The $M_w$ 7.8, 2001 Kunlunshan earthquake: Extreme rupture speed variability and effect of fault geometry

D. P. Robinson, C. Brough, and S. Das

Received 2 November 2005; revised 19 January 2006; accepted 12 April 2006; published 11 August 2006.

By analyzing body wave seismograms, we show that the rupture speed on the Main Kunlun Fault during the $M_w$ 7.8 2001 Kunlunshan, Tibet, earthquake was highly variable and the rupture process consisted of three stages. In the first stage, the rupture propagated for another 150 km at an apparent rupture speed exceeding the $P$ wave speed. In the final stage, the earthquake fault bifurcates, and the rupture front slowed down. Stress drops are found to be higher in regions of higher rupture speeds. The greatest concentration of aftershocks is located near the fault bifurcation zone and hence coincides with the region of highest fault slip, highest stress drop and highest rupture velocity. The fault width is no more than 10 km in most places and is about 20 km in the region of highest slip. This narrow fault width is attributed to the fact that crust below this depth is sufficiently warm not to permit brittle failure to occur. The remarkable similarity of this earthquake with the 1906 California earthquake, in spite of occurring in very different tectonic regimes, is discussed.


1. Introduction

It is now well known that the damage caused by an earthquake depends on the ground velocity and acceleration at the site, which in turn are controlled by the velocity and acceleration of the propagating fault [Madariaga, 1983]. Thus knowledge of the maximum possible speed at which earthquakes can propagate is essential for reliable seismic hazard assessment. In the 1960s, it was believed that earthquake ruptures could not propagate faster than the Rayleigh or shear wave speeds [Broberg, 1999; Das, 2003]. Theoretical rupture propagation studies in the 1970s [Burridge, 1973; Hamano, 1974; Andrews, 1976; Das and Aki, 1977; Burridge et al., 1979; Freund, 1990] suggested that earthquakes could not only exceed the shear wave speed $v_s$ but could actually propagate at the compressional wave speed $v_p$. For a long time, only one observation of supershear rupture speeds was found, namely for the 1979 Imperial Valley, California earthquake [Archuleta, 1984], but recent observations of supershear rupture speeds in laboratory experiments [Rosakis et al., 1999; Samudrala et al., 2002; Xia et al., 2004] rekindled interest in searching for such speeds. Several such observations have now been reported [Bouchon et al., 2001; Bouchon and Vallée, 2003; Antolik et al., 2004; Ozacar and Beck, 2004; Dunham and Archuleta, 2004]. One of these was for the 14 November 2001 $M_w$ 7.8 Kunlunshan, Tibet, earthquake during which an average supershear rupture speed was reported by Bouchon and Vallée [2003]. Since we know that rupture speed varies over the fault, we determine the details of the rupture process for this earthquake, in order to obtain the maximum rupture speed attained.

2. Tectonic Setting of the 2001 Kunlunshan Earthquake

The 14 November 2001 $M_w$ 7.8 Kunlunshan earthquake (Figure 1) ruptured an ~400 km portion of the Kunlun fault, a major strike-slip fault in central Tibet. Left-lateral motion along this fault is believed to accommodate some of the northward motion of the Indian plate under Tibet by lateral extrusion of the Tibetan crust. The rupture length in this earthquake is one of the longest for a strike-slip earthquake (for comparison, the 1906 San Francisco earthquake had a length of 432 km [Yeats et al., 1997]) and is certainly the longest strike-slip faulting earthquake since the advent of modern seismometers. The Kunlunshan earthquake produced surface ruptures, reported from field observations, with displacements as high as 7–8 m [Xu et al., 2002], this large value being supported by interferometric synthetic aperture radar (InSAR) measurements [Lasserre et al., 2005]. Analysis of regional seismograms by Bouchon and Vallée [2003] suggested that the rupture propagated at an average speed of 3.9 km/s, which exceeded the local shear wave speed of 3.5 km/s. The supershear rupture speed was confirmed by analysis of $P$ waves [Antolik et al., 2004;
Ozar and Beck, 2004], which gave an average speed of 3.4–3.6 km/s. In order to study the variability of the rupture speed over the fault, we use $SH$ body waves to resolve the details of the rupture and fault slip histories.

3. Relocated Aftershocks

[4] We have relocated the aftershocks of this earthquake for the 6-month period following it, using the method of Joint Hypocenter Determination [Dewey, 1971, 1983] and $P$ wave arrival times reported by the International Seismological Centre (ISC). The ISC reported 91 earthquakes during this period, with 41 having $m_b \geq 4.0$ and 8 with $m_b \geq 5.0$. Of these 91 earthquakes, we successfully relocated 80, with 73 of these having 90% confidence error ellipses $\leq 30$ km. We consider these 73 earthquakes to be “reliably” relocated. The 80 aftershocks relocated by us for the 6-month period following the earthquake are shown in Figure 2. Only 5 earthquakes in this 6-month period have Harvard centroid moment tensor (CMT) solutions (Harvard CMT solutions for 1983–2002 can be accessed at http://www.seismology.harvard.edu/CMTsearch.html), the largest one on 18 November 2001 having a $M_0$ of 5.6, which is remarkably small for such a large main shock. None of the five CMT solutions have the same mechanism as the main shock, suggesting that they either lie off the main fault or are indicative of small-scale fault plane complexity.

4. Centroid Moment Tensor Solution

[5] Since shallow vertical strike-slip faults have unstable centroid moment tensor (CMT) solutions [Henry et al., 2002], we redo the CMT solution. The Harvard CMT solution was obtained using 44 stations and 123 channels, giving the solution with strike $94^\circ$, dip $61^\circ$, rake $12^\circ$, and a seismic moment $M_0$ of $5.9 \times 10^{20}$ N m (equivalent to $M_w$ of 7.8). Since the Harvard solution was obtained, many additional stations have reported data. We are thus able to recalculate the CMT solution using 89 stations and 243 channels and the same method as used by Harvard [Dziewonski and Woodhouse, 1983]. We find an optimal [Henry et al., 2000] solution of strike $94^\circ$, dip $70^\circ$, rake $25^\circ$, and $M_0$ of $5.7 \times 10^{20}$ N m, close to that found by Harvard. Since shallow crustal earthquakes are not expected to have a significant nondouble-couple component [Henry et al., 2000, 2002; Robinson et al., 2001] we next impose the pure double-couple constraint and find an optimum pure double-couple (OPDC) solution with strike $98^\circ$, dip $90^\circ$, rake $-36^\circ$ (Figure 2), and $M_0$ of $5.4 \times 10^{20}$ N m with steeper dip but similar strike to the

![Figure 1](image-url). (top) Tectonic setting and topography in the region of the Kunlun Fault. The star indicates the epicenter of the 14 November 2001 Kunlunshan earthquake. (bottom) Segments [Van der Woerd et al., 2002] of the Kunlun fault, shown by vertical blue dashed lines, together with the CMT solution obtained in this study (red). The Harvard CMT solution for the 8 November 1997 Manyi earthquake is shown in blue, and the focal mechanisms for two earlier earthquakes in the region with magnitudes $\geq 7$ are shown in black [Triep and Sykes, 1997]. The faulting extent of the Kunlunshan earthquake is shown in red.
previous solutions. Since the dips of shallow strike-slip events are not well constrained [Henry et al., 2000, 2002; Robinson et al., 2001], we then carry out a grid search, for two values of strike (94° and 98°), over many values of dip and rake, with depth fixed, to examine the solution space. For each value of strike, we carried out 651 inversions. Figure 3 shows the misfit space for one value of strike (the other is similar and not shown). We see a broad region of possible solutions with misfit values only marginally higher than the OPDC solution given above. The general shape of this region is similar to such plots for other recent shallow strike-slip earthquakes, namely, the Antarctic Plate [Henry et al., 2000] and Wharton Basin [Robinson et al., 2001] earthquakes. Thus the mantle wave data are not able to provide an unambiguous source mechanism for this earthquake.

5. Analysis of Body Wave Seismograms

To study the detailed faulting process, we use the method of seismogram inversion developed by Das and Kostrov [1990, 1994] and use pure SH waves at 19 stations, as well distributed in azimuth as was available, at epicentral distances from 35°–70°. Outside this distance range, unmodeled phases such as ScS or SS arrive very soon after the S wave making the usable part of the seismogram very short. At distances where no such phase arrives, we terminate the seismogram at 190 s. The source time is ~120 s, and 12 of the 19 stations have seismograms longer than 120 s, 8 of which are longer than 150 s. For similar reasons, the usable part of the pure P wave seismograms are very short, and too few stations were available to be used. To model Green’s functions, we use the crustal model CRUST 5.1 [Mooney et al., 1998] at the source and one modified from CRUST 5.1 for each station, based on known local geological conditions. The seismograms were filtered to contain periods between 2 and 120 s. However, due to attenuation of S waves in the mantle, the seismograms contain very little energy at periods less than 10 s. Thus, at first glance it may appear that we may not be able to image details of the rupture process. In order to understand better the resolution power of data of this type, we have carried out extensive numerical experiments with artificial teleseismic data in the 2–120 s period range, which we have then inverted using the same method as used in this paper for some earthquakes [Henry et al., 2000; Henry and Das, 2002]. Such studies have shown that we are, in fact, able to image details smaller than might be expected. On the basis of this
experience, we have used a spatial cell size of 10 km for the discretization of the fault here. It is interesting to note that similar numerical experiments on artificial strong ground motion data also led to a similar conclusion [Das and Suhadolc, 1996; Das et al., 1996; Sarao et al., 1998], so that this behavior holds at very different length scales.

5.1. Optimal Source Mechanism for Use in Rupture Process Inversions

Since the mantle wave data do not provide an unambiguous source mechanism, we must first obtain this. Attempts at inversions of the \( SH \) wave data using the source mechanisms obtained from our mantle wave study show that the initial parts of the seismogram are poorly fit. An earlier study [Antolik et al., 2004] using \( P \) waves suggested that more than one fault was required to fit that data. We find that at least two faults are needed to fit the data. A third short fault placed in between these two faults, as was done by Antolik et al. [2004], has no perceptible effect on our data fit. We shall refer to these two faults as the Main Kunlun Fault (MKF) and the Secondary Fault (SF), as marked in Figure 2.

5.2. Primary and Preferred Solutions

Using these mechanisms, we obtain our primary solution, in which we use the positivity (no “back slip” on the fault) constraint; do not allow parts of the fault beyond a \( P \) wave from the hypocenter to rupture; and fix the total moment to our CMT solution moment. The rupture propagation with time is shown in Figure 4. Although the data are fitted well, the spatial moment distribution is seen to be very rough. Hence, following Das and Kostrov [1990, 1994], we find a “smeared out” solution in which the maximum moment rate anywhere on the fault at any time is minimized, keeping the fit to the data almost unchanged. This is our preferred solution, discussed from now on and shown in Figure 5. In our interpretation of the rupture process, our general conclusions are seen in both the primary and the preferred solutions, that is, we consider the persistent features of the solution, though it is the preferred solution that is discussed in detail. We use the terms ‘slip’ and ‘moment’ interchangeably in our discussions, as they are proportional to one another, the axes of figures always making it clear which quantity is being shown.

6. Rupture Process

The rupture initiated at the westernmost end of the SF and propagated unilaterally eastward for \( \sim 50 \) km. The rupture then continued along the MKF, with a change of strike from \( \sim 82^\circ \) to \( \sim 100^\circ \), propagating eastward for another \( \sim 400 \) km. Table 1 lists the source parameters.
obtained for the two faults in our study. Figure 5 shows that the rupture process on the MKF consists of three distinct stages. The rupture propagated for the first 120 km on the MKF at an average speed of 3.3 km/s (94% of the local $v_s$). Note that since we have finite cells and time steps in our solution, all distances, times, rupture speeds and stress drops that we obtain are approximate. During the second stage, rupture continued eastward for another 150 km at a very high speed, the front being marked by a dashed line on Figure 5. The trend of the $P$ wave causal zone (indicated by the crosses) helps to show that locally the rupture propagates in this stage at a speed exceeding the $P$ wave speed.

Figure 5. (top) Final slip on the MKF for our preferred solution. (bottom left) Same as Figure 4 but for our preferred solution. The dashed line indicates the approximate rupture front in the second stage of very rapid rupture. For comparison, two lines are drawn indicating the wave speeds of 4.1 (green) and 3.5 (red) km/s. Numerical noise seen adjacent to the $P$ wave causal front is neglected. Arrows indicate the timing of (1) the start of main rupture on the MKF, (2) acceleration of rupture from 3.3 km/s to >6 km/s, (3) termination of rapid slip phase, and (4) beginning and (5) end of third phase of rupture. The arrow at the top indicates the fault bifurcation zone. (top right) Comparison of the $SH$ wave data (black) with the solution synthetics (red), for our preferred solution ($\ell_1$ misfit 0.467). The amplitude at each station is scaled to a constant distance of 60 km, and the maximum amplitude in microns is shown at the beginning of each trace. Tick marks are placed at 10 s intervals along each trace. The $SH$ nodal planes for the MKF for our preferred solution are shown on the stereographic projection of the focal sphere. (bottom right) Moment rate function for the earthquake from our preferred solution.
It is important to note that this does not mean that rupture is acausal, since the front lies in the causal region. What it does mean is that when waves from the earlier parts of the rupture arrive, the rocks over this 150 km long region reach their breaking strength within a very short time interval, so that the apparent rupture speed (~6.7 km/s) exceeds the P wave speed. Thus, in this stage, there is no “classical” rupture front. This possibility had been foreseen by Kostrov [1975] (and is discussed by Kostrov and Das [1988]), who considered an even more extreme case, that it is not impossible, though unlikely, that every point on a fault could reach its critical breaking stress simultaneously, making it appear to have infinite rupture velocity. As we see here, it is certainly possible for small sections of the fault to locally exceed the P wave speed.

Table 1. Source Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>MKF</th>
<th>SF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strike, deg</td>
<td>100</td>
<td>262</td>
</tr>
<tr>
<td>Dip, deg</td>
<td>90</td>
<td>81</td>
</tr>
<tr>
<td>Rake, deg</td>
<td>7</td>
<td>–2</td>
</tr>
<tr>
<td>Moment, (\times 10^{19}) N m</td>
<td>55</td>
<td>4.3</td>
</tr>
<tr>
<td>Length, km</td>
<td>400</td>
<td>50</td>
</tr>
<tr>
<td>Width, km</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Average rupture speed, km/s</td>
<td>4.1</td>
<td>1.3</td>
</tr>
<tr>
<td>Average slip, m</td>
<td>2.2</td>
<td>1.37</td>
</tr>
<tr>
<td>Average stress drop, MPa</td>
<td>2.4</td>
<td>1.5</td>
</tr>
<tr>
<td>Cell size, km</td>
<td>(10 \times 10)</td>
<td>(10 \times 10)</td>
</tr>
<tr>
<td>Time step, s</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>Modelled interval, s</td>
<td>0–120</td>
<td>0–45</td>
</tr>
</tbody>
</table>

Table 2. Comparison With Other Body Wave Studies

<table>
<thead>
<tr>
<th>Study</th>
<th>Number of Stations</th>
<th>Data</th>
<th>L, km</th>
<th>W, km</th>
<th>(u_{\text{max}}), m</th>
<th>(X^*, \text{km})</th>
<th>(A^*, \text{km})</th>
<th>(\Delta \sigma_{\text{av}}, \text{MPa})</th>
<th>(\Delta \sigma_{\text{av}}, \text{MPa})</th>
</tr>
</thead>
<tbody>
<tr>
<td>This study</td>
<td>19</td>
<td>19 SH</td>
<td>400</td>
<td>20</td>
<td>6.5</td>
<td>94.3(^*)\text{E}</td>
<td>50 (\times 20)</td>
<td>2.4</td>
<td>8.0(^{\text{a}})</td>
</tr>
<tr>
<td>Lin et al. [2003]</td>
<td>17</td>
<td>17 P, 7 SH</td>
<td>400</td>
<td>30</td>
<td>6.8</td>
<td>93.1(^*)\text{E}</td>
<td>60 (\times 30)</td>
<td>1.2</td>
<td>7(^{\text{b}})</td>
</tr>
<tr>
<td>Antolik et al. [2004]</td>
<td>37</td>
<td>35 P, 14 SH</td>
<td>400</td>
<td>15</td>
<td>10</td>
<td>93.2(^*)\text{E}</td>
<td>60 (\times 15)</td>
<td>3.75</td>
<td>-</td>
</tr>
<tr>
<td>Ozacar and Beck [2004]</td>
<td>23</td>
<td>23 P</td>
<td>420</td>
<td>15</td>
<td>9</td>
<td>93.1(^*)\text{E}</td>
<td>80 (\times 10)</td>
<td>3.1(^{\text{c}})</td>
<td>-</td>
</tr>
</tbody>
</table>

\(^{a}\)The parameters are \(u_{\text{max}}\) is maximum slip on fault; \(X^*\) is location of main asperity; \(A^*\) is size of main asperity; \(\Delta \sigma\) is average stress drop on fault; \(\Delta \sigma_{\text{av}}\) is stress drop on asperity.

\(^{b}\)Calculated using the same method as Lin et al. [2003].

\(^{c}\)Estimated from published diagrams.

\(^{d}\)Calculated by us using published numbers.
remove the high moment in this region and still fit the data, showing that high moment in the region of the very rapidly rupturing portion of the fault is required by the data.

Finally, we test if our not allowing rupture to occur ahead of the $P$ wavefront from the hypocenter has affected the solution. For this we allow all points to rupture behind a causal front of 9 km/s. This is essentially equivalent to

\[ 2 \times 10^{17} \text{Nm}^{-1} \text{s}^{-1} \]

Figure 6. Schematic showing the final slip on the two faults, for our preferred solution, the maximum slip being 6.5 m. The average rupture speeds in the different stages of rupture are indicated. The Harvard CMT location (grey diamond), the location of the CMT solution obtained in this study (black diamond), and the centroid location of our broadband solution (black star) are shown.

\[ 13 \]

Figure 7. Similar to Figure 5 but for the case when we constrain the average rupture velocity not to exceed the local value of $v_s$ of 3.5 km/s. The inversion is unable to satisfactorily fit the data ($\chi^2$ misfit 0.567) and places the rupture front right up against the causality constraint, indicative of a faster rupture velocity being required.
having no constraint at all on the rupture speed. The moment rate distribution in time is essentially identical to our preferred solution. In particular, no moment is seen ahead of the $P$ wavefront, even though it is permitted. This gives us further confidence in our preferred rupture process.

Theoretical models of supershear rupture propagation suggest that large stresses are generated in the off-fault regions and may cause fresh damage there (H. S. Bhat et al., Off-fault damage patterns due to supershear ruptures with application to the 2001 $M_w$ 8.1 Kokoxili (Kunlun) earthquake, submitted to Journal of Geophysical Research, 2006, hereinafter referred to as Bhat et al., submitted manuscript, 2006). Field observations of anomalous ground cracking have been reported, concentrated in the region off-fault of the very fast rupture speed section of the fault for this earthquake (see photographs of Bhat et al. (submitted manuscript, 2006)).

6.2. Depth Extent of Faulting

In order to determine the depth extent of coseismic faulting, a series of further inversions were carried out in which slip was constrained to occur at various depths. Our inversions showed that it is unlikely that there was any significant slip at a depth greater than 20 km, though in early inversions we permitted slip to occur to 40 km depth. Next, inversions where slip was constrained to occur in only the top 10 km were carried out. These inversions were able to fit the data almost, but not quite, as well as inversions where slip was permitted down to 20 km. However, when slip is constrained to the top 10 km, very large coseismic displacements of the order of 20 m are obtained, which does not agree with the surface observations [Xu et al., 2002; Lasserre et al., 2005]. Finally, a set of inversions were carried out where slip was confined only to depths between 10 and 20 km. These inversions were unable to satisfy the data. Therefore we conclude that slip extended to depths greater than 10 km, but less than 20 km.

For our preferred solution, the final slip on the fault at each of the two cells levels, that averaged over the depth extent, and the measured surface displacements are shown in Figure 9. The slip in the region west of about 93°E is seen to be comparable in the upper and lower cells (Figures 5 and 9). However, when the fault is in the second stage of rupture, though the slip in the lower layer is markedly lower than the upper layer, it is still high compared to other parts of the fault (Figure 5). As the fault enters the third stage of rupture, beyond the bifurcation region, there is little slip in the lower layer, and most of the slip is confined to the top 10 km.

The fact that no slip occurs below 20 km depth has important implications. First, theoretical studies show that on planes with constant properties, ruptures tend to be equidimensional [Das, 1981; Das and Kostrov, 1988; Madariaga et al., 2000]. This earthquake is more than 400 km long yet no more than 20 km wide. So, there must be a very strong change in the properties near 20 km depth to prevent rupture from penetrating deeper. In well studied near-vertical strike-slip regimes such as the San Andreas fault, the seismogenic depth is also about 20 km, but there the reason is well understood, namely, that this is the crustal depth, below which material is too warm to permit brittle failure. The continental convergence zone of Tibet has a double thickness (~60 km) of crust. Therefore, at first glance, one could have expected a much larger fault width here and the fact that this does not occur may appear surprising. Crustal earthquake depths in this region are less than 20 km [Chen and Molnar, 1983], though there is also some evidence that earthquakes occur in the mantle to depths of 100 km [Chen and Yang, 2004]. However, there do not seem to be any earthquakes between about 20 and 70 km in this region. This agrees with recent studies of the temperature structure of the Tibetan crust by Mechie et al. [2004], who showed that the 700°C isotherm is likely to be at a depth of 18 km in this region, as the lower crust has been warmed as a result of burial and radioactive decay in the crust. This would prevent brittle failure below these depths and is thus similar to the seismogenic width of the San Andreas fault.

7. Discussion

The 2001 Kunlunshan earthquake ruptured a clearly resolvable distance of nearly 150 km in a very short time.
interval, leading to an apparent super-$P$ wave rupture speed locally. This is the first observation of this phenomenon, foreseen by Kostrov [1975]. It has been suggested that the large length of the Kunlunshan earthquake was required for the fast rupture speed [Xia et al., 2004]. Figure 1 shows that the MKF is also very straight for the first 270 km, the first and second stages of rupture propagation. Just after passing the bifurcation point, the rupture seems to pause before continuing on for the third stage along the southern fork, the Kunlun Pass Fault. Neither field nor InSAR observations show any coseismic movement on the northern fork, the Xidatan segment of the Kunlun fault [Xu et al., 2002; Lasserre et al., 2005]. This is a very good example of fault rupture being slowed by fault geometry. This fork is also the region with the highest concentration of aftershocks. The rupture terminates at a sharp change of strike (transition from red to black on the MKF in Figure 1), and we interpret this as a geometric barrier [Aki, 1979] which stops the rupture.

[19] Numerical experiments suggest that regions of faster rupture are also stronger [Dunham et al., 2003], while the dimensionless parameter $S$, defined as the ratio of the relative fault strength to stress drop must be small [Hamano, 1974; Andrews, 1976; Das and Aki, 1977]. Our results showing that regions of fast rupture are also regions of higher stress drop show how this is possible for the Kunlunshan earthquake.

[20] Even though the 2001 Kunlun earthquake occurred in a very different tectonic regime than the 1906 California earthquake, they are remarkably similar. Both are vertical strike-slip faults that have very similar lengths and widths, and unusually fast rupture speeds have been reported for both (by Song et al. [2005] for the 1906 earthquake).

[21] The most important and damaging earthquake of recent times was the 26 December 2004 great Sumatra earthquake [Ammon et al., 2005]. It propagated for the first 600 km at an average speed of $\sim$2.5 km/s, with the rupture velocities of later portions still being debated. This is much slower than the average rupture speed of the 2001 Kunlunshan earthquake. It should be noted that real faults rupture in a mixture of the two shear modes, the in-plane and the antiplane mode. The Kunlun earthquake fault was so long compared to its width that it was almost purely an in-plane fault. The 2004 Sumatra earthquake, with a down-dip width of about 200 km, was an antiplane fault along its strike but in-plane along the dip direction. This means that very long subduction zone thrust earthquakes, such as the Sumatra earthquake, could have a very long in-plane rupture front, which, in principle, could reach speeds as high as that for the Kunlun earthquake. However, no observation of supershear rupture speed has yet been reported for any subduction zone earthquake. Whether this is because this has never occurred, or, is due to the fact that seismic stations were not situated in optimal positions to detect this is not known. Obviously, detecting the along-dip rupture velocity is far more difficult than the along-strike speed.

[22] Our finding on how wide a range the rupture speeds of large earthquakes can span has important consequences for seismic hazard assessment, earthquake

Figure 9. Along-strike displacement along the MKF, for our preferred solution, at the upper layer of cells (dotted line), the lower layer of cells (solid line), and averaged over the two layers (dashed line). The measured surface displacements [Xu et al., 2002] are shown as discrete grey circles.
Acknowledgment.


Kostrov, B. V. (1975), Mechanics of the Tectonic Earthquake Focus, 176 Moscow, Moscow.


Rosakis, A. J., O. Samudrala, and D. Coker (1999), Cracks faster than the shear wave speed, Science, 284, 1337–1340.


C. Brough, S. Das, and D. P. Robinson, Department of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK. (christopher.brough@gmail.com; das@earth.ox.ac.uk; davidr@earth.ox.ac.uk)