Article

Study on Low-Frequency Abnormal Signal and Structural Characteristics of 2015 Azuoqi Ms5.8 Earthquake

Shasha Liang 1,2 and Haibin Li 1,*

1 College of Science, Inner Mongolia University of Technology, Hohhot 010015, China; liangshasha200611012@126.com
2 Seismological Bureau of Inner Mongolia Autonomous Region, Hohhot 010010, China
* Correspondence: lhb@imut.edu.cn

Abstract: In this article, the phenomenon of low-frequency abnormal signals before earthquakes, which reflects the three elements of earthquakes and the beneath structure change information, is discussed. Based on the data recorded at the Shizuishan (SZS), Wuhai (WUH) and Dongshenmiao seismic stations around the epicenter of the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015, the low-frequency abnormal signal from the seismic waves before this earthquake is extracted. At the same time, the autocorrelation method is used to extract the reflected waves of the main interface from teleseismic events recorded by the seismic array in the epicenter area, and then the change information from the beneath structure is obtained. It is explained in time and space that the low-frequency abnormal signal before the main earthquake, extracted from the continuous waveform, is directly related to the change in the underground structure near the epicenter, and it can be determined that the wave propagation direction of the crustal stress before the earthquake is from south to north, and it continues to accumulate near the epicenter until the main earthquake occurs.

Keywords: low-frequency signal; autocorrelation; underground structure; amplitude frequency envelope

1. Introduction

A large number of experimental studies show that rocks under high stress will experience the development process of “micro fracture initiation (micro scale and micro energy)—fracture propagation, penetration (small-scale and small energy fracture)—local rock instability (mesoscale and medium energy impact)—rock mass structure failure and large-scale instability (large-scale and large-energy release) [1]”.

One significance of extracting the low-frequency seismic signals is to explore the abnormal information in the above process. After the huge energy generated by the source is released, most of the high-frequency energy is absorbed by the medium or continuously weakened in the form of aftershocks, and a small part of the low-frequency energy has the characteristics of long wave length and slow attenuation, which is not easily absorbed and can be transmitted to thousands of kilometers away. Materials with strong high-frequency absorption will have positive prospects in seismic fortification [2,3].

The low-frequency signal from seismic waves also belongs to a kind of underground pulsation information, which may be the external manifestation of slow earthquakes (also known as intermittent deep tremors, slow slip events, low-frequency earthquakes, etc.) [4,5]. Compared with the conventional seismic radiation energy, the energy radiated from slow earthquakes is less than that from conventional earthquakes [6]. This micro fracture until rock mass instability and macro failure play an important role in the process of seismic nucleation [7,8]. The micro fracture behavior is different from that of rigid structural media [9]. The mechanism explanation of slow earthquakes may be a slow slip. This mechanism explanation has been proved by the rock fracturing in the laboratory [10,11], that is, the stable slip and stick slip phenomena before a rock fracture in the laboratory may correspond to the stable slip and stick slip phenomena near the subduction zone of a
seismic fault. The occurrence of these phenomena mainly depend on the critical stiffness coefficient of the fault [6]. When the fault is unstable or conditionally stable, it leads to the phenomenon of a slow slip. However, the mechanical model of moderately strong earthquakes in nature is very complex and may be triggered by the joint action of many factors [12]. The occurrence of a slow slip event may not necessarily produce a destructive earthquake, but if the slow slip event occurs in the locking area deep within a plate or fault, resulting in a continuous increase in stress accumulation in the locking area, a strong earthquake may be triggered. Extracting the low-frequency abnormal signals from the continuous seismic waveforms is an attempt to find the phenomenon of a slow slip caused by the continuous stress accumulation in this locked area. It is certain that the strongest earthquakes in history showed a locked state underground before the main earthquake: for example, the Wenchuan earthquake of magnitude 8 that occurred in the Chinese mainland on 12 May 2008, and the epicentral area of the Inner Mongolia Left Banner 5.8 earthquake on 15 April 2015. By studying the low-frequency abnormal signals from seismic waves before and after a main earthquake, we can identify the characteristics of a slow slip and of the energy release from the stress accumulation in the locking area [7,13]. Therefore, it is very necessary to make a retrospective study of historical moderate and strong earthquakes. This study can first determine the location of the locking area. If a slow earthquake occurs in the locked section of the fault, when the stress accumulates to the critical value, the locked section of the fault will be unlocked. At the moment of unlocking, the fault friction will suddenly decrease and produce rapid sliding or fracturing, which explains the hypothesis that a slow earthquake in the locked area will trigger the occurrence of a normal earthquake. If slow earthquakes occur in areas with nonlocking or developed fractures, the occurrence of slow earthquakes is only a process of normal stress release equilibrium, which can explain why there are no strong earthquakes in some areas even if there are slow earthquakes. In this way, in order to judge the impact of slow earthquakes or the low-frequency anomaly information about seismic waves on a region, it is necessary to first determine whether the region is a stress locking area. On this premise, the monitoring of slow earthquakes and the low-frequency anomaly information will have more practical significance.

In order to prove that the occurrence of the low-frequency abnormal signal in field seismic waves originates from the change in underground structure, while extracting the low-frequency signal from continuous seismic waveforms, this paper extracts the underground structure information from distant seismic waveforms according to the same frequency period and reflects the change characteristics of an underground structure in different time periods.

2. Extraction of Underground Structure

Formation wave impedance is restricted by two factors: lithology and physical properties. Lithology refers to the characteristics of a rock structure and its components, while physical properties refer to the characteristics of rock porosity, elasticity and fluid in the aquifer. Since the reflection coefficient can reflect the geological elements in the seismic records, we should extract the reflection coefficient series from the seismic records to obtain the geological information, and then use the obtained reflection coefficient series to convert the seismic waveform section into the wave impedance section, so as to comprehensively and accurately display most of the geological information contained in the seismic records [14,15]. The following figures (Figures 1 and 2) show the relationship between the seismic wavelet, reflection coefficient and wave impedance.
is the seismic wavelet, representing the reflection coefficient series and seismic trace. According to Formula (1), the geological information contained in the seismic record will only exist in the only two structural element carriers in the seismic wavelet or reflection coefficient [15,16]. It is impossible to separate every pure instantaneous reflection wavelet from seismic records to study the changes between the top and bottom reflection wavelets in a single target layer to obtain some geological information about the layer. In other words, it is not easy to obtain underground formation information from the seismic wavelet. Therefore, we can only start from another angle: the reflection coefficient series (Figure 2).

The reflection coefficient series depends on the wave impedance [15,16]. It is impossible to separate every pure instantaneous reflection wavelet from seismic records to study the changes between the top and bottom reflection wavelets in a single target layer to obtain some geological information about the layer. In other words, it is not easy to obtain underground formation information from the seismic waveform data with a one-hour window length, where the sampling rate is 0.01 Hz. This

For non-interference seismic traces, the results of a convolution in the seismic wavelet and reflection coefficient series are shown in Formula (1):

\[ s(t) = \omega(t) \ast R(t) \]  
(1)

where \( \omega(t) \) is the seismic wavelet, representing the reflection coefficient series and seismic trace. According to Formula (1), the geological information contained in the seismic record will only exist in the only two structural element carriers in the seismic wavelet or reflection coefficient [15,16]. It is impossible to separate every pure instantaneous reflection wavelet from seismic records to study the changes between the top and bottom reflection wavelets in a single target layer to obtain some geological information about the layer. In other words, it is not easy to obtain underground formation information from the seismic wavelet. Therefore, we can only start from another angle: the reflection coefficient series (Figure 2). The reflection coefficient series depends on the wave impedance difference between the upper and lower media at each formation interface, that is, it is determined by the wave impedance in the formation (the product of the wave velocity and density) [17].

The reflection coefficient of a single interface is expressed by Equation (2):

\[ R = \frac{[\rho_2 v_2 - \rho_1 v_1]}{\rho_2 v_2 + \rho_1 v_1} \]  
(2)

3. Data Processing

Two types of data were used for the study data. One is the continuous seismic waveform data with a one-hour window length, where the sampling rate is 0.01 Hz. This
type of data is used to extract the maximum value of a spectral envelope. The other is the data between 10 s before and 100 s after the arrival of teleseismic P wave, that is, the direct wave and its coda data, which is used to extract the information from the underground structure. In fact, the two different types of data are best recorded from the same dense array. For this article, the continuous seismic wave data are recorded by three fixed stations (DSM, WUH, SZS) around the epicenter, while the teleseismic data are recorded by the China Science Array [18].

3.1. Spectrum Envelope

Intuitively, the continuous waveform recorded by a seismometer with a window length of 1 h is in a basically stable form when no obvious events such as earthquakes or blasting are recorded, and the maximum value of the corresponding spectrum envelope amplitude fluctuates in a basically horizontal state [19]. Therefore, two key features are extracted when processing the continuous waveform. One is to obtain the spectrum from the 1 h continuous seismic waveform through a fast Fourier transform, and the other is to find the maximum amplitude of the spectrum envelope.

\[
x_i = \frac{a_0}{2} + \sum_{k=1}^{m} \left( \sqrt{a_k + b_k} \cos\left(\frac{2\pi ki}{N}\right) + \arctan\left(-\frac{b_k}{a_k}\right)\right)
\]

\[
a_0 = \frac{2}{N} \sum_{i=0}^{N-1} x_i, \quad a_k = \frac{2}{N} \sum_{i=0}^{N-1} x_i \cos\left(\frac{2\pi ki}{N}\right)
\]

\[
b_k = \frac{2}{N} \sum_{i=0}^{N-1} x_i \sin\left(\frac{2\pi ki}{N}\right), \quad k = 1, 2, 3, \ldots, m
\]

\{x_i\} is the continuous waveform data sequence recorded in 1 h, which is the amplitude spectrum. The curve formed by connecting the highest lines of the different frequency amplitudes from the obtained spectrum is called a spectrum envelope, as shown in Figures 3 and 4 [20].
3.2. Autocorrelation Extraction of Reflected Wave

The reflected P wave can be obtained by autocorrelation according to the vertical component of the continuous waveform recorded by the fixed station [21], but the calculation amount from this method is too large when extracting the reflected wave signal, resulting in low efficiency. In this article, the P wave, S wave and coda wave of the teleseismic events are used to extract the reflected wave, and the calculation efficiency is greatly improved. The core idea is that the reflection seismogram with a surface source and surface receiver can be obtained from the autocorrelation of the transmitted waves from the deep source and the receiver at the same position [22], that is, the autocorrelation of the vertically incident transmitted wave is equivalent to the reflected wave. The physical essence of this idea is to extract reflected waves from the multiples between a free surface and a discontinuity using autocorrelation [23]. In this way, the transmission problem of the teleseismic P wave and its coda is transformed into a reflection problem, and then the reflection coefficient of the layered media can be studied [24]. The process of constructing a station superposition autocorrelation map to extract the crustal reflection information can improve the consistency of the crustal reflection coefficient and suppress noise to a certain extent. Compared with an active source exploration, autocorrelation extraction technology, as passive source detection technology, makes up for the shortcomings of an active source exploration in cost and depth [25]. Compared with the traditional receiver function, the autocorrelation method can obtain deeper underground information [26,27].

The teleseismic data recorded by a seismometer includes the source information, transmission path information from a source to discontinuity, reflection and transmission information of multiples from a discontinuity to a free surface, in which the source information is negligible, so a non-interference seismic channel teleseismic record can be expressed in the frequency domain by Equation (6):

$$ W(\omega) = P(\omega)C(\omega) $$

From the relationship between autocorrelation and convolution, it can be seen that after autocorrelation:

$$ A(W(\omega)) = \left[W^T(\omega) \ast [W(\omega)]\right] = C^T(\omega) \ast P^T(\omega) \ast P(\omega) \ast C(\omega) = C^T(\omega) \ast C(\omega) $$

When the teleseismic P wave is vertically incident to the horizontal layer as the transmission wave, the source information as the point source can be ignored, and the
relationship between the transmission response and reflection response can be expressed by Equation (8):

$$T(t) * T^T(t) = (1 + R(t) + R^T(t))$$

The autocorrelation method is used to obtain the seismic reflection wave with the station as the virtual source and receiving point. The processing flow mainly includes four steps: pretrending, autocorrelation, moveout correction and stacked, as shown in Figure 5.

Figure 5. Schematic diagram of data processing flow.

Because this method uses the P wave of the teleseismic seismic events recorded by the vertical component and the seismic phases, such as global seismic phases PKP and PKIKP and their coda, to construct the reflected wave, the most critical step is NMO correction, which is used to eliminate the travel time difference of the seismic reflection phase, caused by the incident slowness, and correct the time difference of the rays to make them incident vertically [28].

According to the Taylor formula, Equation (9) can be obtained:

$$(1 + x)^m = 1 + mx + \frac{m(m - 1)}{2!}x^2 + \ldots + \frac{m(m - 1)\ldots(m - n + 1)}{n!}x^n + \ldots \approx 1 + mx + \frac{m(m - 1)}{2!}x^2$$

The specific determination of the moveout correction is as follows: it is assumed that the interface depth is $h_0$, the velocity of the overlying medium is $v_0$ and the two-way time of the vertical reflection is $\tau_0 = 2h_0/v_0$ when the source is located on the surface. The reflection time in $\tau - P$ domain can be expressed as Equation (10). Let $x = -\frac{v_0^2}{2}p^2$ and $m = \frac{1}{2}$ in Equation (9), so we can find the moveout correlation with the incident slowness $p$ is $-\frac{1}{2}v_0^2p^2$.

$$\tau = \tau_0(1 - \frac{v_0^2}{2}p^2)^{\frac{1}{2}} = \tau_0(1 - \frac{1}{2}v_0^2p^2 + \ldots) \approx \tau_0(1 - \frac{1}{2}v_0^2p^2)$$

4. Results
4.1. Low-Frequency Abnormal Signal of Seismic Wave

The maximum value of the amplitude frequency envelope is tracked for the continuous seismic waveform data recorded at the Wuhai (WUH), Shizuishan (SZS) and Dongsheng-miao (DSM) seismic stations from 0:00 on 1 April to 0:00 on 16 April 2015. The results show that the maximum values of the amplitude frequency envelopes of the three stations have obvious morphological differences (Figures 6–8). On the basis of the relatively gentle changes since 1 April 2015, the curve shape of WUH, closest to the epicenter of the Azuoqi Ms5.8 earthquake, began to fluctuate from 6:00 on 10 April 2015. It is considered that this morphology from WUH corresponds to the theory of fracture nucleation, which may be the manifestation of sub-instabibility. The result from DSM is relatively stable and the same sharp pulse signal as that from WUH before 6 April appears only at some individual times. Moreover, the intensity of this sharp pulse signal weakened after the continuous curve fluctuation at WUH on 6 April. It is speculated that in the process of the rock fracture and nucleation under WUH, the rock fissures are squeezed due to the increasing stress, so the low-frequency signal is blocked and cannot be better transmitted to the area below DSM. The difference in the low-frequency signals tracked by the three stations may be related to the epicentral distance. Assuming that a region is equipped with enough dense seismometers, the epicenter location of moderate and strong earthquakes can be predicted through this difference feature.
4.2. Feature Extraction of Stratigraphic Information

According to the existing seismogenic mechanisms such as slow earthquakes and micro fracture, we try further work to explore whether the medium under the monitoring station has also changed accordingly. When detecting the change in the underground media under the three stations, we use seismic waveform data less than 5 Hz. The data come from the teleseismic data recorded by the Chinese seismological array in western
Inner Mongolia (Figure 9). The analysis data are selected as teleseismic events within the range of 30–90° from the array in three different time periods: October 2013–October 2014, June 2014–April 2015 and April 2015–October 2016 (Figure 10). Select the teleseismic event waveform P wave and its coda ([-10,100] s waveform data), and then perform autocorrelation calculation after band-pass filtering on the selected data at [0.8,2] Hz, [0.5,4] Hz and [1.5] Hz. Finally, stack the autocorrelation results of the teleseismic events with different slowness. The results show that the structural characteristics recorded by the same station during the three different time periods have obvious changes, and the depth information recorded by the stations close to the source on the same survey line has obvious changes (Figures 11–14). In addition, the change degree of the underground structure at the different seismic stations within the same survey is also different. For example, 15673 has the most significant change in Line 4 (L4), which has the most prominent change from June 2014 to April 2015 (Figure 11). The most significant change in survey Line 6 (L6) is 64052, which has shown obvious underground structure changes from October 2013 to July 2014 (Figure 12). The most significant change in survey Line 7 (L7) is 15670, which has shown obvious underground structure changes from April 2015 to October 2016 after the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015 (Figure 13). The most significant change in survey Line 9 (L9) is 15597, which has shown obvious underground structure changes from June 2014 to April 2015 (Figure 14).

Figure 9. Seismic array near the Ms5.8 earthquake in Azuoqi, Inner Mongolia, on 15 April 2015.

Figure 10. Only teleseismic events 30–90° away from the array are selected.
Figure 10. Only teleseismic events 30–90° away from the array are selected.

Figure 11. L4 (from left to right: 15673, 15675, 15701, 15706, 15711, 15716, 15721, 15740 and 15735) reflects the depth results of different frequency bands in three time periods from October 2013 to July 2014, June 2014 to April 2015 and April 2015 to October 2016, respectively.

Figure 12. L6 (from left to right: 15618, 64052, 64054, 64051, 15716, 15717, 15718, 15719 and 15720) reflects the depth results of different frequency bands from October 2012 to July 2014, June 2014 to April 2015 and April 2015 to October 2016, respectively.

Figure 13. L7 (15670, 15675, 15681, 15688 and 15693 from left to right) reflects the depth results of different frequency bands from top to bottom in three time periods: October 2013–July 2014, June 2014–April 2015 and April 2015–October 2016, respectively.
Figure 12. L6 (from left to right: 15618, 64052, 64054, 64051, 15716, 15717, 15718, 15719 and 15720) reflects the depth results of different frequency bands from October 2012 to July 2014, June 2014 to April 2015 and April 2015 to October 2016, respectively.

Figure 13. L7 (15670, 15675, 15681, 15688 and 15693 from left to right) reflects the depth results of different frequency bands from top to bottom in three time periods: October 2013–July 2014, June 2014–April 2015 and April 2015–October 2016, respectively.

Figure 14. L9 (from left to right: 15597, 15711, 15712, 15713, 15714 and 15715) from top to bottom are the results of different frequency bands reflecting depth in three time periods: October 2013–July 2014, June 2014–April 2015 and April 2015–October 2016, respectively.

4.3. Relationship between Low-Frequency Signal and Beneath Structure

As described earlier, the morphological characteristics of the low-frequency signals recorded at DSM, WUH and SZS before the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015, are completely different, and the low-frequency abnormal signals lasting about 100 h before the main earthquake are extracted from the data recorded at WUH (Figures 6–8). Does the low-frequency abnormal signal form from the three seismic stations reflect the state and process of rock micro fracture below the epicenter? The different low-frequency anomaly patterns from the three seismic stations also reflect that the unstable change direction of the stress field is from south to north, that is, the unstable micro fracture from the stress is transmitted from below SZS to below WUH and then blocked under WUH. Until the main earthquake of the Ms5.8 earthquake occurs, the energy continues to be transmitted north to DSM. Therefore, we compared the positions of the temporary stations around the epicenter with DSM, WUH and SZS and found that the erection positions of the stations have the following characteristics: 64052 in L6 is equivalent to SZS, 15597 in L9 is equivalent to WUH, and 15670 in L7 is equivalent to DSM. The underground structure change characteristics of 64052, 15597 and 15670 represent the underground structure change characteristics of SZS, WUH and DSM, respectively. The results show that the underground structure of SZS (64052), located about 60 km from the epicenter of the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015, has changed from October 2013 to July 2014. Until the Ms5.8 earthquake in Azuoqi, Inner Mongolia, its underground morphological characteristics were constantly changing, which was consistent with the characteristics of the low-frequency signal and extreme instability of Shizuishan station (Figures 15 and 16).
4.3. Relationship between Low-Frequency Signal and Beneath Structure

As described earlier, the morphological characteristics of the low-frequency signals recorded at DSM, WUH and SZS before the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015, are completely different, and the low-frequency abnormal signals lasting about 100 h before the main earthquake are extracted from the data recorded at WUH (Figures 6–8). Does the low-frequency abnormal signal form from the three seismic stations reflect the state and process of rock micro fracture below the epicenter? The different low-frequency anomaly patterns from the three seismic stations also reflect that the unstable change direction of the stress field is from south to north, that is, the unstable micro fracture from the stress is transmitted from below SZS to below WUH and then blocked under WUH. Until the main earthquake of the Ms5.8 earthquake occurs, the energy continues to be transmitted north to DSM. Therefore, we compared the positions of the temporary stations around the epicenter with DSM, WUH and SZS and found that the erection positions of the stations have the following characteristics: 64052 in L6 is equivalent to SZS, 15597 in L9 is equivalent to WUH, and 15670 in L7 is equivalent to DSM. The underground structure change characteristics of 64052, 15597 and represent the underground structure change characteristics of SZS, WUH and DSM, respectively. The results show that the underground structure of SZS (64052), located about 60 km from the epicenter of the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015, has changed from October 2013 to July 2014. Until the Ms5.8 earthquake in Azuoqi, Inner Mongolia, its underground morphological characteristics were constantly changing, which was consistent with the characteristics of the low-frequency signal and extreme instability of Shizuishan station (Figures 15 and 16).

![Figure 15](image-url)

**Figure 15.** Low-frequency signal characteristics and 64052 in L6 line = SZS.

![Figure 16](image-url)

**Figure 16.** Variation characteristics of underground structure of 64052 (SZS) in different time periods.
The underground structure morphology of WUH (15597), which is located 42 km from the epicenter of the 2015 Azuoqi Ms5.8 earthquake in Inner Mongolia, did not change as significantly as that of DSM (15670) from October 2013 to July 2014. It only showed significant changes from June 2014 to April 2015. After the earthquake, the medium morphology recovered to a state similar to that of DSM (15670). The SZS appeared earlier than the WUH, in terms of the time when the instability of the underground structure began (Figures 17 and 18).

The underground structure of DSM (15670), which is located 158 km north of the epicenter of the 2015 Azuoqi Ms5.8 earthquake in Inner Mongolia, was basically the same from October 2013 to July 2014 and from June 2014 to April 2015, but there were obvious structural changes from April 2015 to October 2016. In other words, the underground structure of DSM changed significantly under the influence of the earthquake after the Ms5.8 earthquake in Azuoqi, Inner Mongolia, in 2015. In terms of time, the structural changes of DSM, WUH and SZS are different. From the perspective of the time of the structural change under the monitoring stations, SZS changes first, followed by WUH, and finally DSM. This feature shows that the stress transmission direction is from south to north, which is consistent with the northeast compression of the Qinghai Tibet block on the Ordos block in the southwest, and that the stress transmission direction reflected by the maximum tracking results from the seismic wave spectrum envelope (Figures 19 and 20).
In the process of rock fracture and nucleation beneath WUH, due to the increase in stress concentration, the rock fissures are squeezed so that the low-frequency signal is blocked and cannot be better transmitted to the position below DSM. The low-frequency stress concentration, the rock fissures are squeezed so that the low-frequency signal is blocked and cannot be better transmitted to the position below DSM. The low-frequency signal recorded by SZS has always been in a sharp pulse state, indicating the development of underground rock fractures, which is not conducive to stress accumulation and may be the transition zone of the energy transmission process.

In the process of rock fracture and nucleation beneath WUH, due to the increase in stress concentration, the rock fissures are squeezed so that the low-frequency signal is blocked and cannot be better transmitted to the position below DSM. The low-frequency signal recorded by SZS has always been in a sharp pulse state, indicating the development of underground rock fractures, which is not conducive to stress accumulation and may be the transition zone of the energy transmission process.

In the research on the underground structure of the epicenter of the 2015 Azuoqi earthquake with Ms5.8 in Inner Mongolia, on the one hand, it is studied according to the survey line profile, while on the other hand, it is studied one by one according to the stations. Using the data from different time periods before and after the main earthquake, it can be clearly seen that the underground structure has changed significantly and that the change characteristics of the underground structure under the stations at different locations from the epicenter before the main earthquake are different. This difference corresponds well to the low-frequency abnormal signal before the earthquake. The details are summarized as follows:

In the three seismic stations, WUH, SZS and DSM, near the Ms5.8 earthquake in 2015, the low-frequency abnormal signals extracted from the recorded seismic waveforms showed completely different forms before the earthquake, and the station closest to the epicenter, WUH, showed unstable fluctuation forms about 100 h before the earthquake. This may be the manifestation of the nucleation process in the metastable state of the source.

In the process of rock fracture and nucleation beneath WUH, due to the increase in stress concentration, the rock fissures are squeezed so that the low-frequency signal is blocked and cannot be better transmitted to the position below DSM. The low-frequency signal recorded by SZS has always been in a sharp pulse state, indicating the development of underground rock fractures, which is not conducive to stress accumulation and may be the transition zone of the energy transmission process.

According to the low-frequency abnormal signal forms recorded by SZS, WUH and DSM, it is judged that the underground stress transmission direction is from south to north and accumulates below WUH until the main earthquake occurs. From the change time of the underground structure, the results show that 64052 equivalent to SZS is the
station with the earliest underground structure change, followed by 15597 equivalent to WUH, and, finally, 15670 corresponding to DSM, which is consistent with judging the stress change direction of an underground structure according to the form of low-frequency abnormal signal.

Through the retrospective study of the 2015 Azuoqi Ms5.8 earthquake in Inner Mongolia, it is found that the underground structures near the epicenter changed significantly before the main earthquake, and this change even appeared 1–2 years before the earthquake. The low-frequency signal of a seismic wave has not been studied with a completely matched time period on the time scale, so it is impossible to state whether it changed during a longer time before the main earthquake. In particular, the fluctuation of the low-frequency abnormal signal before the earthquake at WUH may be the sub-instability state before instability and may be the extraterrestrial manifestation of a stick slip process in the sub-instability state.

Author Contributions: Conceptualization, S.L. and H.L.; methodology, S.L. and H.L.; software, S.L.; validation, S.L. and H.L.; formal analysis, S.L. and H.L.; investigation, S.L. and H.L.; resources, S.L.; data curation, S.L.; writing—original draft preparation, S.L.; writing—review and editing, S.L.; visualization, S.L.; supervision, H.L.; project administration, H.L.; funding acquisition, H.L. All authors have read and agreed to the published version of the manuscript.

Funding: This research was funded by the National Natural Science Foundation of China (grant number is 11962021) and was funded by the Earthquake Youth International Training Program (grant number is 201904190005).

Institutional Review Board Statement: This research work does not involve human and animal research.

Informed Consent Statement: Informed consent was obtained from all subjects involved in the study.

Data Availability Statement: The seismic waveform data provided by the China Seismic Scientific Exploration Array Data Center at the Institute of Geophysics of China Seismological Bureau for this research work. At the same time, Sichuan Seismological Bureau provided continuous seismic waveform data of 9 fixed seismic stations. Here, I would like to express my thanks.

Acknowledgments: I would like to thank B.L.N. Kennett and Weijia Sun for their guidance and help with this work.

Conflicts of Interest: The authors declare no conflict of interest.

References
1. Feng, D.; Chen, H.; Ding, W. Study on spectral anomaly characteristics of seismic waves before large earthquakes. Earthq. Res. 1994, 17, 319–329.
2. Singh, J.; Singh, C.; Kaur, D.; Narang, S.B.; Jotania, R.B.; Kagdi, A.; Joshi, R.; Sombra, S.; Zhou, D.; Trukhanov, S.; et al. Optimization of performance parameters of doped ferrite based microwave absorbers; their structural, tunable reflection loss, bandwidth and input impedance characteristics. IEEE Trans. Magn. 2021, 57, 2800619-19. [CrossRef]
3. Turchenko, V.A.; Trukhanov, S.V.; Kostishin, V.G.; Damay, F.; Porcher, F.; Klygach, D.S.; Vakhitov, M.G.; Lyakhov, D.; Michels, D.; Bozzo, B.; et al. Features of structure, magnetic state and electrodynamic performance of SrFe12-xInxO19. Sci. Rep. 2021, 11, 18342. [CrossRef] [PubMed]
4. Yin, H.; Zhou, H.; Xu, D.; Zhang, W.; Zheng, X.; Wang, S.; Gao, L. Detrital zircon U-Pb age, provenance and tectonic evolution significance of Yuanbaoshan formation on the southern margin of the Central Asian orogenic belt. Acta Geol. Sin. 2017, 91, 2196–2211.
5. Zhang, C.; Shi, Y.; Ma, L.I. Some problems in the study of slow earthquakes. J. Grad. Sch. Chin. Acad. Sci. 2005, 22, 258–269.
6. Yan, W.; Peng, H. Research status and significance of static/slow earthquakes. Geod. Geodyn. 2011, 31, 51–56.
7. Fei, C.W.; Liu, H.T.; Rhea Pliem Choy, Y.S.; Han, L. Hierarchical model updating strategy of complex assembled structures with uncorrelated dynamic modes. Chin. J. Aeronaut. 2021, in press. [CrossRef]
8. Dieterich, J.H. Earthquake nucleation on faults with rate-and state dependent strength. Tectonophysics 1992, 211, 115–134. [CrossRef]
9. Tang, L.; Li, S.; Su, F.; Sun, W.; Liu, J.; He, X. Study on long-period events before earthquakes—History and current situation. Int. Earthq. Dyn. 2002, 19, 48–57.
10. Han, L.; Wang, Y.B.; Zhang, Y.; Lu, C.; Fei, C.W.; Zhao, Y.J. Competitive cracking behavior and microscopic mechanism of Ni-based superalloy blade respecting accelerated CCF failure. Int. J. Fatigue 2021, 150, 106306. [CrossRef]
11. Ma, J.; Guo, Y. Laboratory evidence and earthquake examples of fault acceleration before instability. *Seismogeology* **2014**, *36*, 547–561.
12. Ma, J.; Sherman, S.; Guo, Y. Identification of metastable stress state before earthquake—Taking the experiment of deformation temperature field evolution of 5° inflection fault as an example. *Chin. Sci. Geosci.* **2012**, *42*, 633–645.
13. Yang, L.; Wang, J.; Feng, J.; Hu, Y.; Chen, J.; Yao, J. Preliminary study on low frequency fluctuation of ground pulsation before Wenchuan earthquake and its application. *China Earthq.* **2009**, *25*, 356–366.
14. Yang, L.; Hao, Z.; Wang, J.; Zhang, S.; Yao, J.; Dong, L. Preliminary study on Microwave phenomena of impending strong earthquakes (II). *China Earthq.* **2018**, *34*, 234–243.
15. Xu, Z. Seismic wavelet analysis. *Pet. Geophys. Explor.* **1982**, *17*, 1–15.
16. Xu, Z. Ultimate geological interpretation of seismic records—On the conversion from seismic waveform profile to wave impedance profile. *Geol. Explor.* **2013**, *33*, 23–28.
17. Xu, Z.; Geng, N.; Mei, S. Acoustic emission m value of rock fracture and rock mechanical properties. *Seism. Res.* **1990**, *13*, 291–296.
18. Kikuchi, M.; Kanamori, H. Inversion of complex body wave. *Bull. Seismol. Soc. Am.* **1982**, *72*, 491–506.
19. China Seismological Array. Waveform data of China Seismological exploration array. *China Seismol. Bur.* **2006**. [CrossRef]
20. Yang, L.; Mei, X.; Jiang, J. Spectral migration method for foreshock or generalized foreshock identification and its application. *China Earthq.* **2015**, *31*, 189–197.
21. Shasha, L.; Lixin, G.; Yong, D.; Gen, G.; Lei, W. Research on the Seismic Wave Characteristics of Low Frequency Signals before the Alxa Left Banner MS 5. 8 Earthquake in Inner Mongolia, China. *Earthq. Res. China* **2018**, *32*, 367–376.
22. Gorbatov, A.; Saygin, E.; Kennett, B.L.N. Crustal properties from seismic station autocorrelograms. *Geophys. J. Int.* **2013**, *192*, 861–870. [CrossRef]
23. Langston, C.A. Structure under mount rainier, washington, inferred from teleseismic bodywaves. *J. Geophys. Res.* **1979**, *84*, 4749–4762. [CrossRef]
24. Sun, W.; Fu, L.; Wei, W.; Tang, Q. A new seismic daylight imaging method for determining the structure of lithospheric discontinuity. *Sci. China Earth Sci.* **2019**, *62*, 473–488. [CrossRef]
25. Lei, C.; Liu, H. Study on the relationship between transmission and reflection response in passive source imaging. *J. Ocean. Univ. China* **2017**, *47*, 112–118.
26. Cheng, F. Passive Source Surface Wave Exploration Method and Its Application in Urban Areas. Ph.D. Thesis, China University of Geosciences, Wuhan, China, 2018.
27. Tauzin, B.; Pham, T.-S.; Tkalčić, H. Receiver functions from seismic interferometry: A practical guide. *Geophys. J. Int.* **2019**, *217*, 1–24. [CrossRef]
28. Kennett, B.L.N. *The Seismic Wavefield: Volume 1, Introduction and Theoretical Development*; Cambridge University Press: Cambridge, UK, 2001.