Fault slip rates, effects of elastic heterogeneity on geodetic data, and the strength of the lower crust in the Salton Trough region, southern California

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Received 19 November 2004; revised 11 March 2005; accepted 29 March 2005; published 2 September 2005.

[1] In southernmost California the Salton Trough sedimentary basin lies between the subparallel San Jacinto and San Andreas faults that together with the Elsinore fault, accommodate ~80% of the 50 mm/yr of Pacific–North American relative motion. The slip rates on the San Andreas and San Jacinto faults have been a matter of recent debate, and one source of uncertainty has been the influence of crustal strength heterogeneity due to elastically weak basin sediments on geodetic data. To evaluate these effects, we have modeled regional kinematics with elastic and viscoelastic finite element models that incorporate laterally and vertically varying crustal material properties. We find that in general, the effects of the sedimentary basin on surface deformation are relatively small and that to explain the geodetic velocity data, the slip rate on the San Andreas must be higher than that on the San Jacinto fault, consistent with traditional geologic estimates. We also conclude that a relatively strong lower crust of viscosity >5 x 10^19 Pa s exists in this region.


1. Introduction

[2] In the Salton Trough region (Figure 1) of southern California the subparallel San Jacinto, San Andreas and Elsinore faults together accommodate ~80% of the ~50 mm/yr of relative Pacific–North American motion [DeMets and Dixon, 1999]. A variety of geologic (e.g., paleoseismology, geomorphology) techniques have been used to estimate the long-term slip rates on these faults to assess the evolution of the plate-boundary fault system, regional kinematics and seismic hazard. “Traditional” estimates suggest that the San Andreas is slipping at approximately twice the rate of the San Jacinto, i.e., approximately 25 and 12 mm/yr respectively, and the Elsinore is far less active at ~4 mm/yr [Keller et al., 1982; Weldon and Sieh, 1985; Rockwell et al., 1990; Petersen and Wescousky, 1994; Humphreys and Weldon, 1994]. Reassessment of previous work [Dorsey, 2003], new geologic [Kendrick et al., 2002] and geodetic data [Johnson et al., 1994; Anderson et al., 2003] suggest a higher slip rate for the San Jacinto fault and a correspondingly slower San Andreas. Since geodetic data are instantaneous (in a geologic sense), a major goal of the community is to relate these geodetic data to long-term fault slip rates. This is often done through the use of simple elastic half-space [e.g., Savage and Burford, 1973; Feigl et al., 1993], or viscoelastic relaxation models [Savage and Prescott, 1978; Savage and Lisowski, 1998; Pollitz, 2001; Dixon et al., 2003].

[3] Interseismic strain is naturally sensitive to the heterogeneous elastic and viscous properties of the lithosphere, complicating interpretation of geodetic observations [e.g., Malservisi et al., 2001]. The Salton Trough, which lies between the San Jacinto and San Andreas faults, is a thick sedimentary basin where strong variations in seismic velocity [Kohler et al., 2003; Magistrale et al., 2000] imply large crustal elastic moduli variations, and high heat flow [Lachenbruch et al., 1985; Bonmer et al., 2003] suggests relatively low lower crust and upper mantle viscosity. Previous work in the Ventura Basin of southern California [Donnellan et al., 1993; Hager et al., 1999] has shown that an elastically weak sedimentary cover can obscure locking depth and slip rates estimates based on geodetic data. Thus lateral and vertical variations in elastic and viscous properties in the Salton Trough may be important in influencing the form of surface deformation [Le Pichon et al., 2005] and therefore our interpretation of geodetic data.

[4] In this paper we use two simple, kinematic earth models to estimate the contemporary slip rates of the San Andreas, San Jacinto and Elsinore faults, and the effects of variable elastic and viscous structure of the Salton Trough lithosphere. We have systematically modeled a number of elastic and viscoelastic crustal structures to determine the range of acceptable models. Our primary results are: (1) the effects of the elastically weak Salton Trough sedimentary basin on geodetic velocity are small, (2) models are most consistent with a higher San Andreas than San Jacinto fault...
Figure 1. (a) Map of study area shown in an oblique Mercator projection about the Pacific–North American Euler pole of DeMets and Dixon [1999]. Solid lines show fault traces (SAF, San Andreas fault; SJF, San Jacinto fault; ELS, Elsinore fault), and the stippled region represents the approximate location of the Salton Trough sedimentary basin as defined by the SVM3 [Kohler et al., 2003]. Symbols show geodetic station locations of the CMM3 (pluses) and the subset used in this study (circles for stations away from the SJF and triangles and open squares for stations near the northern and southern sections, respectively, of the SJF). Velocities are shown in a North America reference frame and with 95% confidence ellipses [Shen et al., 2001]. Note that the faults and the majority of the velocity vectors are subparallel to Pacific–North American relative motion direction, indicating that crustal motion in this region is dominantly fault-parallel. The star shows the approximate location of the 1968 Borrego Mountain earthquake. (b) Pacific–North American parallel surface velocity relative to North America. Error bars give 1σ uncertainty. Shaded bars show fault locations. As in Figure 1a, the San Jacinto fault “south” and “north” stations are shown with open squares and triangles, respectively.
slip rate, and (3) models require a fairly strong lower crust with minimum viscosity of $5 \times 10^{19}$ Pa s.

2. Geodetic Data and Shear Modulus Constraints

[5] We use two data sets provided by the Southern California Earthquake Center to constrain our models, the Crustal Motion Map version 3 (CMM3) [Shen et al., 2001] and the three-dimensional Seismic Velocity Model version 3 (SVM3) [Kohler et al., 2003; Magistrale et al., 2000]. CMM3 is a compilation of various types of geodetic data (e.g., campaign GPS, continuous GPS, electronic distance measurement) realized into a common North American reference frame, convenient for plate-boundary studies. The coseismic and post-seismic effects of recent southern California earthquakes (Landers, 1992; Northridge, 1994; Hector Mine, 1999) have been removed from the data as to best represent the interseismic velocity field [Shen et al., 2001].

[6] We have selected a subset of the CMM3 to model the Salton Trough region. This subset includes all the stations in a swath perpendicular to the relatively simple trace of the San Andreas fault excluding those north of the San Andreas fault, and northwest of the Mecca Hills, near the intersection of the Eureka Peak fault, whose motion may be affected by slip transfer from the southern San Andreas fault to the Eastern California Shear Zone [Meade and Hager, 2005]. Figure 1a shows that the dominant orientations of the faults and most of the chosen velocity vectors are nearly parallel to that predicted by the Pacific–North American Euler pole [Demets and Dixon, 1999], indicating fault normal motion in this region is small.

[7] Crustal deformation in the San Andreas fault system of southernmost California may include three-dimensional complexities though for simplicity we analyze these geodetic data with antiplane models (two-dimensional cross sections perpendicular to the faults, with motion in the third dimension, parallel to the faults). The orientation and sense of slip on faults reflects the relative motion of surrounding crustal blocks and while the southern San Andreas and San Jacinto faults are finite length and not perfectly straight, the nearly parallel orientation of the velocity vectors and faults (Figure 1) suggests 2.5 dimensions are adequate for modeling effectively instantaneous geodetic data. Three-dimensional effects such as the large-scale curvature of the faults and local change in strike of the San Bernardino segment of the San Andreas fault are important for longer timescale kinematic modeling [Humphreys and Weldon, 1994] but should not significantly affect our antiplane analysis.

[8] Shear strain rate is greatest over the San Andreas and San Jacinto faults (Figure 1b). This concentration of strain was recognized early in central California and modeled as fault slip below a locking depth in a uniform elastic half-space, which gives rise to the familiar “arctangent” velocity field near strike-slip faults [Savage and Burford, 1973]. The magnitude of strain rate over a fault is thought to be indicative of the slip rate, though the trade-off between locking depth and slip rate can sometimes make modeling geodetic data nonunique in this regard [e.g., Freymueller et al., 1999].

[9] Geodetic strain rate can also depend on the earthquake history of nearby faults. Near the San Jacinto fault the fault-parallel velocities of the southern stations (open squares) are generally faster/slower than the northern stations (triangles) on the western/eastern side of the fault (Figure 1). This is consistent with the pattern expected following a right-lateral earthquake rupture, and may reflect a transient process related to the spatially coincident 1968 Borrego Mountain event [Allen and Nordquist, 1972; Sanders, 1993].

[10] Geodetic velocities are nearly indistinguishable from stable North America beyond ~45 km east of the San Andreas fault, although the small (~1 mm/yr) component of velocity in the San Andreas parallel direction is likely due to rotation of the Colorado plateau associated with extension in the Rio Grande Rift and southern Basin and Range [Humphreys and Weldon, 1994]. To account for this, we subtract 1 mm/yr, the approximate velocity of eastern California (east of the San Andreas fault) relative to North America resolved in the San Andreas fault parallel direction, from the velocity field shown in Figure 1. This 1 mm/yr represents relative Pacific–North American relative motion that is not associated with strike-slip plate boundary deformation and should not be included in modeling so as to avoid overestimating fault slip rates.

[11] Crustal motion is nearly uniform at ~39 mm/yr beyond ~40 km west of the Elsinore fault although any strain accumulation on faults to the west of the data transect in Figure 1 (e.g., Rose Canyon, Newport-Inglewood) that we do not account for may cause a slight overestimate in slip rates. The low strain rate in the data (Figure 1b), however, suggests this problem is minimal. Thus we have a stable reference frame to the east of the San Andreas fault and low strain rate, coherent crustal movement west of the faults.

[12] The SVM3 provides estimates of seismic velocity ($V_p$, $V_s$) and density ($\rho$) at any point in the southern California lithosphere interpolated from a number of seismic velocity sources. We calculate space-varying shear modulus ($\mu$) from $V_s$ and $\rho$ (see Figure 2). Models that incorporate the Salton Trough as an elastically distinct volume are based on the out-of-basin (solid line in Figure 2) and in-basin (dashed) shear modulus versus depth profiles. For our purposes, we define the basin as 20 km wide and adjacent to the San Andreas fault. Below 10 km it appears that the weak basin sediments are compensated by the elastically strong lower crust (Figure 2) thought to result from intrusive magmatism at this landward extension of the Gulf of California oceanic spreading system [Elders et al., 1972; Lachenbruch et al., 1985].

3. Techniques

[13] Two methods have been widely used in evaluating southern California geodetic data. The model of fault slip in a uniform, elastic halfspace [Okada, 1985, 1992] has been used by many authors to invert geodetic data for long-term slip rates via a back-slip method that accounts for interseismic strain accumulation [e.g., Feigl et al., 1993; Bennett et al., 1996; McClusky et al., 2001; Becker et al., 2005; Meade and Hager, 2005]. This method implicitly assumes a high-viscosity lithosphere, i.e., effectively elastic, such that time-dependent viscous processes are not important on the timescales of the modeled deformation process. The apparent success of this method to match most of the geodetic
data in southern California [Meade and Hager, 2005] suggests it captures the essence of surface kinematics in this region.

[14] Surface deformation is also modeled using an elementary earthquake cycle on a fault embedded in an elastic layer overlying a Maxwell viscoelastic halfspace [Nur and Mavko, 1974; Savage and Prescott, 1978; Thatcher, 1983; Savage, 2000; Pollitz, 2001; Dixon et al., 2003]. The surface velocity near a fault predicted by this method depends on the thickness of the elastic layer, the viscosity structure of the viscoelastic substrate and time of observation within the earthquake cycle.

[15] While the analytic solutions to the elastic halfspace and viscoelastic models are useful, they are limited to a relatively few simple cases with uniform or layered rheologies. We therefore use the finite element method that can accommodate arbitrarily complex elastic and viscoelastic rheologies, faults, and kinematic and dynamic boundary conditions. The finite element method is a robust technique and has been widely used to evaluate regional kinematics [e.g., Williams and Richardson, 1991; Saucier and Humphreys, 1993]; the effects of variations in crustal strength and thickness [Hager et al., 1999; Schmatzle et al., 2003]; fault slip rates and effects of the earthquake cycle [e.g., Dixon et al., 2003; Malservisi et al., 2003]; and earthquake dynamics [e.g., Freed and Lin, 2002; Hearn, 2003]. Specifically, we use the finite element code TECTON [Melosh and Raefsky, 1980; Williams and Wadge, 2000], which is a linear element, three-dimensional code that incorporates both elastic and Maxwell viscoelastic rheology. Below we compare finite element results to analytic solutions to verify the accuracy of the numerical method.

[16] Faults are included in the finite element models in two ways. Slippery nodes [Melosh and Williams, 1989] allow surfaces to be free slip, i.e., zero shear stress, while maintaining the fault surface. Earthquakes are included with the split node technique [Melosh and Raefsky, 1981] that imposes an instantaneous relative offset across a fault surface. This technique is commonly used in finite element earthquake displacements and postseismic relaxation studies [e.g., Hearn, 2003]. The elastic and viscoelastic models, described in detail in the next section, use slippery and split nodes, respectively.

4. Models and Results

[17] In this study we include two sets of kinematic models: the elastic locking depth–deep slip model, and viscoelastic earthquake cycle model. Below each are discussed in terms of setup, specific calculations and results.


4.1.1. Introduction

[18] The locking depth–deep slip model is a kinematic description of crustal blocks sliding past on another accommodated by faults that slip continuously at the tectonic load rate below some locking depth, and are locked above. The locked portion of the fault releases the accumulated strain during great earthquakes. The locking depth is often assumed to be base of seismicity, approximately 15 km depth throughout much of southern California [Williams, 1996; Magistrale, 2002].

[19] Our version of this model is shown in Figure 3a. An elastic block with slippery-node faults below a specified locking depth is driven by holding the right side fixed, and moving the left side and the bottom nodes of each block into the page of Figure 3a. The thickness of the block effectively represents the depth at which full fault slip occurs because in this model the interseismic slip tapers from zero at the locking depth to full slip at the base of the block (Figure 3), in contrast to elastic halfspace models which slip uniformly below the locking depth. Seismic evidence suggests that major faults such as the SAF are approximately vertical and penetrate at least into the lower crust [e.g., Zhu, 2002] so we use a block thickness 30km, the approximate average Moho depth [Zhu and Kanamori, 2000] in this region. Blocks are driven at the bottom because we find that side-only driven models with reasonable locking depths distribute shear strain much too broadly to reasonably match the high strain rates near faults observed in the geodetic data.

[20] The relative velocities imposed at the base of the blocks reflect the slip rates of the block-bounding faults. In this way we are able to forward model the surface velocity resulting from a hypothetical set of slip rates, and invert for the slip rates that produce a velocity field that best matches the geodetic velocity data. Surface velocity resulting from a unit slip rate for each fault is shown in Figure 3a.

[21] To find the best fitting set of slip rates we adopt the standard least squares misfit minimization method by solving

\[
a = (G^TW^{-1}G)^{-1}G^TW^{-1}d,
\]
Figure 3. (a) Elastic locking depth model. Blocks are driven at the bottom to reflect the slip rates of the block-boundary faults. Surface velocity (solid line) is shown for 1 mm/yr on each fault (with equal locking depths of 14 km). The finite element mesh contains 3131 nodes and 3000 quadrilateral elements. ELS, Elsinore fault; SJF, San Jacinto fault; SAF, San Andreas fault. (b) Map view velocity profile for the uniform elasticity crust model showing data (dots with 1σ error bars), best fitting model (solid line) with slip rates as shown at the fault locations (shaded bars), and equal-slip model (ELS = 4 mm/yr, SJF = 18 mm/yr, and SAF = 18 mm/yr; dashed line). Model misfit $\chi^2$ for both cases is shown in the top right corner.

where $\mathbf{a}$ is a vector of slip rates, $W$ is a diagonal weighting matrix $W_{ii} = \sigma^2_i$, and $\sigma_i$ is the standard deviation of the geodetic velocities contained in vector $\mathbf{d}$. $G$ is a Green’s function matrix whose columns contain surface velocities caused by unit slip on each of the three faults calculated at data point coordinates. This formulation minimizes the misfit metric

$$
\chi^2 = \frac{1}{n} \sum_{i=1}^{n} \left( \frac{d\mathbf{v}_i - \mathbf{w}\mathbf{v}_i}{\sigma^2_i} \right)^2 / (n - p),
$$

where $d\mathbf{v}_i$ and $\mathbf{w}\mathbf{v}_i$ are data and model velocities, respectively, $n$ is number of data points ($n = 101$) and $p$ is the number of model parameters, $p = 3$ for these models.

These models are driven such that the bottom of each block moves at the prescribed rate independent of the assigned locking depth. Slipping faults between blocks concentrates strain above the faults, though the effect on surface velocity due to reasonable changes in locking depth are generally small. Therefore in calculating the Green’s function matrix $G$ in (1), we set the locking depths to the approximate average maximum depth of seismicity [Magistrale, 2002] near the faults, i.e., 16, 14, and 10 km for the Elsinore, San Jacinto and San Andreas faults respectively.

Uncertainties on the best fitting slip rates are estimated via a Monte Carlo technique that accounts for data and locking depth uncertainty. During each Monte Carlo sample, each data point is randomly sampled according to a normal distribution defined by its variance, a randomly sampled set of locking depth Green’s functions (in the range of ±2 km) is used to construct $G$, and the slip rates are found by equation (1). After 1000 samples the distribution of each slip rate is mapped out and we use their standard deviations (σ) as formal uncertainties. All the uncertainties reported in this paper are 1σ.

4.1.2. Results

Model results for a uniform elasticity crust are shown in Figure 3b where we compare the surface velocity for the best fitting slip rates of $2.7 \pm 0.7$, $15.2 \pm 0.9$ and $21.4 \pm 0.5$ for the Elsinore, San Jacinto and San Andreas faults, respectively, to a hypothesized equal-slip model which distributes slip rates equally on the San Andreas and San Jacinto faults. To demonstrate the effects of elastic heterogeneity on surface velocity and slip rate estimates, three models are presented in Figures 4–6 where we show the surface velocity for an elastically heterogeneous crust and the difference compared to the uniform elasticity model.

Figure 4 shows the effects of a weak volume extending to 10 km depth. This is a simple representation of a sedimentary basin adjacent to the San Andreas fault (see Figure 1). The effect of the weak volume is to increase the strain rate over the basin, and decrease the velocity near the San Andreas fault. However, the magnitude of this effect is small, $\leq 3$ mm/yr.

Figure 5 shows the effect of a weak volume underlain by a strong volume. This model is a simple representation of the Salton Trough crustal structure where the middle and lower crust is mostly mafic intrusive rocks and the upper crust is sedimentary rock that has been deposited as this area extended. Again the strain rate over the weak zone is increased, although the magnitude is smaller than the previous model (Figure 4) due the compensating effect of the strong zone at depth.

Figure 6 shows the effects of crustal shear modulus structure of Figure 2. This elastic structure generally increases in shear modulus with depth everywhere and varies laterally from within to outside the basin. As in Figures 4 and 5, the effect of the basin is an increase in the strain rate within the basin, a slight decrease in the velocity near to San Andreas fault, and a poorer fit to the geodetic data compared to the uniform elasticity model (Figure 3). We conclude from these models that the elasticity structure in the Salton Trough region has only minor influence on the geodetic velocity, because any “basin signal” is not obvious in the data and the elastically heterogeneous models consistently produce a poorer fit (higher $\chi^2$).
The Monte Carlo approach used in estimating uncertainties naturally accounts for covariance between the estimated model parameters. Figure 7 shows the negative covariance between the estimated San Andreas and San Jacinto fault slip rates with contours of $\chi^2_\nu$. The approximate 95% confidence region, based on an F ratio statistic [e.g., Dixon et al., 2002], is shown with the thick contour.

4.2. Viscoelastic Earthquake Cycle Model

4.2.1. Introduction

The second model considered consists of the elementary earthquake cycle [Savage and Prescott, 1978] on a vertical strike-slip fault cutting an elastic layer overlying a Maxwell viscoelastic halfspace. The time-dependent surface velocity depends on the thickness of the elastic layer $H$, the earthquake recurrence interval $T$ and the time since the last earthquake compared to the Maxwell relaxation time $\tau_m = \eta/m$ where $\eta$ and $\mu$ are the viscosity and shear modulus respectively. Pollitz [2001] derived modified earthquake cycle solutions with constant surface velocity at a finite distance from the fault, to simulate the earthquake cycle in a finite width shear zone. With this approach, the surface velocity of a deforming zone is a consequence of the constant velocity of the sides of the shear zone and periodic earthquakes on the fault followed by viscoelastic relaxation of the earthquake-caused stresses. The surface velocity is thus the sum of steady simple shear and periodic, time-decaying earthquake cycle perturbations, which make for arctangent-like velocity profiles near the faults.

The geodetic velocities ~45 km to the east of the San Andreas fault and ~45 km to the west of the Elsinore fault are nearly constant, indicating that, at least for the effectively instantaneous velocity field of the CMM3, a large fraction of Pacific–North American shear at this latitude is...
accommodated within a finite width zone. These geodetic data alone however are insufficient to discriminate between the infinite width [Savage and Prescott, 1978] and finite width [Pollitz, 2001] earthquake cycle models because both predict zero strain rate at sufficient distances away from faults and the form of interseismic deformation, and therefore interpretation of the geodetic velocity data, near to the fault depends on the model assumed. We follow the Pollitz [2001] finite width shear zone approach in our models (described below) for three reasons. First, the elementary earthquake cycle of Savage and Prescott [1978] is “driven” by periodic earthquakes on a fault in the elastic layer and relaxation of the viscoelastic substrate. While the source of the driving stress does not change the infinite width earthquake cycle solutions [Savage, 2000], the superimposed simple shear and earthquakes of the finite width model [Pollitz, 2001] is more intuitively appealing considering the largely passive role of fault rupture in accommodating far-field relative Pacific–North American motion. Second, the finite width model assumes a very high viscosity, i.e., effectively infinite relaxation time compared to earthquake repeat time, so that viscous relaxation in the viscoelastic substrate outside the deforming zone is negligible and the surface velocity is constant. This seems a reasonable assumption for our study region where we expect a lower viscosity Salton Trough lithosphere compared to the relatively nondeforming adjoining regions of the Peninsular Ranges block to the west and now inactive southern Basin and Range to the east [Pearthree et al., 1983]. Finally, the finite width model is more amenable to numerical modeling, necessary for addressing heterogeneous material properties.

The model setup and finite element mesh is shown in Figure 8. The previous set of elastic models show that the Elsinore fault plays a small role in the total slip budget, so for simplicity we have removed its surface velocity contribution by subtracting from the geodetic velocity the predicted surface velocity due to the Elsinore fault with a slip rate of 2.7 mm/yr, and model only the San Jacinto and San Andreas faults.

The gradients in heat flow and maximum depth of seismicity [Magistrale, 2002] in this region suggest lateral variations in elastic thickness of the seismogenic crust. The surface velocity in a viscoelastic earthquake cycle depends on the elastic thickness so we include this effect by tapering the elastic layer thickness according to the approximate maximum depth of seismicity from 15 km at the San Jacinto fault to 10 km at the San Andreas fault (Figure 8). The total crustal thickness is held constant at 30 km although the actual Moho depth varies by approximately ±5 km in this area. Our results do not strongly depend on the total thickness of the crust (30 km) because the earthquake-caused stresses and postseismic flow are highest in the region just below elastic layer and the viscosity there

Figure 6. Same as Figure 4 with elasticity structure according to Figure 2. The shear modulus structure of the stippled-shaded region follows the dashed line, and the unshaded region follows the solid line of Figure 2. The compensating effect of the strong volume at depth reduces the basin signal (Figure 6c) at the surface (compare to Figure 4).

Figure 7. Contours of model misfit $\chi^2$ with respect to estimated slip rates of the San Andreas and San Jacinto faults. Star shows best fit ($\chi^2 = 0.98$), and the thick line the approximate 95% confidence region.
controls the surface velocity. If the Moho is actually much shallower than the assumed 30 km, our “lower crust” viscosity conclusions are also applicable to the uppermost upper mantle. Furthermore, models with uniform thickness upper and lower crust (15 km each) produce generally similar results.

[33] The elastic layer 45 km east of the SAF (Figure 8) is held fixed to represent stable North America and the elastic layer 45 km west of the SJF is kinematically driven into the page. We find that models with shear zone half width much greater than the 65 km used here distribute strain too broadly to reasonably match the geodetic data and conform to the above observation that away from the faults the crust is not significantly straining.

[34] At periodic intervals $T$ an earthquake occurs on a fault with offset $T_v$, where $v$ is fault slip rate. The surface velocity is then calculated for a number of time steps following each earthquake. The calculations are linear with respect to slip rate for Newtonian rheology and we compute each fault response independently with unit slip rate and scale and sum these Green’s functions to get the total velocity field. Results are shown after a sufficient number of earthquakes such that the system is cycle invariant, i.e., startup elastic transients have decayed to negligible values [e.g., Hetland and Hager, 2003].

[35] The average recurrence interval ($T$) for the southern San Andreas fault is estimated at $\sim 215–260$ years [Fumal et al., 2002; Shifflett et al., 2002; Jackson et al., 1995], and $\sim 260$ years for the San Jacinto fault at Anza [Rockwell et al., 2003]. We use a 250 year recurrence interval for bothfaults and find that models with a possibly shorter San Jacinto repeat time do not significantly change our conclusions.

[36] The last surface rupture of the Coachella segment of the San Andreas is estimated to be in the late 1600s [Sieh and Williams, 1990; Shifflett et al., 2002], slightly over

Figure 8. Viscoelastic earthquake cycle model. The elastic upper crust (UC) is underlain by a variable thickness lower crust (LC) underlain by the effectively semi-infinite upper mantle (UM). The two strike slip faults (solid lines in upper crust) are separated by 40 km, the approximate distance between the San Jacinto and San Andreas faults. The model is driven by holding the shaded region in upper crust fixed to the right of the San Andreas fault and driving the shaded region in the upper crust left of the San Jacinto fault into the page. The bottom, right, and left sides are free. The mesh contains 1449 nodes and 1381 quadrilateral elements.

Figure 9. Comparison of a finite element model (solid line) and analytic solutions (shaded dashed line [Pollitz, 2001]) for a symmetric, finite width shear zone earthquake cycle. The right-lateral fault with unit slip rate cuts an elastic layer (15 km) which overlies a viscoelastic half-space of viscosity of (a) $1 \times 10^{19}$ Pa s and (b) $1 \times 10^{20}$ Pa s. The shear modulus throughout is 30 GPa. Surface velocity is shown for a number of times $t$ during the earthquake cycle as fractions of the earthquake repeat time $T$ (250 years). The half width of the shear zone is 65 km, and beyond 65 km the velocity is constant. The finite element results shown here were calculated with a mesh similar to that in Figure 8 except with a uniform 15 km elastic upper layer. The largest difference between the numerical and analytic results is early in the earthquake cycle and approximately 5% near the fault.
three hundred years ago. The relatively long elapsed time since the last rupture suggests that the fault is quite late in its earthquake cycle. The southern strands of the San Jacinto fault have seen a number of moderate magnitude historic earthquakes [e.g., Sanders, 1993] including the fairly recent Mw 6.7 Borrego Mountain rupture in 1968. To the north the Anza segment has been historically quiescent [Thatcher et al., 1975; Sanders and Kanamori, 1984] and is considered fairly late in its earthquake cycle [Rockwell et al., 2003]. We therefore approximate the San Jacinto as a single fault midway (50% with uncertainty of 30–70%) through its earthquake cycle.

4.2.2. Benchmark Test

[37] To confirm the accuracy of the finite element method we compare the model response of a single, right-lateral fault earthquake cycle to the analytic solutions of Pollitz [2001]. Figure 9 shows the numerical results are nearly identical to the analytic solutions with the largest difference <5%. Dixon et al. [2002] and Malservisi et al. [2001] made similar comparisons and came to similar conclusions.

4.2.3. Results

[38] We show three cases distinguished by different lower crust viscosities. In each, we model two heterogeneous crustal elasticity structures to demonstrate how the surface velocity depends on both the elastic and viscous structure of the lithosphere. In all cases, surface velocity is controlled by the viscosity of the lower crust so the upper mantle viscosity is fixed in these models at 5 × 10^19 Pa s. Unless otherwise noted, the shear modulus m is set to a typical crustal value of 30 GPa.

Figure 10. Viscoelastic earthquake cycle model results. (a) Cross section showing fault locations (thick vertical lines in upper crust), layered viscosity structure, and two different weak zones adjacent to the San Andreas fault (SAF). The shaded region is a weak zone with shear modulus half of the surroundings (15 and 30 GPa). This elastic structure is used for Figures 10c and 10e. The hachured region is assigned the in-basin elastic structure of Figure 2. Outside the hachured area is assigned the solid line of Figure 2. Below 30 km, shear modulus is constant, as elastic variation at that depth will not strongly affect the coseismic offset or postseismic flow. This elastic structure is used in Figures 10d and 10f. (b) Map view of velocity data (dots with 1σ uncertainty) and model results with no elastic weak region in the crust (solid line). Best fitting slip rates and uncertainties are shown for the San Jacinto and San Andreas faults. Model misfit χ² is given in the top right corner. (c) Weak zone model with the shaded weak elastic region (inset). (d) Weak zone model with the hachured weak elastic region (inset). (e) Effect of the shaded weak zone shown as the difference between the weak zone model and the uniform model, i.e., Figure 10c minus Figure 10b. (f) Effect of the hachured weak zone shown as the difference between the weak zone model and the uniform model, i.e., Figure 10d minus Figure 10b.
velocity profile is a straight line connecting the shear zone boundaries, i.e., simple shear. Figures 10 and 11 show that a minimum lower crust viscosity of $5 \times 10^{19}$ Pa s is needed to maintain sufficient arctangent-like signal to fit the data late in the San Andreas fault earthquake cycle. Because geodetic strain rate decays with time, and the last rupture of the San Andreas fault is much longer ago ($\approx 300$ years) than the 225 years (90% of 250) we use, our viscosity estimate of $5 \times 10^{19}$ Pa s provides a lower bound. In fact, Figure 12 shows models with a lower crustal viscosity of $1 \times 10^{20}$ Pa s better match the geodetic data. In all cases the effect of the weak elastic basin is to increase the strain rate within the basin and decrease the velocity near the San Andreas fault, though the effect is small, <3 mm/yr (Figures 10e, 10f, 11e, 11f, 12e, and 12f).

[40] The best fitting San Andreas and San Jacinto fault slip rates in Figures 10–12 are again found using equation (1). These viscoelastic models provide velocity as a function of time through the earthquake cycle and we find the best fitting slip rates by solving (1) with Green’s functions ($G$ in equation (1)) calculated with the San Andreas and San Jacinto faults 90 and 50%, respectively, of the way through their earthquake cycles.

[41] Uncertainties on best fitting slip rates are again estimated with a Monte Carlo approach, with uncertainty in earthquake recurrence of the San Jacinto fault included. During each Monte Carlo sample random noise is added to the geodetic data points according to their uncertainty, a Green’s function for the San Jacinto fault is calculated with randomly selected times between 30–70% through its earthquake cycles to reflect our uncertainty in recurrence interval and time of observation within the earthquake cycle (the San Andreas fault is fixed at 90% of the way through its cycle), and slip rates are calculated with (1). After 1000 samples, the slip rate distribution is mapped out and the standard deviation is reported as the formal uncertainty.

[42] To summarize, for lower crust viscosities $<5 \times 10^{19}$ Pa s the velocity profile shows little arctangent-like signal and poorly matches the data. Only models with lower crustal viscosity $\geq 5 \times 10^{19}$ Pa s (likely closer to $1 \times 10^{20}$ Pa s) can match the large strain rates near the San Andreas fault. In all cases the net effect of the elastically weak Salton Trough basin on the surface velocity is $\lesssim 3$ mm/yr. The best fitting San Andreas and San Jacinto fault slip rates for the model with a lower crust viscosity of $1 \times 10^{20}$ Pa s and with no elastic weak region are 23.1 ± 1.2 and 13.8 ± 1.4 mm/yr, respectively. A contour plot of model misfit (centered on 23.1 and 13.8 mm/yr with minimum $\chi^2$ of 0.88) showing the negative covariance between the estimated slip rates for these models would look very similar to that given in Figure 7. These slip rate estimates are robust with respect to chosen elastic and viscosity structure (compare Figures 10–12).

[43] The slight difference in estimated slip rates between the elastic and viscoelastic models can be understood in terms of the rupture history of the San Jacinto fault and the time dependence of the viscoelastic earthquake cycle model. The surface velocity due to creep in an elastic halfspace (or our elastic models, Figures 3b and 4–6) is approximately the same as the surface velocity due to a viscoelastic
earthquake cycle model at approximately 50% of the way through the last cycle (for $T = 250$ years, $\mu = 30$ GPa, and $\eta = 1 \times 10^{20}$ Pa s). Since the viscoelastic signal decays with time, and the San Andreas fault is well constrained to be late in its cycle ($\geq 50\%$), a higher slip rate in the viscoelastic model is necessary to increase both the arctangent-like signal (slope) around the San Andreas and the magnitude of velocity between the San Andreas and San Jacinto faults. To satisfy block motion constraints, a higher San Andreas rate requires a correspondingly lower San Jacinto rate \cite{Bennett et al., 2004}.

Furthermore, models with a shorter San Jacinto fault recurrence interval or models earlier in the San Jacinto fault’s cycle (or both) have the interesting effect of requiring a higher San Andreas slip rate. For example, being earlier in the San Jacinto earthquake cycle has the effect of increasing strain rate near the San Jacinto fault by accelerating the surface velocity to the west of the fault and depressing it to the east (e.g., see Figure 9). To match the magnitude of the velocities between the faults, the San Andreas rate must therefore be increased.

5. Discussion

5.1. Effect of the Salton Trough Sedimentary Basin

In both the elastic and viscoelastic models the effect of the relatively weak Salton Trough sedimentary basin is to increase the strain rate within the basin and decrease the surface velocity near the San Andreas Fault. The magnitude in all cases is $\leq 3$ mm/yr. Moreover, in all models this effect of the basin is to decrease the overall fit to the geodetic velocity data, which do not show any obvious signal of a change in strain rate near the San Andreas or the western edge of the basin as the models predict. Two explanations seem likely.

First, our models have oversimplified the actual geometry of the basin and crustal rheology. The actual shape of the basin may be such that the effect of the near-surface weak sediments smaller than we have predicted with the simple basin geometry. To some degree this is certainly true considering the width of the basin changes along strike of the San Andreas fault. Second, the stronger, presumably mafic, middle to lower crust plays a more important role in canceling the effects of the weak upper crust sedimentary rocks than we have modeled, possibly due to the simplicity of the SVM3 within the basin that may underestimate the seismic velocity of the lower crustal mafic rocks. In any case, the lack of obvious basin signal in the geodetic data implies the vertically integrated strength of basin crust is similar to crust outside the basin and the geodetic velocity in this strike-slip dominated region is more sensitive to this composite strength than any particular crustal component. Furthermore, the presence of the basin does not bias geodetic data significantly in a way that might explain the
discrepant slip rate estimates for the San Jacinto and San Andreas faults.

5.2. Fault Slip Rates

[57] The elastic and viscoelastic models consistently require a San Andreas fault slip rate greater than the San Jacinto fault, \( \sim 21 \) and 15 mm/yr for the elastic models and \( \sim 23 \) and 14 mm/yr for the viscoelastic models. These rates are consistent with the geologic estimates [e.g., Keller et al., 1982; Petersen and Wesnousky, 1994], evolution of the plate boundary fault system [Powell and Weldon, 1992], and elastic block models constrained by geodetic data [Bennett et al., 1996; Meade and Hager, 2005; Becker et al., 2005]. The higher San Jacinto rates suggested by Kendrick et al. [2002] and Dorsey [2003] are not necessarily inconsistent considering our results reflect contemporary deformation and the San Jacinto fault may have been more active in the past [Bennett et al., 2004; Morton and Matti, 1993]. However, if the southern California fault system is abandoning the San Andreas in favor of the younger and more favorably aligned San Jacinto fault, as to avoid the energy expense of mountain building in the San Gorgonio pass area [Morton and Matti, 1993], our results suggest this process is not clearly evident in the contemporary geodetic data.

[45] Johnson et al. [1994] and Anderson et al. [2003] analyzed geodetic data and found the shear strain rate near the San Jacinto fault to be similar to that of the San Andreas fault, suggesting similar slip rates. Our models slightly under fit the strain rate (i.e., spatial derivative of velocity profile) near to the San Jacinto fault. This cannot be remedied by a much higher slip rate on the San Jacinto (e.g., 18 mm/yr; Figure 3b), which would require a lesser San Andreas rate and cause misfit elsewhere, but can be explained in terms of earthquake related transient signals. As shown in Figure 1b, the fault-parallel surface velocities near the 1968 Borrego Mountain rupture are generally faster/slower on the western/eastern side of the San Jacinto fault, compared to stations to the north, consistent with the notion that the elevated strain rate is possibly a transient signal associated with the recent earthquake.

5.3. High Viscosity Lower Crust

[46] Both the elastic and viscoelastic models require a high-viscosity lower crust to match the geodetic data. The elastic halfspace deep slip method implicitly assumes a high viscosity, i.e., effectively elastic, rheology. The viscoelastic earthquake cycle models also require a lower crust viscosity of \( > 5 \times 10^{19} \) Pa s (and likely higher), to maintain large velocity gradients near the faults late in the earthquake cycle.

[50] Although the elastic halfspace and viscoelastic earthquake cycle surface velocity solutions converge in the high-viscosity limit, and cannot be differentiated based solely on geodetic data [e.g., Savage, 1990], we can eliminate the possibility that steady state creep on faults in a relatively low viscosity lower crust and upper mantle cause the observed high velocity gradients at the surface. Analytic solutions and finite element modeling show that geodetic velocities at the surface of an elastic layer overlying a linear viscoelastic halfspace depend only on transient and not secular velocities in the viscous substrate [Zatman, 2000; Hetland and Hager, 2004]. Thus steady state slip on a fault completely below the elastic layer (i.e., in a low-viscosity lower crust and upper mantle) is not observable at the surface. Therefore, to explain the high surface velocity gradients with deep creeping faults, the faults must be creeping in the elastic upper crust or the high viscosity (effectively elastic) lower crust or upper mantle.

[51] Heat flow in the southern Salton Trough region is very high [Lachenbruch et al., 1985; Bonner et al., 2003]. These high heat flow values, extensive hydrothermal activity and recent extrusive volcanism [e.g., Elders et al., 1972; Robinson et al., 1976] suggest high temperatures at depth. For example, Bonner et al. [2003] estimate 400°C at 10 km depth. Such high temperatures suggest a relatively low viscosity lower crust and upper mantle.

[53] However, our results require a high viscosity lower crust, apparently in contradiction to these thermal arguments. The lower crust here is expected to be mafic in composition, a result of intrusive volcanism and underplating at the spreading center [e.g., Lachenbruch et al., 1985], and hence compositionally stronger [Kohlstedt et al., 1995]. The lack of a significant gravity low over the low-density sedimentary basin rocks [Lachenbruch et al., 1985; Fuis et al., 1982] is consistent with a higher density, mafic, lower crust. Furthermore, since water and partial melt content play an important role in controlling viscosity of mafic rocks [e.g., Hirth and Kohlstedt, 1996; Karato, 1986], we infer that the lower crust may also be dry and has little partial melt content, similar to Lachenbruch et al.’s [1985] conclusion that the crust is mostly solid. Thus the high viscosity of the lower crust probably derives from its dry, mafic composition.

[55] For simplicity we have assumed a Newtonian viscosity lower crust and mantle although postseismic deformation studies of recent earthquakes in the Mojave Desert suggest that more complicated rheologies such as nonlinear [Pollitz et al., 2001; Freed and Bürgmann, 2004] or linear composite viscosities [Pollitz, 2003; Ivins and Sammis, 1996; Ivins, 1996] may describe southern California lithosphere. If true, this would not significantly change our earthquake cycle model conclusions; the lower crust would still have to be sufficiently high viscosity to retain high strain gradients late in the earthquake cycle. Assuming postseismic relaxation is responsible for the long-term arctangent signal, it appears the lower crust exhibits a wide spectrum of viscoelastic relaxation times and effective viscosities from rapid relaxation early after an earthquake [Freed and Bürgmann, 2004; Fialko, 2004] to slower relaxation late in the earthquake cycle (our viscoelastic models).

[56] Other studies of postseismic relaxation following recent strike-slip earthquakes [e.g., Pollitz, 2003], analysis of the state of stress near the seismic-aseismic transition of transform faults [Savage and Lachenbruch, 2003], subsidence due to surface loading of Lake Mead [Kaufmann and Amelung, 2000], coherent deformation of the upper and lower crust [Zhu, 2002] and high strain rates near major faults throughout southern California [Meade and Hager, 2005] also argue for a relatively high-viscosity lower crust. These and our results suggest that the lower crust is an important component in the vertically integrated
strength of the southern California lithosphere [e.g., Jackson, 2002].

6. Conclusions

[55] We have investigated the effects of variable crustal rheology on surface deformation in the Salton Trough region of southern California with two kinematic finite element models. In general we find that the effect on surface velocity caused by the presence of the Salton Trough sedimentary basin is small and the geodetic velocity data are most consistent with a greater San Andreas than San Jacinto fault slip rate, approximately 22.3 ±0.7 and 14.5 ±0.8 mm/yr (average of the two uniform properties models), respectively. We also find that in order to explain the high gradients seen in the geodetic velocity data, the lower crust must have a viscosity of >5 × 10^9 Pa s.

[56] Acknowledgments. This research was supported by the National Science Foundation award EAR-0106892 and the Southern California Earthquake Center. SCEC is funded by NSF cooperative agreement EAR-0106924 and USGS cooperative agreement 02HQAG00008. The SCEC contribution number for this paper is 866. The figures in this paper were created with GMT [Wessel and Smith, 1998]. Comments by Eric Hetland, Ray Weldon, Becky Dorsey, and Vicki Langenheim and reviews by Tim Dixon, an anonymous reviewer, and Associate Editor Kelvin Wang lead to significant improvements in the paper.

References


Jackson, J. (2002), Strength of the continental lithosphere: Time to abandon the jelly sandwich?, GSA Today, 12, 4–10.


