Constraint of fault parameters inferred from nonplanar fault modeling

Hideo Aochi and Raul Madariaga
Laboratoire de Géologie, Ecole Normale Supérieure, 24 rue Lhomond, 75231 Paris Cedex 05, France (aochi@geologie.ens.fr; madariag@geologie.ens.fr)

Eiichi Fukuyama
National Research Institute for Earth Science and Disaster Prevention, 3-1 Tennodai, Tsukuba 305-0006, Japan (fuku@bosai.go.jp)

[1] We study the distribution of initial stress and frictional parameters for the 28 June 1992 Landers, California, earthquake through dynamic rupture simulation along a nonplanar fault system. We find that observational evidence of large slip distribution near the ground surface requires large nonzero cohesive forces in the depth-dependent friction law. This is the only way that stress can accumulate and be released at shallow depths. We then study the variation of frictional parameters along the strike of the fault. For this purpose we mapped into our segmented fault model the initial stress heterogeneity inverted by Peyrat et al. [2001] using a planar fault model. Simulations with this initial stress field improved the overall fit of the rupture process to that inferred from kinematic inversions, and also improved the fit to the ground motion observed in Southern California. In order to obtain this fit, we had to introduce an additional variations of frictional parameters along the fault. The most important is a weak Kickapoo fault and a strong Johnson Valley fault.

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1. Introduction

[2] Since the studies of Harris et al. [1991] and Harris and Day [1993], dynamic rupture propagation along nonplanar faults has been simulated using the finite difference method (FDM) [Kase and Kuge, 1998, 2001; Harris and Day, 1999]. Unfortunately, 3-D finite difference can not model fault branching, fault curvature and other geometrical complexities without major changes in its current formulation. Thanks to rapid progress in the development of boundary integral equation methods (BIEM), originally proposed by Koller et al. [1992], Cochard and Madariaga [1994] and Fukuyama and Madariaga [1995], it is now possible to model complex fault systems consisting of
several subfaults, including branched and bent faults [Kame and Yamashita, 1997; Tada and Yamashita, 1997; Aochi et al., 2000a]. These methods can also be applied to study spontaneous crack propagation in intact material [Kame and Yamashita, 1999]. Based on these recent studies, Aochi and Fukuyama [2002] modeled the 1992 Landers earthquake, using a realistic nonplanar fault geometry learned from field observations, a loading system (remote tectonic stress) derived from geological considerations, and a depth-dependent slip-weakening law. In this paper, we investigate the initial condition of the friction law for the simulation of the Landers earthquake.

[3] For the purpose of understanding earthquakes, we have to study the role of the friction law that controls rupture process. Since rupture process are very complex and spatially heterogeneous, frictional parameters or stress field are probably also very heterogeneous. We know from fracture physics that spontaneous rupture propagation requires stress release in order to propagate, so that we need an opposite mechanism in order to stop the rupture process. The effect of frictional parameters are well understood in numerical experiments [Boatwright and Cocco, 1996]. On the other hand, it is very difficult to determine them quantitatively for a real earthquake in the field. For the Landers earthquake, the stress change during the earthquake was very heterogeneous along the fault as determined from kinematic fault models [Wald and Heaton, 1994; Bouchon et al., 1998a; Day et al., 1998] and dynamic modeling and inversion [Olsen et al., 1997; Peyrat et al., 2001]. The relation between shear stress and fault movement was also determined from kinematic inversions of the 1995 Hyogoken-nanbu, Japan, earthquake [Ide and Takeo, 1997; Guatteri and Spudich, 2000; Guatteri et al., 2001]. But according to these authors, the resolution of these estimations is limited because of the finite frequency band used in the kinematic inversion. Usual kinematic inversions can only determine relative dynamic stress changes during rupture. In recent works, there have been attempts to investigate the absolute level of stress field [Bouchon et al., 1998b; Spudich et al., 1998]. One of the results that appears from these works is that the accumulated stress is much less than the stress extrapolated from laboratory experiments.

[4] In spite of the difficulty and uncertainty of the estimation of fault parameters, we absolutely need them for reproducing the rupture branching phenomena in numerical simulations. In this paper, we focus on the question of how to constrain them from observations of rupture of the Landers earthquake. As shown by Aochi et al. [2000a, 2000b], the absolute level of stress and frictional parameters appear explicitly in nonplanar fault systems, because the external tectonic forces produces heterogeneous stress distribution depending on fault orientation (strike). In a recent simulation, Aochi and Fukuyama [2002] succeeded in reproducing the general rupture propagation of the Landers earthquake without any horizontal heterogeneity of fault parameters. Rupture complexity was only due to a heterogeneous stress field produced by the assumed tectonic loading forces. In the previous model [Aochi and Fukuyama, 2002], an external loading force, whose direction changed regionally according to local tectonics, produced heterogeneity not in fault parameters but in initial stress field along the fault system. That model successfully reproduced realistic rupture transfer between faults. For instance, rupture progressed not along the northern Johnson Valley fault, but propagated into the Kickapoo and the Homestead Valley faults, as shown in the map of Figure 1.

[5] In this paper we will test several rupture scenarios for the Landers earthquake. In a first step we consider the variation of the friction law with depth. As proposed by Scholz [1988], cohesive force is usually assumed to be zero at the ground surface in some seismic rupture simulations [Yamashita and Ohnaka, 1992]; whereas other simulations did not assume any depth-dependency [Olsen et al., 1997; Peyrat et al., 2001]. Here we will discuss the question of how cohesive force may affect rupture propagation and strong ground motion. In the second part, we will discuss the horizontal heterogeneity of frictional parameters. We first assume a uniform external load (remote tectonic stress). We will investigate what kind of heterogeneity of fault parameters is required so that rupture chooses the
correct fault branches, and we will compare with the previous heterogeneous external load case [Aochi and Fukuyama, 2002]. We finally study the mapping of the heterogeneity of initial stress obtained from a planar fault inversion [Peyrat et al., 2001] onto our segmented fault model, and compare rupture propagation and seismic radiation.

2. The Nonplanar Fault Model of Aochi and Fukuyama [2002]

[s] For planar fault simulations, one needs to specify the initial shear stress on the fault and its constitutive parameters [e.g., Olsen et al., 1997]. For nonplanar fault modeling, on the other hand, we need absolute stress values [Aochi et al., 2000a]. Aochi and Fukuyama [2002] constructed a nonplanar fault model based on the surface fault traces [Hart et al., 1993] shown in Figure 1. It is important to remark that the model included both the faults that did break during the 1992 rupture and those that did not; so that the model allowed the rupture to progress along any fault segment. The fault segments that the rupture chose depended on the initial stress field and fault properties.

Figure 1. Fault model (red lines) studied by Aochi and Fukuyama [2002]. Black lines are the trace of observed active faults [Hart et al., 1993]. Arrows represent the assumed directions of the maximum principal stress \(\sigma_1\) (red, spatially rotating principal stress assumed by Aochi and Fukuyama [2002]; blue, uniform stress field used in this study).
For the initial stress field, Aochi and Fukuyama [2002] assumed an external load (remote tectonic stress) based on geological studies of local tectonics [Dokka and Travis, 1990; Unruh et al., 1994; Sowers et al., 1994]. The red arrows in Figure 1 shows the maximum principal stress $\sigma_1$ assumed in their simulation. The direction of $\sigma_1$ points in the NNE direction except in the southern part of the rupture area (Johnson Valley, Kickapoo, and southern Homestead Valley faults) where it is closer to the NE direction. The depth dependence of the assumed external load is shown in Figure 2c. As confining pressure increases with depth, all components of the principal stress are assumed to increase proportionally to depth. Initial shear stress $\tau_0$ and normal stress $\sigma_n^0$ are then given by

$$\tau_0 = \frac{\sigma_1 - \sigma_3}{2} \sin(2\theta) \tag{1}$$
$$\sigma_n^0 = \frac{\sigma_1 + \sigma_3}{2} - \frac{\sigma_1 - \sigma_3}{2} \cos(2\theta) \tag{2}$$

where $\theta$ is the angle of the fault strike with respect to the direction of $\sigma_1$.

Total shear and normal stresses ($\tau_{\text{total}}$ and $\sigma_n^{\text{total}}$) at each point on the fault are written in the following,

$$\tau_{\text{total}}(\xi, t) = \tau_0(\xi) + \Delta \tau(\xi, t) \tag{3}$$
$$\sigma_n^{\text{total}}(\xi, t) = \sigma_n^0(\xi) + \Delta \sigma_n(\xi, t) \tag{4}$$

where $\xi$ and $t$ are position and time. $\Delta \tau$ and $\Delta \sigma_n$ represent increment stress during dynamic rupture process and are in the form of integral equation, a convolution of Green’s function and slip velocity on the fault, calculated based on BIEM [Aochi et al., 2000a]. At the beginning ($t = 0$), $\Delta \tau$ and $\Delta \sigma_n$ equal to zero, whereas $\tau_0$ and $\sigma_n^0$ are given in equations (1) and (2). It should be emphasized that shear and normal stresses in equations (3) and (4) are defined with respect to the slip vector on the fault (strike parallel direction). Other four stress components can be calculated in the same way as in Aochi et al. [2000a].

For the fault parameters, Aochi and Fukuyama [2002] introduced a depth-dependent slip-weakening friction law, shown in Figure 2. When the applied shear stress $\tau$ overcomes peak strength $\sigma_p$, rupture starts following a slip-weakening relationship between fault function $\sigma$ and fault slip $\Delta u$,

$$\sigma(\Delta u) = \tau_r + (\sigma_p - \tau_r) \left( 1 - \frac{\Delta u}{\Delta u_c} \right) H \left( 1 - \frac{\Delta u}{\Delta u_c} \right) \tag{5}$$
Here $\tau_r$ is residual strength and $D_c$ is critical slip-weakening distance. Breakdown strength drop $\Delta \tau_b$ is defined by $(\sigma_p - \tau_r)$. $H(\cdot)$ represents the Heaviside function. The slip-weakening friction law was proposed theoretically and numerically [Ida, 1972; Palmer and Rice, 1973], it was then experimentally observed [Okubo and Dieterich, 1984; Ohnaka et al., 1987], modeled theoretically [Matsu’ura et al., 1992], and inferred from seismological modeling of actual earthquake ruptures [Ida and Takeo, 1997; Olsen et al., 1997; Guatteri and Spudich, 2000]. As clearly seen in the definition of equation (3), this criterion is basically combined with total shear stress $\tau^{\text{total}}$ in equation (5). In our previous paper [Aochi and Fukuyama, 2002] and this study, all parameters, $\tau_r$, $\sigma_p$ and $D_c$, are supposed to be temporally invariable, so that it is enough to follow numerically shear stress increment $\Delta \tau$ for practical use. On the other hand, it is possible to introduce a much more complex criterion instead of equation (5). Aochi et al. [2002] investigated a dynamic Coulomb law, whose parameters are temporally variable according to normal stress increment $\Delta \sigma_n$. In that case, equation (5) must be combined with both equations (3) and (4) at the same time. However, they reported that increment in normal stress $\Delta \sigma_n$ is generally much smaller than its absolute normal stress level $(\sigma_n^0)$ when we consider a realistic situation of high confining pressure in the crust, and that, as a result, the frictional parameters do not change drastically with time. That is why, in this paper, we will suppose each frictional parameter is temporally invariable, and consider the effect of their spatial heterogeneity on rupture process.

The product of $\Delta \tau_b$ and $D_c$ determine the fracture energy, a parameter that is the easiest to invert from seismic observations [Aki, 1979; Guatteri and Spudich, 2000; Peyrat et al., 2002]. The absolute level of stress in the friction law, $\sigma_p$ and $\tau_r$, are very important because, for an assumed external load (remote tectonic stress), we need the absolute stress field on any segmented, nonplanar fault system. However, the estimation of absolute value of fault parameters is still unsolved. Some seismological analyses proposed that they are much lower than that extrapolated from experimental results [Bouchon et al., 1998; Spudich et al., 1998], as well as other geological observation also suggested low stress along the San Andreas fault. Regardless of the uncertainty, the depth variation of the frictional parameters is often used in the simulation, based on the rheology due to the pressure and temperature with depth, as modeled in Sibson [1982], Scholz [1988] and Yamashita and Ohnaka [1992]. Above the depth of 12 km, the peak strength $\sigma_p$ as a function of depth $z$ is given by

$$\sigma_p(z) = \sigma_0 + \mu_f \times (P(z) - P_H(z)),$$

where $\mu_f$ is frictional coefficient, $P$ and $P_H$ are the confining pressure and hydrostatic pressure, respectively. $\sigma_0$ is the cohesive force, but $\sigma_0 = 0$ was assumed in the simulation by Aochi and Fukuyama [2002].

Aochi and Fukuyama [2002] modeled the rupture process of the Landers earthquake using a numerical boundary integral equation method (BIEM) for nonplanar faults embedded in a 3D unbounded, homogeneous elastic medium [Aochi et al., 2000a]. Time step and square grid size were taken as 0.06 s and 750 m, respectively; and P- and S-wave velocities were 6.20 and 3.52 km/s, respectively. We used mirror sources for approximating the effect of the free surface. A calculation of 400 time steps took about $4 \times 10^5$ s of CPU time using 8 to 16 CPUs with fortran90 and MPI on a COMPAQ ES40 Cluster (EV6 500 MHz), although we occasionally ran several jobs for one simulation.

In Animation 1 (available in the HTML version of the article at http://www.g-cubed.org) and Figure A1a, we show a movie of one of the dynamic simulations by Aochi and Fukuyama [2002]. We observe that rupture does not propagate on the northern Johnson Valley fault, but chooses the Kickapoo and Homestead Valley faults instead, and then jumps to the northernmost Camp Rock fault. This is the most important feature of the Landers earthquake that they succeeded in reproducing. In the following sections, we will investigate how important were the assumptions they made for rupture propagation, and then discuss...
how we may constrain the fault properties from the numerical simulations.

3. Constraint of the Depth-Dependent Friction Law

[13] Although the segmented model of Aochi and Fukuyama [2002] reproduced the general features of rupture propagation along the fault system, there are some clear discrepancies between their model and observations. In their simulations, there were several large slip areas along strike in overall agreement with the asperities inferred from kinematic inversion [Wald and Heaton, 1994], but the maximum slip was located at a depth of around 12 km. Thus, their model could not produce large slip near the ground surface, although a slip of more than 5 m was observed in the field [Hart et al., 1993]. The discrepancy is due to the depth dependency of friction assumed in the simulation, the depth of 12 km corresponds to the depth where the breakdown strength drop is maximum, as shown in Figure 2. Clearly the assumption of zero peak strength at the ground surface is incorrect. Thus, we have to modify the depth variation of fault properties adapting finite cohesive force $\sigma_0$ in equation (6), so that we expect that finite stress could be accumulated and released near the ground surface to produce much fault slip.

[14] As discussed earlier, this is in agreement with many observations of the Landers earthquake. For example, the inversion result by Bouchon et al. [1998] showed more than 30 MPa static stress drop at the depth of 4 km. The numerical simulation by Olsen et al. [1997] and Peyrat et al. [2001] required more than 10 MPa breakdown strength drop at the ground surface as well as in the deeper crust. Figure 3 shows the depth-variation of fault parameters we assumed in the numerical simulation for different cohesive forces. In the deeper part of the fault, we assumed a slip-hardening friction law instead of a slip-weakening law. We assumed a residual stress level $\tau_r$ equal to 0 at the surface, and breakdown strength drop $\Delta \tau_b$ to be 20 MPa at the depth of 12 km regardless of the values of $\sigma_0$. Friction coefficient $\mu_f = 0.6$ and hydrostatic pressure were assumed.

[15] Let us first compare the dynamic rupture simulations with $\sigma_0 = 0, 5, 10,$ and 12.5 MPa. Figure 4 shows the final slip distribution on all segments of the Landers earthquake for each value of $\sigma_0$. We observe that slip at shallow depth increases as the cohesive force $\sigma_0$ increases. Fault slip of more than 5 m was observed on the Homestead Valley and the Camp Rock faults [Hart et al., 1993], so that in order to explain the surface fault slip, breakdown strength drop must be larger than 5 MPa. In our simulations, large slip areas appear near the ground surface, while artificial maximum slips at depth disappear.

[16] In Figure 5, we show synthetic seismograms computed using the discrete wave number method (AXITRA) for the same crustal structure as in Olsen et al. [1997] and Aochi and Fukuyama [2002]. For each simulation, we show synthetics at station YER which is located in the forward direction of rupture propagation, and where large amplitudes were observed. We observe that the amplitude of the seismogram improves as the cohesive force increases and the fault slip near surface gets larger. We also observe that the width of the pulse becomes narrower as cohesive force increases. This is because cohesion at shallow depths causes large fault slip and fast rupture velocities.

[17] We conclude that a finite cohesive force of more than 5 MPa is required in order to explain the vertical heterogeneity of slip distribution, especially the large slip near surface. Our simulation results indicate the possibility that $\sigma_0$ may be as large as 10 MPa, since the synthetic seismograms for this case fits the observed results better than those of the other values of $\sigma_0$ (Figure 5). $\sigma_0 = 10$ MPa is consistent with the value of 12.5 MPa determined by Olsen et al. [1997] and Peyrat et al. [2001].

4. Constraints on Horizontal Heterogeneity

4.1. Uniform External Load

[18] Choosing the initial stress field has many degrees of freedom, although the simple assumption in Aochi and Fukuyama [2002] was reasonably based on previous geological studies. They found
that a heterogeneous external load (remote tectonic stress) in the southern and northern parts of the model region was essential for the correct fault selection of the Kickapoo and Homestead Valley faults. The selectivity of rupture along a branching fault system is actually very sensitive to the conditions on each branch [Aochi et al., 2000b, 2002].

Before we discuss stress or frictional heterogeneity, let us examine the role of a hypothetical uniform external load on the Landers earthquake propagation. We assume that the direction of maximum uniform principal stress $\sigma_1$ is NNW, as shown by the blue arrows in Figure 1 which produces heterogeneous stress field on the fault as shown in Figures 6a and 6b. This principal direction was inferred from the focal mechanisms of the seismicity before and after the 1992 Landers earthquake [Hauksson, 1994]. We assumed the same frictional parameters as Aochi and Fukuyama [2002], that is, equation (6) with $\sigma_0 = 0$. In this uniform external load, we found that the rupture did not jump to the Kickapoo fault, but continued along the Johnson Valley fault as shown in Animation 2 or Figure A1b, causing a completely different history of rupture propagation from the one that was actually observed. We get a maximum fault slip of 5.0 m and a seismic moment of $3.4 \times 10^{19}$ N-m. Thus, under a uniform external load, we have to assume that the Kickapoo and/or Home-
stead Valley faults are weaker than the northern Johnson Valley fault in order to reproduce the correct rupture path.

4.2. Heterogeneity Produced by a More Complex Friction Law

For the purpose of modeling a weak Kickapoo fault and a strong northern Johnson Valley fault, let us write the frictional parameters $s_p$ and $t_r$ in the following way:

$$s_p(\xi) = \mu_s \sigma^0_n(\xi)$$  \hspace{1cm} (7)

$$t_r(\xi) = \mu_d \sigma^0_n(\xi)$$  \hspace{1cm} (8)

where $\mu_s$ and $\mu_d$ are static and dynamic friction coefficients, and $\sigma^0_n$ is the applied normal stress. Hereafter $\mu_s = 0.6$ and hydrostatic pressure are assumed. Now the frictional parameters, $s_p$ and $t_r$, are functions of the fault position $\xi$. The initial stress field is shown in Figures 6a and 6b, and possible stress drop in Figure 6c. Normal stress is much lower on the Kickapoo fault than on the northern Johnson Valley fault (Figure 6b), so that it implies a “weak” Kickapoo fault following equation (7). However, the accumulated shear stress (Figure 6a) is also low on the Kickapoo fault, so that it is not possible to produce a positive stress drop on this fault as shown in Figure 6c. Thus rupture cannot propagate along the Kickapoo and Homestead Valley faults in this case.

This feature is very interesting. The Kickapoo fault is relatively weak because of a reduced normal stress, but shear stress is not enough to produce slip due to the unfavorable orientation of fault strike with respect to the external load $\sigma_1$. This difficulty does not go away even if we introduce the normal stress dependency dynamically into frictional parameters; $s_p(\xi, t) = \mu_s \sigma_n(\xi, t)$ and $t_r(\xi, t) = \mu_d \sigma_n(\xi, t)$, where the fault parameters are variable with time according to rupture propagation.

Thus we added further heterogeneity of the initial stress field as shown in Figure 7:

$$t_r(\xi) = \mu_d \sigma^0_n(\xi) + \alpha(\xi).$$  \hspace{1cm} (10)

The heterogeneity introduced in equations (9) and (10) is completely different and independent from the one originally included in equations (7) and (8).

Animation 3 shows rupture propagation in this case. Rupture successfully propagates from the Johnson Valley fault to the Camp Rock fault (See also Figure A1d). Besides we observe that rupture transfers through a jog from the Homestead Valley fault to the Emerson fault in contrast to the previous examples without normal stress dependency (Their comparison is shown in Figure A1). There we did not introduce any supplemental
heterogeneity. As a result, we get a maximum fault slip of 3.67 m and a seismic moment of $5.2 \times 10^{19}$ N·s.

[24] In conclusion, in order to reproduce the rupture transfer between different fault branches, especially from the Johnson Valley to the Kickapoo and Homestead Valley faults, we needed an additional heterogeneity $\alpha(\xi)$ which explicitly indicates a weak Kickapoo fault and a strong northern Johnson Valley fault. This seems to be very unlikely. Rockwell et al. [2000] investigated paleoseismic data from several trenches and reported that the time intervals from previous event for the southern and northern Johnson Valley and Kickapoo faults were about 5000 years, while those of the Homestead Valley, Emerson and Camp Rock faults were more than 7000 years.

4.3. Stress Heterogeneity Derived From Planar-Fault Simulations

[25] Olsen et al. [1997] and Peyrat et al. [2001, 2002] successfully determined the heterogeneous initial stress field of the Landers earthquake for a single planar fault through the inversion of the observed ground motion. Since they assumed a single planar fault, the heterogeneity of the stress field can be transferred into the heterogeneity of frictional parameters as long as the amount of available energy to fracture energy is the same in the two models [Peyrat et al., 2002]. That is, a certain relation between $\tau_0$, $\sigma_p$ and $\tau_r$ that is required stress excess $\Delta \tau_{\text{planar}} (\equiv \sigma_p - \tau_0)$ and possible stress drop $\Delta \tau_{\text{planar}} (\equiv \tau_0 - \tau_r)$ is conserved. In the following, we consider how to map their heterogeneity into our nonplanar fault modeling.

[26] Figure 8 briefly shows the procedure we propose in this section. We start with the uniform external load (remote tectonic stress) as assumed in section 4.1, and indicated by the blue allows in Figure 1. For the purpose of determining the final slip distribution during the earthquake, static stress drop plays a fundamental role. We assume possible stress drop $\Delta \tau$ based on the planar-fault
inversion of Peyrat et al. [2001], as shown in Figure 8 and Figure 9. Thus, we define $\tau_r$ by the following expression:

$$\tau_r(\vec{\xi}) = \tau_0(\vec{\xi}) - \Delta \tau_{\text{planar}}(\vec{\xi})$$  \hspace{1cm} (11)$$

where $\tau_0$ is an assumed uniform external load (remote tectonic stress) and $\Delta \tau_{\text{planar}}$ a heterogeneous stress change inverted from the planar model [Peyrat et al., 2001]. They are functions of fault position $\vec{\xi}$, strike as well as depth. We implicitly prohibit negative friction, so that we set $\tau_r=0$ in that case.

[27] Although it is possible to use the same peak strength $\sigma_p$ as in equation (6), we already know
that this did not work in the case of the uniform external load. Thus, in order for rupture to progress correctly, we also have to change peak strength $\sigma_p$ – or strictly speaking, the stress increase ($\Delta\tau_e$) from the assumed initial stress field ($\sigma_p - \tau_0$). Actually, we not only give the stress increase $\Delta\tau_e^{planar}$ of the planar fault simulation [Peyrat et al., 2001], but also a small additional heterogeneity $\beta(\xi)$.

$$\sigma_p(\xi) = \tau_0(\xi) + \Delta\tau_e^{planar}(\xi) + \beta(\xi)$$ (12)

That is necessary because, without a small $\beta(\xi)$, the rupture does not propagate. In the simulations the

Figure 7. Initial conditions with additional heterogeneity to that in Figure 6. (a) Required stress increase, and (b) possible stress drop. Artificial heterogeneity is given independently from the Coulomb friction law. In this case, the Kickapoo fault is weak enough compared to the applied initial shear stress.
rupture never reached the branching point of the Kickapoo fault, because the nonplanarity of the fault perturbs the stress field and actually reduces shear stress with respect to the planar case. Thus we had to increase the available potential strain energy release by increasing $b(\xi)$. An even more critical problem in nonplanar fault modeling is that we need to introduce additional information so that rupture chooses the Kickapoo fault when it reaches the branching point. As shown in the previous section 4.1, when the frictional parameters ($\sigma_p$ and $\tau_r$) are the same on both subfaults, rupture clearly prefers the northern Johnson Valley fault. Thus, in order to obtain the correct fault selection, a weak Kickapoo fault is necessary. As a result, we assume, in the expression of $b(\xi)$, that the northern Johnson Valley fault is strong, and that the Kickapoo, southern Homestead Valley and Camp Rock faults are weak. Now we find that frictional parameters $\sigma_p$ and $\tau_r$ are different from equations (7) and (8). Furthermore, since we give derivative additional stress $\Delta \tau^\text{planar}$ and $\Delta \tau^\text{e,planar}$ in equations (11) and (12), regional stress given by external load is hidden behind the heterogeneity inverted from the planar fault simulation. Thus we can actually use an almost arbitrary initial stress $\tau_0$, otherwise it is also possible to begin with $\sigma_p$ given by equation (7) depending on applied normal stress, and constrain the other parameters $\tau_0$ and $\tau_r$ with the deviations $\Delta \tau^\text{planar}$ and $\Delta \tau^\text{e,planar}$.

[28] Animation 4 (see also Figure A1d) shows rupture propagation for this model. Rupture successfully propagates along the Kickapoo and Homestead Valley faults, then it transfers to the last segment, the Emerson and Camp Rock faults, and finally it is gradually arrested. The process of rupture propagation is very similar to that of Wald and Heaton [1994] and Peyrat et al. [2001]. However, even though the stress drop $\Delta \tau^\text{planar}$ is the same as theirs, the amount of final slip in our simulation is smaller due to the perturbation of the stress field by the geometry of the fault system. The maximum fault slip is 3.6 m, while theirs reached 6 m. We got a seismic moment of $5.3 \times 10^{19}$ N-m, which is somewhat less than the $7.0 \times 10^{19}$ N-m obtained in the planar simulation [Peyrat et al., 2001].

[29] In Figure 10, we compare synthetic seismograms obtained from our simulation with the
observed ground displacement records and those produced by the previous model of Animation 1 or Figure A1d [Aochi and Fukuyama, 2002]. We show recordings at the YER, SVD, and PFO stations which are located in different directions from the fault. At YER, located in the forward direction of rupture propagation, the amplitude of the synthetics does not fit the observation well, because, as we explained above, fault slip is still underestimated in our model. Ground motion at

Figure 9. Initial stress distribution constrained by the mapping of the planar fault simulation [Peyrat et al., 2001]. (a) Initial shear stress produced by the uniform external loading system. (b) Possible stress drop, the difference between the initial shear stress and the residual stress drop ($\tau_0 - \tau_r$), mapped from the planar-fault simulation. (c) Required stress increase, the difference between the peak strength and the initial shear stress ($\sigma_p - \tau_0$), obtained after adding artificial heterogeneity.
YER is affected by directivity much more than by the details of rupture propagation. On the other hand, at SVD and PFO where rupture directivity is small, we observe that the synthetics have improved from the previous simulation [Aochi and Fukuyama, 2002] and become similar to the observed records. That is because we constrained stress drop ($\tau_0 - \tau_r$) so that the final slip distribution has significantly improved (see also Figures A1a and A1d).

5. Summary

[30] In this paper, we tried to determine frictional parameters for the dynamic rupture simulation of the 1992 Landers earthquake along a nonplanar fault system. We found that a finite cohesive force $\sigma_0$ (order of 10 MPa) is required in order to reproduce large slip areas near the surface. Otherwise, stress is never accumulated or released around the surface, so that it inhibits large slip near the surface which is sharp contrast with geological and geodetic observations.

[31] Modeling heterogeneity along fault strikes is much more difficult. We found that the heterogeneity produced by simple variations of normal stress derived from simple tectonic stress models could not explain the observations without an additional heterogeneity of internal origin. Then we observed that the stress heterogeneity inferred by the planar fault inversions [Peyrat et al., 2001] improved the details of the rupture process and the fit to the seismograms observed around the fault system. In every case, we needed to adjust frictional parameters such as a weak Kickapoo fault and a strong northern Johnson Valley fault in order to obtain the correct selection of rupture propagation. We remark that the improvement obtained by the planar fault modeling is independent of the mathematical expressions, equations (6), (7), and (8), assumed in the other part of this study. Actually, we do not need much information about absolute stress, equations (1) and (2), since we use only the deviation of stress. It is clear that the state of regional stress is hidden by the heterogeneity inverted from the planar fault models. It will be a key point whether or not we...
can find this kind of information for realistic rupture models.

Appendix: Snapshots of Dynamic Simulation

Simulation results are given as movie files in this manuscript. In Figure A1, we further show their snapshots in static images for convenience.

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