Postseismic relaxation across the Central Nevada Seismic Belt

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[1] Two GPS geodetic surveys across the Basin and Range (BR) province of western North America have detected an anomalous compression east of the Central Nevada Seismic Belt (CNSB). The first network, installed by the U.S. Geological Survey along an east-west trending transect across the BR at approximately 40°N, consisted of 63 monuments observed campaign style in 1992, 1996, and 1998. The second network, installed by the California Institute of Technology, consisted of 50 continuously operating GPS receivers, 18 of which were roughly along the same transect as the campaign survey. We have used a viscoelastic postseismic relaxation model of the four largest earthquakes in the CNSB (on the Pleasant Valley fault in 1915, the Cedar Mountain fault in 1932, and on the Fairview Peak and Dixie Valley faults in 1954) to interpret the geodetic velocities observed across the CNSB. A model of postseismic relaxation with a lower crustal Maxwell viscosity of 5–50 × 10^{18} Pa s can explain the apparent compression regardless of the assumed mantle viscosity. The geodetic velocities that we determined after removing the postseismic contribution indicate that within current GPS detection limits the BR is stable east of the CNSB and is undergoing rapid right-lateral shear west of the CNSB.

INDEX TERMS: 1208 Geodesy and Gravity: Crustal movements—intraplate (8110); 1236 Geodesy and Gravity: Rheology of the lithosphere and mantle (8160); 1242 Geodesy and Gravity: Seismic deformations (7205); 8159 Tectonophysics: Rheology—crust and lithosphere;
KEYWORDS: postseismic relaxation, Basin and Range, Central Nevada Seismic Belt, Elsasser model


1. Introduction

[2] The Basin and Range (BR) province of western North America has long been recognized as an area of extension, situated between the relatively stable Colorado Plateau (CP) and NW moving Sierra Nevada (SN), with an extensional stress oriented roughly NW-SE [e.g., Zoback, 1989]. Seismicity reveals a strongly delineated belt in western Nevada, running approximately orthogonal to the direction of extension, called the Central Nevada Seismic Belt (CNSB; Figure 1). On the basis of trilateration surveys it appears that the CNSB currently is accommodating the majority of the deformation in the BR, with a principal extension striking N60°W [Savage et al., 1995].

[3] Two Global Positioning System (GPS) geodetic networks were installed and surveyed to assess the current deformation patterns across the Basin and Range (Figure 1). The U.S. Geological Survey (USGS) established a GPS network trending approximately east-west from the SN to the CP at approximately 40°N latitude [Thatcher et al., 1999]. The network consisted of 63 stations, which were occupied in 1992, 1996, and 1998, on two or more days. In 1996, 18 GPS sites were installed by the California Institute of Technology along essentially the same transect as the USGS network as part of the Basin and Range Geodetic Network (BARGEN), a 50-station, continuous GPS network [Bennett et al., 1999; Wernicke et al., 2000]. In this paper we refer to this subset of the BARGEN network as simply the BARGEN network. While the USGS data provided dense spatial coverage with sparse temporal resolution, the BARGEN network consisted of continual temporal coverage and more limited spatial coverage (Figure 1).

[4] Several discrepancies are apparent in the velocity solutions from these two sets of data (Figure 2a), which to a large extent might be explained by the use of different reference frames in processing. We rotated the USGS velocities 2.7 × 10^{-10} rad/yr about a pole at 61°E, 47°S to get the two velocity solutions more consistent (Figure 2b and Appendix A). One site, NEWP, in the BARGEN solution has a large velocity compared to the velocity of the USGS station collocated at NEWP, as well as compared to the USGS sites adjacent to NEWP.

[5] In the BARGEN solution, one site, identified as LEWI, has an anomalous velocity (Figure 2). The velocity determined at LEWI has a smaller magnitude than the velocities determined at the station to the east of LEWI,
indicating a compression between LEWI and the stations to the east. This apparent compression was interpreted by Wernicke et al. [2000] to be the result of postseismic relaxation resulting from an earthquake that occurred on a nearby fault in 1915. The compression is also apparent in the velocity field determined by the USGS, where stations immediately to the east of the CNSB are moving in a westerly direction at a slower rate than those stations farther to the east (Figure 2). Additionally, owing to the lower velocities of sites to the east of the CNSB, the GPS velocities have a relatively large increase in velocities across the CNSB, indicating a relatively large strain accumulation on the CNSB.

In this paper, we present an analysis of the hypothesis that postseismic relaxation has affected the observed velocities and an investigation of the strain accumulation across the CNSB in light of possible perturbations to the secular velocity field by postseismic deformation. We found that the specific viscosity structure proposed by Wernicke et al. [2000] does not satisfactorily model the observed velocity field, but their claim that in the BR postseismic signals are of the same order as the background signal, and thus relatively easy to characterize, is robust.

2. Recent Large Earthquakes in the CNSB

We considered four earthquakes that occurred in the vicinity of the CNSB in the past century with $M_s$ larger than about $10^{19}$ N m (Figure 1). In 1915, a $7.5M_s$ earthquake [Doser, 1988] ruptured the Pleasant Valley fault (PVF), breaking a 61 km long surface trace, striking on average N24°W [Wallace, 1977]. The earthquake was a normal event and had a maximum vertical surface offset ($v_{max}$) of 6.0 m [Wallace, 1977]. In 1932, a $7.2M_s$ right-lateral strike-slip earthquake occurred in the southern region of the CNSB [Abe, 1981], creating a north striking surface trace over 70 km long with maximum slip of 2.0 m [Bell et al., 1999]. In 1954, a $7.1M_w$ earthquake ruptured the Fairview Peak fault (FPF), followed 4 min later by a $6.8M_w$ on the Dixie Valley fault (DVF) [Doser, 1986]. The FPF event ruptured a 31.6-km-long surface trace with both right-lateral and normal components, with maximum surface offsets of 2.9 m and 3.8 m, respectively [Caskey et al., 1996]. The Dixie Valley fault (DVF) ruptured a 42-km-long trace with a vertical offset of $v_{max} = 2.8$ m [Caskey et al., 1996]. The FPF and the DVF on average strike N15°W and N17°W, respectively [Caskey et al., 1996]. Table 1 summarizes the observed parameters for these faults.

In addition to the main rupture of the FPF, four smaller sections of the Fairview Peak fault zone ruptured in 1954, with a cumulative moment release of less than a third of the main FPF event. Caskey et al. [1996] attributed these events to the FPF event. We discuss the implications of higher and lower moments below, although we do not consider these subevents explicitly in this paper. Nor do we consider two large events which occurred in 1954 west of the CNSB, the Rainbow Mountain and Stillwater events [e.g., Doser, 1986], each with a maximum moment release of one tenth of the larger events presented above.

3. Postseismic Model

Wernicke et al. [2000] hypothesized that an anomalous velocity determined at station LEWI (Figure 2) was the result of postseismic relaxation from the Pleasant Valley earthquake of 1915. Moreover, they noted that a geodetically observed compression to the east of the CNSB was likely a manifestation of postseismic relaxation, since this
area is void of thrust faults or other geologic features of compression. The model Wernicke et al. [2000] used to justify their hypothesis was a simple two-dimensional, analytic model of postseismic relaxation. In this model, the upper crust is assumed to be elastic, while the lower crust is modeled as a viscous channel overlaying a rigid substrate [Elsasser, 1969]. An earthquake is modeled as an instantaneous tensile displacement across a vertical dip-slip fault in the elastic upper crust, and since the model is 2-D, the fault is assumed to be infinite in length [e.g., Melosh, 1976]. The model predicted that with a lower crust viscosity ($\eta_{LC}$) of $5 \times 10^{19}$ Pa s and large tensile opening (4 m) there will be approximately 3–4 mm/yr fault perpendicular velocity at 60 km from the fault 100 years after the earthquake [Wernicke et al., 2000]. While the magnitudes of these velocities corresponded to the observed velocity deviations at LEWI, the proposed model is an oversimplification, because it ignores the elastic properties of the lower crust and mantle, the time-dependent rheology of the mantle, it ignores the finite length of the fault, and it ignores the symmetry of the postseismic velocities that are predicted by models of postseismic relaxation. The symmetry of the postseismic displacements was not discussed by Wernicke et al. [2000], and a fortuitous choice of secular displacement field would be required to obscure such a symmetry in the observed geodetic velocities.

[10] In this study, we modeled postseismic velocity fields from the four largest earthquakes in the northern CNSB during the last century. We assumed that postseismic relaxation is due entirely to viscoelastic flow in the lower crust and mantle. We modeled the postseismic relaxation using a finite element solution to a fully viscoelastic, 3-D model using the finite element code Adina (Adina R&D, Inc.), where an earthquake is modeled as an instantaneous displacement across an internal surface. We discretized our finite element models using 8-node linear hexahedra elements. We used three-layer models, composed of a 15-km-thick elastic top layer (upper crust), 15-km-thick Newtonian Maxwell viscoelastic middle layer (lower crust), and 870-km-thick Newtonian Maxwell viscoelastic bottom layer

Table 1. Surface Measurements of Fault Parameters of the Four Ruptures Considered in This Study

<table>
<thead>
<tr>
<th>Fault</th>
<th>PVF</th>
<th>CMF</th>
<th>FPF</th>
<th>DVF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length, km</td>
<td>61</td>
<td>75</td>
<td>32</td>
<td>42</td>
</tr>
<tr>
<td>Strike, deg</td>
<td>24</td>
<td>0</td>
<td>15</td>
<td>17</td>
</tr>
<tr>
<td>Slip, deg</td>
<td>60</td>
<td>90</td>
<td>50–70</td>
<td>30–50</td>
</tr>
<tr>
<td>$\mu_{min}$, m</td>
<td>6.0 (3.5)</td>
<td>—</td>
<td>3.8 (1.2)</td>
<td>2.8 (0.9)</td>
</tr>
<tr>
<td>$\mu_{max}$, m</td>
<td>—</td>
<td>2.0</td>
<td>2.9 (1.0)</td>
<td>—</td>
</tr>
<tr>
<td>$M_{max}^{\mu}$, $\times 10^{27}$ N m</td>
<td>18.8</td>
<td>4.3</td>
<td>8.6</td>
<td>9.0</td>
</tr>
</tbody>
</table>

*PVF [Wallace, 1977], CMF [Bell et al., 1999], and FPF and DVF [Caskey et al., 1996] are the Pleasant Valley, Cedar Mountain, and Fairview Peak and Dixie Valley faults, respectively. Strike is given clockwise from north. $v$ and $u$ are vertical and strike-slip displacements, respectively, and $M_{max}$ is the maximum moment release calculated from the given parameters. Moments are taken from Caskey et al. [1996], except CMF is from Bell et al. [1999].
(uppermost mantle; Figure 3). This crustal structure corresponds to the seismically determined gross structure in the BR [e.g., Catchins, 1992]. In these models, the elastic responses are those of a Poisson body. To describe the instantaneous elastic properties of the model, we took the shear modulus ($\mu_C$) of the upper two layers (crust) to be 32 GPa, while $\mu_M$ of the lower layer (mantle) was 72 GPa. The value of $\mu_M$ we used is a typical magnitude for the PREM uppermost mantle [Dziewonski and Anderson, 1981] and tomography has shown the BR to have little seismic velocity perturbations from PREM [Humphreys and Dueker, 1994]. To control the time-dependent rheology of the model, we allowed the viscosity of the lower crust ($\eta_C$) and mantle ($\eta_M$) to vary in the ranges $10^{16}$ to $10^{27}$ and $10^{17}$ to $10^{22}$ Pa s, respectively.

[11] Since we assume a Newtonian viscosity, these models are linear; we modeled each rupture event, as well as the strike-slip and normal components of each event, separately and added the velocity fields together to get the total postseismic field. To decrease computational size and time of the finite element solution, we utilized the symmetries about the center of the fault. The appropriate boundary conditions imposed on the symmetry plane of the finite element mesh for a normal fault model are that the fault normal and vertical directions are free and the fault parallel velocity is zero. To model a strike-slip fault, we used fixed boundary conditions in the fault normal and vertical direction and stress free in the fault parallel direction on the symmetry plane. We specified the boundaries parallel to the fault to be 900 km from the fault trace at the surface and we hold the boundaries fixed. We fixed the fault normal boundary 900 km from the fault tip and the bottom boundary at 900 km depth, while the surface of the model is stress free. We ignored gravity in these models. Using a model with an elastic layer overlaying a viscoelastic half-space, Rundle [1982] observed that the effects of gravity on vertical postseismic displacements is small at five Maxwell times ($\tau = \eta/\mu$) but large at 45$\tau$. However, Pollitz [1997] showed that by modifying Rundle’s model to include a rigid substrate forming a viscoelastic channel the same thickness as the overlying elastic layer, the effects of gravity on the horizontal postseismic displacements are still small at 45$\tau$.

In this study, we assumed that the change in gravitational force on the lower crust after normal faulting (as the two sides of the fault change elevation) is small compared to the stress perturbations on the lower ductile layer caused by the rupture.

[12] We assumed all faults to be planar and we modeled the Pleasant Valley and Cedar Mountain faults as 62 km long, the Dixie Valley fault as 42 km long, and the Fairview Peak fault as 32 km long. We set the dip of the PVF, DVF, and FPF faults to be 60° and the CMF to be 90°, consistent with those determined from surface mapping and focal mechanisms. We take the maximum observed surface slips to be a proxy for the partitioning of dip slip and strike slip and then scaled the model slip values in order for the model

Table 2. Fault Properties Used in the Models of Postseismic Relaxation

<table>
<thead>
<tr>
<th></th>
<th>PVF</th>
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<th>DVF</th>
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<tr>
<td>Length</td>
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</tr>
<tr>
<td>Strike, deg</td>
<td>25</td>
<td>0</td>
<td>15</td>
<td>16</td>
</tr>
<tr>
<td>Dip, deg</td>
<td>60</td>
<td>90</td>
<td>60</td>
<td>60</td>
</tr>
<tr>
<td>$v$, m</td>
<td>4.74</td>
<td>—</td>
<td>3.52</td>
<td>2.52</td>
</tr>
<tr>
<td>$u$, m</td>
<td>—</td>
<td>1.43</td>
<td>3.62</td>
<td>2.68</td>
</tr>
<tr>
<td>$M_0$, $\times 10^{19}$ N m</td>
<td>18.8</td>
<td>4.3</td>
<td>8.6</td>
<td>9.0</td>
</tr>
</tbody>
</table>

PVF, CMF, FPF, and DVF are the Pleasant Valley, Cedar Mountain, Fairview Peak, and Dixie Valley faults, respectively. Strike is given clockwise from north, $v$ and $u$ are vertical and horizontal throws, respectively, and $M_0$ is the model moment release.
4. Model Results

The postseismic relaxation velocity fields for the four modeled ruptures and a lower crustal viscosity ($\eta_{LC}$) of $10^{19}$ Pa s and mantle viscosity ($\eta_{ML}$) of $10^{20}$ Pa s are shown in Figure 4. The velocity field for the PVF is for 80–90 years after the rupture, the field for the CMF is 60–70 years after the rupture, and the fields for the FPF and DVF are for 40–50 years after the ruptures. We added the four models together and interpolated the fields between the surface finite element nodes using a linear interpolation to sample the field at the GPS station locations. We then subtracted the velocities of the postseismic model from the observed velocities (Figure 5).

When comparing models of nonsecular velocities to geodetically observed velocity fields, typically, a model of the secular field is removed from the observed field, and the residual velocity field is fit to the nonsecular model. The alternative is to assume a plausible nonsecular model (or a set of them) and to subtract the model velocity field from the observed velocity field, leaving a residual velocity field to be evaluated. In this study, we assumed that, to first order, the velocity field observed across the CNSB is composed of secular and nonsecular components, where the nonsecular component is the postseismic relaxation from large earthquakes in the CNSB. We then tested the hypothesis that the geodetically observed compression, and the associated high strain accumulation across the CNSB, is caused by postseismic relaxation.
As we know that earthquakes have occurred on the CNSB, there should be an extensional strain accumulation on the CNSB (i.e., a negative slope in geodetic velocities across the CNSB), even if very slight.

After we subtracted the postseismic velocities modeled with a \( \eta_{LC} \) of about \( 5 - 20 \times 10^{18} \) Pa·s and \( \eta_M = 10 \eta_{LC} \) from the observed velocities, the apparent compression was effectively removed from the velocity field, as can be seen in both the spline and the residual velocities (Figures 7 and 8). However, the compression is not decreased as much by removing the postseismic models with higher or lower viscosities (Figure 8). Additionally, the velocities to the west of the CNSB are decreased, resulting in a gentler increase in velocities from east to west across the CNSB.

As there are no conclusive arguments that the lower crust is rheologically weaker than the mantle in the Basin and Range, we have explored models with varying relative strengths of the lower crust and mantle. As suggested above, we use the maximum positive slope of the spline fit in order to characterize the amount that the compression to the east of the CNSB was removed by the postseismic model. The compression is removed when the maximum slope is small. We found that for a mantle stronger than the lower crust (\( \eta_M > \eta_{LC} \)), a \( \eta_{LC} \) of \( 5 - 20 \times 10^{18} \) Pa·s predicts

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Figure 5. (a) Total postseismic velocities from the four events at the geodetic stations, using \( \eta_{LC} = 10^{19} \) Pa·s and \( \eta_M = 10^{20} \) Pa·s. (b) Resulting geodetic velocities (black) after removing the postseismic velocities from the observed velocities (gray; Figure 2). Locations of the PVF, CMF, FPF, and DVF are shown for reference. Coordinate system is as in Figure 4.
postseismic relaxation which effectively removes the compression (Figures 9 and 10). If the mantle is weaker than the lower crust \( (\eta_M < \eta_{LC}) \), we found postseismic relaxation calculated from models with \( \eta_{LC} \) of \( 2 - 5 \times 10^{19} \) Pa s can account for the compression equally well. Models with \( \eta_M < \eta_{LC} \) consistently produced residual velocities with the maximum slope of the spline fits slightly larger than the lowest values determined for the cases when \( \eta_M > \eta_{LC} \) (Figure 9). However, when looking at the actual residuals, we could not discern a convincing difference between the amount of compression removed for the cases when \( \eta_M > \eta_{LC} \approx 10^{19} \) Pa s and \( \eta_M < \eta_{LC} \approx 5 \times 10^{19} \) Pa s (Figure 10). We found that the compression was removed for \( \eta_M = \eta_{LC} \approx 10^{19} \) Pa s as well (Figure 10). The residual velocities after removing the postseismic model with \( \eta_M = \eta_{LC} = 2 \times 10^{19} \) Pa s produced a spline fit with the lowest maximum slope (Figure 9); however, this particular model has essentially a zero slope across the CNSB, reflecting zero to very small strain accumulation across the CNSB, which leads us to prefer other models. In general, postseismic models with \( \eta_{LC} < 5 \times 10^{18} \) Pa s or \( \eta_{LC} > 5 \times 10^{19} \) Pa s do not remove the compression appreciably, regardless of the values of \( \eta_M \) (Figures 8, 9, and 10d).

BARGEN station LEWI still has a less anomalous azimuth after removing the postseismic relaxation velocity field (Figure 7). BARGEN station NEWP has a high-velocity residual, while the collocated USGS site has a slightly lower-velocity residual. Additionally, the residual velocity at USGS site B280, located closest to the surface trace of the FPF, is high (Figure 7). The velocity residual could be decreased by moving the centroid of the FPF earthquake to the east, such that the USGS site is farther to the west of the fault, so that the relaxation velocity at that point is negative instead of positive, resulting in a lower residual velocity. In fact, the focal mechanism hypocenter for the FPF event is to the east of the surface trace [Doser, 1986]. However, B280 is the only velocity which was noticeably improved by moving the earthquake centroid; therefore, we have kept the rupture model located at the surface rupture.

Only the postseismic velocities closest to the faults are sensitive to the details of the coseismic rupture geometry. The postseismic velocities away from the faults scale with the coseismic moment of the rupture and a slightly higher moment will require slightly higher viscosities and lower moment will require lower viscosities. For given fault
The largest effect of using higher or lower slip values is the viscosities deduced are within the range presented above. Using much lower slip values (about 20% of the slip values used (Table 2), the geometries, the moments scale with fault slip. Using slip values within ±20% of the slip values used (Table 2), the viscosities deduced are within the range presented above. The largest effect of using higher or lower slip values is the amount the velocities to the west and east of the CNSB are decreased and increased, respectively, after removing the postseismic model. Using much lower slip values (about 50% of the values used above), we could not account for the

Figure 7. Example east-west profile of residual velocity magnitudes and azimuths (black symbols) when post-seismic relaxation velocities modeled with $\eta_{LC} = 10^{19}$ Pa s and $\eta_{M} = 10^{20}$ Pa s are subtracted from the observed geodetic velocities (light gray symbols; Figure 6). BARGEN sites LEWI and NEWP and USGS site B280 are labeled, and thin, dashed line connects NEWP with the USGS station collocated at NEWP. Caption is as in Figure 6.

Figure 8. Splines fit to east-west profiles of velocity magnitudes for the original geodetic velocities and after removing postseismic relaxation velocities from models with various values of $\eta_{LC}$ and $\eta_{M} = 10 \times \eta_{LC}$. 
when we decrease the dip to 30° field. Tests we have done on the dip of the fault show that changes the wavelength of the postseismic relaxation, with a noticeably from those presented above). The dip of the fault 45° postseismic deformation increases about three times. Using In the above models, we specified a dip of the DVF at 60°, which is about 10° higher than the dips determined from surface mapping [Caskey et al., 1996] and the event focal mechanism [Doser, 1986]. If we used a dip of the DVF fault of 45°, the inferred crustal viscosities needed to account for the observed compression in the geodetic velocities are slightly higher, but within the ranges given above. [20] The Maxwell times (τ) associated with these lower crust viscosities are in the range of about 5–50 years. The time from the earliest rupture to the time the geodetic observations were made is about 80–90 years, about 1.5–20τ. At these times, the effect of gravity is still expected to be small, even for a half-space [Rundle, 1982; Pollitz, 1997].

5. Conclusions

[21] In this study, we have tested the hypothesis that the geodetically observed compression east of the CNSB could be due to postseismic relaxation from the large earthquakes on the CNSB in the past century. We found that postseismic
Figure 10. Examples of east-west profiles of residual velocity magnitudes (black symbols) when postseismic relaxation velocities modeled with the specified \( \eta_{LC} \) and \( \eta_M \) are subtracted from the observed geodetic velocities (light gray symbols; Figure 6). Solid line is the spline fit to the residual velocities.

Relaxation predicted from models with lower crustal viscosities of \( 5-50 \times 10^{18} \) Pa s could account for the compression east of the CNSB, as well as decrease the velocities to the west of the CNSB, regardless of the viscosity of the mantle. For the case when \( \eta_M > \eta_{LC} \), we found \( \eta_{LC} \) to be about \( 5-20 \times 10^{18} \) Pa s, and for the case when \( \eta_M < \eta_{LC} \), \( \eta_{LC} \) was found to be about \( 20-50 \times 10^{18} \) Pa s. These values of \( \eta_{LC} \) give lower crustal Maxwell times of about 5–50 years. Previous researchers have argued that the lower crust in the Basin and Range is ductile based on temperature arguments [e.g., Lachenbruch and Sass, 1978] and structural arguments [e.g., Gans, 1987; Block and Royden, 1990; McKenzie et al., 2000]. Block and Royden [1990] constrained the lower crustal viscosity to be less than or equal to \( 10^{17} \) Pa s, while McKenzie et al. [2000] argued it is less than \( 6 \times 10^{19} \) Pa s, both consistent with the values presented here.

Wernicke et al. [2000] used the simple 2-D model of Elsasser [1967], which consists of an elastic upper crust overlying a viscous lower crust overlying a rigid mantle, to interpret the anomalous compression east of the CNSB as resulting from postseismic relaxation. They inferred a viscosity of the lower crust \( \eta_{LC} = 5 \times 10^{19} \) Pa s. In the 3-D viscoelastic models presented here, if we assume a high viscosity mantle, we require models with \( \eta_{LC} \) about 5 times lower than inferred by Wernicke et al. [2000]. It is of interest to determine how much of the discrepancy in the inferred \( \eta_{LC} \) is the result of our more realistic 3-D geometry, and how much results from our allowing the lower crust and upper mantle to be viscoelastic. To assess these effects, we calculated the postseismic relaxation velocities assuming a lower crustal viscosity of \( 10^{19} \) Pa s for a 3-D viscoelastic model, for a 2-D viscoelastic model, and for a 2-D viscous (Elsasser) model (Figure 11). The 3-D model is the model presented in Figure 10a; the 2-D viscoelastic model has the same viscoelastic structure as the 3-D model, with the CNSB approximated as an infinitely long normal fault (dipping 60° to the east, with vertical offset of 3.5 m). The 2-D Elsasser model with a purely viscous lower crust is similar to that used by Wernicke et al. [2000], except that we used a lower value of fault normal extension (2 m, corresponding to a 3.5 m vertical offset on a 60° dipping fault) and an elastic modulus of the upper crust consistent with the models presented here. Since the total postseismic velocity field is affected most by the Dixie Valley events (Figure 4), we calculated the postseismic relaxation velocities 40–50 years after the rupture in the 2-D models. The postseismic velocities predicted by the 2-D viscous model are about twice as large as those calculated in the 2-D viscoelastic model. The velocities from the 2-D viscoelastic model are also significantly larger than those calculated in the 3-D viscoelastic models, as well as lacking the variable azimuth. Thus, for a given viscosity and time after the rupture, the velocities the 2-D viscous model predict are too large because of both geometrical and rheological over-simplifications. Increasing the assumed lower crustal viscosity in the Elsasser model achieves the compensating effect of lowering the postseismic velocities.
slightly lower than those to the south, either an effect of the no preferred structure in the BR with a large strain accumulation, the geodetic velocities show that the region east of the CNSB (Figures 7 and 8). In the central BR, the velocities north of 40°N are slightly lower than those to the south, either an effect of the rotation of the BR relative to North America, or because the BR is wider in the north than the south. To the west, the strain rate increases substantially toward the SN block, with a higher strain rate between 39° and 40°N, where the distance between the CNSB and the SN is smaller than to the north. The region of increased strain rate correlates well with seismic activity (Figure 12) and is consistent with a model where the SN is moving to the NWN with respect to the BR, creating a large right lateral strain rate [Dixon et al., 2000].

[23] After we accounted for postseismic strain accumulation, the geodetic velocities show that the region east of the CNSB is relatively stable, moving to the NW at about 4–5 mm/yr, with relatively little strain accumulation across the CNSB (Figures 7 and 8). In the central BR, the geodetic velocities are fairly uniform, possibly indicating that there is no preferred structure in the BR with a large strain accumulation, a conclusion deduced by Wernicke et al. [2000] as well. East of the CNSB, the velocities north of 40°N are slightly lower than those to the south, either an effect of the

Figure 11. Postseismic relaxation velocities at the geodetic stations calculated using the three-dimensional model presented in this study with $\eta_{LC} = 10^{19}$ Pa s and $\eta_M = 10^{22}$ Pa s (thin black vectors), a 2-D viscoelastic model of the CNSB (light black vectors), and a 2-D viscous model (fat gray vectors). The 2-D models include normal throw on an approximation of the CNSB. The viscoelastic model has the viscoelastic structure of the 3-D model and the viscous model is composed of a viscous lower crust ($\eta_{LC} = 10^{19}$ Pa s) overlying a rigid mantle [Elsasser, 1969].

Figure 12. Cartoon representation of the secular velocity field (gray arrows) in the Basin and Range after removing the preferred models of postseismic relaxation from the geodetic velocities. Dotted lines indicate groupings of sites in the BARGEN and USGS geodetic networks with similar velocity characteristics, and dashed line indicates the approximate extent of the Central Nevada Seismic Belt (CNSB). Gray dots are the epicenters of seismicity from 1932 to 2000.
[24] If characteristically large earthquakes in the CNSB had a recurrence interval of 1000 years with horizontal offset of 3 m on a 60° dipping normal fault, the horizontal velocity increase across the CNSB would be about 1.7 mm/yr. This is reasonably consistent with the model that we determined by removing postseismic strains from geodetic velocities. Between 39° and 40°N, the velocities at sites to the east of the CNSB are about 4–5 mm/yr, while at sites directly to the west, the velocities are about 6 mm/yr (Figure 7). North of 40°N, the velocity increase is not as large across the CNSB, however, the seismicity near the PVF is lower and the repeat time is possibly as long as 10,000 years on the PVF [Wallace, 1977]. These simple arguments of expected strain accumulation only hold for fault systems that have a near regular recurrence interval. Wallace [1987] has argued that large earthquakes in much of the BR do not obey simple models of recurrence, as the large earthquakes tend to migrate onto different faults through time, while ruptures on a given fault cluster in time. The absence of a structure in the central BR apparent in the residual velocity field with large strain accumulation supports this hypothesis. A more detailed kinematic model of the residual velocity field found here needs to be investigated to determine slip rates on all of the potential structures in the BR. Additionally, the effects of postseismic relaxation from other larger earthquakes in the Wasatch Range (border of BR and CP) and the Walker Lane belt (border of BR and SN) need to be investigated.

[25] Without accounting for postseismic strain accumulation across the CNSB, researchers run the risk of misinterpreting the geodetic velocities in the vicinity of the CNSB. This is true in general when modeling geodetic observations across any active fault system, especially when the postseismic deformation is expected to be comparable in magnitude to the geodetic velocities.

Appendix A: Arbitrary Reference Frame Correction

[26] We solved for a reference frame correction to the USGS velocity solution for the USGS and BARGEN velocity solutions to be more consistent. We used a genetic algorithm direct search to find the optimum solution, searching for poles over the entire Earth with rotation rates of \(0–6 \times 10^{-10} \text{ rad/yr}\). We defined the “fitness” (\(\phi\)) of each solution as

\[
\phi = \sqrt{\phi_x^2 + \phi_y^2}
\]

where \(\phi_x\) and \(\phi_y\) are the fitness of velocity magnitude and azimuth, respectively. We took both fitness functions to be the weighted mean, accounting for site separation, of the residuals of either the velocity magnitude or azimuth of nearby sites. As an example, the velocity magnitude fitness function was

\[
\phi_v = \frac{\sum_{i \in B} \left( \frac{1}{\sigma_i} \sum_{i \neq j \in B} \frac{1}{\sigma_{ij}} \left( v_i - v_j \right)^2 \right)}{\sum_{i \in B} \frac{1}{\sigma_i}}.
\]

for \(r_{ij} < r_{\text{max}}\), where \(B\) and \(U\) are the set of BARGEN and USGS sites, respectively, \(x\) and \(y\) are variables for BARGEN and USGS sites, respectively, \(v\) is the magnitude of the geodetic velocity, \(r_{ij}\) is the distance between USGS and BARGEN sites, and \(r_{\text{max}}\) is a threshold distance. The scale factor \(\varepsilon\) is an arbitrary scale factor to avoid singularities when \(r_{ij} \to 0\). We computed the errors (\(\sigma\)) on the velocity magnitude and azimuth via standard error propagation, neglecting data correlations,

\[
\sigma_v^2 = \left( \frac{\partial f}{\partial v_i} - \sigma_i \right)^2 + \left( \frac{\partial f}{\partial v_j} - \sigma_j \right)^2
\]

where \(\xi_i = f(v_i, v_j)\) or \(\alpha = f(v_i, v_j)\). Using \(r_{\text{max}} = 0.2\), we found that rotating the USGS velocity field \(2.7 \times 10^{-10} \text{ rad/yr}\) about a pole located at 61°E, 47°S made the two velocity fields more consistent.

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