 2000 yr seismic moment release accounted for the known right-lateral shear velocity across the Marmara region observed by Global Positioning System measurements. Its average rate is relatively constant and varies between 1.6 and 2.4 cm/yr. The long-term seismicity in the Marmara Sea region shows that large earthquakes are less frequent than predicted from the 100-yr-long instrumental period.

Introduction

The Marmara Sea region is a densely populated and fast-developing part of Turkey, roughly bounded by 39.5° N to 41.5° N and 26° E to 31° E (Fig. 1). One of the most tectonically active on the continents, this region has been unusually active during the twentieth century with two earthquakes of M_S 7.3 and 7.4 occurring 240 km apart at its two extremities in 1912 and 1999 (Fig. 2). A question that must be addressed in any realistic assessment of the earthquake hazard in this populous area that includes Istanbul, a megacity of 11 million inhabitants, concerns the long-term seismicity of the region.

To answer this question, Ambraseys *and* Jackson (2000) examined the long-term seismicity of the region over the last 500 yr. In this article the investigation is extended to the past 2000 yr for which both instrumental and macroseismic events are appraised and historical data relating to earthquakes in previous centuries are critically studied. These studies of past earthquakes, which are still in progress, provide a practical context within which early and modern events in the region can be calibrated on a uniform basis and permit creation of yardsticks in terms of qualitative information against which early earthquakes can be classified according to their size (Ambraseys, 2002). In the process of acquiring and classifying this information, new earthquakes were found and known ones were discarded as spurious.

Background

Although scattered indications of earthquakes in the Marmara Sea region go back as far as the third century B.C., adequate documentary coverage of individual events for location and magnitude does not begin until the first century A.D. The coastal area of the Marmara Sea region was always a major trade route, and for more than 20 centuries it was the crossroads between the West and the East, a region of great importance. The Roman land route called the *Via Egnatia*, the main trade route from Rome to the East, used by the Byzantines and later by the Ottomans (Şol Qol), was built ca. 130 B.C. It crossed the Balkans to Byzantium and followed the Thracian coast of the Sea of Marmara via Thessaloniki, Enez (Enos), and Tekirdağ (Raedestos) to Istanbul. From Istanbul, the trade route ran along the north coast of the Gulf of İzmit to İzmit (Nicomedia), the main route crossing over the Gulf of İzmit at Hersek (Helenopolis) to Iznik (Nicaea). From there, one branch turned to the west, passing from Bursa to Çanakkale and the Dardanelles, thus encircling the Sea of Marmara, with the sea routes connecting most of the ports around the Sea and thus supplementing the communication of information with the interior (*e.g.,* Taeschner, 1926).

At the center of the region, Byzantium, later Constantinople, or Istanbul, was the thriving, populous capital of the Lower Roman, Byzantine, and later, Ottoman Empires, with a sustained tradition of historiography. The historical record of this region over the past 20 centuries, as provided by literary sources, is remarkably full and relatively continuous. The Roman, and in particular, the Byzantine, tradition, was necessarily waned with the Ottoman conquest about six centuries ago but was superseded by the fairly rich, but not so accessible, Ottoman historiography, supplemented by occidental sources. The historical record of the Marmara Sea region regarding information about earthquakes is rivaled only by that of a small part of northeast China. The benefit of being able to use observations over a period 20 times longer than the 100 yr of instrumental seismology is obvious.

Tectonics

The active tectonics of the Marmara Sea region in northeast Turkey is dominated by the right-lateral North Anato-

Figure 1. Location map with place names mentioned in the text; inset shows study areas: A, Marmara Sea region; B, Marmara Basin.

lian fault zone, running from Karliova in the east (41° E) to the Gulf of İzmit (30° E). In the eastern part of the Marmara Sea region, the westward motion of Turkey relative to Europe occurs mostly along the North Anatolian Fault zone. East of about longitude 30° E, the North Anatolian Fault system, which accommodates most of the westward motion of Turkey, has a narrow and localized character, clearly defined by the predominantly strike-slip surface along its entire 1000-km length, which is associated with a series of major earthquakes.

West of about 31° E, the North Anatolian Fault system branches into three subparallel strands; a northern strand passes by Sapanca, enters the Gulf of İzmit, and forms the Marmara Sea Basin, and a southern strand runs toward Lake Iznik and Bursa. Near 30° E, the southern strand splits up again into a middle and southern branch, the former passing

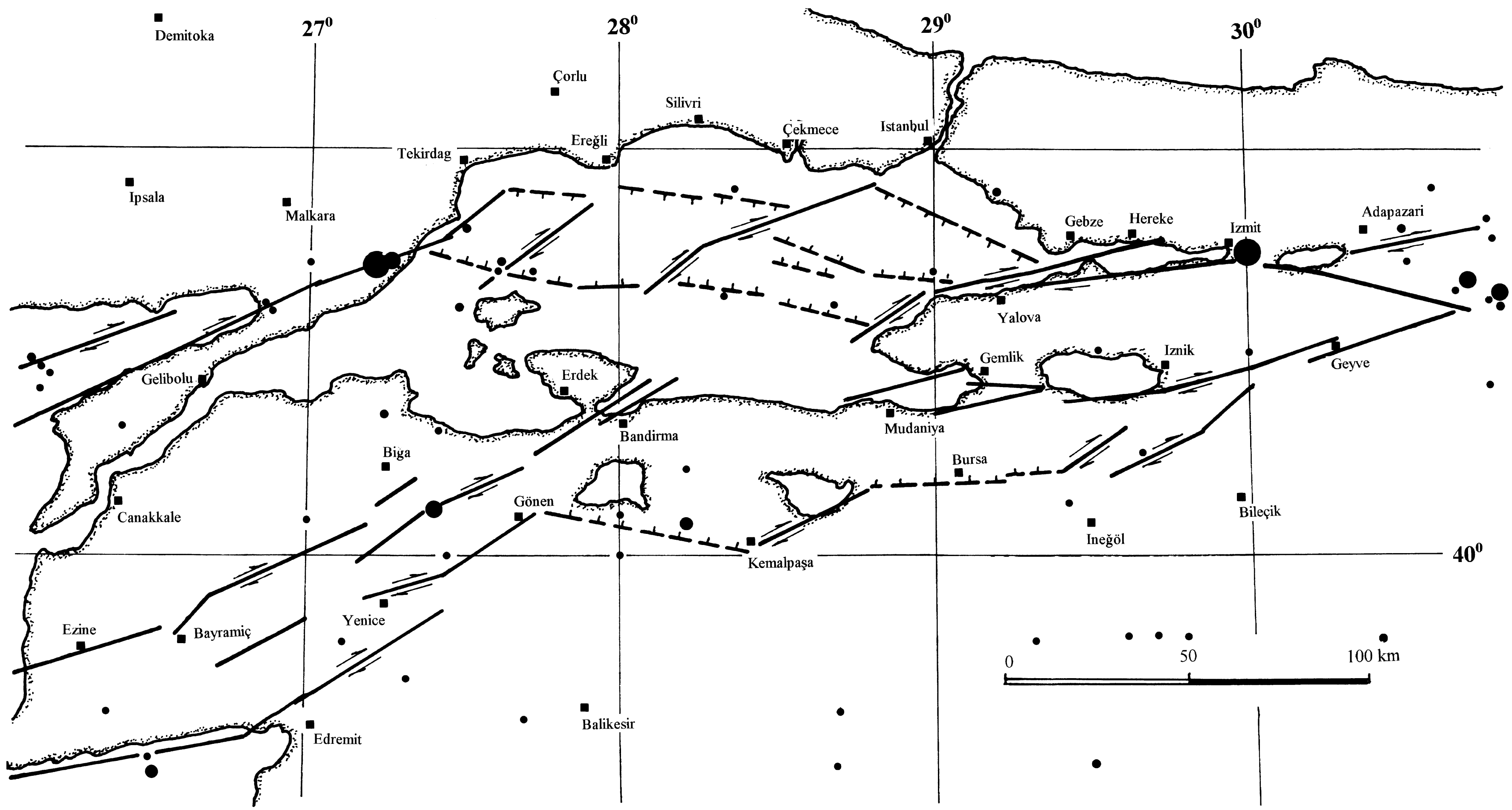

Figure 2. Summary tectonic map of the Marmara Sea region. Solid and dashed lines correspond to strike-slip and normal faults of Barka and Kadinsky-Cade (1988). Superimposed are the epicenters of earthquakes of $M_S \geq 5.0$ from 1900 to 1999.

south of Iznik and the latter south of Bursa, by the Lakes of Apolyont and Manyas into the North Aegean extensional province (Fig. 2).

The Marmara Sea Basin is about 230 km long and 70 km wide with a shallow shelf to the south and a series of subbasins to the north, namely, the Tekirdağ, Central, Çinarcik, Karamürsel, and İzmit basins. Active faulting on land in the region is relatively well documented. However, the pattern of active faulting in the Basin is much less well established. Originally the Basin was considered to be a graben or a structure of right-lateral faults exhibiting an overall normal motion (Barka and Kadinsky-Cade, 1988). Recently it was proposed that the Marmara Sea Basin was controlled by a strike-slip fault that extended between the Gulf of İzmit and the Galipoli Peninsula (Le Pichon *et al.,* 2000). However, its bathymetry, supplemented by faults identified by conventional seismic reflection surveys and focal mechanisms of a few earthquakes (Ambraseys and Jackson, 2000), shows a series of pull-apart basins bounded by a system of relatively short strike-slip and normal faults, which clearly imply significant regional extension responsible for the formation of the Marmara Sea Basin. The depths and steep bathymetric gradients of the subbasins of Tekirdağ in the west and Çinarcik in the east parts of the Basin suggest high local seismic activity (Fig. 3) (Smith *et al.,* 1995; Wong *et al.,* 1995; Parke *et al.,* 2000; Okay *et al.,* 2000).

West of the Marmara Basin (west of 27.5° E), the motion is taken up by the strike-slip Saros zone and by the extension of the crust in and around the North Aegean Sea. In the Aegean the present-day deformation is distributed over a broad region, but in the seismogenic upper crust most of the motion is apparently accomodated by a few subparallel major fault systems including the Saros zone (Taymaz *et al.,* 1991).

The Twentieth-Century Seismicity

Figure 2 depicts all events of $M_S \geq 5.0$ in the period 1900–1999, for which the locations have been reappraised and their magnitudes recalculated from amplitude/period readings of long phases and the Prague formula (Ambraseys and Douglas, 2000). The most active parts of the Marmara Basin are at the terminus of the North Anatolian Fault system in the east and at the junction of the Basin with the North Aegean in the west, regions controlled by strike-slip faults.

Compared with other parts of the North Anatolian Fault, the Marmara Sea region has been one of rather high seismicity in the twentieth century, releasing a total moment of 6.4×10^{27} dyne cm, more than half of which comes from the earthquakes of August 1912 in Ganos and August 1999 in İzmit, with the central Marmara Basin in between contributing only 8% of the total seismic moment released.

Evaluation of Long-Term Seismicity

The use of long-term information, which can best be assessed from an interdisciplinary study, gives a far fuller understanding of earthquake hazard because it is based on human experience of earthquakes over a much greater segment of the geological timescale. Ongoing research in Europe has shown that this is not an easy task (Stucchi, 1993; Albini and Moroni, 1994; Vogt, 1996) and that it requires an enormous amount of work to analyze even one earthquake; the record is held by Castenetto and Galadini's (2001) 788-page book, which describes a single earthquake.

The present study concerns chiefly the long-term seismicity of the Marmara Basin, but my group's investigation extends over the larger area of the Marmara Sea region.

Approach

The macroseismic data retrieved for the historical period have been interpreted following an improved version of the method described by Ambraseys and Melville (1982). This method includes assessment of the observed intensities, epicentral locations, and a calculation of the magnitude of the earthquakes as defined by the intensities assessed at different sites and by their associated source distances.

The source material used is far too long to present in a journal article. The full database gives the various sources that have been found to contain information and relates the record of events to specific earthquakes. For events up to 1500 and after 1800, all the source material, which is in various languages dead and alive, has been translated into English, whereas for events between 1500 and 1800, the full database includes a revised and updated version of the article by Ambraseys and Finkel (1995). The full database of information, which is still under development, is then used to assess epicentral areas, intensities, and magnitudes (Ambraseys, 2001a). The intricacies of the interpretation of the literary information in the sources have been presented elsewhere (Ambraseys, 2002) and some may be mentioned here.

Dating

With historical earthquakes, dating is not always easy. Several calendars are used to date earthquakes over the last twenty centuries, and details of the different dating systems in use may be found, for instance, in Grumel (1958). Dating errors in primary and in particular secondary sources are common, and whether each event reported was really distinct or whether multiple reports of the same event were interpreted as separate events should be examined. Failure to identify and eliminate doublets or amalgamated events, a common error in parametric catalogs, renders assessment of the epicentral area and magnitude particularly problematic.

Simultaneity

Another problem in locating historical events and assessing their size is the simultaneity of the earthquake mentioned in various places. This is not always specifically stated by the early authors, who refer to several places being shaken or damaged in a particular week, month, or year. This confusion occurs chiefly with early chroniclers who mention

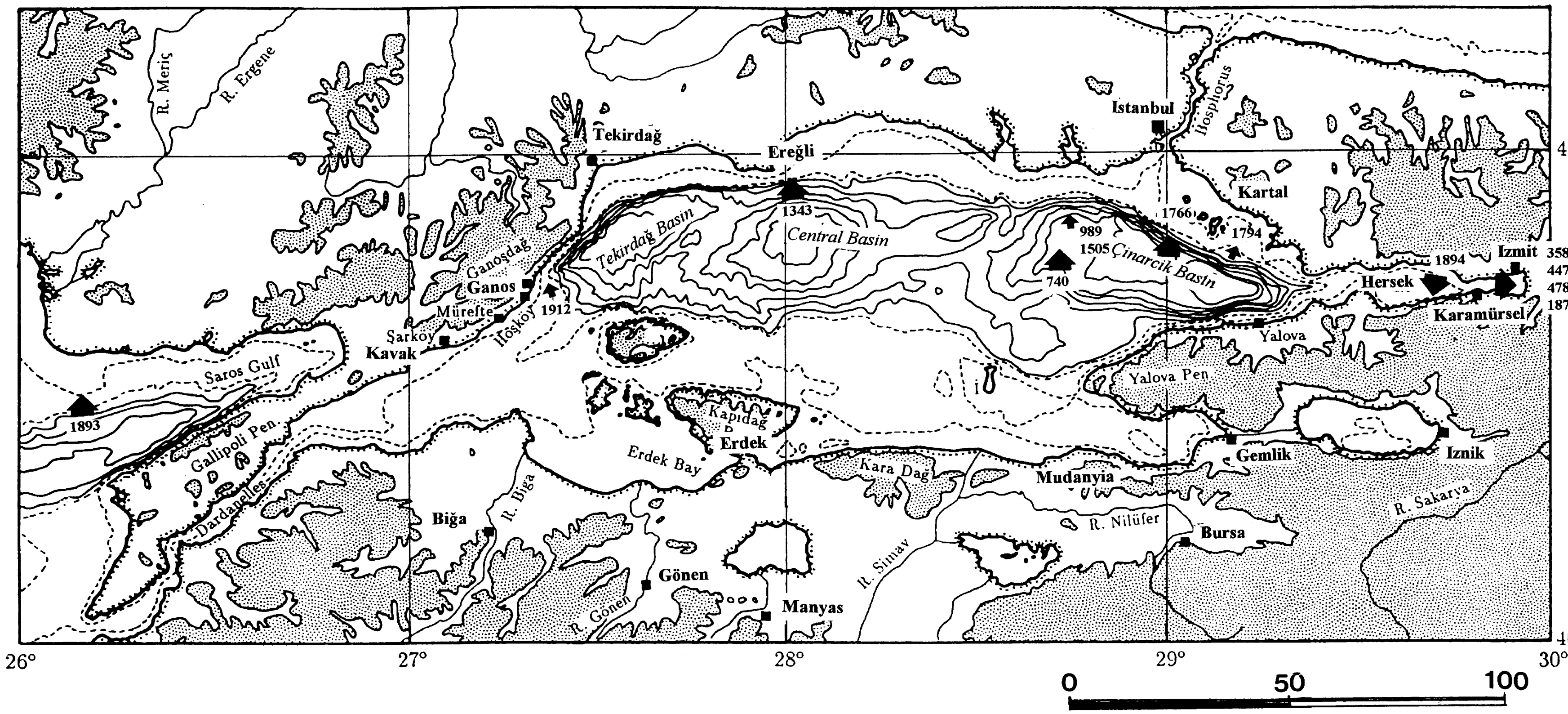

Figure 3. Bathymetric and topographic map of the Marmara Sea region. Below sea level, the bathymetry contours with intervals of 200 m are shown as solid lines; the dashed lines are the 50 m and 100 m bathymetric contours. Topography above 200 m is stippled (based on Smith *et al.,* 1995). Location of topographical features mentioned in the text: I, Izmit subbasin; K, Karamürsel subbasin. Large arrows show the year of the event, and small arrows show the probable location of damaging and felt sea wave, respectively.

all the events of a particular period in a single description. Failure of earthquake compilers to recognize lack of simultaneity often results in spurious doublets or in the amalgamation of different events, which leads to catalog entries of great earthquakes, the sizes of which are grossly exaggerated (Ambraseys and White, 1997).

Earthquake Effects

Assessment of the strength of ground shaking from earthquake effects reported in historical and modern accounts depends largely on the vulnerability of the building stock exposed to the shaking and on the way this information is transmitted in historical documents.

Until relatively recently, houses in Istanbul and in towns and villages around the Sea of Marmara were almost entirely made of wood, and as a consequence, although damaged beyond repair in a severe earthquake, they rarely collapsed and did not cause great loss of life (Refik, 1935). A general observation about the typical timber house is that its inherent strength was considerable but variable and that its vulnerability to earthquakes was rather low. In contrast, stone and brick constructions, which were relatively few in the region, can collapse with great loss of life. Further inland, houses in villages in Thrace, to a great extent of made of mud bricks and covered with flat heavy roofs, were also vulnerable. It is interesting that for Istanbul, damage and loss of life resulting from fires were always higher than from earthquakes (Cezar, 1963).

Quite often, the fact that a few historical monuments made of stone or brick are still to be found in the region in a state of tolerable preservation is interpreted as an indication that either the vulnerability of this class of structures is low or that they have been free from destructive earthquakes. However, our evidence shows that the historical structures still extant have in fact, during their lifetime, been subjected to a number of destructive earthquakes, and they have partly or wholly survived through a process of natural selection. They represent a small sample of the best final design and construction, achieved through the ages by trial-and-error techniques or by chance, and their outcome cannot and should not be used to assess seismic hazard.

Another modern class of man-made structures that seem to have little extra resistance to earthquakes is that of houses built in the last few decades with nontraditional materials, such as reinforced concrete. As the recent earthquakes have shown, in the absence of proper building codes and enforcable regulations, the introduction of new materials and methods of construction has produced a class of highly vulnerable structures, the excessive damage of which tends to inflate intensity grading.

Intensity

As a consequence of these differences in vulnerability of the building stock and its changes with time, it is difficult to assign intensity. We find that damage or collapse of aging houses and public buildings or of city walls is often an indication of the weakness of these structures, which frequently collapse without the help of earthquakes, rather than of the severity of ground shaking, which in the absence of additional corraborating evidence makes it difficult to assess intensity.

Natural exaggeration in the sources, historical and modern, is also a problem; however, it is not difficult to solve. The authenticity of the source, the style of its narrative, internal evidence in the account, and the experience gained by the assessor from processing historical information all combine to permit a realistic, albeit subjective, estimate of the degree of impact or the extent of the area over which earthquake effects were serious. For instance, statements such that after a "destructive" earthquake in Istanbul people took refuge in churches or mosques, or that they organized religious processions in the city, suggest that the shock was not in fact all that destructive. There are also cases in which a historian unintentionally exaggerated damage by reporting collectively effects from more than one earthquake or from fire and thunderstorms. Intentional exaggeration of damage, or syncretization of an earthquake with political or religious happenings, is also known. The substance of such accounts is the same, but the impressions they leave about the seriousness of the event with those who use historical data without scrutiny or at second hand are very different. These problems are discussed elsewhere in connection with the assessment of intensities in the region (Vogt, 1996; White, 2000; Ambraseys, 2002).

Statements referring to fatalities among leading citizens, large-scale reconstruction undertaken by the state, remission of taxes, new coinage, adverse effects on trade, or emigration after an earthquake all suggest a large destructive event. Some of these effects are not always recorded together with or attributed to an earthquake but can be inferred from internal evidence and from the historical context of the event.

Casualty figures are rather difficult to check and are not necessarily indicative of the magnitude or the intensity of an earthquake. Because of the emphasis on effects in major centers, reported figures depend to a large extent on population distribution and density. Even today, with fuller coverage of events, the same bias exists, although this may reflect the genuine situation: quite often the largest proportion of people killed is found in small towns and villages away from urban centers.

The assessment of intensity in this study is made on the Medvedev–Sponheuer–Karnik (MSK) scale, somewhat simplified by disregarding critieria meant for modern structures, which allows for consistent subjective observations and correlation with intensities estimated from twentieth century earthquakes to be made. This simplification is also dictated by the fact that local conditions in the Marmara Sea area, such as siting of settlements and towns, building materials, and techniques, until the early 1900s, had changed little over much of the period under review. Intensity estimates for historical earthquakes, made by the writer and by various independent assessors at different times during the last decade, are accurate only to ± 0.5 intensity unit, in the sense that two assessors assigned intensities differing by one unit and occasionally for some early events are only accurate to ± 1 unit.

Seismic Sea Waves

Little is known in any detail about the 14 seismic sea waves in the Sea of Marmara, six of which were damaging, associated with earthquakes of $M_S \geq 6.8$. It is not possible to say much about the source area of these waves except perhaps that most of them were reported from the northeast coast of the Sea of Marmara and from the Gulf of İzmit, and that the source mechanism of some of them seems to be coastal mass failures and sliding of submarine sediments (Ambraseys, 2001b). In Figure 3, we have assumed sea wave sources to be close to the inferred offshore epicentral area of the associated event.

Epicentral Region

Data provided by historical sources for earthquakes on land are adequate to permit the general location of the general epicentral area, particularly for the larger events, which show some clustering toward large centers of habitation. Early earthquakes are less well located, and it is often difficult to ascertain their true epicentral region, although for the larger events, aided by local tectonics, there is rarely great ambiguity about their general position within the Marmara Sea region. Figure 4 shows the general location of the epicentral regions of historical and modern events.

We find that, allowing for exaggeration, the observation of any damaging or destructive event in the region reflects its seriousness and significance. For the early period, while it is certain that many small-magnitude shocks must be missing from the record, we can reasonably assume that those of which damage details survive were important events.

Study of twentieth-century earthquakes is additionally valuable because it permits correlations to be made between the effects of historical earthquakes and those of the twentieth century. This is possible despite shortcomings in the documentary evidence, for one can, in many cases, determine from internal or contextual evidence such details as the size of the area affected, athe extent and type of damage caused, the duration of aftershocks, and the association of the event with ground deformations, such as surface faulting, landslides, and proximity to known active faults. All such details, ideally found in or inferred from historical sources, when calibrated against similar information derived from the study of modern events, permit assessment of the different intensities experienced throughout the affected area and estimation of the magnitude of the historical earthquake.

For earthquakes that occurred before the 1920s, locations are the centers of the macroseismic regions as defined by the higher intensities. For events with epicenters in the Sea of Marmara, we have attempted to associate earthquakes with active faults. This is often determined based on knowledge of the general location and magnitude of the event and on the nearest known fault in the Marmara Basin that could accomodate the associated seismic moment release.

Classification of Events

All events were classified according to the location of their epicentral region, that is, on land, offshore, or not known, and their quality, that is, according to whether they were well located, less well defined, adopted, or instrumental. They are also classified according to the number of sites used to estimate their magnitude. Depending on maximum effects, earthquakes with epicentral region on land or offshore were classed into those that caused considerable damage, heavy damage, or destruction and social or economic effects resulting in extensive reconstruction. These events are ranked again in terms of the amount of casualties they caused, that is, small, significant, or great. Further classification was made according to whether an event exhibited a small or large area of perceptibility and caused localized or widespread damage. Events accompanied by a seismic sea wave, surface faulting, and ground failures are also flagged.

Table 1 shows the classification of events in 10 fields, the reason for which has at least three purposes: (1) to inspect all events regardless of when and where they happened; (2) to compare the location and size of the events; and (3) to test the completeness of the information. Only in this way might we succeed in getting an overall qualitative evaluation of the historical data before we use them in a final quantitative analysis.

Magnitude

The surface-wave magnitude M_S of pre-1898 earthquakes was estimated from the intensity and epicentral distance using the regional equation

$$\begin{aligned} M_{Si} = {} & -1.54 + 0.65(\mathrm{I_i}) + 0.0029(\mathrm{r_i}) \\ & + 2.14\log(\mathrm{r_i}) + 0.32\mathrm{p} \end{aligned} \tag{1}$$

where M_{Si} is the uniformly recalculated surface-wave magnitude from the Prague formula, in which I_i ($\leq$VIII MSK) is the intensity at a distance r_i (in km) from the seismic source, and p is 0 for mean value and 1 for 84 percentile. Equation (1) was derived from earthquakes in Greece and Western Turkey and is valid for far-field conditions (Ambraseys, 1992). In the near field, source distance is defined as the closest distance to the surface fault rupture, and in the far field, it is defined by the epicentral distance. Site magnitudes M_{Si} were calculated from equation (1), the average value of which is the event magnitude M_S, which, together with the number of station magnitudes used to assess it, is listed in Table 1.

Intensity estimates for historical earthquakes in this study are accurate only to ± 0.5 intensity units at best, and from equation (1) to an uncertainty in M_S of ± 0.35 magnitude units, but for events reported from very few places this uncertainty can be larger.

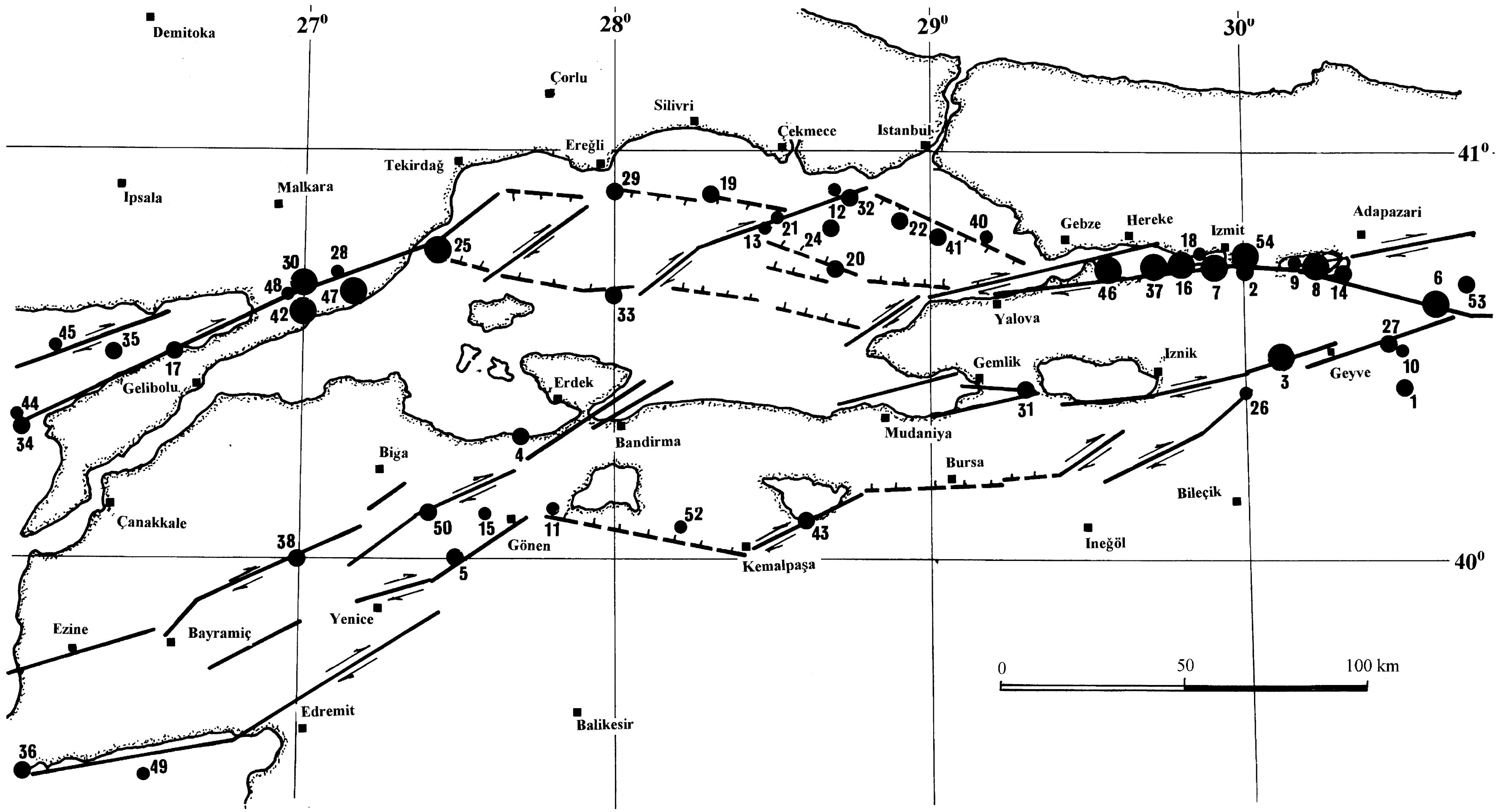

Figure 4. Map showing epicenters of earthquakes after 1899 and location of epicentral regions of historical events of $M_S \geq 6.8$ for the period 1 to 1899. Locations at sea are inferred from macroseismic observations and fault positions. Earthquakes of M_S between 6.8 and 7.4 will have ruptured faults 30 to more than 100 km in length. Numbers refer to entries in Table 1. Solid and dashed lines indicate strike-slip and normal faults, respectively, of Barka and Kadinsky-Cade (1988). Solid circles are graded for $6.8 \leq M_S < 7.0$, $7.0 \leq M_S < 7.3$, and $M_S > 7.3$.

Table 1

Large ($M_S \geq 6.8$) Earthquake in the Marmara Sea Region over the Last 20 Centuries*

									General Effects†										
	Y	M	D	OT	N°	E°	M_S‡	M_0§	1	2	3	4	5	6	7	8	9	10	Region
01	0032	00	00	0000	40.5	30.5	7.0	04.37	1	4	02	2	2	0	2	2	0	0	Nicaea
02	0068	00	00	0000	40.7	30.0	7.2	08.71	1	4	02	2	2	0	2	2	0	0	Nicaea
03	0121	00	00	0000	40.5	30.1	7.4	17.38	1	3	04	3	3	2	2	0	0	0	Nicomedia
04	0123	11	10	2400	40.3	27.7	7.0	04.37	1	3	01	2	2	0	0	0	0	0	Cyzicus
05	0160	00	00	0000	40.0	27.5	7.1	06.17	1	3	06	3	2	0	2	2	2	0	Hellespont
06	0180	05	03	0000	40.6	30.6	7.3	12.30	1	4	02	2	3	0	2	0	0	0	Nicomedia
07	0268	00	00	0000	40.7	29.9	7.3	12.30	1	3	03	3	3	0	2	0	0	0	Nicomedia
08	0358	08	24	0600	40.7	30.2	7.4	17.38	1	3	06	3	3	3	2	2	2	2	Izmit
09	0362	12	02	0000	40.7	30.2	6.8	02.19	1	4	03	2	2	0	1	1	0	0	Izmit
10	0368	10	11	0000	40.5	30.5	6.8	02.19	1	4	02	2	2	0	2	1	0	0	Persis
11	0368	11	00	0000	40.1	27.8	6.8	02.19	1	4	01	2	2	0	0	0	0	0	Germe
12	0407	04	01	1900	40.9	28.7	6.8	02.19	2	4	02	2	1	0	1	1	0	0	Hebdomon
13	0437	09	25	0000	40.8	28.5	6.8	02.19	3	4	01	2	1	1	1	1	0	0	Istanbul
14	0447	11	06	2000	40.7	30.3	7.2	08.71	1	3	05	3	3	3	2	2	2	2	Nicomedia
15	0460	00	00	0000	40.1	27.6	6.9	03.09	1	4	03	4	2	2	2	2	2	0	Cyzicus
16	0478	09	25	0000	40.7	29.8	7.3	12.30	2	3	05	3	3	3	2	2	0	1	Helenopolis
17	0484	00	00	0000	40.5	26.6	7.2	08.71	1	3	08	3	3	2	2	2	2	0	Callipolis
18	0554	08	16	0600	40.7	29.8	6.9	03.09	2	4	04	2	2	2	2	1	2	0	Nicomedia
19	0557	12	14	2400	40.9	28.3	6.9	03.09	2	4	03	2	2	2	1	1	0	0	Silivri
20	0740	10	26	1400	40.7	28.7	7.1	06.17	3	4	05	2	3	2	2	2	0	2	Marmara
00	0823	10	00	0000	00.0	00.0	0.0	00.00	0	0	00	0	0	0	0	0	0	0	Panium
21	0860	05	23	0000	40.8	28.5	6.8	02.19	3	4	02	2	1	1	2	1	0	0	Marmara
22	0869	01	09	0000	40.8	29.0	7.0	04.37	2	4	01	2	2	1	2	2	0	0	CP
23	0967	09	02	0500	40.7	31.5	7.2	08.71	1	3	04	3	3	3	2	2	0	0	Bolu
24	0989	10	25	1900	40.8	28.7	7.2	08.71	3	4	03	2	2	2	2	0	0	1	Marmara
25	1063	09	23	2200	40.8	27.4	7.4	17.38	1	3	05	3	3	3	2	2	0	0	Panio
26	1065	09	00	0000	40.4	30.0	6.8	02.19	1	4	02	2	1	0	1	1	0	0	Nicaea
27	1296	06	01	2400	40.5	30.5	7.0	04.37	1	4	02	2	2	1	2	2	0	0	Bithynia
28	1343	10	18	1200	40.7	27.1	6.9	03.09	1	3	06	3	2	2	1	1	0	0	Ganos
29	1343	10	18	2100	40.9	28.0	7.0	04.37	2	4	03	2	2	1	2	0	0	1	Heraclea
30	1354	03	01	2000	40.7	27.0	7.4	17.38	1	2	07	3	3	3	2	2	1	0	Hexamili
31	1419	03	15	0000	40.4	29.3	7.2	08.71	1	4	03	2	3	2	2	0	2	0	Bursa
00	1489	01	16	0600	00.0	00.0	0.0	00.00	0	0	00	0	0	0	0	0	0	0	Saros?
32	1509	09	10	2200	40.9	28.7	7.2	12.30	2	2	15	2	2	3	1	2	2	2	CP
33	1556	05	10	2400	40.6	28.0	7.1	06.17	1	3	03	3	3	2	0	2	0	0	Gonen
34	1625	05	18	2400	40.3	26.0	7.1	06.17	3	4	05	3	0	0	0	2	0	0	Saros
35	1659	02	17	1900	40.5	26.4	7.2	08.71	2	4	05	2	0	0	0	2	0	0	Saros
36	1672	02	14	0000	39.5	26.0	7.0	04.37	2	4	03	2	2	0	0	2	0	0	Biga
37	1719	05	25	1200	40.7	29.8	7.4	17.38	1	2	17	3	3	3	2	2	2	0	Izmit
38	1737	03	06	0730	40.0	27.0	7.0	04.37	1	3	19	3	3	0	1	2	2	0	Biga
39	1752	07	29	1800	41.5	26.7	6.8	02.19	1	3	17	3	2	2	2	2	2	0	Edirne
40	1754	09	02	2130	40.8	29.2	6.8	02.19	2	3	09	3	2	2	2	2	0	2	Izmit
41	1766	05	22	0500	40.8	29.0	7.1	06.17	2	3	16	2	3	2	2	2	0	1	Marmara
42	1766	08	05	0530	40.6	27.0	7.4	17.38	1	2	20	3	3	3	2	2	2	0	Gonas
43	1855	02	28	0230	40.1	28.6	7.1	06.17	1	2	24	3	3	1	1	2	2	0	Bursa
44	1859	08	21	1130	40.3	26.1	6.8	02.19	2	3	25	3	1	2	1	2	2	2	Saros
45	1893	02	09	1716	40.5	26.2	6.9	03.09	2	3	31	3	2	1	1	2	0	1	Saros
46	1894	07	10	1224	40.7	29.6	7.3	12.30	2	2	81	1	3	3	2	2	2	2	Izmit
47	1912	08	09	0128	40.7	27.2	7.3	12.30	1	2	99	1	2	2	2	2	1	2	Ganos
48	1912	09	13	2331	40.7	27.0	6.8	02.19	1	3	32	1	2	1	1	2	2	0	Ganos
49	1944	10	06	0234	39.5	26.5	6.8	02.19	2	2	67	1	1	1	1	1	2	0	Edremit
50	1953	03	18	1906	40.1	27.4	7.1	06.17	1	2	45	1	2	2	2	2	1	0	Gonen
51	1957	05	26	0633	40.7	31.0	7.1	06.17	1	2	81	1	2	2	2	2	1	0	Abant
52	1964	10	06	1431	40.1	28.2	6.8	02.19	1	2	70	1	1	1	1	1	1	0	Manyas
53	1967	07	22	1657	40.7	30.7	7.2	08.71	1	2	99	1	3	2	2	2	1	0	Mudurnu
54	1999	08	17	0001	40.7	30.0	7.4	17.38	1	1	00	1	3	3	2	2	1	1	Izmit
55	1999	11	12	0000	40.8	31.2	7.1	06.17	1	1	00	1	2	2	2	1	1	0	Duzce

*Further, ongoing work may result in an upgrading of some of the entries in Table 1.

†**1,** Location: 1, on land; 2, offshore; 3, at sea. **2,** Epicentral region: 1, instrumental; 2, well-defined macroseismic; 3, less well defined; 4, adopted. **3,** Number of sites used. **4,** Magnitude: 1, instrumental; 2, macroseismic $M_S \pm 0.5$; macroseismic $\pm$ 0.35. **5,** Maximum effects: 1, considerable damage; 2, heavy damage; 3, destructive, extensive reconstruction, with social and economic repercussions. **6,** Loss of life: 1, small; 2, significant; 3, great. **7,** Extent of damage: 1, local; 2, widespread. **8,** Felt area: 1, small; 2, large. **9,** Ground effects: 1, surface faulting; 2, ground failures and landslides. **10,** Seismic sea waves: 1, damaging; 2, observed.

‡Moderate events $6.8 \leq M_S < 7.0$; large $7.0 \leq M_S < 7.4$; major $M_S \geq 7.4$;

§Moments are given in units of 10^{26} dyne cm, calculated from Ekström and Dziewonski's (1988) global $M_S - M_0$ relation.

Seismic Moment and Moment Magnitude

Relations between surface-wave magnitude M_S and seismic moment M_0, and vice versa, provide suitable functions for the correlation between one source size indicator and the other. Current scaling $\log(M_0) - M_S$ laws for shallow earthquakes have been derived from global or large subglobal datasets for active regions and for stable continental regions. Ekström and Dziewonski (1988) derived global average scaling $\log(M_0) - M_S$ laws in which the independent variable is $\log M_0$, which quite often are assumed to be orthogonal although they are not. These are

$$M_S = -19.24 + \log M_0 \quad \text{for} \quad \log M_0 < 24.5, \tag{2a}$$

$$M_S = -19.24 + \log M_0 - 0.088\,(\log M_0 - a)^2 \quad \text{for } 24.5 \le \log M_0 \le 26.4, \tag{2b}$$

$$M_S = -10.76 + (2/3)\log M_0 \quad \text{for} \quad \log M_0 > 26.4, \tag{2c}$$

These authors then rewrite equations (2) in the form

$$\log M_0 = 19.24 + M_S \quad \text{for} \quad M_S < 5.3, \tag{3a}$$

$$\log M_0 = 30.20 - [92.45 - 11.40\,M_S]^{0.5} \quad \text{for} \quad 5.3 \le M_S \le 6.8, \tag{3b}$$

$$\log M_0 = 16.14 + 1.5M_S \quad \text{for} \quad M_S > 6.8. \tag{3c}$$

However, since equations (2) and (3) are rewritten, formally, they are not the correct relationships for estimating $\log M_0$ from M_S. Regional bias in M_0 does exist, and global average scaling $\log(M_0) - M_S$ laws, such from equation (3), may be inappropriate. To overcome the problem with equations (3) that have been derived by fitting the data with $\log(M_0)$ as the independent variable, and at the same time to take into account regional bias, we derived the following set of bilinear relationships:

$$\log M_0 = 19.08 + M_S \quad \text{for} \quad M_S < 6.0. \tag{4a}$$

and

$$\log M_0 = 19.08 + M_S \quad \text{for} \quad M_S < 6.0 \tag{4b}$$

for the Eastern Mediterranean and the Middle East region up to 70° E, using Harvard centroid moment tensor (CMT) or *P*/*SH* moments from special studies and the corresponding reassessed M_S values from the Prague formula of 577 shallow ($h < 40$ km) earthquakes in the $\log M_0$ range 22.4–27.3; in the regression M_S is the independent variable (Ambraseys and Douglas, 2000). Seismic moments estimated from equation (4) are by a factor 0.7 and 0.85 smaller than from the global relations (3) for $M < 6.0$ and $M > 6$ respectively.

Seismic energy magnitude M_W, moment magnitude **M**, and surface-wave magnitude M_S are defined as a linear transformation of the logarithm of the seismic moment M_0 given by

$$\mathbf{M} = M_S = M_W = (2/3)\log(M_0) - 10.73, \tag{5}$$

in which M_0 is in dyne cm (10^{-7} N m) (Kanamori, 1977), **M** being nothing more than a definition or a transformation of M_0 through equation (5). What is often overlooked is that equation (5) is valid only for events that rupture the entire thickness of the seismogenic zone and that the validity of equation (5) holds only for M_S values larger than about 6.5 (Ekström and Dziewonski, 1988). For smaller magnitudes $\mathbf{M} \neq M_S$. Seismic moments in Table 1 are either Harvard CMT estimates for events after 1977 or values calculated from equation (1) and from Ekström and Dziewonski's (1988) global scaling $\log(M_0) - M_S$ law (??)(3c) for events before 1977.

Discussion

Historical studies should aim to be indicative and to expose points for further analytical clarification rather than prescriptive, since in fact the prescriptions have to be based on rather arbitrary assumptions. It is not sufficient, therefore, merely to lay hold on a few destructive historical earthquakes and use them to model seismicity (Finkel, 2000). It is necessary in addition to develop an intimate knowledge of all aspects of the subject.

The main factors seen to influence the survival of historical data are the quality of the contemporary literary record, the prevailing historical circumstances, the geographical location of events and their magnitude. The total number of earthquakes identified in the Marmara Sea area during the last 20 centuries is 581, of which 408 occurred in the preinstrumental period, before 1900, and 173 in the twentieth century. Of the 408 earthquakes in the preinstrumental period, 301 are small events that caused no damage or great concern, the magnitude of which was not assessed. For the remaing 107 larger earthquakes, magnitudes were assessed from equation (1) with values ranging between 5.0 and 7.4. The magnitude of all 173 earthquakes of the instrumental period, after 1899, were calculated from the Prague formula; these ranged between 4.0 and 7.4. Table 1 lists 54 events from the dataset of $M_S \geq 6.8$, together with their seismic moments, which for earthquakes before 1977 were calculated from equation (3). Table 1 shows also the classification of these events.

For historical earthquakes at sea, locations are necessarily approximate and do not warrant a sophisticated procedure for their determination. Bearing in mind that earthquakes of M_S between 6.8 and 7.4 will have ruptured faults 30 km to more than 100 km in length, locations at sea are necessarily approximate but adequate indicating nothing more than the general location of an event. Indeed, a too sophisticated location procedure carries with it the danger that its weaknesses and assumptions may not be appreciated.

Conversely, a too simple method may be discredited just because it exposes the underlying assumptions too clearly.

Macroseismic Magnitude

For shallow earthquakes of about $M_S < 6.4$, which have source dimensions sufficiently small, there is little difference between source distance and epicentral distance of a site, certainly not larger than the uncertainty in the determination of a twentieth-century epicenter. As the size of an earthquake increases, epicentral distance in the near field loses its meaning as a measure of the proximity of a site to the source, and distance r in equation (1), which has been derived for regional conditions, must be measured from the fault trace.

The use of magnitude calibration relations derived for other regions, using epicentral distance, has an important effect on the assessment of hazard in the Marmara Sea area. As an example magnitudes for some of the historical earthquakes in the Marmara Sea region have been calculated by Parsons *et al.* (2000) from

$$M_I = 1.96 + 0.60(I) + 0.012(d), \tag{6}$$

a calibration formula derived by Bakun and Wentworth (1997) from 22 Californian earthquakes in the magnitude range $4.4 \le M \le 6.9$, in which d is the epicentral distance in km. For $M \ge 6.0$, Parsons *et al.* (2000) assumed that $M_I = \mathbf{M}$, and that for $M_H < 6.0$, M_H is the local magnitude M_L in Hubert-Ferrari *et al.* (2000); d is the median epicentral distance for a particular intensity (I). Equation (6) cannot be compared directly with equation (1) because these two relationships differ in the definitions of distance, magnitude, and regional bias. The only observation that can be made from Figure 5, which shows a plot of equations (1) and (6), is that the latter gives larger magnitudes.

Table 2 lists the values of M_I calculated by Parsons *et al.* (2000) from equation (6) for 13 historical earthquakes in the Marmara Sea region, together with M_H estimates made by Hubert-Ferrari *et al.* (2000) from assumed fault lengths that are judgmental in nature, and the scaling relation of Kanamori and Anderson (1975). In the absence of other relevant information, the length of an active fault is the best guide to the maximum earthquake that might occur along it, although any such guide is a gross approximation at best, particularly with offshore faults for which any assumption regarding their location and length is little more than an arbitrary judgment. For comparison, Table 2 also shows M_S calculated in this study from equation (1). These three datasets are plotted in Figure 6 from which, considering the diverse assumptions made in their derivation in terms of M_S and the small number of observations, the large differences, particularly for $M_S < 7.0$, are to be expected.

Completeness

Much statistical ingenuity has been spent on devising techniques for tackling this problem, but there are doubts about how useful and how well-founded some of these techniques really are, and the statistician here should take a background role. He or she can point to features in the data that look anomalous because they depart from some standard model, particularly when large events are clustered in time. Whether the anomalies are to be ascribed to peculiarities of the tectonic process, or to peculiarities of the process by which the events were recorded in history, is not a question the statistician should be asked to answer. It should be referred back to the historian or the geophysicist. With his-

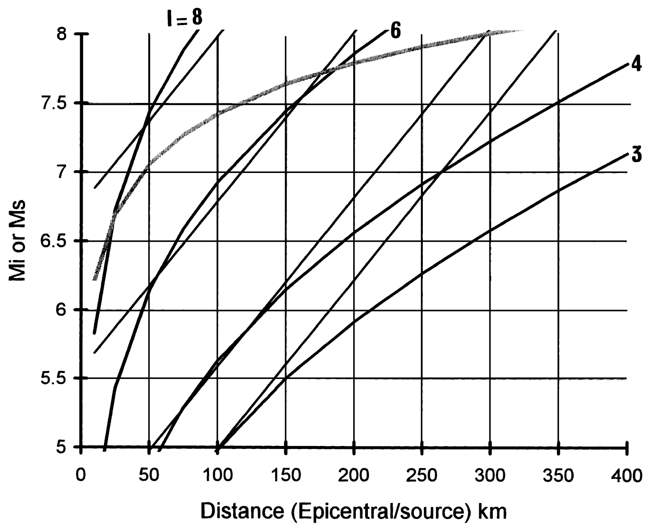

Figure 5. Plot of magnitude M_S from equations (1) shown by thick lines, and M_I from equation (6) drawn with thin lines, for intensities III, IV, VI, and VIII. Site distances for equation (1) are epicentral below the graycurve and source distances above it. For equation (6) source distances are epicentral.

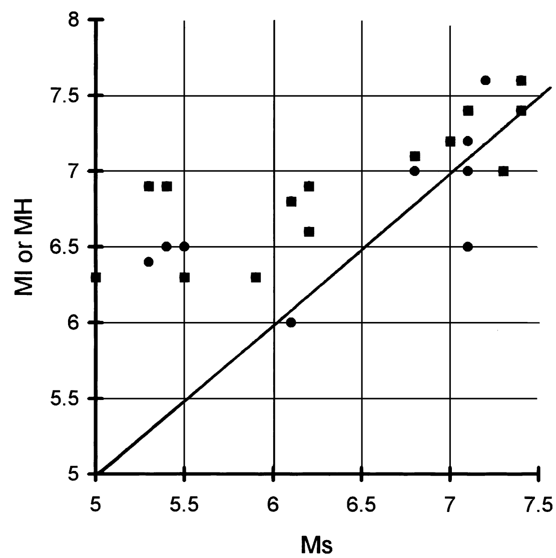

Figure 6. Comparison of M estimated by Parsons *et al.* (2000) (M_I solid circles) and by Hubert-Ferrari *et al.* (2000) (M_H solid squares) with M_S in this study.

Table 2
Comparison of M estimates

Date	M_I*	M_H†	M_S‡
09 September 1509	7.6	–	7.2
10 May 1556	6.5–7.0	–	7.1
25 May 1719	7.6	7.6	7.4
06 March 1737	–	7.2	7.0
02 September 1754	7.0	7.1	6.8
22 May 1766	7.2	7.4	7.1
05 August 1766	7.6	7.4	7.4
29 May 1776	–	6.3	*
08 February 1826	–	6.9	6.2
19 April 1850	6.0–7.0	6.8	6.1
28 February 1855	7.0–7.5	7.4	7.1
11 April 1855	–	6.6	6.2
17 September 1857	6.5–7.0	6.9	5.4
06 November 1863	6.4–7.0	6.9	5.3
13 October 1877	6.0–7.0	6.3	5.5
19 April 1878	–	6.3	5.9
26 October 1889	–	7.0	**
10 July 1894	7.0	7.0	7.3
09 August 1912	7.4	–	7.3
18 March 1953	7.2	–	7.1
06 October 1964	6.9	–	6.9
17 August 1999	7.4	–	7.4

*M_I, Magnitude estimated by Parsons *et al.* (2000).
†M_H, Magnitude estimated by Hubert-Ferrari *et al.* (2000).
‡M_S, This study.
* very small event, ** spurious event

torical data, if an effect is really present it should not take a statistician to bring it out (Vere-Jones, 1987).

Turning to Table 1, it is unlikely that all small and perhaps even moderate historical earthquakes in the interior, away from the coast of the Sea of Marmara, before the middle of the eighteenth century, would have been recorded. It is very probable, however, that many moderate and any large earthquake have been noted, although not necessarily fully identified. After the middle of the eighteenth century, documentary information improved rapidly, further supplemented by occidental sources (consular correspondence, travelers, and the press), so that after the mid-eighteenth century, data about moderate and large magnitude events should be complete. It is reasonable to suppose, therefore, that the available data for the whole region is almost complete only for moderate earthquakes or greater.

We can devise no formal method to test completeness of the data other than by testing their implications. Formal statistical tests are as valid as the distributional assumptions on which they are based. Since these assumptions are rarely likely to be satisfied by historical data, such tests may best be regarded as indicators to the extent to which a particular conclusion would be supported, or not, by the data, if in fact the assumptions were justified, and hence, of the extent to which that conclusion is likely to remain valid despite departure from those assumptions (Lomnitz, 1994).

Contribution of Small Events to the Total Moment

Relatively complete datasets of long-term seismicity in active regions are necessarily restricted to magnitudes greater than 6.8 or 7.0. The contribution of smaller magnitudes to the total moment depends on the piecewise linear frequency distribution of the data and also on the choice of the scaling $\log(M_0) - M_S$ law (e.g., Molnar, 1979; Ambraseys and Sarma, 1999). It is obvious, therefore, that the total moment release and slip rates calculated from such datasets, without consideration of the contribution of smaller magnitudes, will be underestimated by a factor of 1.5 to 2.0.

The contribution of smaller events to the total seismic moment alters the view about regions of high aseismic creep and fits better with direct measurements of velocities derived from geodetic methods. The possibility of aseismic slip in the Marmara Basin is an important question, and solutions to the problem of seismic versus aseismic slip might well come from geodetic observations.

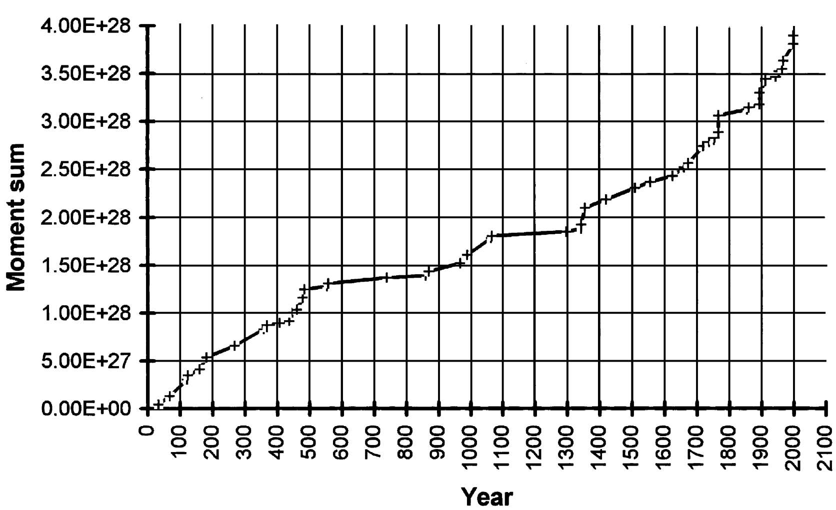

Figure 7. Cumulative seismic moment release in the 420-km-long Marmara Sea region (A) as a function of time for the events listed in Table 1. Total seismic moment released: 3.9 $\times$ 10^{28} dyne cm.

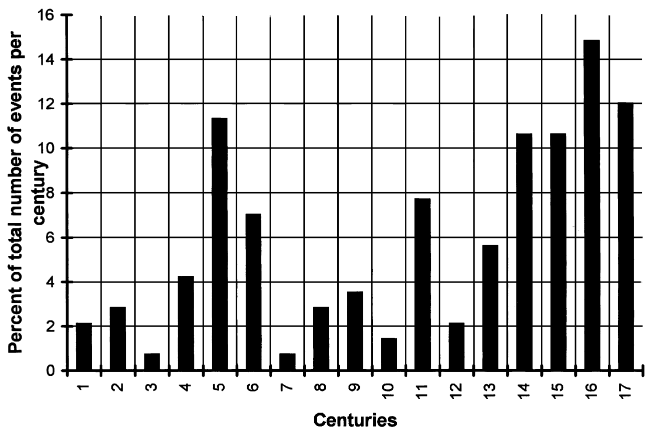

Figure 8. Percentage of the total number of earthquakes per century in the Marmara Sea region in the last 1700 yr.

Seismic Slip Rates

There are several features in the time distribution of earthquakes in the Marmara Sea region that would seem to require discussion. These features show up clearly when presented as a plot of cumulative moment versus time, shown in Figure 7. The most obvious is perhaps the concentration of large events during the fifth century and to a lesser extent in the tenth, fourteenth, and twentieth centuries, as well as the relative seismic quiescence in between these periods that may be thought to be due to lack of information rather than a genuine lack of earthquakes. But from Figure 8, which shows the volume of historical sources, we notice that there is no lack of information about earthquakes in the quiescent periods, and it is rather unlikely that large earthquakes over these relatively well-documented periods have been missed. It may seem unlikely that such features could result from a sudden excess or loss of enthusiasm among local historians during these periods to record earthquakes.

In order to investigate the effect of the size of the study area on clustering, we reduced the size of the study region by taking a 4 times smaller area (e.g., region B in Fig. 1). We find that this removes clustering, and as Figure 9 shows, the rate of moment release over the whole of the last 2000 yr becomes almost constant. Next we estimated the average right-lateral shear velocity through regions A and B by summing the scalar seismic moments of the earthquakes of $M_S \geq 6.8$, using a seismogenic thickness of 12.5 km (10–15 km) and, assuming a value for the rigidity of $\mu = 3.0 \times 10^{11}$ dyne cm^{-2}, each event contributing to this motion. Some events in the Marmara Basin do have normal faulting mechanism, in keeping with the generally short normal fault segments, but we do not think that in these calculations the assumption that all events are strike slip is an important source of error. The main source of uncertainty in the shear velocity is likely to come from the uncertainty in M_0, which is caused by its assessment from global (equation 3c) or from the regional $M_S - M_0$ formulae (equation 4b), and also by the uncertainties in M_S. We therefore estimate the uncertainty in velocity from the uncertainty in observed surface-wave magnitude. Because of the possibility of systematic bias, we do not calculate standard errors, but simply extreme maximum and minimum possible values allowing for all the magnitudes to be underestimated or overrestimated by 0.3 magnitude units and converted into M_0 through equations (3c) or (4b) (see Table 3).

We find that for the Marmara Sea region (A in Fig. 1), which is 420 km long, allowing for missing earthquakes of $M_S < 6.8$ that increase the total moment by a factor of 1.7, earthquakes of the last 2000 yr account for right-lateral velocity that varies with time between 1.5 and 2.5 cm/yr with a mean value of 2.0 ± 0.25 cm/yr, which is shown in Figures 10 and 11. These values are very close to those estimated from Global Positioning System (GPS) measurements by Straub *et al.* (1996) and by Reilinger *et al.* (1997), which are between 2.2 ± 0.3 and 2.6 ± 0.3 cm/yr and correspond to the elastic strain to be accounted for by future earthquakes and aseismic creep (Walcott, 1984).

For the smaller Marmara Basin (B in Fig. 1), which is half as long and half as wide, we find almost identical velocities of 1.9 ± 0.20 cm/yr on average. The faster rates in the first centuries shown in these figures are partly due to the short time intervals used to calculate the ratio at the beginning of the plot. Very similar rates are obtained from the global and regional scaling $\log(M_0) - M_S$ laws (equations 3 and 4), which differ little for $M_s < 6.5$.

Given the uncertainties in (a) the original M_s values, (b) the unaccounted for seismic moment from subevents of large multiple earthquakes, (c) depth, and (d) the $\log(M_0) - M_s$ scaling law, it is difficult to estimate more realistic velocities obtained from seismicity. We consider it likely that the global relation, equation (3), yields an upper bound to the seismic moment release and that the regional relation, equation (4), gives a more realistic estimate. We may then conclude that the major portion, perhaps effectively all of the motion in this region during the last 2000 yr, is probably

Table 3
Average Velocities (cm/yr) Calculated for Earthquakes $M_S \geq 6.8$ in the period 1–1999

	Marmara Sea Region			Marmara Basin		
	Mininum	Probable	Maximun	Mininum	Probable	Maximum
Global (equation 3)	0.73	2.03	5.73	0.67	1.89	5.32
Regional (equation 4)	0.61	1.73	4.88	0.57	1.60	4.53

achieved by seismic slip on faults, and that aseismic creep, is relatively unimportant.

Frequency–Magnitude Distribution

It is usually agreed that regional seismicity is well described by the Gutenberg cumulative frequency–magnitude relation

$$\log(N/ya) = a - bM_{\mathrm{s}}, \tag{7}$$

in which (N/ya) is the annual number of earthquakes of magnitude equal to or greater than M_{s} per unit area (a). However, whether the same distribution is the proper description of seismicity over a long period of time along a fault zone seems to remain a question.

Using our 100-yr-long dataset, which is more complete for $M_{\mathrm{s}} > 4.5$ than the 2000-yr-long dataset, we examined the frequency–magnitude distributions of earthquakes in the Marmara Sea region (A) and separately in the Marmara Basin (B), which are shown in Figure 12. For $M_{\mathrm{s}} < 6.5$, the results are very similar and follow equation (7) with a b-value of about 0.6; but for larger magnitudes they diverge from equation (7) and show a bilinear distribution, a change in slope that is not well defined because the period of observation is too short to provide adequate statistics for earthquakes of $M_{\mathrm{s}} > 6.5$.

If we extend the period of observations from 100 to 2000 yr we notice that, for both regions A and B, the distribution of events at large magnitudes becomes smoothly asymptotic. Also from Figure 13, which is a detail of Figure 12 for $M_{\mathrm{s}} \geq 6.8$, we notice that in the time between large earthquakes, the smaller Marmara Basin (B) is generally quiescent but for the occurrence of a low-level background activity and that the distribution follows what one might call a "characteristic earthquake" model (Wesnousky *et al.,* 1983). Figures 12 and 13 suggest a genuine departure of the observed data from Gutenberg's equation (7) for large magnitudes and an asymptotic behavior of b at the upper end of the recurrence curve where it steepens. This implies that the true nature of the distribution is hampered not only because the twentieth-century record is too short but also because the test area is too small to define the repeat time of large earthquakes and that the characteristic model could be an artifact of incompleteness of data in space and time. This also implies that the shape of the frequency–magnitude distribution for short-term observations cannot be defined at larger magnitudes. In our case the implication of this is that large earthquakes in the Marmara region are less frequent when predicted from the long-term dataset than from the usual 100-yr instrumental period, making the notion of a recurrence time, in its usual definition, questionable. Preliminary results from

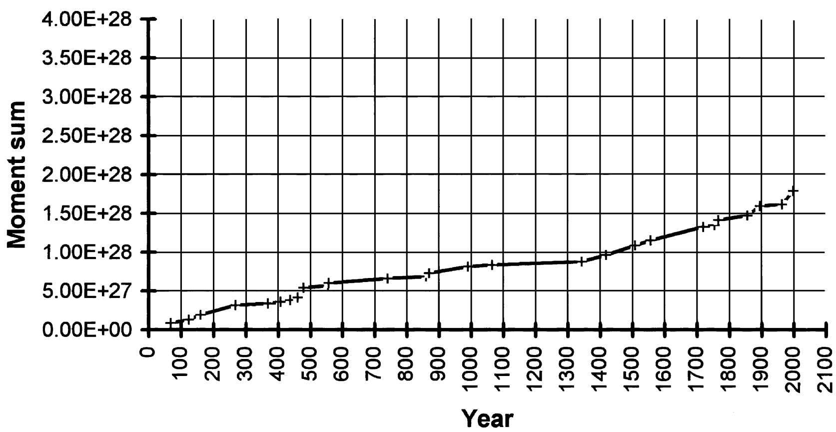

Figure 9. Cumulative seismic moment release in the 210-km-long Marmara Basin region (B) as a function of time for the events listed in Table 1. Total seismic moment released: 1.8×10^{28} dyne cm.

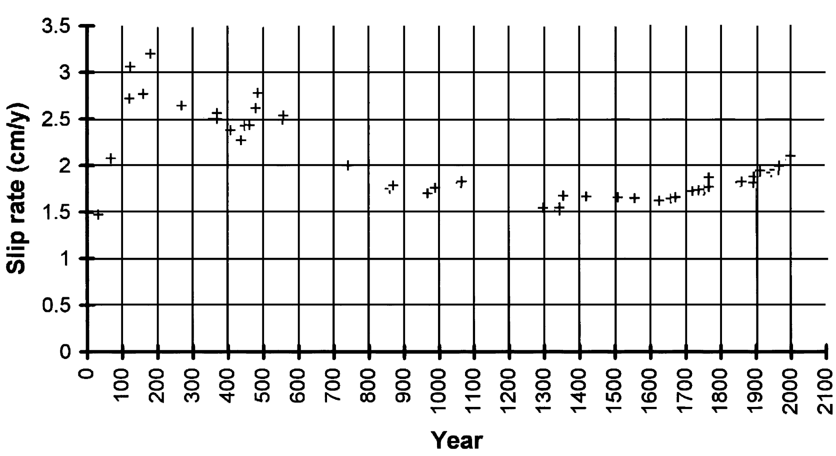

Figure 10. Variation with time of average slip rate in cm/yr calculated for the Marmara Sea region.

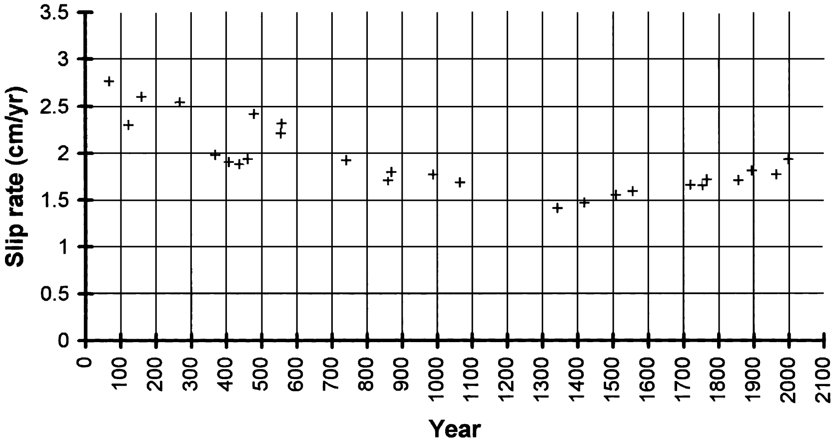

Figure 11. Variation with time of average slip rate in cm/yr calculated for the Marmara Basin region.

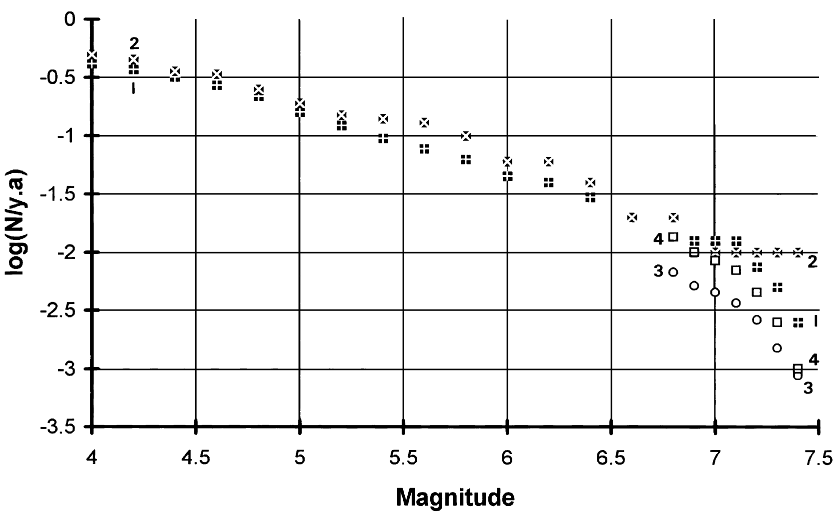

Figure 12. Frequency–magnitude distribution. Values of (*N*/*ya*) have been normalized to annual rate and unit area of 23,000 km^2, that of the Marmara Basin region. 1, crosses: Marmara Sea area, 100 yr; 2, X: Marmara Basin, 100 yr; 3, open circles: Marmara Sea area, 2000 yr; 4, open squares: Marmara Basin, 2000 yr.

the study of a region larger than A in Figure 1, which includes a considerable part of the Anatolian Fault Zone, confirm this and suggest that a greater east–west extent of the Zone needs to be considered (Ambraseys, 2001a).

Figures 12 and 13 show that, in hazard modeling from short-term, 100-yr-long datasets, it is simply not reasonable to ignore the chance that much or all of our records may be from a quiescent or energetic period in the seismic activity. This is one of the possibilities that must be considered when making assumptions with incomplete datasets, and this is the principal reason why statistics alone cannot quickly and simply answer the question of seismic hazard evaluation (Vere-Jones, 1987).

The North-Central Marmara Basin

The question now arises of whether the long-term dataset suggests the existence of a throughgoing fault along the shelf break bordering the Tekirdağ, Central, and Çinarcik subbasins, which could rupture along a length of 150 km or more and produce a major earthquake.

The answer to this question depends much on whether such events have left some significant trace in the bathymetry (Parke *et al.,* 2000). As we have seen, the Sea of Marmara is located where deformation on the North Anatolian Fault changes from being localized strike-slip on a single narrow fault zone striking roughly east–west, to being distributed over a broad region, bearing more to the south, with a substantial normal component of slip in the Basin that is taken up by short fault segments of a few tens of kilometers in length. Faults are generally less continuous offshore than onshore (Wong *et al.,* 1995), and the implication is that rupture of these short faults should be associated with moderate to small earthquakes, which is compatible with both short-period (100 yr) and long-period (2000 yr) evidence shown in Figures 3 and 7. One would expect, therefore, motion transferred from the Anatolian Fault Zone to the Marmara Basin east of 30° E to be taken up by the southern strands of the Anatolian zone, and in particular, with considerable components of normal slip, by the segmented floor of the Basin. This then would have caused the subbasins in the west extremity of the Tekirdağ, and in the east of Çinarcik where the faulting style changes, to show higher activity, which is exactly what we see in Figures 2 and 4.

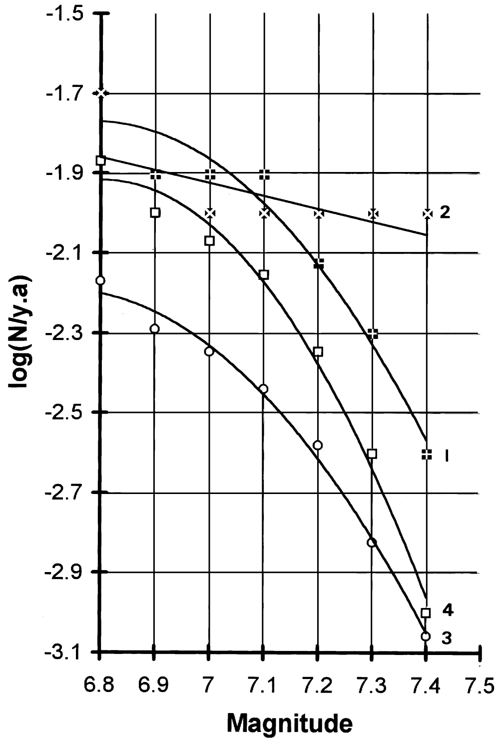

Figure 13. Frequency–magnitude distribution for $M_S \geq 6.8$. 1, crosses: Marmara Sea area, 100 yr; 2, X: Marmara Basin, 100 yr; 3, open circles: Marmara Sea area, 2000 yr; 4, open squares: Marmara Basin, 2000 yr.

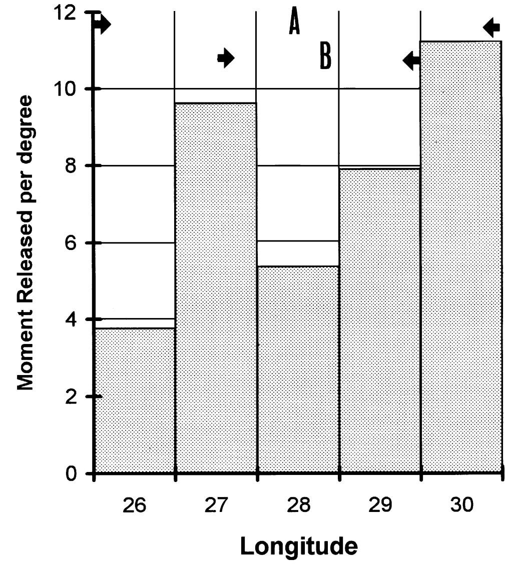

Figure 14. Distribution of the average seismic moment released (without correction for the missing $M_S < 6.8$) per degree of latitude (84 km) in the last 2000 yr. Seismic moment $\times 10^{27}$ dyne cm. A and B are the lengths of the Marmara Sea region and Marmara Basin, respectively. The moment deficiency in region B is not all that large, and it may be partly due to the large number of missing events of $M_S < 6.8$.

This apparent deficiency of large events in the Basin, and consequently of the total long-term seismic moment released is shown in Figure 14, a deficiency that is not all that significant, chiefly caused by the contribution in the total seismic moment of events of $M_s < 6.8$ that are missing from our historical dataset.

The relatively low seismic activity in the Marmara Basin during the last 100 yr is unequivocal (Fig. 2). Whether we can say with confidence the same for the much longer period of 2000 yr depends critically on the correct interpretation of the macroseismic information of the two largest earthquakes of 10 September 1509 and 22 May 1766. We may discuss briefly these two cases for which the interpretation of the supporting macroseismic data have led to debatable conclusions regarding the size and location of some of these events.

The earthquake of 10 September 1509 is of special interest as it has been considered by modern writers (*e.g.,* Parsons *et al.,* 2000) to be a major event of magnitude 7.6 caused by an alleged rupture of a 380-km-long fault along the Marmara Basin extending from near the Dardanelles to İzmit. However, a careful reinterpretation of macroseismic effects in new sources show that this earthquake was not an exceptionally large event (Ambraseys, 2002). It excited widespread contemporary interest because of its location, close to the capital of the Ottoman empire, rather than because of its destructive effects. Its reappraisal shows that, with an epicentral area in the Sea of Marmara, the destruction it caused did not extend very far from Istanbul, certainly not beyond Silivri in the west and Gebze in the east, and that damage at Tekirdağ, Demitoka, and Gelibolu was not more serious than what one would have expected at these distances from an earthquake of M_S 7.2.

Hubert-Ferrari *et al.* (2000), using the data *in* Ambraseys and Finkel (1991), associated the earthquake of 22 May 1766 with a 100-km-long fault rupture offshore the north-central part of the Marmara Basin, which extends from Tekirdağ to near Istanbul, and for which they assume a magnitude of 7.4. However, the epicentral region of this earthquake of magnitude *7.2* is clearly east of Istanbul, in the Çinarcik Basin, a location correctly identified by Parsons *et al.* (2000).

With an historical record 20 times longer than that of the twentieth century, no truly large earthquake of $M_S > 7.5$ in the Marmara Basin has been found extending from Ganos to the Çinarcik subbasins and beyond, of a size comparable to those in the North Anatolian Fault zone, any one of which would account for 700 yr worth of moment release in the Basin. Whether 2000 yr is long enough to obtain a reliable answer to this question is less clear.

Thus, although one cannot rule out the possibility of large strain release in the Marmara Basin, it would not be adequately justifiable to assume at the present time that this can result from rupture of a strike-slip fault that extends from the Gulf of İzmit to the Galipoli Peninsula.

Seismicity of the Iznik–Bursa Area

The south branch of the fault system, in the interior of the mainland south of the Marmara Basin, is less active than the north branch through İzmit and the Marmara Basin. The morphology of the south branch through Geyve, Bursa, and Gönen to the Aegean Sea suggests late Quaternary activity (Barka, 1996) and a long-term seismicity that accounts for an average shear velocity of about 0.3 cm/yr, which is compatible with GPS measurements (Straub, 1996). However, we must admit that the long-term activity of the inland part of this region is not perfectly known.

Seismicity of the Thrace Basin

It is perhaps interesting that in spite of efforts to investigate the long-term seismicity of Thrace, we could find virtually no significant events in the region to the north of the Marmara Basin, except a moderate earthquake, southeast of Edirne in 1752.

We find that, throughout its 17-century-long history, Istanbul has suffered serious damage from near moderate and more distant large offshore earthquakes, originating chiefly from the Çinarcik subbasin. However, the potential high risk today is in great part due to the very rapid growth of Istanbul in the last decades and also due to the increased vulnerability of its modern building stock. Considering that in 1480 the population of Istanbul was just over 100,000 and today has exceeded 10,000,000, with greater Istanbul occupying vulnerable sites that extend along the coast for tens of kilometers, it is obvious that the repetition of one of the damaging historical earthquake today would be the *küçük kiyamat* (Little Apocalypse) of the Ottoman historiographers and cause great losses.

Conclusions

With the data available at the moment, the main conclusions from this study are as follows:

1. From the 2000-yr-long history of the region, there is no macroseismic evidence for a major earthquake that could be associated with rupture of the offshore North Anatolian Fault all along the north coast of the Marmara Basin from the Gelibolu Peninsula to the Gulf of İzmit.
2. The seismicity of the last 2000 yr can account for almost all of the expected 2.2 ± 0.3 cm/yr right-lateral slip in the Marmara Sea region.
3. We find virtually no significant earthquakes in Thrace and a subdued activity in the southern part of the Basin.
4. Large earthquakes in the Marmara region are less frequent when predicted from long-term datasets than from the usual 100-yr instrumental period.
5. Maximum magnitudes from short-time observations are overestimated making the notion of a recurrence time questionable.
6. There is a regional and long-time dependence of seismic activity that renders particularly problematic the assessment of hazard from short-term observations.
7. Clustering of seismic activity must be considered when statistically evaluating hazard in this region.
8. Historical earthquakes in the Basin close to Istanbul have been smaller than those that have occurred east, in the North Analolian Fault zone, and west in the Ganos–Aegean region.

Acknowledgments

I would like to thank Jean Bonnin, Armando Cisternas, James Jackson, and Hans Kahle for revewing this article and for their constructive comments. This work has developed over a period of almost a decade and has owed much to the contribution of individuals too many to mention here.

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Manuscript received 10 December 2000.