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Q_p Structure beneath the Northeastern Japan Arc Estimated from Twofold Spectral Ratio Method

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Abstract: A method for a reliable estimation of the Q structure has been developed, which takes a twofold spectral ratio (TSR) of station pair and source pair located on a straight line. The method embodies very few assumptions for the structure and no supposition for the source spectrum and for the ratio of Q_p to Q_s. The Q_p structure of the uppermost part of the mantle beneath the Tohoku District, Japan, has been obtained by applying the method to earthquake data recorded by the seismic network of Tohoku University. The estimated Q_p structure for the uppermost mantle indicates that the Q value is low in the area between the western coast line and the volcanic front, and high beneath the Japan Sea and the Pacific Ocean.

1. Introduction

The first quantitative estimation of the Q structure beneath the Tohoku District has been carried out by Hasegawa *et al.* (1979). They examined the amplitude ratio of P to S waves and estimated the two-dimensional (2-D) Q_s structure. Umino *et al.* (1981) investigated the 2-D Q_p structure by using the P-wave amplitude ratio between seismic stations. The Q_p and Q_s structures obtained by these two studies have very similar patterns to each other and have a close correspondence to the seismic wave velocity structure beneath this area (e.g. Hasemi *et al.*, 1984 ; Obara *et al.* 1986). The Q value is high in the uppermost part of the mantle on the eastern side of the aseismic front and in the descending slab, while the uppermost mantle on the western side of the volcanic front has a low Q value. Umino and Hasegawa (1984) investigated the three-dimensional (3-D) Q_s structure using an inversion method. Their result revealed that the characteristics of the Q structure mentioned above are basically common in the whole Tohoku District. Hashida (1984) also estimated the 3-D Q_s structure by using seismic intensity data and obtained a similar result.

In the estimation of the Q structure, usually some assumptions are involved in order to simplify the calculation. For example, amplitude source spectra were assumed to be flat in Oliver and Isacks (1967) and Barazangi and Isacks (1971); Hasegawa *et al.* (1979) and Umino and Hasgawa (1984) supposed Q_P/Q_S to be 9/4; Q values in the crust and in the descending plate were a priori given in Umino *et al.* (1981). Moreover, most of the previous works neglected the effects of the site responses (the effects of material at shallow depths beneath the receiver sites).

The corner frequency of the source spectrum should be verified when we intend to use the flat part of the spectrum. It is not well established that Q_P/Q_S is equal to 9/4; some researchers reported Q_P/Q_S is not 9/4 for the upper lithosphere. For example, Rautian *et al.* (1978) found that Q_S is larger than Q_P and $Q_P/Q_S \cong t_P/t_S \cong 1/1.78$ (where t_P and t_S denote the travel times of P and S waves, respectively); Modiano and Hatzfeld (1982) obtained $Q_P/Q_S \cong 1/1.64$; Frankel (1982) and Hasegawa (1982) showed that Q_P is almost equal to Q_S. Although there is no corroborative evidence that Q_P/Q_S is not 9/4 in the asthenosphere as far as we know, we should verify whether Q_P/Q_S is really equal to 9/4 there. Furthermore, we should not neglect the effect of the site response which considerably contaminates the observed spectra of seismic waves (Frankel, 1982). It is desirable to develop a method which requires fewer assumptions for the source spectrum, Q_P/Q_S and site response.

In this paper, we propose a method for estimating Q structure, which embodies very few assumptions for the structure, and no supposition for the source spectrum and for the ratio of Q_P to Q_S.

2. Mothod

The technique developed here is one analogous to the reverse-profile method for the determination of velocity structure in refraction seismology. In general, the displacement spectrum $F(\omega)$ by an earthquake at a certain point is expressed as

$$F(\omega)=GS(\omega)P(\omega)R(\omega)I(\omega)\exp\left\{-\frac{\omega}{2}\int\frac{dl}{QV}\right\}, \tag{1}$$

where ω is angular frequency, G contains the geometrical spreading and any other term which is independent of frequency, and $S(\omega)$ is the source spectrum. $P(\omega)$ is the path effect besides the inelastic attenuation (e.g. $1/\omega$ for a head wave or frequency dependence for a guided wave), $R(\omega)$ is the site response beneath the seismometer and $I(\omega)$ is the instrumental response. V is the seismic wave velocity and $\int dl$ is the integral along the ray. Let us assume that two events (A and B) and two seismic stations (S1 and S2) are located on a straight line as shown in Fig. 1.

The spectral ratio of S1 to S2 for the event A is expersed as

$$\frac{F_A^1(\omega)}{F_A^2(\omega)}\cong\frac{G_A^1R_A^1(\omega)I^1(\omega)}{G_A^2R_A^2(\omega)I^2(\omega)}\exp\left\{-\frac{\omega}{2}\left(\frac{t_1}{Q_1}-\frac{t_m}{Q_m}-\frac{t_2}{Q_2}\right)\right\}, \tag{2}$$

where subscripts and superscripts for F, G, R and I indicate events and stations, respectively. Q_1, Q_2 and Q_m denote the Q values for the ray paths from p to S1, from q to S2, and from p to q shown in Fig. 1b, respectively; travel times for these paths are

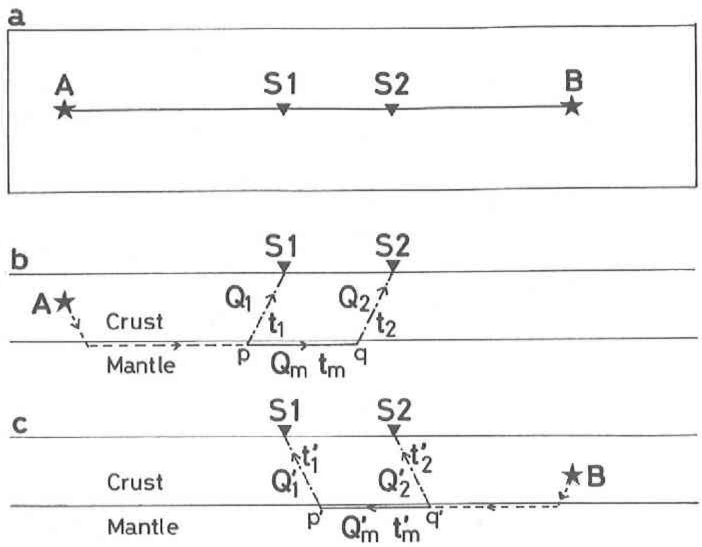

Fig. 1 Typical arrangement of events (A and B) and seismic stations (S1 and S2) to which we can apply the TSR method.

indicated by t_1, t_2 and t_m, respectively. In deriving the above equation, we assumed $P_A^1(\omega) \cong P_A^2(\omega)$, which is quite reasonable if the interval between S1 and S2 are much smaller than the epicentral distance. In the same way, we obtain the spectral ratio for the event B (Fig. 1c) as

$$\frac{F_B^1(\omega)}{F_B^2(\omega)} \cong \frac{G_B^1\, R_B^1(\omega) I^1(\omega)}{G_B^2\, R_B^2(\omega)\, I^2(\omega)} \exp\left\{-\frac{\omega}{2}\left(\frac{t_1'}{Q_1'}+\frac{t_m'}{Q_m'}-\frac{t_2'}{Q_2'}\right)\right\}. \tag{3}$$

If the lateral variation of the structure is not strong beneath the stations, we can consider $t_j = t_j'$, $Q_j' = Q_j'$ $(j=1, 2, m)$ and $R_A^i(\omega) \cong R_B^i(\omega)$ $(i=1, 2)$. Then dividing eq. (3) by eq. (2), we obtain

$$\frac{F_B^1(\omega)}{F_B^2(\omega)} \Big/ \frac{F_A^1(\omega)}{F_A^2(\omega)} = \frac{F_B^1(\omega)\, F_A^2(\omega)}{F_B^2(\omega)\, F_A^1(\omega)} \cong \frac{G_B^1\, G_A^2}{G_B^2\, G_A^1} \exp\left\{-\frac{\omega t_m}{Q_m}\right\}. \tag{4}$$

We call, hereafter, the left side of eq. (4) 'twofold spectral ratio (TSR).' In eq. (4), $G_B^1\, G_A^2 \Big/ G_B^2\, G_A^1$ is independent of frequency, and t_m can be obtained from the mean value of differential travel times between the two stations for the two events. Thus we can estimate the Q value for the upper mantle (Q_m) from the frequency dependency of the TSR if Q_m is independent of frequency.

If there is no velocity gradient in the mantle, Pn wave becomes head wave. The head wave is expressed as $CiS(\omega)/\omega r^2$ (where C is a constant and $i=(-1)^{1/2}$) when the epicentral distance r is much larger than wave length and the thickness of crust (Aki and Richards, p. 212, 1980). Therefore, the head wave can also be expressed in the form eq. (1), namely, $G=iC/r^2$ and $P(\omega)=1/\omega$.

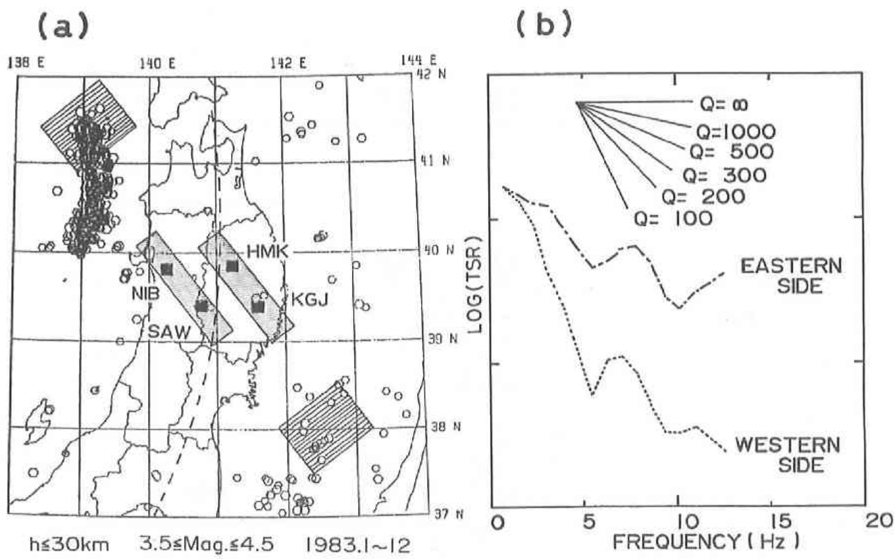

Fig. 2 (a) Examples of locations of seismic stations and events used for the Q estimation by the TSR method. The seismic stations are indicated by solid squares. The areas where Q_p values of the uppermost mantle can be estimated are shadowed, and the events used are in the hatched area of oblique lines. (b) Examples of the TSR's. Dot-dash and dotted lines are the TSR's by using HMK-KGJ and NIB-SAW profiles, respectively. Each TSR is calculated by stacking the data from six pairs of events.

Examples of TSR's for Pn wave are shown in Fig. 2. Figure 2a shows the locations of the studied areas (shadowed) and events (hatched by oblique lines) ; the obtained TSR's are shown in Fig. 2b. Each TSR is calculated by averaging the data from six pairs of events.

One of strong advantages of this method is that we can stack the TSR (eq. (4)). The effect of the stacking is demonstrated in Fig. 3. TRS's for the station pair HMK-KGJ are shown in the figure (dotted and dot-dash lines) ; and stacked (averaged) TSR is the lowest trace (solid line). The TSR is considerably smoothed by the stacking effect. Suzuki (1971) and Sacks and Okada (1974) proposed similar methods to ours. However, they did not adopt the stacking procedure which substantially heightens the reliability of the result. The stacking procedure is highly effective when the frequency band is not very wide and a slight fluctuation in a spectrum considerably affect the result.

Another abvantage of this method is that the site response can be well eliminated when the response is independent of the incident direction of the ray. Figure 4 represents the spectral ratio of HMK to KGJ for the events beneath the Pacific Ocean (dashed line) and that of KGJ to HMK for the events beneath the Japan Sea (dot-dash line). Each trace is obtained by averaging six spectral ratios. The peaks and troughs of the two traces probably indicate the site response ratio. The TSR is the average of these traces (solid line) ; the oscillation of the trace is quite damped.

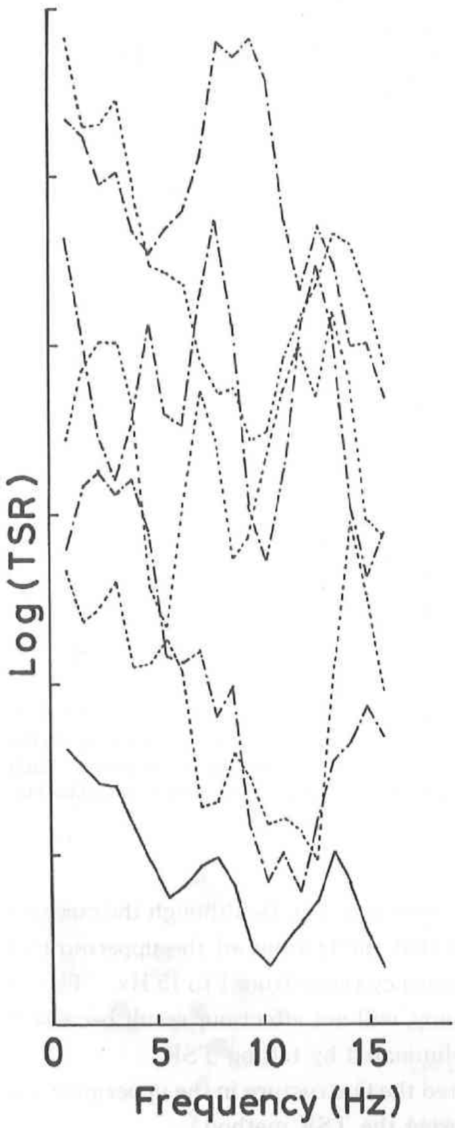

Fig. 3 An example of the stacking effect. Dotted and dot-dash lines indicate the TSR calculated from a single pair of events. Solid line indicates the TSR obtained by stacking the upper six traces. Each trace is shifted arbitrarily in the vertical direction.

We can used the TSR method when two stations and two events lie on a straight line as shown in Fig. 1a, and the sources are sufficiently far from stations. In this method we assume only that the lateral variation of the structure is not strong beneath the stations and the Q value is independent of frequency. Thus, the present method requires the very few assumptions. The first assumption is considered to be reasonable if we select station pairs which are far from tectonic boundaries. It is difficult, prior to analysis, to know whether the second assumption is reasonable or not. However, if Q value really depends on frequency, the TSR curve would not be a straight line on a semilogarithmic scale. The obtained frequency dependencies of TSR's give essentially straight lines on

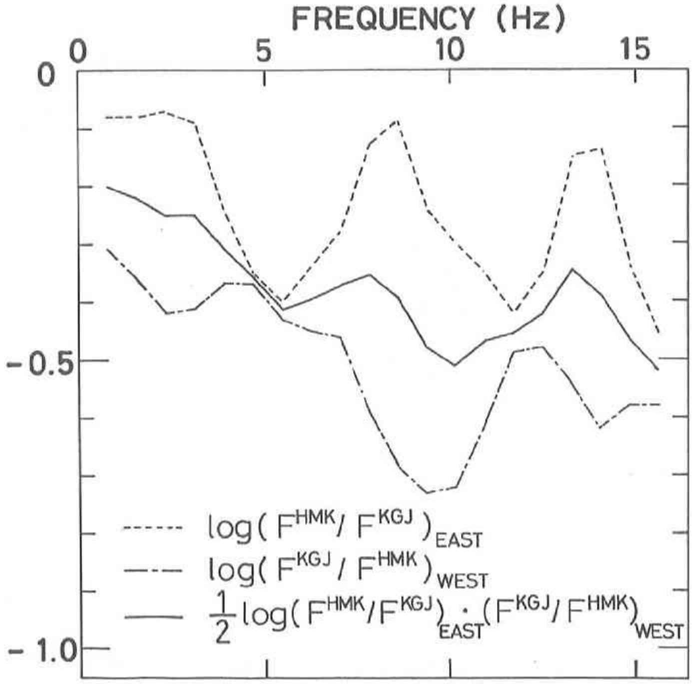

Fig. 4 Comparison between the spectral ratio and the TSR. Dashed and dot-dash lines indicate the spectral ratios of P waves at stations HMK and KGJ for events beneath the Pacific Ocean and those beneath the Japan Sea, respectively. Each trace is obtained by stacking the data from six events. The TSR is calculated by averaging these two traces and shown by solid line.

a semilogarithmic scale as shown in Fig. 2 (see also Fig. 7), although the curves oscillate slightly. Therefore, it can be considered that the Q value of the uppermost mantle is almost independent of frequency in the frequency range from 1 to 15 Hz. The frequency dependency of the Q value in the crust, if any, will not affect our result because it can be treated as the site response and can be eliminated by taking TSR.

In the following section, we investigated the Q structure in the uppermost part of the mantle beneath the Tohoku District by using the TSR method.

3. Q Structure for the Uppermost Part of the Mantle

Figure 5 displays examples of the waveforms for shallow earthquakes beneath the Japan Sea (a) and beneath the Pacific Ocean (b). Pn and Sn waves are clearly attenuated in the eastern part of the Tohoku District for the Japan Sea event (Fig. 5a). It should be noted that Sn wave is fairly attenuated at station SAW which is located west of the volcanic front. On the other hand, station ATM has a clear Sn onset although its epicentral distance is greater than SAW's. Figure 5b shows that Sn wave is not very attenuated as SAW for the event beneath the Pacific Ocean. From these figures, we can estimate qualitatively that the low-Q zone is located between the volcanic front and the

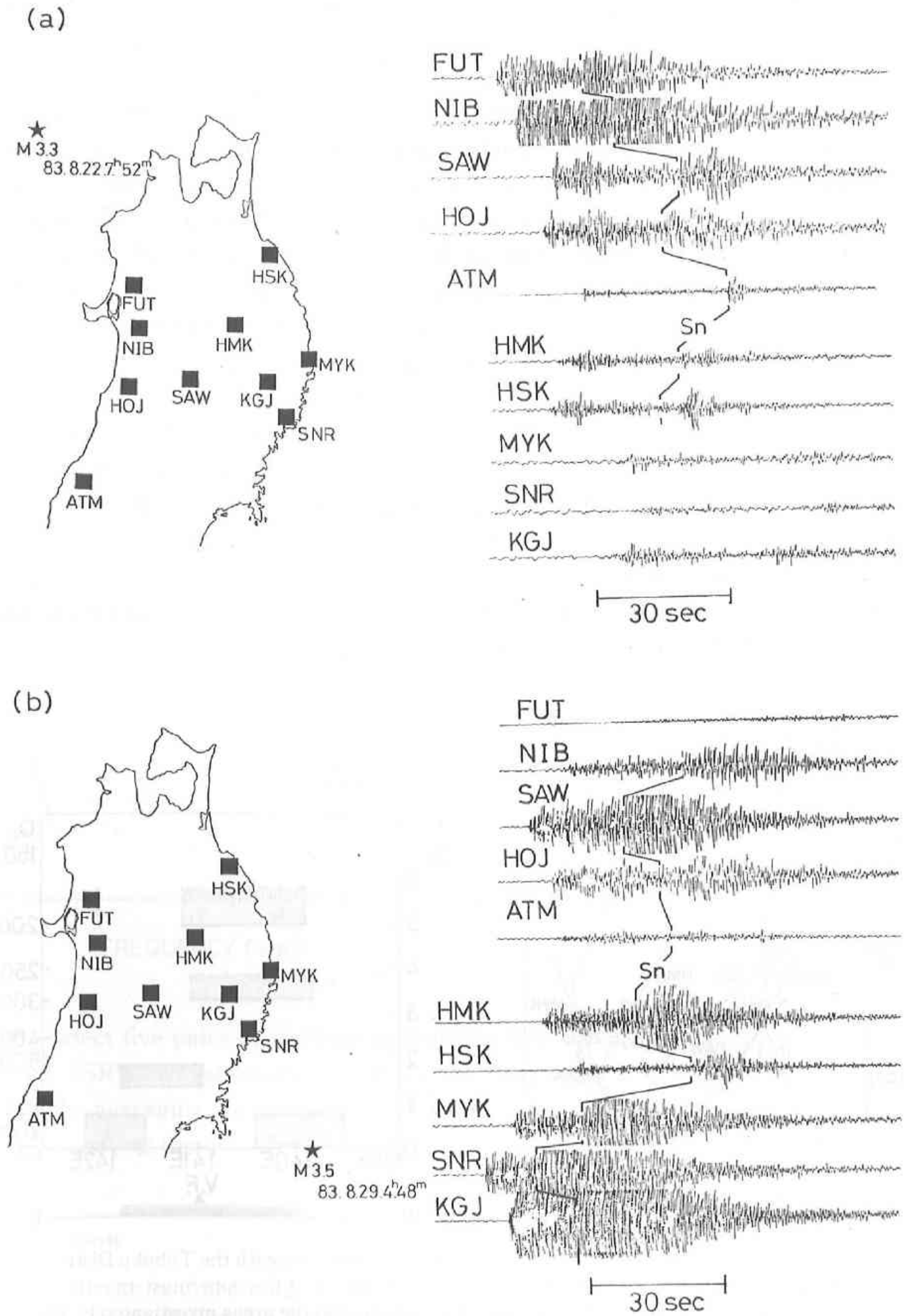

Fig. 5 Examples of vertical-component seismograms for a shallow event beneath the Japan Sea (a) and for that beneath the Pacific Ocean (b). Squares and stars indicate the seismic stations and epicenters, respectively.

coast of the Japan Sea. Quantitative investigation is necessary for estimating the Q value since different mechanisms such as defocusing may contribute to the attenuation (e.g. Matsuzawa *et al.*, 1984).

Usually, S wave amplitude is more sensitive to the change of the structure. Sn wave is strongly attenuated in the trace for station SAW in Fig. 5a. A large phase after the Sn phase is presumably S_MS phase reflected at the Moho discontinuity, taking the amplitude and travel time into consideration (P_MP is also seen about two second after the Pn phase (Umino *et al.*, 1985)). The small Sn phase and the large S_MS phase may indicate the existence of a melting layer at the top of the mantle. Unfortunately, the records at stations near the coastline are saturated for events beneath the Japan Sea when SAW has the large Sn wave amplitude enough to analyze. Therefore, we investigate the Q_p structure in the uppermost part of the mantle beneath the Tohoku District.

Earthquake data recorded by the seismic network of Tohoku University are analyzed in the present study. The seismometers used here are of 1 Hz velocity type and the waveform data are stored in digital tapes with sampling rate of 100 samples/s. We analyze first 1.28 second data after the P arrival for the calculation of spectral ratio and 1.28 second data before the arrival as the reference of noise levels. The spectra are claculated using FFT. Events whose epicentral distances are less than 120 km are not used because their first arrivals may not be Pn waves.

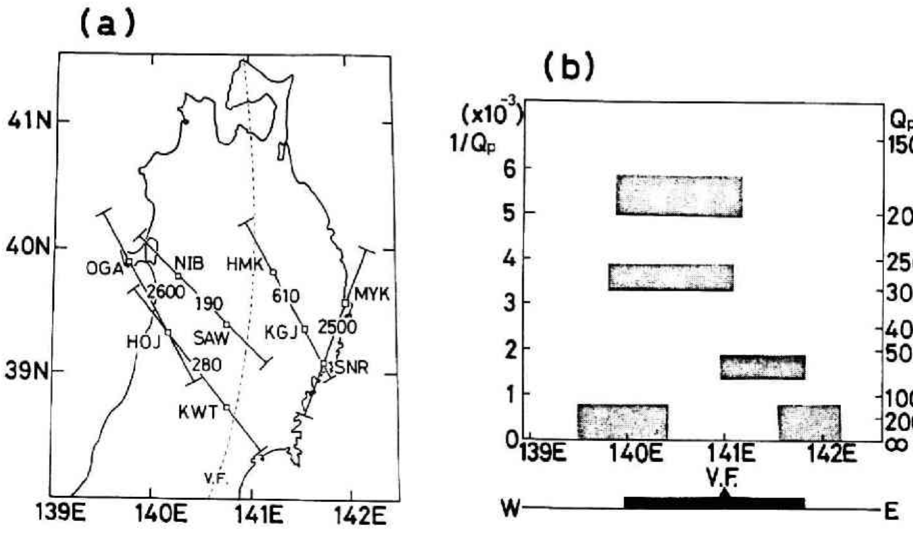

Fig. 6 Estimated Q_p structure for the uppermost mantle beneath the Tohoku District. (a) The locations of profiles used for the estimation of the uppermost-mantle Q_p structure. Squares and bars denote the stations and the areas investigated by the TSR method, respectively. The obtained Q_p values are shown in the middle of each bar. V.F. indicates the volcanic front. (b) The Q_p value for the uppermost part of the mantle projected on the E-W section. Q_p values are shown by shadows with the errors estimated from the least squares method. Thick line and triangle at the bottom of the figure indicate the land area and the volcanic front, respectively.

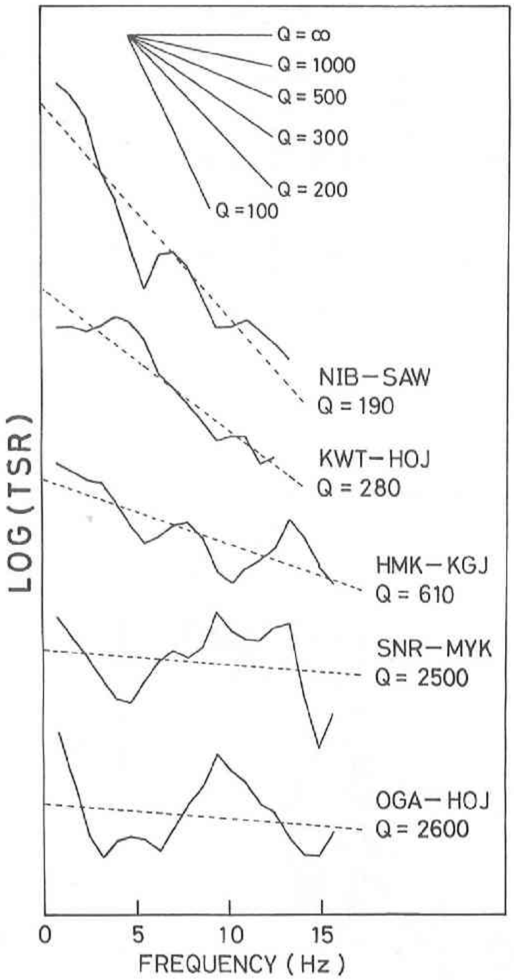

Fig. 7 TSR's and Q_p values estimated for the profiles shown in Fig. 6a. Each TSR shown here is obtained by averaging the TSR's of six pairs of events. Solid lines and dashed lines denote the obtained TSR's and fitting lines estimated from the least squares method, respectively. Each trace is shifted arbitrarily in the vertical direction.

We select five pairs of stations as shown in Fig. 6a, and calculate the TSR's. The obtained TSR's and estimated Q_p values are shown in Fig. 7. Each trace in Fig. 7 is obtained by averaging six independent TSR data, and only the parts with good S/N ratio are shown taking the noise levels into account. The estimated Q_p values are plotted in Fig. 6b.

4. Discussion

Figure 6 shows that the Q_p values in the uppermost mantle on the western side of the volcanic front, on the eastern side of it, and around the coastlines are 170–300, 500–800 and >1500, respectively. The comparison of the present result to the previous results are shown in Table 1. In the table, Q_s values estimated by Hasegawa *et al.* (1979) and Umino *et al.* (1984) are converted into Q_p values assuming the relation $Q_p/Q_s=9/4$. The Q structure estimated in this study agrees well with the results of the previous

Table 1. Q_p structure of the uppermost part of the mantle beneath the Tohoku District

<table>
<tr><th></th><th colspan="2">Coastline of the Japan Sea ▼</th><th colspan="3">Volcanic front ▼</th><th colspan="4">Aseismic front ▼</th></tr>
<tr><td>This study</td><td>>1500</td><td colspan="3">170-300</td><td colspan="3">500-800</td><td colspan="2">>1500</td></tr>
<tr><td>Hasegawa et al. (1979)</td><td colspan="4">230</td><td colspan="3">790</td><td colspan="2">2300</td></tr>
<tr><td>Umino et al. (1981)</td><td colspan="4">200</td><td colspan="3">500</td><td colspan="2">2000</td></tr>
<tr><td>Umino and Hasegawa (1984)</td><td colspan="2">330</td><td>220</td><td colspan="2">800</td><td>1900</td><td colspan="2">1200</td><td>3400</td></tr>
</table>

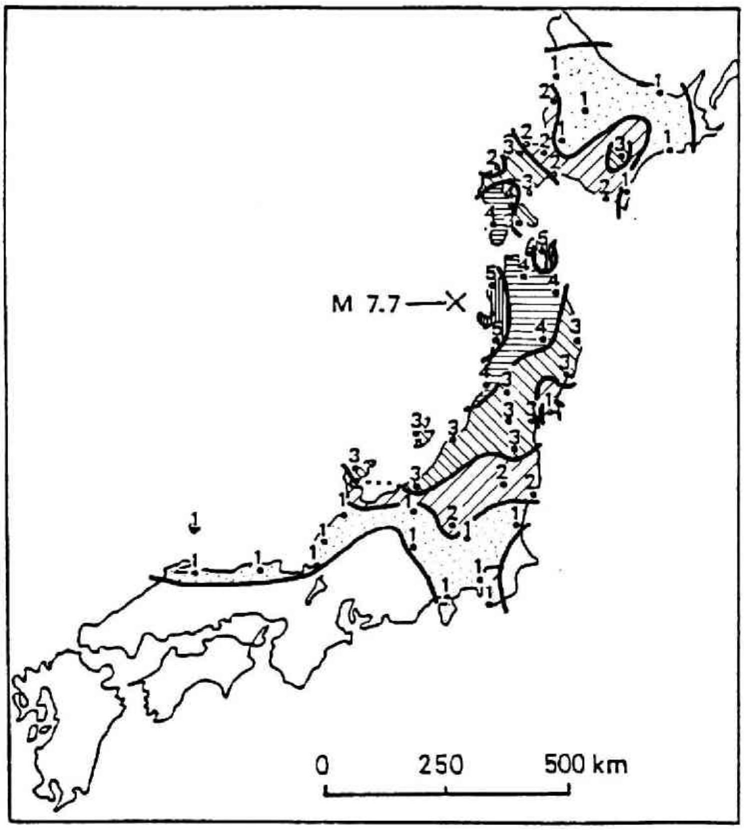

Fig. 8 Seismic intensity map for the 1983 Japan Sea Earthquake (after Japan Meteorological Agency, 1984). Cross and numerals in the map indicate the epicenter and seismic intensities, respectively.

studies except the value around the coast of the Japan Sea. This coincidence of the results probably indicates that Q_P/Q_S is nearly equal to 9/4 in the upper mantle beneath the Tohoku District as assumed in Hasegawa *et al.* (1979) and Umino *et al.* (1984).

Beneath the vicinity of the Japan Sea coast, the Q_P value estimated in this study is much higher than those in the previous studies. In deriving eq. (4), it is assumed that the lateral change in the structure is not strong beneath the stations. The velocity structure considerably changes laterally beneath the coast of the Japan Sea according to Yoshii and Asano (1972). The difference in Q value mentioned above may come from the inappropriate profile OGA-HOJ that crosses the coastline. However, this is not considered to be the main cause for the difference in Q value. The numbers of rays passing through the upper mantle beneath the Japan Sea are very few in the previous studies. Moreover, Figure 5a shows that ATM whose epicentral distance is greater than SAW's has the Sn onset clearer than SAW as mentioned before. Thus, a high-Q zone probably exists at the top of the upper mantle beneath the Japan Sea.

Figure 8 shows a seismic intensity map for the 1983 Japan Sea Earthquake (Japan Meteorological Agency, 1984). The source mechanism of the event is E-W compressional reverse fault type (Umino *et al.*, 1985). The seismic energy radiated toward the south could be small. However, the contours of intensity extend to N-S ; the quake is felt only along the coast of the Japan Sea in the western part of Honshu. The similar patterns of intensity distribution are seen for other shallow events beneath the Japan Sea

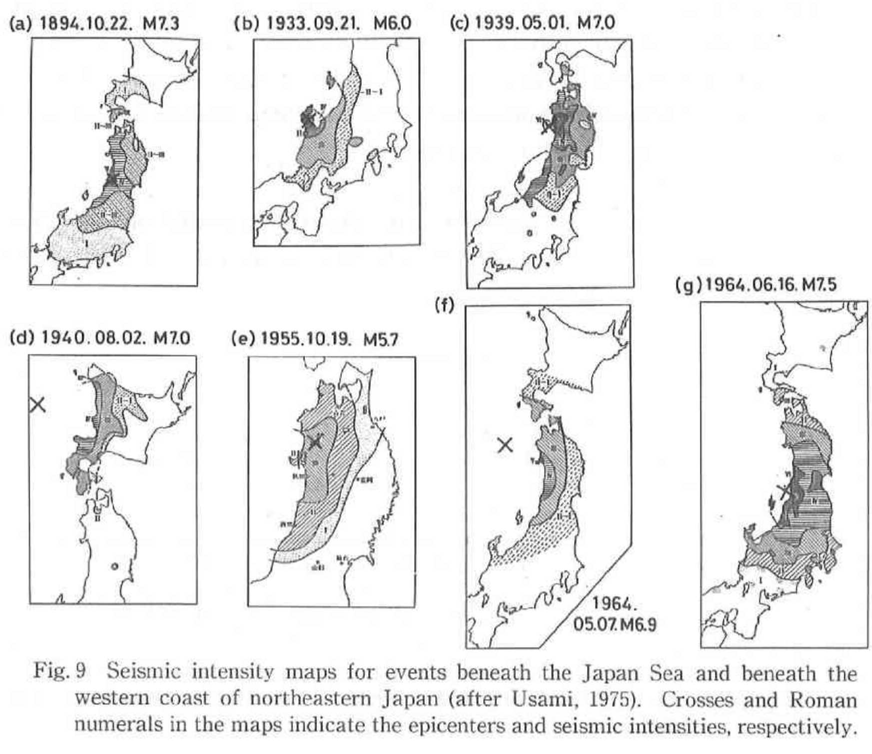

Fig. 9 Seismic intensity maps for events beneath the Japan Sea and beneath the western coast of northeastern Japan (after Usami, 1975). Crosses and Roman numerals in the maps indicate the epicenters and seismic intensities, respectively.

and beneath the west coast of the Tohoku District as shown in Fig. 9. The pattern is not so clear in the case of the 1964 Niigata Earthquake (Fig. 9g), but the area with seismic intensity V still extends along the coastline. Of course, the uppermost-mantle Q_p structurce cannot be estimated simply from the intensity map because the intensity mainly reflects the acceleration amplitude of direct S wave and is affected by the source mechanism, geometrical spreading and site response. Okamoto *et al.* (1984) reported that the waves passing through the upper mantle beneath the Japan Sea contains a plenty of high frequency contents, which is consistent with the intensity distribution. Yamada *et al.* (1984) investigated the Q_s structure beneath the Japan Sea and obtained a high Q_s value although their result is mainly concerned with the crust. These results support that the Q value of the uppermost part of the mantle is probably large beneath the Japan Sea.

The TSR method can be used also for the investigation of Q structure of the crust if the stations are densely distributed. Especially, this technique would be successful for the seismic refraction profile with controlled sources in order to investigate the crustal Q structure.

5. Conclusion

In order to make a reliable estimation of the Q structure, we have developed 'twofold spectral ratio (TSR)' method which embodies the fewest assumptions for the structure, and no supposition for the source spectrum and for the ratio of Q_p to Q_s. This method has been applied to shallow earthquake data to estimate the uppermost-mantle Q_p structure beneath the Tohoku District. The estimated Q_p values for the uppermost mantle beneath the Japan Sea, between the western coast line and the volcanic front, between the volcanic front and the aseismic front, and beneath the Pacific ocean are >1500, 170–300, 500–800 and >1500, respectively.

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References

Aki, K., and P.G. Richards, 1980: *Quantitative seismology; theory and methods*. W.H Freeman and Company, San Francisco, 932 pp.

Barazangi, M., and B. Isacks, 1971: Lateral variations of seismic-wave attenuation in the upper mantle above the inclined earthquake zone of the Tonga Island Arc: deep anomaly in the upper mantle, *J. Geophys. Res.*, **76**, 8493–8516.

Frankel, A., 1982: The effects of attenuation and site response on the spectra of microearthquakes in the northeastern Caribbean, *Bull. Seism. Soc. Am.*, **72**, 1379–1402.

Hasegawa, A., N. Umino, A. Takagi, and Z. Suzuki, 1979: Double planed deep seismic zone and anomalous structure in the upper mantle beneath Northeastern Honshu (Japan), *Tectonophysics*, **57**, 1–6.

Hasegawa, T., 1982: A study on the focal process and the precursory seismic activity for the 1970 Southeastern Akita Prefecture Earthquake, *Thesis*, Tohoku University, Sendai, 229 pp. (in Japanese).

Hasemi, A.H., H. Ishii, and A. Takagi, 1984 : Fine structure beneath the Tohoku district, northeastern Japan arc, as derived by an inversion of P-wave arrival times from local earthquakes, *Tectonophysics*, **101**, 245-265.

Hashida, T., 1984 : Three-dimensional seismic attenuation structure beneath the Japanese Islands as determined from seismic intensity data, *Thesis*, University of Tokyo, Tokyo, 203 pp.

Japan Meteorological Agency, 1984 : *The seismological bulletin of the Japan Meteorological Agency for May 1983*, 173 pp.

Matsuzawa, T., H. Shimizu, A. Hasegawa, A. Takagi and T. Sato, 1984 : Attenuation of seismic waves around the Kyuroku Island, *Abstr. Ann. Meet. Seism. Soc. Japan, 1984 April*, p. 175 (in Japanese).

Modiano, T., and D. Hatzfeld, 1982 : Experimental study of the spectral content for shallow earthquakes, *Bull. Seism. Soc. Am.*, **72**, 1739-1758.

Obara, K., A. Hasegawa and A. Tagaki, 1986 : Three dimensional P and S wave velocity structure beneath the northeastern Japan arc, *Zisin, Ser. II*, **39**, 201-215 (in Japanese).

Okamoto, T., K. Watanabe and A. Kuroiso, 1984 : Property of the attenuation for the seismic waves passing through the upper mantle beneath the southeastern part of the Japan Sea, *Abstr. Ann. Meet. Seism. Soc. Japan, 1984 April*. p. 177 (in Japanese).

Oliver J., and B. Isacks, 1967 : Deep earthquake zones, anomalous structures in the upper mantle, and the lithosphere, *J. Geophys. Res.*, **72**, 4259-4275.

Rautian, T.G., V.I. Khalturin, V.G. Martynov and P. Molnar, 1978 : Preliminary analysis of the spectral content of P and S waves from local earthquakes in the Garm, Tadjikistan region, *Bull. Seism. Soc. Am.*, **68**, 949-971.

Sacks, I.S., and H. Okada, 1974 : A comparison of the anelasticity structure beneath western South America and Japan, *Phys. Earth Planet. Inter.*, **9**, 211-219.

Suzuki, S., 1971 : Anomalous attenuation of P waves in the Matsushiro earthquake swarm area, *J. Phys. Earth*, **20**, 1-21.

Umino, N., and A. Hasegawa, 1984 : Three-dimensional Q_s structure in the Northeastern Japan Arc, *Zisin, Ser. II*, **37**, 217-228 (in Japanese).

Umino, N., A. Hasegawa and A. Takagi, 1981 : Q structure beneath the Tohoku District, *Abstr. Ann. Meet. Seism. Soc. Japan, 1981 October*, p. 28 (in Japanese).

Umino, N, A. Hasegawa, K. Obara, T. Matsuzawa, H. Shimizu, A. Takagi, K. Tanaka and M. Kosuga, 1985 : Hypocenter distribution of foreshocks and aftershocks of the 1983 Japan Sea Earthquake, *Zisin, Ser. II*, **38**, 399-410 (in Japanese).

Usami, T., 1975 : *Materials ; catalogue of the hazard earthquakes in Japan*, University of Tokyo Press, Tokyo, 327 pp. (in Japanese).

Yamada, N., M. Kosuga and H. Sato, 1984 : Q_s structure beneath the northern coast of the Japan Sea in the Tohoku District by using the aftershocks of the 1983 Japan Sea Earthquake, *Abstr. Ann. Meet. Seism. Soc. Japan, 1984 April*, p. 174 (in Japanese).

Yoshii, T., and S. Asano, 1972 : Time-term analyses of explosion seismic data, *J. Phys. Earth*, **20**, 47-59.