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Reflection profiling and velocity structure beneath Mizuho traverse route, East Antarctica

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Abstract: A seismic survey with a high density network was performed by JARE-41 in the austral summer of 1999 to 2000. The seismic line is spread out over 190 km along the Mizuho traverse route. 160 seismic stations programmed on timer operation were installed. A clear reflection is obtained in some shot records. It was described as a PmP reflection in a previous study. This survey geometry enables us to apply reflection analysis. A single fold profile and corresponding velocity structures for the JARE-41 seismic line were obtained in this study. The dipping Moho reflector and dipping reflectivity layer are revealed beneath the full spread of the seismic line in the profile. The Moho reflector is located at approximately 42 km beneath the shot point S-6 and at approximately 29 km below sea level beneath the coastal end of the profile. There is a lateral change in reflection character in the reflective layer in the neighborhood of the shot point S-4, although the velocity of the upper crust from refraction analysis is constant in this region. The structure obtained might be one case of Seaward Dipping Reflector Sequences in the continental margin area of the past Gondwana supercontinent.

key words: seismic survey, Mizuho traverse route, Enderby Land, seismic reflection, Moho discontinuity, Seaward Dipping Reflector Sequence

1. Introduction

Enderby Land, located in eastern Antarctica, has been studied intensively in Japanese Antarctic Research Expedition (JARE) research projects since the 1950's. Enderby Land contains the Napier Complex as a core with surrounding Proterozoic and Paleozoic Complexes. The Lützow-Holm Bay (LHB) area, where Syowa Station is located, had igneous activity at 500 Ma (Hiroi *et al.*, 1991; Shiraishi *et al.*, 1994), suggesting that this area may contain portions with different histories. It is probable that the subsurface structure involves cumulative remnants of ancient geological processes. However, a thick ice-sheet prevents us from direct geological observation of the detailed crustal structure. Only geophysical approaches, such as seismic methods and gravity surveys, address such issues.

JARE-20, -21 and -22 conducted deep crustal seismic surveys along the Mizuho traverse route (Ikami *et al.*, 1983; Ito *et al.*, 1983). Their seismic line is the same as that of this study (Fig. 1). The operations used chemical explosives as a seismic source. The record from JARE- 20, -21 and -22 includes a clear later arrival at distance greater than 100 km from the shot point. Ikami *et al.* (1984) constructed a velocity model down to the Moho discontinuity. Then Ikami and Ito (1986) proposed a refined model to explain the amplitude of the later arrivals. Ito and Kanao (1995) tried to map the depth of the corresponding reflector and discussed its topography. JARE-20, -21 and -22 were the best surveys at the time but were limited to only one shot point for the reflection

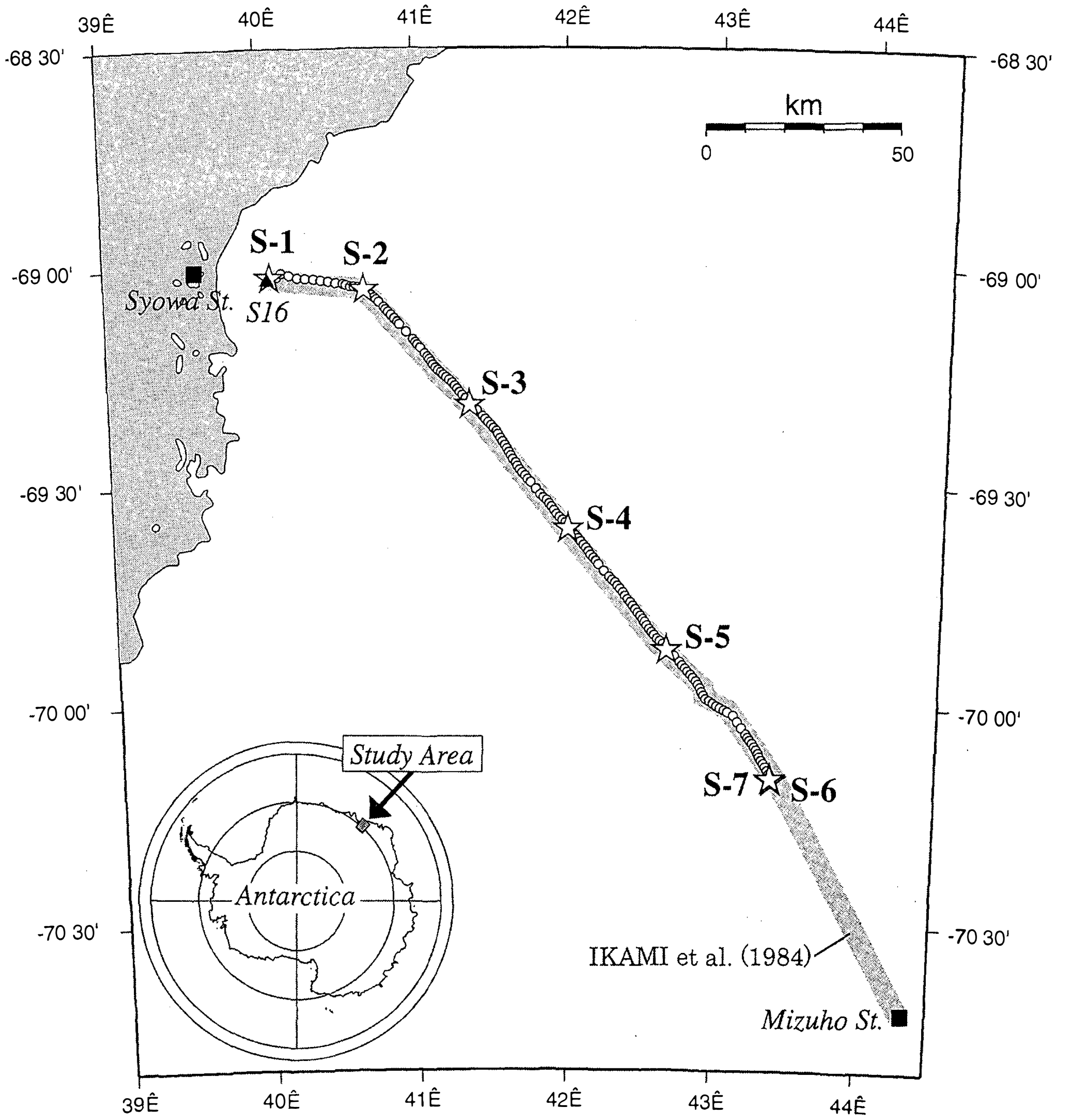

Fig. 1. JARE-41 seismic line (after Miyamachi et al., 2001).
Open stars denote seven shot points and open circles indicate seismic stations except for the line-up stations near each shot point. Distances between the closest shot points are approximately 40 km. This seismic line includes 160 seismic stations with average separation of 1.2 km. The seismic line of Ikami et al. (1984) is plotted with a thick gray line.

analysis.

After this work, there was a long interval before the next chance for a JARE-type survey. JARE-41 was carried out during the austral summer of 1999 to 2000 along the Mizuho traverse route again as part of the Structure and Evolution of East Antarctic Lithosphere (SEAL) transect project. The purpose of JARE-41 was to obtain more precise structure than those estimated from JARE-20, -21 and -22, taking advantage of progress in the recording apparatus technology. The details of the JARE-41 field operation are described in Miyamachi *et al.* (2001) and Murakami *et al.* (2001).

Denser seismic station spacing and more shot points were used in JARE-41 than in JARE-21, and -22. These improvements offered significant advantage for reflection analysis. Velocity structure and single fold profiles from JARE-41 shot records are described and discussed in this paper.

2. Seismic survey and data

The JARE-41 seismic line extended up to 190 km along the Mizuho traverse route from base camp S16. 160 seismic stations were spread out with programmed timer operation. Each station included a single component seismometer (model L22D with 2 Hz vertical motion type by MarkProducts) and a stand-alone seismic recorder (LS8000SH 20 Mbytes capacity by Hakusan Industry Corporation). A Cyclone battery (6V 15Ah) was used as the power source for the recorder. The seismic recorders and batteries were installed in plastic containers with heat insulation.

Each station was programmed to have three data acquisition windows and five time-base calibrations each day with a built-in GPS receiver. Data acquisition and storage for each shot was done at the site. All stations were retrieved after completion of all shooting programs; then the data were transferred from the recorders on the return trip to Japan.

The locations of the stations were measured with a precise GPS receiver (Ashtech Z-FX) after their deployment. An exact location list for the stations is described in Miyamachi *et al.* (2001).

Shot records with clear later arrivals are obtained from the JARE-41 survey. For instance, a shot record for shot point S-6 is shown in Fig. 2a. The vertical axis denotes distance from the shot point in km and the horizontal axis denotes raw travel time in seconds. The arrows mark later arrivals R1 and RS that appear in the range over 100 km. These later arrivals are enhanced by band-pass filtering (2–8 Hz), as shown in Fig. 2b. The move out of these later arrivals are not linear but rather hyperbolic, as expected for a reflector. The same later arrival with the reflection R1 was already reported from previous surveys (Ikami and Ito 1986; Ikami *et al.*, 1984). The later arrival R1 is considered as PmP reflection and the phase RS is expected to be a PmS converted reflection from its arrival time as referred to Ikami *et al.* (1984). The frequency content of the reflection R1 (Fig. 3b) is substantial by lower than that of the seismic record as a whole (Fig. 3a). Thus, band-pass filtering is effective in enhancing the reflection. The corresponding later arrivals with phase R1 are also observed in other shot records of S-2 and S-3.

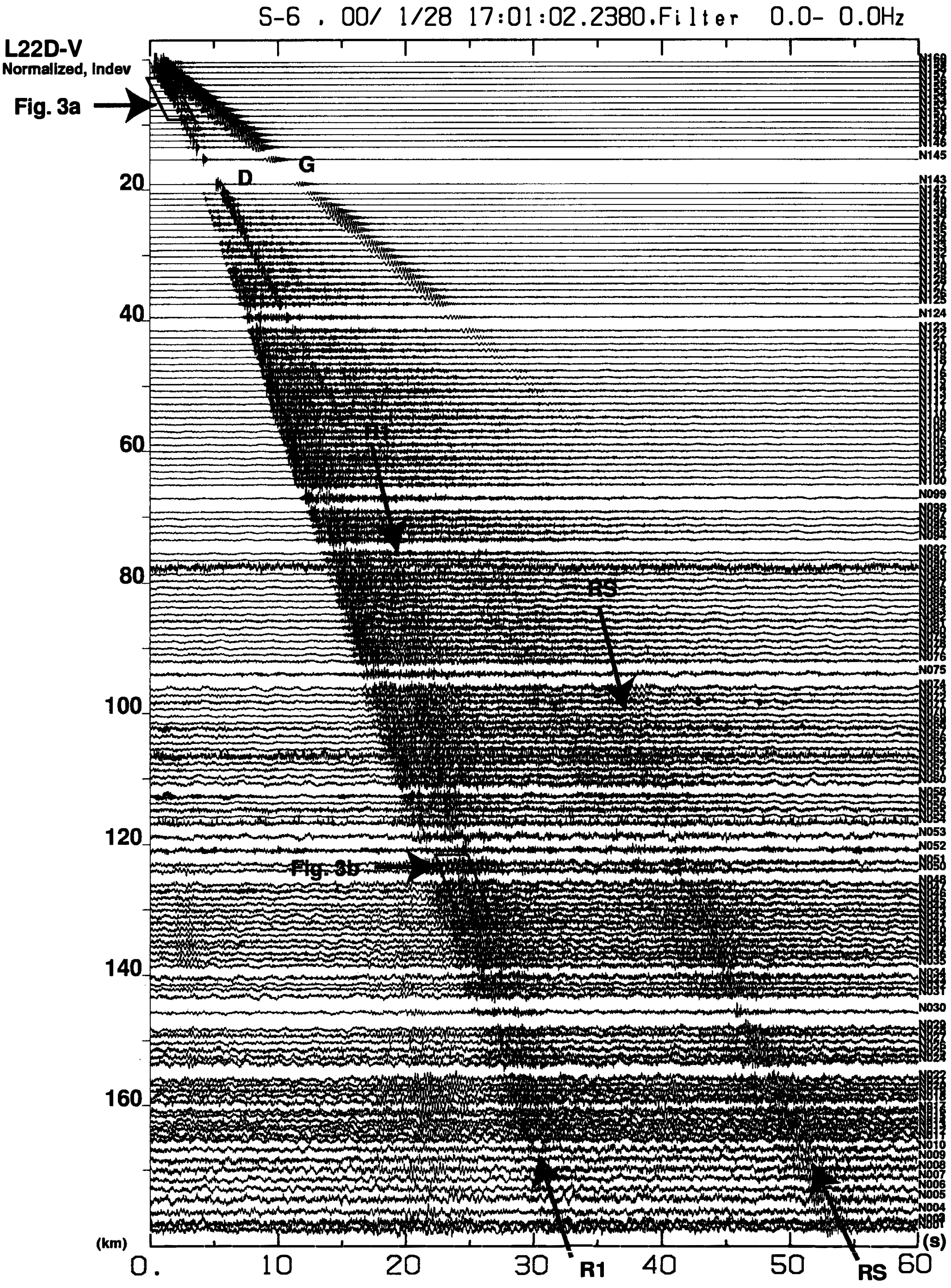

Fig. 2a. A shot record for the shot point S-6. Each trace are normalized by its maximum in the trace. Arrows mark prominent later arrivals R1 and RS. G and D mark surface wave phase and ice-sheet phase, respectively. Time windows for frequency analysis are also indicated with parallelograms.

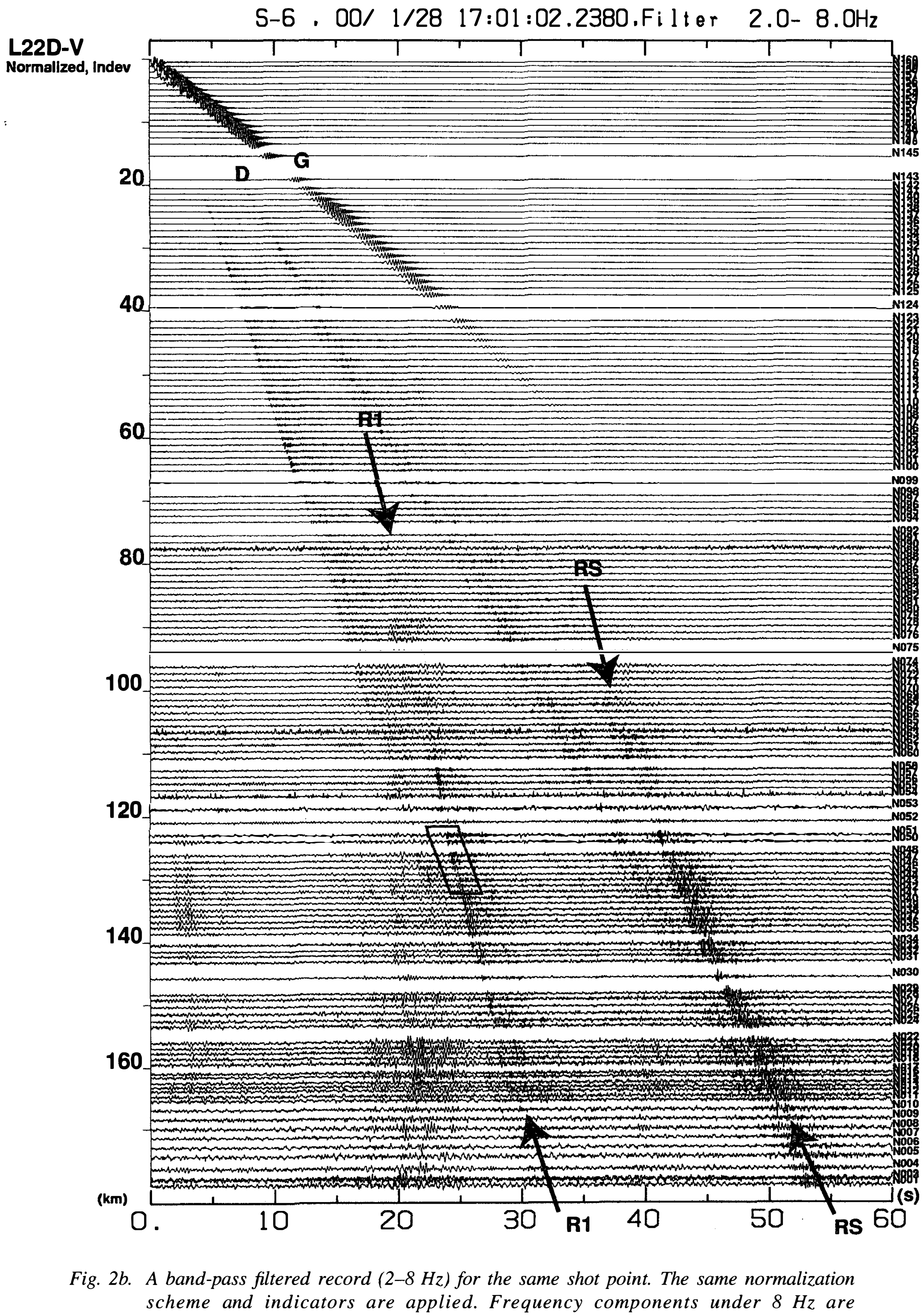

Fig. 2b. *A band-pass filtered record (2–8 Hz) for the same shot point. The same normalization scheme and indicators are applied. Frequency components under 8 Hz are predominant for prominent later arrivals. Markers are treated in the same manner as in previous examples.*

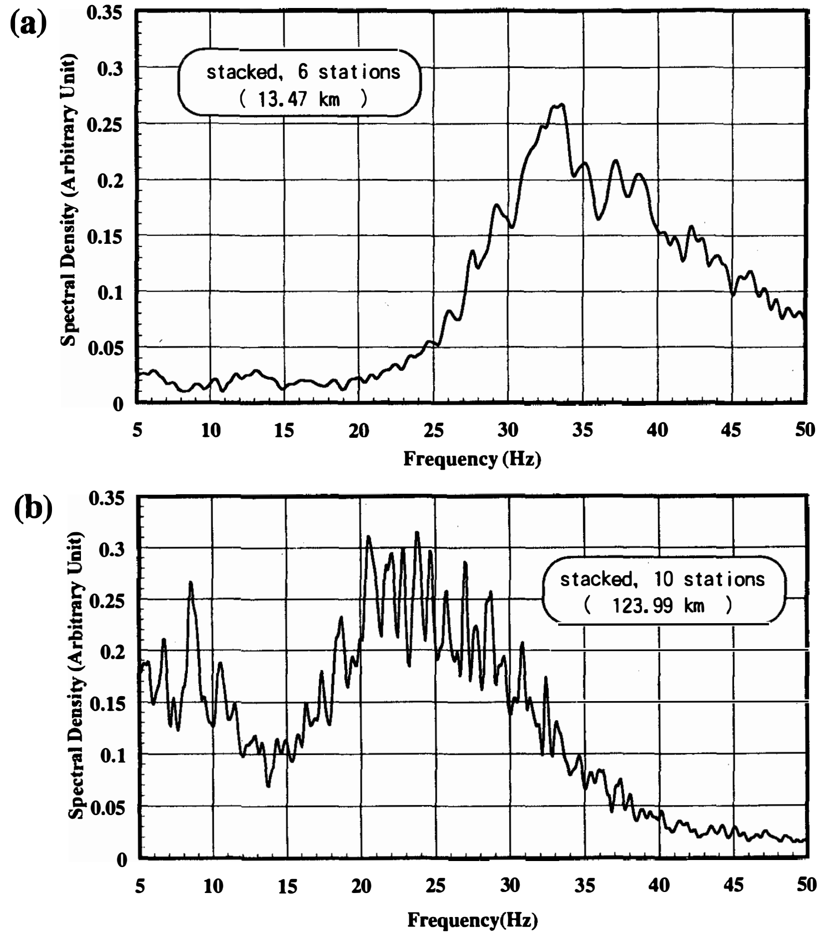

Fig. 3. a: A stacked RMS spectrum around the six first arrivals in the near field of the shot point S-6. A peak frequency is approximately 33 Hz. The mean distance of the stacked members is 13.47 km. b: A stacked RMS spectrum for ten 2 s windows involving the reflection R1. The mean distance of stacked members is 123.99 km. Two peaks at 8 Hz and 21 Hz are observed in the spectrum. Lower frequency components predominate more in the reflection phase R1 than in direct arrivals.

3. Analysis and results

A summary of the data processing used is given in Fig. 4. The data processing is based on conventional reflection analysis. Differences and some important points in this analysis are described as follows.

3.1. Static correction

An elevation variation along a spread causes travel time variations of a reflection. This operation adds the corrections for elevation changes to the travel times. The datum of the correction was chosen to be 2.0 km above the geoid surface of the WGS84 system in order to reduce the unknown factor. The correction time was derived assuming 3.8 km/s, which is an average velocity in the ice-sheet (Tsutsui *et al.*, 2001).

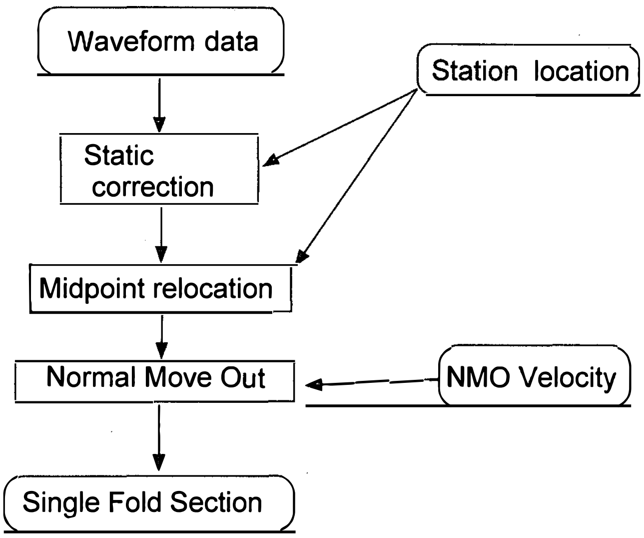

Fig. 4. *Data processing of the reflection analysis. The main processing schemes are based on those of conventional reflection seismology.*

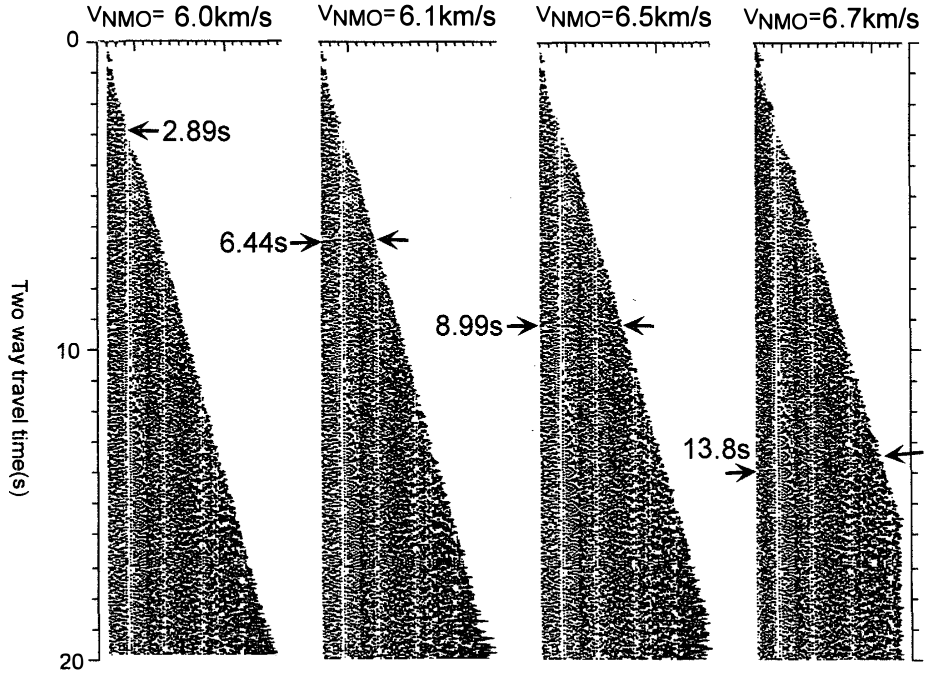

Fig. 5. *NMO trials with various uniform velocity functions for shot record S-6. Panels are ordered from left to right with increasing NMO velocity. An applied NMO velocity is displayed on the top of each panel. Aligned reflections are marked with arrows and their two way times are also denoted with markers. Waveforms are processed with Predictive Deconvolution (Robinson, 1954).*

3.2. Midpoint relocation (Depth point sorting)

In the case of a flat structure, reflection arose at the midpoint between a shot point and a geophone (Mayne, 1962). Seismic traces were relocated to their midpoint under the assumption of flat structure.

3.3. Normal Move Out correction

Normal Move Out (NMO) correction was applied to the relocated records. The NMO correction requires an estimate of the subsurface velocity from trial and error analysis.

An optimum average velocity should be derived for the NMO correction to construct the reflection profile. The optimum velocity yields a linear alignment of reflection events after NMO correction. However, the widely used method proposed by Taner and Koehler (1969) is not applicable in this case. There are too few traces for this method to work in the present case.

Optimum velocities were obtained through NMO correction trials with uniform trial velocities (Fig. 5). Assuming that the structure is smooth beneath the full spread of the JARE-41 seismic line, the optimum velocity gives continuous alignment of the reflection events. The assumption of smooth structure is not a bad assumption for this case, as indicated by the result of Ito and Kanao (1995). The NMO velocity functions obtained are plotted in Fig. 6a. Corresponding interval velocities are also shown, along with those of the refraction result of Ikami and Ito (1986) in Fig. 6b. A significant feature of the velocity structures obtained is varying two way time of the bottom reflection. This fact suggests a dipping bottom reflector.

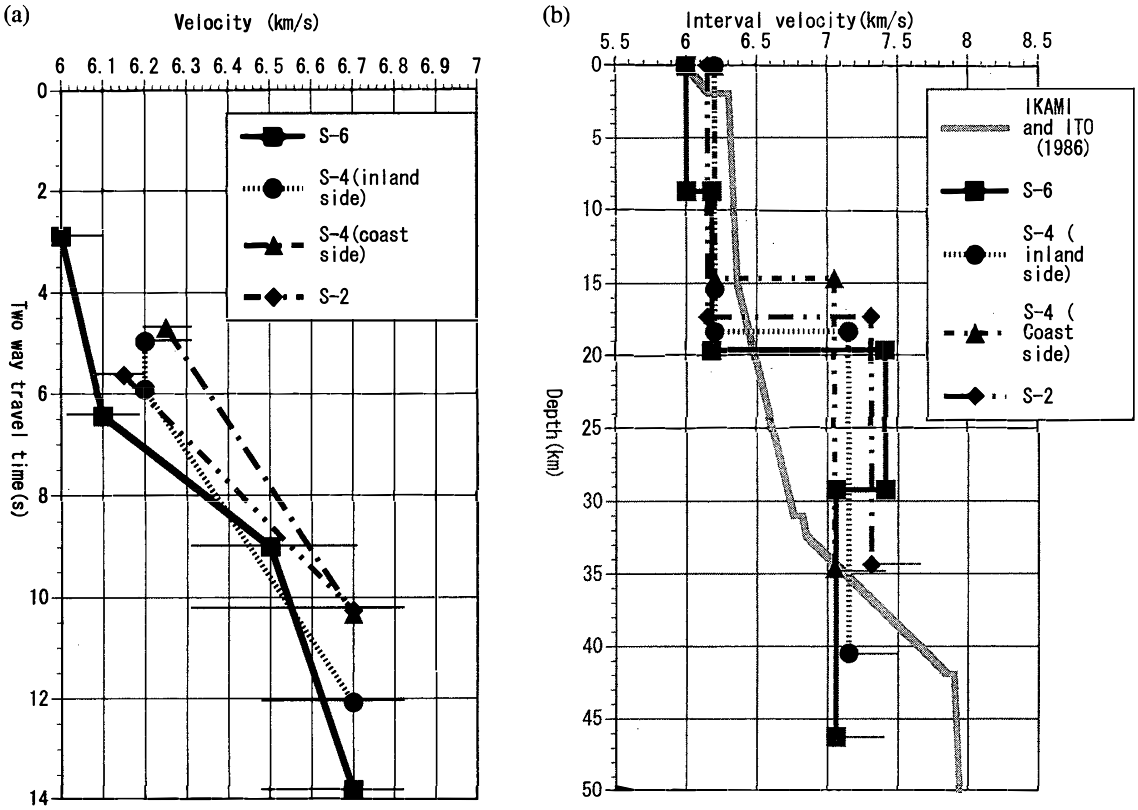

Fig. 6. Obtained velocity structures. The vertical axis describes normal two way time in seconds and the abscissa describes velocity in km/s. Solid boxes, solid circles, solid triangles and solid squares denote obtained optimum velocity at each depth for shot points S-6, S-4 (inland side), S-4 (coast side) and S-2, respectively. a: for NMO velocity function, with bars at markers which indicate the possible range of velocity picking; b: for interval velocities; and the thick gray line denotes a velocity structure from refraction analysis by Ikami and Ito (1984). Horizontal bars mark positions of the bottom reflection.

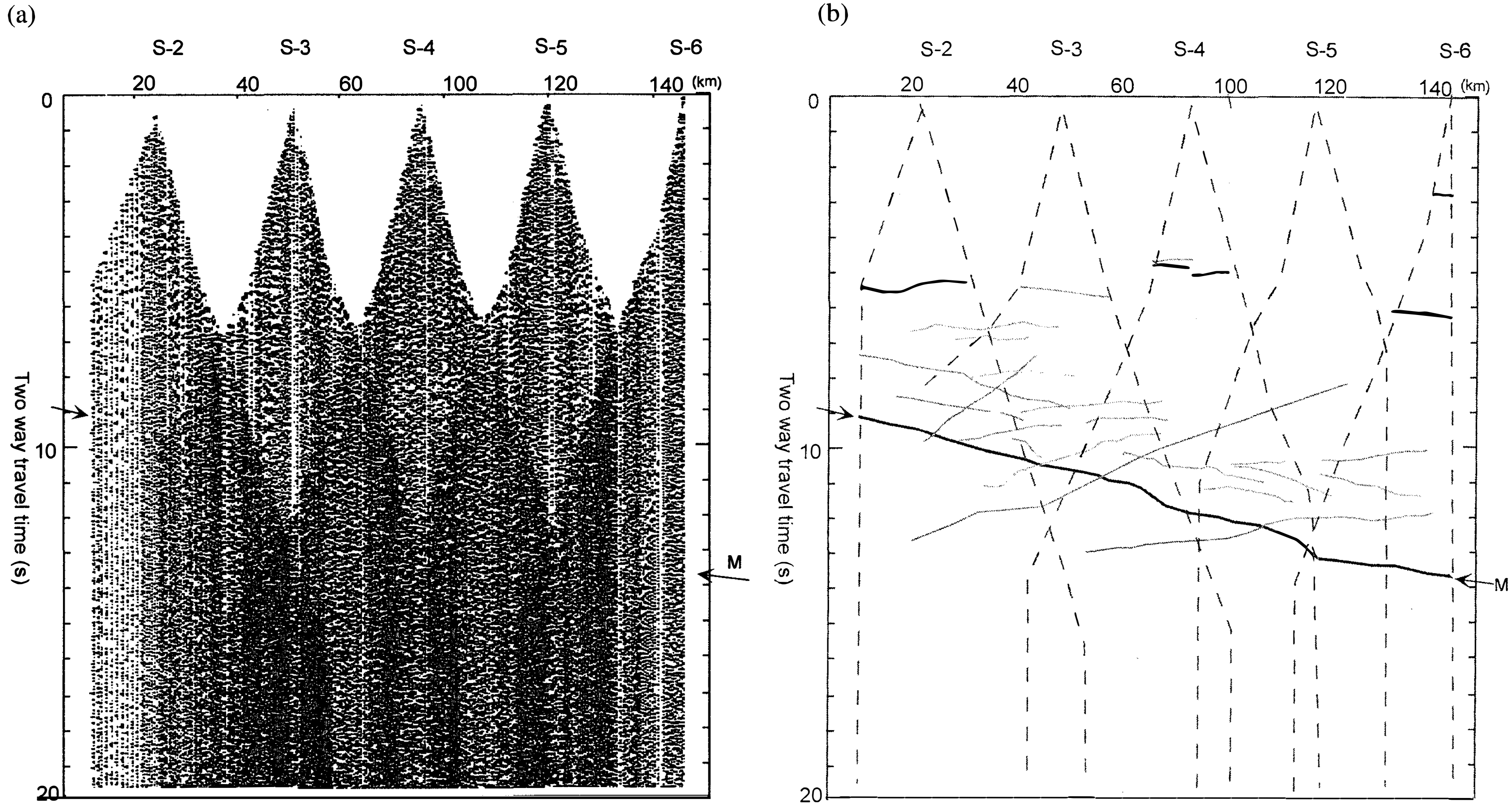

Fig. 7. a: A compiled single fold profile for all shots. The Moho reflection is marked with arrows and is labeled M. b: A distribution of interpreted reflectors that look continuous over a certain extent in the crust and uppermost mantle. Thick lines describe reflections which are identified through the NMO trials. This lines are reflectors which are coherent and are revealed as a result of the processing. Broken lines indicate shapes of each record used to construct the profile.

3.4. Single fold profile

Single fold profiles were produced through the processing described above. The single fold profile is essentially identical with the conventional time section in the seismic reflection method. The Common Mid Point (CMP) sorting in ordinary processing corresponds to the depth point sorting in this case. The main difference is that a relatively small number of traces were included in the depth point gathering in this processing, while typical reflection processing collects many more traces for each CMP. CMP stacking after NMO correction works as a velocity filter that can eliminate undesirable signals with non-reflection moveout. However, such enhancements are not effective in this case. Thus, converted waves are not eliminated in the single fold section. Surgical mutes that aim for rejection of surface waves and direct arrivals can actually be destructive to the appearance of the lowfold profile. This is one reason why a single fold profile can look much poorer.

The single fold profile is shown in Fig 7a. A single velocity structure, as obtained from shot record S-6, was applied in Fig. 7a. As described in Fig. 6a, no significant velocity variation is identified, so that the velocity function from S-6 is considered to be applicable to other shot records as the typical velocity function.

A continuous reflection M is observed in Fig. 7. This reflection M appears clearly over the full spread of the JARE-41 seismic line, with an inclination toward the coast. The reflection M, located at 13.8 s of normal two way time, corresponds to the clear later arrival R1 in Fig. 2b. The depth of the reflector M is estimated to be approximately 44 km at the shot point S-6 and approximately 31 km below the datum at the coast end of the profile (42 km at the shot point s-6 and 29 km at the coast end).

Several coherent reflections are traced and are shown in Fig. 7b. Numerous reflections with small lateral extents are concentrated in the range of 3 s before the reflection M and deepen inland. This layer is what was described as the reflectivity layer in Ito and Kanao (1995) because of its distribution.

4. Discussion

4.1. An optimum velocity structure

There are velocity structures from refraction studies by Ikami *et al.* (1984) and Ikami and Ito (1986) along the same line. The structure from Ikami and Ito (1986) is considered as a reference structure in this region. Interval velocity structures in Fig. 6b are deduced from NMO velocity functions in Fig. 6a by using DIX's formula (Dix, 1955) and are compared with the structure from Ikami and Ito (1986). The obtained velocity structures are somewhat lower than that of Ikami and Ito (1986) down to 15–19 km depth and become higher down to the Moho reflector. However, this is not a significant difference because NMO velocity is not very precise, as shown in Fig. 6a, and there are different bases for velocity determination.

4.2. A discontinuity in the neighborhood of the shot point S-4

Tsutsui *et al.* (2001) identified an undulation in 3.8 km/s layer thickness near shot point S-4 from first arrival analysis. The reflections in this study also indicate some kind of structural junction beneath S-4 (Fig. 7b). Different appearance and different inclination of the reflectors appear on both sides of this location. However, a velocity change at this discontinuity is not detected in this study. Although Tsutsui *et al.* (2001) report that there are no significant changes in the upper crustal velocity, a significant change in the middle to deep crustal structure is suggested in the single fold profile. This discontinuity may be some kind of a joint of geological units or complexes.

4.3. Dipping Moho reflector

Dipping reflectors toward inland in the deep crust are revealed in this study. This result is consistent with the result of a gravity survey by Kanao *et al.* (1994). However, Ito and Kanao (1995) show horizontal reflectors in the deep crust beneath the same region. This difference in reflector topography can be explained by the difference of the NMO velocity applied.

Ito and Kanao (1995) applied 6.3 km/s of NMO velocity in their processing. Although the velocity, 6.3 km/s, was derived from Ikami *et al* (1984), 6.7 km/s is concluded to be optimum for NMO correction in this study. 6.3 km/s of NMO velocity may cause over-correction. If 6.7 km/s is applied to the data set of Ito and Kanao (1995), a similar dip may be obtained for the deep crust reflector.

Moreover, the deep crust reflectors which terminate at the middle of their seismic lines are described in Ito and Kanao (1995). A span of their seismic line bounds an estimated extent of the reflectors because they did not have any data set in the distance range beyond 300 km from the shot point.

It is concluded that the differences described above are not the essential because reflector topography is strongly restricted by line geometry and the analysis parameter obtained from the previous study.

4.4. An interpretation of crustal reflections

As discussed in the previous sections, the crustal structure in LHB has some lateral and vertical variations. They are considered to be related to past regional tectonics, such as metamorphism in the late-Proterozoic to Paleozoic ages (*e.g.*, Hiroi *et al.*, 1991; Motoyoshi *et al.*, 1989). The last break-up of Gondwana, as in the Antarctica/Australia-India rifting (*e.g.*, Anderson, 1994; Storey, 1995), might have created crustal heterogeneity. From the crustal evolution viewpoint, LHB was under compressive stress perpendicular to the thermal axis in the NNW-SSE direction in the last metamorphism age of 500 Ma. Subsequent uplifting of uppermost mantle minerals, such as gabbro, into the crust within the extensional stress condition may have followed at the last breakup at 150 Ma. Seafloor spreading by magnetic anomalies and fracture zone trends south of 63S at this age took place in the WNW-ESE and NNE-SSW directions, respectively (Nogi *et al.*, 1992).

The origin of lower crustal reflectivity is considered from both geologic and geophysical data to have multi-genetic features such as igneous intrusions, lithologic and metamorphic layering, mylonite zones, anastomosing shear zones, seismic anisotropy and fluid layers (*e.g.*, Warner, 1990). Although the cause of lower crustal reflectivity has multi-genetic features, the metamorphic layering in the past regional tectonics might be the principal cause in LHB. Strong reflectivity may be expected from deep crust characterized by layered sequences of mafic and felsic rocks (Goff *et al.*, 1994).

Moreover, it is also created where mafic rocks are interlayered with upper amphibolite and lower granulite facies metapelites but diminishes when mafic rocks are interlayered with high-grade metapelites (Holliger *et al.*, 1993). In some regions, the above "primary" causes were enhanced by ductile stretching during the tectonic extensional process. The predominant reflected layers in the lower crust, in particular, are found in the thin-skinned tectonic areas. The reflective lower crust on the Mizuho Plateau may have also been enhanced under the extensional conditions of the last break-up of Gondwana. Deepening of the lower crustal reflectivity along the Mizuho routes, which are the main characteristic features in this crustal section, may be expected in the typical continental margin area. These dipping crustal structures toward the inland area may also have been affected by the tectonic stresses.

We can also identify the existence of seaward dipping reflections in the lower crustal depths on the Mizuho Plateau, in the entire seismic line. They might be one case of Seaward Dipping Reflector Sequences (SDRS) (Trehu and Wheeler, 1987) in the continental margin area of the past Gondwana super-continent. SDRS are observed in tentative 'Non-volcanic rifting margin' regions under tectonically extensional stress between hot-spot areas, such as the Marion–Kerguelen sea plateau. SDRS are considered to have originated in large igneous provinces (LIPs), which were produced by penetration of the asthenosphere into the thinning lithosphere during the extensional process when continental break-up occurred (Anderson, 1994). However, the actual origins of these reflections cannot be explained adequately by present geophysical and geological data around LHB.

5. Summary

A single fold profile and velocity structure for the JARE-41 seismic line is obtained from reflection analysis. Several clear reflectors are detected down to 14 s of normal two way time; some of these reflections can be traced over the entire JARE-41 seismic line. The reflector which was described in Ito and Kanao (1995) as the Moho discontinuity is also detected. Moreover the dipping Moho reflector is revealed from this study. The Moho reflector is located approximately 42 km beneath shot point S-6 and approximately 29 km below sea level beneath the coastal end of the profile. A reflectivity layer overlays the Moho reflector and the reflectivity layer is also dipping same manner as the overlaid reflector. The crustal velocity obtained in this study increases with depth, but lateral velocity variation in the upper crust in not detected. A discontinuity of the reflection character in the reflective layer is revealed in the neighborhood of shot point S-4. The structure obtained might be one case of Seaward Dipping Reflector Sequences (SDRS) (Trehu and Wheeler, 1987) in the continental margin area of the past Gondwana super-continent.

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