DISCONTINUITY OF BASEMENT ROCK DEPTH IN THE EASTERN COAST OF LAKE BIWA DISCOVERED BY OBSERVATIONS OF MICROSEISMS

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We observed microseisms of the period range from 1 to 4 s with horizontal-motion seismometers on two linear seismic arrays, each about 10 km long, in a low seismicity plain on the south-east coast of Lake Biwa.

Stations on the linear arrays are classified into two groups based on the difference of spectral characteristics of microseisms: one is the stations to the west of the Hino River that exhibited peaks around the period of 3 s in the spectral ratio to a fixed reference station, and the other is those to the east of the Hino River that had no peaks in that period range. The difference of the spectral ratios may be attributed to the lateral change of the depth of basement rock.

In order to determine the depth of the basement rock on both sides of the river, we observed microseisms on both sides with vertical-motion seismometers arranged in two-dimensional arrays. We estimated the subsurface structure from phase velocities of observed microseisms by assuming that the main part of the observed microseisms consisted of the fundamental mode of Rayleigh waves, and compared the inferred structure with a deep-drilling log available in the neighborhood.

The estimated depths of the basement rock are 0.9 km on the west side and 0.1 km on the east side of the Hino River. This is consistent with the existing result of gravitational survey.

1. Introduction

The distribution of earthquakes around Lake Biwa, whose depths are shallower than 30 km, is shown in Fig. 1. There is a low seismicity area to the southeast of the lake, from Moriyama to the Suzuka range, in contrast to the areas of high seismicity to the northeast and the southwest of the lake.

The broken lines in Fig. 1 indicate major faults, and the areas of the high density of the lines correspond to the high-seismicity areas cited above. On the other hand, there are few faults in the above-mentioned low-seismicity area (The Research Group for Active Faults, 1980). This may be caused by thick and unconsolidated deposits, i.e., the Kobiwako group and the Biwako group, that make it difficult to survey outcrops of faults. Our target area, from Moriyama to Ohmi-hachiman, lies...
Fig. 1. Seismicity around Lake Biwa (Jan. 1976-Sep. 1986, focal depths shallower than 30 km) with major faults indicated by broken lines. Provided by Regional Center of Earthquake Prediction, Kyoto University. The rectangle is the target area in this study.

on the western edge of this low seismicity area.

The area is covered by thick and unconsolidated deposits originating from the Yasu and Hino Rivers. Though no topographic feature indicative of fault activity can be found, a steep-gradient belt of the Bouguer anomaly, dipping to the west at the rate of $-3\text{mgal/km}$, lies along the Hino River as shown in Fig. 2 (Horie et al., 1981). The belt exists not only on the shore but also extends toward the center of the lake. At the north end of this steep-gradient zone, Horie (1983) discovered a fault of west-side-down type by means of multi-channel seismic profiling. This fault in the bottom of the lake is located on the intersection of the steep-gradient belt of the Bouguer anomaly and a multi-channel seismic line as indicated by a cross symbol in Fig. 2. However, the strike of the fault was not determined because the fault was found only in one seismic profile.

The west-down displacement of the fault is consistent with the westward inclination of the Bouguer anomaly. The steep-gradient zone of the Bouguer anomaly, extending from the coast to the center of the lake, suggests that a west-down-type fault extends from the shore to the center of the lake. Hence our investigation into the buried faults in this area could be of significance for understanding the tectonics of the area from Moriyama to the Suzuka range.

In this study, seismic explorations by means of microseisms observation were employed for locating and measuring the discontinuity of subsurface structure across the target area around the steep-gradient belt of the Bouguer anomaly.

We will hereafter use the term "microseisms" to represent the seismic ground
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Fig. 2. Bouguer anomaly around Lake Biwa (after HORIE et al., 1981). The solid circles are lake-bottom stations of gravity, and a cross symbol indicates the location of the fault detected through multi-channel seismic profiling by HORIE (1983). The rectangle is the target area in this study.

noises of the period range from 1 to 4 s; this is equivalent to those seismic ground noises called “long period microtremors” by other authors (e.g., NARUSE et al., 1976).

There are various ways for investigating the subsurface structure based on microseisms, i.e., (1) methods in which one utilizes predominant periods (e.g., KANAI et al., 1954), (2) methods in which one evaluates changes in amplitude (e.g., NARUSE et al., 1976), and (3) methods in which one uses phase velocities (AKI, 1957). All these methods were applied in those areas where subsurface structures were horizontally stratified or slowly varying laterally.

In contrast to the above studies, IRIKURA and KAWANAKA (1980) were the first to try to detect lateral discontinuity of subsurface structures from microseisms. They observed microseisms at a site where the subsurface structure had been already determined. They found that spectral ratios to a reference station changed at the discontinuity of the subsurface structure, and then showed by a theory of reflection and transmission of surface waves that the peak frequency of the power spectrum on the basin side became lower with the increase in the distance from the discontinuity.
In this paper we first employ the spectral ratio of microseisms for detecting the discontinuity of subsurface structure, and next we estimate the amount of the basement offset with the method proposed by Horike (1985) by modeling a subsurface structure which satisfies phase velocities of microseisms measured by 2-D seismic arrays.

2. Exploration around the Steep-Gradient Belt of Bouguer Anomaly

We observed microseisms in order to determine the approximate subsurface structure in the area that includes the steep-gradient belt of the Bouguer anomaly.

Two lines, Line 86-5 and Line 86-7, were designed as shown in Fig. 3. Observations on the two lines were made on two different dates; Line 86-5 on May 17–18, 1986 and Line 86-7 on July 28–29, 1986, respectively. Both lines had a common station, OBS-4, as a reference. Our observations were performed at night to avoid short-period artificial noises. Most of the stations except for several reference stations were mobile because of the restriction in the amount of apparatus available. The observations at each station were made as shown by the diagrams in Fig. 4(a) and (b). The starting times and stopping times of recording depended on the watches of the observation crew, which had been adjusted prior to the first observation of the day. OBS-c is absent in Figs. 3 and 4(b) because of trouble with

Fig. 3. Locations of stations and a deep-drilling site. Light-screened area indicates a steep-gradient zone of gravity (more than \(-3\) mgal/km) and thick-screened area indicates the outcrops of the basement rocks. Cross symbol is the deep-drilling site after YokoYama et al. (1979).
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Fig. 4. Operation diagram of observations: (a) for Line 86-5, and (b) for Line 86-7. Numerals and characters in the left column are the station names.

We used portable-type long-period seismometers which were adjusted to the period range of 6.95–7.05 s and a damping factor of 0.7. Seismometers were placed on well-tamped ground or on a firmly fixed basestone to reduce rocking motion of the sensor bodies. The microseisms were recorded on analog tapes after being amplified 1,000–2,000 times. The recording duration was about 45 min or 60 min. Only the N-S component of microseisms was recorded to ease the task of observation.

The analog tapes were played back and digitized at a sampling interval of 0.03 s by a personal computer system in the laboratory. In the analysis, more than five sections of recording that had not been contaminated by transient noises, e.g., traffic noises, were chosen and the root mean squares of the spectrum were calculated from every time window of 30.72 s length in the recorded sections. Then, spectral ratios of each station to the reference station were calculated to eliminate the effect of the variation of sources of microseismic during the observations.

The relative frequency response of all instruments were kept almost flat within the range of about 3 dB over the frequency range of our discussions.

Examples of the data are shown in Fig. 5(a) and (b). Traces from the top to the bottom are arrayed west to east, and the roman numerals below each station name correspond to those in Fig. 4(a) and (b). In Fig. 5(a), low-frequency components of microseisms predominate more significantly in the stations which are located to the west, while there seems to be no distinct predominant low-frequency component in Fig. 5(b).
Fig. 5. Examples of observed microseisms: (a) on Line 86-5, and (b) on Line 86-7.

Fig. 6. Spectral ratios to the station OBS-4. (a) and (b) correspond to Line 86-5 and Line 86-7, respectively.

Therefore, in order to compare both data sets as having similar quality, we calculated the ratios of the spectrum at each station to that at the reference station acquired in the same observation period. The calculated spectral ratios are shown in
Fig. 6(a) and (b). Figure 6(a) is the spectral ratios for the stations along the Line 86-5, and Fig. 6(b) is those along the Line 86-7. The error bars attached to the left end of each ratio indicate the confidence intervals of 95% under the assumption that the spectral ratios follow the F-distribution (JENKINS and WATTS, 1968).

As shown in Fig. 6(a), the spectral ratios appear almost flat in the period range of 0.2 to 1.0 Hz on the stations to the east of OBS-4. In contrast to this, there are distinct peaks in the ratios at the frequency around 0.3–0.4 Hz on the stations to the west of OBS-5. It is evident in Fig. 6(a) that the farther west the station is located, the lower the peak frequency of the spectral ratio becomes, e.g., the peak frequency of the ratio is 0.4 Hz on OBS-5 and 0.3 Hz on OBS-9 at the western end of the Line 86-5. This phenomenon was reported also by IRIKURA and KAWANAKA (1980) and KASUGA (1983). On the other hand, in Fig. 6(b) there are troughs of the spectral ratios around 0.55 Hz on OBS-a and OBS-b, while peaks are observed in the frequency range of 0.3–0.4 Hz on the stations to the farther west of OBS-d.

It should be noted that the stations with peak frequency of 0.3–0.4 Hz in the spectral ratios, OBS-5 and OBS-9 and OBS-d to OBS-g, are located exclusively on the west side of the Hino River. This suggests a difference of subsurface structure between west and east of the river. Moreover, the transient belt of the spectral ratio of microseisms in the frequency range of 0.3–0.4 Hz lies along the Hino River, and its location coincides with the steep-gradient belt of the Bouguer anomaly which is shown as a lightly screened zone in Fig. 3. This is another indication of lateral change of subsurface structure, i.e., the depth of the basement rock, across this river.

It would be expedient to explain here the reason why we did not use raw power spectrum but used spectral ratios to investigate subsurface structure. First, in observations of microseisms, spectrum at a certain station changes considerably with elapse of time. Examples of such a change of spectrum are shown in Fig. 7(a) and (b). It is notable that the spectrum in Fig. 7(a) is flat within 6 dB in the frequency range of 0.2 to 1.0 Hz, but in Fig. 7(b) the spectrum shows smaller values.

![Fig. 7. Averaged spectra of microseisms in the different observation periods at the station OBS-4; (a) represents the averaged spectra during the observations of Line 86-5, and (b) shows those during the observations of Line 86-7. Change of the spectra in different observations is notable.](image)
in the frequency range below 1.0 Hz than those in Fig. 7(a), while it is comparable to that in Fig. 7(a) in the higher frequency range. The difference between Fig. 7(a) and (b) is about 20 dB at 0.3 Hz. Therefore, it is certain that the spectra of incident microseisms changed during the series of observations in the frequency range of 0.3 to 0.4 Hz. The effect of those variations can be removed by using spectral ratios.

Second, the peak frequency reflects not only the subsurface structure around stations but also the characteristics of source and propagation paths. It is, therefore, undesirable to utilize only peak frequencies to discuss subsurface structures around the stations, since this can lead to an unreliable conclusion.

Consequently, spectral ratios seem to be more preferable to raw spectra for separating the effects of subsurface structure around the stations from other effects.

However, in order to conclude that the variations of the spectral ratios reflect exclusively those of the subsurface structure, a further assumption that the spatial characteristics of incident microseisms, e.g., incident direction, are steady with respect to time is required. To verify this assumption, we examined the time variation of the spectral ratios of particular pairs of stations. Figure 8(a) displays the time variation of the spectral ratio of OBS-7 to OBS-4 in Line 86-5, and Fig. 8(b) that of OBS-a to OBS-4 in Line 86-7, respectively. Roman numerals in the spectral ratios indicate the observation time displayed in Fig. 4(a) and (b). Figure 8(a) shows stable ratios over all frequency ranges during all observation times, and Fig. 8(b) also shows stable ratios over the frequencies lower than 1.5 Hz. The spatial

![Fig. 8. Variations of the spectral ratios during the observation series. In (a), spectral ratios of OBS-7 to OBS-4 in Line 86-5 are shown, and those of OBS-a to OBS-4 in Line 86-7 are shown in (b). All the ratios remained steady during the observation series. The roman numerals correspond to those in Fig. 4(a) and (b).](chart.png)
3. Estimation of Subsurface Structures by Two-Dimensional Seismic Array Observations on Both Sides of the Hino River

In the previous section, we pointed out that there is some difference in the depths of the basement rock between the east and west sides of the Hino River. Next, we will estimate the depths of the basement rock based on the phase velocities of microseisms observed with seismic arrays. The locations of the arrays are indicated as “East array” and “West array” in Fig. 3. The observations were carried out on April 24–25, 1987, on the two West arrays of 1 and 0.5 km in diameter (Fig. 9(a)) and on April 27–28, 1987, on the single East array of 0.9 km in diameter (Fig. 9(b)). Each seismic array was composed of five stations, and only the vertical component of microseisms was observed. The seismometers, which were compact and of portable long-period type, were adjusted to the natural period of 6.95 to 7.05 s and to the damping factor of 0.7. Each station was connected to a recording system by cables. The records were stored on floppy diskettes in the form of digital data of sampling frequency at 100 Hz after being passed through a low-pass filter \( f_c = 1.6 \text{ Hz}, 20 \text{ dB/oct} \). More than six recordings as long as 3'24'' were carried out every 6 min at an array. Data were processed in the Data Processing Center of Kyoto University.

In the analysis, the data were selected as in the previous section and equalized for the responses of seismometer. After this procedure, we calculated the phase

![Fig. 9. Geometry of seismic arrays. (a) shows the large and small spreading geometries of the West arrays, and (b) is the East array.](image)
Fig. 10. Examples of observed microseisms on the West (a) and the East (b) arrays, respectively.

Fig. 11. Examples of F-K spectra obtained from (a) West arrays and (b) East array. The $K_x$ axis in (a) oriented toward N33°E, and that in (b) oriented toward right north.

velocities of microseisms by the frequency-wavenumber (F-K) spectrum method by Capon (1969).

Data of sufficient quality for analysis obtained on the East and the West arrays are exemplified in Fig. 10(a) and (b), respectively. The frequency ranges of analysis are 0.40–0.50 Hz in the West arrays and 0.62–0.87 Hz in the East array. These frequency ranges satisfy the following conditions:

1. Coherence is greater than 0.5;
2. Clear dispersion characteristics are noticeable in the phase angles between two stations;
3. The contours of $-1$ dB with respect to the peak in the F-K spectra constitute closed loops within $|K_x| = 1.04$ (km$^{-1}$).
Examples of F-K spectra obtained for the West arrays and for the East array are shown in Fig. 11(a) and (b), respectively. The directions of $K_y$ axis oriented toward N33°E in Fig. 11(a) and right north in Fig. 11(b), respectively. In both figures, contours are plotted every $-1$ dB and the arrows from the origin indicate the directions of propagation of microseisms. The indicated directions of propagations are from south in Fig. 11(a) and from NWN in Fig. 11(b), respectively.

The phase velocities obtained from the F-K spectra are shown in Fig. 12. The solid circles indicate the mean values of phase velocities in the West arrays and the solid squares indicate those in the East array. The error bars which are attached to the symbols in Fig. 12 show the standard deviations of the phase velocities estimated. Both the solid and broken curves are theoretical best fitting dispersion curves which are calculated for the subsurface structures in Table 1.

![Graph showing phase velocities](image)

**Table 1.** Subsurface structure models for both sides of the Hino River.

<table>
<thead>
<tr>
<th>West array</th>
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<tr>
<td>$V_p$ (km/s)</td>
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<td>$V_s$ (km/s)</td>
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<td>3.5</td>
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Fig. 13. (a) Deep-drilling log (after YOKOYAMA et al., 1979) and (b) shear-wave velocity structure models of Table 1 in (b).

Since the frequency ranges in which the phase velocities are estimated are too narrow for inversion analysis, i.e., 0.4–0.6 Hz on the West arrays and 0.62–0.87 Hz on the East array, we estimated the subsurface structures by trial and error. The initial values of $V_p$, $V_s$, and density of the Quaternary layers were assumed by referring to data by HORIKE (1985), and those values of the basement rock by IMAI et al. (1975) and AKI and RICHARDS (1980). The depth of the basement rock was determined as follows: the depth of the basement rock below the West arrays was fixed, because the depth had been known to be about 0.9 km (YOKOYAMA et al., 1979; Fig. 13(a)) from a deep-drilling log at 1.5 km west of the West arrays. $V_p$, $V_s$, and densities were so adjusted by trial and error as to fit the theoretical phase velocities to the observed ones. In the East array, we made an assumption that members of Quaternary layers similar to those on the West arrays lie also below the East array, and then the depth of the basement rock was estimated through a trial-and-error procedure because there was no a priori information about the depth of basement rock.

The shear wave velocities of the models shown in Table 1 are displayed again in Fig. 13(b). The estimated depth of the basement rock is 0.12 km on the East array in contrast to 0.9 km on the West arrays. Hence the difference between the depths on both sides is about 0.8 km. This result is consistent with the gravitational survey around this area (HORIE et al., 1981; Fig. 2). The values of the Bouguer anomaly are $-40$ mgal on the West arrays and $-31$ mgal on the East array, which gives the difference in the basement depth of 0.72 km if we assume $\Delta \rho = 0.3 \text{ g/cm}^3$. 
A geological condition also supports our result that the depth of the basement rock is shallower on the east side than on the west side of the Hino River. Though there are outcrops of the basement rock near the East array, as indicated in Fig. 3 by thick-screened area (ISHIDA et al., 1984), no outcrop of basement rock is found around the west arrays.

Hence we conclude that there is a difference in the basement depth of about 0.8 km between the eastern and western areas of the Hino River.

4. Conclusions

The results of the study are summarized below.

1. The spectrum of microseisms in the period range of 2 to 4 s changes on crossing the Hino River; spectral amplitudes on the west side are greater than those on the east side of the river.

2. The peak frequency of the spectral ratio becomes lower as the distance from the discontinuity is greater on the basin side.

3. A transient belt of the ratio of microseisms of 2 to 4 s period lies along the Hino River, and this belt is coincident with the steep-gradient belt of the Bouguer anomaly dipping westward.

4. The dispersion characteristics of microseisms are different on the east side and the west side of the Hino River, and the difference can be explained by the difference in the depth of basement rock between both sides of the river.

5. The amount of this difference is estimated as 0.8 km, which is consistent with the existing result of a gravitational survey.

6. Hence it is considered that the spectral change of microseisms in the period range of 2 to 4 s reflects a change of basement depth around the area.

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REFERENCES


