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Regional spectral scaling relations of source parameters for earthquakes in the Baltic Shield

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Abstract

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Spectral analysis of regionally recorded Lg waves is performed to determine source parameters such as seismic moment, source radius and stress drop of earthquakes in the Baltic Shield, and to derive regional spectral scaling relations. The data consist of about 350 Lg phases on short-period, vertical-component analog seismograms from earthquakes with magnitude (M_L) ranging from 2 to 5.2. Source radii estimated from the corner frequency show only a slight increase with increasing seismic moment in the range 3×10^{18} to 5×10^{20} dyn cm, while the increase is more rapid for larger events. Baltic Shield earthquakes show increasing stress drops, ranging from about 0.1 to 10 bars, for increasing seismic moment. The relatively low stress drops could, in part, be explained by bias due to band-limited analog data.

The slope of the curve relating seismic moment to corner frequency is steeper than -3 , which suggests a departure from a constant stress drop scaling relation proposed for large earthquakes. The derived relationship between the seismic moment (M_0) and magnitude (M_L) is: $\log M_0 = 16.93 + 1.01 M_L$ (for $2 \leq M_L \leq 5.2$).

1. Introduction

In the Baltic Shield, which extends from the Caledonian Mountains in the west to the north-western U.S.S.R. in the east, the current seismicity is relatively low. Nearly all instrumentally recorded earthquakes are smaller than magnitude 4 (M_L). So far, only a few studies have been performed to determine dynamic source parameters and focal mechanisms (see, e.g., Slunga, 1979; Bungum et al., 1982; Kulhánek et al., 1983; Slunga et al., 1984; Kim et al., 1985).

Recent studies on the excitation and propagation of regional crustal phases have demonstrated that the Lg-wave amplitude spectrum can be used to determine seismic source parameters

such as seismic moment and stress drop (Herrmann and Kijko, 1983; Campillo et al., 1984). The similarity for small earthquakes between the amplitude spectrum of Lg waves, and that of shear waves predicted by Brune's (1970) source model, has been used by Street et al. (1975) and Hasegawa (1983).

The main purpose of the present study is to quantify source parameters of earthquakes in the Baltic Shield and to derive empirical relationships between the source parameters. Attempts have been made to establish spectral scaling relations for earthquakes in the shield, and to compare them with those for other tectonically similar regions, such as eastern Canada and the eastern United States.

TABLE 1
Earthquake data

Date	Origin time			Epicenter		Magnitude M_L	M_0^a	f_0^a	No. obs.
	h	m	s	(° N)	(° E)				
28 Sept. 1962	17	22	1	64.5	20.5	4.0	8.3E20	2.3	4
1 Aug. 1963	16	2	10	62.6	27.9	3.7	3.4E20	3.2	6
3 Oct. 1963	2	5	32	64.7	23.0	3.2	1.5E20	2.9	6
23 Jan. 1965	11	9	39	65.0	23.2	3.3	2.6E20	3.1	7
20 Mar. 1965	2	44	54	67.4	25.9	3.5	1.4E20	3.5	7
22 Jan. 1966	23	46	47	66.7	32.4	3.6	2.2E20	3.4	9
22 June 1966	0	50	38	66.9	18.8	2.2	2.2E19	3.6	2
14 Oct. 1966	16	8	25	65.2	22.6	2.7	3.3E19	3.9	2
4 Jan. 1967	4	44	18	67.9	21.0	3.2	1.4E20	2.9	3
4 Feb. 1967	15	34	56	59.5	13.3	3.7	3.8E20	3.8	4
13 Apr. 1967	8	46	19	68.1	20.8	3.7	2.8E20	3.5	6
13 Apr. 1967	9	3	48	63.2	18.9	3.6	5.3E20	2.5	7
20 May 1967	23	18	12	66.4	33.4	5.2	1.6E22	1.6	2
29 Nov. 1967	9	25	27	60.6	17.7	3.0	5.0E19	3.5	1
1 June 1968	20	29	59	63.4	21.2	2.6	1.8E19	3.8	4
2 Sept. 1968	21	34	40	66.9	23.9	2.2	1.7E19	4.0	1
4 Sept. 1968	17	9	14	66.9	23.8	3.4	5.2E20	2.2	3
24 Oct. 1968	20	13	51	69.1	32.7	3.6	4.0E20	3.1	7
13 Feb. 1969	18	4	5	59.4	13.1	2.7	7.2E19	4.1	2
15 Feb. 1969	4	31	21	65.3	22.9	2.1	3.6E18	5.9	1
23 May 1969	18	40	25	66.0	27.2	3.3	2.3E20	3.1	6
20 July 1969	7	0	53	67.0	22.9	2.8	4.1E19	3.8	4
6 Mar. 1970	12	4	23	65.0	27.5	2.3	1.7E19	5.8	1
12 May 1970	14	14	13	61.0	12.8	3.2	1.6E20	3.0	6
14 June 1970	15	38	13	62.0	17.5	2.0	1.6E19	3.0	3
12 Aug. 1970	19	28	46	61.5	16.4	3.0	7.0E19	3.7	4
24 Feb. 1971	9	6	10	62.3	27.1	2.1	9.3E18	4.3	1
10 July 1971	4	12	13	58.3	13.2	2.5	7.4E19	3.2	2
28 July 1971	23	24	54	62.1	17.3	3.1	7.1E19	3.4	5
10 Oct. 1971	5	29	6	61.9	21.9	2.5	3.6E19	4.9	2
12 June 1972	4	31	33	60.1	14.5	2.8	5.3E19	3.6	4
2 Oct. 1972	19	46	5	68.3	24.2	2.0	9.5E18	6.4	1
16 Dec. 1972	10	9	27	63.5	19.7	2.8	4.1E19	3.8	6
7 Jan. 1973	16	35	11	66.9	31.4	3.7	3.1E20	3.2	10
11 Apr. 1973	5	1	37	58.8	13.4	3.9	5.1E20	2.3	3
17 Apr. 1973	6	17	59	67.9	20.0	3.3	2.7E20	2.9	2
10 Sept. 1973	21	41	9	65.7	28.8	2.7	4.1E19	4.8	1
26 Nov. 1973	21	45	38	62.9	18.5	3.2	1.0E20	3.1	4
10 Dec. 1973	20	3	50	66.6	25.9	3.6	2.6E20	3.5	10
10 Dec. 1973	20	7	56	66.6	25.7	2.9	1.2E20	3.0	4
5 Feb. 1974	22	33	59	58.1	14.0	2.4	3.4E19	3.7	2
21 May 1974	16	51	21	58.3	12.8	3.4	2.8E20	3.1	7
4 June 1974	23	13	51	62.3	17.2	3.6	3.2E20	3.2	11
21 June 1974	6	31	8	66.0	27.3	3.6	3.4E20	3.2	11
11 Aug. 1975	18	28	9	67.5	22.5	3.9	5.3E20	2.5	7
12 Mar. 1976	23	22	21	61.2	17.0	3.3	1.5E20	3.3	8
30 Apr. 1976	12	53	53	64.4	21.1	3.1	1.0E20	3.0	8
12 July 1976	3	29	50	62.8	18.2	2.2	1.8E19	3.9	1
17 Aug. 1976	5	35	54	62.9	17.8	3.5	1.8E20	3.1	10
3 Sept. 1976	4	28	0	58.4	13.9	3.7	5.1E20	2.4	5
3 Oct. 1976	7	26	57	67.8	27.4	2.6	1.7E19	4.2	1

TABLE 1 (continued)

Date	Origin time			Epicenter		Magnitude M_L	M_0^a	f_0^a	No. obs.
	h	m	s	(° N)	(° E)				
25 Oct. 1976	8	39	45	59.3	23.3	4.9	2.3E22	1.3	3
14 Oct. 1977	3	41	49	67.0	23.4	2.9	4.9E19	3.8	2
7 Nov. 1977	0	34	32	68.8	27.5	3.5	3.0E20	3.4	8
2 July 1978	17	48	30	64.0	20.8	2.5	4.1E19	3.6	1
17 Aug. 1978	13	2	44	69.2	26.8	2.4	2.1E19	4.6	2
14 Dec. 1978	14	24	48	63.7	21.1	2.9	5.8E19	3.8	3
17 Feb. 1979	17	31	22	63.2	23.9	3.8	7.3E20	2.1	6
17 Feb. 1979	17	40	58	63.2	23.9	2.6	4.5E19	4.5	4
11 Nov. 1979	23	58	14	61.1	16.9	2.4	3.8E19	4.4	1
23 Dec. 1979	14	9	13	59.6	18.7	3.1	7.5E19	4.2	2
28 May 1980	10	36	45	67.7	24.5	3.1	1.1E20	2.7	4
29 July 1980	9	32	22	68.6	23.6	2.5	3.3E19	4.0	4
19 Aug. 1980	1	13	17	61.5	16.2	2.2	1.7E19	3.8	1
13 Sept. 1980	7	59	57	64.6	20.7	2.6	2.3E19	3.7	4
23 Oct. 1980	10	55	50	65.8	21.7	2.3	2.1E19	3.8	1
25 Oct. 1980	16	45	0	64.4	24.8	2.3	1.3E19	5.4	2
28 Nov. 1980	0	1	33	64.8	29.2	2.2	1.1E19	4.4	1
5 Mar. 1981	7	20	33	66.9	24.2	2.4	2.2E19	4.0	2
10 Apr. 1981	19	43	36	68.8	36.1	4.5	6.0E21	1.4	7
2 June 1981	4	47	27	67.7	23.3	2.1	4.9E19	3.2	1
4 July 1981	9	47	52	65.0	21.0	2.8	3.9E19	3.9	6
10 July 1981	15	10	11	63.6	16.7	2.5	3.9E19	4.0	3
17 Aug. 1981	13	9	17	67.5	21.9	2.6	3.4E19	3.4	2
11 Nov. 1981	2	48	52	57.1	13.1	2.9	5.2E19	3.2	1
3 Feb. 1982	7	8	43	67.3	24.0	2.7	2.9E19	4.0	2
7 May 1982	11	37	32	65.5	21.3	2.1	5.6E18	4.9	1
23 June 1982	1	43	28	66.1	28.3	2.1	2.0E19	5.5	1
21 July 1982	4	59	47	60.9	26.2	2.9	7.9E19	3.9	3
8 Oct. 1982	5	28	56	64.5	20.7	2.9	6.2E19	4.1	5
18 Aug. 1983	6	6	8	65.5	21.8	2.4	3.2E19	5.2	1
29 Sept. 1983	5	3	25	63.8	17.5	4.1	2.3E21	1.8	10
10 Jan. 1984	11	16	23	65.5	22.6	2.2	8.9E18	2.6	1
14 May 1984	8	31	2	65.9	28.5	2.5	2.6E19	4.1	1
25 May 1984	13	18	1	65.9	28.5	3.2	2.0E20	3.2	5
3 Aug. 1984	22	42	9	65.9	28.5	2.8	6.2E19	3.6	4
25 Aug. 1984	19	39	14	67.9	19.5	3.1	1.1E20	3.7	4
24 Dec. 1984	8	2	26	66.0	28.6	2.2	2.5E19	4.0	1
8 Jan. 1985	1	50	33	64.8	20.1	3.0	8.4E19	2.9	4
14 July 1986	14	45	32	58.5	14.0	3.4	2.1E20	3.7	4

^a M_0 denotes seismic moment (dyn cm) and f_0 source-spectrum corner frequency (Hz).

2. Data

Ninety earthquakes in the Baltic Shield, occurring between 1962 and 1986, were selected for analysis (Fig. 1). The magnitude (M_L) of these events is between 2 and 5.2. All events with magnitude greater than 3.5 and occurring between 1962 and 1983 were included, whereas for smaller

events ($M_L < 3.5$), a representative selection was made based on magnitude and location. Event data are given in Table 1. All earthquakes are confined to the crust, but more exact focal depths cannot usually be determined.

About 350 Lg waves recorded on short-period, vertical-component analog seismograms from 23 stations in Fennoscandia were used (Fig. 1). Epi-

Fig. 1. Epicentral distribution of 90 earthquakes with magnitude (M_L) 2.0–5.2 selected for spectral analysis and locations of 23 seismograph stations providing short-period, vertical-component Lg-wave records. The size of each dot is proportional to the magnitude. The inset dot corresponds to magnitude 4.

central distances are in the range of about 50–1500 km. We have made sure that no Sn waves pollute the Lg data used. Twelve stations have a WWSSN-equivalent Benioff seismometer, five Grenet ($T_0 = 1.4$ s, $T_g = 0.7$ s), three Willmore Mark II ($T_0 = 0.8$ s, $T_g = 0.2$ s) and three Geotech S-13 ($T_0 = 0.8$ s). The use of analog seismograms imposes a limit on the highest frequency to be analysed to about 5–10 Hz.

Selected seismograms were photographically enlarged five to ten times and the portions containing the Lg wave were electronically digitized. The digitized data were detrended for linear trend and interpolated for an equal-time interval of 0.05 s using the piecewise continuous cubic polynomial (Wiggins, 1976). In many cases, records from the large events (magnitude usually greater than 3.5) at short distances cannot be used due to missing traces (oversaturation). Two typical digitized seismograms are shown in Fig. 2; a weak Lg trace

with high level of background noise (Fig. 2a), and a strong Lg trace with low level of noise (Fig. 2b).

3. Method of analysis

The digitized seismograms were Fourier transformed to obtain trace amplitude spectra. The time window used for the spectral conversion varies from about 10 to 30 s depending upon the distance and size of each event. Observed Lg-wave amplitude spectra were then corrected for instrument response to obtain estimates of the ground displacement spectral density. To minimize the influence of the noise, the displacement spectra were considered only in a frequency band for which the instrument response is greater than one tenth of the peak magnification. This criterion delimits the available frequency range to about 0.3–8 Hz for most of the instruments used. In some cases, displacement spectra are reliable only up to about 5 Hz.

Fig. 2. Digitized seismograms and corresponding ground displacement amplitude spectra. a. 17 February 1979, $M_L = 2.6$, recorded at Nurmijärvi (NUR) at a distance of 302 km. The maximum peak-to-peak trace amplitude is 0.85 mm. A time window of 12.3 s (between arrows) is used for the spectral conversion. b. 4 September 1968, $M_L = 3.4$, recorded at Kiruna (KIR) at a distance of 179 km. The maximum peak-to-peak trace amplitude is 16.01 mm. A time window of 22.0 s is used. Note that (a) and (b) are plotted with different time scales. In the spectra, selected low-frequency levels and high-frequency asymptotes (slope-2) are indicated by straight lines, and corner frequencies by arrows.

From displacement spectra, the constant low-frequency spectral level (Ω_0) and the corner frequency (f_R) were determined for each event and station. Examples of spectra and corresponding seismograms are shown in Fig. 2. For the plotted spectra, amplitudes at frequencies above f_R fall off with increasing frequency, with a slope of about -2 .

3.1. Determination of low-frequency spectral level, Ω_0

The effect of anelastic attenuation is not considered in determining Ω_0 from the Lg spectrum, since it has insignificant effect upon the estimates of the spectral densities in the low-frequency band ($f \leq 1$ Hz) at which the Ω_0 levels are set in most

cases. The Lg-wave amplitude decrement due to anelastic attenuation is proportional to $\exp(-\gamma R)$, where R is distance and γ is the coefficient of anelastic attenuation, which, in its turn, is related to the quality factor, Q (Nuttli, 1986):

$$\gamma(f) = \pi f / [U(f)Q(f)]$$

where f is the wave frequency and U the Lg-wave group velocity. The reported value of γ for Lg waves at 1 Hz in the central United States (Nuttli, 1973) and southeastern United States (Bollinger, 1979) is 0.0006 km^{-1} . Chun et al. (1987) reported $\gamma(f) = 0.0008 f^{0.81} \text{ km}^{-1}$ for eastern Canada. According to these results, neglecting the effect of anelastic attenuation would introduce a maximum error of a factor of about three at a distance of 1000 km for the 1 Hz spectral component. Most of our observations were made at distances less than 1000 km, in which cases the error would be even smaller. Note that Ω_0 used in the M_0 computation (Section 4.1) is measured from the peak values of spectral densities at low frequencies. When Ω_0 is measured from average values of the low-frequency level (as in Fig. 2), a multiplication factor of two has to be applied (Herrmann and Kijko, 1983).

3.2. Determination of corner frequency, f_0

The correction for the effect of anelastic attenuation upon the observed Lg-wave corner frequency is difficult to make, since the attenua-

tion is not very well known for these waves in the range of observed frequencies (roughly 1–5 Hz). Recent studies indicate that the quality factor, Q , for Lg waves depends upon the frequency and that the dependency is exponential (e.g., Singh and Herrmann, 1983):

$$Q(f) = Q_0 f^\eta$$

where Q_0 is Q at a reference frequency of 1 Hz, and η is a numerical constant. For the eastern United States, Q_0 -values have been computed for several tectonically stable regions. Values of Q_0 vary between 460 and 900, and η between about 0.3 and 0.5 (Singh and Herrmann, 1983; Pulli, 1984; Gupta and McLaughlin, 1987). For eastern Canada, Q_0 varies from about 500 to 1100, and η from about 0.2 to 0.65 (Hasegawa, 1985; Chun et al., 1987; Shin and Herrmann, 1987). If we take $Q_0 = 1100$ and $\eta = 0.19$ (Chun et al., 1987), the 5 Hz spectral component would show an amplitude decrement of a factor of about four at a distance of 500 km, and a factor of about 20 at a distance of 1000 km. Hence, the effect of attenuation upon the high-frequency spectral component at large distances is significant. It is not quite clear whether the attenuation due to scattering is as important as the anelastic attenuation in this frequency band (see, e.g., Aki, 1980; Dainty, 1981).

In Fig. 3 the observed corner frequencies are plotted versus distance for two different magnitude ranges. The data indicate a decrease of the

Fig. 3. Observed corner frequency (f_R) plotted versus epicentral distance, for two groups of events selected on basis of magnitude (M_L).

measured corner frequencies (f_R) with increasing epicentral distance.

The events were then subdivided into four groups according to their seismic moment (Section 4.1). A linear regression was made between the corner frequency and distance for each group. Also, in these cases there is a decrease of corner frequency with increasing distance, and the decrease is smaller the larger the moment (Fig. 4). Hasegawa (1983) reports similar relationships for earthquakes in eastern Canada. The computed slopes were used to normalize the observed corner frequencies (f_R) to zero distance. The normalized corner frequencies are averaged for each event, and the average (f_0) is considered to be representative for the source-spectrum corner frequency. The source-spectrum corner frequencies for the events studied are listed in Table 1.

4. Results

4.1. Seismic moment

The seismic moment (M_0) for the events studied is determined from the low-frequency spectral level (Ω_0) by applying the relation derived empirically by Street et al. (1975) and verified theoretically by Herrmann and Kijko (1983):

$$M_0 = \begin{cases} 4\pi\rho\beta^3 R_0 (R/R_0) \Omega_0, & R < R_0 \\ 4\pi\rho\beta^3 R_0 (R/R_0)^{1/2} \Omega_0, & R \geq R_0 \end{cases}$$

where ρ is the density, β is the shear wave velocity, R is the epicentral distance, and R_0 is the reference distance. For the Baltic Shield earthquakes, we adopt the values $\rho = 2.7 \text{ g/cm}^3$, $\beta = 3.6 \text{ km/s}$ and $R_0 = 100 \text{ km}$. R_0 is used to account for different kinds of Lg-wave geometrical spreading:

Fig. 4. Observed corner frequency plotted versus epicentral distance for four groups of events: seismic moment (M_0 , dyn cm) less than (a) 5×10^{19} ; (b) $5 \times 10^{19} \leq M_0 \leq 5 \times 10^{20}$; (c) $5 \times 10^{20} < M_0 \leq 5 \times 10^{21}$; (d) $M_0 > 5 \times 10^{21}$.

At near distances ($R < R_0$), the spreading is typically of the body-wave type (spherical), while at greater distances ($R > R_0$), it is of the surface-wave type (cylindrical). The above formula is based on the relation of Keilis-Borok (1960) between the seismic moment and the far-field, low-frequency spectral level of S waves: however, the source radiation pattern asymmetry is not considered in our interpretation. The Lg-wave train is composed of a family of rays undergoing subsequent multiple reflections and conversions as well as scattering within the crustal waveguide (see, e.g., Campillo et al., 1984; Kim, 1987). This makes the radiation pattern effect minor in the station record.

The seismic moment for each event is obtained as the average from all station readings. Estimates of the seismic moment for the events studied are listed in Table 1. It may be noted that the computed seismic moment for the second largest event, on 25 October 1976 ($M_L = 4.9$), is in fair agreement with that of Slunga (1979), who obtained a value of $3\text{--}4 \times 10^{22}$ dyn cm using long-period surface waves.

4.2. Source radius and stress drop

Assuming the circular fault model of Brune (1970), the source radius (r) can be estimated from the source-spectrum corner frequency (f_0) through the relation:

$$r = 2.34\beta / [2\pi f_0]$$

Calculated source radii range from 0.2 to 1.0 km (Fig. 5). The source radii for events with seismic moment less than 5×10^{20} dyn cm increase only slightly with increasing moment, whereas the increase is larger for events with seismic moment above 5×10^{20} dyn cm. The source radii for the Meloy earthquake sequence in northern Norway reported by Bungum et al. (1982) also show very modest increase in the seismic moment range 10^{17} to 2×10^{19} dyn cm (see Fig. 5).

The stress drop ($\Delta\sigma$) is estimated from the relation (Keilis-Borok, 1959; Brune, 1970; Kanamori and Anderson, 1975):

$$\Delta\sigma = \frac{7}{16} M_0 / r^3$$

The estimates of the stress drop for most of the

Fig. 5. Seismic moment plotted versus source-spectrum corner frequency and source radius (Brune, 1970 model). Circles indicate results from this study and squares results obtained by Bungum et al. (1982) for the Meloy aftershock sequence in northern Norway. Lines of constant stress drops of 0.1, 1, 10 and 100 bar are indicated.

events fall between 0.1 and 10 bar. The stress drops are depicted in Fig. 5 using lines of constant stress drop, i.e. a slope of -3 for the seismic moment versus corner frequency relation.

The stress drops of the events studied are small compared with those of other regions and studies. They are one order smaller than those computed for earthquakes in eastern Canada with comparable moments (Hasegawa, 1983). The stress drops obtained by Bungum et al. (1982) from the Meloy sequence in northern Norway are 1–2 orders larger than ours (see Fig. 5). Small events in southern Sweden studied by Slunga et al. (1984, see their Fig. 6) in the moment range 10^{18} – 10^{20} dyn cm also show at least one order larger stress drops than our results. Although Slunga et al. (1984) use

Fig. 6. Seismic moment plotted versus magnitude (M_L).

a source model other than that of the present study, and Brune et al. (1986) suggest a possible existence of small low stress-drop events (partial stress-drop model of Brune, 1970) intermixed with high stress-drop events, it is difficult to explain our small values. Possibly the true corner frequencies, particularly for the smaller events, are close to or even larger than the upper limit of the reliable frequency range from the type of recording we use (5–10 Hz), so that our “observed” corner frequencies are too small. This implies an even larger decrease in stress drops (proportional to f_0^3). Note, however, that similar, low stress drops for earthquakes of corresponding size have been obtained in other studies, e.g., by Grosser et al. (1987) for the western Bohemia swarm in 1985–1986.

4.3. Seismic moment – corner frequency relation

The spectral-scaling relation between the seismic moment and the corner frequency has been derived for the moment range 3×10^{18} to 3×10^{22} dyn cm:

$$\begin{aligned} \log f_0 (\text{Hz}) \\ = 3.46(\pm 0.20) - 0.15(\pm 0.01) \log M_0 (\text{dyn cm}) \end{aligned}$$

However, by analogy to the source radius – moment relationship (Section 4.2), the corner frequency – moment dependency is different above and below about 5×10^{20} dyn cm (see Fig. 5), but the slope of the moment–frequency relationship is steeper than -3 also for the larger events. A slope steeper than -3 has also been reported for the central United States (Street et al., 1975), northeastern North America (Street and Turcotte, 1977), and eastern Canada (Hasegawa, 1983).

4.4. Seismic moment – magnitude relation

Figure 6 shows the seismic moment plotted versus magnitude (M_L) for the earthquakes in our study. The empirical relationship between the moment, determined from Lg-wave amplitude spectra, and the magnitude (M_L) reads (Fig. 6):

$$\begin{aligned} \log M_0 (\text{dyn cm}) \\ = 16.93(\pm 0.09) + 1.01(\pm 0.03) M_L, \\ 2 \leq M_L \leq 5.2 \end{aligned}$$

A similar relationship is reported for earthquakes in southern Sweden with magnitudes between 1 and 3.3 (Slunga et al., 1984). The relationship is also comparable to that found for eastern Canada (Hasegawa, 1983), whereas for the western United States an equivalent local magnitude corresponds to a seismic moment one order larger (Wyss and Brune, 1968; Hanks and Boore, 1984). A change in the moment–magnitude relation at a seismic moment between 10^{21} and 10^{22} dyn cm, suggested in several investigations (Street et al., 1975; Hasegawa, 1983; Nuttli, 1983), could not be observed with confidence in this study due to the small numbers of large events.

5. Discussion and conclusions

The source radii for Baltic Shield earthquakes calculated from the corner frequencies are only slightly decreasing with increasing seismic moment in the range 3×10^{18} to 5×10^{20} dyn cm (Fig. 5). Although there are some uncertainties in the measured corner frequencies, possibly due to the analog data used, this weak dependence im-

plies a strong increase of stress drop for increasing seismic moment (Fig. 5).

Nearly constant source radii over some moment range (usually for small events) is suggested in several recent studies employing high-quality digital data at short-distance ranges: e.g., Fletcher et al. (1986) for earthquakes near the Anza array, California; Archuleta (1986) for earthquakes with moment less than 10^{21} dyn cm in Mammoth Lakes, California; and Mueller and Cranswick (1985) (in their Fig. 7) for Miramichi, New Brunswick aftershocks.

The low stress drops obtained for small Baltic Shield earthquakes with seismic moment less than 10^{21} dyn cm may, at least partly, be caused by the analog data used.

The Baltic Shield earthquakes are characterized by a steadily increasing stress drop for increasing seismic moment, which is a consequence of the decrease of corner frequency with increasing seismic moment by a slope steeper than -3 (Fig. 5).

A constant stress-drop, source-scaling relation has been suggested by Kanamori and Anderson (1975) for large shallow earthquakes ($M_0 > 10^{25}$ dyn cm) and by Somerville et al. (1987) for earthquakes with $M_0 > 6 \times 10^{22}$ dyn cm in eastern North America and other intraplate regions. However, a slope steeper than -3 for the curve relating the corner frequency to the seismic moment implies that the similarity assumption proposed for the source-scaling law (Aki, 1967; Kanamori and Anderson, 1975) may not be valid for the small earthquakes considered in this study. Chouet et al. (1978) found such a departure from the similarity assumption for small earthquakes in various regions.

The linear relationship between seismic moment and magnitude is similar to those found for small-magnitude earthquakes in the southern Baltic Shield, and for earthquakes in the Canadian Shield.

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