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Key points:

- A combined mechanism of afterslip and viscoelastic relaxation is proposed for 4-years' deformation after the 2015 Gorkha earthquake
- The combined model is well constrained by Global Positioning System data and validated by independent InSAR observations
- Relative contributions of afterslip and viscoelastic relaxation show significant spatiotemporal variation

Supporting Information:

- Supporting Information S1

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Overlapped Postseismic Deformation Caused by Afterslip and Viscoelastic Relaxation Following the 2015 Mw 7.8 Gorkha (Nepal) Earthquake

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Abstract Studying transient postseismic deformation is one of the key approaches to probe fault kinematics and rheological behavior of rocks surrounding tectonic faults. Previous studies have shown that transient postseismic deformation is dominantly induced by afterslip and postseismic viscoelastic relaxation. However, effects of these two mechanisms are generally overlapped in space and time, making the analysis of the observations challenging. Here, we investigate the overlapping effects of afterslip and viscoelastic relaxation following the 2015 Gorkha earthquake based on the first 4-years' Global Positioning System observations and finite element models. Separated mechanisms either fail to explain the observations or lead to unreasonable model parameters. A combined model that incorporates stress-driven afterslip and viscoelastic relaxation can overcome the limitations of separated mechanisms. Our results indicate that only short-term (<1 year) postseismic deformation in near-to middle-fields is caused mainly by afterslip, whereas that in far-field or long-term are dominated by the viscoelastic relaxation effect. The stress-driven afterslip is mainly confined to a narrow zone adjacent to the coseismic rupture asperities, in contrast to the widespread deep afterslip as suggested by previous studies. The total seismic moment released by the afterslip is ~5% of that released by the Mw 7.8 main shock. The optimal viscosity of the lower crust beneath the south Tibet is estimated to be 2.5×10^{18} Pa-s, which is at least one order lower than that beneath Indian plate, suggesting a significant lateral variation across the Himalayan fault. Validations by using independent InSAR observations further confirm the reliability of the combined model.

1. Introduction

By accommodating ~50% of the India-Eurasia plate convergence, the Main Himalayan Thrust (MHT) is considered to be a seismogenic zone with potential for generating large earthquakes (Avouac, 2003; Bilham, 2004). The recent 2015 Gorkha earthquake partly ruptured a known seismic gap on the central MHT between the 1505 M 8.5 earthquake to the west and the 1934 M 8.2 earthquake to the east (Figure 1). Seismological studies reveal that the rupture of the Gorkha earthquake initiated 15–18 km beneath the Gorkha region, then propagated eastward for ~140 km (e.g., Avouac et al., 2015; Fan & Shearer, 2015; Lay et al., 2017). Geodetic slip models indicate that coseismic slip of the earthquake was concentrated on the subhorizontal décollement at a depth of 10–15 km, whereas the shallow portion of the fault remains unbroken (e.g., Diao et al., 2015; Elliott et al., 2016).

After the event, significant postseismic surface deformation has been observed by Global Positioning System (GPS) and InSAR (Interferometric Synthetic Aperture Radar) techniques (e.g., Mencin et al., 2016; K. Wang & Fialko, 2018). Based on these postseismic observations, several previous studies have investigated two main mechanisms: afterslip and viscoelastic relaxation. Postseismic deformation in early periods is mostly attributed to afterslip at the downdip extension of the coseismic slip (Gualandi et al., 2016; Jiang et al., 2018, 2019; Mencin et al., 2016; Sreejith et al., 2016; K. Wang & Fialko, 2018). To explain long-term and widespread postseismic deformation, combined models that incorporate afterslip and viscoelastic relaxation are considered to be more reasonable (Tian et al., 2020; Zhao et al., 2017).

Although previous models explained most of the observed postseismic displacements, the inferred apparent afterslip mostly distributed far beyond the coseismic slip and aftershock cluster (Figure 1), in contrast to the previous finding that afterslip and aftershocks mainly occur in areas adjacent to the coseismic asperities

Figure 1. Tectonic setting of the 2015 Gorkha earthquake. Shaded ellipses denote the estimated rupture areas of historical earthquakes in this region (Avouac et al., 2015). The red and blue contours show the coseismic slip distributions of the 2015 Gorkha earthquake and its largest aftershock (this study) projected on the surface with intervals of 2 m. Colored lines show locations of deep afterslip inferred by previous studies: a = Z. Jiang et al. (2018), b = Zhao et al. (2017), c = Mencin et al. (2016), e = Wang and Fialko (2018), d = Sreejith et al. (2016), f = Tian et al. (2020). Black triangles represent GPS stations used in this study. Black stars show epicenter of the 2015 event and the Mw 7.4 aftershock. White dots are aftershocks following the Gorkha earthquake (Adhikari et al., 2015). The data of three far-field GPS sites (i.e. XZZB, XZAR, and XZZF) are from Crustal Movement Observation Network of China (CMONOC), whereas the data of other stations are from Nevada Geodetic Laboratory (NGL).

(e.g., Diao et al., 2018; Hsu et al., 2006; Scholz, 1998). Liu-Zeng et al. (2020) found that broad afterslip far beyond the coseismic rupture of the 2015 Gorkha earthquake require the constitutive parameter $(a-b)\sigma$ to decrease northward to ≤ 0.05 MPa and seems unlikely considering the low interseismic coupling and high mid-crustal temperatures beneath southern Tibet. Therefore, they concluded that the widely distributed afterslip far beyond the rupture is physically unreasonable, whereas viscoelastic relaxation in a shear zone at the north extension of the MHT should be more reasonable. Besides, viscoelastic relaxation models utilized by the previous studies either neglect lateral variation of the rheological structure across the Himalaya (Z. Jiang et al., 2018), or do not incorporate the descending slab (Tian et al., 2020; K. Wang & Fialko, 2018; Zhao et al., 2017).

Previous studies have investigated aforementioned issues and found that associated model simplifications may have biased the inferred afterslip or viscosity. Sun and Wang (2015) explored the effect of slab thickness and viscosity structure on inferred afterslip distribution following thrust earthquakes (2011 Mw 9.0 Tohoku earthquake and 2005 Mw 8.7 Nias earthquake), and indicated that these parameters play a major role for the estimated depth of afterslip. Wiseman et al. (2015) suggested that lateral heterogeneity of viscosity across the fault and existence of an elastic slab would significantly affect the simulated deformation due to viscoelastic relaxation following the 2004 Sumatra-Andaman earthquake. Peña et al. (2020) considered a power-law rheology when modeling viscoelastic relaxation following the 2010 Mw 8.8 Maule earthquake and suggested that the power-law rheology can reduce the solved deep afterslip.

Although combined models were proposed already to evaluate the relative contributions of afterslip and viscoelastic relaxation following the Mw 7.8 Gorkha earthquake, the modeling approaches are debatable (Diao et al., 2018; A. M. Freed et al., 2017). The authors of these models first derive an optimal viscoelastic model from the data, and then estimate the afterslip based on the residual data, that is, they consider the

two main mechanisms hierarchically rather than simultaneously. Therefore, the interaction between afterslip and viscoelastic relaxation has been generally ignored for the present case. Recent studies have tried more sophisticated modeling approach to account for the coupling effect of these two main mechanisms (Agata et al., 2019; Diao et al., 2018; Qiu et al., 2018). As the results shown in Diao et al. (2018), viscoelastic relaxation effect due to afterslip can reach up to ~15% of the total postseismic displacements in the first 7 years after the 2008 Mw 7.9 Wenchuan earthquake. Moreover, models proposed by previous studies can generally explain the data used for constraining the model, however, few of them were validated by using independent observations.

In this paper, we probe the afterslip and viscoelastic relaxation based on geodetic data acquired in the first 4 years following the 2015 Gorkha earthquake, utilizing a more reliable viscoelastic structure model. First, we start models by considering afterslip and viscoelastic relaxation as separated mechanisms to investigate how they can explain the observation and whether the inferred model parameters are reasonable. Then, we present a combined model incorporating stress-driven afterslip and viscoelastic relaxation induced by coseismic rupture and time-dependent afterslip. All investigated viscoelastic models incorporate a “ramp-flat-ramp” fault geometry (Elliott et al., 2016) and lateral inhomogeneity of rheological structure. Based on the model results, we evaluate the relative contributions of the two mechanisms, and estimate fault afterslip and rheological behavior of materials surrounding the earthquake. For separated and combined models, we obtain the optimal model parameters by inverting the GPS data, and validate the reliability of these models using the independent InSAR observations. Finally, we discuss the inferred fault afterslip and rheological structure.

2. Postseismic Geodetic Observations

With the GPS stations around the 2015 Gorkha earthquake, postseismic deformation following this event has been measured, which is valuable for probing relevant mechanisms. Here we use the GPS position time series from two data sets to infer the postseismic deformation following this event. One includes GPS stations surrounding the coseismic rupture in Nepal (Mencin et al., 2016; K. Wang & Fialko, 2018). In this data set, most stations had been installed several years before the event (carried out by California Institute of Technology and the Department of Mines and Geology of Nepal). A few stations were installed after the event (Mencin et al., 2016). The time series of those stations were processed and updated by Nevada Geodetic Laboratory (NGL, <http://geodesy.unr.edu/>) utilizing GIPSY-OASIS-II (version 6.1.1) software in the reference frame IGS08 (Tian et al., 2020; K. Wang & Fialko, 2018). The other data set includes continuous GPS stations in southern Tibet that carried out by the CMONOC projects (Crustal Movement Observation Network of China). With GAMIT/GLOBK (version 10.4) software the time series of the CMONOC stations were processed in the reference frame ITRF 08 (<http://www.cgps.ac.cn/>). This data set provides a valuable far-field constraint. As shown by Tian et al. (2020), the obtained displacement time series of these two data sets show good consistency (with differences less than 1 mm) after removing the secular velocities, although different processing approach, model assumptions and reference frame were used. In this paper, we use the two data sets by directly combining them together.

We use the following station-dependent function to describe the time-dependent characters of the observed displacement time series (Diao et al., 2018; Tian et al., 2020),

$$U(t) = U_0 + V_0(t) + \sum_{i=1}^2 \left(C_i \cos \frac{2\pi t}{T_i} + S_i \sin \frac{2\pi t}{T_i} \right) + \left\{ D_c + D_p \left[1 - e^{-\frac{t-t_{eq}}{\tau_p}} \right] \right\} H(t - t_{eq}) \quad (1)$$

where t_{eq} is the occurrence time of the event, U_0 and V_0 represent the preevent station position and the secular velocity of the GPS station, respectively. C_i and S_i are the coefficients describing the seasonal effects (with $i = 1$ and 2 for the annual and semi-annual periods, respectively). D_c is the coseismic displacement offset, D_p is the accumulated postseismic displacement, τ_p is the station-dependent apparent relaxation time, and H is the Heaviside function. For stations with enough preseismic records, the parameters U_0 , V_0 , a , D_c , D_p ,

Figure 2. (a and b) show accumulated horizontal (arrows) and vertical (colored dots) postseismic displacements in two observation periods (2015.31–2017.31 and 2017.31–2019.31). (c–e) represent postseismic displacements in line of sight (LOS) directions from Wang and Fialko (2018). Contours show projected coseismic slip distribution of the main shock and the largest aftershock as shown in Figure 1.

and τ_p are obtained from a nonlinear regression of the observed GPS time series. Coseismic displacements induced by the Mw 7.4 aftershock are calculated using the slip model obtained in Section 3.1, which are removed from the original displacement time series. For stations that only have postseismic records, we use interpolation to estimate the interseismic velocity (V_0) (Shen et al., 2015). The inferred secular velocities are shown in Figure S1. We finally get the postseismic signals by subtracting the long-term linear trend, seasonal variations and coseismic offsets from original position time series (Figure S2).

Figures 2a and 2b show the accumulated postseismic GPS displacements in two observation periods (April 25, 2015–April 24, 2017 and April 25, 2017–April 24, 2019). Similar to the coseismic deformation pattern (e.g., Elliott et al., 2016), postseismic deformation is dominated by southward movements, consistent with the tectonic context of this region. Negligible deformation is captured to the south of the coseismic rupture, implying that the driving sources of the observed postseismic deformation are mainly located at the downdip zones of the rupture. Besides, the rate of the postseismic displacements decays rapidly with time (Figures 2 and S2).

In addition, Wang and Fialko (2018) processed the Sentinel-1 SAR images using the persistent scatter approach, and obtained the postseismic displacements in line of sight (LOS) directions on three tracks (Figures 2c–2e). Detailed processing approaches of the InSAR data can be found in Wang and Fialko (2018). In this study, we will use the InSAR data as an independent observation to validate the postseismic deformation models inferred from the GPS observations.

Figure 3. (a) Structure and mesh configuration of the finite element model used for postseismic viscoelastic relaxation simulations. (b) Fault geometry and detailed setting on a 2D profile (A-A').

3. Modeling of Postseismic Deformation Following the 2015 Gorkha Earthquake

Postseismic deformation following large earthquakes has been explained by afterslip and viscoelastic relaxation (Bürgmann & Dresen, 2008; Diao et al., 2018). Besides, poroelastic rebound of crustal rock can also induce postseismic deformation in near-field areas (Jónsson et al., 2003), which is often ignored due to lack of necessary observations. In most cases, afterslip and viscoelastic relaxation can explain measured postseismic deformation successfully. However, because of their similar deformation patterns, it remains difficult to distinguish these two mechanisms (e.g., Diao et al., 2018). In this section, we first investigate how the observed postseismic deformation can be explained by afterslip and viscoelastic relaxation effect. Then, we evaluate the relative contribution of the two mechanisms based on a combined model. All model parameters are optimized to best fit the GPS data, and then validated using the independent InSAR data.

3.1. Postseismic Viscoelastic Relaxation With a More Reliable Finite Element Model (FEM)

We simulate the viscoelastic relaxation effect by considering a more reliable model configuration, such as the “ramp-flat-ramp” fault geometry (Elliott et al., 2016), laterally variable rheological structure, and the elastic subducting slab. In this study, we model the viscoelastic relaxation using the finite element modeling software—Pylith (Aagaard et al., 2013). Note that, to obtain an end-member estimation of the viscoelastic relaxation effect, we first neglect other mechanisms (e.g., afterslip and poroelastic rebound) that possibly contribute to the observed postseismic deformation, too.

In the FEM, we build a layered structure for each side of the fault with material parameters inferred from seismological studies (Laske et al., 2013; Monsalve et al., 2006). Southern Tibet consists of an elastic upper crust and two underlying viscoelastic layers representing the lower crust and upper mantle, respectively. Beneath the Indian plate, we design one viscoelastic layer representing the upper mantle (Figure 3). We use the Maxwell solid to represent all viscoelastic layers (Diao et al., 2018; A. M. Freed et al., 2017), and set the thickness of the elastic upper crust to 25 km for southern Tibet and 40 km for Indian plate, respectively (Zhao et al., 2017). We configure the extension of India slab based on seismic tomographic imaging results (Duputel et al., 2016; Kosarev et al., 1999; Li et al., 2008; Tilmann et al., 2003). We use seismic velocities (V_p and V_s) and rock density (ρ) to represent the elastic behavior of the assigned layers in the FEM (Figure 3). The lateral boundaries and base of the designed domain are fixed to have zero displacement in horizontal direction and vertical direction, respectively. To suppress artifacts caused by the assigned fixed boundary condition, the FEM extends for 2,000, 1,500, and 800 km in the north, west, and depth directions, respectively (Figure 3a). The mesh size increases smoothly from ~ 1.5 km near the fault to ~ 35 km in far-field areas (Figure 3b).

In the viscoelastic relaxation model, coseismic slips of the Mw 7.8 main shock and the Mw 7.4 aftershock are imposed as the driving source. For internal self-consistency of the fault geometric parameters in modeling the co and postseismic deformation (A. M. Freed et al., 2017), we re-estimate the coseismic slip models by inverting the geodetic data shown in Diao et al. (2015). A constrained least squares method (R. Wang et al., 2013) is used to solve the slip on fault patches. To make the inversion stable, we employ an *a-priori* smoothing constraint on slips of adjacent patches. To balance the model roughness and data misfit, we select the optimal smoothing factor based on the “trade-off” curves shown in Figure S3. The elastic Green’s function matrix is calculated based on an averaged layered earth structure (Laske et al., 2013; Monsalve et al., 2006). The effect caused by neglecting the lateral inhomogeneous earth structure is discussed in Section 3.2. The “ramp-flat-ramp” fault geometry (Elliott et al., 2016) is used for the coseismic slip model inversion (Figure 3b), which are divided into fault patches with a uniform size of 5×5 km. The obtained coseismic slip distribution of the main shock and the Mw 7.4 aftershock (Figure S3) agrees well with previous slip models (e.g., Diao et al., 2015; Elliott et al., 2016).

To validate the accuracy of the results predicted by the FEM, we build a layered earth model and compare the simulated co and postseismic displacements with that calculated based on semi-analytic solutions (PSGRN/PSCMP, R. Wang et al., 2006). In both models, a laterally homogeneous layered structure with a 25-km thick elastic layer and viscoelastic half-space of Maxwell solid ($\eta = 2.5 \times 10^{18}$ Pa·s) are used. As shown in Figure S4, the simulated displacements by using Pylith and PSGRN/PSCMP are in an excellent agreement, which indicate that our FEM simulation is sufficiently accurate.

After the elastic parameters, fault geometry and coseismic slip have been specified, the simulated deformation is controlled by the viscosity of the ductile layers. The postseismic GPS displacement time series are modeled to investigate the model ability of interpreting the time-dependent observations. The optimal value of the viscosity can be estimated by minimizing the data misfit between observations (U) and model predictions (D),

$$F(\eta_{CT}, \eta_{MI}) = \sum_{i=1}^M \sum_{j=1}^{N_i} \sigma_{ij}^{-2} [U_{ij} - D_{ij}(\eta_{CT}, \eta_{MI})]^2 \quad (2)$$

where i is the index of the GPS station, j is the time samples of the observed displacement time series and σ is the observation error of the GPS displacements. η_{CT} is the viscosity of the lower crust beneath the southern Tibet. η_{MI} is the viscosity of the upper mantle underlying the Indian plate. Initial tests suggest that the GPS displacements are not sensitive to the viscosity of the upper mantle beneath southern Tibetan (η_{MT}). We thus do not include η_{MT} as a free parameter in Equation 2, but assume it to be one order of magnitude higher than η_{CT} to reduce the number of unknown model parameters (Zhao et al., 2017).

As shown in Figure 4a, the obtained optimal η_{CT} of 1.5×10^{18} Pa·s is at least one order of magnitude lower than the obtained optimal η_{MI} , suggesting a significant variation of rheological behavior across the fault. Due to the lack of observations to the south of the MHT, only the lower boundary of η_{MI} can be constrained well (Figure 4a). The predicted deformation by the optimal model shows dominant southward movements as revealed by the observations (Figures 4b and 4c). The weighted root mean square (WRMS) residual is ~ 8.3 mm for fitting of displacements in two periods, that is, April 25, 2015–April 24, 2017 and April 25, 2017–April 24, 2019. The optimal viscoelastic relaxation model underestimates postseismic displacements in the early period, but overestimates the displacements in the later period, indicating the limitation of this unitary-mechanism model. Based on the optimal viscoelastic relaxation model, we predict the postseismic displacements in the LOS direction and compare them with the InSAR data. As shown in Figure 5, the viscoelastic relaxation model can capture the main deformation pattern of the InSAR observations, but it underestimates the near-field LOS displacements at the lower edge of the coseismic rupture.

To test the effect induced by the lateral inhomogeneity of the viscosity, we make more comparisons for different combinations of η_{CT} and η_{MI} . As shown in Figure 6, there are northward displacements to the south of the coseismic rupture if using a similar value of η_{MI} and η_{CT} ($\sim 10^{18}$ Pa·s). However, these northward motions diminish, when η_{MI} increases to $\sim 10^{19}$ Pa·s (one order of magnitude larger than η_{CT}), suggesting that lateral variation of viscosity plays an important role in the pattern of simulated deformation. In addition, further simulations are done to test the effect induced by neglecting the subducting slab. As shown in Figure 7, the

Figure 4. Comparison between the observed and predicted postseismic displacements by the optimal viscoelastic relaxation model. (a) Misfit variance as a function of the lower crustal viscosity of the Tibetan Plateau (η_{CT}) and upper mantle viscosity of the India plate (η_{MI}) in the viscoelastic relaxation model. The red star indicates the parameter choice used in the comparisons with data here. (b) and (c) show the comparisons of postseismic displacements in the two periods, respectively. The contours indicate the projected coseismic slip distributions as shown in Figure 1.

simulated horizontal and vertical displacements are significantly biased by neglecting the elastic slab. To reduce the number of unknown model parameters, we fix η_{MT} and the thickness of the slab. To evaluate the potential effects of these two parameters, we carry out two additional tests. For the first test, we fix η_{MI} and η_{CT} at 1×10^{20} Pa·s and 1.5×10^{18} Pa·s, respectively, and simulate the viscoelastic relaxation effect by varying η_{MT} from 1.0×10^{19} Pa·s to 1.0×10^{21} Pa·s. As shown in Figure S5, the predicted deformation remains stable across the fault, indicating that the viscoelastic relaxation of the upper mantle beneath southern Tibet plays

Figure 5. Comparisons between (a–c) observed postseismic InSAR displacements and (d–f) model predictions of the viscoelastic relaxation model. The contours show the projected coseismic slip distributions of the main shock and the Mw 7.4 aftershock as shown in Figure 1.

Figure 6. Ground deformation due to coseismic viscoelastic relaxation in the first two years obtained with variable combination of (η_{CT}, η_{MI}) . (a): $(1.5 \times 10^{18} \text{ Pa}\cdot\text{s}, 5 \times 10^{18} \text{ Pa}\cdot\text{s})$, (b): $(1.5 \times 10^{18} \text{ Pa}\cdot\text{s}, 1.0 \times 10^{19} \text{ Pa}\cdot\text{s})$, (c): $(1.5 \times 10^{18} \text{ Pa}\cdot\text{s}, 5.0 \times 10^{19} \text{ Pa}\cdot\text{s})$ and (d): $(1.5 \times 10^{18} \text{ Pa}\cdot\text{s}, 5.0 \times 10^{20} \text{ Pa}\cdot\text{s})$. η_{CT} and η_{MI} represent viscosities of the lower crust beneath southern Tibet and the upper mantle beneath India plate.

a very minor role. This is because the postseismic deformation is mainly caused by the coseismic stress change, which is concentrated in the lower crust and negligible in the upper mantle (Diao et al., 2015). In the other test, we vary the slab thickness from 30 to 50 km and check the response of the simulated deformation. As shown in Figure S6, only slight change of the deformation ($<3\%$) occurs in the near-field.

Previous studies indicate that the viscosity may reveal to be stress-dependent and increase as the coseismic stress change decreasing from the source area (e.g., A. M. Freed & Bürgmann, 2004; Peña et al., 2020). Here, we further carry out viscoelastic relaxation models with the power-law rheology using the Pylith software. For the viscoelastic relaxation models with the power-law rheology, the same fault geometry, coseismic slip model, mesh configuration, boundary conditions as that in the Maxwell models are used. The only difference is replacing the Maxwell rheology with the power-law rheology for the viscoelastic layers. The elastic property and dislocation creep parameters of the power-law rheology are shown in Figure 3b and Table S1, respectively. The thermal structure across the fault is inferred from Herman et al. (2010) and Jiménez-Munt et al. (2008). A reference strain rate of $0.5 \times 10^{-6} \text{ yr}^{-1}$ is used in the model. Because initial tests indicated that the simulated deformation is not sensitive to the parameter of the upper mantle, we use the parameters shown in Peña et al. (2020) for the upper mantle. We vary the parameters (i.e., preexponent A) of the lower crust beneath southern Tibet to make the model predictions comparable with the observations.

Simulated surface displacements based on the power-law rheology, and comparisons with that predicted by the Maxwell rheology and the observed displacements are shown in Figures S7 and S8, respectively. In the horizontal direction, the displacements predicted by the power-law rheology is similar with that predicted by the Maxwell rheology in the source area, but they decay faster with distance to the north of the fault rupture zone. In the vertical direction, the model with the power-law rheology predicts significant subsidence to the north of the rising region. However, such significant subsidence show clear contrast with the

Figure 7. Surface displacements in the first 2 years caused by viscoelastic relaxation due to coseismic slip. These displacements are calculated from finite element models by considering an elastic slab (a) and without considering the slab (b). (c and d) show SW and vertical displacements on the profile shown in (a and b). For different models, the same combination of ($\eta_{CT} = 1.5 \times 10^{18}$ Pa-s, $\eta_{MI} = 5.0 \times 10^{19}$ Pa-s) is used.

InSAR observation (Figure 2). Moreover, the model predictions of the power-law rheology show a worse fit to the observation than that of the Maxwell rheology, e.g., a larger discrepancy on the orientation of the displacement vectors and a smaller displacement magnitude at far-field areas than observed (Figure S8). We try different model parameters for the power-law rheology model (e.g., reference strain rate and stress exponent), but the pattern of the predicted deformation remains stable. Based on the obtained results, we consider that this stress-dependent rheology might be unsuitable for the present case, therefore use the Maxwell rheology in this study.

3.2. Kinematic Afterslip Model

Kinematic afterslip model has been widely used to explain postseismic deformation in early periods. This mechanism permits kinematic slip on fault patches of extended fault geometry, which is generally capable to explain the data, but the obtained afterslip distribution is not necessarily physically reasonable. Here, we invert the postseismic GPS displacements in the first 2 years (i.e., April 25, 2015–April 24, 2017) to test whether this mechanism can explain data and the obtained afterslip distribution is reasonable. We extend the coseismic fault width along the downdip direction to capture possible deep afterslips. The same inversion method as demonstrated in Section 3.1 are used for kinematic afterslip inversions. The optimal smoothing factor is selected based on the “trade-off” curve between model roughness and data misfit.

We test two different afterslip models: One is constrained by both near- and far-field GPS data (KAF-1); The other one is constrained by near-field GPS data only (KAF-2). It has been found from previous studies that afterslip is more likely to occur around or between the coseismic slip asperities rather than on patches far from them because the enhanced stress mostly occurs adjacent to the coseismic slip (e.g., Hsu et al., 2006;

Figure 8. (a and b) show the inferred kinematic afterslip distribution and data fittings of KAF-1 and KAF-2. (c) is the “so-called” afterslip inferred from the synthetic displacements that predicted by the optimal viscoelastic relaxation model (Section 3.1). The gray circles represent aftershock seismicity as shown in Figure 1.

Scholz, 1998). With these two models, we may check whether the afterslip distribution is reasonable and how it depends on the observation.

The inferred afterslip models and corresponding data fitting are shown in Figure 8. A common and notable feature of the two kinematic afterslip models is the widely distributed afterslip at the downdip extension of the coseismic rupture. Compared with KAF-2, deep afterslip at the downdip edge is required in KAF-1 to fit the far-field displacements. This widespread deep afterslip is isolated and well beyond the depth of the aftershock seismicity, which seems physically unreasonable (Liu-Zeng et al., 2020). A reasonable interpretation is that the driving source of the postseismic deformation partly comes from the viscoelastic relaxation of the deep ductile zone and the apparent deep afterslip is an artifact of ignoring the viscoelastic relaxation effect. In principle, the viscoelastic relaxation will also affect the near-field observation and therefore can bias the afterslip estimation in regions adjacent to the coseismic rupture, too.

To further identify whether the afterslip mechanism was overused, we first calculate the first 2 years’ post-seismic displacements on these GPS sites based on the optimal viscoelastic relaxation model obtained in Section 3.1. Then, we get the “so-called” afterslip by inverting the synthetic postseismic displacement data (Figure 8c). It is interesting to find that both the spatial location and magnitude of the inferred deep afterslip agree well with that shown in KAF-1 and KAF-2. Therefore, we propose that the isolated deep afterslip is actually represent the location of the driving source proxy, where viscoelastic relaxation may play a major role. The LOS displacements calculated from the two kinematic afterslip models are compared with the observation. The inferred displacement residuals are larger than that of the viscoelastic model (Table S2), although so many patches of afterslip are involved as free parameters are involved. Besides, the kinematic afterslip models fail to predict the pattern of the observed postseismic InSAR displacements (Figure 9), suggesting that afterslip cannot be the only mechanism of the observed postseismic deformation.

Note that here we have used a layered earth structure to solve the kinematic afterslip and coseismic slip, whereas the cross-fault elastic inhomogeneity is ignored. To identify the effect, we forward calculate the coseismic displacements using the layered model and the lateral inhomogeneous model, and compare their difference. As shown in Figure S9, the calculated displacements agree very well in both near- and far-field areas, though small difference around 28°N (<5%) is observed. Such difference might be negligible in contrast to other effects, such as the determination of the smoothing factor and uncertainty of the fault geometry at depth. Therefore, considering the huge computation for generating Greens’ function with the FEM model, we use the semi-analytical solution with a layered earth model for simplicity.

Figure 9. Comparisons between (a–c) observed postseismic InSAR displacements and (d–f) model predictions of the KAF-1 model and (h–g) KAF-2. The contours show the projected coseismic slip distributions of the main shock and its largest aftershock with an interval of 2 m. In order to forward calculate the LOS displacements in different time periods, we first obtain the GPS displacements in the corresponding periods of the InSAR data based on Equation 1; then we invert the GPS displacements to get the afterslip in each period; at last, we forward calculate the LOS displacements in each period using the corresponding afterslip model.

3.3. Combined Model of Afterslip and Viscoelastic Relaxation

Considering that both viscoelastic relaxation and afterslip may explain part of the observations, neither of them can simply be ruled out. Actually, a consensus has been reached that a combined model of these two mechanisms should be considered for more reliable estimations of viscosity and afterslip (A. M. Freed et al., 2017; Huang et al., 2014; Tian et al., 2020). By using a new modeling approach (Diao et al., 2018), we combines afterslip and viscoelastic relaxation effects induced by coseismic slip and time-dependent afterslip in one viscoelastic model. In this approach, the spatial pattern of afterslip is correlated to the positive coseismic Coulomb failure stress (CFS) change, whereas the magnitude and temporal evolution of afterslip are to be solved together with the viscosity of ductile layers.

We first calculate the coseismic CFS change on fault patches of the MHT. The CFS change on each fault patch is represented by $\Delta\text{CFS} = \Delta\tau - \mu\Delta\sigma_n$, where $\Delta\tau$ and $\Delta\sigma_n$ are the change of shear stress and normal stress caused by coseismic rupture (Scholz, 1990), and μ is the effective frictional coefficient. Based on the optimal coseismic slip model obtained in Section 3.1, we calculate ΔCFS on each fault patch by using the PSGRN/PSCMP code (R. Wang et al., 2006). The geometric parameters of each patch are the same as that in the coseismic slip model. The rake angle is set to a uniform value of 85° based on the mean rake angle of the inferred coseismic slip vectors. A range values of μ (0.2–0.8) are tested, but the inferred ΔCFS remains stable (Figure S10). Therefore, we choose a uniform μ of 0.5 in this study. Note that the pattern of the stress-driven afterslip correlates to the calculated CFS change, which in turn depends on the inverted coseismic slip model. However, given the dense and high-precision geodetic constraints, inferred coseismic slip models show high consistency in terms of slip magnitude and pattern (Diao et al., 2015; Elliott et al., 2016, and many others), which decreases uncertainty associated with the coseismic model.

Based on the obtained ΔCFS on the fault, we assume the magnitude of afterslip linearly correlates with the positive coseismic CFS change with a scaling factor to be optimized. Given that the fault patches at the updip of the coseismic rupture have negligible afterslip (e.g., Mencin et al., 2016; Zhao et al., 2017), we only consider afterslip on patches deeper than the up edge of the coseismic rupture. The magnitude of afterslip is assumed linearly correlated to the positive CFS change in the form of,

$$\int_0^T \dot{s}(r_c, t') dt' = \alpha \cdot \max[\Delta\text{CFS}(r_c), 0] \quad (3)$$

where T is the observation period (4 years), r_c is the position vector on the rupture fault, and $\dot{s}$ represents the afterslip rate that decays exponentially with time, that is, $\dot{s}(r_c, t) = s_0(r_c) e^{-\frac{t}{\tau}}$, ($t > 0$), where τ is the characteristic decay time of afterslip. Therefore, α and τ are the two key parameters controlling the afterslip time history.

In the present case, we assume that T is large enough for the complete relaxation process of the stress-driven afterslip and Equation 3 can be rewritten to

$$\dot{s}_0(r_c) = \alpha \frac{C_p(r_c)}{\tau} \quad (4)$$

where $C_p(r_c) = \max[\Delta\text{CFS}(r_c), 0]$.

After configuring the magnitude and temporal evolution of afterslip, the postseismic displacements (D) simulated by the combined model can be expressed in the form,

$$D(r, t) = D_{\text{cv}}(r, t) + \alpha \sum_{r_c} \frac{C_p(r_c)}{\tau} \left[D_{\text{ae}}(r, r_c) \tau \left(1 - e^{-\frac{t}{\tau}} \right) + \int_0^t D_{\text{av}}(r, t - t', r_c) e^{-\frac{t'}{\tau}} dt' \right] \quad (5)$$

where r is the position of the GPS stations, D_{cv} represents the postseismic viscoelastic displacement induced by the coseismic slip, D_{ae} and D_{av} are the elastic and viscoelastic displacements Green's functions related to the afterslip. We calculate all these deformation components (D_{cv} , D_{ae} , and D_{av}) based on the same FEM shown in Section 3.1. To further reduce the degrees of freedom in the combined model, we fix η_{MI} to 1×10^{20} Pa·s based on the results obtained in Section 3.1. Finally, the predicted displacements depend on three parameters: the η_{CT} , scaling factor (α) and characteristic time (τ). Considering that the temporal evolution is required in the model, displacement time series rather than accumulated displacements on each GPS station are employed as the model constraint. We optimize these parameters by minimizing the misfit function via grid search,

$$F(\eta_{\text{CT}}, \alpha, \tau) = \sum_{i=1}^M \sum_{j=1}^{N_i} \sigma_{ij}^{-2} [U_{ij} - D_{ij}(\eta_{\text{CT}}, \alpha, \tau)]^2 \quad (6)$$

Figure 10. Search results of the combined postseismic model. (a–c) show 3D misfit variance on the (η_{CT}, α) , (η_{CT}, τ) , and (α, τ) transects, respectively, where η_{CT} is the lower crustal viscosity of the Tibetan Plateau, α is the scaling factors between the afterslip and the positive CFS change for the deep fault part. τ is the characteristic decay time of the afterslip. The red star marks the minimum of the misfit variance, which defines the location of the optimal parameters. (d and e) show comparisons between observed and predicted postseismic displacements in two periods. Colored fault patches draw afterslip distributions. The color dots on each station show the residuals between the observed and predicted vertical displacements. The gray dots represent the aftershock seismicity (Adhikari et al., 2015).

where i represents the indices of the GPS stations, j is the time samples of the observed displacement time series (a time interval of 0.1 year is used), σ is the standard observation error, U and D are the observed and predicted displacements, respectively. The optimal values of η_{CT} is found to be 2.5×10^{18} Pa·s, which is larger than that inferred from the pure viscoelastic relaxation model (Figure 10a). The optimal scaling factor is 0.18 m/MPa with an optimal characteristic decay time of 0.6 year (Figures 10a–10c). Consequently, with a maximum of ~ 0.8 m, the inferred afterslip is mostly distributed on fault patches adjacent to the downdip edge of the coseismic rupture (Figure 10d). The temporal evolution of the stress-driven afterslip suggests that up to 70–90% of the total afterslip was released in the first year (with $\tau = 0.6 \pm 0.2$ year, Figure S11). The seismic moment released by the afterslip corresponds to an Mw 7.0 earthquake, and is $\sim 5\%$ of that released by the Mw 7.8 main shock.

The combined model can explain the GPS displacements better than the pure viscoelastic model with a decreased WRMS residual of ~ 7.4 mm (Figures 10d and 10e). This model can also capture the temporal characteristics of postseismic displacements on most of the GPS stations (Figure S12). The data-fits on stations near the lower edge of the coseismic rupture are significantly improved by incorporating the stress-driven afterslip (Figure 10). The combined model fits the InSAR observations better than separated

viscoelastic relaxation model and kinematic afterslip model in terms of deformation pattern and fitting residual (Figures 5, 9, and 11; Table S2), highlighting the reliability of the combined model. Both viscoelastic relaxation and afterslip contribute significantly to observed LOS displacements in ~ 2 years following the Gorkha earthquake (Figures 11h–11m).

4. Discussion

4.1. Relative Contributions of Viscoelastic Relaxation and Afterslip

To explain the postseismic GPS and InSAR observations in the first 4 years following the 2015 Gorkha earthquake, we investigate separated and combined mechanisms of viscoelastic relaxation and afterslip. The unitary viscoelastic relaxation mechanism can capture the main pattern of the postseismic GPS displacements, but fails to model their magnitude. Kinematic afterslip models explain the GPS observations well, but the obtained deep afterslip is far beyond the coseismic rupture and aftershock seismicity, which is physically unreasonable. A combined mechanism that incorporates afterslip and viscoelastic relaxation can overcome the aforementioned limitations of separated mechanisms. The reliability of the combined model is further supported by the independent InSAR observations.

Based on the optimal combined model, we draw the spatiotemporal variation of the displacement induced by different mechanisms (Figure 12). In the first 1 year, afterslip plays a major role in near-to middle-field, and viscoelastic relaxation contributes comparably with afterslip in far-field areas (Figure 12a). During the second year, the contribution of viscoelastic relaxation increases, and becomes larger than that of the afterslip (Figure 12b). In later periods, the contribution of viscoelastic relaxation is dominant because the stress-driven afterslip was almost released completely (95%) in the first 2 years (Figures 12c and 12d). In the time domain (Figures 12e and 12f), the contribution due to afterslip is mostly accomplished in the initial 1 year, whereas the contribution from viscoelastic relaxation plays a dominant role in later periods. It should be noted that the contribution of viscoelastic relaxation due to afterslip is $\sim 5\%$ of that induced by coseismic rupture.

4.2. Stress-Driven Afterslip Model

To make the inferred afterslip physically reasonable, stress-driven afterslip models have been proposed based on the “rate-and-state” frictional law (Tian et al., 2020; K. Wang & Fialko, 2018; Zhao et al., 2017). With this approach, the generated stress-driven afterslip depends on fault frictional parameters (e.g., the reference slip velocity and rate-and-state parameter of the fault), which are hard to estimate with available observations. We note that Tian et al. (2020), Zhao et al. (2017), and Wang and Fialko (2018) used the same approach to calculate stress-driven afterslip model, but the inferred results are different. The stress-driven afterslip model shown in Tian et al. (2020) indicates significant deep afterslip as that in the kinematic afterslip models (Figure 8), but the stress-driven afterslip in Zhao et al. (2017) and Wang and Fialko (2018) is mainly located on patches adjacent to the coseismic rupture.

In this study, we obtain the stress-driven afterslip model using a simple approach. We assume the afterslip is linearly correlated with the positive CFS change, and solve its magnitude and temporal evolution from the observations, rather than assume empirical fault frictional parameters to the model. The validation by using independent InSAR data and good model-data fitting to the displacement time series suggest that the stress-driven afterslip obtained with the simple approach is perhaps more reliable. The inferred deep afterslip in previous studies may just indicate the location of the driving source where ductile deformation due to viscoelastic relaxation may play a major role (Liu-Zeng et al., 2020).

Note that we only consider the coseismic stress change for solving the stress-driven afterslip in the combined model, whereas the stress caused by the viscoelastic relaxation and early afterslip is ignored. Our previous study on the 2008 Mw 7.9 Wenchuan earthquake (Diao et al., 2018) revealed that the CFS change due to viscoelastic relaxation is less than 7% of that induced by the coseismic rupture. Here, we carry out additional calculation to test the effect of early afterslip on the distribution of ΔCFS . For simplicity, we use afterslip in the first 0.4 year ($\sim 50\%$ of the total afterslip shown in Figure 10) as the early afterslip to calculate

Figure 11. Comparisons between (a–c) observed postseismic InSAR displacements and (d–f) model predictions of the optimal combined model. (h–j) show the contribution of viscoelastic relaxation. (k–m) show contributions of the stress-driven afterslip.

Figure 12. Contributions of different deformation mechanisms to the observed postseismic surface displacement. (a–d) show postseismic displacements in SW direction along the profile shown in Figure 7, calculated from the optimal combined model with different periods. D_{cv} and D_{av} are the viscoelastic effect due to coseismic slip and afterslip, respectively. D_{ae} is the elastic effect induced by the stress-driven afterslip. (e and f) show different displacements components in the time domain on two selected stations (CHLM and DNC4).

the corresponding ΔCFS . As shown in Figure S13, the ΔCFS caused by early afterslip is relatively small in contrast to that caused by coseismic slip. Therefore, the coseismic CFS change should play a dominant role in controlling the spatial pattern of the afterslip, whereas the coupling effect induced by early afterslip and viscoelastic relaxation plays a secondary role.

4.3. Rheological Structure Beneath Southern Tibetan Plateau

The inferred steady-state lower crustal viscosity beneath the southern Tibet is about 2.5×10^{18} Pa·s, which is close to the value obtained by Tian et al. (2020), but is one order lower than that obtained by Zhao et al. (2017). The differences perhaps are induced by the employed viscoelastic model and/or the short observation period. For example, displacements predicted by the viscoelastic relaxation model in Wang and Fialko (2018) show clear difference with the observation, but the displacements inferred from a more reliable viscoelastic model in this study agree well with observations in middle-to far-fields. As the result shown in Section 3.1, the lateral viscosity inhomogeneity and elastic slab can highly effect the simulated postseismic deformation, which in turn bias the estimation of viscosity and afterslip in the combined model. Therefore, an accurate structure model is of great importance for simulating the postseismic viscoelastic relaxation.

To the north of the MHT, a weak middle/lower crust has been identified by various geophysical investigations. For example, magnetotelluric observations revealed a remarkable low resistivity, partial melting and low viscosity zone beneath the southern margin of the Tibet (Lemmonier et al., 1999; Unsworth et al., 2005). Moreover, the corresponding low seismic velocity anomalies (Nelson et al., 1996) and high geothermal flow (Herman et al., 2010) also indicate a ductile layer in this region. In addition, coupled thermal-mechanical models suggest that a low-viscosity crustal channel underlying the southern Tibet is essential for long-term tectonic evolution across the Himalaya range (Beaumont et al., 2001). The postseismic geodetic investigation in this study reveals a relatively low viscosity in the Tibetan crust, which agrees with these previous findings. A ductile lower crust with low viscosities (in the order of 10^{18} Pa·s) has been identified beneath the Tibetan Plateau (e.g., Diao et al., 2018; Huang et al., 2014; Ryder et al., 2014; Wen et al., 2012). This weak lower crust will affect the strength of the whole lithosphere with variable spatiotemporal scales, playing an essential role in dynamic processes and deformation of the Tibetan Plateau (Clark & Royden, 2000; Royden et al., 1997).

In this paper, we use Maxwell model to describe the viscoelastic behavior of the lower crust and upper mantle, and the effect of possible transient viscosity is not included. We test the performance of the power-law rheology for the viscoelastic relaxation model (Section 3.1), however, this complex rheology does not show a better data-fit compared with the simple Maxwell rheology. Given the limited observation and the complexity of the coupled mechanism, we use the Maxwell rheology in this study. Reasonable data fitting suggests that this simple rheological model works well for the present case. Further investigations considering a spatiotemporal variable behavior and its effect on inferred afterslip distribution are worth to be explored in future, especially for regions showing clear lateral inhomogeneity of rheological structure and/or following huge earthquakes ($>M 8$) (e.g., Peña et al., 2020). As shown by Peña et al. (2020), nonlinear rheology will enhance the contribution of viscoelastic relaxation for deep regions and therefore reduce the inferred apparent deep afterslip.

4.4. Postseismic Poroelastic Rebound Effect

Poroelastic rebound may also contribute to postseismic deformation following large earthquakes (e.g., Jónsson et al., 2003). For the 2015 Gorkha earthquake, we evaluate this effect with a simple method that is, widely used in previous studies (Diao et al., 2018; Huang et al., 2014; K. Wang & Fialko, 2018). In this method, the steady-state deformation due to poroelastic rebound can be inferred from coseismic displacements calculated with different Poisson's ratios under undrained and drained conditions in the permeable medium, respectively.

We set the Poisson's ratios for undrained and drained state to 0.28 and 0.25 for an uppermost 4-km-thick permeable layer (K. Wang & Fialko, 2018; Zhao et al., 2017). We apply the combined slip model of the main shock and the Mw 7.4 aftershock obtained in Section 3.1 for the poroelastic modeling. As shown in Figure S14, the simulated postseismic displacement due to poroelastic rebound effect is one order of magnitude smaller than the observations. Moreover, the predicted deformation indicates significant subsidence to the south of the coseismic rupture (Figure S14), where no clear vertical deformation was captured by GPS or InSAR observations (K. Wang & Fialko, 2018; Zhao et al., 2017). Furthermore, Zhao et al. (2017) obtained an even worse data-fit by incorporating the poroelastic rebound effect. With different parameter setting of the permeable layer (e.g., thickness), the simulated displacements can vary, but the magnitude remains at a similar level (Figure S14b). Therefore, we do not consider this mechanism in the combined model.

5. Conclusions

With geodetic observations in the first 4 years following the 2015 Gorkha earthquake, we investigate separated and combined mechanisms to probe afterslip on the central MHT and rheological parameter of surrounding rocks. With a more reliable FEM, viscoelastic relaxation model can fit the pattern of the observed postseismic deformation, but fails to capture the magnitude of the postseismic displacements. Kinematic afterslip model is capable of fitting the observation, but the obtained deep afterslip is physically unreasonable (Liu-Zeng et al., 2020). A combined model that incorporates stress-driven afterslip and viscoelastic

relaxations induced by the coseismic rupture and afterslip can overcome the limitations of separated mechanisms. Based on the combined model, we obtain the stress-driven afterslip near the downdip edge of the coseismic rupture, which is in contrast to the widespread deep afterslip obtained by previous studies. The total seismic moment released by afterslip is $\sim 5\%$ of that released by the Mw 7.8 main shock, corresponding to an Mw 7.0 earthquake. The obtained optimal viscosity of the lower crust beneath southern Tibet is 2.5×10^{18} Pa·s, whereas the viscosity beneath India is at least one order higher, suggesting significant lateral variation across the fault.

Data Availability Statement

The GPS position time series are obtained from Nevada Geodetic Laboratory (NGL, <http://geodesy.unr.edu/>) and Crustal Movement Observation Network of China (<http://www.cgps.ac.cn/>). The InSAR data is obtained from Wang & Fialko. (2018) (<https://igppweb.ucsd.edu/~fialko/data.html>).

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