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Mass Estimates of the Scalar Seismic Moments of Small Earthquake Foci on Southern Sakhalin

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Abstract—The frequency dependence of the function of the seismic wave attenuation was determined for the first time for southern Sakhalin on the basis of seismic coda of local earthquakes using the model of single scattering. The algorithm of the automated definition of the scalar seismic moments was realized for small earthquake foci. Mass estimates of the scalar seismic moments were obtained as exemplified by the aftershocks of the August 17, 2006, Gornozavodsk earthquake (M_W 5.6) and the May–June 2004 Kostromskoe earthquake swarm events, which occurred in South Sakhalin. The dynamic parameters of the earthquake foci were determined from the SH-wave spectra adjusted for absorption and geometrical spreading. The loglinear relationship determined between the seismic moment and the local magnitude is in good agreement with the estimates obtained for other regions and, in a certain sense, does not contradict the average world dependence.

Keywords: *seismic waves, seismic moment, small earthquakes, Sakhalin Island.*

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INTRODUCTION

In recent decades, the southern part of Sakhalin Island experienced strong earthquakes with high aftershock activity, which confirms the link of the modern seismicity with zones of active tectonic dislocations. In particular, the 2001 Takoe earthquake (M_W 5.2) was confined to the active Aprelovskii Fault Zone [6], while the 2004 Kostromskoe Earthquake (M_L 4.8), the 2006 Gornozavodsk earthquake (M_W 5.6) [9], and the destructive 2007 Nevel'sk earthquake (M_W 6.2) [2] were localized in the Western Sakhalin fault zone. The two latter seismic events were accompanied by strong macroseismic effects corresponding to intensities of 6–7 and 8 points on the MSK-64 scale in the epicenter and coastal zone of South Sakhalin, respectively.

Since 2001, the intensification of the seismic swarm activity has successfully coincided with the mounting of a new local network of modern digital seismic stations and the launching of uninterrupted seismic observations in South Sakhalin. At present, the local network (Fig. 1) is sufficiently dense to identify and allocate with confidence earthquakes with $M_L \geq 2.0$ over the entire South Sakhalin territory. However, the calculation of the dynamic parameters of the earthquake sources for the coastal regions, including the focal mechanisms determined from the traditional approaches [3] based on data on the polarities of the first displacements in body waves, provides unstable estimates and invalid solutions. This is related both to the deficit of starting data and to the large "gaps" in the azimuthal coverage of the recording network, as well as to the low signal–noise ratios in the oscillation records.

A number of foreign works published in recent years are dedicated to estimating the dynamic parameters of the sources and focal mechanisms of earthquakes in nonoptimal observation systems [for instance, 25]. Note that the calculation techniques based on the quantitative parameters of the seismic wave form such as the magnitude of the amplitude, the amplitude ratios, and the body wave spectra are steadily being improved. A combination of different methods is often used to increase the reliability of the determinations. For instance, the signs of the first arrival times and the P- to S-wave amplitude ratios provide more reliable estimates of the focal mechanism of earthquakes, including weak earthquakes [22]. These methods are less sensitive to the quality of the records, but they require comprehensive data on the regional and local attenuation model, the Earth's crust structure, and others.

In many seismostatistical tasks, the magnitude of the earthquake energy is estimated in units of the seismic moment, which is calculated either directly from records or via the "magnitude–moment" correlation. In order to obtain reliable estimates of dynamic earthquake parameters (in particular, scalar seismic moments) from digital records, it is first necessary to

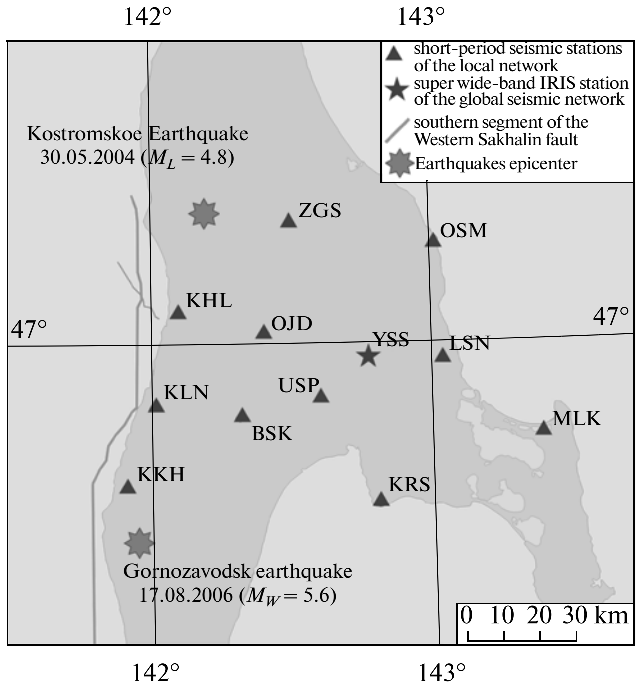

Fig. 1. Location of the seismic stations.

The stations names on the map are shown by abbreviations: (YSS) "Yuzhno-Sakhalinsk," (OJD) "Ozhidaevo," (ZGS) "Zagorskoe," (OSM) "Cape Ostryi," (LSN) "Lesnoe," (MLK) "Mal'kovo," (KRS) "Korsakov," (USP) "Uspenskoe," (BSK) "Belye Skaly," (KHL) "Kholmsk," (KLN) "Kalinino," (KKH) "Kolkhoznoe.

determine the quality factor of the medium characterizing the seismic wave attenuation.

In this work, we report the results of the determination of a local model of the seismic wave attenuation for South Sakhalin. The obtained frequency-dependent absorption function of the medium was used to accomplish an automated algorithm of the mass estimates of the scalar seismic moment from the records of several stations by the example of small earthquakes that occurred in South Sakhalin. The application of the loglinear dependence linking the seismic moment and the local magnitude was justified for small and moderate earthquakes of South Sakhalin.

THE ANALYZED DATA

The work was based on instrumental data obtained during short-period observations in the aftershock zones of the 2004 Kostromskoe (M_L 4.8) and 2006 Gornozavodsk (M_W 5.6) earthquakes (Fig. 1).

Both events occurred in the territory adjoining the southern segment of the West Sakhalin fault. The dynamic seismic regime of the latter is presumably explained by the intense sublatitudinal compression of the Earth's crust, which caused the formation of a dextral strike-slip fault. The style of the deformation is mainly confirmed by the structural geology of South Sakhalin [12] and the satellite monitoring data on the Earth's crust motion [13].

The Kostromskoe earthquake occurred on May 30, 2004, at 2:52 UT in the vicinity of Kostromskoe Settlement in the Kholmsk district near the western coast. The magnitude of the main shock accounted for M_L 4.8. The event was accompanied by earthquake swarms with the magnitude of the strongest aftershock being M_L 3.9. The total amount of recorded earthquakes accounted for more than 100 events.

The Gornozavodsk earthquake took place on August 18, 2006, at 2:20 UT. The epicenter of the main event was located inland near the western coast of the island in the vicinity of the Gornozavodsk Settlement of the Nevel'sk district at a depth of about h = 13 km. The strongest aftershock had a magnitude of M_L 4.5. The aftershock activity was in progress as weak events with amplitudes of M = 2.0–3.0 up to the end of August 2006 and then rapidly stopped, which is not typical of such strong earthquakes. In the course of this activity, the earthquake epicenters were shifted to the northwest toward the Tatar Strait, while the depth of the aftershock epicenters reached 20 km. The aftershock sequence included about 500 earthquakes.

In this work, we analyzed the record fragments of small earthquakes provided by the local network of "Dat" and "Datamark" digital seismic stations (Hakusan Corp., Japan) established in South Sakhalin (Fig. 1). Each station was equipped with a three-component short-period seismic LE-3DLite seismometer (Lennartz Electronics, Germany) with a natural frequency of about 1 Hz and a frequency band up to 80 Hz. The stations operated in a continuous recording regime with sampling at 100 sps.

The initial preparation of the instrumental data (in particular, the selection of the earthquake records with satisfactory signal–noise ratios and epicenter calculations) was conducted by Ch.U. Kim, the leading researcher of the Institute of Marine Geology and Geophysics of the Far East Branch of the Russian Academy of Sciences, who kindly gave us the initial data and the results of their treatment. The additional treatment of the data involved the selection of earthquake records obtained by several stations, the splitting and saving of the data files with fragments of wave forms, and their exporting into the ASCII format suitable for further processing. The main criteria for the selection of the seismic events were a reliable record of the earthquakes by several stations and a good signal/noise ratio (more than 3) in the records of the seismic oscillations. The data of the initial treatment and files prepared from the seismic record fragments were used as the input data for the development of a computer algorithm.

The input data were processed using algorithms coded in FORTRAN, as well as the XDatSegy and Datamark Assist packages applied for the seismological instruments.

The seismic wave attenuation was estimated using selected records of the aftershocks of the 2006 Gornozavodsk earthquake (M_W 5.6).

The scalar seismic moments of the epicenters of the small earthquakes were determined on the basis of the digital records of the aftershocks of the Gornozavodsk earthquake (August 17, 2006, M_W 5.6) and the Kostromskoe earthquake swarm (May 30, 2004, M_L 4.8), which occurred in South Sakhalin and were confidently recorded by no less than three stations. In total, we selected 87 events with magnitudes M_L from 1.4 to 4.0.

THE CALCULATION TECHNIQUE OF THE LOCAL MODEL OF THE ATTENTUATION

The estimation of the seismic wave attenuation is usually applied if the structure and composition of the Earth's crust in the studied region are poorly known. In the general case, the attenuation is characterized by two components: the scattering of the oscillations from the geological inhomogeneties and their absorption in the medium, which leads to the transformation from wave to thermal energies:

$$Q^{-1} = Q_s^{-1} + Q_i^{-1}, \tag{1}$$

where Q^{-1} is the integral quality factor of the medium, Q_s^{-1} are the scattering characteristics, and Q_i^{-1} are the absorption characteristics of the seismic waves.

There are methods to estimate the quality factor of the medium on the basis of the coda wave, the tail portion of the wave form of seismic event. The coda wave is formed mainly from single backscattering S-waves with the significant input of scattered P-waves. The coda wave is also formed in the course of the wave transformations at the lithospheric discontinuities. The coda shape is determined by the wave reverberation in the horizontal layers on the seismic trace, as well as by the lateral inhomogeneities.

According to the simplified model of single scattering proposed in fundamental work [17], the effect of the source process can be neglected for large times (approximately 1.6–2 times more than the travel time of the S-wave), and the coda wave is formed due to single backscattering (refracted) S-waves and its level is independent of the distance to the station [28]. More perfect approaches were developed later on the basis of the numerical modeling of the scattering in a two- and three-dimensional isotropic and anisotropic medium [7, 10, etc.], as well as models with the use of the numerical solution of the nonstationary equation of the spatial transfer of the wave energy [21, 24]. These methods take into account the more realistic model of multiple scattering of seismic waves.

We considered the available approaches and selected a model of single scattering as being comparatively simply implemented and allowing the estimation of the quality factor with sufficiently high accuracy. The model of single scattering was previously successfully applied for estimating the medium absorption in different regions of the earth [26, 29, 33]. Therefore, the obtained results could be compared with similar estimates for the Earth's regions with similar geological settings.

According to the model of single scattering [17], the displacement in the frequency range with the central transmission frequency f is given by the relation

$$A(f, t) = \Omega(f)t^{-\alpha}\exp(-\pi ft/Q_C(f)), \tag{2}$$

where α is a constant depending on the geometrical spreading (for the body waves $\alpha = 1$), t is the time counted from the time in focus, $\Omega(f)$ is the temporal function in the source, and $Q_C(f)$ is the quality factor (from the coda wave).

Taking logarithm (2) yields the required expression:

$$\ln[A(f, t)t] = \ln\Omega(f) - \pi ft/Q_C(f). \tag{3}$$

The slope of equation (3) plotted on the time scale defines the quality factor $Q_C(f)$ for the considered frequency f.

The values Q are sensitive to the choice of such parameters as the time window of the treatment, the initial time of the treatment, the minimal correlation coefficient, and the band width of the filter. The control parameters of the algorithm are required to be known or be similar to compare the results of the determination of the quality factor for different observations. With allowance for works [18, 20, 21, 28, etc.], the data were treated using parameters similar to those in the indicated works.

METHOD OF THE DETERMIANTION OF THE SCALAR SEISMIC MOMENT UNDER CONDITIONS OF LOCAL SHORT-PERIOD OBSERVATIONS

As a rule, the dynamic parameters of the source during teleseismic or regional observations are estimated for strong earthquakes with long wavelengths, whereas records of surface waves, for which the medium in a certain sense is "transparent," serve as the initial data The effect of the medium's quality factor on the recorded spectrum is negligibly low, and the function of the medium's absorption $Q(f)$ is traditionally regarded as frequency-independent. The signal/noise ratio in the surface wave spectrum for small earthquakes often seems to be inadequate, while the accuracy of the estimation of the source dynamic parameters mainly depends on the position and type of the recording apparatus. Therefore, under conditions of local short-period observations, the determination of the scalar seismic moment M_0 [8, 14, 16] from the record of the surface waves is inefficient. The seismic moment of the local earthquakes is determined using the record of the body waves on the basis of the short-period observations: the seismic moment M_0 is determined from the displacement spectra of the P, SV, and SH waves.

At the same time, as was noted in numerous recently published works [10, 33, etc.], $Q(f)$ is proportional to the frequency f (or the degree of frequency) in the frequency range of more than 1 Hz. This must be taken into account for weak earthquakes, when the working frequencies of the seismic signals fall in the range from a few to tens of Hz and the medium quality factor significantly affects the spectral composition of the seismic signals. The spectral model of Brune [19] used in this case represents the source of the tectonic earthquake as the simplified case of a infinite vertical fault with a horizontal trend of the displacement, a constant value of the released shear stress, and without allowance for the finite velocity of the rupture propagation. In addition, this model does not take into account the complex shear geometry, which leads to the enrichment of the spectrum of the transmitted waves in the high-frequency range. The Brune model is schematic, but it provides the fairly accurate estimation of such source parameters as the scalar seismic moment, the released stress, the rupture area, etc., with allowance for the medium absorption of seismic waves.

Let us consider the general case of the amplitude spectrum of the displacements of the seismic body wave $U^{(i)}(f)$ at the ith station:

$$U^{(i)}(f) = G(f)H^{(i)}(f)R^{(i)}(f)S(f), \qquad (4)$$

where $G(f)$ is the amplitude–frequency response (AFR) of the seismograph, $H^{(i)}(f)$ is the response of the medium beneath the station, $R^{(i)}(f)$ is the response of the seismic trace propagation, $S(f)$ is the spectral equivalent of the time function in the source, and f is the frequency.

Taking into account the local ground conditions beneath the observation sites, the function can be written as follows:

$$H^{(i)}(f) - \exp(-\pi\kappa^{(i)}f), \qquad (5)$$

where $\kappa^{(i)}$ is the absorption coefficient beneath the i-th station.

The seismic trace response can be expressed as

$$R^{(i)}(f) = \frac{1}{r_i^{\alpha}}\exp\left\{-\frac{\pi f t_i^*}{Q(f)}\right\}, \qquad (6)$$

where t_i^* is the travel time of the seismic wave from the source to the i-th seismic station, $Q(f)$ is the quality factor, r_i is the hypocentral distance between the source and the observation point, and α is the coefficient that determines the geometrical spreading (for local observations, $\alpha = 1$ at $r \le 100$–150 km).

According to the Brune model [19], the spectral equivalent of the source time function has the following form:

$$S(f) = \frac{M_0 R_{\vartheta\varphi}}{4\pi\rho v^3}\frac{1}{1+(f/f_0)^2}, \qquad (7)$$

where M_0 is the seismic moment, $R_{\vartheta\varphi}$ is the source radiation pattern, ρ is the medium density, v is the shear wave propagation velocity in the source, and f_0 is the corner frequency of the spectral curve cutoff.

Combining all the effects considered above, we have the following formula for the spectral function of the displacement:

$$U^{(i)}(f) = \frac{M_0 R_{\vartheta\varphi}}{4\pi\rho v^3 r_i}\exp\left\{-\frac{\pi f t_i^*}{Q(f)}\right\}\frac{\exp(-\pi\kappa^{(i)}f)}{1+(f/f_0)^2}. \qquad (8)$$

At $f \to 0$ ($f \ll f_0$), expression (8) can be transformed into the following equation:

$$M_0 = \frac{4\pi\rho v^3 r_i}{R_{\vartheta\varphi}}U_0^{(i)}\left(\exp\left\{\frac{\pi f t_i^*}{Q(f)}\right\}\right)\Bigg|_{f\to 0} \times \exp\{-\pi\kappa^{(i)}f\}\big|_{f\to 0}\left(1+\left(\frac{f}{f_0}\right)^2\right)\Bigg|_{f\to 0}, \qquad (9)$$

where $U_0^{(i)} = U^{(i)}(f)|_{f\to 0}$ is the low-frequency region of the displacement spectrum.

If the quality factor of the medium can be ignored, the seismic moment M_0 can be determined from the following relation:

$$M_0 = \frac{4\pi\rho v^3 r}{2R_{\vartheta\varphi}}U_0, \qquad (10)$$

where the coefficient S is caused by the reflection effects from the free surface at the receiver point.

Taking the frequency dependence of the quality factor $Q(f) = Q_0(f)$, we have

$$\exp\left\{\frac{\pi f t^*}{Q(f)}\right\}\Bigg|_{f\to o} = \exp\left\{\frac{\pi t^*}{Q_0}\right\}, \qquad (11)$$

which indicates that the formulas used for M_0's determination omitted the dependence on t^* and Q_0, which, in some cases, may lead to the systematic underestimation of the seismic moment by several times.

Assuming the travel time of the S-wave from the source to the station $t^* = 20$ s and the quality factor $Q_0 = 60$, the correction factor t for the medium absorption (11) is approximately 3. For the parameter $\kappa = 0.05$, the correction for the attenuation beneath station (5) at a frequency of $f \sim 1$ Hz is 1.16. Thus, within the range of the estimated parameters, the attenuation is mainly determined by $Q(f)$. Therefore, the spectra used in this work were corrected only for $Q(f)$, whereas the correction for the local ground conditions was ignored.

M_0 was estimated from the average low-frequency component of the displacement spectrum $U^{(i)}$ corrected for absorption within the frequency range of 0.5–2.0 Hz

$$M_0^{(i)} = \frac{2\pi\rho v^3 r_i}{R_{\vartheta\varphi}} \langle U^{(i)}(f)\exp\left\{\frac{\pi f t_i^*}{Q(f)}\right\}\rangle\Bigg|_{f\in[0.5,\,2.0]}, \quad (12)$$

where $\langle\ \rangle$ is the mean operator, and $M_0^{(i)}$ is the seismic moment estimated for the ith station (with allowance for the coefficient of 1/2 in this formula).

Since the frequency of the LE-3DLite seismic receiver is 1, the AFR of the apparatus was taken into account in the calculations over the entire working interval of the frequencies from 0.5 Hz, including the AFR drop at frequencies below 1 Hz.

THE DETERMINATION RESULTS AND THEIR DISCUSSION

1. The attenuation was calculated using the vertical components of the seismic record. For the analysis of the coda wave and the other calculations, we applied a time window $\Delta T_C = 12$–15 with its origin corresponding to the doubled travel time of the S-wave. The selected value of the time range of the coda treatment corresponds to the scattering area (or more exactly, the scattering ellipsoid) of the seismic waves with a radius of about 50 km at the average distance from the source to the station of ≈50 km. Within the intervals of ΔT_C significantly exceeding 15s, the signal/noise ratio is less than 2, which decreases the significance of the results obtained by the coda processing under large times. The records were filtered by a four-pole two-sided Butterworth filter with central transmission frequencies of 1.5, 3.0, 6.0, and 9.0 Hz and widths of 1.0, 2.0, 4.0, and 6.0 Hz, respectively. Examples of the starting record of the earthquake with the magnitude M_L 3.1, the filtered seismograms with central transmission frequencies of 1.5 and 6.0 Hz, and the fragments of the record and the corresponding codas corrected for the geometrical spreading are shown in Fig. 2.

Then, the quality factor was determined for each of the central frequencies f (Fig. 3) and the form of the frequency-dependent function $Q_C(f)$ was evaluated according to (3).

Due to the low signal/noise ratio in the records of weak earthquakes, the obtained results have low significance and do not allow us to estimate objectively the attenuation function for frequencies of more than 10 Hz. The fragments of the records of strong events with a sufficiently high signal/noise ratio were cut by the amplitude owing to the proximity of the recording stations to the event sources and the low dynamic range of the apparatus, and were not used in the computations.

The quality factor $Q_C(f) \approx 60f$ was found by averaging over all the events. The obtained estimate $Q_C(f)$ in the source zone of the aftershocks of the 2006 Gornozavodsk earthquakes is close to the estimates obtained for the absorption function in Central Japan [33]. Further, this functional dependence was used to estimate the scalar seismic moments of weak earthquakes according to expression (12).

Since the lower boundary of the frequency range used to estimate $Q_C(f)$, is 1.5 Hz, the function of the quality factor was extrapolated to a frequency of 0.5 Hz in order to estimate the average level of the low-frequency part of the displacement spectrums. Such an extrapolation can be substantiated by the fact that the modern studies of the seismic wave attenuation reveal the existence of three frequency ranges: $f < 0.01$, 0.01–10.0, and $f > 10.0$, within which the parameter b for the frequency dependence of the quality factor determined for definite regions of the earth preserves its unchanged value. Since the working frequency range of 0.5–10 Hz is remote from the transition boundaries accounting for, respectively, 0.01 and 10 Hz, the used function of the quality control in the indicated interval quantitatively characterizes the same seismic wave attenuation mechanism. Note also that $Q_C(f)$ was estimated for the frequency range with the lower boundary at 1.5 ± 0.5 Hz. Thus, we may confidently suggest that the view of the frequency dependence of the quality factor remains practically unchanged within the range of 0.5–1.0 Hz.

The applicability of the $Q_C(f)$ estimates is under discussion, because the difference between the quality factors determined from the coda wave and from the spectrum of the body waves remains unclear [21]. The comparison of the quality control values obtained by different methods and for different regions of the earth [23, 36, 27, etc.] shows that, in most cases, these values are of the same order. However, for each certain case with different rheology, source depth, etc., the traditional choice of the time window for the coda analysis with the origin corresponding approximately to the doubled travel time of the S-wave may lead to a systematic error in the Q_0 estimates.

2. Before the quantitative estimation of the spectral parameters, each of the three-component records was adjusted to the senses of the body waves (P, SV, and SH), which approach the receiver point. The calculations were based on the analysis of fragments of the SH-wave record. A time window for each wave form was selected to extract several wave trains, including the "parasitic" effects of converted reflections through the medium interfaces beneath the station. The time window interval was controlled by a parameter of more than 3.0 determinable by the signal/noise ratio. On

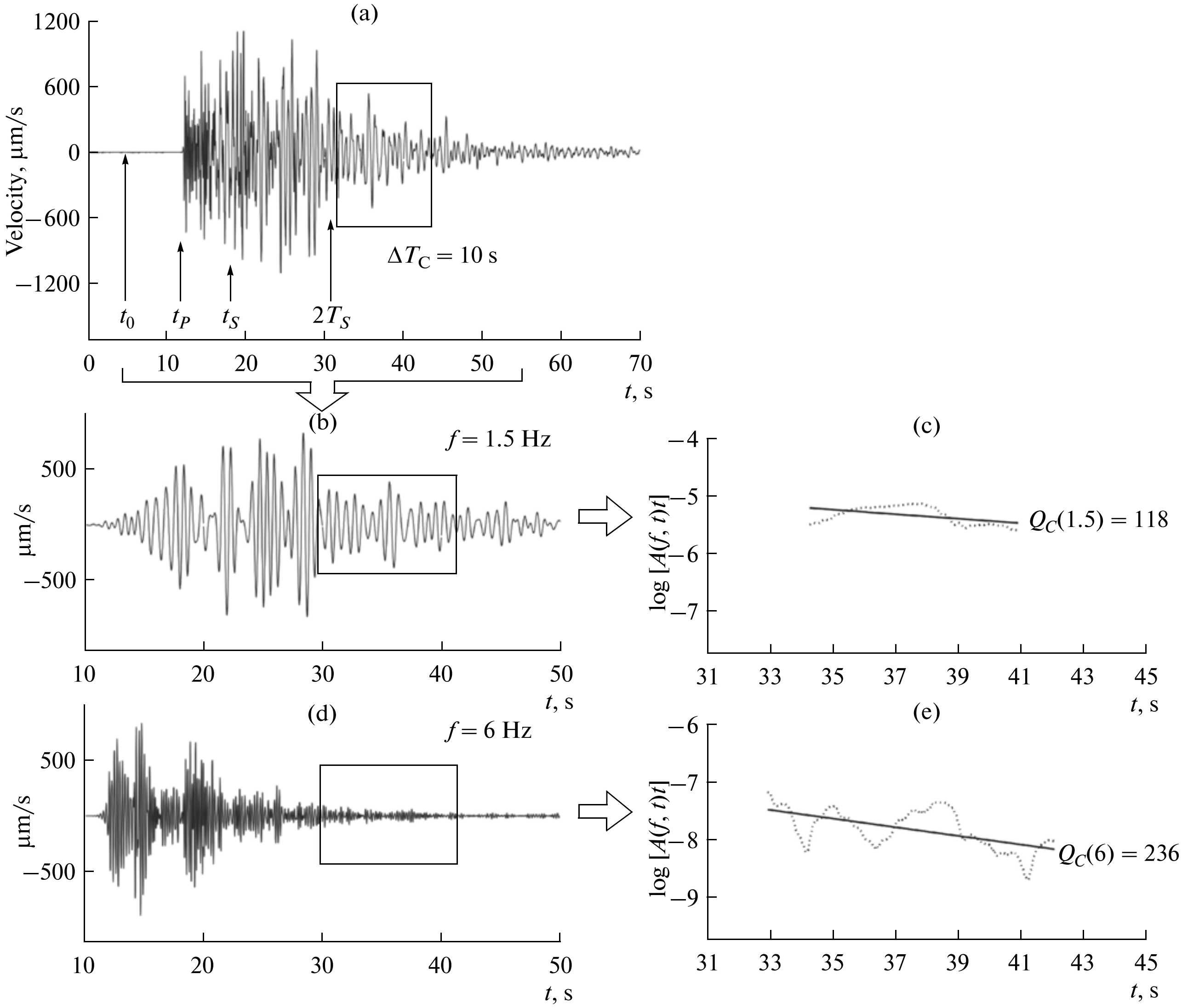

Fig. 2. Initial record of the earthquake with a magnitude of M_L 3.1 (a) and filtered fragments of the record and the corresponding coda waves (with correction for the geometrical spreading) with central transmission frequencies of 1.5 Hz (b, c) and 6.0 Hz (d, e).

average, the time window width for the SH-waves from the moment of the first event was no less than 2 s.

Figure 4 demonstrates the amplitude spectrum of the displacements in the S-wave with correction for the instrumental bias and the medium absorption for an event with a magnitude of M_L 1.8 and the distance to the station of 15 km, as well as the spectrum simulated for the given earthquake according to the Brune model. As is seen from Fig. 4, the observed and simulated (4) spectra practically coincide within the frequency region from 1 to 10 Hz.

Using expression (12) in the computations and taking into account the aforementioned corrections, we subsequently estimated the seismic moment M_0 from the spectral parameters of the earthquake records, and the summary value of the scalar seismic moment was computed by averaging over all the records of the given event:

$$M_0 = anti\log\frac{1}{N}\sum_i \log M_0^{(i)}, \qquad (13)$$

where N is the number of observations.

The calculations were done on the basis of the preliminarily performed data base and wave forms using an automated algorithm coded in FORTRAN for the computation of the spectral parameters. The mean value of the SH-wave radiation pattern in the source $R_{\theta\varphi}$ was taken to be 0.41 for [1], the medium density $\rho = 2400$ kg/m^3, and the S-wave velocity was $v_S =$ 3.0 km/s according to the DSS data [11].

In addition to the seismic moment M_0, the digital records were used to estimate the source radius R. In order to avoid difficulties related to the visual determi-

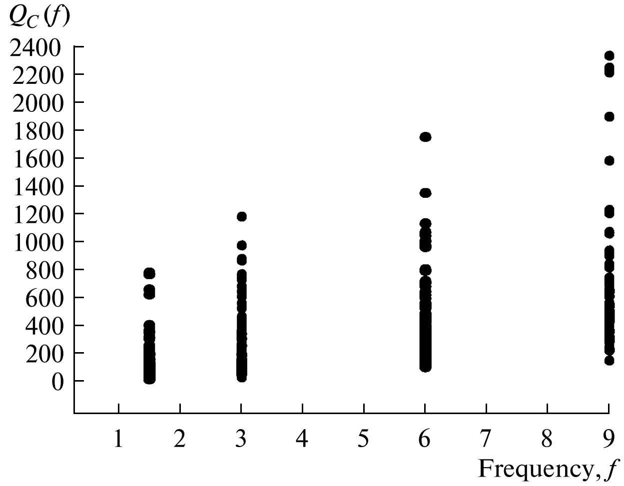

Fig. 3. Values of Q_C for the significant shocks in the aftershock sequence of the 2006 Gornozavodsk earthquake calculated from the digital records of the seismic stations.

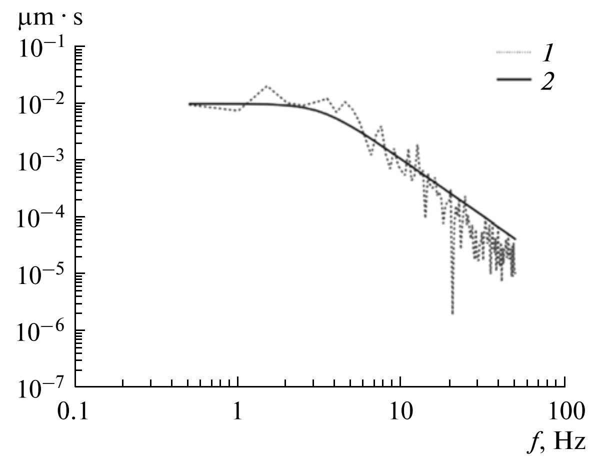

Fig. 4. Amplitude displacement spectra.
(*1*) amplitude displacement spectrum of the observed SH-wave seismograms for the event with M_L 1.8 corrected for the absorption and the amplitude–frequency characteristics of the apparatus, and (*2*) the amplitude displacement spectrum according to the Brune model [18].

nation of the corner frequency of the cutoff f_0, the source size was determined using a Snoke algorithm [31]. The latter is based on the values of the low-frequency part of the displacement spectrum and on the integral of the squared velocity spectrum, which decreases the dependence of the obtained source sizes on the local ground conditions beneath the considered seismic stations:

$$R = 2.34 v_S \left(\frac{\Omega_0^2}{4J}\right)^{1/3}, \tag{14}$$

where $J = \int \{A'(f)\exp[-\pi t f/Q(f)]\}^2 df$ is the released seismic energy, v_S is the *S*-wave propagation velocity, and $A'(f)$ is the spectrum of the displacement velocity.

Figure 5 demonstrates the distribution of the seismic moment M_0 of the sources of the 2004 Kostromskoe earthquake swarm depending on the size (radius) of the source R. The lines in Fig. 5 show the isolines of the released stress $\Delta\sigma$ calculated from the formula for the model of the source in the form of a circular crack [8]:

$$\Delta\sigma = \frac{7M_0}{16R^3}. \tag{15}$$

As is seen from Fig. 5, the source size (radius) does not depend on the seismic moment, since the calculations were not corrected for the effect of the local ground conditions beneath the stations, which provides a constant value of the corner frequency of the cutoff f_0 and, as a result, the source size R.

Let us estimate the average value κ and the correction factor for the absorption beneath the station. The relations between the size (radius) R and the corner-frequency f_0 according to Brune [19] can be written as

$$R = \frac{0.37 v_S}{f_0}. \tag{16}$$

As follows from relation (16) and Fig. 5, f_0 for $R \approx$ 100–200 m is approximately 7 Hz. According to the determination of the corner frequency as the frequency at which the level of the displacement spectrum $U(f)$ accounts for half of the level of the flat area U_0 of the spectral curve, i.e., $\exp(-\pi\kappa f_0) = 0.5$, we obtain $\kappa \approx 0.03$.

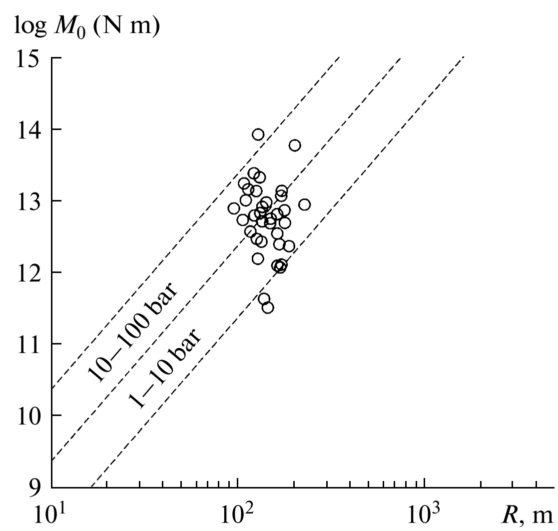

Fig. 5. Distribution of the seismic moment M_0 versus the source size R for the 2004 Kostromskoe earthquake swarm.

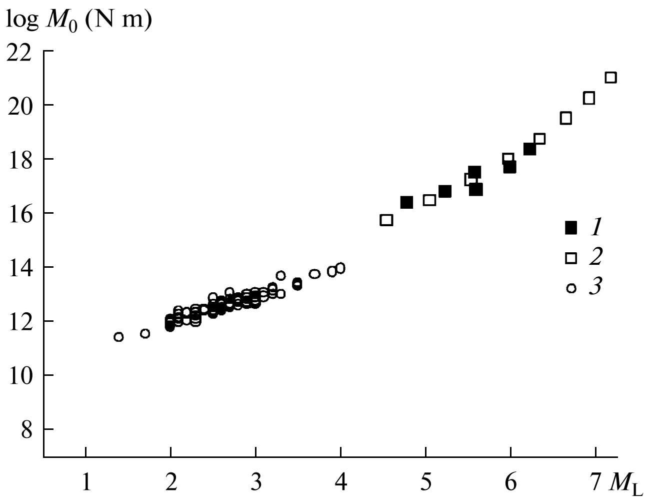

Fig. 6. Dependence of the seismic moment M_0 on the local magnitude M_L.
(*1*) the values M_0 and M_L for the significant earthquakes in Southern Sakhalin during 2001–2010 taken from the NEIC catalogues [34] and the SF GS RAS, respectively; (*2*) the world average dependence from work [4]; and (*3*) the values of M_0 and M_L obtained in this work.

3. The calculated scalar seismic moments were used to construct the correlation dependence M_0 on the local magnitude M_L. The obtained dependence is defined by a loglinear function:

$$\log M_0 \pm 0.08 = 0.95 M_L + 10.18, \quad (17)$$

with a 91% correlation radius, where M_0 is given in N m.

It is seen from Fig. 6 that the average relations between the parameters in the working magnitude range ($1.4 \le M_L \le 4$) in a definite sense are consistent with the average dependence plotted according to the world data [4] for the magnitude range of $M_L > 4.5$. The value of the seismic moment for significant earthquakes that occurred in South Sakhalin within 2001–2010 was taken from the NEIC catalogue [34]. The correlation dependences of M_0 on the local magnitude M_L that were obtained in the works of foreign seismologists for other regions of the earth using this methodology have similar slopes and are approximated by similar values of the approximating straight line and free term: 1.12 and 10.46 for Central Italy [18], 0.99 and 11.10 for Northwestern Greece [29], and others.

The local magnitude of the seismic events M_L was estimated using empirical nomograms that link the energy class according to Rautian (K_p) with the magnitude (M): $K_p = 4.0 + 1.8$ M. A nomogram was developed by the Laboratory of Seismology at the Institute of Marine Geology and Geophysics of the Far East Branch of the Russian Academy of Sciences using the results of instrumental observations in the source zone of the 1996 Neftegrosk earthquake (M_W 7.2) and is recommended for the analysis of earthquakes only on Sakhalin Island. The applied characteristics of the magnitude M is an analogue of the local magnitude M_L [15]. Therefore, the obtained dependence (17) roughly describes the relations between M_0 and M_L.

CONCLUSIONS

The frequency dependence of the quality control $Q_C(f) \approx 60f$ obtained for South Sakhalin is comparable with similar estimates of the absorption for central Japan [33]. Thus, the estimated attenuation function can be applied for the determination of not only the local absorption in the source zone of the 2006 Gornozavodsk earthquake but also for the entire territory of South Sakhalin.

Algorithms described in works [17, 20, 28] can also be used for the identification of the anisotropy effect in the course of the seismic wave attenuation in response to the formation of scattered waves from inhomogeneities of different scales. This is of special importance for the geological medium of Sakhalin Island cut by a system of large submeridionally oriented faults.

An automated algorithm was developed for the first time for the determination of the scalar seismic moment of the source M_0 under Sakhalin Island conditions, and mass estimates of the seismic moment were obtained for weak earthquakes in the south of the island using the frequency function of the quality control calculated on the basis of digital records from several stations. The obtained loglinear correlation between the seismic moments of the source M_0 and the local magnitude M_L within the magnitude range of $M_L = 1.4 \pm 4.0$ is consistent in a certain sense with the mean dependence according to the world data [4] and close to the analogous correlation dependences for other regions [18, 29].

The presented methods and results of the determination of the scalar seismic moment of the source according to the digital records from several seismic stations allow the principle solution of the problem of the moment tensor determination. The knowledge of the seismic moment tensor provides insight into the dynamics of the seismic activity in the light of the deterministic models of the fault interaction, which use, for example, the term of the flowing rock masses or the transfer of the Coulomb interaction [32].

REFERENCES

1. K. Aki and P. Richards, *Quantitative Seismology: Theory and Methods* (Mir, Moscow, 1983; Freeman, San Francisco, 1980).
2. N. F. Vasilenko, A. S. Prytkov, Ch. U. Kim, and X. Takakhashi, "Coseismic Deformations of the Earth's Surface in Sakhalin Related to the August 2, 2007, M_w = 6.2 Nevelsk Earthquake," Tikhookean Geol. **28** (5), 16–21 (2009) [Russ. J. Pac. Geol. **3**, 424–428 (2009)].

3. O. D. Gotsadze, V. I. Keilis-Borok, I. V. Kirillova, et al., *Study of the Earthquake Source Mechanism* (Akad. Nauk SSSR, Moscow, 1957) [in Russian].

4. A. A. Gusev and V. N. Mel'nikova, "Relation between World Average Magnitudes and Those for Kamchatka," Vulkanol. Seismol., No. 6, 55–63 (1990).

5. A. V. Konovalov, Extended Abstract of Candidate's Dissertation in Mathematics and Physics (IMGiG DVO RAN, Yuzhno-Sakhalinsk, 2006).

6. A. V. Konovalov, A. I. Ivashchenko, Ch. U. Kim, and A. S. Sychev, "Structure and Peculiarities of the Seismic Regime in the Source Zone of the Takoe Earthquake on September 1, 2001 (M_W 5.2)," Tikhookean Geol. **26** (2), 93–101 (2007) [Russ. J. Pac. Geol. **1**, 186–193 (2007)].

7. Yu. F. Kopnichev, "On the Role of Multiple Scattering in the Formation of Tail Part of Seismogram," Fiz. Zemli, No. 6, 41–48 (1977).

8. B. V. Kostrov, *Source Mechanics of the Tectonic Earthquake* (Nauka, Moscow, 1975) [in Russian].

9. B. V. Levin, Ch. U. Kim, and I. N. Tikhonov, "The Gornozavodsk Earthquake of August 17(18), 2006, in the South of Sakhalin Island," Tikhookean. Geol. **26** (2), 102–108 (2007) [Russ. J. Pac. Geol. **1**, 194–199 (2007)].

10. V. K. Lemzikov, "Intrinsic Attenuation and Scattering of Shear Waves in the Lithosphere of Kamchatka," Vulkanol. Seismol., No. 3, 50–63 (2007) [J. Volcanol. Seismol. **1**, 185–197 (2007)].

11. M. Kh. Livshits, "Deep Structure of Sakhalin on the Basis of Geophysical Data," Geofiz. Sb. Yuzhno-Sakhalinsk, **24** (2), pp. 16–25 (1972).

12. V. L. Lomtev, S. P. Nikiforov, and Ch. U. Kim, "Tectonic Aspects of the Crustal Seismicity of Sakhalin," Vestn. Dal'nevost. Otd. Ross. Akad. Nauk, No. 4, 64–71 (2007).

13. A. S. Prytkov, Extended Abstract of Candidate's Dissertation in Mathematics and Physics (IMGiG DVO RAN, Yuzhno-Sakhalinsk, 2008).

14. Yu. V. Riznichenko, "Sizes of Source of the Crustal Earthquake and Seismic Moment," in *Studies on the Earthquake Physics* (Moscow, 1976), pp. 9–27 [in Russian].

15. D. A. Safonov, Extended Abstract of Candidate's Dissertation in Mathematics and Physics (IMGiG DVO RAN, Yuzhno-Sakhalinsk, 2008).

16. K. Aki, "Earthquake Mechanism," in *The Upper Mantle Developments in Geotectonics* (Elsevier, Amsterdam, 1972), Vol. 4, pp. 423–446.

17. K. Aki and B. Chouet, "Origin of Coda Waves: Source, Attenuation and Scattering Effects," J. Geophys. Res. **80**, 3322–3342 (1975).

18. D. Bindi, D. Spallarossa, P. Augliera, and C. Cattaneo, "Source Parameters Estimated from the Aftershocks of the 1997 Umbria-Marche (Italy) Seismic Sequence," Bull. Seism. Soc. Am. **91** (3), 448–455 (2001).

19. J. N. Brune, "Tectonic Stress and the Spectra of Seismic Shear Waves from Earthquakes," J. Geophys. Res. **75** (26), 4997–5009 (1970).

20. L. S. Gao, L. C. Lee, N. N. Biswas, et al., "Comparison of the Effects Between Single and Multiple Scattering on Coda Waves for Local Earthquakes," Bull. Seism. Soc. Am **73**, 377–389 (1983).

21. A. A. Gusev and I. R. Abubakirov, "Simulated Envelopes of Non-Isotropically Scattered Waves As Compared to Observed Ones: New Evidence for Fractal Heterogeneity," Geoph. J. Int. **127**, 49–60 (1996).

22. J. L. Hardebeck and P. M. Shearer, "Using SIP Amplitude Ratios To Constrain the Focal Mechanisms of Small Earthquakes," Bull. Seism. Soc. Am **93** (6), 2434–2444 (2003).

23. J. Havskov, S. Malone, D. McClurg, and R. Crosson, "Coda Q for the State of Washington," Bull. Seism. Soc. Am. **79**, 1024–1038 (1989).

24. M. Hoshiba, H. Sato, and M. Fehler, "Numerical Basis of the Separation of Scattering and Intrinsic Absorption from Full Seismogram Envelope—A Monte-Carlo Simulation of Multiple Isotropic Scattering," Pap. Meteorol. Geophys. (Bull. of Meteorol. Res. Inst.) **42**, 65–91 (1991).

25. B. R. Julian and G. R. Foulger, "Earthquake Mechanisms from Linear-Programming Inversion of Seismic-Wave Amplitude Ratios," Bull. Seism. Soc. Am. **86**, 972–980 (1996).

26. N. Kumar, I. A. Parvez, and H. S. Virk, "Estimation of Coda Wave Attenuation for NW Himalayan Region Using Local Earthquakes," Phys. Earth Planet. Inter. **151**, 243–258 (2005).

27. L. Malagnini, R. Herrmann, and K. Koch, "Regional Ground Motion Scaling in Central Europe," Bull. Seism. Soc. Am. **90**, 1052–1061 (2000).

28. T. G. Rautian and V. I. Khalturin, "The Use of the Coda for Determination of the Earthquake Source Spectrum," Bull. Seism. Soc. Am. **68**, 923–948 (1975).

29. Z. Roumelioti, A. Kiratzi, N. Theodoulidis, and C. Papaioannou, "S-Wave Spectral Analysis of the 1995 Kozani-Grevena (NW Greece) Aftershock Sequence," J. Seismol. **6**, 219–236 (2002).

30. J. A. Snoke, J. W. Munsey, A. C. Teague, and G. A. Bollinger, "A Program for Focal Mechanism Determination by Combined Use of Polarity and *SV–P* Amplitude Ratio Data," Earthquake Notes **55** (3), 15 (1984).

31. J. A. Snoke, "Stable Determination of (Brune) Stress Drop," Bull. Seism. Soc. Am **77**, 530–538 (1987).

32. R. S. Stein, "The Role of Stress Transfer in Earthquake Occurrence," Nature **402**, 605–609 (1999).

33. A. L. Stork and H. Ito, "Source Parameters Scaling for Small Earthquakes Observed at the Western Nagano 800-M-Deep Borehole, Central Japan," Bull. Seism. Soc. Am. **94** (5), 1781–1794 (2004).

34. http://earthquake.usgs.gov/earthquakes/eqarchives/epic/epic_rect.php

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