Near surface velocity and Q_s structure of the Quaternary sediment in Bohai basin, China*

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Abstract Heavily populated by Beijing and Tianjin cities, Bohai basin is a seismically active Cenozoic basin suffering from huge loss by devastating earthquakes, such as Tangshan earthquake. The attenuation (Q_p and Q_s) of the surficial Quaternary sediment has not been studied at natural seismic frequency (1–10 Hz), which is crucial to earthquake hazards study. Borehole seismic records of micro earthquake provide us a good way to study the velocity and attenuation of the surficial structure (0–500 m). We found that there are two pulses well separated with simple waveforms on borehole seismic records from the 2006 M_w 4.9 Wen'an earthquake sequence. Then we performed waveform modeling with generalized ray theory (GRT) to confirm that the two pulses are direct wave and surface reflected wave, and found that the average v_p and v_s of the top 300 m in this region are about 1.8 km/s and 0.42 km/s, leading to high v_p/v_s ratio of 4.3. We also modeled surface reflected wave with propagating matrix method to constrain Q_s and the near surface velocity structure. Our modeling indicates that Q_s is at least 30, or probably up to 100, much larger than the typically assumed extremely low Q (~10), but consistent with Q_s modeling in Mississippi embayment. Also, the velocity gradient just beneath the free surface (0–50 m) is very large and velocity increases gradually at larger depth. Our modeling demonstrates the importance of borehole seismic records in resolving shallow velocity and attenuation structure, and hence may help in earthquake hazard simulation.

Key words: Bohai basin; Quaternary sediment; near surface velocity; Q_s ; borehole records

CLC number: P315.3⁺1, P315.9

Document code: A

1 Introduction

The shallow velocity structure and Q value of sediments are important parameters for both seismological study and engineering purpose. The shallow velocity structure of the sediment has a considerable effect on the seismological wave-field by amplifying wave amplitude. However, low Q , which is typically assumed for near surface shear velocity, might suppress the amplification effect because of strong damping. Therefore tremendous efforts have been put into the study of shallow velocity and attenuation structure (Clouser and Langston, 1991; Kinoshita, 2008). Of course, well-logging from borehole samples provides direct and accurate measurements, yet this measurement may be of limited applicability because the frequency range is too large as compared to seismic frequencies of a few seconds to a few Hertz. Anderson et al (1996) found that the attenuation structure in the deeper deposits is a criti-

cal factor that seems to have a comparable influence in strength on the velocity in the upper 30 m. Fortunately, borehole seismic records from local micro earthquakes contain well separated phases which are sensitive to both velocity and Q structure from free surface to several kilometers. Consequently, borehole seismic waveforms have been extensively studied to derive near surface structures in seismic frequency band, which are more relevant to strong motion simulation (Hauksson et al, 1987; Kinoshita, 1994, 2008; Jongmans and Malin, 1995; Blakeslee and Malin, 1991).

However, different groups get fairly different or even contrasting models from borehole seismic records. Yamada and Horiike (2007) proposed that Q_s is negatively frequency dependent as $0.24v_s f^{-1}$, that is, $1/Q_s$ increases linearly with frequency. In contrast, Kinoshita (1994) demonstrated that $1/Q_s(f)$ decreases with frequency as $f^{-0.7/130}$. Recently, Kinoshita (2008) and Stork and Ito (2004) found that Q_s is larger than or equal to Q_p , confirming the tomography results by Hauksson and Shearer (2006). Although Q_p is usually larger than Q_s because shear motion is more dissipated, there are reports of Q_p lower than Q_s (Hauksson and Shearer, 2006).

* Received 25 April 2009; accepted in revised form 1 September 2009; published 10 October 2009.

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Also many other studies indicated that the Q values for both P and S waves showed substantial variability within range from extremely low Q (~ 10) to moderate values of about 25 (Boatwright et al, 1986; Hauksson et al, 1987; Malin et al, 1988; Seale and Archuleta, 1989; Blakeslee and Malin, 1991; Aster and Shearer, 1991). Langston even argued that Q_s must be significantly greater than 30 and is probably around 100 after detailed body wave analysis and surface wave modeling (Langston, 2002, 2003a, b; Langston et al, 2005). He attributed the low Q_s from spectral studies to neglect of complexity in velocity structure (Langston, 2003a, b). Therefore, there should be more studies for different sedimentary basins to understand near surface structure better, toward eventual goal of establishing a reliable attenuation and velocity model for shallow sediments.

Covered with thick Quaternary sediments, Bohai basin in northern China is seismically active with many earthquakes stronger than $M_L 6$ in the past centuries, with the 1976 Tangshan earthquake being the most devastating event. Bohai basin is also heavily populated by mega cities such as Beijing and Tianjin, precise strong motion simulation would be critical for seismic hazard mitigation. Thus, Bohai basin calls for reliable velocity and attenuation parameters, so that meaningful strong motion simulation can be performed. Though various studies have revealed crustal scales Q (scale of tens of kilometers) in northeastern China, few studies focus on the very shallow part of the sedimentary cover (top 1 km or so). Instead, attenuation studies from petroleum industry provides relevant information of Q_p of the sediments (Chen et al, 2001), where Q_p in earthquake frequency band (a few Hertz) turns to be less than 100 for the top 1 km. Fortunately, there are dozens of borehole seismometers installed at depth from 100 m to 500 m in this region of high seismicity, and large amounts of high quality borehole seismograms have become available for modeling near surface structure. As a pilot study in this region, we study local seismic records of the borehole seismometer at station WEA for the 2006 $M_W 4.9$ Wen'an earthquake sequence with generalized ray theory (GRT) and propagation matrix method. We found that Q_s is at least 30 and probably up to 100. Moreover, waveform modeling requires a large velocity gradient just beneath the free surface.

2 Data and analysis

At the station of WEA (Figure 1), a borehole seismometer of three components is installed at a depth of

266 m. The instrument well recorded waveforms of the earthquake on 4 July 2006 with the magnitude of $M_W 4.9$ and distance of about 20 km. The aftershock sequence is also well recorded because of low noise level of borehole seismometers (Figure 1, Table 1). To facilitate waveform analysis, horizontal components were rotated to radial and tangential directions to get SV and SH component. It should be done with caution while rotating the borehole records, due to the existence of miss-orientation or miss-location of the events. In our study, the optimal rotation angle is found by minimizing the tangential energy for the time window of 0.5 s before and 1.0 s after the P wave arrival. Another way to judge the quality of rotation is to check the amplitude of S-P phase converted at the interface between sediments and the basement rock (Langston, 2003a, b). After optimal rotation, S-P phase is hardly visible on SH components.

Figure 1 Earthquake distribution and station WEA. One of the open circles is the aftershock on July 19, 2006, whose waveform is modeled in this study with GRT method to confirm the direct phase and its surface reflected phase.

Table 1 Earthquakes used in this study

a-mo-d	Lat./°N	Long./°E	Depth /km	M_L	Distance /km
2006-05-13	38.92	116.67	2.0	1.8	20.07
2006-07-06	38.92	116.27	7.0	1.7	17.95
2006-07-19	38.88	116.25	5.0	1.9	18.13
2006-08-24	38.87	116.22	17.0	1.5	20.65
2006-10-28	38.58	116.37	17.0	2.4	30.22
2007-01-13	38.83	116.48	9.0	2.2	2.97
2007-04-18	38.58	116.33	30.0	1.8	31.09

Note: The earthquake catalog is from China Earthquake Networks Center (CENC, www.csendmc.ac.cn).

On the tangential components of all the events as shown in Figures 2 and 3, we observe two pulses with comparable amplitudes and the same polarity. The interval between these two pulses is almost identical for different events, suggesting structural origin. It is well established that the first pulse is direct S wave from the earthquake to the receiver while the second is due to surface reflection, here we denote them as S_1 and S_2 , respectively (Hauksson et al, 1987; Fukushima et al, 1992; Jongmans and Malin, 1995). Similarly, the direct P wave (P_1) and surface reflected P wave (P_2) are obvious on vertical components (Figure 3), except for the main shock as shown in Figure 2 where long source duration exceeds their interval.

In order to confirm the direct phases (P_1 , S_1) and surface reflected phases (P_2 , S_2) on the records, we performed a waveform modeling with GRT method. The

Figure 2 The waveforms of the Wen'an 4 July 2006 main shock recorded by WEA station. The waveforms are displayed in radial, tangential and vertical components. The direct waves (P_1 , S_1) and surface reflected waves (P_2 , S_2) are obvious.

Figure 3 Waveforms of direct wave and surface reflected wave of both P and S wave for the Wen'an sequences. Vertical P waveforms (a), radial (b) and tangential (c) S waveforms are shown. The direct waves (P_1 , S_1) and surface reflected waves (P_2 , S_2) are obvious.

GRT method is a technique that calculates seismic phases separately, and then obtains synthetic waveform by summing the individual phases (Helmberger, 1968, 1974). Thus, it has an advantage of identifying seismic phases on seismograms easily. Here we take the after-shock on July 19, 2006 as an example, and use GRT method to confirm the phases such as P_1 , P_2 , S_1 , S_2 and S-P. A simple velocity model is presented in Table 2. Between layers 1 and 2, an artificial discontinuity without velocity jump is set at depth of 266 m for receiver location. Phases such as P_1 , P_2 , S_1 , S_2 and converted phase S-P are calculated separately in forward modeling. Then these individual phases are summed to form syn-

thetic seismogram. From Figure 4, we can find that P_1 , P_2 , S_1 and S_2 are well modeled, though S-P on vertical component is not well modeled in amplitude due to inaccurate source parameters or improper velocity contrast at the basement. Therefore it is sure that the double pulses are S_1 , S_2 for S wave, and P_1 and P_2 for P wave.

After confirming the direct waves (S_1 , P_1) and surface reflected waves (S_2 , P_2), average velocities of both vertically traveling P and S waves are estimated by measuring differential times between P_1 and P_2 , also S_1 and S_2 . The differential time is two-way travel time of P or S wave in the layer above the receiver. Thus the average velocities of both P and S waves can be estimated

Table 2 Velocity model for forward modeling with GRT method

Layer	v_p /km·s ⁻¹	v_s /km·s ⁻¹	ρ /g·cm ⁻³	Thickness /km	Q_p	Q_s
1	1.83	0.42	1.786	0.266 0	100 0	500
2	1.83	0.42	1.786	0.034 0	100 0	500
3	4.2	2.3	2.146	4.300 0	100 0	500
4	5.0	2.9	2.370	8.200	100 0	500
5	6.1	3.5	2.720	7.000 0	100 0	500

Figure 4 Forward modeling of the July 19, 2006 aftershock with GRT method. The waveforms are displayed in radial, tangential and vertical components, with the thin ones as synthetic and thick ones as data. Receiver is below the first layer, and source depth is 12.0 km and the horizontal distance from source to receiver is 18.1 km.

with differential times and burial depth of the receiver. The average velocities from free surface to the depth of 266 m are found to be around 1.8 km/s and 0.42 km/s for vertically traveling P and S waves, respectively, and v_p/v_s is about 4.3. This is based on assumption that ray incidence is nearly vertical beneath receiver. Because velocities in the shallow layers are very low for both P and S waves compared with those in the basement, the upward and reflected waves travel almost vertically following Snell's law (Fukushima et al, 1992). In our study, the seismometer is installed at a much shallower depth and the velocities are even lower than those in the study of Fukushima et al (1992), so that the assumption of vertical rays of P and S waves is valid. From our result, the low velocity layer causes considerable time delay of both P and S waves though the layer is only 266 m in thickness. Thus, attention should be paid to especially

when traveling time of P or S is used to study the deeper structure or locate earthquakes.

Though S_1 and S_2 share the same polarity and similar shapes, we found that positive part of S_2 has much broader shape and smaller amplitude than that of S_1 , while the negative part of S_2 has larger amplitude than that of S_1 as shown in Figure 3. These features maybe originate from shear attenuation and the velocity structure.

The seismological methods including spectral ratio method (Sato, 2006; Kinoshita, 2008) and waveform fitting method (Yamada and Horike, 2007) are employed to study Q value based on borehole records mostly. In this paper, the waveform fitting method is employed to investigate the value of near surface Q_s at the site of WEA. The waveform fitting is based on propagating matrix method with vertical incidence of SH wave. Here we use the direct phase S_1 as source time function and reflected phase S_2 is chosen to be fitted by synthetic reflected wave. Since both the waveforms of direct and surface reflected phases are recorded by the same seismometer, there is no need to correct for instrument response (Kinoshita, 2008).

Considering waveform quality, we take the aftershock occurred on 6 July 2006 for waveform fitting study. A 0.72 s time window of S_1 on SH component is used as incoming planar wave under a layered velocity model, and reflected wave S_2 is calculated in frequency domain then transformed into time domain with FFT. Some typical velocity models are used in the waveform modeling of the reflected SH wave, such as models with near surface velocity gradient and some with low velocity zone (Figures 5, 6 and 8). Then the value of average Q_s (independent of frequency) is chosen from 10 to 120 with increment of 10.

First, the simplest model of constant velocity ($v_s=0.42$ km/s) is used to test the waveform fitting (Figure 5). In forward modeling without attenuation, timing of S_2 is well fitted, indicating that average velocity of vertically traveling S wave is appropriate (Figure 5b). Given the simplicity of the model, the overall waveform of S_2 is not well modeled. Then frequency-independent Q_s is applied, and we find that variation of S_2 with Q_s is obvious as in Figure 5c. From positive part of S_2 , we may estimate that the value of Q_s is at least 30 while neglecting the negative part of the reflected wave. The negative part of reflected wave will be discussed in the following sections.

Waveform fittings with different gradient velocity models are carried out with the same process as described

Figure 5 Forward modeling for the case of constant velocity. (a) Velocity model, the solid line is v_s , the dashed line is density ρ ; (b) Observed waveform (solid line) vs. synthetic waveform (dashed line) for the case of no attenuation; (c) Sensitivity of Q_s to synthetic waveforms of S_2 with Q_s varying from 10 to 120 with increment of 10. Q_s has to be at least 30 to make synthetic seismograms (dashed line) and data (solid line) similar in amplitude. But the large negative part of S_2 requires even higher Q_s .

Figure 6 Examples of waveform modeling with two gradient velocity models. The symbols are the same as in Figure 5. (a) The velocity model with gradient zone of 125 m (left) and its S_2 waveform (right) with Q_s varying from 10 to 120; (b) The gradient zone length is 266 m.

above. Two examples of waveform modeling with gradient model are shown in Figure 6. From these tests, we conclude that Q_s should be at least 30 and probably up to 100.

For these models, the thicknesses of gradient zone are taken as 25, 75, 125, 175 and 266 m. By comparing synthetic waveforms with different gradient thickness, we find that such kind of gradient has no significant effect on the reflected waveform especially when top velocity is fixed with $v_s=0.32$ km/s, and amplitude of negative part of S_2 can not be explained by Q_s alone, as

shown in Figure 7. Probably the velocity models analyzed above are too smooth in velocity change. However, the real model may be more complex. For the effect of gravity consolidation, velocity may increase rapidly with depth just near the free surface, and then at larger depth it increases more slowly. Also, the similarity between S_1 and S_2 and weak wiggles between them suggest that the velocity structure should not be too complex. So, for this situation, waveform of S_2 is mainly affected by the very top velocity structure.

Figure 7 Sensitivity of gradient zone length (25, 75, 125, 175 and 266 m) on S_2 waveforms. The velocity v_s at free surface is set to be 0.32 km/s. Solid line stands for observed waveform and dashed lines for synthetic waveforms according to different lengths of gradient zone. (a) For the case of no attenuation; (b) for the case of $Q_s=30$. Noticing the similarity of synthetics in both figures (a) and (b), S_2 has little sensitivity to the gradient zone lengths used above due to their smooth change in velocity.

Figure 8 Forward modeling with low velocity layer near free surface. (a) The velocity model; (b) Data (solid line) and synthetic (dashed line) S_2 with Q_s varying from 10 to 120. Near surface low velocity layer broadens the positive part of S_2 and increases the amplitude of the negative part.

In order to verify this hypothesis, we carried out some numerical tests, finding that low velocity zone or larger velocity gradient near surface is required to fit the waveform well, and velocity should increase gradually at larger depth. Near the surface, the low velocity layer will broaden positive part and increase amplitude of negative part of S_2 (Figure 8). Also we find that Q_s should be larger than 30 by changing Q_s value from 10 to 120 in the tests.

3 Discussion and conclusions

Borehole seismic records of local micro earthquakes are modeled to study the near surface velocities and Q_s at the site of WEA. First, a forward modeling with GRT method is carried out to confirm the direct wave and its surface reflected wave. Then average velocities of vertically traveling P and S waves are estimated by measuring differential times between direct wave and surface reflected wave. From our study, average velocities of vertically propagating P and S waves are about 1.8 km/s and 0.42 km/s, respectively. Thus v_p/v_s is about 4.3, which is much higher than 1.73 of Poisson solid. Similar ratio of v_p/v_s is also found by Langston (2003a). The near surface velocity of sediment basin is so low that waves will travel much slower in this layer, so attention should be paid to especially when traveling time of P or S is used to study the deeper structure in tomography modeling or locating earthquakes.

Finally, a waveform fitting method is employed to constrain Q_s , based on propagating matrix method with vertical incidence of SH wave. With a series of S_2 waveform fitting, we find that Q_s is at least 30, or probably up to 100, also the velocity gradient just beneath the free surface (maybe 0–50 m) is very large and velocity increases gradually at larger depth. Our result is consistent with the Q_s modeling in Mississippi embayment by Langston (2002), but much larger than the typically assumed extremely low Q (~ 10). And there may be some reasons for such difference of Q_s between different studies: ① For different regions, due to the deposit time or rock type, attenuation may be quite different. For example, Kinoshita (1994) found that $1/Q_s(f)$ decreases with frequency as $f^{-0.7}/130$ in southern Kanto area, while in Osaka basin Yamada and Horike (2007) proposed that Q_s is negatively frequency dependent as $0.24v_s f^{-1}$, that is, $1/Q_s$ is linearly increasing with frequency. ② Different methods may result in different

conclusions. Spectral method which neglects the complexity of velocity structure may get a Q_s value that is affected by complexity of velocity structure. Langston attributes the low Q_s from spectral studies to neglect of complexity in velocity structure, and scattering may be a significant factor in body-wave attenuation study (Langston, 2003a, b). With waveform fitting method which takes velocity structure into consideration, we will get a more reliable result. Through waveform fitting, we find that larger velocity gradient just beneath free surface broadens the waveform of S_2 , which has similar effect of Q_s , thus Q_s attenuation may be overestimated in structure-free attenuation studies. So this considerable effect of structure should be paid more attention to. ③ Some factors, such as fluid and temperature, may affect attenuation. Hauksson and Shearer (2006) believe that $Q_s/Q_p > 1$ suggests partially fluid-saturated crust while $Q_s/Q_p < 1$ may be related to high fluid saturation or elevated crustal temperatures.

In this paper, we just use the station of WEA as a case study, and there are many other borehole seismometers in Capital region of China, which will be used to study Q_s variation in this region to get overall understanding of the Q structure of this region so as to provide realistic parameters for ground motion simulations.

Acknowledgements This study is supported by National Natural Science Foundation of China (No. 40676067) and Knowledge Innovation Program of Chinese Academy of Sciences (No. kzcx2-yw-116-1). We are grateful to two anonymous reviewers for careful review and valuable comments, and also to China Earthquake Networks Center for providing us high quality data. Contribution No. is 2008-008 of Mengcheng National Geophysical Observatory.

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