The evolution of the seismic-aseismic transition during the earthquake cycle: Constraints from the time-dependent depth distribution of aftershocks

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[1] Using the example of the 1992 M 7.3 Landers earthquake, we show that in the aftermath of a large earthquake the depth extent of aftershocks shows an immediate deepening from pre-earthquake levels, followed by a time-dependent postseismic shallowing. We use these seismic data to constrain the change in the depth of the seismic-aseismic transition with time throughout the earthquake cycle. Most studies of the seismic-aseismic transition have focussed on the effects of temperature and/or lithology on the transition either from brittle faulting to viscous flow or from unstable to stable sliding. A strain-rate dependent transient deepening of the brittle-ductile transition following a major earthquake is predicted by geological and laboratory observations. By analyzing the time-dependent depth distributions of aftershocks, we identify and quantify the temporal evolution of this transition. In the example of the Landers earthquake, its depth changes by as much as 3 km over the course of 4 years. *INDEX TERMS:* 7230 Seismology: Seismicity and seismotectonics; 8123 Tectonophysics: Dynamics, seismotectonics; 8159 Tectonophysics: Rheology—crust and lithosphere; 8164 Tectonophysics: Stresses—crust and lithosphere. **Citation:** Rolandone, F., R. Bürgmann, and R. M. Nadeau (2004), The evolution of the seismic-aseismic transition during the earthquake cycle: Constraints from the time-dependent depth distribution of aftershocks, *Geophys. Res. Lett.*, *31*, L23610, doi:10.1029/2004GL021379.

1. Introduction

[2] The mechanical behavior of rocks in the crust is governed by frictional brittle deformation and at greater depth by plastic flow. The nature of the coupling between the brittle and ductile layers and the depth extent and behavior of the transition zone between these two regimes are fundamental questions. Mechanical models of long-term deformation [*Rolandone and Jaupart*, 2002] suggest that the brittle-ductile transition is a wide zone where deformation is accommodated both by slip and ductile flow. In this transition zone, we expect changes in deformation mechanisms during the course of the earthquake cycle. Geologic observations of mutually cross-cutting brittle (fractures and pseudotachylites) and ductile (mylonitic foliation) deformation [*Sibson*, 1986] indicate that there is a transition zone where deformation is accommodated by both brittle and plastic processes. *Trepmann and Stockhert* [2003] use microstructural evidence to infer episodic very high stresses, strain rates and brittle deformation where deformation otherwise occurs by plastic flow. They suggest that stress redistribution during major earthquakes causes a strain-rate dependent downward displacement of the brittle-ductile transition. Thus the brittle-ductile transition is not a sharp and fixed discontinuity in the crust but is expected to vary near the fault with strain rate and therefore to change with time throughout the earthquake cycle.

[3] Geodetic surface deformation measurements provide some insights into the patterns of crustal deformation at depth during an earthquake cycle. SAR interferometric data and GPS measurements give information about the kinematics of fault slip and also reflect the rheologic properties of the Earth's crust and upper mantle [*Pollitz et al.*, 2000]. However, even in the rare instances where high-quality geodetic data span much of the earthquake cycle (e.g., associated with the 1999 Izmit earthquake [*Reilinger et al.*, 2000]), estimates of the locking depth, and the depth distribution of fault slip and viscous flow during coseismic, postseismic and interseismic phases of the earthquake cycle are not well determined. Model resolution problems and the difficulty in identifying the relevant active deformation processes at depth call for a complementary approach. In this paper, we investigate the constraints that micro-seismic activity provide on the behavior of the transition zone.

[4] The maximum depth of seismogenic faulting is interpreted either as the transition from brittle faulting to plastic flow in the continental crust [*Sibson*, 1986] or as the transition from unstable to stable sliding [*Tse and Rice*, 1986; *Scholz*, 1998]. This transition depth depends on four main factors: rock composition, temperature, strain rate, and fluid pressure [*Sibson*, 1986; *Tse and Rice*, 1986]. Spatial variations in the maximum depth of seismicity have been correlated with crustal temperature and/or lithology [*Magistrale*, 2002]. However, little has been done to examine how the maximum depth of seismogenic faulting varies locally, at the scale of a fault segment, and with time during the earthquake cycle. Time-dependent depth patterns of seismicity have been identified in only few previous studies [*Doser and Kanamori*, 1986; *Schaff et al.*, 2002].

[5] New relocation techniques provide a great improvement in the locations of hypocenters and offer an opportunity for more accurate investigations of the time-dependent depth distribution of seismicity. In this paper, we focus on the compelling case of the Landers earthquake. We use the time-dependent depth distribution of aftershocks on the strike-slip Johnson valley fault to constrain the evolution of the seismic-aseismic transition in the crust. We evaluate the deepening of the aftershocks relative to the background

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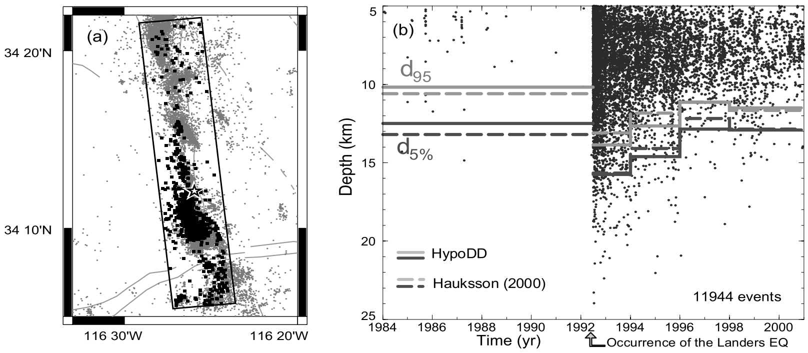

Figure 1. (a) Relocated seismicity around Landers epicenter (star) from 1984 to 2001. The box indicates the seismicity around the surface rupture of the Johnson Valley fault used in this study. Black squares are events deeper than 12 km. Grey lines are active faults. (b) Time-dependent depth distribution of seismicity for the Johnson Valley fault. The dark grey curve shows the statistics for $d_{5\%}$ and the light grey for d_{95} for different time windows: 1984 to June 28 1992 (occurrence of the Landers earthquake), June 28 1992 to 1994, 1994 to 1996, 1996 to 1998, and 1998 to 2001. The dashed lines show the same statistics for the Hauksson (2000) relocations.

seismicity, and the time constant of the postseismic shallowing of the deepest earthquakes.

2. Determination of the Time Dependence of the Depth of the Seismicity

[6] To quantify the depth variation of seismicity, we use two statistical approaches. First, we calculate the d_{95}, the depth above which 95 percent of the earthquakes occur [*Magistrale*, 2002]. This cut-off depth is widely used as an estimate of the depth of the seismic-aseismic transition and we show its variations for different time windows. We also calculate $d_{5\%}$, the average depth of the deepest 5% of the earthquakes, for different time periods and also for constant numbers of events. This statistical approach allows us to demonstrate that the deepening of the seismicity is not biased by the greater numbers of events following the mainshock and the result is not sensitive to a few hypocenters with poor depth control. The average depth of the deepest 5% of the earthquakes provides a robust estimate of the relative depth of the seismic-aseismic transition through time.

[7] The two most important problems in determining the depth of the seismic-aseismic transition from seismicity data are the precision of the hypocenter locations and the completeness of the seismic record. *Richards-Dinger and Shearer* [2000], for example, point out large differences in the depth distributions of events in various published catalogs for southern California. To demonstrate that the temporal trend we identify is reliable and consistent even though the absolute depths vary in different catalogs, we apply the same analysis to two catalogs for which the seismicity is relocated by two different methods. The first catalog of events is relocated using a three-dimensional velocity model based on a joint-hypocenter-velocity approach (JHV) [*Hauksson*, 2000]. We produce the second catalog by relocating seismicity using the double-difference method of *Waldhauser and Ellsworth* [2000]. For the second catalog we use phase data from the Southern California Earthquake Data Center (SCECDC) and a velocity model derived from *Godfrey et al.* [2002], for the velocity structure in the Mojave Desert, to relocate earthquakes in the region of the *M* 7.3 1992 Landers earthquake (Figure 1a). Generating our own catalog allows us to perform various tests to critically assess the effects of catalog completeness, the choice of different subsets of data, and changing the configuration of the seismic network. It also allows us to quantify the depth distribution of seismicity from an independent catalog for comparison with results obtained from that of *Hauksson* [2000]. Details of the relocation procedure are available in auxiliary material[1].

3. Results

[8] We analyze the time-dependent depth distribution of seismicity along the Johnson Valley fault that ruptured in the June 1992 Landers earthquake (Figure 1). We select a 5-km-wide zone around the Johnson Valley fault, as shown in Figure 1a. The background seismicity before the occurrence of the Landers earthquake was sparse (169 events in our relocated catalog), with a $d_{5\%}$ of 12.5 km and 13.2 km determined with the double-difference and joint-hypocenter-velocity techniques, respectively. To better estimate this background level, we calculate the $d_{5\%}$ considering the seismicity in a wider area with 1283 events. We find a similar depth of 12.8 km for this regional estimate.

[9] In the immediate postseismic period, the aftershocks were deeper than the background seismicity. Figure 1b shows the depth of the seismic-aseismic transition for different time intervals. It indicates a deepening by 2–3 km after the mainshock, followed by a time-dependent shallowing. The patterns for the deepest seismicity are very similar for the *Hauksson* [2000] relocations for which we only consider the aftershocks with vertical location errors less than 1.5 km. After the occurrence of Landers, the depth of the seismic-aseismic transition changes by as much as 3 km over the course of 4 years.

[10] To demonstrate the time-depth correlation of the seismic-aseismic transition, we have done calculations for different bins of constant numbers of events. Figure 2 shows the statistics for 1000 events with an overlapping moving

[1]Auxiliary material is available at ftp://ftp.agu.org/apend/gl/2004GL021379.

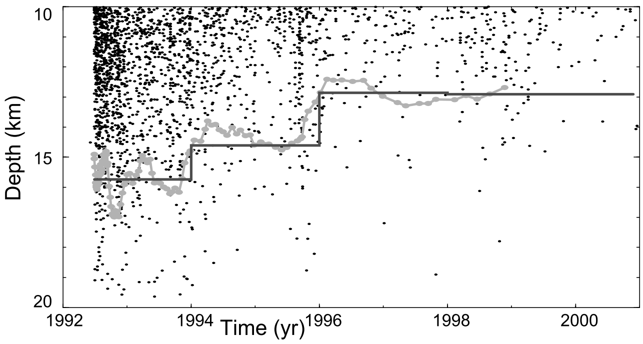

Figure 2. Statistics for the $d_{5\%}$ as in Figure 1b, compared with the same statistics, grey circles, for an overlapping window (90% overlap) for a constant number (1000) of events.

window. Our analysis for a constant number of earthquakes clearly shows that the shallowing of the seismic-aseismic is independent of the variation in the number of earthquakes through time. The decay rate of decrease in the depth of the deep seismicity is very consistent with the previous analysis. It indicates that the transition depth decreases by about 3 km during the first 4 years following Landers. Using a consistent set of stations during the period 1984–2000, the change in depth throughout time is qualitatively the same (see auxiliary material).

4. Discussion

[11] The depth of the seismic-aseismic transition in the crust has mainly been related to rheological parameters controlled by regional strain rate and variations in lithology and crustal temperature [*Williams*, 1996]. However, we have shown that the temporary existence of deep events is correlated in time with the occurrence of a large earthquake, followed by a time-dependent postseismic shallowing. In the case of Landers, the shallowing coincided with a 4 year transient postseismic relaxation indicated by GPS measurements [*Savage et al.*, 2003]. Post-Landers earthquake displacements have been interpreted as viscous relaxation in the lower crust or in the upper mantle [*Pollitz et al.*, 2000; *Freed and Bürgmann*, 2004]. The change in depth of the seismic-aseismic transition reflects the strain-rate dependent changes at the base of the seismogenic zone. The amplitude (a few kilometers) and the duration (a few years) of the depth change argue for the importance of a mechanical control on the temporal distribution of deep earthquakes and on the depth of the seismic-aseismic transition. *Rolandone and Jaupart* [2002] and *Savage and Lachenbruch* [2003], investigating the brittle-ductile transition with respect to secular fault motion, emphasize strong spatial gradients in strain rate near the bottom of the fault and argue that strain-rate dependence is more important than temperature variations in defining the brittle-ductile boundary near active faults.

[12] The characteristics of strain accumulation and release at the bottom of a major fault are still not understood. Possible physical mechanisms, which can give rise to the time-dependent depth pattern of seismicity, are viscous deformation in the lower crust and upper mantle or rate and state-dependent fault properties. Ductile loading, from the crust and/or upper mantle, in response to the occurrence of a large earthquake, may concentrate stresses and trigger aftershocks. To illustrate the effect of the earthquake on the local depth of the brittle-ductile transition, we show in Figure 3 the strength in the frictional regime (8 Mpa km^{-1}, assuming a friction coefficient of 0.75 [*Sibson*, 1986]) and the $d_{5\%}$ we calculated before Landers as the long-term depth of the brittle-ductile transition from the background seismicity (solid grey). This transition can be interpreted as the intersection between the friction law and a flow law in the ductile regime (here a Westerly granite flow law with a heat flow of 52 mW/m^2 [*Williams*, 1996] at a strain rate of 4×10^{-12} s^{-1}). The brittle-ductile transition is deeper after Landers due to high postseismic stresses and strain rates at the base of the seismogenic zone. Following Landers, in the region below the long-term brittle-ductile transition, the stress will exceed the brittle strength and previously ductile material will respond as a Coulomb material subjected to brittle failure (June 1992 to 1994, red, where there are more events at greater depth than in any other time period). Thus, there is a change in deformation mechanism from ductile to brittle. This deepening of the seismicity and of the brittle-ductile transition is followed by postseismic relaxation and a shallowing of the deep seismicity (1994 to 1996, blue, and then 1996 to 2001, green) and of the depth of the transition.

[13] We developed simple models with an elastic layer overlying a Newtonian viscoelastic layer. We calculate the stress changes in the crust from the coseismic displacement and subsequent postseismic viscous flow (Figure 4). High stresses in the ductile layer near the base of the elastic seismogenic zone are favored by high elastic strength, high slip which tapers off rapidly and a fault extending into the viscous layer [*Ellis and Stöckhert*, 2004].

[14] Important questions, beyond the simple stress-change models, relate to the strength profiles for the brittle and ductile crust before the occurrence of a large earthquake, and the impacts of coseismic stress changes on the

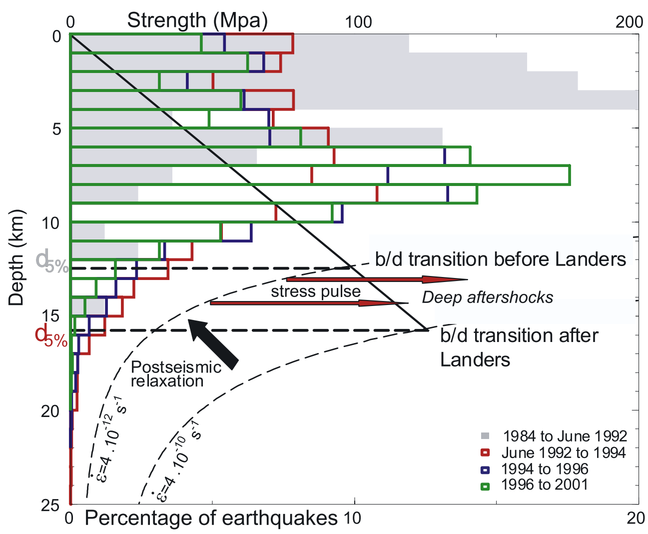

Figure 3. Histograms of the depth distribution of seismicity for different time periods (same intervals as in Figure 1). Overlaid is the strength of the brittle and ductile materials.

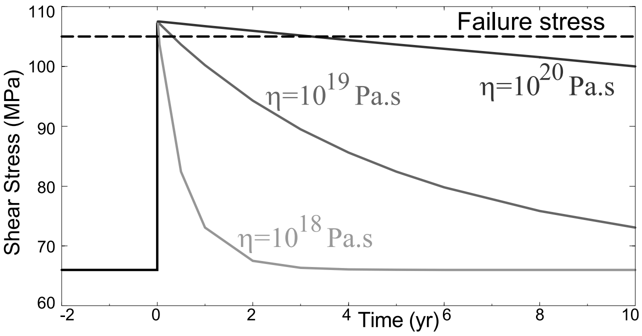

Figure 4. History of stress in the ductile crust, 0.7 km below the long-term brittle-ductile transition, for different crustal viscosities, during synseismic loading and postseismic relaxation. We use the finite element code I-deas. The fault is 12.5 km deep with uniform 5 m of slip.

mode of deformation. The magnitude of the stress pulse is given by the coseismic stress changes but the change in the mode of deformation is determined by the shear stress reaching the frictional strength envelope. The ambient (long-term) viscous shear stress dictated by the strength envelope in the ductile layer plus the stress pulse should exceed the brittle frictional strength. *Savage and Lachenbruch* [2003] argue for a low effective friction coefficient close to the fault in the plastosphere. This would favor a change in deformation mechanisms in the transition zone, because it would decrease the stress increment needed to achieve brittle failure in this region.

[15] The postseismic relaxation depends on the effective viscosity of the crust as indicated in Figure 4 which shows the stress history in the ductile crust. We can therefore relate the duration of the deepening of the seismicity to the effective viscosity of the crust. We consider a linear Maxwell viscosity, but to better understand the postseismic deformation, we need to investigate more complex rheologies such as power law flow in which the effective viscosity changes with stress and therefore with time [*Freed and Bürgmann*, 2004]. Together with geodetic measurements, these seismological observations form the basis for developing more sophisticated models of the partitioning of deformation at depth between brittle faulting and distributed deformation as a function of time.

5. Conclusion

[16] Our investigation of temporal changes in the depth of seismicity aims to increase our knowledge of (1) the depth extent of the transition zone where brittle and ductile deformation alternate, (2) the mechanics of fault zones near the base of the seismogenic zone and (3) the transient deformation processes and their rheological parameters that are active during the postseismic period. We show a strong time dependence of the depth of the deepest aftershocks. After the occurrence of the Landers earthquake, the depth of the seismic-aseimic transition shows an immediate deepening of 2–3 km followed by a time-dependent shallowing by as much as 3 km over the course of 4 years. This change in depth suggests strain-rate dependent changes at the base of the seismogenic zone and implies a mechanical control on the seismic-aseimic transition. Studies of microseismicity near the brittle-ductile transition provide additional constraints on fault zone rheology independent of geodetic data.

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