

Interseismic deformation and geologic evolution of the Death Valley Fault Zone

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Received 31 May 2011; revised 3 April 2012; accepted 20 April 2012; published 8 June 2012.

[1] The Death Valley Fault Zone (DVFZ), located in southeastern California, is an active fault system with an evolved pull-apart basin that has been deforming over the past 6 Myr. We present a study of the interseismic motion and long-term stress accumulation rates to better understand the nature of both past and present-day loading conditions of the DVFZ. Using a 3-D semi-analytic viscoelastic deformation model, combined with geodetic velocities derived from the Mobile Array of GPS for Nevada Transtension (MAGNET) network and the Southern California Earthquake Center (SCEC) Crustal Motion Map version 4 (CMMv4) GPS data, we establish parameters for interseismic slip rate and apparent locking depth for four DVFZ fault segments. Our preferred model provides good fit to the data (1.0 mm/yr and 1.5 mm/yr RMS misfit in the fault-perpendicular and fault-parallel directions, respectively) and yields apparent locking depths between 9.8–17.1 km and strike-slip rates of 3–7 mm/yr for the segments. We also determine subsidence (0.5–0.8 mm/yr) and extension (1.0–1.2 mm/yr) rates in the pull-apart basin region. With these parameters, we construct a DVFZ evolution model for the last 6 Myr that recreates the motion of the fault blocks involved in the formation of the present-day geological structures in Death Valley. Finally, using Coulomb stress accumulation rates derived from our model (0.25–0.49 MPa/100 yr), combined with earthquake recurrence interval estimates of 500 to 2600 years, we assess present-day seismic hazards with calculated moment magnitudes ranging from 6.7–7.7.

Citation: Del Pardo, C., B. R. Smith-Konter, L. F. Serpa, C. Kreemer, G. Blewitt, and W. C. Hammond (2012), Interseismic deformation and geologic evolution of the Death Valley Fault Zone, *J. Geophys. Res.*, 117, B06404, doi:10.1029/2011JB008552.

1. Introduction

[2] The Eastern California Shear Zone (ECSZ) is a tectonically active region of the Basin and Range province located to the east of the San Andreas Fault System (SAFS) (Figure 1). While the SAFS marks the major boundary between the Pacific and North American plates, roughly 20–25% of plate boundary motion is manifested by faults of the ECSZ [Dokka and Travis, 1990; Bennett *et al.*, 1997; McClusky *et al.*, 2001; Knott *et al.*, 2005; Hill and Blewitt, 2006]. Based on geological and geodetic studies, the ECSZ, including the Sierra Nevada microplate, produces between 10–14 mm/yr of displacement [Wernicke *et al.*, 1988; Bennett *et al.*, 1997; Dixon

et al., 2000; McClusky *et al.*, 2001; Bacon and Pezzopane, 2007] distributed along three right-lateral transtensional fault zones: the Death Valley Fault Zone (DVFZ), the Panamint Valley – Hunter Mountain – Saline Valley Fault Zone (PHSFZ) and the Owens Valley Fault Zone (OVFZ). The DVFZ is one of the longest (~310 km in length) and geologically fastest slipping fault systems in the Basin and Range province [Machette *et al.*, 2001].

[3] Of particular interest to this study, the DVFZ is composed mainly of right-lateral strike-slip faults and normal faults accommodating between 2 to 6 mm/yr of motion [Dixon *et al.*, 1995; Bennett *et al.*, 2003; Hill and Blewitt, 2006]. Three major fault sections make up the DVFZ (Figure 1): the Northern Death Valley Fault Zone (NDVFZ), the Black Mountains Fault Zone (BMFZ) and the Southern Death Valley Fault Zone (SDVFZ). The NDVFZ and the SDVFZ are part of a group of northwest trending, right-lateral strike-slip faults found in the southeastern Great Basin [Butler *et al.*, 1988; Dokka and Travis, 1990]. The BMFZ, also referred to as the Central Death Valley fault zone, is a normal and strike-slip fault system along with the NDVFZ and SDVFZ motion [Burchfiel and Stewart, 1966; Hill and Troxel, 1966].

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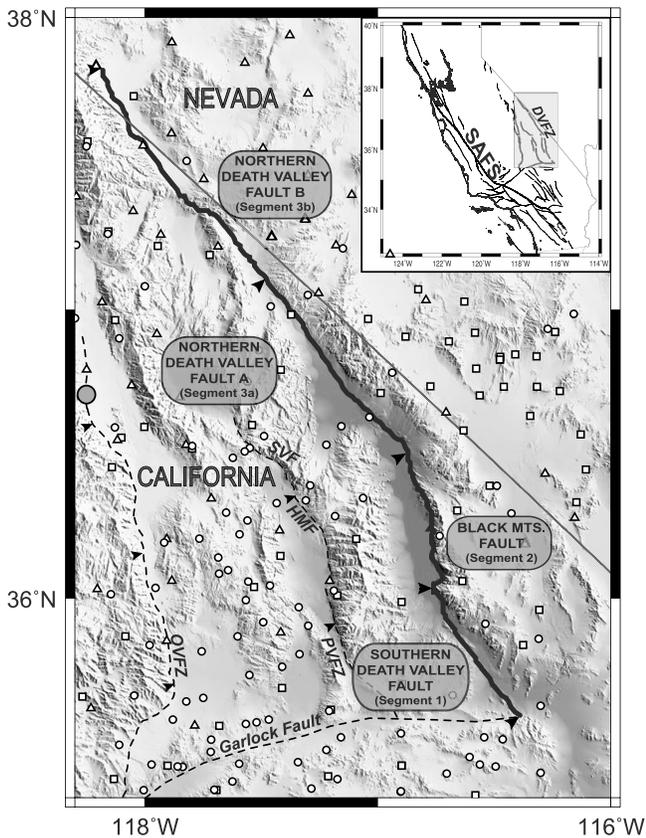


Figure 1. Map showing the location of the Death Valley Fault Zone (DVFZ) and fault segment names used in this study. Dark black line represents the DVFZ, with black arrowheads indicating our model fault segmentation. Dashed lines depict the location of the Garlock fault, Panamint Valley Fault Zone (PVFZ), Hunter Mountain Fault (HMF), Saline Valley Fault (SVF) and Owens Valley Fault Zone (OVFZ). The white circles represent the location of campaign SCEC CMMv4 sites, the white triangles mark the location of the semi-continuous UNR MAGNET GPS stations, and white squares are continuous GPS sites used in this study. Gray circle indicates the place where the 1872 Owens Valley earthquake occurred. Insert: Regional San Andreas Fault System (SAFS) in California. The gray shaded rectangle marks our study area within the Eastern California Shear Zone (ECSZ) in reference to the greater SAFS.

[4] The Death Valley region, located in southeastern California, has been the topic of many geologic and geophysical investigations aimed at understanding the processes involved in its formation and evolution. To reconstruct the Cenozoic history of the region, geological studies [e.g., Burchfiel and Stewart, 1966; Hill and Troxel, 1966; Wright and Troxel, 1966, 1967; Stewart, 1983; Troxel and Wright, 1987; Butler et al., 1988; Serpa and Pavlis, 1996; Knott et al., 1999; Wernicke, 1999] have interpreted fault block motion responsible for the present-day geological structures and deformation features. Geophysical analyses of gravity, magnetic, cosmogenic, and seismic reflection data have aided in the study of the evolution of the DVFZ region [e.g., Serpa et al., 1988; Keener et al., 1993; Blakely et al., 1999], revealing the upper crustal structure, fault block rotations,

and basin development. Several more focused studies on subsections of the DVFZ have also been conducted [e.g., Butler et al., 1988; Miller and Pavlis, 2005; Frankel et al., 2007a, 2007b], revealing the amount and direction of fault displacement, slip rates of the NDVFZ and SDVFZ, and the mechanisms of extension in Central Death Valley. Although there have been many targeted studies of specific Death Valley geologic structures and segments of the fault zone, thus far there has not been a comprehensive geodetic study focused on the entire DVFZ.

[5] The purpose of this investigation is to develop a three-dimensional (3-D) deformation model, constrained by GPS observations, to estimate apparent locking depth and slip rates of the primary fault segments of the DVFZ. From these parameters, we assess variations in stress, strain, and moment accumulation rate for the segments of the DVFZ and place these values in context with regional seismic hazards. We also construct an evolution model of Death Valley to study the formation of the pull-apart basin ~ 6 Ma. Our ultimate goal in this analysis is to utilize both geologic and geodetic data to understand the contemporary slip rate of the DVFZ and relate this to the Cenozoic history to reconstruct the processes involved in its evolution.

2. Kinematics and Geologic Observations of the Death Valley Fault Zone

[6] Based on previous studies, the two strike-slip fault systems in the DVFZ, the SDVFZ and the NDVFZ, appear to be closely related but have large differences in both net slip and modern slip rate [Davis and Burchfiel, 1973; Stewart, 1983; Butler et al., 1988]. Earlier geologic studies have estimated slip rates ranging 3–9 mm/yr [Klinger and Piety, 2000; Frankel et al., 2007a, 2007b; Willis et al., 2008], while geodetic analyses estimate rates spanning 2–8 mm/yr [Bennett et al., 1997; Dixon et al., 2000; McClusky et al., 2001] for the NDVFZ, suggesting that this fault zone likely accommodates most of the motion produced in the northern portion of the ECSZ. In contrast, the SDVFZ has a geologic slip rate between 3–5 mm/yr [Willis et al., 2008; Southern California Earthquake Center (SCEC), 2011a] and a geodetic slip rate of ~ 3 mm/yr [Gan et al., 2000; McClusky et al., 2001]. The BMFZ, which runs through the entire pull-apart basin region, has a geologic strike-slip rate estimated at ~ 4 mm/yr [Willis et al., 2008] and a geodetic slip rate of ~ 3 mm/yr [McClusky et al., 2001].

[7] One of the most characteristic features in the Death Valley area is the pull-apart basin in Central Death Valley, located between the Panamint Range and the Black Mountains blocks [Burchfiel and Stewart, 1966; Hill and Troxel, 1966]. The basin has a highly oblique geometry, a length of approximately 100 km and trends in the north-northwest direction. It has an elevation of approximately 80 m below sea level at its lowest point and includes approximately 3 km of Cenozoic sediments and sedimentary rocks [Mabey, 1963; Serpa et al., 1988; Keener et al., 1993]. The pull-apart basin (Figure 2) was formed in a right stepping bend or gap between the two strike-slip systems [Burchfiel and Stewart, 1966]. According to Stewart [1983], motion on the NDVFZ was initiated prior to the SDVFZ and the ongoing motion of both faults gave rise to the present pull-apart basin system in Central Death Valley during the Miocene [Holm et al., 1994;

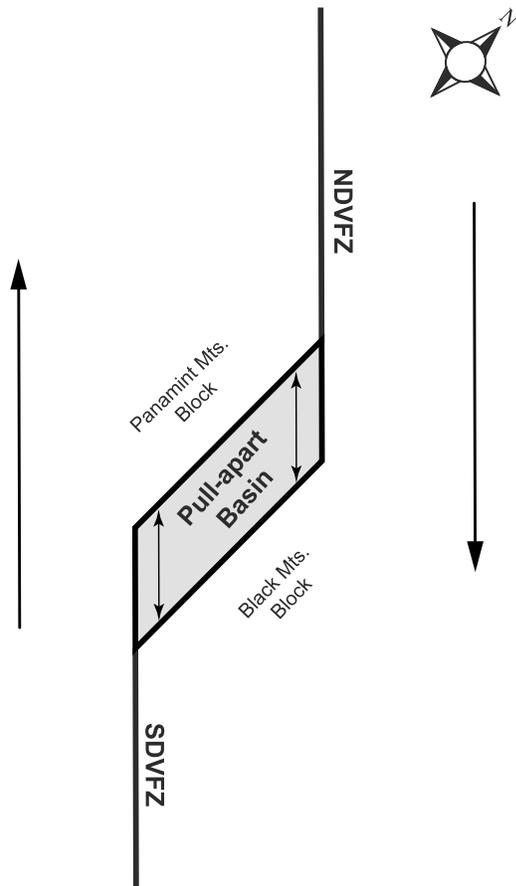


Figure 2. Diagram showing the overall right-lateral strike-slip motion of the DVFZ and the location of the area of tension (gray shaded area) that gave rise to the pull-apart basin in Central Death Valley, California (modified from *Burchfiel and Stewart* [1966]).

Serpa and Pavlis, 1996]. An oblique-slip zone of normal faulting located along the eastern margin of the basin records its extension rate of approximately 1–3 mm/yr [*Burchfiel and Stewart*, 1966; *Hill and Troxel*, 1966; *Serpa and Pavlis*, 1996; *Chávez-Pérez et al.*, 1998; *Klinger and Piety*, 2000; *Knott et al.*, 2005].

[8] The formation and evolution of a pull-apart basin depends on a number of factors [*Aydin and Nur*, 1982]. First, there has to be an offset between two strike-slip faults such that the net displacement across the offset is extensional. The manner in which the basin evolves is highly dependent on the stress field, on the rheology of the rocks around the faults that produce the basin, and on the rotation of the fault trace. When a fault trace is rotated counterclockwise with respect to the far-field velocity vector, the fault will produce a compressional field forming uplift in the region. In the case of the DVFZ, the faults in the system are in a clockwise rotation with respect to the far-field velocity vector, generating an extensional field in the pull-apart basin area, producing a graben structure.

3. 3-D Deformation Model of the DVFZ

[9] To investigate the kinematics of the DVFZ, we have developed a 3-D crustal deformation model for four fault

segments of the fault system (Figure 1 and Table 2). Segment 1 consists of the SDVFZ, segment 2 corresponds to the BMFZ that includes the pull-apart basin, and the two remaining segments represent the NDVFZ (segments 3a and 3b). This last fault zone is divided into approximately two equal-length segments, as this section is the longest fault within the DVFZ (~170 km) and previous studies have suggested that its slip rate may vary along strike [*Frankel et al.*, 2007b; *Ganev et al.*, 2010]. All fault segment traces were digitized at ~1 km resolution, representing the location of the fault system using over 200 linear fault elements. The fault segments were projected into a new coordinate system based on the Pacific-North American plate boundary pole of rotation (PoR) (50.1°N and 285.6°W) [*Wdowinski et al.*, 2007] and embedded in a 1-km grid spanning 500 km by 500 km grid cells in the north-south (y -direction) and east-west (x -direction) directions.

[10] For this study, we use a 3-D semi-analytic linear viscoelastic Maxwell model [*Smith and Sandwell*, 2003, 2004, 2006] that simulates both the elastic [e.g., *Okada*, 1985, 1992] and time-dependent viscoelastic [e.g., *Rundle and Jackson*, 1977; *Savage and Prescott*, 1978] response of vertical strike-slip fault elements to a distribution of body forces. The problem is solved analytically in both the vertical and time dimensions (z , t), while the solution in the two horizontal dimensions (x , y) is developed in the Fourier transform domain to exploit the efficiency offered by the convolution theorem. The model consists of a series of vertical connected faults embedded in a homogeneous elastic plate overlying a viscoelastic half-space (Figure 3) and simulates interseismic strain accumulation, coseismic displacement, post-seismic viscous relaxation of the mantle and complimentary stress behaviors at all stages.

[11] The complete earthquake cycle is modeled with two components: secular and episodic. The secular model simulates interseismic slip that occurs between the fault locking depth and the base of the elastic plate (d to H , Figure 3). We construct this secular model by prescribing fully relaxed slip (assuming infinite time) over the entire thickness of the elastic plate up to the shallow locking depth. In this model component, the fault system is a mature one (geologically evolved), where we analytically sum an infinite number of earthquake cycles to simulate a full secular velocity step across a fault system [*Smith and Sandwell*, 2004]. The episodic model component (or earthquake-response model) prescribes slip over the locked section of each fault segment (0 to d).

[12] Deep slip along faults drives the secular interseismic crustal motions and stress accumulation. Long-term slip rates and locking depths are constrained by contemporary geodetic velocities. The non-secular motion on each fault segment is determined by the earthquake rupture history on that segment. This history requires some knowledge of the timing of major earthquakes over at least the last 1000 years (i.e., an earthquake cycle) and the slip distribution along the segment. Except for the more recent instrumentally recorded events, historical slip distribution is usually unknown and paleoseismic earthquake dates and slip are uncertain. Thus to accommodate realistic earthquake deformation through time, we assume that the amount of coseismic slip for each historical event is equal to the accumulated slip deficit on that segment, estimated by the slip rate and the time since the last

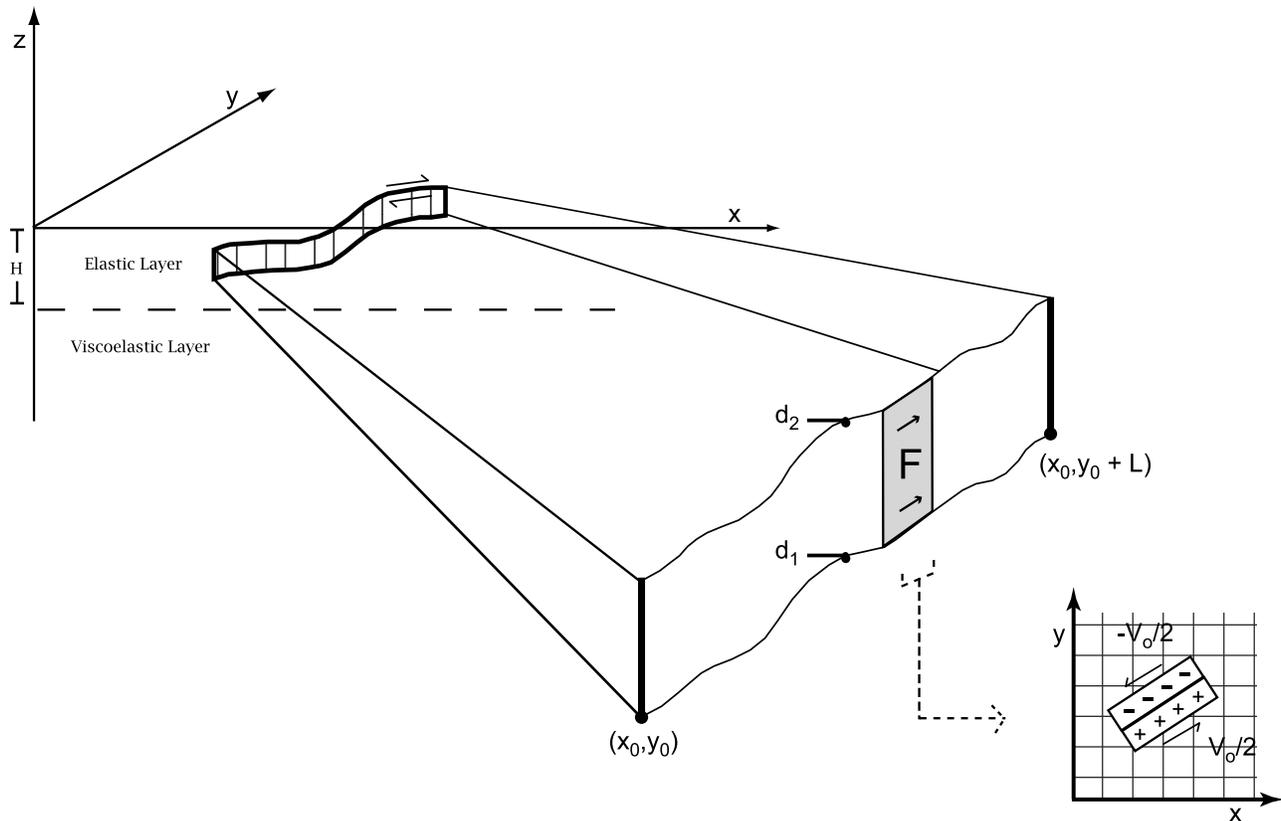


Figure 3. Diagram illustrating the 3-D semi-analytic viscoelastic fault model simulating the response of a body force planar dislocation embedded in an elastic layer overlying a linear Maxwell viscoelastic half-space. Fault elements are embedded in a plate of thickness and extend from a lower depth of d_1 to an upper depth of d_2 . A displacement discontinuity (whose magnitude is determined by the slip rate V_0) across each fault element is simulated using a finite width force couple, F , embedded in a fine grid.

major event. The duration of the viscoelastic response, characterized by the Maxwell time, depends on the viscosity of the underlying half-space and the elastic plate thickness [Smith and Sandwell, 2006]. We assume fixed values (Table 1) for the Young's modulus ($E = 75$ GPa), Poisson's ratio ($\nu = 0.25$), shear modulus ($\mu = 30$ GPa), density ($\rho = 3300$ kg/m³), and gravitational acceleration ($g = 9.81$ m/s²).

[13] In our modeling procedure, we solve for the following model parameters: apparent locking depth for each DVFZ segment (d), horizontal strike-slip rate for each DVFZ segment (s), regional elastic plate thickness (H) and half-space viscosity (η). Our model parameter analysis also includes motion on the PHSFZ and OVFZ, where we adopt slip rates and fault depths from previous studies. The PHSFZ and OVFZ are divided into four fault segments bounded by the paralleling coordinates of the four segments we define for the DVFZ. Strike-slip rates are adjusted to ensure that the sum of input slip rates across the fault system is equal to the far-field estimate of 14 mm/yr in accordance with the upper bound slip motion of the Sierra Nevada microplate located north of the Garlock fault [Wernicke et al., 1988; Bennett et al., 1997; Dixon et al., 2000; McClusky et al., 2001; Bacon and Pezzopane, 2007]. Based on previous measurements [Serpa et al., 1988; Willis et al., 2008; SCEC, 2011a], we assign a slip rate of 4.8 mm/yr for the four fault segments of the DVFZ, which we later adjust to match the geodetic

data. In our starting model, we designate a priori deep slip rates of 5.0 mm/yr for the PHSFZ [Gourmelen et al., 2011] and 4.2 mm/yr for the OVFZ. For each model iteration that searches for the best fitting slip rate on the four DVFZ segments, we adjust the OVFZ slip rate such that the total far-field slip across paralleling segments sums to 14 mm/yr and require that each OVFZ segment slip rate remains within the uncertainty range of the geodetically determined slip of 3.9 ± 1.1 mm/yr [Dixon et al., 1995]. Our starting model also specifies uniform values for locking depth (10 km for all fault segments), and a homogeneous elastic plate thickness of 23 km and viscosity of 1×10^{19} Pa·s.

[14] The time-dependent portion of the model requires information about historical earthquake events and recurrence intervals. The DVFZ study region is an area with low seismic activity in comparison the SAFS. The only

Table 1. Fixed Model Parameters

Property	Name
Young's modulus (E)	75 GPa
Poisson's ratio (ν)	0.25
Shear modulus (μ)	30 GPa
Density (ρ)	3300 kg/m ³
Gravitational acceleration (g)	9.81 m/s ²
Recurrence Interval	1200 yrs

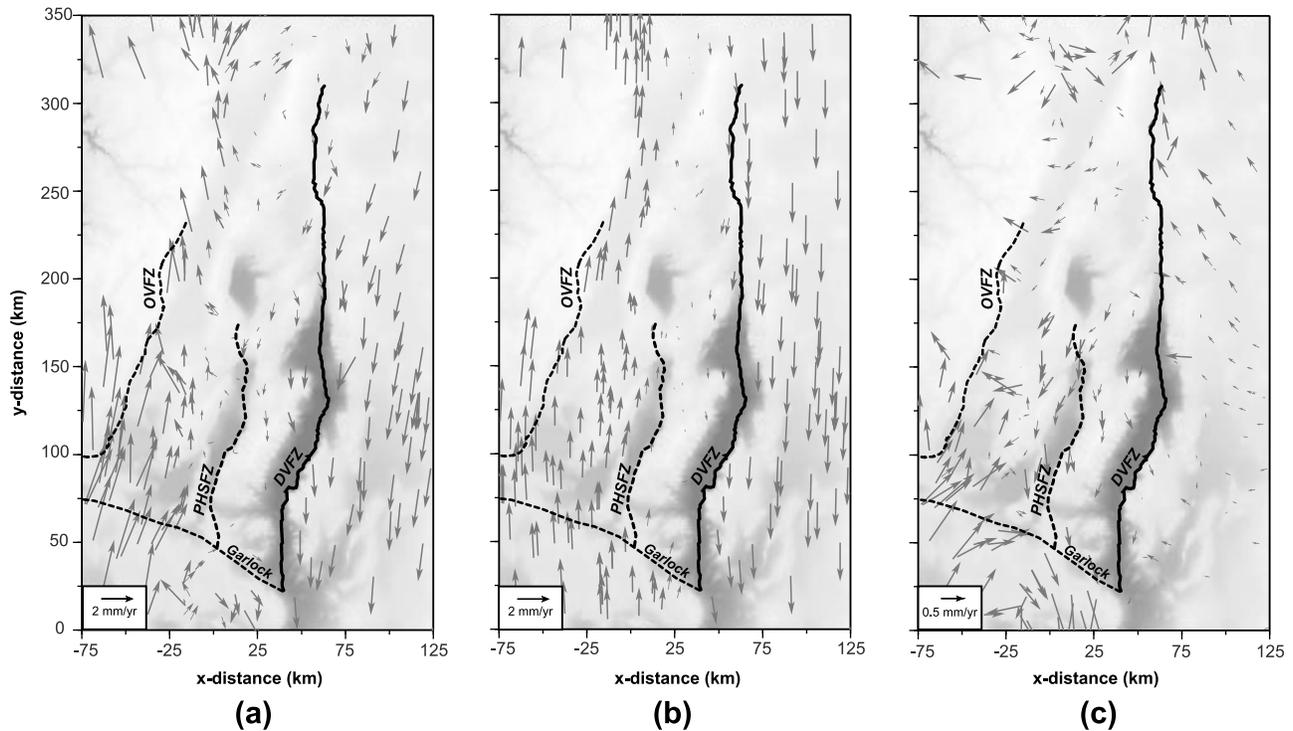


Figure 4. Map view of the (a) GPS velocities used in this study, (b) predicted velocity by our model and (c) the misfit between the observed and modeled velocities. Arrows indicate the direction and magnitude of motion. Solid black line marks the location of the DVFZ and dashed black lines the trace of the additional fault systems considered in the model (OVFZ and PHSFZ) except for the Garlock fault.

significant ($M_w \geq 6$) recorded seismic data available for the general region is from the 1872 $M_w = 7.8$ Owens Valley earthquake (Figure 1). Recurrence intervals are 1750–3100 years for the OVFZ [Dixon *et al.*, 2003], 860–2360 years for the PHSFZ [Zhang *et al.*, 1990], and 500–2600 years for the DVFZ [Wesnowsky, 1986; Klinger and Piety, 2000; Dixon *et al.*, 2003]. Previous modeling efforts have suggested that present-day velocities are not significantly sensitive to seismic events dating back longer than 10 Maxwell times, or ~ 200 years [Smith and Sandwell, 2004], hence we adopt an average recurrence interval of 1200 years for all fault segments in the model. Coseismic fault slip is obtained by multiplying the long-term slip rate of each fault segment by the time between the last major earthquake or by the recurrence interval if no seismic data are known for each respective segment.

4. Geodetic Inversion

[15] We use GPS-derived horizontal velocities within the ECSZ to constrain our model parameters. The velocities apply in this study are derived from a subset of GPS station position time series from a global analysis of $\sim 4,000$ GPS stations collected between 1996 and 2009. To minimize the impact of stations reflecting slip on faults outside the ECSZ, we limit our study to velocities provided by stations located ~ 25 km west to the westernmost part of the OVFZ and ~ 100 km to the east of the DVFZ (Figure 1). This subset contains 240 stations that provide 480 horizontal velocity components whose velocity field is given in Figure 4a. All available continuous GPS stations in the study area with at

least 2.5 years of data are used together with semi-continuous stations of the University of Nevada Reno’s (UNR) Mobile Array of GPS for Nevada Transtension (MAGNET) network [Blewitt *et al.*, 2009], which typically have ~ 4 years of data consisting of 3 campaigns of ~ 1 month data collection. Velocities from the Southern California Earthquake Center (SCEC) Crustal Motion Map version 4 (CMMv4) [Shen *et al.*, 2011] are also included by solving for and applying a transformation between the CMM velocity field and our velocity field for continuous and MAGNET stations.

[16] GPS data are processed using the precise point positioning method of the GIPSY-OASIS II method [Zumberge *et al.*, 1997] with reprocessed fiducial-free GPS orbits and clocks made available by the Jet Propulsion Laboratory (JPL). The GPS observation model includes absolute calibration models for both the station antennas and GPS satellite transmitters. Models of solid earth tides and tidal ocean loading are applied. Tropospheric delay is modeled independently at each station by a zenith parameter and two gradient parameters as a random walk process. Ambiguity resolution is subsequently applied by UNR’s Ambizap3 software [Blewitt, 2008], which exploits a fixed-point theorem and global network estimation filter that operates on the EMST (Euclidean minimum spanning tree).

[17] Daily coordinate transformation parameters into the International Reference Frame (ITRF2005) are provided by JPL. ITRF2005 positions are transformed into NA09, a North America-fixed reference frame developed at UNR, by performing daily transformations into a frame that is defined by minimizing the horizontal velocities of 16 stations across the stable part of the North America continent (away

from areas affected by glacial isostatic adjustments). Common mode errors for this continental scale frame are further reduce by including an additional 35 stations as far away as Greenland, Alaska, Hawaii, and the Caribbean in a daily spatial (7 parameter) filter. We estimate station velocities from the resulting daily time series using the CATS software package [e.g., *Williams, 2003*] while accounting for annual and semi-annual constituents. The same software is used to estimate rate uncertainties given the assumption that the error model is dominated by white noise plus flicker noise.

[18] Our model parameter analysis uses both inverse and forward modeling techniques to fit GPS-derived velocities. To estimate the apparent locking depths for each of the 4 fault segments of the DVFZ we utilize an iterative least square inverse approach based on the Gauss-Newton method. This method solves the set of equations

$$V_{gps}(x, y) = V_m(x, y, \mathbf{d}), \quad (1)$$

where V_{gps} is the geodetic velocity measurement from the GPS stations in terms of the x and y Cartesian coordinate, V_m is the velocity obtained by the model based on a set of locking depths, \mathbf{d} , that minimizes the root mean square residual misfit (RMS). This misfit is calculated using the equations

$$V_{res}^i = \frac{V_{gps}^i - V_m^i}{\sigma^i} \quad (2)$$

$$RMS = \sqrt{\frac{1}{N} \sum_{i=1}^N (V_{res}^i)^2}, \quad (3)$$

where V_{res} is the residual velocity, σ^i is the uncertainty calculated for the i th geodetic velocity measurement, and N is the number of geodetic observations.

[19] We use a Taylor expansion series in terms of locking depth, d , to obtain V_m as

$$V_m(d + \Delta) = V_m(d) + \sum_{j=1}^M \Delta_j \frac{\partial V_m}{\partial d_j} + \dots \quad (4)$$

In equation (4), the partial derivatives are calculated numerically over a 1-km depth range and Δ_j adds a small perturbation in the j th depth parameter. The small model perturbation is calculated using a weighted approach represented by

$$\Delta = (\delta V_m^T \mathbf{C}^{-1} \delta V_m + \lambda \mathbf{I})^{-1} \delta V_m^T \mathbf{C}^{-1} V_{res}, \quad (5)$$

where \mathbf{C} represents the diagonal covariance matrix of GPS uncertainties, λ is the damping parameter, and \mathbf{I} is the identity matrix. Damping parameter $\lambda = 5$, determined empirically [c. f. *Strang, 1986*], is used for each iteration to stabilize residual misfit results of the inversion. While locking depths are estimated using the above inverse methods, we solve for slip rates, elastic plate thickness, and half-space viscosity using a forward modeling approach where we iteratively minimize the residual misfit of the model and data through an incremental parameter search. We performed 2 rounds of iterations for both inverse and forward modeling approaches. The second round of inverse and forward iterations utilized the

best fit results of the first run of iterations as the starting model parameters to ensure model parameter stability.

5. Results

5.1. DVFZ Model Parameters

[20] Our locking depth inversion involves 6 free parameters corresponding to two horizontally shifted velocity components for the GPS data (fault-parallel and fault-perpendicular directions, used to place the geodetic data within the reference frame of the model) and 4 apparent locking depths associated with each DVFZ model fault segment. The horizontally shifted velocity parameters are calculated by removing the mean misfit obtained from the starting model parameters. We iterated 30 times through our inversion algorithm until the locking depth solutions provided stable results with minimal uncertainties. Next, we iteratively modified slip rates, elastic plate thickness, and half-space viscosity (6 free parameters) for a minimized RMS velocity residual. In total, we performed over 70 iterations, scanning the parameter space for locking depths ranging from 1–20 km, slip rates ranging from 1–8 mm/yr, elastic plate thicknesses spanning from 15–50 km, and viscosities varying between from 1×10^{17} – 1×10^{21} Pa·s.

[21] Our best fitting model yielded an RMS velocity misfit of 1.0 mm/yr in the fault-perpendicular direction and 1.5 mm/yr in the fault-parallel direction (Figures 4 and 5). Locking depth and slip rate results for each fault segment are provided in Table 2 and discussed further in section 5.2. Uncertainties in locking depths are determined from the covariance matrix of the final iteration and are reported at 1σ standard deviation. Our best fit model requires an elastic plate thickness of 35 km and half-space viscosity of 1×10^{19} Pa·s. In comparison, results from previous studies have suggested a crustal thickness between 30 and 35 km [*Asmerom et al., 1990; Serpa, 1990*] while other studies assume elastic plate thickness of 15 km [*Hammond et al., 2009, 2010*] for the DVFZ. Prior results have obtained higher crustal viscosities ($10^{19.5}$ Pa·s– $10^{20.5}$ Pa·s) [*Hammond et al., 2009, 2010*] and combined upper mantle viscosities of $10^{18.5}$ – 10^{19} Pa·s [*Pollitz et al., 2001; Hammond et al., 2009, 2010*]. These results suggests that our elastic layer extends to depths consistent with the mid to lower crust thickness and the viscosity of the underlying region is consistent with that of the lower crust and upper mantle.

5.2. Horizontal Velocity Field

[22] The vector velocity field resulting from our best fitting parameter set is illustrated in Figure 4b, with the misfit between GPS and modeled velocities shown in Figure 4c. As the model parameters for the DVFZ segments are specifically optimized to provide a minimized residual, our results show lowest residual differences along the DVFZ corridor. Figure 4 depicts a small amount of rotation to the east of the DVFZ in the GPS velocity field that our model does not account for. This difference may be due to the effect of the pole of rotation applied in this study, which we adopted from previous SAFS model [e.g., *Wdowinski et al., 2007*], thus we expect some regional errors. Complications arising from the interaction of the Garlock fault with southern extensions of the OVfZ, PVfZ, and DVfZ are also evident.

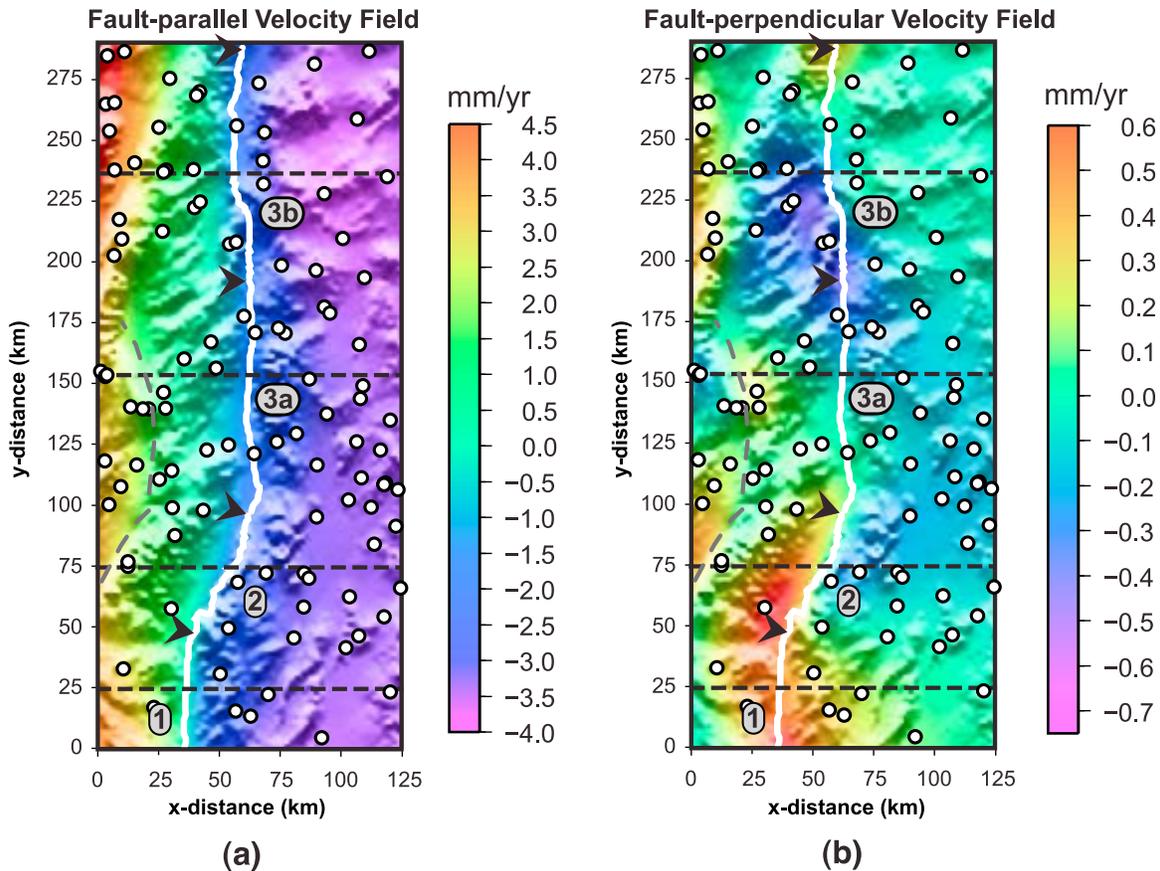


Figure 5. Modeled horizontal velocity fields of the DVFZ with segment locations in the pole of rotation (PoR) x-y coordinate system, showing the (a) fault-parallel and (b) fault-perpendicular directions. For this model projection, we use the PoR of *Wdowinski et al.* [2007] (50.1°N and 285.6°W) and note that the axes are represented in kilometers from a chosen starting position at the southwest corner of the fault zone. Dashed gray line marks the location of the PHSFZ. White circles represent the GPS station locations. The dashed black lines represent fault profile locations of the model represented in Figure 6. The black arrowheads represent fault segment corridor boundaries, from which GPS stations are plotted in Figure 6. Fault labels correspond to segments coinciding with Table 2: (1) SDVF, (2) BMFZ, where the pull-apart basin is located, (3a) southern segment of the NDVFZ and (3b) northern segment of the NDVFZ.

[23] Our resulting horizontal velocity fields are also illustrated in the fault-parallel and fault-perpendicular components, shown in Figure 5. We model a cumulative right-lateral fault-parallel velocity field of 3.8–6.7 mm/yr (2.9–7.3 mm/yr including uncertainties) for the different segments of the DVFZ. In the fault-perpendicular direction, velocities of approximately 0.5–0.6 mm/yr of east trending (positive) deformation span most of the pull-apart basin region, showing

the perpendicular component of motion caused by the right lateral slip of the DVFZ as the fault system bends clockwise. To compare our best fitting model velocities with the GPS velocities, we extract a model velocity profile across the center of each of the fault segment corridors (Figure 6), while the GPS data were binned within the fault corridors and projected onto the perpendicular trace. Fault corridors are of the same length as the division of the fault segments of the

Table 2. Death Valley Fault Zone Model Parameter Results^a

Segment	Name	Slip Rate ^b (mm/yr)	σ^c	Locking Depth (km)	σ^d	Coulomb Stress Rate (MPa/100yr)	Strain Rate (nstrain/yr)	Moment Rate ($10^{14} \times \text{Nm}/100\text{yr}/\text{km}$)	Moment Magnitude
1	SDVF	5.7	0.7	12.7	1.0	0.46	153.3	2.2	7.0–7.4
2	BMFZ	4.8	0.7	9.8	2.6	0.36	119.4	1.4	6.7–7.2
3a	NDVF a	3.8	0.9	17.1	1.3	0.25	82.2	1.9	7.0–7.5
3b	NDVF b	6.7	0.6	13.1	1.0	0.49	161.7	2.6	7.1–7.7

^aThese values are obtained using an elastic plate thickness of 35 km and a half-space viscosity of 1×10^{19} Pa·s.

^bSlip rate refers to the strike-slip rate

^cSlip rate uncertainty at one standard deviation.

^dLocking depth uncertainty.

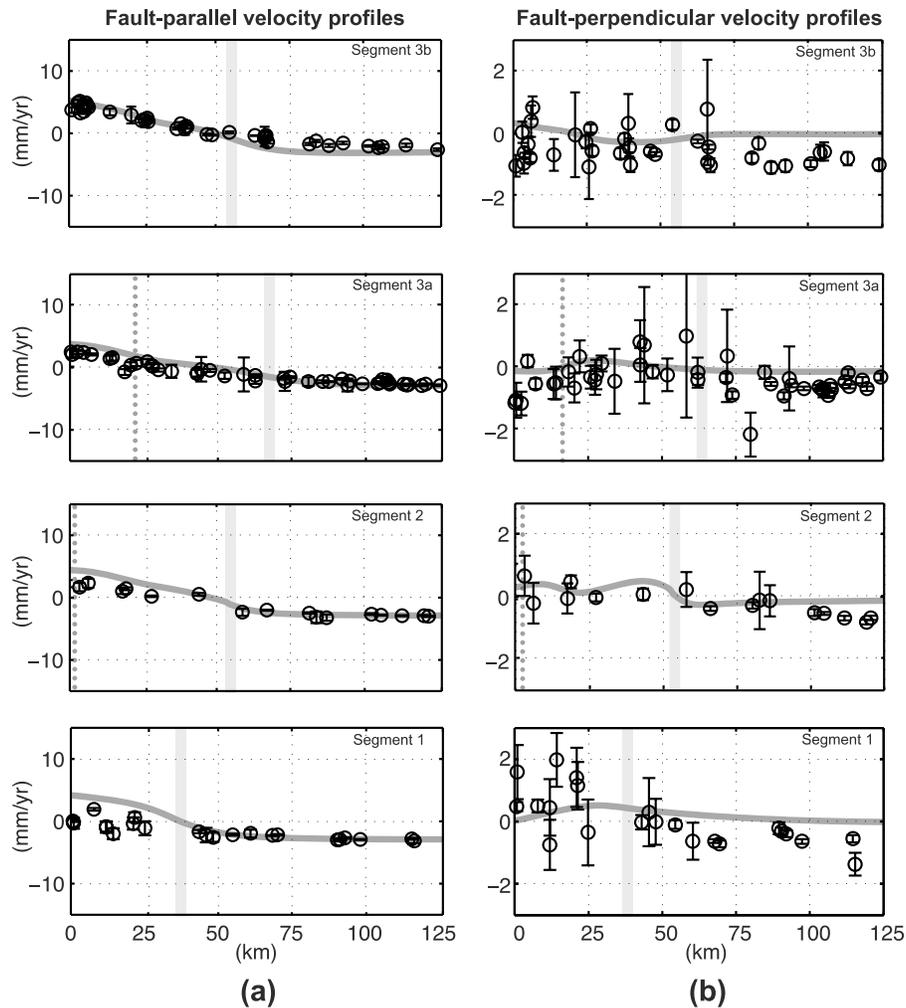


Figure 6. Modeled velocity profiles (gray line) acquired across the center of each fault segment (Figure 5, dashed lines) with GPS velocities (black circles) from each fault segment corridor (Figure 5, black arrowheads) projected onto each profile for visual comparison. Vertical gray box shows the location of the fault segment in profile view and dotted vertical gray line indicates the location of the PHSFZ. The horizontal axis of each plot represents the horizontal (east–west) distance across the model profile. (a) The best fitting fault-parallel velocity model for each fault segment and (b) the fault-perpendicular velocities of the best fitting model for each fault segment.

DVFZ (see Figure 5). Overall, fault-parallel (Figure 6a) and fault-perpendicular (Figure 6b) velocity profiles are in general agreement with the GPS velocities, although there is notable scatter in the data for the fault-perpendicular component. We note that the residual misfit to these data (1.0 mm/yr and 1.5 mm/yr in the fault-perpendicular and fault-parallel directions, respectively) is quite good, however some of the visible scatter in the data, particularly in the fault-perpendicular component, is caused by the projection of station locations onto a single profile.

[24] Our model results (Table 2) yield a locking depth of 12.7 ± 1.0 km and a slip rate of 5.7 ± 0.7 mm/yr for the SDVFZ, shown in segment 1 of Figure 6. The model velocity of this section is primarily constrained by stations found to the east side of the fault segment. Stations located on the west side of the fault produce a noticeable misfit between the model profile and GPS velocities, which is likely due to complexities arising from motion along the

Garlock fault. The GPS velocity field along the southwest quadrant of the DVFZ–Garlock intersection (Figure 4a) reveals both southeast and southwest (y - x space) oriented velocities, while our model simulates simple strike-slip north oriented velocities here, due to the omission of the Garlock fault. This is also reflected in the fault-perpendicular profile (Figure 6b), where the GPS measurements follow the behavior of the model to the east of the fault and depict a great deal of scatter to the west. Both locking depth and slip rate solutions agree with previously published locking depth results ranging from 5–13 km [Peltzer *et al.*, 2001; Dixon *et al.*, 2003; Meade and Hager, 2005; Willis *et al.*, 2008] and slip rate estimates of 3–8 mm/yr [Bennett *et al.*, 1997; Dixon *et al.*, 2003; Hill and Blewitt, 2006; Willis *et al.*, 2008]. These results from previous groups also reflect motion along the BMFZ (segment 2), where we estimate a locking depth of 9.8 ± 2.6 km and a slip rate of 4.8 ± 0.7 mm/yr. The modeled results obtained for segment 2 fit

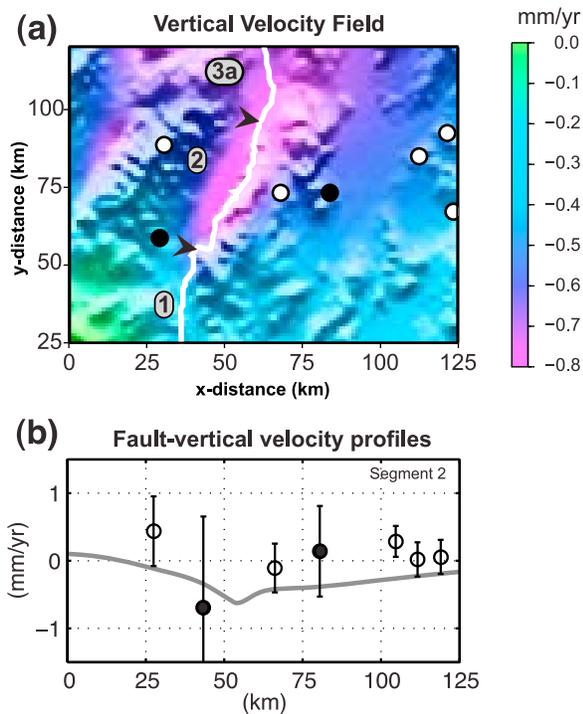


Figure 7. (a) Zoomed in view of the vertical velocity field model highlighting the BMFZ segment of the DVFZ. Negative values represent subsidence. Fault segment labels and symbols are the same as in Figure 5. (b) Modeled vertical velocity profile (gray line) acquired perpendicular to the center of the BMFZ fault corridor showing the subsidence in the pull-apart basin region. GPS velocities for this segment (black open circles) are projected onto the fault-perpendicular profile for visual comparison. Black filled circles represent the GPS stations that have been recording data for less than 5 years.

quite well with the stations located to the east of the fault and one station located to west. The remaining stations available on the west of the fault reflect anomalous motion of the PHSFZ, suggesting that our assumed slip rate [from *Gourmelen et al.*, 2011] for this segment should be less than 5 mm/yr here. The fault-perpendicular profile of segment 2 provides a consistent agreement between the model and the GPS measurements.

[25] Along the area of the two DVFZ northern segments, we have an improved coverage of GPS stations. The best model parameters obtained, given in Table 2, for the NDVFZ (segments 3a and 3b), reflect locking depths of 17.1 ± 1.3 km (3a) and 13.1 ± 1.0 km (3b). Averaging these two segments yield a locking depth of 15.1 km for the entire NDVFZ segment, compared to the 13 km locking depth estimated by *Willis et al.* [2008]. Slip rates derived by previous studies suggest that the NDVFZ segment has a slip rate between 2–9 mm/yr [*Bennett et al.*, 1997; *Dixon et al.*, 2000; *Klinger and Piety*, 2000; *McClusky et al.*, 2001; *Frankel et al.*, 2007a, 2007b; *Willis et al.*, 2008]. Our modeled slip rates (Table 2) of 3.8 ± 0.9 mm/yr and 6.7 ± 0.6 mm/yr for the southern (3a) and northern (3b) segments of the NDVFZ, respectively, are within reasonable agreement of previous results. In the velocity profiles displayed in Figure 6, we note

some larger misfits between GPS and model velocities (fault-parallel and fault-perpendicular directions) for these two segments. As shown in the segment 3a fault-parallel velocity profile (Figure 6a), we overestimate the GPS velocities to the west of the DVFZ, due to an applied high slip rate on the PHSFZ and OVfZ segments, required by our constant far-field velocity constraint (discussed in Section 4). The model also overestimates the fault-parallel velocity of segment 3b west of the fault but it agrees with all of the GPS velocities to the east. In the fault-perpendicular velocity profile of segments 3a and 3b (Figure 6b) we also note a moderate correlation between model and GPS velocities. Our model consistently calculates a higher fault-perpendicular velocity than the observed velocity. This difference between the observed geodetic and model velocities is also revealed in Figure 4c where we note higher residual vectors to the east of the NDVFZ.

[26] We can infer fault perpendicular (or extension rates) for the DVFZ based on the geometry of the faults and the strike-slip rate applied to the model. The fault-perpendicular component is not a parameter we solve for within model parameter search, but rather a consequence of bending fault geometry and fault-perpendicular motion. This is demonstrated in Figure 5b between segments 1 and 2, where our model simulates an increased fault-perpendicular velocity (>0.6 mm/yr) field that corresponds to a net eastward motion in this region due to the transtensional bend in the fault trace. Peak rates of extension are ~ 1.0 – 1.2 mm/yr along the pull-apart basin. Alternatively, between the northern segments (3a and 3b) we also have an area with moderate negative values of fault-perpendicular velocity (~ -0.5 mm/yr), caused by the combination of a small transpressional bend here and secondary effects from the PHSFZ fault geometry to the west.

5.3. Vertical Motion

[27] Vertical motion of the DVFZ is an important calculation in that it provides a first-order rate of subsidence and uplift of the pull-apart basin and surrounding mountain structures in central Death Valley. To develop the model, we assume that the far-field motion is always parallel to the relative plate motion vector. Because the fault segments are not always parallel to this driving force, horizontal motion on free-slipping fault planes has both a fault-parallel and fault-perpendicular component. It is the fault-perpendicular component that drives most of the vertical deformation, thus any vertical deformation features revealed by the model are a direct result of bends in the fault segment geometry. For simplicity, our model does not account for vertical loads due to topography. Furthermore, because the model parameters are constrained only by horizontal GPS velocity measurements, resulting vertical velocity deformation can be compared with both geologically inferred rates and geodetically measured velocities.

[28] Our model yields very minor (<0.2 mm/yr) uplifting and subsiding features along most of the bends of the DVFZ, however our focus is on the major subsiding feature that dominates the interseismic kinematics of the pull-apart basin. The vertical velocity profile along this fault segment (Figure 7b) shows present-day subsidence rates between 0.5 and 0.8 mm/yr across the fault trace. This motion is in agreement with the theory of pull-apart basin formation for a right stepover associated with strike-slip faults [*Aydin and*

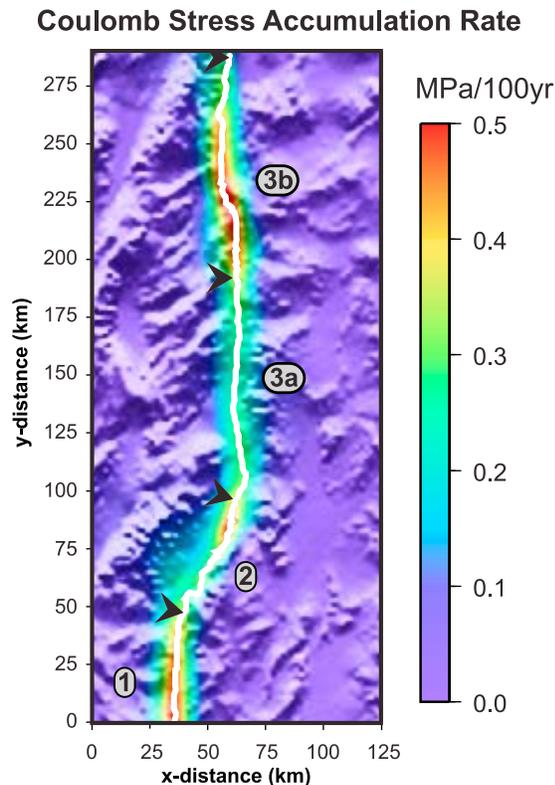


Figure 8. Coulomb stress accumulation of the DVFZ in MPa/100 years. Fault segment labels and symbols are the same as in Figure 5. Color scale is saturated at 0.8 MPa/100 yrs.

Nur, 1982]. The model generates a large oval-shaped depression that represents the subsidence along the entire segment where the pull-apart basin is located. Geologic estimates of the area have calculated a basin of approximately 3 km deep [Keener et al., 1993; Blakely et al., 1999], inferring a constant subsidence rate of 0.5 mm/yr since its formation ~ 6 Ma. Our earthquake cycle model suggests subsidence rates on the order of 0.5 mm/yr are appropriate for the secular portion (last ~ 1000 years) of the earthquake cycle, although this rate can vary from +0.25 mm/yr to -0.5 mm/yr for the first 200 years after a major earthquake. These rates are described further and applied to an evolution model of the DVFZ is discussed in Section 6.3.

[29] As a first-order comparison, we also inspect available vertical GPS velocities along the region of the BMFZ. We again note that these data were not used to constrain our model parameters and should be used with caution due to relatively high measurement uncertainties (0.2–1.3 mm/yr for the DVFZ region). Furthermore, vertical rates have larger uncertainties than horizontal rates and generally are not reliable until the time series include at least 3 years of data. In Figure 7b, we compare our vertical velocity model profile with the GPS velocities from stations that have been recording data for more than 5 years (open circles) and for less than 5 years (solid circles). Our vertical model has an RMS misfit of 0.4 mm/yr and mostly overestimates the subsiding (negative) GPS velocities, although the model profile does lie within 5 out of 7 velocity uncertainties. This

overestimation suggests that the horizontal slip rate we apply to this fault segment could be decreased to better match the vertical data, although doing so would not provide the best fitting velocity model for the fault-parallel and fault-perpendicular components.

5.4. Interseismic Stress and Strain Accumulation Rates

[30] We calculate Coulomb stress accumulation rate [e.g., King et al., 1994; Simpson and Reasenber, 1994; Smith and Sandwell, 2003] along the four fault segments of the DVFZ. This calculation is based on the Coulomb failure criterion, expressed as

$$\dot{\sigma}_f = \dot{\tau} - \mu_f \dot{\sigma}_n, \quad (6)$$

where $\dot{\sigma}_n$ and $\dot{\tau}$ represent the normal and shear stress rates on a failure plane, respectively, and μ_f the effective coefficient of friction. Our model calculates the stress rate tensor from the 3-D vector velocity field in the same manner as Smith and Sandwell [2003, 2006] and we also assume a constant μ_f of 0.6. To calculate the normal and shear stresses, we assume that the strike-slip fault is a vertical fault plane and that right-lateral shear stress is positive. We obtain depth-averaged Coulomb stress rate at a depth equal to half of the locking depth modeled for each fault segment [King et al., 1994; Smith-Konter and Sandwell, 2009]. From Coulomb stress accumulation rates we can also estimate the Coulomb strain rate ($\dot{\epsilon}$) of each fault segment of the DVFZ. This is obtained by taking the ratio between the Coulomb stress rate ($\dot{\sigma}_f$) and the rock shear modulus (μ) expressed as

$$\dot{\epsilon} = \frac{\dot{\sigma}_f}{\mu}. \quad (7)$$

For strain rate calculations we assume a constant shear modulus of 30 GPa and evaluate stress rate at the surface, instead of half the locking depth, to represent the locality in which strain is typically measured by surface strainmeters.

[31] Our model results reveal Coulomb stress accumulation rates (Figure 8) that are relatively high along segments 1 and 3b (peak rates calculated at 0.46 MPa/100 yr and 0.49 MPa/100 yr, respectively) and relatively low along segment 3a (0.25 MPa/100 yr). Representative peak stress and strain rates for each fault segment are given in Table 2. As Coulomb stress accumulation rate is proportional to slip rate and inversely proportional to locking depth, these results illustrate how a fault with a low slip rate and relatively deep locking depth (i.e., segment 3a), accumulates stress at a relatively lower rate. In comparison, segment 2 accumulates stress at a relatively higher rate (0.36 MPa/100 yr) with only a slightly faster slip rate (4.8 ± 0.7 mm/yr versus 3.8 ± 0.9 mm/yr) but a much shallower locking depth (9.8 km versus 17.1 km). Also, contrasting stress accumulation rates along the NDVFZ (3a and 3b) may be due to the fact that the northern segment is closer to the OVFZ and this fault zone may influence the stress accumulating on the northern segment more than the southern segment. In addition, Coulomb stress accumulation rates are dependent on the compressional and extensional fields inherent to fault's orientation with respect to far-field driving force; stress

accumulation rates are slightly decreased (or increased) when the fault's geometry is a compressional (or extensional).

[32] Comparing the DVFZ stress accumulation rates with those of the SAFS (0.2–7.2 MPa/100 yrs) [Smith-Konter and Sandwell, 2009], rates for the DVFZ are mostly lower, but do coincide with stress rates for some segments of the San Jacinto fault and of the Eastern California Shear Zone (north of the Garlock fault). This difference is mainly influenced by larger slip rates (12–40 mm/yr) and, in some cases, more shallow locking depths (6 km) along the SAFS. Likewise, we estimate DVFZ strain rates of 82.2–161.7 nstrain/yr that are lower than the primary San Andreas strain rates derived from large-scale strain rate models developed for the Pacific-North American plate boundary (100–3000 nstrain/yr) [Sandwell et al., 2010; SCEC, 2011b], however these rates are consistent with the ~50–300 nstrain/yr strain rates found locally along the DVFZ [Smith-Konter et al., 2010].

6. Discussion

6.1. Geologic Versus Geodetic Slip Rate Discrepancies

[33] Minor discrepancies between geologic and geodetic slip rates along the DVFZ exist, although they are not significant. For example, for the NDVFZ segment, geologic slip rates have been estimated at 3–9 mm/yr [Klinger and Piety, 2000; Frankel et al., 2007a, 2007b; Willis et al., 2008] and previous reports of geodetic slip rates have been estimated between 2–8 mm/yr [Bennett et al., 1997; Dixon et al., 2000; McClusky et al., 2001]. Such discrepancies might be due to several factors. The earthquake cycle of a fault, for instance, can cause changes in interseismic velocity [Dixon et al., 2003; Meade and Hager, 2005]; fault velocities are faster right after an earthquake and become slower toward the end of the cycle. Considering that there is no record of major events along the DVFZ over the last 1000 years and recurrence intervals are estimated at similar time scales, suggests that the DVFZ might be near the end of its earthquake cycle. Thus we might expect geodetic rates to be lower than geologic rates. However, the SDVFZ and the BMFZ slip rates obtained by our model are slightly higher than the geologic estimates (~1 mm/yr), suggesting that perhaps these two segments are not quite at the end of their respective earthquake cycles. In comparison, the two fault segments of the NDVFZ are within reasonable agreement with geologic values implying that might be nearing the ends of their earthquake cycle. Another possible explanation for our larger slip rates is that we do not take into consideration internal deformation within fault blocks or other faults in the larger system [Cemen and Wright, 1990; Serpa and Pavlis, 1996], which may decrease the velocities of our modeled slip rates.

[34] We also note that there is a high degree of correlation between slip rate and locking depth parameters in geodetic models, as has been demonstrated by several studies [e.g., Segall, 2002; McCaffrey, 2005; Smith-Konter et al., 2011]. The basic premise is that models with faster slip rates and deeper locking depths can provide an equivalent fit to geodetic data as compared with models with slower slip rates and more shallow locking depths. In other words, the slip rate and locking depth parameter estimation problem is ill conditioned. In this analysis, we have done our best to avoid any slip rate or locking depth biases, and have investigated

all optimal combinations of parameters before arriving at the best fit parameters discussed here.

[35] It is also important to note that the total resultant far-field horizontal velocities for all segments are lower than the prescribed slip rates. For example, the input slip rate of segment 2 is 4.8 ± 0.7 mm/yr; however the cumulative far-field velocity revealed in Figure 6a for this segment is 2.7 mm/yr. These phenomena result from the response of a relatively thin elastic plate [Rybicki, 1971; Smith and Sandwell, 2004]. Simple 2-D analytic solutions [e.g., Rybicki, 1971] demonstrate the inherent relationship that exists between the thickness of an elastic plate and the resulting far-field surface velocities; an elastic half-space model (i.e., infinite elastic plate thickness) produces exact far-field surface velocities from input slip rates, on the other hand layered elastic/viscoelastic models can produce far-field velocities with only a fraction of the input slip rates, depending on the thickness of the elastic plate. Moreover, the thicker the elastic plate, the better far-field velocities will agree with input slip rates. Relating this to our DVFZ model, the relatively thin elastic plate (35 km) reduces the resulting far-field velocities. Because of this behavior, our slip rate parameter search tends to prefer larger rates than some that have previously been reported in the literature. While our forward modeling prefers a thin elastic plate, to test the relationship of slip rate and elastic plate thickness, we also performed a simple test using a thicker plate. For this model we forced the elastic plate thickness to be 50 km and solved for the best fitting slip rates. This approach resulted in an overall decrease in preferred slip rates by ~1–3 mm/yr for the DVFZ. The velocity field generated for this model deviated only slightly from the velocity field shown in Figures 5 and 6, but with a higher RMS residual (1.1 mm/yr and 1.7 mm/yr RMS misfit in the fault-perpendicular and fault-parallel directions, respectively). Furthermore, while our results appear to support the use of a thin elastic plate, the elastic plate thickness may play an important role in the reconciliation of geologic and geodetic slip rates.

6.2. Seismic Moment

[36] During the interseismic stage of the earthquake cycle, stress is accumulated over time in the shallow locked zone of a fault. The interseismic phase is completed when this stress is released resulting in an earthquake, or coseismic stress drop. The rate at which stress accumulates on a fault, in conjunction with earthquake recurrence intervals, can be used to make first-order estimates of the stress drop and earthquake magnitude of seismic events. As little is known about the size of past DVFZ earthquakes, this approach can be used to constrain earthquake magnitudes and assess future seismic hazards.

[37] Utilizing Coulomb stress accumulation rates obtained by our model, combined with a DVFZ earthquake recurrence interval of 500–2600 years [Wesnousky, 1986; Klinger and Piety, 2000; Dixon et al., 2003], we estimate minimum and maximum stress drops ($\Delta\sigma$) for DVFZ segments, assuming a constant stress accumulation rate. We then estimate seismic moment (M_0) by incorporating the dimensions of each fault segment,

$$M_0 = \frac{w^2 L \pi \Delta\sigma}{2} \quad (8)$$

where w is the locking depth and L is the length of each segment. The moment magnitude (M_w) is obtained using the relationship

$$M_w = \frac{\log M_0}{1.5} - 6.03 \quad (9)$$

We can also make simple estimates of seismic moment rate from our resulting model parameter analysis. The seismic moment accumulation rate is estimated from the locking depth (d_j), slip rate (s_j), and μ values of each fault segment. In this analysis, we estimate the seismic moment accumulation rate per unit fault length (I) as

$$\frac{\dot{M}_j}{I} = \mu d_j s_j. \quad (10)$$

We calculate earthquakes moment magnitudes for the DVFZ ranging from 6.7–7.1 (based on the lower recurrence interval of 500 years) and 7.2–7.7 (based on the upper recurrence interval of 2600 years). We also calculate seismic moment rates for the DVFZ ranging from 1.4 – 2.6×10^{14} Nm/100 yr/km (Table 2). Because there is a linear relationship between seismic moment accumulation rate, locking depth, and slip rate, fault segments with deeper locking depth and higher slip rates produce a higher moment accumulation rate. Conversely, as moment magnitude is derived from stress drop, which is inversely proportional to locking depth, it is also possible to have larger earthquake magnitudes from more shallow fault depths. This is reflected in our results obtained by our model and moment magnitude calculations. For example, the highest moment accumulation rate, 2.6×10^{14} Nm/100 yr/km, is calculated for the NDVFZ (b) segment. This fault segment has the fastest slip rate, one of the deeper locking depths (17.1 km) and is poised to generate the biggest moment magnitude (7.1–7.7 for the lower and upper recurrence intervals, respectively). In contrast, the BMFZ segment has the lowest moment rate (1.4×10^{14} Nm/100 yr/km) and smallest moment magnitude (6.7–7.2 for the lower and upper recurrence intervals).

[38] While moment accumulation rates for the DVFZ are low in comparison to moment rates estimated for the primary SAFS (8.7 – 7.7×10^{14} Nm/100 yr/km) [Smith-Konter *et al.*, 2011], these rates are comparable to rates derived for some segments of the San Jacinto fault and the ECSZ, which are ~ 2 – 4×10^{14} Nm/100 yr/km [Smith-Konter *et al.*, 2011]. Considering the ECSZ hosted two significant earthquakes in the last 20 years (1992 M7.3 Landers earthquake and the 1999 M7.1 Hector Mine earthquake), it is certainly possible that the DVFZ is capable of generating a large event. Previous studies have reported a broad range of plausible earthquake magnitudes for the DVFZ (6.5–7.9) [Wesnowsky, 1986; Field *et al.*, 2009; SCEC, 2011a] but the new magnitudes derived here further limit this range of magnitudes to 6.7–7.7. Community-derived seismic hazard models like the Uniform California Earthquake Rupture Forecast V2 (UCERF2) establish the probability of having a seismic event of a magnitude greater or equal to 6.5 along the DVFZ during the next 30 years between 5–7% [Field *et al.*, 2009], largely based on relatively low slip rates of the fault zone. Moreover, although seismic events along the DVFZ have not dominated the historic earthquake record and are currently thought to

have a low probability of occurring in the next 30 years, these results imply that when a large earthquake does occur along the DVFZ, it has the potential to release large amounts of energy and significantly alter the regional stress field.

6.3. Death Valley Evolution Model

[39] Due to the well exposed structures in the Death Valley area, there is a general agreement on the right-lateral sense of motion, the northwest trend, and that the displacement of both the NDVFZ and the SDVFZ produced the pull-apart basin [Burchfiel and Stewart, 1966; Hill and Troxel, 1966; Wright and Troxel, 1967, 1970; Machette *et al.*, 2001], however several of these studies do not agree on the amount of displacement on the fault zones. This disagreement is directly related to a controversy about the structural interpretation of the area and the amount of tectonic displacement. Some studies have suggested that the NDVFZ has a right-lateral offset of ~ 40 to 100 km [Stewart, 1967, 1983; McKee, 1968; Stewart *et al.*, 1968; Snow and Wernicke, 1989], while the SDVFZ has also been interpreted to have as little as ~ 8 km of displacement [Wright and Troxel, 1967, 1970; Davis, 1977] or as much as 20 to 80 km [Drewes, 1963; Stewart, 1983; Wernicke *et al.*, 1988; Holm *et al.*, 1992; Applegate, 1995; Snow and Wernicke, 2000]. The difference in the amount of inferred displacement in the fault zones is due to the lack of clear piercing points. Most are based on interpreted correlations of faults and features such as the distribution of Precambrian rocks, stream channels, and sedimentary facies boundaries, all of which carry ambiguities that lead to alternative interpretations.

[40] The slip rates obtained by our model for the NDVFZ (segments 3a and 3b) and the SDVFZ (segment 1), allow us to calculate an approximate amount of displacement of each of the fault zones. The NDVFZ amount of displacement is estimated by taking its time of formation, ~ 15 Ma [Wernicke *et al.*, 1988], and by making a simple assumption that the fault zone has had a constant slip rate since then. Using this, we estimate that the southern NDVFZ fault (segment 3a and slip rate of 3.8 ± 0.9 mm/yr) has an offset of 57 km and the northern NDVFZ fault (segment 3b and slip rate of 6.7 ± 0.6 mm/yr) has an offset of 100.5 km. This distance calculated for the southern NDVFZ segment is within the amount of displacement (40–100 km) obtained in the previously mentioned studies, while the offset for the northern section of the fault is slightly higher. For the SDVFZ, Stewart [1983] suggests that this region was formed after the NDVFZ but before the pull-apart basin ~ 6 Ma because the SDVFZ is involved in the pull-apart basin formation. If we assume that the SDVFZ was created approximately 10–14 Ma and that it has been slipping at a constant rate of 5.7 ± 0.7 mm/yr, then we estimate an offset of approximately 57–80 km. This calculation places our results within the range of displacement (20–80 km) suggested by previous analyses.

[41] Using the horizontal motion of the fault segments obtained by our model, we are able to reconstruct a possible evolution scenario of the DVFZ over the last 6 Myr (Figure 9). The evolution model is reconstructed to 6 Ma because the past 6 Myr of history in the area is better constrained by other geological studies and there is an overall agreement in the proposed models for the evolution of the region [e.g. Burchfiel and Stewart, 1966; Serpa and Pavlis, 1996; Knott *et al.*, 2005]. In this evolution model, we assume that the east side

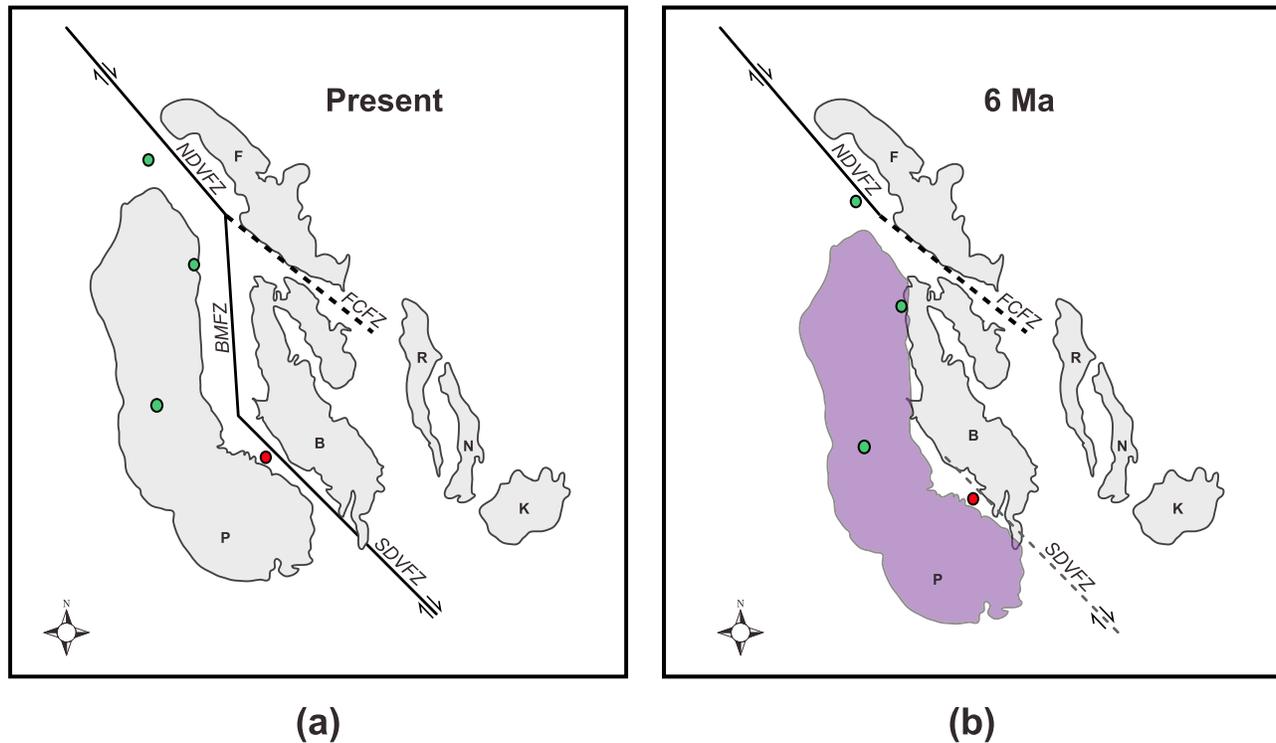


Figure 9. Simplified map of the DVFZ that shows the location of the principal fault zones and describes its evolution over the last 6 Myr. Modified from *Wright et al.* [1991]. The figure labeling is as follows: Black Mountains (B), Funeral Mountains (F), Kingston Range (K), Nopah Range (N), Panamint and Owshead Mountains (P) and Resting Spring Range (R). Here we include the Furnace Creek Fault Zone (FCFZ) for reference although it is not included in this study. Green dots represent the site of the locations taken to reconstruct the motion of the Death Valley fault blocks; red dot shows the location of the cinder cone used to constrain the extension rate of the evolution model.

of the DVFZ remains stationary, while the west side is stepped back in time to the calculated positions using the horizontal velocity field results. We selected four locations (green and red dots in Figure 9) within the Panamint and Owshead Mountains, on the west side of the fault zone, that are then relocated to their previous location based on the strike-slip rates provided in Table 2 and an extension rate of 1 mm/yr. These reference points were selected as representative structures west of the fault zone that can be mapped back to a closed basin with appropriate fault block motion. The extension rate for the reconstruction was obtained from our model results and correlated with rates calculated by *Ganev et al.* [2010] for the northern DVFZ ($\sim 0.7 \pm 0.3$ mm/yr) and other geodetic studies that measured ~ 1 mm/yr of extension [*Bennett et al.*, 2003; *Wesnousky*, 2005]. It is also constrained by dividing the approximate 100 m offset [*Wright and Troxel*, 1984] of a cinder cone (red dot in Figure 8) located in the northern part of the SDVZF, by its age. The age of this cinder cone is poorly resolved due to extensive contamination by extraneous ^{40}Ar , but the best estimate of the ages is between 100 Ka to 300 Ka (T. Pavlis, personal communication, 2010). This calculation yields extension rates of 0.3–1.0 mm/yr, from which we adopt the greater value.

[42] Using this model reconstruction, we interpret the evolution of the DVFZ in its last 6 Myr (Figure 9b), where the Panamint Mountains block and the Black Mountains block close the pull-apart basin. The closing of the pull-apart

basin at this time based on our model-derived rates is an important result because it agrees with the age of the pull-apart basin [*Cemen et al.*, 1982; *Snow and Lux*, 1999]. This age is also correlated by sedimentary records, which support basin sediments no older than this age [*Wright et al.*, 1999; *Knott et al.*, 2005]. Our 6 Ma model is also in agreement with geologic observations of the region that indicate that the area was dominated by transtensional systems during this entire time interval [*Burchfiel and Stewart*, 1966; *Wright et al.*, 1991; *Serpa and Pavlis*, 1996] and with the formation of the Black Mountains turtlebacks (detachment fault surfaces) which provide the best record of the pre-Pliocene extension in the region [*Mancktelow and Pavlis*, 1994]. In addition, our reconstruction model implies movement of the Panamint Mountains block over the Black Mountains fault block. Previous reconstruction models of the DVFZ have the Panamint Mountains fault block overriding the Black Mountains but placing it further to the east around the Resting Spring and Nopah Ranges [*Wernicke et al.*, 1988; *Snow and Wernicke*, 2000]. This disagreement may be caused by the fact that we do not take into account the internal deformation and rotation of the Death Valley fault blocks.

[43] Our evolution model also includes the vertical development of the pull-apart basin in Central Death Valley (Figure 10) that was developed by transtensional systems in the area [*Burchfiel and Stewart*, 1966; *Serpa and Pavlis*,

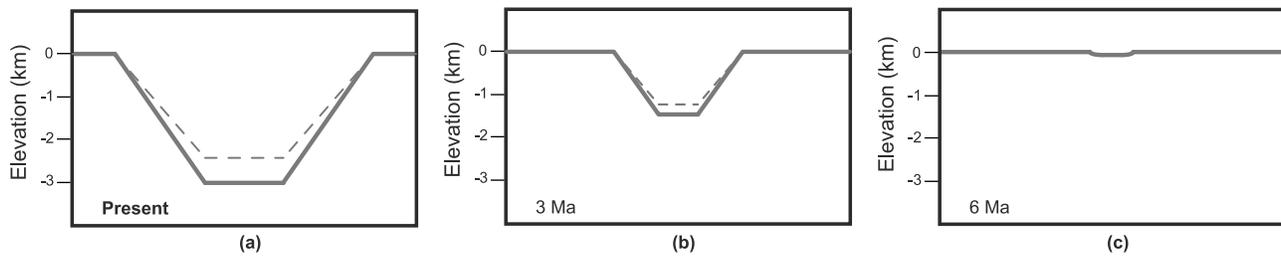


Figure 10. Cross-section schematic of the BMFZ showing the evolution of the pull-apart basin in Central Death Valley from present (Figure 10a) to its formation approximately 6 Ma (Figure 10c). Solid line represents cumulative subsidence based on a secular (constant) rate of 0.5 mm/yr and the dashed line represents cumulative subsidence that accounts for velocity changes throughout the earthquake cycle due to the transient postseismic motions. Note that the horizontal axis is not to scale.

1996] approximately 6 Ma. From its formation up to present-day, the transtensional forces have produced a graben ~ 3 km deep and sediments that have covered most of it [Keener *et al.*, 1993; Blakely *et al.*, 1999]. The pull-apart basin model is generally accepted for the past 6 Myr history of the valley because during that time frame there does not appear to have been a significant change in the regional stress field [Stewart, 1967, 1986; Wright and Troxel, 1967; Serpa and Pavlis, 1996; Snow and Wernicke, 2000]. Our vertical evolution model suggests two important components (coseismic/postseismic and secular) of vertical displacement along the pull-apart basin for each earthquake cycle. During the coseismic and postseismic episode (describing the \sim first 200 years of the earthquake cycle), vertical rates along the fault vary from $+0.25$ mm/yr to -0.5 mm/yr, yielding a net vertical displacement of about 5 mm of uplift. During the remaining 1000 years of the cycle (assuming a 1200 year cycle), the velocity levels out to a constant -0.5 mm/yr, yielding a net vertical displacement of 500 mm. Thus for each earthquake cycle, ~ 495 mm of subsidence is expected, with 1% of this deformation derived from transient motion during the first 200 years of the coseismic and postseismic stages. If we assume a basin age of 6 Myr, with an earthquake recurrence interval of ~ 1200 years, then ~ 2.5 km of subsidence would have resulted from 5,000 earthquake cycles spanning this time period. Alternatively, if we assume a constant subsidence rate of the pull-apart basin over the last 6 Ma, approximated by the secular rate (0.5 mm/yr), we obtain a basin depth of 3 km. Thus transient postseismic motion may reduce the total subsidence of the pull-apart basin by as much as 0.5 km (Figure 10). We note that our model rates likely contain a small source of error due to the omission of low angle normal faulting [Wernicke, 1985; Keener *et al.*, 1993; Mancktelow and Pavlis, 1994], which would increase the rate of subsidence in the basin.

7. Conclusions

[44] In this study, we use a 3-D semi-analytic viscoelastic deformation model and new geodetic data of the Basin and Range to analyze the present-day crustal deformation of the Death Valley Fault Zone. Our results yield apparent locking depths between 9.8–17.1 km, horizontal strike-slip rates of 3.8–6.7 mm/yr, and vertical deformation rates (subsidence) of 0.5–0.8 mm/yr for the DVFZ. These model rates are in good agreement with geologic measurements of fault motion

that imply offsets between 40–100 km in the northern segments, and 20–80 km in the south, and a 7–15 km basin width for the Death Valley pull-apart basin. Coulomb stress and strain accumulation rates range from 0.25–0.49 MPa/100 yr and 82.2–161.7 nstrain/yr, respectively, with the highest values along the southern and the northern sections of the DVFZ. We also calculate seismic moment accumulation rates per unit fault length of $1.4\text{--}2.6 \times 10^{14}$ Nm/100 yr/km and moment magnitude of events that span 6.7–7.7. While seismic hazard models estimate a fairly low probability of earthquake occurrence along the DVFZ in the near term (5–7% for a earthquake of magnitude 6.5+ over the next 30 years), we find that the DVFZ accumulates stress and strain at rates proportional to fault segments of the San Jacinto fault and of the ECSZ. Moreover, although seismic events along the DVFZ have not dominated the historic earthquake record, these results suggest that when a large earthquake does occur along the DVFZ, it has the potential to release large amounts of energy, comparable to some of the greatest earthquakes in California and Nevada over the last 100 years.

[45] We correlate resulting slip rates with a geologic model over the last 6 Myr describing 1) the function, and evolution of the DVFZ and 2) the development of the Central Death Valley pull-apart basin since its formation. Our modeled slip rates for the DVFZ are consistent with geological estimates and provide a basis for constructing an evolution model spanning the last 6 Myr for Death Valley and the pull-apart basin. This evolution model closes the pull-apart basin in 6 Myr by bringing together the Panamint Mountain fault block and the Black Mountains fault block. This model also concurs with the formation age of the Black Mountains turtlebacks [Mancktelow and Pavlis, 1994].

[46] While our model is capable of reproducing the deformation of the DVFZ in most areas, differences between the results obtained here and other analyses may be attributed to both parameterization of the fault model and lack of geodetic data in some regions of the DVFZ, especially along the southern segments. Alternative reconstruction models for the evolution of the DVFZ place the Panamint Mountains fault block overriding the Black Mountains fault block, but these models place its original position further east of the Black Mountains. This discrepancy may be due to fact that we are not including internal deformation and rotation of the Death Valley fault blocks. From this we suggest that further analysis of the fault zone using other modeling techniques

may be required. We are developing a complimentary deformation model of the DVFZ using a finite difference approach, which will enable us to further investigate the rheology and stress behavior of the area along with a dipping fault geometry for strike-slip faults. Moreover, this first order investigation of fault slip rates and apparent locking depths of the DVFZ has provided us with solid geophysical constraints of the present-day motion of the DVFZ enabling the construction of an evolution model that recreates the motion of the fault blocks involved in the deformation and formation of structures in Death Valley.

[47] **Acknowledgments.** We thank the Associate Editor and two anonymous reviewers for their suggestions to help clarify this manuscript. We also thank Terry Pavlis for his useful discussions and for providing a careful in-house review of a draft version of the manuscript and Nicolas Pingitore for his assistance with the statistical calculations. This research was supported by the National Science Foundation (EAR-0847499, EAR-083852 and EAR-0844389) and by the Southern California Earthquake Center. SCEC is funded by NSF Cooperative Agreement EAR-0106924 and USGS Cooperative Agreement 02HQAG0008. The SCEC contribution number for this paper is 1514.

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