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Los Angeles

Theories and Applications of Enhanced Earthquake Back-projection Imaging

A dissertation submitted in partial satisfaction

of the requirements for the degree

Doctor of Philosophy in Geophysics and Space Physics

by

Ailin Zhang

2019

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ABSTRACT OF THE DISSERTATION

Theories and Applications of Enhanced Earthquake Back-projection Imaging

by

Ailin Zhang

Doctor of Philosophy in Geophysics and Space Physics

University of California, Los Angeles, 2019

Professor Lingsen Meng, Chair

Earthquake source imaging is an efficient tool to reveal the kinematic rupture process of earthquakes. The accuracy and robustness of source imaging is of vital importance to constrain source properties and help discriminate among potential physical mechanisms. In this work, I develop array-based approach: enhanced back-projection and coherence method to provide better constraints on kinematic source imaging.

Slowness correction is proposed to improve the source imaging of complicated rupture in conjugate fault systems. The correction is capable of mitigating 3D structural effects that commonly exist for M8 earthquakes.

For deep earthquakes, I introduce 3D back-projection which provides depth resolution of the rupture front. Resolving ruptures in 3D space helps us to understand the rupture process and physical mechanisms of deep earthquakes. Furthermore, coherence model is established to relate earthquake source size with inter-station cross-correlation. This method creates independent measures to validate and constrain the source sizes of deep earthquakes.

Aforementioned array-based approaches have been successfully applied to recently occurred earthquakes, and I will take this chance to revisit the rupture mechanisms of the events revealed by enhanced source imaging.

The dissertation of Ailin Zhang is approved.

Paul M. Davis

Gilles F. Peltzer

Yingnian Wu

Lingsen Meng, Committee Chair

University of California, Los Angeles

2019

To My Family

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Zhang, A., & Meng, L., Teleseismic Array Waveform Coherence Controlled by Earthquake Rupture Dimension (In revision).

CHAPTER 1

Introduction

Earthquakes correspond to transient energy release from the Earth's interior. An earthquake can produce significant ground shaking, leading to devastating natural hazards, building damages and even casualties. Studying source properties of an earthquake (e.g. rupture extent, speed, duration, etc..) is of great importance to understand the physics of the rupture and facilitate hazard assessment and damage mitigation. Numerous efforts have been devoted into deploying seismic instrumentation and developing methodologies to resolve earthquake source complexities. For example, earthquake catalogue, magnitude, and centroid moment tensor can be acquired almost in real time ([Storchak et al., 2013](#)), delivering fundamental information on the active tectonics. Currently, those basic parameters are routinely documented and maintained by a few agencies, such as NEIC, ISC, GFZ, JMA, etc. While the point-source based methods provide instant and robust estimations of the earthquake and tsunami source, in some scenarios, earthquake can be more complex and diversified. Earthquakes can navigate and branch through multiple fault segments, and display distinct energy release patterns in the source area. Rupture complexities are broadly reported for $M > 7$ quakes. Point source models underestimate the complexity and yield partial knowledge of the earthquakes. Therefore, a more detailed description of rupture complexity is desirable which will promote better understanding of rupture dynamics. Among all the strategies, Inversion-based and array-processing based techniques are most frequently adopted to resolve the finite source effects of complex ruptures. A brief introduction of the prevalent methodologies is provided in the following section.

1.1 Inversion based method

According to the representation theorem (Aki & Richards, 2002), the ground-motion $U_n(x, t)$ recorded at location x can be written as

$$U_n(x, t) = \int d\tau \iint_{\Sigma} [s_i(\xi, \tau)] c_{ijpq} v_j * G_{np,q}(x, t - \tau; \xi, 0) d\Sigma \quad (1.1)$$

Where rupture process is parameterised by $[s_i(\xi, \tau)]$: slip velocity recorded at location ξ and time τ . c_{ijpq} , v_j and $G_{np,q}$ are elastic tensor, normal vector of the fault plane and the spatial derivative in x_q direction of the Greens function respectively.

Slip histories can be discretized at each subfault and it can vary over the rupture area. In equation 1.1, the product of s_i , c_{ijpq} , and v_j yields a unit of moment per unit area, which is defined as moment density m_{pq} :

$$m_{pq} = s_i c_{ijpq} v_j \quad (1.2)$$

We can reformat equation 1.1 as the spatial integral of the convolution between the moment density and the Greens function:

$$U_n(x, t) = \int_{\Sigma} m_{pq} * G_{np,q} d\Sigma \quad (1.3)$$

Considering the rupture finiteness, a large event can be decomposed into the summation of multiple subevents with different moment tensors. Equation 1.1 and 1.3 set the theoretical basis to invert for source parameters from geophysical data. The two equations evolve into two major inversion based methods: finite fault inversion (Olson & Apsel, 1982; Hartzell & Heaton, 1983; Ji et al., 2002) and multiple point source inversion (Duputel et al., 2012; Y. Chen et al., 2014; Duputel & Rivera, 2017).

However, the inversion is often underdetermined, and it often yields remarkable differences for the same event. For example, the 1999 Izmit earthquake has been studied by multiple groups and the results have large variances (Ide et al., 2005). The non-uniqueness is originated from different parameterizations, simplification of the earths structure, imbal-

anced data coverage, etc. The difference between source models are evaluated and documented by the Source Inversion Validation (SIV) project. The physical interpretations of the inconsistency remains unclear.

1.2 Array-processing based method

The biases from source inversions encourage the whole community to develop new methods to obtain more reliable source properties. With the recent progress in deploying large-scale dense seismic arrays, array processing based methods starts to play an important role in source observations. Seismic array is a dense configuration of seismic stations, with a spacing of tens of kilometers, which is originally designed for military purposes such as nuclear monitoring (Rost & Thomas, 2002). The dense spatial sampling becomes superior in enhancing seismic signals (such as improving the signal-to-noise ratio through waveform stacking), which benefits both earth structure (Saltzer et al., 2001; Lin, Li, et al., 2013; Vidale & Earle, 2000) and earthquake source studies (Bolt et al., 1982; Goldstein & Archuleta, 1991). Back-projection (BP) analysis is one of the most successful applications in resolving earthquake source complexities with array processing strategies. Unlike aforementioned inversion based approach which relies on parametrization and simplifications, BP is data-driven and provides robust information about rupture complexity.

According to Kiser and Ishii (2017), the governing equation of back-projection power in the source grid j at time t can be written as:

$$s_j(t) = \sum_{i=1}^N \omega_i f_i(t + \tau_{ij}) \quad (1.4)$$

Where ω_i is the weighting coefficient, $f_i(t)$ is the waveform and τ_{ij} is the travel time from the i^{th} station to the j^{th} source grid. Among all the terms in equation 1.4, only τ_{ij} requires further calculation. So far, the most prevalent approach to calculate τ_{ij} is proposed by Ishii, which is approximated by travel times T_{ij}^{1d} from a one-dimensional reference model (IASPEI91, PREM, etc..) plus an empirical correction $\delta\tau_i$ for 3D structural effects:

$$\tau_{ij} = T_{ij}^{1d} + \delta\tau_i \quad (1.5)$$

The correction term is deemed as a constant for a given station, which is acquired by cross-correlation the first window of P-arrivals (Ishii et al., 2005).

The most intuitive approach is to linearly stack the recorded waveforms according to 1.4. This has been first adopted in studying the 2004 M 9.2 Sumatra earthquake (Ishii et al., 2005). BP imaging extracted from Hi-Net provides accurate estimation of rupture length, duration, and speed.

BP provides robust resolution at each time frame, since it only utilizes array processing and doesn't make assumptions on fault geometry or Greens function. Dense regional seismic arrays also provides coherent signal up to a relatively high frequency ~1Hz, resulting in high spatial resolutions of the BP images. Those advantages made BP an outstanding approach in kinematic source imaging that has been widely adopted in recent studies of large earthquakes (Walker et al., 2005; Vallée et al., 2008; Kiser & Ishii, 2012; Yagi et al., 2012; Okuwaki et al., 2014; Fan & Shearer, 2015; Nissen et al., 2016; D. Wang, Takeuchi, et al., 2016; Fan et al., 2017; Meng et al., 2019).

On the other hand, scientists devote extensive efforts to improve the BP approach. One of the major goals is to improve the data quality. This requires signal processing strategies to enhance the signal to noise ratio. Nth root stacking, MUSIC and correlation stacking are all successful applications in enhancing seismic signals (Y. Xu et al., 2009; Meng et al., 2011; Fukahata et al., 2013). It is also worth noting that BP may be subject to swimming effect due to imbalanced sampling of stations. Swimming effect refers to the trade off between location and time due to poor azimuthal coverage from regional array (Y. Xu et al., 2009). Reference window strategy (Meng, Ampuero, Sladen, & Rendon, 2012) and relative BP (Zhang & Ge, 2010) are proposed to mitigate smearing artifact. Meanwhile, Yao et al. (2011) proposed Compressive Sensing BP which focuses on the frequency dependent behavior during a large earthquake. Yagi et al. (2012) seek to add more physical interpretations of BP powers by deconvolving the signal with realistic Greens function in Hybrid back projection. And Fan

and Shearer (2015) utilize global array to improve the azimuthal coverage of BP.

Most of the aforementioned techniques are either focused in signal processing or physical interpretations, which utilize only the P wave train of a given event. There is still potential of improving BP accuracy by incorporating more events in the region or adding additional phases. The depth resolution is also poor in the aforementioned BP studies. Because P wave cannot independently constrain depth variation. Lack of depth resolution could be more problematic where fault geometries are generally unknown. In such circumstances, we might fail to constrain fault plane orientation, which is a fundamental question of earthquake rupture mechanism. Furthermore, BP is established upon point source assumptions at each time frame, and there is chance to incorporate finite source based models. The limitation of current BP approach motivates the research topics of my PhD work.

1.3 Motivation and structure of the dissertation

The motivation of my PhD research is to improve the accuracy and robustness of current back-projection and provide reliable benchmarks/ parameters for inversion-based source imaging. In the following chapters, I demonstrate how BP contributes to developing a more plausible physics-based earthquake source models and sheds light into the physical mechanism. The kinematic source imaging and dynamic rupture models will jointly improve the understanding of rupture complexities and provide physical constraints of the ruptures. In particular, I will focus on two types of complexities: earthquakes in conjugate fault systems and deep earthquakes. During my PhD, I enhance conventional BP for each case study and combine other source imaging techniques, to reveal high-resolution details of the rupture process and guidance in the earthquake source physics. The structure of the dissertation is briefly explained as following.

In Chapter 2, the enhanced BP is carefully explained. Two strategies are introduced: Slowness enhanced BP and 3D BP. 1) By calibrating slowness, BP is customized to mitigate systematic spatial biases for large scale ruptures of $M \geq 8$ events. 2) By combining direct P

and depth (pP) phases, depth resolution can be reasonably constrained for deep earthquakes. Applications of enhanced BP are discussed as case studies from Chapter 3 to Chapter 6. Chapter 3 is the source analysis for the Mw 7.8 Kaikoura earthquake, which serves as an example of how BP investigates complex ruptures in conjugated fault systems. The Kaikoura earthquake is well recorded by multiple geophysical datasets including local and teleseismic recordings, GPS time series, and satellite imaging which allows us to develop kinematic source models with BP and finite source inversions and discuss the mechanisms by dynamic rupture models. This chapter is incorporated from two manuscripts: Zhang, A., Wu, B., Meng, L., Kyriakopoulos, C., Oglesby, D., & Fang, L. Unraveling the Rupture Complexity of the 2016 Mw 7.8 Kaikoura, New Zealand Earthquake with Slowness-enhanced Back-Projections and Dynamic Rupture Simulations. (Submitted to JGR). and Xu, W., Feng, G., Meng, L., Zhang, A., Ampuero, J. P., Brgmann, R., & Fang, L. (2018). Transpressional rupture cascade of the 2016 Mw 7.8 Kaikoura earthquake, New Zealand. *Journal of Geophysical Research: Solid Earth*, 123(3), 2396-2409. I lead the study of BP part in both papers, and lead the discussion of the first paper.

Chapter 4 provides another example of complicated dynamic ruptures in the conjugate fault systems. The source complexity of the 2017 M7.9 Kodiak earthquake is discussed. This chapter is modified from Ruppert, N. A., Rollins, C., Zhang, A., Meng, L., Holtkamp, S. G., West, M. E., & Freymueller, J. T. (2018). Complex faulting and triggered rupture during the 2018 MW 7.9 offshore Kodiak, Alaska, earthquake. *Geophysical Research Letters*, 45(15), 7533-7541. I am the contributing author in this study to perform BP analysis and interpret the rupture triggered by ambient stress field. The BP in this study also provides prior knowledge of fault plane parameterization for the finite fault inversions.

Chapter 5 studies the 2015 Izu-Bonin deep earthquake. The combination of BP and Multiple Source Inversion (MSI) jointly shed light on the cascade failures of deep ruptures on multiple nearby faults. This chapter is modified from Chen, Y., Meng, L., Zhang, A., & Wen, L. (2018). Source complexity of the 2015 Mw 7.9 Bonin earthquake. *Geochemistry, Geophysics, Geosystems*, 19(7), 2109-2120. I am the contributing author to conduct the 3D

back-projections and estimate the uncertainty in this approach.

Chapter 6 documents the effort to study deep earthquakes in Tonga with the 3D BP projections. To the first order, the fault plane ambiguity for deep earthquake can be addressed with 3D BP. We provide detailed rupture process in 3D space for the Mw 8.2 and Mw 7.9 earthquake respectively and discuss the interaction between competing mechanisms phase transformation and shear instability.

Chapter 7 reports a new array-based strategy that can potentially resolve rupture dimensions of deep earthquake. Unlike BP that extracts information from waveforms, or source inversion that aims to minimize the residual between waveforms and synthetic data, here we compute waveform coherence between station pairs to extract source properties such as source dimensions. An example of modeling earthquake source sizes in Okhotsk is provided in this chapter. This chapter is modified from Zhang, A.& Meng, L., *Teleseismic Array Waveform Coherence Controlled by Earthquake Rupture Dimension*, which is in revision.

Chapter 8 is the concluding chapter to summarize the BP enhancements, source complexities and the coherence method. I will discuss key messages from these topics and provide future research directions.

CHAPTER 2

Theories in Enhanced BP

Abstract

This chapter introduces the theories in enhanced BP. For large ruptures at shallow depths, we observe systematic spatial bias of conventional BP, therefore I develop slowness correction to improve BP accuracy. For deep earthquake, 3D BP is developed to resolve the 3D spatial resolution of the rupture.

[2.2.1](#) is modified from the method part from Meng, L., Zhang, A., & Yagi, Y. (2016). Improving back projection imaging with a novel physicsbased aftershock calibration approach: A case study of the 2015 Gorkha earthquake. *Geophysical Research Letters*, 43(2), 628-636. I am the contributing author to perform the calibration and provide synthetic validations.

[2.2.2](#) and [2.2.3](#) is modified from the supplementary information from Zhang, A., Wu, B., Meng, L., Kyriakopoulos, C., Oglesby, D., & Fang, L. Unraveling the Rupture Complexity of the 2016 Mw 7.8 Kaikoura, New Zealand Earthquake with Slowness-enhanced Back-Projections and Dynamic Rupture Simulations. (Submitted to JGR). I am the joint first author of this paper to perform Slowness-enhanced Back-Projections.

2.1 Introduction

Back projection (BP) is a successful application of array processing technique to study earthquake source properties. Owing to its independence from fault geometry and rupture speed, BP is prevalently adopted to study the rupture process of large earthquakes ([Ide et al., 2005](#); [Y. Xu et al., 2009](#); [Zhang & Ge, 2010](#); [Ishii, 2011](#); [Kiser & Ishii, 2011](#); [Meng et al., 2011](#);

Meng, Ampuero, Stock, et al., 2012; Fan & Shearer, 2015; Zhang et al., 2017) However, BP may suffer from artifacts that leads to uncertainties and spatial biases. Arbitrary interpretation of high frequency radiators will yield wrong findings and produce erroneous conclusions. Therefore, it is essential to address the uncertainties and biases of BP.

Despite the lack of explicit equations to quantify the uncertainty, there exists some extent of approximations. Y. Xu et al. (2009) introduced Array Response Function (ARF) to evaluate the effect of array geometry on the resolution of BP. ARF at a grid of interest can be expressed as:

$$ARF = \left| \frac{1}{n} \sum_{j=1}^n \omega_j e^{2\pi i f \Delta t_j} \right|^2 \quad (2.1)$$

Where n is the number of stations, j is the weighting factor imposed on each station, f is frequency, and Δt_j is the travel time difference between the grid and the hypocenter at station j . Ideally, ARF should have a shape of delta function centered at the hypocenter within the first window of the initial P arrival. Therefore, the size of the major lobe in ARF can be used to justify the resolution of BP images for a given station setting. The smaller and more concentrated the lobe is, the higher the resolution. The rule of thumb of ARF is that global subarray (large scale array) tends to perform better than small local arrays, and the depth resolution is less well-constrained compared with 2d horizontal variation. ARF is calculated with the assumption of perfect waveform coherence, therefore it only represents the uncertainty embedded for a particular array. The second approach, complementary to ARF, is to perform sampling techniques to quantify the uncertainty embedded in seismic data. Sampling based techniques are non-parametric, which requires repeatedly resample subsets of the whole data space and perform BP on each subsets. Resampling methods, such as bootstrapping and jackknife, are adopted in evaluating the variances of parameters (uncertainties) in BP. For BP imaging, a comparison between uncertainties derived from ARF and sampling is shown in Figure 2.1.

When an array has good coherence, uncertainty will be more dependent on ARF, which is determined by array geometry. But if data coherence decays, uncertainty will rely on both ARF and data permutation. Another example of resampling with jackknife is documented in

([Fan & Shearer, 2016](#)) to resolve the variances in rupture speed estimations, which improves our understanding in statistical significance of BP. Besides ARF and resampling, it is worth noting that theoretical basis to address the biases hasnt been well established. Spatial biases in BP correspond to location offset from the BP radiator to the true location. However, during the rupture process of a large earthquake, it is hard to infer true locations. Thus, a quantitative and reliable factor to measure spatial biases merits investigation.

The motivation of improving conventional BP comes from the aforementioned resolution limits in depth and the spatial biases. In this chapter, I will first demonstrate the spatial biases we have extensively observed in BP and propose a metric to quantify the biases. After documenting the biases, I will explain the origins of the biases and introduce the slowness calibrations. Another enhancement to conventional BP is the 3D BP of deep earthquakes. The idea of 3D BP is to include additional phases (depth phase) that can help constrain the ARF in the vertical dimension.

2.2 Slowness enhanced back-projection

Despite the vacancy of reliable benchmarks about rupture finiteness of a large major earthquake, we are still able to utilize moderate events in the source area. Events with magnitudes larger than 7 can display predominant finite source effects and cannot be simply approximated by a point source. Yet moderate earthquakes (M5- M6) can be properly treated as a point source at teleseismic scales. Waveforms released from moderate earthquakes also have good signal to noise ratio at teleseismic distances. The coherence of waveforms ensures a clear and robust BP imaging. We use the exact same set of parameters as the mainshock to back-project moderate events, therefore HF radiators of those events can be regarded as subevents of the mainshock. Furthermore, considering the event locations determined with local seismic network is much more accurate than those inferred from teleseismic BP, the relative difference between ISC and NEIC catalog is typically at an order of 5 km ([Figure S2.1](#)). Therefore, we can treat the catalogue locations of aftershocks as ground truth. Meanwhile BP inferred locations are derived from the average BP location from the first 5s for

each event. The average offset between true locations and BP inferred locations is calculated to denote the spatial bias embedded in BP. We first observed such biases in the 2015 Mw 7.8 Nepal Gorkha earthquake (See Figure 2.2). Averaged spatial biases are 20 km, 21 km and 28 km respectively for EU, NA and AU Array.

We define the offset as biases rather than uncertainties because the pattern is systematic and non-random within an array. For example, according to Figure 2.2, inferred BP locations (diamonds) are drifting toward the EU array. Similarly, we observe a systematic bias of 24 km on average for the Illapel earthquake (Meng et al., 2018) and 27 km for the Kaikoura earthquake.

Such spatial bias can originate from insufficient hypocenter correction. The conceptual sketch is shown in Figure 2.3 as proposed by Fan and Shearer (2017) and theoretical equations are derived as below. To begin with, equation for the conventional beamforming BP can be written as:

$$BP(\xi, t) = \sum_j u_j(t + T_j^{mod}(\xi) + \delta T_j(\xi)) \quad (2.2)$$

Where $BP(\xi, t)$ is the beamforming power as a function of source location ξ and time t , $T_j^{mod}(\xi)$ is the theoretical travel time from source location ξ to j^{th} receiver derived from a reference model, and $\delta T_j(\xi)$ is a 3d travel time error term that accounts for the difference between true travel time and the theoretical travel time from source location ξ to j^{th} receiver. Under the assumption that the travel time error term is weakly dependent on the source region, we can approximate any $\delta T_j(\xi)$ with the travel time error measured at the hypocenter $\delta T_j(\xi_h)$:

$$\delta T_j(\xi) \approx \delta T_j(\xi_h) = T_j(\xi_h) - T_j^{mod}(\xi_h) \quad (2.3)$$

$\delta T_j(\xi_h)$ can be obtained by aligning the initial P-phase with cross-correlations assuming that the first arrival originates from the hypocenter. It has been proposed by (Ishii et al., 2005) for the 2004 Mw 9.2 Sumatra earthquake. In theory, the approximation is valid only in the vicinity of the hypocenter ($|\xi - \xi_h|$ is small enough). The approximation is likely to break down if the rupture length is larger than 100 km, which is the case for all M 8 events.

2.2.1 1D calibration

The overarching goal of the calibration is to more accurately quantify $\delta T_j(\xi)$ for given $\delta T_j(\xi_h)$:

$$\begin{aligned}\delta T_j(\xi) &= T_j(\xi) - T_j^{mod}(\xi) \\ \delta T_j(\xi_h) &= T_j(\xi_h) - T_j^{mod}(\xi_h)\end{aligned}\tag{2.4}$$

Therefore,

$$\delta T_j(\xi) - \delta T_j(\xi_h) = T_j(\xi) - T_j^{mod}(\xi) - T_j(\xi_h) - T_j^{mod}(\xi_h)\tag{2.5}$$

Based on the far-field approximation (Aki & Richards, 2002), differential travel time $T_j(\xi) - T_j(\xi_h)$ can be obtained by the product between difference between their distances projected along the ray direction r_j and the slowness s_j (Sketched in Figure 2.4).

$$T_j(\xi) - T_j^{mod}(\xi) = s_j r_j (\xi - \xi_h)\tag{2.6}$$

where r_j is ray direction near the source, s_j is the apparent slowness (ray parameter) evaluated from the hypocenter to the station j.

Here we treat the slowness s , as a scalar along the ray direction, which holds true if the first order deviation of the velocity model is a 1d perturbation. Therefore,

$$\delta s_j = \frac{\delta T_j(\xi) - \delta T_j(\xi_h)}{r_j(\xi - \xi_h)}\tag{2.7}$$

Equation 2.7 can be verified by a synthetic test with known δs_j . We construct the true model as a Gaussian velocity perturbation with a half width of 250 km and a peak value of 1 km/s centered at depth 1000 km from the 1D reference model (PREM) (Figure 2.4). We make the perturbation very large so that distinctions between theoretical and calculated δs_j can be captured easily. The perturbation is visualized in Figure 2.4. We also compute synthetic seismograms with the true travel time. If s_j is a scalar along the ray direction, δs_j derived from equation 2.7 will be identical to δs_j that is explicitly known for given models. The comparison is shown in Figure 2.4(c), which proves that the two quantities are equal.

We further implement equation 2.7 for real cases. For the 2015 Nepal-Gorkha earthquake. We pick moderate events as M5 - M6 aftershocks that are listed in Table S2.1. δT_j can

be obtained by multichannel waveform cross-correlations (VanDecar & Crosson, 1990) at aftershock locations. Multichannel cross correlation is favorable because it provides robust and accurate estimation of differential arrival times between stations. After measuring δT_j , we will be able to solve and incorporate δs_j to back-project the moderate aftershocks, i.e. BP is performed under:

$$BP(\xi, t) = \sum_j u_j(t + T_j(\xi_h) + (s_j + \delta s_j)r_j(\xi - \xi_h)) \quad (2.8)$$

As shown in Figure 2.4, the 1d calibration significantly mitigates the spatial bias of BP. The bias reduces from 20 km to 9 km for EU array, 21 km to 5 km for NA array and 28 km to 15 km for AU Array. After the correction, BP inferred location is more consistent with the catalog locations. Moreover, corrected results from the three arrays agree with each other better. The remaining discrepancy between arrays is on the order of 10 km, which can be interpreted as catalog location uncertainties of BP.

The correction improves the aftershock BP significantly, hence it is a more accurate BP for the whole source region and has the potential to improve mainshock BP. It is clear from Figure 2.2 that the mainshock BP is also improved by the 1d calibration. Multiple array BPs are now in better consistence with each other.

2.2.2 2D calibration

1D calibration treats the slowness correction term as a scalar along the ray path. The calibration can be deemed as back projecting the waveforms with a refined 1d model, where all the time correction is along the ray path. However, there are more complex scenarios that slowness correction term can be a vector pointing to arbitrary directions. This will correspond to 3D heterogeneities of the true earth model. Thus, we also propose a 2D calibration which work better for a more complex setting.

The modification is made on (2.7), where δs_j is parameterised as a vector rather than a scalar,

$$\delta T_j(\xi) = \delta T_j(\xi_h) + \delta s_j(\xi - \xi_h) \quad (2.9)$$

The product of δs and $\xi - \xi_h$ can be considered as an inner product of two vectors. Because teleseismic P wave travel time has low sensitivity to depth, we simplify the equation by ignoring $\frac{\partial T}{\partial \xi_{UD,j}}$:

$$\begin{aligned}\delta T_j(\xi) &= \delta T_j(\xi_h) + \delta s_{NS,j}(\xi - \xi_h)_{NS} + \delta s_{EW,j}(\xi - \xi_h)_{EW} + \delta s_{UD,j}(\xi - \xi_h)_{UD} \\ &= \delta T_j(\xi_h) + \delta s_{NS,j}(\xi - \xi_h)_{NS} + \delta s_{EW,j}(\xi - \xi_h)_{EW}\end{aligned}\quad (2.10)$$

Where EW, NS, UD denote three orthogonal directions of East-West, North-South, and Up-Down respectively. The horizontal slowness vector $(\delta s_{NS,j}, \delta s_{EW,j})$ can be inverted with data from at least two aftershocks.

According to equation 2.10, for each aftershock, ξ is the aftershock hypocenter from earthquake catalogue. $\delta T_j(\xi_h)$ is the hypocenter correction addressed by a multichannel cross correlation (VanDecar & Crosson, 1990). Note the time residuals represents relative time differences between stations yet it doesn't have an absolute reference (e.g. origin time). Although the absolute reference can be acquired from catalogues, it is not necessary for BP, and we are only interested in relative differences. Concepts of relative time differences are more rigorously demonstrated as follow.

We introduce a reference station *ref* and the travel time error $\delta T_{ref}(\xi)$:

$$\delta T_{ref}(\xi) = \delta T_{ref}(\xi_h) + \delta s_{NS,ref}(\xi - \xi_h)_{NS} + \delta s_{EW,ref}(\xi - \xi_h)_{EW}\quad (2.11)$$

The choice of *ref* can be arbitrary, because the differential travel time is self-consistent in multichannel cross correlation. We then take the difference between Equation 2.10 and 2.11:

$$\begin{aligned}\Delta T_j(\xi) &= \delta T_j(\xi) - \delta T_{ref}(\xi) \\ &= \Delta T_j(\xi_h) + \Delta \delta s_{NS,j}(\xi - \xi_h)_{NS} + \Delta \delta s_{EW,j}(\xi - \xi_h)_{EW}\end{aligned}\quad (2.12)$$

Where Δ denote the relative differences with respect to the reference station, equivalent to differences of time residuals in cross-correlation. Then, two unknown terms $\Delta \delta s_{NS,j}$ and $\Delta \delta s_{EW,j}$ can be solved with a linear regression.

$$\Delta T = [\xi - \xi_h] \cdot \Delta \delta s \quad (2.13)$$

Equation 2.13 is a matrix form of the inverting $\Delta \delta s$. To avoid overfitting, a ridge regression is introduced, which incorporates an additional term $\lambda ||\Delta \delta s||^2$ to penalize extremely large $\Delta \delta s$. Therefore, our ultimate goal is to minimize R:

$$R(\Delta \delta s) = ||\Delta T - [\xi - \xi_h] \cdot \Delta \delta s||^2 + \lambda ||\Delta \delta s||^2 \quad (2.14)$$

Where R is the residual function, and λ is the regularization factor that controls the strength of ridge regression. Once we solve $\Delta \delta s$, we can update differential travel time for station j at any source grid ξ to be:

$$\Delta T_j(\xi) = \Delta T_j(\xi_h) + \Delta \delta s_{NS,j}(\xi - \xi_h)_{NS} + \Delta \delta s_{EW,j}(\xi - \xi_h)_{EW} \quad (2.15)$$

The 2D calibration is performed on the Kaikoura earthquake sequence, we use 7 Mw > 5.5 aftershocks (Table S2.2), located in the mainshock rupture area. BP images are shown in Figure 2.5. The spatial bias is 27 km, motivating the calibration. To maintain accurate timing measurements, we only keep stations with high mean correlation-coefficient (> 0.7) for at least 4 aftershocks. Around 140 stations are included for the slowness calibration.

The determination of λ in the ridge regression follows a process of cross-validation. It generally follows the leave-one-out strategy. Each time, we put one earthquake in the testing group, and the others in the training group. Slowness will be purely regressed with the data in the training group, and we can therefore calculate the error for the two groups with the inverted data. For this study, we have 7 earthquakes, thus this will be repeat 7 times by putting different earthquakes in the testing group. We then plot the averaged training error and testing error as a function of λ . λ will be chosen when testing error reaches minimum. Selection of λ is similar to L curve used in inversion to estimate parameter of interest.(See Figure 2.6)

Once we solve for the optimal slowness vectors, we can apply the correction to the aftershocks. Figure 2.5 (d) and (e) compare BP inferred and catalogue locations before and

after calibration. Before calibration, RMS spatial bias is 27 km. After calibration, it reduces to 6 km. The calibration mitigates nearly 80% of the spatial bias, hence improving the accuracy of BP.

2.2.3 Indirect measurement

The ray-theory based slowness calibration mitigates spatial biases that are embedded in conventional back-projections due to insufficient timing correction. The calibration utilizes the time error between mainshock and aftershock alignment, which requires accurate differential travel time measurement for the two datasets. To ensure accuracy, we set a cut-off threshold to only keep coherent aftershock seismograms with high correlation coefficient (0.7). Such a threshold limits the use of aftershocks with smaller magnitudes, the seismograms of which tend to be less coherent due to lower signal to noise ratio.

Nevertheless, we propose an indirect slowness calibration approach which is less restrained in terms of data selection. In the alternative approach, we use a proxy to represent the direct measurement of time errors for aftershocks. We first back-project the aftershock data with the timing correction derived from mainshock hypocenter alignment. The BP-inferred aftershock location provides the optimal time-shifts that maximize the beamforming power. Thus, the differential travel-time from the mainshock hypocenter to the BP inferred location in the 1D reference model becomes a substitute for direct time-lag measurements from cross correlation. This proxy is similar to the grid searching method to locate tremors and extract the travel time (Wech & Creager, 2008). Tremor signals are usually weak and embedded in the noise. Direct time-lag measurements from data may suffer large uncertainties. Therefore, Wech and Creager (2008) consider the waveform coherence as the raw data instead of the seismogram itself. By calculating waveform coherence at all grids with time-shifts calculated from a reference model, the optimal location maximizes the coherence function, which is also informative in providing robust relative travel time pattern for all the stations.

Therefore, we can replace direct measurement of $\delta T_j(\xi) - \delta T_{ref}(\xi)$ in equation. 2.11 with

a BP inferred time error:

$$\Delta T_j(\xi) = \delta T_j(\xi) - \delta T_{ref}(\xi) = [T_j^{mod}(\xi') - T_j^{mod}(\xi_h)] - [T_{ref}^{mod}(\xi') - T_{ref}^{mod}(\xi_h)] \quad (2.16)$$

where $T_j^{mod}(\xi)$ is the travel time from source location ξ to j^{th} receiver derived from the reference model. For each aftershock, ξ is the catalogue location, and ξ' is the BP inferred location. $\delta T_j(\xi)$ of arbitrary stations can be calculated using equation 2.16 regardless of data quality.

Once we get the proxy of the time errors, we can reproduce the inversion of equation 2.14 and 2.15. Figure 2.5. (d) and (e) compare calibrated aftershocks of the direct and indirect slowness calibration. Both of the methods are effective in mitigating spatial biases. For the direct calibration, average spatial bias reduces from 27 km to 6 km. For the indirect calibration, average spatial bias reduces from 27 km to 9 km.

For this particular case study, the direct calibration is slightly more accurate than the indirect approach. However, we expect indirect calibration to be more widely used in areas with few large aftershocks. Kiser and Ishii (2013) point out that BP can image aftershocks as small as Mw 4. One potential application for the indirect approach is for aftershock detection and relocation in continuous data. We can first apply a crude BP on the whole wave train, then pick up high-energy bursts that match with the existing aftershock catalogue. The spatial biases between the catalogue location and the BP inferred location can be used to invert for slowness vectors. We can then reproduce a calibrated BP on the continuous data, which not only detects hidden aftershocks (similar to Kiser and Ishii (2013)) but also refines the spatial accuracy for the entire earthquake sequence.

2.3 3D Back-projection(BP)

P wave recorded by a seismic array is generally insensitive to depth variation at teleseismic distance. In order to improve the depth resolution of BP, the combination of depth phases will be beneficial. At teleseismic distance, the ray paths of direct P and pP differ in take-off angles. P wave has a downward propagation and pP has an upward propagation (See

Figure 2.7). We can back project the two phases respectively and then stack the powers. The combination of upward and downward phases improves the depth resolution, primarily because their ray parameters and ARFs are different. Adding up two dissimilar ARFs strengthens the region with the most coherent signals and allows us to constrain vertical variation. This approach is also analogous to multiarray BP. Combining multiple arrays help reduce the lobe imaged by ARF. Follow the same strategy for evaluating 2D BP, we will discuss the ARF, Uncertainty and spatial biases embedded in 3D BP with real examples.

2.3.1 ARF of 3D BP

3D BP is the sum of contribution from both P and pP phases, which requires that P and pP are both identifiable in the seismic recording. P wave is the first clear wave train from the seismic recording, which tends to have higher signal to noise ratio. For pP phase, it can be either contaminated by coda waves due to scattering or mixed with other phases that share similar ray parameters. Such a limitation makes 3D BP more plausible for deep earthquakes. Deep earthquakes have simpler waveforms compared with its shallow counterparts. At deeper depths, pP is separable from other phases. The depth resolution is further investigated by ARF. We compute ARFs of USArray for 2015 Mw 7.9 Bonin earthquake in Figure 2.8.

It is obvious that the depth resolution is the highest for combined P and pP wave. Stacking two phases helps to emphasize the joint lobe, which is more likely to be the actual source of HF radiator. Combining multiple phases also improves the horizontal resolution. ARF provides solid evidence that deep earthquakes can be better imaged by 3D BP.

2.3.2 Uncertainties of 3D BP

To address the location uncertainties of 3D BP, we adopt the resampling technique. It is performed with bootstrapping. The idea is to provide a three-dimensional uncertainty ellipsoid for the BP radiator at each frame. The ellipsoid will be crucial to understand the trade-offs between horizontal and vertical scales and set the confidence region for each radiator. We also introduce random noises to the waveforms to better mimic real uncertainties.

First, a reference point-source location in each snapshot is selected as the peak of the back-projection image. Seismograms are then aligned and stacked with the time shift predicted by the reference location, which becomes our signal at this time window. Noise for each station is the difference between waveform and signal. For each realization, we take 80 % of stations with random noise to perform BP. The noise is first shuffled by randomizing its Fourier phase spectrum, then added back to the original waveform. One hundred realizations are repeatedly to obtain the 95% confidence bound on the true peak location.

Uncertainty of the 2014 Mw 8.3 Okhotsk earthquake is shown in Figure 2.9. From the 2D horizontal map view, HF radiators generally agree with previous studies (Meng et al., 2014; Ye et al., 2013; Zhan et al., 2014; Wei et al., 2013). The first segment is a rupture along the slab contour (AA'), but the major power is released on a secondary rupture which extensively span a wide area BB' . HF radiators have distinct patterns on AA' and BB' , which suggests different mechanisms on the two segments. The 95% confidence bound suggests an interval on the order of ~ 10 km for every dimension, similar to what we have observed in 2D BP.

2.3.3 Spatial biases in 3D BP

In order to address the spatial bias, we inherit the same strategy to calculate the biases embedded in moderate events near the mainshock. We chose a M5.5 and a M6.0 deep-earthquake that are closely spaced with each other (150 km horizontally and 50 km vertically away) near Izu-Bonin island. We treat the M 6.0 as mainshock and adopt the same set of parameters to back project the M 5.5 events. Thus M 5.5 can be deemed as a subevent of the M 6.0 quake. BP images are shown in Figure 2.10.

According to the figure, bias between the BP inferred and catalog location of M5.5 event is on the order of 15 km, considerably smaller than shallow regions as discussed in slowness calibrations.

2.4 Discussion

2.4.1 Slowness Calibration

The Spatial biases is notable for large earthquakes, and quite a few research groups identified the spatial biases by examining seismicity near a large earthquake (D. Wang et al., 2018; Zhang et al., 2017). The bias is reported to be 30 km for whole Japan island receive be USArray (Fan & Shearer, 2017), and 20 km for the 2010 Mw8.8 Maule earthquake (Palo et al., 2014). While the source of the bias is acknowledged by the community, the development of effective calibration has been lagged behind. Ishii et al. (2007) first proposed an empirical correction based on a weighted linear interpolation of travel time difference on each grid from 110 aftershocks for Mw 9.2 Sumatra earthquake. The minimum magnitude is 5 in the calibration. The calibrated BP contains significant incoherent artifacts that are less smooth than the original BP. Although this is an attempt to improve travel time errors, it introduces additional biases from uncertain measurement introduced from aftershocks. Palo et al. (2014) introduced both a static smoothing and a dynamic kriging term to correct the travel-time errors. He utilized 23 aftershocks to calibrate the 2010 Mw 8.8 Maule events. The calibration is similar to Ishii et al. (2007)s because it is interpolation bases, but kriging and smoothing play an important role to reduce data uncertainties and remove outliers. A significant drawback for the interpolation-based calibration is that interpolation requires a dense spatial distribution of aftershocks as well as high data quality. So far, such empirical aftershock calibrations have not been routinely integrated in BP method, because for many events whose large aftershocks are sparse, improper interpolation would bias the BP results and produce additional artifacts. The slowness calibration is robust because we refine the model to capture the first-order slowness deviation. It is more generalized and has been more widely used (Meng et al., 2018; Sangha et al., 2017; Bao et al., 2019) to mitigate the spatial bias. Another effective calibration is to incorporate 3D Global tomography model (Z. Liu et al., 2017). Unlike the aforementioned that requires accurate knowledge about aftershocks, the 3D model BP doesnt rely on aftershocks. 3D model BP is a universal model that can be

performed to any area. The more accurate the 3D model is, the better BP will perform. On the other side, 3D Global tomography model is smoothed and regularized to fit all the data, this will remove some regional small-scale heterogeneity. Therefore, 3D model BP may not be capable of extracting all the details during a rupture, but it exhibits improvement from the 1d reference model.

To sum up, in order to acquire a more accurate BP imaging, we will favor the slowness calibration as long as there are a few aftershocks in the area.

2.4.2 Evaluation of 3D BP

3D BP creates a chance to study the rupture process of deep earthquakes. Due to deficit of geophysical instruments at depths, physical mechanisms of deep earthquakes are still under debate. Reliable observations of the kinematic rupture process are crucial to identify the governing mechanism. Therefore, 3D BP becomes a powerful tool to constrain the rupture process. It is encouraging that 3D BP provides reasonable resolution and accuracy, allowing us to interpret the BP radiators with confidence. 3D BP may also suffer from the model heterogeneity, but we don't introduce slowness correction for 3D BP, as the spatial biases observed in the aftershock test is reasonably small. Furthermore, 3D BP requires slowness correction in 3D space, and this will introduce additional artifacts when observational data is sparse or suffers uncertainty. According to [2.3.2](#) and [2.3.3](#), the spatial bias is similar to the size of uncertainty bound. Also, deep earthquakes have large variations in depth, it is very likely that there is no available moderate event or events are too sparse to be taken advantage of.

2.5 Conclusion

In this chapter, enhanced BP imaging techniques are proposed, validated and successfully applied. For shallow earthquakes, both 1d and 2d calibrations are derived. They will fit regions with different extents of structural heterogeneity. We also propose two types of

measurements to accommodate different data quality. For events with clear waveforms and high signal to noise ratio, we can directly measure time errors with multichannel cross-correlation. On the other side, we can infer time errors from the spatial biases, which will be beneficial when multichannel cross-correlation breaks down for low signal to noise ratio. For deep earthquakes, 3D BP proves to be an effective approach that will have high resolution on all the dimensions. The uncertainty and bias are reasonable, enable us to interpret rupture process and resolve the mechanisms for deep earthquakes.

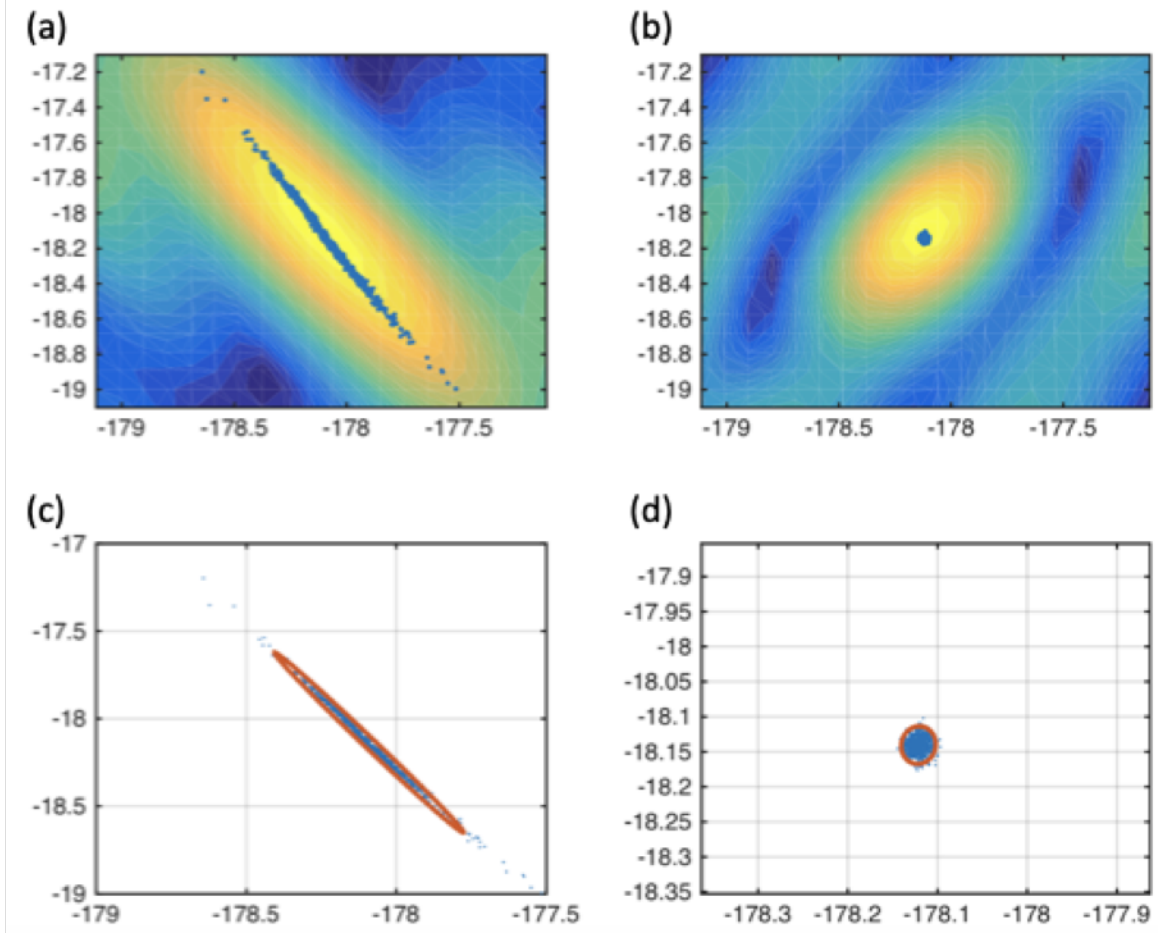


Figure 2.1: Comparison between array geometry and data induced uncertainty. Basemap in (a) and (b) are maps of ARF for Hinet and North America array for the 2018 Mw 8.2 Tonga earthquake. Each Dot indicates a realization of the location inferred from the beamform power. The distribution of the dots denotes the uncertainty derived from sampling. (c) and (d) is a zoomed map of the spatial uncertainty from sampling. We use an ellipse to enclose 95% of the dots to quantify the uncertainty.

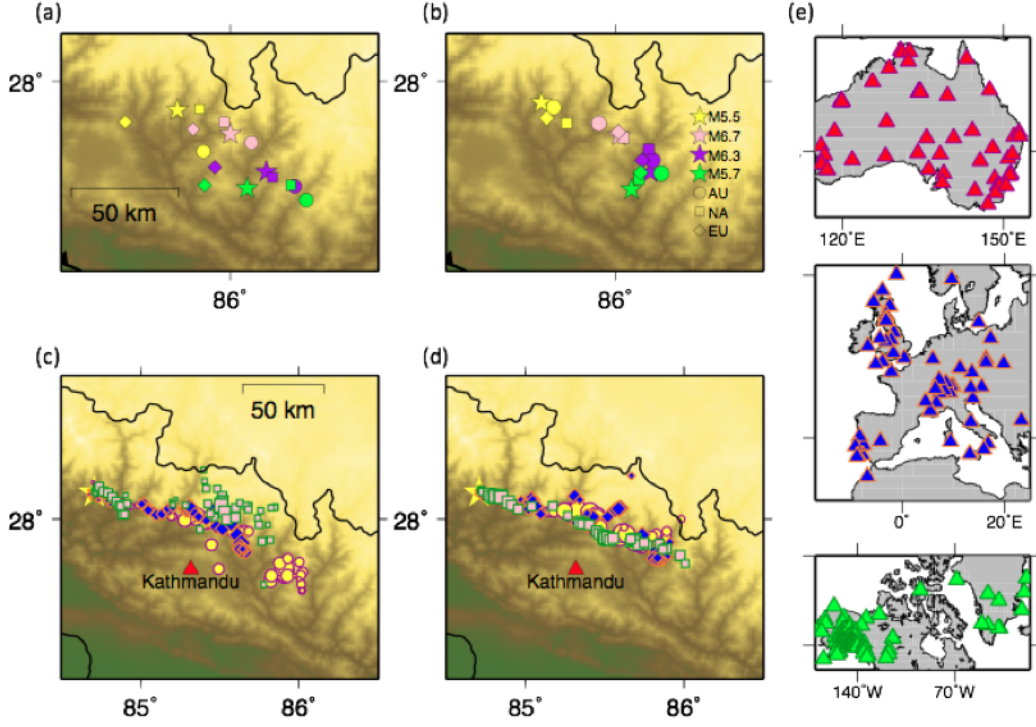


Figure 2.2: BP radiators for mainshocks and aftershocks in the rupture area and station settings. (a) shows aftershock locations inferred from BP with catalog location and (b) shows the aftershock locations using the slowness correction. The stars show NEIC catalog locations of four moderate-size (M5–7) aftershocks, with colors corresponding to four aftershocks. The symbols (circle, square and diamond) denote apparent BP locations imaged by the AU, NA and EU array. (c) shows the mainshock BPs of the three arrays before the correction and (d) shows the mainshock BP after the corrections are applied. (e) represents station distribution of each array.

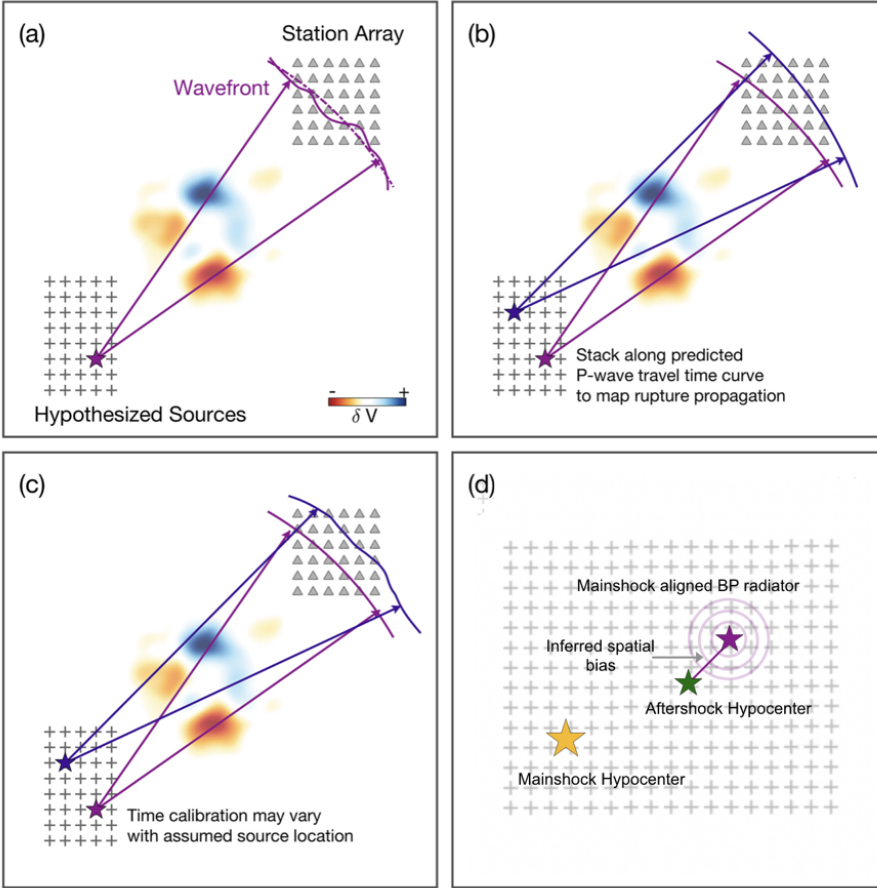


Figure 2.3: A conceptual sketch of the origin of the spatial biases due to empirical hypocenter correction. Modified from Fan et al., (2017). (a) demonstrates the empirical hypocenter correction at the hypocenter (purple star). The distorted wavefront (solid purple line) is produced from the heterogeneity of the 3D structural effects. Empirical time correction is performed to each station so that the wavefront can be calibrated to match the dashed purple line. (b) Ideally, if another source grid (blue star) shares the same 3d travel time error as the hypocenter, if we apply the same empirical time correction, the wavefront from that location can be accurately predicted by a 1d travel time different. (c) However, the empirical time correction may not be uniform in the source area. In reality the wavefront from the source grid can be distorted slightly from that predicted in (b). (d) Such a error in time domain will be convert in to biases in spatial domain. For an aftershock with known hypocenter (purple star), BP inferred location is biased from its true location due to the error in empirical time correction.

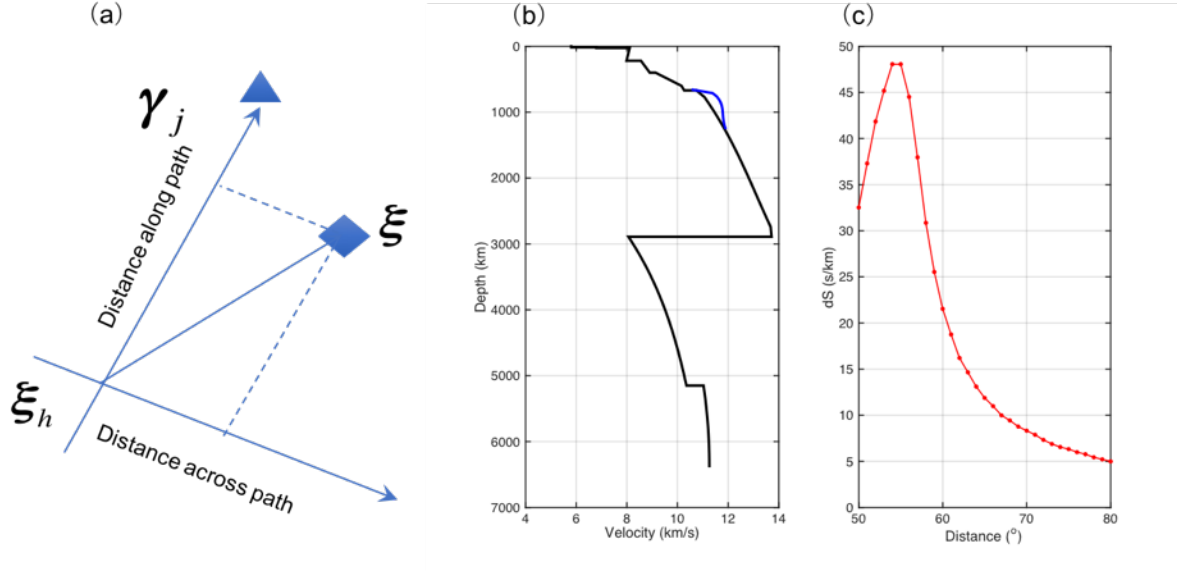


Figure 2.4: Model setup of 1d calibration. (a) is a sketch for far-field approximation which demonstrates mainshock hypocenter, aftershock hypocenter, and ray vector point toward the array. (b) Model comparison between the true model and the PREM model. The difference between true model and PREM is marked by blue, the rest of the model is plotted by black solid line, which is the same for both models. (c) Theoretical slowness error inferred from model is marked by red line, and slowness error calculated for some random stations are marked by red dots. The two quantities are identical, validating equation 2.6

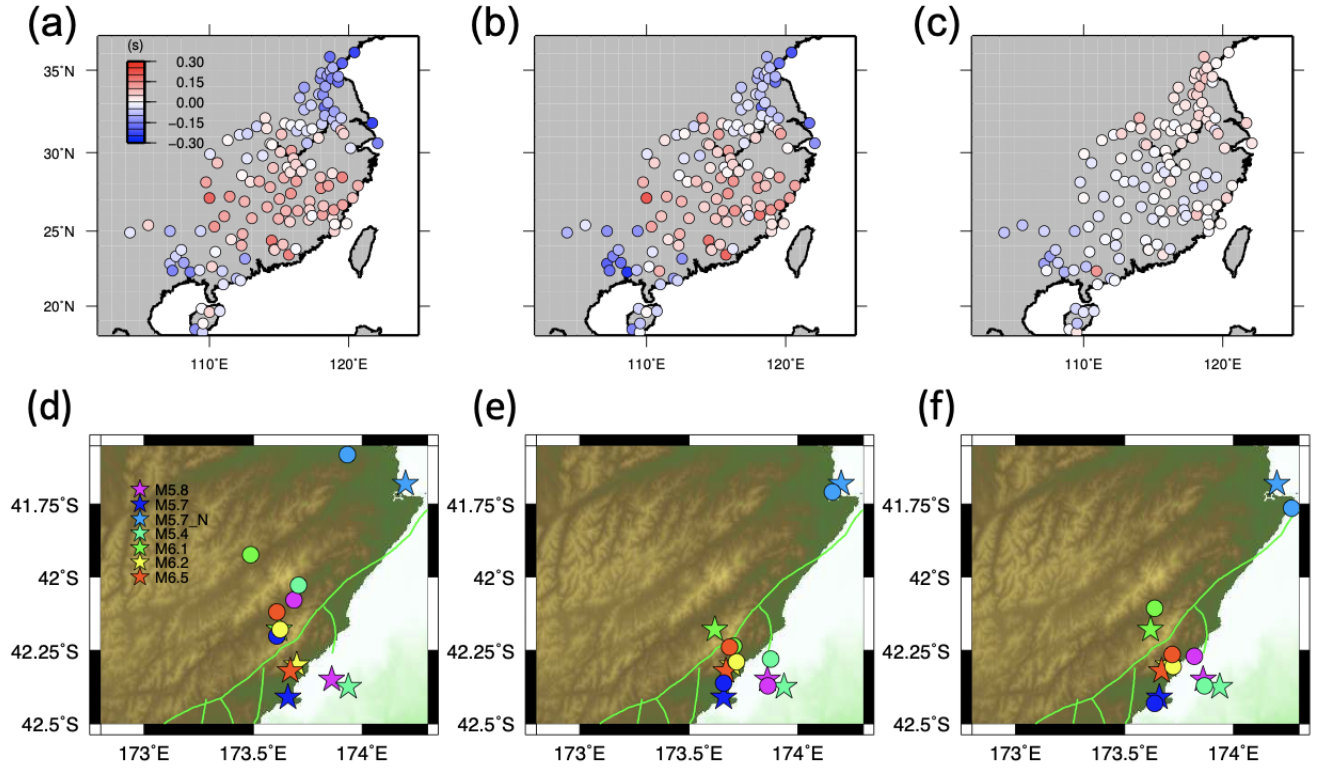


Figure 2.5: 3D travel time errors of China Array and spatial biases recorded in the Kaikoura earthquake rupture area. (a) - (c) are mapviews of relative differential time errors $T_j(\xi)$. (a) Time errors observed from the M 5.7 event at each station in China Array. Dots represent stations adopted for calibration and are color-coded by time error. (b) Time errors predicted by the slowness calibration. (c) Residual between the observed and predicted time errors. (d) - (f) compare spatial biases between catalogue location (star) and BP inferred location (circle). Events are differentiated by colors, and other information can be found in Table S2.2. In details, (d) represents spatial bias without calibration, (e) remaining biases after slowness calibration and (f) remaining biases after indirect slowness calibration. Yellow star is the mainshock hypocenter from NEIC catalogue, and empty circles are background seismicity in that region.

uclathes-2.0/./Picture6.png

Figure 2.6: Training error and testing error of the slowness regression. Red dots are averaged MSE of training data. And the blue dots are for testing data. The Training error reaches minimum at $\lambda = 40$.

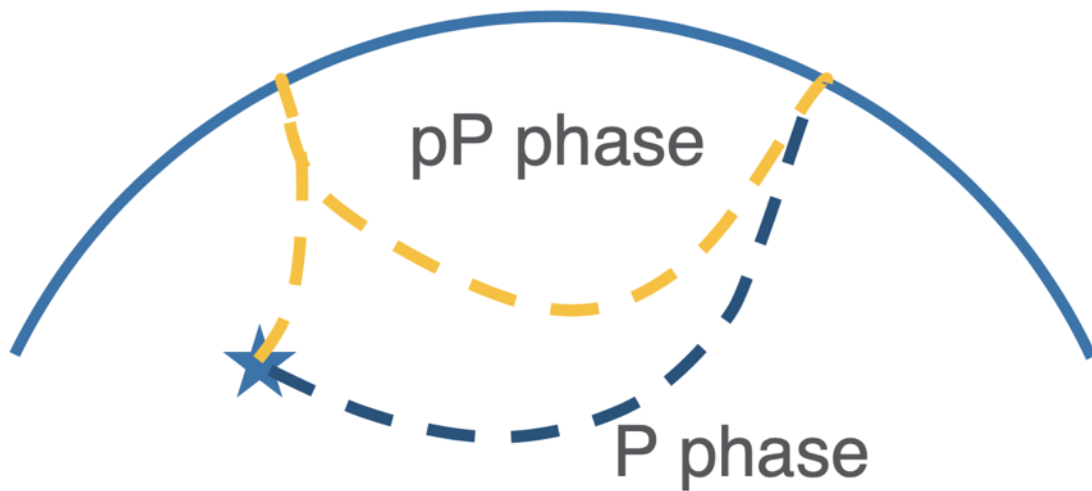


Figure 2.7: Sketch of ray paths of P and pP phases. Dashed lines are simplified representation of ray paths in the earth interior. P phase is propagating downward, and pP phase is propagating upward from the hypocenter.

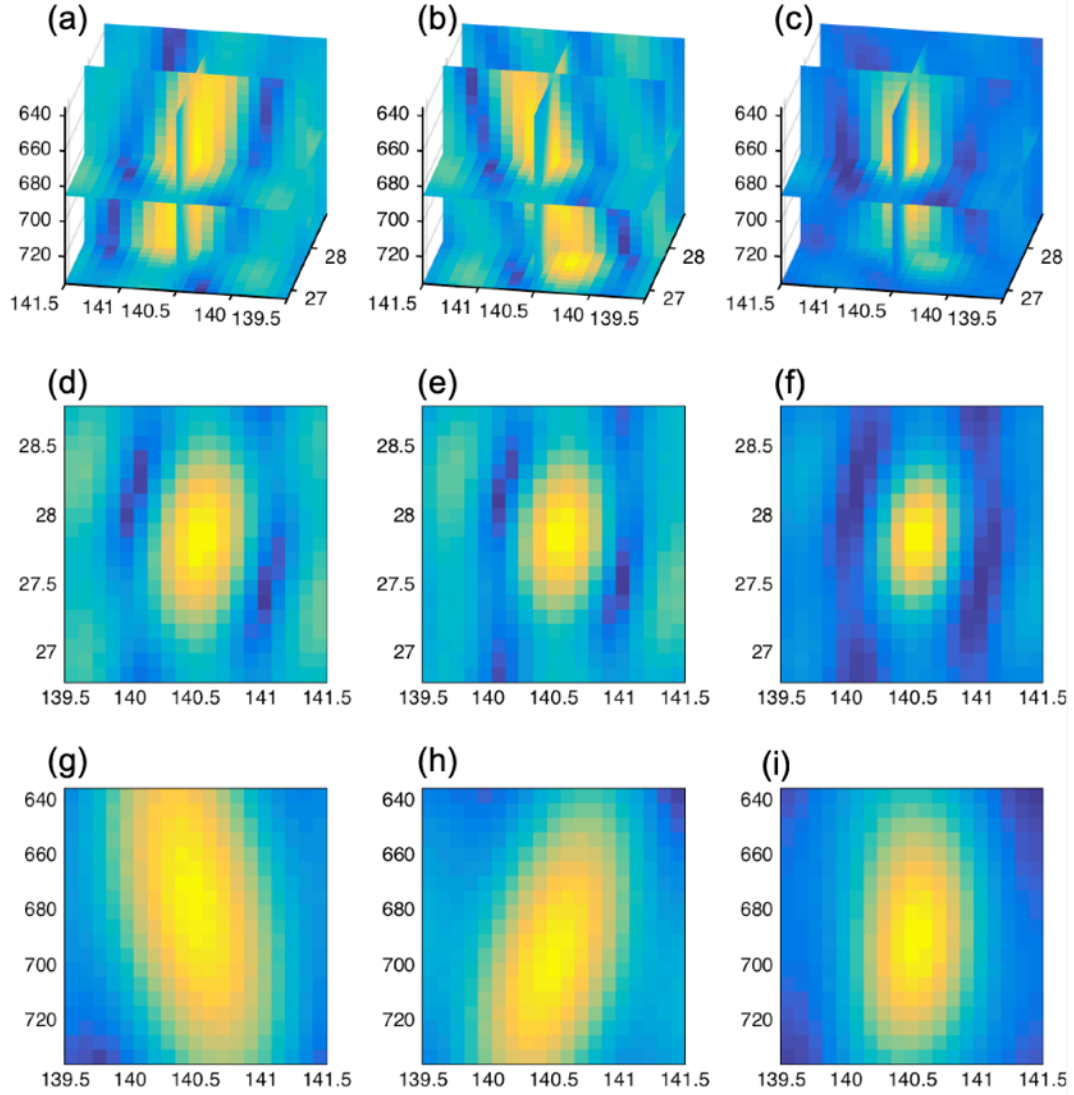


Figure 2.8: Array Response Function (ARF) of USArray for the 2015 Mw 7.9 Izu-Bonin Earthquake. (a), (d), and (g) illustrate the ARF of P wave, which correspond to visualization from different dimensions. (a) is the 3D view, with cross sections right at the hypocenter. (d) is a horizontal mapview right at the hypocenter depth, where X axis is longitude and Y axis is latitude. (g) is a vertical cross-section that goes through the hypocenter, where X axis is longitude and Y axis is depth. (b), (e), and (h) are similar visualizations for the ARF of pP phase. And (c), (f), and (i) are visualized for the combination of P and pP phases.

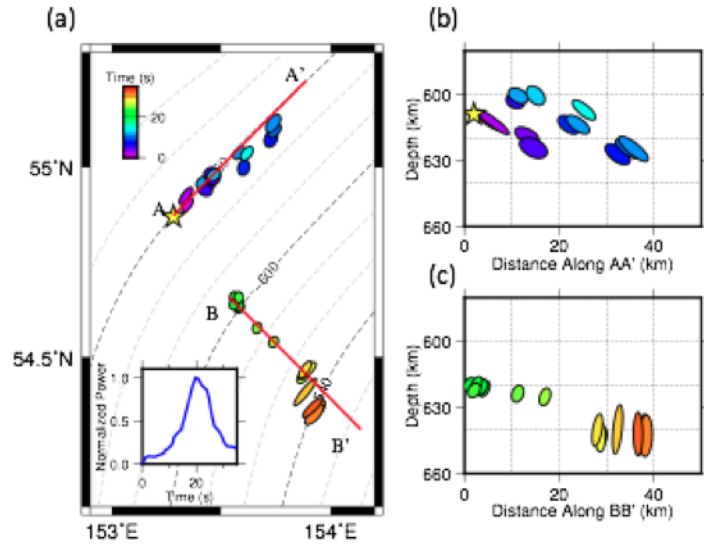


Figure 2.9: 3DBP image using European (EU) network for the 2014 Mw 8.3 Okhotsk earthquake. (a) BP locations in map view. (b) Cross-section of BP images along AA' in (a). (c) Cross-section of BP images along BB' in (a). The colors and shape of the ellipsoids demonstrate the timing and the 95% confidence interval on the peak locations.

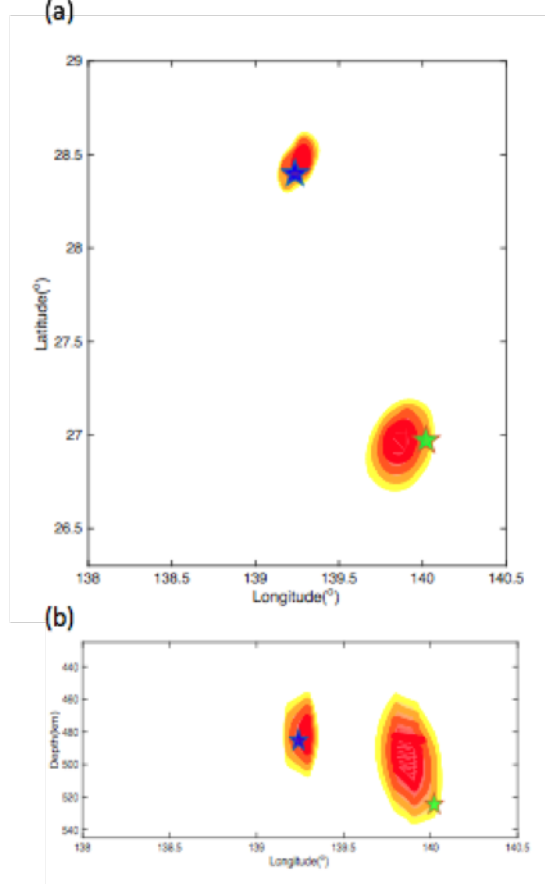


Figure 2.10: Test of spatial biases in 3D back projection. Blue and green stars are M6 deep earthquakes occurred beneath Bonin Islands at 2010-11-30 and 2011-01-12 respectively. We back-projected the two events with a uniform timing correction acquired from cross-correlation of the blue star. This process is similar to imaging aftershocks with continuous data as described by Kiser and Ishii (2013). (a) denotes a horizontal map view of the 95% beam-forming power contour and hypocenter locations. (b) denotes a vertical cross section of the 95% beam-forming power contour and hypocenter locations.

2.6 Supplementary material

Time	Latitude	Longitude	Magnitude
2015-05-12T07:17:00	27.88	85.75	M5.5
2015-04-16T07:09:00	27.78	86	M6.7
2015-12-05T07:36:00	27.62	86.17	M6.3
2015-04-25T06:18:10	27.55	86.08	M5.7

Table S2.1: NEIC catalogue of major aftershocks in Nepal used in the calibration.

Time	Latitude	Longitude	Magnitude
2016-11-14T06:47:52	-42.353	173.8551	5.8
2016-11-13T11:12:39	-42.4063	173.6606	5.7
2016-11-15T06:30:30	-42.35	173.94	5.4
2016-11-13T11:52:45	-42.1762	173.6227	6.1
2016-11-13T13:31:26	-42.2972	173.6997	6.2
2016-11-13T11:32:08	-42.2377	173.6075	6.5
2016-11-13T11:41:49	-41.6877	174.2061	5.7

Table S2.2: NEIC catalogue of major aftershocks in New Zealand used in the calibration.

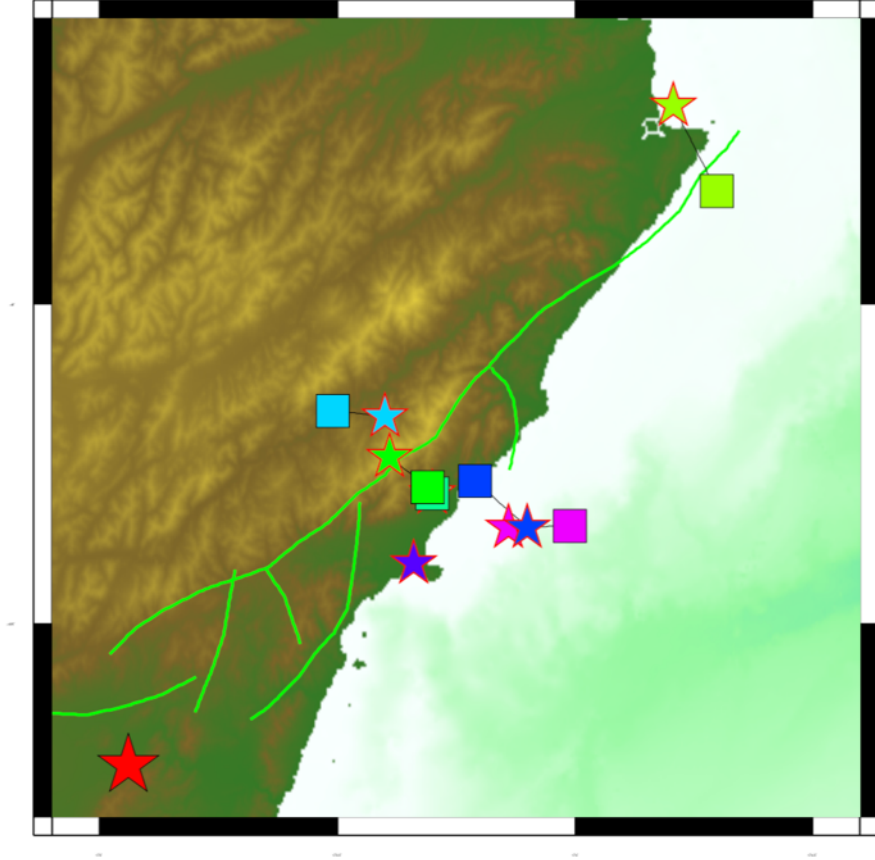


Figure S2.1: Mapview of relative aftershock locations for the Kaikoura source area. The hypocenter of the mainshock from GeoNet and NEIC catalog are shifted to 173.06 E, 42.72 S, and aftershocks are aligned within each catalogue correspondingly. In this manner, we can compare relative catalog differences between the two networks.

CHAPTER 3

Unraveling the Rupture Complexity of the 2016 Mw 7.8 Kaikoura, New Zealand Earthquake with Slowness-enhanced Back-Projections and Dynamic Rupture Simulations

Abstract

Characterizing the complexity of earthquake rupture is of great importance to understanding the physical mechanisms of large earthquakes. The 2016 Mw 7.8 Kaikoura earthquake is one of the most geometrically complex earthquakes, involving cascade failures and wide stepovers. In this work, we study the rupture process by back-projecting the teleseismic P-waves recorded by broadband stations in southern China. We develop a back-projection (BP) technique improved by slowness correction that reduces 80% of the spatial biases in aftershock tests. The calibrated BP reveals that the earthquake initiates to the south of the Hope fault and propagates northeastward via stepping and branching on at least six distinct fault surfaces, with an overall slow apparent rupture speed of 1.4 km/s for the first 70 s. Most of the high-frequency seismic radiation involves three conjugate faults (Conway-Charwell, Hundalee, and Papatea) trending NW-SE, with a rupture speed of 2 km/s on each fault. We conduct a set of dynamic rupture simulations on a generic fault geometry that represents the Malborough Fault system. Our dynamic simulations suggest that rupture can jump an apparent wide step-over between two segmented thrust faults due to their somewhat closer proximity at depth, as well as a strong dynamic unclamping effect. The

jumping rupture would transfer negative Coulomb stress to the Hope fault, inhibiting its slip. Additionally, the earthquake rupture on segmented thrust faults tends to take an indirect zig-zag route instead of a straight route, which could possibly explain the anomalously long duration of the Kaikoura event.

This Chapter is modified from Zhang, A., Wu, B., Meng, L., Kyriakopoulos, C., Oglesby, D., & Fang, L. Unraveling the Rupture Complexity of the 2016 Mw 7.8 Kaikoura, New Zealand Earthquake with Slowness-enhanced Back-Projections and Dynamic Rupture Simulations. (Submitted to JGR).

I conducted kinematic source imaging part. Wu, B., Kyriakopoulos and C., Oglesby, D., performed dynamic rupture simulation. Fang, L helped with data acquisition.

3.1 Introduction

3.1.1 Source observations of the Kaikoura earthquake

An Mw 7.8 earthquake struck the northern South Island of New Zealand on November 14, 2016 NZDT (11:02 on November 13 UTC). According to GeoNet (<https://www.geonet.org.nz>), the epicenter is located south of the Hope fault inside the Marlborough Fault system, 60 km southwest of the city of Kaikoura. The earthquake caused two deaths and over \$900 million of business claims. The 2016 quake is the largest historical event in the Marlborough system, exceeding the 1848 Mw 7.5 Blenheim earthquake.

The Marlborough Fault system consists of a series of dextral strike-slip faults, linking the transition from Hikurangi subduction on the east to oblique transpressional collision along the Alpine fault to the west (Little & Jones, 1998). Immediately after the earthquake, both GCMT and USGS W-phase solutions reported a significant non-double couple component, implying that the earthquake cannot be simply represented by a rupture on a single plane. The geometrical complexity is directly revealed from field observations of at least 12 faults with surface ruptures, with the total rupture length exceeding 150 km (Litchfield et al., 2016; Kaiser et al., 2017). The Kaikoura earthquake is recorded by multiple geophysical datasets of

seismic waveforms, geodetic images, field surveys, and tsunami recordings, which enable the detailed study of the complex rupture process using a variety of techniques and constitute to a special issue in BSSA (Berryman et al., 2018).

Multiple-point source inversion (Duputel & Rivera, 2017) suggests a weak initiation of strike slip faulting followed by a major moment release of dip-slip rupture. In order to incorporate finite source effects, both Hollingsworth et al. (2017) and Zhang et al. (2017) develop two-fault models based on the two major energy concentrations inferred from teleseismic back-projection (BP): one near the hypocenter, and the other near the Kekerengu faults. Hollingsworth et al. (2017) parametrize a shallow dipping thrust fault and a simplified Kekerengu fault (ignoring the Papatea fault) to extract a slip distribution by fitting teleseismic data. Zhang et al. (2017) interpret the slip distribution as a rupture sequence occurring on the Humps and Hundalee faults, followed by slip on a deeper thrust plane near the Kekerengu fault.

More complex models are constrained by the local strong-motion, GPS, InSAR and tsunami data (Cesca et al., 2017; Hamling et al., 2017; T. Wang et al., 2018; W. Xu et al., 2018). A three-fault model is constructed by Cesca et al. (2017), which includes two strike-slip faults with different orientations and a thrust fault in accordance with principle moment tensors of aftershocks in the area. The rupture is reported to occur sequentially on the Humps, the Leaders, and the Kekerengu-Needles fault zone. The role of the slip on the subduction plate interface is controversial. Hamling et al. (2017) provide a static slip model by parameterizing 12 major faults from joint inversions of GPS and InSAR data, yet they ignore the Papatea fault. Their model suggests that the major slip occurred on crustal faults rather than the plate interface. In contrast, T. Wang et al. (2018) combine seismic, geodetic, and tsunami data to invert for the slip distribution on 12 faults and favor a large portion of moment release on the plate interface.

The source models generally agree in the following points (1) The earthquake initiates weakly on the Humps fault, and subsequently branches and jumps among multiple faults. (2) Major moment release occurs in the Papatea-Jordan-Kekerengu fault system. (3) The

rupture process is slow, with overall rupture speeds of 1.4-1.7 km/s (Zhang et al., 2017; Hollingsworth et al., 2017; Holden et al., 2017). Most of the aforementioned studies are developed using finite source inversions. Different parameterizations due to the complex geometric pattern of faults and data selections render significant discrepancies in slip distribution between models produced by various groups, especially in the southern part of the rupture which accounts for less than 30% of the total moment. While the major moment release occurs in the northern part of the rupture (in the Papatea-Jordan-Kekerengu fault system), the southern part involves slip partitioning on more fragmented and seemingly disconnected faults. It is therefore essential to understand what role fault branching and jumping play in navigating the rupture through the complex fault system of this earthquake. For the Papatea-Jordan-Kekerengu fault system, the slip on the Papatea fault remains controversial due to the change in dip angles and geometrical complexity. While W. Xu et al. (2018) and T. Wang et al. (2018) reconcile InSAR or tsunami data to resolve slip of around 10 meters on the Papatea fault by assuming a shallow dipping plane (<45 degrees), both Holden et al. (2017) and Hamling et al. (2017) underline the difficulty to constrain the slip on Papatea fault with a subvertical plane (75-80 degrees) as indicated by significant uplift observed in the field.

3.1.2 The back-projection source imaging

In order to acquire a detailed kinematic rupture process especially for the complex southern part of the rupture, we apply the back-projection (BP) analysis using the teleseismic P-waves recorded by 297 broadband stations in southern China. BP tracks high frequency (up to 2 Hz in this study) radiators from coherent phases of a seismic array, and it does not require any prior assumptions of fault parametrization or regularization (Ishii et al., 2005). BP also incorporates travel time variations from the three-dimensional Earth structure using an empirical hypocenter correction (Ishii et al., 2005).

Such a static hypocenter correction has been widely adopted under the assumption of uniform 3D travel time errors across the entire rupture area (Kiser & Ishii, 2017). While

this approximation is valid in the vicinity of the hypocenter, it gradually breaks down as the rupture front moves away from the hypocenter, which is problematic for BP of large earthquakes. Substantial efforts have been devoted to improving the accuracy of large-scale BP. [Ishii et al. \(2007\)](#) proposed an empirical correction based on a weighted linear interpolation of travel time difference on each grid from aftershocks. [Palo et al. \(2014\)](#) introduced both a static smoothing and a dynamic kriging term to correct the travel-time errors. However, both [Ishii et al. \(2007\)](#) and [Palo et al. \(2014\)](#) employ an empirical interpolation, which requires a dense spatial distribution of aftershocks as well as high data quality. So far, such empirical aftershock calibrations have not been routinely integrated in BP method, because for many events such as the Kaikoura earthquake whose large aftershocks are sparse, improper interpolation would bias the BP results and produce additional artifacts. Therefore, a more generalized time calibration is essential. For the Kaikoura event, [Zhang et al. \(2017\)](#) indicate the spatial bias of their BP with the South America Array could be as large as 32 km from cross checking of an Mw 6.5 aftershock (67 km away from the hypocenter). Considering the fault spacing in this region can be as dense as 10 km, spatial bias introduced from static hypocenter correction merits improvement.

In this study, we first demonstrate the necessity of applying the slowness correction to BP. We compare the difference between non-calibrated BP-inferred aftershock locations and the NEIC catalogue locations. Our results show that there is a systematic westward bias of BP-inferred locations (RMSE = 27 km, RMSE: Root Mean Square Error) due to an insufficient approximation of the three-dimensional velocity structure using a one-dimensional model. We develop a slowness vector calibration from previous work ([Meng et al., 2016](#)). We find that with the calibration, the spatial bias of the aftershocks can be significantly reduced from an RMSE of 27 km to 6 km. We present both direct and indirect time corrections for aftershock calibration and find that the directly calibrated BPs are more accurate, and thus become our preferred results. The indirect calibration is an improved version of that shown in [W. Xu et al. \(2018\)](#) because we include more aftershocks for calibrations and introduce a ridge regression to avoid overfitting. We apply the same calibration to the Mw 7.8 Kaikoura

mainshock and image the first 70s of the rupture. With improved accuracy, BP imaging reveals that the Mw 7.8 Kaikoura event is a complex earthquake, with major high-frequency energy release on three conjugate fault segments: the Conway-Charwell, Hundalee, and Papatea faults, sequentially. Calibrated BP also provides a refined spatial-temporal evolution of fault stepping and branching during the earthquake: rupture speed on each segment is around 2 km/s, while the overall average speed is only 1.4 km/s. We conduct dynamic rupture simulations to investigate the wide rupture step-over and overall slow rupture propagation that are observed in the 2016 Kaikoura event. Our dynamic simulations suggest that rupture can jump an apparent wide step-over between two segmented thrust faults due to their somewhat closer proximity at depth, as well as a strong dynamic unclamping effect. Earthquake rupture favors a "zig-zag" route when it jumps over a stepover, since the Coulomb stress perturbation is stronger from locations on neighboring faults that are not necessarily closely overlapping in a linear fashion. As a result, an event on multiple stepover segments would have an effectively slow migration speed as simply estimated along the average strike of the fault, while on each segment the rupture speed is relatively fast.

3.2 Enhanced Back-projection with slowness calibration

To evaluate the accuracy of static hypocenter correction, we first applied the same "hypocenter correction" of the main shock to the aftershocks, then back-projected the aftershock waveforms in the same fashion as the main shock. This procedure is designed to inspect the bias of the BP since the location of the aftershock is known according to the NEIC catalogue. In this study, we chose seven moderate size ($M_{5.4} \sim 6.5$) aftershocks listed in Table S2.2. Moderate aftershocks are favorable because they have good data quality with sufficient energy in the high frequency band (0.5 - 2 Hz). For larger ($M > 6.5$) events, centroid locations of the finite source cannot be considered as collocated with the epicenter location. Conversely, moderate events can be simplified as point sources at teleseismic distance. BP inferred locations are derived from the average BP location from the first 5s for each aftershock. A comparison between the BP inferred location and NEIC catalogue

location is presented in Figure 3.1 (a). According to Figure 3.1 (a), there is a systematic spatial bias (27 km, on average) between BP-inferred and catalogue locations. For all the events, we observe a northwestward trend from each catalogue location to the corresponding BP location. To clarify if the spatial bias is simply a reflection of uncertainties from the catalogue, we inspect the relative difference between NEIC and GeoNet catalogues. Relative locations of aftershocks with respect to the mainshock hypocenter is compared between catalogues, and the averaged spatial difference is less than 10 km, which is considerably smaller than the bias in BP. The spatial biases range from 20 - 30 km in distance, comparable to the rupture stepping distance. Considering the complex fault setting, such offsets could completely change the rupture history; it is a significantly large bias that needs to be mitigated. The error can also be visualized in time domain as the travel time error recorded at different stations for a Mw 5.7 event in Figure 3.2.

Hence, we introduce 2d slowness correction to calibrate the BP result. The methodology of 2d slowness calibration is discussed in 2.2.2 and 2.2.3. Both direct and indirect 2d slowness calibration are effective in mitigating spatial biases. For the direct calibration, average spatial bias reduces from 27 km to 6 km. For the indirect calibration, average spatial bias reduces from 27 km to 9 km. Since the direct calibration has the advantage of smaller misfit, therefore, we will focus on the BP results from the direct calibration in further discussions.

The calibration is applied to the mainshock BP for a refined kinematic rupture process as described in Figure 3.3. The mainshock waveforms are aligned by initial P arrivals under a low-frequency filter (0.1 - 0.5 Hz) with window length of 10 s, followed by a high-frequency filter (0.5 - 2 Hz) of a 7 s window (Figure S3.1). The BP is performed at 0.5 - 2 Hz with a moving window of 10 s. The earthquake initiates near the Humps fault zone and propagates northeastward for ~ 100 km on a complex fault system with wide-steptovers. Coherent high frequency radiators can be captured on multiple major faults, including the Humps and Conway-Charwell fault zones, the Hundalee fault zone, the Jordan fault, and the Papatea fault. In the first 10s, BP radiators appear to be propagating with a small energy burst, indicating a weak initiation on the Humps fault zone. From 10 to 20 s, the rupture extends

along the Conway-Charwell fault zone thrust at a speed of approximately 1.7 km/s. An abrupt jump occurs at around 20 s, when the radiators step from the Conway-Charwell fault zone to the Hundalee fault with an accelerated rupture speed of 2 km/s (Figure 3.3c). The sudden jump also corresponds with the first energy burst released along the Hundalee fault zone (Figure 3.3d). Propagation along the Hundalee fault lasts for another 10s. At 32 s, the rupture moves away from the Hundalee fault. Unlike the previous segments, the radiators are off the existing fault traces and appear to be more scattered. At around 44 s the rupture steps back to the Jordan fault and branches to the Papatea fault almost simultaneously at a speed of 2 km/s. The branching generates another energy burst on the Papatea fault, which is consistent with the large amount of surface displacement observed in the field in this area (Hamling et al., 2017). The radiators on the Papatea fault segments are propagating along dip direction of the major fault, while appears to be stagnant along the strike direction. The rupture likely goes beyond 70 s, breaking the Kekerengu fault segment (T. Wang et al., 2018; W. Xu et al., 2018). At this stage, BP radiators become scattered and do not conform to a particular fault trace, possibly due to the dominance of coda waves and the change of radiation patterns (Zhang et al., 2017). While the overall rupture speed along the Hump-Hope-Jordan-Kekerengu faults of the Marlborough fault system is relatively slow, around 1.3-1.4 km/s (Figure 3.3c), the speed along each individual segment is significantly faster, reflecting the geometrical complexity of multi-fault rupture.

The calibrated BP of Mw 7.8 Kaikoura earthquake reflects unprecedented geometrical complexity of this event. It is consistent and robust with distribution of selected aftershocks and hypocenter from different agencies. (Figure S3.2 and S3.3). The timing and location of the subevents inferred from BP also agree with near-field strong motion observations (Figure S3.4). Our BP generally agrees with previously published BP analysis (Hollingsworth et al., 2017; D. Wang et al., 2018; Zhang et al., 2017) in rupture speed and length, except that due to our calibration method we are able to provide a more detailed geometrical pattern which highlights fault stepovers and branches.

3.3 Dynamic simulation

3.3.1 Objectives

BP observations highlight two interesting features: (1) The apparent rupture jumps (referred to hereafter as jumps, but one should keep in mind that we cannot fully exclude the possibility of continuous rupture along some un-imaged connecting structure) twice across wide stepovers with a distance of ~ 20 km. (2) The overall rupture speed of ~ 1.4 km/s is significantly slower than the local shear wave speed, which is about 3km/s based on the 3D velocity model (Eberhart-Phillips & Bannister, 2010). The recondite features are uncommon among earthquakes. Both geological and theoretical studies suggest that it is difficult for earthquake ruptures to jump a distance larger than few kilometers on multi-segmented strike-slip faults (Wesnousky, 2006; Harris & Day, 1999) and thrust faults (Magistrale & Day, 1999). Furthermore, earthquake rupture speed as low as half of the shear wave speed is unusual, since dynamic stress wave that pushes the rupture front forward typically travel at elastic wave speeds of multiple km/s. Two mechanical models have been published to explain the wide jump and low overall rupture propagation speed in the Kaikoura event. Ulrich et al. (2019) conduct a set of dynamic rupture simulations to model the Kaikoura earthquake. They propose that there are unmapped faults connecting the gap between segmented thrust faults, and thus the Kaikoura earthquake rupture propagated on a zigzagging fault system. In this case, the overall rupture speed would be slow because the total rupture length is increased. Ando and Kaneko (2018) also set up a dynamic rupture model for the Kaikoura event. In their model, rupture jumps from the Hundalee fault onto the upper Kowhai fault; the re-nucleation process and rupture detour slow down the overall rupture speed. Both of the models highlight the importance of complex fault geometry in the earthquake rupture process. However, the focus of these models is to reproduce the Kaikoura rupture process as close as possible with stress state and fault geometry that are estimated from geology, geodetic and seismic observations. They do not discuss the physical processes involved in wide rupture jump and slow propagation in a more general sense.

Here we present an alternative spontaneous dynamic rupture model that focuses on the two particular phenomena. We use a piecewise-planar fault geometry to represent the complex Kaikoura fault system. The stress state on each segment is homogeneous and is adjusted by a trial-and-error method to reproduce the final slip that is inferred from the model reported by [W. Xu et al. \(2018\)](#). Our model reproduces the wide jump and slow rupture propagation features from the current back-projection results (and those of other researchers) even though it is only tuned to match the observed final slip. We find that earthquake rupture can jump a large distance over segmented thrust faults due to a dynamic unclamping effect. More importantly, we find that the spatial pattern of the stress interaction between two segmented thrust faults intrinsically favor a zigzagging rupture route, which consequently reproduces an overall slow rupture speed. Our simple model helps to illuminate the physical processes that may have taken place in this earthquake, with implications for fault dynamics in general, and therefore is complementary to previous models that tend to reproduce the observation closely.

3.3.2 Method and settings

The dynamic rupture simulation is accomplished using the Finite Element code FaultMod ([Barall, 2009](#)). This code allows for the full 3-D simulation of dynamic rupture occurring along a fault surface, including the coupled off-fault wave propagation. In the model, earthquake rupture grows spontaneously after artificial nucleation at the hypocenter, and the following spatial-temporal earthquake evolution process is calculated based on initial conditions and basic physical interactions. A linear slip weakening friction law ([Ida, 1972](#); [Andrews, 1976](#)) is used on all the fault segments in our model (Figure S3.5), so that the faults will be triggered once the shear stress is higher than its yielding stress and a critical slip patch size for self-sustaining rupture ([Day, 1982](#)) is reached. FaultMod has been extensively tested and validated under the SCEC/USGS benchmarking exercise ([Barall, 2009](#); [Harris et al., 2009, 2018](#)).

We simplify the Kaikoura fault system as several piecewise-planar faults based on the

geodetic model in [W. Xu et al. \(2018\)](#). In total, we have 7 fault segments, corresponding to the Humps, the Conway-Charwell, the Hope, the Hundalee, the Papatea, the Jordan, and the Kekerengu faults (Figure 3.4, map view of the faults). We do not include the Stone Jug fault and the Point Kean fault of the [W. Xu et al. \(2018\)](#) model into our fault geometry in this model either because the slip on those segments is modest, or the rupture area is not well constrained. We have nevertheless conducted some simulations with the Point Kean Fault, and they will be discussed later in the discussion section. We also do not include the subduction interface in our model, as it is not observed or constrained by BP. We point out that by not having these faults, we are not ruling out the possible participation of these faults; Our focus in this study is just the interaction between segmented thrust faults.

3.3.3 Model results

A dynamic simulation with the parameters in Figure 3.4 is performed, and snapshots of fault slip evolution are shown in Figure 3.5. The rupture nucleation is seen as a circular concentration of slip on the Humps fault. This region expands at a speed of 2 km/s until the radius reaches 5 km at roughly 4 s after rupture initiation; thereafter, rupture propagates spontaneously. At 10 s, the rupture front reaches the Conway-Charwell fault and propagates northeasterly. At 24 s, the Conway-Charwell fault has fully ruptured. Between 29 and 42 s, rupture re-nucleates in the middle of the Hundalee fault and propagates across this entire segment. At 39 s, before the rupture of the Hundalee fault terminates, a new rupture area appears at the bottom of the Papatea fault, which subsequently propagates towards the Jordan and Kekerengu faults. At 52 s, the rupture reaches the Papatea-Jordan-Kekerengu junction. The downward motion of the Papatea footwall increases the shear stress on the Kekerengu fault, while the upward motion of the hanging wall decreases the shear stress on the Jordan fault. For that reason, rupture initially propagates to the Kekerengu fault but not the Jordan fault. At 60 s, the Jordan fault finally starts to rupture due to increased shear stress from the Kekerengu. At about 72 s, rupture terminates on the Jordan fault, and the whole system reaches its final state. Maximum slip in our dynamic rupture model is about

9 m on the Humps, Conway-Charwell, and Hundalee faults, about 12m on the Papatea fault, about 16m on the Jordan fault, and about 18m on the Kekerengu fault, which are generally consistent with the observed slip distribution in previous studies (Hamling et al., 2017; W. Xu et al., 2018; T. Wang et al., 2018). The Hope fault does not rupture in our simulation.

A schematic diagram summarizing the overall rupture process is shown in Figure 3.6, which illustrates a zigzag rupture route traveling from south to north. The complex geometry, with its multiple overlapping segments and re-nucleations at branches and stepovers, requires increased rupture time compared to a typically slipping planar fault, reducing the apparent overall rupture speed down to 1.6km/s in our model. However, rupture speeds can be as fast as 2 km/s on each individual fault. This result agrees with the BP observation. Figure 3.6b shows the moment rate function of our simulated event. It displays multiple pulses with a total duration of around 75s, with a peak moment rate at about 60 s, and an amplitude of $5.94 \times 10^{19} Nm/s$. This model generally matches the USGS moment rate function (first 75 s in Figure S3.6) in terms of both timing and relative peak amplitude, except that the moment rate function in our model is shorter by 10 s at the beginning compared to the kinematic source (noted that we do not include Needle fault in our model). This difference is possibly due to the artificial nucleation process common in most dynamic rupture models, which will be discussed in Section 3.4.

3.3.4 Wide jumps and slow zig-zag rupture

Our model results can be used to explain how the wide step-over ruptures happen. High frequency radiators from BP are broadly utilized to approximate rupture fronts of earthquakes (Ishii et al., 2005; Krüger & Ohrnberger, 2005). Guided by seismic observations, dynamic rupture simulations offer physical insight into how the rupture is able to propagate across the two large stepovers in this complex fault system. Stress histories for the nucleation points on the Hundalee fault (point A) and the Papatea fault (point B) in our simulations are shown in Figure 3.6. The Hundalee fault is triggered by an increase in Coulomb stress

of 2 MPa. Shear stress on point A is almost the same as its initial value when it ruptures; the Coulomb stress increase is primarily accomplished by a 4 MPa drop of normal stress due to slip on the Conway-Charwell fault. A key factor that allows the Hundalee to be triggered is that its shear stress takes a turn from decreasing to increasing at about 21 s. The stress transfer speed (i.e., S wave speed) in the medium is about 3 km/s, and the distance between the Hundalee and Conway-Charwell is about 15 km. It suggests that starting from about 16 s, the slip on the Conway-Charwell starts to generate positive shear stress that will be received by the Hundalee fault about 5 s later. Snapshots in Figure 3.5 shows that at 16 s the Conway-Charwell fault is about half-way ruptured, which indicates that the part of the Conway-Charwell near the Hope transfers positive shear stress to the Hundalee, while the part of the Conway-Charwell far away from the Hope transfers negative shear stress to the Hundalee. This result explains why, in our simulated model, there is a large time gap between the Conway-Charwell rupture and the Hundalee rupture: Rupture on the Hundalee does not have enough shear stress to start until the whole the Conway-Charwell ruptures. According to Figure 3.6d, the Papatea fault is triggered by an increase Coulomb stress of about 1.5 MPa. Similar to point A, the shear stress at point B at its rupture time is almost the same as its initial value, while its normal stress is decreased by about 2 MPa, suggesting that unclamping is a main factor in triggering the Papatea fault.

The unclamping effects observed here can be qualitatively explained with the rotation of the middle block in between two stepover thrust faults. As shown in Figure S3.7(a), the left thrust fault pulls the middle block down, which reduces the shear stress on the right thrust fault; however, the left thrust fault also rotates the middle block anti-clockwise, which unclamps the right thrust fault. The above analysis suggests a possible reason for the wide apparent jumping among stepping faults. The Conway-Charwell, Hundalee, and Papatea fault are in a favorable orientation for rupture jumping. Though separated perpendicular to strike, their dip angles are different, and adjacent faults have large areas of overlap. Since they all slip primarily in the thrust direction, the normal stress interaction among them is relatively high. With a favorable geometrical setting and high initial shear stress relative to

normal stress, the stress transfer between stepping faults is large enough to allow apparent rupture jumps as large as ~ 20 km.

Our dynamic rupture model can also explain why the overall rupture speed is slow while the rupture speed on each individual fault is fast. As our model shows, rupture propagates in a zigzag manner in such a complex fault system. This is because the maximum triggering stress comes from the diagonal direction in between two stepover thrust faults. For example, as shown in Figure S3.7 (b), the southern part of the Conway-charwell fault (away from the Hope fault) sends negative shear stress to the Hundalee fault, while the northern part of the Conway-charwell fault (close to the Hope fault) send positive shear stress to the Hundalee fault. Since rupture on the Conway-charwell fault propagates from the southern part to the northern part, the overall rupture route is zigzag-like. Therefore, despite the short distance between the start and end points, the total length of the rupture path could be large. This effect is amplified by the slow buildup of slip required for the static triggering of one of the faults. If this is the case, the apparent rupture speed obtained from a finite fault inversion using a planar fault and fixed rupture speed would be slow, while the actual rupture speed on each individual fault would in fact be large. The rupture speed of 2 km/s also agrees with systematic study of shallow earthquakes ([D. Wang, Mori, & Koketsu, 2016](#)).

3.4 Discussion

3.4.1 Triggering mechanisms of wide stepovers

While the dynamic simulation explains the apparent wide jumps, it is important to point out that there are other potential mechanisms that could explain observations, due to the non-uniqueness of numerical modeling. Delayed triggering by either static stress or dynamic stress is another possible mechanism responsible for the slow jumping over a large distance. [Hamling et al. \(2017\)](#) reports a positive 0.75 MPa static coulomb stress transfer on the northern (Kekerengu-Papatea) faults due to the static slip from southern part (Humps - Conway Charwell - Hundalee) which is potentially enough to cause delayed triggering. Possible slip

on the underlying subduction interface could also be a solution: slip on southern faults could trigger slip on the interface, which could then trigger the northern faults. In this case, rupture traveling downward and upward on the structures could make the time gap between southern fault rupture and northern fault rupture larger. However, there are discrepancies between current studies on how much of the rupture comes from the dipping plate boundary interface. In our BP, we observe off-fault radiators that cannot be associated with known fault traces. This effect could be an artifact of the wide step over between two faults, as BP utilizes a continuous time window that tends to generate a seemingly connected rupture process. However, our synthetic test illustrated in Figure S3.8 shows that if the artifact exists, the radiators between stage 2 and 3 are much fewer than the observation. Thus, the scattering between stage 2 and 3 cannot be totally explained by artifacts, but associates with ruptures. One of the possibilities is slip on the subducting plate interface, as the low angle thrust could introduce the scattering in BP. Another piece of supporting evidence for potential plate interface slip is that the radiators collocate with plate interface slip patches from source inversions (W. Xu et al., 2018; Bai et al., 2017) as well as the off-fault aftershock activities. Therefore, our BP radiators between stage 2 and 3 are also likely to represent rupture on unmapped faults, but unfortunately, BP provides limited resolution on depth variation. Without additional constraints on depth, the interpretation of unknown crustal fault is just one of the many possibilities. More direct constraints on rupture depth will be necessary, yet the source locations observed in this study could potentially provide spatial constraints on the deformation on the plate interface.

3.4.2 The effects of the zero-slip assumption on the Hope fault

Our BP suggests that the Conway-Charwell rupture triggered the far-away Hundalee fault instead of the nearby Hope fault. This effect is achieved in our dynamic model via a low initial shear stress setting on the Hope fault. However, one may wonder how this event would have been different if the Hope fault had indeed ruptured. To help answer this speculative question, we run additional numerical simulations, which reveal that a lack of slip on the

Hope fault could be fundamental in controlling the behavior of the entire event. We test three cases of the Hope fault with higher initial shear stresses than in our preferred model, which thus (artificially) allow slip on the Hope fault: 2 MPa, 3 MPa, and 8 MPa, respectively. Figure 3.7 compares the final slip and moment rate function of these three simulations. As shown in Figure 3.7a, the 2MPa case behaves similarly to our preferred model where initial shear stress on Hope is 0 MPa, except that there is a minor amount of slip on the Hope fault. However, when we increase the initial shear stress on the Hope fault to 3 MPa, the jump between the Conway-Charwell fault and the Hundalee fault disappears, as shown in Figure 3.7b. That is because the higher stress on the Hope fault causes the Hope fault to slip, and subsequently transfer negative Coulomb stress to the Hundalee fault, which in turn prevents the jump between the Conway-Charwell and Hundalee. When we increase the Hope fault shear stress to 8 MPa, the entire Hope fault is weak enough to allow spontaneous rupture, and the rupture of southern faults propagates to northern faults through the Hope fault. In this case, rupture releases most of its energy at around 40 s, and the apparent overall rupture speed is as fast as ~ 2 km/s, contradicting our observations. These three synthetic scenarios imply that significant slip on the Hope would either prevent the jump from the Conway-Charwell fault to the Hundalee fault, or make the Hope fault a connection between northern and southern faults, which in turn shortens the total duration. In summary, with our current model parameters, zero slip on the Hope fault not only is necessary to match the observed lack of slip there, but is also the key in ensuring that the Kaikoura event has a relatively long duration, with rupture jumping among multiple stepping segments, in agreement with observations.

3.4.3 Moment rate functions of dynamic rupture models

As mentioned in the Section 3.3.3, the moment rate function in our model is shorter by 10s at the beginning compared to the USGS kinematic moment rate function (Figure S3.6). It is probably introduced by artificial nucleation processes. At the beginning of the simulation we artificially raise the initial shear stress above the yielding level within an expanding circular

area (nucleation radius) to nucleate the event. This type of forced nucleation process makes the initiation of rupture in our simulation faster than models constrained by real data. This is not the most accurate way to start a numerical event compared with a long-time-frame model of earthquake nucleation using rate-and-state friction. However, it is a numerically efficient method and it is used by many dynamic rupture modelers (Day, 1982; Harris & Day, 1993; Oglesby et al., 1998). Since the focus of our model is the evolution of rupture and dynamic interaction across the different faults, we do not claim to accurately model the first 10 s or so of (slow) rupture, and take advantage of this freedom to shift our source time function by approximately 10 s to match the subsequent rupture evolution. Future models may well include this slow rupture evolution at the start of the event.

3.4.4 BP-inferred rupture kinematics and the strong-motion recordings

In this study, we are only capable of resolving the first 70s of the rupture due to the coda wave contamination and less waveform coherence in the later rupture. The BP results up to 80 s are shown in Figure S3.9. We are also not able to image the rupture on the Kekerengu and Needles fault because the change of the radiation pattern from the northern rupture produces much weaker waveforms than the southern subevent (Figure S3.10). In addition, the BP-inferred subevents south of the Kekerengu fault are consistent with the prominent energy release in the local strong motion recordings (Figure S3.4). Therefore, we choose to interpret the BP results in the first 70 s and south of the Kekerengu fault. The timing of the rupture on the Kekerengu fault can be somewhat discussed based on the local strong motion record. The strong motion recording produces weak power release beyond 85s, indicating that the Kekerengu fault breaks before 85s. According to Figure S3.11, the two strong-motion pulses can be associated with rupture on Kekerengu faults that occurred between either 45 to 55 s, or 65 to 75s. If the Kekerengu fault corresponds to the first pulse, a fast rupture speed of 4 km/s is required between Jordan (grey star in Figure. S3.4b) and Kekerengu faults, as is reported by Holden et al. (2017). If the Kekerengu rupture corresponds to the second pulse, it can be explained by a relatively slow overall rupture speed of 1.6 km/s, which is grossly

consistent with slow rupture velocity inferred from BP.

Another interesting observation of the HF radiators is that in the first 20 s, HF radiators tend to have larger beam power, while the finite source inversion results capture relatively low moment release in the same area. The discrepancy has been discussed by [Zhang et al. \(2017\)](#), and we note that the HF energy has better correlation with the peak ground acceleration map. We infer that in the first 20s, abrupt rupture velocity or slip rate variations are probable, leading to enhanced HF radiation ([Madariaga, 1977](#)). The frequency-dependent behavior of rupture properties is also a motivation to incorporate HF radiators from BP in hazard assessment and ground motion prediction.

3.5 Conclusion

In this study, slowness calibration is shown to improve the spatial accuracy from conventional BP. The calibrated result sheds light on the first 70 s of the Mw 7.8 Kaikoura earthquake, providing a reliable estimate of the spatio-temporal evolution of the rupture process. The overall rupture is trending SW-NE, but the observed HF radiators are located near three conjugate faults (Conway-Charwell, Hundalee, and Papatea) trending SE-NW. The slow averaged rupture speed (~ 1.4 km/s) is also a confirmation of the complex rupture process. However, on each conjugate fault, the rupture speed is faster (~ 2 km/s). The observed extensive jumping and branching is reproduced in a 3D dynamic simulation on the complex fault system. Our dynamic simulations suggest that rupture can jump an apparent wide step-over between two segmented thrust faults due to a shorter distance at depth, as well as a strong dynamic unclamping effect; these jumps would transfer negative Coulomb stress on the Hope fault. Additionally, the earthquake rupture on segmented thrust faults tends to take a zigzagged rather than straight route, which could possibly explain why the Kaikoura event has an anomalously long duration. This case study highlights the role of the fault jumping and stepping in navigating dynamic rupture through fault systems of complex geometry.

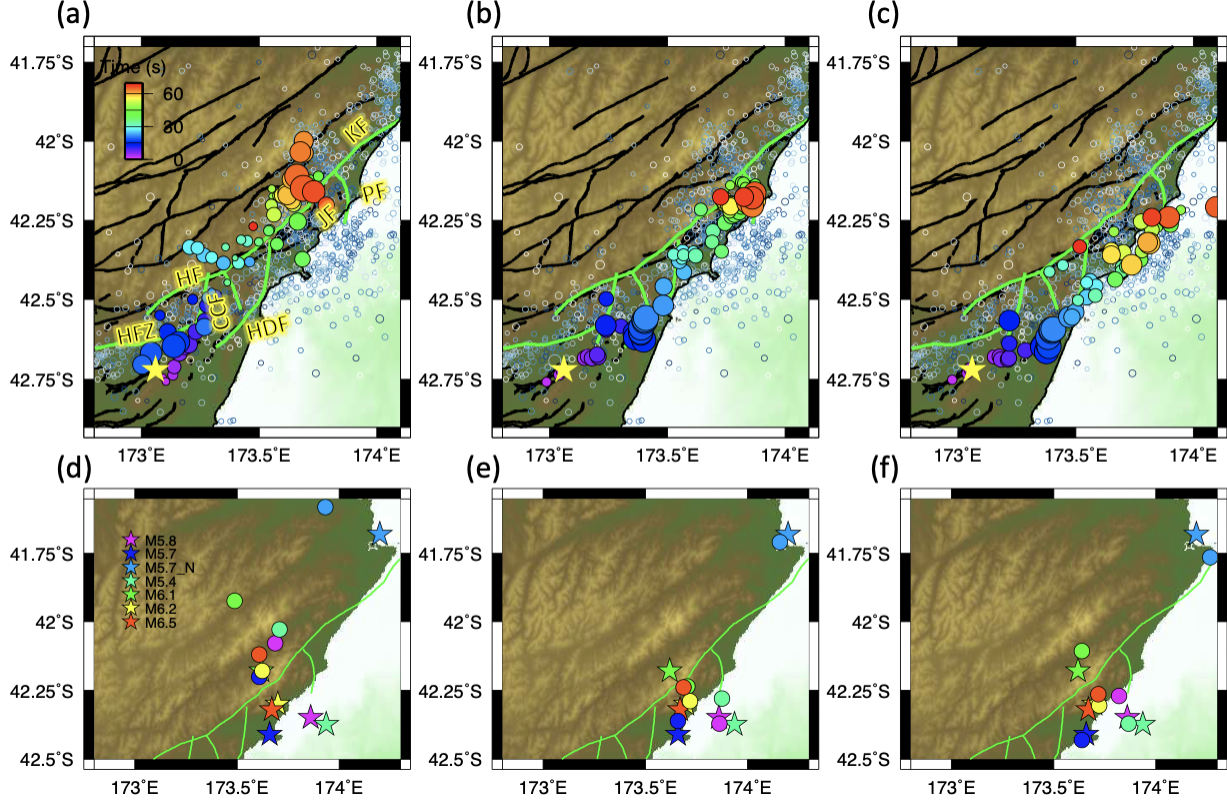


Figure 3.1: Mainshock and aftershock BPs. Back projection results for the mainshock (upper panel) (a) without calibration (b) with slowness calibration (c) with indirect slowness calibration, respectively and 6 aftershocks (lower panel) for (d) without calibration (e) with slowness calibration (f) with indirect slowness calibration. For the mainshock BP, dots are HF radiators sized by power and color-coded by time. Yellow star is the mainshock hypocenter from NEIC catalogue, and empty circles are background seismicity in that region. In aftershock BP images, stars are catalogue locations of different aftershocks, with magnitudes listed in the legend of panel (a) and detailed event information in Table S2.2. HFZ: Humps fault zone, HF: Hope fault, CCF: Conway-Charwell fault, HDF: Hundalee fault, JF: Jordan fault, PF: Papatea fault, KF: Kekerengu fault.

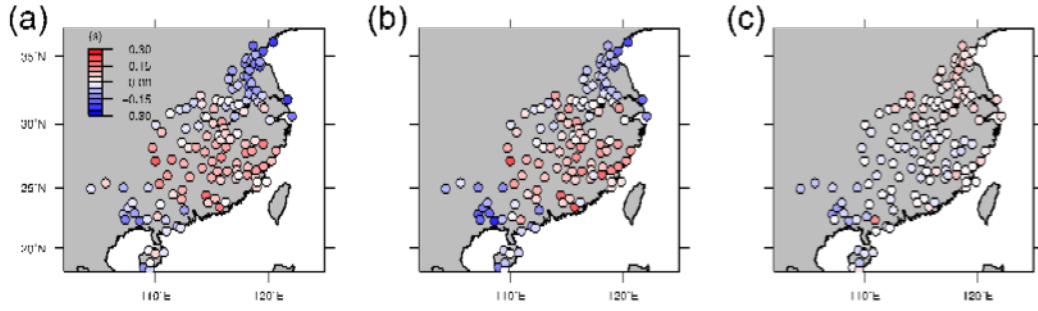


Figure 3.2: Map view of relative differential time errors $\Delta T_j(\xi)$ for the M 5.7 event. (a) Time errors observed from the M 5.7 event at each station. Dots represent stations adopted for calibration and are color-coded by time error. (b) Time errors predicted by the slowness calibration. (c) Residual between the observed and predicted time errors.

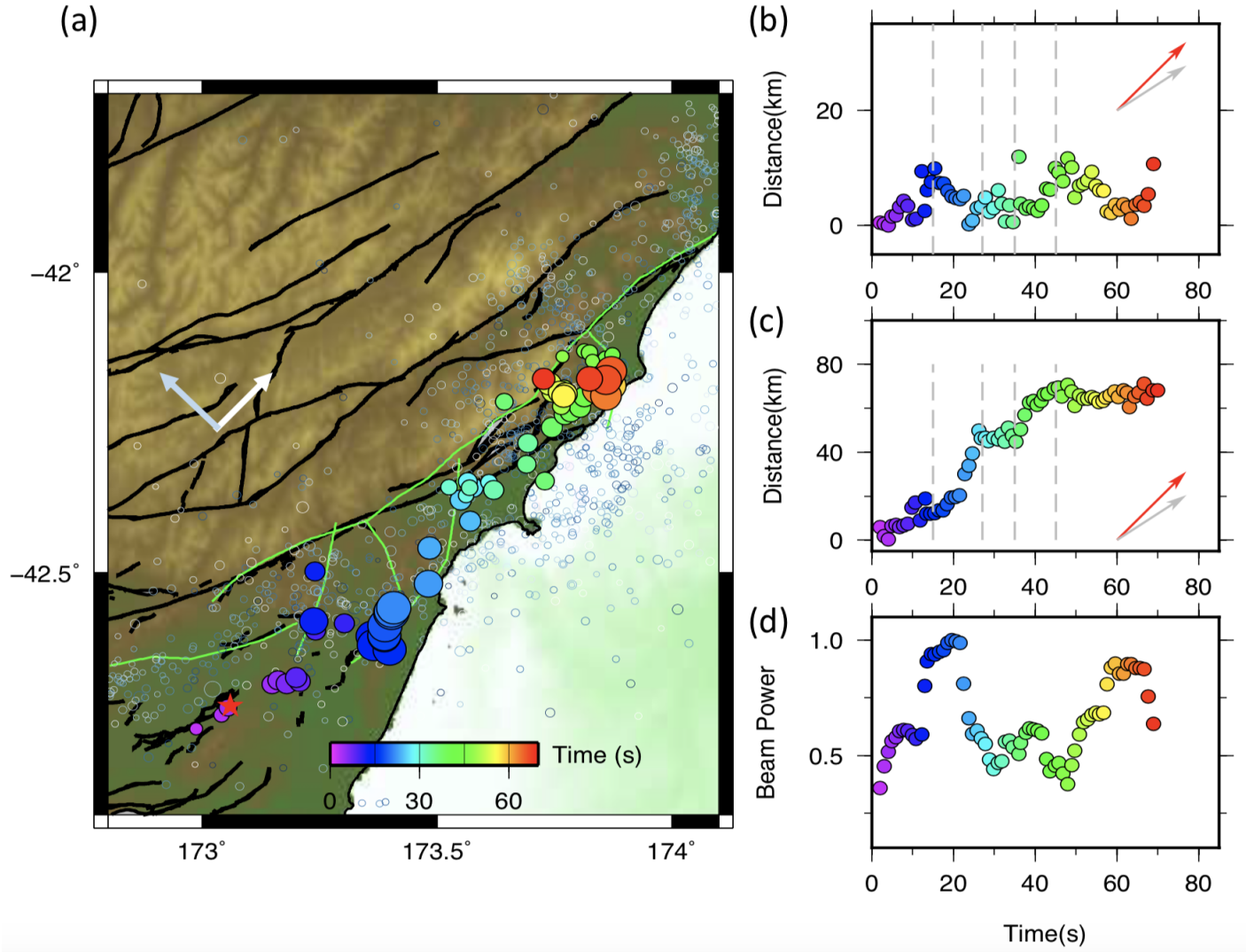


Figure 3.3: Calibrated BP of the mainshock. (a) Mainshock HF radiators (colored dots) imaged by the slowness calibrated BP, with the circle sized by power and colored by time. The red star denotes the National Earthquake Information Center (NEIC) epicenter. (b) Rupture distance as a function of time perpendicular to the direction (light blue arrow in (a)) of the Hope-Jordan-Kekerengu fault system. The grey and red dashed arrows denote rupture speeds of 1.4 and 2 km/s, respectively. Relative fast rupture segments are marked by grey dashed lines. (c) Rupture distance as a function of time along the direction (white arrow in (a)) of the Hope-Jordan-Kekerengu fault system. All the notations are same as (b). (d) Relative beam power as a function of time. HFZ: Humps fault zone, HF: Hope fault, CCF: Conway-Charwell fault, HDF: Hundalee fault, JF: Jordan fault, PF: Papatea fault, KF: Kekerengu fault.

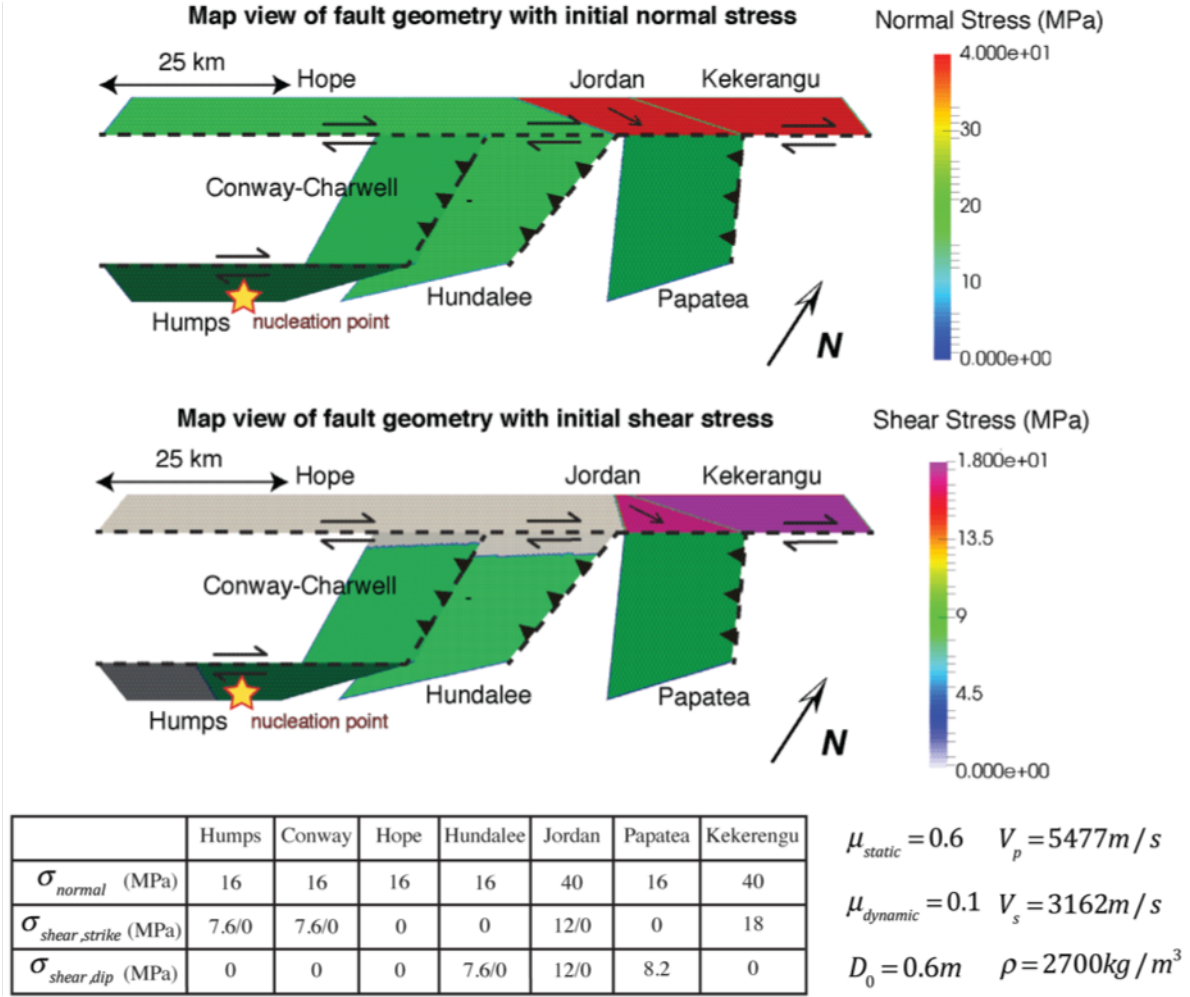


Figure 3.4: Parameter setting of the simulation. (a) and (b): Map view of fault geometry in our rupture dynamic model with initial stress condition. Dashed line shows the surface trace of each fault segment. Initial stress magnitude is shown by different colors and initial stress direction is shown by arrows. The yellow star on Humps fault is the nucleation point in our model, which is set to be hypocenter of the Kaikoura earthquake. Note that in order to show 3-dimensional features better, we add shadow to the geometry based on a light source from north, resulting in the Humps fault being slightly darker than the Conway-Charwell fault under similar stress conditions. (c). Detailed initial conditions. Table shows normal stress, shear stress along strike, and shear stress along dip. Faults with two stress values indicate that there are two different stress conditions on this fault, which can also be seen in Figure 4(a) and (b). Other parameters in our model: static friction coefficient, dynamic friction coefficient, critical slip weakening distance, P wave velocity, S wave velocity, and material density.

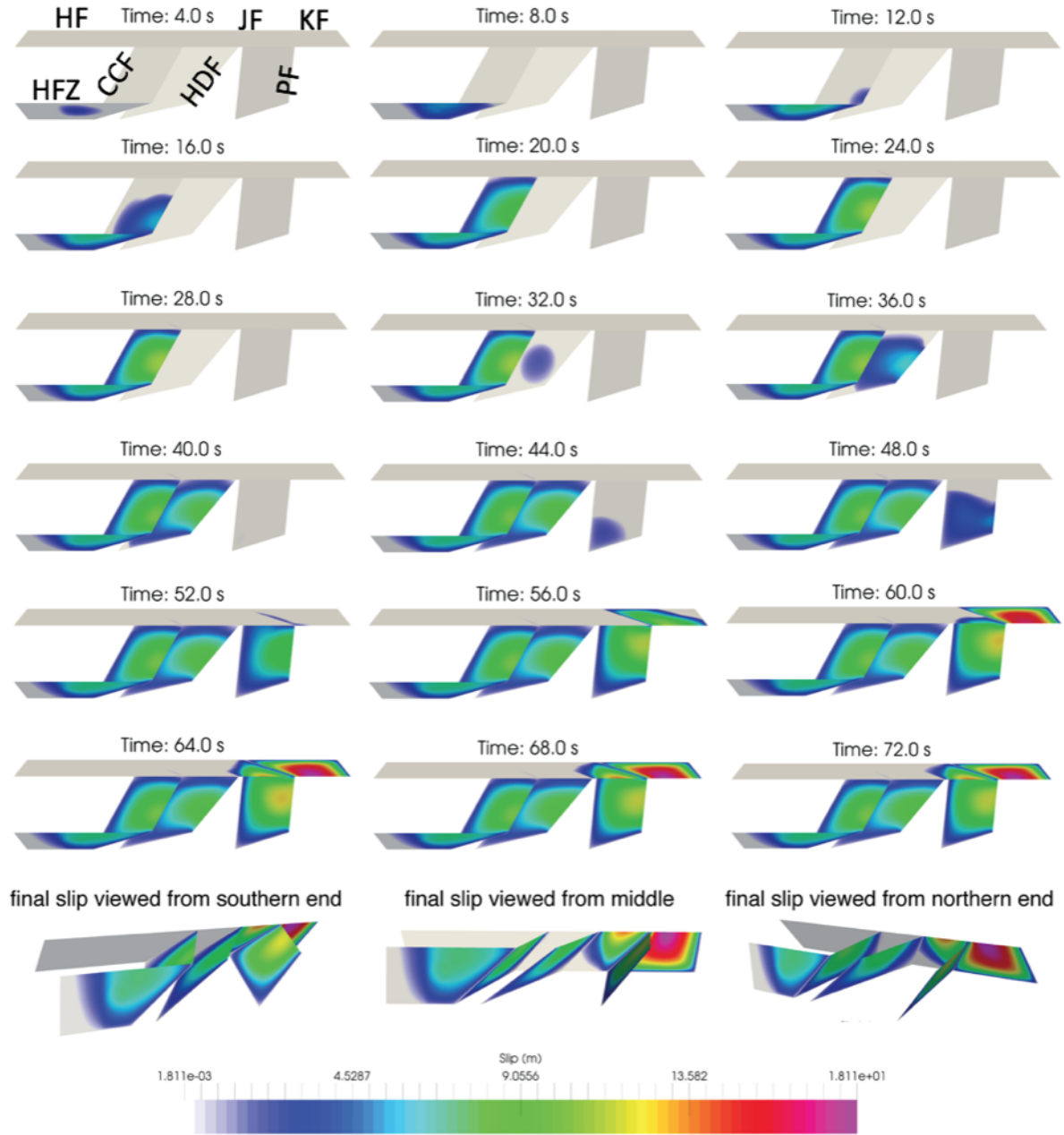


Figure 3.5: Snapshots showing rupture process with slip evolution. Physical parameters of this simulation are shown in Figure 3.4. The last 3 panels show final slip viewed from different angles for clarity.

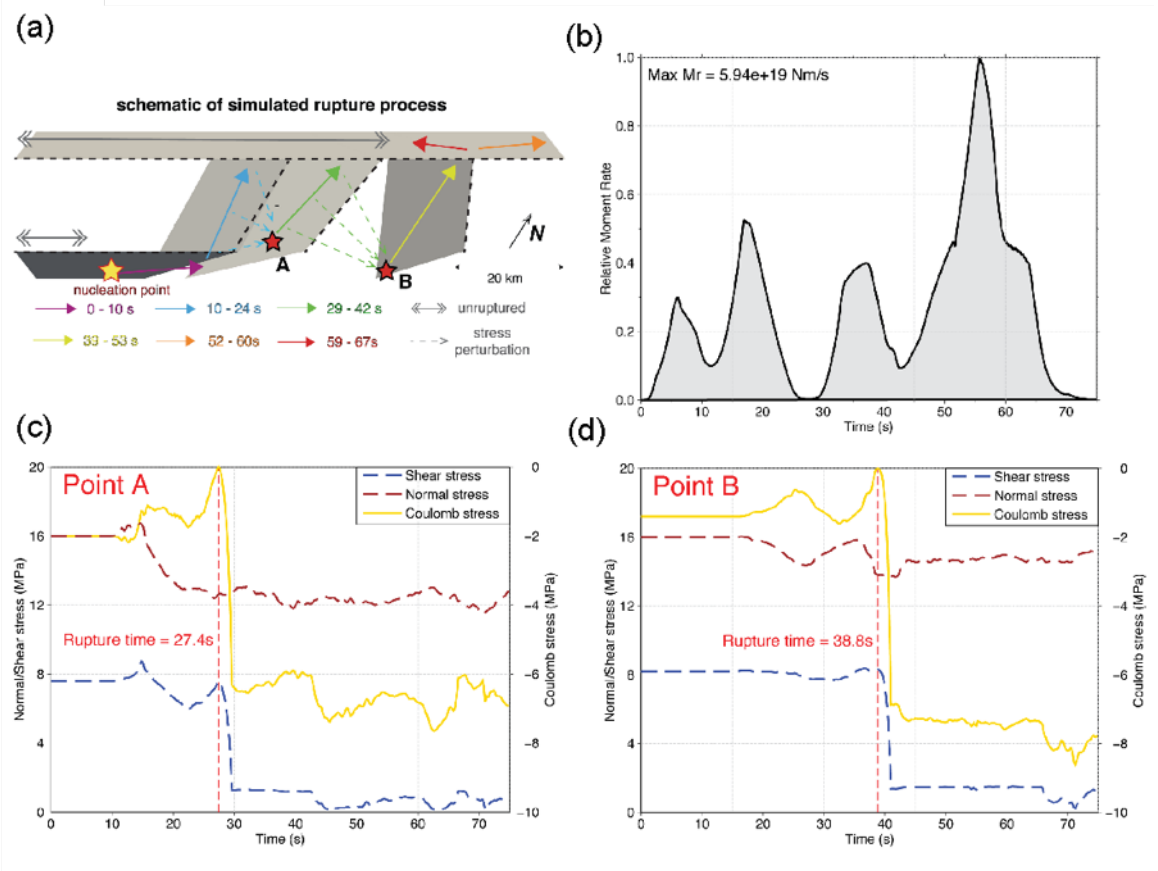
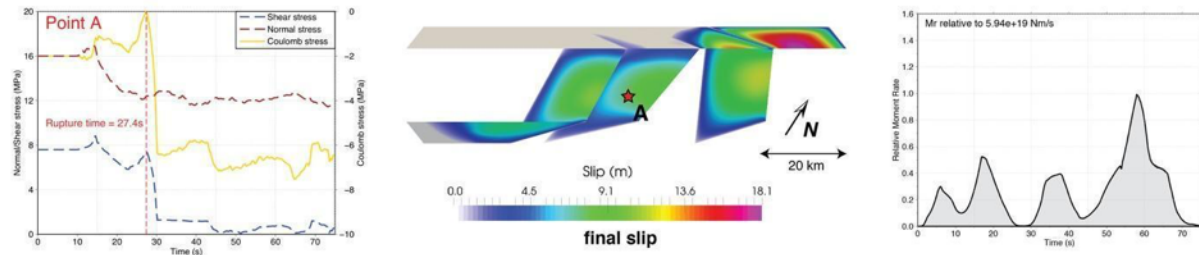
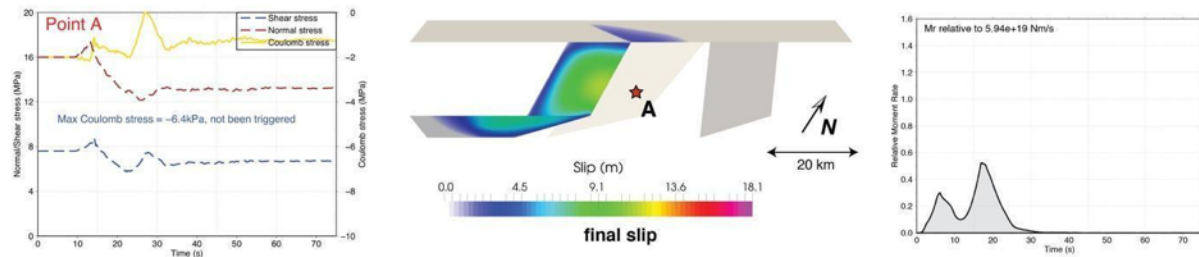


Figure 3.6: Slip history revealed from dynamic simulation. (a). Schematic representation of simulated rupture process. Yellow star shows the hypocenter. Red stars show the renucleation locations on the Hundalee fault (point A) and the Papatea fault (point B). (b). Moment rate function of simulated event. (c) and (d) stress histories of points A and B, respectively. Blue dashed line represents shear stress, brown dashed line represents normal stress, gold solid line represents Coulomb stress, and vertical red dashed line represents the rupture time of this point. Coulomb stress is calculated by $\tau - \mu\sigma$, in which τ is shear stress, μ is static friction coefficient, and σ is normal stress. Note that the vertical scale for the Coulomb stress is different from the vertical scale for the shear and normal stresses.

(a) Hope fault initial shear stress = 2 MPa



(b) Hope fault initial shear stress = 3 MPa



(c) Hope fault initial shear stress = 8 MPa

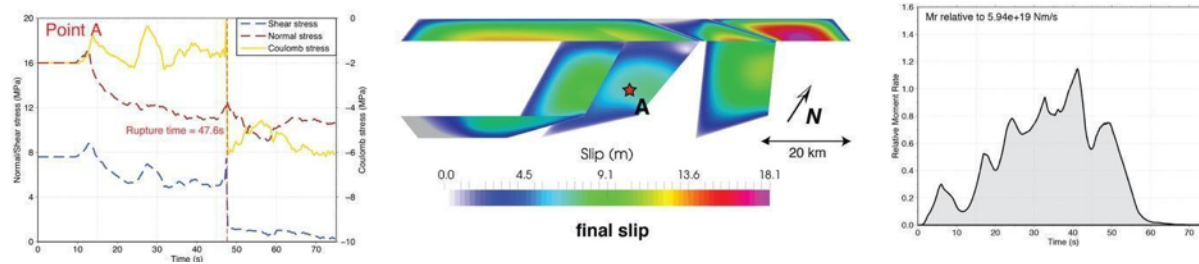


Figure 3.7: Stress history of point A (left), final slip (middle), and moment rate function (right) for 3 different simulations. Initial shear stress on Hope fault is (a). 2MPa (b). 3MPa (c). 8MPa.

3.6 Supplementary material

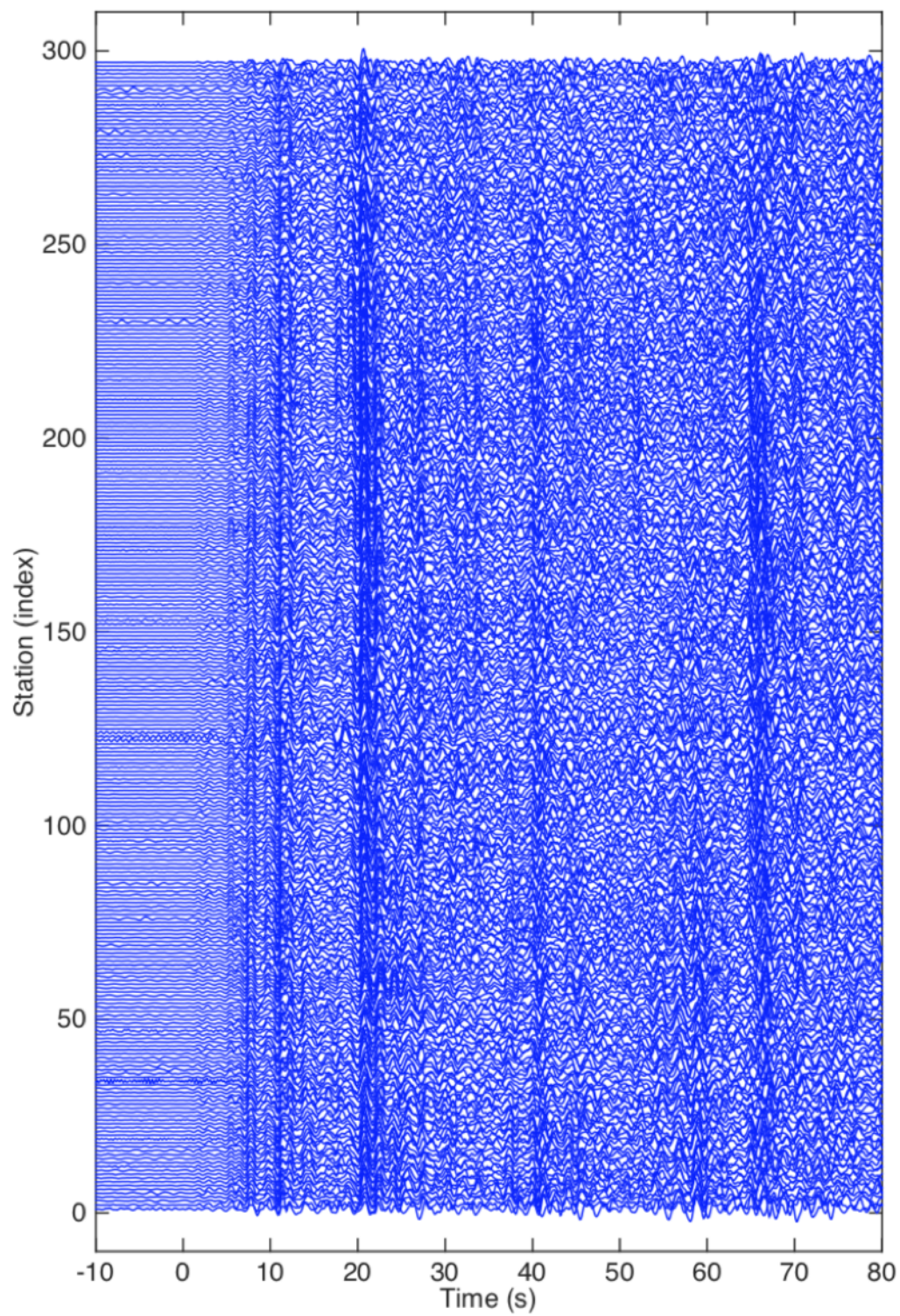


Figure S3.1: Aligned filtered (0.5 - 2 Hz) waveform of the mainshock. Time 0 is referenced to begin time of back-projection.

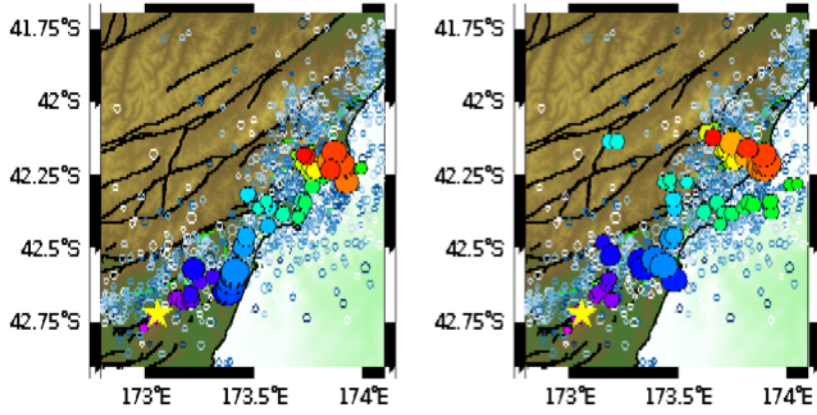


Figure S3.2: Comparison of calibrated BP with different aftershocks. Left panel is the results using all 6 events in table S1 and right panel is the results of only using the last 4 events in table S1.

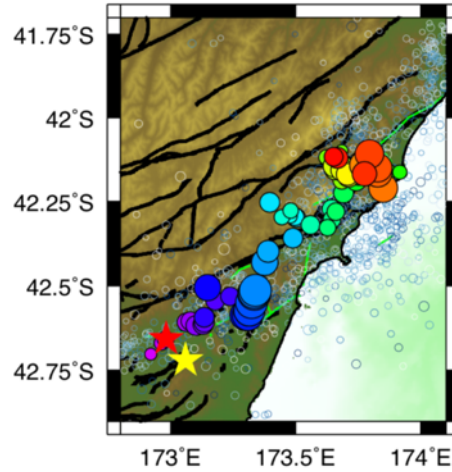


Figure S3.3: BP imaging based on relocated hypocenter (red), and the yellow star is the old hypocenter location acquired from NEIC catalogue.

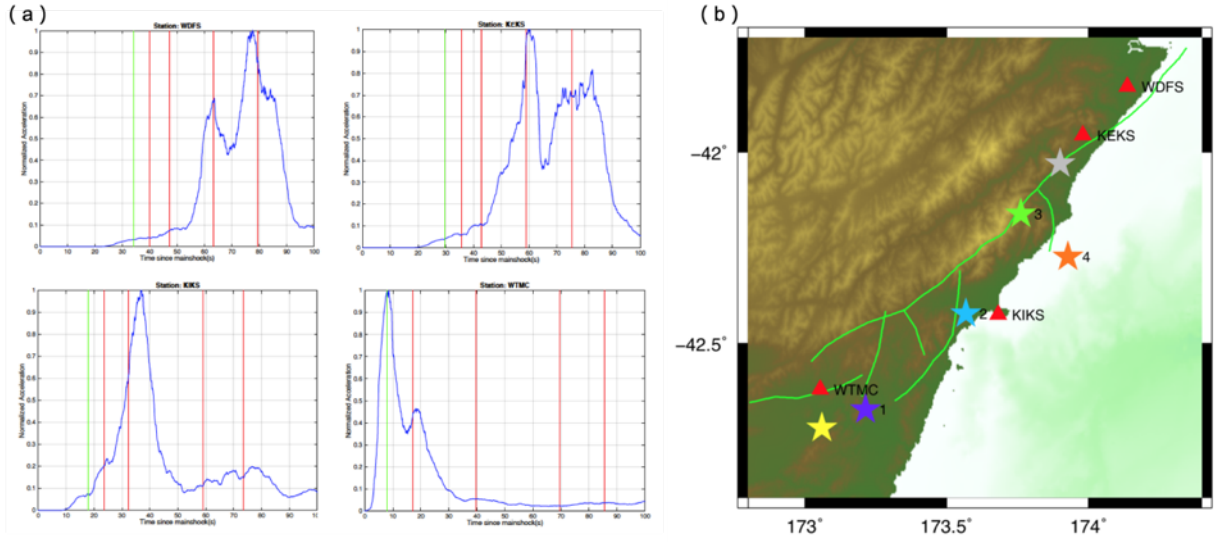


Figure S3.4: Near field ground motion observations. (a) Smoothed Envelope of accelerations at near field stations. The green vertical line is the predicted arrival time of the wave radiated from hypocenter ($-42.72, 173.06$). And the red solid vertical lines are the predicting timing of 4 subevents (location map in b) occurred at the Humps-Conway Charwell, the Hundalee, the Papatea faults and scattered radiators near the Jordan fault respectively. (b) Location of subevents and strong motion stations. The stations are labeled by the name, and the subevents are marked by occurrence order. The yellow star is the hypocenter. The grey star is the hypothesized location of the subevent occurred on the Kekerengu fault.

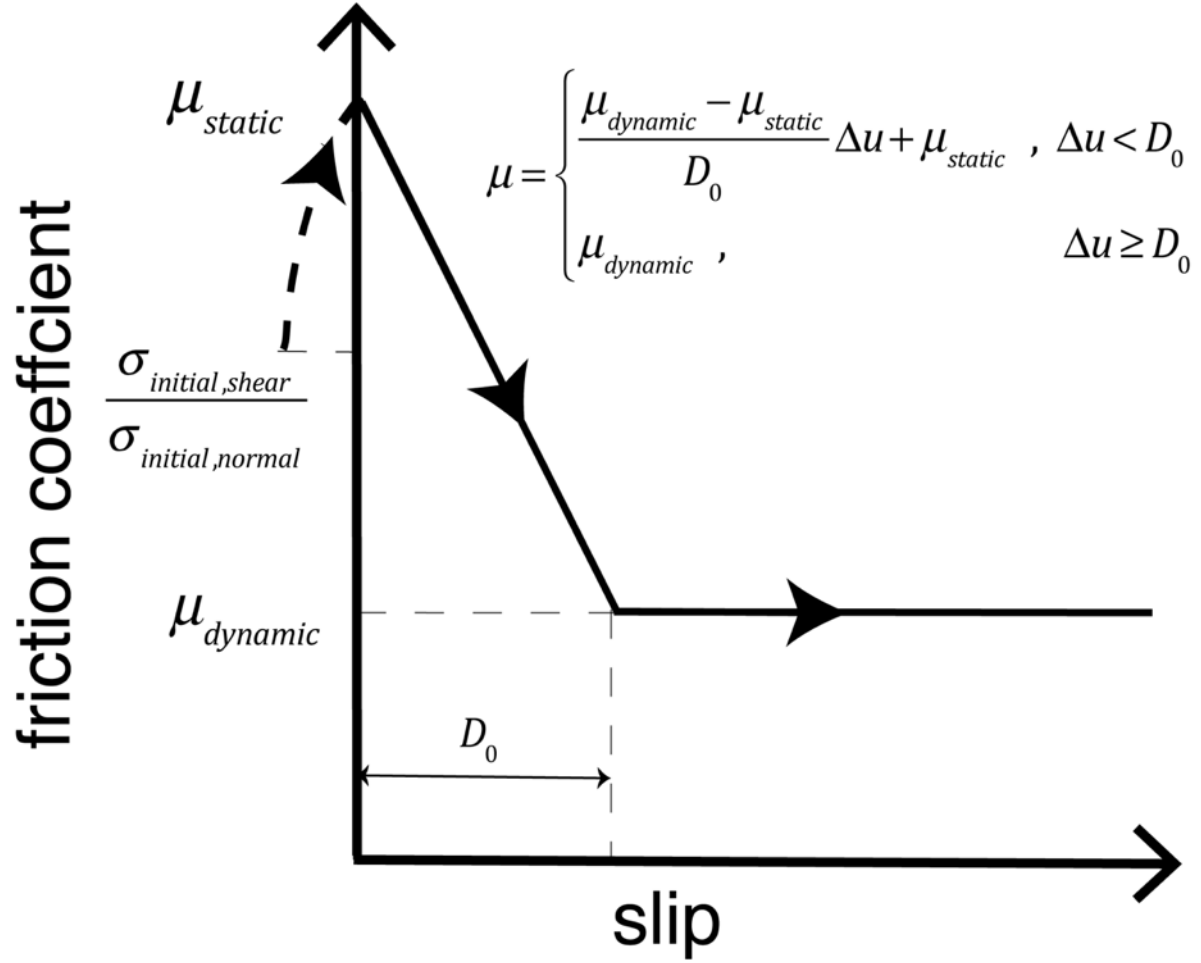


Figure S3.5: Schematic diagram of slip weakening friction law. Fault starts to slip when friction coefficient (shear stress to normal stress ratio) exceed static coefficient μ_{static} ; friction coefficient drops linearly as slip increase until fault slips a critical slip weakening distance, D_0 . After that, friction coefficient stays at a stable value, dynamic coefficient $\mu_{dynamic}$.

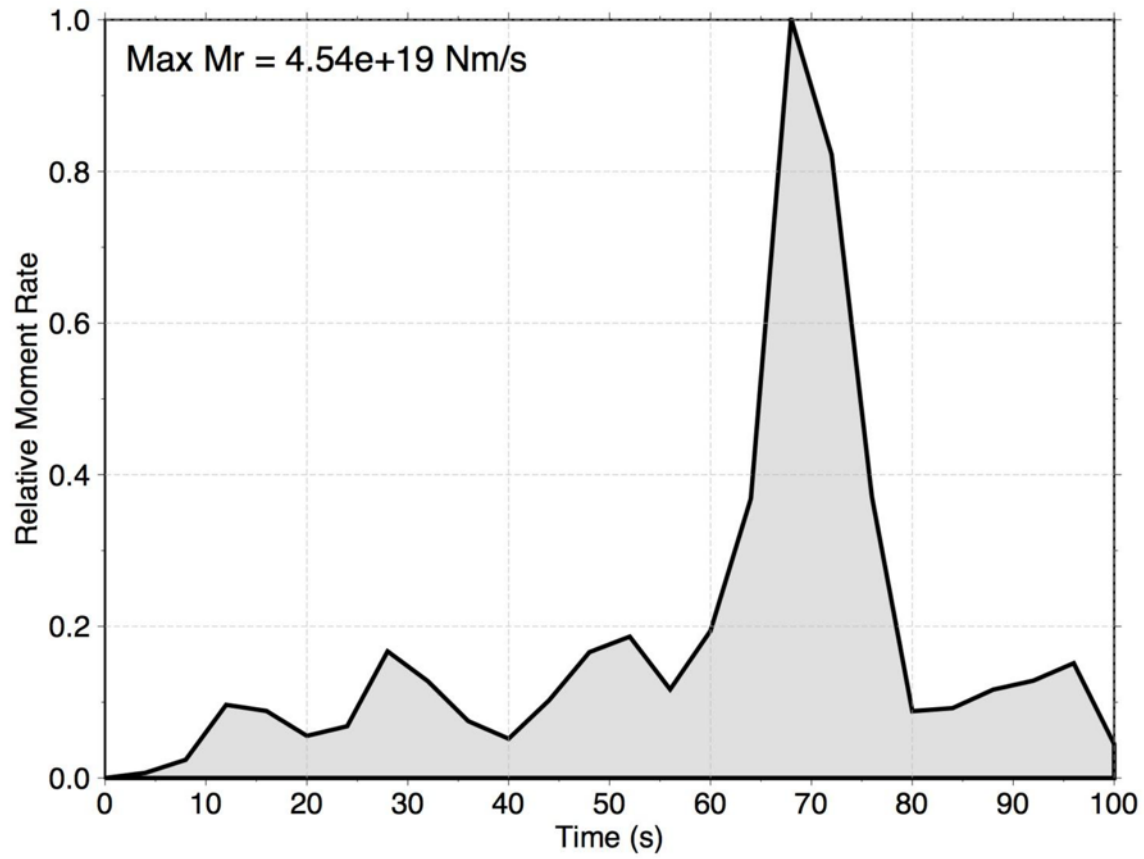


Figure S3.6: Moment rate functions of the M7.8 Kaikoura earthquake obtained by USGS, describing the rate of moment release with time after earthquake origin, relative to the peak moment rate (listed in the top left corner of the plot).

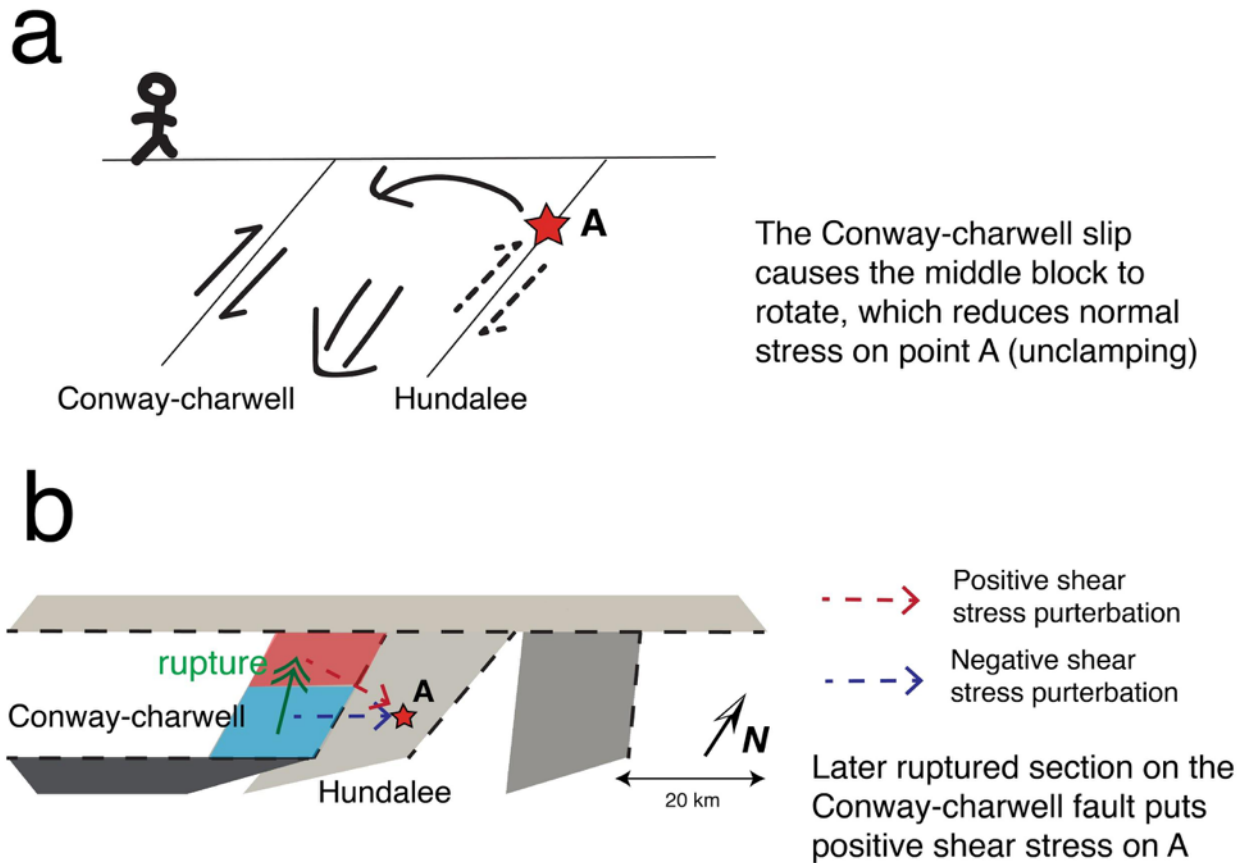


Figure S3.7: Conceptual cartoon that explains the rupture process. 2D schematic showing how the Conway-charwell fault slip unclamps the Hundalee fault. The Conway-charwell fault slip rotates the block in the middle, which reduces the normal stress on point A. (b) 3D schematic showing why the rupture would take a zigzag route. Green arrow represent rupture direction. The lower part of the Conway-charwell fault (blue region) sends negative shear stress to the Hundalee fault, while the upper part of the Conway-charwell fault (red region) sends positive shear stress to the Hundalee fault.

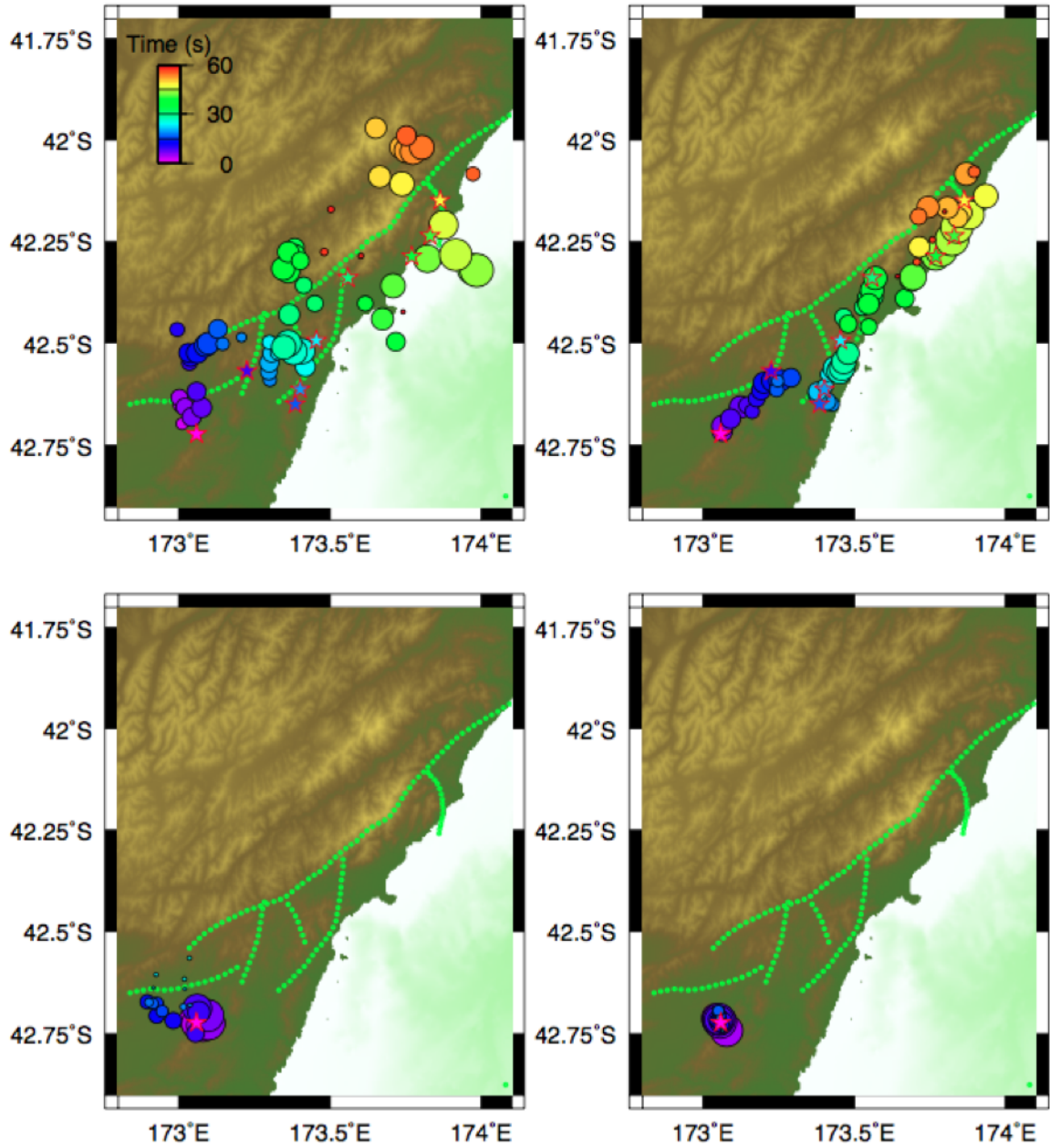


Figure S3.8: Synthetic resolution test of back projection. Input subevents are marked with stars with red edges and color coded by time. The resolved BP radiators are plotted in circles sized by power and colored by time. (a) is the BP image using synthetic eGF with realistic structural effects. The eGF has a BP image as shown in (c). And (b) is the BP image using ideal eGF (simplified as a perfect point source). The eGF has a BP image as shown in (d).

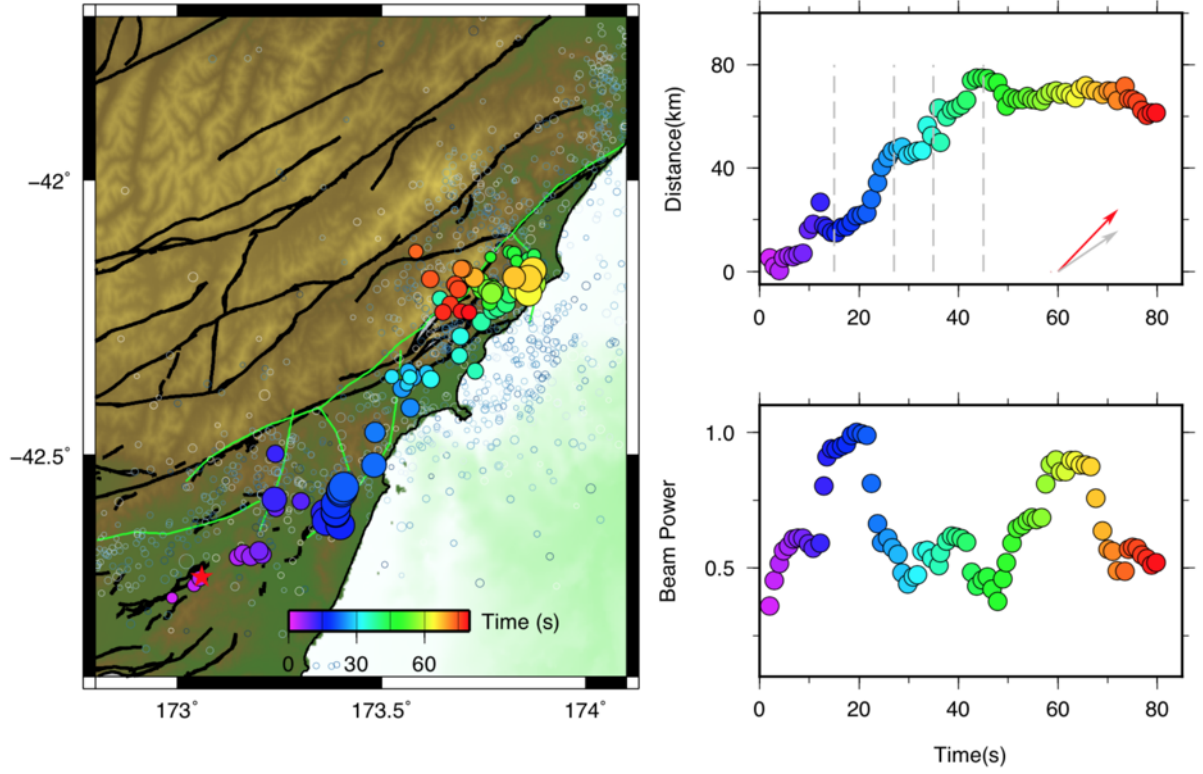


Figure S3.9: Backprojections of the Kaikoura earthquake for the first 80 s. (a) First 80 s of mainshock HF radiators (colored dots) imaged by the slowness calibrated BP, with the circle sized by power and colored by time. The red star denotes the National Earthquake Information Center (NEIC) epicenter. (b) Rupture distance as a function of time along the direction of the Hope-Jordan-Kekerengu fault system. The grey and red dashed arrows denote rupture speeds of 1.4 and 2 km/s, respectively. Relative fast rupture segments are marked by grey dashed lines. (c) Relative beam power as a function of time. HFZ: Humps fault zone, HF: Hope fault, CCF: Conway-Charwell fault, HDF: Hundalee fault, JF: Jordan fault, PF: Papatea fault, KF: Kekerengu fault.

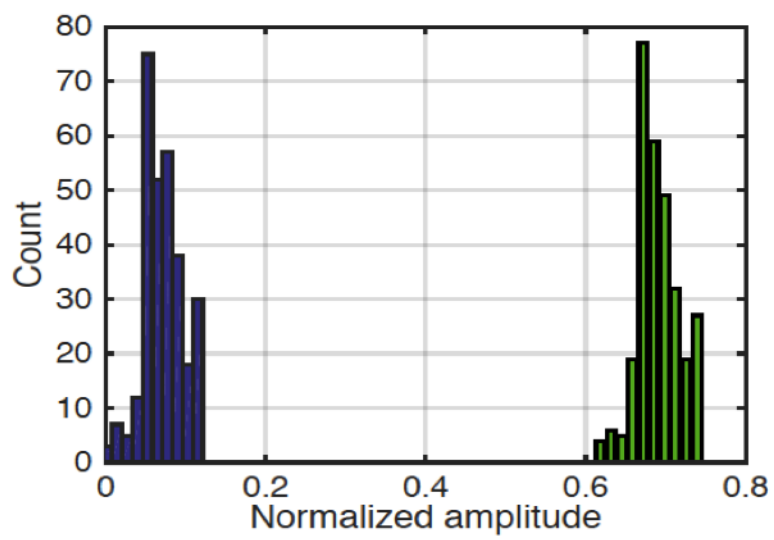


Figure S3.10: Histogram of normalized amplitudes. The figure shows normalized amplitudes of all the stations for the subevent on the Kekerengu fault (blue) and on the southern part (green).

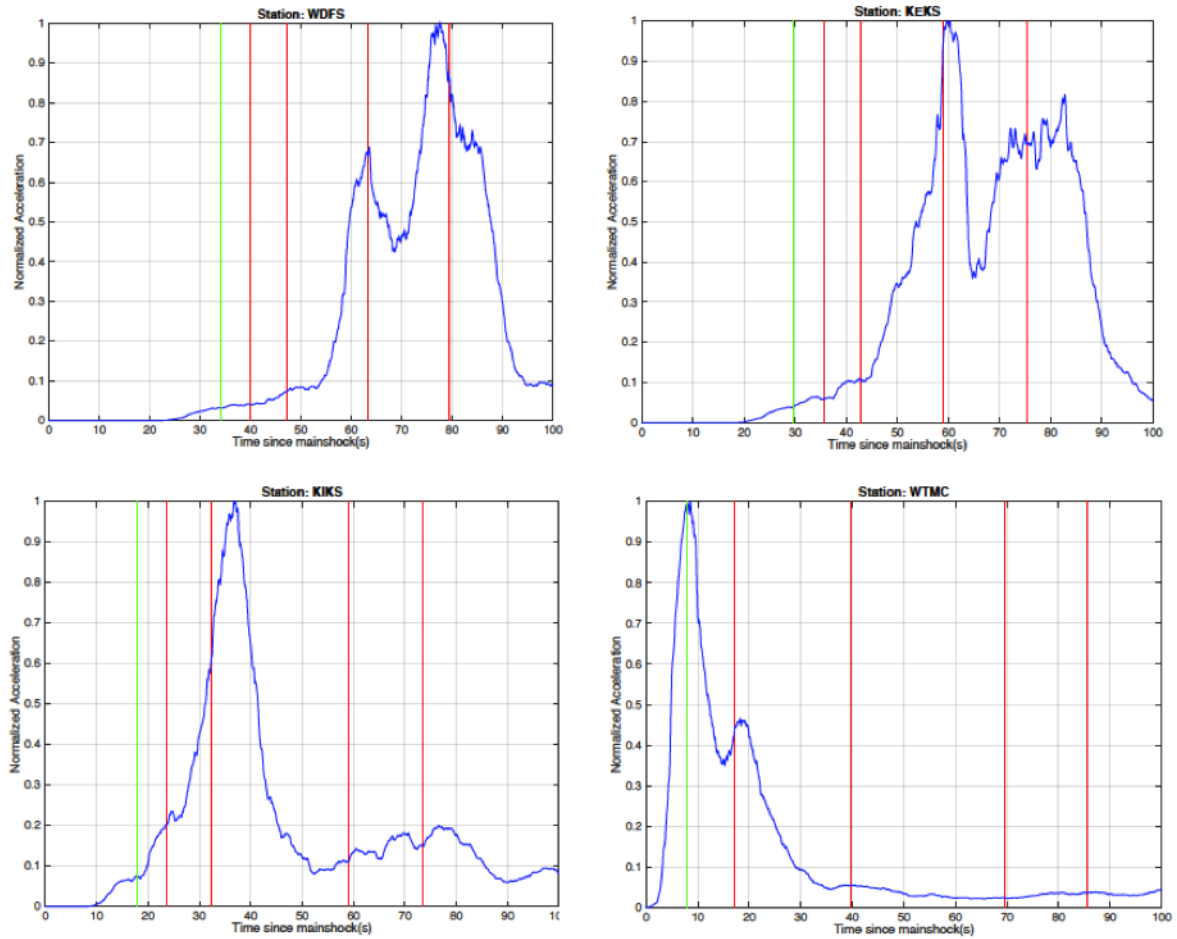


Figure S3.11: Smoothed Envelope of accelerations at near field stations. The green vertical line is the predicted arrival time of the hypocenter $(-42.72, 173.06)$. And the red solid vertical lines are the predicting timing of 4 subevents occurred at the Humps-Conway Charwell, the Hundalee, the Papatea faults and scattered radiators near the Jordan fault respectively.

CHAPTER 4

Complex Faulting and Triggered Rupture During the 2018 MW 7.9 Offshore Kodiak, Alaska Earthquake

Abstract

We study the rupture process of the 2018 MW 7.9 Offshore Kodiak earthquake with aftershock relocations, source mechanisms, back projection and slip inversions. We observe rupture branching from the S-to-N to the E-to-W segments. The lack of internal connectivity within the rupture suggests a high degree of dynamic triggering and a complexity consistent with other recent large oceanic-plate earthquakes. Overall, left-lateral faulting on conjugate SW-trending fault planes released most of the energy dominating the high-frequency radiated wavefield and the resulting crustal deformation.

This Chapter is modified from Ruppert, N. A., Rollins, C., Zhang, A., Meng, L., Holtkamp, S. G., West, M. E., & Freymueller, J. T. (2018). Complex faulting and triggered rupture during the 2018 MW 7.9 offshore Kodiak, Alaska, earthquake. *Geophysical Research Letters*, 45(15), 7533-7541.

I am the contributing author to conduct back projection and analyze the potential of dynamic triggering. Ruppert, N. A., Rollins, C., Holtkamp, S. G., West, M. E., and Freymueller, J. T. conducted aftershock relocation and source inversions.

4.1 Introduction

In the early morning hours of January 23, 2018, a magnitude 7.9 earthquake struck ~ 300 km offshore Kodiak Island, Alaska in the outer rise region of the Alaska-Aleutian subduction zone. While the moment tensor for the earthquake suggests predominantly strike-slip faulting, the true complexity of the source has only become evident through analysis of multiple datasets.

The Alaska-Aleutian subduction zone extends for $\sim 3,800$ km from the Gulf of Alaska westward to the Komandorsky Islands and accommodates relative motion between the Pacific and North American plates. The characteristics of this tectonic boundary, such as convergence direction and rate, trench separation and degree of coupling, vary widely along its extent (Freymueller et al., 2008). The nature of outer rise seismicity also changes along the arc: typical normal faulting outer rise events are more common in the central and western Aleutians (west of 160W) and nearly absent in the eastern Aleutians and southern Alaska (Christensen & Ruff, 1988). Very few earthquake source mechanisms are available for the outer rise region offshore Kodiak Island. Those that do exist do not indicate a uniform pattern of faulting. An Mb 5.4 event occurred a few months after the great 1964 Mw 9.2 Alaska earthquake and featured typical outer rise normal faulting (Stauder & Bollinger, 1966). An Mw 6.0 earthquake in 1999 east of the current sequence was a reverse faulting event and an Mw 4.9 earthquake in 2017 was a strike-slip event with a similar focal mechanism to the 2018 Mw 7.9 earthquake (Figure 4.1).

A series Mw 7.2-7.8 strike-slip events occurred in the northeast Gulf of Alaska in 1987-1988 on a system of conjugate and parallel faults in an area with very little historic seismicity (Lahr et al., 1988; Hwang & Kanamori, 1992). However, seismic reflection data provides clear evidence that these faults were active long before the earthquake sequence began (Reece et al., 2013). These events were followed by more examples of complex faulting in oceanic lithosphere. The 1998 Mw 8.1 Balleny Islands (Antarctic Plate) earthquake ruptured faults that did not correlate with any known features on the ocean floor and may have involved sub-events on different fault planes (Henry et al., 2000). The 2000 Mw 7.8 Wharton Basin

earthquake (off Sumatra) possibly involved ruptures of both NS- and EW-trending fault planes during the mainshock ([Robinson et al., 2001](#); [Abercrombie et al., 2003](#)) The 2004 Mw 8.1 Tasman Sea earthquake reactivated old structures with a new mechanism, rupturing an old spreading ridge in a strike-slip event ([Robinson, 2011](#)).

More recently, the 2012 Wharton Basin sequence involved multiple $M_{\text{L}}8$ sub-events within a complex set of faults intersecting at highly oblique angles ([Meng, Ampuero, Sladen, & Rendon, 2012](#)), possibly including reactivated fracture zones ([Wei et al., 2013](#)). High-resolution bathymetry and seismic reflection data in the source region of the 2012 earthquakes illuminate left-lateral strike-slip faults along reactivated oceanic fracture zones and conjugate shear zones expressed by sets of normal faults ([Singh et al., 2017](#)). In this light, the faulting process of the Mw 8.6 event was reinterpreted as an en echelon faulting featuring 75-100-km-long fault segments along the fracture zones and shorter segments in the direction of the shear zones.

These examples are a reminder that the far-field observations are often an aggregate over a series of sub-faults. The studies above suggest this may be particularly common in oceanic crust, where the preexisting fabric may not necessarily align with the dominant stress directions. In these cases, stress release might be most easily accommodated through a complex set of sub-fault ruptures instead of occurring along a single canonical fault line.

Several fracture zones have been identified in the vicinity of the 2018 Offshore Kodiak Island earthquake such as the Aja Fracture Zone to the south ([Pitman & Hayes, 1968](#); [Matthews et al., 2011](#)). In addition, seismic reflection and high-resolution bathymetry have revealed several N-S trending faults ([Reece et al., 2013](#)), which were interpreted as preexisting fractures in the Pacific Plate spreading fabric that have been reactivated as predominately normal plate bending faults by Pacific Plate subduction. This sea floor fabric bears striking resemblance to that imaged in the Wharton Basin region ([Singh et al., 2017](#)).

4.2 Methods and Results

4.2.1 Aftershock Relocations and Source Mechanisms

Over 2800 aftershocks are produced until the end of March, 2018. With a first glance, aftershocks are likely to locate in multiple faults, that demonstrate a complex geometry. In order to acquire a more accurate aftershock catalog, we employ the earthquake relocation technique GrowClust ([Trugman & Shearer, 2017](#)).

The overall aftershock distribution shows several sharp N-S and broad NE-SW trending faults (Figure 4.2). We identified 4 distinct aftershock groups associated with possible fault planes (S1-4 in Figure 4.2). The most prominent is the 110-km-long N-S trend with the mainshock epicenter located near its southern end (S1). Neither catalog nor relocated aftershocks indicate a single, well-defined $\sim$ EW-striking trend. Instead, epicenters in the region west of S1 are distributed over a broad NE-SW trending zone (S2) and form several clear N-S streaks $\sim$ 10 km long (Figure 4.2 inset). Focal mechanisms for aftershocks along these streaks show right-lateral strike slip faulting. There is also a broad 70-km-long NS-aligned cluster northeast of the main S1 segment (S4). However, no connecting structure between the main segment and this northeast segment is clear in the aftershock locations (S3).

The focal mechanisms for aftershocks within S1 and S4 segments (NEIC ComCat) predominantly indicate right-lateral strike-slip motion. A cluster of reverse-faulting aftershocks is concentrated near the southern end of the main S1 lineation. These are located just north of the identified fracture zone and may indicate rupture interaction with this structural barrier. One of the only two normal faulting aftershocks is associated with the S4 aftershock cluster and indicates NS-fault planes (it was the largest aftershock). The second normal-faulting aftershock is located in the middle of the S2 segment and indicates a W- or SW-trending fault plane. Its location is close to the Mb 5.5 normal faulting event in 1964.

4.2.2 Back-projection Results

We back-projected teleseismic P waves recorded at 298 broadband stations in continental Europe to image the kinematic rupture process of the MW 7.9 mainshock (Figure 4.3). The back-projection technique utilizes very few prior assumptions on fault parametrization, Greens functions, or rupture velocity, and thus can robustly constrain the spatio-temporal evolution, rupture extent and rupture speed of large complex earthquakes (Ishii et al., 2005; Meng, Ampuero, Sladen, & Rendon, 2012; Zhang et al., 2012; Fan & Shearer, 2015; D. Wang, Mori, & Koketsu, 2016). In this study, we adopted the Multitaper-MUSIC array processing technique that provides superior resolution to separate closely spaced sources (Meng et al., 2011). We collected coherent waveform data from the ORFEUS data center (<http://orfeus-eu.org/webdc3/>). The seismograms are filtered in the band of 0.5 to 2 Hz, and empirically aligned with the first 5 seconds of the P-wave arrivals. This correction procedure assumes that the first arrivals originated from the hypocenter (Ishii et al., 2005) and provides a static correction of travel time due to 3D path effects. The aligned data are then back-projected using a 1D reference velocity model to pinpoint when and where difference source regions were radiating energy.

The results suggest complex conjugate ruptures and segmentation along several faults trending $\sim$ NS and $\sim$ EW observations that are remarkably consistent with the aftershock locations (Figure 4.3). The earthquake initiated as a relatively low-energy bilateral rupture along a NS-trending fault (S1) near its southern end with the first notable energy burst at around 10 seconds. Between 20 and 30 seconds, the rupture leaps $\sim$ 60 km NW from the earthquake epicenter (S1.5 segment). From 30 to 50 seconds, the dominant release occurs along a continuous E-to-W trending path with a rupture speed of 3 km/s (S2 segment). Beyond 50 seconds, radiators are less energetic. This late energy appears well constrained to a region north and east of the rest of the rupture. It likely represents another fault segment near the northern end of S1 with the strike in the westerly direction and predominant propagation path from east to west (S3). The rupture speed on S3 is around 3 km/s, similar to that on S2. Beyond 75 seconds, back-projection is no longer viable because the teleseismic

P waves lack coherency. The overall rupture extent is ~ 100 km in the NS direction, which is consistent with the aftershock zone extent. The average NS rupture speed is (~ 1.4 km/s), is lower than the two EW ruptures. The rupture pattern is overall highly complex, as inferred for the previous oceanic-plate events.

4.2.3 GPS Modeling Results

A five-segment fault model is proposed based on the distribution of relocated aftershocks and BP radiators, consisting of three vertical fault segments striking exactly north-south and two striking 253° (Figure 4.4b). Coseismic slip is inverted from the horizontal and vertical GPS displacements recorded in Plate Boundary Observatory (PBO) using a Monte Carlo method.

Inverted slip distribution is consistent with the distribution of high-frequency radiators. Relatively little slip is inferred on the two fault segments striking north-south, particularly the easternmost segment S4. Most of the slip is inferred on the segments S2 and S3 striking 253° , consistent GPS displacements. The total coseismic moment release in the GPS-based model (computed as the sum of $SA\mu$, where S is the slip, A is the area of each patch and μ is the shear modulus along each patch, here 45 GPa as per the heterogeneous Earth model) is equivalent to a $M_w=7.81$ earthquake. We find that this slip distribution would have imparted a Coulomb stress increase of ~ 0.5 bars to the Alaska subduction zone along the nearby trench (Figure 4.4).

4.3 Discussion

The M_w 7.9 Offshore Kodiak earthquake is another example of a complex rupture of the oceanic lithosphere. The source complexity is evidenced by the aftershock distribution, which extends over a $> 100 \times 100 - km$ region and includes several well-defined N-S lineations but no clear E-W trends. At the same time, GPS modeling of crustal deformation (this study) and finite source modeling with teleseismic recordings(<https://earthquake.usgs.gov/>

earthquakes/eventpage/us2000cmy3#finite-fault) favor left-lateral faulting on faults striking south of west. The back-projection results presented in this study also support complex rupture model featuring rupture in both N-S and E-W directions.

P-wave first-motion analysis, location of the epicenter and back-projection results all point to the rupture initiation as a right-lateral strike-slip earthquake on a N-S oriented fault plane, with predominant propagation to the north, but also some propagation to the south. However, instead of simply continuing along this faulting plane, the rupture bifurcated into a left-lateral strike-slip conjugate fault plane westward. The most energetic back-projection radiators coincide with the location of the centroid moment tensor (<http://www.globalcmt.org>) near the point where the two proposed planes would intersect (Figure 4.3). The GPS modeling also required significant motions on this westerly-trending fault plane (Figure 4.4). At the same time, a number of smaller right-lateral strike-slip ruptures were triggered along NS oriented plate bending faults (Figure 4.2, inset).

Additional complication of the rupture process occurred between 20-30 seconds (Figure 4.3). The high-frequency radiators suddenly shifted 60 km NW of the epicenter to a region that is seemingly disconnected from the main rupture propagation. Following this (30-50 seconds) an NE-SW left-lateral rupture crossed this area once again. Later in the sequence (50-70 seconds), there was another $\sim$ E-to-W rupture propagation on S3 that again would have been triggered either during or after the main rupture propagation along S1. The aftershock locations clearly image a 70-km-long NS-oriented fault segment (S4), which corresponds to high-frequency radiators at around and after 65 seconds. This last fault segment has not significantly contributed towards resulting crustal displacements.

To test triggering mechanisms during the mainshock rupture we compute static Coulomb and dynamic stress changes between various ruptured sub-faults. The Coulomb stress changes are computed using the elastic component of RELAX (Barbot & Fialko, 2010a, 2010b) assuming a homogeneous halfspace Earth model and an effective friction coefficient of 0.4 (King et al., 1994). We assume right-lateral and left-lateral strike-slip faulting on northerly and westerly trending fault segments, respectively. Results indicate that static

stress changes along S1-3 sub-faults are either negligible or clamping, while the stress changes along S4 are significant enough to promote the faulting (Figure 4.5 a-d).

To address the effect of dynamic triggering from S1 to S1.5 and S2, we compute the synthetic three-component velocity seismograms using the frequencywavenumber integral method (Zhu & Rivera, 2002). Moment tensor of S1 and the fault orientation of S2 are taken from the slip inversion (Figure 4.3). Figure 4.5e is an example of the seismogram at (55.95N, 149.82W), where we see the cluster of high frequency (HF) radiators on S1.5 and S2. The amplitudes of the radial component are more pronounced than the transverse component, suggesting that Rayleigh waves play a dominant role of triggering the rupture. Moreover, the timing of the radiators on HF matches with the pulses of the seismogram. According to (Jaeger et al., 1979), peak dynamic stress can be inferred from the velocity:

$$\sigma = Gu/v_s \quad (4.1)$$

where G is the shear modulus, u is the peak velocity and v_s is the phase velocity. In this study, we use a shear modulus of 30 GPa and a constant Rayleigh waves velocity of 3.5 km/s (Miyazawa & Brodsky, 2008). The peak ground velocity is about 0.78 cm/s, and the corresponding peak stress change is around 0.67 bar, above the threshold of triggering of around 0.1 bar (Brodsky & Prejean, 2005; Hill & Prejean, 2015). Therefore the HF radiators observed on S1.5 and S2 can be dynamically triggered by the Rayleigh wave radiated from S1. In summary, we hypothesize that the dynamic stress change played a significant role in triggering N-S sub-fault rupture S1.5 ahead of the main rupture front reaching that area and promoted westerly bifurcation of the rupture onto a conjugate fault plane S2. The static stress changes were significant enough to possibly promote the rupture of S4.

4.4 Conclusion

Complex faulting pattern of the 2018 Mw 7.9 Offshore Kodiak earthquake indicates a complex rupture on a system of N-S and NE-SW-oriented conjugate faults. The N-S faults correlate with plate bending faults imaged by seismic imaging and high-definition bathymetry

([Reece et al., 2013](#)). The NE-SW faulting trends do not coincide with locations of oceanic plate fractures, but correlate with orientations of fault planes of previously recorded earthquakes in the area. Rupture propagation during the earthquake did not follow a continuous path and was partially driven by dynamic and static stress triggering. The back-projection and GPS modeling results indicate that left-lateral strike-slip segments released most energy thus dominating the radiated wavefield and the resulting crustal deformation.

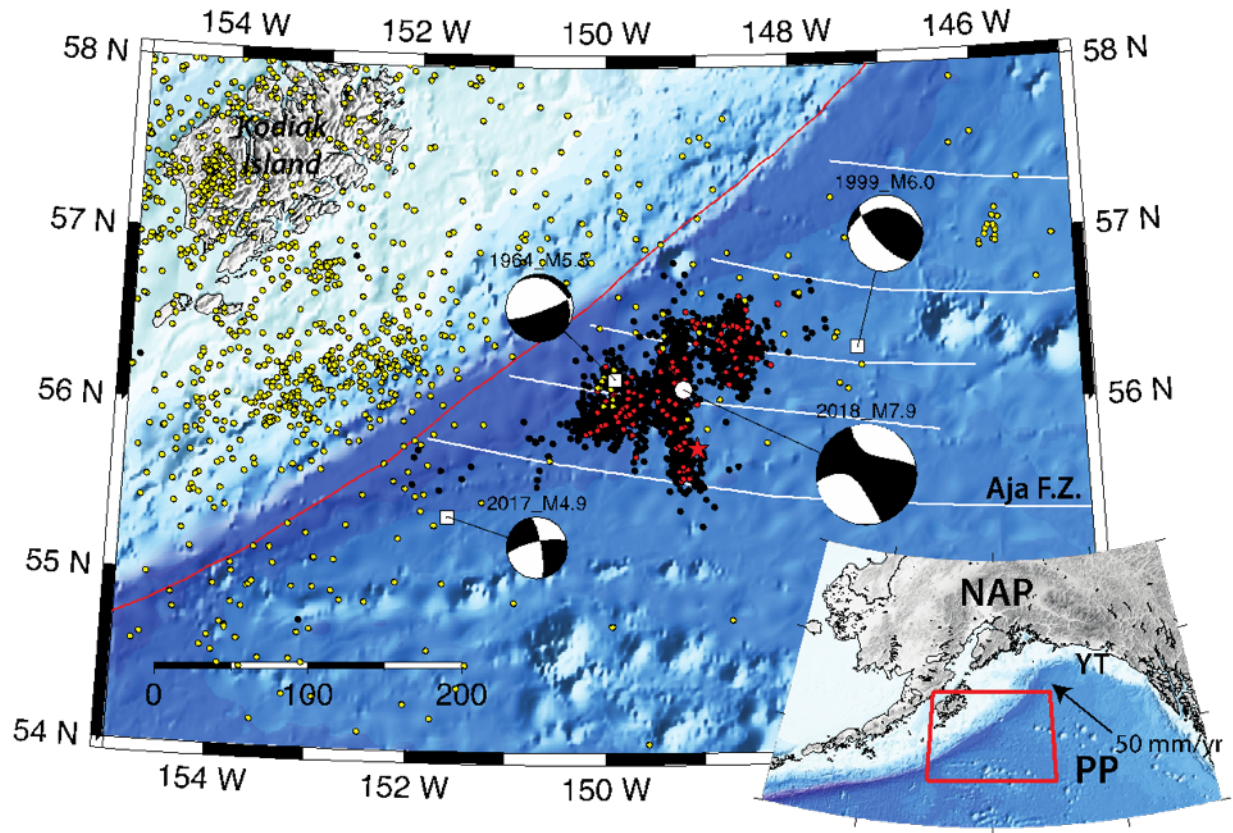


Figure 4.1: Regional map showing background seismicity ($M \geq 3$, 1990-2017, yellow circles), 2018 aftershocks ($M \geq 3$ and 4, black and red circles, respectively), moment tensors and locations of previously recorded events in the outer rise region (white squares) and the 23 January 2018 MW 7.9 earthquake (circle indicates location of the centroid). Inferred fracture zones are shown by white lines and the megathrust is marked by red line. Inset figure shows location of the study area within the regional frame (NAP North American Plate, PP Pacific Plate, YT Yakutat Terrane).

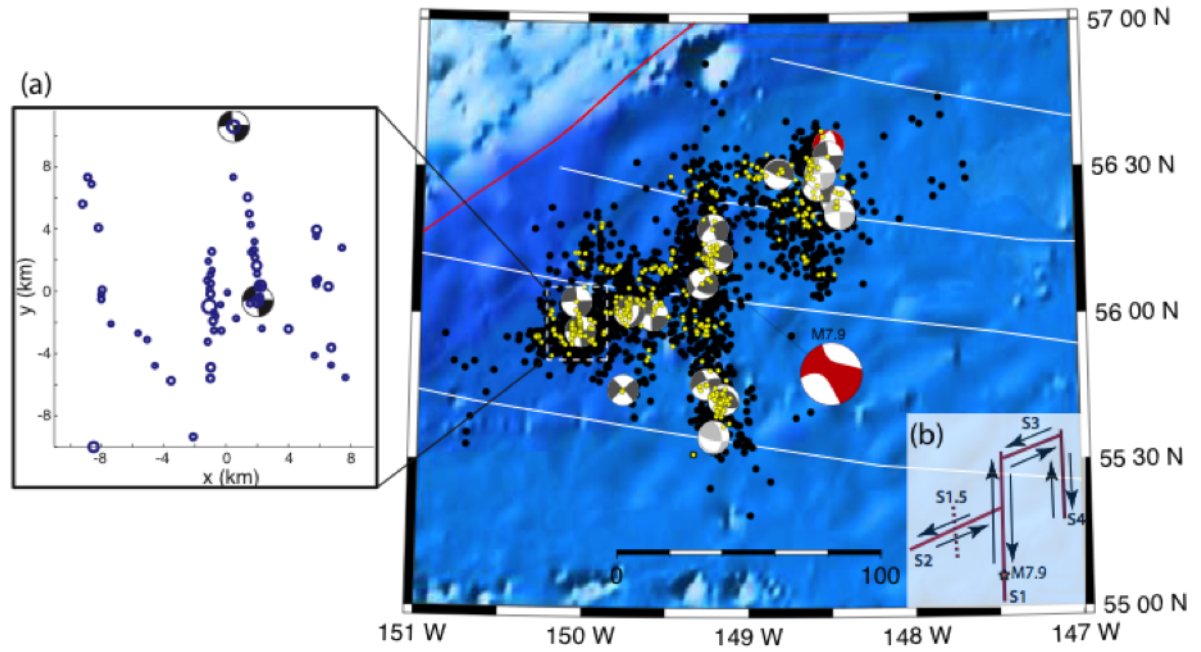


Figure 4.2: Map of the aftershocks and moment tensors: black $M \geq 3$ aftershocks from Alaska Earthquake Center catalog, yellow relocated aftershocks. Grey beach balls are from NEIC ComCat catalog and red beach balls are the centroid moment tensors from GCMT for the mainshock and the largest aftershock (24 January 00:04 UTC). Inset map (a) shows a subset of relocated aftershocks that demonstrates right-lateral faulting along NS trending faults. Inset (b) shows a proposed fault segment model.

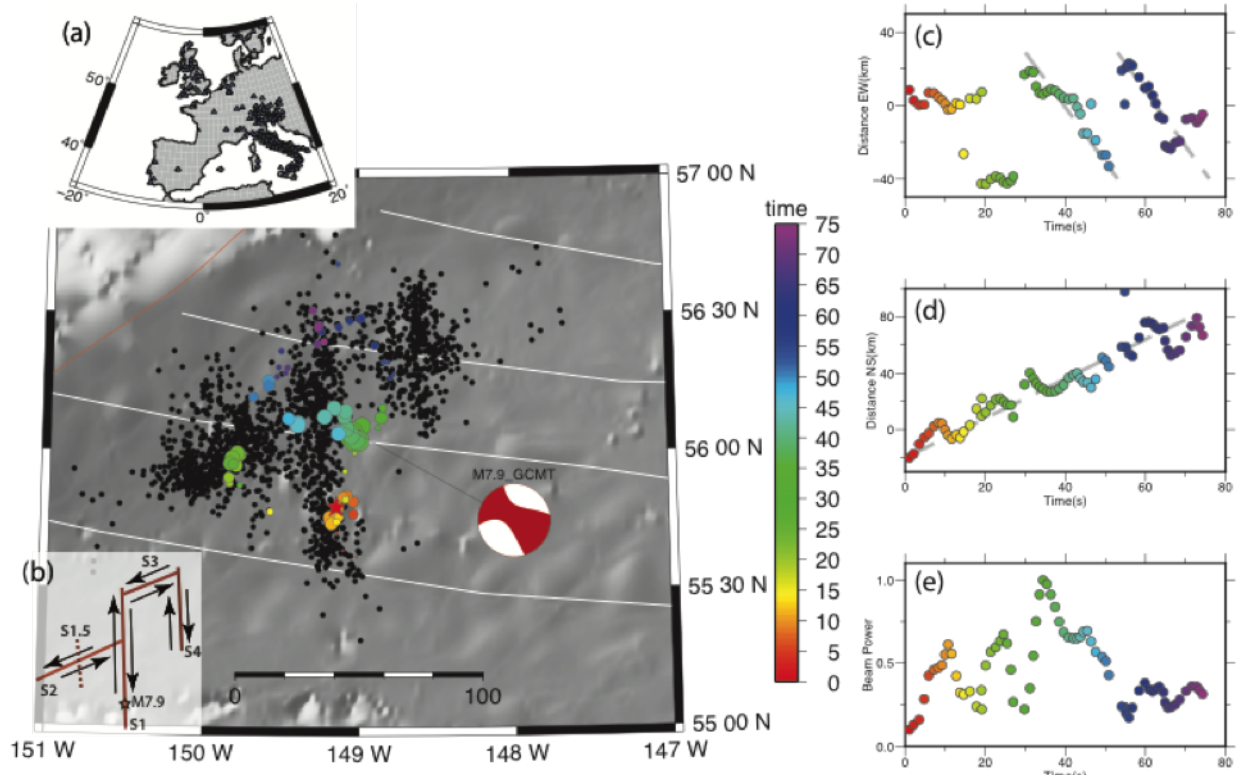


Figure 4.3: Kinematic rupture process revealed by back-projection (BP). The circles are color-coded by time and sized by power to represent the location of high frequency radiators at different time frame. Red star is the MW 7.9 epicenter and black circles are $M_L=3$ aftershocks. Inset map (a) shows distribution of stations used for BP. Inset (b) shows a proposed fault segment model. Right top panel (c) shows rupture distance along the EW direction (positive distance indicates rupture to the east of the hypocenter); dashed lines are reference rupture speed of 3 km/s. Right middle panel (d) shows rupture distance along the NS direction (positive distance indicates rupture to the north of the hypocenter); the dashed line indicates reference rupture speed of 1.4 km/s. Right bottom panel (e) shows beam power of the radiators as a function of time.

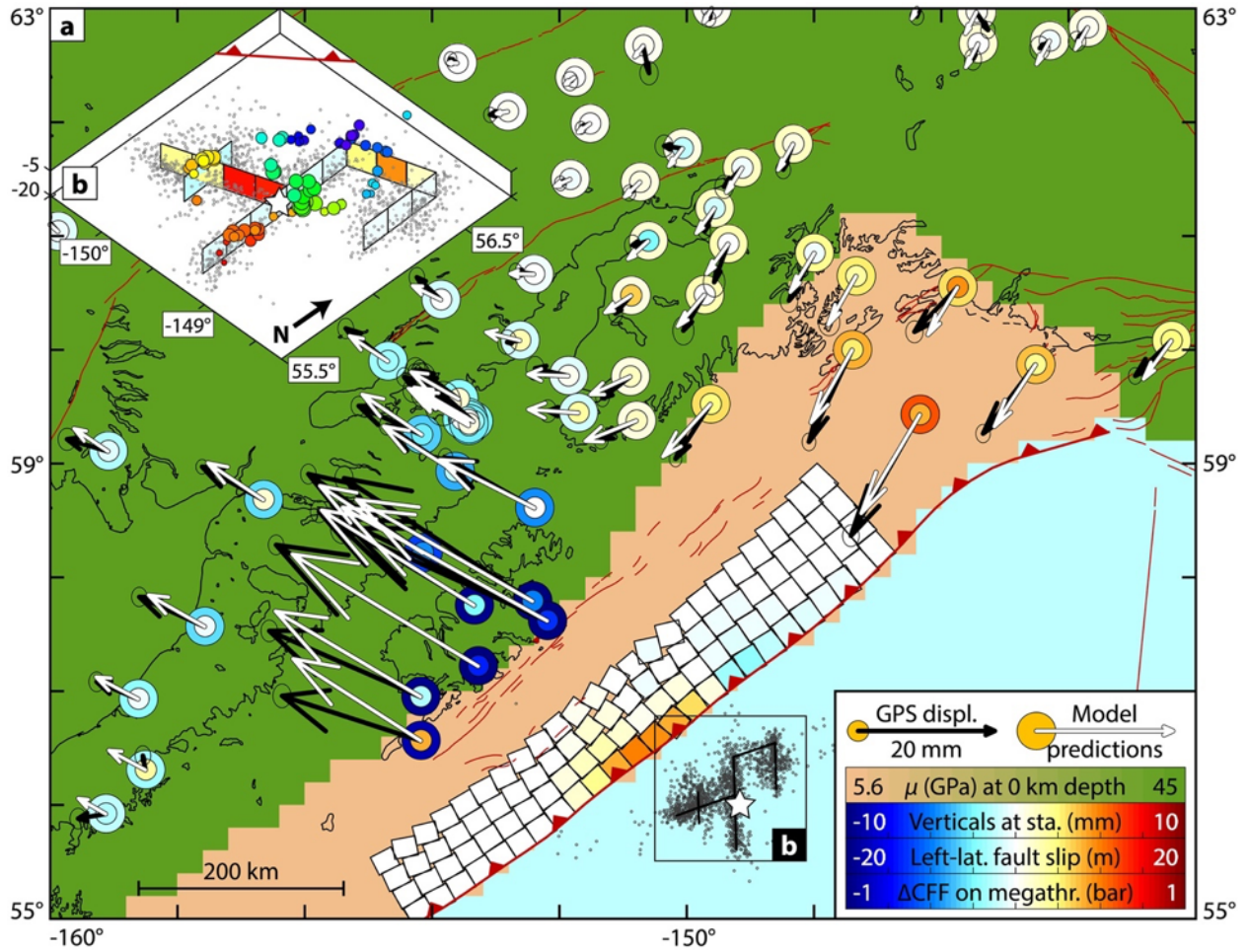


Figure 4.4: Observed and modeled static GPS displacements (a) and estimated slip on ruptured fault segments (b). Black arrows and small colored circles are observed horizontal and vertical displacements. White arrows and large colored circles are model-predicted horizontal and vertical displacements. Warm colors indicate uplift at GPS stations and left-lateral slip on segments in the inset; cool colors indicate subsidence and right-lateral slip. Dark gray circles are $M \geq 3$ aftershocks. Background beige color represents accretionary prism in the model; gray represents more typical lithosphere. Red lines are faults. Also shown are Coulomb stress changes imparted to the megathrust.

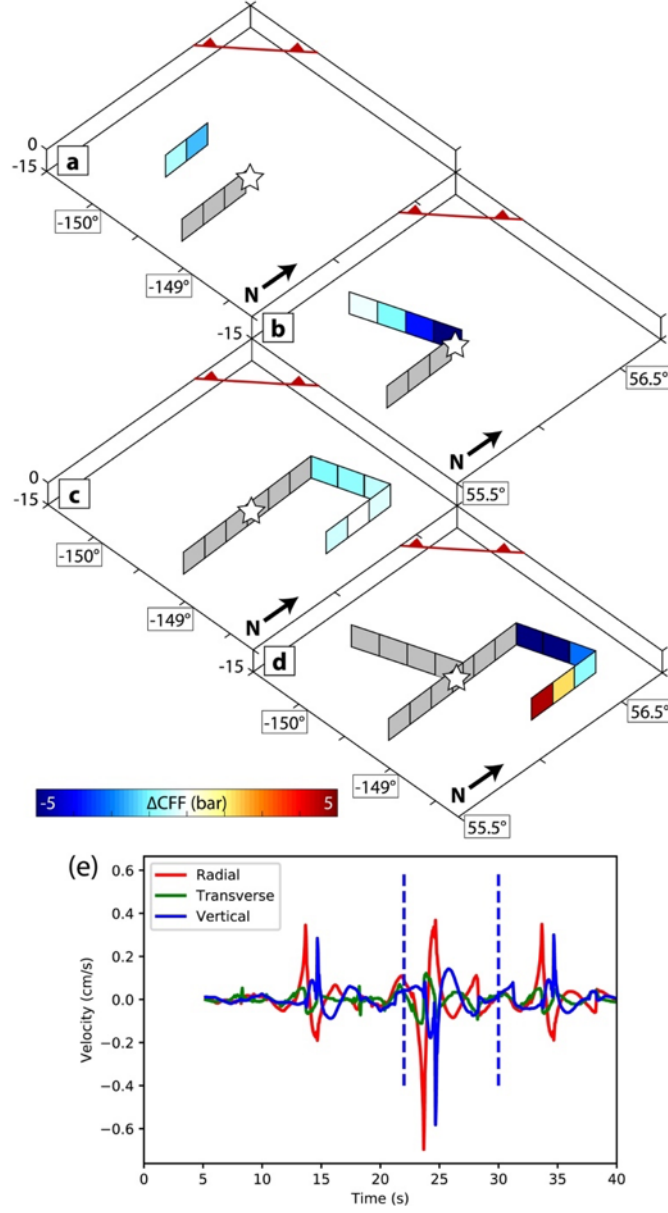


Figure 4.5: Results of the static and dynamic stress triggering tests. Progressive Coulomb stress changes throughout the progression of the rupture. (a) Stress changes imparted by slip on the southern half of S1 to S1.5. (b) Stress changes imparted by slip on the southern half of S1 to S2. (c) Stress changes imparted by slip on the full length of S1 to S3 and S4. (d) Stress changes imparted by slip on S1 and S2 to S3 and S4. (e) Synthetic three-component velocity seismograms recorded at 55.95N, 149.82W. The blue dashed line marks the timing of the high frequency radiators of back projection on sub-fault S2.

CHAPTER 5

Source Complexity of the 2015 Mw 7.9 Bonin Earthquake

Abstract

We combine BP and Multiple Source Inversion (MSI) to study the rupture complexity of the 2015 Mw 7.9 Bonin Earthquake. Both BP and MSI suggest the earthquake can not be parameterized as a planar fault rupture. The rupture is initiated as a bilateral SW-NE propagation, and followed by a NW source extension. This event is potentially a cascade failure of deep ruptures on multiple nearby faults.

This Chapter is modified from Chen, Y., Meng, L., Zhang, A., & Wen, L. (2018). Source complexity of the 2015 Mw 7.9 Bonin earthquake. *Geochemistry, Geophysics, Geosystems*, 19(7), 2109-2120. I am the contributing author to conduct 3D back projection and analyze the uncertainty related testes. Chen, Y., et al., performed multiple source inversions and led the discussion.

5.1 Introduction

Since deep-focus earthquakes were first discovered in the 1920s, the physics of deep-focus earthquakes is still under debate. Deep-focus earthquakes cannot be facilitated by the brittle-frictional processes like their shallow counterparts, because increasing pressure and temperature with depth strongly promotes ductile deformation and precludes brittle failure (Scholz, 2019). Various hypotheses have been proposed to explore the seismogenic mystery of deep-

focus earthquakes. Two physical mechanisms, transformational failure and shear thermal instability, are widely considered as candidates to induce deep-focus ruptures. The transformational faulting mechanism assumes that metastable olivine wedges exist in subducted slabs. Phase transformation with volume shrink and heat release probably leads to an abrupt breakdown of the meta-stable olivine (Green, 2007; Green & Houston, 1995; Houston, 2015; Kirby, 1987; Y. Wang et al., 2017). L. Liu and Zhang (2015) proposed that differential volume changes between the crust and mantle lithosphere is able to cause shear strain accumulation to trigger and induce deep earthquakes. Shear thermal instability hypothesizes that a non-linear positive feedback between temperature-dependent rheology and shear heating promotes shear localization that is able to achieve a catastrophic earthquake failure (Hobbs & Ord, 1988; Karato et al., 2001; Kelemen & Hirth, 2007; Ogawa, 1987; Kanamori et al., 1998; Prieto et al., 2013).

Previous investigations provided important constraints on source processes and physical mechanisms of deep-focus earthquakes. The fault plane orientation analysis of deep-focus earthquakes exhibited a complex rupture system of deep-focus earthquakes with both sub-horizontal and sub-vertical fault planes in the Tonga, Japan-Kuril-Kamchatka and Izu-Bonin-Mariana subduction zones, suggesting that local stress fields may control the fault plane orientation (Myhill, 2012; Warren et al., 2015, 2007). The source dimensions of some large earthquakes were proposed to extend out of the geometry of the metastable olivine zones (Y. Chen & Wen, 2015; Y. Chen et al., 2014; Silver et al., 1995; Wiens et al., 1994; Zhan et al., 2014). However, W.-P. Chen (1995) suggested that the 1994 Mw 8.2 Bolivia earthquake could be interpreted as En echelon faults that could be fitted into the metastable olivine zone. Rupture parameters are important for understanding deep earthquake physics. Ye, Lay, Kanamori, et al. (2016) reported diverse rupture parameters for the 2015 Peru deep earthquake doublet, indicating different physical mechanisms. Ruiz et al. (2017) proposed that the doublet do not show significant different rupture parameters. It is also crucial for determining the detailed source process whether deep-focus earthquakes fail on a rupture plane (Antolik et al., 1996, 1999; Ye et al., 2013), several asperities (Silver et al., 1995; Wei

et al., 2013; Zhan et al., 2014) or different rupture planes triggering each other (Y. Chen & Wen, 2015; Y. Chen et al., 2014). In summary, different observations resulted in different conclusions. Elaborate analyses are needed to further constrain deep-focus earthquakes

The recent 2015 Bonin earthquake (Figure. 1a) is the third largest deep-focus events ever recorded. It can provide a unique opportunity to investigate the source properties and dynamics, physical mechanisms of large deep-focus earthquakes, and even the seismic structure of the Izu-Bonin slab. The earthquake is reported with a moment magnitude of Mw 7.9 and a depth of 677 km by the Global Centroid Moment Tensor solutions (GCMT). The USGS slab 1.0 model (Hayes et al., 2012) and the tomography results (Fukao & Obayashi, 2013) showed that the Pacific slab subducts to the west beneath the Asian plate, and the International Seismological Centre (ISC) EHB seismicity (Engdahl et al., 1998) above 600 km is well consistent with these models (Figure. 1b, 1c). However, the Bonin earthquake is below the assumed slab illuminated by the background seismicity (Figure. 1c). An important approach to distinguishing the physical mechanism is if the earthquake occurred in the core of the slab (Silver et al., 1995; Wiens et al., 1994). The slab morphology is unclear and controversial as the previous studies reported (Obayashi et al., 2017; Okal & Kirby, 2016; Porritt & Yoshioka, 2016; Takemura et al., 2016; Yang et al., 2017; Ye, Lay, Zhan, et al., 2016; Zhao et al., 2017). Those studies tried to explain why the 2015 Bonin earthquake is unusually deep and isolated from the background seismicity (Figure. 1b, 1c). Zhao et al. (2017) conducted P-wave tomography and suggested that the earthquake occurred within the subducting Pacific slab penetrating into the lower mantle without buckling. Obayashi et al. (2017) suggested a shoe-shaped stagnant slab, and the 2015 event occurred in the heel part of this shoe-shaped slab. Porritt and Yoshioka (2016) observed multiple conversions from receiver functions within and below the transition zone, and proposed slab pileup in the mantle transition zone. Yang et al. (2017) proposed that the 2015 event could be explained by the slab buckling. Okal and Kirby (2016) recommended that the 2015 event occurred in a detached and fragmented slab. Although the slab geometry is still unclear, Takemura et al. (2016) examined propagation characteristics of P waves for the Bonin earthquake and

compared them with the synthetics simulated with the earthquake located at the top, middle and bottom of the slab. They conclude that the Bonin earthquake occur at the bottom of the slab. The occurrence at the bottom of the slab is not consistent with the metastable-olivine wedge. Finite-fault inversion with a single fault plane was applied to study the earthquake (Ye, Lay, Zhan, et al., 2016). The work was based on a source model derived with a simplified planar fault plane. Tomography and dynamic simulation of the slab geometry and dynamics displayed the complex morphology in the slab (Obayashi et al., 2017; Porritt & Yoshioka, 2016; Yang et al., 2017; Zhao et al., 2017). Structural and stress heterogeneity of the slab may result in source geometrical complexity. Therefore, we re-investigate the earthquake with more detailed analyses.

We apply a new 3D back-projection (3DBP) and our multiple-source inversion (MSI) to study the spatio-temporal distribution and focal mechanism of the sub-events, and constrain the three-dimensional source geometry and kinematic rupture process of the 2015 Bonin earthquake. Our study displays a non-planar spatial distribution and focal mechanisms of the sub-events. The variation of focal mechanisms along the location can be explained by either rupture on a curved fault plane or sub-events crossing multiple sub-parallel fault interfaces. The result can be explained by the slab stress or structure heterogeneity.

5.2 3D Back-projection

Back projection (BP) has been recognized as an important observational approach to imaging large deep earthquakes at teleseismic distance (Ishii et al., 2005). Deep earthquake waveforms are typically simpler than their shallow counterparts as the velocity structure near the source is less heterogeneous at depth (Frohlich, 2006), which allows BP imaging of more coherent wavefields at higher frequencies than previously achieved in BP studies of shallow earthquakes (Meng et al., 2014). In our previous work, the BP imaging of the 2013 Mw 8.3 Okhotsk earthquake allowed for comparison of the rupture with local geometry of the subducting Pacific slab and evaluated the contributions of different physical mechanisms in a single earthquake (Meng et al., 2014).

A key limitation of the standard back-projection approach is the lack of depth resolution. Since a trade-off of source location and timing exists along the ray-path of the seismic wave, a predefined fault plane is thus required to reduce the spatial uncertainty. At teleseismic distance, the seismic rays travel almost vertically. The BP usually achieves good resolution on sub-horizontal fault planes such as the shallow-dipping subduction interface. However, in the case of the sub-vertical fault planes, the location trade-off remains significant, as the fault plane is sub-parallel to the ray-path. In other words, the BP method usually provides a better horizontal resolution than the depth resolution, similar to that of the earthquake location problems. One recent development in BP imaging improves the depth resolution by combining multiple seismic phases. These phases intersect in space such that it overcomes the trade-offs of individual rays. (Kiser & Ishii, 2011) developed a modified BP approach to study the mechanisms of intermediate-depth earthquakes. The direct P phase and depth pP phase are included in the back-projection analysis, which has potential to determining the energy distribution with respect to depth and time. This capability of 3D imaging is also important to accurately estimate the rupture speeds. In conventional 2D imaging, since the orientations of fault planes need to be assumed, the source dimension is projected on the fault plane. An incorrect depth would cause the trade-off of source occurring time, and the wrong choice of the plane may promote very large uncertainty of rupture speed. Here, we further develop a full-3D back-projection (3DBP) approach that resolves source locations in 3D volume.

Taking advantage of the recently developed large regional seismic arrays, we are able to image the source process by back-projecting the teleseismic data recorded by the Australian (AU), European (EU) and North American (NA) networks (Figure 5.1a). For both direct P and pP phases, the seismograms are filtered between 0.25 to 1 Hz, the highest frequency band with adequate signal to noise ratio and waveform coherency. We align the initial P-wave and pP-wave arrivals (Figure. S5.1, S5.2) separately with a multi-channel cross-correlation technique (VanDecar & Crosson, 1990). For the BPs of both phases, we adopt a 10-s long sliding window to balance the temporal resolution and robustness. For each

phase, we normalized the seismograms by their initial arrivals so that the waveforms have uniform amplitude at all stations. This procedure removes the path and site effect on the waveform amplitude since the BP is based solely on the coherent phase information. To fix the absolute location, the first arrival is assumed to come from the National Earthquake Information Center (NEIC) hypocenter location (140.49 E, 27.83 N, depth of 677 km). Two independent BP processings of the P and pP phases determine the 3D location of the later high frequency (HF) sources based on the differential travel time relative to the hypocenter. The BP of the two phases are first normalized by their peak power within each time frame and are then summed up to combine the resolutions of the two phases. This normalization ensures equal weighting on both phases regardless their amplitude difference which are affected by the attenuation effect. The relative strength of each source is represented by the sum of beamforming power of the P wave and pP wave with equal weight to reflect energy signatures of the earthquake.

Our 3DBP suggests that the source process can be reliably imaged in the first 20s of the 2015 Bonin earthquake. The 3DBP spatio-temporal histories of the HF radiation of the Mw 7.9 mainshock imaged with the AU array, EU array and NA array are presented in Figure. 5.2, S5.3 and S5.4, respectively. We consider the 3DBP result using AU array to be more confident, because take-off angles of seismic rays from EU array are close to the nodal planes (Figure. S5.5a, S5.5b), and the radiation pattern further indicates that the amplitudes (hence the signal to noise ratio) of both P and pP phases of AU are significantly larger than both EU and NA. (Figure S5.5c). The size of the HF rupture is on the order of 40 km horizontally and 20 km in depth, rather compact compared to shallow earthquakes of similar magnitudes. The speed of the HF migration was overall slow, at approximately 2 km/s. The earthquake comprised two stages according to its rupture directions. In the first stage, the rupture began with a near-horizontal propagation in the first 10 s, following a linear path bilaterally in the NE-SW direction. This pattern is clearly imaged by the AU array but is less prominent in the images produced with the EU and NA arrays (Figure. 5.2, S5.3, S5.4). In the second stage, a strong NW propagation took over and produced

the dominant directivity (Figure. 5.2). The focal depth at this stage becomes shallower as the rupture propagate northwestward (Figure. 5.2). Such vertical propagation is generally consistent with the shallow dip angle 25° of sub-horizontal plane inferred by GCMT (Figure. 5.2b). This two-step source propagation is consistent with the locations inverted using multiple-source inversion (stars in Figure. 5.2, introduced in section 5.3).

One concern of the 3DBP is whether the trade-off between horizontal locations and depths result in large spatial uncertainty. These error bounds at each time step can be determined by bootstrapping the back-projections with random noise added to the signal. The noise level can be estimated based on the incoherent part of the waveforms. First, a reference point-source location in each snapshot is defined by the peak of the back-projection image. At each time step, seismograms are then aligned and stacked with the time shift predicted by the reference location. The noise at each station is computed as the waveform residual with respect to the array stacked waveform. The noise is shuffled by randomizing its Fourier phase spectrum, then added back to the original waveform. Note that the time shift is designed to estimates the incoherent noise and is not applied to the original waveform. One hundred synthetic realizations of the stacked array recordings plus noise are then back-projected at each time step. The ellipsoids in Fig. 2 show the 95% confidence bound on the true peak location. Another source of the location uncertainty comes from the variation of the path effects. BP relies on the hypocenter time-correction to calibrate the travel time error due to 3D Earth structure. If there are substantial structural variations near the source domain, the hypocenter time-correction may be less valid to be applied to a distant subsource away from the hypocenter. Detailed methodology is documented in Section 2.3.3. The resolution test indicates that the difference between the BP and catalog location of M5.5 event is on the order of 15 km, considerably smaller than the source dimension of the mainshock. The location uncertainty due to path effect grows if the subsource is further away from the hypocenter. Since the spatial separation from the two earthquakes are larger than the source dimension of the Bonin earthquake, we expect 15 km as the upper bound of the location uncertainty due to path effect.

To further understand the performances of our 3DBP resolving complicated kinematic sources, we also conduct earthquake scenarios involving bilateral ruptures forming L-shape branches. This scenario is inspired by subevent locations imaged by multiple-source inversions (section 5.3). We use the waveforms of a M 6.5 deep earthquake (2015-06-23 12:18:30 UTC) that occurred in the Bonin area as empirical Greens functions. We adopt the seismograms recorded by the NA array which have high signal-to-noise ratio and adequate coherence. We created two rupture scenarios comprising 3 moving sources with subevent time interval of 5 s, and 5 moving sources with interval of 2.5 s, to represent discrete and more continuous ruptures, respectively. We back-projected the synthetic seismograms at two frequency bands, from 0.25 Hz to 1 Hz with a window length of 10 s, the same processing parameters as used in the mainshock, and a higher frequency band of 0.5 to 2 Hz with a window length of 5 s. According to Fig. S5.6, 3DBP is overall capable of resolving the spatio-temporal pattern of the input source. In the case of the continuous sources, 3DBP reasonably reproduces the rupture extension, timing and subevent locations (Fig. S5.6). For the discrete sources, 3DBP almost identically recovers the input source at the frequency range of 0.25 Hz to 1 Hz. The higher-frequency (0.5 - 2 Hz) result is slight noisier and shows more scatters, possibly due to the smaller waveform coherency. Note that the BPs of the discrete sources at both frequency bands produces minor connections between the first and second subevents. This is because the window length is comparable with or longer than the temporal separations between subevents, so that a single window includes multiple subevents. Overall, the synthetic tests demonstrate that 3DBP is capable of imaging the input kinematic scenarios.

5.3 Multiple-source inversion

To analyze the spatio-temporal distribution and focal mechanisms of the energy packets of the seismograms, we adopt a multiple-source inversion method to invert the direct and near-surface reflected waveforms (Y. Chen et al., 2014).

The earthquake begins with strong sub-events during the first ~ 15 s, followed by a weak

energy release lasting ~ 5 s. The MSI result exhibits that the first sub-event is to the northeast of the initiation point (the azimuth with the earliest arrival of the P waves). The second and third sub-events are to the south and the southwest, respectively. The fourth and fifth sub-events propagate to the northwest. The inferred best-fitting sub-events are shown in Figure. 5.3(a) (red beach balls) and Table S5.1. The MSI results in reasonably good waveform fitting between synthetics (red traces in Figure. 5.3b) and seismic data (black traces in Figure. 5.3b) for P, pP, SH and sSH waves. The inverted sub-events align closed to the GCMT sub-horizontal fault plane as a flipped T shape (stars in Figure 5.2, red dots in Figure 5.3a), well consistent with the AU back-projection result (ellipsoids in Figure 5.2). The first three sub-events propagate bilaterally in the direction NE-SW with moment magnitudes of 7.5, while the fourth and fifth sub-events propagate to northwest near-perpendicular to the first three sub-events, with moment magnitude of 7.5 and 7.1, respectively. The focal mechanisms (strike/dip/rake) of the sub-events are 47/25/329, 40/30/321, 30/29/316, 21/22/310 and 10/14/308, respectively. The summed moment tensor of the inferred sub-events yields an overall focal mechanism of 35/26/321, almost identical to the GCMT solution (37/26/322) for the mainshock. The predicted seismic moment is $7.1 \times 10^{20} Nm$, slightly lower than $7.7 \times 10^{20} Nm$ in the GCMT solution. We also found that the sub-event strike and dip angles decrease as it moves away from the first sub-event.

5.4 Discussion

Our case study of the Bonin earthquake indicates the importance of studying the deep earthquake source process in full 3D space with 3DBP and MSI that include both the direct and depth phases. The depth information is particularly useful to constrain the geometry of deep earthquakes. For example, the 2013 Mw 8.3 Okhotsk earthquake was first considered to be confined on the GCMT sub-horizontal plane (Meng et al., 2014; Wei et al., 2013; Ye et al., 2013). However, the inversion with the direct waves and the near-surface reflected waves exhibited that the earthquake propagated to the southwest with a downdip extension, which is inconsistent with either the GCMT sub-horizontal or sub-vertical planes (Y. Chen et al.,

2014).

MSI exhibits the similar spatio-temporal pattern as 3DBP does. The sub-events do not align on either of the GCMT planes. We note that seismic attenuation and heterogeneity would generate systematical bias of the absolute location; however, our result is based on relative timings, which can minimize the errors from the attenuation and heterogeneity (Y. Chen et al., 2014). MSI can also determine focal mechanisms of the sub-events, which gives rise to better understanding of the source complexity than previous studies (Ye, Lay, Zhan, et al., 2016). In this study, the earthquake can be modeled by five sub-events based on reasonably good fitting of the seismic P, pP, S and sS waveforms. The MSI reveals focal-mechanism variations of the sub-events, indicating that the earthquake cannot be modeled by a single planar rupture. Our inverted source process is generally consistent with Ye, Lay, Zhan, et al. (2016), but reveals information that is more detailed. We compare the MSI results of the Okhotsk earthquake and the Bonin earthquake. The Okhotsk earthquake has a relatively larger magnitude, and a significant depth extension. The focal mechanisms of those sub-events are inconsistent with their spatial distribution; therefore, the Okhotsk earthquake cannot be explained by one or two ruptures (Y. Chen et al., 2014). This Bonin earthquake has a relatively smaller magnitude. The spatio-temporal distribution and focal mechanisms of the sub-events vary slightly, but are still close to the GCMT sub-horizontal fault plane. Such spatially variable focal mechanisms within one earthquake may be explained by continuous rupture along a single curved fault interface or sub-events crossing multiple sub-parallel fault planes (Cai & Wiens, 2016; Y. Chen & Wen, 2015; Poli et al., 2016). In the former case, focal mechanisms varying along the location are consistent with a curved rupture process. The curvature required for the single fault plane is about $1 \times 10^{-5} m^{-1}$, which is not abnormal for fault surface traces (Bergbauer, 2004; Mynatt et al., 2007), but whether such curvature exist for deep-focus rupture remains to be verified. In the latter case, multiple sub-parallel faults with moderately varying strikes could be connected in a single macroscopic rupture with step over on rupture propagation. The required step-over distance between the sub-faults are in the order of 10 km, which is a comfortable step-over distance given a seismogenic width of 50

km (Bai et al., 2017). Multiple-parallel faults can also be explained by sub-events triggering each other (Y. Chen et al., 2014; Y. Chen & Wen, 2015; Poli & Prieto, 2016).

Our investigation indicates that the inverted sub-event focal mechanisms are generally consistent with the GCMT solution. The focal mechanisms vary along the location, which indicates strong heterogeneity in the slab, either stress heterogeneity (Yang et al., 2017) or varying pre-existing weak zones (Y. Chen & Wen, 2015). The slab heterogeneity may be caused by slab stagnation, buckling, pileup and thickening near 660 km (Obayashi et al., 2017; Porritt & Yoshioka, 2016; Yang et al., 2017). Those processes can generate stress heterogeneities (Myhill, 2012; Poli et al., 2016), and/or distort the preexisting weak zones to complex geometries, giving rise to sub-events of various focal mechanisms (Ruiz et al., 2017; Zahradnik et al., 2017). Such stress can trigger either phase transformation (Y. Wang et al., 2017) or shear thermal instability (Ogawa, 1987). The step-over distance is a critical factor for the step-over model of sub-events crossing multiple fault interfaces. For shallow earthquakes, the seismogenic depth is limited by the thickness of the brittle crust (~ 10 km). For deep earthquakes, if the thermal shear instability is the driving mechanism, the seismogenic width is predicted to be 40–60 km (Karato et al., 2001), which is much larger than the metastable olivine wedge.

5.5 Conclusion

We develop a novel back projection technique that allows source imaging in full three-dimensional space with a high-depth resolution. Our results indicate an initial SW-NE bilateral source propagation followed by a northwest source extension. The multiple-source inversion reveals a two-step source process with propagating directions nearly perpendicular to each other, consistent with the 3-D back projection result. The spatial distribution and focal mechanisms of the subevents cannot be modeled by a single planar rupture, which may display a curved rupture plane or subevents crossing multiple fault interfaces. The complex source process can be best explained by stress or structure heterogeneity within the deep slab.

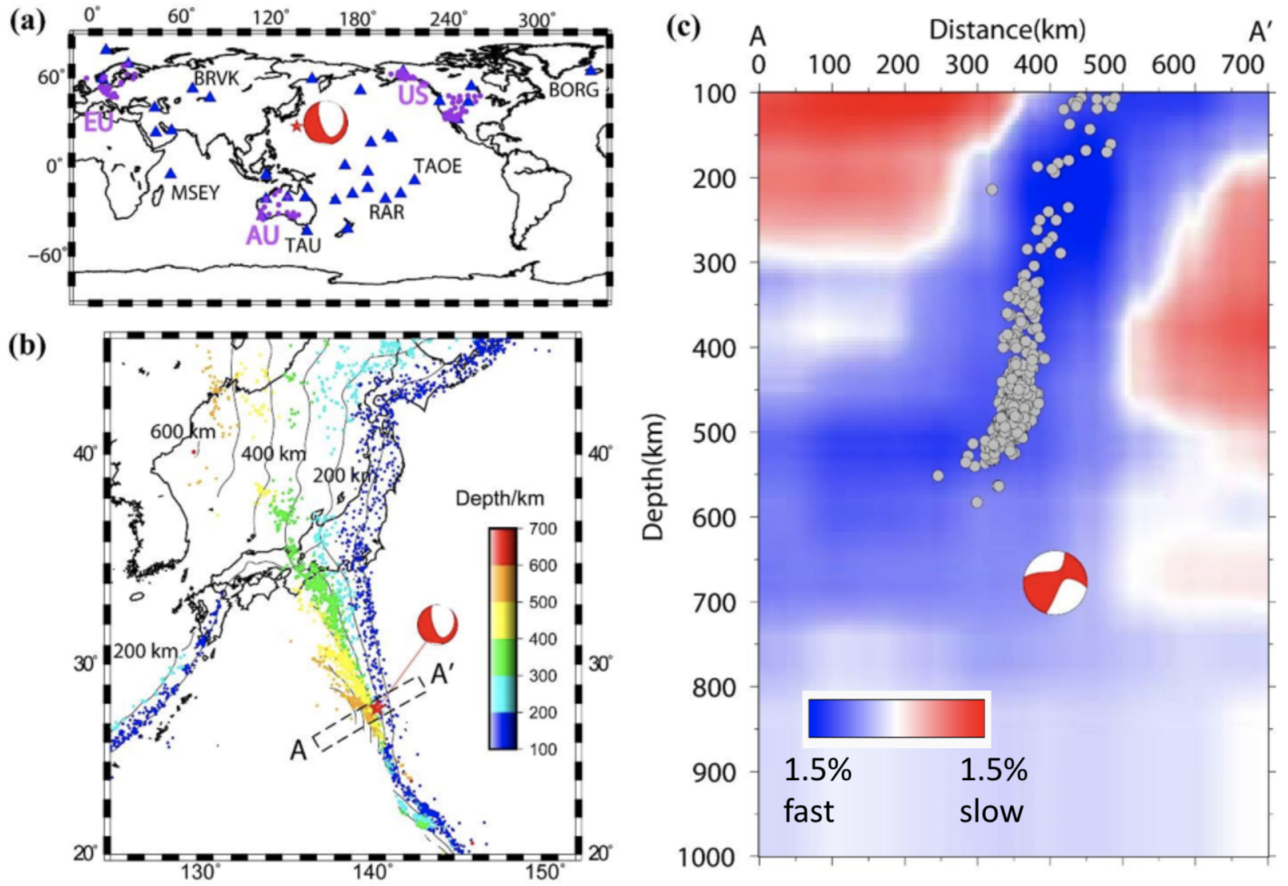


Figure 5.1: Tectonic settings for the source area. (a) Location of the 30 May 2015 earthquake (red star) and seismic station coverage. Purple circles represent seismic arrays used in 3D back-projection. Blue triangles represent seismic stations used in multiple-source inversion. Red beach ball describes the GCMT focal mechanism of the earthquake. (b) Map view of the seismicity at the adjacent region. The circles represent the seismicity, color-coded by depth. The depth contours labeled with slab depth are from the USGS slab 1.0 model (Hayes et al., 2012). The red star represents the location of the 2015 Bonin earthquake. (c) Cross-section view of the seismicity and tomography model (Fukao & Obayashi, 2013) along the rectangle AA' in (b). Red beach ball represents the mainshock focal mechanism projected in side view.

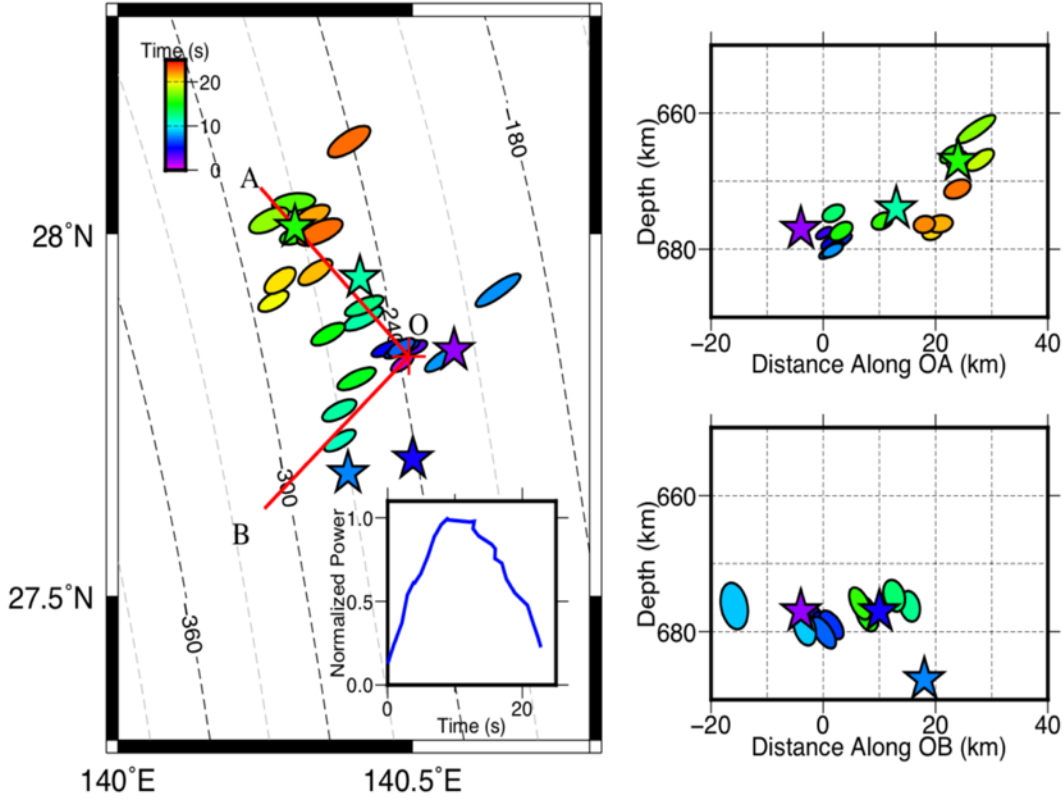


Figure 5.2: 3DBP image using Australian (AU) network for the 2015 Mw 7.9 Bonin earthquake. (a) BP locations in map view. (b) Cross-section of BP images along OA in (a). (c) Cross-section of BP images along OB in (a). The colors and shape of the ellipsoids demonstrate the 3DBP timing (with a interval of 1 s) and the 95% confidence interval on the peak locations. The red cross indicates the hypocenter location. The colors and shapes of the stars show the MSI inferred timing and location in Table S5.1 and Figure 5.3 for comparison.

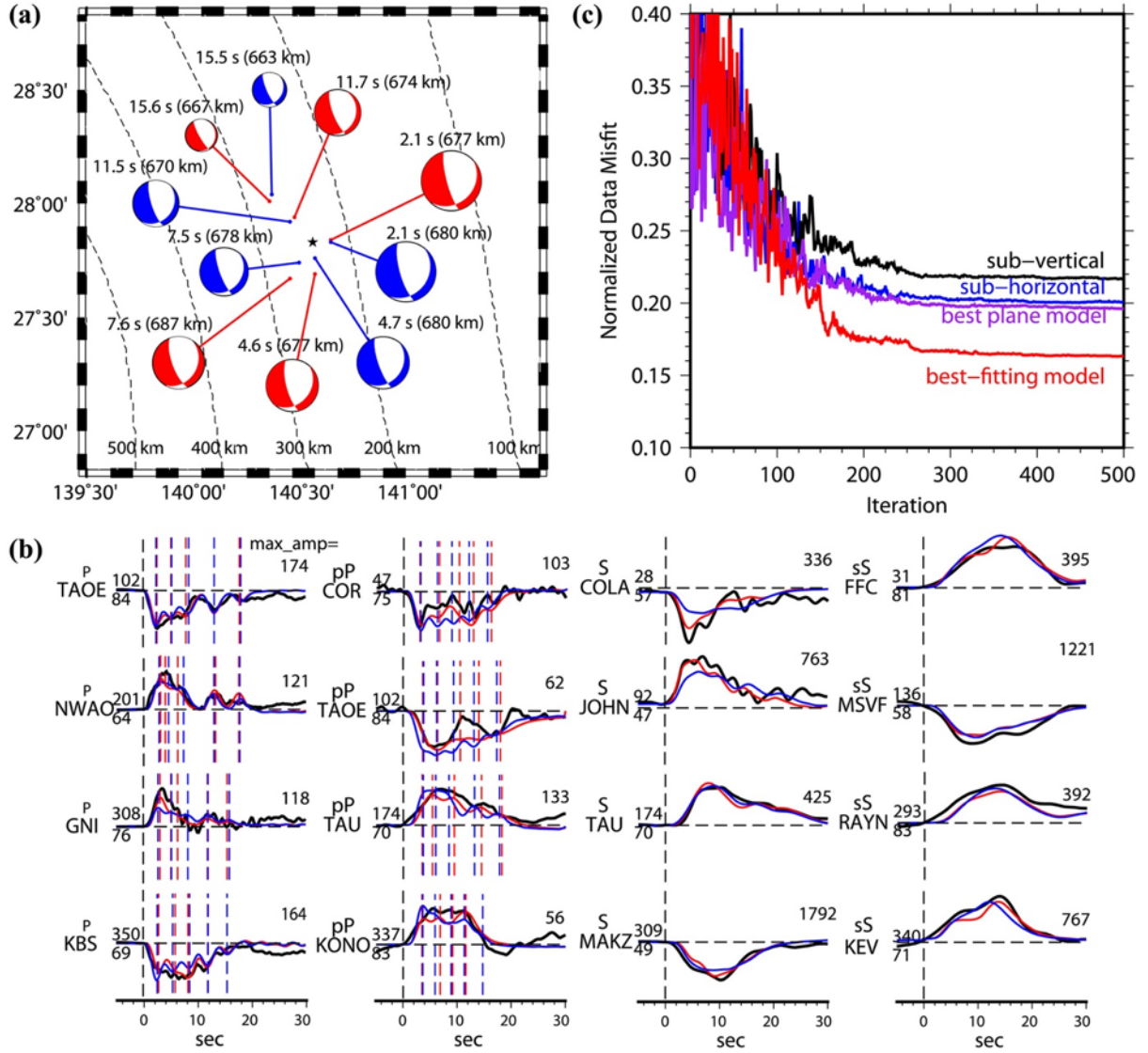


Figure 5.3: Multiple source inversions of the Bonin earthquake. (a) Locations of the best-fitting sub-event model (red points) and the sub-event model compelled onto the sub-horizontal plane (blue points), along with slab depth contours (black dashed traces). Red and blue beach balls represent inferred focal mechanisms of the sub-events of the best-fitting sub-event model (Table S5.1) and the best planar model (Table S5.2), labeled with the sub-event peak times/moment magnitudes/depths. (b) Selected seismograms (black traces) in four quadrants comparing with synthetics predicted based on the best-fitting sub-event model (red traces) and synthetics based on the sub-horizontal model (blue traces). Seismic records are labeled with station name, phase, along with azimuth, epicentral distance, and the maximum amplitude of the data. (c) The normalized data misfits of 500 iterations for the best-fitting (red), the sub-horizontal (blue), and the sub-vertical (black) models.

5.6 Supplementary material

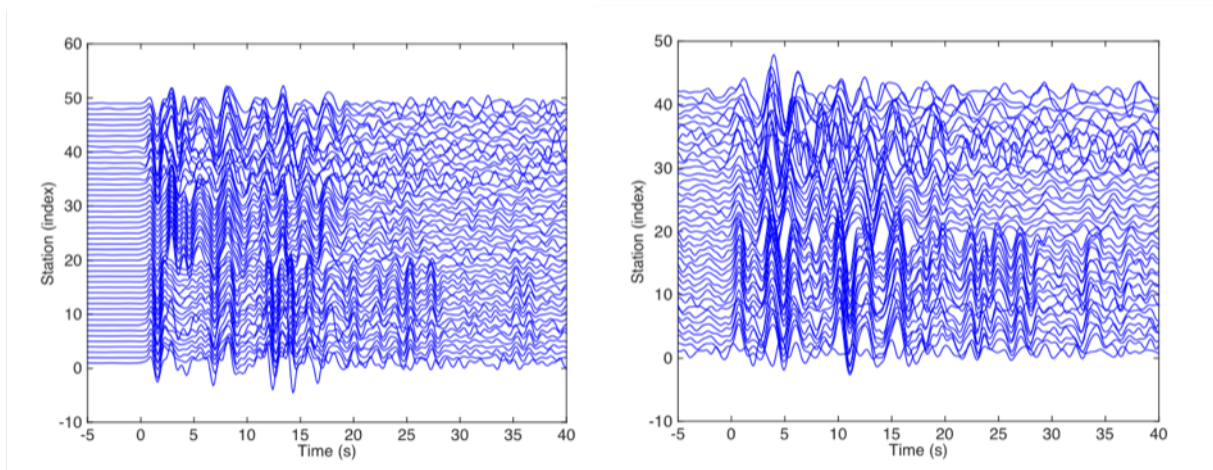


Figure S5.1: Mainshock waveform recorded at Australian array filtered within 0.25 1 Hz. (Left panel: P, right panel: pP) SNR for P: 43 dB. SNR for pP: 15 dB.

Sub-event	dt (s)	dn (km)	de (km)	dz (km)	duration (s)	moment (Nm)	strike (°)	dip	slip (°)
1	2.1	0.7	7.7	0.0	3.6	2.0E+20	46.5	25.1	329.1
2	4.6	-16.0	1.3	0.0	3.8	1.4E+20	40.0	30.3	320.5
3	7.6	-18.0	-10.1	10.0	4.4	1.7E+20	30.1	29.3	316.1
4	11.7	11.7	-8.2	-3.0	4.0	1.4E+20	21.2	22.3	310.2
5	15.6	20.0	-18.2	-10.0	4.0	6.0E+19	10.0	13.8	312.0

Table S5.1: Source parameters of the best-fitting five sub-events of the Bonin Earthquake inferred by multiple-source inversion, dt presents occurring time, dn, de, and dz represent event location in north, east and vertical direction, duration represents seismic duration, strike, dip and slip represent strike, dip and slip angles for focal mechanism.

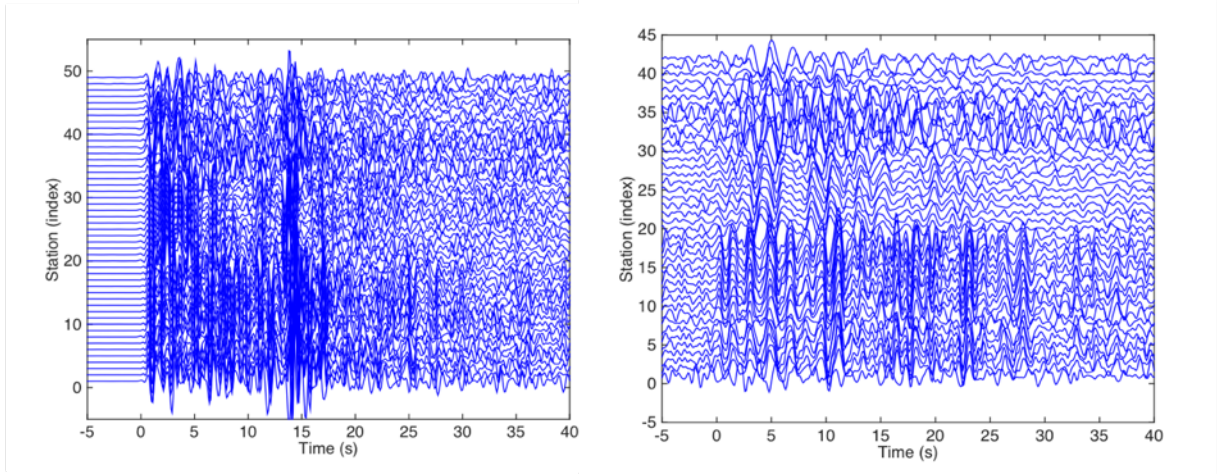


Figure S5.2: Mainshock waveform recorded at Australian array filtered within 0.5 2 Hz. (Left panel: P, right panel: pP) SNR for P:37 dB. SNR for pP: 7 dB.

Sub-event	dt (s)	dn (km)	de (km)	dz (km)	duration (s)	moment (Nm)	strike (°)	dip	slip (°)
1	2.1	0.0	7.8	3.0	3.1	2.0E+20	37.0	26.0	320.0
2	4.7	-7.4	0.5	2.4	4.4	1.6E+20	37.0	26.0	323.0
3	7.5	-10.4	-6.1	0.7	4.4	1.4E+20	37.0	26.0	324.2
4	11.5	10.3	-9.8	-6.9	4.0	1.3E+20	37.0	26.0	322.1
5	15.5	24.1	-17.4	-13.9	4.0	7.0E+19	37.0	26.0	328.0

Table S5.2: Source parameters of the sub-event model with the five sub-events compelled onto the GCMT sub-horizontal plane.

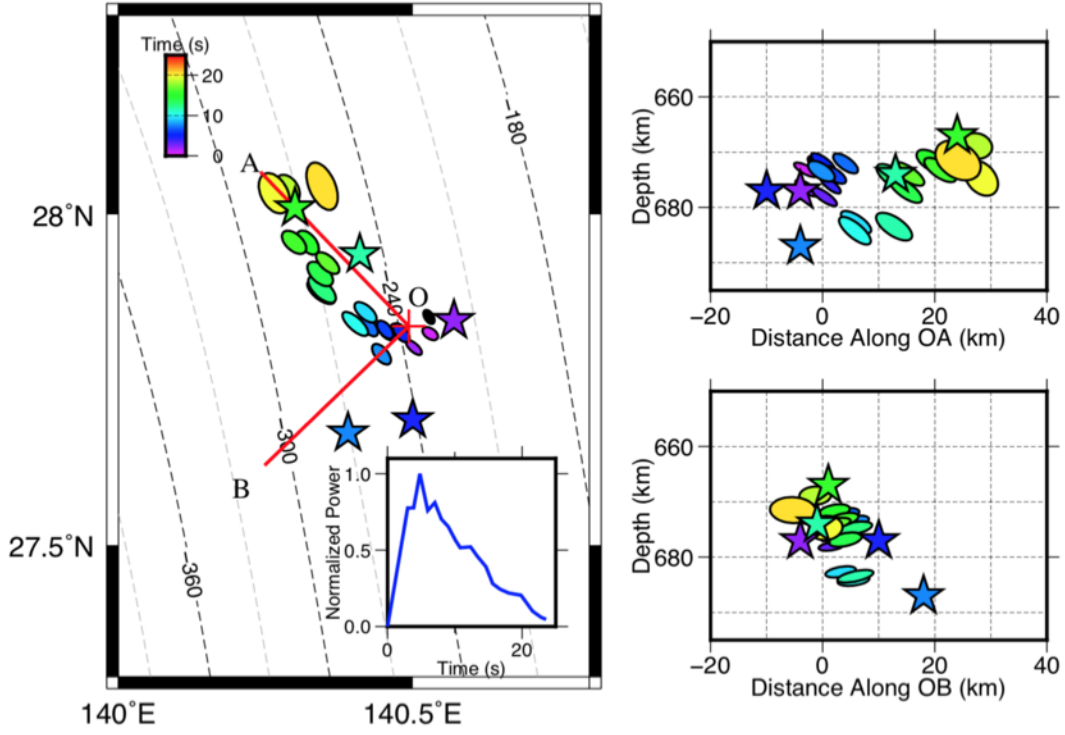


Figure S5.3: 3DBP image using European (EU) network for the 2015 Mw 7.9 Bonin earthquake. (a) BP locations in map view. (b) Cross-section of BP images along OA in (a). (c) Cross-section of BP images along OB in (a). The colors and shape of the ellipsoids demonstrate the timing and the 95% confidence interval on the peak locations. The colors and locations of the stars show the MSI inferred times and locations in Table S5.1.

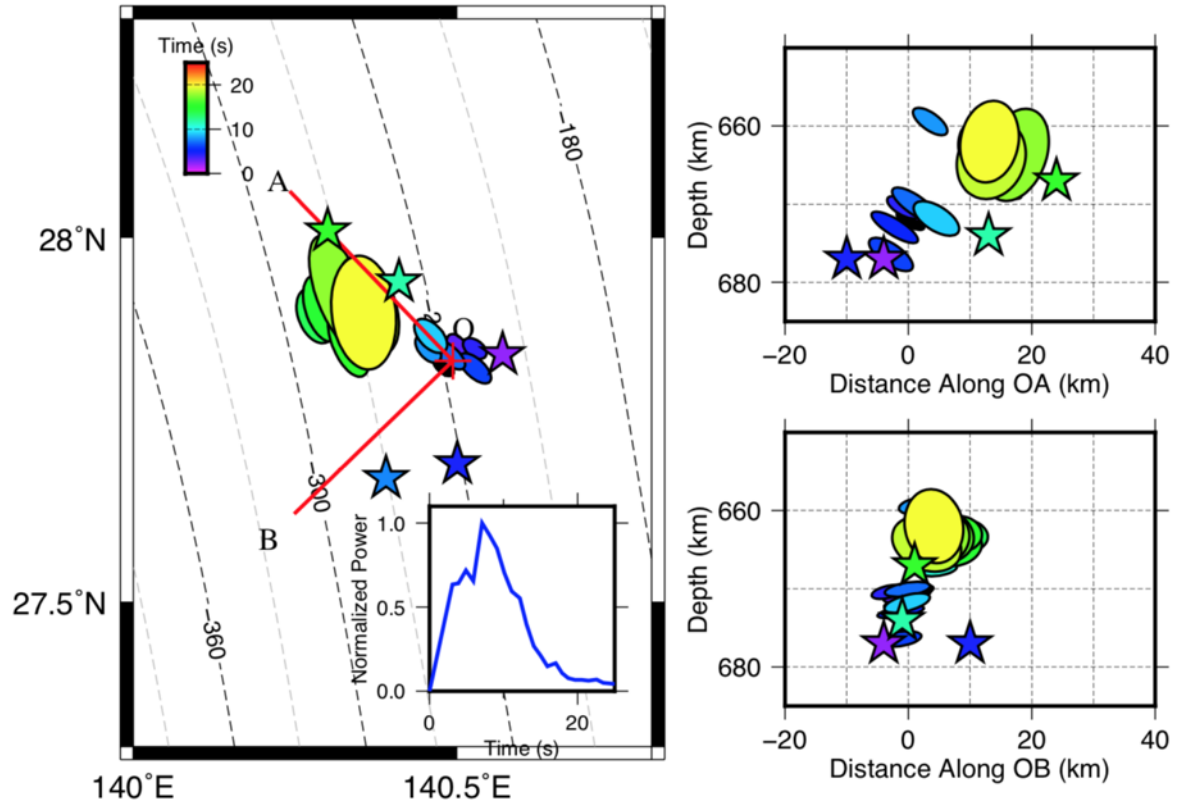


Figure S5.4: 3DBP image using North American (NA) network for the 2015 Mw 7.9 Bonin earthquake. (a) BP locations in map view. (b) Cross-section of BP images along OA in (a). (c) Cross-section of BP images along OB in (a). The colors and shape of the ellipsoids demonstrate the timing and the 95% confidence interval on the peak locations. The colors and locations of the stars show the MSI inferred times and locations in Table S5.1.

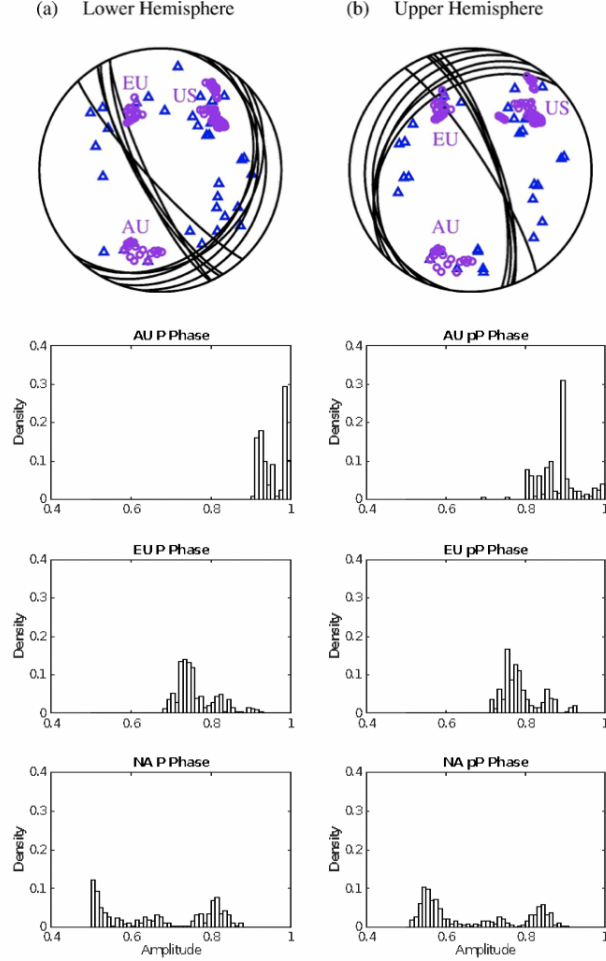


Figure S5.5: Take-off angles for the seismic ray from the AU, EU, and US arrays and the GSN stations shown on (a) upper hemisphere for P and SH waves and (b) lower hemisphere for pP and sSH waves. The purple circles represent take-off angles for (a) P and (b) pP waves for the stations of the three arrays, and the blue triangles are represents take-off angles for (a) P/SH and (b) pP/sSH waves for the GSN stations.(c) Density histogram of amplitude of P and pP phases for all networks.

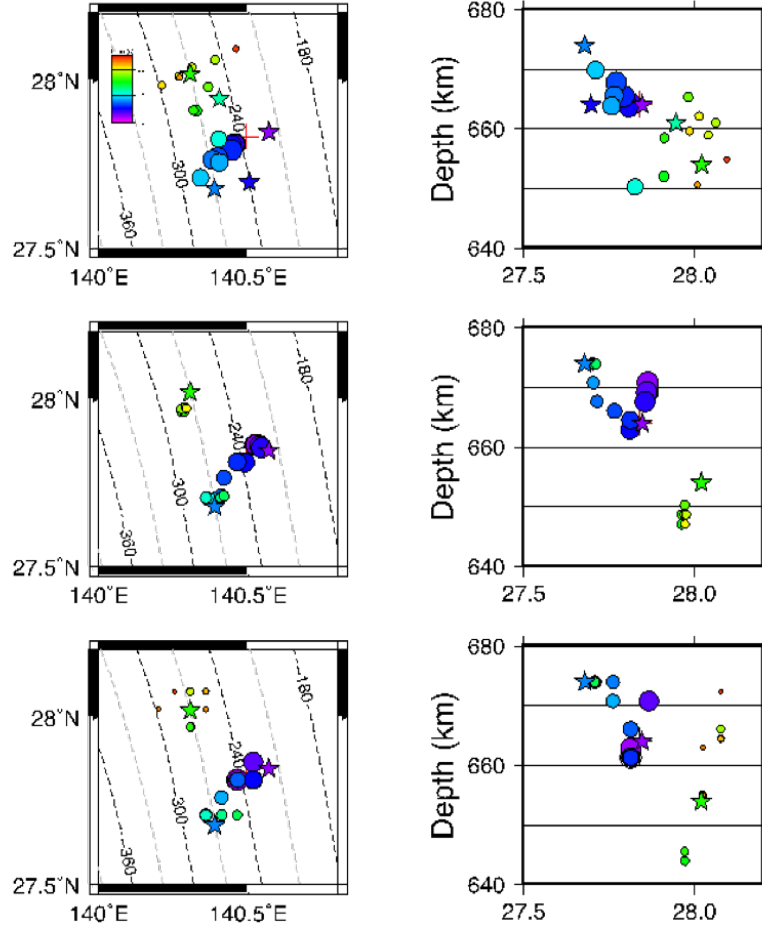


Figure S5.6: Synthetic test BP resolvability of continuous and discontinuous ruptures. (a) Continuous ruptures with 5 sources as listed in Table S5.1. Stars are input sources, cross is the hypocenter location and dots are BP radiators processed in 0.25 - 1 Hz. (Time scale is the same as Figure S5.3 S5.4) (b). Discontinuous ruptures with 3 separate sources (stars). The time separation is 5 s in the case and dots are resolved BP radiators in 0.25 - 1 Hz. (c) Same input setting as (b), but the BP is performed in 0.5 - 2 Hz.

CHAPTER 6

Rupture Complexity and Variability of the 2018 Tonga-Fiji Doublet

Abstract

In this study, 3D BP is performed to study the rupture process of the 2018 Tonga doublet. The two events display different faulting patterns and favor different mechanisms. The Mw 8.2 is a fast rupture in an extensive region, but the Mw 7.9 ruptures slowly in a compact area. Both of the events are closely related to local thermal conditions, and the rupture pattern is thermally controlled.

This is an independent work by myself. The manuscript is in preparation. We will collaborate with Yu Chen on MSI in the paper.

6.1 Introduction

Mechanisms of deep earthquakes are not well understood due to limited observations. There are a few competing mechanisms that are currently under debate, for example: 1) dehydration embrittlement ([Raleigh & Paterson, 1965](#); [Meade & Jeanloz, 1991](#)) 2) phase transformation ([Kirby, 1987](#); [Green II & Burnley, 1989](#); [Kirby et al., 1991](#)) and 3) thermal shear instability ([Ogawa, 1987](#); [Hobbs & Ord, 1988](#); [Karato et al., 2001](#)). In order to distinguish these proposed mechanisms, accurate and detailed descriptions of the kinematic process becomes crucial.

On 00:19:40 (UTC) of 19 August 2018, a magnitude 8.2 earthquake struck the Tonga-Fiji

islands. The nearest residential region is ~ 450 km east of the epicenter, and the earthquake is too deep to generate severe ground shaking or damages. 18 days later, another magnitude 7.9 earthquake occurred ~ 300 km to the west of the Mw 8.2. As shown in Figure 6.1, both of the events occurred within the Tonga subduction zone. The Tonga subduction is a convergent boundary where the Pacific plate is subducting beneath the Australian plate at a speed of 10- 20 cm/yr (Bevis et al., 1995). Tonga subduction zone hosts the majority of deep earthquakes worldwide. The two large events create a unique opportunity for investigating rupture mechanisms of deep earthquakes

In the first place, their proximities in space and time suggests a causal correlation, while their background seismicity and aftershock sequences display significantly different patterns. The Mw 8.2 source area has dense background seismicity and ~ 281 aftershocks are generated within 1 month. In contrast, the Mw 7.9 occurs in a previously aseismic region with few documented events and fewer aftershocks (~ 6) are generated. Detailed source analysis becomes essential to illuminate the rupture complexities in this area. Aftershock relocation, directivity analysis and conventional 2d backprojection are performed to study the Tonga-Fiji doublet (Fan et al., 2019). It suggests that the doublet are two end members of deep earthquakes: the Mw 8.2 has a fast rupture speed and abundant aftershock activities. It is likely to be thermally controlled, whereas the Mw 7.9 is slow and absent of aftershock sequences, favoring other mechanisms such as shear instability.

In the study of Fan et al. (2019), 2D backprojection and directivity analysis assume a horizontal plane, thus there is a deficiency in depth resolution. For the doublet, rupture dimension in vertical direction is inferred from aftershocks distributions. However, such inference is subjected to large ambiguities. In this study, we seek to provide direct constraint of kinematic rupture process in 3D space. The doublet has been well recorded by multiple seismic arrays: Western US, Alaska, and Hinet, which allows us to perform a kinematic source imaging using 3D backprojection (3DBP). 3DBP utilizes downward propagating P phase and upward propagating pP phase to jointly constrain the depth of the seismic source locations. 3D BPs of the doublet confirm distinct rupture patterns, and provide insight of

rupture mechanisms and slab deformation.

6.2 Data and Method

6.2.1 Two Dimensional Teleseismic P wave back-projection

2D teleseismic P-wave back-projection (BP) is a widely applied technique which extracts waveform coherence to resolve the rupture process of large earthquakes (Ishii et al., 2005; D. Wang & Mori, 2011; Koper et al., 2011; Zhang et al., 2012; Yagi & Okuwaki, 2015; Ruppert et al., 2018). BP doesn't require assumptions of rupture speed, priori fault geometries or Greens function, thus ensures its robustness in estimating source parameters (e.g. rupture lengths, rupture speeds and subevents) (Yao et al., 2011; Kiser & Ishii, 2012; Fan & Shearer, 2015).

In this study, seismic array data is collected from all accessible large regional arrays in North America and Japan (Hinet) that are located at epicentral distances between 60 to 90°. We split the stations in North America into 2 arrays: Alaska and Western US, because each individual group contains better waveform similarity within the group. The data is processed independently for every array. The P waves are empirically aligned by initial P arrivals for the first 10s window (Ishii et al., 2005). The alignment is first performed in low frequency band of 0.05 - 0.5Hz and then in high-frequency band of 0.5 - 2 Hz to achieve higher timing accuracy. The 2D BP is performed at a constant depth equivalent to the hypocentral depth. Therefore we resolve only the lateral variations of the BP radiators at different frequencies (0.5 - 2 Hz, 0.25 - 1 Hz and 0.05 - 0.5 Hz). We adopt the Multitaper-MUSIC array processing technique proposed by (Meng et al., 2011), which improves spatial resolution for closely spaced sources. P wave trains are then back-projected onto the 2D source plane according to the travel time table derived with the 1D reference velocity model. We also check the network quality within every network by visually examining the waveform coherence and the calculated relative amplitude (Figure S6.1, S6.2) based on the source radiation pattern. For the M7.9 earthquake, the polarity of the Alaska Array flips when the

stations cross the nodal plane. This effect introduces additional artifacts (Fan & Shearer, 2015). Therefore, we consider the BPs of the M7.9 event of the Hinet and the Western US arrays are more trustworthy than that of the Alaska array.

6.2.2 3-D Back Projection (3D BP)

While 2D P-wave backprojection provides horizontal constraint of the radiators, it doesn't provide satisfactory depth resolution as there exists significant trade-off between depth and time. This is primarily because the ray parameter of P wave is not sensitive to depth changes at teleseismic distances (Kiser & Ishii, 2017).

We can nevertheless combine an additional seismic wave phase (pP) in the back projection. Velocity structures tend to be less heterogeneous at depth, and deep earthquakes generally produce more coherent waveforms than their shallow counterparts. As depth increases, the pP can be separated from other phases due to a lag in arrival time. pP becomes visually identifiable as P-wave code is less significant. The Aforementioned benefits allow us to combine P and pP phases. 3D BP significantly mitigates the trade-offs of individual ray paths, enhancing the depth resolution of conventional BPs (Kiser & Ishii, 2011; Y. Chen et al., 2018). We closely follow the 3D BP method described in Y. Chen et al. (2018), with the key parameters demonstrated as follows.

Similar to processing P waves, pP phase is pre-aligned by theoretical arrivals and empirically corrected by cross-correlating the first 10 s window at 2 frequency bands: 0.05 - 0.5 Hz and 0.5 - 1 Hz. Both P and pP amplitudes are normalized by their initial arrivals to remove the potential path or site effects which could bias the results. We set up a new 3D grid search space (latitude, longitude and depth) around the hypocenter and fix the absolute location of the hypocenter according to ISC catalog. P and pP phases are first independently back-projected in the 3D space at 0.25 - 1 Hz. This is the highest frequency limit set by pP due to the attenuation at near-surface structure along its ray path (Kiser & Ishii, 2017). Beyond 1Hz, waveforms appear to be less coherent due to the velocity heterogeneity. Normalized BP power of the two phases are summed up to represent the power of the 3D BP.

As discussed in [Y. Chen et al. \(2018\)](#), the take-off angles of both p and pP can not be too close the nodal plane. In other words, 3D BP imaging is more reliable for arrays that receive large amplitudes of both P and pP phases. Based on these criteria, western US becomes the ideal candidate to image the Mw 8.2 earthquake, and Hinet for the Mw 7.9 event.

6.3 Results

6.3.1 Mw 8.2 Fiji earthquake

2D BP images for different arrays at different frequency bands are shown in Figure 6.2. For the Mw 8.2 earthquake, all of the three arrays provide reliable BP images and they are consistent in the mapview. We are able to image the first 30 s of the rupture, where coherent waveform ensures reliable distribution of BP radiators. According to Figure 6.2 (a) and 6.2 (b), at 0.5 - 2 Hz and 0.25 - 1Hz, the rupture propagates northward from the hypocenter and follows the slabs contour. HF radiators rotate approximately 45 degrees in azimuth, which is consistent with the curvature of the slab. At lower frequency (0.05 - 0.5 Hz), the radiators appear to be more scattered (Figure 6.2 c). Overall, we don't observe significant frequency dependent behaviour except more scatteredness at the lowest frequency band.

The 2D BP imaging set constraints of horizontal rupture extent and duration, nevertheless we can further investigate the rupture process with 3D BP. We interpret the result of the kinematic rupture process of the Mw 8.2 with the western US Array.

For the Mw 8.2 event, we are able to capture the first 30s of the rupture. From Figure 6.3, The rupture front suggests a unilateral northward rupture and gradually migrates to shallower depth. There is a gap between radiators before and after 25s. HF radiation after 25s is likely to be dynamically triggered. The horizontal mapview of the 3D BP radiators is consistent with 2D BP, with a rupture length of 70 km. Vertical rupture dimension is around 60 km. The HF radiators can be well constrained by a plane with a strike angle of 13 degree and dip angle of 80 degree, favoring the subvertical nodal plane according to USGS (Strike 13 degree, dip 70 degree). The fault plane suggests a trenchward dipping geometry, which

is consistent with a downdip-compressional stress field inside the slab (Warren et al., 2007). The HF radiators are migrating at a speed of 3.3 km/s.

6.3.2 Mw 7.9 Suva earthquake

For the M7.9 earthquake, we present the HF radiators in Figure 6.2. (d) -(f). We also image the first 30 s with robust BP images. We decide not to interpret Alaska array because it encounters polarity flipping. The other two arrays, Western US and Hinet, produce consistent rupture process: At 0.5- 2 Hz band, the quake is propagating NE following the geometry of the slab contour in the first ~ 15 s. After 15s, the rupture front is propagating to the north, deviating from 600 km slab contour. At lower frequency, radiators become more scattered, but the overall pattern is consistent. Both of the doublets display change to azimuth during rupture propagation, but the Mw 8.2 follows the slab contour more closely, whereas the Mw 7.9 deviates the contour at later stage.

3D BP from Hinet is our preferred result for the Mw 7.9 event. For the Mw 7.9 event, 3D BP is capable of imaging the first 20 s of the rupture. Beyond 20 s, pP loses coherence which fails to construct phase interference of the subsequent rupture, and the BP radiators become too scattered to be realistic. The rupture initiates along the slab contour and propagated unilaterally in the first 15s. In vertical axis, radiators tend to move to shallower depths. After 15s, the rupture rotated to near perpendicular direction trending NS, and the radiators migrates to deeper part. It is quite obvious that rupture mechanism before and after 15 s are distinct. Change in strike of rupture also provides evidence of large non-DC component. The horizontal size of the 3D BP rupture is about 25 km, and the vertical size is 20 km. We also fit a plane to accommodate the radiators released from the first 15 s of the rupture. Here, we only focus on the first segment since we have shown the latter part doesn't necessarily conform with the same plane. The optimal plane has a strike angle of 274 degree and a dip angle of 42 degree, which yields gross consistency with NP1 (strike: 305 degree, dip: 57 degree) of the USGS moment tensor. The overall rupture speed inferred from BP radiators is ~ 2.4 km/s.

6.4 Discussion

6.4.1 Source processes of the Mw 8.2 and Mw 7.9 earthquakes

In this study, we perform multi-array BPs to ensure robustness and constrain uncertainties. The 2D BPs from different networks are generally consistent, allowing us to constrain the source duration and horizontal rupture dimensions. In order to evaluate the results of 3D BPs, we adopt the strategy as discussed in [Y. Chen et al. \(2018\)](#). The key of a successful 3D BP lies in prominent P and pP amplitudes. We compute the expected relative amplitudes for P and pP for each array and visually inspect the waveforms. According to Figure S6.1 and S6.2, we choose to interpret 3D BP of western US array for the Mw 8.2 and the 3D BP of Hinet for the Mw 7.9.

3D BP provides direct depth constraints of the rupture process, which enables more accurate measurements of source properties than indirect inference ([Fan et al., 2019](#)). Due to lack of depth resolution, indirect inference from aftershocks and other benchmarks could be subject to large ambiguity. According to our observation, the rupture speed of the Mw 8.2 is 3.3 km/s, significantly faster than that of the Mw 7.9 (2.4 km/s). According to 2D BP, the rupture duration of the Mw 7.9 is comparable to the Mw 8.2, but the rupture dimension of Mw 7.9 is smaller. We estimate a rupture length of ~ 100 km for the Mw 8.2 within an area of 80 km (Horizontal) $\times$ 60 km (Vertical). The radiators of the Mw 7.9 can be confined in an area of 40 km (Horizontal) $\times$ 20 km (Vertical). Our 3D BP reveals similar rupture speed estimation as reported by [Fan et al. \(2019\)](#), but smaller source area. Moreover, the Mw 8.2 produces abundant aftershock activities, exceeding the amount of aftershocks generated by the 1996 Mw 7.4 Tonga earthquake. The source area of the Mw 7.9 is much quieter, with very few aftershocks. Despite of the spatial and temporal proximities for the doublet, BP analysis reveals that they have dissimilar source properties.

The sharp contrasts between the doublet can be associated with temperature of the slab ([Wiens et al., 1994](#)). The slab temperature is a vital factor for aftershock activities ([Wiens & Gilbert, 1996](#); [Zhang et al., 2017](#)): colder slabs can host more aftershocks and warmer slabs

inhibit aftershock sequences. The temperature condition also influences the rupture process of large deep earthquakes: fast rupture and large rupture dimension are more prevalent for earthquakes initiated in the cold core of the slab, whereas warmer slabs produce earthquakes with slower rupture speeds and smaller sizes (Wiens, 2001). Our source observations indicate that the Mw 8.2 nucleates in the cold core of the slab, and the Mw 7.9 occurs in a warmer region.

6.4.2 Non-double-couple component of Mw 7.9 earthquake

There is sufficient evidence to support that Mw 7.9 is ruptured on multiple fault planes. USGS moment tensor solution suggests that the Mw 7.9 has a very significant non-DC component (64%) as opposed to 11% for the Mw 8.2. Significant CLVD is more common for deep earthquakes than shallow ones (Kuge & Kawakatsu, 1993). Ruptures on non-parallel fault planes and varying slip directions are used to explain the large CLVD component (Green & Houston, 1995; Frohlich, 2006). Both 2D and 3D BPs illustrate the change of strikes, suggesting participation of multiple fault segments. Moreover, the Mw 7.9 earthquake is isolated from background seismicity. Regardless the physical mechanisms (Phase transformation or thermal instability), since the seismogenic zone is likely to be thermally controlled and bounded by the cold core of subducting slabs, it becomes narrower as depth increases (Houston, 1993). The narrow seismogenic zone reduces the probability of creating a single large and continuous fault plane. Therefore, earthquakes are more probable to rupture on several small non-parallel planes and lead to large non-DC component. To accommodate the regional stress field, different planes appear to share a common compressional axis (Houston, 1993; Frohlich, 2006).

6.4.3 Relationship between the deep-focus earthquakes and slab geometry

The rupture processes of the two earthquakes reflects the slab geometries in the deep mantle. Specifically, the rupture path of the M8.2 event rotates along with the slab curvature while the M7.9 event initially follows the slab strike before it eventually propagate in the updip

direction. our findings indicate the slab contours is a proxy of the temperature gradient that controls and confines the rupture kinematics.

The Tonga subduction zone hosted quite a few M 7 events, which creates a chance to compare the 2018 doublet with historical events. The 1994 Mw 7.6 Tonga earthquake resembles the Mw 8.2 earthquake. They both have fast rupture speeds and produce active aftershock sequences (McGuire et al., 1997). Therefore, their physical mechanisms could be consistent. The rupture path suggests that metastable olivine wedge plays an important role in the initiation and propagation of the Mw 8.2 event. At termination stages, we observe a spatial gap that separates HF radiators before and after 25 s. This is probably a fault segmentation involving heterogeneous stress distribution, discontinuation of slab geometry and thermal gradient. One of the possibilities is the thermal halo effect (McGuire et al., 1997; Wiens et al., 1994). The thermal halo contains warm materials that inhibit earthquake nucleation, but can potentially sustain ruptures to propagate into this region under a different physical mechanism (Fan et al., 2019).

Meanwhile, the Mw 7.9 shares common features with deep earthquake in South America and Bonin island: 1994 Mw 8.3 Bolivia earthquake, 1998 Mw 7.1 Bonin earthquake, and 2000 Mw 7.3 Bonin earthquake (Tibi, Bock, & Wiens, 2003). They all suggest dissipative rupture with low speed and efficiency. It is probably generated through shear instabilities rather than phase transformation. The thermal shear instability involves feedback between heat and grain size reduction, which results in shear localization (Ogawa, 1987; Kanamori et al., 1998; Karato et al., 2001). Thus, we conclude the doublet represents distinct rupture process and favors different mechanism.

The 2018 Tonga-Fiji doublet shares somewhat similarity with the 2002 doublet (Tibi, Wiens, & Inoue, 2003). Second events in both doublets are reported to occur in regions with sparse background seismicity. The 2002 doublet is separated by 7 minutes and 300 km, and the 2018 doublet is separated by 18 days and 300 km. Considering the similar tectonic setting and source properties, they are supposed to share similar triggering mechanisms. According to Fan et al. (2019), both static and dynamic triggering promote the earthquake

failure. But it remains mysterious why the 2018 doublet has an anomalously long triggering duration.

6.5 Conclusion

We perform comprehensive back-projection analysis to the 2018 Tonga-Fiji doublet. While the Mw 8.2 ruptured an extensive area (>100 km) with a fast rupture speed of 3.3km/s, the Mw 7.9 has a more compact rupture area with slow rupture speed 2.4 km/s that involves multiple fault segments. Such distinctive source properties that can be attributed to different thermal conditions. We observe fast and extensive rupture for the Mw 8.2. It is likely to be initiated through transformational faulting and terminated by the thermal shear instability mechanism. The rupture front tends to follow the strike of the slab contour as well as penetrates to shallower depths. Contradictingly, the Mw 7.9 has a much slower rupture speed and smaller size. Shear instability is used to explain the mechanism of the Mw 7.9. Both of the ruptures favor downdip-compressional stress field, and the faults are newly generated rather than re-activated from pre-existing weak zones. The stress field in the source area can be heterogeneous.

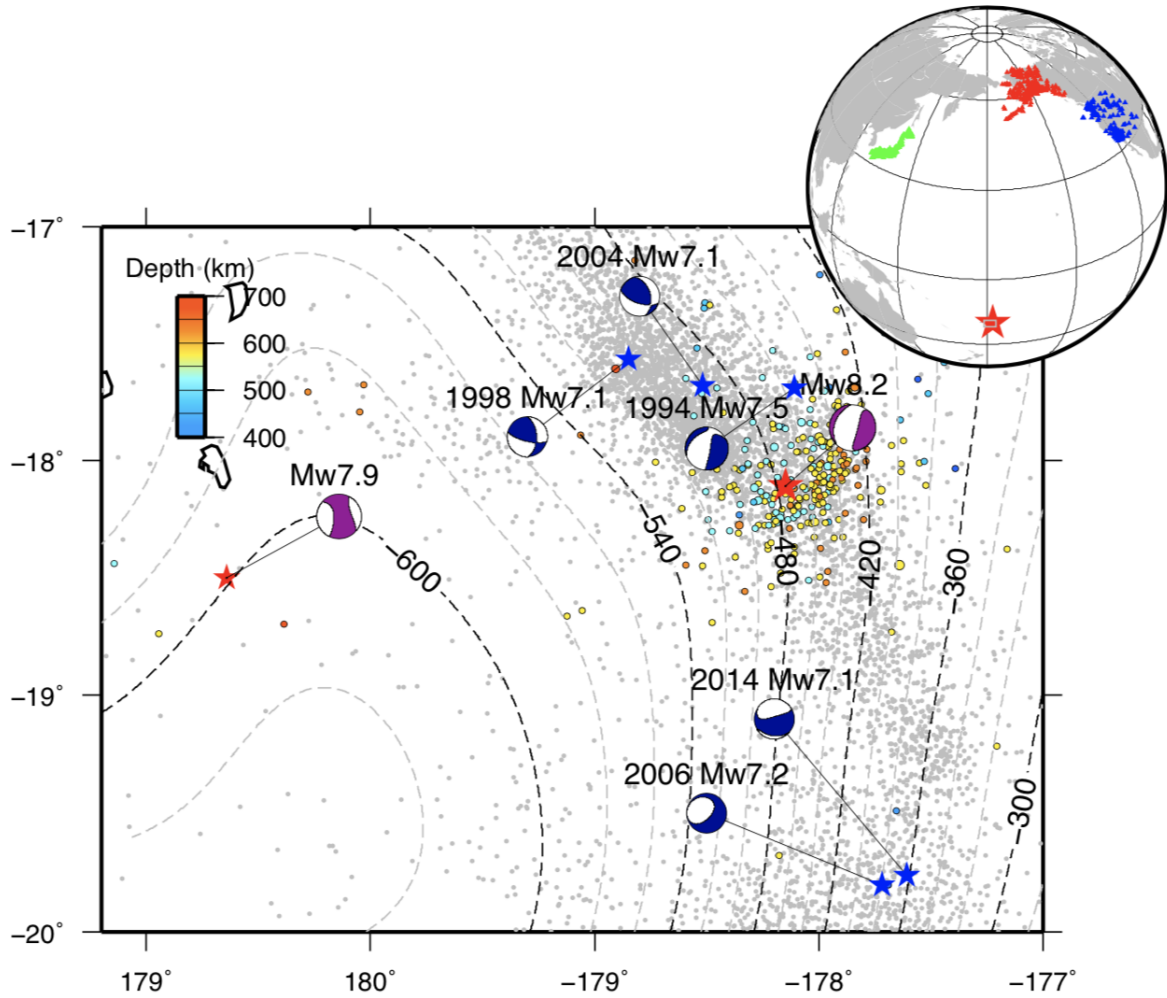


Figure 6.1: Seismo-tectonics of the Fiji-Tonga Region. The red stars are the hypocenters of studied earthquakes, and corresponding focal mechanisms acquired from GCMT ([Ekström et al., 2012](#)) are also plotted. Grey dots are background seismicity since 1990 above magnitude 5.5. One-month aftershocks for each events are labeled by colored dots. Different colors represent different depths. Background contour shows the slab depth, which is from slab contour. The inset map in the right corner provides a mapview of seismic stations used in the study. Blue, red and green triangles denote North America, Alaska, and Hinet Array respectively. Red star in the inset map represents the source area. Previous events in this area with magnitude above 7 are labeled by blue stars and moment tensors

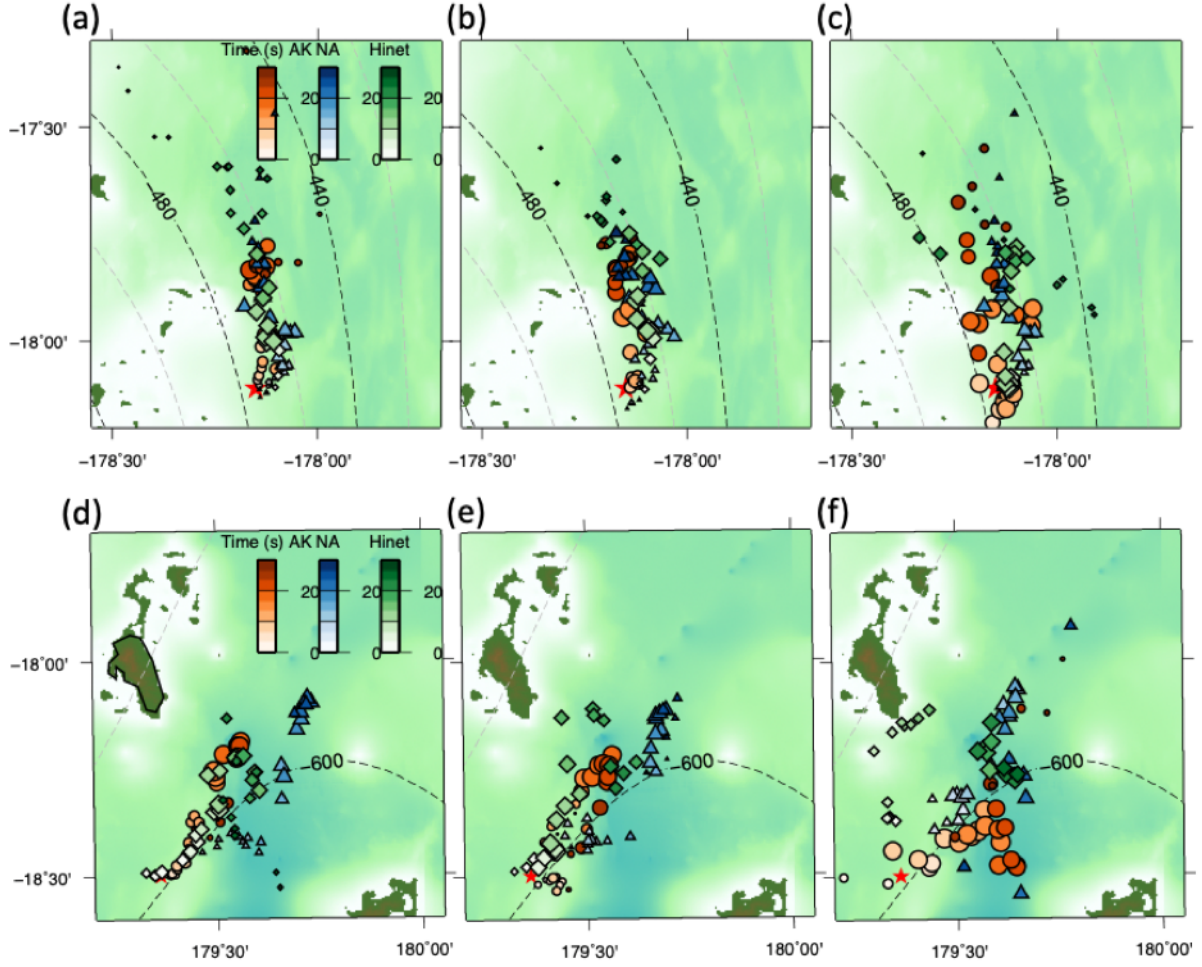


Figure 6.2: 2D backprojections from different arrays and frequencies. Upper panel describes the BP radiators of Mw 8.2 earthquakes. Circles, rectangles, and diamonds are symbols of BP from Alaska, western US, and Hinet respectively. Radiators are colored by time and sized by power. (a), (b) and (c) correspond to 0.5 - 2 Hz, 0.25 - 1 Hz, and 0.05 - 0.5 Hz. Lower panel shares the same notation as upper panel. It represents the BP of Mw 7.9 earthquake. (d), (e), and (f) are BPs performed under 0.5 - 2 Hz, 0.25 - 1 Hz, and 0.05 - 0.5 Hz correspondingly

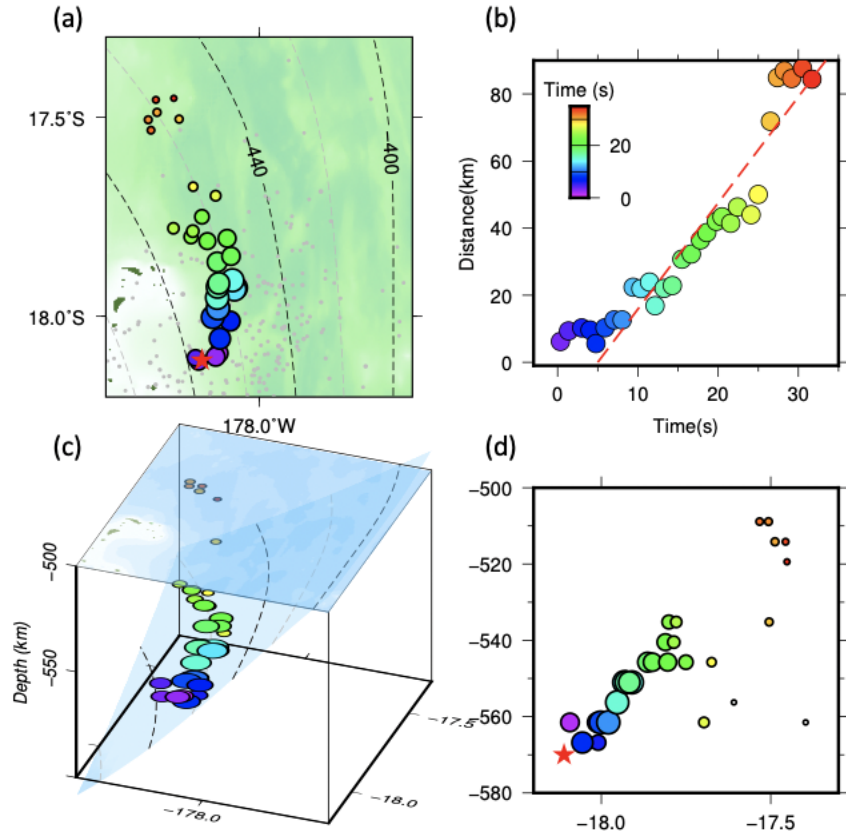


Figure 6.3: 3D BP results of the Mw 8.2 earthquake. (a) Mapview of 3D BP radiators imaged from western US network. Dots are sized by power and color-coded by time. (b) Rupture distance as a function of time and the notations are the same as (a). (c) 3D view of the rupture the transparent plane denotes the plate contour (d) Depth profile of radiators as a function of latitude for North America stations, and the notations are the same as (a).

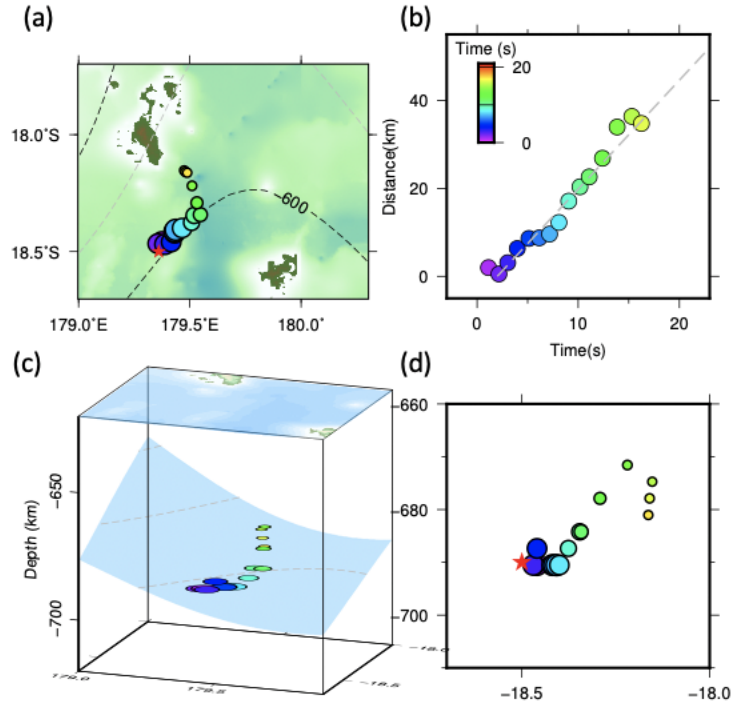


Figure 6.4: 3D BP results of the Mw 7.9 earthquake. (a) Mapview of 3D back projection radiators of the M7.9 earthquake imaged from Hi-net. Dots are sized by power and color-coded by time. (b) Rupture distance as a function of time and the notations are the same as (a). (c) 3D view of the rupture the transparent plane denotes the plate contour (d) Depth profile of radiators as a function of latitude for North America stations, and the notations are the same as (a).

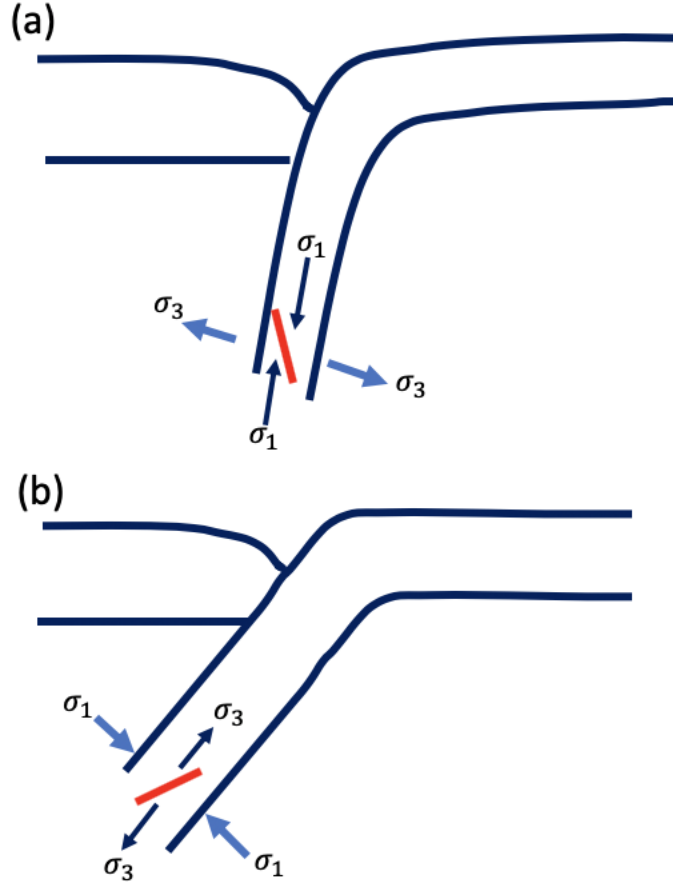


Figure 6.5: Conceptual sketch of the fault plane and cross section of slab. (a) describes the realistic setting of the Mw 8.2 earthquake and (b) describes the Mw 7.9 earthquake, dipping angles are acquired from slab 2.0. Red solid lines denote the rupture plane inferred from 3D-BP radiators. The relationship between stress and earthquake rupture is shown in each panel.

6.6 Supplementary material

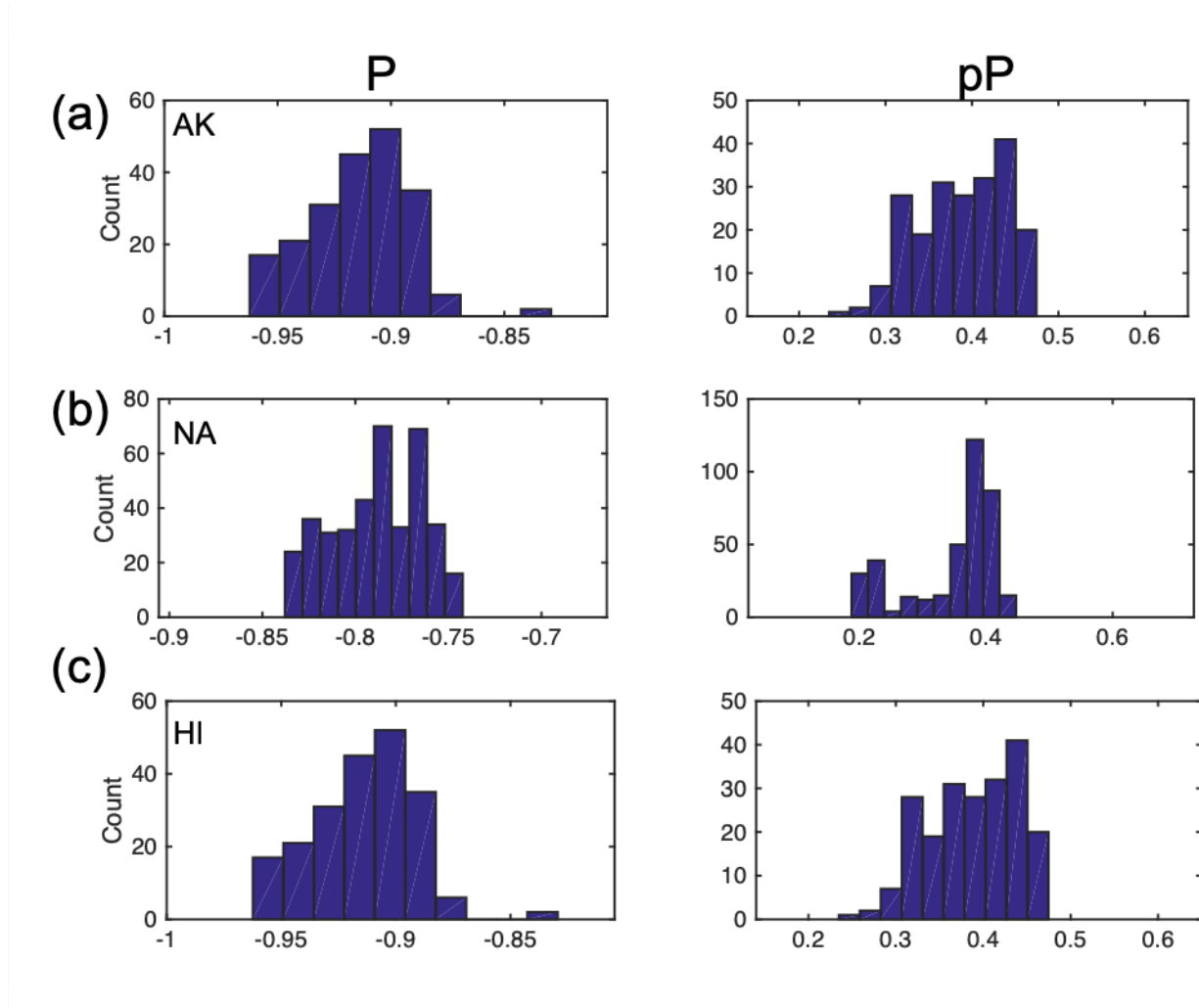


Figure S6.1: Density histogram of amplitude for P and pP phases of the Mw 8.2 event.(a)-(c) correspond to AK, NA, and Hinet respectively. In each panel, left figure is the amplitude histogram for P and the right is for pP.

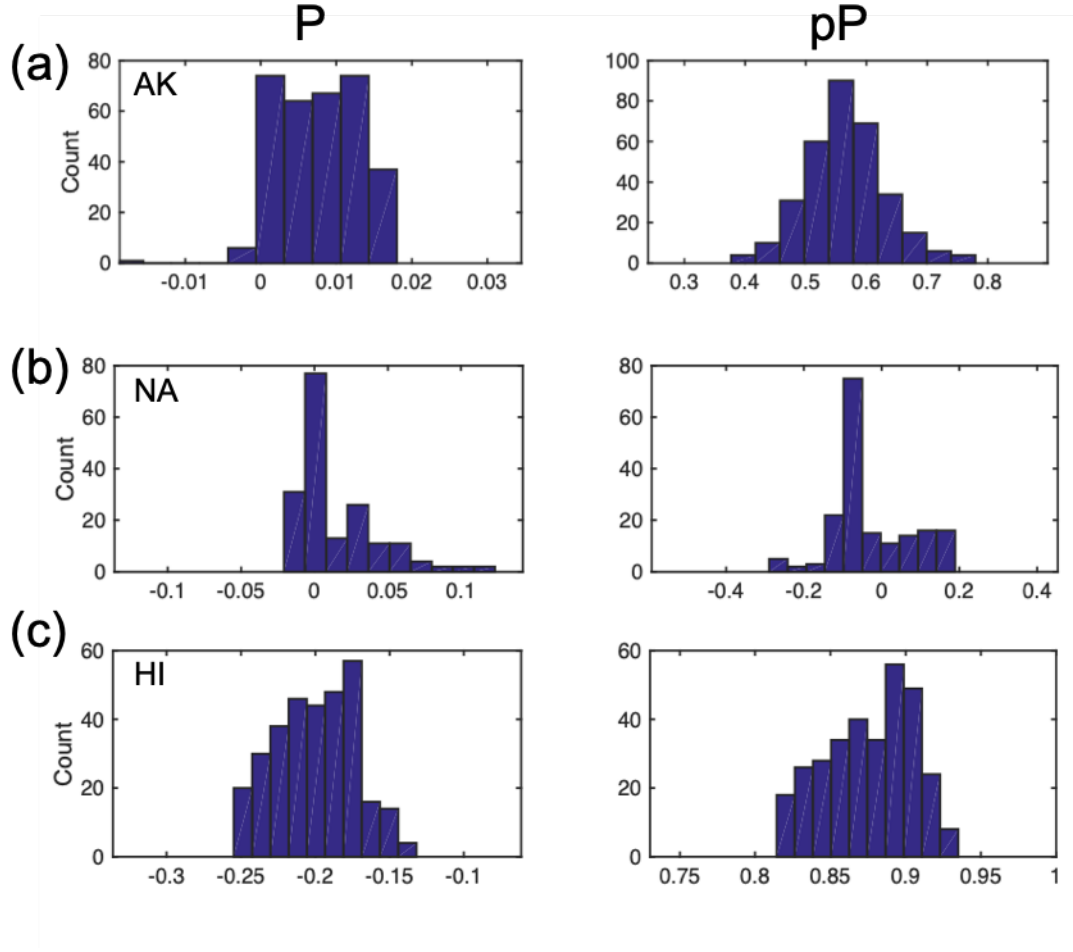


Figure S6.2: Density histogram of amplitude for P and pP phases of the Mw 7.9 event.(a)-(c) correspond to AK, NA, and Hinet respectively. In each panel, left figure is the amplitude histogram for P and the right is for pP.

CHAPTER 7

Teleseismic Array Waveform Coherence Controlled by Earthquake Rupture Dimension

Abstract

Earthquake source dimension is one of the most fundamental source parameters but remains difficult to be constrained. Here we report a new observation which relates the decay rate of array coherence to source size. We analyze waveform coherence of deep earthquakes under the sea of Okhotsk as a function of inter-station distance in USArray and the Australian array. We find that the coherence falls off faster with distance for earthquakes of larger magnitudes. The magnitude dependence of the coherence decay rate can be explained by a finite source effect. Increasing the source size causes larger phase delay between seismic radiations at different stations, resulting in a breakdown of waveform coherence. A quantitative representation of inter-station coherence as a function of source size and station distribution is established. The phase differences between the seismograms are proportional to the source dimension times the angular separation of the stations. Our synthetic tests indicate that while other parameters, such as rupture speed and directivity, can modulate the coherence function, the decay rate of array coherence is to the first order controlled by the source size. We reproduce the coherence pattern of Mw 7.7 Okhotsk earthquake using the rupture dimension inferred from teleseismic back-projections. Our observation captures a unique pattern of array coherence and illustrate the potential of utilizing waveform coherence to study earthquake source parameters.

This is an independent work that has been submitted to GRL, and we are now revising

the manuscript and prepare for re-submission

7.1 Introduction

Over last two decades, the development of large-aperture seismic arrays such as USArray, ORFEUS array (European network) and Hi-net (Japanese network) has enabled a broad spectrum of seismological studies. The dense spatial sampling of an array promote significant progresses in the studies of Earths interiors and Earthquake source, such as constraining 3D mantle structures beneath North America (Burdick et al., 2014), detecting deep earth discontinuities (Lin, Tsai, et al., 2013; Poli & Prieto, 2016), and imaging earthquake kinematic rupture process using array-back projections (Kiser et al., 2011; Meng et al., 2018). Seismic array processing methods can also improve the signal-to-noise ratios, allowing the detection of smaller earthquakes using template matching techniques (Gibbons & Ringdal, 2006; Peng & Zhao, 2009; Hawthorne & Ampuero, 2017). These coherence-based detections are also applied to non-volcanic tremors (Shelly et al., 2007; Obara, 2002), and anthropogenic signals (Riahi & Gerstoft, 2015).

While waveform coherence has been widely applied in earthquake detection, there is limited contribution toward constraining earthquake source dimensions. Current methods to study source parameters are mostly based on deterministic waveform fitting. Finite source inversion can be applied to teleseismic recordings to resolve slip distributions and rupture dimensions (Ji et al., 2002). Second-degree moments can be also inverted to address the rupture directivities, dimensions and durations of intermediate to small events recorded by local array (McGuire et al., 2001). While waveform fitting approaches provide useful constraints on earthquake source properties, the requirement of the deterministic Greens function limits the approaches to relatively low frequencies ($f < 0.1$ Hz), in which the crustal structure is roughly known. The high-frequency ($f \sim 1$ Hz) information that carries the finer details of the source process is traditionally thought to be too noisy and uncertain to be modeled deterministically.

Here, we report a novel observation to relate earthquake rupture dimension with array waveform coherence in a broad frequency range. The idea of exploring array waveform coherence is originally motivated by the purpose of seismic array processing. The spatial coherence length (maximum interstation distance with adequate waveform coherence) of an array is a key constraint on array design, which determines the maximum spatial scale to apply a back-projection analysis. Previous work indicates that waveform coherence decays dramatically in an exponential fashion with inter-station distance and frequency (Luco & Wong, 1986; Zerva & Zervas, 2002). These findings are based on local seismic phases, and the waveform incoherence tends to reflect the scattering effect from the heterogeneous crustal velocity structure. Nevertheless, owing to the fact that the path effect of teleseismic waveforms are rather simple, we have a unique opportunity to study the contribution from earthquake sources. We compute the inter-station coherence (cross-correlation coefficient between waveforms recorded by a pair of stations) of deep earthquakes beneath the sea of Okhotsk and Bonin islands recorded by USArray at teleseismic distance. For the same earthquake, we find that the Coherence function decays as a function of frequency and inter-station distance. For different earthquakes, the inter-station coherence falls off at a faster decay rate as the magnitudes of earthquakes increase. We hypothesize that the coherence pattern of an array is governed by a finite source effect. If the source size of an earthquake is larger, the phase delay of the ruptures on different subfaults are larger, resulting in a faster decay rate of coherence as a function of inter-station distance. A quantitative relationship accounting for rupture finiteness is established to relate the source size to the coherence pattern. We explain the coherence pattern of Okhotsk earthquake based on the rupture dimension inferred from teleseismic back-projections.

7.2 Array Coherence of Deep earthquakes

In our initial effort, we study four deep earthquakes under the sea of Okhotsk recorded by the USArray to understand relationship between coherence and source size. Deep earthquakes are chosen in our initial study because their waveforms are relatively simple compared with

their shallow counterparts, as the velocity structure near the source is less heterogeneous at depth. The choice of the deep earthquake ensures a minimal effect on the coherence due to source structure complexity. We study four deep earthquakes under the sea of Okhotsk recorded by the USArray to understand relationship between coherence and source size. The deep events beneath the sea of Okhotsk have been well-documented in previous studies. Finite source inversions, multiple source inversions, back-projections, and directivity analysis have been performed for earthquakes in this region (Ye et al., 2013; Wei et al., 2013; Meng et al., 2014; Zhan et al., 2014; Y. Chen et al., 2014; Park & Ishii, 2015). The source parameters reported in these studies serve as benchmarks in our coherence method.

We collected the data of deep earthquakes recorded by USArray at teleseismic distances from Incorporated Research Institute for Seismology (IRIS) data center. The data was then band-pass filtered into different frequency bands (0.125-0.5, 0.25-1, 0.5-2 Hz). The frequency bands are selected to ensure adequate signal-to-noise ratio. To mitigate travel time error due to the 3D velocity structural effects, we adopted a multi-channel cross-correlation technique to align the filtered waveforms with their initial P arrival. Interstation coherence between all the station pairs was then calculated at different frequency bands. The time window starts from initial P arrival and lasts for a fixed duration for each event according to the W Phase source duration. Such window lengths are sufficiently long to record the whole rupture process, and yet not too long to involve the degrade of coherence due to coda waves. For USArray, we include over 200 stations in the coherence measurement of each earthquake, composing over 40,000 unique station pairs. To address the statistical pattern in coherence, station pairs are grouped into distance bins with an interval of 50 km. We plot the median coherence of each bin against the interstation distance. For each distance bin, we estimate the uncertainty by bootstrapping the median coherence of 85% randomly selected station combinations in 100 realizations.

Figure 7.1 (a)-(d) show the coherence functions of Mw 6.7, 7.3, 7.7, and 8.3 deep earthquakes under the sea of Okhotsk, respectively. They occurred during 2008-2013 and were recorded by Earthscope USArray. The four deep earthquakes have similar path effects; there-

fore, coherence differences are purely controlled by the source effect. First, in order to isolate the structural effect on waveform coherence across large-scale arrays, we examine the inter-station coherence of relatively small M6.7 deep earthquakes with the minimal source effect. We find that the coherence is above 0.9 at all frequency bands across the USArray over inter-station distances > 10 wavelengths (Figure 7.1). This suggests that the contribution from the 3D structural effects is negligible on waveform coherence of the direct teleseismic P arrivals across USArray. However, as the magnitudes and therefore earthquake rupture sizes increase, we observe significant coherence decay as a function of both interstation distance and frequency. At 0.5 to 2 Hz, the coherence of the Mw 7.3 earthquake degrades from 0.8 to 0.7 over a distance range of 1000 km. The Mw 7.7 and Mw 8.3 events show even faster inter-station coherence decay from 0.7 to 0.2 and from 0.9 to 0.5, respectively. Although the Mw 8.3 earthquake is larger than the Mw 7.7 earthquake, their coherence decay rate is similar. This is because the bilateral rupture of the Mw 8.3 event reduces the coherence decay rate due to the source size (discussed in section 3). It's also worth to note that coherence at close distance (< 50 km) varies for different events. The earthquakes that occurred in different time period are recorded by USArray transversing different parts of the continental US. The M 7.3 and M 7.7 earthquakes were recorded by stations in the western US, while M 6.7 and M 8.3 were recorded by stations located in the eastern US. The initial coherence of the M 6.7 and M 8.3 events are higher than that of the M 7.3 and M 7.7 events, likely due to more homogeneous structure in the eastern US than the western US. Overall, by analyzing the coherence pattern of the deep earthquakes in Okhotsk, we conclude that the coherence decay rate increases with earthquake magnitudes. This magnitude dependence of the coherence decay rate can be explained by a finite source effect. From classical representation theory, the far-field body wave seismograms of a finite-fault rupture are the integration of source functions across the fault plane convolved with the Greens functions on each sub-fault ([Aki & Richards, 2002](#)). The travel times of the Greens functions depend both on source and receiver locations. Therefore, increasing either the source size or station distance causes larger phase delay between Greens functions, resulting in a breakdown of waveform coherence.

7.3 Synthetic Test of Coherence Dependence on Source Size

To verify the hypothesis that source size controls the decay pattern of inter-station coherence, we derived analytical relationship for inter-station coherence as a function of source size. We adopt an elongated rectangular fault ($L \gg W$) embedded in a 2D homogeneous whole space as sketched in Figure 7.2 (a). We follow the setting of Haskell's model: a unilateral rupture with a constant speed of V_r along the L direction. A uniform linear receiver array is perpendicularly oriented with respect to the fault. This simplistic setting is intended to isolate the effect of the source size on the waveform coherence.

The far-field displacement Ω at receiver position x can be expressed in the form of sum of the contribution of the sub-sources along the fault plane, thus can be written as

$$\Omega(x, t) = \frac{R}{4\pi\rho c^3} \frac{\mu W}{r} \int \dot{D}(t - \frac{x}{V_r} - \frac{r - x \cos \theta}{c}) dx \quad (7.1)$$

Where c refers to P wave speed, t is time variable, ρ is density, R is radiation pattern, r is the hypocentral distance, V_r is rupture speed, $\dot{D}(x, t)$ is the slip rate function, and θ is the angle between rupture direction and seismic ray. Thus, waveform in frequency (ω) domain can be written as,

$$\Omega(\gamma, \omega) = \frac{R}{4\pi\rho c^3} \frac{\mu W}{r} \dot{D}(\omega) e^{-i\omega r/c} \int_0^L \exp(-i\omega(\frac{x}{V_r} - \frac{x \cos \theta}{c})) dx \quad (7.2)$$

After integration, equation 7.2 yields:

$$\Omega(\gamma, \omega) = \frac{R}{4\pi\rho c^3} \frac{\mu W}{r} \dot{D}(\omega) e^{-i\omega r/c} \frac{\sin(Z)}{Z} \exp(iZ) \quad (7.3)$$

Where

$$Z = \frac{\omega L}{2} (\frac{1}{V_r} - \frac{\cos \theta}{c}) \quad (7.4)$$

If we define the coherence between the seismograms received at station A and B as the normalized cross-correlation function. The coherence at a single frequency ω can be written as,

$$Coherence(\theta_A, \theta_B, \omega) = \frac{Real(\Omega_A(\gamma, \omega) \bullet \Omega_B^*(\gamma, \omega))}{|\Omega_A(\gamma, \omega)| \bullet |\Omega_B(\gamma, \omega)|} = \cos(\frac{\omega L}{2c} (\cos \theta_A - \cos \theta_B)) \quad (7.5)$$

Equation 7.5 suggests that the coherence depends on the product of the source size L and station angular separation $\cos \theta_A - \cos \theta_B$. In teleseismic setting, the station azimuthal difference of a regional array is within a few degrees. For small azimuth differences, coherence is monotonically decreasing with the difference between θ_A and θ_B , which is projected into interstation distance of the linear array (Figure 7.2a). Equation 7.5 also indicates that a larger rupture length L results in smaller coherence for given θ_A and θ_B . Therefore faster coherence decays are expected for larger earthquakes. The complete form of the coherence equation requires the integration over the frequency range of interest.

$$C(\theta_A, \theta_B) = \frac{\text{Real}(\int \Omega(\theta_A, \omega) \bullet \Omega^*(\theta_B, \omega) d\omega)}{\sqrt{\int \Omega(\theta_A, \omega) \bullet \Omega^*(\theta_A, \omega) d\omega} \sqrt{\int \Omega(\theta_B, \omega) \bullet \Omega^*(\theta_B, \omega) d\omega}} \quad (7.6)$$

Note that integral in equation (7.6) cannot be evaluated in simple analytical form due to the integral of the Sinc function ($\frac{\sin Z}{Z}$). But we can nevertheless compute the coherence numerically. We compute the synthetic coherence for various source sizes and frequency (Figure 7.2b), which decreases monotonically as a function of interstation distance (angular separation). The coherence also decays faster at higher frequency. For the same frequency range, the larger rupture length results in faster coherence decay. This result is consistent with the coherence pattern of the deep earthquakes recorded by USArray. Its worth to note that the interstation coherence is independent of the rupture speed in equation (7.5). However, the frequency integral in equation (7.6) introduces second order terms involving the rupture speed. To understand the effect of rupture speed on the coherence curve, we test synthetic ruptures with a fixed rupture length of 70 km and varying rupture speeds from 2.5 km/s to 5 km/s. Figure 7.3 indicates that faster rupture speed produces slightly slower coherence decay. But the effect of the rupture speed is minor compared with the role of the rupture size. The rupture directivity seems to have much more significant effect on the coherence fall-off. We compare the coherence pattern of for both unilateral (Figure 7.3a) and bilateral (Figure 7.3b) scenarios. The waveform of a bilateral rupture with length $2L$ can be regarded as a summation of contributions from two separate branches of length L rupturing in the opposite directions:

$$\Omega(\gamma, \omega) \propto H_Z \exp(iZ) + H_Y \exp(iY) \quad (7.7)$$

$$Z = \frac{\omega L}{2} \left(\frac{1}{V_r} - \frac{\cos \theta}{c} \right), Y = \frac{\omega L}{2} \left(\frac{1}{V_r} + \frac{\cos \theta}{c} \right), H_z = \frac{\sin Z}{Z} \quad (7.8)$$

$$\begin{aligned} & \text{Coherence}(\theta_A, \theta_B, \omega) \\ &= \frac{H_{Z_\alpha} H_{Z_b} \cos(Z_\alpha - Z_b) + H_{Y_\alpha} H_{Y_b} \cos(Y_\alpha - Y_b) + H_{Z_\alpha} H_{Y_b} \cos(Z_\alpha - Y_b) + H_{Y_\alpha} H_{Z_b} \cos(Y_\alpha - Z_b)}{\sqrt{(H_{Z_\alpha}^2 + H_{Y_\alpha}^2 + 2H_{Z_\alpha} H_{Y_\alpha} \cos(\frac{\omega L \cos \theta_a}{c}))} \sqrt{(H_{Z_b}^2 + H_{Y_b}^2 + 2H_{Z_b} H_{Y_b} \cos(\frac{\omega L \cos \theta_b}{c}))}} \quad (7.9) \end{aligned}$$

The numerator of Coherence $(\theta_A, \theta_B, \omega)$ function of the bilateral rupture involves four complicated terms without simple analytical form: The first two terms represent the phase difference of the same branch between the station pair A and B, and the latter two terms are the cross terms of phase difference between the two branches. Our numerical test suggests that the cross terms does not significantly decrease with station angular separations. Only the first two term contribute to the coherence decay. Therefore, the effective size that controls the coherence decay function is L instead of 2L. This effect is illustrated by our numerical tests in Figure. 7.3b. The coherence decay rate of the bilateral rupture of length 2L is similar to that of the unilateral rupture of length L in Figure. 7.3a. Our tests suggest that bilateral rupture directivity effectively controls the coherence decay rate.

The coherence function of the 1D fault trace in a 2D medium can be generalized into a setting of 2D fault plane embedded in a 3D space. In the case, the fault plane aspect ratio can be arbitrary, thus the waveform equation 7.2 is revised to be

$$\Omega(\gamma, \omega) e^{-i\omega r/c} = \frac{R}{4\pi\rho c^3} \frac{\mu W}{r} \dot{D}(\omega) G(\omega) \int_0^L \int_0^W \exp(i\omega(\frac{x \cos \beta + y \sin \beta}{V_r} + \frac{x \cos \theta + y \cos \alpha}{c})) dy dx \quad (7.10)$$

Where $G(\omega)$ is the spectra of the empirical Greens function, x is along strike direction with a length of L, and y is along dip direction with a length of W, θ is the angle between ray vector and strike direction, and α is the angle between ray vector and dip direction and β is the rupture direction angle, which is measured counterclockwise with respect to the strike direction. Similar to the 1D model, coherence will be modulated by rupture length L, width

W, frequency, and angular separations between stations. Angular separation is parametrized as a combination of θ and α , which is comparable to the difference of azimuths and take-off angles for a station pair. Increasing either azimuthal or incidental difference in station pairs will lead to a larger interstation distance. Thus, we can present coherence as a function of interstation distance as a proxy of angular separation.

7.4 Forward modeling of Coherence for the Okhotsk events

With the equation provided in the previous section, we are able to model the coherence pattern observed for deep Okhotsk earthquakes. At the sea of Okhotsk, the Pacific plate underthrusts beneath the Okhotsk plate (part of North America plate) along the Kuril-Kamchatka subduction zone. Deep section of the Pacific slab beneath the Sea of Okhotsk has hosted several M7+ earthquakes since 2008. We are interested in reproducing observed coherence pattern with source dimensions inferred from teleseismic back-projections. Unlike shallow earthquakes whose source dimensions and geometries can be estimated through well-developed aftershock and geodetic data, deep earthquakes are generally deficient in such activity (Wiens & McGuire, 1995). Therefore back-projection (Meng et al., 2014) is used to extract the rupture size and speed in this study. Rupture dimensions are parameterised into rupture length along strike and width along dip direction.

Back-projection also indicates that the Mw 7.7 is a unilateral event and the Mw 8.3 is bilateral. Fault plane geometry is determined by the W Phase solution. As we have mentioned, M7.7 and M8.3 events are recorded by two distinct sets of stations, thus M7.3 and M6.7 events act as empirical Green's functions accordingly.

For the Mw 7.7 event, we set the rupture length to be 50 km according to the back-projection results. Rupture width is constrained by the width of metastable olivine wedge (~ 15 km), therefore the rupture along dip direction is set to be 30 km. For the Mw 8.3 event, we set the rupture length to be 115 km and width to be 60 km in order to include all the radiators. Rupture speed for the Mw 7.7 is set as 3.0 km/s (similar with Park and

Ishii, 2015), and rupture speed for Mw 8.3 is 4.0 km/s, and it is a bilateral rupture with hypocenter in the center of the fault plane. With those source parameters, we can compute the synthetic coherence curve for the Mw 7.7 and 8.3 as shown in Figure 7.4.

The forward model with realistic source parameters provides similar coherence pattern as observational data. The coherence curve falls off as a function of interstation distance. The decay rate is slower for lower frequency bands, but becomes larger at higher frequencies. The coherence decay is mainly modulated by source size, as we are able to reproduce similar amount of coherence fall off for both events. The predicted coherence curve further validates that estimated source dimension is reliable. The forward modeling has the potential to identify the optimal source dimension inferred from other waveform-based models.

Another factor that controls coherence decay is the directivity effect. We run additional forward models for the Mw 7.7 case with different rupture speeds and directivities as shown in Figure 7.5. Directivity effect (unilateral or bilateral) has a more significant effect on coherence than rupture speed. With known directivity, coherence doesn't vary too much as we change rupture speed from 2.5 km/s to 4.5 km/s. Such a rupture speed range is large enough to include a broad spectrum of deep earthquakes. Therefore, we can conclude that rupture speed doesn't dominate the coherence curve. However, with fixed rupture speed, the coherence curve varies a lot between bilateral and unilateral ruptures. Overall, we can conclude that both directivity effect and source size are first order factors that control the coherence curve.

7.5 Discussion

We systematically studied the inter-station waveform coherence of deep earthquakes under the sea of Okhotsk recorded by USArray. A salient feature is that the decay rate of the coherence curve depends on the earthquake source size and directivity. We constructed analytical relations to explain the coherence behavior governed by the earthquake source size.

With realistic source parameters, we are able to forward model the coherence curve, which shares similarity with the observation. While the coherence model capture key features in the observation, we acknowledge that we cannot reproduce the exact same coherence curve that minimizes the residual. Presumably, coherence should converge to 1 when interstation distance is close to 0. But from observations, the coherence at small interstation distances can vary from 0.7 to 1. Second order factors, such as path effect and ambient noise can play a role in the coherence curve. Therefore, the scope of this study is focus on source size and directivity. Our result is an effort to demonstrate such relations and set the stage for future studies to constrain source dimension using waveform coherence. About rupture speed, we show that the effect of the rupture speed is minimal in controlling the decay rate of coherence. Therefore, we think it is reasonable to use even the super-shear M6 event as the Greens function of the larger event for forward coherence modeling.

Although it is premature to directly invert for source size from coherence curve, we can cross validate source sizes from other studies with the coherence approach. For deep earthquakes, source dimension is conventionally inferred from source inversions or back projections. Source inversions generally suffer from non-uniqueness, and back projections assume a point source at each time frame. Hence, extracted source sizes are subject to large variabilities. We can utilize coherence curve as an alternative to cross-validate the inferred source size. If the size is realistic and reasonable, synthetic coherence curve will be closer to the observation.

Through seismic waveform interferometry, we develop a reliable resort to add additional constraint on earthquake source dimension, which is conventionally difficult to be accurately constrained. Such observations potentially provide new constraint in discriminating physical models derived from numerical simulations and laboratory experiments. For example, the rupture length and width in our case studies shed light on the mechanisms of deep earthquakes. The M7.7 event is confined in a narrow rupture width on a horizontal plane (~ 25 km), comparable with the thickness of the metastable olivine wedge (~ 15 km) (Marone & Liu, 1997). Such scenario can be interpreted as transformational failure (Green & Hous-

ton, 1995) within metastable olivine wedge. However, the M8.3 event spans a more extensive widths as $\sim 40\text{km}$, beyond the plausible width of the metastable olivine wedge, which requires alternative mechanisms such as thermally induced shear instability (Meng et al., 2014).

Another potential benefit from better-constrained rupture dimension is to reduce the variability of stress drop estimations. Under assumption of circular rupture (Eshelby, 1957), the relationship between stress drop ($\Delta\sigma$), moment release (M_0) and corner frequency (ω_c) can be written as:

$$\Delta\sigma = \frac{7}{16} \left(\frac{\omega_c}{k\beta} \right)^3 M_0 \quad (7.11)$$

Where β is the shear wave velocity and k is a constant dependent on the theoretical source model. The two most commonly used models are the, with $k = 0.37$ ($Vr = 0.9\beta$) and $k = 0.21$ ($Vr = \text{inf}$) respectively, where Vr is the rupture speed.

However, the biased sampling of the focal sphere is unavoidable in reality. This uncertainty of ω_c contributes to the apparent variability of the stress drop estimates (Kaneko & Shearer, 2014).

For the rectangular rupture of a unilateral Haskell source, the corner frequency ω_c is related to the rise time T and rupture time L/V , with L being the rupture length and V being the rupture speed. Therefore, the corner frequency uncertainty can be directly related to the uncertainty in rupture dimension. We can further develop the stress drop estimation from rupture dimension, by considering moment release as:

$$M_0 = \mu D W L \quad (7.12)$$

Where D is the average displacement. For an embedded strike-slip fault, the stress drop is estimated by:

$$\Delta\sigma = \frac{4}{\pi} \mu \left(\frac{D}{W} \right) = \frac{4}{\pi} \frac{M_0}{W^2 L} \quad (7.13)$$

Once we can include a direct estimate of the rupture dimensions W and L , the stress drop can be directly estimated from equation 7.10. As this relationship circumvents the model assumptions on corner frequencies, it would in principle provides more reliable stress-drop

estimates if the rupture dimensions L and W can well constrained. Based on our forward models, we estimate the stress drop of the M7.7 event as 6 MPa and the M8.3 event as 12 MPa. Published stress drops of M8.3 event are reported as 12-15 MPa (Ye et al., 2013) and 8 MPa (Wei et al., 2013). Our estimation also yields consistency with previous studies.

7.6 Conclusion

In this study, we provide a novel approach to observe the finite source effect. We compute interstation coherence at different frequency ranges as a function of interstation distance. We observe that coherence is predominantly modulated by source size and directivity effect. Coherence decays faster as the source size becomes larger. Unilateral ruptures tend to have a faster coherence decay than bilateral ruptures. The magnitude dependence of the coherence decay rate can be explained by a finite source effect. Increasing the source size causes larger phase delay between seismic radiations at different stations, resulting in a breakdown of waveform coherence. A quantitative representation of inter-station coherence as a function of source size and station distribution is established. We use forward models with realistic settings to analyze the interactions between source parameter on coherence. Our observation captures a unique pattern of array coherence and illustrate the potential of utilizing waveform coherence to study earthquake source parameters.

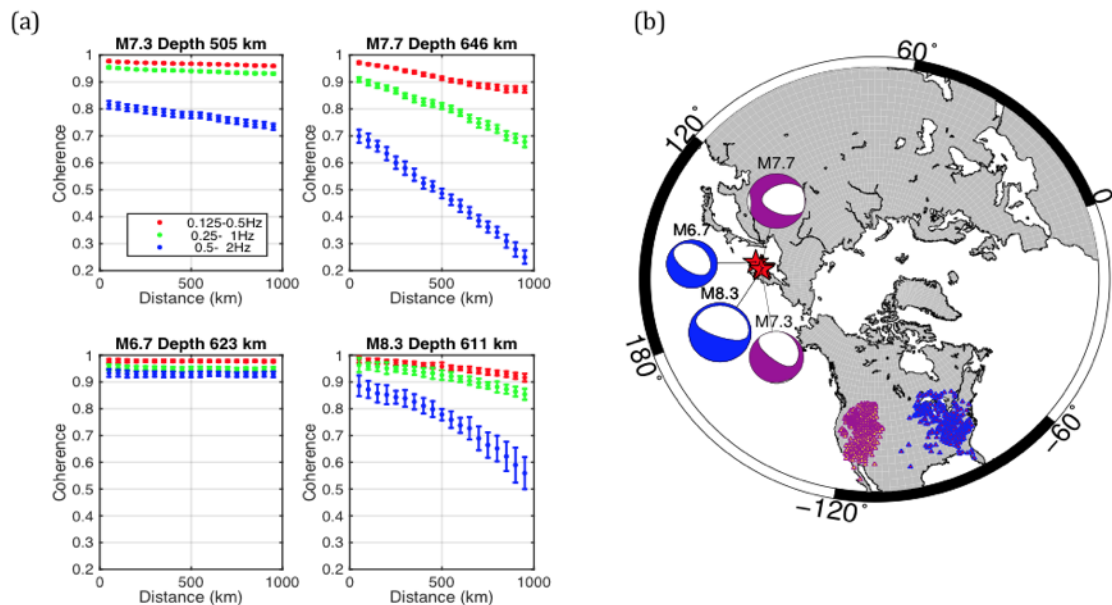


Figure 7.1: Coherence pattern for deep earthquakes in the Sea of Okhotsk. (a) Observed coherence as a function of interstation distance of the four Sea of Okhotsk deep earthquakes (M6.7, M7.3, M7.7, M8.3) recorded at USArray. The dots represent the median coherence value of each interstation distance bin (50 km) color-coded by the frequency band, with the error bar delimiting 95% confidence interval. Four panels in (a) compare the coherence decay of deep earthquakes with different magnitudes. (b) Map of station distribution and focal mechanisms of earthquake. Triangles denote USArray stations used in coherence measurement. Red stars are the centroid location of earthquakes, and moment tensor solutions are taken from W phase.

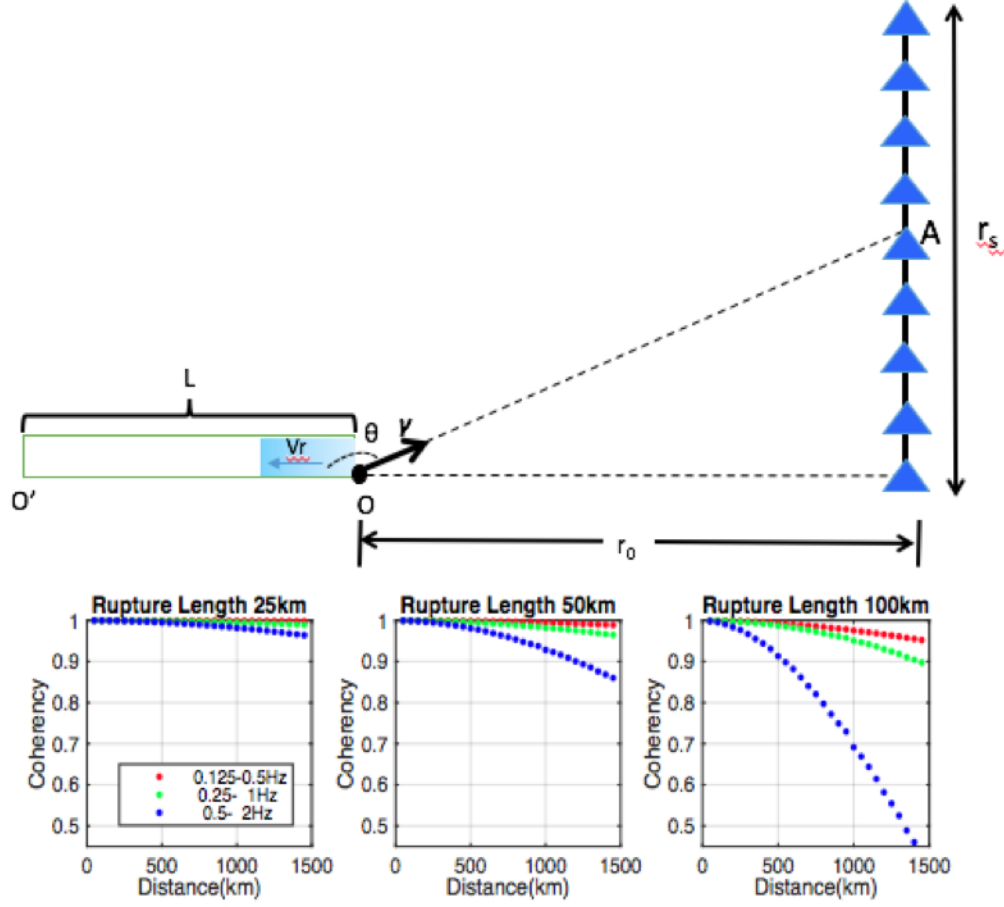


Figure 7.2: Conceptual sketch and synthetic coherence function.(a) A sketch of the source and receiver settings for the synthetic derivation, with L being rupture length, W being rupture width, θ being the orientation for stations, γ being the unit vector pointing from the fault origin (O) to the receivers, $r_0 = 6000$ km being the distance from the source to the bottom end of the receiver array, and $r_s = 1500$ km, being the size of a uniform linear array. (b) Synthetic coherence function for a rectangular fault embedded in 2D medium inferred from setting in (a). Figures from left to right compare distance-coherence curves for different rupture lengths, L (25, 50, and 100km), and uniform rupture speed 3km/s.

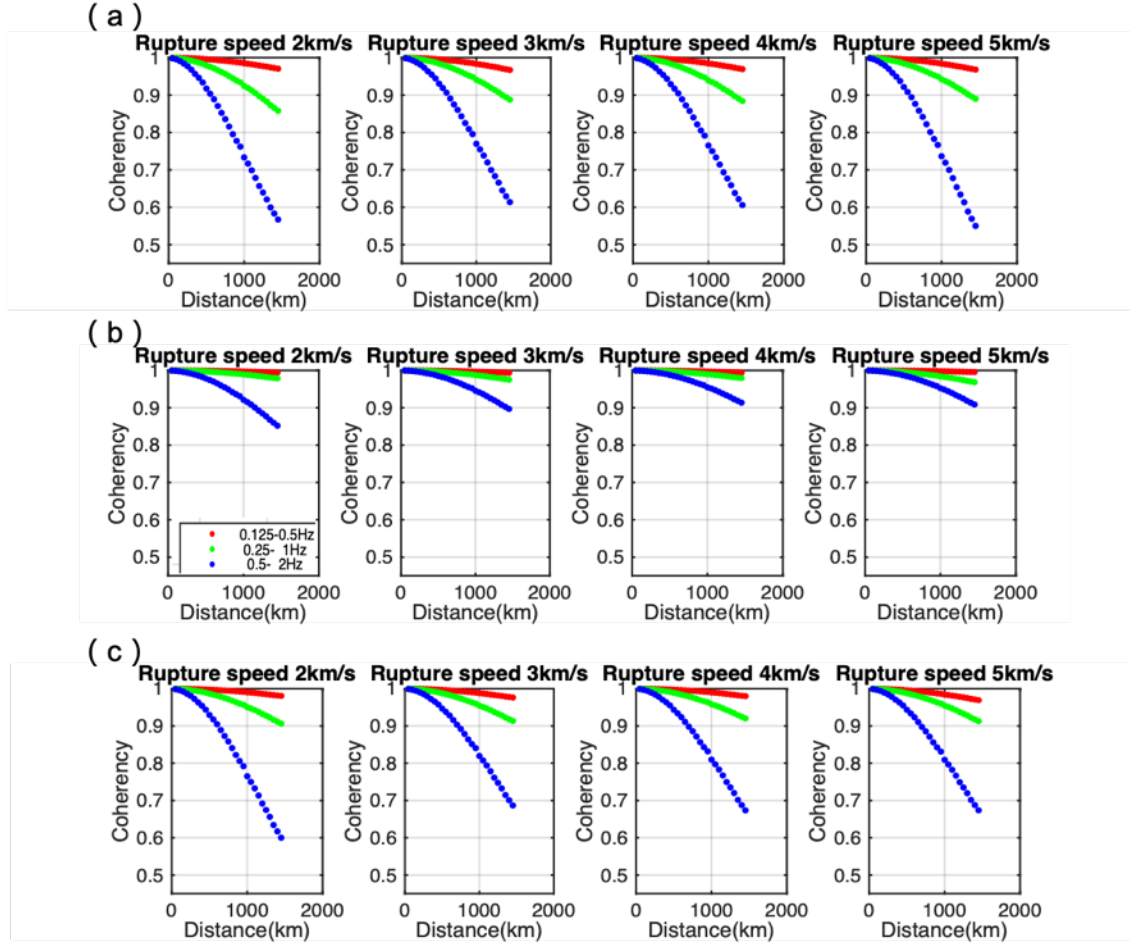


Figure 7.3: Coherence function for unilateral and bilateral scenarios. (a) Coherence curve of unilateral rupture with different rupture speeds. Coherence is presented as a function of interstation distance. Station and fault setting are the same as Fig 2, and the rupture length is fixed as 70 km. Rupture speeds vary from 2 km/s to 5 km/s from left to right. (b) Coherence curve of bilateral rupture with different rupture speeds. Other parameters are the same as (a). (c) Coherence curve of bilateral rupture with different rupture speeds. Other parameters are the same as (a) and (b), except that the rupture length is 140 km.

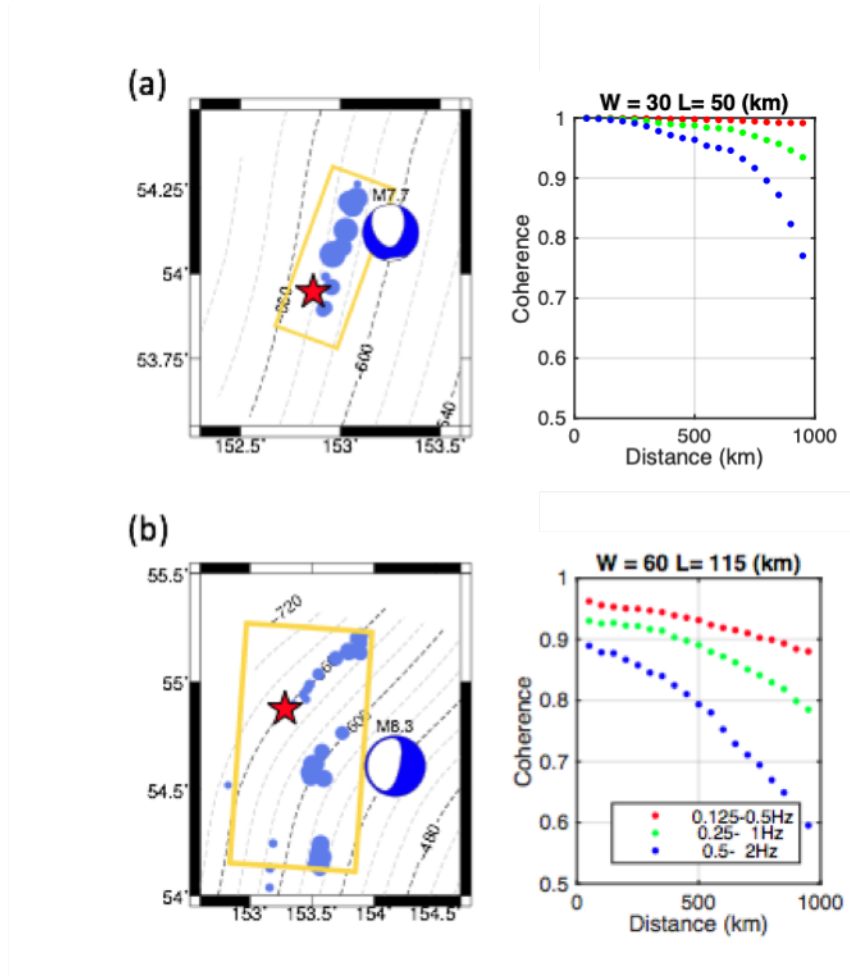


Figure 7.4: Forward models of Okhotsk earthquakes. (a) Mapview and coherence curve of the Mw 7.7 earthquake. Back-projection radiators are marked with blue dots and the rectangle with solid yellow lines is the fault plane which can enclose the dimension of the radiators. Red stars are the NEIC catalogue location of the and blue dots are mapping the back-projection trace. Focal mechanisms are plotted as a blue beach ball. The right panel is the modeled coherence curve colored by frequency for an earthquake with the rupture dimension of the rectangle, width and length are displayed in the title. (b) Mapview and coherence curve of the Mw 8.3 earthquake. All the notations are same with (a).

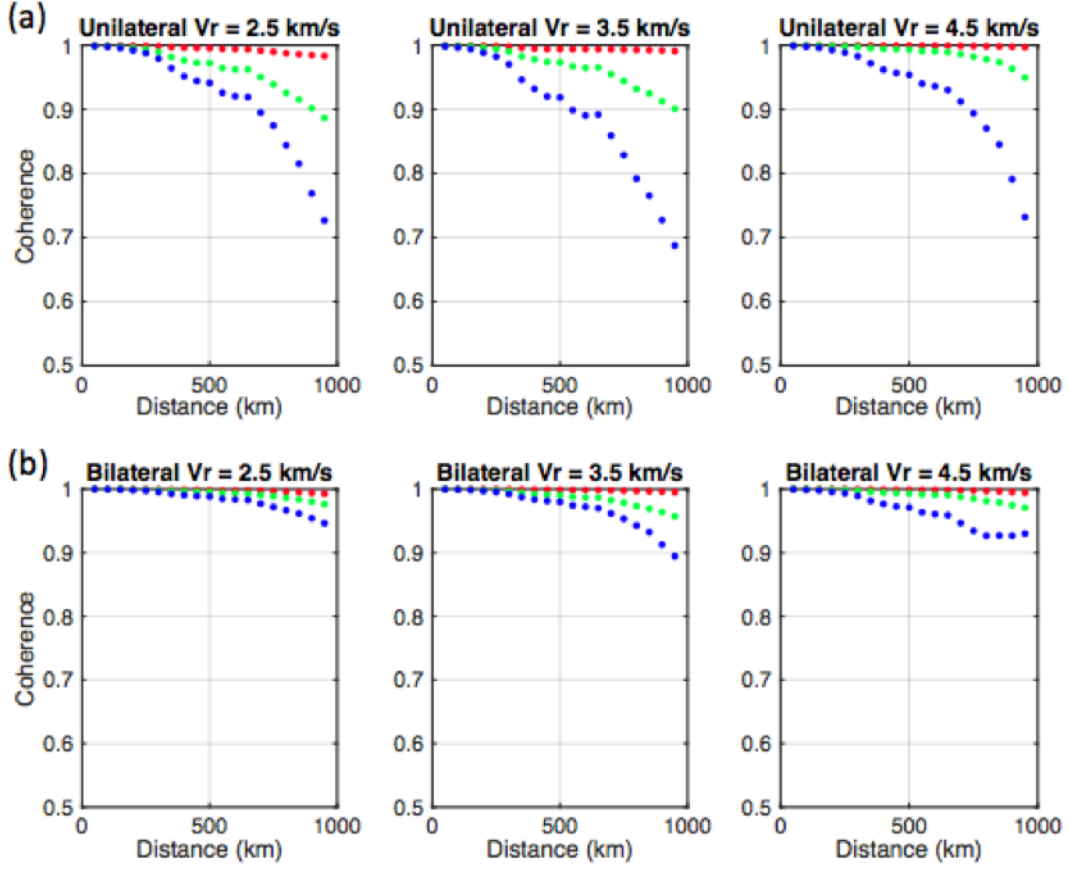


Figure 7.5: Coherence curves in the setting of the Mw 7.7 events with different rupture speeds and directivity effects. We fix the rupture dimension as 30 km along dip and 50 km along strike. (a) assumes a unilateral rupture and each panel corresponds to different rupture speeds (2.5, 3.5 and 4.5 km/s) (b) assumes a bilateral rupture, with the same rupture speed as selected in (a).

CHAPTER 8

Conclusion

Through my PhD, my research focus is on developing enhanced source imaging techniques to provide more accurate constraints on kinematic rupture process. To accomplish this goal for a wide spectrum of earthquakes, we have developed multiple strategies. For shallow earthquakes in conjugate fault systems, we propose back projections with slowness corrections to mitigate spatial biases that is a common issue in conventional back-projections. The slowness correction has been successfully applied to ruptures in conjugate fault systems on and off shores. For the 2016 Mw Kaikoura earthquake, we observe cascade ruptures on multiple conjugate faults that are perpendicular to the major fault (the Hope fault). High frequency radiators are observed to take a zig-zag path during propagation. The rupture speed on each segment is above 2 km/s, whereas the overall migration speed is as slow as ~ 1.4 -1.7 km/s. Such a kinematic rupture process can be reproduced by a generic dynamic simulation. The fault branching and stepping can be explained by dynamic unclamping effects with preferred geometrical orientation. Similar rupture pattern is also observed offshore in the Kodiak region. The 2018 Mw 7.9 Kodiak earthquake is imaged by teleseismic P wave backprojection. High frequency radiators are observed to be clustered on EW trending conjugate faults. On each segment, rupture is propagating at a speed of 3 km/s, and the overall rupture speed is slow (~ 1.7 km/s). The refined source imaging enables a unique chance for us to understand the interaction between static and dynamic triggering in a complex rupture.

On the other hand, deep earthquakes generally occur at a depth that is larger than 300 km. At such depth, fault orientation is poorly constrained due to limited aftershock activities and large uncertainty in depth resolution. Limited observations also bring open questions to rupture process of deep earthquakes. We develop 3D back projection that resolve

spatial-temporal evolution of the rupture in 3D space. With 3D BP, we observe rupture complexities of the 2015 Izu-Bonin earthquake and 2018 Tonga-Fiji doublet. Both events are consistent with MSI, and 3D BP is very helpful in illuminating the physical mechanisms of deep earthquakes. Besides back projection, we also propose a novel approach that utilizes array waveform coherence to investigate finite source effect. We observe directivity and source size are two the most important factors in modulating the coherence curve. This method creates independent measures to validate and constrain the source sizes of deep earthquakes.

I vision that there is still large potential to improve the kinematic source imaging in future studies. In particular, array-based approach will earn more prevalence with extensive instrument deployment. The technique used in this study is more physics based, which are derived from classical ray theories. Physics-based models are generalized and has the benefit of simplicity and effectiveness. However, prior assumptions and simplification are made to create a condensed physical model, which can be violated in realistic cases. Development in array seismology and big volumes of data will enable data-driven models to address open questions in earthquake source physics. For example, there is effort in the community to cluster source time functions only from the data point of view. Then physical based simulations are performed to interpret each cluster and address the generation of source time functions in this cluster. Data-driven models requires more efficient data mining strategies that can create correlations between earthquake source properties and observations. With appropriate data-mining, there is chance to extract more representative features from the seismic data which has more direct constraints on fundamental source parameters. Data-driven model will create a chance for us to discover hidden patterns beneath data, that has been ignored or simplified in the physics-based models.

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