Seven years of postseismic deformation following the 1999, M = 7.4 and M = 7.2, Izmit-Düzce, Turkey earthquake sequence

The MIT Faculty has made this article openly available. Please share how this access benefits you. Your story matters.

| As Published | http://dx.doi.org/10.1029/2008jb006021 |
| Publisher | American Geophysical Union (AGU)/Wiley |
| Version | Author's final manuscript |
| Accessed | Sun Jul 30 14:46:26 EDT 2017 |
| Citable Link | http://hdl.handle.net/1721.1/106538 |
| Terms of Use | Creative Commons Attribution-Noncommercial-Share Alike |
| Detailed Terms | http://creativecommons.org/licenses/by-nc-sa/4.0/ |
Seven Years of Postseismic Deformation following the 1999, M = 7.4, and M = 7.2, Izmit-Düzce, Turkey Earthquake Sequence

S. Ergintav¹, S. McClusky², E. Hearn³, R. Reilinger², R. Cakmak⁴, T. Herring², H. Ozener⁴-⁶, O. Lenk⁵, and E. Tari⁶

1. TUBITAK, Marmara Research Center, Earth and Marine Sciences Institute, Gebze, Kocaeli, Turkey.
2. Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA, USA.
3. Department of Earth and Ocean Sciences, University of British Columbia, Vancouver, BC, Canada.
4. Kandilli Observatory and Earthquake Research Institute, Bogazici University, Istanbul, Turkey.
5. General Command of Mapping, Ankara, Turkey.
6. Istanbul Technical University, Istanbul, Turkey.

Abstract.

We report the results of seven years of postseismic deformation measurements using continuously recorded and survey mode GPS observations for the 1999 Izmit-Düzce earthquake sequence. Resolvable, time-dependent postseismic changes to the pre-earthquake interseismic velocity field extend at least as far as the continuous GPS station in Ankara, ~200 km SE of the Izmit rupture. Seven years after the earthquake sequence, the relative postseismic velocity across the North Anatolian Fault (NAF) reaches ~10 mm/yr, roughly 50% of the steady-state interseismic rate, with the highest postseismic velocities within 40 km of the coseismic ruptures. We use a sequence of logarithmic time functions to fit GPS site motions. Up to three characteristic time behaviors with time constants of 1 day, 150 days, and 3500 days are necessary to fit all the transient motion observed at the continuous GPS stations. The first term is mainly required for the component of site motion parallel to the NAF at near-field continuous GPS sites, strongly implicating rapid, shallow afterslip. The intermediate and longer-term postseismic velocity components reflect more broadly distributed strain, with a symmetric double-
couple pattern suggestive of either localized, deep afterslip or viscoelastic relaxation of the upper mantle and/or lower crust. In two areas this pattern is superimposed on north-south extension centered on the NAF. In the Marmara Sea, the north-south extension may result from triggered aseismic slip along the Princes Island and other normal faults. This slip could be the response of coseismically weakened faults to background extensional stress in the Marmara Sea region.

1. Introduction
Plate-boundary earthquakes are understood to be episodic, “instantaneous” displacements of the Earth’s upper crust, accommodating the roughly steady relative motions of tectonic plates, or blocks comprising the plate boundary. Geodetic observations constrain the rates of relative motion across faults, and are used to deduce kinematic fault characteristics such as locking depth (Savage and Burford, 1973). How the relative motions of adjacent blocks are accommodated below the upper crust remains a subject of debate, in large part because of our limited knowledge of the rheology and structure of the middle and lower crust, and upper mantle, around active faults.

Coseismic offset along a fault imparts an instantaneous stress change on the adjacent lithosphere, including near-field fault segments. Monitoring the spatial and temporal character of deformation following an earthquake therefore provides constraints on the mechanical processes by which this instantaneous coseismic stress is dissipated in space and time. Because coseismic slip is in general not uniform along the fault surface (e.g., Mai, 2007), a component of the postseismic motions may result from afterslip on coseismically loaded sections of the fault, that is, along patches that slipped in the earthquake much less than adjacent areas of the fault. Coseismic stresses in the lower crust or upper mantle may relax via viscous creep, loosening the grip of relaxing material on its stiffer surroundings (such as the base of the elastic upper crust). Thus, measurements of postseismic deformation can help constrain both fault zone characteristics (e.g., frictional or viscous parameters) and crustal/upper mantle rheology. It remains a challenge to separate the effects on surface motions from these fundamentally different mechanical processes.
Understanding the mechanics responsible for postseismic surface deformations is also important for seismic hazard estimation since postseismic strain release along one segment of the fault may push adjacent segments closer to failure (e.g., King et al., 1994; Stein et al., 1997). The M=7.2, 1999 Düzce earthquake that occurred approximately three months after the Izmit event and extended the Izmit coseismic fault 40 km to the east may well have been induced by coseismic and postseismic stress increases from the Izmit event (Hearn et al., 2002). Parsons (2004) documents the increase in probability of large earthquakes in the Marmara Sea region due to Izmit coseismic and postseismic stress changes.

In this paper we present the results of seven years of geodetic monitoring of postseismic motions for the 1999, Izmit-Düzce earthquake sequence. This data set is particularly instructive because the wealth of pre-earthquake geodetic constraints on strain accumulation in the region surrounding the fault (Kahle et al., 1998; Ayhan et al., 2002; McClusky et al., 2000; Straub et al., 1997). These data allow us to characterize the velocity field late in the earthquake cycle, that is, during the ten years just prior to the Izmit earthquake sequence. Here, we assume that crustal strain accumulation around the fault will decay to these same spatial patterns and rates at some future time, presumably prior to the next earthquake on this fault segment. We justify this assumption by noting the relative insensitivity of GPS velocity profiles across the North Anatolian Fault (NAF) to time since the most recent major earthquake (e.g., Ayhan et al., 2002; Reilinger et al., 2006). This assumption allows us to estimate longer-term time behavior than could be done with the existing short record of postseismic motions (i.e., compared to the estimated 250 to 400-year earthquake repeat time [e.g., Ambraseys, 1970, Hartleb et al., 2006]).

2. The Izmit-Düzce Earthquake Sequence

The Izmit-Düzce earthquake sequence (M = 7.4, 17/08/99, and M = 7.2, 12/11/99) occurred along the western segment of the NAF, east of the Marmara Sea (Figure 1). The NAF is a major right-lateral, strike slip, vertical fault, extending about 1200 km from the
Karliova Triple Junction (KTJ) in eastern Turkey to the northern Aegean Sea, accommodating the westward motion of the Anatolian plate relative to Eurasia (e.g., Sengor et al., 2005). The fault has an approximately uniform present-day slip rate of ~25 mm/yr along its entire 1200 km length in Turkey (McClusky et al., 2000, Reilinger et al., 2006). The present-day GPS slip rate is approximately 20% greater than geological estimates of the Holocene slip rate along the central part of the NAF (Kozaci et al., 2007), although it is not known if this difference is due to real rate changes or to the fact that coseismic surface offsets are observed to be smaller than offsets on deeper parts of the fault (e.g., Feigl et al., 2002), presumably because of inelastic deformation near the fault.

The Izmit earthquake represents the latest in a series of major (M > 7) 20th-century earthquakes that broke the entire length of the NAF, excluding the segment in the Sea of Marmara (Ambraseys, 1970, Toksoz et al., 1979) (Figure 1). Surface offsets along the exposed Izmit segment of the NAF vary along strike, reaching a maximum of around 5-6 m (Barka et al., 2002). The Izmit event was followed three months later by the Düzce, Turkey earthquake (M=7.2) that extended the Izmit surface break about 40 km to the east with right lateral surface offsets comparable to those for the Izmit event (Ayhan et al., 2001). Together these events filled about a 150 km segment of the Marmara seismic gap as identified prior to the 1999 events (Toksoz et al., 1979; Stein et al., 1997). The remaining seismic gap in the Sea of Marmara is a major source of concern for earthquake hazards because of the close proximity to Istanbul (e.g., Parsons, 2004).

3. Interseismic and Izmit-Düzce coseismic deformation

Interseismic deformation around the NAF has been monitored with GPS networks since 1988 (e.g., Kahle et al., 1997; McClusky et al., 2000; Reilinger et al., 2006). Elastic models based on these data indicate a slip rate of ~25 mm/year along the central NAF (McClusky et al., 2000), and 24 and 4 to 10 mm/yr on the northern and southern strands of the NAF in the Marmara Sea region (Meade et al., 2002). Figure 2 shows, and Table S1 lists, pre-Izmit earthquake GPS site velocities in a Eurasia-fixed reference frame, predicted by the Reilinger et al. (2006) elastic block model for all sites in the postseismic
study region. These model velocities are based entirely on measurements made before the Izmit earthquake sequence. Also shown on Figure 2 are the residuals to the block model fit in this region. The broader velocity field for the E Mediterranean region indicates that Anatolia is rotating counterclockwise with respect to Eurasia at $1.2^\circ \pm 0.02^\circ$/Myr about a pole located at $30.8^\circ \pm 0.8^\circ$N, $32.1^\circ \pm 0.7^\circ$E (Reilinger et al., 2006). This block rotation results in westward motion of northwest Anatolia at ~20 – 25 mm/yr relative to Eurasia and NE-SW oriented extension of the Marmara Sea region, likely related to forces associated with the Hellenic trench (e.g., Oral et al., 1993, Meijer and Wortel, 1997, McClusky et al., 2000).

Coseismic displacements for the 1999 Izmit earthquake were measured at 51 GPS sites (Reilinger et al., 2000). The displacements, that reached magnitudes of up to 2 m, are consistent with right-lateral slip along the NAF. Layered elastic models suggest slip to a depth of 20 km, and a total geodetic moment of $2.3 \times 10^{20}$ N m (Hearn and Bürgmann, 2005). Most of the coseismic strain is concentrated within 100 km of the rupture, with maximum displacements at the closest GPS sites.

4. Izmit earthquake postseismic deformation

The Izmit earthquake was immediately followed by observable, transient deformation of the Earth’s surface within about 100 km of the rupture (Ergintav et al., 2002). By the time of the Düzce event (87 days later), this postseismic deformation had released the equivalent of 20% of the estimated coseismic moment (Reilinger et al., 2000). Maximum postseismic site displacements during this time interval were about 50 mm. The time-dependence of the postseismic site motions was modeled using simple exponential functions with a decay constant of 57 days (Ergintav et al., 2002). The early postseismic deformation has been attributed to afterslip on the NAF, in areas that were loaded by the patchy coseismic slip (Bürgmann et al., 2002; Hearn et al., 2002). Subsequent postseismic velocities exceed values that would be consistent with afterslip driven solely by the coseismic stress (Hearn et al., 2006 and 2008).

4.1. Izmit-Düzce postseismic GPS measurements
Figure 2 shows names and locations of all of the GPS sites where Izmit postseismic velocity estimates have been made. Twenty-eight of these are permanent, or temporary, continuously operating GPS sites (CGPS sites), and the others are survey-mode GPS (SGPS) sites, where the position of a monument is surveyed at irregular intervals. GPS control is available along the entire coseismic rupture, but is most dense along the central segment (29.5 to 30.5° E) and quite sparse along the eastern segment. GPS sites extend only about 50 km north of the coseismic fault before reaching the Black Sea; coverage extends further from the fault to the south, but many more distant sites have more limited postseismic survey observations. The longest CGPS time series are from the sites in Ankara (ANKR) and Istanbul (KANT; Kandilli Observatory) from which we obtain the best-constrained pre-earthquake velocities. Other CGPS sites have either shorter pre-earthquake records (e.g., DUMT, KANT, MADT, MER1, TUBI) or were installed after the earthquake (e.g., ULUT, KART, ERDT, BOZT, ISTA). Several temporary CGPS sites were removed after one year and therefore mainly constrain early postseismic motions (BEST, MURT, UCG2, HAMT). ABAT and YIGT were installed after the Düzce earthquake and only operated for just over six months.

Post-earthquake survey observations (Figure 2) were started as early as one week following the Izmit event at several SGPS sites near the rupture, but this was too late to capture the earliest, rapid postseismic motions detected by the near-field CGPS sites. Table S1 gives details of the GPS data used in this study, including the postseismic occupation time intervals and the pre-earthquake modeled velocities of these sites.

In summary, the GPS data available along the central NAF have widely varying spatial and temporal characteristics making it difficult to compare results from different locations and to integrate the results to provide a coherent picture of the earthquake deformation cycle. In spite of these limitations, the geodetic dataset presented here is among the most extensive available for an earthquake deformation cycle.

To isolate the component of site motion due to postseismic effects, we subtract the estimated pre-earthquake site velocity as determined from the elastic block model for the
Marmara region (Reilinger et al., 2006). This is necessary because pre-Izmit velocities are not available at all of the GPS sites with postseismic velocities. The block model fits GPS site velocities to within their 95% confidence ranges at most sites, and fits velocities at 95% of the sites to within 2 mm/yr. Overall, the block model WRMS for sites in the Marmara Sea region is ~2 mm/yr. There are significant residuals at some sites (Figure 2) though these do not form a systematic pattern. For example, the observed southward velocity at TUBI is 4.1 mm/yr greater than predicted by the block model, and the north component of the FIS1 GPS velocity exceeds the model prediction by 5 mm/yr. In the central and western Marmara region, most GPS velocity residuals are within the 95% confidence interval. A notable exception is MER1, with a GPS velocity that is 6 mm/yr more westerly than the block model velocity. These seemingly small discrepancies become important when we address the later postseismic velocity field in the Marmara region.

4.2. Analysis of postseismic GPS data

We analyzed the GPS data from the Izmit/Düzce region using the GAMIT/GLOBK software (King and Bock, 2004, Herring, 2004). We used the GPS phase observations to estimate station coordinates, the zenith delay of the atmosphere at each station, and orbital and Earth orientation parameters (EOPS). The data were analyzed in daily segments, saving the loosely constrained geodetic parameters estimates and their associated covariances as quasi-observations. We provided orbital control and tied these regional measurements to an external global reference frame by including in the regional analysis data from 5-10 near-by continuously operating IGS stations for each day. The regional quasi-observations were then combined with quasi-observations from an analysis of GPS phase data from over 250 stations performed by the Scripps Orbital and Permanent Array Center (SOPAC) at UC San Diego (Bock et al., 1997).

The reference frame for our position estimates was defined by applying generalized constraints [Dong et al., 1998] while estimating a 6-parameter transformation (6 components of translation and rotation). Specifically we defined the reference frame by
minimizing the daily horizontal positions of 22 IGS stations within ~2000 km of the Izmit/Düzce region with respect to the IGS00 realization of ITRF2000 NNR frame (ITRF2000I) (Ray et al., 2004). The average daily WRMS fit of our solutions to the ITRF200I reference frame is ~1.2 mm. Finally we rotated our ITRF2000I station positions into a Eurasian fixed frame using the Eurasia - ITRF2000I Euler vector estimated in a simultaneous least squares solution using 32 stations located on the Eurasian plate in the ITRF2000I velocity solution and its associated full variance covariance matrix. The WRMS fit of the 32 sites to a non-deforming Eurasian plate is ~0.5 mm/yr. The positions of sites in the Izmit/Düzce region in this Eurasian fixed reference frame and their associated 1-sigma standard deviations are given in Table S2. North and east station positions are provided for each week of data observed between 1999 and 2006. Table S3 lists the times of jumps in the position-time series caused by equipment changes or physical damage to the antenna monument.

4.3. Modeling postseismic GPS site velocities.

We model the north and east components of the GPS time series as a linear combination of a secular inter-seismic velocity and logarithmic terms with up to three “characteristic” time constants.

\[
P_t = v.t + \sum_{i=1,3} A_i \ln \left(1 + \frac{t}{\tau_i}\right) \quad \text{1}
\]

\[P_t = \text{North and east component of position at time } t \text{ after the earthquake.}
\]

\[A_i = \text{Amplitude of the } i\text{th logarithmic term.}
\]

\[\tau_i = \text{the time constant of the } i\text{th logarithmic term.}
\]

\[t = \text{time since the earthquake.}
\]

\[v = \text{North and east component interseismic velocity of the site, from the block model (Reilinger et al., 2006).}
\]

To obtain instantaneous velocities, we take the derivative of this expression with respect to time:
\[ V_t = v + \sum_{i=1,3} \left[ \frac{A_i}{(1+t/\tau_i) \cdot \tau_i} \right] \quad \text{Equation 2} \]

Equation 2 reduces to:

\[ V_t = v + \sum_{i=1,3} \left[ \frac{A_i}{\tau_i + t} \right] \quad \text{Equation 3} \]

\[ V_t = \text{Station component (N or E) velocity at time } t \]

The amplitudes of the different log term coefficients \((A_1, A_2, \text{ and } A_3)\) give an estimate of the power of each characteristic time behavior in the GPS time series.

The time series for the position of the continuous station in Ankara (ANKR; Figure 3) shows an apparent change in slope (i.e., change in velocity) at the time of the Izmit earthquake. This change in slope requires a deformation process (or processes) that changes slowly in comparison to the time since the earthquake (7 years). Because our observations do not extend long enough after the earthquake to fully characterize this long-term behavior, we estimate a long-term time constant by assuming that the velocity of the Ankara CGPS station will return to its pre-earthquake value at some future time, presumably prior to the next earthquake. As indicated in Figure 3, a time constant of about 10 yrs (3500 days) provides the best fit to the time series and results in a long-term velocity that matches the well-determined pre-earthquake velocity of the Ankara station. We further assume that this same long-term time constant is appropriate for fitting the time series for all other sites in the network with sufficient observation records, and that these sites will also return asymptotically to their pre-earthquake velocities.

The characteristic times, \(\tau_1, \tau_2, \text{ and } \tau_3\), give the temporal behavior of each component during the 7-year interval of our post-earthquake observations. This representation gives a very good fit to the velocity time series, with rms deviations of similar order to the data uncertainties (Table S1). On the other hand, this functional representation of the time series is not unique. Other functions (e.g., exponential and spline functions) also provide
acceptable fits. Because of this, we put no special physical significance on a logarithmic representation. In addition, the specific “characteristic times” used here are not strongly constrained; they can vary by a factor of two without significantly degrading the fit to the time series. However, data from most of the continuous GPS sites do require all three time constants, with values in the range of a few days, hundreds of days, and a few thousand days. This is illustrated in Figure 4, where residual displacement-time data for the CGPS site TUBI are shown in sequence, after each of the three logarithmic function components is successively removed. Figure 4 clearly shows that removing one or even two of the terms is insufficient for removing a systematic time-dependence from the data (in both horizontal components).

Table S1 lists the coefficients of the different log terms for the east and north components of motion for each station observed. We emphasize again that the data that forms the basis for these plots varies widely in terms of time span and number of observations. Accordingly, the apparent absence of a short time constant term at some sites may be due to the absence of early postseismic observations, rather than evidence that such deformation did not occur at that location. Most of the SGPS sites are missing early observations and their velocities are fit by just a single logarithmic term with a characteristic decay time of ten to 100 days, consistent with SGPS observations from other large, continental strike-slip earthquakes (e.g., Owen et al., 2002 and Shen et al., 1994).

Likewise, stations with short total records have insufficient observations to constrain the longest time constant terms. Even CGPS stations with a full 7-year record of postseismic motions are insufficient to constrain the longest time constant term. We can only estimate this term by requiring that the site velocity approaches the pre-earthquake velocity at some time following the earthquake.

4.4. Time dependence of GPS postseismic velocity field

With the functional representations for the time behavior of GPS sites during the postseismic period, it is possible to generate a consistent set of site velocities at any time
after the earthquake sequence. Figures 5a, 5b, and 5c show snapshots of postseismic, GPS site velocities one week, six months, and six years after the Izmit earthquake. (Figures showing postseismic velocities after one month, three months, one year, and three years, are included in the online supplement.) Since we do not know the physical mechanism(s) driving the observed deformation, we do not extrapolate our post-earthquake velocity estimates beyond the time of our observations. The earliest postseismic velocities are largest near the Izmit rupture, and the east (fault-parallel) component is dominant (as it is for most of the near-field coseismic displacements, Reilinger et al., 2000). This confirms that the early postseismic deformation is due to afterslip (Ergintav et al., 2002, Hearn et al., 2002, Bürgmann et al., 2002), and at least some of this afterslip occurs in the top few kilometers of the crust. Ten days after the Izmit earthquake, the east velocity component at GPS site KOS1 was 1.1 m/yr, ± 48 mm/yr. The next-highest east velocities are 550 to 600 mm/yr, at GPS sites BEST, KOPI, and UCGT, with one-sigma errors from 10 to 80 mm/yr.

After six months (Figure 5b), the postseismic velocity field is somewhat broader in scale, with larger fault-normal velocity components at many sites. This suggests a deep-seated process, such as afterslip localized below the central part of the rupture, viscoelastic relaxation of the lower crust and/or upper mantle, or a combination. Large fault-parallel velocities in the near field suggest continuing, shallow afterslip. Maximum amplitudes of the east component are 80 ± 7 and 61±1.6 mm/yr at BEST and SEFI. East velocity amplitudes at SISL, AGUZ, and KAZI are all 54 ± 1 to 3 mm/yr. (Site KOS1 was no longer being monitored six months after the Izmit earthquake.)

Six years after the Izmit earthquake, the maximum velocities (absolute values of the east component) are 5.7 to 7.4 ±1 to .3 mm/yr, at sites TUBI, SEFI, AGUZ, YANT, and UCGT (Figure 5c). Several other sites have velocities in the 4 to 5 mm/yr range. This indicates up to about 12 mm/yr of relative motion across the NAF, in excess of what was recorded prior to the Izmit earthquake.
The southward motion at GPS sites on the Armutlu Peninsula (Figure 5c) cannot be attributed to errors in the block model used to correct the postseismic velocities. The block model produces velocities comparable to GPS values, except at site FIS1, where the modeled velocity is more southwardly oriented than the GPS velocity (Figure 2). As a result, too little southward motion is removed from the postseismic velocity, and the magnitude of the southward velocity at this site is probably greater than indicated on Figure 5c.

East- to southeastward motion dominates in the western Marmara Sea region, at GPS sites located both north and south of the NAF (Figure 5c). Although many of these sites started operation in 2003 or later, and have large formal errors, several (MER1, SELP, MADT, and KRDT) have long time series. Velocities at the latter four sites clearly indicate eastward motion at 1-5 mm/yr. Velocities at these four sites also suggest about 1 to 3 mm/yr of opening (north-south extension) across the Marmara Sea, and this is corroborated by velocities at the other GPS sites in this region. In general, the eastward velocities cannot be attributed to our use of the Reilinger et al. (2006) block model, rather than GPS data, to correct the postseismic velocities for secular deformation. One exception to this is site MER1, where the block-modeled velocity was more westerly than the GPS velocity. The eastward velocity at this site six years after the Izmit earthquake (Figure 5c) may result solely from the secular velocity correction.

The maximum, total postseismic displacement after six years is about 150 mm, which is attained at two GPS sites that are 10 to 15 km from the rupture (SEFI and KAZI). About 2/3 of the displacement at these two sites accumulates during the first year, and by year 6, these values are increasing by just 5 mm/yr. Sites 30 km from the rupture (e.g., KANR) displaced about 90 mm in six years. (Extrapolated displacements from GPS sites which were operating during only part of the interval from 1999 to 2006 are not considered here.) For comparison, the maximum Izmit coseismic displacement was 2 meters. Table S4 lists the modeled north and east velocities and uncertainties for each site with postseismic observations, at one-week intervals. Velocities are only given within the time span of the actual GPS measurements.
5. Kinematic afterslip models

GPS site velocities were inverted for afterslip rate along the NAF, and elastic dislocation models incorporating the best slip rate solutions were used to forward model GPS site velocities, at several postseismic time intervals. We do not suggest that all postseismic deformation is literally due to afterslip; the dynamic processes causing Izmit postseismic deformation are addressed by Hearn et al. (2002, 2006, and 2008). Kinematic afterslip is treated here as a proxy for postseismic deformation due to afterslip or viscoelastic relaxation (Savage, 1990). We attempt to model the postseismic deformation as afterslip along the NAF in part to look for deformation which must result from other processes (such as triggered slip on un-modeled, dip-slip faults, e.g. Bonhoff et al., 2007).

In the inversions, afterslip was allowed at depths shallower than 40 km, and along strike to 50 km beyond the east and west ends of the Izmit and Düzce ruptures. We assume that slip occurs on a continuous, curved fault embedded in a uniform elastic halfspace. More realistic, layered elastic models would result in a higher total moment of afterslip, and increased afterslip in the middle and lower crust; Hearn and Bürgmann, 2005. The Izmit rupture geometry is from Feigl et al. (2002) and the Düzce rupture geometry is from Bürgmann et al. (2002). The Izmit rupture is assumed to be vertical and the Düzce rupture dips 51 degrees to the north. In the Marmara Sea, the NAF is extended to the west, approximately following the trace of the Main Marmara Fault (Demirbag et al., 2003; also the southern MMF trace of Imren et al., 2001). East of the Düzce rupture, we extend the fault to the east and assume that the northward dip is the same as for the Düzce rupture. The fault surface is divided into 4 km square patches, with 65 patches along strike and ten down dip.

Using linear least-squares inversion methods, we find a slip distribution that minimizes misfit to measured coseismic displacements weighted by measurement errors (i.e., the weighted residual sum of squares, or WRSS) while preserving smoothness of the slip distribution (e.g., Du et al., 1992; Bürgmann et al., 2002). We use the bounded variable least-squares (BVLS) method (Stark and Parker, 1995) to impose non-negativity (that is,
right-lateral slip, normal dip slip, or opening only). A finite-difference approximation of the Laplacian represents the roughness of the estimated slip distribution (Harris and Segall, 1987; Du et al., 1992), and the sum of this roughness (scaled by the smoothing parameter $b$) and the WRSS is minimized in the inversion. Hence, the extent to which smoothness is enforced at the expense of fit to surface displacements can be adjusted by varying $b$. We use a Tikhonov plot to select an optimal value of $b$, that is, a value that suppresses spurious structure in the slip distribution while not smoothing out salient features. The inversion requires as input a data kernel $G$, which is a matrix of Green’s functions relating slip on individual fault patches to surface deformation at all surface observation points. $G$ is calculated using the Okada (1985) analytical solution.

Figure 6 shows the distribution and rate of afterslip along the NAF that best fits the GPS velocity field one week, 180 days, and 6 years after the Izmit earthquake. A week after the Izmit earthquake, afterslip is clearly occurring in the top four km of the NAF. (Due to the distribution of GPS sites at this time, deeper afterslip cannot be ruled out.) The maximum afterslip velocity, 5.5 m/yr, occurs along the Golcuk segment. Forty-five days after the Izmit earthquake, the shallow slip rate has decreased to 1.4 m/yr, but it is joined by apparent, deeper afterslip below the eastern Golcuk and western Sapanca segments, and below the central Karadare segment, at a maximum rate of 2 m/yr and centered on a depth of 20-24 km. Later on, the east (Karadare) patch dominates, apparently slipping at 400 mm/yr and 300 mm/yr 3 and 6 years after the Izmit earthquake, and centered a bit deeper (at a depth of about 28 km). This deep “slip” patch probably is an artifact. Viscoelastic relaxation of the mantle or lower crust causes surface deformation in the form of a diffuse “double couple” pattern at the surface. Inversion of such surface velocities for fault slip will yield a deep patch of localized, apparent slip. A finite-element modeling study based on the GPS data presented here suggests that the first 2.5 years of Izmit postseismic deformation is due to a combination of afterslip and viscoelastic relaxation of the mantle and possibly the lower crust [Hearn et al., 2008].

Our early afterslip patterns and rates (particularly for deep afterslip) are comparable to the results shown by Bürgmann et al. (2002) for the first 80 days after the Izmit
earthquake. Our results are also qualitatively similar to patterns of total afterslip from the first thirty days after the Izmit earthquake, inferred from SAR interferograms by Cakir et al. (2003). Cakir et al. place the patch of maximum early afterslip about 20 km to the west of our western, high afterslip rate patch (Figure 6b; the location of this patch remains unchanged throughout the postseismic time interval). We obtain a maximum, cumulative afterslip of about 2 meters, beneath the western Sapanca segment. Most of this accrues during the first year after the Izmit earthquake.

According to Omer Emre (personal communication with Hearn, August 17, 2006), 40 centimeters of postseismic slip accrued over five years along the Golcuk segment of the Izmit rupture (based on successive measurements of the relative displacement of a stone wall at the Navy base in the Yuzbasilar district of the town of Golcuk). The maximum integrated slip over five years in this part of our model (about 20 km east of Golcuk) is about 70 mm, with nearly all accruing in the first 80 days. At Golcuk, we estimate about 15 cm of cumulative surface afterslip in the first 45 days (and none thereafter). Our estimates of surface afterslip are substantially less than reported at this one site, possibly because of the coarse representation of the NAF in our model, but the observed post-earthquake fault offset supports our contention that shallow afterslip is at least partially responsible for the observed postseismic deformation.

Table 1 shows how well dislocation models using the above slip solutions fit the GPS site velocities. In general, elastic dislocation models fit the east (fault-parallel) component of the velocities better than the north component. The dislocation models perform more poorly at later time epochs, particularly for the northward velocity component. Much of this misfit is due to apparent north-south extension across the Lake Sapanca region and the Marmara Sea, which are described below.

The vertical GPS site velocity component is excluded from the following discussion, though our GPS velocity solution includes the vertical component. For the dislocation models, the fit to vertical velocities is worse than it would be for a model with no vertical motion at all. However, due to the large uncertainties, the contribution of the vertical
velocity component to the total WRSS is consistently about a factor of ten smaller than
the contribution from the north component, and it is insensitive to changes in the
distribution and smoothness of slip.

Table 1

<table>
<thead>
<tr>
<th>Model/time</th>
<th>WRSS_E</th>
<th>WRSS_N</th>
<th>WRSS_all</th>
<th>%reduct_E</th>
<th>%reduct_N</th>
<th>%reduct</th>
</tr>
</thead>
<tbody>
<tr>
<td>10 days</td>
<td>556</td>
<td>254</td>
<td>810</td>
<td>92</td>
<td>63</td>
<td>89</td>
</tr>
<tr>
<td>45 days</td>
<td>818</td>
<td>403</td>
<td>1221</td>
<td>94</td>
<td>81</td>
<td>93</td>
</tr>
<tr>
<td>180 days</td>
<td>11260</td>
<td>371</td>
<td>1497</td>
<td>92</td>
<td>90</td>
<td>91</td>
</tr>
<tr>
<td>3 yrs</td>
<td>18573</td>
<td>5750</td>
<td>2.4323e4</td>
<td>72</td>
<td>56</td>
<td>70</td>
</tr>
<tr>
<td>3yrs_a</td>
<td>18233</td>
<td>3961</td>
<td>2.2194e4</td>
<td>73</td>
<td>70</td>
<td>73</td>
</tr>
<tr>
<td>3yrs_b</td>
<td>15657</td>
<td>3387</td>
<td>1.9044e4</td>
<td>78</td>
<td>71</td>
<td>77</td>
</tr>
<tr>
<td>3yrs_c</td>
<td>10155</td>
<td>4223</td>
<td>1.4378e4</td>
<td>85</td>
<td>68</td>
<td>82</td>
</tr>
<tr>
<td>3yrs_d</td>
<td>9329</td>
<td>2941</td>
<td>1.2270e4</td>
<td>86</td>
<td>78</td>
<td>85</td>
</tr>
<tr>
<td>6 yrs</td>
<td>10228</td>
<td>2426</td>
<td>1.265e4</td>
<td>66</td>
<td>25</td>
<td>62</td>
</tr>
</tbody>
</table>

a -- opening along the NAF and MMF
b -- PIF only - 20 degree S dip
c -- PIF and Armutlu dip slip plane - 40 degree S dip
d -- PIF and Armutlu dip slip plane - 20 degree S dip

Figures 7a, 7b, and 7c show observed and residual (observed minus modeled) GPS site
velocities one week, six months, and six years after the Izmit earthquake. As described
above, dislocation models fit the east velocity component better than the north velocity
component. The color background for these plots shows shear strain rate ($\varepsilon_{yx}$), illustrating
that most of the postseismic shear strain may be explained (kinematically) in terms of
afterslip on the ruptures and their lateral and down-dip extensions. (In these plots, the
positive x direction is east and the positive y direction is north.) Figures 7a, 7b, and 7c
also illustrate that shear strain is approximately symmetric around the NAF - that is, there
is no sign of a dramatic contrast in elastic properties of the crust across this segment of
the NAF.
Most of the east-west oriented normal strain rate ($\varepsilon_{xx}$) is also explained by these elastic dislocation models. Figure 8 shows the velocities and residuals six years after the Izmit earthquake, superimposed on a color background of $\varepsilon_{yy}$. This figure illustrates that the elastic dislocation model cannot explain $\varepsilon_{yy}$ (extension across the NAF). Subtracting the dislocation-modeled velocities does nothing to reduce $\varepsilon_{yy}$. This steady accumulation of fault-normal tensional strain is seen in data from all time epochs, but it is most obvious in the later postseismic velocity fields because velocities due to other postseismic processes are smaller. For infinite, strike-slip faults, elastic dislocation models should be able to reproduce velocity fields produced by layered, viscoelastic relaxation models (e. g., Savage, 1990). In the 3D case, this is not necessarily the case, but the residuals should be antisymmetric and there should never be extension or contraction centered on the fault. Note that Figure 8 does not include GPS sites further to the west, which are discussed in section 4.4 and shown on Figure 5c, and show 1-3 mm/yr of north-south extension across the Marmara Sea six years after the Izmit earthquake.

To address this extension in the eastern Marmara, we developed one simple model with fault-normal opening along the NAF (the Main Marmara Fault) in the Marmara Sea, and several models with dip slip on the Princes Island Fault (PIF) and on east-west oriented, north dipping normal faults bounding the Armutlu peninsula (Saroglu et al., 1992; Parke et al., 2002). Dips of 20 to 45 degrees were tested on all of these faults. With these models we attempted to explain the residual velocities three years after the Izmit earthquake (we chose three years because of the large number of occupied GPS sites near the Marmara Sea at that time). Results of these models are summarized on Table 1.

The NAF opening model explains the north-south strain across the Marmara, if 60 mm/yr of opening is allowed (in the top 4 km) on the easternmost 8 km of the MMF. Oddly, this model also requires 140 mm/yr of opening along a 40-km-long zone at a depth of 26 to 32 km below the Yalova and Golcuk segments. This model reduces misfits to the north residual displacement component, but by only a third (Table 1). Shallow dip slip on the PIF (5 to 10 mm/yr) explains the GPS residual site velocities at ISTA, KANT, YANT and BADT (north of the Marmara Sea) but not the residual southwestward velocities at sites
CINA, FIS1, and KUTE (on the Armutlu Peninsula). Thus, this slip does not explain all of the NAF-normal extension in the eastern Marmara region.

Dip-slip on a south dipping surface approximately 10 km below the Armutlu peninsula explains most of the NAF-normal extension in this region and significantly improves model performance for both the east and north velocity components. However, normal faults in this area are thought to dip north (Saroglu et al., 1992; Parke et al., 2002). None of the models with north-dipping faults on the Armutlu Peninsula could fit the GPS velocities in this region better than a model in which no slip was allowed on these features.

6. Discussion

6.1 Limited postseismic deformation and slip
Accelerated postseismic deformation accounts for relatively little of the total GPS site motion, or the total fault slip, over the course of an earthquake cycle. The limited contribution of accelerated postseismic deformation, and the similarity of strain profiles around parts of the NAF that have failed in major earthquakes at different times, suggests that interseismic GPS site velocities do not vary much interseismically. Hence, interseismic GPS velocities should be comparable to velocities predicted by block models based on geologic fault slip rates (e.g., Meade et al., 2002).

Most geologically-determined fault slip rates for the NAF are smaller than the geodetic slip rate (e.g., Kozaci et al., 2007, Hartleb et al., 2006). Given the insensitivity of shear strain rates along the NAF to time since the last large earthquake (60 to over 200 years), it seems unlikely that this discrepancy is due to accelerated postseismic deformation from 20th-century NAF earthquakes. Earthquake cycle models suggest that if earthquake recurrence times are irregular, GPS site velocities may differ from the predictions of block models based on long-term, geologic slip rates (Meade and Hager, 2004; Hetland and Hager, 2005). Changes in fault slip rates over geologically short timescales have been documented (e.g., tens of thousands of years, Oskin et al., 2007, Peltzer et al., 2001; and
thousands of years, Dolan et al., 2007, Weldon et al., 2004), and these changes are associated with changes in earthquake recurrence times (i.e., clustering). There is no evidence of clustering of large NAF earthquakes, sufficient to explain the difference between the GPS slip rate and very recent (2500 year) geological slip rates for the central NAF (Hartleb et al., 2006). Other causes of the slip rate discrepancy should be investigated. Some possibilities include systematically low fault slip rate estimates from geologic observations, or a recent acceleration in the NAF slip rate due to slowly evolving loads (Hetzel and Hampel, 2005), pore pressures (Chery and Vernant, 2006), or changes to fault zone rheology.

6.2 Generally symmetric velocity fields and strain rates
Postseismic velocities are roughly symmetric across the NAF, except in the eastern Marmara Sea region where the velocity field is complicated. This implies that broad-scale, lateral variations in the elastic structure are minor in the vicinity of the Izmit rupture. This is inconsistent with a factor-of-ten contrast in crustal elasticity across the NAF in the Marmara Sea region (Le Pichon et al., 2003). If even a smaller elasticity contrast were present in the study area, both the coseismic and the early postseismic data would have reflected this in an obvious manner.

Geophysical observations suggest a contrast in mantle viscosity across the NAF (e.g., Sandvol et al., 2001, Gok et al., 2000). This could yield a contrast in later postseismic and interseismic velocities on opposite sides of the NAF. Models of time-averaged deformation (ignoring earthquake cycle effects) suggest that it should be possible to infer a contrast in effective viscosity of one to two orders of magnitude across the NAF, from asymmetry of the secular GPS velocity field (Fischer, 2006). However, the postseismic GPS velocities (Figures 5 and 8) do not suggest asymmetry in fault-parallel velocities around the Izmit rupture.

6.3 Postseismic velocity changes that cannot be explained with our dislocation models
The late postseismic velocity field has two puzzling features. First, extension across the Marmara Sea is occurring 1-3 mm/year faster than before the Izmit earthquake. Second, many GPS sites in the western Marmara region (on both sides of the NAF) are moving more slowly toward the west (by up to 5 mm/yr) than they were before the Izmit earthquake. These observations are hard to explain unless (1) there has been a change in the distribution of strain among faults in the Marmara region or (2) there were systematic errors in the assumed pre-earthquake GPS site velocities.

One potential explanation for (2) would be that the block model used to estimate GPS site velocities was systematically biased. To explain the southeastward motions on Figure 5c, the north and west velocity components of the block-modeled velocities at GPS sites south and southwest of the Marmara would have to be larger than the GPS velocities. As indicated in Figure 2 and in Section 4.4 (see also Table S1 and Figure 8 in Reilinger et al., 2006) this is not the case. Although residuals (observed minus block-modeled velocities) exceed the 95% confidence limits at several sites in the region, no systematic bias is evident.

### 6.3.1 Lake Sapanca Extension

In the Lake Sapanca region, apparent extension across the NAF may result from a slight northward dip of the fault, which we modeled as vertical. In their InSAR studies of Izmit coseismic and early postseismic deformation, Cakir et al. (2003) found that they could not model the coseismic InSAR range change data in this same area, unless there was some component of dip slip (i.e., a rake of 176 degrees) and the NAF was assumed to dip 85 degrees to the north. Thus, afterslip might also include some component of dip slip. Geophysical confirmation of the NAF dip in this area is needed. If the NAF is vertical in the area of Sapanca Lake, then another explanation must be sought for the fault-normal extension in this area (for example, strain due to changes in the properties of sediments or damaged material in the stepover).

### 6.3.2 Marmara Sea Extension
A small coseismic contribution of dip slip on steeply dipping fault segments (as proposed for the Lake Sapanca region by Cakir et al., 2003) could potentially drive enough postseismic dip slip to explain the Marmara Sea extension. This would require coseismic slip extending well into the Marmara Sea along the MMF, with some dip-slip component. GPS and InSAR data do not suggest such coseismic slip (e.g., Reilinger et al., 2000) though there has been debate on this point (Karabulut et al., 2002). If we accept coseismic slip estimates based on GPS (Reilinger et al., 2000; Hearn and Bürgmann, 2005) and a suite of space geodetic methods (Feigl et al., 2002), it is unlikely that sufficient dip slip sufficient to cause 1 to 3 mm/yr of extension across the entire Marmara Sea could be driven by coseismic stresses. The maximum coseismic plus early postseismic shear stress loading available to drive dip slip is less than 0.1 MPa at the west end of the Yalova fault (based on modeled stresses from Hearn et al., 2002). Further into the Marmara Sea, along the MMF, coseismic and early postseismic stress changes (all components) are negligible.

If we assume that the Marmara - central NAF crustal stressing rate is approximately steady over time (i.e., consistent with tectonic features, Allmendinger et al., 2007), then the minimum principal stress ($s_3$) axis is oriented NNE in the Marmara Sea, and the region is in a tensional stress state (becoming transtensional as one travels east). Under this background stress, dip slip may occur across normal faults parallel to the NAF fault if the “strength” or loading of the fault decreases coseismically. Such slip would be consistent with some observations of microseismicity indicative of normal faulting in the Marmara Sea after the Izmit earthquake (Karabulut et al., 2002, Bonhoff et al., 2006).

Although dynamic stressing may explain rheology changes in the top few kilometers of the crust (due to microcracking), a coseismic change in fault zone rheology at greater depths, on faults far from the coseismic rupture, is harder to explain. The alternative way to redistribute the rate of slip along Marmara faults would be through changes in pore pressures and hence effective normal stresses (e.g., Chery and Vernant, 2006). We have insufficient data to distinguish between these two hypotheses.
Postseismic changes in the creep rates of local faults in the area of a major earthquake have been noted before, in the Eastern California Shear Zone (Hudnut et al., 2002). Hudnut et al. find that a 30 by 50 km block of crust which was essentially stationary relative to North America before the 1992 Landers earthquake sequence, began moving in a sense consistent with being “entrained in flow along with the Pacific Plate”, as a result of a coseismic change in the locus of shear strain, that is, in the distribution of creep rates among parallel, strike-slip faults. Something similar could be occurring with small, fault-bounded blocks in the Marmara Sea (e.g., Flerit et al., 2003).

6.4 The western Marmara postseismic anomaly
The eastward, late postseismic motion at GPS sites around the central and western parts of the Marmara Sea (e.g., most sites west of 29 degrees longitude on Figure 5c) is difficult to explain. Coseismic stresses are negligible in most of this region, and any accelerated slip on the NAF would result in increased westward (or southwestward) velocities at sites to the south. In the Marmara Sea region and to the south, the lithosphere is hotter and the crust thinner than in the rest of the study area (e.g., Pfister et al., 1998). However, preliminary viscoelastic models suggest that a locally low mantle viscosity in this area cannot cause such a large deviation from the classic, double-couple postseismic velocity field.

7. Conclusions
We report the results of seven years of postseismic deformation measurements using continuously recording and survey mode GPS (CGPS and SGPS) observations, for the 1999 Izmit earthquake sequence. Geodetic monitoring in the epicentral region was initiated more than ten years before this earthquake sequence, providing constraints on the rate and spatial pattern of pre-earthquake strain accumulation and coseismic motions. Geodetic monitoring was intensified following the first event and continues to the present time.
Postseismic motions show a similar spatial pattern to coseismic motions, being symmetric across, but more broadly distributed around, the coseismic ruptures. Resolvable postseismic changes to the velocity field extend at least as far as the CGPS station in Ankara, 200 km SE of the Izmit rupture. Seven years after the earthquake sequence, deviations from the interseismic velocity field observed late in the earthquake cycle (i.e., during the ten-year period just prior to the earthquake) are largest within about 40 km of the fault, reaching ~10 mm/yr, roughly 50% of the pre-earthquake rate. Present-day postseismic motions decrease further from the fault to a value of ~3 mm/yr at Ankara (~15% of pre-earthquake motion rate).

Postseismic site velocities decrease monotonically with time following the earthquake sequence at all sites. We use a sequence of logarithmic functions to characterize GPS site motions and find a good fit using three characteristic time behaviors with time constants of 1 day, 150 days, and 3500 days. We constrain the longest-term time behavior by requiring that site velocities decay to their pre-earthquake rates at some future time, presumably prior to the next earthquake on that segment of the fault. CGPS sites that were operating at the time of the Izmit earthquake, and that are within about 50 km of the rupture, require the one-day decay time term. This term is mainly required for the component of site motion parallel to the NAF, strongly implicating rapid afterslip on and possibly below the coseismic fault (as noted by Reilinger et al., 2000, Ergintav et al., 2002, Bürgmann et al., 2002, and Hearn et al. 2002). Most of the CGPS sites require all three characteristic time terms. However, the velocity of the Ankara continuous GPS station located 250 km from the epicenter is fit well with only the longest time series term.

While the earliest postseismic deformation is associated with shallow afterslip, based on its spatial character, the later postseismic velocity fields are more broadly distributed. The spatial pattern of later postseismic deformation is suggestive of either localized afterslip in the lower crust, or viscoelastic relaxation of the upper mantle and lower crust (i.e., a distributed double couple pattern). Finite-element modeling confirms that coseismic stress-driven afterslip and distributed viscoelastic relaxation of the mantle and/or lower
crust can explain the first 2.5 years of Izmit postseismic deformation (see Hearn et al., 2008). Gravity data from the eastern Marmara Sea also suggest that viscoelastic relaxation of the lower crust and/or upper mantle was occurring during the time interval covered by our survey (from 2003 to 2005; Ergintav et al., 2007).

The north-south component of postseismic deformation is not well explained by elastic dislocation or viscoelastic finite-element models. This is because north-south extension, centered on the Marmara Sea and Lake Sapanca, is superimposed on the distributed, “double-couple” postseismic velocity field. In the Marmara Sea region, some of this extension is consistent with triggered creep along the Princes Island normal fault. Most of the extension cannot be linked to dip slip on a particular fault. It may be due to slip on dipping NAF segments in the Marmara Sea that were weakened by the earthquake, driven by the background stress rather than the coseismic stress change.

**Acknowledgments.** Funding for this project was provided by TUBITAK CAYADAG Project No:103Y100, EU 6, Frame FORESIGHT Project (Contract no: 511139), and TUBITAK TARAL 1007 Project No: 105G019. This research was also supported by NSF grants EAR-0337497, EAR-0305480, and INT-0001583 to MIT, and by NSERC Discovery Grant RGPIN 261458-07 to Elizabeth Hearn at UBC. We gratefully acknowledge all of the research groups who have participated in measuring surface deformation around the NAF since 1988.

**Figure Captions**

**Figure 1.** Tectonic setting of the North Anatolian Fault (NAF), modified from Reilinger et al. (2000). Numbers along the NAF indicate years of large, 20th-century earthquakes, and red arrows indicate the associated rupture lengths.
Figure 2. GPS site names and pre-Izmit velocities. Blue circles indicate SGPS sites, red triangles represent temporary CGPS sites, and green diamonds indicate other CGPS sites. Black arrows indicate block-modeled velocities (Reilinger et al., 2006). Pink arrows are residuals, at GPS sites were pre-Izmit velocity estimates were available. Pink ellipses indicate 95% confidence limits for the pre-Izmit velocities.

Figure 3. ANKR displacement time series. Top two panels show north and east position time series, with weekly positions and one-sigma errors (68% confidence limits). Symbols are colored orange after August 17, 1999. Bottom two panels show these time series after removal of a single logarithmic term with a 10-year decay time (see text).

Figure 4. TUBI displacement time series, after sequential removal of the 1-day, 180-day, and 3500-day logarithmic function terms. The top top two panels in Figure 4a show the north and east displacement time series, and the bottom two panels show the result of subtracting the 3500-day logarithmic term from these time series. The top two panels of Figure 4b show the result of subsequently subtracting the 150-day logarithmic term from the time series. The east component clearly shows a remaining, rapidly decaying term. The bottom two panels of Figure 4b show the displacement time series after subtraction of the one-day logarithmic function term from the displacement time series shown in the top two panels.

Figure 5. GPS site velocities (a) one week, (b) six months, and (c) six years after the 1999 Izmit earthquake. These velocities have been corrected for “secular” deformation using the pre-Izmit velocity field shown on Figure 2. More velocity field snapshots are included in the online supplement (for one month, three months, one year, and three years after the Izmit earthquake). On Figure 5a, MMF is the Main Marmara Fault (based on Demirbag et al., 2003), AP is the Armutlu Peninsula, and PIF is the Princes Island Fault.
Figure 6. Kinematic afterslip models. Postseismic slip rates for 1 week, 180 days, and six years after the Izmit earthquake are shown. The slip rates shown here are solely the postseismic perturbation: they do not include secular aseismic slip. Yellow dashed lines show outlines of coseismic slip from Reilinger et al. (2000). Y, G, W, E, K, and D indicate the Yalova, Golcuk, East and West Sapanca, Karadare, and Düzce segments.

Figure 7. Modeled velocities and residuals, from elastic dislocation models using the slip rates shown in Figure 6. Figures 7a, 7b, and 7c show modeled (blue) and GPS (yellow) velocities (top panels) and residuals (bottom panels) for 1 week, 180 days, and 6 years after the Izmit earthquake. Background colors indicate shear strain rate ($\varepsilon_{yx}$), estimated from the GPS velocities (top panels) and from the residuals (bottom panels). Green lines show faults, including the north-dipping Düzce segment.

Figure 8. Modeled and GPS velocities, and residuals (as shown in Figure 7c) after six years, with background colors showing NAF-normal strain rate $\varepsilon_{yy}$. Areas of north-south extension in the Marmara Sea and around Sapanca Lake are not affected by subtraction of the modeled velocities. This means they cannot be modeled with afterslip along the NAF as it is represented in our dislocation models.

References


Bürgmann, R., M. E. Ayhan, E. J. Fielding, T. J. Wright, S. McClusky, B. Aktug, C. Demir, O. Lenk and A. Türkezer (2002), Deformation during the 12 November


Kozaci, O., J. Dolan, R. Finkel and R. Hartleb (2007), Late Holocene slip rate for the North Anatolian fault, Turkey, from cosmogenic $^{36}$Cl geochronology: Implications for the constancy of fault loading and strain release rates Geology, 35 (10), 867-870; DOI: 10.1130/G23187A.1.


Reilinger, R., N. Toksoz, S. McClusky, and A. Barka (2000), 1999 Izmit earthquake was no surprise, GSA Today, 10, 1-6.


