Along-dip variation of teleseismic short-period radiation from the 11 March 2011 Tohoku earthquake (Mw 9.0)

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[1] Locations of coherent short-period seismic wave radiation from the 11 March 2011 Tohoku earthquake (Mw 9.0) are imaged by back-projecting teleseismic P waves recorded across North America for a series of narrow, overlapping passbands centered at 8s, 4s, 2s, 1s, and 0.5s. Initially the energy release for all five passbands migrates slowly down-dip, however over time the two longer-period passbands show coherent energy release systematically shifted up-dip of the shorter-period source regions. Back-projection images of P waves from ten (point-source-like) aftershocks do not show a frequency-dependent trend, implying that the frequency dependence observed for the main shock is not an artifact created by 3D earth structure, depth phase interference, or some other deficiency. We conclude that the unstable sliding properties along the megathrust are segmented, with faster moment rate variations in the down-dip region and relatively smooth sliding further up-dip. Citation: Koper, K. D., A. R. Hutko, and T. Lay (2011), Along-dip variation of teleseismic short-period radiation from the 11 March 2011 Tohoku earthquake (Mw 9.0), Geophys. Res. Lett., 38, L21309, doi:10.1029/2011GL049689.

1. Introduction

[2] The Mw 9.0 Tohoku earthquake of 11 March 2011 is one of the few earthquakes that appears to have ruptured the entire width of the seismogenic megathrust, from trench to down-dip decoupling zone [e.g., Simons et al., 2011; Lay et al., 2011; Hayes, 2011; Yoshida et al., 2011; Ide et al., 2011]. Previous studies have established that there are depth-varying earthquake rupture processes on the Japan megathrust, including moment-scaled source duration [e.g., Bilek and Lay, 2000], rupture velocity [e.g., Kanamori, 1972; Bilek and Lay, 2002], and frequency content [e.g., Fukao and Kanjo, 1980]. It is thus not surprising to find that the total megathrust rupture of the 2011 Tohoku event appears to involve composite behavior with regions of large, coherent short-period energy release, as defined by back-projection of teleseismic P waves, located far down-dip of high-slip regions, as defined by inversion of longer-period teleseismic data [Koper et al., 2011].

[3] Quantifying the depth-varying properties of the Tohoku rupture is important for understanding the influence of pressure, temperature, fluid content, mineral phase, and other conditions on frictional behavior along the megathrust, and is relevant to estimating seismic hazard and strong ground motions from future great events. Here we extend analysis of the spatial variability in the spectrum radiated from the great Tohoku event by exploring the frequency dependence of back-projection images created in a series of narrow, overlapping passbands. Unlike our earlier work [Koper et al., 2011], we do not rely on comparisons between back-projection images of beam power and space-time models of slip, but only on properties of the back-projection images themselves. These images are equally affected by the assumptions and limitations of the technique, hence if frequency dependent results are obtained it lends further credence to the idea that differences between back-projection images and slip models are not artifacts of the different approximations made in the two methods, but rather reflect actual properties of the earthquake rupture.

2. Back-Projection of Varying Period P Waves

[4] One approach to characterizing the spatial distribution of the short-period portion of the seismic radiation spectrum involves back-projection of teleseismic body waves recorded across a regional or continental scale array [e.g., Ishii et al., 2005; Krüger and Ohrnberger, 2005]. This technique is most effective at short periods (<5 s) where the signal can be treated as a flux of energy with slowly changing phase that can be exploited to find space-time source locations of coherent radiation despite ignorance of the specific Green functions. It remains unclear precisely how such back-projection images are related to the total source rupture process and the degree to which depth phase interference corrupts the images. For a large event the short-period spectral amplitude is much lower than at long periods so seismic moment and total slip cannot be recovered. However, rupture aspects that are poorly resolved by long periods such as rupture front expansion rate, localized regions of relatively high particle slip velocities, and patches of high stress drop may be preferentially sensed by the short-periods and manifested in back-projection images. Back-projection appears to be rather robust, with several groups finding consistent results for different algorithms applied to short-period teleseismic P waves from the Tohoku earthquake [Ishii, 2011; Koper et al., 2011; Meng et al., 2011; Simons et al., 2011; Wang and Mori, 2011; Zhang et al., 2011].

[5] Back-projection imaging tends to smear beam power in time and space across the source region toward the receiver array, thus the preferred configuration of teleseismic seismometers for observing along-dip variations in the Tohoku rupture is an array of stations at azimuths along
Figure 1. Station locations for the 92 P waves used in the back-projection. The color indicates the average correlation coefficient for each waveform as determined by a multi-channel cross-correlation algorithm [VanDecar and Crosson, 1990]. The inset shows the shifted and normalized velocity waveforms.

The trench strike (≈22°N). Stations in North America are at azimuths of 21°–50°N and have a dense station distribution including the EarthScope Transportable Array, so we focus on this configuration rather than other geometries such as provided by large numbers of European stations (at azimuths of ≈290°–350°N).

Station locations for the 92 North American P waves that we ultimately used are shown in Figure 1. The traces have been aligned and normalized using the multi-channel cross-correlation algorithm of VanDecar and Crosson [1990]. The resulting time shifts represent station corrections that account for the 3D Earth structure between the source and receiver regions. These corrections are relative to a hypocenter of 38.19°N, 142.68°E, and 21 km [Chu et al., 2011], and with respect to the AK135 reference model [Kennett et al., 1995]. The 92 stations were selected from a total of about 650 candidate stations because of high coherency (minimum mean correlation coefficient of 0.88 when considered as a distinct group) and quasi-uniform geographical distribution. We experimented with a denser group of 378 P waves, with approximately the same aperture and similarity as shown in Figure 1, and found nearly identical results to what we present here (See Figures S1–S4 in the auxiliary material). We prefer the smaller geometry because the main-lobe of the array response function is slightly sharper and more symmetrical (Figure S5), leading to less along-dip smearing of the true energy maxima.

Our back-projection procedure closely follows that described by Xu et al. [2009] and Koper et al. [2011]. After the unfiltered P waves were aligned and normalized, we applied a 4-pole Butterworth filter centered at 0.5 s, 1 s, 2 s, 4 s, or 8 s. As shown in the auxiliary material (Figure S6), the combined amplitude responses of these filters cover the range of periods in which teleseismic back-projection is generally viable. At longer periods the spatial resolution provided by a regional configuration of seismometers becomes too poor to resolve rupture properties, while at shorter periods the P waves become too incoherent to interfere constructively when stacked over a large aperture array. Next the traces were back-projected to a 2D grid of points spaced at 0.1° intervals and spanning 139°E–145°E and 35°N–41°N. The depth was fixed at the hypocentral value of 21 km, and travel times were calculated using AK135 [Kennett et al., 1995]. Beam power was computed at 1 s intervals starting 20 s before the origin time and continuing until 200 s after. The power was measured from a tapered beam segment created with fourth-root stacking.

Figure 2 presents the locations of beam-power maxima at every fourth time step (i.e., every four seconds) for the back-projections computed in each of the five pass-bands. Circle size is proportional to beam power and color indicates time. These locations should not all be considered resolved sources of short-period energy because of the smearing effect of the array response, however this presentation provides a clearer sense of rupture propagation than would a time-integrated map of beam power. Animations of the back-projections are shown in the auxiliary material, Animation S1. Processing was identical in each case except for two parameters: the window length was set to three times the dominant period of the pass-band in order to equalize the number of cycles averaged in beam power calculation, and the length of time averaged during the post-processing was set at 5 times the lower end of the pass-band in order to mitigate the space-time streaking effect of the array response. For example, 14 frames (seconds) were averaged for the 2 s band, 7 frames (seconds) were averaged for the 1 s band, and so on. This is necessary because the time component of the array response becomes longer as period is increased.

As shown in Figure 2, locations of peak beam power for the three shorter-period bands are westward and down-dip of the corresponding locations for the two longer-period bands, especially after a lag time of 80 s. While the spread of beam energy is greater at longer periods, due to diminished resolution, the frequency dependent behavior is consistent with the expectation that the up-dip region slipped without strong short-period radiation, even while longer period seismic signals are coherent from the up-dip region. This is further emphasized in Figure 3, in which we compare locations of beam power maxima from broader bands of 0.7–3 s, and 3–11 s (see auxiliary material, Animation S2). The back-projection images in these bands are more robust than those calculated in the narrower bands of Figure 2 because the increased width of the pass-band leads to higher resolution in the time domain. Again a clear offset in the beam power maxima is observed, with shorter-period energy locating distinctly down-dip of the longer-period energy. This is true not just for the dominant northern segment of the rupture but also holds for the weaker and later southern segment. The time-integrated maximum of the short-period energy is located ~50 km to the NW of its longer-period counterpart, approximately in the down-dip direction (Figure 4).

3. Robustness Test

The longer period portion of the 0.35–11 s period range for which back-projection is viable overlaps the period range in which deterministic modeling of waveforms is successful. For periods longer than ~5 s, teleseismic Green functions can be computed with sufficient accuracy that space-time distributions of slip can be recovered from observed data. Such slip inversions provide more complete models of the rupture than back-projection images at longer period (~5–10 s) because they account for actual propaga-
tion effects between source and receiver. All current back-projection schemes for shallow events assume direct P (or PKIKP) propagation from source to receiver, and ignore energy associated with the free surface such as depth phases, near source crustal reverberations, and water layer multiples, all of which have similar slowness to the primary reference phase and therefore may interfere constructively in the stacks. Furthermore, back-projection imaging does not account for attenuation variations or radiation pattern variations across the network signals. No matter how sophisticated the array processing technique used to back-project the data, these problems persist, and one should view back-projection as somewhat blunt imaging of concentrations of short-period radiation rather than very precise imaging of attributes of the slip distribution.

[11] For these reasons one must be concerned about the possibility of interference effects producing an apparent frequency dependence of the source spectrum from varying depths. We test whether the observed frequency dependent pattern could involve some bias from interference effects by back-projecting P waves recorded at the same 92 North America stations from ten aftershocks that occurred within the first five hours after the main shock. The processing is exactly the same as before with the station corrections.

Figure 2. (a–e) Back-projection results for five narrow pass-bands showing beam power maxima for bands centered on 8 s, 4 s, 2 s, 1 s, and 0.5 s respectively. Every fourth beam power maximum is shown (i.e., every four seconds), with circle size proportional to beam power. The white star is the epicenter from Chu et al. [2011].

Figure 3. Comparison of back-projection results for a shorter-period band of (a) 0.7–3 s and a longer-period band of (b) 3–11 s. We show beam power maxima at every fourth time step (4 s) with circle size proportional to beam power. The white star is the epicenter from Chu et al. [2011].

Figure 4. Comparison of back-projection results for a shorter-period band of 0.7–3 s and a longer-period band of 3–11 s. We show time-integrated beam power for both cases with contours separated by 10% and maxima indicated by triangles. The white star is the epicenter from Chu et al. [2011].
determined for the main shock used for all the aftershocks. Figure 5 shows the concentrated beam peaks in two passbands (centered at 0.5-s and 4-s) for each aftershock. No systematic difference in time-integrated beam power (a proxy for the source centroids of these moderate size events) is found between the 0.5-s and 4-s bands for the aftershocks; however, the main shock retains the clear offset in this quantity. We infer that the corresponding frequency dependence in Figures 2, 3, and 4 is not the result of interference from 3D wave scattering or systematic variations in Green functions with source depth. If so, we would have seen a similar westward offset in the short-period beam power contours of the aftershocks.

[12] The other robustness issue we consider is the sensitivity of the back-projection results to the precise location of the epicenter. Because the MCCC-derived station corrections are defined relative to theoretical travel times from an assumed epicenter, our back-projection images are effectively tied to the assumed epicenter. We performed a parallel analysis to that shown above for the USGS/NEIC hypocenter of 38.322°N, 142.369°E, 24.4 km (about 30 km to the NW of the Chu et al. [2011] epicenter) and found that the resulting beam power is comparably shifted to the NW, but otherwise the relative patterns, including the frequency dependent offset, are unaffected. We infer that other hypocentral estimates for the Tohoku earthquake [e.g., Zhao et al., 2011] would likewise yield consistent back-projection results. The source location mainly enters into absolute positioning the short-period source locations relative to fault slip, which is not our focus here.

4. Discussion and Conclusions

[13] The seismic source spectrum is influenced by rupture dynamics and kinematics, with far-field signals sensing the convolutional interactions of rupture processes that make it challenging to isolate contributions from rupture expansion, particle dislocation and acceleration, fine-scale roughness, fracture front nonlinearities, etc. These source characteristics are likely to vary spatially for all megathrust events, and reconstructing how separate regions of the fault plane slipped and radiated seismic energy is a major undertaking. Advances in data quality and inversion strategies have enabled various combinations of seismic, GPS, INSAR, and tsunami data to be inverted for longer time scale and static motions of recent large earthquakes. Applications to the 2011 Tohoku earthquake for seismic wave periods from ~5 s up to static motions indicate massive slip, averaging 15–20 m across the megathrust, with a concentrated region of 40–60 m slip up-dip of the epicenter close to the trench [e.g., Ammon et al., 2011; Hayes, 2011; Ide et al., 2011; Lay et al., 2011; Simons et al., 2011; Yoshida et al., 2011]. These methods have limitations, but give relatively consistent solutions for large-scale slip distribution because reasonably accurate Green functions can be used for deterministic analysis of longer period signals.

[14] For periods less than about 10 s, the accuracy of Green functions decreases due to our ignorance of fine structure and approximations made in their calculation, and typically by about 5 s period there is very limited ability to account reliably for the seismic spectrum with deterministic models for large, offshore events. At these periods alternative methods such as back-projection imaging are required. For the 11 March 2011 Tohoku earthquake back-projection indicates that coherent shorter-period energy (<3 s) was primarily radiated from deep portions of the megathrust beneath the Honshu coastline, whereas longer-period (~3–11 s) energy was radiated from shallower depths up-dip to the east and near the hypocenter. The along-dip transition does not appear to be a gradual function of period and instead seems abrupt, although further work is needed to verify this observation and determine whether the apparent threshold in period is related to the structural heterogeneity in the source region. The largest slip, at even shallower megathrust depths, appears not to be accompanied by sufficient coherent teleseismic short-period seismic energy to be imaged by back-projection.

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