Postseismic deformation due to the 2019 Ridgecrest earthquakes and its impact on the CGM

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Introduction

The Community Geodetic Model (CGM) is built on the complementary strengths of temporally dense GNSS data and spatially dense InSAR data. Unlike any of the other SCEC CXM models, one unique feature of the CGM is its time dependence, meaning that the CGM is expected to characterize not only the average velocity but also the temporal evolution of the ground deformation in Southern California.

Atmospheric noise remains one of the major error sources in InSAR measurement. To mitigate this problem, some type of temporal smoothing or filtering is generally applied in the InSAR time series analysis. This is mostly fine for slow and steady deformation processes, including the interseismic deformation. Yet in the case of a sudden change in the strain rate followed by a gradual and slow transient signal with unknown temporal evolution (i.e., coseismic deformation followed by postseismic transients), temporal smoothing or filtering would blend the deformation from one process to the other, making the interpretation of the resulting time series less straightforward. For this reason, the current effort of the CGM development uses Sentinel-1 InSAR data only from before the 2019 Ridgecrest earthquakes, meaning that the current CGM would only cover the time period from early 2015, when the Sentinel-1 mission started to collect data over Southern California, to the time before the 2019 Ridgecrest earthquake. To maintain a continuous CGM that goes beyond the 2019 Ridgecrest earthquake sequence, one must find ways to separate and correct for the co- and postseismic deformation due to the earthquake, especially if the interseismic strain is determined to be the primary target of the CGM.

This project is aimed at deriving the postseismic deformation ~1-2 years after the 2019 Ridgecrest earthquake and using the measured surface deformation to constrain postseismic relaxation models that can eventually be used to correct for the postseismic transients of the 2019 Ridgecrest earthquake sequence. This will allow for the CGM to finally be determined to cover a longer time period spanning beyond the 2019 Ridgecrest earthquake sequence. In addition to directly contributing to the development of the CGM, postseismic relaxation models derived in this study will shed light on the mechanical properties of fault zones, including frictional properties of the fault in the case of afterslip, the effective viscosity of the lower crust and upper mantle in the case of viscoelastic relaxation, and hydrological properties of rocks surrounding the fault in the case of poroelastic rebound.

Data processing

GNSS

There are more than 400 continuous GNSS stations within 300 km from the 2019 Ridgecrest earthquake. In this project, we analyze the time series at those stations ~one year after the mainshock. To minimize the potential contamination due to postseismic deformation from the 1992 Landers and the 1999 Hector Mine earthquakes, we use data from two years before the 2019 Ridgecrest earthquake to estimate the preseismic velocity at each station and subtract it from the postseismic recordings. The horizontal displacement field revealed by the GNSS observations is characterized by south-to-southeastward motion on the eastern side of the fault and north-to-northwestward motion on the western side of the fault, a deformation pattern similar to the coseismic displacement field. The postseismic deformation is clear as far away as the Mojave Desert, which is ~100 km south of the 2019 Ridgecrest earthquake. We find that postseismic deformation at those greater distances can be well modelled by viscoelastic relaxation.
primarily in the upper mantle using a Burger’s rheology with a transient viscosity of $\sim10^{18}$ Pas and a steady-state viscosity of $4\times10^{19}$ Pas (Liu et al., 2020).

Figure 1. InSAR observations of postseismic deformation following the 2019 Ridgecrest earthquake sequence. (a) and (b) cumulative line-of-sight (LOS) displacements 0.9 year after the mainshock derived from Sentinel-1 data of the ascending track ASC64, and descending track D071, respectively. (c) and (d) show the cumulative postseismic displacements along the East-West and vertical directions, respectively, by decomposing the LOS displacements from both the ascending and descending tracks. (e) zoom-in of the vertical displacement near the mainshock epicenter. Magenta triangles represent the campaign-mode GNSS sites deployed shortly after the 2019 Ridgecrest earthquake (Brooks et al., 2020). (f) Time series of the LOS displacement at a point near the mainshock epicenter (magenta square in (e)) with different strategies for mitigating the atmospheric noise.

**InSAR**

The InSAR analysis in this project primarily relies on the C-band Sentinel-1 observations. We use the ERA-5 weather model in combination with Common-Scene-Stacking (CSS) (Tymofyeyeva and Fialko, 2015) to correct for the atmospheric noise. We find that these methods are effective to reduce the scatter and variations in the resulting time series, while still maintaining the temporal decay feature associated with the postseismic transient (Figure 1f). Combining data from the two looking directions, we derive the displacements along the E-W and vertical directions. Similar to the GNSS measurement, the InSAR-based E-W displacement field is characterized by primarily eastward motion on the eastern side of the fault, and westward motion on the western side. Such a deformation pattern is qualitatively consistent with what one would expect from afterslip. The vertical displacement field however is rather complex, with most deformation being concentrated around the rupture tips and the fault geometric complexities, including the releasing-bend fault step-over near the mainshock epicenter and the fault junction between the Mw6.4 foreshock and the Mw 7.1 mainshock.

**Modeling of postseismic deformation**

Physical models of surface deformation due to postseismic relaxation mechanisms (e.g. afterslip, viscoelastic relaxation, and poroelastic rebound) should reflect the input coseismic stress perturbation. For afterslip, here we only kinematically invert for the distribution of afterslip using available surface displacement, without invoking any dynamics and constitutive law that may control the afterslip evolution in space and time (e.g. rate-and-state friction), because the 2019 Ridgecrest earthquake rupture is rather complex and current data may not be sufficient to robustly constrain the mechanical properties (e.g. friction) associated with afterslip. For viscoelastic relaxation and poroelastic rebound, our
simulations of surface deformation is based on stress changes from the coseismic slip model of Wang et al. (2020).

**Viscoelastic relaxation**

To avoid the potential trade-off between afterslip and the viscoelastic relaxation, in the current study, we do not use the surface deformation to constrain the rheological parameters. Instead, we use a rheological structure inferred by modeling the postseismic transients >10 years after the 1992 Landers and the 1999 Hector Mine earthquakes (Liu et al. 2020). We find that the rheological model we adopt here can satisfactorily explain the GNSS displacements in the far field to the south of the rupture, where the contribution from afterslip becomes negligible. However, the model predicts substantially smaller displacements at several sites around Owens Lake to the north. The difference in misfit between the observation and model may be indicative of spatial variation of the rheological structure around the region.

**Afterslip**

The horizontal GNSS and InSAR E-W displacements after correcting for the viscoelastic relaxation are then used to invert for the afterslip. Here we only use the horizontal displacements in the afterslip inversion because as discussed above, the vertical displacement field seems to be mostly controlled by poroelastic rebound around fault geometric complexities and rupture tips. We use a fault geometry similar to the coseismic slip model of Wang et al. (2020), but extend the fault segments to a depth of 45 km. The faults are discretized into patches whose sizes increase with depth.

![Figure 2](image.png)

Figure 2. Distribution of cumulative afterslip 0.9 years after the mainshock. Grey dots in panels to the right represent the aftershocks during ~2 weeks after the mainshock (Shelly, 2019). Blue contours represent the coseismic slip contours at 1 meter interval (from Wang et al., 2020). Red star represents the hypocenter of the mainshock. Green star denotes the approximate location of the Mw 5.5 aftershock on 06/30/2020.

Figure 2 shows the cumulative afterslip 0.9 year after the mainshock. One striking feature of the afterslip model is that it is similar to the coseismic rupture, most of the afterslip moment concentrates in a relatively shallow depth range (0-10 km). In addition, the maxima of afterslip seem to be mostly collocated at or close to areas of high coseismic slip. We run the inversion with different model regularizations and data selections, e.g., smoothing and relative weighting between the GNSS and InSAR datasets. The resulting afterslip models are overall similar to what is shown in Fig. 2, despite some small-scale variations. While horizontal displacements predicted by the afterslip model shown in Fig. 2 match the observations reasonably well, the afterslip model predicts the opposite sign of the vertical displacements in the areas of major vertical deformation, suggesting that the vertical displacement is mainly controlled by other relaxation mechanisms. While the sign of vertical deformation due to deep-seated viscoelastic relaxation is also expected to be opposite to that from afterslip (e.g., Pollitz et al., 2001), the wavelength of surface deformation due to the viscoelastic relaxation is much longer compared to what the data reveal. In addition, the magnitude of vertical displacement calculated from the
viscoelastic relaxation model based on the rheological structure of Liu et al. (2020) is <5 mm during the observation period, which is about one order smaller than the maximum vertical displacement inferred from the InSAR observation. Therefore, we suggest that vertical postseismic deformation observed in this study is mainly due to poroelastic rebound.

**Poroelastic rebound**

Signals of poroelastic rebound have been observed following several moderate to large earthquakes. Yet, to date, most previous studies have modeled the poroelastic rebound by simply differencing the surface displacements calculated from a coseismic slip model with different Poisson’s ratios (corresponding to undrained to drained conditions) (e.g., Peltzer et al. 1998; Jónsson et al., 2003; Fialko, 2004). In this study, we model the surface deformation due to the poroelastic rebound using the software package PEGRN/PECMP developed by Wang and Kümpel (2003), which solves the fully coupled poroelastic diffusion problem in a layered half-space. We attempt to constrain the hydrological properties, specifically the hydraulic diffusivity of the shallow crust, by comparing the time series of the surface deformation between observations and model predictions.

As a start, we focus on the area close to the mainshock epicenter (Figure 1e). We use GNSS time series recorded by a campaign network deployed shortly after the earthquake (Brooks et al., 2020), and the InSAR LOS displacement time series from the ascending track Asc64 after correcting for the contributions from viscoelastic relaxation and afterslip for each time epoch. For the GNSS displacements, we only used the vertical components at stations within the ‘uplift’ feature near the mainshock epicenter (clr1, clr2, clr7), as the horizontal displacements may include significant contributions from the afterslip. For the same reason, one must first correct for afterslip contributions in the LOS displacement time series at each SAR image acquisition time. To do so, we assume that afterslip everywhere on the fault follows the same temporal evolution pattern, so its temporal characteristics can be approximated by that of the surface displacements. We perform a Principal Component Analysis (PCA) to the GNSS time series at stations of intermediate distance (<80 km) from the rupture. We then compute the LOS displacement at each SAR acquisition epoch using the afterslip model shown in Figure 2 whose temporal evolution follows best-fitting logarithmic function of the PC1 temporal response.

In this report, we focus on the poroelastic rebound process in the area around the mainshock epicenter. The magnitude of the poroelastic rebound depends on the hydraulic diffusivity structure of the poroelastic medium, specifically the depth to which the crust is allowed to accommodate poroelastic rebound and the hydrological properties of the layer. Here we assume that only the uppermost several km of the Earth’s crust is sufficiently porous and dominates the observed surface deformation (e.g., Ingebritsen and Manning, 1999). The thickness of this porous layer (2km in this study) is obtained by comparing the wavelength of surface deformation across the uplift feature near the mainshock epicenter between poroelastic rebound model predictions and observations. Under the assumption of a single porous layer with uniform porous properties, the only parameter that controls the speed (i.e. temporal evolution) of the poroelastic diffusion and the resulting surface deformation is the hydraulic diffusivity of the layer. Focusing on the temporal pattern, for each model we scale the resulting surface displacement time series to ensure that they best match the total displacements at the end of corresponding GNSS and InSAR observation periods.
Figure 3. Modeling of the poroelastic rebound in the area close to the mainshock epicenter as a function of hydraulic diffusivity. The top row is based on the vertical displacement time series at the GNSS site clr1. The bottom row is based on the average LOS displacement time series of pixels around the mainshock epicenter uplift feature. Magenta lines for both cases represent the best-fitting exponential functions to the respective displacement time series.

Figure 3 shows the comparison of displacement time series in the area of clear postseismic uplift near the mainshock epicenter between observations and model predictions due to poroelastic rebound as a function of the hydrological diffusivity. Both the GNSS and InSAR LOS displacement time series best match with the model with a hydraulic diffusivity on the order of 0.01 m^2/s, which is comparable to in-situ estimates of the fault zone diffusivity based on tidal response data (Xue et al., 2013; Xu et al., 2016), but is at least one order lower than that estimated with the propagation of seismicity accompanying fluid pressure diffusion through fractured rocks (e.g., Shapiro and Dinske, 2009; Yu et al., 2019).

The current postseismic observations also indicate possible heterogeneities in the hydrological properties along the fault. In particular, the InSAR observations reveal a large area of significant uplift near the fault junction between the Mw 6.4 foreshock rupture and the Mw 7.1 mainshock rupture, where afterslip produces subsidence. However, the poroelastic rebound model using the hydrological properties constrained by data in the uplift area near the mainshock epicenter produces only modest and localized surface uplift west of the fault junction, regardless of the coseismic slip model used in the calculation. We are still exploring what could explain the heterogeneity in hydraulic diffusivity and other hydrological parameters along the fault.

In summary, in this project, we derived postseismic deformation time series using InSAR and GNSS observations collected ~1 year after the mainshock. Using these measurements, we developed postseismic relaxation models that can satisfactorily explain the data during the observation period. However, data used in this project span over one year after the earthquake, and the data quality is still relatively low. In particular, current InSAR data are still too noisy to robustly distinguish the contributions from the viscoelastic relaxation. In addition, as we discussed above, current data are indicative of spatial heterogeneity in the mechanical properties, e.g., hydrological properties, along the fault. We still need to quantitatively characterize these heterogeneities and associate them with local geology. We anticipate that with more SAR image acquisitions over the next year, and an updated algorithm we have recently developed to account for the long-wave atmospheric noise in InSAR measurements using the GNSS zenith delay products, the results, including both measurements of the surface deformation and model constraints will be further improved.
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References


