High-frequency rupture properties of the M_w 9.0 off the Pacific coast of Tohoku Earthquake

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The devastating M_w 9.0 earthquake on March 11, 2011 is one of the most complex earthquakes of all the recent large events. Its source region is compact for an earthquake of this size, but it has highly variable amount of energy release from different segments. These conditions prevent conventional back-projection analysis to reveal the details of the rupture process. We incorporate a new metric to assess coherency as a part of back-projection analysis to ensure identification of these weak features. The main features obtained with this new back-projection approach are consistent with previous back-projection results, with strongest energy release down-dip and close to the epicentral location. The main rupture propagation is along strike, in southwestern direction giving rise to the length extent of the earthquake. The new coherency function also allows us to investigate rupture characteristics at the beginning of the earthquake, resolving initial up-dip propagation from the epicentral location. Furthermore, some of very weak energy in the stacks are identified with high coherency. These additional source regions extend the area of the earthquake farther south and north than the region that has been imaged by other back-projection studies.

Key words: Back projection, rupture properties, high frequency, frequency dependence.

1. Introduction

The giant earthquake on March 11, 2011 (off the Pacific coast of Tohoku) released high-frequency seismic energy that are strongly dependent on both space and time. This poses a challenge for back-projection analysis (e.g., Ishii *et al.*, 2005), because high-amplitude energy may obscure weak energy release. This was an issue when the direct P waveforms are combined with depth phases for intermediate-depth earthquakes (Kiser *et al.*, 2011). In this case, the high-amplitude P arrivals conceal depth-phase arrivals, and the authors apply a taper to suppress P -wave artifacts. This approach would be difficult to apply in the current case where the P waveform itself is rich in amplitude variation. To overcome this problem, and to capture weak, and yet potentially important, energy releases, we introduce coherency function where the coherency of individual traces are monitored as a function of location and time. In this paper, we combine the coherency function and back-projection stacks to investigate the detailed high-frequency rupture characteristics of the March 11, 2011 M_w 9.0 main-shock.

2. Method

The back-projection technique takes advantage of waveform similarity recorded by a large-aperture seismic array to monitor changes in azimuth and distance of the energy source from the array (e.g., Ishii *et al.*, 2005). Seismograms

recorded at each station of the array, $u(t)$, are stacked after time-reversed to grid of locations around the hypocentral location, i.e.,

$$s_i(t) = \frac{1}{W} \sum_{k=1}^K w_k u_k(t - t_{ik}^0 + \Delta t_k),$$

where $s_i(t)$ is the stacked seismogram at the i 'th grid location, the sum is over all stations k , t_{ik}^0 is the predicted travel time between grid i and station k using a one-dimensional model, and Δt_k is station-dependent time correction. This time correction is obtained by cross correlating the initial few seconds of a well-recorded event in the source region. It accounts for wavefront distortion due to lateral variations and station error (e.g., timing issue). The cross correlation procedure also yields relative amplitude, A_k , and polarity, p_k , of each trace with respect to a reference waveform. In order to ensure that no single trace dominates the stack, the seismograms are normalized by weighting factor $w_k = p_k/A_k$. Sum of all the weighted seismograms are divided by the sum of weighting factors $W = \sum w_k$ to define the stack at the source $s_i(t)$. The cross correlation values obtained from Δt_k determination are also used to identify incoherent or noisy seismograms.

If an earthquake releases energy nearly evenly with time (e.g., 2004 Sumatra-Andaman earthquake; Ishii *et al.*, 2005), the back-projection stacks $s_i(t)$ provide reliable estimates of the source locations. However, if the energy release is highly time dependent, artifacts from strong energy release may obscure weak energy release. This effect can be

Fig. 1. Distribution of stations in North America used in the back-projection analysis of the M_w 9.0 March 11 mainshock and M_w 7.1 April 7 aftershock. Green triangles show 453 stations that contributed to low-frequency analysis, and red triangles show 433 stations that are used for high-frequency analysis. Note that stations that are more than 92 degrees from the April 7 aftershock location are not included, as well as seismograms with correlation values less than 0.7 (Ishii *et al.*, 2005).

circumvented by defining a coherency function $x_i(t)$, i.e.,

$$x_i(t) = \frac{1}{K} \sum_{k=1}^K \frac{p_k \sum_{\tau=t}^{t+T} [u_k(\tau - t_{ik}^0 + \Delta t_k) \cdot s_i(\tau)]}{\sqrt{\sum_{\tau=t}^{t+T} u_k^2(\tau - t_{ik}^0 + \Delta t_k)} \sqrt{\sum_{\tau=t}^{t+T} s_i^2(\tau)}}.$$

This gives average cross correlation value of individual seismogram and stacked trace as a function of space (grid point i) and time. The time window T over which cross correlation is calculated, should reflect how long energy may radiate from a given location as well as ensuring that there is enough waveform within this window for meaningful cross correlation calculation. This implies that the coherency function is most useful for high-frequency data, and not so much for low-frequency analysis.

The value of the coherency function is expected to be large for identifying the location and time of energy release. This is because the coherency function is not affected by the strength of various energy release, hence weak energy release can be identified as robustly as strong energy release. On the other hand, the coherency is expected to degrade as one moves away from location where the seismograms have been aligned due to changes in lateral structure. Furthermore, if the source is distributed over a diffuse area, the coherency function will be low. Despite these issues, rupture locations are determined more reliably with the coherency function, hence we use this function to estimate spatio-temporal variations in rupture using high-frequency data. We rely on the stacked traces to obtain relative amplitude information.

3. Data

We take advantage of the availability of data from the Transportable Array in the U.S. with nearly 400 stations. This array is a part of the EarthScope USArray project, and is currently located in the central US (Fig. 1). In order to improve azimuth and distance coverage, we complement the Transportable Array data with additional stations in the US, i.e., stations from Southern California

Seismic Network (operated by Caltech and USGS), ANZA Seismic Network (operated by the Institute of Geophysics and Planetary Physics, Scripps Institution of Oceanography), Advanced National Seismographic System (operated by USGS), Pacific Northwest Seismograph Network (operated by University of Oregon), University of Utah Regional/Urban Seismic Network (operated by University of Utah), Berkeley Digital Seismic Network (operated by UC Berkeley), and stations in the US that are part of the Global Seismographic Network (operated by USGS). Finally, seismograms recorded by the Canadian National Seismograph Network provided by Geological Survey of Canada are included. To avoid complications associated with later arrivals, we apply the back-projection technique to the first P -wave arrivals using the vertical component recorded at above stations. This large array spanning almost the entire North America provides excellent resolution. Synthetic tests using a Ricker wavelet with central frequency of 1 Hz (corresponding to the high-frequency case presented in this manuscript; e.g., Ishii *et al.*, 2007) give resolution area of about 500 km².

One important component of back-projection analysis is the station-specific correction (Ishii *et al.*, 2005) that accounts for wavefront distortion due to lateral variations and station error (e.g., timing issue). If this correction is not properly determined, spurious features can be introduced. Because of the narrow frequency filter used and complexity of the initial few seconds of P waveform, cycle skips are commonly observed with the cross correlation procedure for the M_w 9.0 mainshock. Instead of relying upon such questionable alignment, we utilize waveforms of the M_w 7.1 aftershock from April 7, 2011. This earthquake has clean first arrival, and can be easily cross correlated and aligned. Consequently, all results presented in this paper are with respect to the hypocentral location of this aftershock, i.e., 38.253°N 141.640°E and 49.0-km depth as provided by the National Earthquake Information Centre (NEIC).

4. Results

To investigate if the rupture of the 2011 Tohoku earthquake shows frequency dependence as observed for the 2010 M_w 8.8 Maule, Chile earthquake (Kiser and Ishii, 2011a), we filter the data to two narrow frequency bands. We refer to the two frequencies as high (0.8 to 2 Hz) and low frequency (10 to 20 second period). The low-frequency back-projection results do not have spatial or temporal resolution to identify details of the rupture process due to long period and long wavelength. We, therefore, analyze the high-frequency results to characterize the rupture, then compare main features of high-frequency results to low-frequency results. Note that the back-projection approach provides location and timing of relative energy release during an earthquake, but cannot constrain parameters such as dip of the fault or absolute slip (e.g., Ishii *et al.*, 2005, 2007).

4.1 High-frequency back-projection results

Assuming that the epicentral location of the earthquake used to obtain the station-specific corrections Δt_k is correct, we can use the coherency function to estimate the epicentral location and time of the mainshock. The epicentral location from back-projection analysis is at 142.8°E and 38.0°N , and the event begins at 05:46:28 UT. We estimate that the uncertainty in longitude and latitude to be about $\pm 0.2^\circ$, and uncertainty in time to be about ± 7 seconds. The hypocentral time of the earthquake from both NEIC and Japan Meteorological Agency (JMA) is 05:46:23, within uncertainty of the back-projection estimate. The two catalogues give somewhat different epicentral locations of 142.369°E 38.322°N (NEIC) and 142.86°E 38.103°N (JMA). The back-projection location is closer to the JMA determination than the NEIC solution.

The rupture characteristics of this earthquake can be divided into six episodes, four of which involve significant area of rupture and two appear nearly as a point source. The first subevent of this earthquake lasts for about 40 seconds in total, and is simple, i.e., similar to other large earthquakes we have studied (e.g., Kiser and Ishii, 2011b). It starts out with slow propagation from the epicentre toward the trench in the initial 20 seconds. When the rupture reaches the updip extent, it transitions into a bilateral rupture where the updip energy remains but with energy starting to propagate downdip (Fig. 2). Most of the energy release during this episode is very weak, and hence typically this behaviour is not captured in existing back-projection studies (e.g., Hutko, 2011; Kiser, 2011; Wang and Mori, 2011; Yao *et al.*, 2011).

The second subevent, spanning a time between 40 to 120 seconds from the beginning of the event and releasing the largest amount of energy, is the most complex of all the subevents. Much of the energy release occurs in the first 60 seconds of this subevent within a large area between about 37.5°N and 39°N (Fig. 2). This area covers much of the seismogenic zone at this location, and includes the area that ruptured during the first subevent around the epicentre. Energy release is highest within the northern portion of this area and downdip of the epicentral location. The complexity of this subevent is not obvious from Fig. 2, but if the energy release locations are monitored, they move around

Fig. 2. Different subevents during the mainshock are indicated by different contours, and the background colour shows the relative amounts of energy released at each location (warmer colours indicating more energy release). The cyan contour shows the area ruptured during subevent 1 which occurs within the first 40 seconds, and covers an area around the epicentre (red star with black outline). The two white contours show areas associated with the main part (dashed curve) and later bilateral rupture (solid curve) of subevent 2. The solid magenta and yellow contours are the ruptures that propagate updip during subevents 3 and 5, respectively. The dashed magenta and yellow contours are subevents 4 and 6, respectively, that behave nearly as a point source. The aseismic patch imaged using the coherency function is located between the white outline of the subevent 2 and magenta outline of subevent 3, and clearly shows up as a gap between northern and central parts of imaged high-frequency rupture region. The yellow dots are aftershocks between March 11 and 17 as given by the JMA catalogue. The black and red curves show the coastline and the trench location, respectively. The range of colours used in this plot is shown on the bottom right corner.

considerably, sometimes rupturing a region that have already released energy (e.g., first subevent area). Near the end of this subevent, starting around 100 seconds since the event initiation, the back-projection results show bilateral rupture propagation. This propagation is nearly parallel to the trench, and one front moves northeast, up to about 39.5°N , and the other front moves southwest, down to about 37.5°N (Fig. 2). The last part of the northeastern propagation is somewhat suspicious. The propagation of back-projected energy is very similar to that expected from imperfect data coverage. Even though coherency is above the background level, part of this northeastern finger may be a spurious feature, and that the actual northeastern rupture may stop at about 39°N . The southern propagation of the bilateral rupture encounters a patch that does not participate in the rupture during the entire mainshock (Fig. 2). This patch is very cleanly imaged in the coherency function, but is also visible in the stack. The rupture literally moves around this aseismic patch, and initiates rupture in a region south of the second subevent.

The third subevent, starting with the rupture that propagates around the aseismic zone, is much weaker compared

Fig. 3. Comparison of high- and low-frequency results. (a) The relative amplitude of high- (blue) and low-frequency (red) back-projection results as a function of time. (b) Spatial comparison of energy locations for two frequencies. The background colour (red/warm colours indicate large energy release locations and blue/cold colours show low energy release regions) shows the low-frequency energy distribution around 80 seconds after the event initiation. The white contour shows the high-frequency energy release location between 50 and 110 seconds. The range in time window used for obtaining the high-frequency contour allows us to compare high-frequency results (with very fine time resolution) with lower-frequency results (poorer time resolution). The white star shows the epicentral location determined by high-frequency coherency function. (c) Same as in (b) except for the 180-second peak for the low-frequency result. The high-frequency contour is calculated between 150 and 210 seconds. The colour scale has been normalized, but the peak amplitude for this 180-second energy is about 23% of the peak shown in (b).

to the second subevent (Fig. 2), and is a very brief episode with duration of about 25 seconds. The initial energy release south of the aseismic patch is the strongest, followed by very diffuse energy propagating slightly updip. This subevent is followed by about 15 seconds of quiescence where no significant coherency is observed within the studied area.

At about 140 seconds, there is a subevent that has the characteristics of a point source within the limitation of the current technique and data set (Fig. 2). It occurs within the rupture area of preceding subevent 3, at about 141°E and 37°N . It may be the very first aftershock of the mainshock.

The fifth subevent, and the last event involving significant area, starts about 145 seconds and lasts for about 35 seconds. This event covers an area south of subevent 3 with practically no overlap (Fig. 2), and has very weak, diffuse energy release that is difficult to identify in normal back-projection stacks. However, the coherency is above the background level, suggesting that the weak energy seen in the back-projection stacks are indeed real. The main propagation direction during this subevent is updip, toward the trench. In fact, it appears that it may even reach the trench location, although the amplitude of high-frequency energy release is almost at the background level by the time the energy arrives at the trench.

The high-frequency mainshock rupture is nearly done with the fifth subevent except for another point-source-like event at about 195 seconds. This final event occurs at the edge of the rupture areas of subevents 3 and 4, at about 37°N and 142.8°E . Again, because of its size and proximity to previously ruptured area, it may be another early aftershock.

Using the normal back-projection stack, we can show the relative amplitude of the subevents as a function of time (Fig. 3(a)). Even though the episodes are identified based upon examination of the spatio-temporal behaviour of the coherency function, each episode has a complementary peak in the amplitude-time plot. Based upon this result, and

the analysis of the coherency function, the duration of the mainshock is around 210 seconds.

4.2 Frequency dependence

The relative source-time function at high and low frequencies considered in this paper are, in general, consistent with one another (Fig. 3(a)). The large energy release occurs at about 80 seconds into the event, and the duration is about 210 to 220 seconds. Despite the fact that the two frequencies are not forced to have the same epicentral time, the timing at which the amplitude increases to above background noise level agrees well.

The low-frequency source-time function is characterized by initial increase in energy, the main peak which decays away rather rapidly, and a second peak at the end of the event. The main energy release around 80 seconds is better defined than at high frequency. This is a clear peak with duration of about 70 seconds whereas the high-frequency energy increase is achieved in steps with multiple peaks associated with various subevents. Based upon this observation, some subevents are more effective at releasing high-frequency energy than low-frequency energy. The opposite happens at the end of the mainshock where the relative low-frequency amplitude at about 180 seconds is larger than the relative high-frequency amplitude.

Because the spatial resolution is poor at low frequency due to larger wavelength and longer period, detailed rupture characteristics, such as that revealed at high frequency, cannot be obtained. Figure 3(b) shows the 80-second peak at low frequency, and the lack of lateral resolution compared to high-frequency results is immediately obvious. For this peak in energy release, the locations of high- and low-frequency results match well, although the low-frequency peak is slightly south of the high-frequency region. On the other hand, the second peak at about 180 seconds is significantly offset between the high- and low-frequency energy release locations (Fig. 3(c)). The low-frequency energy release is occurring around the trench location, most likely at

Fig. 4. Comparison with plate interface coupling models. (a) High-frequency relative energy release (background colour within rupture areas) compared with slip-deficiency model based upon GPS data by Hashimoto *et al.* (2009) (blue contours). The red star with black outline indicates the epicentre as determined by the back-projection technique, and red dots show the locations of the aftershocks between March 11 and 17 (JMA catalogue). (b) Same as in (a) except that the blue contours show coupling strengths inferred by Loveless and Meade (2010). (c) Same as in (a) except that the comparison is made to source regions of historical tsunamigenic earthquakes provided by Hatori (1987).

the trench, while most of the high-frequency energy release does not reach the trench. This peak at low frequency follows the updip propagation of high-frequency energy that happens during subevent 5, eventually reaching the trench (Fig. 2). This suggests that once rupture propagates into the trench, the slip becomes slower, effectively releasing energy at lower frequency.

If one performs back-projection analysis at even lower frequency with additional data for improved spatial resolution (Kiser and Ishii, manuscript in preparation), the largest energy release appears updip and northeast of the epicentral location (Kiser and Ishii, manuscript in preparation). This location is consistent with seafloor GPS observations (e.g., Sato *et al.*, 2011), suggesting that major slip is associated with such low-frequency energy release.

5. Discussion

The total relative energy release map obtained using high-frequency data (Fig. 2) are compared to slip deficiency (Hashimoto *et al.*, 2009) and plate interface coupling (Loveless and Meade, 2010) models that have been derived based upon dense GPS observations in Japan (Fig. 4(a) and (b)). The peak energy release location imaged by back-projection analysis to be downdip of the epicentral location agrees well with areas of largest slip deficiency and highest coupling. The general outline of the rupture area also matches with areas that are expected to break seismically.

The high-frequency back-projection results can also be compared to source regions of past tsunamigenic earthquakes, and the general agreement is good (Fig. 4(c)). The main high-frequency energy release occurs in the area where many tsunamigenic earthquakes have repeated in the past. This highly locked zone has been producing magnitude 7.5 to 8.0 earthquakes with the recurrence time of about 37 years (The Headquarters for Earthquake Research Promotion, 2011). The three sources shown in Fig. 4(b) that lie within the northern rupture area are inferred source ar-

eas of 1915, 1936, and 1978 events (Hatori, 1987). The 1978 source region, which is the most downdip source along the coastline, are known to have ruptured in 1897, 1861, and 1835 (e.g., Hatori, 1987). To the south of this area, there have not been as many earthquakes with damaging tsunamis, but a portion of the southern rupture area imaged by the back-projection analysis overlaps with the source region of the 1938 tsunamigenic earthquake (e.g., Hatori, 1987; Fig. 4(c)). The southern part that broke during the March 11, 2011 mainshock is also not unknown to large earthquakes, and rupture regions of such earthquakes inferred from aftershock distribution have updip extent (Uchida *et al.*, 2009) that roughly matches with the broad updip extent of subevent 5 (Fig. 2).

Finally, recent studies (e.g., Minoura, 2008) have scientifically confirmed the earthquake of 869 A.D. (Jogan) and associated tsunami that devastated similar region as that affected by the event in 2011. Identification of old tsunami deposits has shown extensive inundation in areas around the Sendai plain from the Jogan earthquake (e.g., Sawai *et al.*, 2008). Tsunami simulations that attempt to produce similar inundation have shown that the earthquake source is unlikely to be at the outer rise or shallowest part of the plate interface (e.g., Satake *et al.*, 2008). The preferred source is within the seismogenic zone, with depth ranging somewhere between 15 and 50 km (Satake *et al.*, 2008). The southern part of the area that broke during subevent 2 in 2011 (Fig. 2) is within the proposed source region for the 869 A.D. Jogan earthquake, and may explain the level of tsunami damage in the Sendai area.

The back-projection results suggest that many segments of the plate interface that have broken independently in the past have moved during the mainshock of March 11, 2011. The rupture area shows involvement of at least three large patches with minimal overlap. This multi-segment process, as well as nearly complete breakage of the seismogenic zone, are similar to other recent great earthquakes

(e.g., 2010 Maule, Chile; Kiser and Ishii, 2011a, b). If we exclude the outer-rise activity, the distribution of after-shocks is also mostly consistent with inferred rupture area (Fig. 2), although the match is not as good as those observed for other great earthquakes (Kiser and Ishii, 2011b). One feature that has been observed in some finite-fault models (e.g., Ammon *et al.*, 2011; Hayes, 2011; Shao *et al.*, 2011) and not in the back-projection results (e.g., Hutko, 2011; Kiser, 2011; Wang and Mori, 2011; Yao *et al.*, 2011) is the large slip updip of the epicentre, extending almost to the trench. Even using the coherency function, no sign of high-frequency energy release from this region is observed, i.e., the gap in energy release close to the trench in Fig. 2 is real. In fact, this updip gap in high-frequency energy release seems to be where the finite-fault models indicate highest slip. This may mean that the slip mechanism in this region is very different from any of the recent great earthquakes, i.e., deformation or slip without any high-frequency energy release. In contrast, high-frequency energy release is observed up to the trench at the southern end of the main-shock rupture area. This area also releases energy at lower frequency, suggestive of possible near-trench deformation.

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