Kinematic rupture process of the 2014 Chile $M_w$ 8.1 earthquake constrained by strong-motion, GPS static offsets and teleseismic data

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SUMMARY

On 2014 April 1, a magnitude $M_w$ 8.1 interplate thrust earthquake ruptured a densely instrumented region of Iquique seismic gap in northern Chile. The abundant data sets near and around the rupture zone provide a unique opportunity to study the detailed source process of this megathrust earthquake. We retrieved the spatial and temporal distributions of slip during the main shock and one strong aftershock through a joint inversion of teleseismic records, GPS offsets and strong motion data. The main shock rupture initiated at a focal depth of about 25 km and propagated around the hypocentre. The peak slip amplitude in the model is ~6.5 m, located in the southeast of the hypocentre. The major slip patch is located around the hypocentre, spanning ~150 km along dip and ~160 km along strike. The associated static stress drop is ~3 MPa. Most of the seismic moment was released within 150 s. The total seismic moment of our preferred model is $1.72 \times 10^{21}$ N m, equivalent to $M_w$ 8.1. For the strong aftershock on 2014 April 3, the slip mainly occurred in a relatively compact area, and the major slip area surrounded the hypocentre with the peak amplitude of ~2.5 m. There is a secondary slip patch located downdip from the hypocentre with the peak slip of ~2.1 m. The total seismic moment is about $3.9 \times 10^{20}$ N m, equivalent to $M_w$ 7.7. Between the rupture areas of the main shock and the 2007 November 14 $M_w$ 7.7 Antofagasta, Chile earthquake, there is an earthquake vacant zone with a total length of about 150 km. Historically, if there is no big earthquake or obvious aseismic creep occurring in this area, it has a great potential of generating strong earthquakes with magnitude larger than $M_w$ 7.0 in the future.

Key words: Earthquake ground motions; Earthquake source observations; Body waves; Subduction zone processes; Crustal structure; South America.

1 INTRODUCTION

Subduction zones, resulting from the convergent movement between two tectonic plates, are locations where megathrust earthquakes occur. These megathrust earthquakes post tremendous hazards to local inhabitants and global populations through destructive ground shaking and earthquake-generated tsunami, as exemplified by the 2004 $M_w$ 9.15 Sumatra–Andaman and 2011 $M_w$ 9.0 Tohoku-Oki earthquakes. Hence, studies on the nature of megathrust earthquakes in subduction zone are of great importance and interest to both earthquake relief efforts and seismogenic studies. To the west of Chile, the Nazca Plate is subducting beneath the South America Plate at about 65 mm yr$^{-1}$ (DeMets et al. 2010). Due to the rapid convergent rate between these two plates, numerous big thrust earthquakes have occurred, such as the 1960 $M_w$ 9.5 Valdivia earthquake (Plafker & Savage 1970; Kanamori & Cipar 1974), and the 2001 $M_w$ 8.5 Arequipa earthquake (Okal et al. 2006). In recent years, InSAR and GPS studies (Chlieh et al. 2011; Moreno et al. 2011; Métois et al. 2012; Béjar-Pizarro et al. 2013) have identified a large locked patch near the coastline of the north Chile, roughly 500 km in length along the strike direction (Fig. 1). Moreover, this area has not experienced any big earthquakes since 1877, which makes it possible to accumulate huge fault slip equivalent to an earthquake with magnitude of $M_w$ 8.0–8.8 (Comte & Pardo 1991; Lomnitz 2004; Chlieh et al. 2011).

On 2014 April 1, an $M_w$ 8.1 thrust earthquake ruptured the region of Iquique seismic gap in northern Chile. To date, several rupture models of this event have been constructed based on teleseismic data (Lay et al. 2014; Yagi et al. 2014), tsunami observations (An et al. 2014), and joint inversion of multiple data sets (Lay et al. 2014; Schurr et al. 2014). However, there can be found significant discrepancies between the inverted models, because of the effect of low frequency filtering resulting from the long distance propagation of teleseismic data, the resolution of teleseismic data significantly reduces, and these data can only constrain the ruptured area close to the hypocentre (rupture initiation point). In contrast, near-field
data sets, such as strong-motion data, static and high-rate GPS, and InSAR data, can provide strong constraints in precisely and accurately locating the fault rupture (Delouis et al. 2004; Zhang et al. 2014).

As it is well known, an accurate slip model is critically important to estimate the ground shaking (e.g. Ren et al. 2013; Nakashima & Lavan 2014) and the Coulomb stress change on surrounding faults, and to assess seismic hazards and analyse the possible post-seismic mechanisms. Thus, collecting near field data and using them to build a reliable and accurate rupture process model is vital to understand the seismogenic structure. In this study, we retrieved the spatial and temporal distribution model of this $M_w$ 8.1 mega earthquake and its $M_w$ 7.7 aftershock through a joint inversion of strong-motion, GPS static offsets and teleseismic data. The goal is to constrain the rupture history to better understand the mechanics of the causative fault and to discuss the increasing seismic hazard after this earthquake.

2 DATA

As for the main shock, we selected strong-motion waveforms recorded at 22 stations from the Integrated Plate Boundary Observatory Chile (IPOC) network (http://www.ipoc-network.org/, Fig. 1). The accelerograms were bandpass filtered between 0.01 and 0.25 Hz by two zero-phases, second-order Butterworth filters, and then integrated to ground velocity data. The bandpass filters were chosen in order to avoid the interference by long-period integration noise and circumvent the inadequacies of the theoretical Green functions at higher frequencies. Although near-field strong motion waveforms are very sensitive to earthquake rupture process, all those strong motion stations used in this study are located to the east side of the source region with a limited azimuthal coverage of 152° ranging from 24.4° to 176.4°. Moreover, studies of finite-fault inversion for large megathrust events with one-sided (land side only) station distributions have demonstrated that the slip resolution will reduce in the regions relatively far from offshore rupture zones. However, this limitation can be partially compensated by joint inversion with teleseismic data sets (Ammon et al. 2011; Koketsu et al. 2011; Wei et al. 2012; Yue et al. 2013).

To improve the azimuthal coverage, we further incorporated 29 broad-band teleseismic $P$ waves and 21 $S$/$H$ waves. These seismic data were downloaded from the Incorporated Research Institutions for Seismology (IRIS) network, the instrument responses of these data were deconvolved from the original recordings to obtain ground velocities to improve the spatiotemporal resolution of the inversion (Wald et al. 1996). The seismograms were then bandpass filtered at 0.002–0.5 Hz. Compared with seismograms, static displacements have been proven particularly useful for defining the slip distribution for large complex ruptures, such as the $M_w$ 7.2 El Mayor-Cucapah event (Wei et al. 2011). Thus we also added seven GPS static offsets from the IPOC network to our joint inversion.

For the $M_w$ 7.7 aftershock on 2014 April 3, we selected 13 strong motion records, 4 GPS offsets in the near field and 22 teleseismic $P$ waves with high signal-to-noise ratio and good azimuthal distribution as demonstrated in Fig. S1. All the data processing procedures are the same as those used in the main shock inversion.

3 CHECKERBOARD TESTS

In seismology, checkerboard tests can provide a direct visualization of the resolution of inversion by using different data sets. Here we conducted checkerboard tests to investigate the resolution of each type of data set and to explore the potential advantage of joint inversion.

Similar to what we did for the real data, we designed a model of a fault plane striking 357° and dipping 16°, and divided it into a series of smaller subfaults with a size of 12 km × 10 km. The synthetic data were calculated for the same stations from which we obtained the real data (Fig. 1) by using the same Green’s functions as used in the real data inversions. The slip patches contain 5 × 4 subfaults with grid size of 60 km × 40 km for each subfault, each subfault has slip amplitude of 4 m, rake angle of 109°, average rupture velocity of 2.5 km s$^{-1}$ and rise time of 12 s. The synthetic displacement and velocity waveforms were computed for geodetic and seismic stations respectively. Then we did separate inversions using single kind of data, like seismic, geodesic, strong motion, or synthetic data, and also did joint inversion by combing all the data sets. During the inversion, the slip amplitude could vary from 0 to 8 m, the rake angle could change from 49° to 169°, the rise time could vary from 2.0 to 20 s with an interval of 2.0 s and the rupture velocity could change from 1.5 to 3.5 km s$^{-1}$.

Our testing results are shown in Fig. 2 with the Fig. 2(a) plotted as the input model. We found that the resolution for the teleseismic body wave is poor (Fig. 2b) and is insufficient to resolve the rupture process in detail. Such low resolution of the teleseismic inversion was also demonstrated by Delouis et al. (2010) for the 2010 Maule

![Figure 1](http://gji.oxfordjournals.org/)
earthquake. Since all the GPS stations are located to one side of the epicentre, the GPS static offsets only provide generally good resolution for downdip slip, but poor resolution of slip near the trench. Neither the shallow slip pattern nor the slip amplitude is well constrained by the inversion solely using GPS static offsets (Fig. 2c). For the \( M_w 8.1 \) event, most strong-motion stations are located in the south along the fault plane (Fig. 1), which makes it hard to well constrain the slip to the north (Fig. 2d). Based on the testing results, we found that each individual data set only provides limited resolution of inverted slip models and no single data set can resolve the input model well. Compared with the inversion using a single kind of data set, the joint inversion can resolve the rupture process much better, overcome the limitations of separate inversions and ideally achieve good resolution across the entire fault model, especially the slip near the trench (Fig. 2e).

Although it is very intuitive and convenient to show the inversion resolution using the checkerboard test, we applied the same Green’s function for generating synthetic data and inversion in the checkerboard test, for which the effect of the Green’s function discrepancy will not be tested. The fault geometry is an important parameter to determine the Green’s function, for which the radiation pattern and depth phase vary with different fault geometry, and may cause some uncertainties in our inversion result. In our checkerboard test, the dip angle is fixed (16\(^\circ\)), which is inconsistent with the real case. The crustal velocity model is another factor that can take effects on the Green’s function for the inversion. In order to reduce the disturbance from the crustal velocity model, the Green’s functions were generated using a 1-D crustal model derived from local studies in this region (Lüth 2000; Table 1). So the disturbance caused by the inaccuracy of the crustal velocity model should be small. Thus, we can conclude that the joint inversion is capable of providing high resolution and reliable rupture model for the Chile earthquake.

### 4 INVERSION SETUP

#### 4.1 The 2014 April 1 \( M_w 8.1 \) Earthquake

In this study, we modelled the fault geometry using a 300 \( \times \) 200 km rectangular plane (Fig. 1). We initially tried the strike of 357\(^\circ\) and dip of 18\(^\circ\) from GCMT and allowed small perturbations to better fit the data. When fixed the strike angle, we found that a dip of 16\(^\circ\) does

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**Table 1.** 1-D velocity model used for inversion.

<table>
<thead>
<tr>
<th>No.</th>
<th>Thickness (km)</th>
<th>( V_P ) (km s(^{-1}))</th>
<th>( V_S ) (km s(^{-1}))</th>
<th>Density (g cm(^{-3}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>4.00</td>
<td>2.299</td>
<td>2.500</td>
</tr>
<tr>
<td>2</td>
<td>8</td>
<td>6.05</td>
<td>3.477</td>
<td>2.910</td>
</tr>
<tr>
<td>3</td>
<td>13</td>
<td>6.28</td>
<td>3.609</td>
<td>2.956</td>
</tr>
<tr>
<td>4</td>
<td>10</td>
<td>6.39</td>
<td>3.672</td>
<td>2.978</td>
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<tr>
<td>5</td>
<td>35</td>
<td>6.51</td>
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<tr>
<td>6</td>
<td>23</td>
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<td>4.368</td>
<td>3.220</td>
</tr>
<tr>
<td>7</td>
<td>60</td>
<td>8.10</td>
<td>4.655</td>
<td>3.320</td>
</tr>
</tbody>
</table>

Note: \( V_P/V_S \) ratio 1.78.
Figure 3. (a) The surface projection of the inverted rupture slip distribution, grey dots indicate the aftershocks, red dots indicate the aftershocks of the 2007 November 14 $M_w$ 7.7 Antofagasta, Chile earthquake, all the aftershocks from the NEIC catalogue. The slip (>0.5 m) distribution of the $M_w$ 7.7 aftershock is outlined with red-dashed line. The elliptical area shows the seismic gap. (b) The inverted slip model, the direction of the fault plane is indicated by the long black arrow on the top of the figure, and the star indicates the hypocentre, the colour bar scales the slip amplitude, the white arrows represents the slip directions and contours display the rupture initiation time in second. The coseismic slip >1.5 m regions are outlined with solid yellow curves. (c) The moment-rate function of the main shock.

4.2 The $M_w$ 7.7 aftershock on 2014 April 3

For the finite-fault inversion of the aftershock on 2014 April 3, the fault plane was parametrized with 19 and 18 subfaults along strike (357°) and dip (18°), respectively, the total area of fault model is 152 km × 144 km. The rupture initiation was set at the NEIC hypocentre (20.518° S, 70.498° W, 31 km), and the same velocity model as used in the main shock was applied to approximate the structure in the source region. During the inversion process, the slip value for each subfault could vary from 0 to 5 m, and the rake angle was searched from 41° to 161° with a search step of 3°, the average rupture velocity was set to vary in a range from 1.0 to 3.0 km s$^{-1}$ with an interval of 0.1 km s$^{-1}$, the rise time was allowed to change from 1.2 to 12.0 s with time step of 1.2 s.

5 INVERSION RESULTS AND DISCUSSION

The preferred model of the main shock constrained by the joint inversion is shown in Figs 3(a) and (b). The major slip patch is located around the hypocentre, spanning ~150 km along strike and ~160 km along dip. The average slip of the well-resolved primary slip region (slip > 1 m) is ~3 m. The total seismic moment of our preferred model is 1.72 × 10$^{21}$ N m, equivalent to $M_w$ = 8.1. The inverted peak slip is ~6.6 m, located at the southeast of the hypocentre. Taking this earthquake as a circular fault with a radius (R) of 65 km and an average slip ($\Delta D$) of 2.0 m, the estimate of static stress drop based on the expression (Kanamori & Anderson...
The dominant slip in intraplate earthquakes have a median of around 3.3 MPa (Kanamori, 1975) which is consistent with global studies showing that stress drops of offset and teleseismic observations are given in Figs S3-S5. The corresponding fits to the local strong motion waveforms, GPS static teleseismic data. The preferred model is shown in Fig. S2(a). The joint inversion, the moment of this offshore rupture patch which is a stable feature of many inversions we conducted. Based on the joint inversion, the moment of this offshore rupture patch is approximately 18 per cent of the total moment, and the waveform contributions from this patch are significant for the amplitude in the various data signals. The seismic moment of the updip slip patch is \( \sim 3.1 \times 10^{20} \) N m, equivalent to an \( M_w \) 7.5 earthquake. The waveform and static displacement fits of all data sets are shown in Figs 4–6. Our preferred model can well explain both recorded waveform and static displacement fits of all data sets.

Although most of the observed data were fitted very well, relatively big misfits are seen for some strong motion stations, such as the EW component of station PB04, PB05, PB06 and NS component of station PB09, particularly in the later segments of the records. These misfits are presumably due to the path effects caused by the heterogeneities of the earth’s internal structures. Nevertheless, this joint inversion achieved good waveform fits to the various data sets.

We next constrained the slip distribution of the \( M_w \) 7.7 aftershock on 2014 April 3 using the selected strong-motion, GPS offsets and teleseismic data. The preferred model is shown in Fig. S2(a). The corresponding fits to the local strong motion waveforms, GPS static offsets and teleseismic observations are given in Figs S3–S5. The slip mainly occurred in a relatively compact area, and the amplitude of peak slip in the model is \( \sim 2.5 \) m, located around the hypocentre. There is a secondary slip patch located downdip from the hypocentre with the peak slip of \( \sim 2.1 \) m. The total seismic moment of our preferred model is \( 3.9 \times 10^{20} \) N m, equivalent to \( M_w \) 7.7. The rupture velocity of the major slip patch is relative slow, only 1–2 km s\(^{-1}\). The rupture propagated slowly mainly along downdip direction in the first 10 s, and then accelerated to \( \sim 3.0 \) km s\(^{-1}\) (85 per cent of shear wave velocity) along downdip direction during the next 10 s. For the moment rate function of the rupture model, our rupture model shows that the major rupture occurred in the first 30 s, and most of the seismic moment are released in the first 80 s (Fig. S2b).

Comparing the slip distributions between the main shock and its largest aftershock, we observe strongly complementary slip distributions. We suspect the \( M_w \) 7.7 aftershock was triggered by the main shock. Furthermore, we also should pay attention to the uncracked area in the rupture fault. Although the \( M_w \) 8.1 main shock and its aftershock ruptured in the Iquique seismic gap, and the 2007 November 14 Antofagasta events occurred in the southern edges of the gap (Fig. 3a), the remaining unbroken zone is still 150 km long along the strike. Thus, it is capable of generating an earthquake with magnitude larger than \( M_w \) 7.0, and may pose a high seismic risk in the near future.

In order to analyse the reliability of our rupture model and figure out some new characteristics of the 2014 Iquique, Chile \( M_w \) 8.1 earthquake, we compared our joint rupture model with results of other studies. Lay et al. (2014) obtained two rupture models: one is constrained by teleseismic body waves only and the other is obtained by inverting teleseismic body waves and tsunami data jointly; An et al. (2014) built the rupture model by tsunami data; Yagi et al. (2014) inverted the rupture model by teleseismic P-wave data. The teleseismic-based models including the finite fault model posted on USGS’ website have similar slip patterns, mainly located in the southeast of the hypocentre, which is consistent with our joint rupture model. However, obvious differences in the slip distribution close to the trench can be found between these rupture models. In
the model of Yagi et al. (2014), there is no significant slip near the trench; however, the teleseismic-based rupture model of Lay et al. (2014) and finite fault model posted on USGS’ website have a slip of about 4.0 m near the trench. Both An et al. (2014) and Lay et al. (2014) used tsunami records to constrain the rupture model and did not find significant slip near the trench. In our model, there is an obvious updip slip above the hypocentre.

We also compared our result with the model constrained by joint inversion with both seismic and geodetic data. Schurr et al. (2014) investigated the Iquique earthquake by joint inversion of local strong

Figure 5. Comparison of the strong-motion records (black line) and synthetic seismograms (red line) derived from our model. Both data and synthetics are aligned by the first P arrivals. The number at the first of each seismogram indicates the station name, and the number at the right top is the maximum velocity of the records in cm s⁻¹.
Figure 6. Comparison of teleseismic velocity records in black and synthetic seismograms in red predicted by the slip model, the seismograms are bandpass filtered with frequency band of 0.002–0.5 Hz. Both data and synthetic seismograms are aligned on the P (a) and SH (b) arrivals. The number at the end of each trace is the peak velocity of the data in micrometres per second. The azimuth and distance in degrees are shown at the beginning of each record with the azimuth on top.

motion, GPS static displacements and teleseismic data. Comparing with their result and our work, the majority of the rupture slip is quite similar, with the peak rupture located southeast to the epicentre. In the model of Schurr et al. (2014), the shape of the major rupture slip is also like an ellipse with long axis of about 140 km along latitudinal direction and short axis of 120 km along longitudinal direction. Similar to the case of teleseismic data inversion and the tsunami data inversion, the biggest difference also mainly
comes from the rupture slip near the trench. There’s no significant slip near the trench in the model of Schurr et al. (2014), most rupture slip concentrated around the hypocentre, and the cumulative slip shows a simple ‘bullseye’ pattern with a peak slip of about 4.4 m, which is smaller than our model (about 6.5 m in our model). The other difference from the secondary asperity, there is a secondary asperity located east to the epicentre in our preferred model, which makes the whole rupture fault like a ‘dumbbell’ while there is no such kind of asperity in the model of Schurr et al. (2014).

The reasons for these differences are presumably due to the following factors: (1) The weights of data sets used in their model and this work might be different. Viewing from model of Schurr et al. (2014), we can find that the static model constrained by static GPS data is quite similar to the joint model, but the rupture model inverted by teleseismic and strong motion data is quite different with the final joint inversion model, and the peak slip is only half of that in the joint inversion model. It means that the near field static GPS data play dominant role in the inversion, and might be set too heavy weigh during the inversion. On the other hand, the checkerboard in our model shows that each data set plays its own role in the joint inversion, and combining all of the observations the rupture slip can be recovered quite well. (2) The inversion methods are different in these two works. In the model of Schurr et al. (2014), the temporal evolution of slip is modelled using multiple time windows in the inversion procedure. The benefit of such an approach is that it results in a linear system of equations; however, the main drawback is that it introduces many more degrees of freedom into the solution space, which potentially adds to the non-uniqueness of the result. In our approach (Ji et al. 2002), we perform a joint, non-linear inversion to solve for the rise time, rupture initiation time and slip for each subfault simultaneously. The benefit of this approach is the reduction in the number of free parameters. (3) The data sets are different. The teleseismic data used by Schurr et al. (2014) only involved P wave data, while both P and SH waves are used in our work, which may provide more constrains to the rupture process. (4) The filter-band is different. In our work the frequency bands of the bandpass filter (0.01–0.25 Hz for strong motion data, 0.002–0.5 Hz for teleseismic data) are higher than that of Schurr et al. (2014) (0.01–0.1 Hz for both strong motion and teleseismic data), as we know, the lower the frequency, the lower the resolution, that’s why the slip pattern of Schurr et al. (2014) model is like a simple ‘bullseye’, and no detailed image of local asperities.

Moreover, another aspect that needs to be considered is about the null-space problem. Usually there is trade-off between neighbour subfaults close to the trench, if one slip patch near the trench is smaller and another slip patch in the deeper depth is larger, the two waveforms in one receiver could be similar, thus can cause some null-space problem. When only teleseismic data are used, or the local stations are mainly concentrated in a small area with unevenly azimuthal distribution, this kind of problem may be especially serious. Fortunately, in our joint inversion, both teleseismic and local data are applied, and the local stations are distributed relatively evenly surrounding the source region except in the NW side, thus the null space problem should not be serious, but it may still cause some ambiguity in the rupture process model, this could be another reason for the difference in the slip near the trench between this model and the other studies. The models developed by tsunami data (An et al. 2014; Lay et al. 2014) demonstrate that there is no obvious slip near the trench, which is a little different with our model. As we know, the tsunami data sets are sensitive to the shallow slip, some near trench in our model may be caused by the null space effect. Thus, we should pay more attention to the deeper patches near and below the hypocentre area.

Anyway, considering the data coverage, checkerboard test and the differences between our work and other models, we conclude that our preferred rupture model of the $M_w$ 8.1 Iquique earthquake is reliable and accurate in the deeper part of the fault.

6 CONCLUSIONS

By jointly inverting strong motion data, GPS offsets and teleseismic data, we investigated the source rupture process of the $M_w$ 8.1 earthquake occurring on 2014 April 1 and its $M_w$ 7.7 aftershock in north Chile. Compared with previous models constrained either by teleseismic data alone, by teleseismic data and tsunami data jointly, or by teleseismic data, strong motion data and static GPS data jointly, our model is quite reliable and exhibits some new characteristics, including some updip slip near the trench. The main shock rupture initiated at about 25 km depth and propagated around the hypocentre, and the peak slip amplitude in the model is $\sim$6.6 m, located in the lower left of the hypocentre. The main large slip patch was located around the hypocentre, spanning $\sim$150 km along dip and $\sim$160 km along strike. The associated static stress drop is $\sim$3 MPa. Most of the seismic moment was released within first 150 s after the occurrence of the earthquake. The total seismic moment of our preferred model is $1.72 \times 10^{21}$ N m, equivalent to $M_w$ 8.1. As for the $M_w$ 7.7 aftershock, the slip mainly occurred in a relatively compact area, and the peak slip amplitude in the model is $\sim$2.5 m. The total seismic moment of our preferred model is $3.9 \times 10^{20}$ N m, equivalent to $M_w$ 7.7. From the slip distribution model of the main shock and aftershock, we observe strongly complementary slip distributions, suggesting a triggering relationship between them. Based on the rupture models of the Chile $M_w$ 8.1 earthquake, its $M_w$ 7.7 aftershock and the earthquake sequence of the 2007 November 14 Antofagasta earthquake, it can be observed that there is a 150 km unruptured gap, which could pose high earthquake risk in the future and where extra monitoring attention is needed.

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REFERENCES


Rupture process of the 2014 Chile $M_w$ 8.1 earthquake


SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this paper:

Figure S1. Map view of the near-field data coverage of the $M_w$ 7.7 aftershock. Red star and red beachball denote the epicentre and GCMT focal mechanism of the main shock and its largest aftershock, grey dots indicate the aftershocks from NEIC catalogue. Yellow inverted triangles denote the strong motion stations. Black arrows are GPS coseismic observations. Upper insert panel shows the distribution of teleseismic stations. The black rectangle outlines the fault plane of the finite fault inversion in this work.

Figure S2. (a) The aftershock slip distribution calculated from the joint inversion. The direction of the fault plane is indicated by the long black arrow on the top of the figure, and the star indicates the hypocentre, the colour bar scales the slip amplitude, the white arrows represents the slip directions and contours display the rupture initiation time in second. (b) The moment-rate function of the $M_w$ 7.7 aftershock.

Figure S3. Comparison between the observed displacements and those predicted from our model. Panel (a) is the horizontal displacement and (b) is the vertical displacement. The black arrows are the GPS data, the red arrows are synthetics. The rectangle is the assumed fault plane; the beachball shows the epicentre of the $M_w$ 7.7 aftershock.

Figure S4. Comparison of the strong-motion records (black line) and synthetic seismograms (red line) derived from our model. Both data and synthetics are aligned by the first $P$ arrivals. The number at the first of each seismogram indicates the station name, and the number at the right top is the maximum velocity of the records in cm s$^{-1}$.

Figure S5. Comparison of teleseismic velocity records in black and synthetic seismograms in red predicted by the slip model, the seismograms are bandpass filtered with frequency band of 0.002–0.5 Hz. Both data and synthetic seismograms are aligned on the $P$ (a) and SH (b) arrivals. The number at the end of each trace is the peak velocity of the data in micrometres per second. The azimuth and distance in degrees are shown at the beginning of each record with the azimuth on top (http://gji.oxfordjournals.org/lookup/suppl?doi:10.1093/gji/ggy214/-/DC1).

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