Depth constraints on nonlinear strong ground motion from the 2004 Parkfield earthquake

Justin L. Rubinstein and Gregory C. Beroza

Department of Geophysics, Stanford University, Stanford, California, USA

Received 12 April 2005; revised 7 June 2005; accepted 22 June 2005; published 27 July 2005.

[1] We use the two target repeating earthquake sequences of SAFOD to identify time varying properties of the shallow crust in the Parkfield area at the surface and in shallow boreholes. At the surface, we find that the 2004 Parkfield earthquake caused direct S wave delays exceeding 7 ms, and S coda delays exceeding 15 ms. We attribute these delays to cracks formed or opened during the strong shaking of the Parkfield earthquake. Observations at depth show that the direct S wave arrival time was much less affected by the Parkfield earthquake. This provides evidence that damage caused by strong shaking (nonlinear strong ground motion), is limited to the very near surface (<100 m). Citation: Rubinstein, J. L., and G. C. Beroza (2005), Depth constraints on nonlinear strong ground motion from the 2004 Parkfield earthquake, Geophys. Res. Lett., 32, L14313, doi:10.1029/2005GL023189.

1. Introduction

[2] A number of studies have identified nonlinear site response to strong ground motion [e.g., Field et al., 1997; Frankel et al., 2002; Wen et al., 1994]; typically using spectral ratios. Spectral ratios highlight the differences in the frequency response of a site with respect to other sites (to identify linear site response) or with respect to itself (to identify nonlinear site response). Rock sites are typically used as reference sites in spectral ratios, under the assumption that they don't have a significant site response. This assumption has been challenged by Steidl et al. [1996], who showed that surface rock sites can have a significant site response. To avoid this possible bias, some studies use borehole seismometers as reference sites [e.g., Huang et al., 2005; Wen et al., 1994]. Laboratory and theoretical studies have shown that susceptibility to nonlinear wave propagation decreases with increasing compressive stress (increasing depth) [Ostrovsky et al., 2000; Zinszner et al., 1997], implying that borehole sites are reasonable reference sites for studying nonlinear strong ground motion; however, there is scant corroborative evidence from field observations.

[3] In this study, we use an alternative technique to identify the effects of nonlinear strong ground motion with the aim of testing whether strong ground motion is linear at shallow borehole depths (75-350 m). We use repeating earthquakes (multiplets) to observe subtle, widespread changes in seismic velocity immediately following

Copyright 2005 by the American Geophysical Union. 0094-8276/05/2005GL023189\$05.00

the 2004 Parkfield earthquake. Widespread coseismic velocity reductions have been shown to be indicative of damage induced by strong shaking of a number of earthquakes [*Rubinstein and Beroza*, 2004a, 2004b; *Schaff and Beroza*, 2004; Z. Peng and Y. Ben-Zion, Temporal changes of shallow seismic velocity around the Karadere-Duzce branch of the North Anatolian Fault and strong ground motion, submitted to *Pure and Applied Geophysics*, 2005]. This technique provides a direct measurement of the effects of nonlinear wave propagation, and therefore allows us to determine if wave propagation is linear at shallow depths.

[4] Other studies have looked for changes in wavepropagation in the Parkfield region, with mixed results. For this same earthquake, a significant decrease and recovery of seismic velocities within the fault zone has been observed using fault zone trapped waves (Y. G. Li, personal communication, 2005). Others have looked for changes in wave propagation in Parkfield during periods of seismic quiescence (no M6 earthquakes). *Karageorgi et al.* [1992, 1997] found small changes in seismic velocity corresponding to changes in fluid levels, fault creep, and variations in microseismicity. *Nadeau et al.* [1994] found no significant variation in seismic velocities using repeating earthquakes. *Niu et al.* [2003] identified the movement of scatters associated with an aseismic transient.

2. Data

[5] We study two repeating microearthquake sequences that are the "target events" of the San Andreas Fault Observatory at Depth (SAFOD) to monitor the time dependence of seismic velocity in the Parkfield region (Figure 1). Cross-correlation measurements reveal that the member events of both repeating earthquake sequences are located within meters of each other. The two repeating earthquake sequences are separated by approximately 60-70 m along the San Andreas Fault [Nadeau et al., 2004]. Both events repeated approximately one year prior to the Parkfield earthquake, on October 21 and 22, 2003. Since the Parkfield earthquake, one has repeated twice, the other three times: they both repeated on September 28, 2004 (two days after the mainshock) and one sequence repeated October 24, 2004, and January 23, 2005, while the other repeated December 8, 2004. We examine these events using two seismic networks: the Northern California Seismic Network (NCSN), a network of high gain, short period, surface seismometers that record at 100 samples per second and the High Resolution Seismic Network



Figure 1. Map of the study region.

(HRSN), a network of short period, shallow borehole seismometers (\sim 70–350 m depth) that record at 250 samples per second.

3. Method

[6] We use a moving window cross-correlation technique to measure the relative arrival time of seismic phases of one event relative to a reference event in the same repeating earthquake sequence (Figure 2). The late/early arrivals indicate temporal changes in seismic velocity at the near surface. We examine unclipped, 1.28 second, Hanning tapered windows of zero-phase filtered vertical component seismograms. All the seismograms for each repeating earthquake sequence at each station are initially aligned to subsample precision to the manually picked P arrival of the reference event for that multiplet-station pair. As a result, the delays computed by the moving window cross correlation reflect the change in the S (or other phase) minus P times from the reference event to the data events.

[7] Although we find that delays are largest in the S coda (Figures 2b and 3), we choose to examine the delays of the direct S arrival. This provides a measure of the change in seismic velocity near the station that is relatively insensitive to scattering, unlike coda measurements. The arrival time of the direct S is determined under the assumption of a Poisson solid, given a precise computation of the P travel time that uses a manual P pick and precise origin times of the reference event determined by relative relocation [Waldhauser et al., 2004]. For each event-station pair, we compute the delay in the arrival of the S wave, relative to its reference event from October 2003. The delay of the S wave is computed as the median of the delay for windows centered on a time period spanning 0.1s before and 0.45 seconds after the theoretical S arrival time. To ensure data quality, we enforce a minimum correlation coefficient of 0.8 for both the P and S arrivals and a minimum signal to noise ratio of 4:1 at the P arrival.

[8] We treat the S delays observed for the repeats immediately following the Parkfield earthquake as the coseismic change in delays. In doing this, we make the assumption that between October 2003 and the 2004



Figure 2. (a) Repeating earthquake sequence 2, as recorded by NCSN station PHF. Vertical, dashed line indicates *S* Arrival. (b) Delay of September 2004 repeat of earthquake sequence 2 at PHF relative to October 2003 repeat.

Parkfield earthquake there was no significant change in seismic velocities. Processes associated with aseismic transients have been shown to influence wave propagation [Niu et al., 2003], however, no such transients were observed in this area between October 2003 and September 2004 (J. R. Murray, personal communication, 2005). In this time period, the 2003 San Simeon earthquake also occurred nearby, so its influence must be considered. Unfortunately, we don't have the temporal resolution to measure any effect of the San Simeon earthquake, which implies that our "coseismic" measurements could be overprinted by a postseismic effect following San Simeon. However, the strong shaking of the San Simeon earthquake was much weaker than the shaking of the Parkfield earthquake at our sites, so we expect that its effect on seismic velocity will be significantly weaker. Furthermore, a number of studies have previously shown that earthquake induced seismic velocity changes heal logarithmically with time [Rubinstein and Beroza, 2004a, 2004b; Schaff and Beroza, 2004]. This indicates that any effect that the San Simeon earthquake had on local seismic velocities would be mostly healed



Figure 3. Delay of September and October 2004 repeats of repeating earthquake sequence 1, relative to October 2003 repeat at (a) NCSN station PMM and (b) HRSN station EAD. Vertical dashed lines represent S arrivals. Shaded area marks window over which median S delay is computed.



Figure 4. Coseismic *S* delay of each station plotted against peak ground velocity experienced in the 2004 Parkfield earthquake. The *S* delay is the mean of the measurements computed for the September 2004 repeats of both multiplets relative to their October 2003 repeats.

by the time the Parkfield earthquake occurred. For these reasons, we believe that the change in seismic velocities that we observe will be related to the Parkfield earthquake. In fact, the delays we observe should be considered a lower bound on the delays caused by the Parkfield earthquake as healing will have progressed for the first two days after the mainshock, for which we have no observations.

4. Observations

4.1. Surface Stations

[9] At many of the NCSN stations, we find significant delays caused by the Parkfield earthquake. The delays vary in strength throughout the seismogram (Figures 2 and 3a). The largest delays are in the S coda, exceeding 25 ms at PMM. Delays of the direct S arrival can exceed 7 ms (e.g., PMM). In the second repeat of both repeating earthquake sequences after the Parkfield earthquake, we find the delays have decreased significantly throughout the seismogram (Figure 3a). This implies that the local damage is healing with time.

4.2. Borehole Stations

[10] Our observations at the HRSN stations are significantly different than those for the surface seismometers (NCSN). Typically, there is little or no delay (<2 ms) in the *S* arrival following the Parkfield earthquake (Figures 3b and 4). Similar to the NCSN stations, at many of the HRSN stations we observe delays in the *P* and *S* codas that increase with time into the coda (Figure 3b). For those borehole stations that observe delays in the *P* and *S* codas, we find that the coda delays show healing between the first and second repeats (Figure 3b).

5. Discussion

[11] The different behaviors of the delays observed at NCSN (surface) stations and HRSN (borehole) stations

suggests that the upper 100 m of the Earth's crust responds differently to strong ground motion than do deeper materials. The relation between strong ground shaking and S delays for the two networks accentuate this point, as we see a clear scaling between strong ground motion and S delays for the surface stations and no scaling of S delays to strong ground motion for the borehole stations (Figure 4). We don't observe a scaling of S delays to strong ground shaking at depth because the borehole records are from far below the shallow layers damaged by the Parkfield earthquake.

5.1. Physical Model and Interpretation

[12] To explain the delays observed at the NCSN stations, we appeal to a model in which the strong shaking of the Parkfield earthquake caused cracks to grow and/or open near the surface, effectively damaging the medium (nonlinear wave propagation). We have observed similar phenomena for the Loma Prieta, Chittenden, and Morgan Hill earthquakes [*Rubinstein and Beroza*, 2004a, 2004b; *Schaff and Beroza*, 2004]. The behavior of the delays induced by the Parkfield earthquake parallel the delays induced by the Loma Prieta and Morgan Hill earthquakes: *S* delays scale with strong ground shaking, decrease with time following the mainshock, and are largest in the coda.

[13] Unlike the surface stations, we find that the *S* arrival is not delayed at the HRSN borehole stations. Because the *S*-*P* time stays consistent before and after the Parkfield earthquake, we believe that there are no velocity reductions local to the HRSN borehole stations. This implies that the strong shaking of the Parkfield earthquake is not causing damage below depths of ~100 m. What delays are present in the *S* and *P* codas, we attribute to scattered energy that is coming from nearer the surface where nonlinear strong ground motion has reduced seismic velocities. This suggests that even the minimal amount of pressure that rocks are under in the shallow boreholes of the HRSN is enough to prevent damage during strong ground motion.

[14] Li et al. [1998, 2003] also appeal to earthquake induced cracking to explain velocity reductions following the Landers and Hector Mine earthquakes. Their observations are made much closer to the fault than ours (<500 m). With this data, they see significant velocity changes extending much deeper than we do (3 km for Landers and 5 km for Hector Mine) (Y. G. Li, personal communication, 2005). This difference can be explained by either 1) the presence of a different damage mechanism in the fault zone (e.g., shearing induced damage) or 2) differing conditions (e.g., high fluid pressures within the fault zone increasing the susceptibility to damage, or stronger shaking resulting in deeper damage).

[15] Some suggest that damage caused by the passage of seismic waves, the same phenomenon that we study here, is responsible for triggering of earthquakes at large distances (J. Gomberg and P. Johnson, Dynamic deformation scaling and earthquake triggering, submitted to *Nature*, 2005; P. A. Johnson and X. Jia, Nonlinear dynamics, granular media and dynamic earthquake triggering, submitted to *Nature*, 2005). We find that strong shaking does not damage earth materials detectably, even at modest depths (100–300 m), suggesting that for this triggering

model to work, pore fluid pressures would have to be nearly lithostatic, such that the effective stresses were comparable to those at 100 m depth. A related observation was made on the Landers fault, which was shown to be damaged by the strong shaking of the nearby Hector Mine earthquake [*Vidale and Li*, 2003]. This suggests that the high fluid pressures needed for this triggering mechanism are plausible.

5.2. Outliers?

[16] In general, the borehole stations do not have a significant response to the Parkfield earthquake, with the exception of VCA and RMN that have coseismic S delays of 4.3 and 6.9 ms respectively (Figure 4). Because these sites do not experience particularly strong ground shaking, relative to the other HRSN stations, it may be that fluid pressures at these sites were particularly high. Raising fluid pressures would increase susceptibility to strong ground motion induced damage, allowing for the large coseismic S delays we observed. RMN is also the most shallow of the HRSN stations (73 m), which might contribute to a greater susceptibility to damage due to a lower overburden. The coupling at RMN is also known to be somewhat poor (R. Nadeau, personal communication, 2005), which could provide an alternative explanation to the anomalously high delays observed there.

[17] For the surface stations, we observe a trend of increasing S delays with increasing strong motion. PCA and PST lie significantly above and below the trend, respectively (Figure 4). We have previously appealed to variations in rock strength to explain scatter in the correlation between strong shaking and velocity reductions [Rubinstein and Beroza, 2004a]. This does not explain the observations at PCA and PST as their site geologies are not significantly weaker or stronger than the average site. These variations must then come from our incomplete understanding of subsurface geology or from other limitations in our analysis. A likely source of these variations is our parameterization of strong ground motion. Our strong ground motion parameters for each site are spatially interpolated from ShakeMap. ShakeMap is a routinely produced map of strong motion parameters for earthquakes M3.5 and larger (method described by Boatwright et al. [2003]). The interpolation immediately introduces uncertainty into our measurement. The accuracy of ShakeMap's measurements is limited by the number and proximity of strong motion observations used in its computation. Specifically for PCA, there are not many nearby strong ground motion stations. ShakeMap also cannot account for localized effects (e.g., topographic effects, resonances, etc.) that may cause increased or decreased strong shaking. Although we have shown previously that strength of shaking correlates well with coseismic velocity reductions, factors other than the peak ground motion (e.g., duration of strong shaking) may control the ultimate amount of near-surface damage caused by an earthquake.

6. Conclusions

[18] We have used repeating earthquakes near Parkfield to identify near surface reductions in seismic velocity. Specifically, we identify delays in *S* arrival times at surface stations, and the general absence of delays at shallow borehole seismometers (depths $\sim 100-300$ m). Previous studies have shown that strong shaking of earthquakes damaging rocks can cause delays in *S* arrival times. The depth dependence of the *S* delays therefore implies that the pressure at the depth of shallow boreholes prevents strong shaking from damaging rocks at depth. This allows us to conclude that nonlinear wave propagation and the damage that it induces is limited to the very near surface or to regions of particularly high pore fluid pressure.

[19] Acknowledgments. We are grateful to F. Waldhauser and R. Nadeau for providing the catalog information for the repeating earthquake sequences used in this study. We thank D. Oppenheimer, H. Macbeth, and D. Neuhauser for their assistance and expedience in providing access to data. M. Thayer kindly provided the preliminary results of his mapping of the Parkfield area. This paper was improved by the suggestions of J. Vidale, Z. Peng, A, Zollo, and an anonymous reviewer. Conversations with P. Johnson and J. Gomberg improved this work. All data used in this study were made available through the NCEDC. NCSN data was provided by the U.S. Geological Survey, Menlo Park. HRSN data was provided by the Berkeley Seismological Laboratory of UC Berkeley. Research supported by the U.S. Geological Survey (USGS), Department of the Interior, under USGS award number 05HQGR0007. The views and conclusions contained in this document are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the U.S. Government.

References

- Boatwright, J., H. Bundock, J. Luetgert, L. Seekins, L. Gee, and P. Lombard (2003), The dependence of PGA and PGV on distance and magnitude inferred from Northern California ShakeMap data, *Bull. Seismol. Soc. Am.*, 93, 2043–2055.
- Field, E. H., P. A. Johnson, I. A. Beresnev, and Y. Zeng (1997), Nonlinear ground-motion amplification by sediments during the 1994 Northridge earthquake, *Nature*, 390, 599–602.
- Frankel, A. D., D. L. Carver, and R. A. Williams (2002), Nonlinear and linear site response and basin effects in Seattle for the M6. 8 Nisqually, Washington, earthquake, *Bull. Seismol. Soc. Am.*, 92, 2090–2109.
- Huang, H.-C., S.-W. Huang, and H.-C. Chiu (2005), Observed evolution of linear and nonlinear effects at the Dahan Downhole Array, Taiwan: Analysis of the September 21, 1999 M7.3 Chi-Chi earthquake sequence, *Pure Appl. Geophys.*, 162, 1–20.
- Karageorgi, E., R. Clymer, and T. V. McEvilly (1992), Seismological studies at Parkfield II. Search for temporal variations in wave propagation using vibroseis, *Bull. Seismol. Soc. Am.*, 82, 1388–1415.
- Karageorgi, E., T. V. McEvilly, and R. Clymer (1997), Seismological studies at Parkfield IV: Variations in controlled-source waveform parameters and their correlation with seismicity, 1987 to 1995, *Bull. Seismol. Soc. Am.*, 87, 39–49.
- Li, Y. G., J. E. Vidale, K. Aki, F. Xu, and T. Burdette (1998), Evidence of shallow fault zone strengthening after the 1992 M7.5 Landers, California, earthquake, *Science*, 279, 217–219.
- Li, Y. G., J. E. Vidale, S. M. Day, D. D. Oglesby, and E. Cochran (2003), Postseismic fault healing on the rupture zone of the 1999 M7.1 Hector Mine, California, earthquake, *Bull. Seismol. Soc. Am.*, 93, 854–869.
- Nadeau, R., M. Anatolik, P. A. Johnson, W. Foxall, and T. V. McEvilly (1994), Seismological studies at Parkfield III: Microearthquake clusters in the study of fault-zone dynamic, *Bull. Seismol. Soc. Am.*, 84, 174–263.
- Nadeau, R. M., A. Michelini, R. A. Uhrhammer, D. Dolenc, and T. V. McEvilly (2004), Detailed kinematics, structure, and recurrence of micro-seismicity in the SAFOD target region, *Geophys. Res. Lett.*, 31, L12S08, doi:10.1029/2003GL019409.
- Niu, F., P. G. Silver, R. M. Nadeau, and T. V. McEvilly (2003), Stressinduced migration of seismic scatterers associated with the 1993 Parkfield aseismic transient event, *Nature*, 426, 544–548.
- Ostrovsky, L. A., P. A. Johnson, and T. J. Shankland (2000), The mechanism of strong nonlinear elasticity in Earth solids, in *Nonlinear Acoustics at the Turn of the Millennium: ISNA 15*, edited by W. Lauterborn and T. Kurz, pp. 75–84, Am. Inst. of Phys., College, Park, Md.
- Rubinstein, J. L., and G. C. Beroza (2004a), Evidence for widespread nonlinear strong ground motion in the Mw 6.9 Loma Prieta earthquake, *Bull. Seismol. Soc. Am.*, 94, 1595–1608.

- Rubinstein, J. L., and G. C. Beroza (2004b), Nonlinear strong ground motion in the ML 5. 4 Chittenden earthquake: Evidence that preexisting damage increases susceptibility to further damage, *Geophys. Res. Lett.*, 31, L23614, doi:10.1029/2004GL021357.
- Schaff, D. P., and G. C. Beroza (2004), Coseismic and postseismic velocity changes measured by repeating earthquakes, J. Geophys. Res., 109, B10302, doi:10.1029/2004JB003011.
- Steidl, J. H., A. G. Tumarkin, and R. J. Archuleta (1996), What is a reference site?, Bull. Seismol. Soc. Am., 86, 1733–1748.
- Vidale, J. E., and Y. G. Li (2003), Damage to the shallow Landers fault from the nearby Hector Mine earthquake, *Nature*, 421, 524–526.
 Waldhauser, F., W. L. Ellsworth, D. P. Schaff, and A. Cole (2004), Streaks,
- Waldhauser, F., W. L. Ellsworth, D. P. Schaff, and A. Cole (2004), Streaks, multiplets, and holes: High-resolution sptaio-temporal behavior of Parkfield seismicity, *Geophys. Res. Lett.*, 31, L18608, doi:10.1029/ 2004GL020649.
- Wen, K.-L., I. A. Beresnev, and Y. T. Yeh (1994), Nonlinear soil amplification inferred from downhole strong seismic motion data, *Geophys. Res. Lett.*, 21, 2625–2628.
- Zinszner, B., P. A. Johnson, and P. N. J. Rasolofosaon (1997), Influence of change in physical state on elastic nonlinear response in rock: Significance of effective pressure and water saturation, J. Geophys. Res., 102, 8105–8120.

G. C. Beroza and J. L. Rubinstein, Department of Geophysics, Stanford University, 397 Panama Mall, Stanford, CA 94305–22215, USA. (justin@pangea.stanford.edu)