

# Surface creep on the North Anatolian Fault at Ismetpasa, Turkey, 1944-2016

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## Abstract

We re-evaluate the 72 year history of surface slip on the North Anatolian Fault at Ismetpasa since the  $M_w=7.4$  1944 Bolu/Gerede earthquake. A revised analysis of published observations suggests that days after the earthquake the fault had been offset by 3.7 m, and 6 years later by an additional 0.74 m. Creep was first recognized on the fault in 1969 as a 0.13 m offset of a wall constructed in 1957 that now (2016) has been offset by 0.52 m. A carbon-rod creep-meter operated across the fault in the past two years confirms results from an invar-wire creep-meter operated 1982-1991 that surface slip is episodic. Months of fault inactivity are interrupted by slow slip ( $\leq 10 \mu\text{m/day}$ ) or multiple creep events with cumulative amplitudes of 2-10 mm, durations of several weeks, and with slip-rates briefly exceeding  $>2.5 \text{ mm/hour}$ . Creep events accommodate 80% of the surface slip and individually release  $\approx 10^{-6}$  shear-strain on the flanks of the uppermost 3-7 km of the fault. GPS and InSAR methods yield a current fault slip rate of  $7.6 \pm 1 \text{ mm/yr}$  suggesting that creep-meters incompletely sample the full width of the surface shear zone. The slip rate has slowed from  $>10 \text{ mm/yr}$  in 1969 to 6.1 mm/year at present, 4.65 mm/yr of which appears to be due to steady interseismic creep driven by plate boundary stressing rates. We calculate that a further 1 m of aseismic surface slip will precede the next major earthquake on the fault assuming a  $\approx 260$  year mainshock recurrence interval on this segment.

## Introduction

The North Anatolian Fault in Turkey is one of several major transform faults known to slip both in earthquakes and, in some segments, aseismically in the form of surface and subsurface creep. The discovery of surface creep was made accidentally during a 1969 field trip to inspect the surface rupture of the 1944  $M_w=7.3$  Bolu/Gerede earthquake. Clarence Allen, Nicholas Ambraseys and Alkut Aytun were among the participants of this April 1969 field trip, and each published photographs subsequently documenting evidence for fault creep near the railway station of Ismetpasa (Figure 1). They reported that the eastern wall of a road maintenance compound had been offset by 24 cm (Ambraseys, 1970), (or 18 cm, Aytun, 1982) in a dextral sense by aseismic slip in the 13 years since its construction in April 1957. The aseismic slip rates determined from these approximate numbers for the interval 1957-1969 were 1.5-2 cm/yr, a faster rate than recorded by a 155-m-aperture geodetic alignment array ( $\approx 10$  mm/yr) measured periodically in the following decade by Aytun (1982) suggesting that the 1957-1969 offset may have captured rapid afterslip following the 1944 earthquake, the  $M_w=6.9$  earthquake in 1951 (Eyidogan et al., 1991), or the  $M_s 7.1$  Abant/Bolu earthquake of 25 May 1957. An  $\approx 800$  m fault-normal, 1.2-km-along-strike, trilateration/triangulation network measured between 1972 and 1980 also revealed rates of  $\approx 10$  mm/yr (Eren, 1984) a rate that was adjusted to 9.6 mm/yr by the incorporation of a further decade of data (Deniz et al., 1993). For part of this time (1982-1990) a USGS-designed creep-meter installed near the wall recorded a slower rate of 7.4-7.7 mm/yr (Altay and Sav, 1991). With the development of GPS, InSAR and LIDAR, numerous investigations have confirmed rates of 6-11 mm/yr and have mapped the spatial location and depth distribution of aseismic slip along the fault (Cakir et al., 2005; Ozener et al., 2013; Kaneko et al., 2013; Cetin et al., 2014).

The early reports of a significant slowing in creep rate over decades suggested that slip processes at Ismetpasa were initiated as afterslip following the 1944  $M_w=7.3$  Bolu/Gerede earthquake (Cakir et al., 2005). In this article we show that these earliest estimates of slip between 1957 and 1969 were inadvertently too large, suggesting pre-1969 slip rates of 15-20 mm/yr, and that this has influenced

subsequent interpretations of afterslip on the fault. Photographs of the offset-wall taken throughout the past 59 years indicate that slip rates reduced from  $\approx 10$  mm/yr in 1969 to less than 7.5 mm/yr currently.

In contrast to these inflated estimates of 1957-1969 afterslip, we deduce that estimates of coseismic slip and pre-1957 afterslip were considerably underestimated. Following the 1944 earthquake and prior to construction of the wall in 1957, two offsets of the adjacent 1934 railroad track were reported. The first estimate of 1.5 m related to repairs to the railroad track after the earthquake in 1944, and the other in 1950 of 30 cm related to the removal of a growing kink in the track caused by afterslip. To reconcile this 1.8 m of cumulative post-seismic railroad offset with the  $5 \pm 1$  m post-1944 offsets of cultural features evident nearby (Kondo et al., 2011) we conclude that the  $24^\circ$  obliquity of the railroad track was not taken into account, and that dextral fault offset amounted to 4.4 m.

Continuous slip measurements across the fault at Ismetpasa recorded in the 1980's and in the past two years reveal episodic creep - slip quiescence interrupted by brief intervals of slow steady creep at 2-5 mm/yr, or sudden multiple creep events with initial rates of slip exceeding 3.5 mm/hour. The episodic nature of surface slip means that geodetic observations undertaken at intervals of less than several years may yield higher or lower rates due to the existence of infrequent creep events on the fault with cumulative amplitudes of more than 15 mm in two months (e.g. Kutoglu et al., 2010). Creep-meters installed across the fault record lower rates than those recorded by the wall or by geodetic methods, due to their incomplete sampling of the  $\approx 12$  m wide shear zone centered on the fault.

The existence of steady creep on the North Anatolian fault since  $\approx 1980$  confirms the similarity of the central Anatolian and San Andreas faults that has been remarked upon elsewhere (Allen, 1970; 1982; Kaneko et al., 2013; Bohnhoff et al., 2015). Afterslip followed on the North Anatolian fault was also documented following the M7.4 1999 Izmit earthquake (Ergintav et al., 2009; Cakir et al., 2003 & 2012).

The article first presents a precise analysis of extant photographs at Ismetpasa to develop the record of long-term (1957-2016) slip, and concatenates it with several previous studies of fault slip in the time domain. This differs from previous studies of creep at Ismetpasa that have considered aseismic slip in the velocity domain. Data from continuous measurements of aseismic slip at Ismetpasa using Invar and carbon-fiber creep-meters are then analyzed to distinguish both steady-state and episodic components of surface creep to deduce the depth range over which each occurs. Finally we develop an afterslip history between 1944 and 2016 in an attempt to resolve the transition from afterslip, to present-day plate-boundary stressing in the context of the earthquake cycle on the Ismetpasa segment of the North Anatolian Fault.

### **The 1957 wall at Ismetpasa**

Our study of surface slip at Ismetpasa was initially motivated by a numerical discrepancy in the 1957-1969 offset of the eastern wall of the maintenance compound where creep was first reported. Ambraseys (1970) describes the wall as being initially discovered in April 1969 with an offset of 24 cm, yet during a 14 October 1969 return visit Aytun (1982) recorded an offset of only 18 cm (Aytun, 1982). Ambraseys qualified his observation with the statement “.. one can debate about the accuracy of this figure...” suggesting that there may have been some observational discussion in the field. We resolve this measurement uncertainty by examining contemporary photographs using the width of the wall to scale the 1969 offset, which when combined with its azimuth relative to the strike of the fault provides a precise measure of dextral slip.

We describe below how we quantify the offset of the wall from published photographs. The accuracy of the measurements of published photos is limited 10-15 mm by the resolution of the dot matrix in the printed pages. From Ambraseys' Figure 13, Aytun's Figure 3 and Allen's Figure 5 it is possible to average the measured offset of each side of the wall as  $114 \pm 15$  mm,  $119 \pm 10$  mm and  $122 \pm 17$  mm respectively, using the current mean wall-width as a scale. The mean  $695 \pm 21$  mm width of the wall's saddle-coping capstone is based on the average of 31 measurements in 2016.

To convert photographically determined wall offset to fault-parallel dextral slip, the obliquity of the wall to the fault is required (Figure 2), which was not specified by these early authors except in the form of sketch maps. Ambraseys' (1970, Figure 12) sketch shows the azimuth of the wall as N27°E and Aytun's (1982, Figure 2) shows it at N24°E. Ambraseys' and Allen's photos of the wall, when projected southwards intersect the approximate center of a house 100 m south of the fault. This house is visible both on the 26 August 1965 Corona 2-m-resolution imagery (Scene DS1023-2135DF107), and on Google Earth 2013 imagery, indicating that the azimuth of the wall was then, and is now, N15±2°E. Although Ambraseys' and Aytun's azimuths may thus be discarded, their implied azimuths may have influenced the reported estimates of dextral slip calculated from their measured offsets of the wall. Alternatively, the published values for offset may have been influenced from considerations of the overthrust south-west wall whose offset was obscured by crushing and which, following early repairs, has now been replaced by a drain.

To estimate the local strike of the fault we measured the fault azimuth 2 km eastward from Ismetpasa to an offset cowshed wall at Hamamli (N83.5°E), and 2.2 km westward (N78°W) to an offset curb at a petrol station, and adopted a mean strike of the fault at Ismetpasa of N81°E with an uncertainty of about 2 degrees due to the local change in strike of the fault. The obliquity of the fault to the wall (66°) requires the observed offsets of the wall to be multiplied by  $1/\sin(66) = 1.09 \pm 0.02$  (Figure 2). Hence, the above derived offsets from analysis of the wall yields dextral slip estimates of 126, 130 and 134 mm respectively (Table 1), significantly less than the 240 mm dextral slip on April 1969 reported by Ambraseys, but close to Aytun's 180 mm estimate made in October 1969. These discrepancies are too large to be accounted for by the apparent errors in wall obliquity attributable to these early reports. Ambraseys would have needed an obliquity of 26° but his sketch indicates 56° (close to the complementary angle). Aytun would have needed an obliquity of 39°, but his sketch indicates 58°. An error in our assumed width of the concrete capstone to the wall (caused by possible repairs since 1969) would scale both values linearly, however, cracks in the 1969 photographs of the concrete saddle-coping that caps the masonry wall are evident to this day, and uniquely-identifiable,

field-stone blocks within the wall remain in their original positions. We conclude that the present wall has not been reconstructed in the past five decades.

Because the accurate measurement of the offset in 1969 is central to constraining early creep rates at Ismetpasa we explain our analysis in detail. Two of the original May 1969 35 mm negatives provided by Clarence Allen, were scanned at 4000 dpi (6  $\mu\text{m}$  pixel sampling of the 35 mm image) and loaded into a standard graphic analysis software (*Eazydraw*) order to numerically quantify the deviation of each edge of the wall from a straight line at  $\approx 1$  m intervals. The following procedure was adopted to obtain offsets with uniform sample spacing along the wall from oblique perspective photographs (Figure 4). Straight lines were first aligned by eye with each edge of the wall converging at a vanishing-point on the eye-line horizon. Points of equal wall width were next constructed horizontally across the digitized photograph to act as a template for lines converging at an arbitrary, but spatially separated, second vanishing-point, also on the eye-line horizon. The intersection of these converging lines with the constructed line following the east side of the wall identified the locus of equal distances along the wall for measurements of wall-width and offset. The distances from the sub-pixel east- and west-edges of the wall to the guide lines ( $x_e$  and  $x_b$ ) were measured at each of these points and an average offset derived in arbitrary units on the screen. This mean offset was converted to a dimensionless ratio by dividing each average by the local wall-width measured on the photographic image ( $x_{\text{wall}}$ ). To convert these photographically derived ratios to millimeters of true local offset, we multiplied each derived ratio by the mean wall-width measured directly in 2016 (695 mm). The correct positioning of the artificial eye-horizon and the two wall-parallel reference lines is somewhat subjective, but does not influence the accuracy because linear trends in the numerical series are removed prior to estimating fault offset, as explained below.

To determine the offset of the wall where it crosses the fault we fit straight lines in a least squares sense to the north and south halves of the numerical series derived above. A  $< 2$  mm one-standard deviation misfit was typically obtained for each of these alignments. The lateral separation of these two lines provides an objective offset, but since the two lines are not precisely parallel, the offset

depends on the point chosen to evaluate it. The chosen point is the midpoint of the shear zone of the fault,  $\approx 12$  m from the southern end of the wall. The two original 35 mm photographs yielded dextral offsets of  $129 \pm 6$  mm (viewed north) and  $119 \pm 6$  mm (viewed south) a difference that we attribute to local roughness in the edge of the capstone. The average derived from scanned images of the three published 1969 photographs was  $130 \pm 9$  mm (Table 1).

For the 2016 data, and for all inferred offsets 1970-2016, the east edge of the capstone south of the fault is a misleading indication of offset because it has been pulled clockwise relative to body of the wall by the superstructure of the steel chain link fence across the fault (Figure 3). To correct for capstone rotation in 2016 we subtracted the measured eastward overlap, and a short westward underlap (see Figure 3e), of the east edge of the capstone from the underlying wall at 1 m intervals along the entire wall (Electronic supplement Table S1), and subtracted this from the photographically derived offsets of the eastern edge of the capstone. A least-squares fit to the resulting series yields a measure of the straightness and azimuth of the wall both south and north of the fault, independent of capstone misalignment. The formal uncertainty in the offset of the resulting north and south least squares fits was  $\pm 2.5$  mm, but we recognize that the accuracy of the conversion from capstone-offset to wall-offset depends on the value of the original 1957 overlap south of the fault for which we have no direct measure. To estimate this we determined the mean and standard deviation of the eastward overlap of the capstone on the northern half of the wall ( $90.8 \pm 8.6$  mm) and derived a combined offset and uncertainty of  $90.8 \pm 11$  mm. This value represents the correction needed to compare the eastern edge of the northern *capstone*, with the eastern edge of the underlying southern *wall* to overcome errors in cap rotation. LIDAR scans of the wall taken in 2007 and 2009 (Karabacak et al., 2011) indicate that the average roughness of the fieldstone wall exceeds the uncertainty attributed to the top edge of the wall alone (Figure 4c). Several photographs of the 2016 offset were analyzed yielding slightly different values depending on whether points closer than 3 m to the fault were included in the analysis (Table 1).



The wall was surmounted by a brush-thicket fence in 1970, and subsequently by pillars and a chain-link fence, making sequential photographs showing both east and west edges of the wall and the crest of its center-line, similar to those shown in Figure 3a-d, impossible. Photos in 1970, 1980 and 1984 show only the eastern edge of the wall, and in photographs this is sometimes obscured by weeds and bushes. The photographs (see electronic supplement) show the wire chain-link fence progressively tensioned by fault slip. The increasing tension eventually resulted in the detachment of the southern capstone from the underlying wall, and clockwise rotation of the capstone. Because all photos of the wall were taken from above, they must be corrected for this rotation. In 2016 precise measurements of this rotation indicated that the east edge of the capstone and the east edge of the wall coincide at a point midway between the southern end of the wall but this crossover point is expected to vary with time, because the axis of rotation is near the center of the wall. Where the full width of the wall was not visible we used the crest of the saddle-coping as a linear feature (width/2) to calibrate offset of the eastern edge of the wall, or calibrated blemishes in the wall from other photos to scale the offset. A reported offset of 40 cm by Şaroğlu & Barka (1995) is not accompanied by a photograph, an uncertainty, or an indication of what obliquity correction was applied, and consequently to this we assign a large uncertainty ( $\pm 50$  mm). Photographically derived offsets of the wall are listed in Table 1 and are plotted in Figure 5.

LIDAR measurements were obtained in 2007-2009 (Karabacak et al., 2011), and also in 2015. These data were decimated by retaining all but the first arriving offsets consisting of a 40-mm-wide band describing wall roughness. The residual data were detrended using a least squares fit to data on the northern half of the wall and the offset across the fault measured between points  $\pm 3.5$  m of the fault. The offsets so determined were assigned uncertainties of 11 mm for 2007 and 35 mm for the 2015 measurement which more comprehensively characterizes the roughness of the northern half of the wall.

The weighted mean of all five 1969 estimates of wall offset is  $125.4 \pm 4.3$  mm yielding a mean 1957-1969 rate of  $10.45 \pm 0.35$  mm/yr. The most reliable measured offset of the wall (avoiding capstone

rotation) since then is  $516 \pm 15$  mm derived in 2016, a mean 1957-2016 rate of  $8.75 \pm 0.09$  mm/yr. The formal 1957-2016 rate uncertainty is underestimated in that it does not account for roughness of the underlying wall imaged by LIDAR.

The mean 1969-2016 rate is  $8.8 \pm 0.3$  mm/yr, a 25% slowing in rate compared to pre-1969 rates. The 1957 to 1969 rate lies within the uncertainties of the  $10.7 \pm 0.7$  mm/yr 1969-1979 rate observed by Aytun (1982), and the 9.6 mm/yr rate 1972-1990 reported by Deniz et al., (1993). A linear weighted least squares fit to all the photographically determined offsets, constrained to pass through the origin (wall construction in 1957) indicates a mean creep rate of  $9.2 \pm 0.2$  mm/yr (Figure 5b).

No perturbations in slip-rate are evident in the data following several nearby earthquakes on the fault, or the 1999 earthquakes (Kutoglu et al., 2008 and 2010), however, wall offsets are constrained between 1984 and 2007 by a single point in 1995 devoid of a photographic record or stated uncertainty (Saroğlu & Barka, 1995). We discuss next the occurrence of large creep events recorded by creep-meters at Ismetpasa whose unrecorded occurrence would result in apparent accelerations in rate, that will have escaped notice in the sparse photographic record.

### **Continuous Measurements of Creep using Invar and Carbon-rod Creepmeters**

In 1982, a 20-m-long Invar wire creep-meter was installed at Ismetpasa (Altay and Sav, 1991). In their Figure 3 they show the azimuth of the wall as  $N12^\circ E$ , the fault as  $N76^\circ E$  and their creep-meter as  $N48^\circ E$ , implying an obliquity of  $28^\circ$  close to the  $30^\circ$  claimed in their text. If we assume that they used uncorrected magnetic bearings to determine azimuths, their corrected wall azimuth ( $15^\circ E$ ) and their fault strike would be  $N79^\circ E$  consistent with that adopted in this article. This would imply their creep-meter azimuth was  $N51^\circ E$ , which when combined with our adopted fault strike ( $N81^\circ E$ ) is consistent with their stated obliquity of  $30^\circ$  (Figure 6).

The Invar wire of the Altay and Sav instrument was freely suspended in a catenary within a buried pipe crossing the fault at a depth of 1.5 m. The wire was anchored at its western end, and held in tension at its eastern end by a mass-balance rotating on a hinge (Burford and Harsh, 1980). The rotation of the mass was recorded by an LVDT (Linear Variable Differential Transformer) whose voltage output was registered by a strip-chart recorder with a chart-recording speed of 15 cm/day, a sample-rate of 12 s, and full-scale recording span of approximately 3.3 mm. Mechanical adjustments were required to bring the record back on scale when the 3.3 mm recording range was approached or exceeded, but the absolute datum was simultaneously indicated by means of a scale attached to the rotating mass. Thus although continuous data were lost on five occasions, the cumulative creep record was preserved by a sequence of manual readings before and after gaps in recording. These manual offsets were not published.

Data from Altay and Sav's Invar wire creep-meter are reproduced from their hand-drafted cumulative plot in Figure 7a. A digital listing of these data will be found in the supplementary materials. The mean dextral slip rate averaged over 8 years is stated to be 7.4-7.7 mm/yr, but a least squares fit to a digitized version of their plotted data yields a rate of  $7.34 \pm 0.10$  mm/yr, possibly resulting from the different temporal sampling adopted in our digitization of their data. The data consist of multiple creep events with amplitudes of up to 2.5 mm separated by intervals of slow creep or zero slip. Infrequent slow sinistral displacements were recorded, presumably due to soil contraction in summer months.

Chart records of four creep events (Figure 7b). that together form part of the largest observed sequence in their data, are reproduced in their article with amplitudes of 0.66 mm to 2.54 mm, with decay time constants of 1-2 hours, and with durations of less than 30 hours. Throughout their data Altay and Sav recognize a bimodal distribution of creep events, distinguishing between those with amplitudes of  $< 0.7$  mm, and others with amplitudes of  $\geq 1.5$  mm. They note periods of steady slow slip on the fault. Some of their events end abruptly, while others continue as slow slip. Operation of the Invar-wire creep-meter was discontinued as a result of new construction inside the compound.

In May 2014 we installed a carbon-fiber rod creep-meter outside the compound to the east of the wall (Fig. 6, Fig. 8). A continuous 4-mm-diameter, 16.5-m-long, carbon-fiber rod slides within a 12-mm-internal-diameter, PVC-conduit joined with O-ring expansion slip joints buried 33° obliquely across the fault below a depth of 35 cm. The western end of the rod is anchored rigidly to a stainless steel tripod driven to 2 m depth with one vertical rod and two rods inclined at 45° riveted at the top. A second identical buried tripod arrangement installed 16.6 m to the east on the far side of the fault, supports a displacement sensing unit. The free eastern end of the rod is held in tension by a flexible wire fastened to a 4.5 N constant-tension spring motor via a 90° elbow that feeds the wire to the surface (Figure 8). Two transducers measure the extension caused by fault motion: a subsurface 12 mm range LVDT, and a Hall-effect transducer installed in an electronics housing 50 cm above the surface. The Hall effect rotary transducer is attached to a 100-mm-circumference-wheel around which the flexible wire is wrapped. The wire is held in tension by a spring motor with a range of approximately 1.5 m. The arrangement is designed so that should the LVDT reach the end of its range, a record of slip of up to 1.5 m is retained by the Hall transducer. Displacement data are recorded along with temperature and battery health information by a 12-bit data logger providing least-count resolutions of 3  $\mu\text{m}$  and 25  $\mu\text{m}$  for the LVDT and Hall effect transducers, respectively. The data are simultaneously converted to dextral slip in millimeters and transmitted via the Iridium satellite with a maximum latency of 30 minutes. A correction factor for obliquity and sensor calibration (1.271) has been applied to the transmitted data (see supplementary materials for these digital data) to convert these displacement data to dextral slip (Figure 9) but no other editing was required. In April 2016 a second data logger was added with a 5 s sampling rate to record the details of creep events. This second logger has a memory capacity of 3 weeks and must be downloaded manually before the creep-event data are overwritten.

Despite burial at only 35 cm depth, the graphite rod and LVDT sensor are not significantly influenced by annual surface temperature changes of up to 60°C (Figure 9). A periodic daily signal of 0.6  $\mu\text{m}/^\circ\text{C}$  is evident in the LVDT signal, but this signal is more pronounced ( $\approx 100 \mu\text{m}/^\circ\text{C}$ ) in the

above-ground Hall effect sensor, which is influenced by the mounting arrangement of the solar panel and Iridium transmitter. In August of each year we recorded a 28  $\mu\text{m}$  cumulative slow left-lateral displacement that we attribute to thermoelastic processes similar to the  $\approx 0.1$  mm left-lateral seasonal signal noted by Altay and Sav (1991).

The continuous record from April 2014-June 16 creep (Figure 9) exhibits the same features as exhibited in the 1982-1990 creep data: periods of no slip are interrupted by slow slip at rates of 2-6 mm/yr, and/or by abrupt creep events, each consisting of a double creep event. In 2015 double creep events were separated by 120-180 hours but the larger events evident in Altay and Sav's creep data appear to be separated by several weeks. Amplitudes of individual recent creep events are 1.3-2.5 mm, similar to those recorded in the 1980's.

The creep events in the past 21 months at Ismetpasa have contributed  $\approx 7.2$  mm of the total of total dextral slip, with the remaining 2.6 mm occurring in the form of slow slip. On the Superstition Hills fault, California, the existence of slow uneventful slip punctuated by distinct creep events was attributed to a variation in rheology with depth (Bilham and Behr, 1992). Steady slip was considered to be confined to shallow layers within 100 m of the surface, with episodic events occurring between the surface and depths of  $\approx 3$  km (Du et al., 2003; Wei et al., 2013). The recent uneventful creep at Ismetpasa constitutes 27% of the total slip signal. For the 1982-1990 creep data of Altay and Sav (1991), we estimate at least 19% of the total creep comes from slow slip.

In Figure 10 we illustrate the variation of surface slip rate with depth of creep anticipated by an elastic model with imposed regional shear of  $2.2 \times 10^{-7}$ /year ( $22 \pm 3$  mm/yr at  $\pm 50$  km, McClusky et al., 2000;  $24 \pm 2$  mm/yr, Reilinger et al., 2006), and a shear rate of 22 mm/yr below 10 km depth (Cetin et al., 2014). We used a 3D boundary element elastic half-space model (Gomberg and Ellis, 1994) in which shear tractions are imposed  $\pm 50$  km from and parallel to a locked vertical fault between a shallow depth  $d$ , and the base of the seismogenic zone  $D$ . We varied the depth of the freely slipping frictionless fault above depth,  $d$ , to quantify the resulting surface slip. Bohnhoff et al., (2016)

estimate a mean depth,  $D$ , to the base of the seismogenic zone for the Anatolian fault of 12-14 km, Assuming a seismogenic depth of 12 km at Ismetpasa, the  $\approx 8.8$  mm/yr slip rate recorded by the offset wall across the fault in recent decades (Figure 5) corresponds to creep from the surface to  $5.5 \pm 0.1$  km (Figure 10c), similar to that derived from 1992-1999 InSAR studies based on a calculated rate of  $8 \pm 2$  mm/yr at Ismetpasa (Cakir et al., 2005; Cetin et al., 2014). The creep-meter data with its current rate of 5.9 mm/yr suggests that this shallow component of creep now extends to 3.6 km depth, and that the 25% contribution from slow slip (Figure 9) is confined to depths shallower than 1.8 km (c.f. Figure 8, Bilham & Behr, 1992).

The surface creep rate indicated by creep-meter data acquired 2014-2016 (5.9 mm/yr from Figure 9) is lower than that indicated by the interpolated rate from the offset wall (6.1 mm/yr), and 22% slower than that observed from nearby GPS observations 2005-2011 (Ozener et al., 2012), but samples an insufficiently long history of slip to be certain of the significance of its mean rate. However, the much longer 8-year-long-rate measured by the creep-meters in the 1980's (7.3-7.7 mm/yr), was then 16% lower than the mean rate recorded geodetically (Kutoglu and Akcin, 2006). We attribute these lower rates to the narrower fault-normal sampling aperture of the creep-meter (10 m) compared to the deformed width of the wall ( $\approx 20$  m). Cracks indicating the rotation of wall segments suggest that to completely sample the deforming zone, the length of  $30^\circ$ -oblique creep-meters would need to be 30 m (Figure 6), which presents difficulties with the current installation, because it would require us to extend the instrument beneath a road.

The lower rate sampled by the creep-meters has important implications for the interpretations of fault zone rheology and strain release during creep events. The mean rate measured by the creep-meter 1982-1991 ( $7.6 \pm 0.1$  mm/yr, Altay and Sav, 1991) was 82% of the mean slip geodetic slip rate (9.3 mm/yr, Deniz et al., 1993) in approximately the same time interval, although re-evaluated rates for the creep-meter ( $7.3 \pm 0.1$  mm/yr, Figure 7a) and the geodetic data (8.7 mm/yr, Kutoglu and Akcin, 2006) permit a range of 78-87%. Distributed shear tens of meters from a primary fault trace is common on creeping faults, and is occasionally related to slip on contiguous traces (Bilham & Whitehead, 1997;

Galehouse and Lienkaemper, 2003; Cakir et al., 2012; Lindsey et al., 2014). Hence the amplitude of the continuous creep-meter data should be multiplied by 1.2-1.4 to correct for this incomplete sampling of the shear zone. The rates from recent creep data must be increased by a factor of  $\approx 1.25$  for it to agree with the recent geodesy, however, the geodetic estimates average rates obtained from somewhat different along-strike segments of the fault. This amplification factor applies both to long term creep rates and to the amplitude of recorded creep events measured by the creep-meter. To prevent confusion, and because the scaling factor is currently an approximation, this scaling factor is omitted in the discussion of creep event amplitudes and velocities in the following section except where specified.

## **Creep events**

Due to the currently uncertain amplitude-scaling factor of the creep events caused by the creepmeters' incomplete span of the Ismetpasa shear zone, we leave to a future study the determination of the rate- and state-dependent properties of the fault zone, and here examine the evolution of surface slip during the creep events of 1987 and 2015. Although the measured amplitudes of the creep events differ in the two creep-meter installations due to their different lengths, the evolution of slip with time in each creep event should be unaffected. In what follows we analyze the observed data without correction for these unknown amplitude errors. Thus estimates for displacement, strain and velocity mentioned in this following section underestimate their true values by 1.2-1.3.

The six creep events available for analyses resemble each other in general appearance and presumably represent a small patch, or several patches, of the fault slipping aseismically with neither rigorous time-predictable nor slip-predictable behavior. In the 8 years of instrumental data reproduced by Altay and Sav (1991), slip events on the fault occur within clearly defined limits; slip on the fault apparently never lags more than 6 mm behind the mean creep rate, and never overshoots this mean rate by more than 4 mm. Our current data, which conform to a narrower  $\pm 2$  mm band (Figure 9), thus

may be atypical of the longer term creep behavior of the fault. To illustrate this apparent difference we have superimposed our recent creep data on a part of the 1982-1990 creep data it most resembles (Figure 7a). The similarity between the 1988/9 and 2014/5 data emphasizes that sequences of cumulative creep events, with large amplitudes, are separated by longer periods of quiescence than we have recently recorded.

The 10 mm range of displacements evident between the slip limits shown in Figure 7a can be used to deduce approximate maximum strains in the uppermost 5 km of the fault zone. They correspond to maximum shear strains of  $2 \times 10^{-6}$  if strain accumulation and release is confined to one side of the fault only, or as is more likely, antisymmetric shear strain of  $10^{-6}$  on each flank of the fault ( $\approx 5$  mm in 5 km). In contrast, our current data lag or lead the past two years' mean creep rate by only 2 mm (Figure 9), which implies an antisymmetric strain storage and release budget of  $2 \times 10^{-7}$  on each flank of the fault. The largest discrete creep events (2.5 mm) on the fault are associated with strain release of  $2.5 \times 10^{-7}$  strain.

The maximum slip velocity we recorded in 2015 was 0.54 mm/hr, but the maximum recorded by Altay and Sav in May 1987 was 3.54 mm/hr (Table 2). This large difference is not due to the different spatial sampling widths of the two creep-meters but to the difference in the temporal sampling rate. Our telemetered 30 minute sampling interval underestimates maximum fault slip velocities, because these velocities have short duration and are manifest in the first several minutes of each creep event only. The 1980 strip chart data sampling once every 12 seconds captured these fast initial slip rates. The duration of each creep event exceeds several hundred hours but most of the slip occurs in the first 5 hours.

We first compare the creep events recorded in 1987 with those recorded in 2015 using a double exponential decay curve of the form:

$$\text{slip}(t) = A_0 + A_1 e^{-t/\tau_1} + A_2 e^{-t/\tau_2} \text{ mm} \quad (1)$$



where  $t$  is the time after the start of the event. The parameters in this expression describe the amplitude and decay of the creep event, but have no physical significance. They can be fit to the data quite precisely (misfits are  $<\pm 50\mu\text{m}$ ), permitting a simple comparison of creep event characteristics separated by three decades (Table 2). The 1987 data were first corrected for their non-linear chart scale using a *Rustrak* pen radius of 62 mm (F. Wyatt, personal communication). To avoid biasing the comparisons by using data of different duration we truncated all the data to the same 0.5-30 hour window before solving for the parameters in (1). Each creep event can be characterized by a rapidly decaying signal with decay rate  $1 < \tau_1 < 3.3$  hours, followed by a longer-period decaying process with decay rate  $6 < \tau_2 < 37$  hours.

Although  $\tau_2$  averages  $19\pm 11$  hours for creep events considered in 1987 and 2015, the initial time constant,  $\tau_1$ , for the 1987 data is approximately half that for the 2015 data. We note that the 1987 creep events constitute the largest cumulative sequence of events observed by Altay and Sav (1991) in 8 years of recording and therefore may have been anomalous. The sequence started 8 May, followed by minor events on 24 and 28 May with amplitudes of 1.8 mm, 0.66 mm and 0.68 mm respectively (written by hand on the charts digitized in Figure 7b) bringing the total slip in May (excluding slow slip between these events) to 3.05 mm. For June they summarize total monthly slip as 8.4 mm in their text, only 2.54 mm of which is accounted for in the 1 June creep event. This 1 June event was associated with the most rapid initial slip, the shortest time constant, and the largest amplitude, causing the record to go off-scale. An additional 5.86 mm of creep (2-3 missing creep events) thus occurred in June, which these authors either recorded as micrometer offsets, or chose not to reproduce in their article.

We next fit the creep-event data to an expression for a static (non-propagating) creep event proposed by Wesson (1988).

$$u(t) = u_{\max} \left\{ 1 - [ Ct^{(b-1)} u_{\max}^{(b-1)} + \Delta t + 1 ]^{-1/(b-1)} \right\} \quad (2)$$

Where  $u(t)$  is the slip on the fault at time  $t$  (for  $b > 1$ ) and we introduce  $\Delta t$  to adjust for the data sampling interval to account for the unknown initiation time of the creep event relative to the internal clock in the data logger. The solution includes the final amplitude  $u_{\max} = \sigma d_e / \alpha \mu$  (mm) of the creep event, and a rheological parameter  $C = [(\alpha \mu / d_e)^b / k]$   $\text{mm}^{(1-b)}/\text{s}$ , where  $\sigma$  is the shear stress applied to the fault,  $d_e$  is the depth of slip,  $\alpha$  is determined by the aspect ratio of the fault slip event ( $\alpha = 0.61$  if the total length of the slipping zone is  $d_e$ ;  $\alpha = 0.5$  if the slipping zone is infinitely long) and  $\mu \approx 30$  GPa. For a viscous fault  $b = 1$  and  $k = \eta/W$ , the ratio of the viscosity of the fault zone to its width.

The slip histories of the six events can be approximately fit to equation 2 (misfits are shown in Figure 11c) yielding an average value for the exponent  $b = 2.7$  ( $1.3 < b < 4.5$ ) confirming that simple viscosity is inappropriate. This average value is twice that derived by Wesson (1988) for the northern segment of the creeping zone of the San Andreas Fault in central California. The solutions for  $C$ ,  $u_{\max}$  and  $b$  are listed in Table 3. The shear stress at failure,  $\sigma = \alpha \mu u_{\max} / d_e$ , is determined by the geometry of the slipping zone and by the strain released by the slipping patch. If we adopt the shallowest inferred depth of the creeping zone as  $d_e = 5$  km we derive an average value of  $\sigma = 0.009 \pm 0.002$  MPa, consistent with the low strain release ( $10^{-6}$ ) inferred from depth and slip amplitude above. From the mean value of  $C$  and  $b$  we derive a mean value for  $k$  of  $3.5 \times 10^{-12}$   $\text{kg m}^{-2} \text{s}^{-1}$ .

An improved fit to the data ( $\pm 20$   $\mu\text{m}$ ) can be obtained from equation 3 which assumes that the two depths for quasi steady slip ( $u_{\text{slow}} \leq 1$  km) and event-like slip ( $u_{\text{event}} \leq 5$  km) are associated with equations of similar form but different slip parameters:

$$u(t) = u \left\{ 1 - [C t^{(b_0-1)} u^{(b_0-1)} + 1]^{-1/(b_0-1)} \right\} + u/5 \left\{ 1 - [C_1 t^{(b_1-1)} (u/5)^{(b_1-1)} + \Delta t + 1]^{-1/(b_1-1)} \right\} \quad (3)$$

In practice, the final amplitude,  $u = u_{\max}$  in equation 3 for the 2015 creep events can be fixed by inspection (e.g., from Figure 9) leaving five unknowns (Table 4). The misfits resemble those shown in

Figure 11b, i.e. they do not show the systematic misfit characterized by Figure 11c. As anticipated, the rheological parameter  $b_1$  for the second term in equation 3 is smaller in each case than the first term  $b_0$  that describes the onset of each creep event ( $b_1/b_0 \approx 0.6$ ). For the 1987 events the value of the  $b_1$  parameter is close to that describing purely viscous slip.

## Afterslip

In the preceding sections we presented evidence for aseismic offset of the fault since 1957 in the form of repeated measurements of alignment arrays, steady offset of the Ismetpasa stone wall, and in the form of two multiyear segments of continuous creep-monitoring. In this section we extend the fault slip record to the time of the 1944 earthquake (05:22 local time, 1 February 1944), first by extrapolating the apparent slowing in rate 1957-2016 to the time before construction of the Ismetpasa wall, and then by invoking additional estimates of slip reported as coseismic slip in 1944, and as inferred afterslip in 1950.

Seven years prior to wall construction, in 1950, the nearby 1934 railroad track was reported to have been offset by 30 cm (Ambraseys, 1970) a value that lacks both an estimate of uncertainty and, as we discuss later, an indication of whether or not a correction for fault obliquity has been applied. We later discuss the implications of correcting this and the coseismic slip value stated by Ambraseys(1970) to dextral slip, but we first adopt this hitherto accepted afterslip value of 30 cm to demonstrate fits to possible time dependent decay rates. Using an exponential expression for the decay in afterslip displacement with time

$$\text{afterslip} = k + Ae^{-t/\tau} \quad (4)$$

where  $k$  is a linear interseismic velocity, and  $A$  is the amplitude of transient afterslip decaying at a rate of  $t/\tau$  at time  $t$  following the mainshock, the best fit in a least squares sense solving for incremental

displacements in the periods 1944-1950, 1957-1969 and 1969-2016 yields  $A=40$  mm,  $k=9.8$  mm and  $\tau =10$  years. If we assume this correctly characterizes the entire interseismic cycle, cumulative estimated interseismic surface slip will amount to 2.4 m assuming an earthquake recurrence interval of  $\approx 200$  years (Figure 5c).

We also fit the data to AFTER (equation 5), an afterslip decay curve that has been invoked to describe slip on California strike-slip faults that slip both seismically and aseismically (Boatwright et al., 1989, Lienkaemper et al, 2016).

$$\text{afterslip} = u_f \left( \frac{t/T}{t/T + 1} \right)^c \quad (5)$$

The empirical expression invoked by AFTER includes a final slip term,  $u_f$  that represents the sum of coseismic and total afterslip displacement, that we initially set to 4.4 m (Kondo et al., 2005), a characteristic exponent,  $c$ , that determines the decay rate, and a variable  $1/T$ , where  $T$  is the effective duration of the afterslip process. In solutions that fit the observed wall offset data we determined  $0.8 < c < 0.9$  and  $100 < T < 550$  days. However, all solutions required pre-1950 slip to have been less than 115 mm. If we ignore this apparent misfit, we would conclude that for a 6 m seismic cycle (releasing  $\approx 276$  years of interseismic plate motion), afterslip in the interseismic interval amounts to  $\approx 1.4$  m followed (and preceded) by coseismic slip of 5.6 m (Figure 3c). In Figure 12 we illustrate graphically that the pre-1957 reports of offset at Ismetpasa may have been misinterpreted, which would render the above values for interseismic slip extrapolated from the 1957-1969 data considerably in error.

The railroad lines at Ismetpasa are stated to have been offset coseismically by 1.5 m in the 1944 earthquake (Ambraseys, 1970). We here question whether this stated number represents dextral slip, or their lateral offset from a former straight line (Figure 12). Ambraseys' statement "*displaced the Zonguldak railroad line by about 150 cm*" has been construed to quantify dextral slip (Kondo et al., 2005), and Ambraseys certainly implied this in the context of his article. However, the 1.5 m of observed offset is significantly lower than the 3.6 m-6 m offsets Kondo et al., (2005) measure in the

form of offset rows of trees, irrigation canals, dirt roads and field boundaries 1 km to the west, and 2.5 km to the east of the railroad (Figure 13b). If we suppose that the reported railroad offset of  $\approx 1.5$  m were not of dextral slip (Figure 12a), but a report of misalignment of the railroad lines following their repair (Figure 12b), the dextral offset must be increased by  $1/\sin 24^\circ$  to  $\approx 3.7$  m.

Likewise the interpretation of the second adjustment to the rails in 1950, following the initial 1.5 m re-alignment in 1944 requires re-evaluation. Here Ambraseys (1970) states that in 1950 "*the railroad tracks were again distorted at the same point showing a displacement of 30 cm*". A buckle in the rails had developed that required trains to slow as they approached Ismetpasa station from the west. To correct a quite minor rail-normal misalignment of  $\approx 12$  cm the railroad engineers would have needed to shorten the rails by removing a 32-cm-length of steel track (Figure 12c). Alternatively, if there were truly a misalignment of the line of 30 cm (Figure 12d) it would have required the dextral fault afterslip accumulated since 1944 to have been 74 cm.

Thus it appears to us very probable that the 1944 earthquake resulted in 3.7 m of dextral slip when the railroad tracks were repaired and an additional 0.74 m of afterslip by 1950. We admit the possibility that the numbers recorded by Ambraseys (1970) may have been rounded to decimeters suggesting no great accuracy should be ascribed to them. If we assign  $\pm 20$  cm to the 1.5 m 1944 estimate and  $\pm 10$  cm to the 30 cm misalignment in 1950, the combined fault slip and its uncertainty 1944-1950 becomes  $4.4 \pm 0.7$  m (Table 5).

Treating the precise time of the initial 3.7 m of inferred dextral offset as a variable and using the constraints in rows 1, 2, 4 & 5 of Table 5 in equation (5) we solve for the time of the initial repair to the railroad. The railroad was constructed a decade before the 1944 earthquake, however, we ignore 1934-1944 late interseismic slip on the fault which would reduce coseismic slip by approximately 6 cm were it similar to current rates, an amount that is embraced by the  $\pm 20$  cm uncertainty estimated for the coseismic slip. We calculate that the reported "1.5 m railroad offset" corresponds to the second day after the earthquake, a plausible time for engineering repairs to the railroad.

The values in Table 5 potentially permit afterslip to be distinguished from the presence of steady-creep on the fault. The accuracy with which this can be achieved depends on several assumptions, including the precise timing of the initial railroad repair, that constrains the 1944-1950 afterslip interval. We consider it probable that the 3.7 m deduced to have distorted the rail line at Ismetpasa, would not have been manifest within seconds of rupture as co-seismic slip, but instead was manifest entirely as afterslip reaching the surface starting minutes or hours after the mainshock. This appears to us plausible because surface slip has been delayed in several Anatolian earthquakes by many hours. For example, Ambraseys (1970) recounts that in the 22 July 1967 Mudurnu earthquake at Acemler, paths were offset and an irrigation ditch flooded the village many hours after rupture. Examples of delayed surface slip are common on creeping faults in California. No surface occurred coseismically for five hours on the **M**6.8 1987 Superstition Hills earthquake (Bilham, 1989), nor on the southern segment of the **M**6.0, 2004 Parkfield rupture for three days (Bilham, 2004; Langbein, 2005), nor in the Napa earthquake (Lienkaemper et al., 2014) in California. Thus we suppose that in 1944 the "coseismic" slip and afterslip constitute different stages of an evolving afterslip process that may have been largely completed by 1950, and which now contributes <25% to the observed aseismic slip rate at Ismetpasa.

The rapid decay in afterslip prior to 1957 suggests that the currently observed creep on the Ismetpasa segment is now driven largely by a plate-boundary stressing process similar to that occurring in central California. Accordingly we fit the data numerically to the sum of the right hand side of equation (2) and a linear slip term  $vt$ , where  $t$  is the time after the earthquake and  $v$  is a steady interseismic velocity constrained by the parameters listed in Table 5 to be 4.65 mm/yr. Residual afterslip plus this background velocity yields a current (2016) slip rate of 6.1 mm/y (Figure 14a) and suggests that an additional slip of  $\approx 1$  m will accumulate before the next earthquake supposing that the next Bolu-Gerede earthquake recurs in  $\approx 2220$  CE. Ambraseys (1970) reports no rail realignments in the period 1950-1969 during which time we estimate 24 cm of dextral slip, corresponding to 9.8 cm of rail misalignment (Figure 14a). The 1951 earthquake (Eyidogan et al., 1991) does not appear to have contributed significantly to surface slip despite mention of surface cracks (Ambraseys, 1970) and

minor buckling of the rails (Ergin et al., 1967; Aytun, 1982). Slip during the 1950-1957 interval (before wall construction) is calculated to have been 114 mm corresponding to a mean slip velocity of 16 mm/yr, slowing to 10.4 mm/yr between 1957 and 1969, and 8.7 mm/yr between 1969 and 2016, consistent with rates calculated from the offset wall. Adopting the lower or higher range of uncertainties for the railroad offsets increases the misfit significantly and shifts the estimated time of the first railroad measurement, but does not change the interseismic background velocity. No solution was possible for a calculation using the raw-railroad offsets (1.5 m in 1944 and 0.3 m in 1950) as input.

## Discussion

The 150-km-length of the creeping zone on the North Anatolian fault (Figure 13a from Cetin et al., 2014) is comparable to that of the creeping central San Andreas fault, however, earthquakes that have ruptured the ends of the creeping zone in central California have been relatively modest (e.g. Parkfield  $M \sim 6.0$ ), with major earthquakes ( $M_w \geq 7.6$ ) in 1906 and 1857 abutting, rather than overlapping, those segments currently exhibiting surface creep. The combination of creep and seismic behavior of the Ismetpasa segment more closely resembles that on the Hayward fault, where apparently steady creep rates of 5-9 mm/yr are interrupted by  $M_w \leq 7$  earthquakes at  $\approx 160$  year intervals (Lienkaemper et al., 2012). At least since 40 years after the 1868 earthquake mean creep rates at Fremont on the Hayward fault have been steady, as documented by a foundation wall that has now been offset in Fremont by  $\approx 1$  m. This time interval represents 80% of the earthquake cycle but it is probable that rapid initial (but undocumented) afterslip occurred prior to wall construction in Fremont (Lienkaemper et al., 2016). The duration of observed surface creep data at Ismetpasa starting 13 years after the 1944 earthquake, is currently only 26% of the inferred  $\approx 276$ -yr earthquake cycle for major earthquakes here (1944-1668), and early afterslip is evident in the numerical data. However, in contrast to previous studies at Ismetpasa, the decadal-averaged creep rate at Ismetpasa has slowed by only  $\approx 12\%$  since creep was first discovered in 1969. In the following sections we discuss the implications of the

transition from coseismic slip to what appears now to be linear but episodic surface creep, on the interpretation of 1994 earthquake magnitude, and on the role of creep in the interseismic cycle.

### *Current creep rates near Ismetpasa*

The velocities plotted in Figure 14b show a bimodal distribution. Those with a fault-normal aperture <20 m (wall offsets, LIDAR and creep-meter data) must be multiplied by approximately  $\approx 1.25$  to bring them into agreement with triangulation, GPS and InSAR data with their considerably larger observational footprints. The correction factor is common on creeping faults where surface shear is sometimes distributed tens of meters from a discrete offset that may accommodate 75-90% of the shear displacement manifest across the surface fault (Galehouse and Lienkaemper, 2003). We allude to this correction factor in our discussion of the results from creep-meters, where we argue that incomplete sampling of the surface shear zone by creep-meters requires observed creep event parameters to be increased in amplitude by a factor of 1.2-1.4. Some of the creep rate estimates include velocities from segments of the fault to the east where creep rates are known to be higher (e.g. Cetin et al., 2014). A further discrepancy between near-fault slip measurements and geodetic data acquired far from the fault is that, in addition to the near fault strain-field caused by slip on the surface fault down to a locking depth of 5-7 km, these distant measurements increasingly sample a regional strain-field with an arc-tan signature caused by creep below the seismogenic zone.

### *Accelerations in slip rate?*

Apart from the episodic rate changes associated with surface creep events we find no evidence to support changes of rate on the fault related to distant seismicity. However, some authors have claimed significant changes in rate based on the apparent outliers evident in Figure 14b. These are labeled K8 (Kutoglu et al., 2008) and K10 (Kutoglu et al., 2010), that we now briefly discuss. The two time intervals with high velocities recorded are 5 and 6 years duration with velocities of  $12.0 \pm 1.3$  and  $15.1 \pm 4.1$  respectively, preceded by a contiguous 10 year interval with  $6.7 \pm 1.1$  mm/yr velocity



(K6), suggesting an acceleration in slip rate in the interval 1992-2008 (Table 5). This anomalous acceleration has been discussed subsequently by Kutoglu et al., (2012) and by Gormus et al., (2013). The weighted mean velocity for these three epochs would be  $8.9 \pm 5$  mm/yr, similar to InSAR measurements in the early part of this window (1992-2002;  $8 \pm 3$  mm/yr; Cakir et al., 2005) and GPS measurements near the end of this time window (2005-2011;  $7.6 \pm 1$  mm/yr; Ozener et al., 2012) which do not register this increase in rate. We note that short segments of geodetic data yield erratic velocities due to the influence of creep-events on the fault with slip amplitudes of as large as 16-20 mm, however, such events would influence 5 year sampling interval velocity estimates by only 2 mm/year. We therefore cannot account for reported velocities exceeding 10 mm/yr 1999-2008. The cumulative offset of the wall indicates that no unusual increment in slip occurred between 1999 and 2016, which therefore refutes any long-term contribution to cumulative slip from K6, K8 or K10.

Cetin et al., (2010, Figure 8) document the effects on the fault of the 6 June 2000,  $M_w=6.0$  Orta earthquake, a normal faulting event that occurred  $\approx 35$  km SE of Ismetpasa (Taymaz et al., 2007; Cakir and Akoglu, 2008) and was accompanied by numerous aftershocks. It is possible that this earthquake may have locally shifted the control points analyzed by Kutoglu et al., (2008, 2010 & 2013).

### ***Magnitude of the 1 February 1944 Bolu/Gerede earthquake?***

Kondo et al., (2005) estimate that the 1944 earthquake as  $M_w=7.6$ , based on its 175 km long rupture and their mean observed surface slip of 3.4 m. They correctly note that their observed surface slip includes afterslip, but we argue above that *all* of the surface slip in the Ismetpasa segment may be a combination of afterslip combined with steady creep. This differs from the model presented by Kaneko et al. (2013), who assume 50% coseismic surface slip, 20% afterslip and 30% interseismic creep. The distinction is important because if no coseismic surface slip occurred shallower than say, 5 km, the moment magnitude would need to be adjusted downward by 0.2 magnitude units to  $M_w=7.4$ . If velocity strengthening inhibited coseismic slip in the uppermost 7 km of the fault, the resulting magnitude would be closer to the magnitude  $M_s=7.3$  determined by Dewey (1976). However, our

$M_w=7.4$  magnitude estimate may err on the low side because creep currently decays westward (Cakir et al., 2005), admitting the possibility that as much as 2/3 of the western rupture may have been entirely coseismic from the base of the seismogenic zone to the surface. The absence of shallow present-day creep presently to the west, however, does not mean that rate and state friction moderated afterslip did not occur in the near surface rupture of segments west of Ismetpasa following rupture.

### *Interseismic creep and the earthquake cycle*

In Figure 14a we illustrate the inferred development of 6.1 m of surface slip during the entire earthquake cycle ( $276\pm 40$  years) based on measured slip in the past 72 years. By definition, this 6.1 m equals the seismogenic slip potential of the seismogenic zone at the end of the earthquake cycle. If we assume that a substantial fraction of the afterslip in the uppermost 5-7 km of the fault was largely complete by 1969, 25 years after the mainshock, the implication is that at the current stage in the interseismic period (72 years after the mainshock) and for the remainder of the interseismic interval, surface creep now responds to plate boundary stressing processes. This observation is significant because it implies that the base of the surface creeping zone is no longer responding to stress from the last major earthquake. Instead, the upper surface of the seismogenic zone that will eventually rupture in a future earthquake is now being incrementally stressed by creep events. The monitoring of creep events thus has utility in monitoring processes that may occur as a prelude to an approaching future earthquake rupture. Based on a 276-yr historical recurrence time, an additional 1 m of surface creep is anticipated before the next major earthquake here (Figure 14a).

Of the  $5.9\pm 0.1$  mm/yr slip rate presently recorded by the Ismetpasa creep-meter we calculate from equation [5] that 1.25 mm/yr (21%) may be attributable to residual afterslip, and 4.65 mm/yr (79%) from linear creep driven by present-day loading. With the base of the seismogenic zone at 12 km the inferred base of this linear creep would lie at a depth of  $\approx 2.9$  km, whereas the total 5.9 mm/yr creep signal would be inferred to extend to a depth of  $\approx 3.6$  km. The formulation thus implies that the slowing in surface creep is caused by a geometric shallowing in the depth of creep events on the fault.

In contrast, the velocity strengthening formulation investigated by Kaneko et al., (2013) predicts that the depth of creep should increase during the interseismic interval. In Figure 14b we compare our corrected afterslip rate curve with two models published by Kaneko et al., 2013 for locking depths of 3.2 km and 4 km using friction properties (a-b) of 0.01 and 0.026 respectively. Their curves steepen in the early part of the afterslip process to fit Ambraseys' (1970) erroneous x2 slip rate but otherwise provide a reasonable fit to our data. Their model, however, differs substantially in the first year of the earthquake during which they envisage coseismic surface rupture followed by afterslip, whereas we infer zero surface slip during the mainshock and a surface slip signature that develops entirely post-seismically. Clearly, it would be desirable to examine physically-based afterslip models using the new early constraints supplemented by additional data from creep-events.

Assuming a current creep rate of 6.1 mm to 7.6 mm year (Figure 10c) a depth-averaged shear strain is accumulating at 1.5-1.7  $\mu$ strain/year on the uppermost 3.6 km to 5 km of the fault, corresponding to 0.75-0.85  $\mu$ strain/yr. An implication is that in the 220-276 years between major earthquakes on this segment, the antisymmetric shear strain in the uppermost 5 km of the fault prior to rupture will have incremented by 165-230  $\mu$ strain. If we exclude inferred afterslip and use instead the inferred linear creep rate of 4.65 mm/yr above 2.9 km depth, the depth averaged strain accumulation rate is 0.8 antisymmetric  $\mu$ strain/year. The implied strain at failure is twice that anticipated on flanks of the Hayward fault ( $83 \pm 34$   $\mu$ strain) that slips for most of its length at 5 mm/yr down to a creeping depth of 5 km between major earthquakes (Lienkaemper et al., 2012).

The 1982-1991 creep-meter data suggest that variations in surface creep rate occur with a multi-year cycle. The cause of these long term variations may be that somewhat deeper patches of the fault occasionally slip, or that pulses of slip propagate along the fault. For example, in 1987 and 1990 the surface creep accelerates following an interval of lower rates formed from slow slip and relatively modest creep events (Figure 7). Slip in 1982 may have followed such an event, and our 2014-2016 creepmeter data followed a nearby  $\approx 20$  mm slip event on the fault (Rousset et al., 2016). These 3-5 year cycles perturb the mean creep rates derived from infrequent geodetic observations. The time

history of each creep event suggests that at least two decay constants are involved describing the slip process: initial slip has a characteristic decay time of 1-3 hours, subsequent slip has a time constant of 6-24 hours. Individual creep events release shear strain of less than  $10^{-6}$  on the flanks of the fault, but the shear strain released by the larger creep sequences in 1987 and 1990 (up to 16 mm in two months if we include a 1.4 scaling factor) may involve larger strain release. This depends on whether the creep events extend to greater depth during these sequences, or whether they involve a propagating pulse of slip limited to uniform depth. A laterally, along-strike propagating 2-mm, surface-displacement limited to 3.6 km depth would produce  $\approx 0.6 \times 10^{-6}$  of flank shear strain. A larger surface creep event scaled to greater depth would involve similar strain release but would release strain over a larger volume.

Creep rates in central California are occasionally retarded or accelerated by nearby earthquakes (Burford and Harsh, 1980; Schultz, 1982; Titus et al., 2005; 2006). We see no evidence for perturbations in the creep rate at Ismetpasa resulting from earthquakes on nearby or distant fault segments, although we concede that changes in rate may have occurred that would have escaped detection because of the absence of a continuous creep data for 42 years of the 59-year-long record. The largest slip episode recorded in May/June 1987 (12 mm observed, 16.8 mm corrected for sampling aperture) is associated with no known local earthquakes, although a 48 km distant  $M_w=4.5$  earthquake occurred five months prior to its onset. A yet larger creep event ( $\approx 20$  mm) is inferred to have been recorded by InSAR data in 2013, one year before the carbon-rod creep-meter started recording (Rousset et al., 2016). The 2013-2016 data may thus resemble the creep sequence recorded by Altay and Sav (1991) in 1987-1989 when a large multiple creep event was followed by minor episodes of creep in the following two years (Figure 7a).

## Conclusions

Numerous previous studies of creep on the North Anatolian Fault have accepted Ambraseys' (1970) estimate that the wall constructed at Ismetpasa in 1957 had been offset by 24 cm when he visited it with others in 1969. However, the photograph provided in his 1970 article, and those taken the same day by Aytun and Allen and reproduced in their 1982 articles, indicate that the actual offset ( $\approx 13$  cm) was considerably smaller, and that its rate of offset since 1957 had been approximately 10 mm/yr, half the rate deduced from Ambraseys' original value, but similar to slip rates 1969-1979. These and additional photographs and LIDAR measurements, provide an absolute measure of the wall's progressive offset indicating a slowing in its rate from more than 10 mm/yr in 1969 to its current (2016) creepmeter offset rate of  $6.1 \pm 0.2$  mm/yr. Recent triangulation, trilateration, InSAR and GPS rates that average displacements over a wider region near the fault indicate a current creep rate of  $7.6 \pm 1$  mm/yr approximately 1.25 times faster, indicating that the creep-meters and the wall sample creep from a wider shear zone than accessible to their 5-12 m fault-normal apertures.

In contrast to the stable long-term decadal creep rate, measurements with creep-meters show that inter-annual surface slip is episodic, and consists of periods of no slip (47% of the time in the past two years), interrupted by months of slow-slip (44% of the time in the past two years) at rates of  $\approx 3$  mm/yr, or by abrupt slip events with transient velocities exceeding 3 mm/hour, slip durations of many days, and, in the case of multiple events, with cumulative amplitudes of many millimeters. To correct for the incomplete spatial sampling of the surface shear-zone by the creep-meters, their recorded displacements and velocities must be multiplied by a factor of 1.2-1.4. If we include this correction factor, the largest sequence of creep events (recorded in May/June 1978) cumulatively accounts for 16.8 mm of slip on the fault, similar in amplitude to a creep episode reported to have occurred to the east of Ismetpasa in 2013 (Rousset et al., 2016).

Because creep events occur at several-month intervals on the fault, all measurements of surface fault slip suffer from temporal aliasing errors in estimating long-term creep rates. Typically Ismetpasa creep events occur in pairs with cumulative amplitudes of  $\approx 4$  mm ( $\approx 5$  mm when corrected for incomplete fault zone sampling), hence if they occurred regularly in time would occur approximately yearly. In practice they occur erratically with no slip in some years followed by multiple sequences of events every 3-5 years. Hence all methods require at least a decade of data if they are to yield estimates of average slip rate accurate to 10%. We see no perturbation to the long term slowing in creep rate despite several nearby earthquakes. Recent earthquakes may have triggered minor slip when the creep-meters were not operating, but this has resulted in no obvious perturbation to the apparently slow decrease in decadal surface creep rates.

The observed offset rate of the Ismetpasa wall when combined with the remotely observed shear strain rate indicates that creep initially extended to 5-7 km depth, less than half way to the base of the seismogenic zone, but that it now extends to 3.6 km, less than one third to the depth of the base of the seismogenic zone. Episodes of slow uneventful slip on the fault account for approximately 25% of the total slip but occur for 44% of the time, and we conclude that this slow slip process is confined to depths shallower than 1.8 km. From the inferred superposition of shallow slow slip during abrupt creep events at deeper depth we surmise that the rheological control of these slower processes is closer to that expected from viscous materials. Observed creep events are associated with strain accumulation and release of the order of  $10^{-6}$ , and with onset velocities  $>3$  mm/hour and with decay time constants of the order of 1-3 hours. The largest discrete creep events, when corrected for limited fault-zone sampling exceed 3.4 mm, and are often followed a few days or weeks later by a similar sized creep event. Although these double events are likely to signify propagating creep along the fault, with data from one isolated creep-meter it is not possible to distinguish between spontaneous slip of a discrete patch, or longitudinal creep propagation.

If Ambraseys' (1970) reported offsets to the railroad in 1944 and 1950 were statements of rail misalignment, their correction for fault obliquity requires 4.4 m of slip on the fault in the 6 years

following the earthquake. With the addition of subsequent surface creep reported in this article, the resulting  $\approx 5.0$  m of surface slip accords with measured 1944 offsets on contiguous segments of the fault by Kondo et al., (2005). We conjecture, based on an empirical afterslip decay rate that 1944 surface slip occurred entirely as afterslip starting hours after the mainshock, and that residual afterslip at Ismetpasa (currently 1-2 mm/yr) is imposed on a linear background creep of  $\approx 4.65$  mm/yr. A further  $\approx 1$  m of surface slip is anticipated before the next major earthquake on the fault assuming an interseismic cycle similar to the 1668-1944 cycle (Bohnhoff et al., 2016) releasing a  $\approx 6$  m of plate boundary slip deficit. The inferred absence of surface slip in the  $M_s=7.3$  1944 mainshock has the effect of reducing the geometric moment magnitude derived from its mean slip and surface rupture length from  $M_w=7.6$  to perhaps as low as  $M_w=7.4$ . Of importance to seismic hazard studies is the observation that creep processes are currently responding to loading caused by plate motion, and that episodic surface creep events incrementally increase shear strain near the upper surface of the seismogenic zone that will eventually fail in a future major earthquake.

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AGU Supplementary Materials

1. High resolution scans of archival photographs

May 1969	Published photographs by Allen, Ambraseys and Aytun
May 1969	High resolution photographs Clarence Allen
October 1970	High resolution photograph Roger Bilham
April 1980	High resolution photograph Wayne Thatcher
October 1984	High resolution photograph James Jackson

Table S1 Wall offset geometry 2016

2. Ismetpasa carbon fiber creep-meter data. Text file. 15 min. samples March 2014-Sept 2016.

3. Altay and Sav (1982) Text file. Invar creep-meter data digitized from published record

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Table 1 Estimated and photographically-quantified wall offset and dextral slip

Source	Wall Azimuth	Fault Strike	Obliquity	pub. mm	photo mm	Dextral mm	epoch
Ambraseys (1970, Fig.13)	N27E	N83E	(56°)	240	114±15	126±17	1969.4
Aytun (1982, Fig. 3) 1969	N24E	N82E	(58°)	180	119 ±10	130±12	1969.8
Allen (1984, Fig.5)	N15E	N82E	66°	-	122±17	134±19	1969.4
Allen 1969 north photo*	-	-	66°	-	117±6	128.6±7	1969.4
Allen 1969 south photo*	-	-	66	-	109±6	119.4±7	1969.4
<b>weighted 1969 offset*</b>						<b>125.4±4.3</b>	<b>1969.4</b>
Bilham 1970 north photo	-	-	-	-	147±25	161±30	1970.7
Aytun(1982, Fig.4) 1978	-	-	-	-	186±13	204±14	1978.7
Thatcher 1980 photo	-	-	-	-	214±20	234±22	1980.4
Jackson 1984 photo 1	-	-	-	-	252±15	276±20	1984.4
Jackson 1984 photo 2	-	-	-	-	247±30	270±32	1984.8
Şaroğlu & Barka (1995)	-	-	-	400	-	400±50	1995.5
Lidar 2007	-	-	-	-	416±10	455±12	2007.7
21 May 2014 2.8 mm lens	N15±0.5°E	N81±2E	66±2.5°		465±25	509±27	2014.4
Lidar 2015	-	-	-	-	500±30	547±35	2015.6
25 April 2016 20 cm lens**	-	-	-		482±10	528±11	2016.3
25 April 2016 50 mm lens**	-	-	-	-	458±10	504±11	2016.3
<b>weighted 1957-2016 offset</b>						<b>516±15</b>	<b>2016.3</b>

\* From least squares fits to east and west edges of the capstone. \*\* least squares fits to east edge of wall. The LIDAR 2007 data are from Karabacak et al (2011). Figure numbers in parenthesis in column 1 refer to those in cited publications. The penultimate column indicates dextral slip from wall offset assuming 66° wall/fault obliquity.

Table 2. Creep event amplitudes and velocities, with parameters fit to a double exponential (1) for the interval 0.5 to 30 hours following the start of each creep event. Slip in column 2 is estimated from data shown in Figures 7 & 9. The values for displacement and velocity shown in Table 2-4 should be multiplied by 1.2-1.4 to account for incomplete sampling of the fault zone.

Event	slip mm	$V_{\max}$ mm/h	A	$-A_1$ mm	$\pm$	$\tau_1$ hours	$\pm$	$-A_2$ mm	$\pm$	$\tau_2$ hours	$\pm$	$\chi^2$
8-May-87	1.8	3.54	1.77	0.62	.07	1.8	0.3	0.47	.05	14.6	5.6	.003
1-Jun-87	2.5	1.57	2.31	0.96	.04	1.1	0.1	0.77	.02	11.4	1.3	.009
17-Apr-15	2.4	0.54	2.29	1.59	.06	2.4	0.1	0.8	0.1	24.3	9.2	.03
24-Apr-15	1.9	0.40	1.52	0.76	.18	2.1	0.3	0.5	0.2	6.3	1.7	.006
10-Dec-15	1.3	0.36	1.39	0.81	.01	2.9	.07	0.53	.01	37.5	.94	.03
15-Dec-15	1.3	0.18	0.94	0.56	.02	3.3	.01	0.34	.05	15.5	.01	.007

Table 3. Creep event parameters derived for equation 2 with their uncertainties. The curves systematically misfit the rise time and early slip in the data (see Figure 11c) resulting in large uncertainties in some of the C solutions.  $\Delta t$  (s) is an offset caused by the data logger sampling interval.

Event	u, mm	$\pm u$	C	$\pm C$	b	$\pm b$	$-\Delta t$	$\pm \Delta t$	$\chi^2$
8-May-87	1.91	.07	<b>0.29</b>	.05	<b>2.7</b>	.4	.24	.07	.005
1-Jun-87	2.80	0.75	<b>0.064</b>	.02	<b>3.8</b>	0.25	.68	.15	.004
17-Apr-15	2.14	0.1	<b>0.19</b>	0.01	<b>1.7</b>	0.1	.16	.03	.060
24-Apr-15	1.52	.01	<b>0.33</b>	.01	<b>1.3</b>	0.1	0.0	.05	.006
10-Dec-15	1.3	0.05	<b>0.23</b>	0.02	<b>2.2</b>	0.5	.03	.02	.008
15-Dec-15	1.52	0.17	<b>0.052</b>	0.68	<b>4.5</b>	0.01	.45	.01	.019



Table 4. Parameter fits to equation 3 assuming that slow slip processes shallower than 1 km are active during a deep creep event extending to >3km depth.

Event	u, mm	C <sub>0</sub>	±C <sub>0</sub>	C <sub>1</sub>	±C <sub>1</sub>	b <sub>0</sub>	±b <sub>0</sub>	b <sub>1</sub>	b <sub>1</sub> ±	-Δt	Δt±
8-May-87	1.6	<b>0.33</b>	0.48	<b>.01</b>	.23	<b>1.6</b>	0.7	<b>1.01</b>	0.02	.01	.01
1-Jun-87	2.54	<b>0.29</b>	.01	<b>.003</b>	.001	<b>2.7</b>	.05	<b>1.02</b>	0.9	.001	1.4
17-Apr-15	2.3	<b>0.23</b>	.01	<b>.019</b>	.001	<b>1.59</b>	.04	<b>1.3</b>	.2	.33	.01
24-Apr-15	2.1	<b>.069</b>	.006	<b>.006</b>	.002	<b>3.67</b>	.07	<b>2.1</b>	.3	.09	.06
10-Dec-15	1.3	<b>.258</b>	.009	<b>.156</b>	.005	<b>2.54</b>	.05	<b>1.36</b>	.01	.01	.01
15-Dec-15	1.3	<b>.248</b>	.009	<b>.016</b>	.005	<b>2.54</b>	.05	<b>1.36</b>	.18	.01	.01

Table 5 Coseismic slip, afterslip, wall offset and estimated uncertainties at Ismetpasa. Column 5 assumes published values refer to rail offsets and are corrected by  $1/\sin 24^\circ$  to convert to dextral slip. \*stated by Ambraseys (1970); \*\* reported by Kondo et al., (2005). The slip potential estimate assumes a plate velocity of  $22\pm 1$  mm/yr. We calculate fault slip in the interval 1950-1957 as 114 mm using the stated 1944, 1950, 1969 and 2016 offsets (see text). The interseismic residual is the difference between the slip potential (276 yrs at  $22\pm 1$  mm/yr) and the current surface slip (2016) assuming a similar interseismic renewal time. Bold text (row 6) indicates calculated surface slip to 2016. Rows 7 and 8 (italics) indicate inferred interseismic slip and current residual. Rows in italics are estimated or calculated rather than observed.

row	epoch	offset		dextral		
		published m	process	slip mm	uncertainty mm	time interval ,years
1	1-Feb-44	1.5*	coseismic	3700	490	0 to 30 days
2	1944-1950	0.3*	afterslip	740	25	20 days to 6 years
3	<i>1950-1957</i>	<i>none*</i>	<i>data gap</i>	<i>≈114</i>	<i>≈15</i>	<i>6 to 13 years</i>
4	1957-1969	wall photos	wall offset	125.4	10	13-25 years
5	1969-2016	wall photos	wall offset	516	15	13 to 72 years
6	<b>1944-2016</b>	<b>3.6-5±1**</b>	<b>cumulative slip</b>	<b>4990</b>	<b>555</b>	zero to 72 years
7	<i>1668-1944</i>	<i>6.072</i>	<i>slip potential</i>	-	276	<i>276 years</i>
8	<i>2016-2220</i>	-	<i>interseismic residual</i>	<i>1081</i>	380	<i>204 years</i>

Table 6 Ismetpasa slip velocities from wall offsets, creep-meters, LIDAR, triangulation, trilateration, GPS and InSAR. Numbers in parenthesis in the last two columns refer to a revised analysis of Erin (1984) and Deniz et al., (1993) by Kutoglu and Akcin, 2006 (K6). Those marked with an asterisk are derived for stated time periods.

source	published mm/yr	method	interval	duration years	aperture m	adopted mm/yr	1 $\sigma$ mm/yr
Ambraseys, 1970	20	photo	1957-1969	12	47 m	10.4	0.4
Aytun, 1982	15	photo	1957-1969	12	47 m	10.45	0.35
Aytun 1982	10	triangulation	1969-1979	10	160 m	10.8	0.4
Eren,1984	$\approx$ 10	trilateration	1072-1982	10	800 m	10.2 (10.6)	0.6 (1.1)
Deniz et al., 1993	9.6	trilateration	1982-1992	10	800 m	9.3 (8.7)	0.7 (1.3)
K6-Kutoglu & Akcin, 2006	7.8	GPS	1992-2002	10	800 m	7.8 (6.7)	0.5 (1.1)
Altay and Sav, 1991	7.6 $\pm$ 0.1	creep-meter	1982-1991	9	15 m	7.3	0.1
K8-Kutoglu et al., 2008	12 $\pm$ 1.3	GPS	2002-2007	5	800 m	12.0	1.3
K10=Kutoglu et al., 2010		GPS	2002-2008	6	800 m	15.1	4.1
Cakir et al., 2005	8 $\pm$ 3	inSAR	1992-2002	10	>50km	8.0	3.0
Cetin et al., 2014	8.35 $\pm$ 0.24	InSAR	2003-2011	8	>50 km	8.35	0.24
Karabacak et al., 2011	8.4 $\pm$ 1.6	LIDAR	2007-2009	2	20 m	8.4	1.6
Ozener et al., 2012	7.6 $\pm$ 1	GPS	2005-2011	6	800 m	7.6	1.0
Kaneko et al., 2013	$\approx$ 9	InSAR	2007-2011	4	>50km	6	0.5
1969-2016 wall offset*	from Fig 5b	photo	1969-2016	47	47 m	8.3	0.1
1984-2016 wall offset*	from Fig 5b	photo	1984-2016	33	47 m	7.1	0.3
1969-1984 wall offset*	from Fig 5b	photo	1969-1984	15	47 m	9.9	0.3
2014-2016 continuous*	from Fig 9	creep-meter	2014-2016	2	8 m	5.9	0.1
2014-2016 afterslip fit*	from Fig 14a	offset wall	2014-2016	2	47m	6.1	1.0



Figure 1. Google Earth view of the Isetpasa maintenance yard showing the North Anatolian Fault and locations of invar and carbon-fiber creep-meter (40.8698°N, 32.6250°E) and nearby geodetic alignment array points 13-15 (Aytun, 1984), and the eastern wall of the maintenance yard, cross the fault at 66°. The railroad crosses the fault at 24° in compression and has required occasional re-alignment and shortening to accommodate creep.

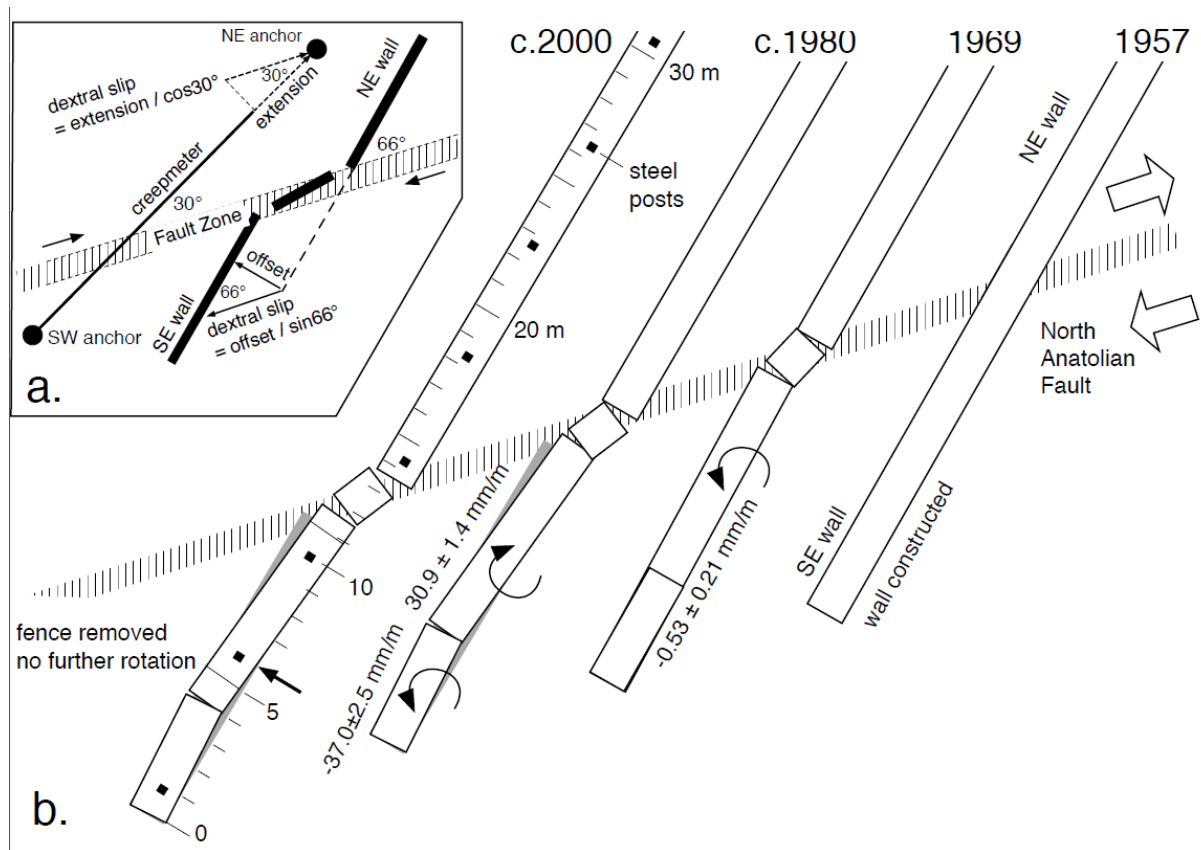


Figure 2a. Relations between dextral slip, wall offset and creep-meter extension. 2b In 1970 a chain link fence was fastened to the capstone of the wall, which caused the progressive clockwise rotation of the southernmost 11.7 m of the concrete cap from the underlying wall. Measurements in 2016 confirm the two halves of the underlying wall remain parallel. Rotation of the capstone ceased when the chain link fence was replaced by a free standing fence inside the wall. Throughout the northern half of the wall the saddle-coping presently overhangs the wall by  $90.8 \pm 8.6$  mm to the east. This overhang, when applied to the point where the east edge of the wall and capstone intersect) provides an unambiguous measure of offset for photographs taken from above the wall between 1970 and 2010. In 2016 this point lies midway between the southern end of the wall and the fault offset.

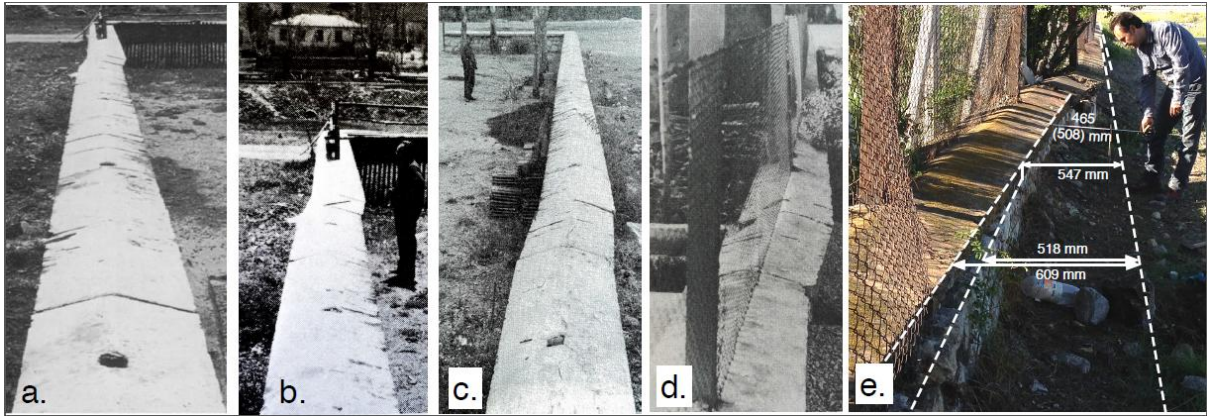


Figure 3. Published photographs of the Isetpasa wall. a) Ambraseys (1970) view south April 1969, b) Allen (1982) view south April 1969. c) view north (Aytun, 1982) 14 Oct 1969 and d) view north (Aytun, 1982) 11 July 1978. The house visible in (b) can be readily identified in 1965 and 2013 satellite imagery and constrains the 1969 azimuth of the wall to  $N15^{\circ}E$ . e) view north, May 2014. Distances from a projected alignment to the northern half of the wall to the southern wall and its damaged capstone are indicated where capstone rotation has exposed the top of the wall.

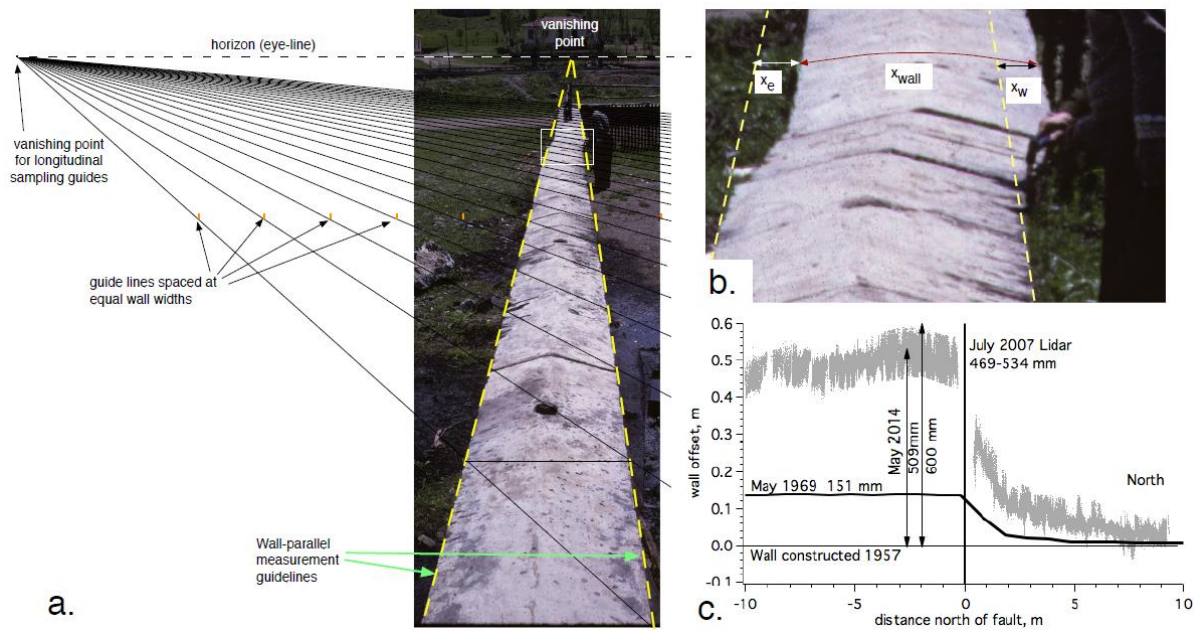


Figure 4a. Geometrical projection used for photo analysis of the Ismetpasa wall showing enlarged view of fault offset (4b) and measured parameters:  $x_e$  = offset east edge,  $x_w$ =offset west edge,  $x_{wall}$ = horizontal width of wall co-linear with  $x_e$  and  $x_w$ . At each longitudinal point the dimensionless ratio  $(x_e+x_w) / (2*x_{wall})$  was determined numerically and multiplied by the mean measured wall-width (695 mm) to determine the local offset in millimeters. 4c The capstone offset (black line May 1969) is compared with the 2007 LIDAR capstone and wall offset (grey shading indicates point cloud), and the April 2016 computed wall offset (red line with uncertainties).

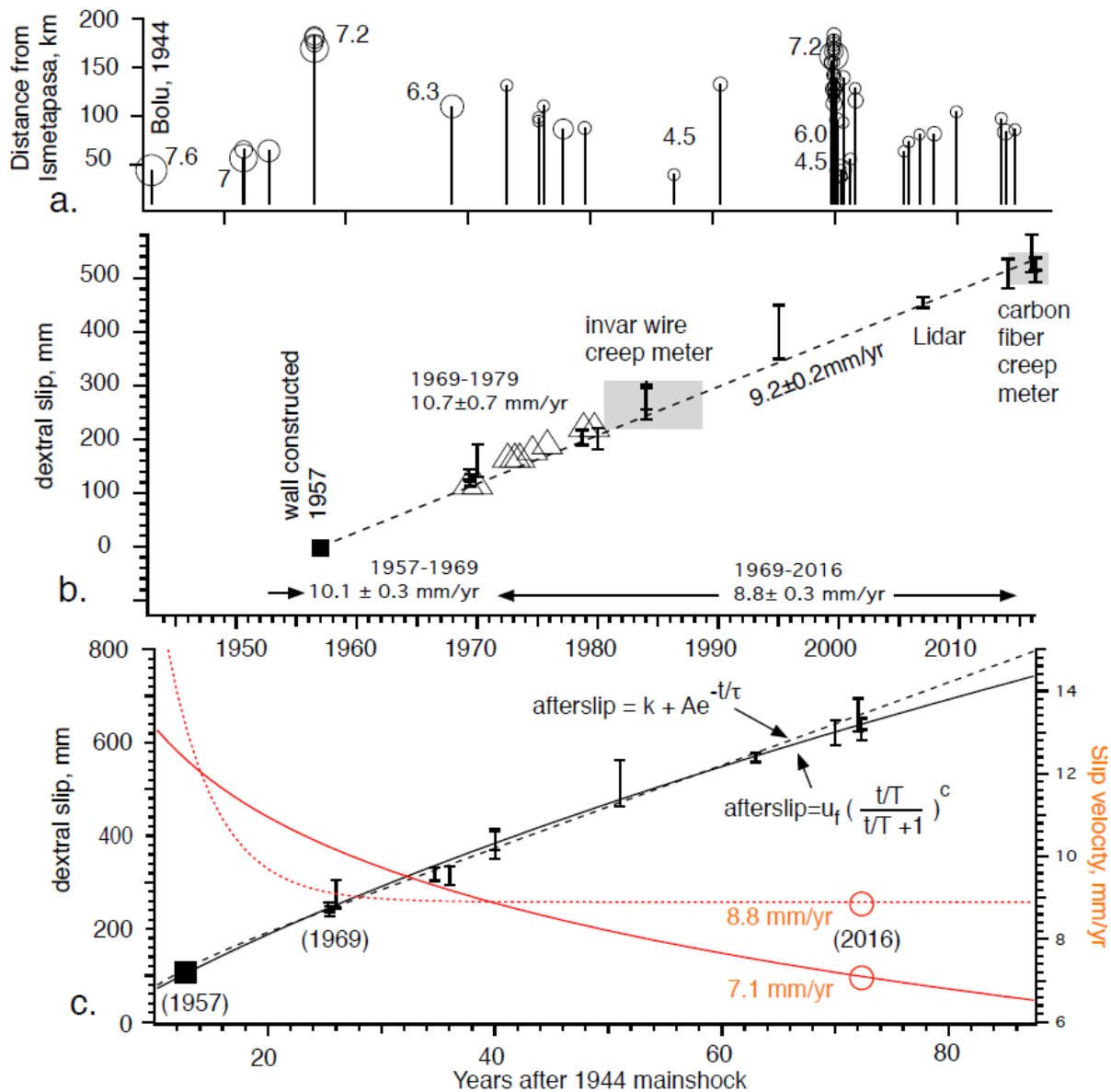


Figure 5a Epicenters of significant earthquakes as a function of distance from Ismetpasa. The Bolu Gerede earthquake occurred 05:22 local time 1 February 1944. 5b. Dextral surface creep at Ismetpasa since wall construction in 1957 (Table 1) fit to straight line segments. The weighted least squares fit to all the wall-offset data yields  $9.2 \pm 0.2$  mm/yr, but least-squares fits to early data yield a higher rate. Triangles indicate ten years of 155-m-aperture alignment array measurements with a mean rate of  $10.7 \pm 0.7$  mm/yr (Aytun, 1982), and grey areas indicate the times of operation of continuous creep-meters. 5c. Synthetic afterslip curves that equally fit the data shown in 5b but predict different early afterslip and cumulative interseismic creep (arbitrary absolute datum). The dashed exponential fit rapidly approaches a linear interseismic slip rate, after an initial decay with a  $\approx 10$  year decay constant. Present day velocities predicted by each curve are circled on the slip velocity curves (red).

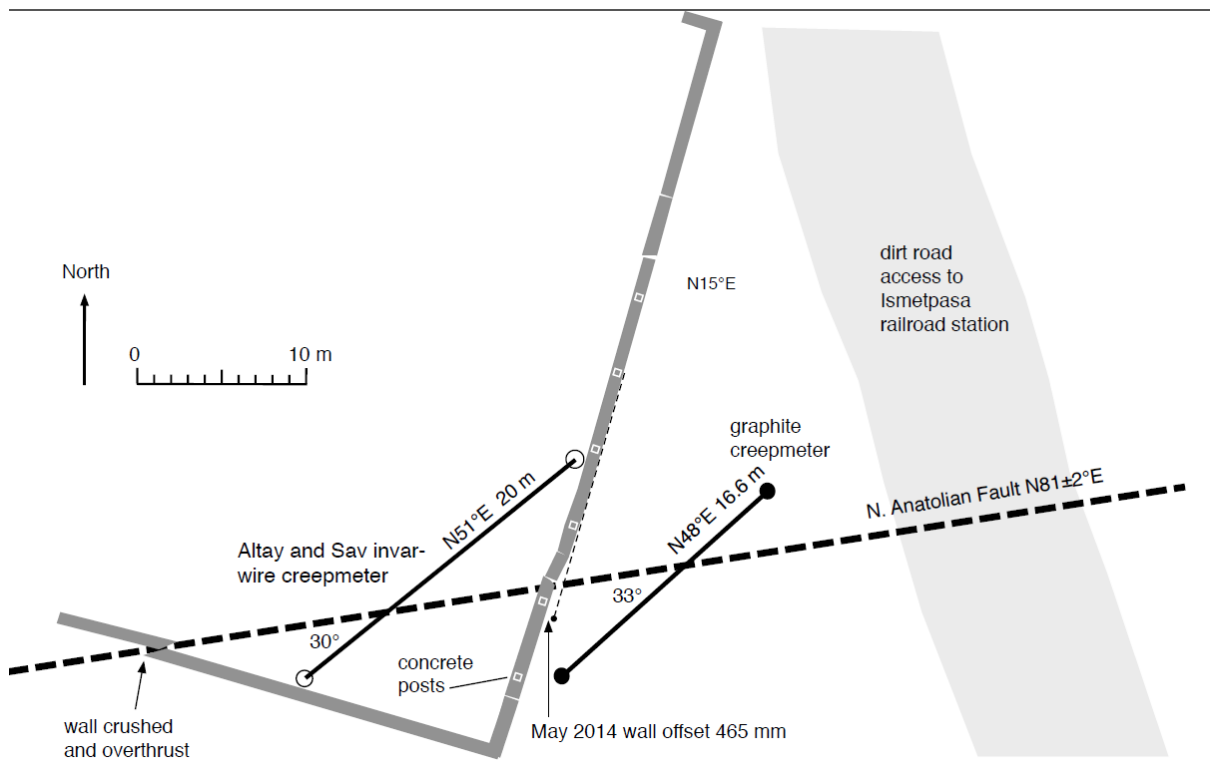


Figure 6. Location of creep-meters relative to the 47-m-long offset wall discussed in the preceding section.



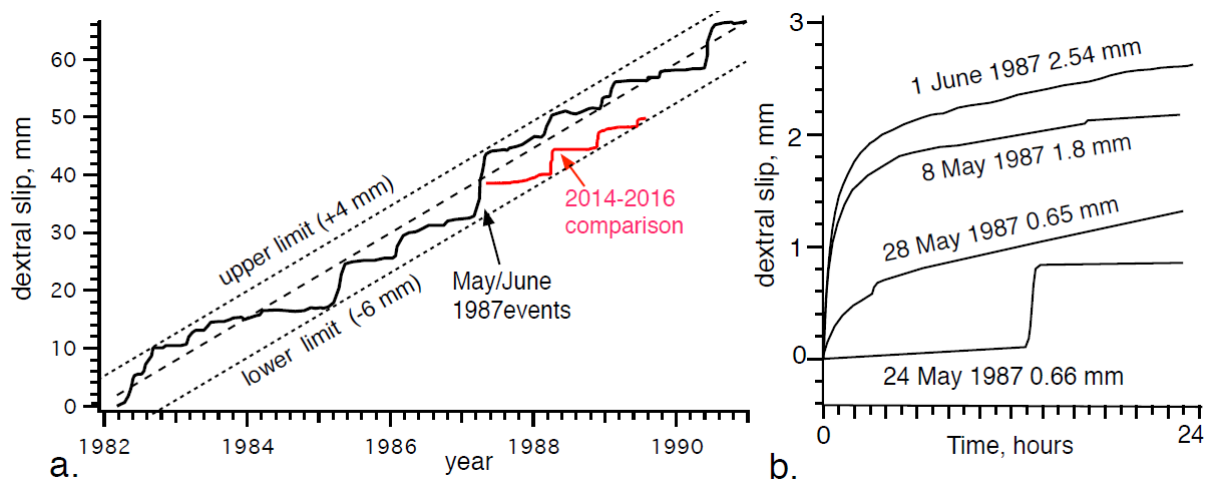


Figure 7a. Cumulative creep recorded by Altay and Sav (1991) with (7b) creep events reproduced from their article. The authors reported a mean creep rate of 7.46-7.7 mm/y; the dashed least-squares-fit line shown has slope  $7.3 \pm 0.1$  mm/yr. Creep events occur when surface creep lags 6 mm below the mean creep rate, and the relaxed surface slip state is no more than 4 mm above it. Recent 2014-16 creep data with arbitrary time and displacement datum are superimposed on the 1987-89 data for comparison purposes.

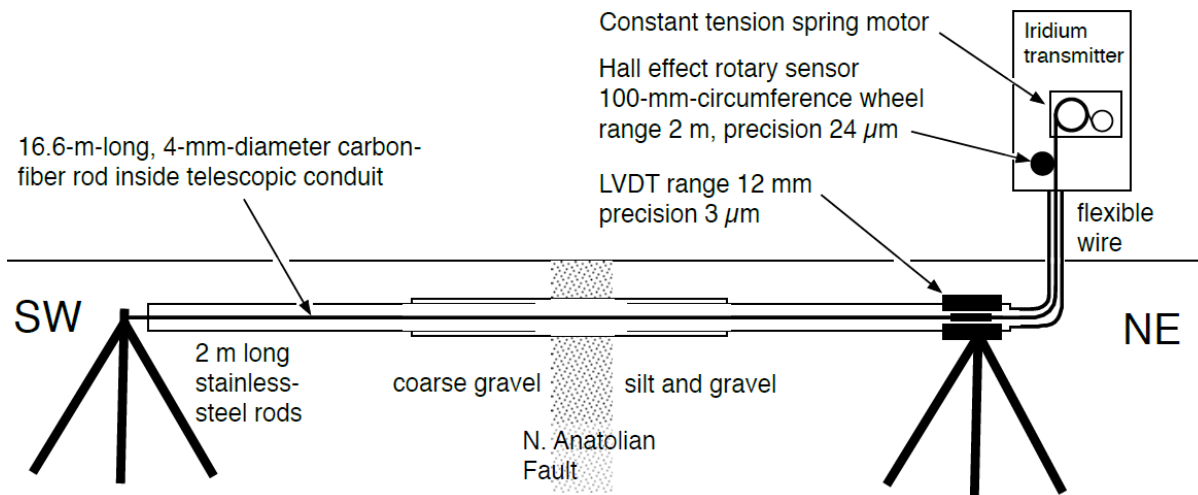


Figure 8. Schematic cross section of the 16.6-m-long Ismetpasa creep-meter. The rotary sensor is driven by a single turn wrapped around a precision 100 mm circumference wheel. The rod slides inside a PVC conduit which results in occasional "stiction" with amplitudes 3-20  $\mu\text{m}$ .

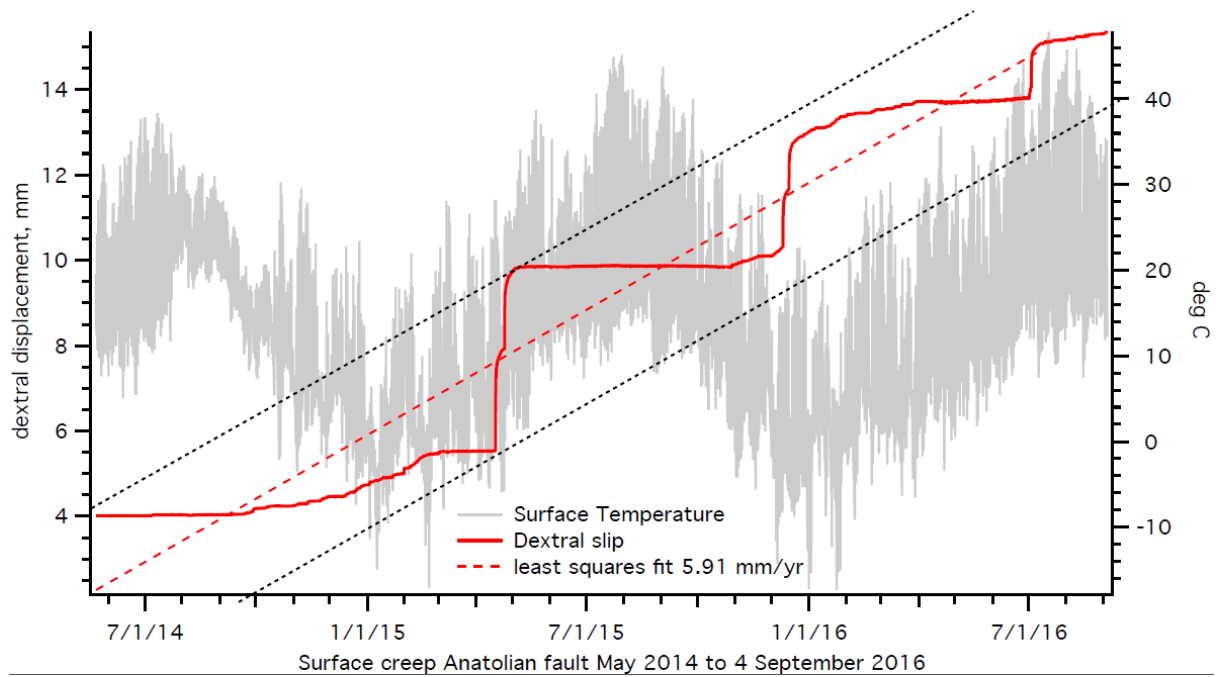


Figure 9 Ismetpasa creep data to May 2014 to September 2016 from the carbon-fiber rod creep-meter. The trace starts after a ~20 mm Aug-Sep 2013 creep episode reported by (Rousset et al., 2016) 10 km to the east of here and stops following a third creep episode that occurred too late for analysis in the current manuscript. The black dashed lines are parallel to, and  $\pm 2$  mm from, the least squares fit to these data ( $5.91 \pm 0.01$  mm/yr). In this time interval the surface trace spends 47% of its time locked, and 44% of its time creeping at a slow rate of  $3.5 \pm 2$  mm/yr. This slow slip contributes 25% of the total slip of 10.86 mm in 2 years. The creep trace is insignificantly influenced by environmental temperature effects (the grey trace records  $60^\circ\text{C}$  annual, and  $5\text{-}30^\circ\text{C}$  daily, surface temperature changes).

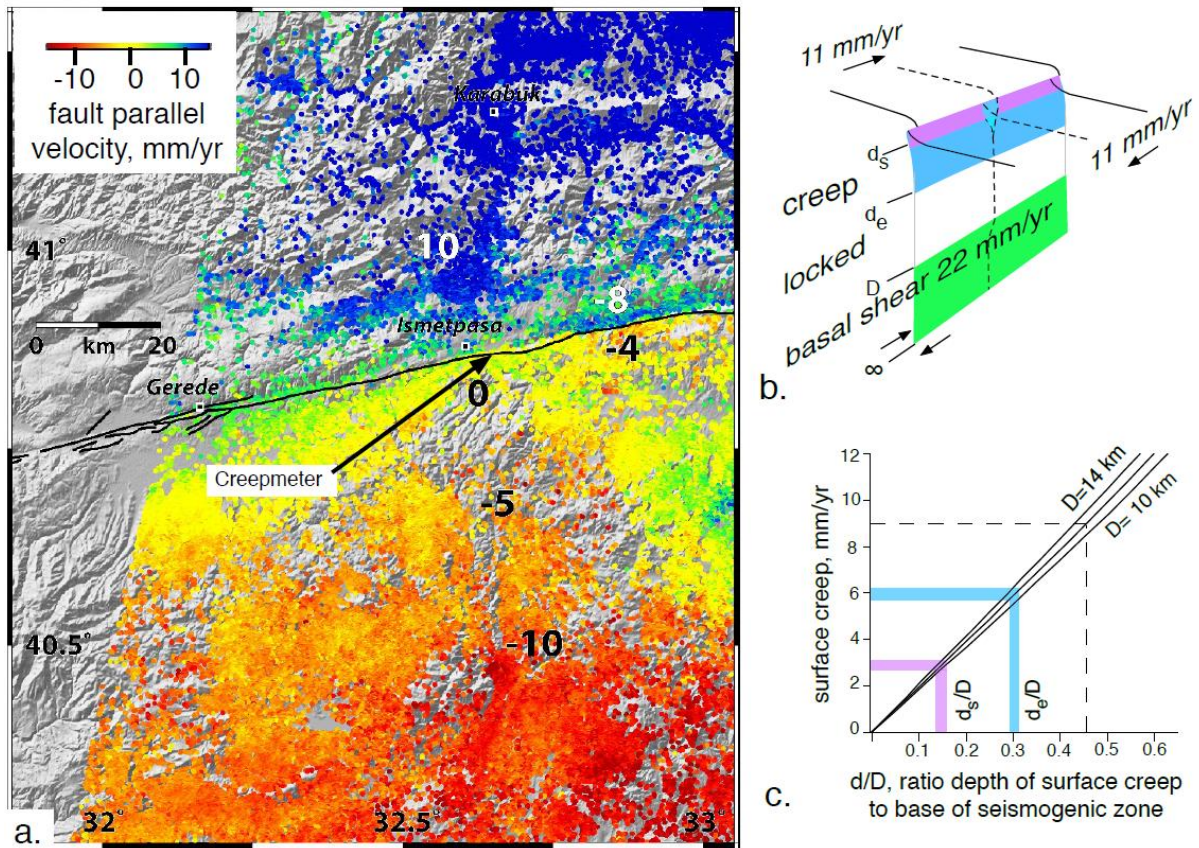


Figure 10a Point InSAR velocity field (2003-2010) from Cetin et al., (2014) with b. schematic view of steady and episodic creep on fault above the locked seismogenic zone. c. An elastic model of (b.) showing dependence of surface slip rate on the depth of creep for  $D=10, 12$  and  $14$  km. For the data shown in Figure 9, slow uneventful slip (violet) occurs at depths shallower than  $d_s = 1.8$  km, and episodic creep (blue) occurs to a depth of  $d_e = 3.6$  km for  $D = 12$  km.

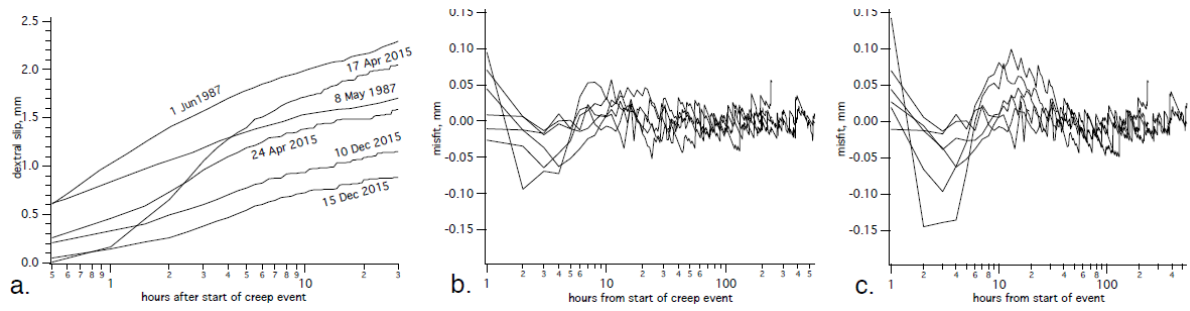


Figure 11 **a.** Log-linear plots of creep events from 1987 and 2015. Misfits to **b.** double exponential equation (1), and **c.** a three parameter physical description of surface slip (equation 2) that less precisely emulates the initial slip of each creep event. The step-like nature of the creep signal is caused by "stiction" in the creep-meter with an amplitude of 20-30  $\mu\text{m}$ .

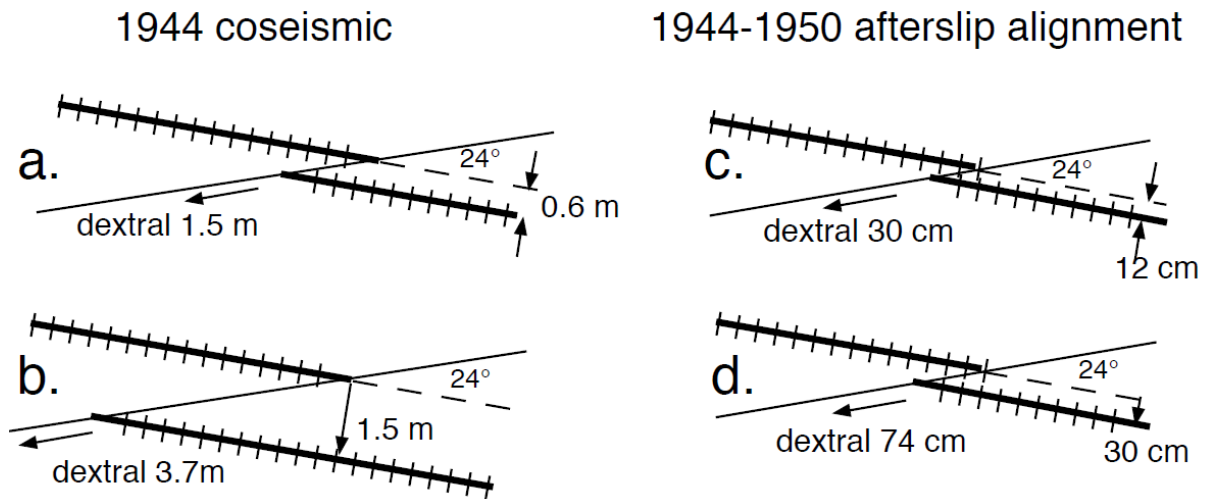


Figure 12. Alternative interpretations for rail offset and dextral fault slip. Line offsets were considered to describe dextral offsets (a. and c.) by Ambraseys (1970) and by subsequent authors. However, if the railroad engineers were describing offsets in the alignment of the rails, the implied dextral fault slip is increased by a factor of  $1/\sin 24^\circ$ . The cumulative coseismic slip (b.=3.7 m) and afterslip (d.=0.74 m) is consistent with the 3.6-6 m of slip documented by Kondo et al., (2015) nearby for the 1944 earthquake.

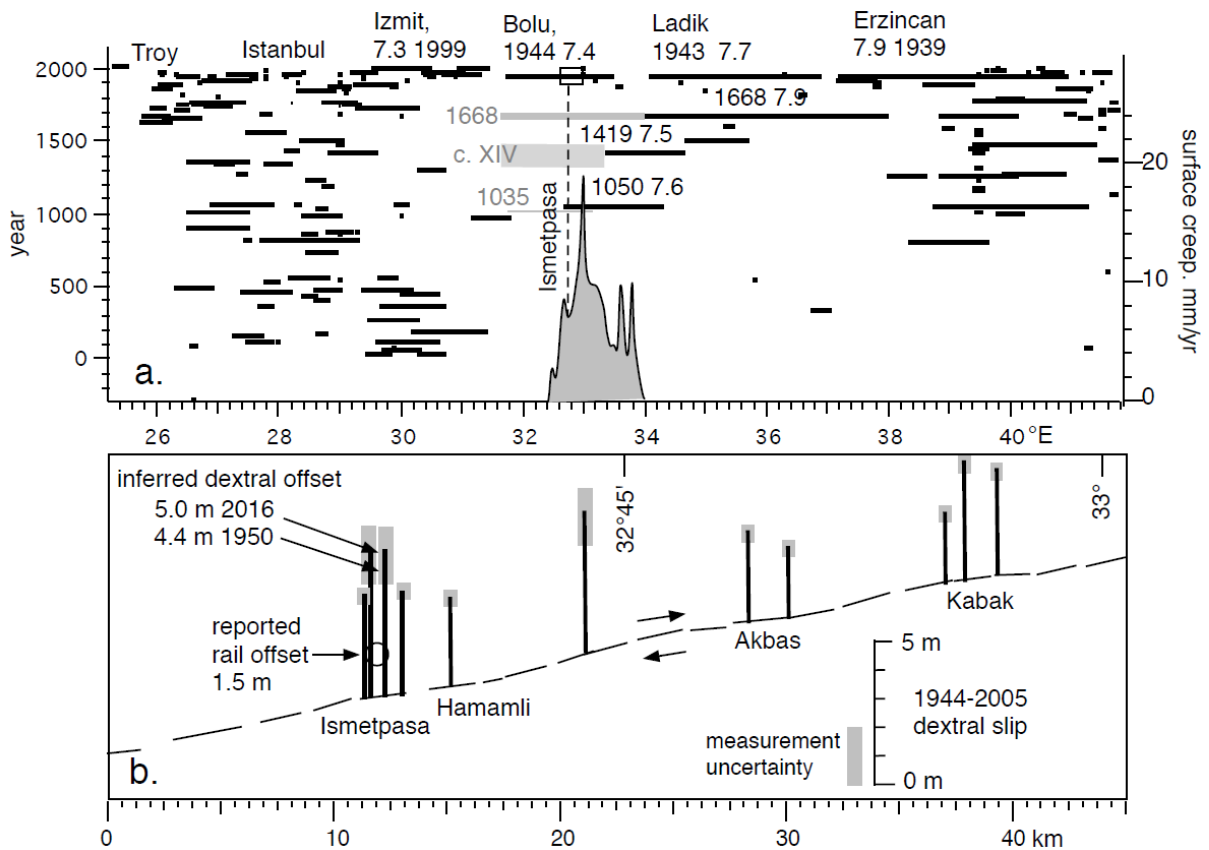


Figure 13a. Earthquake rupture space-time diagram along the North Anatolian fault for the past 2000 years (Bohnhoff et al., 2016) including inferred earthquakes (grey bars) near Ismetpasa in the Bolu/Gerede segment from Kondo et al., (2010). Surface creep (Cetin et al., 2014) is shaded grey, and the rectangular box on the Bolu/Gerede earthquake rupture is enlarged in the lower panel. 13b. Map view of Ismetpasa segment showing field offsets from Kondo et al., (2005) compared to reported rail offset (circle) from Ambraseys (1970). The rail offset is much lower than neighboring offsets, but is close to the 4.4 m dextral slip (arrowed) 1950 offset calculated at Ismetpasa assuming that the rails cross the fault obliquely at  $24^\circ$ . Nearby field offsets are indistinguishable from the  $\approx 5$  m of cumulative slip including post 1950 afterslip.

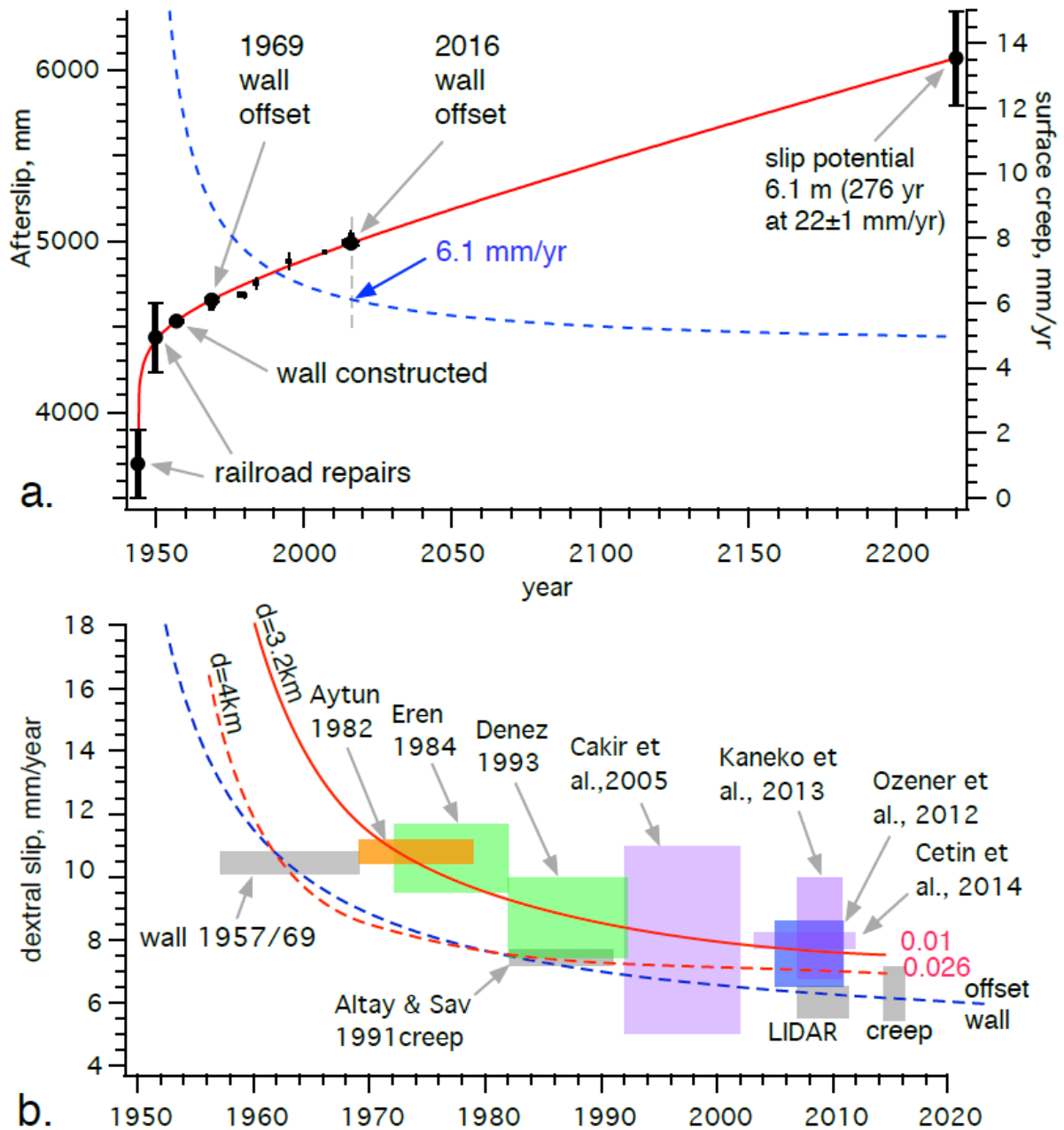


Figure 14a. Railroad and wall offsets (from Table 5) as a function of time. The best fit red line to the displacement data is obtained by simultaneously solving for a linear creep velocity (4.65 mm/yr) and afterslip predicted from Equation (2). **b.** Compilation of observed fault slip velocities from Table 6 (K6, K8 and K10 are identified in Table 5 and discussed in the text). While continuous creep measurements and LIDAR (Karabacak et al., 2011) generally agree with velocities derived from wall offsets (grey boxes and blue dashed line from Figure 14a), most geodetic estimates indicate a  $\approx 1.25$  faster velocity as depicted by the solid red line (predicted velocity decay curve for  $a-b=0.01$  and



locking depth of 3.2 km from Kaneko et al., 2013). The dashed red line corresponds to their calculation for locking depth of 4 km and  $a-b=0.026$ . Measurements with narrow aperture indicate a present-day (2016) creep rate of 6.1 mm/yr whereas large scale measurements (orange, 155-m alignment array; green and blue  $\approx 800$  m GPS array, violet  $>30$  km InSAR) favor a rate closer to 7.6 mm/yr. Some of the geodetic measurements include velocity estimates from regions to the east where slip rates are higher.