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# Microseismicity in Local Areas: The Southeastern Part of the Fennoscandian Shield

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**Abstract**—This paper summarizes the observations of microseismic emissions by these authors in several areas of the Russian part of the Fennoscandian Shield to assess the potential of microseismicity to determine the present-day activity of local features in the upper part of the geologic medium. We give amplitude–frequency characteristics and the space–time distribution of naturally occurring microseismic events that are hypothesized to be of endogenous origin. We discuss the relationships these characteristics have to the regional geodynamic setting, average dimensions, and petrographic composition of active rock blocks.

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## INTRODUCTION

Local microseismicity is a direct piece of evidence that is useful when attempting to unravel the deformation and faulting processes in the upper crust. The authors have been studying the space–time distribution and amplitude–frequency characteristics (AFC) of microseismic events (ME) since 1994 in areas of the Russian part of the Fennoscandia in order to investigate the potential of microseismicity observations for assessing the present-day activity of faulted blocks and structural tectonic blocks from a few hundred meters to a few kilometers across. It is believed that the radiation of MEs of an endogenous tectonic origin under the conditions of crystalline Fennoscandian rocks is initiated by relative displacements, fractures, and rotations of rock blocks. The process is due to the elastic rebound mechanism as formulated by Harry Reid (1911) and developed later in models (Rice, 1979; Shebalin, 1984; Spivak, 1994, among others).

Observations of microseismicity are widely used in Russia and abroad to deal with many problems, such as safety issues in the mining industry, intensification of oil extraction, assessment of landslide stability, and the injection of liquid and gaseous waste into deep crustal horizons. Microseismic events are commonly classified within a rather wide magnitude range from events with moment magnitudes equal to –2, –3 (Chen et al., 2005) to $M$ ~ 4 events (Nemati et al., 2013). Microearthquakes, rock bursts, and collapse events in mines are all treated as MEs. The *microearthquake* concept appeared in the publications of Japanese seismologists as early as in the late 1940s (Asada and Suzuki, 1949). K. Kasahara (1981) defined the concept of an *ultramicroearthquake* in the 1960s, while Teng and Henyey (1981) were the first to describe events that these authors called *nanoearthquakes*. According to the classification of Lee and Stewart (1981), microearthquakes are those with $M$ between 1 and 3, while ultramicroearthquakes have $M \leq 1$. The most complete ME classification can be found in (Levin et al., 2010) with the following classes: small earthquakes, with $1 \leq M \leq 3$; microearthquakes, with $M$ between 0 and –4, and microcracks (or nanoearthquakes), with $M \leq -5$. The assessment of local activity shown by geodynamic processes using microseismicity can be of help for choosing sites for large-scale and long-term engineering facilities. These include nuclear power stations, burial locations of radioactive waste, and areas containing intersections of major fault zones with backbone gas and oil pipelines.

We studied the AFCs of local naturally occurring MEs in local areas within the Russian part of Fennoscandia (Fig. 1) for events that mostly fall in the classes of micro- and nano-earthquakes, according to the classification of Levin et al. (2010). Most of these areas were within the paleoseismic structures as identified by Lukashov (2004), with some of these being zones of major neotectonic faults that have clear expression in the present-day relief. As an example, the Paanayarvi area is situated in the 45-km long Paanayarvi–Kukasozero fault zone; this area has surface expression as a dense network of discontinuous disturbances that have displaced various geological bodies, mostly in the right-lateral sense of movement (Systra, 1991; Zykov, 2001). The Zaonezhskii area is situated in the Putkozero trough graben zone, which is over 60 km long and

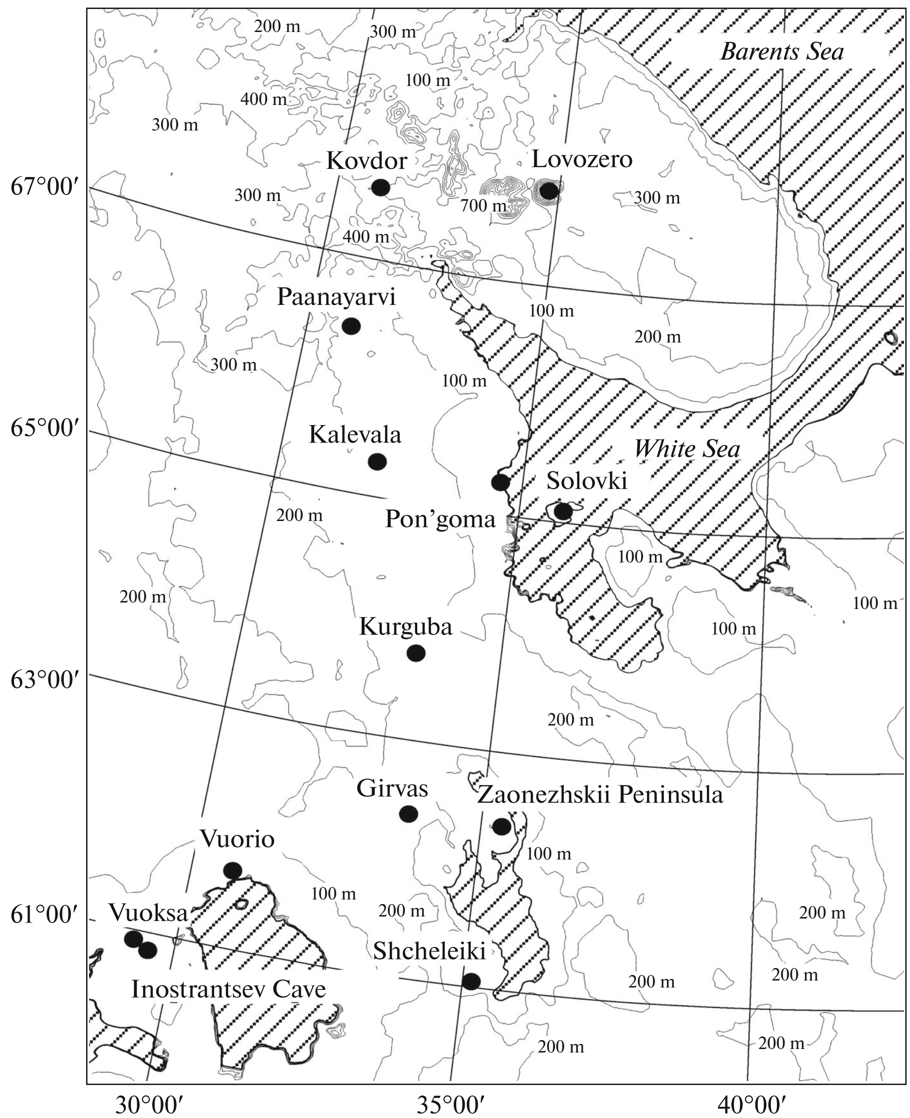


**Fig. 1.** A map that shows the areas where microseismic emissions were recorded.

has surface expression as steep scarps that reach heights of 30 m in the present-day relief (Fig. 2) that are separated by numerous fractures with a dominant normal sense of movement (Makarov and Shchukin, 2007). The Kalevala and Inostrantsev Cave areas were monitored by a seismic network that was deployed in a zone of intensive post-glacial seismic faults as described in (Lukashov, 2004; Systra, 2014; Nikonov, 2014, among others). Only the Pon’goma area is in a zone of comparatively low present-day activity.

The petrographic compositions of rocks in the observation areas are different. The Zaonezhskii area is composed of diabase and gabbro diabase. The other areas include basic volcanic rocks and komatiite basalts (the Kalevala area), quartz porphyries (the Paanayarvi); the Valaam area is dominated mostly by diabases. The Vuoksa and Inostrantsev Cave areas are composed of granite, the Vuorio, Girvas, and Kurguba areas are composed of sedimentary–volcanogenic schist, gneisses, basalts, and basaltic lava, the Pon’goma area is composed of migmatite, gneiss, and granite gneisses, while the Shcheleiki area is composed of dolerites.

Microseismic emissions were recorded by a mobile network of seismic stations deployed on the ground surface. Most stations were at considerable distances from sources of industrial noise, such as towns, highways, and railways. Our estimates of the spectral content of seismic noise for windless nighttime (Spungin, 2016) in many areas show that it is nearly identical with the noise spectrum at the NORESS regional small-aperture array, Norway (Bungum et al., 1985).

Along with endogenous MEs, we also studied signals of exogenous, technological, and anthropogenic

**Fig. 2.** A fragment of the Putkozero trough graben near the Zaonezhskii area.

origins that have AFCs similar to those for endogenous events, including seismic signals due to passing traffic, thunderstorms, impacts of falling objects, etc. The AFCs for such noise signals were studied in order to suppress them before detection of endogenous MEs. These results were partly published in (Spungin, 1997, 1999, 2007, 2011). The present paper describes results that were mostly obtained during the current century in areas of the southeastern Fennoscandian Shield.

## METHODS OF STUDY

Microseismic emission was recorded by a mobile local network of 3–4 geophones deployed at intervals of 100–300 m at the ground surface. The network included one or two three-component geophones and two geophones that recorded the vertical component alone. The geophones were SM-3KV seismic receivers with external amplifier units or SM-3KVE receivers with in-built amplifiers. The transfer function of a seismic channel for both geophone types was $\sim 4 \cdot 10^5$ V/m/s in the 0.5–40 Hz band. The geophones were installed on exposures of crystalline bedrock and were fastened to the rock by cement. Signals were transmitted from all geophones via connecting cables to the central station where the records of all channels were synchronized. Multichannel analog tape recorders were used in the 1990s for recording. Since 2001 the recording was carried out on a notebook with a 12-bit analog-to-digital converter, with a 14-bit instrument being used since 2006. The sampling rate was 200 Hz per channel.

The recording duration differed from area to area and gradually increased with improvements in the equipment. The observations in the Lovozero, Vuorio, Girvas, and Valaam areas were conducted before 2002; the duration of the recording did not exceed 3 days, with the recording being continuous during 15–30 min since the start of each even-numbered hour. The subsequent observations were conducted continuously, with the duration of the recording reaching 5–17 days. In two areas (Kalevala and Zaonezhskii) the observations were repeated in order to assess the stability of the AFCs and of the spatial distribution of MEs. A special journal was used during the recording to mark the times at which noise signals occurred due to technogenic, anthropogenic, and natural sources: passing traffic, axes hitting wood, thunder, lightning, etc. The total duration of continuous observation in all the areas shown in Fig. 1 was approximately 78 days.

The velocities of elastic waves were determined by the first exciting calibration signals at three to four sites in each area, 50–250 m from the geophones. The signals were excited by 5–10-kg fragments of crystalline rock that were dropped from a height of 2 m and hit bedrock.

Preliminary processing involved detection of MEs that were recorded by at least three geophones. Signal identification was visual, generally on original seismograms, more rarely using band filtering. If a signal had an AFC that was not similar to any of the known noise signals due to technogenic, anthropogenic or exogenous sources, its origin was treated as endogenous. The identified signals were saved as files and were used for location of ME epicenters and for computing the ME energy at the source.

The location was carried out for the MEs whose signals were distorted very little by noise; a method was used to minimize the time residuals of arrival times at the stations following a program developed by P.B. Kaazik at the Institute of Geosphere Dynamics for a homogeneous half-space. The first step was to specify (in interactive mode) the velocities of body waves (Vp or Vs), the position of the origin in the coordinate system used in this calculation (C), and the size of the spatial region (R); the program then determined the most probable position of the epicenter. Reliable determination of epicenter coordinates was possible for approximately 30–50% of the total number of identified MEs. A result was judged to be reliable when the direction of polarization in the ME signal at a three-component station was approximately the same as the direction to the resulting epicenter. Location uncertainties for epicenters due to insufficient incorporation of inhomogeneities in the geological section are estimated to be ±(10–20)% of the distance between the epicenter and the center of the network.

The hypocenter was located using the angle of emergence of seismic waves and the distance to the epicenter. Because the ME signals have low amplitudes, considerable errors can occur while trying to determine the angle of emergence. For this reason we specified the same value of the hypocentral depth for all MEs at the start of the calculation (150 m).

The hypocenter depth in the Zaonezhskii area was calculated twice for all MEs and separately for each of the three-component seismic sensors. The discrepancy between depth determinations based on measurements of different geophones did not exceed 90% of the average depth, with the average discrepancy being 33% of the average depth. The depth of ME hypocenters resulting from these determinations in the Zaonezhskii area was in the 10–150 m range, with the average depth for the area being 39 m (Spungin, 2011).

For the MEs that were thought to be located reliably, we estimated the energy released at the source (E, J) using the well-known relationship that connects it with the parameters of seismic waves at the recording site (Savarenskii and Kirnos, 1955):

$$E = \pi \rho v u_{\mathrm{m}}^2 R^2 n / f\alpha,$$

where $\rho$, kg/m$^3$ is the rock density; $v$, m/s is the average velocity of elastic waves in the earth; $u_{\mathrm{m}}$, m/c is the absolute value of the maximum ground-motion velocity vector; $R$, m is the distance between the ME hypocenter and the recording site; $n$ is the number of swings in the recorded wave train; $f$, Hz is the dominant frequency of motion in the train; $\alpha$ is the fraction of total energy released by the source and transmitted as seismic energy. Depending on the prevailing rock composition in an area of observation, the average earth characteristics were found in the following limits: the density was 2600–2800 kg/m$^3$; the velocities of compressional and shear waves were within 5000–5500 m/s and 2800–3200 m/s, respectively. The ratio $\alpha$ was assumed to be 0.0025.

The geological investigation included field documentation of local discontinuities and making maps of block divisibility of the areas using a morphotectonic analysis of topographic maps of scale 1 : 25000–1 : 200000. The field documentation of local structural discontinuities was compiled for the Vuorio, Paanayarvi, Kalevala, and Zaonezhskii areas. We made maps of block divisibility for these areas, as well as for Lovozero, Pon'goma, Inostrantsev Cave, and Vuoksa. We determined the present-day state of stress and identified zones with geodynamic settings of compression and tension for the Paanayarvi area. We recorded the air pressure for five areas and the temperature and wind velocity in the near-ground air layer for two areas to investigate the influence of these exogenous factors on the behavior and intensity of microseismic emissions. The method and results of these studies can be found in (Spungin, 2007, 2011, 2013).

## RESULTS AND A DISCUSSION

MEs of endogenous genesis were recorded in all of the observation areas. The typical record has the shape of a radio-impulsive wave train consisting of three to ten swings of different amplitudes and frequencies. The signals lasted from 0.2 to 1.5 s, occasionally reaching 3–5 s; the amplitude ranged between a few tens to a few hundred nm/s. The ME signals generally show a dipping angle of emergence (mostly within 30°–60°) and are clearly seen in the velocigrams of vertical and horizontal components (Fig. 3). The signals are obviously a mixture of body and surface seismic waves where individual types of waves are difficult to identify because of their superposition. Signals whose sources were 300–500 m from the recording sites (these were the majority of the signals) have arrival times for compressional (*P*) and shear (*S*) waves at recording sites that differ by a few hundredths of a second, which is much smaller than the periods of seismic waves due to MEs. The bulk of the energy of a recorded signal for most MEs was carried by shear waves, or by a surface wave if the source was shallow. The ME signals are dominated by shear or surface waves, as is also shown by epicenter location, both for naturally occurring MEs and for calibration signals created by humans. The most reliable results were attained when the seismic wave velocity was set as ~(2.8–3.2) km/s in the location program; this is the velocity of shear and surface waves for the Fennoscandian bedrock.

ME signals were only occasionally observed in which both compressional and shear waves could be identified (Fig. 4) based on an appreciable difference in their arrival times and in the direction of polarization at a three-component station relative to the station–source direction. The delay of the *S* arrival time behind *P* was 0.35 s in Fig. 4. If we assume the compressional velocity to be 5 km/s in our case and the velocity ratio to be between *P* and *S* to be 1.7, then the epicenter of the source of that signal was at a distance of ~2.5 km from the three-component station. Judging from the comparatively low amplitude of *P* on the vertical component record (compared with the horizontal components), the source of that ME was near to the ground surface. If it was of endogenous origin, it was most likely to have been caused by a strike-slip movement, since the *P* amplitude on the horizontal components was comparable with that of *S*.

The ME signals differ in their spectral content, dominant frequency, the envelope slope, and wave polarization, both in different areas and within a single area. Signals that involve a shear wave of mostly vertical polarization (SV) were encountered, whose radiation could be caused by normal or reverse block micromovements at the ME source. ME signals whose shear wave is dominated by horizontal polarization

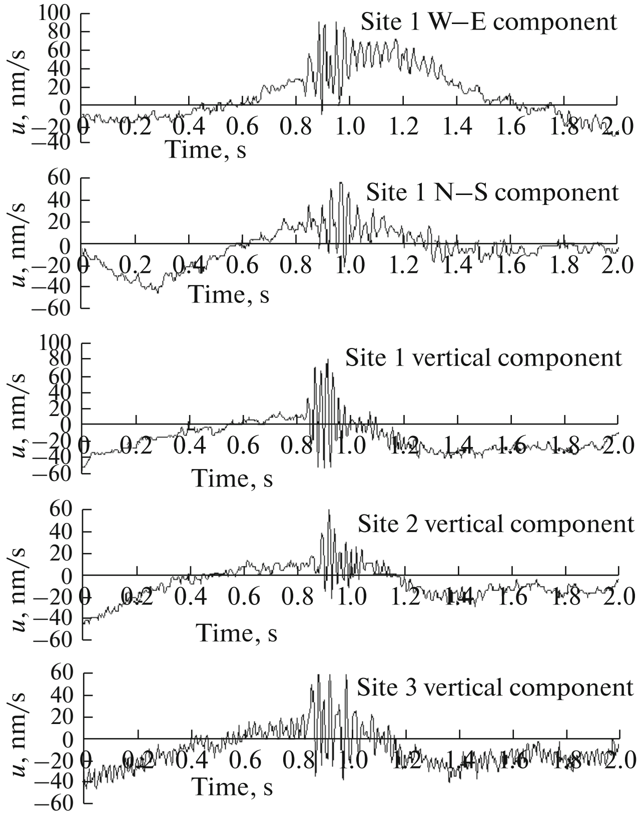


**Fig. 3.** Velocigrams of an endogenous ME recorded in the Pon'goma area at 04:27 GMT on August 13, 2011. The positions of seismic recording sites are shown in Fig. 6.

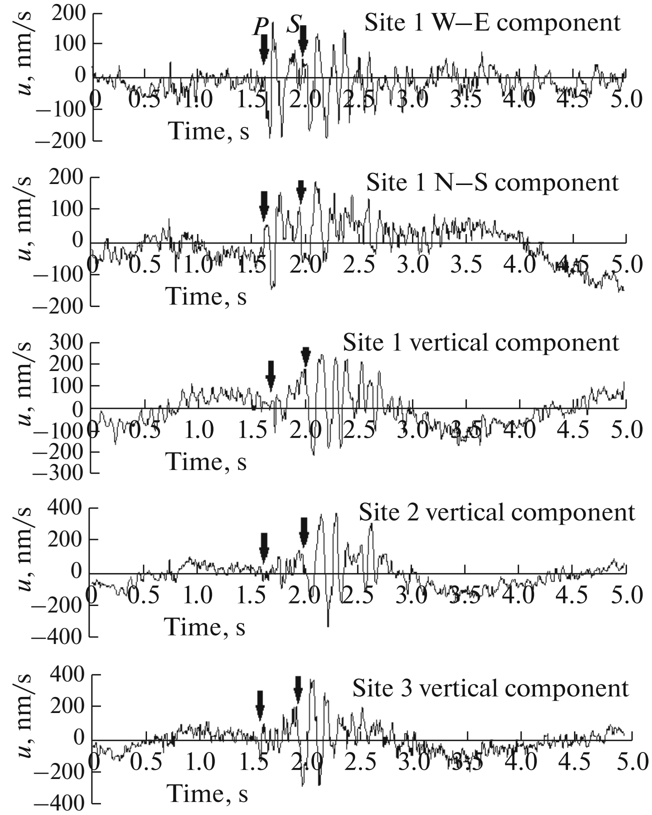


**Fig. 4.** Velocigrams of an endogenous ME recorded in the Vuoksa area at 07:36 GMT on August 18, 2013. The arrows show the onsets of compressional and shear waves.

(SH) were also observed, which were obviously caused by strike-slip block movements. However, most signals have the inclined polarization ellipse of mixed normal-oblique block displacements, with either of the slips being dominant.

The frequency spectrum of the MEs is diverse, covering the entire recording range. The dominant frequencies and shapes of ME signals differ substantially from area to area. Figure 5 shows variation curves of the dominant (modal) spectral frequency of ME signals for the six areas of our observation. Its is seen that this parameter varies in all areas within a wide band and mostly occupies nearly the entire operating range of the seismic channel. At the same time, the modal values of variation curves differ substantially for different areas. They are near 5–6 Hz for Vuoksa and the Inostrantsev Cave areas, near 15 Hz for the Pon'goma area, and near 30 Hz for the Kalevala area, while the dominant frequencies of ME signals for the Paanayarvi area are characterized by a bimodal distribution with modes at 5 and 12 Hz. Since the frequency of a signal is largely controlled by the size of the rock block that initiated an ME, this indicates that the sizes of the active blocks that emit larger numbers of MEs appreciably differ in the studied areas.

A relationship of the active block size to petrographic rock composition and to the present-day geodynamic setting is seen in the observed areas. The active blocks are the largest in the Vuoksa and Inostrantsev Cave areas, and to some extent in the Paanayarvi area. According to the dependence of the characteristic frequency of impulsive motion on the size of the relevant active block (Kocharyan and Koby-

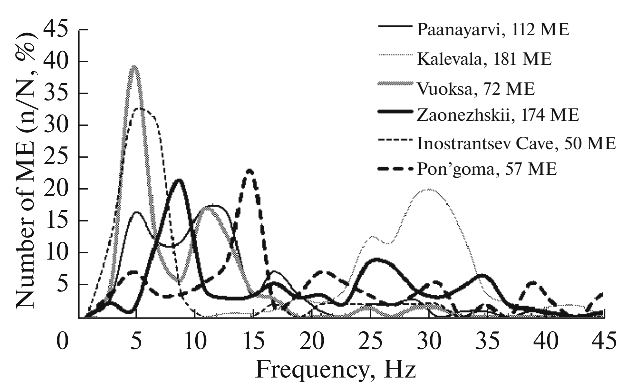


**Fig. 5.** Variation curves of the dominant frequency in MEs recorded in different areas of observation.

chenko, 2003), they are approximately 10 m across. All these areas are composed of acid rocks (the two first consist of granite, while the Paanayarvi area consists of quartz porphyries) and are situated near major regional zones of recent strike-slip movement. The Vuoksa and Inostrantsev Cave areas are in the zone affected by the dynamic influence of the Vuoksa fault zone, which consists of several branches that strike northwest and extend for a total of approximately 70 km, while the Paanayarvi area is situated in the zone of the Paanayarvi–Kukasozero fault that extends for 45 km. It is known that rocks contain more cracks and are fragmented to a greater degree in the zones affected by the influence of major faults. It is probable that because the rocks in such zones are less dense, the faces of small blocks concentrate stresses that would be sufficient to initiate MEs less frequently; it is for this reason that the rate of high-frequency MEs is relatively low in such locations.

The smallest sizes of active blocks, approximately 1 m, occur in the Kalevala area, which is composed of basic rocks (volcanic rocks and komatiite basalt) and is in a zone of abnormally high concentration of horizontal compressive stresses that squeezed blocks of crystalline rocks upward during Holocene time (Systra, 2014). Under these conditions, the relatively large blocks are more compressed; they have limited possibilities of movement and the relaxation of present-day stresses occurs during movements of mostly smaller blocks.

The Zaonezhskii area also contains many active blocks with relatively small sizes. This area is also composed of basic rocks, that is, diabases and gabbro diabases. At the same time, this area, as well as the Vuoksa, Inostrantsev Cave, and Paanayarvi areas, is situated in the zone of a regional fault (the Putkozero trough graben); however, this fault is not of the strike-slip type like the former ones, but involves a large normal component (see Fig. 2). It is possible that the density deficit in zones affected by the dynamic influence of normal faults is different compared with the case of strike-slip faults, so that these features control both the sizes of active blocks and the character of relaxation of present-day stresses in a blocky earth.

The total energy of most MEs recorded in this study varied between a few tenths and a few hundred Joules. It was very rare that $M_L \approx 1$ was recorded. The average energy of MEs and the range of this variation were appreciably different in different areas. The maximum values (up to a few hundred Joules) were recorded in the Paanayarvi and Inostrantsev cave areas. The Zaonezhskii, Lovozero, and Girvas areas did not generate ME energies above 100 J; the values in the Kalevala, Vuorio, and Pon'goma areas barely reached 50 J, while they were below 10 J in the Valaam area. The energies of approximately 80% of all MEs that were recorded in the Kalevala, Vuorio, Girvas, Zaonezhskii, and Pon'goma areas did not exceed 20 J.

It is known that the seismicity of any area is characterized by dispersed (over the area) and concentrated (localized in individual zones) components (Shebalin et al., 1991). The former type of seismicity is an indicator of background geodynamic activity in an area of the geologic medium or a structural tectonic block, while the latter reflects the activity of discontinuities (faults and cracks of various hierarchical orders). According to the self-similarity of the seismic process (Sadovskii and Pisarenko, 1991), this also applies to microseismicity. The areas of the southeastern Fennoscandia studied here showed persistent concentration of ME epicenters in zones of local discontinuities. ME epicenters mostly mark zones or individual parts of small fault zones and larger crack zones ranging between some tens of meters and a few kilometers in length; these have hierarchical orders of VI–V according to the classification of (*SNiP ...,* 1988). Most of them can also be identified from morphotectonic features in the present-day relief (block boundaries and secondary (in importance) intrablock fractures) (Spungin, 1997, 1999, 2007, 2011).

The area where the ME epicenters were recorded by our network of three to four geophones varies within ~0.4–4.0 km$^2$ in different areas and is approximately 1 km$^2$ on average, which is obviously determined by the block structure of specific areas and by the intensity of regional geodynamic processes. The largest area of the ME epicenters was observed in the Paanayarvi and Lovozero areas, with the smallest areas occurring in the Zaonezhskii and Pon'goma areas. The concentration of the epicenters of the smallest MEs with energies below 10 J occurs in zones of block boundaries or intrablock fractures near the observation network, at distances of 100–150 m from the observing stations. The zones of discontinuities that are 200–300 m from the center of the network are marked by larger MEs with energies of 10–100 J (Figs. 6 and 7). The degree of concentration of ME epicenters per unit fault length is different. We did not detect any correlation between the expressiveness of fault zones based on morphotectonic features and the degree of concentration of ME epicenters within them. As an example, the ME epicenters in the Zaonezhskii and Vuorio areas marked mostly northeast striking fault zones that have comparatively low surface expression, while the concentration of ME epicenters was lower by a factor of a few times within northwest striking fault zones involving tectonic scarps as high as 30–60 m (Spungin, 2011).

It should be noted that not all discontinuous disturbances identified via geomorphology are marked by ME epicenters, even the largest ones. The same thing can be said about the discontinuous disturbances that are situated in the vicinity of the local seismic network. Only some of these faults generated MEs, or some areas within these faults. As an example, not a single ME was recorded in the zone of a secondary discon-

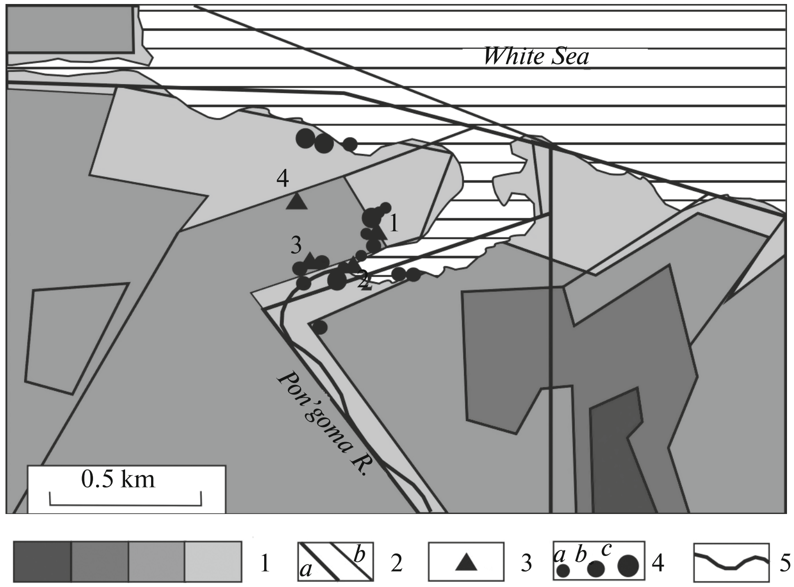


**Fig. 6.** A map of the block divisibility for the Pon'goma area and the spatial distribution of ME epicenters recorded in August 5–19, 2011. (1) height gradation of the upper surfaces of morphostructurally expressed blocks with differentiation along the height at intervals of ~5 m; (2) faults that separate the blocks; the faults are expressed as large relief forms (*a*), and interblock faults expressed as small morpho-sculptural forms in the relief (*b*); (3) the site of seismic observation and its identification number; (4) ME epicenters with the following energies at the source: <1 J (*a*), 1–10 J (*b*), and >10 J (*c*); (5) the bed of the Pon'goma R.

tinuous disturbance that strikes northeast–east near station 4 in the Pon'goma area. At the same time, three MEs were recorded along another fault, the northwest–west fault that is similar as to hierarchy and which was more than 100 m from the same station 4 north of the observing network (see Fig. 6).

In addition, the propagation of ME signals initiated by local fault zones can be shielded by zones of the same or lower hierarchical level, if that zone is between the ME source and the network. As an example, no ME was recorded in the Pon'goma area south of the northeast-striking block boundary that passes south of the observing network along the Pon'goma R. valley (see Fig. 6). Turning to the Paanayarvi area (see Fig. 7), we see that a single ME only was recorded there with the epicenter southeast of the seismic network beyond the axis of the regional Paanayarvi–Kukasozero fault.

We did not record a high concentration of ME epicenters along the strikes of major regional faults that are a few tens of kilometers long (of order III according to the classification of (*SNiP ...,* 1988)) in any of the studied areas. As an example, microseismic emissions were only observed at feather faults that strike northeast around the north-west trending Putkozero trough graben, which is well expressed in the present-day relief (see Fig. 2) and passes in the Zaonezhskii area near the seismic stations (Spungin, 2011). A similar picture was observed earlier in the Vuorio and Paanayarvi areas, as well as in the South Alps, where similar surveys were conducted during a longer time (Adushkin, 1993; Spungin, 1999). This seems to be due to certain features in the present-day geodynamics of these zones. V.P. Solonenko (1986) noted that "The existence of major faults is not sufficient to invest them with a high seismic potential... Potential seismicity depends on the activity of neotectonic regional structures that accumulate stresses."

The observed spatial distribution of ME epicenters in the Paanayarvi area (see Fig. 7), which is situated in the zone of the active regional Paanayarvi–Kukasozero fault, allows us to relate the distribution to the conditions of the present-day state of stress, in particular, to the geodynamic settings of compression and tension. A total of 112 MEs were recorded in the Paanayarvi area between July 7 and 11, 2004. The distribution of their dominant frequencies was polymodal, with the best-expressed modes occurring at 5 and 11 Hz (see Fig. 5). Characteristically, events of different frequency contents made up two sets that were isolated spatially and included approximately equal numbers of MEs. The MEs of relatively lower frequencies were concentrated in the set that was linearly elongate northeast–east at an azimuth of ~70°. The second events made up an ellipsoidal cloud whose longer axis strikes north–northeast at an azimuth of ~25° (see Fig. 7). Both of these sets formed an implicitly en echelon row that seems to reflect the present-day geodynamics of this area, which can be explained by invoking bulk displacement of a blocky medium.

The geodynamic type of the present-day state of stress in the Paanayarvi–Kukasozero fault zone consists in right lateral strike-slip movements. When such movements occur in a rock mass they make individual


**Fig. 7.** A map of the block divisibility of the Paanayarvi area and the spatial distribution of ME epicenters recorded in July 4–11, 2004. (1) height gradation of the upper surfaces of morphostructurally visible blocks with differentiation along height at intervals of ~10 m; (2) zones of local tension—separations that can act as asperities during relative block movements; (3) faults that separate the blocks; the faults are expressed as large relief forms; (4) intrablock faults that are expressed as small morpho-sculptural forms; (5) ME epicenters of energy over 100 J: with dominant signal frequency 2–9 Hz (*a*) and 9–40 Hz (*b*); (6) ME epicenters with source energy below 100 J: with dominant signal frequency 2–9 Kz (*a*) and 9–40 Hz (*b*); (7) site of seismic observation and its identification number.

blocks rotate and crack into separate parts during their mutual interaction. The process involves the formation of asperities at block boundaries, which may be both under compression and under tension during rotation, depending on the evolving situation at contacts with the adjacent blocks. In the map shown in Fig. 7 such asperities can be seen as separations at block edges. Observations of large earthquakes show that such earthquakes differ in the frequency range in which the maximum energy is radiated. The earthquakes that are confined to divergent tectonic zones generally have longer periods compared with those in convergent zones. It was also found that reverse-slip earthquakes generally radiate energy at higher frequencies compared with the strike-slip events (Lyskova, 1999). Proceeding by analogy, it can be concluded that the patch of concentrated MEs of low frequency in the Paanayarvi area reflects an area of activated present-day tension, while the concentration of high-frequency events reflects a tension area. The overall structure of this fault system can be interpreted as a series of evolving separation cracks, judging by its diagonal position relative to the main fault strike.

Figure 8 shows a variation curve for the azimuths of discontinuous disturbances that are marked by ME epicenters based on the observations in all the areas under study (see Fig. 1). Each value at the curve corresponds to a single ME that was recorded either near a block boundary zone or near an intrablock boundary of crack origin that strikes in the relevant direction. In those cases in which an ME was recorded in the junction zone or in a zone of intersection between two boundaries with different azimuths, that ME was plotted twice in the diagram, for both directions. The azimuths of faults and block boundaries that are marked by ME epicenters coincide with the main directions of fault tectonics as found in Archean and Proterozoic rocks. However, while the latter are dominated by northwest striking features (various areas of the Karelian craton contain faults and dikes that strike

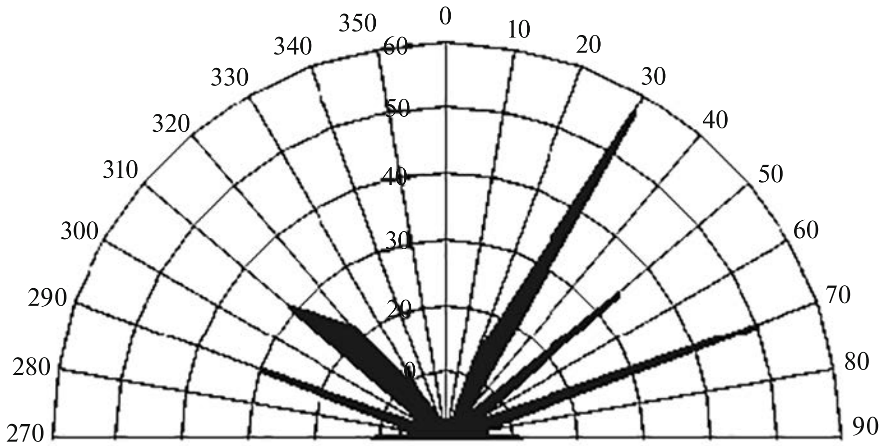


**Fig. 8.** A diagram of the strikes of discontinuities, as recorded by ME epicenters in the areas of microseismic observation.

280°–305°; 315°–330°; and 325°–335° (Systra, 1991)), the ME epicenters mostly mark discontinuous disturbances that strike northeast and northeast–east, which are orthogonal to them. The number of MEs that were recorded in the fault zones that strike 25°–35° and 65°–75° is approximately two time larger than that of the MEs that mark the discontinuous disturbances that strike 285°–295° and 305°–325°. Viewed kinematically, these faults correspond with strike-slip separations that form under the principal compressive axes that strike west—northwest; this is consistent with the idea of Quaternary and earlier stress macrofields in the Fennoscandian Shield controlled by North Atlantic spreading (Yudakhin et al., 2003).

## CONCLUSIONS

Studies of naturally occurring microseismicity were carried out for the first time in the Russian part of Fennoscandia in areas far from sources of strong technogenic and anthropogenic noise. We studied ME signals of endogenous tectonic origin in areas with different rock compositions, structural tectonic structures, and regional geodynamics. We assessed the potential of a mobile seismic network for recording MEs under the conditions of Fennoscandia. In fact, this study is a pioneering survey. During this work we revised our goals and improved the instrumentation and observational techniques.

It was found that under the conditions that prevail in Fennoscandia, where crystalline bedrocks are exposed at the ground surface, a local network consisting of three to four geophones with 100–300 m between the geophones can record very-low-amplitude MEs of endogenous origin, whose energies at the source are a few tenths of a Joule ($M$ from ≈ –4). MEs with energies below 10 J can be recorded at distances of up to ~100 m from the observing stations; those with energies of 10–100 J can be recorded at distances of below ~200 m, while events with energies of 100–1000 J can be recorded at distances of approximately 1 km. The AFC, the average energy of the MEs, and the intensity of microseismic emission are not the same in different areas and depend on the size of the active block and on the regional geodynamic setting. Most ME signals consist of a mixture of body and surface waves. The number of events for which one can detect the onset times of compressional and shear waves is very limited. Events of $M_L \approx 1$ were recorded in rare cases.

The mode of radiation that produces MEs is similar to ordinary seismicity; both periods of high activity and quiescent periods occur. The quiescences did not last longer than 6–10 h in most areas, with the maximum duration (56 h) being observed in the Pon'goma area, which is in a region that involves rather low occurrences of present-day geodynamics.

One notes that the dominant signal frequency and ME energy depend on the composition of the rocks that compose the area of generation, which appear to control the average size of active blocks. MEs of higher energy are observed in areas that are composed of acid rocks, such as granite and quartz porphyries. The energy of most MEs that are emitted from areas composed of basic rocks (basalts and diabases) is approximately an order of magnitude lower. The areas composed of sedimentary–volcanogenic schist and gneisses occupy an intermediate position. Since the sizes of those blocks of the highest activity are controlled by the properties of the constituent rocks and by the geodynamic setting in the area of observation, the dominant ME signal frequency characterizes the geological structure of the area.

In all areas, the ME epicenters mark zones of discontinuous disturbances or parts of them where the most intensive rock deformation is obviously occurring during the present-day (current) phase of geological evolution. The concentration of ME epicenters in the most active zones is an order of magnitude greater than within the blocks that are separated by these zones. Considering that the energy of most MEs is low, the propagation of their signals can be hampered by comparatively small local faults, from order IV downward, after the classification in (*SNiP ...,* 1988). The area of ME occurrence that was recorded by our network within an area of observation is not the same everywhere and varies within the 0.4–4.0 km$^2$ range, which obviously depends on the block structure of the area and on the regional geodynamic setting.

It was found that microseismicity can be used to assess the geodynamic setting within individual local zones. The MEs that occur in zones of compression and tension can differ in their dominant signal frequency. It is likely that normal and strike-slip movements can be identified using the polarization of ME signals.

The practical use of microseismicity has prospects during the reconnaissance phase of engineering geological surveying for a rough assessment of the present-day (current) activity in local zones of discontinuous disturbances. In contrast to the geodetic and strainmeter techniques, which are commonly used along lines of observation or at observation sites, observations of microseismicity allow one to assess the activity of local features in an area, as well as to roughly identify individual zones of discontinuous disturbances that are more or less active during the current phase of geological evolution that should be studied later using geodetic and strainmeter techniques. Under favorable conditions, microseismicity allows one to assess the geodynamic settings within individual local faults and to identify zones of compression, tension, and shear.

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