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Key Points:

- A two-step strategy with multi seismic datasets is proposed to perform multi-scale finite fault inversion in the wavelet domain
- For the 2016 Mw 6.5 Norcia earthquake, the higher-frequency slip appeared near the edges of the lower-frequency slip patch
- The frequency-dependent rupture of this earthquake may be relative to fault complexities and unevenly distributed pore fluid pressure

Supporting Information:

Supporting Information may be found in the online version of this article.

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Frequency-Dependent Rupture Characteristics of the 30 October 2016 Mw 6.5 Norcia, Italy Earthquake Inferred From Joint Multi-Scale Slip Inversion

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Abstract The relationship between multi-scale rupture characteristics and different frequency contents of the seismic waveforms has been revealed for moderate to great earthquakes by previous researches. To better understand such a relationship, we propose a two-step strategy to perform finite fault inversion in the wavelet domain with multi seismic datasets, involving the seismic observations at both teleseismic and near-field distances. In our two-step strategy, the larger-scale pattern of the seismic rupture is first constrained by using the lower-frequency seismic waveforms and then based on the lower-frequency rupture pattern, the higher-frequency rupture features are obtained by performing inversion with higher-frequency data. We apply this two-step strategy with multi seismic datasets to study the 30 October 2016 Mw 6.5 Norcia, Italy earthquake, which is the largest event of the 2016–2017 central Italy seismic sequence. By analyzing the recovered slip patterns from different frequency bands, we notice the frequency-dependent rupture features of this earthquake. The lower-frequency slip concentrated at the shallower depths above the source, and the higher-frequency slip appeared near the edges of the lower-frequency slip patch, working as “barriers”, due to fault complexities in central Italy. The frequency-dependent slip pattern of the 30 October 2016 Norcia earthquake shows a relation to the unevenly distributed pore fluid pressure, as indicated by Vp/Vs tomography from previous researches. Our results suggest that fault complexities and unevenly distributed fluids contribute to the frequency-dependent rupture of this earthquake.

1. Introduction

Unveiling the temporal and spatial characteristics of the rupture process is crucial to understand the mechanism of earthquakes. Many previous studies have found that seismic waveforms at different frequency bands show different characteristics of the earthquake rupture, which indicates a relationship between the frequency content of seismic data and the “apparent” rupture process (e.g., Uchide et al., 2013; Wang & Mori, 2011; Yao et al., 2011; Yin et al., 2017). Such a relationship is called the frequency-dependent rupture characteristic. Generally, lower-frequency seismic waveforms are more sensitive to larger-scale rupture characteristics (such as smooth and large slip on the fault plane), while higher-frequency seismic signals are imprints of smaller-scale characteristics (such as sudden changes in rupture properties, e.g., rupture velocities and strong frictional variations, and a heterogeneous 3-D velocity structure hosting the fault plane) (e.g., Lay et al., 2012; Meng et al., 2011; Sato, 1994; Yamashita, 1983; Yao et al., 2011, 2013). In previous studies, the frequency-dependent rupture characteristic has been revealed for moderate to great earthquakes (Mw > 5.5), such as the 1989 Mw 7.1 Loma Prieta earthquake (Zeng et al., 1993), the 1997 Mw 5.9 Yamaguchi earthquake (Ide, 1999), the 2000 Mw 7.3 western Tottori earthquake (Suzuki & Iwata, 2009), the 2004 Mw 9.2 and 2005 Mw 8.6 Sumatra earthquakes (Yao et al., 2013), the 2010 Mw 7.2 El Mayor-Cucapah earthquake (Uchide et al., 2013), the 2010 Mw 8.8 Maule earthquake (Wang & Mori, 2011; Yao et al., 2013), the 2011 Mw 9.0 Tohoku earthquake (Wang & Mori, 2011; Yao et al., 2011, 2013), and the 2015 Mw 7.8 Nepal earthquake (Yin et al., 2017).

Finite fault inversion is one of the most used methods to reveal the earthquake rupture processes, which can resolve the heterogeneous coseismic slip distribution along the fault plane (Ide, 2007). There are many

Figure 1. Seismicity in central Italy. Dots show earthquakes with a magnitude between 3 and 5 detected by the Istituto Nazionale di Geofisica e Vulcanologia (INGV) from January 1990 to January 2017, and stars show $M > 5$ events. Beach balls show the focal mechanisms of $M > 5$ earthquakes, with events before 2000 from the European-Mediterranean Regional Centroid-Moment Tensors (RCMT) catalog (Pondrelli, 2002) and events after 2000 from the Time Domain Moment Tensor (TDMT) catalog (Scognamiglio et al., 2006). The color of dots, stars, and beach balls is scaling with time. White and black lines show the major thrusts and normal faults in central Italy, respectively (MCT: the Mt. Cavallo thrust, OAST: the Olevano-Antrododo-Sibillini Mts. thrust, GST: the Gran Sasso thrust, VBFS: the Mt. Vettore-Mt. Bove fault system, LMFS: the Laga Mts. fault system) (modified after Pizzi & Galadini, 2009). Red lines indicate the coseismic surface rupture trace of the 30 October Mw 6.5 Norcia earthquake (from Villani et al., 2018).

approaches to perform finite fault inversion, including time-domain (i.e., the objective function of the inversion is constructed in the time domain, such as the multi-time-window method) (e.g., Hartzell & Heaton, 1983; Kikuchi & Kanamori, 1982; Olson & Apsel, 1982), frequency-domain (the objective function is constructed in the frequency domain) (e.g., Cotton & Campillo, 1995; Fan et al., 2014; Olson & Anderson, 1988), and wavelet-domain (the objective function is constructed in the wavelet domain) (e.g., Ji et al., 2002; Suzuki & Iwata, 2009) methods. Although all these methods can well resolve the earthquake rupture processes and have been successfully applied to many notable earthquakes (e.g., Avouac et al., 2015; Uchide & Ide, 2007; Wei et al., 2012; Ye et al., 2016; Yokota et al., 2011), among them only wavelet-domain methods can analyze both the time and frequency information of seismic waveforms at different scales by the usage of the wavelet transform. Ji et al. (2002) proposed a wavelet-domain method based on the simulated annealing (SA) approach to perform finite fault inversion with seismic waveforms of different frequency contents (i.e., lower- and higher-frequency signals), trying to recover both larger- and smaller-scale characteristics of the rupture processes simultaneously.

In our previous research (W. Liu & Yao, 2018), we proposed an alternative two-stage SA strategy to recover larger-scale characteristics first and then smaller-scale features, based on a single dataset of teleseismic waveforms. Such a strategy can work efficiently at converging to the global minimum solution with regard to the choice of inversion parameters (e.g., the temperature-decreasing scheme), and reveal the potential to resolve multi-scale characteristics of the seismic rupture. Benefited from the development of seismic observation techniques, multi seismic datasets (e.g., teleseismic data, near-field strong motion data, and tsunami wave data) can be accessed nowadays. In this paper, to better understand the frequency-dependent rupture characteristic for moderate size earthquakes, we extend this two-stage strategy to multi seismic datasets, including the seismic observations at teleseismic (epicentral distance of 30° – 90°) and near-field (epicentral distance less than a few fault lengths) distances.

In recent years, the normal fault systems in central Italy have been ruptured by many moderate-size earthquakes, such as the 1997 Colfiorito seismic sequence (e.g., Chiaraluce et al., 2003), the 2009 L'Aquila seismic sequence (e.g., Hermann et al., 2011), and the most recent 2016–2017 Amatrice-Visso-Norcia seismic sequence (e.g., Bignami et al., 2019; Chiarabba et al., 2018; Chiaraluce et al., 2017; Improta et al., 2019) (Figure 1). The Apennines belt in central Italy is a seismically active orogenic wedge, which was formed by the subduction of the continental Adriatic plate beneath the Eurasian plate since ~ 30 Ma (Agostinetti & Faccenna, 2018; Faccenna et al., 2014; Sun et al., 2019). Subduction of a continental lithosphere can produce an orogenic wedge by scraping off crustal material from the downgoing plate and accreting them to the upper plate. After the collision, a post-orogenic extension was superimposed on the belt during Quaternary (Pizzi & Galadini, 2009), which may be related to rollback and/or delamination of the subducting slab (Agostinetti & Faccenna, 2018; Chiarabba & Chiodini, 2013; Faccenna et al., 2014). As a consequence, complex normal fault systems were produced in central Italy (Bonini et al., 2019). Due to this environment, the rupture of moderate size earthquakes that occurred in central Italy possessed complicated rupture patterns, which produced notable surface ruptures (e.g., Villani et al., 2018) and caused great damage to human society (e.g., Chiaraluce et al., 2017). Understanding the frequency-dependent rupture feature of these damaging moderate size earthquakes would be critical for future earthquake physics and seismic hazard analysis in central Italy.

Here we study the largest event of the most recent 2016–2017 seismic sequence, the 30 October 2016 Mw 6.5 Norcia earthquake, which occurred at 42.8303°N and 13.1092°E with a focal depth of 10.0 km according to the Istituto Nazionale di Geofisica e Vulcanologia (INGV). According to the Time Domain Moment Tensor (TDMT) catalog (Scognamiglio et al., 2006) published by INGV, the 2016–2017 seismic sequence contains 9 events with moment magnitude larger than 5, beginning on 24 August 2016 with an Mw 6.0 event (the 24 August 2016 Amatrice earthquake) and ending on 18 January 2017 with 4 events (Figure 1). The 30 October 2016 Norcia earthquake has been studied by many researches (e.g., Bonini et al., 2019; Cheloni et al., 2017; Chiaraluce et al., 2017; Galli et al., 2019; Pizzi et al., 2017; Scognamiglio et al., 2018), which were carried out with static geodetic data and/or near-field strong motion waveforms and focused on revealing the rupture pattern. Yet, the frequency-dependent characteristic of the rupture has not been investigated. In this paper, we first introduce the two-stage strategy of finite fault inversion and expand it to include multi seismic datasets, and then we apply this strategy to analyze the 30 October 2016 Norcia earthquake.

2. Methods

The seismic rupture of one earthquake can be expressed as the temporal and spatial distribution of slip on the fault plane where the earthquake ruptures (Ide, 2007). In finite fault inversion, the fault plane can be discretized into a regular gridwork of M (along-strike) by N (down-dip) subfaults, then the seismic waveform recorded at one station, $u(x,t)$, can be written as the superposition of the contribution from all the $M \times N$ subfaults (e.g., Hartzell & Heaton, 1983; Ji et al., 2002; Olson & Apsel, 1982):

$$u(x,t) = \sum_{m=1}^M \sum_{n=1}^N \mu_{mn} A_{mn} D_{mn} \left[G_{mn}^1(V_{mn}, t) \cos(\theta_{mn}) + G_{mn}^2(V_{mn}, t) \sin(\theta_{mn}) \right] * S_{mn}(t), \quad (1)$$

in which m and n denote the m th along-strike and n th down-dip subfault; μ_{mn} and A_{mn} are the shear modulus and area of the subfault mn , which are known once given the velocity model of the source region and the discretization of the fault plane; G_{mn}^1 and G_{mn}^2 are the Green's functions of the subfault mn in the along-strike and down-dip directions, respectively; D_{mn} , θ_{mn} , V_{mn} , and $S_{mn}(t)$ are the slip amount, rake angle, averaged rupture velocity from the hypocenter to the subfault mn , and normalized slip-rate function, respectively. To simplify the slip history, the normalized slip-rate function $S_{mn}(t)$ can be represented by a trigonometric function, such as an asymmetric cosine function suggested by Ji et al. (2003):

$$S(t) = \begin{cases} [1 - \cos(\pi t / t_s)] / (t_s + t_e) & 0 < t < t_s \\ [1 + \cos(\pi(t - t_s) / t_e)] / (t_s + t_e) & t_s \leq t \leq t_s + t_e \\ 0 & t > t_s + t_e \end{cases} \quad (2)$$

in which t_s is called the starting time (i.e., the duration from the rupture initiation to the time with the maximum slip velocity), t_e is called the ending time (i.e., the duration from the time with the maximum slip velocity to the rupture termination), and $t_s + t_e$ is the time length of the slip-rate function (i.e., the whole slip duration). These two parameters can help to model the asymmetric slip history (Ji et al., 2003). Therefore, we can use D , θ , V , t_s , and t_e to describe the slip distribution on the fault plane, i.e., the unknown parameters to be solved in finite fault inversion.

Since $S(t)$ is a nonlinear function of t_s and t_e (Equation 2), Equation 1 becomes a nonlinear problem, which can be expressed as:

$$\mathbf{d}^{obs} = G(\mathbf{m}), \quad (3)$$

in which $\mathbf{d}^{obs}$ is the vector of observed seismic waveforms and $\mathbf{m}$ is the vector of the unknown parameters (D , θ , V , t_s , and t_e).

In our previous research (W. Liu & Yao, 2018), a two-step strategy was introduced to the dataset of teleseismic waveform inversion (see Text S1), which had successfully obtained the multi-scale rupture process of the 2015 Mw 7.8 Nepal earthquake. Since the rapid development of geophysical observation techniques,

many other datasets are made available for earthquake source studies, such as the near-field strong motion waveforms, Global Navigation Satellite System (GNSS) and Interferometric Synthetic Aperture Radar (InSAR) observations, and tsunami waveforms. Based on the nature of these observations, they have different characteristics and sensitivities to the earthquake rupture parameters. The teleseismic waveforms are dominated by relatively lower-frequency ground motion, while the near-field strong motion waveforms contain more higher-frequency radiation from the source. The geodetic data (e.g., GNSS and InSAR) can well resolve the surface deformation produced by one earthquake, but do not have a temporal resolution of the rupture. Jointly involving multi datasets into finite fault inversion can therefore improve the spatial and temporal resolution to the rupture process (Yue et al., 2014). Guided by a similar idea, we expand our two-step strategy to include another key dataset, the near-field strong motion waveforms, to jointly constrain the rupture process of earthquakes in finite fault inversion.

By including the near-field strong motion data, Equation 3 becomes:

$$\begin{pmatrix} \mathbf{d}_{tel}^{obs} \\ \mathbf{d}_{str}^{obs} \end{pmatrix} = \begin{pmatrix} G_{tel} \\ G_{str} \end{pmatrix}(\mathbf{m}), \quad (4)$$

in which $\mathbf{d}_{tel}^{obs}$ and $\mathbf{d}_{str}^{obs}$ are the observed teleseismic and near-field strong motion waveforms; G_{tel} and G_{str} are the forward operators to generate synthetic teleseismic and strong motion waveforms for a given set of model parameters.

As this is a nonlinear inversion problem, we use a Simulated Annealing (SA) method (Rothman, 1986; Sen & Stoffa, 2013; Tarantola, 2005) to find the global optimal solution in the wavelet domain, i.e., we construct the objective function in the wavelet domain and search for the solution with the global minimum of the objective function. The initial model is generated randomly. At the i th iteration, the objective function $\varphi^i(\mathbf{m})$ can be expressed as:

$$\varphi^i(m) = \sum_{j=j_{min}}^{j=j_{max}} W_{j;tel}^i \times \|\alpha_{j;tel}^{obs} - \alpha_{j;tel}^{syn}\|_L + W_{j;str}^i \times \|\alpha_{j;str}^{obs} - \alpha_{j;str}^{syn}\|_L + Constraints, \quad (5)$$

in which $W_{j;tel}^i$ and $W_{j;str}^i$ are the weights of the j th scale in the wavelet domain for the teleseismic ($\alpha_{j;tel}$) and near-field strong motion ($\alpha_{j;str}$) data, respectively; j_{min} and j_{max} are the minimum and the maximum wavelet scales used. Here we use the combination of the $L1$ and $L2$ norms to measure the data misfit in the wavelet domain, as suggested by Ji et al. (2002):

$$\|\alpha_j^{obs} - \alpha_j^{syn}\|_L = \frac{1}{2^j} \sum_{k=0} \left| \alpha_{j,k}^{obs} - \alpha_{j,k}^{syn} \right| + \sqrt{\frac{1}{2^j} \sum_{k=0} (\alpha_{j,k}^{obs} - \alpha_{j,k}^{syn})^2}. \quad (6)$$

In our two-step strategy, these weights ($W_{j;tel}^i$ and $W_{j;str}^i$) can be varied with the iteration index i (see Text S1). Compared with equation S4, involving the near-field strong motion dataset is equivalent to introduce one term that fits the strong motion data in the objective function (Equation 5). Therefore, we can conveniently involve any other time-series datasets (e.g., tsunami waveforms), the procedure is not much different from involving the near-field strong motion waveforms.

To study the frequency-dependent characteristic of the seismic rupture, the slip model should be resolved with seismic waveforms in different frequency bands (e.g., Ide, 1999; Suzuki & Iwata, 2009; Zeng et al., 1993). In our two-step strategy, the lower- and higher-frequency slip models can be obtained conveniently by modifying the weights of different wavelet scales ($W_{j;tel}^i$ and $W_{j;str}^i$). In the first stage of our strategy, the weights of higher wavelet scales (corresponding to higher frequency bands) are set to 0 and only lower-frequency waveforms are considered to resolve a lower-frequency slip model. Then in the second stage, the weights of higher wavelet scales are increasing to 1 (while the weights of lower wavelet scales are decreasing to 0) to resolve a higher-frequency slip model. Following Suzuki and Iwata (2009), in the second stage of our strategy, we resolve the slip pattern (D , θ , t_s and t_e) while fixing the rupture velocity (V) to the lower-frequency slip model obtained in the first stage.

Figure 2. Examples of the wavelet transform and data analysis. (a)–(d) The wavelet transform applied to the teleseismic waveforms (a)–(b) and the strong motion waveforms (c)–(d). Waveforms for each scale are shown in (a) and (c), and the wavelet coefficients for each scale are shown in (b) and (d). Bottom rows show the original waveforms, with the station name indicated to the right. The other rows show the waveforms on different scales. The frequency range of the j th scale is shown to the left, and the maximum amplitude is indicated to the right. (e) Comparison among the maximum amplitudes of the waveforms at different scales, with cyan lines indicating the teleseismic waveforms (TEL) and gray lines denoting the strong motion waveforms (STR). Notice that TEL and STR have different units (TEL: $\mu\text{m/s}$, STR: cm/s). Bold cyan and black lines show the average values in different scales of TEL and STR, respectively.

3. Application to the 30 October 2016 Mw 6.5 Norcia Earthquake

3.1. Datasets

To understand the frequency-dependent rupture characteristics of moderate size earthquakes, we applied the two-step strategy to the 30 October 2016 Mw 6.5 Norcia earthquake, for which both teleseismic and near-field strong motion data are available with high quality. The teleseismic P waves recorded by Global Seismographic Network (GSN) stations were downloaded from the Incorporated Research Institutions for Seismology (IRIS). After we removed the instrument responses to velocity measurements, the teleseismic P waveforms were then filtered to 0.02–0.625 Hz and resampled with a 0.2-s sampling interval.

The near-field strong motion waveforms in three components (north, east, and vertical) recorded at stations of the Rete Accelerometrica Nazionale (RAN) and the Rete Sismica Nazionale (RSN) were downloaded from the Italian Accelerometric Archive (ITACA). The near-field strong motion waveforms provided by ITACA were in accelerations. After integrated into velocity measurements, the strong motion waveforms were then filtered to 0.02–0.625 Hz and resampled with a 0.2-s sampling interval.

By applying the Meyer-Yamada wavelet (Ji et al., 2002; Yamada & Ohkitani, 1991; Yomogida, 1994), the waveforms can be transformed into different scales, from lower to higher frequency ranges (Figure 2). We compared the maximum amplitudes of all the waveforms at different scales (Figure 2e) and found that for most teleseismic waveforms, the amplitudes in scales 5–6 (corresponding to 0.078–0.312 Hz) are much larger than other scales, while for most strong motion waveforms, the amplitudes in scales 6–7 (corresponding to 0.156–0.625 Hz) are much larger. This is consistent with our knowledge that the teleseismic waveforms are dominant in lower-frequency information, while the strong motion waveforms contain more

higher-frequency information. By jointly using the seismic observations at teleseismic and near-field distances, a more detailed rupture model is expected to be obtained.

3.2. Velocity Model

Here we adopted Multitel3 (Qian et al., 2017) and FK (Zhu & Rivera, 2002) methods to calculate the theoretical Green's functions at teleseismic and near-field distances, respectively. Multitel3 is a hybrid method to calculate the theoretical Green's functions at teleseismic distances, which is based on the propagator matrix (for the source and receiver responses) and ray theory (for the Earth's mantle response) (see Qian et al., 2017 in detail). FK is a widely used frequency-wavenumber integration method to calculate the theoretical Green's functions at near-field distances. Here, a 1-D velocity model of central Italy from Ameri et al. (2012) (Figure S1) was adopted in the FK method, as well as used to calculate the source response in the Multitel3 method. The receiver and Earth's mantle responses in the Multitel3 method were calculated from IASP91 (Kennett & Engdahl, 1991).

To testify the reliability of the 1-D velocity model used in the source region, we chose 3 M–5 events nearby (Figure S2a), determined their moment tensors by gCAP (Zhu & Ben-Zion, 2013) method, and compared observed data and synthetic waveforms in the frequency band used in the finite fault inversion (see Text S2). The moment tensor solutions we obtained are similar to those from the TDMT catalog (Scognamiglio et al., 2006) (Figure S2 and Table S1) and the fitting of waveforms is reasonably well (Figures S3–S11), suggesting the 1-D velocity model we adopted can synthesize the near-field waveforms well in the target frequency band (0.02–0.625 Hz). Based on the waveform fitting and azimuthal distribution, 19 near-field strong motion stations with epicentral distances less than 60 km were selected to further study the rupture process of the 30 October Mw 6.5 event (Figure 3a) (see Text S2). For the teleseismic recordings, 31 P waveforms were selected according to the signal-to-noise ratio and the azimuthal distribution (Figure 3b).

3.3. Inversion Settings

Previous researches suggest that the 30 October 2016 Norcia earthquake ruptured two fault planes, the main fault plane is associated with the Mt. Vettore-Mt. Bove fault system (VBFS in Figure 1) and the ancillary fault plane is associated with the Olevano-Antronoco-Sibillini Mts. thrust (OAST in Figure 1) (e.g., Bonini et al., 2019; Cheloni et al., 2017; Scognamiglio et al., 2018). Referring to Scognamiglio et al. (2018), we considered the geometry of the two causative faults as the main plane (hereinafter named as the F1 plane) with the strike of 155° and dip of 47° and the ancillary plane (hereinafter named as the F2 plane) with the strike of 210° and dip of 36° (Figure 3a). These two fault planes were divided into 17 × 11 and 7 × 8 subfaults with dimensions of 2 km along strike and 2 km along dip, respectively. During the inversion, the slip amplitude was allowed to vary from 0 to 3 m, and the rake angle was from 240° to 300° for the F1 plane and 320°–380° for the F2 plane. The starting and ending time of the slip rate function (t_s and t_e) were allowed to vary from 0.4 to 2.0 s at a 0.2-s interval, respectively. The default rupture velocity was set to 2.5 km/s and was allowed to vary from 1.0 to 3.0 km/s. The rupture was set to start on the main plane, propagating from the source of this event, and the initial rupture on the ancillary plane was set to 0.8 s delayed, referring to Scognamiglio et al. (2018).

According to the sampling interval, the signal-to-noise ratio, and the estimated duration of the rupture, as well as considering the potential local site effects in central Italy (Bindi et al., 2011; Luzi et al., 2019; Tinti et al., 2016), we chose scales 3–7 (corresponding to 0.02–0.625 Hz) for both the teleseismic and strong motion waveforms (Figure 2). We then split these scales ($j = 3$ –7) to lower and higher frequencies based on the analysis of the amplitudes of the waveforms in different scales. As shown in Figure 2e, for the teleseismic waveforms, the amplitude is largest in scale $j = 6$ and decreases quickly in scales $j > 6$, indicating the energy is mostly concentrated in scales $j \leq 6$. While for the strong motion waveforms, the amplitude is also the largest in scale $j = 6$ but decreases not so fast in scales $j > 6$. Therefore, we considered scales $j = 3$ –6 as the lower frequencies, while scales $j = 7$ as the higher frequencies.

In our two-step strategy, we first used scales less than 7 ($j = 3$ –6, corresponding to 0.02–0.312 Hz) of both datasets to perform the lower-frequency inversion (i.e., for both datasets, the weights of scales 3–6 were set to 1 and the weights of scale 7 were set to 0). After the inversion converged (i.e., the variation of the objective

Figure 3. Stations used in finite fault inversion. (a) Inverted triangles show the 19 near-field strong motion stations. The yellow star indicates the epicenter of the 30 October Mw 6.5 earthquake. Colored planes indicate the fault geometry adopted in finite fault inversion, with the color scaling to the depth. Also shows the coseismic surface rupture trace (red lines), the major thrusts (white lines), and normal faults (black lines) (see Figure 1 in detail). (b) Distribution of the 31 teleseismic stations selected.

function $\left| \phi^i(\mathbf{m}) - \phi^{i-1}(\mathbf{m}) \right| / \phi^{i-1}(\mathbf{m})$ is less than 0.05% after 50 iterations), we then considered higher scales ($j = 7$, corresponding to 0.312–0.625 Hz) to perform the higher-frequency inversion (i.e., the weights of scale 7 were set to 1 and the weights of scales 3–6 were set to 0).

To stabilize the inversion, we also adopted three types of model constraints: the first is to minimize the difference between the slip along adjacent subfaults with a Laplacian operator (Ji et al., 2002); the second is to minimize the difference between the total moment calculated from the resolved slip model and a priori one (Hartzell et al., 1996), such as the Global Centroid-Moment-Tensor (GCMT) solution; and the third is a temporal constraint to compress the irresolvable roughness of the rupture front (Shao et al., 2011). Referring to previous applications (e.g., Ji, et al., 2003; W. Liu & Yao, 2018, 2020; Shao et al., 2011), the weight between these three constraints and the fitting of waveforms was set to 0.1.

4. Results

Figure 4 shows the recovered slip models from the first (lower-frequency inversion) and the second (higher-frequency inversion) stages of our strategy, respectively, jointly using both the teleseismic and near-field strong motion waveforms. The distribution of slip amount, slip duration, and rupture velocity is also shown in Figure 5. The lower-frequency slip model (hereinafter named as the LF model) shows a slip concentrated at the shallow depths (0–10 km) above the source on both F1 and F2 planes, with the maximum slip amplitude of 1.7 m (Figures 4a, 4d, and 5a). The slip distributed along the strike on the F1 plane (Figure 4a), consistent with the coseismic surface rupture trace collected by Villani et al. (2018). However, the slip pattern revealed by the higher-frequency slip model (hereinafter named as the HF model) shows some slip distributed at two sides of the lower-frequency slip (Figures 4b, 4e, and 5b), with the maximum slip amplitude

Figure 4. Map view of jointly inverted slip models of the 30 October Mw 6.5 earthquake recovered on the F1 and F2 planes. (a), (d) Slip model was recovered by jointly using lower-frequency (0.02–0.312 Hz) teleseismic and strong motion waveforms (the LF model). (b), (e) Slip model was recovered by jointly using higher-frequency (0.312–0.625 Hz) teleseismic and strong motion waveforms (the HF model). (c), (f) Contours of the LF (cyan) and HF (purple) models at a 0.3-m interval. For (a), (b), (d), and (e), the moment rate function is shown on the lower left as an inset (the black line indicates the total moment rate from the two planes, and the blue filled region indicates the moment rate from one single plane). For each panel, the magenta star indicates the epicenter of the 30 October Mw 6.5 earthquake, boxes indicate the fault planes (with the magenta edge show the upper limit of the fault plane). White squares indicate the location of some towns in this region. Gray dots show the seismicity after the 30 October Mw 6.5 earthquake until January 2017. Also we show the coseismic surface rupture trace (red lines), the major thrusts (white lines), and normal faults (black lines) (see Figure 1 in detail).

of 0.9 m. The contours of both LF and HF models at a 0.3-m interval (Figures 4c and 4f) indicate that the higher-frequency slip is mostly distributed in the region with less lower-frequency slip, which is consistent with the frequency-dependent rupture characteristics revealed in other moderate earthquakes that the higher-frequency slip is usually adjacent to the lower-frequency slip (e.g., Ide, 1999; Nakahara et al., 2002; Zeng et al., 1993). Another interesting feature is that the lower-frequency slip region presents a faster rupture velocity (>2.8 km/s) than the surrounding regions, as shown in Figure 5e.

Comparison among observed data and synthetic waveforms generated from the LF model in the lower frequency band (0.02–0.312 Hz) shows good waveform fitting, with the root-mean-square (RMS) ~ 0.098 for teleseismic waveforms (Figure S13) and ~ 0.111 for strong motion data (Figure S14). In the higher frequency band (0.312–0.625 Hz), the predicted waveforms generated from the HF model are acceptable, although the RMS values are higher than in the lower frequency band (~ 0.171 for teleseismic waveforms, Figure S15; ~ 0.273 for strong motion data, Figure S16).

We have also obtained the slip models from using one single dataset, that is, only using the teleseismic dataset (Figure S17) and only using the near-field strong motion dataset (Figure S18), respectively. The

Figure 5. Slip models recovered by performing jointly finite fault inversions with both teleseismic and near-field strong motion waveforms. The left column (a, c, e) shows the slip model obtained by using lower-frequency (0.02–0.312 Hz) waveforms, while the right column (b, d, f) shows the slip model obtained by using higher-frequency (0.312–0.625 Hz) waveforms. The two slip models presented here share the same colorbar. (a)–(b) Distribution of slip amount. White arrows indicate rake angles calculated clockwise from the direction of the strike. Contours indicate the time of rupture fronts at a 2-s interval. (c)–(d) Distribution of slip duration. Contours indicate the slip duration at a 0.8-s interval. Note that the smoothness of slip duration is not considered in the inversion, so the distribution of slip duration is heterogeneous although the slip distribution is smooth. (e)–(f) Distribution of rupture velocity. Contours indicate the rupture velocity at 0.4-km/s interval. For each panel, the magenta star indicates the source of the 30 October 2016 Mw 6.5 Norcia earthquake, and the magenta dot indicates the initial rupture point on the F2 plane. The X-axis is the distance along the strike, and the Y-axis is the depth from the surface.

lower-frequency slip patterns revealed in these LF models look similar, sharing the feature that concentrates at the shallower depths (Figures 5a, S17a, and S18a). However, the higher-frequency slip pattern obtained from only using the teleseismic dataset is different from using the near-field strong motion dataset (Figures 5b, S17b, and S18b), especially at the right side on the F1 plane, where the model from only using the teleseismic dataset shows less slip. We consider the difference presented in the three HF models may be due to the difference of resolution between the teleseismic and near-field strong motion waveforms, especially in the higher frequency band (>0.3 Hz) (e.g., Wald & Somerville, 1995; Yokota et al., 2011; Yue et al., 2014). Based on the recovered slip models, we generated and compared the synthetic waveforms at both teleseismic and near-field distances (Figures S19–S22). In the lower frequency band, all the three LF models can reasonably predict the waveforms (Figures S19 and S20), as they shared a similar slip pattern. While in the higher frequency band, due to the limited data resolution, only when both datasets are included in the inversion, the recovered HF model can predict the waveforms better (Figures S21 and S22). Hence in the following discussion, we focus on the recovered models from both the teleseismic and near-field strong motion waveforms (Figure 5).

5. Discussion

5.1. Evaluation of the Uncertainty of the Recovered Slip Models

Before we further discuss the different rupture characteristics revealed in the LF and HF models, we need to evaluate the uncertainty and robustness of these slip models. In the first stage of our two-step strategy, we let the inversion terminate and start the second stage automatically after it reaches a convergence, i.e., the variation of the objective function is less than 0.05% after 50 iterations. Similarly, the inversion also stops automatically when it converges in the second stage. For each stage, we can obtain a subset of the last 10,000 slip models that have been found during the SA procedure. Such a subset can be used to estimate the uncertainty of the inversion, to see what features can be revealed reliably. We calculated the standard deviation of each inverse parameter for both subsets of slip models in the two stages. Besides, we also used a normalized standard deviation to show the uncertainties more directly. For each subfault l , we weighted the standard deviation of all the five inverse parameters (i.e., slip amount, rake angle, rupture velocity, and the starting and ending time of the slip rate function) to obtain a normalized standard deviation $\bar{\sigma}_l$ by (W. Liu & Yao, 2020):

$$\bar{\sigma}_l = \frac{1}{5} \sum_{j=1}^5 w_j \sigma_{lj}, \quad (7)$$

here σ_{lj} is the j th inverse parameter on the l th subfault, and w_j is the weight of the j th inverse parameter, which was set to be half the range of the j th inverse parameter.

The calculated standard deviations for both stages are shown in Figures S23 and S24, respectively. The results present acceptable uncertainties in most parts of the model, except near the edges of the model, especially at the shallowest and deeper depths. Moreover, the standard deviations obtained in the first stage are larger than in the second stage, this is because the first stage started from a randomly generated initial model, while the second stage started from the best model obtained in the first stage.

To test the robustness of the slip models, we have conducted the bootstrap procedure to subsample the datasets. For each recovered slip model $\hat{\mathbf{m}}$ (the LF and HF models), we generated a dataset $\mathbf{d}_b$ containing the synthetic teleseismic and strong motion waveforms for all the stations we used by:

$$\mathbf{d}_b = \mathbf{G}(\hat{\mathbf{m}}), \quad (8)$$

then we produced 100 different data subsets independently based on $\mathbf{d}_b$. First, we randomly selected the teleseismic waveforms 40 times, picking one waveform from the total 31 P teleseismic waveforms at a time. Then we obtained 40 P teleseismic waveforms, containing some repeated waveforms. We checked and removed these repeated ones. Similarly, we randomly selected the near-field strong motion stations 25 times, picking one station from the total 19 stations at a time, and checked and removed the repeated ones. After that, we constructed a subset $\mathbf{d}_i$ from $\mathbf{d}_b$. Repeating these steps 100 times, we obtained 100 different data subsets. The histograms of the number of waveforms contained in these 100 data subsets are presented in Figure S25.

Then we performed finite fault inversion with these 100 different data subsets to obtain 100 slip models $\mathbf{m}_i$. Then, following Aster et al. (2013) we constructed a $q \times N$ matrix $\mathbf{A}$ with:

$$\mathbf{A}_{i,\cdot} = \mathbf{m}_i^T - \bar{\mathbf{m}}^T, \quad (9)$$

here $q = 100$, N is the number of inverse parameters, and $\bar{\mathbf{m}}$ is the average slip model. Finally, we obtained an estimate of the covariance matrix $\mathbf{C}$ by:

$$\mathbf{C} = \text{Cov}(\mathbf{m}) = \frac{\mathbf{A}^T \mathbf{A}}{q}, \quad (10)$$

and the standard deviation of each inverse parameter σ_k ($\sigma_k = \sqrt{\mathbf{C}_{k,k}}$).

By performing this procedure, we obtained an estimate of the uncertainty of the recovered slip model $\hat{\mathbf{m}}$ (the LF and HF models), as shown in Figures S26 and S27, respectively. The estimated uncertainties from the bootstrap procedure share a similar distribution with those estimated from the subsets of slip models presented in Figures S23 and S24, which show acceptable robustness except at the shallower and deeper depths. The standard deviations obtained from the bootstrap procedure are much larger, indicating that the data distribution can greatly affect the recovered slip model. Nevertheless, the major features obtained from the bootstrap procedure (Figure S28) are similar to those presented in Figure 5, suggesting that the recovered features are robust. Therefore, in the following discussion, we focus on the major features of the LF and HF models.

5.2. Frequency-Dependent Rupture Characteristics of the 30 October 2016 Mw 6.5 Norcia Earthquake

The slip models obtained from lower and higher frequency bands in this study show the frequency-dependent rupture characteristics of the 30 October 2016 Norcia earthquake. The major feature of the slip pattern recovered from the LF model shows a major slip patch at the shallower depths above the source (Figures 4 and 5), which is consistent with models proposed by previous studies with geodetic and/or strong motion data (e.g., Cheloni et al., 2017; Chiaraluce et al., 2017; C. Liu et al., 2017; Pizzi et al., 2017; Scognamiglio et al., 2018). The new point of our study is that we also obtained the distribution of higher-frequency slip, which is mostly adjacent to the region of lower-frequency slip (Figure 4).

To further discuss the temporal and spatial evolution of the rupture, we calculated the slip at different times. The snapshots of the slip at a 1-s interval are presented in Figure 6. We can find that after nucleation, the rupture propagated bilaterally to the left (northwestward) and the right (southeastward). The lower-frequency slip appeared at the shallow depths (<10 km) on both F1 and F2 planes, while the higher-frequency slip was widely distributed at depths. After ~ 4 s, the leftward rupture seemed to stop further leftward at ~ 10 km from the source but continued toward shallower depths, which seems like it met a “barrier”. Moreover, the higher-frequency slip appeared at the edges of the lower-frequency slip (Figure 4), which is consistent with the location of the “barrier”. On the other hand, the rightward rupture continued until ~ 10 s and ruptured both F1 and F2 planes. On the F1 plane, the lower-frequency rupture stopped at ~ 15 km from the source, while a large higher-frequency slip appeared at the edges of the lower-frequency slip. On the F2 plane, the lower-frequency slip is mostly concentrated within ~ 5 km along the strike from the initial rupture point, while the higher-frequency slip is mostly distributed right to the lower-frequency slip (Figure 6).

The higher-frequency slip mostly appeared at which the lower-frequency rupture had faded, such as the strong higher-frequency slip located at ~ 10 – 20 km from the source on the F1 plane (Figures 5 and 6). As suggested by analytic and numerical experiments, the stopping of rupture can generate great high-frequency energy radiation (e.g., Madariaga, 1977). This feature has also been revealed in some other earthquakes, such as the 2015 Mw 7.8 Nepal earthquake (Qin & Yao, 2017). Another feature is an asymmetrically bilateral rupture on the F1 plane. The northwestward rupture propagated ~ 10 km in ~ 4 s from the source, while the southeastward rupture propagated ~ 15 km in ~ 8 s (Figure 6). Moreover, the higher-frequency slip lasted until ~ 10 s and is stronger than that on the northwestern part (Figures 4 and 5).

Based on these features, the rupture of the 30 October 2016 Norcia earthquake can be summarized into several parts (Figure 7): the lower-frequency slip concentrated on the shallow parts of the F1 and F2 planes (named as the LF patch), and the higher-frequency slip appeared on the northwestern part of the F1 plane (named as the HF1 patch), appeared on the southeastern part of the F1 plane (named as the HF2 patch), and appeared at the edges of the LF patch on the F2 plane (named as the HF3 patch). The rupture nucleated and propagated quickly and steadily on the LF patch, radiating abundant lower-frequency energy. The higher frequency generated from the HF1, HF2, and HF3 patches may be associated with barriers to stop the rupture. Moreover, the HF2 patch may also be due to fault complexities, as it locates near the junction of the F1 and F2 planes (Figure 7). Previous researches have suggested that geometric discontinuities within complex fault systems can work as geometric barriers, which can contribute to the generation of high-frequency energy radiation (e.g., Huang & Yao, 2019; Okuwaki & Yagi, 2018).

Figure 6. Snapshots of the temporal and spatial evolution of the rupture obtained in this study. For each snapshot, red patterns show the lower-frequency (0.02–0.312 Hz) rupture, blue contours show the higher-frequency (0.312–0.625 Hz) rupture at a 20-cm interval. The gray dashed contour denotes the position of a reference rupture front with a rupture velocity of 2.5 km/s. The magenta star indicates the source of the 30 October Mw 6.5 earthquake, while the magenta dot indicates the initial rupture point on the F2 plane.

Mildon et al. (2019) simulated the seismic cycles and the Coulomb stress distribution in the central Italy normal fault systems. They suggested that there are two stress barriers on Mt. Vettore-Mt. Bove fault system (i.e., the F1 plane), which can restrain the propagation of seismic rupture. Figure 7 shows the rupture features of the 30 October event along with the rupture regions of two previous events of the 2016–2017 central Italy seismic sequence (i.e., the 24 August Mw 6.0 Amatrice earthquake and the 26 October Mw 5.9 Visso earthquake). The two stress barriers, as suggested by Mildon et al. (2017), locates nearby the boundaries between the 30 October event and the two previous events. These two previous events ruptured the southeastern and northwestern parts of the F1 plane, respectively, indicating that they released the stresses accumulated on these two regions during the interseismic period. In contrast, the central part of the F1 plane (i.e., the LF patch) remained not being ruptured during these two previous events due to the two stress barriers (Mildon et al., 2017, 2019). The lower-frequency slip inferred from our models filled in this

Figure 7. Schematic map of the frequency-dependent rupture characteristics of the 30 October 2016 Mw 6.5 Norcia earthquake. Orange and yellow blocks indicate the lower- and higher-frequency patches, respectively. Purple and cyan blocks indicate the rupture regions of the 24 August 2016 Mw 6.0 Amatrice earthquake (E1) and the 26 October 2016 Mw 5.9 Visso earthquake (E2), respectively (modified after Cheloni et al., 2017). Green lines denote the location of the two stress barriers suggested by Mildon et al. (2017). Magenta, purple, and cyan stars indicate the epicenter of these three events. Also we show the coseismic surface rupture trace (red lines), the major thrusts (white lines), and normal faults (black lines) (see Figure 1 in detail).

ation (Noda & Lapusta, 2010), while the HF1, HF2, and HF3 patches prefer to rupture in smaller slip and radiate more higher-frequency energy (Figure 8). Although the HF2 patch also possesses a higher pore fluid pressure, as it locates at the junction of the F1 and F2 planes, it can also generate higher-frequency energy radiation due to fault complexities. Furthermore, from the Vp/Vs distribution on the F1 plane (Figure 8c), we can find two lower Vp/Vs anomalies lying beside the LF patch, nearby the HF1 and HF2 patches, which implies an association with the two stress barriers to restrain the rupture propagation. This suggests that the heterogeneous distribution of pore fluid pressure may also be caused by fault complexities (Di Luccio et al., 2010). However, further investigations are needed to figure out this conjecture.

The unevenly distributed pore fluid pressure may also be associated with various hydraulic properties, such as permeability and diffusivity (Di Luccio et al., 2010; Malagnini et al., 2012). One of the possible mechanisms to produce heterogeneous hydraulic properties is the various rupture histories of different patches (Noda & Lapusta, 2010). As suggested, the hydraulic diffusivity and permeability within an undamaged region (e.g., the LF patch, which is not ruptured during previous earthquakes) are lower than within a fresh ruptured region (e.g., the HF1 and HF2 patches, which is located within the rupture regions of previous earthquakes) (Di Luccio et al., 2010; Malagnini et al., 2012).

The frequency-dependent rupture feature associated with heterogeneous hydraulic properties has been indicated in great earthquakes, such as the 2011 Tohoku-Oki earthquake (Wei et al., 2012). Wei et al. (2012) suggested the shallower fault portion (where larger smooth slip and lower-frequency radiation were

blank, suggesting that it released the accumulated stress on the LF patch. Moreover, the HF1 and HF2 patches locate nearby the edges of the LF patch are also consistent with the location of the two barriers (Figure 7).

The formation of the stress barriers may be associated with the variation of the fault geometry (Mildon et al., 2019). Mildon et al. (2019) investigated the along-strike fault bends and found that the variation of the fault geometry can significantly affect the simulated stress distribution. On the other hand, the variation of the fault geometry can also produce high-frequency energy radiation (e.g., Huang & Yao, 2019; Okuwaki & Yagi, 2018). Hence, the consistency between the higher-frequency slip patches and the barriers indicates that fault complexities may contribute to the frequency-dependent rupture of the 30 October 2016 Norcia earthquake.

5.3. Possible Mechanisms Associated with the Frequency-Dependent Rupture Features

In central Italy, the occurrence of seismic sequences is suggested to be associated with fluids (e.g., Chiarabba et al., 2018; Di Luccio et al., 2010; Malagnini et al., 2012). It is thought that CO₂-rich fluids uprising from a metasomatized mantle wedge (Chiarabba & Chiodini, 2013; Chiodini et al., 2011) can be trapped in regions confined by structural discontinuities, such as preexisting thrusts and normal faults in central Italy (Di Luccio et al., 2010). These fluids can cause an increment in pore fluid pressure within trapped regions and finally lead to seismic rupture.

Figure 8 shows the distribution of Vp/Vs in the rupture region of the 30 October 2016 Norcia earthquake. We can find that the LF patch locates in the region with higher Vp/Vs anomalies, while the higher-frequency patches mostly lie within the lower Vp/Vs region, or near the boundaries between the higher and lower Vp/Vs anomalies. Chiarabba et al. (2018) suggested that in this region, high Vp/Vs may indicate the existence of fluids, which is associated with a high pore fluid pressure. It is plausible that the LF patch sustains a higher pore fluid pressure and hence a weaker fault strength to produce larger slip and lower-frequency energy radiation

Figure 8. Comparison between the frequency-dependent rupture features of the 30 October 2016 Mw 6.5 Norcia earthquake and the Vp/Vs tomography from Chiarabba et al. (2018). (a)–(b) Two horizontal profiles at 6- and 9-km depths, respectively. Orange and yellow blocks indicate the lower- and higher-frequency patches, respectively. Purple and cyan blocks indicate the rupture regions of the 24 August 2016 Mw 6.0 Amatrice earthquake (E1) and the 26 October 2016 Mw 5.9 Visso earthquake (E2), respectively (modified after Cheloni et al., 2017). Green lines denote the location of the two stress barriers suggested by Milton et al. (2017). Magenta, purple, and cyan stars indicate the epicenter of these three events, respectively. (c) The distribution of Vp/Vs on the F1 plane (with strike of 155° and dip of 47°). Dashed orange and yellow lines indicate the lower- and higher-frequency patches, respectively. Green lines denote the location of the two stress barriers suggested by Milton et al. (2017). The magenta star indicates the source of the 30 October event, while purple and cyan stars show the projection of the source of the two previous events on the F1 plane, respectively.

observed) of the 2011 Tohoku-Oki earthquake may possess a lower hydraulic diffusivity to allow thermal pressurization to drive strong weakening, while the deeper portion with higher hydraulic diffusivity presented a rougher slip and higher-frequency radiation.

Our results seem to favor the idea that the frequency-dependent rupture is likely associated with heterogeneous hydraulic properties, suggesting that fluids may play an important role in the nucleation and rupture of earthquakes in the central Italy normal fault systems (Chiarabba et al., 2018; Di Luccio et al., 2010; Malagnini et al., 2012). Nevertheless, we could not rule out other factors, such as temperature variation due to frictional heating, and local lithology within the rupture region, which also determines the frictional

properties within the rupture zone (Noda & Lapusta, 2010). Further researches on dynamic rupture and tomography of fine fault structures may help to issue these questions.

6. Conclusions

Here we applied the two-step strategy of finite fault inversion proposed in our previous study (W. Liu & Yao, 2018) to multi datasets of the seismic observations at teleseismic and near-field distances. By the usage of the two-step strategy, here we studied the slip pattern of the 30 October 2016 Mw 6.5 Norcia earthquake, the largest event of the 2016–2017 central Italy seismic sequence, in different frequency bands. Our results reveal frequency-dependent rupture characteristics of this event, i.e., the lower-frequency slip concentrated on the shallower parts of the faults above the source, while the higher-frequency slip appeared near the edges of the lower-frequency slip region. The higher-frequency slip patches work as barriers, due to fault complexities in central Italy. Moreover, such a frequency-dependent rupture feature may also be associated with the unevenly distributed pore fluid pressure, as indicated by Vp/Vs tomography. Our results suggest that fault complexities and unevenly distributed fluids may contribute to the frequency-dependent rupture of this earthquake.

Data Availability Statement

The teleseismic waveform data are provided by Data Management Center (DMC) in Incorporated Research Institutions for Seismology (IRIS) (<https://ds.iris.edu/ds/nodes/dmc/data/>). The near-field strong motion waveform data are provided by the Italian Accelerometric Archive (ITACA) (<http://itaca.mi.ingv.it/>). The moment tensor solutions are provided by the European-Mediterranean Regional Centroid-Moment Tensors (RCMT) catalog (<http://rcmt2.bo.ingv.it/>) and the Time Domain Moment Tensor (TDMT) catalog (<http://inside.rm.ingv.it/en/tdmt>). The earthquake catalogs are from the Istituto Nazionale di Geofisica e Vulcanologia (INGV) (<http://inside.rm.ingv.it/en>). Topography data used in all the figures are from SRTM v4.1 (Jarvis A., H.I. Reuter, A. Nelson, E. Guea, 2008, Hole-filled seamless SRTM data V4, International Center for Tropical Agriculture (CIAT), available from <http://srtm.csi.cgiar.org/>). All the figures are plotted by Generic Mapping Tools (GMT) v4.5.13 (<http://gmt.soest.hawaii.edu/>).

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