Viscoelastic Block Models of the North Anatolian Fault:  
A Unified Earthquake Cycle Representation of Pre- and Postseismic Geodetic Observations

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Abstract  Along the North Anatolian fault (NAF), the surface deformation associated with tectonic block motions, elastic strain accumulation, and the viscoelastic response to past earthquakes has been geodetically observed over the last two decades. These observations include campaign-mode Global Positioning System (GPS) velocities from the decade prior to the 1999 \( M_w \) 7.4 İzmit earthquake and seven years of continuously recorded postseismic deformation following the seismic event. Here, we develop a 3D viscoelastic block model of the greater NAF region, including the last 2000 yrs of earthquake history across Anatolia, to simultaneously explain geodetic observations from both before and after the Izmit earthquake. With a phenomenologically motivated simple two-layer structure (schizosphere and plastosphere) and a Burgers rheology (with Maxwell viscosity \( \log_{10} \eta_M \approx 18.6–19.0 \) Pa·s and Kelvin viscosity \( \log_{10} \eta_K \approx 18.0–19.0 \) Pa·s), a block model that incorporates tectonic plate motions, interseismic elastic strain accumulation, transient viscoelastic perturbations, and internal strain can explain both the pre- and post-Izmit earthquake observations with a single unified model. Viscoelastic corrections to the interseismic GPS velocity field with the unified model reach magnitudes of \( \sim 2.9 \) mm/yr. Geodetically constrained slip-deficit rate estimates along the central NAF and northern strand of the NAF in the Sea of Marmara vary nonmonotonically with Maxwell viscosity and change by up to 23% (\( \sim 4 \) mm/yr) for viscosities ranging from \( 10^{18} \) to \( 10^{23} \) Pa·s. For the best-fit viscosity structures, central NAF slip-deficit rates reach 22 mm/yr, increasing to 28 mm/yr in the Sea of Marmara. Along the central NAF, these rates are similar to the fastest geologic slip-rate estimates. The fastest slip-deficit rate estimates along the entire fault system (\( \sim 27–28 \) mm/yr) occur less than 50 km from Istanbul, along the northern strand of the NAF in the Sea of Marmara.

Electronic Supplement: Figure of sensitivity of viscoelastic block model slip-deficit rate estimates and contour plot of mean residual improvement.

Introduction

Existing models of earthquake cycle deformation are largely focused on either explaining postseismic geodetic observations from the first 1–10 yrs after large \( M_w > 6.5 \) earthquakes (e.g., Reilinger et al., 2000; Bürgmann et al., 2002; Ergintav et al., 2002, 2009; Hearn et al., 2002; Ryder et al., 2007, 2011), or explaining observations of nominally interseismic deformation occurring long after (> 10 yrs) an earthquake (e.g., Savage and Burford, 1973). To gain a more complete understanding of the earthquake cycle, postseismic and interseismic geodetic observations must be integrated and explained with a unified model that can provide a physical explanation for both rapid postseismic deformation and more slowly varying interseismic deformation (Hetland, 2006; Hearn et al., 2009; DeVries and Meade, 2013, Meade et al., 2013; Yamasaki et al., 2014; Hearn and Thatcher, 2015).

Geodetic observations across strike-slip faults from both before and after large earthquakes exist along the San Andreas fault (SAF) at Parkfield (e.g., Bakun and Lindh, 1985), across the Kunlun fault in Tibet (Bell et al., 2011; Zhang et al., 2004, 2009; Ryder et al., 2007, 2011), and along the North Anatolian fault (NAF) in Turkey (Fig. 1; Reilinger et al., 1997, 2006; Bürgmann et al., 2002; Ergintav et al., 2002, 2009).
pre- and postseismic observations represent an opportunity to constrain models of the earthquake cycle on strike-slip faults worldwide. The Global Positioning System (GPS) observations across the NAF exhibit dense spatial and temporal coverage. The easternmost and westernmost sections of the NAF were widely recognized as seismic gaps prior to the 1999 $M_w$ 7.4 İzmit earthquake (Toksöz et al., 1979; Barka, 1992, 1996; Stein et al., 1997) and as a result, survey-mode GPS campaigns were conducted in eastern Turkey and the Sea of Marmara region in the decade before the earthquake (Reilinger et al., 1997, 2006; Straub et al., 1997; McClusky et al., 2000). In addition, several permanent GPS stations were installed to continuously record deformation (Ergintav et al., 2002). These observations together revealed a localized velocity gradient with a differential velocity of $\sim 24$ mm/yr over a 200-km-wide transect across the western strand of the NAF in the Sea of Marmara region prior to the 1999 earthquake (Fig. 1; Bürgmann et al., 2002; Meade et al., 2002). Elastic block models incorporating these pre-earthquake velocities provide fault-slip-deficit rate constraints (sometimes termed geodetic slip-rate constraints) of $\sim 24$–30 mm/yr along the NAF, with rates increasing westward along the main northern strand of the NAF in the Sea of Marmara region (Reilinger et al., 1997, 2006; Straub et al., 1997; McClusky et al., 2000). In addition, several permanent GPS stations were installed to continuously record deformation (Ergintav et al., 2002). These observations together revealed a localized velocity gradient with a differential velocity of $\sim 24$ mm/yr over a 200-km-wide transect across the western strand of the NAF in the Sea of Marmara region prior to the 1999 earthquake (Fig. 1; Bürgmann et al., 2002; Meade et al., 2002). Elastic block models incorporating these pre-earthquake velocities provide fault-slip-deficit rate constraints (sometimes termed geodetic slip-rate constraints) of $\sim 24$–30 mm/yr along the NAF, with rates increasing westward along the main northern strand of the NAF in the Sea of Marmara region (Reilinger et al., 1997, 2006; McClusky et al., 2000; Meade et al., 2002).

Postseismic GPS displacements immediately following the 1999 İzmit earthquake have been effectively explained by aseismic afterslip down-dip of the coseismic rupture zone with maximum slip of about 43 cm over a 75-day period (Reilinger et al., 2000). Time-dependent analysis of GPS position time series in the 87 days after the İzmit earthquake suggests that maximum aseismic slip rates may have reached 2 m/yr (Bürgmann et al., 2002). A weakly velocity strengthening $(a-b = 0.19$ MPa) frictional afterslip model has also been used to explain the GPS displacements in the 80 days immediately after the İzmit earthquake (Hearn et al., 2002). Over a longer observation period from 1999 to 2003, three logarithmic decay constants have been used to fit the displacement time series observed at continuous GPS stations (Ergintav et al., 2002, 2009). Postseismic time series from the four years immediately after the İzmit earthquake have also been explained by a combination of frictional afterslip during the first few months after the earthquake, and visco-elastic deformation with a Maxwell rheology with viscosity $\eta_M = 2-5 \times 10^{19}$ Pa·s for the remainder of this observational interval (Hearn et al., 2009). Another study (Wang et al., 2008) incorporates a standard linear solid rheology to explain the İzmit postseismic observations.

To understand the NAF earthquake cycle to an even greater extent, both the pre- and postearthquake data have to be examined and explained simultaneously. This was first recognized after the İzmit earthquake (Hetland, 2006), and in 2D, GPS data from both before and after the 1999 $M_w$ 7.4 İzmit earthquake have been successfully explained with a two-layer Burgers model with a transient (Kelvin) relaxation timescale ($\tau_K = \eta_K/\mu_K$) of 2–5 yrs and a steady (Maxwell) timescale ($\tau_M = \eta_M/\mu_M$) of more than 400 yrs (Hetland, 2006). More recently, models incorporating shear zones (24–40 km in width) extending to midcrustal depths with low viscosities ($\sim 2 \times 10^{18}$ Pa·s) in the shear zone and higher viscosities ($> 2 \times 10^{20}$ Pa·s) in the surrounding medium have been used.
to explain GPS observations in the Sea of Marmara region from both before the İzmit earthquake and six months of post-seismic deformation after the event (Yamasaki et al., 2014). In another recent study (Hearn and Thatcher, 2015), crust-to-lithosphere-scale viscous shear zones with effective viscosity-per-unit width in the zones increasing from \((\eta_{\text{visc}}/w) \approx 10^{15} \text{ Pa}\cdot\text{s}/\text{m}\) in the crust to \((\eta_{\text{visc}}/w) \geq 5 \times 10^{16} \text{ Pa}\cdot\text{s}/\text{m}\) below the Moho were found to explain pre- and postearthquake deformation. In these preferred models, the shear zones cut through a strong \((\eta > 10^{20} \text{ Pa}\cdot\text{s})\) lithosphere and a weaker asthenosphere \((\eta < 10^{19} \text{ Pa}\cdot\text{s})\) (Hearn and Thatcher, 2015).

The previous earthquake cycle models of the NAF described above (Hetland, 2006; Yamasaki et al., 2014) have taken into account repeated earthquakes on the İzmit strand of the western NAF. Here, we develop a 3D viscoelastic block model of the greater NAF, taking into account the last 2000 yrs of earthquake history (e.g., Barks, 1992, 1996; Ambraseys and Jackson, 2000; Ambraseys, 2002) to assess in an internally consistent manner whether or not both pre- and postearthquake deformation along the NAF can be simultaneously explained with a single rheological model. We also examine the sensitivity of geodetic slip-deficit rate estimates to the assumed viscosity structure and compare these slip-deficit rates from geodetically constrained viscoelastic block models with Holocene slip-rate estimates along the NAF based on offset geomorphic features (e.g., Hubert-Ferrari et al., 2002; Kondo et al., 2004; Kozaci et al., 2007; 2009; Pucci et al., 2008; Dolan, 2009; Üçarkuş [unpublished thesis, 2010; see Data and Resources]; Meghraoui et al., 2012).

**Viscoelastic Block Model**

In classical elastic block theory (Mats’ura et al., 1986; McCaffrey, 2002; Meade and Hager, 2005; Meade and Loveless, 2009), interseismic velocities \(v_{\text{i}}\) are modeled as the sum of the velocities due to long-term block motion \(v_{\text{B}}\), the velocities due to the accumulated slip deficit \(v_{\text{SD}}\), the viscoelastic effects of the most recent earthquake \(v_{\text{VE}}\) along the fault segment, and the mean velocity throughout the earthquake cycle velocity \(v_{\text{VE}}\). For simplicity, the velocities due to homogenous internal strain of the blocks \(v_{\text{i}}\) are not included in this schematic diagram.

\[
v_{\text{i}} = v_{\text{B}} - v_{\text{SD}} + v_{\text{VE}}.
\]

To incorporate the viscoelastic earthquake cycle effects of past earthquakes into 3D kinematically consistent block models (Sato and Mats’ura, 1988; Smith and Sandwell, 2006; Johnson et al., 2007; Pollitz et al., 2008, 2010; Hilley et al., 2009; Chuang and Johnson, 2011; Hearn et al., 2013; Tong et al., 2014), a viscoelastic correction must be taken into account. The correction consists of two components: (1) the viscoelastic effects of the most recent earthquake \(v_{\text{VE}}\) along the fault segment and (2) the mean velocity throughout the earthquake cycle \(v_{\text{VE}}\). In a viscoelastic block model framework, modeled interseismic velocities are composed of five components (Fig. 2):

\[
v_{\text{i}} = v_{\text{B}} - v_{\text{SD}} + v_{\text{VE}}(\eta_{\text{M}}, \eta_{\text{K}}, t - t_{\text{eq}})
- \dot{v}_{\text{VE}}(\eta_{\text{M}}, \eta_{\text{K}}, T).
\]

The first term of the viscoelastic correction, the viscoelastic effect of the most-recent earthquake, depends on the time since the most-recent earthquake \(t - t_{\text{eq}}\) and the assumed viscosities \(\eta_{\text{M}}\) and \(\eta_{\text{K}}\). The mean earthquake cycle velocity term, the second term in the correction, depends on the assumed recurrence interval \(T\) and viscosities \(\eta_{\text{M}}\) and \(\eta_{\text{K}}\). This construction ensures that over many earthquake cycles, the displacements everywhere are equal to the long-term block displacements. In classic block theory (Mats’ura et al., 1986; McCaffrey, 2002; Meade and Hager, 2005; Meade and Loveless, 2009), this is referred to as kinematic consistency.
Fault-slip-deficit rates constrained by an elastic block model are kinematically consistent because they are linearly proportional to differential block motions.

However, in a viscoelastic block model framework, the concept of kinematic consistency is more complicated and may be divided into two distinct components, which we term type I and type II kinematic consistency. To satisfy type I kinematic consistency, fault-slip-deficit rates must be linearly proportional to differential block motions, just as in elastic block models. Type II kinematic consistency is related to the assumptions implicit in the viscoelastic correction term \( v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T) \) in equation (2). To calculate this viscoelastic correction, we assume a characteristic slip \( s \) and recurrence interval \( T \) for the earthquake and therefore implicitly assume a long-term slip rate of \( s/T \) on the fault. In order for a viscoelastic block model to satisfy type II kinematic consistency, the final slip-deficit rate estimate from viscoelastic block models must be equal to the long-term slip rate \( s/T \) that we assume in the viscoelastic corrections. Constructing a viscoelastic model that satisfies type II kinematic consistency and is as consistent as possible with the timing and estimated magnitudes of historic earthquakes would require an iterative method, in which we would assume a long-term slip rate consistent with geologic estimates of \( s \) and \( T \) and then incrementally iterate these values until we approach \( s_{\text{II}} \) and \( T_{\text{II}} \), the values of \( s \) and \( T \) for which \( s/T \) is equivalent to the final viscoelastic block model slip-deficit rate estimates. In a viscoelastic block model that satisfies the conditions of both type I and type II kinematic consistency, the assumed long-term slip rate \( s_{\text{II}} / T_{\text{II}} \) may or may not be consistent with geologic constraints on \( s \) and \( T \). The viscoelastic block models in this article automatically satisfy type I kinematic consistency (e.g., Meade and Loveless, 2009), but here we choose to be as consistent as possible with geologic estimates of \( s \) and \( T \) along the NAF (Barka, 1992, 1996, 1999; Ambroseys and Jackson, 2000; Ambroseys et al., 2002) at the cost of type II kinematic consistency.

To show the construction of the viscoelastic block model in more detail, we can rewrite the right side of equation (2) in a block model framework:

\[
v_1 = [G_B - G_{SD}]\Omega + G_\epsilon \dot{\epsilon} + v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T).
\]

The matrix–vector products give the contributions of block rotation, slip-deficit, and homogenous internal block strain to the velocity field. Assuming that the viscoelastic contributions are functions of past earthquake activity only, we can rewrite equation (3) as

\[
v_1 = v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) + \bar{v}_{\text{VE}}(\eta_M, \eta_K, T) = [G_B - G_{SD}]\Omega + G_\epsilon \dot{\epsilon}.
\]

Generalizing this framework to incorporate the viscoelastic effects of \( N_{\text{eq}} \) periodic earthquakes on different fault segments requires only a summation over all of the individual earthquakes, because the rheologies considered here are linear:

\[
v_1 = \sum_{i=1}^{N_{\text{eq}}} [v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T_i)] = [G_B - G_{SD}]\Omega + G_\epsilon \dot{\epsilon}.
\]

This can be rewritten as

\[
v_s(\eta_M, \eta_K) = [G_B - G_{SD}]\Omega + G_\epsilon \dot{\epsilon},
\]

in which

\[
v_s(\eta_M, \eta_K)
= v_1 - \sum_{i=1}^{N_{\text{eq}}} [v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T_i)].
\]

Equation (6) is a classic block model problem (e.g., Meade and Loveless, 2009) with a velocity field \( v_s(\eta_M, \eta_K) \) modified to correct for the viscoelastic effects of \( N_{\text{eq}} \) sets of periodic earthquakes. With this framework, it is straightforward to construct a viscoelastic block model incorporating the effects of many ancient and historic earthquakes: we calculate the viscoelastic effects from previous earthquakes \( \sum_{i=1}^{N_{eq}} [v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T_i)] \) and then combine these viscoelastic velocities with the observed interseismic GPS velocities to calculate \( v_s(\eta_M, \eta_K) \) as

\[
v_s(\eta_M, \eta_K)
= v_{\text{GPS}} - \sum_{i=1}^{N_{\text{eq}}} [v_{\text{VE}}(\eta_M, \eta_K, t - t_{\text{eq}}) - \bar{v}_{\text{VE}}(\eta_M, \eta_K, T_i)].
\]
To approximate finite rupture sources, we integrate over point sources using a 2D Legendre–Gaussian quadrature rule (e.g., Hildebrand, 1987). The deformation due to each point source can be simply summed with Gaussian weights to obtain the deformation due to a finite source. The number of earthquake sources \( n^2 \) required to accurately represent a particular finite source is based on a heuristic scaling \( n = cL/d \) (Klöckner et al., 2013), in which \( d \) is the distance of the closest observation point to the finite source, \( L \) is the length of the finite source, and \( c \) is a constant. We use \( c = 3 \), a value at which the mean percentage difference in displacement magnitude from an equivalent elastic finite source solution (Okada, 1992) is <0.2% immediately after an earthquake in a square area of \( 2L \times 2L \) around the fault at the surface.

Here, we use an idealized two-layer rheology structure (schizosphere and plastosphere), in which the upper elastic layer is 15 km thick, and earthquakes rupture through this entire layer. The lower layer is a Burgers viscoelastic half-space with a transient Kelvin viscosity \( \eta_K \) and a long-term Maxwell viscosity \( \eta_M \) (Fig. 3). Elastic moduli of \( \lambda = \mu = 3 \times 10^{10} \text{ Pa} \) are assumed in the elastic layer and in the Burgers viscoelastic layer. More complicated models incorporating three rheological layers and viscous shear zones have been considered in the previous studies of post-seismic deformation along the NAF (e.g., Hearn et al., 2009; Yamasaki et al., 2014; Hearn and Thatcher, 2015). Here, the use of a simple two-layer model is phenomenologically motivated, as we are interested in finding the simplest model that can explain the available data. In 2D, this two-layer model can explain the agreement between geodetic slip-deficit rates and geologic slip-rate estimates across 15 strike-slip faults worldwide (Meade et al., 2013).

To do these calculations in practice, we rotate each fault segment into an oblique Mercator projection so that its strike is parallel to the \( x \) axis and calculate the viscoelastic effects of the earthquakes along individual fault segments \( \nu_{VE}(\eta_M, \eta_K, t - t_{eq}) \) (Meade and Loveless, 2009). The mean earthquake cycle velocities \( \nu_{VE}(\eta_M, \eta_K, T_i) \) are given by the difference between the displacements \( d \) at the beginning and end of the earthquake cycle divided by the duration of the earthquake cycle \( T_i \):

\[
\nu_{VE}(\eta_M, \eta_K, T_i) = \frac{d_{VE}(\eta_M, \eta_K, T_i) - d_{VE}(\eta_M, \eta_K, 0)}{T_i}. \tag{9}
\]

The North Anatolian Fault

Geodetic Observations of Preseismic and Postseismic Deformation

In anticipation of potential future earthquakes, survey-mode GPS campaigns were conducted around the western part of the NAF in the Sea of Marmara region in the decade before the Izmit earthquake. The pre-earthquake velocity field used here consists of 122 stations (Fig. 1; Reilinger et al., 2006). For simplicity, we treat this campaign-mode velocity field observed from 1988 to 1999 as representative of the instantaneous velocity field on 1 January 1997. Additionally, a few permanent GPS stations were recording deformation at the time of the Izmit earthquake (Ergintav et al., 2009) and continued to record postseismic deformation after the earthquake. Immediately after the earthquake, more continuous stations were deployed (Ergintav et al., 2009). A 7-yr position time series has been published from GPS station TUBI. This station was moving steadily prior to the earthquake (Fig. 4a,b; from fig. 4a of Ergintav et al., 2009). In the weeks immediately after the earthquake, the station moved rapidly (>100 mm/yr) to the east, but its eastward velocity decayed to ~10 mm/yr after two years (Fig. 4).

Because a linear interseismic trend has been subtracted from the TUBI time series (Ergintav et al., 2002, 2009) (Fig. 4a), this time series records a postseismic perturbation to steady interseismic motion. In other words, if we integrate equation (2) in time, we obtain displacements

\[
d(t) = v_B t - v_{SD} t + v_i t + \int_0^t \nu_{VE}(\eta_M, \eta_K, t - t_{eq}) dt - \nu_{VE}(\eta_M, \eta_K, T_i) t. \tag{10}
\]

The linear terms \( v_B t, v_{SD} t, v_i t, \) and \( \nu_{VE}(\eta_M, \eta_K, T_i) t \) have been subtracted from the time series (Fig. 4b), so we compare the TUBI position time series in the east direction with \( d_{VE}^{\text{east}}(t) = \int_0^t \nu_{VE}^{\text{east}}(\eta_M, \eta_K, t - t_{eq}) dt \), the eastward displacements due to only the viscoelastic perturbation term. We do not take into account the (nonlinear) viscoelastic effects of the historic earthquakes in Figure 5 on the TUBI time series for simplicity and because the 1999 Izmit earthquake dominates the nonlinear viscoelastic signal. Across all viscosity structures tested, the maximum east displacement due to...
For segments of the NAF that have ruptured multiple times over the past two millennia, we use the timing of the repeated ruptures to estimate characteristic recurrence intervals. There are records of earthquakes occurring near the 1999 Izmit earthquake in 68, 268, 478, 1719, and 1894 (Ambraseys, 2002; Drab et al., 2015). Based on the more recent events, we use a recurrence interval of $T = 100$ yrs for these Izmit-like earthquakes. For fault segments in the Sea of Marmara region without a recorded pattern of repeated ruptures, we assume $T = 500$ yrs. The exceptions are the 1 B.C.E. and 121 C.E. earthquakes, for which we assume recurrence intervals of 2050 yrs, and the 1296 and 1419 events, for which we assume recurrence intervals of 1000 yrs, assuming that they may reoccur soon but have not done so yet. Along the central NAF, prior to the 1999 event, seven earthquakes of $M_w > 6.8$ have ruptured the fault in 1939, 1942, 1943, 1944, 1951, 1957, and 1967, largely from east to west. Based on earthquake history along the central NAF (Barka, 1992, 1996), we use $T = 400$ yrs for these most recent earthquakes.

The average slip magnitude estimates for earthquakes that occurred before 1939 are based on estimated surface-wave magnitudes (Ambraseys and Jackson, 2000; Ambraseys, 2002). We treat these surface-wave magnitudes as proxies for moment magnitudes and calculate average slip using a shear modulus of $\mu = 3 \times 10^{10}$ and the fault geometry in Figure 1. For the 1894 earthquake, we assume the moment magnitude of the 1999 Izmit earthquake ($M_w = 7.4$; Barka, 1999; Ambraseys, 2001). For the earthquakes from 1939 to 1967, we calculate the average slip for each earthquake using the seismic moments reported in previous viscoelastic studies of the NAF (Lorenzo-Martín et al., 2006). We assume uniform slip along the entire length of each historical rupture (Table 1).

We have also included hypothetical ancient earthquakes at 1 B.C.E. on two segments of the two southern strands of the NAF in the Sea of Marmara (Fig. 5; Table 1). The removal of these hypothetical earthquakes from the models does not substantially change the conclusions in this article. However, because these fault segments are in a tectonically active region and to our knowledge there is no evidence that they accommodate substantial creep or have ruptured in recent historical earthquakes (Ambraseys and Jackson, 2000; Ambraseys, 2002); we assume that they have ruptured within the last few thousand years and may now be late in the earthquake cycle.

Viscoelastic Modeling of the NAF Earthquake Cycle

Modeling the Postseismic Data from GPS Station TUBI

We found the viscosity structure that best explains the postseismic GPS data from TUBI, a Maxwell viscosity $\eta_M = 10^{18.6}$ Pa·s and a Kelvin viscosity $\eta_K = 10^{18.0}$ Pa·s (Fig. 4a), by a grid search over Maxwell and Kelvin viscosities in the $10^{17.0}$ to $10^{23.0}$ Pa·s range. We present residuals in terms of a mean residual improvement (MRI) percentage over an elastic model:

$$d_{\text{VE}}(t = 2007)$$ from the penultimate earthquake, in 1967, is 2.4 mm, $<3\%$ of the total displacement at TUBI.

Earthquake History along the NAF

Because the region was colonized by the Greeks in the seventh century B.C.E. and was the seat of the Roman, Byzantine, and Ottoman Empires since the fourth century C.E., the earthquake history along the western NAF over the past 2000 yrs is relatively well known from written records (Ambraseys, 2002). In the Sea of Marmara region in particular, 55 $M_s \geq 6.8$ earthquakes have been identified since 1 B.C.E. (Ambraseys, 2002). For the purposes of this article, we have included the effects of the earthquakes that likely occurred along the strands included in the block model geometry (Fig. 5; Ambraseys and Jackson, 2000; Ambraseys, 2002). The block geometry (Fig. 1) is based on a detailed fault map of Turkey (R. Reilinger, personal comm., 2013). The historic earthquakes included in the viscoelastic block models and their parameters are listed in Table 1. The effects of the 1949, 1992, 1971, and 1966 earthquakes are not included because they occurred to the east of our study area, beyond the eastern extent of the 1939 rupture (Barka, 1992, 1996; Stein et al., 1997).
\[ \text{MRI}(\eta_M, \eta_K) = -\frac{[r_{VE}(\eta_M, \eta_K) - r_E]}{r_E} \times 100 \]  

(Fig. 6a), in which \( r_{VE}(\eta_M, \eta_K) \) are the residuals for the viscoelastic models with viscoelastic parameters \( \eta_M, \eta_K \), and \( r_E \) are the residuals for an elastic model. Models that explain the data better than an elastic model will have a large and positive MRI; models that do not explain the data as well as an elastic model will have a negative MRI. In the case of the TUBI postseismic data, \( r_{VE} \) is the mean of the absolute values of the displacements (67.5 mm) because the displace-

<table>
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<th>Average Slip (m)</th>
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*Hypothetical ancient earthquakes that are not in the historic record.
and we enforce three tensile slip-rate constraints on the cen-
residual displacement of 13.31 mm), and a decrease to
fit Maxwell viscosity constant, an increase in Kelvin viscosity
for Maxwell and Kelvin viscosities ranging from
account the viscoelastic effects of historical earthquakes
(minimum seismic time series with a mean residual displacement of
MMM2 (mean residual displacement of 3.27 mm, which corresponds to an MRI of 95.16%. (b) Contour plot of MRI (equation 11) as a function of assumed viscosity structure for the interseismic GPS data from 1988 to 1999 (Reilinger et al., 2006), before the İzmit earthquake. The viscoelastic block models best explain the GPS velocity field with a Maxwell viscosity \( \eta_M = 10^{19.0} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{19.0} \) Pa·s. This best-fit model is indicated by a red circle and explains the corrected interseismic velocity field \( v_r(\eta_M, \eta_K) \) (equation 8) with a mean residual velocity of 3.14 mm/yr, which corresponds to an MRI of 2.73%.

1. In the models considered here, fault-slip-deficit rate estimates and block model residual velocities are most sensitive to variations in Maxwell viscosity (Fig. 9). Slip-deficit rate estimates along the central and eastern NAF may vary by as much as \( \sim 23\% \) (4–5 mm/yr), depending on the assumed Maxwell viscosity (Fig. 9), whereas variations in Kelvin viscosity in the \( 10^{17.0} \) to \( 10^{23.0} \) Pa·s range lead to changes of less than \( 3\%–4\% \) (\( < \sim 1 \) mm/yr) in estimated slip-deficit rates along the central NAF. The relative insensitivity of these results to Kelvin viscosity is likely due to the timing of the historic earthquakes; the most recent earthquake included in the block models occurred in 1967, and the effects of the transient Kelvin viscosity have largely abated after 30 yrs. Because of the difference in sensitivity, we present estimated fault-slip-deficit rates as functions of Maxwell viscosity, at a fixed \( \eta_K \) of \( 10^{19.0} \) Pa·s (Fig. 9).

2. Slip-deficit rate estimates do not vary monotonically with increases or decreases in assumed Maxwell viscosity (Fig. 9). Assuming a Maxwell viscosity \( \eta_M = 10^{20} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{19} \) Pa·s, slip-deficit rate estimates along the central NAF are \( \sim 22 \) mm/yr. A decrease in assumed Maxwell viscosity to \( \eta_M = 10^{19} \) Pa·s lowers slip-deficit rate estimates to \( \sim 21 \) mm/yr, and a further decrease in assumed Maxwell viscosity to \( \eta_M = 10^{17} \) Pa·s raises the estimated slip-deficit rate to 24 mm/yr (Fig. 9). The reasons why slip-deficit rate estimates are nonmonotonic functions of Maxwell viscosity can be seen from equation (4). The viscoelastic correction added to the GPS velocity field for a single set of periodic earthquakes is

\[
- \nu_{vE}(\eta_M, \eta_K, \tau - \tau_{eq}) + \tilde{v}_{vE}(\eta_M, \eta_K, T)
\]

The earth-

Figure 6. (a) Contour plot of mean residual improvement (MRI; equation 11) as a function of assumed viscosity structure for the postseismic GPS data from TUBIT. The viscosity structure that allows the model to fit the data best, a Maxwell viscosity \( \eta_M = 10^{18.6} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{18.0} \) Pa·s, is highlighted by a red circle. This model fits the time series with a mean residual displacement of 3.27 mm, which corresponds to an MRI of 95.16%. (b) Contour plot of MRI (equation 11) as a function of assumed viscosity structure for the interseismic GPS data from 1988 to 1999 (Reilinger et al., 2006), before the İzmit earthquake. The viscoelastic block models best explain the GPS velocity field with a Maxwell viscosity \( \eta_M = 10^{19.0} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{19.0} \) Pa·s. This best-fit model is indicated by a red circle and explains the corrected interseismic velocity field \( v_r(\eta_M, \eta_K) \) (equation 8) with a mean residual velocity of 3.14 mm/yr, which corresponds to an MRI of 2.73%.

ments due to the postseismic perturbation only for an elastic model are 0.

The model with the best-fit viscosity structure (\( \eta_M = 10^{18.6} \) Pa·s and \( \eta_K = 10^{18.0} \) Pa·s) explains the postseismic time series with a mean residual displacement of 3.27 mm, which corresponds to an MRI of 95.16% (Figs. 4a and 6a). Holding the best-fit Kelvin viscosity constant, an increase in Maxwell viscosity to \( \eta_M = 10^{19.0} \) Pa·s decreases the MRI to 79.71% (mean residual displacement of 13.96 mm), whereas a decrease in Maxwell viscosity to \( \eta_M = 10^{18.0} \) Pa·s decreases the MRI to 37.25% (mean residual displacement of 42.36 mm) (Fig. 6a). Overall, mean residual displacements are less sensitive to changes in Kelvin viscosity; holding the best-fit Maxwell viscosity constant, an increase in Kelvin viscosity to \( \eta_K = 10^{19.0} \) Pa·s corresponds to an MRI of 80.27% (mean residual displacement of 13.31 mm), and a decrease to \( \eta_K = 10^{17.0} \) Pa·s corresponds to an MRI of 90.92% (mean residual displacement of 4.00 mm) (Fig. 6a).

Modeling the Pre-İzmit Earthquake Interseismic Velocity Field

We consider 3D block models (Fig. 1) taking into account the viscoelastic effects of historical earthquakes for Maxwell and Kelvin viscosities ranging from \( 10^{17.0} \) to \( 10^{23.0} \) Pa·s. These block models also incorporate homogenous internal block strain (e.g., Meade and Loveless, 2009), and we enforce three tensile slip-rate constraints on the central and southern strands of the NAF in the Sea of Marmara to damp fault normal motion. We summarize the viscoelastic block model results (Figs. 7–10) in six major points.
quakes we take into account in these models (Fig. 5; Table 1) took place between 1 B.C.E. and 1967, and therefore \(30 < t - t_{eq} < 1997\) yrs. For low viscosities (\(\eta_M < 10^{19}\) Pa·s; \(\eta_K < 10^{19}\) Pa·s), after \(\sim 100\) yrs, the first term \(-v_{VE}(\eta_M, \eta_K, t - t_{eq}) \approx 0\) because all stresses are rapidly relaxed, and the mean velocity term \(v_{VE}(\eta_M, \eta_K, T)\) dominates the viscoelastic correction. As a result, the viscoelastic correction added to the GPS velocity field is in the same direction as the near-fault velocities, increasing their magnitudes and leading to slip-deficit rate estimates that are generally faster than elastic block model estimates (Fig. 9).

For high viscosities (\(\eta_M > 10^{20}\) Pa·s; \(\eta_K > 10^{20}\) Pa·s), the magnitudes of the viscoelastic corrections are small compared to the magnitude of the observed GPS velocities. The mean magnitude of the viscoelastic corrections across all stations for viscosity structures with \(\eta_M > 10^{20}\) Pa·s and \(\eta_K > 10^{20}\) Pa·s is less than 0.4 mm/yr (<3% of the mean magnitude GPS station velocity). As a result, the estimated slip-deficit rates approach elastic block model slip-deficit rates for these high viscosities (Fig. 9).

The slip-deficit rate estimates assuming midrange viscosities (\(10^{18}\) Pa·s < \(\eta_M < 10^{20}\) Pa·s, \(10^{18}\) Pa·s < \(\eta_K < 10^{20}\) Pa·s) are perhaps the most interesting (Fig. 9). These slip-deficit rate estimates depend on the relative magnitudes of the two terms in the viscoelastic corrections for each earthquake. For earthquakes that occurred more than a few centuries ago, \(-v_{VE}(\eta_M, \eta_K, t - t_{eq}) \approx 0\) and \(v_{VE}(\eta_M, \eta_K, T)\) are the larger magnitude terms. In this case, for the same reason as low-viscosity cases, the slip-

\[\text{Figure 7.} \] (a) Observed (blue) and modeled (red) interseismic GPS velocity fields for the best-fit viscosity structure highlighted in Figure 6b (Maxwell viscosity \(\eta_M = 10^{19.0}\) Pa·s and a Kelvin viscosity \(\eta_K = 10^{19.0}\) Pa·s). The observed velocity field shown here is the corrected interseismic velocity field \(v_{c}(\eta_M, \eta_K)\) (equation 8). (b) Residual velocities for the case shown in (a). Note the different scale from (a).
Figure 8. The components of the modeled interseismic velocity field (equation 2) for the best-fit viscoelastic block model (Maxwell viscosity $\eta_M = 10^{19.1} \text{ Pa} \cdot \text{s}$ and a Kelvin viscosity $\eta_K = 10^{19.6} \text{ Pa} \cdot \text{s}$), with velocity magnitude and direction shown with white arrows and the logarithm of velocity magnitude shown in color to emphasize subtle spatial variations. (a) Velocities due to long-term block motion, (b) the viscoelastic effects of the most recent earthquake along the fault segment, (c) the velocities due to the accumulated slip deficit, (d) the mean velocity throughout the earthquake cycle, and (e) the velocities due to the homogeneous intrablock strain. Interseismic velocities (f) are modeled as the sum of the velocity components in (a), (b), (c), and (e) with the mean velocity throughout the earthquake cycle velocity (d) subtracted to ensure that over many earthquake cycles, the displacements everywhere are equal to the long-term block displacements.

Figure 9. Sensitivity of viscoelastic block model slip-deficit rate estimates to variations in assumed Maxwell viscosity at a fixed Kelvin viscosity of $\eta_K = 10^{19} \text{ Pa} \cdot \text{s}$ for selected fault segments. Dotted gray lines indicate slip-deficit rate estimates from an elastic model with internal strain. In each panel, elastic slip-deficit rates do not coincide exactly with slip-deficit rates at high Maxwell viscosities because the Kelvin viscosity $\eta_K$ is fixed at $10^{19} \text{ Pa} \cdot \text{s}$.
deficit rate estimates would likely be faster than elastic block model estimates. However, the viscoelastic corrections for the more recent 1939–1967 earthquakes are dominated by $v_{VE} (\eta_M, \eta_K, t - t_{eq})$, and therefore the viscoelastic corrections added to the GPS velocity field for these earthquakes are in the opposite direction as the near-fault velocities, decreasing their magnitudes and leading to slip-deficit rate estimates that are slower than elastic block model results (Fig. 9).

3. The incorporation of homogeneous internal block strain has a substantial effect on slip-deficit rate estimates along the entire fault system. Along the central NAF, slip-deficit rate estimates from an elastic block model with no internal strain are ~27–28 mm/yr, and after incorporating internal strain, these estimates decrease by ~20% to 21–22 mm/yr (Fig. 10). Slip-deficit rate variations due to viscoelastic effects assuming the best-fit viscosity structures (Fig. 6) are comparatively small (~1–2 mm/yr; Fig. 10) along the central and eastern NAF. In contrast, the addition of internal block strain leads to an ~0.5–2.5 mm/yr increase in slip-deficit rate estimates relative to an elastic block model along the northern strand of the NAF in the Sea of Marmara (Fig. 10). Variations in slip-deficit rate due to viscoelastic effects across the best-fit models along this northern strand are of comparable magnitudes (Fig. 10).

4. In the Sea of Marmara region, the variations in estimated slip-deficit rates are complex functions of viscosity structure (Figs. 9 and 10). After incorporating the effects of previous earthquakes with the best-fit viscoelastic structures (Fig. 6b), slip-deficit rate estimates decrease by 1–2 mm/yr along the central NAF, but increase on the northern strand of the western NAF by a more substantial ~2–3 mm/yr (Fig. 9). The slip-deficit rate differential between the northern Sea of Marmara strand and the central NAF is the largest after taking into account both internal block strain and viscoelastic effects; slip-deficit rate estimates on the northern strand, less than 50 km from Istanbul, are ~20% faster than slip-deficit rate estimates along the central NAF for viscoelastic block models with the best-fit viscosity structures (Fig. 6b) but only ~15% faster for an elastic model with internal strain (Fig. 10). In addition, the senses of slip on the two southern Sea of Marmara strands reverse depending on assumed viscosity structure. In the elastic limit, the middle strand of the NAF that runs along the southern coast of the Sea of Marmara is right lateral with a slip-deficit rate of ~5 mm/yr and the southernmost strand of the NAF is left lateral with a slip-deficit rate of <1 mm/yr. For Maxwell viscosities $\eta_M < 10^{20.0}$ Pa·s, the sense of slip on the southernmost strand becomes left lateral, and for Maxwell viscosities $\eta_M < 10^{18.7}$ Pa·s, the sense of slip on the middle strand becomes right lateral.

Figure 10. Viscoelastic block model slip-deficit rate estimates along strike assuming the best-fit viscosity structures (Fig. 6): $\eta_M = 10^{19.0}$ Pa·s, $\eta_K = 10^{19.0}$ Pa·s (blue), and $\eta_M = 10^{18.6}$ Pa·s, $\eta_K = 10^{18.0}$ Pa·s (red). For comparison, slip-deficit rates from an elastic block model incorporating internal strain are plotted in gray, and those for an elastic block model with no internal strain are in green. For simplicity, only the slip-deficit rate estimates on the northernmost segments of the fault in the Sea of Marmara region are shown. White circles labeled (a) indicate Meghraoui et al. (2012), (b) Dolan et al. (2009) and Uçarkuş (unpublished thesis, 2010; see Data and Resources), (c) Pucci et al. (2008), (d) Kozaci et al. (2007), (e) Hubert-Ferrari et al. (2002), (f) Kozaci et al. (2009), and (g) Kondo et al. (2004), and corresponding error bars represent geologic slip-rate estimates and reported uncertainties.
becomes right lateral. The trade-off in slip sense between the two southern strands is largely due to the complicated fault geometry in this region and sparse geodetic observations due to the location of the Sea of Marmara (Fig. 1). The most distinct gradient in GPS velocities occurs across the northern strand of the NAF (Fig. 1); farther south, the velocity gradient is less distinct, allowing a trade-off in slip sense between the two southern NAF strands.

5. Overall, the viscosity structure that best explains the interseismic GPS data from 1988 to 1999 (Reilinger et al., 2006), before the Izmit earthquake, is a Maxwell viscosity \( \eta_M = 10^{19.0} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{19.0} \) Pa·s (Figs. 6b, 7, and 8). This model fits the corrected interseismic velocity field \( \mathbf{v}_s(\eta_M, \eta_K) \) (equation 9) with a mean residual velocity of 3.14 mm/yr, which corresponds to an MRI (equation 10) of 2.73% (Figs. 6b and 7).

6. Slip-deficit rate estimates and changes in block model residuals are most sensitive to the earthquakes that occurred since 1500 C.E. Removing the effects of the pre-1500 C.E. earthquakes does not change the major conclusions of this article. Estimated slip-deficit rate changes by less than \( \sim 2 \) mm/yr (Fig. S1, available in the electronic supplement to this article) and the best-fit viscosity structure for the interseismic GPS data changes slightly from \( \eta_M = 10^{19.0} \) Pa·s and \( \eta_K = 10^{19.0} \) Pa·s to \( \eta_M = 10^{18.8} \) Pa·s and \( \eta_K = 10^{19.0} \) Pa·s (Fig. S2).

Discussion

We have systematically modeled both pre- and post-Izmit observations with two-layer Burgers viscoelastic models with Maxwell and Kelvin viscosities ranging from \( 10^{17.0} \) to \( 10^{23.0} \) Pa·s. The viscosity structure that best explains the TUBI postseismic data is a Maxwell viscosity \( \eta_M = 10^{18.6} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{18.0} \) Pa·s (Fig. 6a), and the viscosity structure that best explains the interseismic data is a Maxwell viscosity \( \eta_M = 10^{19.0} \) Pa·s and a Kelvin viscosity \( \eta_K = 10^{19.0} \) Pa·s (Fig. 6b).

A comparison of the MRI values for the pre- and post-earthquake data as a function of viscosity structure (Fig. 6) reveals that the two best-fit models are remarkably similar, and both sets of data may be explained with a single geodetically constrained Burgers rheology of \( \eta_M = 10^{18.6} \) to \( 10^{19.0} \) Pa·s and \( \eta_K = 10^{18.0} \) to \( 10^{19.0} \) Pa·s. The variation in slip-deficit rate estimates along the central NAF for viscosities within this best-fit range is \( \sim 1 \) mm/yr (Fig. 10). An elastic block model that incorporates homogenous internal block strain, but does not incorporate viscoelasticity, yields slip-deficit rate estimates that are 1.0–1.5 mm/yr faster than the best-fit viscoelastic models along the central NAF (Fig. 9). For comparison, an elastic block model that does not incorporate both homogeneous strain and viscoelasticity leads to slip-deficit rate estimates that are 5 mm/yr faster along the central NAF than an elastic model incorporating internal strain (Fig. 9).

The viscoelastic block model results (Figs. 9 and 10) suggest that slip-deficit rate estimates along the central NAF (31° E–38° E; Fig. 1) with a geodetically constrained rheology of \( M_w = 10^{18.6} \) to \( 10^{19.0} \) Pa·s and \( \eta_K = 10^{18.0} \) to \( 10^{19.0} \) Pa·s are only 1–2 mm/yr (5%–10%) slower than those of an analogous elastic block model that incorporates internal block strain (Fig. 10). Conversely, on the northern strand of the NAF in the Sea of Marmara, slip-deficit rate estimates with the geodetically constrained rheology are significantly faster (2–3 mm/yr or 10%–15%) than slip-deficit rates from an elastic model that incorporates internal strain. Slip-deficit rate estimates along the northern strand of the NAF, after accounting for viscoelastic effects with the geodetically constrained rheology, are the fastest in the entire fault system (27–28 mm/yr; Fig. 10). In other words, the viscoelastic effects of historical earthquakes, when not accounted for, may partially mask very fast (27–28 mm/yr) slip-deficit rates along the northern strand of the NAF. The slip-deficit rate estimates along the northern strand of the NAF are especially significant, because it runs less than 50 km from Istanbul, a city with more than 14 million people.

The geodetically constrained rheology can explain both the pre-Izmit nominally interseismic (late in the earthquake cycle) GPS velocities and the postseismic (early in the earthquake cycle) displacements from GPS station TUBI (Fig. 6). The goal of explaining geodetic observations from across the earthquake cycle is to develop unified earthquake cycle models that can simultaneously provide a physical explanation for both rapid postseismic deformation and more slowly varying, nominally interseismic, deformation (Hetland, 2006; Meade et al., 2013). Previous viscosity structures estimated based on both pre- and postearthquake geodetic observations along the NAF (transient relaxation timescale \( \tau_K = \eta_K/\mu_K \) of 2–5 yrs, corresponding to a Kelvin viscosity \( \eta_K \) of \( \sim 0.8 \times 10^{18} \) to \( \sim 1.9 \times 10^{18} \) Pa·s for the shear modulus \( \mu \) used here, and a steady timescale of \( \tau_M = \eta_M/\mu_M \) of more than 400 yrs, corresponding to a Maxwell viscosity \( \eta_M \) of \( \sim 1.5 \times 10^{20} \) Pa·s; Hetland, 2006) may be slightly higher than the best-fit viscosities found here (Fig. 6) but are not directly analogous to the present study, because they are based on 2D modeling without a block model framework or internal block strain. To explain both the postseismic deformation and the pre-earthquake velocity gradient across the NAF observed before 1999, a subsequent study suggested that a combination of afterslip and a Burgers rheology with two viscosities (2–5 \( \times 10^{19} \) Pa·s and at least \( 2 \times 10^{20} \) Pa·s) might be necessary (Hearn et al., 2009).

Geodetic observations across strike-slip faults from both early and late in the earthquake cycle now exist at several locations worldwide. In Tibet, Interferometric Synthetic Aperture Radar (InSAR) and GPS measurements across the Kunlun fault in Tibet reveal localized velocity profiles with differential velocities of \( \sim 3 \) and \( \sim 12 \) mm/yr across the faults prior to the 1997 \( M_w 7.6 \) Manyi (Bell et al., 2011) and 2001 \( M_w 7.8 \) Kokoxili (Zhang et al., 2004) earthquakes, respectively. After these Tibet earthquakes, GPS and InSAR data recorded postseismic motions up to 10 times larger in magnitude than pre-earthquake velocities (Ryder et al., 2007,
In 2D, geodetic data from both early and late in the 1999 İzmit earthquake cycle in Tibet have been modeled with layered Maxwell models (DeVries and Meade, 2013), and more globally, geodetic data from across 15 strike-slip faults worldwide are consistent with a 2D, two-layer Burgers model (Meade et al., 2013). More recently, localized shear zones have been used to explain representative velocity profiles before and after large strike-slip earthquakes (Hearn and Thatcher, 2015) as well as data from before and after the İzmit earthquake (Yamasaki et al., 2014). Along the central and eastern NAF, previous geodetic slip-deficit rate estimates (e.g., Reilinger et al., 1997, 2006; McClusky et al., 2000) were ~1–10 mm/yr faster than geologic slip-rate estimates (Fig. 10; e.g., Hubert-Ferrari et al., 2002; Okumura et al., 2003; Kondo et al., 2004; Kozaci et al., 2007, 2009). This discrepancy may be partially explained if the geologic slip-rate estimates are considered to be minimum bounds (Kozaci et al., 2007; Dolan, 2009). However, here, after taking into account internal block strain and the viscoelastic effects of historic earthquakes along the NAF, the discrepancy between geodetic slip-deficit rate and geologic slip-rate estimates decreases along the central NAF; indeed, the geodetic slip-deficit rate is, within error, the same as the fastest geological slip rate of $21.5 \pm 5.5$ mm/yr (Kozaci et al., 2007) along the central NAF (Fig. 10), suggesting that there may be no discrepancy between the different rates along this part of the system.

Finally, the NAF is often considered to be a seismically active analog of the SAF system in California; the fault systems are of similar size and both delineate major transform plate boundaries. Recent viscoelastic block modeling studies focused on California have suggested that faults that are late in the earthquake cycle may have higher slip rates than previously estimated in classic block models (Johnson et al., 2007; Chuang and Johnson, 2011; Hearn et al., 2013; Tong et al., 2014). These results are perhaps analogous to the results of the present study along the western NAF, where slip-rate estimates from a viscoelastic block model are 2–5 mm/yr higher than those from an elastic model (Fig. 10).

**Conclusions**

GPS data from both before and after the 1999 İzmit earthquake may be simultaneously explained with a two-layer Burgers rheology incorporated into 3D block models with $\eta_M = 10^{18.6} - 10^{19.0}$ Pa·s and $\eta_K = 10^{18.0} - 10^{19.0}$ Pa·s in the lower layer. Viscoelastic block models of the interseismic velocity field observed prior to the 1999 earthquake fit the GPS data best with a viscosity structure of $\eta_M = 10^{19.0}$ Pa·s and $\eta_K = 10^{19.0}$ Pa·s, and the TUBI postseismic data are best explained by a two-layer viscoelastic model with $\eta_M = 10^{18.0}$ Pa·s and $\eta_K = 10^{18.0}$ Pa·s. In addition to a unified description of surface deformation prior to and after a large strike-slip earthquake, these viscoelastic block model results suggest that: (1) the fastest slip-deficit rate estimates along the entire fault system (~27–28 mm/yr) occur along the northern strand of the NAF in the Sea of Marmara, less than 50 km from Istanbul; (2) slip-deficit rate estimates do not vary monotonically with Maxwell viscosity along the central and eastern NAF; (3) the sensess of slip on the two NAF strands in the southern Sea of Marmara region reverse depending on assumed Maxwell viscosity; (4) slip-deficit rate estimates from viscoelastic block models with a geothermally constrained rheology along the central and eastern NAF are 1–2 mm/yr slower than equivalent elastic models; and (5) after taking into account internal block strain and the viscoelastic effects of historic earthquakes with the best-fit viscosities estimated here, the discrepancy between the geodetic slip-deficit rate estimates and geologic slip-rate estimates decreases along the central NAF.

**Data and Resources**

All data used in this article came from published sources listed in the references. The computations in this article were run on the Odyssey cluster supported by the Faculty of Arts and Sciences (FAS) Division of Science Research Computing Group at Harvard University. This study used the following numerical inverse Laplace transform written by K. J. Hollenbeck in 1998, INVLPAM: A MATLAB function for numerical inversion of Laplace transforms by the de Hoog algorithm (https://www.mathworks.com/matlabcentral/answers/uploaded_files/1034/invlap.m, last accessed November 2016). Finally, the detailed citation for the reference Üçarkuş (unpublished thesis, 2010) in the main text is G. Üçarkuş (2010), Active faulting and earthquake scarp along the North Anatolian fault in the Sea of Marmara, unpubl. Ph.D. Thesis, Istanbul Technical University, Istanbul, Turkey.

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