MECHANICAL INTERACTION OF EARTHQUAKE FAULTS IN NORTHWEST TURKEY

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I certify that I have read this dissertation and that, in my opinion, it is fully adequate in scope and quality as a dissertation for the degree of Doctor of Philosophy

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Abstract

With a predominant westward direction, the migration of earthquakes makes the North Anatolian fault in Turkey an ideal natural laboratory for studying how earthquakes interact along major fault systems. This study explores the mechanical interaction and stress triggering relationships between past earthquake ruptures and evaluates the future rupture potential in the Marmara Sea region.

The stress perturbation associated with the 1967 Mudurnu Valley earthquake helped trigger the 1999 Izmit earthquake. However, test of different potential rupture geometries of the 1967 event show that only if this rupture extended in the subsurface towards the 1999 Izmit segments, do the Izmit faults receive greater stress loading than all other neighboring faults.

The likelihood for an earthquake rupture to propagate along obliquely-oriented intersecting fault segments, as observed in the 1999 Izmit earthquake, depends on the relative orientations of the faults. Static Coulomb stress changes indicate that when a secondary fault strikes counterclockwise relative to a master strike-slip fault, a dipping secondary fault will fail more readily than a vertical fault. For a clockwise change in strike, a vertical fault will fail more readily than a dipping fault. The pre-existing fault segment geometries at the eastern end of the Izmit earthquake helped terminate rupture there and fault intersections of several major faults in the Marmara Sea favor composite rupture geometry.

Incorporating geophysical data with geomechanical modeling shows that, of several proposed configurations, a Marmara Sea fault geometry composed of several large east-west trending strike-slip strands with multiple fault strands bounding individual basins best fits the observed seafloor morphology and pattern of stratigraphic horizon deformation.

Coulomb stresses changes due to earthquakes since 1900 are increased on the Princes’ Island, Çinarcık, and Armutlu fault segments within the eastern Marmara Sea. In four of the six plausible configurations for the western termination of the 1999 Izmit rupture and the location of the 1963 Yalova earthquake, the Çinarcık fault receives the greatest average stress change. In another cases, the stress increase on the Princes’
Islands fault is greatest. Although it is difficult to conclude which rupture scenario is most plausible, the results identify the consequences of each configuration.
Since 1912, the North Anatolian fault (NAF) in Turkey has produced thirteen M>6.7 earthquakes with cumulative rupture of over 1100 km of its entire 1500 km length (Barka, 1996; Dewey, 1976; Toksoz, 1979). With a predominant westward direction, the migration of mainshock epicenters along this fault provides the world's clearest example of temporal and spatial organization of large earthquakes, making the NAF an ideal natural laboratory for studying how earthquakes interact along major fault systems. Ideally, improved understanding of the underlying physical interactions of these events would allow scientists to better predict the spatial and temporal window for expected future earthquakes. In this dissertation I explore the mechanical interaction between faults in northwest Turkey and potential effects of this interaction on earthquake occurrence.

One of the most important recent observations in the study of earthquake rupture processes is that the accumulated stress perturbations associated with previous earthquakes may constrain the locations of future earthquakes (Stein, 1999). Aftershock locations (Das and Scholz, 1981; Stein and Lisowski, 1983; Oppenheimer et al., 1988; King et al., 1994; Harris, 1998), changes in the rate of seismicity (Reasenberg and Simpson, 1992; Parsons et al., 1999), and the occurrence (Stein et al., 1992) or suppression (Bakun, 1999) of subsequent large mainshocks have been shown to correlate with the spatial distribution of Coulomb stress produced by preceding earthquakes. This stress triggering mechanism is evident along the NAF, as nine of ten epicenters of the M>6.7 earthquakes from 1939-1992 occur where previous events increased the Coulomb stress on the fault (Stein et al., 1997).

On August 17, 1999, the $M_w = 7.4$ Izmit earthquake occurred in a region of increased stress change (Stein et al., 1997) and extended the western limit of NAF rupture another 80 km west, terminating within the Marmara Sea. This event, and the $M_s = 7.4$ Ganos earthquake which ruptured its western coast in 1912, now define the Marmara Sea, with Istanbul and its 12 million residents on its northern coast, to be the most hazardous seismic gap along the NAF. Unlike the fairly straight and continuous trace of the portion of the NAF that ruptured from 1939-1967, the fault pattern within the submarine setting...
of the Marmara Sea is more complex and has been under great debate. The coincidence of the last seismic gap with a region of structural complexity within the fault zone therefore led me to investigate the relationship between structural heterogeneity and stress triggering.

This dissertation includes four chapters that investigate and constrain fault configurations in northwest Turkey and evaluate stress changes and earthquake triggering in this region. There is a temporal progression in the chapters – the first investigates most recent example of stress triggering along the NAF, the second evaluates the effects of obliquely-oriented fault segments on coseismic rupture as seen in the 1999 Izmit earthquake, the third constrains the complex fault pattern in the Marmara Sea where the next large earthquake is likely to occur, and the fourth focuses on stress changes on the faults in the eastern Marmara Sea due to recent earthquakes.

In chapter one, I take a detailed look at the potential stress triggering relationship between the 1967 Mudurnu Valley earthquake and the 1999 Izmit earthquake to determine if the stress triggering relationship holds for this pair of events. I evaluate if the Izmit fault received the greatest stress increase of all neighboring faults to determine if it could have clearly been identified as the most susceptible fault after the earthquake in 1967. I find that for three potential geometries of the 1967 Mudurnu Valley rupture, there are positive Coulomb stress changes at the hypocenter of the 1999 Izmit earthquake. In addition, only if the rupture in 1967 extended in the subsurface towards the Izmit segments, do the Izmit faults receive the greatest stress loading of all the neighboring faults in the region. This study was supported by three weeks of field mapping of faults in the overlap region between the 1967 and 1999 ruptures and 3-D boundary element modeling. This paper is co-authored by Atilla Aydin and Frantz Maerten. Atilla Aydin helped in conceptual development of the project, assisted the field work in Turkey, and provided manuscript review. Frantz Maerten provided assistance with numerical computations and the Poly3D computer program. The manuscript entitled “Investigating the transition between the 1967 Mudurnu Valley and 1999 Izmit earthquakes along the North Anatolian fault with static stress changes” has been published in Geophysical Journal International in 2003.
In chapter two I investigate the tendencies of rupture propagation along discontinuous and obliquely oriented fault segments within a single earthquake according to static stress changes. Large earthquakes, including the 1999 Izmit earthquake, have large rupture lengths and typically require rupture along multiple fault segments. To evaluate which fault segment intersection geometries most and least favor throughgoing rupture, I conducted numerical experiments that evaluated static stress changes and slip distributions on intersecting fault segments. The results show that the pre-existing fault segment intersection geometries at the eastern end of the Izmit earthquake would favor the observed termination of rupture. It also shows that several intersections of major fault strands in the Marmara Sea favor throughgoing rupture. This paper is co-authored by Atilla Aydin who provided advice and manuscript review. The manuscript “Rupture Progression along discontinuous oblique fault sets: Implications for the Karadere rupture segment of the 1999 Izmit earthquake, and future rupture in the Sea of Marmara” is in press for special issue of Tectonophysics entitled “Active Faulting and Continental Deformation in the Eastern Mediterranean Region.”

In chapter three I use geomechanical modeling to constraint several of the different fault geometries proposed for the Marmara Sea. I demonstrate that this modeling approach offers an independent means of evaluating the mechanical feasibility of different fault geometries that are derived by different authors, in several cases, from the same data set. I test the fault models in terms of their ability to reproduce the observed seafloor geomorphology and the structural deformation pattern of the stratigraphic horizon defining the transition from pre-transform to syn-transform deposition at selected locations within the Marmara Sea. This paper is co-authored with Atilla Aydin who helped obtain the bathymetry and seismic reflection data and provided conceptual guidance and manuscript review. The manuscript “Using Geomechanical Modeling to Constrain the Fault Geometry within the Marmara Sea, Turkey” has been submitted to the Journal of Geophysical Research.

In chapter four I apply the concepts of chapters one and two and the best-fitting fault geometry from chapter three to evaluate the stress changes on faults in the eastern Marmara Sea due to earthquakes since 1900. I consider several scenarios based on different proposed locations of the western termination of 1999 Izmit earthquake rupture
and the location of the 1963 Yalova earthquake and evaluate the most likely future rupture scenarios based on the stress change magnitudes and the potential effects of fault geometry on westward propagation. This manuscript is co-authored with Atilla Aydin and Tim Wright of Oxford University. Atilla Aydin provided conceptual guidance and manuscript review. Tim Wright performed slip inversions from InSAR surface deformation data for the 1999 Izmit earthquake fault that I used in the stress change calculations.

In addition to these four chapters I contributed as a co-author to the paper entitled “Slip on a system of shallow thrusts imaged by SAR interferometry in the Shahdad thrust belt, SE Iran” by Eric Fielding, Tim Wright, Jordan Muller, Barry Parsons, and Richard Walker currently in press for publication in Geology. To this manuscript I contributed calculations of surface deformation from thrust fault slip and several prepared figures. Eric Fielding and Tim Wright inverted SAR interferometry data to obtain the fault slip solutions that I used for the surface deformation calculations. Because I am not the first author on this manuscript, it is not included in this dissertation.

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Chapter 1

Investigating the transition between the 1967 Mudurnu Valley and 1999 Izmit earthquakes along the North Anatolian fault with static stress changes

Abstract

The last major rupture along the North Anatolian fault (NAF) prior to the Izmit and Düzce earthquakes in 1999 was the $M_s = 7.1$ July 22, 1967 Mudurnu Valley Earthquake which ruptured approximately 80 km along the surface of the west-central NAF. The geographic continuity of the Mudurnu Valley and Izmit ruptures corresponds with the historical westward progression of major earthquakes along the NAF and encourages us to test the stress triggering effects of the 1967 earthquake on the 1999 earthquake. Although the triggered earthquake sequence from 1939-1999 has been investigated by earlier workers (Stein et al., 1997), this study addresses several outstanding problems related to this most recent earthquake interaction in light of more detailed fault configurations and a *posteriori* knowledge of the 1999 events. The questions we hope to answer are: 1) Did the stress perturbation caused by the 1967 Mudurnu Valley earthquake promote failure at the 1999 Izmit earthquake hypocenter and along Izmit fault segments? 2) Why did the 1999 Izmit earthquake rupture along faults in the center of the Izmit-Akyazi depression rather than along the western continuation of the Mudurnu Valley fault that ruptured in 1967? and 3) What is the rupture configuration at the western end of the 1967 earthquake fault that would most favor slip on the August 17, 1999 Izmit earthquake fault? Using a 3-D boundary element method, we test three possible rupture configurations for the 1967 Mudurnu Valley earthquake. Coulomb stress changes are calculated on major faults within the region due to non-uniform slip associated with the 1967 Mudurnu Valley earthquake. We find that for all three potential 1967 Mudurnu Valley rupture geometries there are positive Coulomb stress changes at the hypocenter of the 1999 Izmit earthquake. We also find, however, that only when subsurface rupture of the 1967 Mudurnu Valley earthquake deviates from the mountainfront fault and extends towards Lake Sapanca does an Izmit rupture segment
(Sakarya segment) receive greater Coulomb stress changes than the mountainfront fault west of the Mudurnu rupture.

**Introduction**

Since 1939, the 1500 km long North Anatolian fault (NAF) in northern Turkey has produced twelve M > 6.7 earthquakes that have propagated both eastward and westward, with a cumulative rupture length of over 1100 km (Dewey, 1976; Stein et al., 1997; Toksoz, 1979) (Fig. 1.1). On August 17, 1999 the largest and most damaging earthquake in Turkey in 60 years extended the historical westward progression of large earthquakes along the NAF to within 100 km of Istanbul, a city of 12 million people (Barka, 1999) (Fig. 1.1). The M_w = 7.4 Izmit (Kocaeli) earthquake represents the latest event within the historical sequence and its 125 km long surface rupture brings the western limit of recent NAF rupture into the Bay of Izmit in the Sea of Marmara (Fig. 1.2).

The last major earthquake prior to the 1999 Izmit event was the M_s = 7.1 (Pinar et al., 1996) Mudurnu Valley earthquake of July 22, 1967 that brought the previous western termination of NAF rupture near the town of Kanlıçay (Fig. 1.3). After 1967, in accordance with the historical earthquake progression, many researchers turned their attention towards the NAF segments west of the Mudurnu Valley. Investigations included monitoring seismic activity (Crampin and Evans, 1986; Evans et al., 1985; Gürbüz et al., 2000; Iio, 1991; McKenzie, 1972), measuring ground surface displacements with a GPS network (Reilinger et al., 1997; Straub et al., 1997), and predicting areas of increased Coulomb stress with fault dislocation models (Nalbant, 1998; Stein et al., 1997). In a study of the cumulative rupture along the NAF, Stein et al. (1997) showed that nine out of ten ruptures along the main fault trace since 1939 were brought closer to failure by preceding earthquakes. In combination, these research efforts created an increased awareness of the potential triggering effect caused by the perturbed stress field associated with prior earthquakes and lead to general agreement that the northern strand of the NAF in this region was highly susceptible to future failure. Indeed several researchers correctly identified the Izmit region as among the most hazardous
Figure 1.1. (a) Tectonic map of Turkey (modified from Holzer, 2000). The NAF system is defined as one dominant trace before splitting into three strands west of the Mudurnu Valley. (b) Westward progression of earthquakes along the NAF since 1939 (modified from Stein et al., 1997). Moment magnitudes are shown beneath year of event. Ruptured portions of the fault are shown with thick white line.
Figure 1.2. a) Rupture traces for the 1967 Mudurnu Valley, 1999 Izmit, and 1999 Düzce earthquakes along the NAF. Also included are approximate alignment of the 1943 Hendek-Adapazari and 1963 Yalova earthquake ruptures (from Parsons et al., 2000). Background 30m resolution Digital Elevation map processed by Eric Fielding, Oxford University available at http://www.earth.ox.ac.uk/~geodesy/izmit.html. Box shows location of inset map. b) Inset of transition area between 1967 Mudurnu Valley earthquake and the 1999 Izmit earthquake. White arrows mark approximate location of Izmit rupture segments. Black arrows mark approximate location of mountainfront fault.
Figure 1.3. a) Fault map and selected Quaternary deposits in NW Anatolia. Geology from Ternek (1964) and Tokay (1964). Faults from compilation by Aydin and Kalafat (2002). The surface traces of the last three major earthquake ruptures along the NAF are shown with thick lines. Descriptions of ruptures at each circled letter (a)-(e) are provided in the text and Fig. 3b. Town names are abbreviated as follows: Iz = Izmit, Gc = Gölcük, Ar = Arifiye, Sp = Sapanca, Dg = Dogançay, Kl = Kanlicay, Kr = Karapürcek, Ak = Akyazi, Dk = Dokurcun, Gy = Gölyaka, and Dz = Düzce. Box shows location of Fig. 3b. b) Inset map showing secondary surface ruptures associated with the 1967 Mudurnu Valley earthquake (modified from Ambraseys and Zatopek, 1969). Lateral and vertical offsets are provided. c) Inset map of secondary faults (thin lines) between the western termination of the Mudurnu Valley earthquake rupture and Lake Sapanca. Box shows location of Fig. 3c. Secondary faults were mapped from geomorphology provided by 1:25000 topographic maps and field reconnaissance by Muller and Aydin. Faults are dashed in areas of increased uncertainty.
after 1967 (Crampin and Evans, 1986; Gürbüz et al., 2000; Stein et al., 1997; Toksoz, 1979).

Labeling the Izmit region as the most susceptible to future failure after the 1967 earthquake was fairly intuitive based on the historical consistency of westward propagating ruptures that were confined to the single main trace of the NAF (Fig. 1.1b), and on the local microseismic swarm near the eastern termination of the Bay of Izmit (Crampin and Evans, 1986; Crampin et al., 1985; Evans et al., 1985). The increasingly complex geometry of the NAF as it branches into multiple strands west of the Mudurnu Valley (Fig. 1.1a), however, increases the difficulty of predicting the failure of specific faults. This paper investigates the potential for stress transfer between the 1967 Mudurnu Valley and 1999 Izmit earthquakes on a finer scale. We calculate the stress effects of the Mudurnu Valley earthquake on the future Izmit epicenter and maximum slip regions. There are two general questions we ask. One, did the 1967 Mudurnu Valley earthquake promote failure of the 1999 Izmit earthquake fault segments? And, two, did the Izmit earthquake segments accommodate the greatest Coulomb stress changes of all neighboring faults?

**Regional fault geometry**

On a regional scale, a change from purely dextral strike-slip to a combination of strike-slip and normal faulting focal mechanisms occurs along the NAF west of approximately 30° E (Jackson and McKenzie, 1984; McKenzie, 1972). This transition has been attributed to the interaction of the strike-slip fault system with the extensional regime of the Aegean Sea (Crampin and Evans, 1986; Gürbüz et al., 2000). The region accommodating strike-slip motion also broadens in northwestern Anatolia and the NAF splits into three zones west of the Mudurnu Valley (Armijo et al., 1999) (Fig. 1.1a). The southern zone extends towards Bursa south of Uluabat and Kuş Lakes and intersects the Aegean Sea near Edremit. From the Mudurnu Valley, the central zone defines the southern margins of Iznik Lake and the Marmara Sea to the Erdek peninsula. The northern zone extends west to Lake Sapanca, and then can be traced into the Bay of Izmit and the Marmara Sea. These three east-west and northeast-southwest trending zones, which seem to have accommodated most of the large historical earthquakes in the region,
lie within a regional pattern of east-west and northeast-southwest trending right-lateral strike slip faults (Fig. 1.3). GPS measurements (Straub et al., 1997) indicate that the three branches combine to accommodate 22±3 mm/yr of dextral strike-slip motion of Anatolia relative to Eurasia. The northern zone is the most active with approximate slip rates of 10-15 mm/yr (Straub et al., 1997). The region between the northern and southern branches shows 5-10 mm/yr of distributed deformation without well-defined slip along the central fault strand. The southern fault strand is active and accommodates approximately 2-4 mm/yr of dextral slip.

These active fault systems appear to be inherited from older structures (Yilmaz et al., 1997), the most prominent of which is the topographic (Fig. 1.2) and lithologic (Fig. 1.3) boundary that extends from the Mudurnu Valley west to Gölcük, which we refer to as the mountainfront fault. This major boundary, along which the 1967 Mudurnu Valley earthquake ruptured, separates pre-Tertiary basement to the south and Neogene deposits to the north (Sarogolu et al., 1992), and delineates the southern margin of the Izmit-Sapanca depression. The mountainfront fault, mapped as being active (Sarogolu et al., 1992; Woith et al., 2000), exits the Mudurnu Valley and separates the more mountainous region to the south from dissected hills to the west-northwest for ~10 km (Fig. 1.2b) before trending east-west to Dogancay. West of Dogancay, the fault trends to the northwest towards the western end of Lake Sapanca. West of Lake Sapanca the fault trends approximately east-west and continues towards Gölcük. The rupture of the eastern portion of this fault in 1967 within the Mudurnu Valley and the observation of triggered slip on the fault south of Gölcük (Barka et al., 2000) and just south of Lake Sapanca during the 1999 Izmit earthquake (Aydin and Kalafat, 2002) suggest that the mountainfront fault is indeed active. Based on the locations of sequential earthquakes during the 1939-1967 progression (Toksoz, 1979), the position of the mountainfront fault, immediately ahead of the 1967 Mudurnu Valley rupture, suggests that this fault, in addition the Izmit earthquake faults, would be a likely candidate for the next NAF earthquake after 1967. In this study we therefore compute stress changes on the mountainfront fault as well as the Izmit fault segments. For three potential 1967 Mudurnu Valley earthquake rupture configurations, we attempt to identify cases in which the Izmit fault receives a larger stress trigger than the mountainfront fault.
Earthquake surface ruptures and relay structures

1967 Mudurnu Valley earthquake

The Mudurnu Valley earthquake is reported to be $M_s=7.1$ (Ambraseys and Zatopek, 1969; Pinar et al., 1996) and $M_w=7.0$ (Stein et al., 1997) with an approximate hypocentral depth of 10 km (Pinar et al., 1996) and an epicentral location of 40.7° N, 30.8° E. In detail, Pinar et al. (1996) inverted teleseismic body waves and found six subevents. Three distinct right-lateral sub-events at a depth of 10 km with inferred slip planes striking 275.7°, 93.0°, and 281.6°, dipping 89.3°, 89.4°, and 89.1°, and with rakes -179.5°, -178.4°, and -178.8°, respectively, comprise 84% of the total seismic moment. A left-lateral subevent at a depth of 6 km, with inferred slip plane striking 156.6°, dipping 87.4°, and with rake 175.5° comprises 9% of the total moment. The remaining 7% of total moment is released from two dip-slip subevents with different slip mechanisms determined to have depths of 18 km. Pinar et al. (1996) attribute the dip-slip events to local areas of tension or compression associated with segmentation of the dominant right-lateral fault system.

The main surface trace of the Mudurnu Valley rupture is slightly arcuate, trending almost east-west for much of its 80 km length (Fig. 1.3) (Ambraseys and Zatopek, 1969). The easternmost 25 km of the fault break lies in a zone ruptured in the 1957 $M_s=7.0$ Abant earthquake (McKenzie, 1972) but the later rupture does not precisely follow the surface trace from 1957 (Ambraseys and Zatopek, 1969). Along most of its length, the surface rupture occupies a broad 1-3 km wide shear zone with distributed secondary tensional and compressional structures. Figure 1.4b shows the distribution of surface offset measured for the Mudurnu Valley earthquake (Ambraseys and Zatopek, 1969; Güclü, 1969). The sense of lateral motion in the Mudurnu Valley earthquake is right-lateral with a maximum horizontal offset of 1.9 m. A normal slip maximum of 1.2 m with the north side downthrown occurs toward the western end of the rupture. There are, however, inconsistencies in the reported vertical offsets along the fault that approach tens of centimeters and differ in sign at several locations. This suggests that the reported vertical slip values are most likely associated with local discontinuities in a predominantly right-lateral fault structure. The dominant right-lateral focal mechanisms,
Figure 1.4. (a) Fault configuration in the numerical model for three Mudurnu Valley rupture scenarios. The western termination of the Mudurnu Valley earthquake fault differs in each case. (b) The surface offset along the Mudurnu Valley rupture measured by Güclü (1969) and Ambraseys and Zatopek (1969). Mudurnu Valley earthquake fault 3-D slip distributions for (c) case 1 (d) case 2 and (e) case 3. The surface slip in each model case matches a 4th-order polynomial fit through the measured right-lateral surface slip. The slip maximum at depth varies such that the calculated seismic moment for each case equals the estimated value of $2.7 \times 10^{19}$ Nm. In cases 2 and 3 there is no surface slip west of the termination point in case 1. In all cases the model fault slip is pure right-lateral.
together with the lack of large or consistent dip-slip focal mechanisms and surface offsets, suggest that the Mudurnu Valley earthquake may be well represented as pure right-lateral slip event in our modeling.

1999 Izmit earthquake

The Izmit rupture is composed of six segments that we refer to, from west to east, as the Karamürsel, Sapanca, Sakarya, Akyazi, Karadere, and Gölyaka segments (Fig. 1.4). The Karamürsel segment extends to the west from a fault stepover near the epicentral region of Gölcük [epicenter located 40.70° N, 29.99° E (Tibi et al., 2001)] and continues along the southern coast of the Bay of Izmit. The largest surface offset along this segment, and of the entire Izmit surface rupture, reaches a maximum of 4.8 m near Gölcük (Fig. 1.5). The Sapanca and Sakarya segments trend east-west and extend from just east of Gölcük into the north side of Lake Sapanca (~30 km) and then from the southern portion of Lake Sapanca to a few kilometers west of Akyazi (~20 km), respectively. The maximum surface offsets on these segments are 3.0 m and 4.5 m, respectively. The Akyazi segment, approximately ~6 km long with 0.5 m of offset, trends roughly west-northwest. The Karadere segment trends N 65° E for ~40 km along a geomorphically well-defined fault zone from east of Akyazi to southeast of Gölyaka. On this segment, slip reaches a maximum of 1.4 m. The Gölyaka segment extends ~ 6-8 km from Gölyaka east towards Lake Melen and has only 0.2-0.5 m of right-lateral surface offset.

Local structures between the 1967 and 1999 ruptures

To understand the transfer of earthquake slip from the Mudurnu Valley earthquake fault to the Izmit earthquake fault, we must consider the region between the western termination of the Mudurnu Valley earthquake rupture and the segments of the Izmit fault rupture within the middle of the Izmit-Sapanca depression (Fig. 1.2). If the Mudurnu Valley fault continues along the mountainfront fault, as the topography suggests, then the transfer of slip onto the Izmit-Sapanca fault segments would require a lateral jump northward of several kilometers. If the Mudurnu Valley and Izmit-Sapanca
Figure 1.5. Surface slip for the 1999 Izmit earthquake corresponding to each rupture segment. Measured by Aydin and Kalafat (2002).
faults, however, were actually linked at depth, then the rupture of the Izmit-Sapanca segments in 1999 would be the extension of a pre-existing rupture front.

Using a 30m resolution DEM image (Fig. 1.2a), one can trace the westward extension of the Mudurnu Valley fault into the Adapazari plain and along the mountainfront fault separating the more mountainous region to the south from dissected hills to the WNW for ~10 km as it trends towards Lake Sapanca. The mountainfront fault is not well exposed at the surface in this region. Perhaps for this reason, Ikeda et al., (1991) suggest, on the basis of stereographic analysis of SPOT imagery, that there is no geomorphic evidence for Quaternary faulting along this section of the mountainfront fault. Rather, they argue that there are many faults and folds of short length forming a wide deformation zone within the Izmit-Sapanca depression to the north. Field reconnaissance and mapping of geomorphology by the authors indicate northwest trending subsidiary faults southeast of Lake Sapanca (Fig. 1.3c). From surface-based evidence alone, however, it remains unclear whether these structures link the mountainfront fault and the Izmit earthquake fault.

Within this deformation zone, Ambraseys and Zatopek (1969) map several small intermittent ruptures caused by the 1967 event in a zone from Kanlicay to the shores of Lake Sapanca as summarized below and shown in Fig. 1.3b. The main continuous Mudurnu Valley earthquake surface rupture terminates about 3 km southeast of Karapürçek with 25 cm of right-lateral offset and 10 cm of vertical offset (north side downthrown) (Point a). Proceeding west-northwest, there were a series of increasingly discontinuous ground ruptures with vertical offsets decreasing to 5-10 cm and barely discernable lateral offset. Further north and west near Kanlicay a ~2 km rupture crosses two streams showing 70 cm of right-lateral offset and 30 cm of throw to the north (Point b). Between Kanlicay and Dogançay there is no reported evidence of surface rupture. Near Dogançay, a rail line was damaged by a 40 cm vertical offset along a lineation trending N70°W (Point c); however, no lateral offset was observed. This feature near Dogançay represents the westernmost surface rupture along the mountainfront fault that can be associated with the 1967 rupture.

From Dogançay north to Sapanca there is no reported evidence of surface rupture. North along the Sakarya River there were several small surface ruptures with only
vertical offsets reported (near Point d). However, vertical offset together with their proximity to the river suggests the possibility of landsliding or settlement due to shaking and liquifaction. Further north and west from this point, the only reported surface ruptures are on the shores of Lake Sapanca. On the southern shore of the lake there was a wide zone of ground ruptures (Point e) trending N30°W extending for up to 400 m inland and accommodating right-lateral displacements of 20 cm and a throw to the northeast of 40 cm. On the northwest projection of these cracks, emerging from the north side of the lake, a crack displayed barely perceptible lateral offset and 15 cm of throw to the northeast. While Ambraseys and Zatopek (1969) cannot directly attribute this array of secondary surface ruptures to a tectonic origin, they form a compelling lineation between the 1967 and 1999 surface rupture traces and suggest the potential for further rupture in the subsurface during the Mudurnu Valley earthquake.

Coulomb stress changes and model

We now investigate the potential for triggering of the 1999 Izmit earthquake due to different 1967 subsurface rupture geometries. The three different 1967 rupture geometries (Fig. 1.5a) are based on surface deformation reported right after the 1967 earthquake and on geological and geomorphological observations. In the first case, the western termination of the Mudurnu Valley fault is the same as the western termination of continuous surface rupture in 1967 near the town of Karapürcek. The geometry of case 2 allows for the possibility of slip extending at depth along the mountainfront further than Karapürcek, as may be suggested by the minor surface ruptures near Dogancay (Fig. 1.3a, point c). In case 3, the Mudurnu Valley earthquake fault extends towards Lake Sapanca through the zone of secondary surface ruptures mapped by Ambraseys & Zatopek (1969).

Based on Eq. 1 and the elastic stress distribution about a strike-slip fault, the Coulomb stress change on any ‘observation’ fault will depend on the coefficient of friction and the locations and orientations of the observation faults relative to the fault that has slipped (Muller and Aydin, in review). Given that the orientations of the Izmit and mountainfront fault segments are fixed, and the coefficient of friction is not likely to change from one configuration to the next, then the location of the greatest Coulomb
stress increase is controlled by the location and proximity of the western end of the Mudurnu Valley rupture.

We calculate the change in the static right-lateral Coulomb stress on the Izmit segments due to coseismic slip on a 3-D dislocation model of the Mudurnu Valley rupture within an elastic half-space. Coulomb stress analysis has previously been used to investigate the distribution of aftershocks with spatial patterns of static stress change, as well as to resolve static stress changes on neighboring fault planes. Harris (1998) provides an excellent review of the previous studies utilizing Coulomb stress principles to calculate static stress changes caused by slip on faults during earthquakes. Several of these studies have investigated NAF earthquake sequences. Stein et al. (1997) investigated the sequential changes in Coulomb stress due to the events in the period 1939-1967 and Hubert-Ferrari et al. (2000), Parsons et al. (2000), and King et al. (2001) calculated the cumulative static Coulomb stress changes in the Marmara Sea region since 1700. On a larger scale, Nalbant et al. (1998) investigated stress coupling between 29 M ≥ 6.0 events in northwest Turkey and the north Aegean Sea since 1912. This study focuses on a smaller scale than those previously mentioned in both the time window and the spatial area of interest. In addition, because we are interested in regions that are close to fault tips, we use interpolated 3-D slip distributions that taper to zero at the lateral and bottom fault edges rather than uniform slip on rectangular dislocations.

To calculate Coulomb stresses on the Izmit rupture segments we employ Poly3D (Thomas, 1993) which is a boundary element code based on the analytical solution for the elastic boundary value problem of an angular dislocation in a half space (Comninou and Dunders, 1975; Jeyakumaran et al., 1992). Model faults are composed of triangular dislocation elements, which allow for curved 3-D surface and tipline geometries. The model faults are vertical and extend to a depth of 15 km. Due to a lack of sufficient data, we assume the 15 km maximum depth of coseismic faulting in the Mudurnu Valley earthquake based on the maximum depth of slip in the Izmit earthquake. For the Izmit earthquake, this coseismic fault depth corresponds to the approximate depth to which aftershocks extend following the Izmit earthquake (Özalaybey et al., 2002; Tibi, 2001) as well as the maximum depth for coseismic slip for the Izmit earthquake inverted from
GPS measurements (Reilinger et al., 2000) and GPS, InSAR, and SPOT data (Feigl et al., 2002).

The change in right-lateral Coulomb stress, \( \Delta \sigma_{ri} \), on the fault segments is defined as the difference in \( \sigma_{ri} \) before and after the Mudurnu Valley coseismic slip:

\[
\Delta \sigma_{ri} = \Delta \tau_{ri} + \mu' \Delta \sigma_n
\]

(1.1)

where \( \mu' \) is the effective coefficient of friction (including the effects of pore pressure) and \( \Delta \sigma_n \) and \( \Delta \tau_{ri} \) are the changes in the normal and right-lateral shear tractions, respectively, associated with the stress perturbation of the Mudurnu Valley earthquake. Tensile stresses are positive and \( \tau_{ri} \) is set to be positive for right-lateral slip. There is an isotropic normal (lithostatic) stress gradient of 25 MPa/km with depth, and a regional tectonic compressive stress of 10 MPa applied at an orientation of N 55° W (Stein et al., 1997).

We use \( \mu' = 0 \) and \( \mu' = 0.6 \) as endmember values for the effective coefficient of friction in our models. Since faults within intraplate regions (Townend and Zoback, 2000) are suggested to have greater friction strength than interplate faults [e.g., the San Andreas fault (Townend and Zoback, 2001)], we use the observed coefficient of friction for intraplate faults of \( \mu' = 0.6 \) as an upper bound for the interplate faults in our model.

Following the arguments presented in section 3.1, we model the Mudurnu Valley earthquake fault with a right-lateral slip boundary condition. The net slip on the Mudurnu Valley fault in each case is applied such that it equals a 4th order function of the measured coseismic right-lateral offset at the surface (Ambraseys and Zatopek, 1969) and is zero along the lateral and lower edges (Fig. 1.4c-e). In cases 2 and 3 the surface slip west of Karapürcek is prescribed to be zero. There is also a local slip maximum defined at the hypocenter and on the rest of the fault is calculated as a linear interpolation from this maximum to the prescribed values along the tiplines. The magnitude of the local slip maximum at depth is constrained such that the average slip over the fault corresponds with the calculated seismic moment for the Mudurnu Valley earthquake. Estimates of the seismic moment of the earthquake vary widely between authors. Stein et al. (1997) provides the lowest reported value of \( 2.7 \times 10^{19} \) Nm for the event based on uniform slip on faults that extend to a 12.5 km depth while Pinar et al. (1996) sum the calculated moments of six subevents for the largest reported total of \( 1.1 \times 10^{20} \) Nm. We
use the minimum estimate of seismic moment for our analysis ~$2.7 \times 10^{19}$ Nm (Stein et al., 1997) so the slip magnitudes we apply at depth in each case may underestimate the actual Mudurnu Valley coseismic slip magnitudes.

Unlike other models of Coulomb stress transfer (King et al., 2001; Stein et al., 1997), we do not include interseismic stress loading via a deep dislocation or viscous shear zone below the depth of coseismic faulting. We do this for several reasons. First, it is not understood how the mountainfront or the Izmit earthquake faults serve to accommodate the deep plate motion. Adding secular slip beneath either fault would influence which fault would accommodate the largest Coulomb stress change since 1967. Estimates of the differential plate motion from GPS studies along this portion of the NAF system are 21-22 cm/yr (Reilinger et al., 1997; Straub et al., 1997). Applying this secular slip rate to a dislocation beneath the coseismic fault segments would load the coseismic fault segments by approximately 0.015 MPa/yr for a fault with a locking depth of 12.5 km (Stein et al., 1997). Over the period 1967-1999 secular slip at depth could therefore be expected to load the coseismic fault segments by approximately 0.5 MPa. These secular stress changes are up to several times the magnitude of the static coseismic stress changes calculated in our models. Unfortunately, the positions of the GPS stations in this region (Straub et al., 1997) does not allow one to partition relative displacements between the closely-spaced, subparallel fault strands (mountainfront fault and the Izmit earthquake fault), leaving the question of which may have been the more active fault before 1999, or where to apply slip via deep dislocations beneath the coseismic faults, unresolved. However, as suggested by observations of seismicity rate changes (Stein, 1999) and the concepts of rate and state friction (Dieterich, 1994), the sudden coseismic stress increases can have a much larger effect on earthquake triggering than secular stress loading that we omit.

Second, if we were to add deep dislocations beneath both faults with equal slip rates, it would superimpose a fairly uniform stress distribution upon the perturbation of the Mudurnu Valley coseismic slip and would not serve to help differentiate either fault as more susceptible to failure after 1967. Due to the proximity (<10 km) of the two faults, it is also likely that applying slip via a broad (several km) viscous shear zone at depth along the plate boundary would supply comparable loading to the mountainfront
and Izmit earthquake faults within the crust above. A deep viscous shear zone would therefore not preferentially load either fault, and would not help us assess which fault strand was more susceptible to failure after 1967.

Results

If the extent of 1967 Mudurnu Valley earthquake slip at the surface marks the western termination of slip at depth (Fig. 1.4a, case 1), the largest increase in right-lateral Coulomb stress, on any fault in the model, is located along the westward continuation of the Mudurnu Valley fault along the mountainfront fault (Fig 1.6 and Table 1.1). This is the true for both $\mu' = 0$ and $\mu' = 0.6$. For the case of $\mu' = 0$, $\Delta \sigma_{rl}$ reaches 0.57 MPa on the mountainfront fault. Of the Izmit earthquake fault segments, the Sakarya segment receives the most loading for right-lateral failure with $\Delta \sigma_{rl}$ approaching 0.06 MPa. For $\mu' = 0.6$, the mountainfront fault is still most favored overall, and the Sakarya segment is still most favored of all the Izmit segments, although $\Delta \sigma_{rl}$ is reduced on both the mountainfront fault (0.50 MPa) and the Sakarya segment (0.03 MPa). The lack of dependence on $\mu'$ for the Coulomb stress changes indicates that the shear stress changes are much larger than the normal stress changes on these segments for this model configuration.

Coulomb stress changes on the Karadere segment, however, differ substantially for different values of $\mu'$. The case of $\mu' = 0$ shows that the shear stress changes (equivalent to $\Delta \sigma_{rl}$ for $\mu' = 0$) reduce the tendency for right-lateral slip. A reduction in the normal stress acting across the Karadere segment, however, leads to the increased Coulomb stress change for the case of $\mu' = 0.6$. Therefore, unclamping normal stress changes dominate the Coulomb stress changes on the Karadere segment. Interestingly, the strike of the Karadere segment is oriented 25° towards the northeast relative to the western Izmit rupture segments. This orientation is at a high angle relative to N55°W, the most compressive regional stress direction, and also forms a significant discontinuity for the eastward propagating rupture to transition. The Coulomb stress changes due to the 1967 earthquake may therefore have eased the coseismic rupture progression in 1999 by reducing the normal stress across the Karadere segment.
Figure 1.6. Right-lateral Coulomb stress changes on Izmit and mountainfront fault segments for Case 1 in which subsurface slip in 1967 does not extend past the limits of the observed continuous surface rupture associated with the Mudurnu Valley earthquake. Red areas correspond to regions of increased tendency for right-lateral failure. a) $\mu' = 0$. b) $\mu' = 0.6$. 
Table 1.1. Maximum Coulomb stress changes ($\Delta \sigma_H$ in MPa) on the largest Izmit fault segments, at the Izmit earthquake epicenter, and on the mountainfront fault for the three Mudurnu Valley fault rupture configurations and the two endmember effective friction ($\mu'$) values.

<table>
<thead>
<tr>
<th>Case</th>
<th>$\mu'$</th>
<th>Izmit epicenter</th>
<th>Karamürsel</th>
<th>Sapanca</th>
<th>Sakarya</th>
<th>Karadere</th>
<th>Mountain-front</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0</td>
<td>0.011</td>
<td>0.011</td>
<td>0.044</td>
<td>0.060</td>
<td>-0.017</td>
<td>0.574</td>
</tr>
<tr>
<td>1</td>
<td>0.6</td>
<td>0.009</td>
<td>0.009</td>
<td>0.030</td>
<td>0.033</td>
<td>0.174</td>
<td>0.507</td>
</tr>
<tr>
<td>2</td>
<td>0</td>
<td>0.016</td>
<td>0.018</td>
<td>0.076</td>
<td>0.038</td>
<td>-0.245</td>
<td>0.185</td>
</tr>
<tr>
<td>2</td>
<td>0.6</td>
<td>0.014</td>
<td>0.014</td>
<td>0.041</td>
<td>0.037</td>
<td>0.109</td>
<td>0.155</td>
</tr>
<tr>
<td>3</td>
<td>0</td>
<td>0.017</td>
<td>0.018</td>
<td>0.141</td>
<td>0.179</td>
<td>-0.026</td>
<td>0.127</td>
</tr>
<tr>
<td>3</td>
<td>0.6</td>
<td>0.014</td>
<td>0.015</td>
<td>0.084</td>
<td>0.166</td>
<td>0.119</td>
<td>0.122</td>
</tr>
</tbody>
</table>
If the western extent of subsurface rupture during the 1967 Mudurnu Valley earthquake follows the mountain front fault to the west (Fig. 1.4a, case 2), then the potential for right-lateral failure is again greater along the mountainfront continuation of the Mudurnu Valley fault than on the 1999 Izmit earthquake fault segments (Fig. 1.7) for both $\mu'$ values (Table 1.1). In this case, however, the Sapanca segment is the most favored of the Izmit earthquake segments. For $\mu' = 0$, the maximum $\Delta \sigma_{r,l}$ is 0.19 MPa and 0.76 MPa for the mountainfront and Sapanca fault segments, respectively. Increased normal stress clamping on the Sapanca segment, however, reduces the magnitude of $\Delta \sigma_{r,l}$ for the case of $\mu' = 0.6$. Again, the Karadere segment has negative right-lateral Coulomb stress changes for $\mu' = 0$ and positive right-lateral Coulomb stress changes for $\mu' = 0.6$.

If the western extent of the subsurface rupture during the 1967 Mudurnu Valley earthquake diverges from the fault along the mountain front and follows the alignment of the reported minor surface failures (Fig. 1.4a, case 3), then the segment accommodating the greatest right-lateral Coulomb stress change depends on the coefficient of friction assigned for the faults. For $\mu' = 0$, the maximum $\Delta \sigma_{r,l}$ is greater on both the Sapanca and Sakarya segments than it is on the mountainfront fault (Fig. 1.8 and Table 1.1). The large positive values of $\Delta \sigma_{r,l}$ on the Sakarya segment are restricted to only the western edge of the fault within several kilometers of the surface; therefore, we would suggest that the Sapanca segment is more favored for failure than the Sakarya segment in this case. For $\mu' = 0.6$, while the Sakarya segment again has large increases in $\sigma_{r,l}$ on its western edge, the results suggest that the mountainfront fault is most favored to fail. The change in sign of $\Delta \sigma_{r,l}$ on the Karadere segment according to $\mu'$ is comparable to the other two cases.

In summary, all three scenarios investigated in this study provide increasing right-lateral Coulomb stresses along the western segments of the Izmit earthquake fault (Karamürsel, Sapanca, and Sakarya) and at the Izmit hypocenter (0.009-0.017 MPa); therefore, we conclude that the Mudurnu Valley earthquake may have served as a trigger for the Izmit earthquake. However, all scenarios, except case 3 with $\mu' = 0$, provide Coulomb stress increases along the mountainfront fault that exceed the stress increases on the Izmit earthquake fault segments. For the case 3 configuration (Fig. 1.4a) and $\mu' = 0$, there are larger Coulomb stress increases on the Sapanca segment than on the
Figure 1.7. Right-lateral Coulomb stress changes on Izmit and mountainfront fault segments for Case 2 in which 1967 rupture extends along the continuation of the mountainfront fault west of the Mudurnu Valley. a) $\mu' = 0$. b) $\mu' = 0.6$. 
Figure 1.8. Right-lateral Coulomb stress changes on Izmit and mountainfront fault segments for Case 3 in which 1967 rupture in the subsurface extends towards Lake Sapanca. a) $\mu' = 0$. b) $\mu' = 0.6$. 
mountainfront fault. Even in this case, however, the location of the increasing Coulomb stress is on the eastern part of the Sapanca segment and the magnitude of the maximum stress change is slightly less than the maximum on the Sakarya segment. In addition, all three scenarios also result in increasing right-lateral Coulomb stress on the Karadere segment in the eastern part of the Izmit earthquake fault for only the greater coefficient of friction.

Discussion

Our results indicate that the 1967 Mudurnu Valley earthquake promoted right-lateral failure on all segments of the Izmit earthquake fault west of the potential termination of the 1967 rupture. This result is consistent with the notion of a westward progression of sequential failure along the NAF (Dewey, 1976; Stein et al., 1997; Toksoz, 1979). Our results suggest, however, that only for a case in which the 1967 Mudurnu Valley coseismic slip extended in the subsurface towards Lake Sapanca (case 3) with \( \mu' \) close to zero would the static Coulomb stress changes favor rupture on the Izmit earthquake fault over the mountainfront fault. This suggests that the geometry of the western portion of the 1967 rupture was closer to that represented in case 3. Parsons et al. (2000) use a configuration of the Mudurnu Valley rupture similar to our case 3 for their stress triggering study for the Marmara Sea region, yet they do not provide justification for locating the western termination of 1967 rupture near Lake Sapanca. We feel that our modeling results and the observation of shear offsets of tens of centimeters at the surface near Lake Sapanca in 1967 justify this location for the western termination of Mudurnu Valley earthquake rupture.

In the absence of any knowledge of the pre-strength condition on the Izmit earthquake fault segments, our results also suggest that the active NAF may be rather weak, with a low effective coefficient of friction. The notion of a low coefficient of friction has been proposed for other large plate-boundary strike slip-faults, such as the San Andreas fault (Townend and Zoback, 2001), due to measurements of high angles between the local most compressive stress direction and the fault strike. A focal mechanism inversion of earthquakes in the Marmara Sea region suggests that the most compressive stress is about 55° from the strike of the Izmit segments (Gürbüz et al.,
This angle is less than the $60^\circ$-$90^\circ$ angles measured along much of the San Andreas (Townend and Zoback, 2001), but is greater than the $\sim 30^\circ$ that would be predicted for a strong fault (i.e. $\mu' = 0.6$). More in situ stress measurements would be necessary to evaluate the frictional strength along this portion of the NAF.

For the Mudurnu Valley earthquake fault configuration of case 3 and $\mu' = 0$, the area of greatest increase in right-lateral Coulomb stress is located near the overlap of the Sapanca and Sakarya segments near Lake Sapanca. This location is nearly 35 kilometers from the instrumental epicenter located near Gölcük (Toksoz et al., 1999), along the western part of the Sapanca segment. The observed mismatch of the hypocentral location of the Izmit earthquake and the location of the greatest right-lateral Coulomb stresses increases in all three of our model cases may therefore be interpreted in several ways.

One interpretation is that the western termination of the Mudurnu Valley rupture could be inaccurate and it actually occurred further east, closer to Gölcük. This is unlikely due to the lack of any surface rupture observations west of Lake Sapanca associated with the Mudurnu Valley earthquake.

A second interpretation is static stress perturbations have a greater effect on the magnitudes of eventual slip rather than on the process of slip initiation. The initiation process could rather be dictated by other heterogeneities in geometry or local fault strength that are not represented in our model. An apparent relationship between static stress changes and slip maxima has been noted for other large earthquakes sequences (Du and Aydin, 1993; Perfettini et al., 1999). In our model, this relationship is supported as a slip maximum near Lake Sapanca, the location of greatest Coulomb stress increase, has been computed from joint inversions of InSAR, GPS, and seismic data (Delouis et al., 2002) and inversion of InSAR, GPS, and other satellite imagery (Feigl et al., 2002). The surface slip (Fig. 1.5) does not correlate as well with the locations of the computed stress increases in our case 3, although the Sapanca and Sakarya segments do have the greatest right-lateral surface slip apart from the epicentral region.

A third interpretation is that other historical earthquakes before 1999 have preferentially loaded the Izmit fault, and specifically the Gölcük region, for failure. Potential earthquakes affecting the pre-1999 stress condition on the Izmit fault include the 1943 Hendek-Adapazari earthquake ($M_s = 6.4$) near Hendek north of the Sapanca
segment, and the 1963 Yalova earthquake ($M_t = 6.4$) within the Marmara Sea to the west of the Bay of Izmit and north of Yalova (Fig. 1.2a). The 1943 right-lateral strike slip event reduced the Coulomb failure stress in on the Izmit fault (Parsons et al., 2000), and to a lesser degree on the mountainfront fault, suggesting that this event would not have preferentially loaded the Izmit earthquake fault, or the epicentral Gölcük region, for failure. The 1963 pure normal faulting event should have increased the Coulomb failure stress on both the Izmit and mountainfront faults. However, due to the proximity of these two faults and their distance from the Yalova rupture, the stress increases for both faults would be small and of comparable magnitude. The location of the 1963 event, west of the Izmit rupture, however, could have possibly contributed to the Izmit hypocenter being located west of the maximum Coulomb stress changes predicted by our model.

Whereas previous studies of stress transfer along the NAF have focused on the regional Coulomb stress changes associated with multiple ruptures over hundreds of kilometers and tens to hundreds of years (Hubert-Ferrari et al., 2000; Stein et al., 1997), this study focuses on a more detailed fault configuration for a single interacting earthquake pair. In the stress studies of Stein et al. (1997), the Mudurnu Valley rupture stops at the western extent of 1967 surface rupture and the maximum and mean Coulomb stress changes on the Sapanca fault segment (the future Izmit earthquake fault) due to the preceding $M > 6.7$ earthquakes are positive on the order of 0.01 MPa. In the fault model of Parsons et al. (2000), the 1967 rupture extends all the way to the south side of Lake Sapanca and the Gölcük segment and the Izmit hypocentral region receive the greatest increases in Coulomb stress. However, in our model with more accurately constrained fault geometries, the stress perturbation caused by the 1967 Mudurnu Valley earthquake resolved greater Coulomb stresses on a fault other than the Izmit earthquake fault for two of three potential 1967 rupture configurations. This suggests that while calculations of Coulomb stress changes on optimally oriented planes (Stein et al., 1997) place Izmit within a region of stress increase after 1967, Coulomb stress changes resolved on faults within that broad zone of stress increase would likely have failed to predict the particular fault most susceptible after 1967.
Conclusions

For three potential locations of the western termination of rupture in the 1967 Mudurnu Valley earthquake, we show that the 1999 Izmit hypocenter received positive right-lateral Coulomb stress changes due to the 1967 rupture. In addition, the mountainfront fault and all Izmit earthquake fault segments west of the 1967 rupture termination also received positive right-lateral Coulomb stress changes. However, only when the 1967 rupture terminates near the southern shore of Lake Sapanca, and the faults have an effective coefficient of friction of zero, do our results show that the Izmit fault segments receive greater Coulomb stress increases than the mountainfront fault. These results highlight the fact that minor changes in rupture geometry and slip distribution can have a significant impact on identifying which fault strand or segments within a larger fault system may be most susceptible to future failure. The ability to identify which segment or strand within a fault zone is most susceptible to failure is crucial information for evaluating future rupture scenarios. Incorporating additional data such as paleoseismic recurrence intervals and recent microseismicity in conjunction with Coulomb stress models may therefore aid in characterizing the seismic hazard of individual faults within a complex regional fault configuration.

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Chapter 2

Rupture progression along discontinuous oblique fault sets: Implications for the Karadere rupture segment of the 1999 Izmit earthquake, and future rupture in the Sea of Marmara

Abstract

Large earthquakes in strike-slip regimes commonly rupture fault segments that are oblique to each other in both strike and dip. This was the case during the 1999 Izmit earthquake, which mainly ruptured E-W-striking right-lateral faults but also ruptured the N60°E-striking Karadere fault at the eastern end of the main rupture. It will also likely be so for any future large fault rupture in the adjacent Sea of Marmara. Our aim here is to characterize the effects of regional stress direction, stress triggering due to rupture, and mechanical slip interaction on the composite rupture process. We examine the failure tendency and slip mechanism on secondary faults that are oblique in strike and dip to a vertical strike-slip fault or “master” fault. For a regional stress field well-oriented for slip on a vertical right-lateral strike-slip fault, we determine that oblique normal faulting is most favored on dipping faults with two different strikes, both of which are oriented clockwise from the strike-slip fault. The orientation closer in strike to the master fault is predicted to slip with right-lateral oblique normal slip, the other one with left-lateral oblique normal slip. The most favored secondary fault orientations depend on the effective coefficient of friction on the faults and the ratio of the vertical stress to the maximum horizontal stress. If the regional stress instead causes left-lateral slip on the vertical master fault, the most favored secondary faults would be oriented counterclockwise from the master fault. For secondary faults striking ±30° oblique to the master fault, right-lateral slip on the master fault brings both these secondary fault orientations closer to the Coulomb condition for shear failure with oblique right-lateral slip. For a secondary fault striking 30° counterclockwise, the predicted stress change and the component of reverse slip both increase for shallower-angle dips of the secondary fault. For a secondary fault striking 30° clockwise, the predicted stress change decreases but the predicted component of normal slip increases for shallower-angle dips of the secondary fault. When both the vertical master fault and the dipping secondary fault are
allowed to slip, mechanical interaction produces sharp gradients or discontinuities in slip across their intersection lines. This can effectively constrain rupture to limited portions of larger faults, depending on the locations of fault intersections. Across the fault intersection line, predicted rakes can vary by >40° and the sense of lateral slip can reverse. Application of these results provides a potential explanation for why only a limited portion of the Karadere fault ruptured during the Izmit earthquake. Our results also suggest that the geometries of fault intersection within the Sea of Marmara favor composite rupture of multiple oblique fault segments.

**Introduction**

Within a single strike-slip earthquake there are commonly changes in strike and dip between rupture segments on a variety of scales. On a centimeter- to kilometer-scale, fractures or faults bridge stepovers (Aydin and Nur, 1982; Segall and Pollard, 1980) or form jogs (Sibson, 1985) between larger parallel faults. On a kilometer to tens of kilometer scale, strike-slip or dip-slip faults with meters of offset in individual earthquakes form bends or double-bends (e.g., Barka and Kadinsky-Cade, 1988). While smaller-scale secondary faults and fractures could potentially be created during an earthquake rupture event, large oblique fault segments typically rupture pre-existing faults. Adjacent faults within a composite rupture can vary in strike by up to 30° and in dip by up to 60°. Examples that exhibit this behavior include the 1976 Songpan, China (Jones et al., 1984), the 1999 İzmit, Turkey (Aydin and Kalafat, 2002), and the 1999 Hector Mine, California (Scientists from the U.S. Geological Survey, 2000) earthquake ruptures, illustrated in Figure 2.1.

We define "composite rupture" as meaning the coseismic rupture process with discontinuous rupture of long (tens of km) fault segments oriented oblique to each other, whether by multiple events or cascading subevents. For example, the 1970 Tonghai, China earthquake ruptured through a double bend and was composed of three to four seismic events over ~50 s (Zhou et al., 1983). Similarly, the 1999 İzmit earthquake also ruptured through a double bend before terminating along the Düzce fault and was determined to have multiple seismic subevents each with $M_w > 6.9$ (e.g., Tibi, 2001; Li et al., 2002).
Figure 2.1. Examples of composite earthquake ruptures. (a) 1976 Songpan, China (modified from Jones, et al., 1984); (b) 1999 Izmit, Turkey (modified from Aydin and Kalafat, 2002); and (c) 1999 Hector Mine, California (modified from Scientists from USGS, SCEC, and CDMG, 2000). The regional maximum principal stress directions are shown with large black arrows, interpreted from Mueller et al. (2000) in (a) and (c) and from Gürbüz, et al. (2000) (N35°W) in (b).
When the obliquity in strike between segments is large, the rupture of the bend section of the fault trace may involve a different mechanism from the pure strike-slip mainshock. For instance, the three fault segments of the August 1976 Songpan earthquake (Fig. 2.1a) each ruptured with events greater than magnitude 6.7, with the second shock five days after the first and the third thirty hours later (Jones et al., 1984). This sequence included reverse-oblique left-lateral strike-slip events on subparallel faults at the ends of the structure and an almost pure reverse event within the 12 km long restraining double-bend in between (Jones et al., 1984). Such observations of seismic subevents along obliquely-oriented rupture segments suggest that a smooth and continuous rupture is typically unable to propagate through significant changes in fault strike and dip. Additional seismic events comparable to the initial mainshock may instead be necessary to overcome these obstacles.

The North Anatolian fault (NAF) in northwest Turkey offers insight into composite rupture processes involving fault intersections at oblique angles. In the 1999 İzmit earthquake (Fig. 2.1b), the E-W striking and vertically-dipping Gölcük to Akyazi rupture segments intersect the S60°W-striking Karadere segment that dips NNW at ~70-80° (as interpreted from the aftershock locations in Seeber et al., 2000). The Karadere segment, in turn, intersects the approximately E-W trending segment of the Düzce fault, a small part of which ruptured in both the August 17 İzmit and November 12 Düzce earthquakes. The documented surface slip along the Karadere segment involved no significant systematic dip-slip component, and ~1-2 m of right-lateral slip, less than the 4-5 m slip maxima on the Gölcük and Sakarya segments further west (e.g., Aydin and Kalafat, 2002). Although the slip on the Karadere segment was almost exclusively right-lateral, an $M_w$ 4.9 earthquake along the Karadere segment approximately one month earlier involved almost pure dip-slip (Fig. 2.1b) (Seeber et al., 2000). While it is possible that this event did not occur on the Karadere fault itself (the strike of the steeper nodal plane is subparallel to the Karadere fault), this event suggests the possibility that a fault segment may rupture with a different mechanism depending on whether this rupture is isolated or a subevent within a larger earthquake. The location of the termination of İzmit earthquake rupture, ~10 km east of the Karadere segment, and the apparent local temporal variability in focal mechanisms, led us to investigate the role of oblique-angle
fault intersections in strike-slip rupture sequences. We set out to quantify effects of the \textit{regional stress direction}, stress \textit{triggering due to rupture}, and the \textit{mechanical interaction of slip} to characterize the failure tendency and slip mechanism on secondary faults that are \textit{oblique in strike and dip to a strike-slip “master” fault}. Using three-dimensional boundary element modeling, we show that strike-slip on a vertical master fault will affect the state of stress on such secondary faults, in a manner that depends on their dip and relative strike. We also predict, through analyses of \textit{computed slip distributions}, that the focal mechanisms of subevents that may rupture the secondary faults have increasing components of oblique slip that increase with decreasing fault dip, which has implications for the ability of continued rupture propagation past the secondary fault. We then discuss how our results relate to past and future ruptures in northwest Turkey. The Izmit rupture may thus serve as an analog for future rupture in the Sea of Marmara, where adjacent fault segments intersect at strongly oblique angles, so understanding the interactions of such fault segments may be important for assessing potential rupture scenarios there.

\textbf{Effects of regional stress loading on composite rupture}

\textit{General analysis}

The tendency for an earthquake rupture to propagate along an obliquely-oriented secondary fault is enhanced when the regional stress field produces favorable conditions for failure on both fault orientations. This effect depends on the orientations and frictional properties of the faults and the orientations and relative magnitudes of the three principal stresses. For composite rupture to occur as a result of regional stress loading, both fault segment orientations must satisfy the condition for shear failure, which can be expressed as the condition for positive Coulomb stress,

\[
\sigma_c = |\tau| + \mu' \sigma
\]  \hfill (2.1)

where \(\sigma\) is the normal stress across the fault plane, and \(\tau\) is the resolved shear stress in the direction of predicted shear failure, and \(\mu'\) is the effective coefficient of friction (Jaeger and Cook, 1979). Our calculations follow the convention in which compressive normal stress is negative, hence the (+) sign; an increase in compressive normal stress will inhibit shear failure. The effective coefficient \(\mu'\) is used to relate pore pressure to the normal
stress acting on the fault surface and can be defined as \( \mu' = \mu(1-B) \) where Skempton’s coefficient \( B \) theoretically ranges from 0 for fully drained conditions to 1 for fully saturated conditions (Harris, 1998). However, \( \mu' \) is not physically well-constrained because we commonly do not know the appropriate value of \( B \) to employ for the fault rocks in the seismogenic zone and because pore pressures can change with time (Scholz, 2002). Studies of stress changes produced by earthquakes suggest that \( \mu' \) can range from 0 to 0.75, with values of \( \mu' = 0.4-0.5 \) appropriate in earthquake triggering investigations (e.g., Harris and Simpson, 1996; Harris et al., 1995; Stein et al., 1997). Deng and Sykes (1997) found that the earthquake stress interactions in southern California are fairly insensitive to different values of \( \mu' \) and they suggest the range 0-0.6 is reasonable. In our models, we use \( \mu' = 0.5-0.6 \). These values are towards the upper limit suggested by Deng and Sykes (1997) and they allow for the normal stresses on the oblique secondary faults in our models to have a greater affect on the calculated Coulomb stress magnitudes. While the tendency to slip depends on the magnitude of \( \sigma_c \), the sense of any slip depends on the direction of the maximum resolved shear stress.

Here we briefly analyze the regional stress and fault strength conditions that promote composite failure of vertical master faults and 60\textdegree-dipping secondary faults. We define a homogeneous regional stress field that will favor both strike-slip and normal faulting. The principal stresses are assumed orthogonal to the Earth’s surface, the vertical stress \( S_v \) and most compressive horizontal stress being equivalent in magnitude and exceeding the magnitude of the least compressive horizontal stress \( S_{\text{hmin}} \) (i.e., \( S_{\text{hmax}} = S_v > S_{\text{hmin}} \)). To make this model approximate to northwest Turkey, we take note of the regional stress field inversion by Gürbüz et al. (2000) using teleseismic waveforms to determine earthquake focal mechanisms. They determined a shear regime (i.e., with the intermediate principal stress vertical) close to extension, with \( (S_v - S_{\text{hmin}})/(S_{\text{hmax}} - S_{\text{hmin}}) = R = 0.93 \), and with the maximum principal stress oriented N35\textdegreeW. This \( R \) value suggests that \( S_v \) and \( S_{\text{hmax}} \) should be comparable in magnitude. To keep our analysis general, however, we test a range of \( S_{\text{hmax}} \) orientations. We then consider the specific \( S_{\text{hmax}} \) direction and fault orientations in northwest Turkey in section 2.2.
We estimate the vertical stress $S_v (= \rho_r g z$, where $\rho_r$ is the assumed density of crustal rock, 2500 kg m$^{-3}$, and $g$ is the acceleration due to gravity, 9.8 m s$^{-2}$) at $z = 5$ km depth to be 123 MPa and the hydrostatic pore pressure $P_p (= \rho_w g z$, $\rho_w$ being the density of water, 1000 kg m$^{-3}$) to be 49 MPa. The minimum horizontal stress $S_{\text{hmin}}$ is calculated such that the vertical fault will be at a critical state for Mohr-Coulomb shear failure for an assumed coefficient of friction, this condition being given by

$$S_{\text{hmin}} = (S_{\text{hmax}} - P_p)/\mu + P_p,$$

(2.2)

where

$$\mu = [(\mu^2 + 1)^{1/2} + \mu]^2$$

(Jaeger and Cook, 1979). For $\mu^2 = 0.6$, $S_{\text{hmin}}$ is 73 MPa at 5 km depth. The effective stresses subtract the normal stresses due to the hydrostatic pressure, so that effective $S_v$, $S_{\text{hmax}}$, and $S_{\text{hmin}}$ are 74 MPa, 74 MPa, and 24 MPa, respectively.

Because this stress state is critical for $\mu^2 = 0.6$, only vertical faults oriented $\pm \tan^{-1}(1/\mu^2)/2$ (i.e., 29.5° for $\mu^2=0.6$) relative to $S_{\text{hmax}}$ will fail according to the Mohr-Coulomb criterion (Fig. 2.2). While Coulomb stress calculations within a uniform regional stress field will not predict failure of faults in non-optimal orientations, they will tell us how the regional stress is contributing to their eventual failure during an earthquake. We are not interested in the ability for regional stresses alone to initiate rupture, but rather we are interested in how the contribution of regional stress affects the tendency for rupture along obliquely-oriented faults. Therefore, to consider failure along multiple secondary fault orientations, we arbitrarily assume a strength drop to $\mu^2 = 0.5$ on the faults to allow for $\sigma_c > 0$ on a range of fault orientations. Increased pore pressures along a fault would reduce the effective coefficient of friction, but here we do not attribute this strength drop to a particular physical mechanism. Strictly speaking, conditions with $\sigma_c > 0$ should never occur, because shear failure will occur on an optimally-oriented fault in the crust before such conditions can be reached. In our regional stress analysis, however, faults with $\sigma_c > 0$ do not physically slip, and we use positive values of $\sigma_c$ to indicate the orientations of pre-existing faults that would be most favored to slip within the defined regional stress field.
Figure 2.2. State of stress used for regional stress analysis. Stress conditions on all faults surfaces are calculated at 5 km depth. $S_{Hmax}$, the maximum horizontal principal stress, is set equal to the lithostatic stress $S_L$ of 123 MPa at 5 km depth (calculated for a crustal density of 2500 kg m$^{-3}$). Pore pressure $P_p$ is set as 49 MPa (assuming pore water saturation). $S_{Hmin}$ is set for critical shear failure according to equation (2). The angle $\alpha$ is that between $S_{Hmax}$ and the strike of a vertical strike-slip fault (clockwise positive). The assumed stress field has a uniform orientation throughout the model, with all stress tensor elements increasing at the same rate with depth.
The resolved shear stress on the model faults is computed by using a three-dimensional transformation of the homogeneous regional stress tensor into fault-parallel and fault-normal co-ordinates \((x'_1, x'_2, x'_3)\). In this local coordinate system, \(x'_1\) is along strike, \(x'_2\) is normal to the fault, and \(x'_3\) is along dip. In tensor notation, the fault local stress components are given by

\[
\sigma_{ij} = l_{ij} \sigma_{pq},
\]

with summation over repeated indices (Means, 1976). The \(l_{ij}\) terms represent the direction cosines between the new \(x'_i\) axes and each old axis. The maximum resolved shear stress on the model fault, \(\tau_{\text{max}}\), is given by

\[
\tau_{\text{max}} = (\sigma_{1'2'}^2 + \sigma_{2'3'}^2)^{1/2}.
\]

The direction of the maximum resolved shear stress on the model fault surface, our proxy for the predicted rake \(\lambda\) on the fault, is, using the rake convention of Aki and Richards (1980),

\[
\lambda = \tan^{-1}(\sigma_{2'3'}/\sigma_{1'2'}) + \pi/2 \text{ for } \sigma_{1'2'} > 0 \text{ and } \sigma_{2'3'} > 0,
\]

\[
\lambda = \tan^{-1}(\sigma_{2'3'}/\sigma_{1'2'}) - \pi/2 \text{ for } \sigma_{1'2'} > 0 \text{ and } \sigma_{2'3'} < 0,
\]

\[
\lambda = \tan^{-1}(\sigma_{2'3'}/\sigma_{1'2'}) - \pi/2 \text{ for } \sigma_{1'2'} < 0 \text{ and } \sigma_{2'3'} > 0,
\]

\[
\lambda = \tan^{-1}(\sigma_{2'3'}/\sigma_{1'2'}) + \pi/2 \text{ for } \sigma_{1'2'} < 0 \text{ and } \sigma_{2'3'} < 0,
\]

where \(\sigma_{1'2'}\) is positive for right-lateral slip, \(\sigma_{2'3'}\) is positive for reverse slip, and \(\sigma_{2'3'}\) is positive in tension. The maximum Coulomb stress \(\sigma_{\text{c max}}\) on the fault plane is therefore

\[
\sigma_{\text{c max}} = |\tau_{\text{max}}| + \mu' \sigma_{2'2'}.
\]

The normalized Coulomb stresses on faults striking at angle \(\alpha\) (counterclockwise positive) relative to \(S_{\text{th max}}\) are shown to be positive for limited angular ranges (Fig. 2.3a), for both vertical and 60° fault dips. Under such conditions, the vertical fault will fail in pure strike slip, and the dipping fault will fail in oblique slip in the rake directions shown in Fig. 2.3b. Where the ratio of the resolved fault-plane horizontal and down-dip tractions is zero, the dipping fault would fail in pure dip-slip. Where the ratio of the horizontal and down-dip components of the resolved shear stress on the dipping fault is \(\pm 1\), its slip would involve equal proportions of dip-slip and strike slip, indicating, for normal slip, a rake of -135° or -45°. As illustrated in Fig. 2.3, under the specified
Figure 2.3. (a) Normalized Coulomb stress acting on vertical (solid) and 60° dipping (dashed) faults for all strikes \( \alpha \) (counterclockwise positive) relative to the direction of \( S_{Hmax} \). The range a marked on the left-hand side of the figure gives \( \alpha \) for which a vertical fault is favored for right-lateral slip. The range b gives \( \alpha \) for which a 60° dipping fault is favored to fail. The range c gives \( \alpha \) over which composite failure of both faults would be favored. (b) The rake direction, \( \lambda \), of the maximum shear traction on the 60° dipping fault planes. For \( \alpha > 0 \), there would be oblique right-lateral and normal slip. For \( \alpha < 0 \), there would be oblique left-lateral and normal slip. The region between the vertical dashed show where the Coulomb stress is positive and failure is likely within the defined remote stress field.
conditions a vertical fault with $\mu' = 0.5$ will fail left-laterally if its strike satisfies $-47^\circ < \alpha < -17^\circ$ with a varying rake shown in Fig. 2.3b.

The combination of fault strikes where both the vertical and dipping faults are both favored to fail is more limited. These are combinations for which the regional stress would favor independent ruptures in a composite pattern. For vertical faults that strike at angle $\alpha$ relative to $S_{H\text{max}}$ and 60°-dipping faults that strike at angle $\beta$ (counterclockwise positive) relative to the vertical faults (Fig. 2.4a), we determine the combinations of $\alpha$ and $\beta$ that favor composite rupture. We examine different values of $\mu'$ and different ratios of $S_v$ to $S_{H\text{max}}$.

We restrict our analysis to the case where the shear traction on the vertical fault favors right-lateral slip (i.e., to $0^\circ < \alpha < 90^\circ$), given the sense of faulting in northwest Turkey. The shaded regions in Fig. 2.4b define the combinations of $\alpha$ and $\beta$ for which the Coulomb stresses are positive on both the vertical and dipping faults. For the case of $\mu' = 0.5$ and $S_v / S_{H\text{max}} = 1$ (first panel in Fig. 2.4b), $\alpha$ for composite failure ranges from $17^\circ$ to $47^\circ$ (as before) and $\beta$ has a $74^\circ$ range with limits that vary with $\alpha$. For example, this solution is thus consistent with right-lateral slip on a vertical fault striking between N77°W and S73°W for a maximum principal stress oriented N60°W. The boxed diagrams in Figure 2.4c below each case show the ranges of $\beta$ for $\alpha = 30^\circ$ (the optimal case for right-lateral shear failure on the vertical fault). Increasing $\mu'$ to from 0.5 to 0.55 keeping other parameter values constant causes the range of both $\alpha$ and $\beta$ to decrease (see second panel of Fig. 2.3b), as does decreasing the ratio $\sigma_v / \sigma_{H\text{max}}$ from 1 to 0.9 or 0.8 (third and fourth panels). In each case, oblique normal faulting is most favored on faults of two orientations diverging clockwise from the strike-slip fault. The dipping fault closer in strike to the vertical fault will have right-lateral oblique normal slip but, because it is oriented clockwise relative to $S_{H\text{max}}$, the more strongly oblique fault would have left-lateral oblique slip.

*Application to the Karadere Fault*

We now compare our analysis with the regional stress conditions and fault orientations in northwest Turkey. For the N35°W orientation of $S_{H\text{max}}$ (Gürbüz et al.,
Figure 2.4.  (a) The vertical strike-slip fault is oriented $\alpha$ degrees (counterclockwise positive) from $S_{H\text{max}}$ and the strike of the dipping fault is oriented $\beta$ from the vertical fault. The results are independent of the fault dip direction. (b) Four cases of composite slip potential with right-lateral shear stress on the vertical fault and varied values of $\mu'$ and $S_v / S_{H\text{max}}$. Shaded regions define the combinations of $\alpha$ and $\beta$ for which the Coulomb stresses are positive on both the vertical and dipping faults. (c) For each case in (b), the dipping faults with strike directions within the gray regions have positive Coulomb stress. The thin lines show the strike directions of the most-favored dipping faults (Coulomb stress is a maximum) oriented $\beta_{\text{max}}$. In each case, the $b_{\text{max}}$ orientation rotated clockwise relative to $S_{H\text{max}}$ is left-lateral / normal oblique slip and the orientation rotated counterclockwise is right-lateral / normal oblique slip.
our solution is consistent with right-lateral slip on a vertical fault striking between N52°W and N82°W (for 17° < α < 47°). For the N82°W limit, which is closest to the observed strike of the NAFZ, the 60°-dipping fault with strike between limits of N72°W and N2°E is most favored to slip. These limits do not correspond to the observed N60°E strike of the Karadere fault (Fig. 2.1). The Karadere fault is not favored to slip because it is oriented subperpendicular to the N35°W assumed direction of the maximum principal stress. This result implies that the regional stress field, if homogeneous and with the orientation and ratio of principal stresses that we have assumed, would discourage rupture of the Karadere fault during the İzmit earthquake. One possible explanation for this apparent contradiction is that the orientation and/or relative magnitude of the principal stresses varies in the vicinity of the Karadere fault segment. In a kinematic analysis of compressional fault stepovers, Westaway (1995) suggests spatial variation of the principal stress field in the vicinity of a stepover such that the maximum principal stress rotates to be strongly oblique to, or even subperpendicular to, the strike of the secondary fault within the stepover. Spatial variation of the local stress field in this manner; however, would also not promote failure of the Karadere fault because the local orientation of the most compressive stress still prohibits slip on the fault. Another explanation is that the perturbed stress magnitudes and directions associated with coseismic slip on the western İzmit earthquake segments triggered the Karadere segment to failure. We will investigate this in the next major section of this paper.

**Static stress triggering**

We now test the effects of stress triggering on composite rupture. We examine the changes in stress on an obliquely-oriented secondary fault due to slip on a vertical master fault, with different fault intersection geometries. This test is set up as a simplified investigation of rupture along the Karadere fault segment in the 1999 İzmit earthquake (Fig. 2.1). We use the numerical boundary element code Poly3D (Thomas, 1993) to calculate the quasi-static stress fields in a linear elastic, homogeneous and isotropic half-space. Quasi-static means that the solution for the elastic boundary-value problem that is calculated by our code does not consider the time-dependent, or dynamic, stresses that are associated with the propagation of seismic waves during earthquake
rupture, just the effects resulting from the total fault slip on each segment. Fault surfaces embedded in this model half-space consist of planar, triangular elements composed of superposed angular dislocations (cf. Comninou and Dundes, 1975). The displacement discontinuity, or fault slip, is constant on each element, but multiple elements can be used to model an arbitrary number of mechanically-interacting fractures or faults with non-uniform slip distributions (Crouch and Starfield, 1983).

This modeling approach differs from kinematic models of fault stepovers (Westaway, 1995) in that the only geometric assumptions in our model are the initial geometry of the faults, and the orientation of the regional stress. Each element in the model considers the stress perturbation associated with slip on all other elements in the model; the stress, strain, and displacement fields after our model faults slip are thus heterogeneous, being constrained only by the equilibrium and compatibility equations of linear elasticity and the specified elastic moduli (cf. Barber, 1992). Elastic solutions have previously been used to predict coseismic slip from surface deformation (Bürghmann et al., 2002; Feigl et al., 2002), the location of triggered large earthquakes (Stein, et al., 1997) and aftershocks (King et al., 1994). However, linear elasticity theory restricts our model to cases in which the fault slip is small compared with the fault length and to timescales that are short compared with the Maxwell time of the deepest lower crust. We therefore restrict our study to coseismic slip and deformation, and do not address the structural evolution of the fault stepovers over interseismic or geologic time scales (cf. Westaway, 1995) where time-dependent material rheology must be considered. Time-dependent deformation could produce a heterogeneous pre-stress condition on faults composing a stepover, which could affect the propensity for rupture of obliquely-oriented faults. However, to isolate the effects of fault orientation in coseismic rupture, we restrict ourselves to the case of uniform regional stress loading on the faults.

The model vertical master fault is 100 km long (Fig. 2.5a), approximating the length of the E-W trending Gölcük, Sapanca, and Sakarya rupture segments of the İzmit earthquake, west of the Karadere fault. The width, or down-dip depth limit, of the master fault is 15 km, corresponding to the depth limit of İzmit aftershocks. The obliquely-oriented secondary observation "fault" consists of a line of grid points at which the stress tensor is calculated after slip on the master fault. The three en echelon segments of this
Figure 2.5. (a) A model fault intersection configuration in which the angle $\beta$, the difference in strike between master and secondary fault segments, measured counterclockwise, is 30°. The stress tensors are calculated at nodes (model node spacing is 1 km) on the secondary fault and then transformed into fault-parallel and fault-normal coordinates to calculate Coulomb stress changes for different assumed fault dips. (b) Predicted Coulomb stress change (black) and rake (gray) for dips of 30° (dashed lines) and 90° (solid lines) for the secondary fault oriented with $\beta = 30°$, with $\alpha = 30°$. Rake values $\pm 180°$ indicate a maximum shear stress direction favoring pure right-lateral slip on the secondary fault. Rake values of 90° and -90° corresponds to pure reverse slip and pure normal slip, respectively.
master fault (shown schematically in Fig. 2.1b) are modeled as one fault for two reasons. First, aftershock locations near the Gölcük stepover (between the Gölcük and Sapanca segments; Fig. 2.1b) are too diffuse to clearly define whether or not this stepover fault structure extends to significant depths (Özalaybey et al., 2002). Second, small stepovers near Sapanca and Gölcük would have minimal effect of the stress field near the eastern tip of the Akyazı fault, due to their large distance from this point.

Model faults are assumed to have a boundary condition of zero shear traction and zero fault-normal displacement discontinuity. This produces a complete stress drop on the model faults. While stress drops calculated for earthquakes indicate that coseismic ruptures typically do not relieve all of the shear stress in the model, we believe that our endmember frictionless fault assumption will nonetheless provide insight into the kinematics of rupture along obliquely-oriented faults. The complete stress drop criterion means that the calculated magnitudes of the shear displacement discontinuities do not depend on the magnitudes of the normal tractions acting on the fault surface.

The remote direction of maximum principal stress in this model is oriented 30° clockwise from the master fault strike (i.e., toward N60°W given the E-W fault strike), this choice being based on the optimal direction for failure when \( \mu = 0.6 \), although this is not in accordance with the results of Gürbüz et al., (2000). The effective vertical remote stress, \( S_v \) of 109.5 MPa is applied, to account for the effective overburden above the 7.5 km depth of the observation points. The maximum and minimum horizontal remote stresses are set to \( S_h \pm 3.5 \) MPa (i.e., 106 MPa and 113 MPa, respectively). This magnitude of differential stress is applied so that the frictionless slip computed on the model master fault produces a comparable seismic moment to that observed in the İzmit mainshock. Under this horizontal differential stress of 7 MPa, the calculated average slip on the master fault is 3.1 m. Given the area of this model master fault (1500 km²; i.e., 100 km × 15 km) and a typical shear modulus of 30 GPa, 3.1 m of average right-lateral slip provides the estimated scalar seismic moment of \( 1.4 \times 10^{30} \) Nm for the İzmit mainshock (Toksoz et al., 1999). The magnitudes of the regional differential stresses do not affect our calculations of stress conditions on the secondary faults because on those faults we are calculating the Coulomb stress changes due to master faults slip and not the absolute stress magnitudes. For stress changes to trigger slip requires that the stress
perturbation is of sufficient magnitude to exceed the strength of the secondary fault. Stress changes of the order of 10 to 100 kPa appear to trigger earthquake aftershocks (Reasenberg and Simpson, 1992; King et al., 1994). In these cases, the magnitude of the stress trigger is much smaller than the stress drop in each aftershock, suggesting that the faults on which the aftershocks occur are already close to their failure threshold (Harris, 1998). In our study, we focus on how variations in fault orientation affect the trigger magnitudes without specifically defining if the threshold of failure has been reached in each case.

Stress changes are examined for model secondary faults, with a range of dips, striking at $\beta = \pm 30^\circ$ relative to the master fault. Black lines in Fig. 2.5b show the Coulomb stress change on a secondary fault with $\beta = +30^\circ$ (i.e., in the contractional quadrant of the right-lateral master fault), caused by slip on the master fault. Gray lines show the direction of the maximum resolved shear stress on the secondary fault plane, indicating the rake with which this secondary fault is predicted to slip if its Coulomb failure condition is exceeded as a result of this stress perturbation. For a vertical model secondary fault (solid lines), the predicted Coulomb stress changes are positive throughout its length, exceed 0.4 MPa, and act in a direction to favor predominantly right-lateral slip, with $-180^\circ < \lambda < -170^\circ$. For secondary faults that dip at $30^\circ$ (dashed lines), the Coulomb stress changes are also positive, but are greater close to the fault intersection than for the vertical fault case. The sense of triggered slip predicted for this secondary fault dip varies from indicating similar-magnitude components of reverse and right-lateral strike slip near the fault intersection to predominantly strike-slip farther from this intersection.

Figure 2.6a shows that for $\beta = 30^\circ$ the predicted Coulomb stress change, spatially averaged across the 40 km long secondary fault, increases for shallower dips of this secondary fault. A $30^\circ$ dipping secondary fault is thus nearly 1.5 times as likely to experience triggered slip as a vertical secondary fault. The predicted slip sense on this secondary fault also involves an increasing reverse slip component for decreasing dip (Fig. 2.6b).

For a secondary fault in the extensional quadrant of the right-lateral master fault ($\beta = -30^\circ$) (Fig. 2.7), the Coulomb stress changes for vertical and $30^\circ$ dipping secondary
Figure 2.6. (a) Spatially-averaged percentage change in Coulomb stress for a secondary fault with different dips relative to a vertical secondary fault for $\beta = 30^\circ$. (b) Spatially-averaged rake along a secondary fault with different dips.
Figure 2.7. (a) A model fault intersection configuration in which the angle $\beta$ is $-30^\circ$. (b) Predicted Coulomb stress change (black) and rake (gray) for dips of $30^\circ$ (dashed lines) and $90^\circ$ (solid lines) for the secondary fault oriented with $\beta = -30^\circ$, with $\alpha = 30^\circ$. 
faults are positive, but larger than in the case with $\beta = 30^\circ$. However, unlike the previous case, the maximum Coulomb stress change is greater on the vertical secondary fault than on the $30^\circ$ dipping secondary fault. The predicted sense of triggered slip is again predominantly right-lateral for the vertical secondary fault, whereas for the $30^\circ$ dipping secondary fault a combination of right-lateral and oblique normal slip is triggered. Decreasing the dip of the secondary fault results in reduced predicted changes in Coulomb stress (Fig. 2.8a) and increased predicted components of normal slip (Fig. 2.8b).

These results suggest that triggered slip on a secondary fault in the contractional quadrant of the master fault is favored more for secondary faults that are dipping than for those that are vertical. The opposite is the case if the secondary fault lies in the extensional quadrant (Table 2.1). This conclusion is supported by the rupture geometries of the Songpan and Hector Mine earthquakes (Fig. 2.1a,c). In both these cases, the rupture was able to propagate through, and well beyond, the oblique segment. For the Songpan earthquake (Fig. 2.1a), the secondary fault was located in the contractional quadrant of the master fault, had a relatively low-angle dip, and rupture was able to propagate through this secondary fault. For the Hector Mine earthquake (Fig. 2.1c), the secondary fault was located in the extensional quadrant of the master fault, had a vertical dip, and rupture was able to propagate through this secondary fault. For the Izmit earthquake, the secondary fault (the Karadere segment) was located in the contractional quadrant of the master fault, had a steep (~70-80°) dip, and rupture was able to propagate past this secondary fault for several kilometers only. Our results are consistent with this observation: they suggest that this Karadere segment does not have the optimum geometry for triggered slip; potentially explaining why slip in the Izmit earthquake did not propagate more than several kilometers further east.

**Effects of slip interaction**

The stress triggering analysis in the previous section describes the stress loading conditions on secondary faults that intersect with master faults. We will now characterize the slip that would occur in response to those stresses. The slip computed on the master and secondary faults in the model occurs in response to the regional stress applied in the
Figure 2.8. (a) Spatially-averaged percentage change in Coulomb stress for a secondary fault with different dips relative to a vertical secondary fault for $\beta = -30^\circ$. (b) Spatially-averaged rake along a secondary fault with different dips.
Table 2.1. Summary of stress triggering model results.

<table>
<thead>
<tr>
<th>Secondary fault orientation</th>
<th>30° counterclockwise from master fault</th>
<th>30° clockwise from master fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quadrant if master fault is right-lateral</td>
<td>Contractional</td>
<td>Extensional</td>
</tr>
<tr>
<td>$\sigma_c$ change with dip</td>
<td>Increases for shallower dips</td>
<td>Decreases for shallower dips</td>
</tr>
<tr>
<td>Triggered slip sense</td>
<td>Right-lateral with increasing reverse component for shallower dips</td>
<td>Right-lateral with increasing normal component for shallower dips</td>
</tr>
<tr>
<td>Vertical fault assumption</td>
<td>Underestimates magnitude of triggering effect</td>
<td>Overestimates magnitude of triggering effect</td>
</tr>
</tbody>
</table>
stress triggering analysis in section 3. These slip distributions will be heterogeneous along the fault planes as they interact with each other to mutually satisfy their prescribed traction boundary conditions. Predictions of different magnitudes of slip on secondary faults will be used to infer the likelihood of whether slip can propagate through the structure or is terminated by it, and predicted directions of slip on the secondary fault will indicate the type of mechanism any triggered secondary event will have. We again consider the transfer of slip through the oblique fault intersections of the 1999 İzmit earthquake rupture.

Quasi-static slip calculations have previously been used to show how parallel discontinuous faults interact and to predict the development of secondary structures in the intervening relay zones (cf. Segall and Pollard, 1980). The same numerical procedure as is adopted in our study has also previously been used to calculate slip distributions on intersecting normal faults (Maerten et al., 1999), to calculate paleostress directions from the deviation of slickenlines on interacting normal faults (Maerten, 2000), and to predict fault patterns in conjunction with seismic reflection data (Maerten et al., 2000). In this model there is no temporal sequence of slip—all faults slip simultaneously to satisfy the no normal displacement and zero shear stress boundary condition imposed on all faults.

The model set-up for our generic intersection tests is identical to that for the Coulomb stress tests (e.g., Fig. 2.5a) except the line of observation points in the previous tests is replaced by a model fault that is permitted to slip (Fig. 2.9a). We specify the strike of this secondary fault as 30° counterclockwise from the master fault, and investigate secondary faults with dips in the range from 30° to 90°. The length of the model master fault at its mid-depth is 100 km, with its upper and lower edges being longer and shorter, respectively, to provide a clean intersection with the secondary fault.

The entire model secondary fault slips in a right-lateral reverse oblique sense due to its orientation relative to $S_{\text{imax}}$. However, there is a steep slip gradient on the secondary fault as a result of fault interaction (Fig. 2.9b). The greatest slip occurs past the intersection line (within the contractional quadrant of the master fault) on the upper part of the secondary fault; slip is relatively diminished behind the intersection line (within the extensional quadrant of the master fault). At all points on the secondary fault the slip is reverse-oblique right-lateral with a $\sim 5^\circ$ change in rake toward increased reverse slip.
Figure 2.9. (a) Slip interaction model set-up. $S_Y = 109.5$ MPa, $S_{H_{\text{max}}}$ and $S_{\text{hmin}}$ are $S_Y \pm 3.5$ MPa, respectively. For this test, $S_{H_{\text{max}}}$ is oriented at $\alpha = 30^\circ$. Within this stress field the master fault has an average right-lateral slip of 3.1 m. Figures 2.11, 2.12, and 2.14 show the slip components along the observation line at the mid-depth of the secondary fault. (b) Slip magnitude and directions on secondary fault striking $30^\circ$ counterclockwise from master fault and dipping $60^\circ$ towards this master fault. Arrows indicate amount and direction of motion of far (north) side of fault relative to near side. Positions $>0$ km correspond to parts of the secondary fault beyond the end of the master fault. Positions $<0$ km correspond to parts of the secondary fault forming an acute angle with the master fault on the other side of their intersection line. Slip contour interval is 0.5 m.
past the fault intersection. The greatest displacement of material (not shown) in the model occurs along the north (far) side of master fault in Figure 2.9a and along the north (far) side of the secondary fault to the east (right) of the fault intersection line. Likewise, the smallest displacements occur in the acute angle overlap region between the two faults; the particular geometry of the faults in this region does not promote dextral slip.

The entire model secondary fault slips in a right-lateral reverse oblique sense due its orientation relative to $S_{Hmax}$. However, there is a steep slip gradient on the secondary fault as a result of fault interaction (Fig. 2.9b). The greatest slip occurs past the intersection line (within the contractual quadrant of the master fault) on the upper part of the secondary fault; slip is relatively diminished behind the intersection line (within the extensional quadrant of the master fault). At all points on the secondary fault the slip is reverse-oblique right-lateral with a $\sim 5^\circ$ change in rake toward increased reverse slip past the fault intersection. The greatest displacement of material (not shown) occurs along the north (far) side of master fault in Figure 2.9a and along the northwest (far) side of the secondary fault to the east (right) of the fault intersection line. Likewise, the smallest displacements occur in the acute angle overlap region between the two faults; the particular geometry of the faults in this acute angle region does not promote dextral slip. This suggests that the part of the secondary fault to the left of the fault intersection line would be unlikely to slip with the addition of frictional resistance on the faults.

Figure 2.10 shows how the components of right-lateral and reverse slip, and the associated rake, vary along the mid-depth line of the secondary fault with $\beta = 30^\circ$ as a function of its dip. The slip distributions for a secondary fault that dips at $90^\circ$ and slips in response the regional stress field alone is also shown for comparison; there is no dip-slip and the right-lateral slip has a smooth elliptical profile. Both the right-lateral and reverse components of slip are greater beyond the intersection line with the master fault (from 0 to 40 km) than in the region behind this intersection (from -40 to 0 km). The dip of the secondary fault does not have a significant effect on the strike-slip component, but the magnitude of dip-slip substantially increases for shallower secondary fault dips.

Figure 2.11 shows that the character of the interaction between intersecting faults also depends on their difference in strike. For this test the secondary fault is assumed to be vertical, and striking at $\pm 15^\circ$, $\pm 30^\circ$, and $\pm 45^\circ$ relative to the vertical master fault.
Figure 2.10. Slip along secondary fault as a function of its dip. (a) Model configuration. (b) Right-lateral strike-slip component. (c) Dip-slip component. (d) Rake, 180° indicating pure right-lateral slip; 90° indicating pure reverse slip. Line styles and labels correspond to different dips of the secondary fault (in °).
Figure 2.11. Slip on vertical secondary fault for different strikes, β. Slip values along observation line. (a) Strike-slip component. (b) Dip-slip component.
Changes in strike of >45° are uncommon for significant adjacent rupture segments along the North Anatolian fault (e.g., Barka and Kadinsky-Cade, 1988) so we do not test past that range. For vertical strike slip faults (Fig. 2.11), the regional stress field promotes right-lateral slip on the master fault and slip on the secondary fault that is right-lateral and reverse for positive (counterclockwise; \(\beta > 0\)) changes in fault strike. For negative changes in fault strike (clockwise; \(\beta < 0\)), vertical secondary faults with \(|\beta| < \alpha\) have right-lateral normal oblique slip and faults with \(|\beta| > \alpha\) have left-lateral normal oblique slip. For increasing positive strike changes, the magnitude of the predicted right-lateral slip decreases and that of the reverse component increases. The fault interaction increases the right-lateral component, or decreases the left-lateral component, as one moves along the secondary fault past the intersection point (towards +40 km) relative to sites behind the intersection point (towards -40 km). In all cases the dip-slip component increases behind the intersection point due to the fault interaction.

On a 60° dipping secondary fault striking 30° clockwise to the master fault (Fig. 2.12), the slip again increases ahead of the intersection line and decreases behind it. In this case, the slip vectors show predominantly normal slip. This occurs because this secondary fault is parallel to \(S_{\text{Hmax}}\), reducing the predicted relative magnitude of the horizontal resolved shear stress. The reduced shear stress contribution from the regional stress enables the master fault to induce a minor component of right-lateral slip on the secondary fault to the right of their intersection line and oblique slip with roughly equal components of left-lateral and normal slip to the left of it. For 60° dipping secondary faults with a range of strikes relative to the master fault (Fig. 2.13), the strike-slip component of slip behaves as in the vertical case (Fig. 2.11). However, the dip-slip component is increased for all strike changes relative to that for a vertical secondary fault. If it strikes at +15° relative to the master fault the secondary fault is predominantly right-lateral, but it is nearly pure reverse dip-slip for a +45° strike change. However, the dip-slip component varies little for negative changes in strike.
Figure 2.12. (a) Slip interaction model set-up. Slip magnitude and directions for secondary fault oriented $\beta = 30^\circ$ clockwise from the vertical master fault and dipping $60^\circ$ towards master fault.
Figure 2.13. Slip on 60° dipping secondary fault for different strikes, $\beta$. (a) Strike-slip component. (b) Dip-slip component.
Discussion

Implications for the Karadere segment

The Karadere fault segment can be regarded as forming a double bend relative to adjacent segments of the NAFZ, as it strikes oblique to these E-W trending fault segments (Fig. 2.14b). The fault on which this Karadere rupture segment occurred continues for several kilometers beyond either end of the 1999 rupture, as evidenced by the topography (Fig. 2.14a), the distribution of aftershocks (lio, et al., 2002), and field mapping (Emre et al., 2003). Indeed, Barka et al. (2002) report 20 cm of right-lateral offset and surface cracking along the northeast extension of the fault 5 km past the termination of the main rupture segment. Whether or not these northeast and southwest extensions of the Karadere fault are presently active, it is compelling to note that the section of the Karadere fault that slipped is confined to the region between projected intersections with the neighboring rupture segments of the Izmit earthquake (segments 4 and 6 in Fig. 2.14b). This suggests the potential for mechanical interaction among the fault segments. The relative timing of failure of each of the three fault segments is not discernable from seismic records, and the rupture configuration is somewhat more complicated than the master fault / oblique secondary fault configurations that we investigated earlier. Nevertheless, our modeling approach allows us to investigate how slip may be partitioned among the three fault segments and how the combined effect of two fault intersections may have effectively aided in constraining rupture of the Karadere fault to only limited central portion of its length.

The distribution of right-lateral slip in Fig. 2.14c compares the predicted slip on the Karadere rupture segment when the secondary fault segments (segments 4 and 6) are (dashed line) and are not (solid line) included. In both cases, all faults are assumed to slip without friction in response to the same remote stress field (with maximum principal stress assumed oriented N60°W) applied in sections 3 and 4. The E-W trending Gölcük, Sapanca, and Sakarya segments are also included in this model. The lack of fault friction assumed in this model means that normal stress changes do not affect fault slip; therefore, our interpretations are qualitative, and would be affected if the faults were not assumed to be frictionless. In both cases all fault segments slip right-laterally. Inclusion of the secondary faults 4 and 6 enhances the right-lateral slip along the central portion of the
Figure 2.14. (a) Digital elevation map of Karadere region with Izmit ruptures segments shown as black lines and the Düzce rupture as a gray line. (b) Configuration of surface rupture segments of the 1999 Izmit, Turkey earthquake, adapted from Aydin and Kalafat (2002). Karadere segment is labeled as segment 5. Dotted box shows region for (a). (c) Right-lateral slip on segment 5 for models that include segments 1-6 (dashed) and include segments 1, 2, 3 and 5 (solid). Slip values normalized by maximum right-lateral slip when segments 4 and 6 are not included. Vertical lines delineate the intersections with segments 4 and 6. (d) Difference in slip between the two models in (c).
Karadere fault (Fig. 2.14c dashed line) and reduces the right-lateral slip near its ends. However, in both cases, the entire Karadere fault slips, including the parts that are inferred to exist beyond the E-W-striking faults at both ends of the observed rupture segment (Fig. 2.14d). This model thus demonstrates that slip can be partitioned by oblique intersections between faults, being enhanced between the fault intersections and diminished outside them. However, this model is unable to reproduce the observation of zero slip during the İzmit earthquake beyond these intersection points, which as already noted may be due to the assumption that faults are frictionless.

Possible future rupture in the Sea of Marmara

Characterizing the effects of pre-existing fault geometry on the propagation of coseismic rupture is a critical question in northwest Turkey where the North Anatolian fault branches into a complex pattern of faults oriented obliquely relative to each other (cf. Armijo et al., 1999; Bozkurt, 2001). By analogy with the İzmit rupture, large earthquakes in the Sea of Marmara are likely to involve multiple rupture segments. The complex structure of the Sea of Marmara indicates long-term interaction between obliquely oriented faults, revealed for instance by the Çınarcık, Central Marmara, and Tekirdağ basins while, overall, the entire Marmara Sea pull-apart basin has accommodated an estimated 85 km of right-lateral slip (cf. Westaway, 1994; Armijo et al., 1999). However, the question remains which of the faults reported within the Sea of Marmara will rupture in potential future earthquakes.

Interpretations of the fault geometry within the Sea of Marmara differ (cf. Le Pichon et al., 2001; Armijo et al., 2002; Sato et al., 2004; Seeber et al., 2004; Okay et al., 2004). However, regardless of these differences, it seems clear (Fig. 2.15) that any rupture of significant length (>70 km) in the Sea of Marmara will have to rupture through intersections between faults whose strikes and dips differ by up to 30°. The instrumental seismicity supports this interpretation in that while there has been predominance of normal and strike-slip events for $M_s \leq 6.7$, only strike-slip events have occurred at greater magnitudes (Fig. 2.16).

Pending a more focused investigation, we can use the results of the present study to make some tentative inferences about potential future fault rupture characteristics in
Figure 2.15. Fault map of the Marmara Sea adapted from Armijo, et al. (2002). Thick segmented lines represent potential rupture scenarios at the western termination of the 1999 Izmit earthquake. These faults intersect the Northern Boundary fault, Inner Boundary fault, and Southern Boundary fault forming oblique fault intersections.
Figure 2.16. Rake versus magnitude for earthquakes $M_s > 5$ in the Sea of Marmara region since 1943. Focal mechanisms from Eyidogan (1988). For events labelled 1, moment magnitudes, $M_w$, have been used rather than $M_s$. 
the eastern Sea of Marmara, using the assumption (as before) that this region’s state of stress involves a maximum principal stress oriented N35°W. Although the transition of slip from the Sakarya segment to the Karadere segment in the İzmit rupture does not seem to be favored within the observed regional stress field, we do predict favorable composite rupture fault geometries elsewhere in northwest Turkey. For example, at the transition in the eastern Sea of Marmara from the vertical E-W-striking İzmit fault to the SSW-dipping fault running along the northern margin of the Çınarcık basin to central Marmara fault between the Çınarcık and Central Marmara Basins (Armijo et al., 2002; İmren et al., 2001; Le Pichon et al., 2001; Sato et al., 2004). Similarly, the vertical fault segments approaching the Central Marmara Basin from either direction intersect with dipping basin-bounding faults within their extensional quadrants.

The mapping by Okay et al. (2000) and Armijo et al. (2002) shows three major fault segments, oblique to the general E-W trend of the NAFZ, which could accommodate rupture propagating westward from the western end of the İzmit earthquake rupture. Using the nomenclature of Okay et al. (2000) these are, from north to south (Fig. 2.15): the Northern Boundary fault (NBF) south of the Princes’ Islands, which strikes N60°W and dips at 65° toward S30°W; the Inner Boundary fault (IBF) along the southern margin of the Çınarcık Basin, which strikes N76°W and dips at 70° toward N14°E; and the Armutlu fault or Southern Boundary fault (SBF), which strikes S68°W and dips at 85° toward N22°E. The Coulomb stresses resolved on these faults due to the regional stress field suggest that the Northern Boundary fault is the most favored for failure \( \sigma_c / (S_{Hmax} - P_p) = 0.05 \), followed by the Inner Boundary fault \( \sigma_c / (S_{Hmax} - P_p) = 0.01 \) while the South Boundary fault \( \sigma_c / (S_{Hmax} - P_p) = -0.33 \) is not favored as it is vertical and almost parallel to \( S_{Hmax} \). These tentative conclusions rely on the regional stress field orientation determined by Gürbüz et al. (2000), and would be subject to modification following any future improved stress inversion.

The potential for rupture of these obliquely-oriented fault segments can also be qualitatively evaluated in terms of static stress changes associated with the western termination of the İzmit rupture. If rupture were to initiate west of the İzmit rupture and propagate westward onto the NBF (thick dotted line in Fig. 2.15), an immediate clockwise bend of 30° would be required. Like other studies (Pinar et al., 2001; Çakir et
al., 2003), we infer that the NBF is well-oriented for triggered right-lateral strike-slip failure. If the rupture was instead to propagate along the Yalova and Armutlu fault segments and then branch onto the IBF (thick dashed line in Fig. 2.15) or SBF (thick dot-dashed line in Fig. 2.15), clockwise bends of 25° and 45°, respectively, would be required. Our stress triggering analyses suggest that the Coulomb stresses due to rupture approaching the bends from the east would promote failure for all three strike and dip changes, with the greatest change corresponding to the 45° bend onto the SBF.

If rupture were instead to initiate on the Central Marmara fault (CMF) and propagate eastward, the oblique intersection with the NBF would be favorable for composite rupture, as suggested by the results plotted in Figure 2.7 that approximate this geometry. If the rupture were to propagate west on the CMF there are several fault branches diverging clockwise to the north from this fault that are also available for rupture branching into the extensional quadrant of the CMF.

Static versus dynamic stress

Field evidence shows that fault discontinuities and oblique intersections have consistent geometries across a wide range of scales, from small outcrop-scale sheared joints to large earthquake-generating fault systems (e.g., Aydin and Nur, 1982; Mann et al., 1983; Sibson, 1986). The development of these fault discontinuities, or the triggering of slip along them, has been successfully examined using static stress analyses (e.g., Aydin and Du, 1995; Aydin and Schultz, 1989; Segall and Pollard, 1980; Simpson and Reasenberg, 1994). In terms of earthquake faults, recent large ruptures (e.g., Songpan, 1976; Landers, 1992; Izmit, 1999 and Hector Mine, 1999) show that rupture jogs or branches select pre-existing faults, and that the geometries of these pre-existing fault patterns are typically consistent with static stress predictions for fault interactions. Therefore, while dynamic effects may impact near-fault rupture processes (Poliakov et al., 2002) and aftershock distributions (Kilb et al., 2000), rupture branching onto secondary oblique fault segments seems to be primarily controlled by fault intersections and terminations that have developed under longer-term static stress loading. We believe that our static analyses provide a useful simple tool for identifying the most favorable...
composite rupture paths along a pre-existing fault configuration based on the regional stress field and the stress perturbation associated with previous earthquakes.

Conclusions

The geometry of oblique fault intersections affects the mechanics of composite earthquake rupture in a strike-slip environment. The feasibility of composite rupture for a given regional stress state depends on the obliquity of the secondary fault (the difference between its strike and that of the master strike-slip fault), its dip, the strengths of the faults (i.e., their coefficients of friction), and the relative magnitudes of the vertical and horizontal principal stresses. Composite rupture is most favored as a consequence of the regional state of stress in northwest Turkey when a vertical master fault strikes at an angle of \( \tan^{-1}(1/\mu')/2 \) relative to \( S_{H_{\text{max}}} \) and a dipping secondary faults strikes clockwise relative to the adjacent master strike-slip fault. The range of strikes for failure along a secondary fault decreases for increasing \( \mu' \), the effective coefficient of friction on a fault, and for decreasing \( S_v / S_{H_{\text{max}}} \). For each value of \( \alpha \), the angle between the maximum horizontal principal stress and the strike of the master fault, there are two optimal strikes of 60° dipping secondary faults, \( \beta_{\text{opt}} \), which take up right- or left-lateral oblique normal slip. For increasing \( \mu' \), the range of \( \beta \) for composite failure decreases and the values of \( \beta_{\text{opt}} \) for right- and left-lateral oblique failure approach \(-\alpha\). For decreasing \( S_v / S_{H_{\text{max}}} \) the range of \( \beta \) for composite failure reduces and separates into two different ranges of \( \beta \) for right- and left-lateral failure.

Coulomb stresses resolved on secondary fault segments due to slip along a vertical master fault vary according to the difference in strike of these faults and the dip of the secondary fault. For a counterclockwise change in strike from the master fault to the secondary fault, oblique right-lateral reverse slip is promoted. In addition, a dipping secondary fault will fail more readily than a vertical secondary fault. For a clockwise change in strike, oblique right-lateral normal slip on the secondary fault is facilitated and a vertical secondary fault will fail more readily than a dipping one. Mechanical interaction between intersecting faults during slip may produce partitioning of slip on either side of the fault intersection line. Rupture of the Karadere segment in the İzmit earthquake, which strikes obliquely to the main strike-slip fault segments that also
ruptured, may have been facilitated by mechanical interaction across fault intersections. A preliminary assessment suggests that some of the oblique intersection geometries between faults in the Sea of Marmara may favor composite rupture.

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Chapter 3

Using geomechanical modeling to constrain fault geometries proposed for the northern Marmara Sea

Abstract

In recognition of elevated seismic hazard potential along the North Anatolian fault within the Marmara Sea, geophysical surveys have recently been conducted to collect high-resolution bathymetry and several seismic reflection data sets. Despite the increased quality and quantity of data, fault interpretations by different authors are varied and a conclusive fault geometry within the Marmara Sea remains elusive. This paper presents a geomechanical modeling method for constraining the proposed fault geometry interpretations. In the model, crustal faults within the Marmara Sea slip in accordance with GPS-constrained slip rates along deep plate boundary dislocations. We test three fault models proposed by other authors and show that a fault interpretation with a series of pull-apart basin structures along a master strike-slip fault best produces the observed deformation pattern within the Marmara Sea. The locations of the basins along the northern Marmara trough and their relative subsidence magnitudes are well matched. The computed slip vectors on the faults in the best-fitting model indicate that, except for small faults bounding the Tekirdağ basin, the faults have dominantly right-lateral strike-slip, with rakes less than 15°. On the local scale, this suggests that interaction between the segmented strike-slip faults is the dominant mechanism for producing the observed vertical deformation in the Marmara Sea. On the regional scale, if the Marmara Sea faults behave in this manner over geologic time scales, then right-lateral plate boundary motion, without an additional component of north-south extension associated with the Aegean Sea, is sufficient to produce the observed Marmara Sea pull-apart basin morphology.

Introduction

On August 17, 1999 the largest and most damaging earthquake in Turkey in 60 years extended the historical progression of large earthquakes along the right-lateral, strike slip North Anatolian fault (NAF), to within 100 km of Istanbul, a city of 12 million
people (Barka, 1999). Since 1939, the 1500 km long NAF in northern Turkey (Fig. 3.1) has produced twelve $M > 6.7$ earthquakes that have propagated predominantly westward, with a cumulative rupture length of over 1200 km (Dewey, 1976; Toksöz et al., 1979; Barka, 1996). The $M_w = 7.4$ İzmit earthquake in August 1999 extended the western limit of rupture another 80 to 100 km west and into the Marmara Sea (Fig. 3.1).

The westward NAF earthquake progression through 1999, together with the 1912 $M_s = 7.4$ Saros earthquake on the Gelibolu peninsula (Fig. 3.1) (Ambraseys and Finkel, 1991), now define the Marmara Sea, recognized as a large composite pull-apart basin, as the only significant portion of the NAF that has not ruptured in the past 200 years (Reilinger et al., 2000). For such an obvious seismic gap, the fundamental question has shifted from whether or not an earthquake will occur to where and how large the next damaging earthquake will be. Researchers have proposed likely scenarios for a future Marmara Sea earthquake. Hubert-Ferrari et al. (2000) suggest that two events of magnitude equal to, or greater than the İzmit earthquake may be expected within the next decades within the Marmara Sea. Le Pichon et al. (1999) suggest that the possibility for the entire length of the Marmara Sea to rupture in a single event.

There are several outstanding issues that make such projections difficult. One is that the fault configuration within the Marmara Sea is still open to interpretation (Fig. 3.2). Models that have been proposed range from one continuous fault strand (Aksu et al., 2000; İmren et al., 2001; Le Pichon et al., 2001), to a zone with many subsidiary fault strands (Armijo et al., 2002), to en echelon patterns of smaller faults (Barka, 1992; Wong et al., 1995). A second issue is how the rupture geometry in a future event will depend on the present fault configuration and the interaction among its components. It may be likely that the future event would utilize discontinuous fault segments in en echelon configurations or at oblique orientation, especially within a region where the interaction between strike-slip and normal faults is so apparent.

Resolving the Marmara Sea fault geometry therefore remains of key importance. In terms of earthquake hazard, understanding the fault geometry in the Marmara Sea is essential because fault geometry can control the ability for rupture to propagate along multiple fault segments (Wesnousky, 1988; Harris and Day, 1999; Muller and Aydin, in press). In terms of geological development, understanding the fault geometry and its
Figure 3.1. (a) Tectonic map of the Anatolian peninsula and its vicinity (modified from Holzer, 2000). The NAF system is defined as one dominant trace before splitting into three strands west of Bolu. Note the general northeast-southwest trend of the NAF strands to the east and west of the Sea of Marmara, and the east-west trend of the NAF strands in the immediate vicinity of the Marmara Sea. Box shows area of lower map. (b) Faults in the Marmara Sea region (from Barka, 1999) and rupture traces for the most recent large earthquakes along the North Anatolian fault east of this region.
Figure 3.2. Active fault patterns proposed for the Marmara Sea. Interpretations with similar major fault locations have been combined in (a) and (b). (a) Fault interpretation of Okay et al. (2000) (solid lines) and Parke et al. (1999) (dashed where substantially differing from Okay et al., 2000). (b) Fault interpretation of Le Pichon et al. (2001) (solid lines) and Imren et al. (2001) (dashed where substantially differing from Le Pichon et al., 2001). (c) Fault interpretation of Armijo et al. (2002). Topography DEM derived from Gtopo30 global dataset. High-resolution bathymetry (Rangin et al., 2000) with depth scale next to (c). In (a), TB = Tekirdag basin; CMB = Central Marmara basin; KB = Kumburgaz basin; CB = Çinarcik basin; IB = Imrali basin.
related deformation may shed light on the structural evolution of the NAF within the Marmara Sea. In this study we use a numerical model to place mechanical constraints on the Marmara Sea fault geometry. We test different proposed fault models in terms of their ability to reproduce the observed seafloor geomorphology and the structural deformation pattern of the stratigraphic horizon defining the transition from pre-transform to syn-transform deposition at selected locations within the Marmara Sea. This technique enables one to discriminate on a mechanical basis among competing fault models, which otherwise may be equally justified in terms of bathymetric and seismic reflection data.

Marmara Sea composite pull-apart basin

For most of its 1500 km trace, the NAF follows a fairly linear small circle trace about the pole of rotation of the Anatolian plate (Reilinger et al., 1997). The major deviation to this trend occurs at the location of the Marmara Sea, where the NAF splits into three strands that locally trend roughly east-west (Fig. 3.1). The northern strand defines the northern margin of the Marmara Sea and may in itself be multiple segments that bound basins within the northern Marmara trough. West of the sea, the northern strand strikes more southwesterly as the Ganos fault and then exits the Gulf of Saros to bound a series of basins in the northern Aegean (Koukouvelas and Aydin, 2002). The central strand defines the southern margin of the Marmara Sea. The southern strand trends more southwesterly through Bursa before entering the Aegean Sea (Fig. 3.1).

GPS measurements indicate that the three branches combine to accommodate 20-25 mm/yr of dextral strike-slip motion of Anatolia relative to Eurasia (Reilinger et al., 1997; Straub et al., 1997). The northern zone is the most active with an approximate slip rate of 10-15 mm/yr (Straub et al., 1997). The region between the northern and southern branches shows 5-10 mm/yr of distributed deformation without well-defined slip along the central fault strand. The southern fault strand accommodates approximately 2-4 mm/yr of dextral slip.

The NAF fault strands to the east and west of the Marmara Sea region trend northeast-southwest, whereas the NAF within the Marmara Sea region trends east-west (Fig. 3.1). Slip along this regional fault configuration creates an extensional jog within
the Marmara region, producing local extension and subsidence and contributing to formation of the Marmara Sea composite pull-apart basin. On a local scale, the northern branch of the NAF within the Marmara Sea itself breaks down to define several fault-bound basins. Localized vertical deformation within a dominantly strike-slip faulting regime suggests a significant degree of interaction between the larger strike-slip fault segments and the subsidiary faults that locally bound the basins.

High-resolution bathymetry shows the Çınarcık, Central Marmara, and Tekirdağ basins are each 1100-1200 m deep and are separated by two shallow sills reaching 600 m of water depth (Fig. 3.2). Combining the thickness of sedimentary fill in the central basin of at least 4 km (Hirm et al., 2003) with the 600-800 m bathymetric relief gives a rough estimate of at least 5 km of maximum structural relief within the Marmara Sea. This vertical offset is approximately 6-14% of the estimated 35-85 km of total lateral offset (Şengör, 1979; Barka and Kadinsky-Cade, 1988; Koçyiğit, 1989) along the NAF in the Marmara Sea region since the late Pliocene. Similar subsidence has occurred within basins in the southern Marmara Sea; however, these basins are filled with sediment due to a larger sediment supply from the southern Marmara fluvial systems and have yet to be well imaged.

The formation of these basins within a strike-slip faulting environment has typically been attributed to interplay between the right-lateral NAF transform motion and north-south oriented extension in western Anatolia and the Aegean region (Eyidoğan, 1988; Wong et al., 1995; Le Pichon et al., 2001; Parke et al., 2002). Le Pichon et al. (2001) suggest that the northern Marmara trough and the basins therein were formed in an extensional environment during the late Miocene. Then during the Pliocene, when right-lateral shear strain dominated, the trough was cut though by a main strike-slip fault. They argue that the motion of the main Marmara fault is not suitable to create the observed trough and basin structure. Conversely, Armijo et al. (2002) argue that the stepover geometry of the NAF fault strands alone is sufficient to form the trough and basin pattern, and that the asymmetric partitioning of normal- and strike-slip on these faults has persisted since the inception of the NAF in the Marmara region. In many places scarps visible in high-resolution bathymetry mimic the local basin geometry, and in some places show nested (composite) pull-apart features that suggest long-term
activity (Armijo et al., 2002). These hypotheses on the structural evolution of the Marmara Sea clearly are contrasting. Our quantitative modeling approach allows us to address the Marmara Sea fault configuration and evolution controversy from a mechanical perspective by relating fault slip and basin formation in the Marmara Sea with regional plate boundary loading.

Proposed Marmara Sea fault geometries

Previously proposed fault geometries for the Marmara Sea differ in the location and orientation of the faults as well as the continuity of the different fault segments (Fig. 3.2). Early models proposed a series of northeast-southwest trending strike-slip faults and northwest-southeast trending normal faults that bound the individual basins within the Marmara trough (Barka, 1992; Ergün and Özel, 1995; Wong et al., 1995). After collection of 2-D deep seismic reflection data, Parke et al. (1999) and Okay et al. (2000) proposed models with northern and southern fault strands bounding the northern Marmara trough with different fault patterns between the two bounding structures (Fig. 3.2a). Aksu et al. (2000) proposed a similar model that included a master strike-slip fault beneath the central axis of the trough. After collection of high-resolution bathymetric data in 2000, İmren et al. (2001) and Le Pichon et al. (2001) proposed a single strike-slip fault system cutting lengthwise through the trough (Fig. 3.2b). Conversely, Armijo et al. (2002) concluded that no evidence for a single, continuous, purely strike-slip fault exists and that fault segments bounding pull-apart features at a range of scales remain active (Fig. 3.2c). Although many of these interpretations were derived from the same data set, there is significant variation among them, calling for a mechanics-based quantitative evaluation of the relationship between faulting and basin formation.

Geomechanical model framework

Our mechanical modeling approach evaluates the consistency of the proposed fault geometries with the observed Marmara sea geomorphology and structural deformation. If slip on faults in a given geometry produces surface deformation that resembles the morphology of the Marmara Sea, then that fault geometry is considered to be mechanically consistent with the geomorphology. In general, our method applies slip
along plate boundary dislocations to load the upper crust in the Marmara Sea region, and lets the shallower Marmara Sea faults slip in response to that loading. We then compute surface deformation produced by the slip on the Marmara Sea fault. The modeling software we use to calculate the displacement fields in a linear-elastic, homogeneous and isotropic half-space is the numerical boundary element code Poly3D (Thomas, 1993). Fault surfaces embedded in the half-space are composed of planar, triangular elements composed of superposed angular dislocations (Comninou and Dunders, 1975). The displacement discontinuity, or fault slip, is constant on each element, but multiple elements are used to model an arbitrary number of mechanically-interacting fractures or faults with non-uniform slip distributions (Crouch and Starfield, 1983). After solving for the displacement discontinuities on the fault elements, we calculate displacements within the body or at the free surface at specified observation point locations.

The Marmara Sea fault configurations that we test extend from the surface to 15 km depth. We assume 15 km to be the base of the seismogenic portion of the crust based on the depth limit of aftershocks in the Marmara Sea region following the 1999 İzmit earthquake (Karabulut et al., 2002). The boundary conditions on these faults are prescribed as zero shear traction and zero relative normal displacement between the fault faces. To drive slip on the Marmara Sea faults, we apply the loading via plate boundary dislocations (Fig. 3.3). Savage (1990) shows that applying depth-dependent slip along a vertical dislocation in an elastic half-space can duplicate the surface deformation of transform faulting in an elastic plate overlying a viscoelastic aesthenosphere. Simplifying Savage’s model, we assume the plate boundary structures have uniform slip. We constrain the location and slip magnitudes of our plate boundary structures by our ability to match GPS surface displacement measurements (Flerit et al., 2003). The deep dislocations in our model extend from 15 km to 1000 km depth, essentially infinite compared to the horizontal dimensions of our Marmara Sea region of interest, and have the projected surface traces shown in Figure 3.3.

In the model, we apply slip magnitudes to the plate boundary structures that equal the amount of slip accumulated in one year. The slip magnitudes that we use are derived from slip rates determined from GPS studies (McClusky et al., 2000; Meade et al., 2002). The regional structures and rates that we consider are the single NAF trace east of the
Figure 3.3. Map projection of plate boundary dislocations that provide stress loading for shallow Marmara Sea faults. The plate boundary faults extend from 15 km depth to 1000 km. The slip rates on the NAF strands and the convergence rate on the Hellenic Arc are marked. Rates are assumed to be constant with depth.
Mudurnu Valley (30 mm/yr), the northern (NNAF) and southern (SNAF) strands of the NAF in the Marmara Sea region (24 and 6 mm/yr, respectively), and the Hellenic subduction zone in the southern Aegean Sea (30 mm/yr). All of the plate boundary structures are vertical. The displacement discontinuities applied to the NAF structures are pure right-lateral slip and the Hellenic rate is arc-normal convergence.

The slip below 15 km depth, on our plate boundary structures, is able to reproduce the observed annual horizontal surface displacements with mean errors (2.61 mm) that are comparable to the mean errors of the GPS measurements (1.75 mm) (Fig. 3.4). The greatest errors are associated with the southwestern data points. This region is just north of the northernmost major normal fault that is accommodating the north-south extension in the Aegean Sea and southwestern Anatolia. Including a deep extensional structure in this region may reduce the errors in this region; however, the close correspondence of the model and the observations in the local Marmara Sea area allows us to accept the somewhat larger errors in this region. Regardless, the residuals (Fig. 3.4b) show consistently small magnitudes and lack of consistent orientation within the Marmara Sea region, indicating that our rather simple fault model may include the necessary major structures and their contribution to the overall slip rates. The annual stress loading at 7 km depth reach > 0.01 MPA on the NNAF and > 0.005 MPa on the SNAF (Fig. 3.5).

Several limitations of the model should be discussed. First, the model incorporates an elastic half-space and the deformation that we calculate represents that which would be produced by earthquakes relieving the plate boundary loading accumulated over the course of one year. We do not explicitly consider time-dependent viscoelastic (e.g. Nur and Mavko, 1974), viscoelastic-gravitational (e.g. Rundle, 1982), or those effects in combination with loading from sedimentation and erosion (e.g. King et al., 1988). In general, the long-term effective elastic thickness of the crust controls the width of the viscoelastic-gravitational deformation, and the sediment loading controls the asymmetry of the distribution of uplift and subsidence (King et al., 1988). In viscoelastic-gravitational models of a thrust fault cutting an elastic plate that overlies a viscous substrate, Rundle (1982) shows that the maximum magnitude of the surface displacements due solely to the time-dependent response of the viscoelastic-gravitational stress relaxation can approach 50% of the maximum magnitude of the coseismic surface
Figure 3.4. (a) Annual surface displacements measured by GPS (black arrows) (Meade et al., 1999) and computed from plate boundary loading in our model (gray arrows). Displacements are shown relative to station DEMI on the Eurasian plate. (b) Residual displacement vectors between the GPS and computed vectors shown in (a).
Figure 3.5. Annual Coulomb stress accumulation at a depth of 7 km due to plate boundary dislocation loading. Coulomb stresses magnitudes are calculated for hypothetical optimally-oriented fault planes and represent the maximum Coulomb stresses available to produce slip on the Marmara fault segments.
displacements. At times at least 100 times greater than the viscoelastic relaxation time of the coupled system, $T_r = \eta/\mu$, where $\eta$ is the viscosity of the aesthenosphere and $\mu$ is the shear modulus of the elastic plate, the highly asymmetric coseismic surface uplift and subsidence equilibrate to reach comparable magnitudes (P. Segall, pers. comm.). In general, at long times, as elastic stresses relax, gravitational restoring forces will become increasingly important.

While these viscoelastic, gravitational, and sediment loading effects can clearly affect the magnitude of surface displacements over geologic time, the general locations of uplift versus subsidence relative to the fault locations do not change when compared with the purely coseismic solution. This notion is supported by the geologic record of relatively stable locations of the northern Marmara trough basins as evidenced by at least 4 km of fairly continuous basin fill within the Central Marmara and Çınarcık basins (Him et al., 2003) since the late Pliocene. In this regard, the model we use may predict the general locations of basins and uplifts, but should not be considered an attempt to predict absolute magnitudes of structural relief. The kinematics of the elastic solution we employ capture the first-order surface deformation due to crustal faulting and we show that this solution in itself is useful in differentiating the various fault models proposed for the Marmara Sea by other authors, although including time-dependent viscoelastic and gravitational effects may result in a more precise solution.

Second, the absence of a frictional criterion in our model implies that slip magnitudes on the shallow faults depend only on the resolved shear stress contributed from the remote loading and the interaction of slip on neighboring shallow faults. Differences in resolved normal stresses on the faults, most pronounced for faults that are most oblique to the plate boundary structures, do not affect the fault slip magnitudes. Inclusion of a frictional fault slip criterion would affect the magnitudes, and to a more limited extent, the directions, of fault slip, but we do not constrain the effects of fault friction in this study.

**Tested Marmara Sea fault configurations**

We have selected three proposed fault models to evaluate. We selected the Armijo et al. (2002) and Le Pichon et al. (2001) models because they were both
interpreted using the latest high-resolution bathymetry. Although it also used the latest bathymetry, we do not test the Imren et al. (2001) model because it so closely resembles the Le Pichon et al. (2001) model. We also test the Okay et al. (2000) model to represent the models that include northern and southern boundary faults (e.g. Parke et al., 1999).

With the exception of the northern and southern boundary faults in the Okay et al. (2000) model, all Marmara Sea faults dip at angles between 70° and 90°. Faults that were interpreted to be major strike slip-slip structures dip at 90° based on seismic reflection profiles (Parke et al., 2003). Smaller basin-bounding faults were interpreted to dip 70° or 80° toward the center of the basin. We assign steep dips to the basin-bounding faults based on recent seismic reflection data that reveals the Central Marmara basin to be box-shaped to at least 4 km depth (Hirn et al., 2003) and the Çınarcık basin to be at least 6 km deep. Assuming these faults extend to the base of the seismogenic portion of the crust and do not merge at depth (Hirn et al., 2003), then the dips must be greater than 60°. Dips of 70° or 80° prevent the faults from merging at a more shallow depth than the maximum seismogenic depth of 15 km.

**Computed Marmara Sea deformation patterns**

It is important to recognize that lateral motion along strike-slip faults produces a characteristic pattern of vertical surface deformation (Fig. 3.6). Surface deformation patterns produced by strike-slip faulting will vary according to the geometry of the individual faults and the relative location of neighboring faults (Chinnery, 1961). The clockwise change in strike of the NAF as it enters the Marmara Sea region from the east or west creates a large-scale extensional jog. The regional change in fault strike may have led to the localization of subsidence in the Marmara Sea region since 3-5 Ma, as continued right-lateral strike-slip motion will produce crustal extension and subsidence in the bend region (Fig. 3.7). On a smaller scale within the Marmara Sea, interaction among a series of echelon fault strands may be responsible for formation of the smaller basins. The modeling produces the observations of local basin subsidence overprinted on broader regional subsidence.

The primary results of models are vertical surface displacement rates generated by plate boundary slip below 15 km depth and Marmara Sea fault slip above 15 km depth.
Figure 3.6. Computed pattern of vertical surface deformation above overlapping, right-stepping, right-lateral strike slip faults. Solid contours denote uplift, dashed contours denote subsidence. The displacement values are normalized by the maximum vertical surface displacement. Greatest subsidence occurs in the overlap region of this extensional fault stepover. In this simple model, the faults are approximately vertical, have length $L$, extend to depth $0.4L$, are separated by $0.2L$, and overlap $0.2L$. 
Figure 3.7. Vertical displacement rate in the Marmara Sea region to right-lateral slip on the plate boundary dislocations shown in Figure 3.3.
For the three proposed fault geometries tested here we evaluate their ability to reproduce the observed location and relative magnitudes of subsidence along the northern Marmara trough.

Okay et al. (2000) model

Figure 3.8 shows the vertical surface deformation produced by slip along the fault geometry proposed by Okay et al. (2000) with the boundary conditions described earlier. For this model, the northern boundary fault dips 15° to the south and merges at approximately 6 km with the vertical strike-slip fault running longitudinally through the sea as proposed by Okay et al. (2000). The displacement rates are everywhere negative, indicating both regional and localized subsidence in the Marmara Sea region. This fault geometry successfully produces some of the large-scale basin structures within the northern trough (blue areas in Fig. 3.8) in the general location of the Çınarcık, Central Marmara, and İmralı basins (compare Fig. 3.2a and 3.8b). The model best predicts the location and subsidence pattern within the Çınarcık basin. Interestingly, the predicted subsidence rate of ~5 mm/yr would produce 1.5-2.5 km of vertical displacement were these faults active over the 3-5 Ma history of the NAF within the Marmara Sea. These subsidence magnitudes, when extrapolated, are reasonable in that they are greater than the bathymetric depth but less than the observed vertical structural offset (Hirn et al., 2003). While we suggest that the locations of the basins produced by one year’s worth of fault slip are likely to persist through time, we hesitate to speculate on the applicability of the absolute subsidence magnitudes due to inelastic behavior of the crust over the scale of geologic time.

The location of the Central Marmara basin is predicted; however, this model predicts an asymmetric basin that is deepest along the main NAF strand and then shallows to the north. This is in contrast to the relatively uniform block subsidence within the Central Marmara basin observed in seismic reflection profiles (Wong et al., 1995; Armijo et al., 1999; Parke et al., 1999; Demirbağ et al., 2003). This discrepancy implies that the geometry and/or the slip magnitudes on the basin bounding faults are not consistent with the observed deformation.
Figure 3.8. (a) Fault geometry from Okay et al. (2000). Fault dips marked on the figure are constant with depth. (b) Vertical deformation rates predicted by the boundary element model in mm/yr for the fault pattern in (a). Color bar on the right-hand side: blue regions indicate greater subsidence, red regions indicate lesser subsidence.
The most anomalous prediction of the Okay et al. (2000) fault model is the relative uplift in the vicinity of the Tekirdağ basin. Contrary to the observed morphology, this model produced greater relative uplift in the location of the Tekirdağ basin than in the location of Ganos Mountain farther west. This casts doubt on their interpreted intersection of a northern boundary fault with the main NAF just offshore of the Gelibolu peninsula.

*Le Pichon et al. (2001) model*

The mechanical model using the fault geometry proposed by Le Pichon et al. (2001) produces localized subsidence at the locations of the Çınarcık, Central Marmara, and Tekirdağ basins (compare Fig. 3.2b and 3.9b). The Tekirdağ basin is well-formed as the plate boundary loading produces a normal slip component on the dipping fault strands that bound the basin on three sides. Similarly, the pull-apart structure at the Central Marmara basin localizes subsidence between the overlapping fault strands, although subsidence extends to the north of the actual basin. This asymmetry of subsidence is not seen in the seismic reflection data.

The minor change in strike of the main vertical fault east of the Central Marmara basin produces local subsidence south of the fault that corresponds with the location of the Kumburgaz basin (see Fig. 3.2b for location). This localized basin is part of a larger region of subsidence that is greatest in the hanging wall above the 70° dipping Princes' Islands fault and steadily decreases to the south. Since the Le Pichon et al. (2001) model proposes no active faults south of the Princes Island fault, the observed localized subsidence in the location of the İmrahi basin is not reproduced.

The fault model of Le Pichon et al. (2001) represents an improvement over the model of Okay et al. (2000) in that the locations of the Çınarcık, Central Marmara, and Tekirdağ basins are predicted and they align to form a linear trough of basins separated by local highs. Key features that this model fails to produce; however, are a well-defined southern boundary to the Çınarcık basin, as well as localized subsidence in the location of the İmrahi basin.
Figure 3.9. (a) Fault geometry from Le Pichon et al. (2001). Fault dips marked in figure are constant with depth. (b) Vertical deformation rates predicted by the boundary element model in mm/yr for the fault pattern in (a). Color bar on the right-hand side: blue regions indicate greater subsidence, red regions indicate lesser subsidence.
Armijo et al. (2002) model

In addition to all of the structures of the Le Pichon et al. (2001) model, the fault model of Armijo et al. (2002) contains active faults along the northern margin of the Armutlu peninsula, two sets of overlapping fault strands at the Central Marmara basin, and faults bounding the Tekirdağ basin on four sides (Fig. 3.10a). The subsidence pattern produced by this model (Fig. 3.10b) is similar to the Le Pichon et al. (2001) model with some notable exceptions that make this model more compatible of the observed Marmara Sea morphology. First, the Tekirdağ basin is better defined and is of comparable subsidence rate to the Central Marmara and Çınarcık basins. Second, the high separating the Tekirdağ and Central Marmara basins is more pronounced. Third, the dimensions of the basin predicted at the location of the Central Marmara basin better match the seismic reflection observations. Fourth, the Çınarcık basin is bounded to the south and reproduces the observed uniform subsidence of the basin floor, in contrast to the tilted subsidence predicted in the Le Pichon et al. (2001) model. Finally, there is local subsidence in the location of the İmralı basin in the form of a southward tilting basement reflector (Fig. 3.10b).

Comparison with pre-transform / syn-transform stratigraphic horizon morphology

Comparisons of the surface displacements computed in our mechanical models with the bathymetry of the northern Marmara Sea provide an opportunity for evaluating the spatial distributions of deformation expected from each fault model. However, the bathymetric surface represents erosion and deposition of sediments superposed on the structural deformation due to faulting, and therefore is subject to processes in addition to fault-related deformation. In the Marmara Sea, the morphology of the seafloor and the morphology of the underlying structures differs most in the southern part of the sea, where a large influx of fluvial sediment has filled the structural basins in the region (Aksu et al., 1999; Parke et al., 2003). Therefore, to make comparisons with structural deformation alone, we compare our model displacements with the displaced and offset stratigraphic horizon marking the boundary between the Pliocene-Quaternary syn-transform sediments and the Miocene and older pre-transform strata (Okay et al., 1999; Parke et al., 1999; Okay et al., 2000; Parke et al., 2002). This horizon should record all
Figure 3.10. (a) Fault geometry from Armijo et al. (2002). Fault dips are marked in the figure are constant with depth. (b) Vertical deformation rates predicted by the boundary element model in mm/yr for the fault pattern in (a). Color bar on right-hand side: blue regions indicate greater subsidence, red regions indicate lesser subsidence.
of the NAF-related deformation in the Marmara Sea. Comparison of model deformation results with the deformation observed in lines 5 and 33 of the 1997 MTA seismic reflection survey (Fig. 3.11 and 3.12, respectively) (Parke et al., 2003), which trend north-south through the Çınarcık, İmralı, and Tekirdağ basins, supports the results from the modeling and bathymetry presented earlier.

The Le Pichon et al. (2001) and Armijo et al. (2002) models predict fairly uniform subsidence of the Tekirdağ basin floor (Fig. 3.11) along the location of seismic line 5. This subsidence is produced by slip along the main vertical fault strand bounding the basins to the south and along the subsidiary east-west trending fault bounding the basin to the north. While the location of maximum subsidence predicted by the Le Pichon et al. (2001) and Armijo et al. (2002) models is in agreement with the observed deformation, both models fail to predict the 17° ± 5° southward dip (Parke et al., 2002) of the pre-transform/syn-transform contact north of the southern bounding fault (Fig. 3.11c). The dip of this depositional surface, and the southward thickening of sediments overlying it, suggest that the Tekirdağ basin has developed as a half-graben rather than a full-graben structure. This suggests that the northern bounding fault in the Le Pichon et al. (2001) and Armijo et al. (2002) models, while it corresponds to the location of a scarp on the seafloor and small offsets of seismic horizons, is a rather recent structure that has not accommodated significant offset.

Along line 33 (Fig. 3.12), which trends north-south just west of the Armutlu peninsula, the Okay et al. (2000) and Armijo et al. (2002) models produce localized subsidence at the locations of the Çınarcık and İmralı basins (Fig. 3.12a) – matching the observed shape of the pre-transform/syn-transform contact (Fig. 3.12c). The Le Pichon et al. (2001) model, because it is composed of a single fault in this region, is unable to reproduce the relatively uniform subsidence pattern within the Çınarcık basin. Similarly, this model does not reproduce subsidence associated with the İmralı basin. The Armijo et al. (2002) model is more consistent with observations along this line than the Okay et al. (2000) model because it predicts a nearly horizontal Çınarcık basin floor, the observed high between the Çınarcık and İmralı basins, and the southward dip of the İmralı basin floor.
Figure 3.11. (a) Vertical surface displacement rates computed by the boundary element model along position of seismic line 5 from 1997 MTA seismic reflection survey (Parke et al., 2003). Inset map shows location of line 5. (b) Uninterpreted seismic reflection profile 5. (c) Interpreted profile. Bold line is the interpreted pre-transform / syn-transform stratigraphic contact. Dashed lines indicate faults or fault zone boundaries. Dashed line marked "M" indicates location of seafloor multiple.
Figure 3.12. (a) Vertical surface displacement rates computed by the boundary element model along position of seismic line 33 from 1997 MTA seismic reflection survey (Parke et al., 2003). Inset map shows location of line 33. (b) Uninterpreted seismic reflection profile 33. (c) Interpreted profile. Bold line is the interpreted pre-transform/syn-transform sediment contact. Dashed lines indicate faults or fault zone boundaries. Dashed line marked "M" indicates location of seafloor multiple.
Discussion

Comparing and constraining the proposed fault interpretations

Of the proposed fault geometries that we tested, the Armijo et al. (2002) model best reproduces the observed morphology and vertical structural offsets within the Marmara Sea. In addition to the interpreted active major Marmara fault proposed by Le Pichon et al. (2001), the Armijo et al. (2002) fault geometry includes active structures to the south of the Çınarcık and İmrâli basins and larger bounding faults to the north and south of the Central Marmara basin. Where the major strike-slip fault passes through a basin, the active scarps define the northern side of the Çınarcık basin, both the north and south sides of the Central Marmara basin, and the southern margin of the Tekirdağ basin – we conclude it does not cut through the center of the basins. Therefore, if this major fault is active, it should follow that the basins themselves, and likely the other subsidiary faults, should also be active. While sandbox models of pull-apart basins show that the strike-slip component of slip on the initial basin-bounding faults decreases after strike-slip faulting initiates in the center of the basin (Dooley and McClay, 1997; Rahe et al., 1998), this model prediction would not apply to the Marmara Sea basins because the major strike-slip fault forms along the margin, and not the center, of each of the three sub-basins.

Thus, we propose that all the basin-bounding faults should be considered active. The Armutlu fault can be interpreted as being active based on shallow seismic reflection profiles that show faults offsetting the youngest sediments on the southern margin of the Çınarcık basin (Smith et al., 1995). Modeling results show that, indeed, the southern Çınarcık fault must be active to best reproduce the observed basin morphology. The activity of the Armutlu fault suggests that the other subsidiary basin-bounding faults mapped by Armijo et al. (2002) should be considered active and capable of accommodating earthquake rupture.

The fault models that we evaluate focus on the Marmara trough region and do not extend to the southern Marmara Sea region where there are known to be active faults that cut the youngest sediments and produce scarps just north of the Kapıdağ Peninsula (Smith et al., 1995). These structures were imaged to a depth of <100 m and, therefore, the depth to which these faults extend is unknown. Unlike in the northern trough,
however, significant structural relief in this area would likely be masked by the greater fluvial sediment influx into this region (Aksu et al., 1999). Our results should therefore be viewed as an evaluation of the fault geometry along the northern strand of the NAF, with the caveat that there are major fault strands producing structural relief in the southern Marmara Sea whose impact on the seafloor geomorphology remains to be considered. However, from a seismic hazard standpoint, these faults may be less critical in that they are not direct extensions of the faults that ruptured in the 1912 Ganos and 1999 İzmit earthquakes.

*Slip kinematics of Marmara Sea faults*

To this point, we have evaluated the models in their ability to produce the observed seafloor morphology and stratigraphic horizon deformation within the Marmara Sea. Another test of each model's ability to simulate the kinematics of Marmara Sea deformation is to compare observed and computed fault slip vectors. We compare slip vectors of the faults within the Armijo et al. (2002) model configuration with the observed earthquake focal mechanism solutions. If the fault slip vectors in the model match the observed recent earthquake focal mechanisms, then it would provide a link between the Armijo et al. (2002) fault configuration, the formation of the northern trough basins, and the active coseismic deformation.

The distributions and directions of slip on the faults in our models are primarily controlled by the loading imposed by the right-lateral NAF plate boundary structures, and are controlled to a lesser degree by the mechanical interaction with neighboring faults. With the exception of small fault segments bounding the Tekirdağ basin to the north and east, the computed rakes on the faults in the Armijo et al. (2002) model show predominantly right-lateral slip (Table 3.1 and Fig. 3.13). This reflects the dominant effect of the deep plate-boundary loading.

Figure 3.14 compares the slip on our model faults to Marmara Sea earthquake focal mechanisms. We represent the slip on model fault segments in the form of lower hemisphere focal mechanisms solutions (Table 3.1). We compute the moment tensor composed of a double couple of equivalent forces representing the distributed slip on each model fault (Jost and Herrmann, 1989). The focal mechanisms for each model fault
Table 3.1. Fault planes and rakes for each (Armijo et al., 2002) Poly3D model fault segment (numbering shown in Figure 3.13a). Model faults are not planar and have heterogeneous slip. For each fault, a moment tensor is computed and the best-fit double couple focal mechanism solution is determined (Jost and Herrmann, 1989). In calculating the best-fit fault planes and rake directions, fault elements that have greater slip receive greater weight; therefore, the strike, dip, and rake can differ slightly from the model fault orientation and rake.

<table>
<thead>
<tr>
<th>Fault segment</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>263</td>
<td>81</td>
<td>178</td>
</tr>
<tr>
<td>2</td>
<td>283</td>
<td>80</td>
<td>176</td>
</tr>
<tr>
<td>3</td>
<td>245</td>
<td>71</td>
<td>-159</td>
</tr>
<tr>
<td>4</td>
<td>303</td>
<td>70</td>
<td>-157</td>
</tr>
<tr>
<td>5</td>
<td>118</td>
<td>70</td>
<td>-168</td>
</tr>
<tr>
<td>6</td>
<td>263</td>
<td>88</td>
<td>174</td>
</tr>
<tr>
<td>7</td>
<td>255</td>
<td>88</td>
<td>-168</td>
</tr>
<tr>
<td>8</td>
<td>90</td>
<td>82</td>
<td>-175</td>
</tr>
<tr>
<td>9</td>
<td>89</td>
<td>89</td>
<td>-178</td>
</tr>
<tr>
<td>10</td>
<td>288</td>
<td>80</td>
<td>-165</td>
</tr>
<tr>
<td>11</td>
<td>287</td>
<td>69</td>
<td>-179</td>
</tr>
<tr>
<td>12</td>
<td>76</td>
<td>88</td>
<td>174</td>
</tr>
<tr>
<td>13</td>
<td>85</td>
<td>88</td>
<td>-175</td>
</tr>
<tr>
<td>14</td>
<td>85</td>
<td>88</td>
<td>171</td>
</tr>
<tr>
<td>15</td>
<td>220</td>
<td>80</td>
<td>-74</td>
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<td>90</td>
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<td>-140</td>
</tr>
<tr>
<td>17</td>
<td>51</td>
<td>88</td>
<td>163</td>
</tr>
<tr>
<td>18</td>
<td>49</td>
<td>88</td>
<td>-174</td>
</tr>
</tbody>
</table>
Figure 3.13. (a) Individual fault segments representing the Armijo et al. (2002) fault configuration within the Poly3D model. (b) Lower hemisphere equal area stereonet projection of the rakes on the fault segments. The fault planes and rakes are best-fit focal mechanism planes (slip-weighted average) computed for the heterogeneous slip pattern on each non-planar fault segment. Solid and open triangles indicate predominantly reverse faulting and normal faulting, respectively.
Figure 3.14. (a) Solid circles and letters indicate locations of M > 4 earthquakes in the Marmara Sea. (b) Earthquake focal mechanisms (letters, upper row) and lower hemisphere equal area stereonet projection of the rakes on the fault segments in our model (numbers, lower row). Each earthquake focal mechanism is paired with the nearest model fault segment. Numbering of model fault segments is shown in Figure 3.12a. Fault segment slip mechanisms are given in Table 3.1. Focal mechanism data are given in Table 3.2.
are then computed from the moment tensors (Jost and Herrmann, 1989). We compare our computed focal mechanisms for the Armijo et al. (2002) model with nine composite earthquake focal mechanisms (Taymaz et al., 2002; Sato et al., in press) and 17 $M_s > 4$ earthquake focal mechanisms (Table 3.2) (Eyidoğan, 1988; Taymaz et al., 1991; Örgülü and Aktar, 2001; Özalaybey et al., 2002; Pinar et al., 2003).

Slip on the model faults matches the earthquake focal mechanisms well over most regions of the Marmara Sea. The model fault segments and the earthquakes exhibit predominantly right-lateral strike-slip. Earthquakes ($d$, $h$, and $j$) located close to the fault along the southern margin of the Çınarcık basin (fault 3) all exhibit dip-slip normal faulting and in this region the model does not fit the observed seismicity well. The nearby earthquakes on land along the Yalova coast are also not reproduced in our model. Again, these events have a greater dip-slip component than our nearest model fault. The mismatch in these areas could perhaps be attributed to the relatively large distance between the events and the nearest fault (> 5 km). Interestingly, this region of along the Yalova coast also corresponds to the location of a microseismicity cluster following the 1999 İzmit earthquake. Depending on the western termination location of the İzmit earthquake rupture, dip-slip earthquakes in this region could be located within a region of decreased normal stress associated with the stress perturbation of the İzmit earthquake.

The focal mechanism marked $j$ corresponds to a $M_s = 6.4$ earthquake with primarily normal slip (Taymaz et al., 1991) located near the western termination (Wright et al., 2001) of the almost pure right-lateral 1999 İzmit earthquake rupture. This is the largest event in our catalog, and its location and slip mechanism raise several interesting questions. First, can the local stress field vary enough over tens of kilometers on the İzmit-Armutlu fault system to produce vastly different $M > 6$ earthquakes in 1963 and 1999? Second, could this fault zones slip with different slip mechanisms in $M \approx 6$ earthquakes versus in large $M > 7$ ruptures? We cannot answer these questions in this study and, unfortunately, the answers to both questions depend upon the applicability of comparing our fault slip vectors with earthquake focal mechanisms.

While the fault slip vectors in our model cannot match some of the earthquake focal mechanisms in Figure 3.14, in general, the model fault slip directions correspond with the predominantly right-lateral strike-slip seismotectonics within the northern trough
Table 3.2. Earthquake catalog composed of nine composite earthquake focal mechanisms (Sato et al., in press) and 17 $M \geq 4$ earthquakes located in the Marmara Sea region. Events are sorted by referenced publication. The strike, dip, and rake define the focal mechanism solutions for the events.

<table>
<thead>
<tr>
<th>ID</th>
<th>Date</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
<th>Magnitude</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>4/2002-7/2002</td>
<td>40.81</td>
<td>27.72</td>
<td>0</td>
<td>80</td>
<td>-10</td>
<td>composite</td>
<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>b</td>
<td>4/2002-7/2002</td>
<td>40.82</td>
<td>27.84</td>
<td>70</td>
<td>90</td>
<td>-180</td>
<td>composite</td>
<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>e</td>
<td>4/2002-7/2002</td>
<td>40.76</td>
<td>29.13</td>
<td>0</td>
<td>60</td>
<td>-30</td>
<td>composite</td>
<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>g</td>
<td>4/2002-7/2002</td>
<td>40.59</td>
<td>29.01</td>
<td>190</td>
<td>65</td>
<td>-40</td>
<td>composite</td>
<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>h</td>
<td>4/2002-7/2002</td>
<td>40.72</td>
<td>29.02</td>
<td>120</td>
<td>40</td>
<td>-65</td>
<td>composite</td>
<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>i</td>
<td>4/2002-7/2002</td>
<td>40.76</td>
<td>28.03</td>
<td>70</td>
<td>90</td>
<td>-180</td>
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<td><em>Taymaz et al.</em> [in review]</td>
</tr>
<tr>
<td>j</td>
<td>9/19/1963</td>
<td>40.90</td>
<td>28.11</td>
<td>356</td>
<td>71</td>
<td>-11</td>
<td>5.0 Ms</td>
<td><em>Eyidoğan</em> [1988]</td>
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<tr>
<td>l</td>
<td>3/24/2001</td>
<td>40.86</td>
<td>28.88</td>
<td>105</td>
<td>78</td>
<td>-170</td>
<td>4.0 Ml</td>
<td><em>Özalaybey et al.</em> [2002]</td>
</tr>
<tr>
<td>m</td>
<td>10/20/1999</td>
<td>40.80</td>
<td>29.03</td>
<td>127</td>
<td>75</td>
<td>-170</td>
<td>4.5 Ml</td>
<td><em>Özalaybey et al.</em> [2002]</td>
</tr>
<tr>
<td>n</td>
<td>8/17/1999</td>
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<td>175</td>
<td>4.6 Ml</td>
<td><em>Özalaybey et al.</em> [2002]</td>
</tr>
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<td>o</td>
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<td>40.70</td>
<td>29.17</td>
<td>72</td>
<td>75</td>
<td>153</td>
<td>4.0 Ml</td>
<td><em>Özalaybey et al.</em> [2002]</td>
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<td>-167</td>
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<td><em>Özalaybey et al.</em> [2002]</td>
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<tr>
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<td>85</td>
<td>63</td>
<td>-161</td>
<td>5.0 Ml</td>
<td><em>Örgülü and Aktar</em> [2001]</td>
</tr>
<tr>
<td>s</td>
<td>8/17/1999</td>
<td>40.77</td>
<td>29.07</td>
<td>208</td>
<td>82</td>
<td>-27</td>
<td>4.2 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>t</td>
<td>8/17/1999</td>
<td>40.75</td>
<td>29.11</td>
<td>202</td>
<td>68</td>
<td>-1</td>
<td>4.7 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>u</td>
<td>8/17/1999</td>
<td>40.76</td>
<td>29.11</td>
<td>112</td>
<td>88</td>
<td>170</td>
<td>5.2 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>v</td>
<td>7/7/2000</td>
<td>40.84</td>
<td>29.19</td>
<td>142</td>
<td>39</td>
<td>-32</td>
<td>4.2 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>w</td>
<td>9/9/1999</td>
<td>40.69</td>
<td>29.16</td>
<td>69</td>
<td>80</td>
<td>148</td>
<td>4.0 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>x</td>
<td>4/13/1995</td>
<td>40.86</td>
<td>27.67</td>
<td>92</td>
<td>46</td>
<td>-137</td>
<td>5.0 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>y</td>
<td>2/8/1995</td>
<td>40.82</td>
<td>27.77</td>
<td>33</td>
<td>42</td>
<td>-137</td>
<td>4.5 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
<tr>
<td>z</td>
<td>4/18/1995</td>
<td>40.80</td>
<td>27.84</td>
<td>20</td>
<td>70</td>
<td>133</td>
<td>4.5 Mw</td>
<td><em>Pinar et al.</em> [2003]</td>
</tr>
</tbody>
</table>
of the Marmara Sea. The low-angle rake values calculated in our model are also consistent with formation of localized deep basins over the 3-5 million years of the NAF activity within the Marmara Sea. For example, the 11° rake on the Princes Islands fault produces normal slip that is 19% of the right-lateral strike-slip in each slip event. Assuming that the Princes Islands fault has been active during accumulation of the 35-85 km of cumulative right-lateral offset (Şengör, 1979; Barka and Kadinsky-Cade, 1988; Koçyiğit, 1989) over the past 3-5 Ma, there should be 7-17 km of dip-slip on the Princes Islands fault. These estimates exceed the estimated 8 km depth of the Çınarcık basin (Hirn et al., 2003), but they show that the relatively shallow fault rakes computed in the Armijo et al. (2002) are more than sufficient to produce the observed magnitudes of basin subsidence.

**Implications for the tectonic evolution of the Marmara Sea**

Le Pichon et al. (2001) note that the location of their proposed main Marmara fault could not produce the observed basin pattern in the Marmara Sea, and from this assumption, they suggest a two-stage evolutionary model for the Marmara Sea. They propose that north-south extension in the middle Miocene enlarged the Marmara Sea basin and formed the Marmara trough. Major NAF strike-slip motion beginning in the early Pliocene then formed cross-basins in a NNE-SSW extensional system. While the possibility of this interpretation cannot be categorically ruled out, the modeling results in this paper suggest that a two-stage model with an extensional regime predating the modern strike-slip regime is not necessary in explaining the formation of the Marmara trough and its basins. The regional deformation associated with strike slip motion and the change in orientation of the NAF in the Marmara region (Fig. 3.7) supports the concept that subsidence can be localized in this region without regional extension. Modeling shows that the Armijo et al. (2002) fault pattern, to a certain degree the Le Pichon et al. (2001) model, both produce subsidence in isolated locations consistent with the observed basins geometry. This implies the possibility for steady-state basin subsidence associated with continued, predominantly right-lateral, fault slip since inception the NAF in the Marmara region. While the sedimentary depositional record in seismic reflection profiles does suggests variation in the rates of sedimentation or
subsidence over time (Parke et al., 2002), there is not evidence for major changes in basin location.

Many authors have proposed that north-south oriented extension in the Aegean and southwestern Turkey may still be active (Eyidoğan, 1988; Wong et al., 1995; Le Pichon et al., 2001; Parke et al., 2002). We conclude that the modern development of the Marmara Sea morphology could result solely from fault slip interaction along the right-lateral NAF plate boundary without requiring a contribution of north-south extension from the Aegean region or southwestern Turkey. This conclusion is based on the following evidence:

1) GPS studies show that the north-south trending extension in western Turkey is confined to regions south of the southernmost NAF strand in the Marmara Sea region (Reilinger et al., 1997; McClusky et al., 2000).

2) Kahle et al. (2000) show that contractional strain is resolved on most of the major fault segments within the Marmara Sea, with the exception of the Princes Islands fault. The Princes Island fault is a local extension jog relative to the strike the NNAF to the east and west – supporting the notion that strike-slip fault interaction can produce local extension.

3) Our results and those of Flerit et al. (2003) show that plate boundary slip rates with no net opening across the Marmara Sea perpendicular to the NAF can reproduce the GPS measurements of horizontal surface displacements.

Conclusions

There exist several interpretations for the fault geometry in the Marmara Sea, in some cases based on identical data sets. In this paper, we have sought to evaluate the representative fault models by establishing a mechanical link between the proposed fault structures and the associated deformation that accompanies fault slip. We drive slip on the proposed fault geometries with deep plate boundary slip constrained by interseismic GPS velocity measurements. Our models, however, are not unique. Fault dips for all models are interpretations, the precise nature and location of the plate boundary beneath the NNAF is unknown, and extrapolation of our elastic models to geologic time is arguable. What we have done is to compare the consistency of the calculated Marmara
Sea morphologies and structural offset patterns with those of the observed modern day Marmara Sea visible in bathymetric maps and seismic reflection profiles. A high degree of consistency suggests that the tested fault geometry has been active throughout the development of the basin morphology. If the basins are actively subsiding today, then the faults in the tested geometry should be considered active. A low degree of consistency suggests that either the faults in the proposed geometry may not have been active throughout formation of the basin morphology or that the proposed fault pattern has been incorrectly interpreted.

We argue in this paper that, of several tested geometries, the Marmara Sea fault geometry proposed by Armijo et al. (2002) produces the best match to the observed Marmara Sea basin morphology and the offset and tilt pre-transform / syn-transform stratigraphic horizon. This, perhaps, is not surprising because the Armijo et al. (2002) model contains the most recent fault structure proposed by Le Pichon et al. (2001) and İmren et al. (2001). From a seismic hazards standpoint, this model is also attractive because it takes a more conservative approach by mapping more faults as being active. This is an important consideration because damaging earthquake along strike-slip earthquakes often occur on subsidiary faults that are previously unmapped or assumed inactive (e.g. the $M_w = 6.9$ 1989 Loma Prieta, CA and $M_w = 6.7$ 1994 Northridge, CA earthquakes). From a geological evolution point of view, our results suggest the possibility of steady-state right-lateral strike-slip faulting and basin formation within the Marmara Sea without requiring a two-stage model of regional extension in the Miocene followed by a transform regime since the Pliocene.

Acknowledgements

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Chapter 4

Static Coulomb stress change scenarios for earthquake ruptures in the eastern Marmara Sea

Abstract

Ambiguous locations of the western termination of the 1999 Izmit rupture and the location of the 1963 Yalova earthquake add uncertainty to any single prediction of the earthquake-induced stress changes in the eastern Sea of Marmara. We use a historical earthquake catalog, mapped Marmara Sea fault traces, and fault slip distributions for the 1999 Izmit earthquake to determine the stress change scenarios that result from six plausible configurations for the western termination of 1999 Izmit earthquake rupture and the location of the 1963 Yalova earthquake. Coulomb stresses calculated using geomechanical models are increased on the Princes’ Island, Çinarcık, and Armutlu fault segments in each case. In four of the six plausible rupture configurations, the Çinarcık fault receives the greatest average stress change. In one other configuration, the average stress increase on the Princes’ Islands fault is greatest. In another, the stress changes on the Çinarcık and Princes’ Islands fault are comparable. Rupture initiating on either the Princes’ Islands or Armutlu faults would be favored to propagate onto the Central Marmara, or Imrali fault, respectively, based on their favorable geometries of the respective fault intersections. Whereas rupture initiating on the Çinarcık fault would be limited to a much shorter rupture based on its mapped western termination. Therefore, while the earthquake-induced stress changes may, in most cases, be greatest on the Çinarcık fault, an earthquake initiating on this fault may produce a shorter cumulative rupture compared to rupture initiation on the two other major eastern Marmara Sea fault segments.

Introduction

The cascading sequence of large earthquakes along the North Anatolian fault from 1912-1999 has led scientists to an intuitive, yet troubling, conclusion: the next large earthquake is most likely to rupture along the northern portion of the Sea of Marmara and within tens of kilometers of Istanbul, a city of 14 million inhabitants (Fig. 4.1) (Le
Figure 4.1. Seismolectric setting of the Marmara Sea and the surrounding region. Surface traces of M2 events since 1900 (adapted from various sources).

Boundaries of Fig. 2a.

 jego figure shows the regional setting of the color figure. The large rectangle in color figure shows offshore locations of major strands of the NAF. Marmara Sea fault geometry is modified from Armitage et al. September 18, 2012. The different rupture scenarios interpreted in other publications. This figure shows 6.4 earthquakes since 1900 are plotted in red. The ruptures 1953a and 1963b refer to the same event on M2.2.
To address this threat, earth scientists aim to provide geologically-constrained probabilities of future rupture scenarios (e.g., Stein et al., 1997). Following the 1999 Izmit earthquake, Parsons et al. (2000) combined earthquake records over the past several centuries, earthquake recurrence intervals, and computed stress changes associated with recent earthquakes to estimate that there is a $62 \pm 15\%$ probability for strong shaking to occur in Istanbul before 2030. The contribution of the computed stress changes to this probability estimate is large — without the stress changes associated with the 1999 Izmit earthquake, the probability is reduced by $\approx 15\%$. Realistic future rupture scenarios and effective incorporation of earthquake interaction in probability studies therefore requires accurate calculation of stress changes on a Marmara Sea fault geometry that is well-constrained.

Several recent studies have calculated stress changes on vertical faults with uniform or smoothly varying strike directions throughout the Sea of Marmara due to large earthquakes within this century (Nalbant et al., 1998; Hubert-Ferrari et al., 2000; Parsons et al., 2000; Cakir, et al., 2003). This “optimal-orientation” approach is only accurate to the degree that the local fault orientation in the Sea of Marmara matches the assumed regional fault orientation in the model. This approach therefore introduces inaccuracies in regions, such as the eastern Sea of Marmara, where faults that bound the Çınarcık basin dip $60^\circ$ or shallower and deviate in strike from the assumed plate boundary by up to $30^\circ$. Others have computed stress changes resolved onto individual Sea of Marmara fault segments (Hubert-Ferrari et al., 2000; Pinar et al. 2001). In general, the calculated stress increases range from 0.1 to 5 bars depending on the proximity of the fault segment to the nearby modeled earthquake ruptures. Not surprisingly, the greatest stress increases are located in the eastern Sea of Marmara region, to the west of the presumed termination of the 1999 Izmit earthquake. In this region, the northern North Anatolian fault (NNAF) splits into three strands (Fig. 4.2), and Pinar et al. (2001) suggest that the Yalova-Hersek fault segment (0.45 MPa) was subjected to a greater stress change due to the Izmit earthquake than either the Princes’ Islands (0.18 MPa) or Çınarcık-Yalova (0.09 MPa) fault segments.

The recent earthquakes in this region, however, have ambiguous rupture configurations, making it difficult to constrain a single stress change scenario. Both the
Figure 4.2. a) Location of major 1999 Izmit earthquake rupture segments (yellow bounded by black) and epicenter (yellow star), Izmit earthquake aftershocks (red) (Ozalaybey et al., 2002), and faults within the Sea of Marmara (solid black) (Armijo et al., 2002). TB = Tekirdag basin, CMB = Central Marmara basin, CB = Çınarcık basin. b) Configuration of eastern Sea of Marmara faults and the three different westernmost Izmit earthquake rupture segment geometries (pink, orange, and green) and two different 1963 earthquake rupture locations (blue and cyan) tested in our study. Rupture scenario 1a includes orange and cyan ruptures. Scenario 1b includes orange and blue ruptures. Scenario 2a includes pink and cyan ruptures. Scenario 2b includes pink and blue ruptures. Scenario 3a includes green and cyan ruptures. Scenario 3b includes green and blue ruptures. Small yellow stars indicate epicentral locations of 1963 Yalova earthquake proposed by 1-Taymaz et al. (1991), 2-Jackson and McKenzie (1987), and 3-Ambraseys (1988).
location of the western termination of 1999 Izmit earthquake rupture and the rupture of the 1963 eastern Marmara earthquake (commonly referred to as the “Yalova” earthquake) are debated within the literature. Various authors have suggested the 1999 Izmit earthquake has either terminated east of the Hersek Delta within the Bay of Izmit, or propagated anywhere between 12 and 33 km further west (Fig. 4.2b). The 1963 earthquake has been thought to occur on both the northwest-trending Princes’ Islands fault (Taymaz, 1991; Parsons et al. 2000) and the west-northwest-trending southern Çinarcik fault (Jackson and McKenzie, 1987; Nalbant et al, 1998). The stress changes that are calculated within the eastern Sea of Marmara will depend significantly on the configurations of these ruptures. We aim to evaluate how the future rupture scenarios, as projected by Coulomb stress changes, depend on the locations of these recent eastern Sea of Marmara earthquake ruptures. We argue that future rupture is most favored along either the southern Çinarcik fault or the Princes’ Islands fault, depending on the location of earlier ruptures. The difference between these two scenarios has important implications for the potential hazard posed to region by a future earthquake due to the proximity of the faults to Istanbul. The rupture lengths that may develop for an earthquake initiating on either fault are likely to be different. The Çinarcik fault is mapped is mapped to have an abrupt western termination whereas the Princes’ Island fault may more effectively transfer slip as it links with the Central Marmara fault that continues tens of kilometers further west.

1963 earthquake location

The size of the $M_s=6.2-6.4$ 1963 earthquake (Table 4.1) suggests that the fault segment ruptured in this earthquake released at least several decades of stress accumulation, and would require decades of plate-motion related stress build-up to rupture again. Thus, the local distribution of stress change on the eastern Sea of Marmara fault segments depends significantly on the location of this event. Seismological studies have determined epicentral locations up to tens of kilometers apart (Fig. 4.2b) (Jackson and McKenzie, 1988; Ambraseys, 1988; and Taymaz et al., 1991). Relative to the locations of the Princes’ Islands and Çinarcik faults (Fig. 4.2b), Taymaz et al (1991) determine the event to be to north of both faults, Jackson and McKenzie (1988) place it
Table 4.1. Location of the 1963 earthquake in seismological and stress modeling studies

<table>
<thead>
<tr>
<th>Lon (°)</th>
<th>Lat (°)</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>29.09</td>
<td>40.71</td>
<td>268</td>
<td>70</td>
<td>235</td>
<td>Jackson and McKenzie (1987)</td>
</tr>
<tr>
<td>29.1</td>
<td>40.6</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>Ambraseys (1988)</td>
</tr>
<tr>
<td>Princes’ Island fault</td>
<td>296</td>
<td>~60</td>
<td>-</td>
<td>-</td>
<td>Parsons et al. (2000)</td>
</tr>
<tr>
<td>29.2</td>
<td>40.9</td>
<td>304</td>
<td>~56</td>
<td>-82</td>
<td>Taymaz et al. (1991)</td>
</tr>
<tr>
<td>Çinarcik fault</td>
<td>304</td>
<td>~56</td>
<td>-82</td>
<td>-</td>
<td>Nalbant et al. (1998)</td>
</tr>
</tbody>
</table>
between the faults, and Ambraseys (1988) provides a location south of both faults. Based on the ambiguity of these locations, previous stress change studies have modeled the 1963 rupture either on the southwest-dipping Princes’ Islands fault (Parsons et al., 2000) or the north-dipping southern Çinarcik fault (Nalbant et al., 1998; Hubert-Ferrari et al., 2000).

Taymaz et al. (1991) provide the most comprehensive source determination and compute almost pure normal slip on south- and north-dipping nodal planes that approximately match the orientations of the Princes’ Islands and southern Çinarcik faults, respectively. However, their resolved epicentral location is not near to the major faults in the region (Fig. 4.2a), it is ~15 km northeast of the Princes’ Islands fault and ~25 northeast of the southern Çinarcik fault. Damage from the earthquake was localized on the southern coast of the Sea of Marmara, leading Nalbant et al. (1998) to attribute the earthquake to rupture of the southern Çinarcik fault, and the north-dipping nodal plane of Taymaz et al. (1991). Parsons has attributed the event to rupture of the Princes’ Island fault, presumably based on its closer proximity to the epicenter location, and the southwest-dipping nodal plane of Taymaz et al. (1991). The combination of epicentral location, nodal plane solutions, and damage reports does not clearly define which fault ruptured in 1963. Therefore, we test and evaluate separate scenarios of rupture on both the Princes’ Islands and the southern Çinarcik faults.

Possible locations for western termination of 1999 Izmit rupture

Previous slip inversions

Since most of the Izmit surface rupture west of the epicenter occurred underwater within the Bay of Izmit, we rely upon a variety of geophysical data sets help constrain the rupture location. These include GPS and InSAR geodetic data, seismic waveform data, and aftershock locations. Inverting for Izmit earthquake fault slip, (Reilinger et al., 2000) show that GPS geodetic measurements do not permit slip in excess of 0.6 m along the Yalova segment of the Izmit rupture more than 10 to 15 km west of Hersek Delta. Synthetic Aperture Radar interferometry (InSAR) indicates that the best-fitting length of rupture beyond the Hersek Delta is 12 km, with good fits to the interferograms for fault lengths between 9 and 18 km (Wright et al., 2001). At fault lengths beyond 12 km west
of the Hersek Delta, the increase in misfit remains lower than that for lengths shorter than 12 km. This allows for the possibility that rupture extended further than 12 or even 18 km west of Hersek Delta. However, the small increases in misfit may be attributed to the west end of coverage of the interferogram data, reduced data coverage as the Gulf of Izmit widens, and a lack of coherence on the southern shore of the Gulf of Izmit in the interferograms (Wright et al., 2001).

Seismic waveform data reveals different interpretations of the western rupture termination depending on the data set and processing method. Yagi and Kikuchi (2000) use near-field strong motion data and teleseismic body wave data to invert for the slip distribution and show that subsurface slip extended no further than 25 km west of the epicenter near Gölcük (15 km east of Hersek Delta). Gülen et al. (2002) invert teleseismic body waves from 25 seismic stations to model the source process. When they include the latter phases of the seismograms (60-100s) in their inversion, they find a rupture subevent west of the Hersek delta corresponding to 1.3 m of slip approximately 10 km west of Hersek Delta. They propose that this subevent corresponds to rupture of the Yalova segment and is responsible for the lineation of surface cracking nearby onshore, and the alignment of aftershocks extending along the coast and into the Marmara Sea. Aydin and Kalafat (2002) argued for a similar southwest extension of the rupture based on the field observation of earthquake related faults and increased damage in the Yalova region. Based primarily on a lineation of aftershocks, Gülen et al. (2002) also suggest the possibility that the Izmit rupture propagated west along a more northerly fault segment extending from the Hersek Delta to the southern portion of the Princes' Islands fault. However, this rupture scenario is hard to constrain because it is well within the Sea of Marmara, placing it further from coastal sites and geodetic measurements.

Based on the ground motion from five near-fault accelerometers, Bouchon et al. (2002) compute fault slip that sharply decreases from a local surface slip maximum of 2 m near the Hersek Delta on a fault that trends east-west into the Marmara Sea. Also using near-fault strong motion waveforms, Sekiguchi et al. (2002) suggest that slip extended west of the Hersek Delta for at least 5 km.

Several authors use joint inversion of different types of geodetic data (Feigl et al., 2002), or combined geodetic and seismic data (Delouis et al., 2002) to constrain the
western rupture termination. Similar to the Bouchon et al. (2002) results, the solutions of both Delouis et al. (2002) and Feigl et al. (2002) include less than a meter of slip about 50 km west of the epicenter that decays to zero about 15 km west of Hersek Delta. In summary, the various slip inversion studies suggest that the western termination of the 1999 Izmit earthquake rupture is located 10-20 km west of the Hersek Delta (50-60 km west of the epicenter), although the latitudinal position is not well constrained because the inversion models presuppose a fault trend.

Aftershock location studies

The distribution of aftershocks following the Izmit earthquake extend approximately 75 km west of the epicenter (Karabulut et al., 2002; Özalaybey et al., 2002) and are clustered in three general regions within and west of the Bay of Izmit (Fig. 4.2a). One cluster is a linear trend of aftershocks that extend about 35 km west of Hersek with an uninterrupted E-W trend parallel to the axis of Izmit Bay. The vast majority of the hypocenters is located in the 5-17 km depth range (Karabulut et al., 2002; Özalaybey et al., 2002) and have predominantly strike-slip focal mechanisms (Karabulut et al., 2002).

Further west, the cluster branches to the NW and extends to 10 km south of the Princes Islands. Coseismic slip in this region is not confirmed or ruled out by the geodetic or strong motion inversions and may be suggested by teleseismic body wave inversion (Gülen et al., 2002). Focal mechanisms of three $M_l \geq 4.0$ events that occurred along this lineation show right-lateral strike-slip mechanisms along WNW-trending fault planes (Özalaybey et al., 2002), while many smaller events show normal faulting mechanisms (Karabulut et al., 2002).

In contrast, the third cluster of aftershocks, mostly onshore and near the city of Yalova, show 45°-60° north-dipping normal faulting mechanisms (Özalaybey et al., 2002) that cluster along a N50°-dipping zone of seismicity. The number of aftershocks within the Yalova cluster significantly increases two days after the mainshock. This delayed onset, together with the dip-slip focal mechanisms, indicate that this cluster represents triggered seismicity and is not likely part of the mainshock rupture; however
the localized cluster suggests a local stress concentration that could be related to nearby termination of Izmit rupture.

Pinar et al. (2001) concluded that the Izmit rupture terminated east of 29.6° E. They argue that this rupture termination location, as opposed to one further west at 29.1° E, places a greater number of aftershocks within regions of increased Coulomb stresses. Several observations complicate this approach. First, the vertical faults with an optimally-oriented strike direction along which Coulomb stress were calculated in their model is less accurate in view on the variation of the focal mechanism nodal planes in this region. Second, in addition the regions ahead of the lateral fault tips, aftershocks can be triggered along the upper and lower tiplines of ruptures. In map view, the aftershocks along the upper or lower tiplines would plot along the main trace of the rupture, resulting in a rather continuous lineation of aftershocks along the fault and well past its tips. This makes it difficult to clearly discern where within this linear cluster the lateral terminations of rupture are located. Indeed, there are aftershocks all along the rupture trace for both locations of western termination that Pinar et al. (2001) evaluate, suggesting that aftershocks are a poor constraint for determining the lateral limits of this rupture.

Western terminations of Izmit earthquake rupture in our model

Review of the geophysical data and investigations suggests four possible western termination locations for the Izmit rupture (Fig. 4.2b):

1) Rupture terminating just east of Hersek Delta as suggested by Pinar et al. (2001).

2) Rupture continuing 15 km west-southwest of the Hersek Delta along the southern strand of the NNAF within the Bay of Izmit as suggested by slip inversions from seismic and geodetic data.

3) Rupture continuing 20 km west of the Hersek Delta along the northern strand of the NNAF towards the Princes’ Islands fault as suggested by aftershocks and slip inversions by Gülen et al. (2002).
4) Rupture continuing 30 km west-southwest of Hersek Delta along the southern strand of the NNAF as suggested by aftershock clustering and strong shaking in Çınarcık and Yalova (Özalaybey et al., 2002).

To evaluate the stress changes that each of these different rupture scenarios would produce on faults in the eastern Sea of Marmara, we first obtain slip distributions for the different Izmit rupture geometries. We use InSAR ground surface displacement data to invert for distributed slip on all segments of the Izmit rupture (Wright et al., 2001), with the geometry of the westernmost segment being different in each case. Solving for the distributed slip, rather than simply applying a uniform slip magnitude to an entire fault surface, is important because faults with uniform slip produce stress changes up to several times greater than that of tapered slip within a half-length of the ruptured fault (Fig. 4.3). Using well-constrained distributed slip solutions will be particularly important for the faults in the eastern Sea of Marmara because the NNAF splits into three different strands within a half-length of the major fault segments of the Izmit rupture (Fig. 4.2a).

We solve for Izmit earthquake fault slip using a downhill simplex inversion technique (Press et al., 1992) and ground surface displacements from InSAR interferometry (Wright et al., 2001) and GPS data (Reilinger et al, 2000). The interferogram used in the inversion is an ERS-2 ascending interferogram computed from satellite passes on August 13 and September 17, 1999. The western edge of the interferogram is located about 25 km west of Hersek Delta. The computed fault slip values will therefore be less well-constrained approaching or exceeding the western edge of the interferogram. The GPS data were acquired within the first two months of the earthquake and were recorded at 51 sites (Reilinger et al, 2000).

The fault segments used in our inversion are rectangular and we specify the dip, dip direction, and fault height. Slip is fixed at zero along the lateral and bottom edges and bottom of the faults. Our inversion uses a smoothing parameter to constrain the variation in slip between each slip patch its neighbors and also a positivity constraint to ensure that slip within each entire fault segment is right-lateral. Surface displacements for each slip solution are calculated, with each fault patch modeled as a rectangular dislocation in a uniform elastic medium. The final slip solution is the one that provides
Figure 4.3. Comparison of stress change magnitudes as a function of distance from the fault edge for uniform versus distributed slip. Uniform slip produces Coulomb stress changes that are several times greater than stress changes from distributions with either maximum slip at the surface (solid) or at depth (dashed). Faults with slip distributions are shown in gray with the average slip in both cases (1 m) matching the case of uniform slip.
the minimum misfit between the observed geodetic surface displacements and those computed using the model slip solution.

The multi-segment slip distributions for the four inversions, projected onto a single fault for display, are shown in Figure 4.4. For the most part, they are very similar. In each case, a slip maximum of 6.7-6.8 m is located west of the epicenter, along the eastern portion of the Bay of Izmit. Slip decreases to less than 3 m within 10 km of east Hersek Delta and just west of Lake Sapanca. The misfits between the modeled and observed InSAR and GPS surface displacements indicate that the scenario with no faulting west of Hersek Delta cannot reproduce observed displacements as well as cases with slip extending further west (Fig. 4.5, Table 4.2). The three cases of rupture west of Hersek Delta; however, all fit the data equally well and cannot be differentiated. This is due to the proximity of the western edge of the InSAR interferogram and the lack of data within the Bay of Izmit. These results require us to test the three different Izmit rupture configurations where rupture extends west of Hersek Delta in our stress change calculations.

**Marmara Sea earthquakes since 1900**

In addition to the 1963 Yalova and 1999 Izmit earthquakes, we include in our models all $M_s \geq 6.4$ earthquakes in the Sea of Marmara region since 1900 (Fig. 4.1; Table 4.3). Besides the 1963 Yalova and 1999 Izmit events, these earthquakes contribute stress changes on the order of 0.01-0.02 MPa to the faults in the eastern region of the Sea of Marmara, which represent $\sim$5-10% of the total stress changes when the 1963 Yalova and 1999 Izmit earthquakes are included in the model. The pattern of stress change in the Marmara Sea region produced by all of the earthquakes is shown in Figure 4.6. Here, as a rough approximation to the geometry of the major faults within the Marmara Sea (Parke et al. 1999; Okay et al., 2000; Le Pichon et al., 2001; Armijo et al. 2002), the Coulomb stress is calculated on vertical, east-west trending planes. Nearly the entire northern trough region of the Marmara Sea has accommodated increased Coulomb stress change with the greatest concentration located west of the Izmit rupture and east of the 1912 Ganos rupture. Here we assume the 1912 rupture terminated several kilometers into the Sea (Nalbant et al., 1998), although this location is in debate (Armijo et al., 2003).
Figure 4.4. a) 3-D geometry of Izmit earthquake rupture segments used for slip inversion showing the three different western termination geometries. The gray segments with black outline are the same for all four inversions. The three different westernmost segments outlined in red and numerically labeled correspond to the segments included for inversion in scenario 1, 2, and 3. The inversion smooths slip across the discontinuous segment boundaries. b) Computed slip distribution for scenario 1 (fault extends 15 km southwest of Hersek Delta, along the Bay of Izmit coast). Slip values are projected onto a planar surface for presentation only. c) Slip distribution for scenario 2 (fault extends 20 km west towards Princes’ Islands fault). d) Slip distribution for scenario 3 (fault extends 30 km southwest along Yalova/Çınarcık coast to intersect Armutlu fault).
**Figure 4.5.** Computed satellite line-of-sight surface displacement interferograms (left column) and residual interferograms (right column) for endmember Izmit fault geometries. Residuals are calculated by subtracting the model interferogram from the actual interferogram. In upper panels, there is no Izmit rupture segment west of Hersek Delta. In the lower panels, the fault geometry corresponds to scenario 3 in Figure 4.4. Note the concentration of fringes indicating large residuals in the western end of the Izmit rupture when there is no fault segment west of Hersek Delta (upper right panel).
Table 4.2. Root-mean-squared misfits between modeled and observed InSAR and GPS data at the observation locations.

<table>
<thead>
<tr>
<th>Western termination of Izmit rupture</th>
<th>InSAR residual (cm)</th>
<th>GPS residual (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hersek Delta</td>
<td>3.45</td>
<td>4.99</td>
</tr>
<tr>
<td>20 km west of Hersek Delta, at intersection with Princes' Islands fault</td>
<td>3.30</td>
<td>4.50</td>
</tr>
<tr>
<td>15 km west of Hersek Delta, along southern coast of Bay of Izmit</td>
<td>3.29</td>
<td>4.59</td>
</tr>
<tr>
<td>30 km west of Hersek Delta, along southern coast of Bay of Izmit</td>
<td>3.27</td>
<td>4.60</td>
</tr>
</tbody>
</table>
Table 4.3. Earthquakes since 1900 included in the stress change model.

<table>
<thead>
<tr>
<th>Date</th>
<th>$M_s$</th>
<th>$M_0$ $(10^{19}$ N-m)</th>
<th>Area of model fault† $(10^8$ m$^2$)</th>
<th>Dip of model fault (°)</th>
<th>Slip‡ (m)</th>
<th>Rake on model fault§ (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan. 19, 1912</td>
<td>7.4</td>
<td>12.6°</td>
<td>11.7</td>
<td>90</td>
<td>3.60</td>
<td>180</td>
</tr>
<tr>
<td>Feb. 4, 1935</td>
<td>6.4</td>
<td>0.4°</td>
<td>3.4</td>
<td>60N</td>
<td>0.39</td>
<td>-90</td>
</tr>
<tr>
<td>June 20, 1943</td>
<td>6.4</td>
<td>0.4°</td>
<td>2.4</td>
<td>90</td>
<td>0.54</td>
<td>180</td>
</tr>
<tr>
<td>Feb. 1, 1944</td>
<td>7.3</td>
<td>8.9°</td>
<td>19.4</td>
<td>90</td>
<td>1.53</td>
<td>180</td>
</tr>
<tr>
<td>March 18, 1953</td>
<td>7.2</td>
<td>6.3°</td>
<td>9.0</td>
<td>90</td>
<td>2.35</td>
<td>180</td>
</tr>
<tr>
<td>May 26, 1957</td>
<td>7.0</td>
<td>3.2°</td>
<td>3.0</td>
<td>90</td>
<td>3.57</td>
<td>180</td>
</tr>
<tr>
<td>Sept. 18, 1963</td>
<td>6.4</td>
<td>0.1°</td>
<td>3.0°†</td>
<td>70S°†</td>
<td>0.10°†</td>
<td>-90</td>
</tr>
<tr>
<td>Oct. 6, 1964</td>
<td>6.9</td>
<td>0.4°</td>
<td>7.8</td>
<td>60N</td>
<td>0.18</td>
<td>-90</td>
</tr>
<tr>
<td>July 22, 1967</td>
<td>7.1</td>
<td>11.0°</td>
<td>1.2</td>
<td>90</td>
<td>2.96</td>
<td>180</td>
</tr>
<tr>
<td>Nov. 12, 1999</td>
<td>7.2</td>
<td>6.3°</td>
<td>7.8</td>
<td>60N</td>
<td>2.69</td>
<td>180</td>
</tr>
</tbody>
</table>

a Scalar seismic moment computed by assuming $M_w = M_s$ and using the relationship $M_0 = 10^{(1.5M_w+9)}$ (Shearer, 1999).

b Scalar seismic moment from Taymaz et al. (1991).

c Scalar seismic moment from Pinar et al. (1996).

d Rupture on the Princes' Islands fault (e.g. Parsons, et al., 2000).

e Rupture on the Çınarcık fault (e.g. Nalbant et al., 1998).

† Faults have lengths shown in Figure 4.1, dips in column 5, and extend to 15 km depth.

‡ Average slip, $u = M_J/\mu A$, where $\mu = 30$ GPa, and $A$ is fault area (Shearer, 1999).

§ Rake assumed to be pure dip-slip or strike-slip.
Figure 4.6. Coulomb stress changes on east-west and N30°W trending vertical fault planes in the Marmara Sea region due to all $M_s \geq 6$ earthquakes since 1900 (Table 4.3). For this example, the 1963 Yalova earthquake is located on the Çinarçık fault and the Izmit rupture termintes 15 km west of the Hersek Delta. Red areas denote stress increases, blue areas denote stress decreases. Stress changes are calculated at a depth of 7.5 km.
This pattern of greatest stress change supports strain calculation from GPS measurements (Straub et al., 1997) and aftershock locations (Taymaz et al., 2001) that indicate the NNAF is the most active of the three of the NAF and is the most susceptible to future earthquake rupture.

**Stress changes after 1999**

To gain an accurate measure of the Coulomb stress change in the eastern Marmara Sea for each scenario, we resolve the shear and normal stress on the fault surfaces at a depth of 7.5 km. Faults dips and dip directions are shown in Figure 4.2b. The Coulomb stress change is defined as

\[ \sigma_c = |\Delta \tau| + \mu' \Delta \sigma \]  

(4.1)

where \( \Delta \tau \) is the change in the shear traction on the fault surface, \( \Delta \sigma \) is the change in the normal traction (compression is negative), and \( \mu' \) is the effective coefficient of friction. We set \( \mu' = 0.4 \), for direct comparison with other Coulomb stress change studies along the NAF (e.g. Stein et al., 1997; Hubert-Ferrari et al., 2000; Pinar et al., 2001). It is necessary to keep track of the sign of \( \tau \) (traction inducing right-lateral slip is positive) because shear tractions of different sign, but equal magnitude, on adjacent areas of a fault surface could produce a uniform positive Coulomb stress change, while this situation would not physically lead to failure in any given direction. Such a scenario is unlikely to occur, but should be considered in the eastern Marmara Sea region where perturbations from dip-slip and strike-slip earthquakes in close proximity are modeled.

Stress increases associated with secular plate motion are hard to accurately constrain in the Marmara sea region. Due to the lack of GPS survey measurements close to the fault strands, it is unclear how the multiple upper crustal fault strands in this region connect with deep plate motion beneath the locked seismogenic portion of the lithosphere. Studies suggest different locking depths and plate boundary locations in the Marmara Sea (Meade et al., 2002; Hubert-Ferrari et al., 2000). In addition, due to uncertainties in rupture location, it is unclear if the 1719, 1754, or 1766 event was the last major earthquake in this region (Hubert-Ferrari et al., 2000; Parsons et al., 2000). Regardless, based on these dates and the slip rate of the NNAF it is clear that the eastern Marmara Sea is in the late stages of an interseismic cycle and the secular stress
accumulation should be large. Using the most conservative penultimate earthquake date of 1766 and estimates of secular stress loading in these faults for a conservative locking depth of 15 km, the Princes’ Islands and Çinarcık faults will have accumulated at least 3-4 MPa of tectonic stress (Muller and Aydin, in review). Since this accumulated stress is an order of magnitude greater than the stress changes due to nearby earthquakes, the location, orientation, and locking depth of deep plate boundary slip are important factors in determining the Coulomb failure conditions in the eastern Marmara faults. However, until more accurate geophysical constraints on the tectonic loading conditions are available, we focus on the stress perturbations associated with recent earthquakes.

We compute stress changes using Poly3D (Thomas, 1993), a boundary element code that can compute stress changes at any point in an isotropic and homogeneous elastic half-space due to slip on fault surfaces composed of triangular elements. Each element has constant slip, but many elements with different slip magnitudes can compose a fault surface. The slip distributions in Figure 4.4 are applied to the İzmit fault segments in our different tests. We test permutations of the three different western terminations of the 1999 İzmit rupture and the two possible locations of the 1963 Yalova earthquake, for a total of six different eastern Marmara rupture scenarios (Fig. 4.7a-f, col. 1).

**Scenarios 1a and 1b**

If the İzmit earthquake ruptured 15 km southwest of Hersek Delta along the southern coast of the Bay of İzmit, and the 1963 earthquake ruptured along the Çinarcık fault (scenario 1a, Fig. 4.7a), the East Çinarcık segment receives the greatest stress change, with a maximum of over 1 MPa. This magnitude of Coulomb stress change is greater than the average increase of 0.4 MPa calculated at the epicenters of large earthquakes triggered during the 1939-1992 earthquake sequence along the NAF (Stein et al., 1997). The next largest average increases occur on the West İzmit and Princes’ Islands fault segments. Based on the stress changes, it is difficult to assess the most likely future rupture for this scenario. To produce an earthquake rupture of at least 40 km, westward propagating rupture on the East Çinarcık fault would have to either propagate through the portion of the fault assumed to have ruptured in 1963, or it would have to propagate onto the Armutlu fault. Therefore, throughgoing rupture along the
Figure 4.7. Rupture scenarios and Coulomb stress changes for the eastern Marmara Sea faults. West Izmit fault and Princes’ Island fault = green; East and West Çinarcık fault = red; Armutlu fault = purple. Western segment of Izmit earthquake rupture is shown as thick orange, magenta, or light green line corresponding to Figure 4.2b. 1963 earthquake rupture is shown as light blue and dark blue corresponding to Figure 4.2b.
West Izmit and Princes' Islands faults or initial slip on the East Çinarcık segment that branches onto the Armutlu fault may be more plausible rupture scenarios than rupture west along the Çinarcık fault.

If instead the 1963 earthquake occurred on the Princes' Islands fault (scenario 1b, Fig. 4.7b), the East Çinarcık fault is most favored for future slip. In this case, there is no impediment to throughgoing rupture on the Çinarcık fault because the fault hasn't ruptured recently, so complete rupture of the Çinarcık fault is possible. In this case, although the West Izmit fault segment has a 30% greater average stress increase than in scenario 1a, the location of the 1963 rupture would likely impede throughgoing rupture along this fault segment.

Scenarios 2a and 2b

If the Izmit earthquake ruptured 20 km directly west of the Hersek Delta to the intersection with the Princes' Islands fault segment, and the 1963 earthquake was on the Çinarcık fault, the East Çinarcık and Princes' Islands faults receive comparable maximum and average Coulomb stress increases (scenario 2a, Fig. 4.7c). Following earlier arguments, while the stress changes equally favor rupture initiation on either segment, throughgoing rupture is more likely on the Princes' Islands fault because, in the this case, it is not assumed to have ruptured in 1963. Branching rupture onto the Armutlu fault is again plausible, but the average stress increase on the Princes' Islands fault is at least twice as large as that on the Armutlu fault.

If the 1963 earthquake ruptured the Princes' Island earthquake (scenario 2b, Fig. 4.7d), the East Çinarcık fault again receives the greatest stress changes. The stress changes on the West Çinarcık and Armutlu faults are comparable, which suggests that rupture initiating on the East Çinarcık fault could be equally likely to propagate onto either segment.

Scenarios 3a and 3b

If the Izmit earthquake ruptured 30 km west of Hersek Delta along the southern coast of the Bay of Izmit, and the 1963 earthquake ruptured along the Çinarcık fault (scenario 3a, Fig. 4.7e), the Princes' Island fault receives the greatest stress change, with
a maximum of 0.3 MPa. The alignment of the western part of the Izmit rupture with the Armutlu fault in this case intuitively suggests that the Armutlu fault should receive the greatest Coulomb stress changes due to the large shear stress concentrations ahead of the Izmit rupture tip. In this case, although the rupture extends 30 km west of Hersek Delta, the amount of slip resolved in the slip inversion is less than 20 cm within several kilometers of the western edge of the Izmit rupture. This, combined with the reduced normal stresses acting on the Princes’ Islands fault due to its location within the extensional deformation quadrant of the Izmit rupture, are the likely causes for the greatest Coulomb stress increases being located on the Princes’ Islands fault. If the Princes’ Island fault ruptured in the 1963 earthquake (scenario 3b, Fig. 4.7f), initiation and continued westward rupture along the Çinarcık fault is the most likely future rupture scenario according to the stress changes.

The results show that, in all cases, all of the Marmara faults receive increases in Coulomb stress due to the combined effects of all earthquakes since 1999. However, the fault segment receiving the greatest stress change is not the same for each scenario (Fig. 4.7, col. 2, Table 4.4). In four of the six scenarios, the Çinarcık fault receives the greatest stress change. In one scenario, the Çinarcık and Princes’ Islands faults receive approximately equal stress increases, and in another the Princes’ Islands fault receives the greatest increase. In all tests, the Armutlu fault receives stress increases less than those on the Çinarcık and Princes’ Islands faults, but of the same order of magnitude. The direction of the maximum shear stress vector along the faults, in all cases, never deviates by more than ±20° from pure right-lateral shear.

**Future westward rupture propagation**

Having analyzed which of the faults in the eastern Marmara Sea are likely to fail for each scenario, we now focus on the potential for the initial rupture to extend further west. We consider geometry of fault mapped further west and calculate Coulomb stress changes produced by the scenarios of rupture in Figure 4.7 that most favor failure of each of the three different major eastern Marmara Sea fault segments (Princes’ Islands, Çinarcık, and Armutlu faults). In the future failure scenario for each eastern Marmara fault, we apply uniform slip from 0 to 15 km depth with a magnitude derived from an
Table 4.4. Maximum and average Coulomb stress changes on eastern Marmara Sea faults. Fault locations and rupture scenarios are shown in Figure 4.7. Boldface values indicate the greatest stress change in each test.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>1a</th>
<th>1b</th>
<th>2a</th>
<th>2b</th>
<th>3a</th>
<th>3b</th>
</tr>
</thead>
<tbody>
<tr>
<td>Izmit rupture western termination</td>
<td>15 km west of Hersek Delta, along coast</td>
<td>15 km west of Hersek Delta, along coast</td>
<td>20 km west of Hersek Delta, up to Princes' Islands fault</td>
<td>20 km west of Hersek Delta, up to Princes' Islands fault</td>
<td>30 km west of Hersek Delta, along coast</td>
<td>30 km west of Hersek Delta, along coast</td>
</tr>
<tr>
<td>Fault ruptured in 1963 event</td>
<td>Çınarcık</td>
<td>Princes' Island</td>
<td>Çınarcık</td>
<td>Princes' Island</td>
<td>Çınarcık</td>
<td>Princes' Island</td>
</tr>
</tbody>
</table>

**Coulomb stress change (MPa)**

<table>
<thead>
<tr>
<th></th>
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</tr>
</thead>
<tbody>
<tr>
<td>West Izmit</td>
<td>0.32</td>
<td>0.29</td>
<td>0.81</td>
<td>0.38</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Princes' Islands</td>
<td>0.32</td>
<td>0.16</td>
<td>0.17</td>
<td>0.11</td>
<td>0.30</td>
<td>0.16</td>
<td>0.18</td>
<td>0.11</td>
<td>0.35</td>
<td>0.16</td>
<td>0.18</td>
<td>0.11</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East Çınarcık</td>
<td>1.16</td>
<td>0.54</td>
<td>1.17</td>
<td>0.50</td>
<td>0.29</td>
<td>0.17</td>
<td>0.26</td>
<td>0.16</td>
<td>0.35</td>
<td>0.16</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>West Çınarcık</td>
<td>0.14</td>
<td>0.07</td>
<td>0.19</td>
<td>0.10</td>
<td>0.13</td>
<td>0.07</td>
<td>0.09</td>
<td>0.07</td>
<td>0.16</td>
<td>0.08</td>
<td>0.28</td>
<td>0.13</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Armutlu</td>
<td>0.21</td>
<td>0.06</td>
<td>0.09</td>
<td>0.06</td>
<td>0.22</td>
<td>0.06</td>
<td>0.12</td>
<td>0.06</td>
<td>0.25</td>
<td>0.07</td>
<td>0.16</td>
<td>0.07</td>
<td></td>
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</tbody>
</table>
empirical relationship between surface rupture length and moment magnitude (Table 4.5) (Wells and Coppersmith, 1994).

The potential rupture length of an earthquake rupture initiating on the Princes’ Island fault segment would depend on the ability for the rupture continue onto the vertical, nearly east-west trending Central Marmara fault (Muller and Aydin, in press). Coulomb stress changes on the Central Marmara fault, calculated for either a strike-slip or normal-slip earthquake on the Princes’ Islands fault (see Table 4.5), are greatly increased near the intersection of this fault with the Princes’ Islands fault and are positive for the entire fault segment (Fig. 4.8a). For this hypothetical scenario, the uniform slip on the Princes’ Island fault exaggerates the magnitudes of the stress change close the fault intersection; therefore we consider the results to be qualitative. Nevertheless, it is clear that the relative orientations of the Princes’ Islands fault and the Central Marmara fault favor throughgoing rupture from the standpoint of static stress change.

The length of rupture associated with a Çinarcik fault earthquake would strongly depend on the accuracy of the fault configuration showing no mapped connection between this fault and the Central Marmara fault further west (Armijo et al., Le Pichon et al., Okay et al., Parke et al.). Strike-slip earthquake ruptures have been not jump between fault segments separated by more than 5-10 km (Harris and Day, 1993), so it would be unlikely to propagate further west. Rupture of this Çinarcik fault geometry would also decrease the Coulomb stresses acting on the Central Marmara and Imrali fault further west (Fig. 4.8b; Table 4.6), perhaps impeding or delaying the continued migration of slip through the central and western portions of the NNAF within the Marmara Sea.

In many of our stress change scenarios, the Armutlu fault receives stress changes greater than 0.2 MPa, even though in no case is it the segment with the greatest stress increase. The scenario in which İzmit rupture has propagated along the coast past Yalova (Fig. 4.7e) shows the greatest possibility for future failure along the Armutlu fault. Despite the large change in fault strike, the Coulomb stress is significantly increased along the Imrali fault (Fig. 4.8c). This is primarily because the fault is located in what would be the extensional quadrant of the Armutlu rupture segment. Transition of slip between these fault segments would produce a rupture with a combined length exceeding 80 km.
Table 4.5. Rupture parameters for the eastern Marmara faults used in potential future rupture scenarios.

<table>
<thead>
<tr>
<th>Fault</th>
<th>Area* (km²)</th>
<th>M_w†</th>
<th>Mo‡ (10¹⁹ Nm)</th>
<th>Ave. slip§ (m)</th>
<th>Rake (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Princes' Island (SS)</td>
<td>507</td>
<td>6.74</td>
<td>1.29</td>
<td>0.85</td>
<td>180</td>
</tr>
<tr>
<td>Princes' Island (NS)</td>
<td>507</td>
<td>6.69</td>
<td>1.08</td>
<td>0.71</td>
<td>-90</td>
</tr>
<tr>
<td>Çinarcık</td>
<td>1230</td>
<td>7.13</td>
<td>4.95</td>
<td>1.34</td>
<td>180</td>
</tr>
<tr>
<td>Armutlu</td>
<td>588</td>
<td>6.80</td>
<td>1.58</td>
<td>0.90</td>
<td>180</td>
</tr>
</tbody>
</table>

* Faults extend from surface to 15 km depth along surface traces in Fig. 4.8. Fault dips are given in Figure 4.2b.
† Calculated from empirical relationship for strike-slip earthquakes $M_w = 3.98 + 1.02 \times \log(A)$, where $A$ is rupture area (Wells and Coppersmith, 1994).
‡ Calculated from empirical relationship for normal-slip earthquakes $M_w = 3.93 + 1.02 \times \log(A)$ (Wells and Coppersmith, 1994).
§ Scalar seismic moment, $M_o = 10^A (1.5 \times M_w + 9)$ (Shearer, 1999).
§ Average fault slip, $u = M_o / (\mu A)$ (Shearer, 1999).
Figure 4.8. Coulomb stress changes on faults in the central Marmara Sea due to potential future eastern Marmara ruptures. Central Marmara fault = thin green line. Armutlu fault = thin purple line. Imralı fault = thin blue line. a) A case in which the initial future rupture occurs on the Princes' Island fault (thick green line). Stress plot on right shows stress changes for strike-slip Princes' Islands earthquake. Solid lines show results for the strike-slip mechanism and dashed lines for a normal dip-slip mechanism (Table 4.5). b) A case in which the initial future rupture occurs on Çınarcık fault (thick red line). c) A case in which the initial future rupture occurs on Armutlu fault (thick purple line).
Table 4.6. Maximum and average Coulomb stress changes on Marmara Sea faults west of potential eastern Marmara Sea ruptures. Fault locations and rupture scenarios are shown in Figure 4.8.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>1</th>
<th>1</th>
<th>2</th>
<th>3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Future rupture segment</td>
<td>Princes' Island fault (strike-slip offset)</td>
<td>Princes' Island fault (normal-slip offset)</td>
<td>Çinarcik fault</td>
<td>Armutlu fault</td>
</tr>
<tr>
<td>Central Marmara</td>
<td>1.42</td>
<td>0.15</td>
<td>1.28</td>
<td>0.06</td>
</tr>
<tr>
<td>Armutlu</td>
<td>0.38</td>
<td>0.08</td>
<td>0.02</td>
<td>0.00</td>
</tr>
<tr>
<td>Imrali</td>
<td>0.01</td>
<td>-0.05</td>
<td>0.06</td>
<td>0.00</td>
</tr>
</tbody>
</table>
Discussion

By considering the uncertainties in the rupture geometries of the 1963 Yalova earthquake and the western termination of the 1999 Izmit earthquake, our analysis represents a systematic investigation of the Coulomb stress changes on eastern Marmara Sea faults due to the most plausible scenarios. Assuming that static stress changes do provide a measure of the future failure tendency for large earthquakes (e.g. King et al., 1994; Stein et al., 1997), our results show that predicting future rupture scenarios depends on accurately constraining the recent earthquake locations and slip distributions (Muller et al., 2003). For the majority of our stress change scenarios, the Çinarcik fault receives a greater stress increase than the Princes’ Islands fault segment. In an analysis of the stress changes from the Izmit earthquake alone, Pinar et al. (2001) reach the same conclusion. This result does not support the notion that combined failure along the Princes’ Island and Central Marmara faults is the only, or most likely future rupture scenario (e.g. Le Pichon et al., 1999) or the suggestion that the stress changes on the Princes’ Islands and Çinarcik fault are essentially identical (e.g. Hubert-Ferrari et al. 2000). Our results instead show that several future scenarios are possible and that, in the case of the Çinarcik fault rupture, a future earthquake may not be as hazardous because of the limited length of such a rupture if the fault is not connected to other Marmara Sea fault strands further west.

One way to evaluate the effects of different stress change magnitudes on different faults is to compare the magnitude of the stress change with the long-term stress loading rate on each fault due to plate motion. GPS measurements (Reilinger et al., 1997) and elastic modeling (Flerit et al., 2003; Muller and Aydin, in review) of slip along the plate boundary structures in northwest Turkey suggests that 30 mm/yr of right-lateral slip on the NAF east of the Marmara Sea and 24 mm/yr on the NNAF and 6 mm/yr on SNAF within the Marmara Sea region are appropriate long-term slip rates. These slip rates applied to plate boundary with a locking depth of 15 km (Muller and Aydin, in review) provides approximate stressing rates of ~1.5x10^{-2} MPa/yr at 7.5 km on the faults of interest. The stress changes from scenarios la and lb have therefore advanced the rupture time by approximately 30 years for the East Çinarcik fault. In the other scenarios, the averages stress changes shorten the time until rupture on the Princes’
Islands and West Çinarcık fault by approximately a decade relative to their long-term stressing rate.

Several factors that we have not included in our model could affect the future rupture scenario in the eastern Marmara Sea. First, the location of the 1894 earthquake that caused widespread damage in the Bay of İzmit between the cities of Yalova and Sapanca is not very well constrained. From extensive review of historical damage reports, Ambraseys and Finkel (1991) suggest that the rupture occurred in the Bay of İzmit and further east. Hubert-Ferrari et al. (2000) support this location and suggest the event ruptured the mountainfront fault between the 1967 Mudurnu Valley earthquake and the 1999 İzmit earthquake. Parsons et al. (2000) suggest, however, that the rupture occurred within the Bay of İzmit and extended further west along the Armutlu fault. The latter interpretation would have relieved some of the long-term stress accumulation in the eastern Marmara Sea, whereas the earlier interpretation would have added an increased stress perturbation. The locations of damage reported in this earthquake, including İstanbul, are similar to those reported in the İzmit earthquake; therefore, it does not appear that rupture extending far along the Armutlu fault is necessary to explain the damage pattern. We therefore hesitate to rule out the Armutlu fault as a potential candidate for future rupture based on the 1894 rupture location interpreted by Parsons et al. (2000).

Second, the partitioning of loading from deep slip along the plate boundary loads the faults onto the shallower eastern Marmara Sea faults is not well-constrained. GPS data is able to show that slip below the locked shallow faults is several times greater along the NNAF than along the SNAF (Straub et al, 1997), but within the shallow NNAF strands themselves, it is unclear if certain faults are being loaded with a greater stressing rate. A shallow fault preferentially aligned above the focus of deep plate boundary motion would likely fail more frequently than faults further from the plate boundary.

In this paper we do not investigate the stress changes resolved on the fault geometry in the western Marmara Sea. In focusing on the eastern Marmara Sea region, we are by no means implying that the next large earthquake in the Marmara Sea could not occur further west, in the region of increased Coulomb stress east of the 1912 Ganos earthquake (Fig. 4.6). We recognize that the stress change scenario in this region will be affected by
uncertainty in the location of the eastern termination of the Ganos earthquake. This $M_r = 7.4$ earthquake rupture only 40 km on land before entering the Gulf of Saros to the west and the Marmara Sea to the east, suggesting that a significant portion of its rupture occurred offshore to the east or west. Armijo et al. (2003) suggest that a set of surface breaks extending 60 km east in to the Marmara Sea from the onshore rupture, as observed from a submarine remote operated vehicle, may be associated with the 1912 rupture. Evaluation of earthquake-induced stress changes in the western Marmara Sea region would therefore need to consider the ambiguities in the 1912 rupture geometry as well as the complex fault geometries bounding the Tekirdag and Central Marmara basins (Fig. 4.2a).

**Conclusions**

We find that earthquake-induced stress changes of several tenths of an MPa load faults in the eastern Marmara Sea for each possible different location of the 1963 Yalova earthquake and western termination of the 1999 Izmit earthquake. In four of the six permutations of these past rupture locations that we test, a portion of the Çınarcık fault receives the greatest average stress change. In one case, when Izmit earthquake slip extending up to 30 km west of the Hersek Delta and the 1963 rupture occurred on the Çınarcık fault, the Princes’ Islands fault receives the greatest stress increase. For the case where the Izmit rupture propagated straight west and the 1963 event was on the Çınarcık fault, the stress increases on the Princes’ Island and portions of the Çınarcık fault are comparable. The magnitudes of the stress increases due to the recent perturbations relative to the long-term tectonic stressing rate suggest that the time before future rupture is advanced by approximately 10 to 30 years.

Based on the mapped fault configuration extending west from these faults, rupture along the Princes’ Island fault is likely to be the most hazardous due to its proximity to Istanbul and because its intersection geometry with the Central Marmara fault favors continuous propagation of westward rupture. This scenario would result in a cumulative rupture length of at least 80 km before the rupture intersected the Central Marmara basin. In contrast, rupture of the Çınarcık fault would not likely exceed 50-60 km based on its isolated western termination and therefore may represent a scenario more likely in terms
of stress change, but with slightly less associated hazard to Istanbul. Earthquake-induced average stress changes on the Armutlu fault, while positive, do not exceed those on the other eastern Marmara faults for any tests. If, however, rupture were to initiate on this segment, then static stress changes would favor the transition of westward rupture onto the northwest-trending I Mrali fault segment, thereby producing a large combined rupture length.

Our results show that identification of the fault segment receiving the greatest earthquake-induced stress increase can vary depending on minor changes in local rupture geometry and slip distribution, thus highlighting the need for accurate surface mapping or active fault and earthquake rupture geometries as well as quality geodetic or seismological data for modeling earthquake slip distributions.

Acknowledgements

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