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Determination of Earthquake Parameters

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The paper describes the determination of the earthquake parameters on the basis of the macroseismic data and the determination of the source parameters on the basis of the microseismic data.

В статье описывается определение параметров землетрясений на основе макросейсмических данных и определение очаговых параметров на основе микросейсмических данных.

Článek popisuje určení parametrů zemětřesení na základě makroseismických údajů a zdrojových parametrů na základě mikroseismických dat.

1. Determination of Earthquake Parameters from Macroseismic Observations

The geographical coordinates of the macroseismic epicentre, i.e. of the centre of gravity of the isoseismal of the highest intensity can be read directly from the isoseismal map. For the determination of the epicentral intensity, the focal depth and the absorption coefficient the Kövesligethy formula and the Blake formula are used i.e.

$$I_0 - I_n = 3 \log (D_n/h) + 3(D_n - h) \alpha \log e \tag{1}$$

assuming the model of energy propagation

$$E = E_0 \exp(-\beta D_n) D_n^{-no} ;$$

$$I_0 - I_n = k \log (D_n/h) \tag{2}$$

assuming the model of energy propagation

$$E = E_o D_n^{-n} ,$$

where I_0 is the epicentral intensity, $D_n = \sqrt{(r_n^2 + h^2)}$, where r_n is the isoseismal radius corresponding to intensity I_n, h is the focal depth, α the absorption coefficient, k is the numerical parameter, E_0 is the seismic energy radiated from the focus, E is

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the seismic energy in the hypocentral distance D_n and β, no, n are numerical parameters.

Both formulae assume the relationship between the seismic energy and the intensity in the form $\log E = k_1 + k_2 I$, where E is the seismic energy, I the intensity and k_1, k_2 the numerical constants. Considering this relationship we can derive the parameters characterizing the absorption of seismic energy from the parameters characterizing the absorption of intensity.

For the model described by the Blake's formula we obtain $n = k \,.\, k_2$ and for the model described by the Kövesligethy's formula we obtain $no = 3k_2$, $\beta = 3\alpha k_2$.

The Kövesligethy formula is a transcendent equation with unknown parameters h, α, I_0. It can be solved numerically by the following method: for each earthquake with a number of isoseismal radii greater than or equal to three, the mean square deviations from the curve $f(I_0, I_n, r_n, h, \alpha) = 0$ for all $\alpha \in (0.001, 0.050)$ with the step 0.001, for all $h \in (1\text{ km}, 60\text{ km})$ with the step 1 km in the case of shallow earthquakes and $h \in (1\text{ km}, 200\text{ km})$ for intermediate earthquakes and for all $I_0 \in$ (the highest observed I_n, 12) with the step of 0.5° MSK-64 are determined. The values of I_0,h and α with the smallest mean square deviations are chosen as representative values. Very often I_0 is greater that the maximum observed intensity I_n by 0.5° of the intensity scale; when we put I_0 equal to the maximum observed intensity, we obtain a high value of the focal depth that does not agree with the instrumental data.

From Eq. (1) it follows for the absorption coefficient:

$$\alpha_n = (I_0 - I_n - 3\log(D_n/h)) : (3(D_n - h)\log e)\,; \tag{3}$$

$$\mathrm{d}\alpha/\mathrm{d}h = (3r_n^2 \log e + (I_0 - I_n - 3\log(D_n/h))\, D_n h) : (3(D_n - h)\, D_n^2 \log e)\,.$$

α is a continuous function of h for $r_n > 0$. For $(I_0 - I_n - 3\log(Dn/h)) \geqq 0$ and in the case of the opposite inequality for $(3\, r_n^2 \log e + (I_0 - I_n - 3\log(D_n/h))\, D_n h) > 0$, $\mathrm{d}\alpha/\mathrm{d}h > 0$; it means that α is an increasing function of h. For $(3\, r_n^2 \log e + (I_0 - I_n - 3\log(D_n/h))\, D_n h) = 0$, $\mathrm{d}\alpha/\mathrm{d}h = 0$; it means also that α remains constant with increasing h. For $(3\, r_n^2 \log e + (I_0 - I_n - 3\log(D_n/h))\, D_n h) < 0$, $\mathrm{d}\alpha/\mathrm{d}h < 0$, α is a decreasing function of h. All classes are observed in seismological practice.

In the Blake formula the value of h represents the mean value of h_n determined for individual I_n if we consider $I_0 \in$ (maximum observed intensity I_n, 12) with the step of 0.5° of the intensity scale and requesting the smallest mean square deviation from the curve $f(I_0, I_n, r_n, h, k) = 0$. On average I_0 is by 0.5° greater than the maximum observed intensity. From Eq. (2) it follows for the parameter k:

$$k_n = (I_0 - I_n) : \log(D_n/h)\,; \tag{4}$$

$$\mathrm{d}k/\mathrm{d}h = (I_0 - I_n)\, r_n^2 : h\, D_n^2 (\log(D_n/h))^2\,, \quad \text{i.e.}$$

k is a continuous function for $h > 0$; $\mathrm{d}k/\mathrm{d}h > 0$ indicates that k is an increasing function of h.

Assuming $k = 3 + k_1$ we obtain a relation between the absorption coefficient α and the parameter k:

$$\alpha = [(\log (D_n/h)) : (3 (D_n - h) \log e)] \, k_1 .$$

This relationship shows that if h is constant or changes only slightly, the absorption coefficient α is directly proportional to the parameter k.

In order to obtain the same numerical values of the focal depth by the formulae (1) and (2) it is necessary to determine the parameter k for individual focal regions. We assume h(Blake)-h(Kövesligethy) $= a + bk$, where a, b are the numerical parameters; a and b are determined by the least-squares method and $k = -a/b$ is the resulting value of the parameter k.

In individual focal regions the anomalous propagation of the seismic energy in different azimuths can be described by the values of α and k in those azimuths. Eqs (3) and (4) hold for $n = 1, 2, \ldots, s$; r_n corresponding to I_n and $I_1 > I_2 > \ldots$ $\ldots > I_s$. α_n and k_n as a rule increase with increasing n; $\alpha_n = 0$ corresponds to $k_n = 3$; $\alpha_n < 0$ corresponds to $k_n < 3$; α_n and k_n are too great for r_n to be comparable with the focal depth h; for $r_n > 2.5h$ the values of α_n and k_n have practically a small dispersion (the observed variations are connected with the geological structure) and, therefore,

$$k = \sum_{i=j}^{s} k_i/(s - j + 1) \tag{5}$$

$$\alpha = \sum_{i=j}^{s} \alpha_i(s - j + 1) \tag{6}$$

for $r_{j-1} \leqq 2.5h$ and $r_j > 2.5h$:

2. Methodical Questions of Source Parameters Determination

Theoretical research of the earthquake source allows to use the amplitude spectra of the P and S waves for determination of the source parameters: seismic moment, stress drop, source dimension, average relative displacement of two sides of a fault, energy emitted from the focus in the form of seismic waves, etc.

In the case of stronger earthquakes, especially shallow ones, the surface waves, particularly the Rayleigh waves are clearly recorded. The amplitude spectra are clear as well, Fig. 1 [1, 2]. If we use the same relationship for the computation of the seismic moment on the basis of the surface waves as for the body waves and if we use the phase velocity of the surface waves we obtain approximately: $M_0(L) : M_0(B) \doteq 10$ for shallow earthquakes and $M_0(L) = M_0(B)$ for intermediate ones. Using this result we can correct the numerical results obtained on the basis of the surface waves. The relationship for the computation of the source dimension from amplitude spectra of surface waves was statistically derived in the form: $r = 0.273 v_L/f_0$, where r is the source dimension, v_L the phase velocity and f_0 the corner frequency [2].

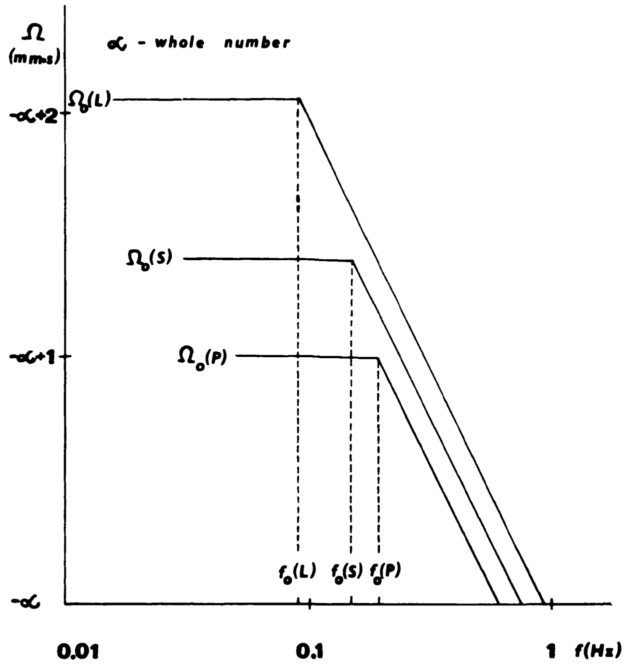

Fig. 1. Comparison of the amplitude spectra of the P, S and surface waves for the European shocks.

Fig. 2 shows the amplitude spectra of the coda waves. It is notworthy that in this case the long-period spectral level is evident. Several authors interpreted the coda waves as the reflected *S* waves and found a relation between coda and body waves for individual focal regions. They use also coda waves for the source parameters determination. It is necessary to derive the relation for Europe. The numerical values of the source parameters depend on the formula used for computation [1, 2].

When computing source parameters we must take the following into acount:

(a) The sampling period of the input signal must be chosen sufficiently small so that the highest frequencies in a given signal are not lost. The sampling period must contain at least ten points for the smallest period [1, 2].
(b) The choice of the sample length can also influence numerical values of the computed amplitude spectra. The sample length must contain the whole wave group and allow the determination of the long-period spectral value. With increasing the sample length the long-period spectral amplitude increases and the corner frequency decreases [2]. The substantial changes occurred in cases in which the longer sample contains the next wave group; if short and long samples contain only one wave group, the changes of the long-period spectral amplitude and the corner frequency are negligible.

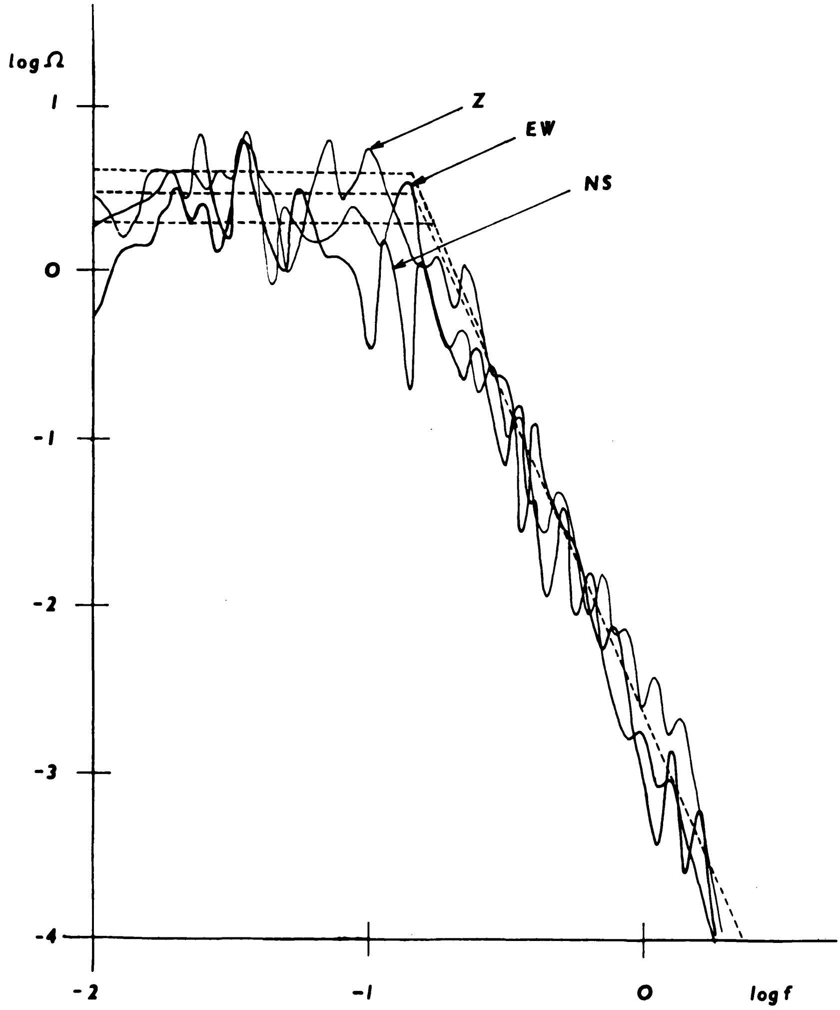

Fig. 2. Amplitude spectra of coda waves of Vrancea earthquake of March 4, 1977 calculated on the basis of the record of the FBV seismograph located on the seismic station Kašperské Hory.

(c) For the calculation of amplitude spectra several numerical methods can be used; it is generally known that spectra can differ not only by numerical values but also by the form [2]. For computation of amplitude spectra the FFT and the Fillon methods are available.

(d) Seismic waves and their spectra depend on the azimuth between the seismic station and the fault, on the geological structure in the source region and below the station, on the properties of the medium in which waves propagate and on the recording instrument. The instrumental response can be corrected by the relation $\Omega_E(f) = \Omega_C(f)/H(f)$, where $\Omega_C(f)$ is the computed amplitude spectrum, $H(f)$

is the amplitude-frequency response of the instrument and $\Omega_E(f)$ is the corrected spectrum. For calculation of spectra the records of broad-band instrument are most convinient.

The influence of the inelastic attenuation of seismic waves on the amplitude spectrum is given by the relationship $\Omega_F(f) = \Omega_D(f) \exp(D\pi f/Qv)$, where $\Omega_F(f)$ is the spectrum radiated from the focus, $\Omega_D(f)$ the spectrum in the hypocentral distance D, v the wave velocity and Q the attenuation coefficient. In [1, 2] it is shown that the numetical values of the amplitude spectra up to 0.1 Hz do not change practically, it means that also the numerical value of the long-period level remains almost constant. However, the numerical value of the corner frequency may change: its shift to higher values causes a decrease of the source dimension; the magnitude of this shift depends on the hypocentral distance, the wave velocity and the attenuation coefficient in the above formula. The theoretical calculations show that orientation of the fault in respect to seismic stations does not practically influence the long-period spectral values, but it influences the corner frequency; the smallest corner frequency should be observed in the direction in which the fracture is spreading; it means that in principle, azimuthal variation of the amplitude spectra provides information on the direction of the fault propagation. According to the results of several seismologists it seems that in distances greater than 20°,the amplitude spectra become smoother and do not show any azimuthal variations.

(e) The inequalities among the spectral levels found for the components NS, EW, Z repeat for the P, S and L phases and they are observed in a reverse sense for the corner frequencies, e.g. $\Omega_0(EW) > \Omega_0(NS) > \Omega_0(Z)$ and $f_0(EW) < f_0(NS) < f_0(Z)$ (2). The ratio among the components EW, NS and Z depends probably on the orientation of the fault in respect to the seismic station and on the mechanism. From the physical point of view it is necessary to consider the whole vector in the calculation of source parameters and not the individual components. The amplitude spectrum of the whole vector is given by the relationship $\Omega(f) = \sqrt{([\Omega_{EW}(f)]^2 + [\Omega_{NS}(f)]^2 + [\Omega_Z(f)]^2)}$, where $\Omega_{EW}(f)$, $\Omega_{NS}(f)$ and $\Omega_Z(f)$ are the amplitude spectra of the individual components, the details are in [2].

(f) Before the calculation of the spectrum the data must be corrected for the zero line variation. The minimum step is to subtract the mean amplitude value so that we obtain a stationary series with zero mean value and avoid the distortion of the spectrum and the occurrence of very high spectral values [2].

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