Fault slip velocities inferred from the spectra of ground motion

Nergis Ani Anil-Bayrak
Iowa State University

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Fault slip velocities inferred from the spectra of ground motion

by

Nergis Ani Anil-Bayrak

A thesis submitted to the graduate faculty
in partial fulfillment of the requirements for the degree of

MASTER OF SCIENCE

Major: Geology
Program of Study Committee:
Igor A. Beresnev, Major Professor
Neal Iverson
Sri Sritharan

Iowa State University
Ames, Iowa
2008

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To the memories of my dad Dr. Ali Anil (1955-2006),
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Nergis Ani Anil-Bayrak
ABSTRACT

Better understanding of earthquake-source properties is an important goal in seismology. Dynamic fault theories and practices of ground-motion prediction need independent information about source characteristics obtained directly from recorded data.

The maximum slip velocity on a rupturing fault is the parameter that controls the strength of an earthquake’s high-frequency radiation and the properties of its Fourier spectra. We therefore have tested an empirical method for determining the peak slip velocities for a number of well-recorded earthquakes using such spectral information.

High-quality ground-motion data from small-to-moderate earthquakes in Japan were collected, and Fourier transforms of the accelerograms were computed for both horizontal and vertical components of the data. Regional parameters (site effects and path effects) that distort the true source spectra were investigated and separated from the recorded spectra. The obtained source terms following the classic “$\omega^2$” spectral model were used to determine the corner frequency that carries the information about the fault’s maximum velocity.

The results indicate that the maximum slip velocity of the selected Japanese earthquakes ranged from approximately 0.2 to 0.6 m/sec. Direct observation-based slip velocity determinations provide valuable physical information about earthquakes that
can be used for constraining dynamics theories of faulting or in ground-motion prediction.
1. GENERAL INTRODUCTION

1.1. Introduction

Seismology is an important branch of earth sciences that provides knowledge about the structure of Earth (Doyle, 1995). Earthquake recording, determination of earthquake locations and magnitudes, seismic-wave propagation, and earthquake effects on buildings are some of the interest areas of seismology. In addition to these, understanding the earthquake mechanism is one of the most important problems of modern seismology (Kasahara, 1981).

There is much practical need in obtaining independent information about earthquake-source dynamic properties directly from observable data. The maximum slip velocity during earthquake rupture is one such parameter. Such observations can be made from recorded data. For instance, the Fourier spectrum of the recorded ground-motion accelerograms provides significant information about earthquake-source dynamic parameters. However, recorded accelerograms include considerable distortions in the waveform during the wave propagation along the path to the receiver or due to the site conditions. Attenuation causes a frequency-dependent reduction in amplitude, scattering produces complicated wave superpositions, and reverberation in shallow sedimentary layers causes frequency-dependent resonance amplification. Therefore,
the recorded spectrum includes unwanted distortions (Scherbaum, 1994). However, all these factors can be isolated from the recorded ground-motion spectrum.

Beresnev (2001, 2002) and Beresnev and Atkinson (2002) argue theoretically that the maximum slip velocity on a rupturing fault is the source parameter that controls the fault’s high-frequency radiation, thus determining the level of seismic hazard. It has also been recently suggested that the peak slip velocity places an upper bound on the peak ground motion experienced during an earthquake (McGarr and Fletcher, 2007), highlighting the importance of this dynamic source parameter. In addition, Beresnev (2001, 2002) and Beresnev and Atkinson (2002) also provided the framework from which this parameter can be directly calculated from the corner frequencies of the shear-wave spectra. Their study is based on the fact that small-to-moderate size earthquakes radiate the particle-acceleration Fourier spectra in the far field that follow the “$\omega^2$-shape”,

\[
u_{ff}(\omega) = C M_0 \omega^2 [1 + (\omega / \omega_c)^2]^{-1}, \tag{1.1}
\]

where $C$ is a constant, $M_0$ is the seismic moment, and $\omega_c$ is the spectrum’s “corner frequency” at which the spectrum changes character (Reiter, 1990) from an increasing level at lower frequencies to a constant level at higher frequencies. By choosing small-to-moderate earthquakes observed in the far-field, we assume that the earthquake sources behave as point sources and follow the “$\omega^2$-shape”. Equation 1.1, representing the Fourier spectra in the far-field, is, strictly speaking, valid for point sources. This is
why we had to select small earthquakes observed far from hypocenters. As long as an
earthquake source can be considered a point source, the rupture-spreading effect and
the rupture surface are not variables in the analysis. Figure 1-1 shows an example of a
theoretical source spectrum and its corner frequency. Such a spectrum displays two
distinct slopes: one equal to 2 on the log-log scale (explaining the name of the “\(\omega^2\)-
spectrum”) and one equal to zero (the constant part) (as can be illustrated by Figure 1-
1). The cross-over frequency between these two slopes is the corner frequency. Thus
this character change occurs due to the source model we use, which has been validated
by years of seismological observations. Note that the third, high-frequency, slope seen
in Figure 1-1 is due to the near-surface attenuation (“kappa”) effect (see Equation 1.11)
and is not a source effect.

![Corner Frequency](image)

**Figure 1-1. Source Spectrum and Corner Frequency**
The functional form of the shear-dislocation displacement time history that radiates this spectrum is

\[ u(t) = U[1 - (1 + t/\tau) \exp(-t/\tau)], \quad (1.2) \]

where \( U \) is the dislocation’s final displacement and \( \tau = 1/\omega_c \). The maximum value \( v_{\text{max}} \) of the slip velocity \( u'(t) \) is

\[ v_{\text{max}} = \omega_c U/e, \quad (1.3) \]

where \( e \) is the base of the natural logarithm. With the definition of the seismic moment,

\[ M_0 = \mu UA, \quad (1.4) \]

where \( \mu \) is the shear modulus and \( A \) is the rupture area, and the relationship,

\[ V_S = \left( \frac{\mu}{\rho} \right)^{1/2}, \quad (1.5) \]

where \( V_S \) is the shear-wave velocity and \( \rho \) is the density, Equation 1.3 takes the form

\[ v_{\text{max}} = (2\pi/e)(M_0 / \rho V_S^2 A) f_c, \quad (1.6) \]
where \( f_c = \omega_c/2\pi \) (Beresnev, 2002, Equation 7). Equation 1.6 allows estimation of the maximum slip velocity from the corner frequency and other observable data.

A recorded spectrum’s common model is given by

\[
\text{Recorded Spectrum} (R, f) = \text{Source} (f) \times \text{Path} (R, f) \times \text{Site} (f),
\]

(1.7)

where \( R \) is the distance to the observation point. As seen from Equation 1.7, the recorded spectrum is formed from three multiplicative terms: the source spectrum itself, the path effect, and the site effect. The term Source \((f)\) is represented by Equation 1.1. The path effect is typically approximated as

\[
\text{Path} (R, f) = \exp\left(-\pi f R_0 / Q(f) V_s R / R_0 \right),
\]

(1.8)

where \( R_0 \) is a reference distance and \( Q(f) \) is the quality factor that characterizes anelastic attenuation. The \( 1/(R/R_0) \) factor in Equation 1.8 defines the geometric spreading of the waves, which is a decay in the wave amplitude with distance caused by the seismic energy spreading over greater and greater spherical surfaces (in the approximation of a spherical wavefront) as the wave propagates away (Lay and Wallace, 1995).
On the other hand, anelastic attenuation is the energy loss due to the processes of internal friction during the wave propagation, which also diminishes the amplitudes. During wave propagation, the successive conversion of potential energy as particle position to kinetic energy as particle velocity is not entirely reversible. For example, movements along mineral dislocations or shear heating at grain boundaries have effects on wave energy. These processes are collectively described as “internal friction” (Lay and Wallace, 1995), although it is generally agreed upon that the true physical reasons for the energy absorption are still not exactly understood (Sheriff and Geldart, 1995). The dimensionless quantity $Q$ describes the strength of the attenuation. $Q$ is defined in terms of the fractional energy loss per one cycle of deformation,

$$\frac{1}{Q(\omega)} = \frac{-\Delta E}{2\pi E},$$  \hspace{1cm} (1.9)

where $E$ is the peak strain energy available in the cycle and $-\Delta E$ is the energy lost during the cycle. Large values of $Q$ imply therefore small attenuation (Shearer, 1999).

The quality factor is usually expressed in a frequency-dependent relationship

$$Q = Q_0 f^n,$$  \hspace{1cm} (1.10)

where $n$ typically varies from 0 to 1 and $Q_0$ is a constant. $Q$, $Q_0$ and $n$ are functions of rock properties and may be different for different locations (Reiter, 1990).
The site effect on strong ground motion is a widespread phenomenon that often affects the degree of the damage caused by earthquakes. Understanding and removing the site effect is crucial for determining the source spectra. Assuming a hard-rock site condition (no significant local response), the site effect can be written as

\[ Site(f) = Crustal \ Amplification(f) \exp(-\pi \kappa f) , \]  

(1.11)

where the Anderson-Hough kappa \( \kappa \) describes the high-frequency spectral roll-off (Anderson and Hough, 1984).

The amplification is the function of shear-wave velocity and density of the medium. For a particular frequency, the amplification is given by the square root of the ratio between the seismic impedance (velocity times density) at the depth of the source and the seismic impedance averaged over a depth corresponding to a quarter wavelength (Boore and Joyner, 1997). This equation is given by

\[ A(f) = \sqrt{\left( \rho_s V_s \right)_s / \bar{\rho}(z) \bar{V}_s(z)} , \]  

(1.12)

where the small subscript “s” represents the values in the vicinity of the source, and \( \bar{\rho}(z) \) and \( \bar{V}_s(z) \) are the time-averaged values of the density and shear-wave velocity over the depth interval, respectively. Of course, because the averaging is performed over a quarter-wavelength, the amplification function \( A(f) \) is frequency-dependent.
As Equation 1.7 demonstrates, the recorded ground-motion spectra include the path and site effects; therefore, for studying earthquake-source parameters, isolating the source spectrum from these effects is the first step. This topic has been addressed by several investigators. Prejean and Ellsworth (2001), Chen and Atkinson (2002), and Ottemöller and Havskov (2003) analyzed the site and path effects separately to remove them from the recorded spectra within the model exemplified by Equation 1.7. To achieve the same goal, we will need to specify the regional parameters entering Equations 1.8 and 1.11 for the region of our interest.

Different earthquake databases from various regions may use different types of magnitude definitions. Although $M_s$, $M_b$, and $M_L$ (the surface-wave, the body-wave, and the local magnitudes, respectively) are some commonly used scales, Japan earthquakes that we will use in this study are quantified by yet another definition, the magnitude $M_{JMA}$ (JMA stands for the Japan Meteorological Agency). Therefore, the relationships between the different types of magnitudes, or at least relationships between the moment magnitude (which is the direct measure of the size of the rupture) and $M_{JMA}$ should be established to convert the latter to more accepted units. We will provide the necessary conversion equations in a later discussion related to our estimation of the moment $M_0$ and fault area $A$ entering Equation 1.6.

The main purposes of the following material in this thesis are removing the site and path effects from the recorded spectra to convert them to the source spectra and determining the maximum slip velocities on faults during earthquakes from the corner
frequencies of the source spectra, following the approach illustrated by Equations 1.6 to 1.8 and Equation 1.11.

Direct observation-based determinations of this type will provide valuable physical information about the in-situ faulting processes that can be used for constraining the dynamics theories of faulting or in ground-motion prediction.

1.2. Thesis Objectives

The main objective of this study is the direct observation of the maximum fault-slip velocities from the recorded strong-ground-motion data. The study is based on the analysis of the recorded earthquake spectra, determining and removing the unwanted site and path effects and obtaining the source spectra. Beresnev (2001, 2002) and Beresnev and Atkinson (2002) showed that the maximum slip velocity on a rupturing fault is the only source parameter that can be directly determined from the corner frequencies of the source spectra. By calculating the source spectra, corner frequencies can be identified and used for the maximum-slip-velocity calculations.

Because the maximum slip velocity is the parameter that controls the strength of earthquakes' high-frequency radiation, the results of this study can be used for the improvement in the quality of ground-motion prediction.
2. STRONG-MOTION DATA SETS AND ANALYSIS

2.1. Data Description

For the purposes of the study, we have opted to utilize modern high-quality ground-motion data from multiple small-to-moderate earthquakes in Japan recorded on rock sites, in the magnitude range from 4.0 to 6.0. The original records were obtained from the website of KiK-Net, the Japanese digital strong-motion accelerographic network (www.kik.bosai.go.jp). The website provides downhole and surface records for all three components of the accelerograms. It is well known that seismic noise and site effects are significantly reduced in a borehole compared to a surface recording (Abercrombie, 1998). Therefore, to minimize the effect of near-surface weathering and noise on the records, the data from the downhole accelerometers were used.

From all available events in the 4-6 magnitude range, we selected the earthquakes (1) that produced recordings at at least two different rock sites and (2) whose spectra followed the assumed “$\omega^2$-shape”. The former criterion is needed to estimate possible variability in the corner frequencies depending on the azimuth from the source to a recording site. The latter retains the validity of the underlying spectral model (Equation 1.1): although the $\omega^2$-model is commonly observed and used in seismology, there is no reason to believe that a variety of source processes would be exhausted by this only possibility. Rock stations were chosen for the analysis to minimize the site effects. The lithologies under the selected strong-motion stations are demonstrated in
alphabetical order in the cross-sections in Figure 2-1. Also, all sites with their corresponding geology and geographic coordinates are tabulated in Table 2-1. Detailed information about the earthquakes and sites that recorded them is listed in Table 2-2.

Figure 2-1. Lithologies of Stations
Figure 2-1 (continued). Lithologies of Stations

<table>
<thead>
<tr>
<th>Station</th>
<th>Lithology</th>
<th>Depth</th>
<th>Log</th>
</tr>
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<tr>
<td>NARH06</td>
<td>weathered granite</td>
<td>0-20</td>
<td></td>
</tr>
<tr>
<td>OKYH02</td>
<td>granite</td>
<td>20-30</td>
<td></td>
</tr>
<tr>
<td>OKYH04</td>
<td>granite</td>
<td>30-40</td>
<td></td>
</tr>
<tr>
<td>OKYH07</td>
<td>granite</td>
<td>40-50</td>
<td></td>
</tr>
<tr>
<td>OKYH09</td>
<td>granite</td>
<td>50-60</td>
<td></td>
</tr>
<tr>
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</tr>
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<td>granite</td>
<td>10-20</td>
<td></td>
</tr>
<tr>
<td>YMGH11</td>
<td>granite</td>
<td>20-30</td>
<td></td>
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Table 2-1. Rock Sites that Recorded the Events Selected

<table>
<thead>
<tr>
<th>Site Code</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>Geology</th>
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<tr>
<td>EHMH03</td>
<td>33.9121</td>
<td>133.6523</td>
<td>100</td>
<td>Black Schist</td>
</tr>
<tr>
<td>FKOH03</td>
<td>33.5575</td>
<td>130.5522</td>
<td>100</td>
<td>Granite</td>
</tr>
<tr>
<td>FKOH04</td>
<td>33.5479</td>
<td>130.7475</td>
<td>100</td>
<td>Granodiorite</td>
</tr>
<tr>
<td>GIFH07</td>
<td>35.4147</td>
<td>136.4376</td>
<td>100</td>
<td>Slate</td>
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<tr>
<td>HRSH01</td>
<td>34.3701</td>
<td>133.0259</td>
<td>205</td>
<td>Orthoclase Biotite Granite</td>
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<tr>
<td>HRSH07</td>
<td>34.2850</td>
<td>132.6436</td>
<td>102</td>
<td>Granite</td>
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<td>NARH06</td>
<td>34.6381</td>
<td>136.0540</td>
<td>101</td>
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<td>34.7468</td>
<td>134.0728</td>
<td>200</td>
<td>Granite</td>
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<td>OKYH04</td>
<td>34.6397</td>
<td>133.6888</td>
<td>100</td>
<td>Granite</td>
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<tr>
<td>OKYH07</td>
<td>35.0461</td>
<td>133.3196</td>
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<td>133.6792</td>
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<td>Granite</td>
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<tr>
<td>OKYH11</td>
<td>35.0700</td>
<td>134.1189</td>
<td>200</td>
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<tr>
<td>YMGH02</td>
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Table 2-2. Earthquake and Station Information from Kik-Net Database

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<th>Longitude</th>
<th>Magnitude*</th>
<th>Depth(km)</th>
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* Japan Meteorological Agency (JMA) magnitude.
2.2. Data Analysis

2.2.1. Distribution of Earthquakes and Recording Stations

The geographic distribution of both earthquakes (stars) and ground-motion stations (balloons) used in the analysis is shown in Figure 2-3. Both epicenters and recording sites are spread over a large territory with varying azimuths to the sources, where the ray-path coverage is extensive. This will necessitate our comparison of the path-effect corrections carried out with alternative regional Q-models, because no single determination of \( Q(f) \) is available for the entire region shown.

Figure 2-2. Earthquakes (stars) and stations (balloons) used in this study
2.2.2. **Magnitude Scales**

The intensity of an earthquake is a crude way to describe its size. Determination of the intensity is based on the damage of structures, the presence of fractures, cracks, and landslides of the ground, and the perceptions of people who experienced the event. Since the intensity is an indirect measure of the size of an earthquake, it is not an exact quantitative indicator. For example, despite a relative small size, a very shallow earthquake can produce high intensity. Therefore, a measurement of the size of a seismic event must include terms of energy releases at its focus. The first energy-release-based scale of magnitude was created by Charles Richter in 1935; it initiated a tradition of assigning magnitudes (such as $M_s$, $M_b$, and $M_L$) according to largest amplitude in a certain portion of a seismogram.

The “moment magnitude”, introduced by Hanks and Kanamori (1979), introduced a modern standard approach to quantifying the earthquake size. It is based on the concept of seismic moment, which reflects the amount of permanent slip on an earthquake fault,

$$M = \frac{2}{3} \log M_0 - 10.7,$$

where $M_0$ is measured in dyne-cm (Shearer, 1999; Udias, 1999).
Different magnitude definitions may be used for different regions as discussed earlier. We have dealt with Japanese earthquakes that are defined by the measured Japan-Meteorological-Agency Magnitude ($M_{JMA}$). Therefore, we needed conversion of $M_{JMA}$ to the more common moment magnitude. Equation 2.2 gives an empirical relationship between $M_{JMA}$ and $M_0$, which was determined by Moya et al. (2000).

\[
\log M_0 = 1.54M_{JMA} + 15.8.
\] (2.2)

Accordingly, we used Equation 2.2 to obtain the seismic moment, and then substituted this moment into Equation 2.1 to obtain the moment magnitude.
3. DATA PROCESSING

3.1. Calculation of Fourier Acceleration Spectra

Seismological data that are produced by earthquakes can be recorded at a point on Earth’s surface as ground displacement, velocity, or acceleration. These data represent time histories, and the Fourier transform is a useful tool that allows one to change the time series from time domain to the frequency domain or vice versa (Udias, 1999).

The representation of the recorded spectrum in form 1.7 is an example of a frequency-domain approach. The path and site effects can then be viewed as frequency responses of certain filters applied to the earthquake signal. If this signal is $S(t)$, then the effect of such a filter in the time domain is the mathematical “convolution” of $S(t)$ with the filter’s impulse response $I(t)$, where $I(t)$ may represent either the path or site distortion (Lay and Wallace, 1995). If the recorded time series is $g(t)$, the mathematical definition of the convolution operator (*) is

$$g(t) = S(t) * I(t) = \int_{-\infty}^{\infty} S(\tau)I(t-\tau)d\tau,$$                   (3.1)

which corresponds to the multiplication of the Fourier spectra in the frequency domain,

$$g(\omega) = S(\omega)I(\omega).$$                                                   (3.2)
Equation 3.2 corresponds to the multiplicative spectral model 1.7 that we use. As seen from Equation 3.2, filtering effects on a recorded spectrum can be removed from it by spectral division in the frequency domain (Chen, 2000). This is the approach we have taken for the corrections for the site and path effects.

Since the same displacement time history controls the far-field radiation of both S and P waves from a displacement-discontinuity source, both body waves will have the same spectral shape and the same corner frequencies (Beresnev, 2002). However, regional-propagation properties of shear waves are much better studied in strong-motion seismology, because, being more destructive during earthquakes, the shear waves represent more significant practical interest. Since the S waves are better studied, the properties needed for our corrections, such as the kappa-value or the quality factor, are sometimes simply unavailable for P waves. Therefore, we analyzed the spectra of the recorded shear-waves. The shear-wave window, containing the main arriving energy, had therefore to be identified first for all records. All three components (East-West, North-South, and Up-Down) of the records were used to determine the shear-wave window with its maximum length set to 10 s. Figure 3-1 shows an example of the window determination. The accelerograms belong to the 2007/04/26 earthquake recorded by station HRSH01.

After determining the shear wave window, its Fourier amplitude spectrum was computed by a FORTRAN-code. East-West and North-South components of the calculated spectrum were arithmetically averaged. Figure 3-2 shows the horizontal-
component-spectra of the shear-wave window shown in Figure 3-1, and their average spectrum is presented in Figure 3-3.

Figure 3-1. Three Components of Ground Acceleration Recorded by Station HRSH01 during the 2007/04/26 event. Peak Accelerations are Indicated.
Figure 3-2. Raw Fourier Spectra of the Horizontal Components of the Shear-Wave Window Shown in Figure 3-1.
3.2. Determining the Source Spectra

3.2.1. Recorded Spectra

The recorded spectra themselves (seen in Figure 3-3, for example) include site and path effects as seen in Equation 1.7, however these factors are needed to be removed from recorded spectra in order to obtain the source spectra, which are then used in the subsequent analysis. We therefore proceed to the description of such spectral corrections.

3.2.2. Path Effect Corrections

Equation 1.8 was applied to the recorded spectra for the path-effect corrections. As introduced previously, this equation includes the quality-factor term which is typically determined by geophysical inversions of the recorded data for all unknown regional
terms in this equation. A relatively new alternative to such inversions is the genetic (grid-search in the entire parameter space) algorithm. Since the earthquakes used in our study occurred at various regions (Figure 2-3), the quality factor $Q(f)$ needed to be selected for the corresponding regions.

Several studies determined $Q(f)$ for Japanese regions (Kinoshita, 1994; Moya and Irikura, 1998; Moya et al., 2000; Petukhin et al., 2003). We have chosen two definitions to compare their influence on the path-effect correction and ascertain whether a precise determination was needed. Figure 3-4 shows the geographic position of these regions relative to the earthquake epicenters: they all belong to the central and south-western Japan.

The first $Q$-model that we used is for the Kanto region, for the frequency range of $f = 0.5-16$ Hz (Kinoshita, 1994),

$$Q(f) = 130f^{0.7}. \quad (3.3)$$

(it was also used for the entire region of Japan by Chen and Atkinson, 2002, Table 3). The other is for the Kinki region, for the frequency range of $f = 1-35$ Hz (Petukhin et al. 2003),

$$Q(f) = 180f^{0.7}. \quad (3.4)$$
The results of the application of these two alternative models and the analysis of the ensuing uncertainty in the corner frequency/slip velocity determination will be addressed in a later section.

Figure 3-4. The Earthquakes and the $Q(f)$ Values
The crustal shear-wave velocity enters the path-effect calculation in Equation 1.8. We have assumed the value $V_S = 3.6$ km/s, following Chen and Atkinson (2002) who used this value for Japan. The reference distance $R_0$ was set to 1 km.

3.2.3. Site Effect Corrections

Since the amplitude of an earthquake wave can be increased or decreased by both the properties and configuration of the near-surface material through which the wave propagates (Reiter, 1990), the geological condition of the site is a crucial issue for the site-response computation. The importance of site effects on strong-motion records were emphasized in various studies (e.g., Chen and Atkinson, 2002; Boore and Joyner, 1997).

Equation 1.11, consisting of a crustal-amplification function and an exponential term, was applied to the recorded spectra for the site-effect corrections. In order to minimize this effect, only records on rock sites (Table 2-1) were used.

The amplification coefficients at particular frequencies for western North American generic rock sites were introduced by Boore and Joyner (1997) (see Table 3-1). Their coefficients were computed by using the quarter-wavelength approximation presented by Joyner et al. (1981). From the comparison of large quantities of recorded data, Chen and Atkinson (2002) reached a conclusion that the generic crustal amplifications for California and Japan rock and shallow-soil sites appeared to have the same shape;
this allowed us to use the generic rock-amplification model developed for western North America by Boore and Joyner (1997, Table 3). Since this model provides amplifications at tabulated specific frequencies, the values for other frequencies were computed by linear interpolation.

The kappa value is a parameter of the exponential term in Equation 1.11. In numerous previous studies, specific spectral-decay-parameter values, kappa, have been determined and used for specific regions (e.g., Boore and Joyner, 1997; Chen and Atkinson, 2002). Chen and Atkinson (2002) obtained six specific κ results for six different world regions, and Japan is one of them. Their result for Japan is \( \kappa = 0.035 \text{s} \). Since there is not any kappa range in the literature for Japan, this specific value from Chen and Atkinson’s (2002) was used in our study as well.

Table 3-1. Crustal Amplifications

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<tr>
<td>0.09</td>
<td>1.10</td>
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<td>6.05</td>
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<td>3.13</td>
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</table>
3.2.4.  Source Spectrum

By removing the path and site effects, according to the seismogram model of Equation 1.7, the source spectrum was obtained. The source acceleration spectra were assumed to have the classic $\omega^2$-shape (Aki, 1967; Brune, 1970, 1971) with a single corner frequency. In order to keep the data within the limits of the general validity of the $\omega^2$-model, this study focused on small-to-moderate events only.

3.3.  Calculating the maximum slip velocity

3.3.1.  Corner Frequency

Figure 3-5 shows the isolated source spectrum after the site- and path-effect corrections have been applied to the raw spectrum in Figure 3-3. After all corrections have been applied to the recorded spectrum, two different slopes can usually be clearly identified. In our work, first the approximate intersection point of these two slopes was determined by eye, and, second, the linear-regression fitted lines were plotted to identify the intersection of the slopes defining the corner frequency $f_c$ (Savage, 1972). The more accurate corner-frequency values were determined by equaling the equations of the slopes to each other, but of course such a determination still carries an inherent degree of ambiguity. This method is nevertheless considered standard (e.g., Prejean and Ellsworth, 2001; Chen and Atkinson, 2002; Ottemöller and Havskov, 2003). To estimate an uncertainty in such determinations, we used at least two stations for each
earthquake to compare the inferred corner frequencies. We thus had a chance to evaluate the reliability of the selected corner frequencies. All corrected source spectra for all records listed in Table 2-2 are shown in Figures 3-6 to 3-41. The earthquake and station names are indicated on the Figures. The individual spectra were respectively calculated for two different regional $Q$-models (Equations 3.3 and 3.4).

![Corrected Average Spectrum](image1)

Figure 3-5. The spectrum from Figure 3-3 corrected for the path and site effects. The corner frequency and the fitted lines are shown

![Corrected Spectrum](image2)

Figure 3-6. Corrected Spectrum of EQ:2006/05/28, Sta: YM GH02 ($Q(f) = 130f^{0.7}$)
Figure 3-7. Corrected Spectrum of EQ 2006/05/28, Sta: YMGH04 ($Q(f) = 130f^{0.7}$)

Figure 3-8. Corrected Spectrum of EQ: 2006/05/28, Sta: YMGH11 ($Q(f) = 130f^{0.7}$)

Figure 3-9. Corrected Spectrum of EQ 2006/07/11, Sta: FKOH03 ($Q(f) = 130f^{0.7}$)
Figure 3-10. Corrected Spectrum of EQ 2006/07/11, Sta: FKOH04 \((Q(f) =130f^{0.7})\)

Figure 3-11. Corrected Spectrum of EQ 2006/07/11, Sta: YMGH08 \((Q(f) =130f^{0.7})\)

Figure 3-12. Corrected Spectrum of EQ: 2007/01/22, Sta: GIFH07 \((Q(f) =130f^{0.7})\)
Figure 3-13. Corrected Spectrum of EQ: 2007/01/22, Sta: NARH06 \((Q(f) = 130f^{-0.7})\)

Figure 3-14. Corrected Spectrum of EQ: 2007/04/26, Sta: EHMH03 \((Q(f) = 130f^{-0.7})\)

Figure 3-15. Corrected Spectrum of EQ: 2007/04/26, Sta: HRSH01 \((Q(f) = 130f^{-0.7})\)
Figure 3-16. Corrected Spectrum of EQ: 2007/04/26, Sta: OKYH02 \( (Q(f) = 130f^{0.7}) \)

Figure 3-17. Corrected Spectrum of EQ: 2007/04/26, Sta: OKHY07 \( (Q(f) = 130f^{0.7}) \)

Figure 3-18. Corrected Spectrum of EQ: 2007/04/26, Sta: OKYH09 \( (Q(f) = 130f^{0.7}) \)
Figure 3-19. Corrected Spectrum of EQ: 2007/04/26 Sta: OKYH11 ($Q(f) = 130f^{0.7}$)

Figure 3-20. Corrected Spectrum of EQ: 2007/05/13, Sta: HRSH07 ($Q(f) = 130f^{0.7}$)

Figure 3-21. Corrected Spectrum of EQ: 2007/05/13, Sta: OKYH04 ($Q(f) = 130f^{0.7}$)
Figure 3-22. Corrected Spectrum of EQ: 2007/05/13, Sta: YMGH02 \((Q(f) = 130f^{0.7})\)

Figure 3-23. Corrected Spectrum of EQ: 2007/05/13, Sta: YMGH04 \((Q(f) = 130f^{0.7})\)

Figure 3-24. Corrected Spectrum of EQ: 2006/05/28, Sta: YMGH02 \((Q(f) = 180f^{0.7})\)
Figure 3-25. Corrected Spectrum of EQ: 2006/05/28, Sta: YMGH04 \(Q(f) = 180f^{0.7}\)

Figure 3-26. Corrected Spectrum of EQ: 2006/05/28, Sta: YMGH11 \(Q(f) = 180f^{0.7}\)

Figure 3-27. Corrected Spectrum of EQ: 2006/07/11, Sta: FKOH03 \(Q(f) = 180f^{0.7}\)
Figure 3-28. Corrected Spectrum of EQ 2006/07/11, Sta: FKOH04 \((Q(f) = 180f^{0.7})\)

Figure 3-29. Corrected Spectrum of EQ: 2006/07/11, Sta: YMGH08 \((Q(f) = 180f^{0.7})\)

Figure 3-30. Corrected Spectrum of EQ: 2007/01/22, Sta: GIFH07 \((Q(f) = 180f^{0.7})\)
Figure 3-31. Corrected Spectrum of EQ: 2007/01/22, Sta: NARH06 ($Q(f) = 180f^{-0.7}$)

Figure 3-32. Corrected Spectrum of EQ: 2007/04/26, Sta: EHMH03 ($Q(f) = 180f^{-0.7}$)

Figure 3-33. Corrected Spectrum of EQ: 2007/04/26, Sta: HRSH01 ($Q(f) = 180f^{-0.7}$)
Figure 3-34. Corrected Spectrum of EQ: 2007/04/26, Sta:OKYH02 \((Q(f) = 180f^{0.7})\)

Figure 3-35. Corrected Spectrum of EQ: 2007/04/26, Sta: OKYH07 \((Q(f) = 180f^{0.7})\)

Figure 3-36. Corrected Spectrum of EQ: 2007/04/26, Sta: OKYH09 \((Q(f) = 180f^{0.7})\)
Figure 3-37. Corrected Spectrum of EQ: 2007/04/26, Sta: OKYH11 ($Q(f) = 180f^{0.7}$)

Figure 3-38. Corrected Spectrum of EQ: 2007/05/13, Sta: HRSH07 ($Q(f) = 180f^{0.7}$)

Figure 3-39. Corrected Spectrum of EQ: 2007/05/13, Sta: OKYH04 ($Q(f) = 180f^{0.7}$)
3.3.2. Maximum Slip Velocity

Finally, Equation 1.6 allows estimating the maximum slip velocities from the corner frequencies inferred and other parameters. The crustal density was taken as $\rho = 2.8$ g/cm$^3$ (Chen and Atkinson, 2002).
For the calculation using Equation 1.6, we also need to know the rupture area of the earthquake. Wells and Coppersmith (1994) developed an empirical approach to the rupture-area determination for different types of earthquake slip. They indicated that the spatial pattern of aftershocks, occurring within a few hours to a few days of the mainshock, generally define the maximum extent of the co-seismic rupture, and this method was used to estimate the rupture areas. They related the rupture area to the moment magnitude of an earthquake. Following Beresnev and Atkinson (2002), we therefore determined $A$ through Wells and Coppersmith’s (1994, Table 2A and Figure 16A) empirical formula as $\text{km}^2$,

$$\log A = -3.49 + 0.91M,$$  \hspace{1cm} (3.5)

where $M$ is the moment magnitude.
4. RESULTS and DISCUSSION

Tables 4-1 and 4-2 summarize two different inferred corner frequencies and maximum slip velocities corresponding to the two different $Q(f)$ models (Equation 3.3 and 3.4) for all five earthquakes and their stations. As seen in the Tables the obtained results for each quality factor are very close to each other. The standard deviations in the corner frequency and calculated maximum slip velocity are also listed for each quality factor. These standard deviations have been obtained from the recordings of the same earthquake at several rock stations as shown in the Tables.

The maximum slip velocities calculated for these small and moderate earthquakes range between approximately 0.2 and 0.6 m/s. Rice (2007) and Brown et al. (2007) reported (oral communications) the typical seismic slip rates in the range of 0.1-0.8 and 0.5-2 m/s, respectively. These values are entirely compatible with our direct measurements. On the other hand, McGarr and Fletcher (2007) summarized peak slip velocities for eight large earthquakes from different world regions ranging from 2.3 to 12 m/s. The discrepancy with our results is obvious, in that their values are systematically larger. McGarr and Fletcher’s results have been obtained from published finite-fault slip inversions and not directly from ground-motion records. Such inversions are not necessarily reliable and should be interpreted with caution; there is presently no established way of assessing their quality (see Beresnev, 2003, for a review). More direct determinations should be preferred.
Table 4-1. Corner Frequencies and Maximum Slip Velocities for $Q(f) = 130f^{0.7}$

<table>
<thead>
<tr>
<th>Earthquake &amp; Stations</th>
<th>Corner Frequency (Hz)</th>
<th>Mean Corner Frequency (Hz)</th>
<th>Standard Deviation of Corner Frequency (Hz)</th>
<th>Mean Maximum Slip Velocity (cm/sec)</th>
<th>Standard Deviation of Maximum Slip Velocity (cm/sec)</th>
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Table 4-2. Corner Frequencies and Maximum Slip Velocities for $Q(f) = 180f^{0.7}$

<table>
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<th>Earthquake &amp; Stations</th>
<th>Corner Frequency (Hz)</th>
<th>Mean Corner Frequency (Hz)</th>
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The direct method for peak-slip-velocity calculation that we have tested provides valuable information for the studies of in-situ dynamic fault properties and supplies observational constraints needed for the development of the theories of dynamic faulting.
5. CONCLUSIONS

Seismic ground-motion records supply hidden but the only available information about earthquake-source properties. Understanding the source mechanism provides important improvements in our ability to predict seismic motions from future events. Inferred source properties also provide fundamental knowledge about physical processes in Earth’s interior. Peak slip velocity on rupturing faults is one of the important earthquake properties that can be determined from the observations.

During wave propagation along the path and in the shallow subsurface, the seismic waveform is distorted. The recorded spectrum, therefore, includes this distortion as well as the desired source properties. However, these unwanted effects can be separated from the recorded spectra to isolate the source properties.

In this study, events with spectra that followed the assumed $\omega^2$-shape were selected, and recordings of these events obtained from rock sites to minimize site effects. After removing all distortions, the refined spectra approximately follow the $\omega^2$-shape with a single corner frequency. In order to determine the uncertainties in the corner-frequency determination, we used multiple recordings of the same earthquake and also compared effects of alternative equations of the anelastic-attenuation operator $Q(f)$. Specific input parameters such as the kappa value, the shear-wave velocity, and the crustal density were taken from previous analogous studies. There is no information in the literature
about the uncertainties in these values, and we followed the approach of the previous investigators (e.g., Chen and Atkinson, 2002). Peak slip velocities and their uncertainties were calculated by using the corner frequencies of the source spectra. The typical peak velocities of slip for moderate-size Japanese earthquakes turned out to be of the order of 0.2-0.6 m/s.

This study has tested the method for a direct determination of the fault-slip velocities from the recorded data. This information provides valuable insight into understanding the in-situ dynamic fault behavior and will be useful to constrain models of ground-motion prediction.
REFERENCES


