GEOLOGY | THE GEOLOGICAL SOCIETY OF AMERICA®

https://doi.org/10.1130/G47729.1

Manuscript received 2 April 2020
Revised manuscript received 26 August 2020
Manuscript accepted 29 August 2020

Thin crème brûlée rheological structure for the Eastern California Shear Zone

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ABSTRACT

Since the occurrence of the 1992 CE Mw 7.3 Landers and 1999 Mw 7.1 Hector Mine earthquakes in the Mojave Desert (California, USA), postseismic deformation following both earthquakes has been intensively studied, and models with a strong crust overlying a low-viscosity mantle asthenosphere have been favored. However, we recently found that the near-field postseismic transients after the two earthquakes have lasted longer than previously thought, which requires a revision of the postseismic modeling. Our new modeling results based on the revised postseismic transients show that: (1) the effective viscosity of the lower crust beneath the Mojave region at the decadal time scale is ~2×10^{20} Pa·s (transient viscosity ~2×10^{19} Pa·s), i.e., only ~5 times that of the underlying mantle asthenosphere, and (2) the transient viscosity of the upper mantle exhibits a time-dependent increase, providing fresh geodetic evidence for frequency-dependent rheology (e.g., Andrade or extended Burgers rheology). The inferred transient rheology for the first year agrees well with that obtained for the July 2019 Mw 6.4 and Mw 7.1 Ridgecrest earthquakes ~180 km north of the two Mojave events. Our modeling results support a thin crème brûlée model for the Eastern California Shear Zone (part of the Pacific-North America plate boundary) in which both the lower crust and the upper mantle exhibit ductility at decadal time scales.

INTRODUCTION

A long-standing debate in geodynamics is focused on the first-order variability in the rheology and strength of Earth's outermost layers with depth and, in particular, whether the long-term strength resides in the crust or the lithospheric mantle. Two end-member models of the rheological structure—the jelly sandwich and crème brûlée models—have been proposed (Chen and Molnar, 1983; Burov and Watts, 2006). The jelly sandwich model invokes a weaker lower crust that is sandwiched between a strong upper crust and lithospheric mantle (e.g., Burov and Watts, 2006). The crème brûlée model for some actively deforming regions has many variants, as either the upper crust or the entire crust can be strong, but they all include a weak upper mantle (Bürgmann and Dresen, 2008; Jackson et al., 2008). Where the load-bearing layer resides seems to also depend on the considered time scale of loading and deformation (Burov and Diament, 1995). Hence, a time-dependent crème brûlée model was proposed for actively deforming regions (e.g., back-arc regions), in which the lower crust is strong at the time scale of the seismic cycle (<10^4 yr) but appears significantly weaker at longer time scales (Thatcher and Pollitz, 2008).

The 1992 CE Mw 7.3 Landers and 1999 Mw 7.1 Hector Mine earthquakes occurred within the Eastern California Shear Zone (ECSZ; Mojave Desert, California, USA; Fig. 1) and produced some of the best-studied continental earthquake cycle deformation to date (Massonnet et al., 1993; Shen et al., 1994; Deng et al., 1998; Peltzer et al., 1998; Pollitz et al., 2001; Fialko et al., 2002; Pollitz, 2003, 2015; Fialko, 2004a; Freed and Bürgmann, 2004; Freed et al., 2012). In the past 20 years, most studies of the postseismic deformation following the two Mojave earthquakes concluded that the steady-state viscosity of the lower crust is at least an order of magnitude greater (>10^{20} Pa·s) than that of the upper mantle (e.g., Freed et al., 2012; Pollitz, 2015), which implies that the lower crust essentially behaves as an elastic material at decadal time scales. Hence, the rheological structure of the Mojave region serves as a prime example of the time-dependent crème brûlée model (Thatcher and Pollitz, 2008; Freed et al., 2012; Pollitz, 2015).

In this study, we explored the time-dependent rheology of the lower crust and upper mantle beneath the Mojave Desert at time scales of months to decades, based on new insights gained from modeling the postseismic deformation of three M >7 earthquakes in the ECSZ (Fig. 1), and we found evidence for a thin crème brûlée rheological structure with resolvable viscous relaxation below the brittle upper crust.

ENDURING LATE-STAGE POSTSEISMIC TRANSIENTS

An accurate characterization of the preseismic background deformation is required for obtaining reliable postseismic transients (Savage and Svarc, 2009; Liu et al., 2015). Although intensively studied, the near-field post-Landers GPS data were not properly corrected for the secular deformation. Most studies (e.g., Spinler et al., 2010; Herbert et al., 2014) obtained the alleged interseismic GPS velocities consistent with the solution by Shen et al. (2011) without calibration. Some recent studies derived the secular velocities from the postseismic GPS time series with a time span >7 yr (Freed et al., 2012; Pollitz, 2015) but without constraints from the pre-Landers data. This method works well for the far-field data, because the postseismic transients there are subtle. We used pre-Landers triangulation and trilateration data (Sauber et al., 1994) to establish the secular deformation field (Liu et al., 2015). We found that the nominally interseismic GPS velocities derived from the post-Landers GPS data (Shen et al., 2011)

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CITATION: Liu, S., et al., 2020, Thin crème brûlée rheological structure for the Eastern California Shear Zone: Geology, v. 49, p. , https://doi.org/10.1130/G47729.1

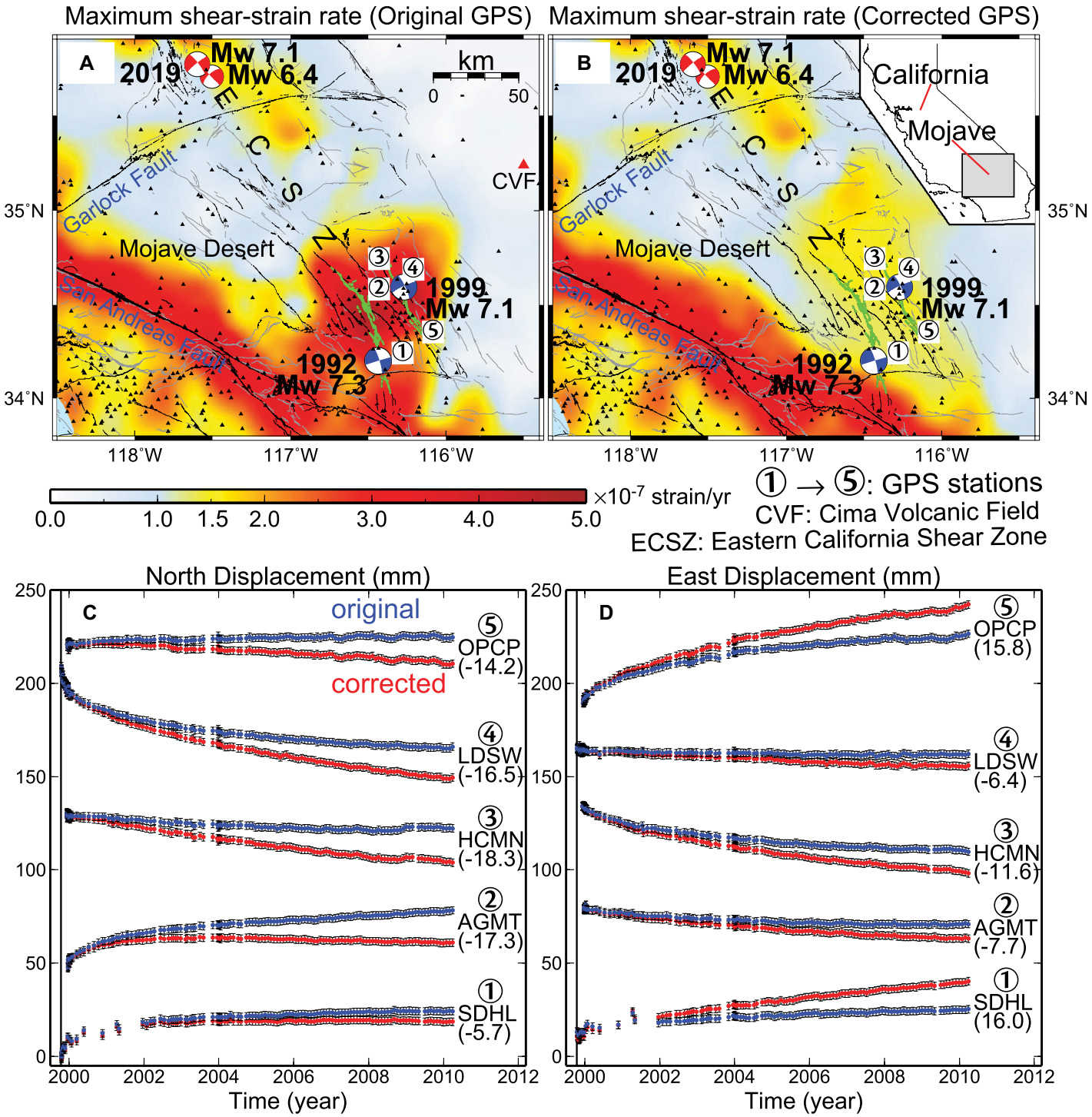

Figure 1. GPS-derived strain rate and postseismic GPS time series. (A–B) Maximum shear-strain rate derived from (A) original (Shen et al., 2011) and (B) corrected (Liu et al., 2015) GPS velocities. Blue beach balls are focal mechanisms of the 1992 and 1999 CE Mojave events in the Mojave Desert Eastern California Shear Zone (ECSZ), red beach balls are the 2019 Ridgecrest events in the central ECSZ, green lines are coseismic ruptures, and black triangles show positions of all GPS sites. (C–D) Original (blue dots) and corrected (red dots) near-field postseismic displacement after 1999 event. Numbers in parentheses below GPS site names (A and B) indicate excess (new minus original) postseismic displacements at the time of 2010.26.

correspond to abnormally high maximum shear-strain rates in the near field (>3 × 10^{-7} strain/yr; Fig. 1A), comparable to those across the nearby San Andreas fault. The near-field velocity solution from Pollitz (2015) is similar to ours but is not constrained by the pre-Landers data in the southern Mojave region. Thus, the previously estimated GPS velocities may contain biases in the form of residual postseismic transients, leading to an excess of 2–3 mm/yr of right-lateral shear across the 1992 and 1999 ruptures. After the velocity correction, the maximum shear-strain rates across the 1992 and 1999 ruptures are reduced to ~1.5 × 10^{-7} strain/yr (Fig. 1B), consistent with the observed pre-Landers strain rates (Sauber et al., 1994; Liu et al., 2015). Thus, the near-field postseismic transients are larger and more

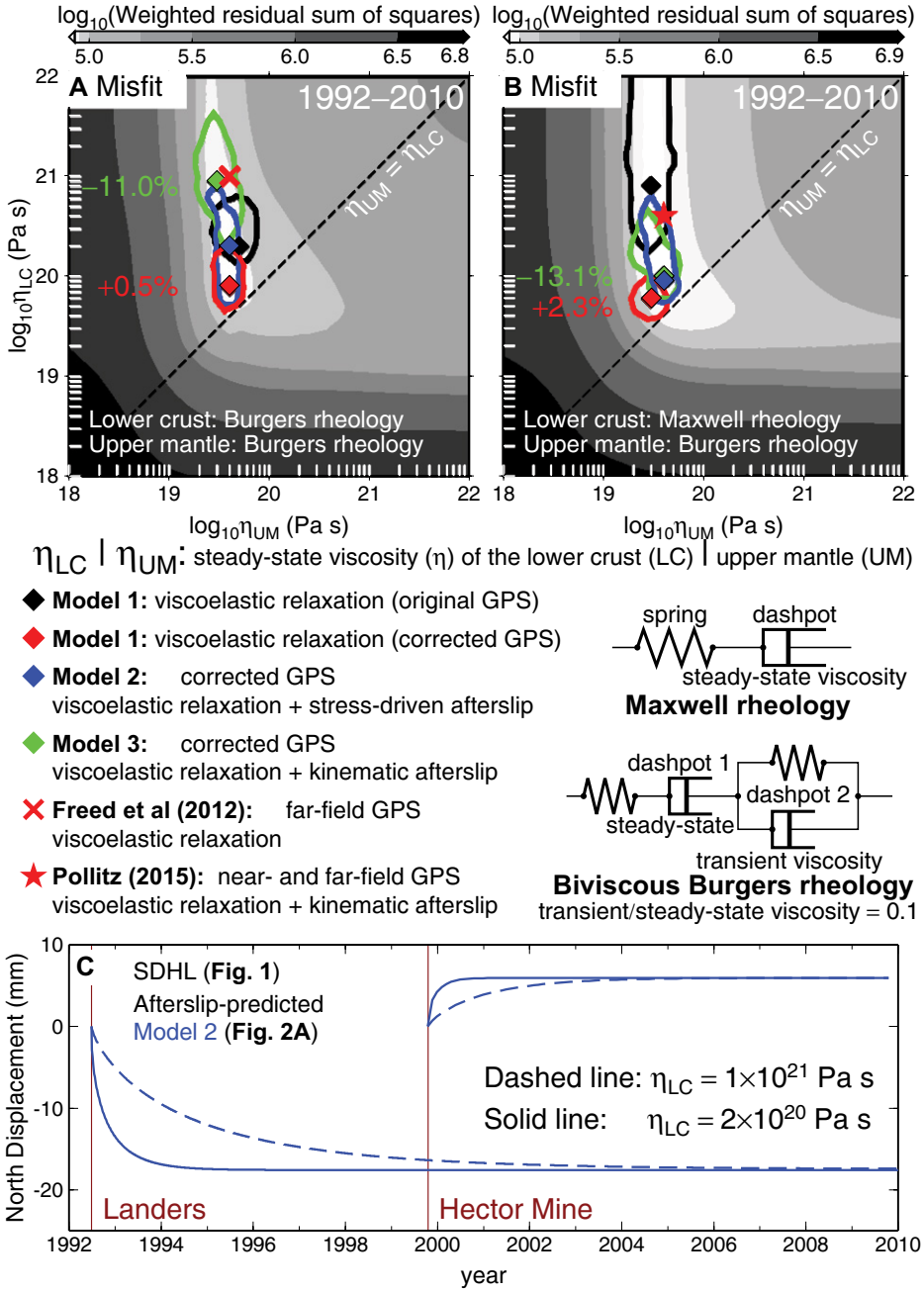

Figure 2. Model misfit to 1992–2010 CE GPS data as a function of steady-state viscosities of the upper mantle (η_{UM}) and lower crust (η_{LC}): (A) Burgers and (B) Maxwell rheology for lower crust. Colored lines show statistical 95% confidence regions for weighted residual sum of squares (WRSS) of models relative to the best-fit models marked by color-coded diamonds of Model 1 (red), Model 2 (blue and gray shades), and Model 3 (green). Black diamond and contour mark results of Model 1 without GPS velocity correction. Red and green numbers indicate percentage difference of WRSS for the best-fit Model 1 and Model 3 relative to the best-fit Model 2. (C) Stress-driven afterslip-induced northward displacement at SDHL (Fig. 1).

enduring than previously determined (Figs. 1C and 1D), which suggests that we must revisit the postseismic deformation following the two Mojave earthquakes.

POST-LANDERS AND HECTOR MINE EARTHQUAKE DEFORMATION MODELING

We considered poroelastic rebound (fluid flow in response to earthquake-induced pressure changes; Peltzer et al., 1998), afterslip (aseismic slip surrounding the rupture and on its deeper extension; Shen et al., 1994), and viscoelastic relaxation (postseismic flow of ductile rocks below the upper crust; Pollitz et al., 2001) to explain the corrected 1992–2004 campaign GPS (Shen et al., 2011) and 1999–2010 continuous GPS (Blewitt et al., 2018) time series from stations <300 km from the epicenters of the two Mojave events (see the Supplemental Material[1]). We used a simple three-layer Earth structure with an elastic upper crust underlain by viscoelastic lower crust and upper mantle. To ensure consistency, we used coseismic slip models (Simons et al., 2002; Fialko, 2004b) derived in the same layered elastic structure to drive the modeled deformation.

When only viscoelastic relaxation is considered (Model 1), if Maxwell rheology is applied, the optimal viscosities of the lower crust and upper mantle are ~6×10^{19} Pa·s and ~3×10^{19} Pa·s, respectively (Fig. 2). If biviscous Burgers rheology is applied to both layers, the preferred steady-state viscosities of the lower crust and upper mantle are ~8×10^{19} Pa·s and ~4×10^{19} Pa·s (transient/steady-state viscosity = 0.1, unless specified otherwise), respectively, and the misfit to the data decreases by 6.0%–20.4% (red diamonds in Fig. 2). If the interseismic velocity correction is not considered, the optimal steady-state viscosity of the lower crust increases to >2.0×10^{20} Pa·s, while that of the upper mantle remains consistent with our new estimates.

When both viscoelastic relaxation and stress-driven or kinematic afterslip are considered (i.e., Models 2 and 3), the steady-state viscosity of the upper mantle remains at ~4×10^{19} Pa·s, but that of the lower crust depends on the selected bulk rheology (Fig. 2). If Burgers rheology is applied (Fig. 2A), the preferred steady-state viscosity of the lower crust is ~2×10^{20} Pa·s and ~9.0×10^{20} Pa·s for Model 2 and Model 3, respectively. If Maxwell rheology is applied (Fig. 2B), the optimal viscosity of the lower crust is 9–10 × 10^{19} Pa·s, irrespective of the considered afterslip model (stress-driven or kinematic).

Shallow poroelastic rebound in the upper 2.5 km (hydraulic diffusivity ≥1.0 m^2/s), rather than upper crustal-scale poroelastic rebound, is required to reduce the data misfit (Fig. 3; for Model 2, the inclusion of this process reduces the misfit to the post–Hector Mine GPS data at the 89.4% confidence level), which is consistent with evidence from near-field interferometric synthetic aperture radar (InSAR) data (Peltzer et al., 1998). However, considering poroelastic rebound does not affect the estimates of viscosities of the lower crust and upper mantle, which are constrained by more far-reaching deformation.

TRADE-OFFS AMONG CHOICES OF COSEISMIC RUPTURE MODEL, AFTERSLIP, AND TRANSIENT RHEOLOGY OF THE LOWER CRUST

Estimates of the lower-crustal viscosity are affected by the magnitude of stress changes in

[1]Supplemental Material. Supplemental details on data analysis, modeling methods, and modeling results, as well as supplemental figures and table. Please visit https://doi.org/10.1130/GEOL.S.12964937 to access the supplemental material, and contact editing@geosociety.org with any questions.

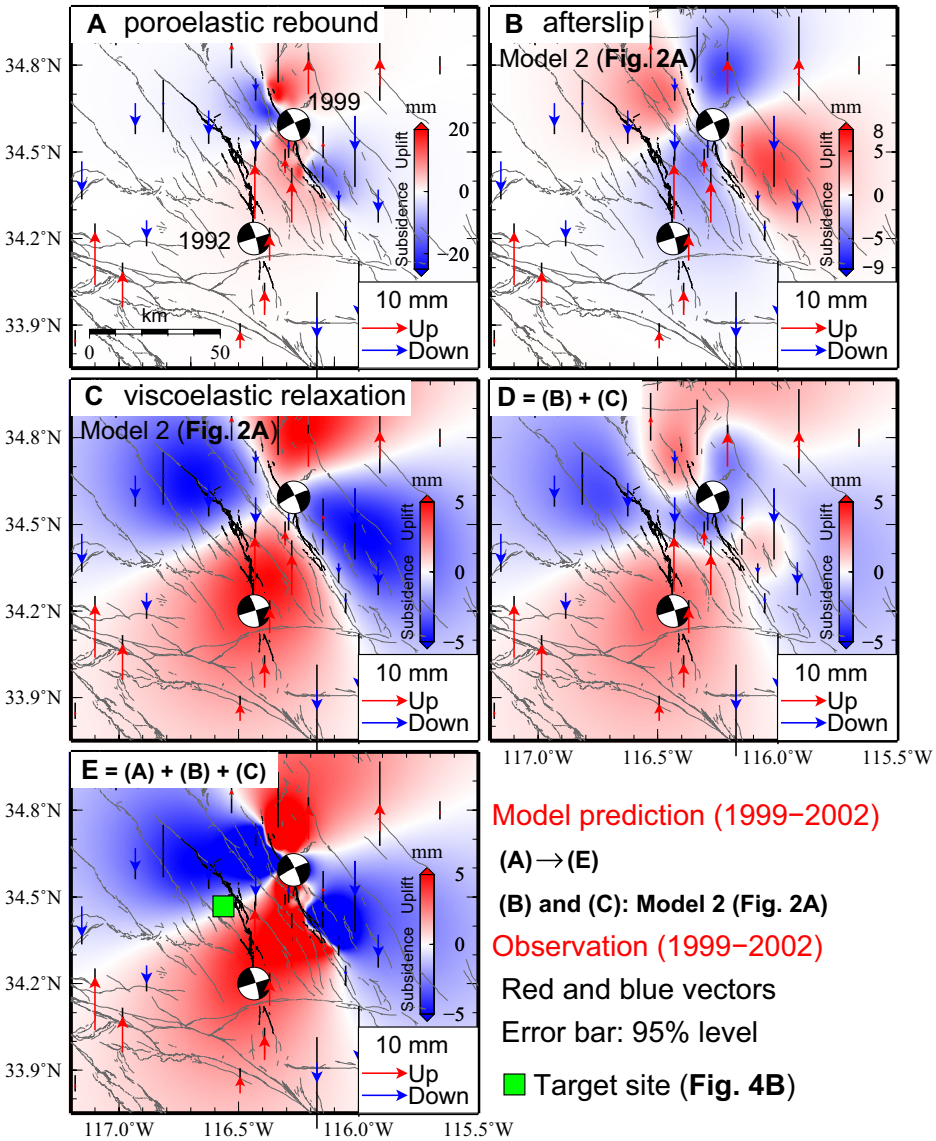

Figure 3. Observed and modeled vertical displacements (1999–2002 CE).

the lower crust imparted by the coseismic rupture. Rupture models derived in a homogeneous half-space (e.g., Xu et al., 2016) have slip at 10–15 km depth that is ~20% smaller than the slip models from layered Earth models (Simons et al., 2002; Fialko, 2004b), and so they induce ~20% smaller differential stress in the lower crust, and the estimated lower-crustal viscosity decreases by ~20% (Model 1).

Both deep afterslip and transient viscoelastic relaxation in the lower crust generate similar near-field horizontal postseismic transients, making our estimates of the lower-crustal viscosity sensitive to the trade-off between the two processes (Fig. 2). Assuming Burgers rheology for the lower crust, we find that: (1) the best-fit steady-state viscosity of the lower crust is ~2 × 10^{20} Pa·s (Fig. 2A), and (2) afterslip ends in the first 2 yr (Fig. 2C). A steady-state viscosity of the lower crust close to 10^{21} Pa·s requires afterslip lasting for >3 yr (Fig. 2C). This model can fit the observed vertical displacements, but the data misfit of the horizontal displacement is worse than the best-fit Model 2 at the 99% confidence level.

If Burgers rheology is applied to the lower crust (Fig. 2A), the transient viscosity is ~2 × 10^{19} Pa·s (Fig. 4A), suggesting that lower-crustal rocks exhibit ductility at the decadal time scale (even if afterslip is allowed). The earthquake-induced differential stress at 25 km depth can diminish by ~30% and ~45%, respectively, 20 and 50 yr after the 1992 event (Fig. 4B); approximately twice as much as that for the case with Maxwell rheology. If not (Fig. 2B), more enduring afterslip is required to explain the temporal evolution of the near-field signals (similar to the dashed lines in Fig. 2C), but the best-fit steady-state viscosity still agrees with the case with Burgers rheology. We prefer the best-fit Model 2 with Burgers rheology, which fits the data better than the Maxwell rheology model at the 99% confidence level.

EVOLVING TRANSIENT RHEOLOGY OF THE UPPER MANTLE

Our estimates of the transient viscosity of the upper mantle, considering different time spans after the main shocks, exhibit a time-dependent increase (Fig. 4A). For the Mojave case, we fix the steady-state viscosity to the value derived from the full 18 yr of GPS data (4 × 10^{19} Pa·s), then the optimal transient mantle viscosities are found to be ~4 × 10^{17} Pa·s and ~9 × 10^{17} Pa·s for the data in the first month and in the first year, respectively. For the 2019 Ridgecrest case (Fig. 1), using the same steady-state viscosity, we find that the transient viscosity of the upper mantle is ~3 × 10^{17} Pa·s and ~1 × 10^{18} Pa·s in the first month and first year, respectively (Section S3.4 in the Supplemental Material).

Our findings provide prima facie observational evidence for a frequency-dependent transient rheology of the upper mantle, similar to a recent theoretical argument (Lau and Holtzman, 2019) but at a shorter time scale. So far, models with a biviscous Burgers body (Pollitz, 2003), power-law rheology (Freed and Bürgmann, 2004), or both (Freed et al., 2012) have been employed, but with only one mode of transient bulk rheology. Burgers body models, with a transient viscosity 10% of the steady-state viscosity, underpredict the far-field deformation in the first year (Freed et al., 2012, their figure 9; Pollitz, 2015, his figure 3). Post-Landers, Hector Mine, and Ridgecrest earthquake GPS data support more than one characteristic time of transient deformation in the upper mantle, showing that a nonlinear power law, Andrade, or extended Burgers rheology is required to fit the data. Yet, the caveat is that a pronounced depth/temperature-dependent rheology could provide an alternative explanation for such time-varying viscosities (Freed et al., 2007; Riva and Govers, 2009).

THIN CRÈME BRÛLÉE MODEL FOR THE ECSZ

Most previous studies of the postseismic deformation in the Mojave Desert have found a crust-to-mantle viscosity ratio ≥10 and support a time-dependent crème brûlée model (Pollitz et al., 2001; Freed et al., 2007, 2012; Pollitz, 2015): the short-term lithospheric strength resides in the whole crust, but the lower crust may lose strength in the long term, making the upper crust the only load-bearing layer at geologic time scales (Thatcher and Pollitz, 2008). Models constrained by far-field GPS data, which are incapable of tightly resolving the lower-crustal viscosity, favor a crust-to-mantle steady-state viscosity ratio close to 25 (Freed et al., 2012), and we also found a similar ratio when using only the far-field data. Pollitz (2015) used post-Hector Mine GPS data similar to ours and fixed the ratio to 10. In contrast, we obtained

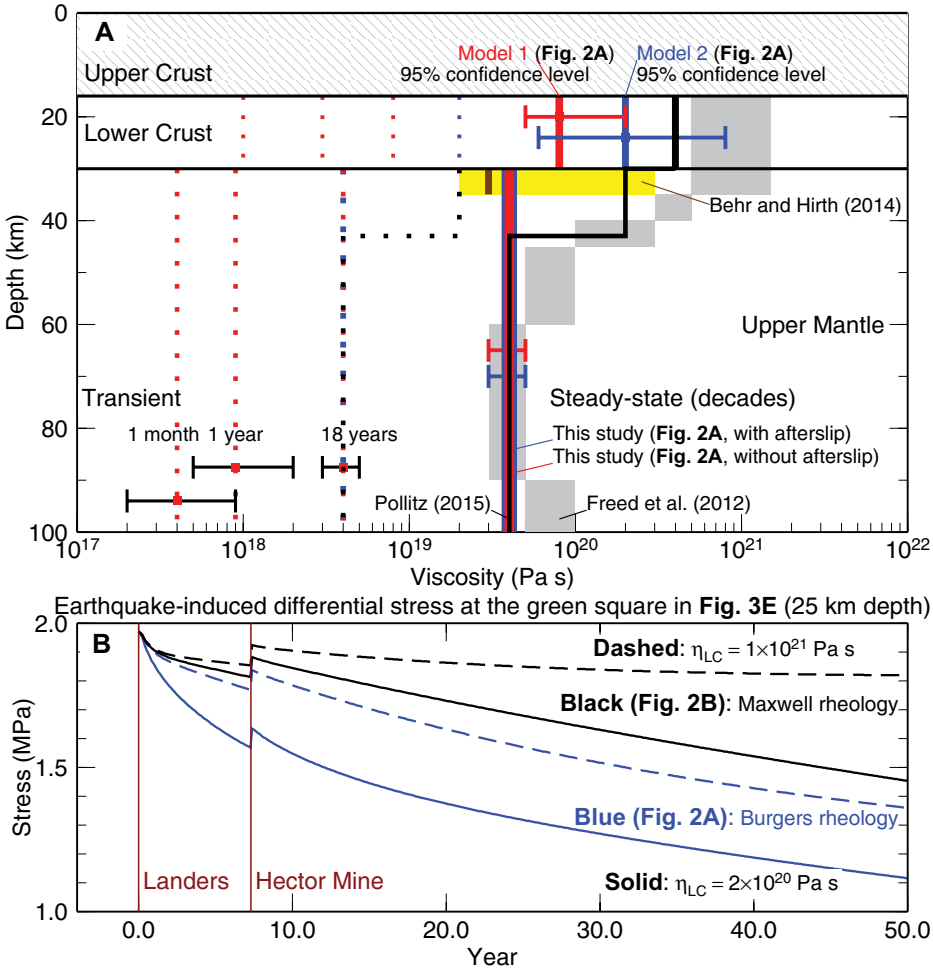

Figure 4. (A) Inferred rheological structure for the Eastern California Shear Zone (ECSZ). Dotted/solid lines are transient/steady-state viscosities (red and blue—this study; black—Pollitz, 2015). Acceptable steady-state viscosities for the lower crust and upper mantle: red/blue horizontal bars for the case without/with afterslip (Fig. 2A). Black bars indicate range of transient viscosities of the upper mantle. Other previous results are shown by gray shading based on far-field GPS data (Freed et al., 2012) and by yellow shading (brown line—optimal value) based on mantle peridotite xenoliths from the Cima volcanic field (eastern California; Behr and Hirth, 2014). (B) Evolution of differential stress at 25-km-deep target (Fig. 3E); η_{LC}—viscosity of lower crust.

a preferred ratio of ~5 and found that the ratio proposed by Pollitz (2015) lies near the upper bound of ours (Fig. 2B). To explain the crust-to-mantle viscosity ratio ≥10, the rock-forming minerals were assumed to be wet plagioclase or mafic rock for the lower crust and hydrated olivine for the upper mantle (Behr and Hirth, 2014; Pollitz, 2015). However, the steady-state viscosity of wet plagioclase can be ~2×10^{20} Pa·s (Getsinger et al., 2013), i.e., lower than previous geodetic results but consistent with ours (~2×10^{20} Pa·s). Also, the lower crust may contain quartz (Lowry and Pérez-Gussinyé, 2011), implying that the lower crust might be weaker than that found in previous geodetic (Freed et al., 2012), geophysical (Shinevar et al., 2018), and petrological (Behr and Hirth, 2014) studies.

We propose a thin crème brûlée model for the rheological structure of the Mojave Desert at earthquake-cycle time scales. As revealed by 18 yr of GPS data, the lower crust did experience substantial stress relaxation at the decadal time scale (Fig. 4B). The similar spatial distribution of faults (forming a 60–90-km-wide shear zone), cumulative strike-slip rate (~12 mm/yr; Meade and Hager, 2005), effective elastic thickness (<10 km; Lowry and Pérez-Gussinyé, 2011), and surface heat flow (>70 mW/m^2; Blackwell and Richards, 2004) north of the Garlock fault (Fig. 1), and the transient viscosities that we derived from the post-Ridgecrest earthquake data lead us to propose a similar rheological structure for the southern and central ECSZ. Because the effective elastic thickness in the region is found to be <10 km (Lowry and Pérez-Gussinyé, 2011), i.e., ~2–3 km less than the thickness of the seismogenic layer (Rolandone et al., 2004), both the lower crust and upper mantle might be weak at geodetic and geologic time scales.

ACKNOWLEDGMENTS

We appreciate constructive reviews by Greg Hirth, Fred Pollitz, and James Jackson. We thank Sylvain Barbot for sharing the Unicycle code; Rongjiang Wang for his guidance on the PSGRN/PSCMP software; and Xiaohua Xu and Kang Wang for sharing their rupture models. Liu acknowledges support from the Institute of Geology, China Earthquake Administration. This study was funded by King Abdullah University of Science and Technology (KAUST, Saudi Arabia) award BAS/1/1353–01–01. This is Southern California Earthquake Center publication 10779.

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