A Magma Body at a Volcanic Front as Seismic Emission Source, Honshu, Japan

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Previous seismic research revealed a well-defined S-wave reflector in the mid-crust near Shirane Volcano, Japan; it was interpreted as a thin body of partially molten magma [9]. We studied this unusual feature using a modified method of emission tomography [10] based on three-component records of microseisms and the codas from local earthquakes to construct images of endogenic sources of seismic emission. An analysis showed that this crustal heterogeneity radiated weak emission signals whose intensity and composition varied over time. The time behavior of the emission may be controlled not only by processes that are going on inside the magma body and near its surface but also by deeper processes: the presence of deep-seated magma conduits beneath the reflector is indicated by the fact that this and similar bodies are confined to volumes generating deep low-frequency earthquakes and low-velocity zones which extend from the upper crust into the mantle as deep as 100-150 km.

INTRODUCTION

The Japan Islands are a typical subduction zone. The subduction rate is rather high there (about 10 cm/yr), resulting in high seismic activity in the region; earthquakes occur as far as 1400 km inland from the trench. Numerous shallow earthquakes occur in the upper crust due to large horizontal stresses caused by relative plate motion: 95% of all shallow events do not occur deeper than 15 km, this depth being interpreted as a brittle-ductile boundary. It is known [15] that partial melting occurs at 100-200-km depth and generates magma that rises to the surface and is erupted to form long chains of volcanoes above Benioff zones. The volcanic front in the northeastern Japan Arc passes through the middle of Honshu Island from north to south parallel to the Japan Trench some 250-300 km from it. The tomographic studies of the region [7] revealed that the volcanic front involved
zones of lower $P$-wave velocity extending from the upper crust to a 100–150-km depth into the mantle. The low velocity volumes dip west approximately parallel to the downgoing oceanic plate and are interpreted as regions heated by the mantle processes that produce the deep-seated sources of the volcanoes.

Seismic observations in volcanic areas have revealed high-contrast heterogeneities in the mid-crust observed as strong $S$-reflectors: a reflected $S\times S$ phase can be seen on records of local earthquakes occurring near and above such reflectors, the amplitude of this phase occasionally being larger than that of a direct $S$ wave; a clear $S\times P$ phase sometimes appears on records of some stations [6]. Similar reflectors have so far been identified at five localities in the northeastern Japan Arc; all of them are situated near active volcanoes and confined to deep-seated low-velocity zones [5]. Detailed seismological investigations revealed that these heterogeneities were thin extensive regions filled with partially molten magma. We studied a reflector in the southernmost part of the northeastern Japan Arc not far from Nikko-Shirane Volcano [9]. Our analysis of the distribution of $S\times S$ travel times over the area showed that the velocity-contrastind crustal heterogeneity had a very uneven top boundary and manifested itself as a set of reflectors scattered over a $15 \times 15$ km area at depths of 8 to 15 km. Peaks in the spectral ratio of direct to reflected $S$ waves were used to estimate the thickness of the reflecting body. This body was found to consist of at least two layers having a total thickness of about 100 m and dipping 30–45° away from the volcano’s summit [9].

This paper presents the first results of our study. We used a method of emission tomography based on the analysis of three-component records of the coda from local earthquakes and microseisms which contain seismic emission noise radiated by the crust. The data were recorded by a network of seismic stations deployed in the area of Shirane Volcano on the side of the crustal reflector. The analysis was done by visualizing the endogenic sources concerned using a phased seismic array which revealed a weak emission signal related to an existing geologic anomaly [1], [2], [3], [10].

METHOD OF STUDY

In view of the fact that earthquake coda and microseisms are generated by multiple sources, the resulting wave field has a pattern of a stochastic process. When there are no pronounced individual sources, the distribution of instantaneous amplitudes is close to normal [11], [13], so that a suitable sensor spacing makes noise almost uncorrelatable at different stations. And when even a weak spatially coherent signal appears, the responsible source can be detected and located, and its intensity determined, by using records of a directional seismic array [1], [10]. The observed waves are recorded at the earth’s surface by a seismic array, and the records can be analyzed by statistical identification of signals arriving from different regions in the earth. This is done by summing the records of all
array sensors using time delays corresponding to the time-distance curve of an assumed signal from the region scanned. The summed seismogram is then used to calculate a power spectral estimate that quantitatively characterizes the radiation intensity coming from the volume around the point being aimed at. The exceedance of a detection threshold (the confidence interval for pure noise) indicates the presence of a seismic source. In this way, scanning over points of a 3-D grid gives a 3-D image, in which active sources and scatterers of transmitted seismic waves are seen as bright spots.

In the case under consideration, the object of study was far enough from the recording array: the distance from the heterogeneity to the center of the array was equal to a few times the array size. With this distance we could not expect a satisfactory spatial resolution along the source-receiver line. For this reason we did not determine the exact coordinates of the source, but merely found its azimuth based on a spatially coherent component and plotted the wave field in the angle of emergence – source azimuth coordinates, that is, the adjustment was based on the time-distance curve for a plane wave over a grid of azimuths and angles of emergence. Essentially, this approach is similar to the method of controlled directional reception used in seismic prospecting [2], [3]. Our records had three components, and this made it possible to adjust the array not only by time-distance curve but also by signal polarization. A seismic source usually radiates both $P$ and $S$ waves. The $P$ wave and both components of $S$ ($SV$ and $SH$) corresponding to one ray have different polarizations in the far zone, the displacements at a recording site being parallel to three different orthogonal coordinate axes. In our analysis we used the projection of the instantaneous total three-dimensional amplitude onto the direction of the relevant axis and produced images of the $P$, $SV$, and $SH$ wave patterns. The distribution of intensity in the image was numerically reconstructed using the following similarity measure known in seismology [14]:

$$S = \frac{\sum_{j=1}^{T} \left( \sum_{i=1}^{k} f_{ij}(\tau_i) \right)^2}{\sum_{j=1}^{T} \sum_{i=1}^{k} f_{ij}^2(\tau_i)},$$

where $f_{ij}$ is three-component instantaneous amplitude projected onto the relevant direction; $i$ is sensor number; $j$ is the time sample number after applying a time delay $\tau_i$; $k$ is the number of sensors; $T$ is the number of independent time samples. Each value of $S$ in a grid square of the image corresponds to adjustment at a certain signal arrival direction; i.e., sensor delays are assigned in accordance with the time-distance curve of a plane wave, whose azimuth and angle of emergence are determined using a uniform sampling grid.

When an identical signal is present at all sensors, it is easy to show that the mean value of the similarity measure is [1]
\[ \langle S \rangle = k \frac{E_s + E_n/k}{E_s + E_n} , \]

where \( E_s \) and \( E_n \) denote a signal and a noise energy, respectively. This formula explains the physical meaning of a coherent estimate, which is proportional, for a large number of sensors, to the ratio of the signal energy to the added signal and noise energy, being actually an estimate of the energy contribution of the coherent signal into the total wave field. In the absence of a coherent signal the similarity measure has an asymptotically normal distribution with a mean and a variance [1], being respectively,

\[ \langle S \rangle = 1 \]

and

\[ \sigma^2 = (2/T)(1 - 1/k). \]

As the signal/noise ratio grows larger, the similarity measure increases but does not exceed the number of sensors. Visual images of the wave field were reconstructed by plotting the \( S \)-1 value distribution.

The assumed velocities of body waves were \( V_p = 5.20 \text{ km/s} \) and \( V_s = 3.06 \text{ km/s} \) [12]. The azimuth was varied at intervals of \( 7.5^\circ \) between 0 and \( 360^\circ \), the respective values for angles of emergence being \( 3.75^\circ \) between 0 and \( 90^\circ \).

**DATA BASE**

We used the results of seismological observations conducted in the Nikko-Ashio area in northern Kanto. A closely-spaced seismic array was operated there during 1.5 months from October 15, 1993, to November 31, 1993. It was adjusted to record small local earthquakes of \( K \geq 3 \) (\( K \) is earthquake energy class). The NIKKO array (Fig. 1) had a size of 7 by 7 km and consisted of 195 three-component stations which were mostly installed on large rocks or concrete cubes and in part buried into the ground. We used sensors of three types: one with a lower cutoff frequency of 2 Hz and the two others with 4.5 Hz. During data processing we used records of the 4.5 Hz stations which had identical instrument response curves. The array was deployed in mountains at heights of 650–1300 m above sea level to ensure a low (for Japan) man-induced noise. The average spacing between the stations was 70 m (200 m at the most).

It should be noted that our observation geometry had serious drawbacks as to the application of emission tomography: the sensor locations were not uniform and depended on roads for transporting the instruments through the mountains; the rough topography and velocity variations that could not be accounted for strongly distorted the waveforms, especially at higher frequencies; in addition, we had very short samples of microseismic noise at our disposal. However, in spite of these serious drawbacks, we succeeded to get
Figure 1 Sample records (a) of a local earthquake (Nov 9, 1993, 17 h 42 min) and array geometry (b); A to G - subarray names; A to G with numerals are station names. Type of record: UD vertical, NS north-south horizontal, EW east-west horizontal; 1 - location of earthquakes used in this study; horizontal brackets mark S×S reflections.
Table 1 Earthquake information.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time, h-min</th>
<th>Depth, km</th>
<th>Long. °E</th>
<th>Lat. °N</th>
<th>Number of recording stations</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>before sorting</td>
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<td></td>
<td></td>
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<td>after sorting</td>
</tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>coda</td>
</tr>
<tr>
<td>11.09.1993</td>
<td>17-42</td>
<td>8.99</td>
<td>139.540</td>
<td>36.625</td>
<td>147</td>
</tr>
<tr>
<td>11.11.1993</td>
<td>03-19</td>
<td>7.26</td>
<td>139.452</td>
<td>36.578</td>
<td>165</td>
</tr>
<tr>
<td>11.17.1993</td>
<td>00-39</td>
<td>7.50</td>
<td>139.478</td>
<td>36.640</td>
<td>153</td>
</tr>
<tr>
<td>11.18.1993</td>
<td>14-46</td>
<td>7.69</td>
<td>139.482</td>
<td>36.632</td>
<td>169</td>
</tr>
<tr>
<td></td>
<td>02-10</td>
<td>7.65</td>
<td>139.407</td>
<td>36.591</td>
<td>174</td>
</tr>
</tbody>
</table>

a well-focused image of the source. This was facilitated by a few positive factors. The array had a large number of instruments — 195 three-component stations; the raw records were of high quality. For various reasons the real number of stations that recorded an event was smaller, but this (about 100) was sufficient to provide an enhanced signal/noise ratio. The terrain where the array was deployed consisted of bedrock without a sedimentary layer to speak of. Apparently, the velocity heterogeneities were not very sharp, and the topographic effect was not drastic enough to damage the spatial correlation and polarization of seismic waves. The wave field image was well-focused also because the object of study was prominent because its microseism intensity was much higher than the relatively quiet background; the rough topography and inaccessibility of the area made for a low level of man-induced noise.

Generally our records had a fixed length of 20 s with a sampling rate of 200 Hz, but we used data with every second reading deleted, the sampling interval thus being \( D_t = 0.01 \) s. We processed records of six local earthquakes of magnitude \( M = 2.5 - 3 \) (see Table 1). They occurred at about the same depth (7.3-9 km) around the array (Fig. 1). A seismogram from each earthquake consisted of three segments: about 4 s of microseismic noise before the earthquake, first \( P \) and \( S \) arrivals, and the coda. The \( P \) to \( S \) time difference is about 1.5 s, corresponding to an epicentral distance of 10-12 km. For each earthquake we analyzed the pre-event microseisms in a window of 3.6 s and the coda in the time interval 4-16 s after the first \( P \) arrivals. The time window for microseisms was chosen to be as long as possible, but before the first arrivals; this was strictly checked. We did not excessively diminish the coda time window, because the \( S \times S \) phase had a large enough amplitude, even though the reflector was far from the array; this ensured a good signal/noise ratio, although the time window was significantly longer than the reflected phase. Figure 1 shows sample records for one of the earthquake listed in Table
1, the farthest from the reflector; the $S \times S$ reflection can be seen at times of about 7 s after the first arrivals.

The first processing step consisted in inspecting all records to reject those with strong noise, abnormal instrumental time shifts, and strong distortions in the first arrivals. Preliminary analysis of the data showed that the rms amplitude, both for microseisms and the coda, might differ by as much as a factor of more than 40 between the stations. After the first arrivals the mean value was significantly displaced for 50% of the instruments. A mean-level displacement in an electrodynamic instrument can be caused by various factors: "percolation" of a large-amplitude low-frequency seismic signal, nonlinear effects in the coil or magnet support system, or dry friction effects which may occur in some recorders. Therefore, the effect of site conditions was eliminated by converting the records of each instrument to a zero mean and dividing by the rms three-component amplitude for a given station, separately for the coda and for microseisms.

RESULTS AND DISCUSSION

Figure 2 shows a map [9] of the study area indicating the location of the array and the locations of the reflectors that make up the surface of the crustal $S$-reflector near the volcano. The reflector has an uneven surface dipping southeast from the center of the volcano and consists of two parts: a smaller upper and a deeper and horizontally extensive lower part. The reflector boundaries are seen in the azimuth range 260–340° looking from the array center. The azimuths for the upper part are 305–325°.

Figure 3 shows images for the codas of three earthquakes listed in Table 1. The fact that the 250–350° azimuth region is the "brightest", for all coda samples indicate that the bulk of the reflected energy arrived from these directions whatever the epicenter locations. Taking the event closest to the reflector (Fig. 3), one clearly discerns a reflected signal in the form of a $P$ wave arriving from the same direction as $S$. This seems to be an $S \times P$ phase, which can occasionally be seen by the eye along with an $S \times S$ phase [6]. The azimuth range 300–350° and the angles of emergence 20–50° stand out as the brightest for the coda samples.

Figure 4 shows images of the wave field for microseismic noise samples before two earthquakes. A coherent signal can clearly be seen in Fig. 4A arriving from about the same direction as the largest reflected phase: 290–340° for the azimuth and 20–60° for the angle of emergence. Whereas differences in the coda images can be explained by different directions of seismic "illumination", the cause of this phenomenon for microseisms lies in the effect of multiple, in particular, surface noise sources, the intensity and locations of which may vary strongly in time. For instance, a sample in Fig. 4B shows that a coherent signal is noticeable in $P$ and $SV$, but is lost in $SH$ because of the dominant contribution of surface sources.
Figure 2  Location map of the array area [9] near Shirane Volcano; 1 – array location; 2, 3 – locations of reflections in the crustal S-reflector; 4 – seismic stations.

Figure 5 shows averaged images of the microseismic wave field, the intensity at every point of these being the arithmetic mean for six microseism samples. A coherent signal is clearly seen in the $SH$ component with the brightest spot at $300-320^\circ$ (azimuth) and $40-50^\circ$ (angle of emergence). As to $P$ and $SV$, the signal is lost in a higher activity of another stable endogenic source which shows as bright spots in the azimuth range of $60-220^\circ$. This phenomenon calls for a special study.

This remarkable coincidence of the arrival directions of the brightest coda reflection and the coherent component of microseismic noise indicates their common origin. The
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Figure 3.4
Figure 3  Image of a wave field in the azimuth–angle of emergence coordinates for three orthogonal body wave components (a – P, b – SV, c – SH) reconstructed from earthquake codas (date/time): A – Nov 14, 1993/07 h 46 min; B – Nov 18, 1993/02 h 10 min; C – Nov 17, 1993/00 h 39 min; d – array geometry; e – intensity scale; I – earthquake epicenter. Horizontal brackets below mark the locations of the brightest spots.
Figure 4A
Figure 4 Wave field image in the azimuth–angle of emergence coordinates for three orthogonal body wave components plotted from microseismic noise samples: A - Nov 9, 1993/17h 42 min; B - Nov 14, 1993/07h 46 min. For the other notations see Fig. 3.
image based on the microseisms comes from the region that generated the seismic signal. The coda contains diffracted (reradiated) waves mostly derived from incident compression- al and shear waves and waves that were radiated by internal sources. The latter might be initiated by transmitted seismic waves. The coherent component of the microseismic noise identified cannot have been a reflected signal from any surface sources, because no intensive surface sources were observed for the sample of Fig. 4.4 or for the average image at the azimuths of the coherent component; the most intensive component in the SH phase of the average image is that at azimuth 315°. The endogenic source observed in the azimuth range of 60–220° was too weak and too far from the reflector to provide enough energy for a deep reflection. It can thus be inferred that there was an active emission source, e.g., a magma body, which can be detected with the help of a phased seismic array, but is not seen on individual records. It should be noted that the arrival azimuths of the coherent component are fairly well consistent with the concentration of reflections in the top of the reflector (Fig. 2); that is, the emission signal seems to be generated by the upper of the two parts that make up the crustal heterogeneity.

This identification of an emission signal rediated by a crustal reflector is a new but not an unexpected result. The crustal heterogeneity is an elongated thin body that has an extensive complex interface between a solid host rock and a hot molten magma which is slowly cooling and solidifying. Evidently, the interface is a region of the active circulation of gas fluids and hot brines, chemical transformations, and phase changes. These processes involve changes in the volume, pressure, and fracturing of the rocks and entail changes in the stress field, which may generate weak seismic signals. The fact that all of the five crustal reflectors identified in the northeastern Japan Arc are located near the zones of lower P-wave velocity extending to 100–150-km depths [6] suggests the existence of active channels that connect these crustal heterogeneities with upper mantle magma sources. Magmatic activity near the S-reflectors is indicated by a specific deep-focus low-frequency earthquakes that occur in these areas near active volcanoes and low velocity zones and can be attributed to magmatic activity in the upper mantle and lower crust: these earthquakes occur at depths of 20–50 km corresponding to plastic flow regions and are distinguished by low frequencies (1–5.5 Hz), and magnitudes below or equal to 2.2 [8]. Magma rise toward the surface accompanied by gas liberation and abrupt pressure changes produced in the intermediate chamber by tectonic forces may result in pressure variations in the crustal magmatic heterogeneity, which generate seismic signals in a wide frequency range.

These results show that the direction of a weak seismic signal source can be determined using arrays with nonuniformly spaced stations and, in the case of rough topographies, even high frequencies (5–50 Hz). Because high frequencies are more sensitive to velocity variation and topography, the image seems to be largely formed by low frequencies. Further research should be done to investigate the structure of images in different frequency ranges to optimize frequency choice.
Figure 5 Wave field image reconstructed for three orthogonal body wave components, averaged over six noise samples. Notations as in Fig. 3.
CONCLUSIONS

The analysis of microseismic noise records proved emission tomography to be an efficient seismic noise tool in studies of the structure and dynamics of rocks in volcanic areas.

A coherent component was discovered in the microseismic noise, which arrives from the same direction as does the largest reflected $S \times S$ phase in the codas of local earthquakes.

The relative intensity and composition of the coherent signal vary over time. The bulk of the signal energy is in the form of shear waves, though some $P$ waves can clearly be seen in some noise samples.

The coherent component of microseismic noise is not a reflection: no bright surface sources could be detected at its azimuths, and microseismic noise is known to decrease rapidly with depth.

The coherent noise component seems to be a weak emission signal reflecting the processes occurring in and above a thin magma. The restriction of coherent noise to areas of long-period earthquakes and zones of low $P$-wave velocity extending into the upper mantle is evidence of possible active channels that connect crustal magma bodies with deeper intermediate magma sources. The dynamics of processes that operate in them is capable of affecting the character of emission radiation as well.

The chief methodological result of this study is that emission tomography can be applied to data obtained in poor observational conditions (nonuniform spacing of seismic stations over the area, surface recording in areas of rugged topography, and the like) and does not require large amounts of data: a few record segments of about 4 s are sufficient. Special observations oriented at emission tomography can yield much more definite results: a longer recording time will enable the use of a longer time window which will diminish statistical fluctuations of brightness and produce clearer and more detailed images.

Observations of seismic emission in seismic areas of Tadjikistan and Turkmenia revealed that high-frequency seismic noise (tens of hertz) was an indicator of the nonequilibrium state of rocks medium: noise level increased greatly before earthquakes [4]. It is expected that seismic emission can be used as an indicator of nonequilibrium states in a magmatic system, a fact useful for predicting volcanic eruptions. An example is Shirane Volcano near which the study reported here was carried out. The last eruption took place there in 1952.

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