Earthquake activity related to seismic cycles in a model for a heterogeneous strike-slip fault

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Abstract

We investigate the evolution of seismicity within large earthquake cycles in a model of a discrete strike-slip fault in elastic solid. The model dynamics is governed by realistic boundary conditions consisting of constant velocity motion of regions around the fault, static/kinetic friction and dislocation creep along the fault, and 3D elastic stress transfer. The fault consists of brittle parts which fail during earthquakes and undergo small creep deformation between events, and aseismic creep cells which are characterized by high ongoing creep motion. This mixture of brittle and creep cells is found to generate realistic aftershock sequences which follow the modified Omori law and scale with the mainshock size. Furthermore, we find that the distribution of interevent times of the simulated earthquakes is in good agreement with observations. The temporal occurrence, however, is magnitude-dependent; in particular, the small events are clustered in time, whereas the largest earthquakes occur quasiperiodically. Averaging the seismicity before several large earthquakes, we observe an increase of activity and a broadening scaling range of magnitudes when the time of the next mainshock is approached. These results are characteristics of a critical point behavior. The presence of critical point dynamics is further supported by the evolution of the stress field in the model, which is compatible with the observation of accelerating moment release in natural fault systems.

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1. Introduction

The complexity of spatiotemporal earthquake occurrence on natural fault systems has been discussed and quantified in numerous studies (for an overview see Rundle et al., 2000b). While some seismicity patterns show an almost universal behavior, others are observed less frequently and follow no obvious law. For example, aftershocks according to the modified Omori law are observed after almost all large earthquakes in continental lithosphere (Utsu et al., 1995; Utsu, 2002), whereas
foreshocks are a rare phenomenon (Wyss, 1997). Moreover, the Parkfield experiment in California (Roeloffs and Langbein, 1994) demonstrates that even if a fault segment shows relatively regular behavior over several decades, a prediction of future activity can still fail. One reason for this is that the available data sets are too limited to provide enough information for the understanding of the complex relationships between the various physical mechanisms in the earth’s crust. This emphasizes the importance of model simulations which cover many seismic cycles and allow to study the dependences of seismicity on the underlying parameters and evolving stress field.

Several conceptual models have been proposed to describe properties of observed seismicity, e.g. spring-block models and cellular automata, which reproduce the frequency–size distribution of earthquakes. These models are mainly based on tectonic loading and the coseismic stress transfer (Burridge and Knopoff, 1967; Bak and Tang, 1989). Others have implemented additional mechanisms, like viscoelastic relaxation in the fault zone (Dieterich, 1972; Hainzl et al., 1999), fault strengthening or weakening after a block slip (Ito and Matuzsaki, 1990), pore fluid flow (Nur and Booker, 1972), rate-state friction (Dieterich, 1994) and damage rheology (Ben-Zion and Lyakhovsky, 2006; Shcherbakov and Turcotte, 2004). Although most of these models reproduce certain observed phenomena, the underlying frameworks are often abstract and the application to an actual fault system with its spatiotemporal complexity of earthquake occurrence remains questionable.

In this study, we analyze earthquake catalogs generated by the model of Zöller et al. (2005a) for a large heterogeneous strike-slip fault. The model develops further the framework of Ben-Zion and Rice (1993) and Ben-Zion (1996) for a discrete fault in a surrounding elastic solid, and it combines computational efficiency with realistic physical properties. The simulations cover 1000s of years and allow us to reproduce brittle as well as postseismic and interseismic creep deformation. The model ingredients, including laws for brittle and creep deformation, stress transfer, and boundary conditions, are compatible with empirical knowledge. Several relationships between imposed model parameters (e.g., frictional parameters, creep velocities, spatial heterogeneities) and observed seismicity quantities like frequency–size distributions and aftershock clustering, have been previously quantified (Zöller et al., 2004, 2005a,b). In this paper, we will focus on the temporal evolution of seismicity during seismic cycles. Many observational and theoretical studies have shown that seismicity is largely time-dependent. In particular, the seismic moment release appears to accelerate prior to some large earthquakes (Bufo and Varnes, 1993; Bowman et al., 1998; Jaumé and Sykes, 1999). This phenomenon suggests that seismicity is a process that may be characterized by critical point behavior in terms of upcoming long-range correlations prior to a large earthquake (Sornette and Sornette, 1990; Zöller et al., 2001; Zöller and Hainzl, 2002). Analyzing several stress and seismicity functions over seismic cycles, Ben-Zion et al. (2003) conclude that mainshock occurrence is associated with a period where the stress-field evolves toward a critical level of disorder. In this state, the stress field heterogeneities are characterized by many size scales and a brittle failure can evolve into a large earthquake. It is, therefore, an important question whether the evolution of stress and seismicity in our model reflects such critical point behavior.

2. Model

In this section, we give a brief description of the fault model. Further details can be found in Zöller et al. (2004, 2005a,b) and references therein.

Our model includes a single rectangular fault embedded in a 3D elastic half space (Fig. 1). A fault region of 70 km length and 17.5 km depth is covered by a computational grid, divided into 128 × 32 uniform cells, where deformational processes are calculated. Tectonic loading is imposed by a motion with constant velocity \( v_{pl} = 35 \text{ mm/year} \) of the regions around the computational grid. The space-dependent loading rate provides realistic boundary conditions. Using the static stress transfer function for slip in elastic solid, the continuous tectonic loading for each cell on the computational grid is a linear function of time and plate velocity \( v_{pt} \). Additional loadings on a given cell occur due to brittle and creep failures on the fault. While the loading produces an increase of stress on the fault, the local stress may be reduced by interseismic creep and coseismic slip. As in Zöller et al. (2005a), all cells on the computational grid undergo creep deformation, while brittle deformation during an earthquake is only possible for “brittle cells” (white cells in Fig. 1). The aseismic “creep cells” (black cells in Fig. 1) are generated by near-vertical random walks from the free surface to depth accounting for offsets and step-over regions in the fault zone.

The ongoing creep motion on the grid is implemented by a local constitutive law corresponding to lab-based dislocation creep (Ben-Zion, 1996):

\[
\dot{u}_{\text{creep}}(x,z,t) = c(x,z)\tau(x,z,t)^3, \tag{1}
\]
Earthquake is thus given by $c(x, z; t)$ is the local stress at time $t$, and $c_t(x, z)$ and $c_b(x, z)$ are time-independent coefficients, which are different for the creep cells and the brittle cells. As shown in Zöller et al. (2005a), this design can produce aftershock sequences, where the ratio of the mean creep coefficients in the creep cells and in the brittle cells $\frac{c_t}{c_b}$ determines the exponent $p$ of the modified Omori law for the rate $n(t)$ of aftershocks:

$$n(t) = \frac{c_1}{(c_2 + t - t_M)^p}$$

where $u_{\text{creep}}(x, z, t)$ is the creep velocity of the cell with coordinates $(x, z)$, $\tau(x, z, t)$ is the local stress at time $t$, and $c(x, z)$ are time-independent coefficients, which are different for the creep cells and the brittle cells. As shown in Zöller et al. (2005a), this design can produce aftershock sequences, where the ratio of the mean creep coefficients in the creep cells and in the brittle cells $\frac{c_t}{c_b}$ determines the exponent $p$ of the modified Omori law for the rate $n(t)$ of aftershocks:

$$n(t) = \frac{c_1}{(c_2 + t - t_M)^p}$$

In (2), $c_1$ and $c_2$ are time-independent numbers, and $t_M$ is the mainshock occurrence time. For this work, we choose creep coefficients leading to $p=1$: The creep parameters for the brittle cells $c_b$ are randomly distributed in the interval $[0.9(c_b); 1.1(c_b)]$, where $c_b = 10^{-7}$ ms$^{-1}$ MPa$^{-3}$. The values for the creep parameters $c_c$ are calculated accordingly with $c_c = 10^{-2}$ ms$^{-1}$ MPa$^{-3}$. These choices correspond to model A in Zöller et al. (2005a). Eq. (1) results in a system of $128 \times 32$ coupled ordinary differential equations, which is solved numerically using a Runge–Kutta scheme.

The coseismic processes are governed by static/kinetic friction. An earthquake is initiated if the local stress $\tau(x, z; t)$ exceeds the static friction $\tau_s(x, z)$. Then the stress drops in cell $(x, z)$ to the arrest stress $\tau_d(x, z)$ and the strength drops to the lower dynamic value $\tau_d$. The brittle failure envelope $\tau_d(x, z, t)$ during an earthquake is thus given by:

$$\tau_d(x, z, t) = \begin{cases} \tau_s(x, z) : & \text{cell } (x, z) \text{ failed not during this event} \\ \tau_d(x, z) : & \text{cell } (x, z) \text{ already failed during this event} \end{cases}$$

If $\tau(x, z; t) < \tau_d(x, z; t)$ for all cells on the grid, the earthquake is terminated. Then, the brittle strength recovers back to the static level for all cells: $\tau_f(x, z; t) = \tau_s(x, z)$. The dynamic friction is calculated from the static and arrest stress levels in relation to a dynamic overshoot coefficient $D$:

$$\tau_d = \frac{\tau_s - \tau_d}{D}.$$  

Following Madariaga (1976), we use $D=1.25$. The static strength is constant, $\tau_s(x, z) = 10$ MPa, and the arrest stress is chosen randomly from the interval $\tau_d(x, z) \in [0; 1]$ MPa.

A quasidynamic stress transfer due to coseismic slip and creep motion is calculated by means of the three-dimensional solution $K(x, z; x', z')$ of Chinnery (1963) for static dislocations on rectangular patches in an elastic Poisson solid with rigidity $\mu = 30$ GPa:

$$\Delta \tau(x, z; t) = \sum_{(x', z') \in \text{grid}} K(x, z; x', z') \Delta u(x', z'; t - r/\nu_s),$$

where $\Delta u$ is the slip, $r$ is the spatial distance between the cells $(x, z)$ and $(x', z')$, $\nu_s$ is a constant shear wave velocity, and the kernel $K$ decays like $1/r^3$. The value of $\nu_s$ defines the event time scale but has no influence on the earthquake catalogs, as long as $\nu_s > 0$.

3. Results

We analyze an earthquake catalog covering about 5000 years, which contains 200,000 earthquakes with moment magnitudes $M$ between 4.0 and 6.8. The seismic potency $P_0$ (moment $M_0$ divided by rigidity $\mu$) is calculated as the integral of slip over the rupture area during an earthquake. The range of magnitudes depends on the segmentation of the grid into computational cells, e.g. smaller cells would allow to simulate smaller earthquakes.
On the other hand, a larger fault would result in a higher maximum magnitude. Fig. 2 shows an earthquake sequence (magnitude vs. time) from the catalog over a period of 60 years.

### 3.1. Aftershocks

The most obvious feature in Fig. 2 is the aftershock clustering following the largest events. As mentioned above, the aftershock decay is compatible with the modified Omori law, Eq. (2), with an exponent $p = 1$. Furthermore, it has been demonstrated that the aftershocks are predominantly concentrated at the margins of the fault segments (Zöller et al., 2005a), which is good agreement with observational studies.

For the following investigation, we define a mainshock to be the largest event within ±1 month and an aftershock to be an earthquake occurring up to one month after the mainshock somewhere on the computational grid. An important question is the dependence of the aftershock sequences on the mainshock magnitude $M_{\text{main}}$, expressed by the value of $c_1$ in Eq. (2). The exponent $p$ is found to be almost independent of $M_{\text{main}}$ in agreement with Utsu (1962, 2002). For the rate of aftershocks, Reasenberg (1985) assumes the relation

$$c_1 \sim 10^{3.5M_{\text{main}}}.$$  \hspace{1cm} (6)

The number of aftershocks as a function of the mainshock magnitude is given in Fig. 3a. It is clearly visible that the scaling law Eq. (6) is fulfilled in our model. A second scaling relation is observed, if the number of aftershocks is plotted as a function of the mainshock rupture area, leading to $N_A \propto A^{1/2}$ and thus to $N_A \propto R$, where $R$ is the rupture length (Fig. 3b). Combining both scaling laws, results in a relation $P_0$ (or $M_0$) $\propto A^{9/8}$ between seismic potency (or moment) and rupture area $A$ of the mainshock. Plotting $P_0$ directly versus $A$ of the simulated earthquakes shows (Fig. 4) that the slope of the log ($P_0$) versus $A$ for the small
events is about 1 and that the slope increases for the larger events ($P_0 \propto A^{9/8}$). This is compatible with simulation results of Ben-Zion and Rice (1993) and high resolution observations (Ben-Zion and Zhu, 2002). A scaling relation between the log ($P_0$) and area $A$ with a slope close to 1 is expected for earthquakes in a dynamic regime close to criticality (Fisher et al., 1997).

3.2. Interevent times

In recent studies, it has been suggested that the distribution of interevent times can be described by a universal law. In particular, the distributions from different tectonic environments, different spatial scales (from worldwide to local seismicity) and different magnitude ranges collapse, if the time $\Delta t$ is rescaled with the rate $R_{xy}$ of seismic occurrence in a region denoted by $(x, y)$ (Corral, 2004):

$$D_{xy}(\Delta t) = R_{xy} \cdot f(R_{xy} \Delta t),$$

where $D_{xy}$ is the probability density for the interevent time $\Delta t$, and $f$ can be expressed by a generalized gamma distribution

$$f(\theta) = C \frac{1}{\theta^{\gamma+1}} \exp \left( - \frac{\theta^\delta}{B} \right)$$

with parameters $C$, $\gamma$, $\delta$, and $B$, which have been determined by a fit to several observational catalogs (Corral, 2004).

In Fig. 5, we compare $D_{xy}(\Delta t)$ from Eq. (7) with two earthquake catalogs: (1) the ANSS catalog of California (catalog ranges are given in the caption) and (2) the model catalog. Due to the universality of Eq. (7) with respect to different spatial scales, the comparison of our model simulating a single fault of 70 km length with a region of hundreds of kilometers including several faults in California is straightforward without coarse graining the ANSS catalog. We note that due to the numerical procedure, especially the Runge–Kutta integration of the creep law (Eq. (1)), the interevent times in the model have a finite lower limit. However, in the region where the interevent times are calculated, the agreement of the three curves is remarkable. For small values of $\Delta t$, Eq. (7) deviates from the California data; for high values the model has a slightly better correspondence with the observational data than Eq. (7). Thus our results generally support the recent findings of Corral (2004). A detailed analysis of the small deviations from Eq. (7), and their possible origin will be addressed by means of a refined parameter space study and additional statistical testing in future studies.

The degree of temporal clustering of earthquakes can be estimated by the coefficient of variation $CV$ of the interevent time distribution

$$CV = \frac{\sigma}{\mu},$$

where $\sigma$ is the standard deviation and $\mu$ the mean value of the interevent time distribution. High values of $CV$ denote clustered activity, while low values represent quasiperiodic occurrence of events. The case $CV = 1$ corresponds to a random Poisson process. Ben-Zion (1996) and Zöller et al. (2005b) have found that the clustering properties of the large events depend on the degree of quenched spatial disorder of the fault. Here we show that $CV$ as a function of the lower magnitude cutoff has a characteristic shape (Fig. 6). The values of $CV$ are higher than 1 (clustered) for small and intermediate earthquakes ($M \leq 5.4$) and smaller than 1 (quasiperiodic)
for larger earthquakes. This corresponds to the case of a low degree of disorder in Zöller et al. (2005b), because the brittle cells which participate in an earthquake have no significant spatial disorder. We note that this behavior resembles the seismicity on the Parkfield segment of the San Andreas fault, which is characterized by a quasiperiodic occurrence of mainshocks.

A different behavior is observed on the San Jacinto fault in California, where the largest events occur less regularly and have overall smaller magnitudes. As discussed in Ben-Zion (1996) and Zöller et al. (2005b), this can be modeled by imposing higher degrees of disorder leading to a broader range of spatial size scales, e.g. by using a higher number of near-vertical barriers. While barriers provide a simple and physically motivated way to tune the degree of disorder, other types of heterogeneities may work as well, as long as they are able to produce strong enough variations of the brittle fault strength.

### 3.3. Foreshocks

In Zöller et al. (2005a) and in Section 3.1, it is shown that our model produces aftershock sequences, and that the parameter space of the model includes regions where these sequences are quantitatively in good agreement with observed aftershock activity. Foreshock activity is, however, less present in observed data. Although it is documented that the increase of activity prior to a mainshock follows an “inverse Omori law” \( n(t) \sim (c + t - t_m)^{-q} \) with an exponent \( q \approx 1 \) (Hainzl et al., 1999), the overall smaller number of foreshocks requires very long data sets including many mainshocks to detect a clear foreshock signal. Stacking the activity before 225 mainshocks with \( M \geq 6 \), shows a moderate but clearly visible increase of the earthquake rate before a mainshock (Fig. 7). A stable estimation of the exponent \( q \) is, however, not possible from Fig. 7 due to the small number of extra events. The observation of an increasing event rate is encouraging for the applicability of the model to real fault zones.

### 3.4. Critical point behavior

The interpretation of seismicity in terms of critical point dynamics has opened new perspectives for the analysis and the understanding of earthquake data.

![Fig. 6](image_url)  
*Fig. 6. The temporal earthquake occurrence quantified by the coefficient of variation as a function of the lower magnitude cutoff. Values larger than 1 indicate clustering, whereas lower values point to quasiperiodic behavior.*

![Fig. 7](image_url)  
*Fig. 7. The stacked and averaged activity prior to mainshocks (225 sequences). Here, mainshocks are \( M \geq 6 \) events which are the largest earthquakes within \( \pm 10 \) years.*

![Fig. 8](image_url)  
*Fig. 8. The averaged activity and stress evolution relative to mainshocks (\( M \geq 6 \) events which are the largest earthquakes within \( \pm 10 \) years).*
These include self-organized criticality (Bak and Tang, 1989; Hainzl et al., 1999), growing spatial correlation length (Zöller et al., 2001; Zöller and Hainzl, 2002) and accelerating moment release (Bufe and Varnes, 1993; Jaumé and Sykes, 1999). A key question is whether the approach to the critical point can be measured by stress or seismicity functions as addressed by Ben-Zion et al. (2003). Their analysis of catalogs from three end-member models leads to the conclusion that large earthquakes are preceded by a period of criticality characterized by a highly disordered stress field with a broad range of size scales. In this section, we search for signals of criticality in our more realistic version of the model.

Fig. 8 (bottom) shows the average earthquake rate before and after a large event as a function of time, based again on stacking of 225 large event ($M \geq 6$) sequences. The top panel of Fig. 8 is the corresponding mean stress (normalized) on the computational grid. The mainshock itself is characterized by a significant stress drop. In the following period of aftershocks, lasting for about 1.2 years, the mean stress still decreases before it recovers and increases again. This stress increase is almost linear and lasts until the next mainshock occurs.

In previous studies (Ben-Zion, 1996; Ben-Zion et al., 2003; Zöller et al., 2005b), it was demonstrated that the degree of disorder of the stress field is correlated with the frequency–size distribution of the seismicity: a smooth stress field produces characteristic earthquake distributions with a frequently occurring characteristic event, almost no intermediate earthquakes, and small events following a truncated Gutenberg–Richter distribution; in contrast, the seismicity from a disordered stress field is characterized by a Gutenberg–Richter distribution over a broad range of magnitudes. Therefore, we expect that the frequency–size distribution before a mainshock has a broad scaling region, whereas the overall seismicity follows a characteristic earthquake law. Fig. 9 shows the cumulative frequency–size distributions for foreshocks, aftershocks, and all earthquakes in our model. Due to the small number of foreshocks, the corresponding curve is less smooth than the curves for aftershocks and all events. However, it is clearly visible that the foreshock activity shows the smallest deviations from a scaling law (denoted as solid lines) compared to the other curves. Although the overall frequency–size distribution of earthquakes follows a characteristic earthquake law, the periods before mainshocks are characterized by Gutenberg–Richter type statistics, with a broad scaling range pointing to a disordered stress field with a wide range of spatial size scales.

The second indication for critical point behavior is the acceleration of seismic moment release, which is assumed to follow the power law

$$\Delta M(t) = A + B(t - t_f)^m,$$

where the cumulative Benioff strain ($\Sigma \Omega(t)$) is calculated from the moment releases $M_0(t)$ of earthquakes at times $t' \leq t$ by $(\Sigma \Omega(t)) = \int_0^t \sqrt{M_0(t')^2} dt'$. The time $t_f$ is the mainshock occurrence time, and $A$, $B$, and $m$ are parameters. Analyzing eight large earthquakes in California, Bowman et al. (1998) find values of the exponent $m$ in the range $0.1 \leq m \leq 0.55$. The theoretical analysis of Rundle et al. (2000a) of a critical point process leads to $m=0.25$. The observational study of Bufe and Varnes (1993) and the theoretical damage model of Ben-Zion and Lyakhovsky (2002) suggest the value $m=0.3$. 

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Fig. 9. The frequency–magnitude distribution of all earthquakes, foreshocks and aftershocks, respectively. Foreshocks and aftershocks are defined to be earthquakes occurring within one month before and after a mainshock, where the mainshock definition is the same way as before. The dotted lines refer to $b$-values of 1 and 2.
Due to the noisy character of single earthquake sequences, it is difficult to detect accelerating moment release before individual mainshocks. Instead, we use a more robust representation by plotting the release of potency $P_0$ as a function of the (normalized) stress level on the fault in Fig. 10. We note that the stress increase on the fault is linear (see Fig. 8). Consequently, the abcissa represents effectively the time between two mainshocks. The curve shows a clear power law dependence of the (non-cumulative) potency release on the stress level with an exponent of $s = -1.5$. This results in a power law for the cumulative Benioff strain according to Eq. (10) with $m = 0.5 \cdot s + 1 = 0.25$, which is in good agreement with empirical and theoretical findings (Bufe and Varnes, 1993; Bowman et al., 1998; Rundle et al., 2000a; Ben-Zion and Lyakhovsky, 2002). In sum, our results are compatible with the hypothesis of accelerating moment release and provide further support for the presence of critical point dynamics in the simulated seismicity.

4. Discussion and conclusions

In the present work, we have analyzed an earthquake catalog from a recently developed model for a large heterogeneous strike-slip fault. As shown in a previous work, this model reproduces various characteristics of observed seismicity (Ben-Zion, 1996; Zöller et al., 2005a). Using different parameters, like frictional values and creep rates, the model can be tuned towards observed cases, e.g. a fault which produces clustered seismicity obeying Gutenberg–Richter statistics or a fault with quasiperiodically occurring mainshocks following the characteristic earthquake distribution. Here, we have analyzed an earthquake catalog from the latter situation, which resembles many features of seismicity on the Parkfield segment of the San Andreas fault in California, including quasiperiodically occurring mainshocks, which are followed by aftershock sequences. In contrast to short observational data sets covering some years or decades, we consider an observational period of about 5000 years of simulated seismicity. This allows us to use a stacking procedure in order to unveil properties, which are not visible in short and noisy data sets.

The stacking of many aftershock sequences shows a scaling relation between the mainshock magnitude and the number of aftershocks. This gives in combination with the modified Omori law an excellent explanation of aftershock activity. In addition, the less frequently occurring foreshock activity is also visible in our model. The temporal occurrence of earthquakes is quasiperiodic for large events and clustered for intermediate and small events with an overall interevent time distribution that follows a recently proposed universal law, and is very similar to the interevent time distribution of California seismicity. While these results demonstrate a high correspondence of the model output with observed seismicity, we also find support for the hypothesis that seismicity is associated with critical point dynamics. In particular, we have shown that the scaling range of the frequency–size distribution becomes broader when the time of a mainshock is approached. This indicates that the stress field has reached a certain degree of disorder, where a single failure can evolve into a large event. Furthermore, the moment release accelerates in interseismic periods in good agreement with observational and theoretical results.

In sum, we have demonstrated that our model provides a very good reproduction of natural earthquake activity and gives new insights to various aspects of the “critical earthquake concept.” Continuing joint study of model simulations and data associated with large fault zones may improve the ability to predict large earthquakes in these faults.

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