SEISMOLOGICAL REGIONALITY OF THE MIDDLE JAPAN TRENCH (FUKUSHIMA-OKI) REVEALED BY OCEAN BOTTOM SEISMOGRAPHY

海底地震観測による日本海溝中部（福島県沖）の地震学的地域性

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ABSTRACT

A series of intensive ocean bottom seismographic observations were conducted during 1982 and 1985 in the middle Japan Trench area (36.5 - 38 ^{o}N; Fukushima-Oki) to study the regionality of the upper-lithospheric structure and the hypocenter distribution in the trench region. A large amount of continuous magnetic tape records obtained from the observations were processed with a specially designed computer-aided system.

The seismic refraction profiling performed on the landward slope is 200 km long and parallels the trench at 3 - 4 km water depth (60 - 70 km from the trench axis). Records of 76 explosive shots by 8 OBSs (ocean bottom seismometers) on the profile were analyzed through a two-dimensional ray-tracing. The profiling revealed a southward-thickening crust, in which its upper part has V_p = 5 - 6 km/s and its lower part is about 6.8 km/s. The Moho is 21 - 23.5 km deep and its V_p = 8.2 - 8.3 km/s. This result is supported by existing gravity data. It also suggests that the dip of subduction in the shallow part steepens southward along the Japan Trench by an angle of several degrees at around 38 ^{o}N. The 6.8 km/s layer seems to have rather uniform thickness (5 - 6 km) throughout the length of the Japan Trench. This layer is regarded as a part of the subducting oceanic plate.

Two earthquake observations with 5- and 6-OBS arrays which covered the landward side of the trench located about 200 local events with errors less than 10 km in the total period of about 50 days. The subducting lithosphere slab is almost aseismic up to 80 km landward from the trench axis in Fukushima-Oki. The upper plate also shows low seismicity within 50 km from the trench axis, while it has high seismicity farther landward. These features form a striking contrast to the situation of Sanriku-Oki region where foci are distributed rather uniformly over the trench region.

The southward-deepening of the subducting Pacific plate along the Japan Trench below the inner slope seems to have been prepared beforehand as indicated by several hundred meters of water-depth contrast lasting eastward as long as 1000 km on the northwest Pacific Ocean floor. The trenchward low seismicity in Fukushima-Oki is interpreted as an indication of weak mechanical interaction (coupling) between the two plates due to the lateral deepening of the subducting plate along the trench.

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1 INTRODUCTION

1.1 GENERAL

The northern Honshu arc (or the northeastern Japan arc) is one of the most intensively studied subduction zones of the earth. Seismologically, its land area is covered with sensitive microearthquake networks and its sea area has been studied through repeated temporary OBS (ocean bottom seismograph) observations. This arc is therefore often referred to as an example of a well-investigated arc-trench system. However, the well-studied zone is restricted to its northern part because the most of the intensive OBS works have been concentrated to this part. As one of the farther steps in the refinement of the plate tectonics, studies of regional variations within a subduction zone, as well as the comparative studies of subduction zones, should become important.

Among many geological, geophysical and geochemical means to investigate subduction zones, seismological ones offer some of the most direct approaches to understand the mechanical process of subduction; for example, seismic exploration with controlled sources gives the highest resolution of crustal or lithospheric structure; earthquakes in a subduction zone represent plate motions and present mechanical state of the relevant region.

Although it is known from the land earthquake networks that the seismicity along the Japan Trench shows certain regionality, debates on the subject have been difficult for the lack of definite seismological data, namely the lithospheric structure, precise hypocenter distribution and focal mechanism.

Since seismic methods utilize seismic waves propagated through or from the region of interest, geometrical proximity to the target is essentially important to gain the resolution of observation. Therefore seismological studies of trench area is most effectively made by using OBSs. OBSs have been playing a central role in marine refraction experiments and earthquake observations. Precise hypocenter distributions in sea area including subduction zones have been obtained by many researchers and it has been shown that detectability and locatability of earthquakes are greatly improved by the use of an OBS array (e.g. Yamada, 1980; Hirata *et al.*, 1983). In refraction experiments, OBSs offer better signal-to-noise ratio than hydrophones (Asada and Shimamura, 1976).

The present study deals with such subjects through intensive OBS observations in the middle Japan Trench (Fukushima-Oki) area. In Part 1, geophysical settings are reviewed with reference to the past studies. Part 2 stresses the technical aspect of OBS observations and describes the instruments used in the experiments. A data playback system newly developed for processing of data from the OBSs are presented in Part 3. In Part 4, the field and data reduction procedures of OBS observations conducted in 1982 and 1985 are described. The data are analyzed and results are derived in Part 5. The seismo-tectonic regionality along the Japan Trench is discussed on the basis of the results in Part 6.

1.2 GEOPHYSICAL SETTING

NORTHERN HONSHU ARC AND ITS SEISMOLOGICAL REGIONALITY

The Japan Trench is situated at the northwestern rim of the Pacific Ocean, connecting with the Kurile Trench at its north end and with the Izu-Bonin Trench and the Sagami Trough at its south end. Here, we divide conventionally the 700 km long Japan Trench into three parts (**Fig. 1.1**); the northern part or Sanriku-Oki region (38 - 41 ^{o}N), the middle part or Fukushima-Oki region (36.5 - 38 ^{o}N), and the southern part or Boso-Oki region (34.5 - 36.5 ^{o}N). The trench axis changes its curvature slightly at the boundary between Sanriku-Oki and Fukushima-Oki regions. Since the seismicity of Boso-Oki region is made complicated by the subduction of the Philippine Sea plate at the Sagami Trough, it is not appropriate to simply compare its seismicity with those of the other two regions.

Figure 1.2 shows the seismicity of large earthquakes ($M_S \geq 7.4$) along the Japan Trench during the last 100 years compiled by Kawakatsu and Seno (1983). In Fukushima-Oki area, the 1896 M_S = 7.4 event, the 1905 M_S = 7.8 event and the most recent five events with M_S > 7 in 1938 are known to have occurred (Utsu, 1979; Abe, 1977).

Kawakatsu and Seno (1983) studied the regional variation of seismicity along the Japan Trench using the ISC data by determining focal depths from *pP-P* time intervals with the *pwP* phase correction after Yoshii(1977). Accuracy of the focal depths were somewhat improved by this correction, however, earthquakes of m_b < 4.0 could not be treated. They proposed to separate the zone of low angle thrust earthquakes at the plate interface into two; the shallow thrust zone (0 - 40 km deep) and the deep thrust zone (40 - 60 km deep). These two zones correspond approximately to the areas beneath the seaward and landward halves of the landward trench slope (between the trench axis and the coast), respectively. They characterized the seismicity of Fukushima-Oki by less activity in the shallow thrust zone than those of the other regions and periodical occurrence of large earthquakes in the deep thrust zone.

The distribution of small earthquakes monitored with a land microearthquake network is also in accordance with such a regional variation of seismicity as shown in **Fig. 1.3** (Faculty of Science, Tohoku University, 1983). In the figure, epicenters are concentrated landward in Fukushima-Oki, while in Sanriku-Oki, they are distributed more homogeneously below the area landward the trench. Although studies on such regional variation of seismicity need precise hypocenter distribution and seismic velocity structure of the relevant area, existing data regarding them are too little for Fukushima-Oki; land-based microearthquake networks can hardly resolve the focal depth in the trench area, i.e. far off the coast.

In the following two sections, we review the past studies on seismicity and structure by OBSs in Fukushima-Oki and its adjacent regions.

PAST STUDIES ON SEISMICITY BY OBS'S

Hypocenters in offshore areas located by land seismographic networks have inherent errors which are especially large for the focal depths. Since these errors are caused by improper geometry of the observation sites relative to foci, they can be reduced drastically by observing earthquakes with an OBS array which covers the epicentral area. Yamada (1980) determined the aftershock distribution of the 1978 Miyagi-Oki earthquake occurred in the southern Sanriku-Oki. Nagumo *et al.* (1976 and 1984) and Hirata *et al.* (1983 and 1985) have conducted microearthquake observations in northern Sanriku-Oki (39 - 41 ^{o}N) with OBS arrays consisting of three to eight OBSs and obtained detailed microseismicity of both landward and seaward the trench. These observations have produced precise microearthquake distributions unobtainable by land observations. The results indicate that the seismicity of the trench area is mostly in the upper part of the lithosphere and it smoothly continues to the upper plane of the double seismic zone. Shallow seismicity seaward the trench axis and the existence of a narrow seismicity gap below the inner trench wall in the northern Sanriku-Oki also have been confirmed from the results.

On the other hand, there has never been an OBS observation for seismicity in Fukushima-Oki (nor Boso-Oki). Therefore, seismicity surveys in these regions with OBS arrays are necessary for comparative studies of seismological regionality along the Japan Trench based on accurate hypocenter distribution of microearthquakes. In this study, we carried out OBS observations in Fukushima-Oki.

PAST STUDIES ON STRUCTURE

Marine seismic surveys of the middle and the southern Japan Trench regions are not many. In Fukushima-Oki, a single- and a multi-channel seismic reflection profilings have been made across the trench by Murauchi and Asanuma (1970) and Sakurai *et al.* (1981), respectively. The latter reported on a 270-km-long profile which ran across the trench and through the site of OBS S2 of the present study. Suyehiro *et al.* (1984) performed a 40-km-long refraction profiling with an airgun-OBS system at 50 km southwest of OBS S17 of the present study where water depth is 200 - 300 m. No refraction survey has been made farther offshore in Fukushima-Oki, so that little has been known about its lithospheric velocity structure.

In Boso-Oki, Kanazawa *et al.* (1985) have made an explosion-OBS refraction profiling at a water depth of 3 km on the landward trench wall paralleling the trench (**Fig. 1.4**). They preliminarily reported that the Moho in Boso-Oki is significantly deeper (23 km) than that in the corresponding profile of Sanriku-Oki.

Structure of Sanriku-Oki has been investigated fairly well. Crustal and uppermost-mantle structure seaward the trench and the upper-crustal structure landward the trench have been obtained by Ludwig *et al.* (1966), Nagumo *et al.* (1980), Murauchi and Ludwig (1980) and Suyehiro *et al.* (1985) in the profiles paralleling the trench. Down-to-sub-Moho structure of the inner trench slope of Sanriku-Oki has been studied through explosion-OBS refraction experiments by Asano *et al.* (1981) and Kanazawa *et al.* (1985). The profile of the former is approximately along the isobath of

1500 m, and the latter have preliminarily reported on the profile along the 3 and 5 km isobaths. The positions of these profiles are shown in **Fig. 1.4** and a structural section across the trench compiled by Kanazawa et al. (1985) are shown in **Fig. 1.5.** Besides these refraction experiments, several multichannel seismic profiles have been conducted across the trench in Sanriku-Oki (e.g. Nasu *et al.*, 1980). These marine seismic data, combined with the well-defined seismicity and structure obtained for land area, have made Sanriku-Oki area one of seismologically the most understood subduction zones of the earth. Other geophysical data have also been obtained in detail for the northern Honshu arc (e.g. Yoshii, 1977).

We have conducted an OBS-refraction profiling on the inner trench slope of Fukushima-Oki along the water depth of 3 - 4 km where no structural survey had been made. The crust and uppermantle structure landward the trench is important for accurate hypocenter location in that area, and is also important in discussing the subduction process because it includes the descending plate underlying the seaward edge of the overlying plate. This study's survey enabled the comparison of the lithospheric structure and detailed hypocenter distribution along the Japan Trench.

2 INSTRUMENTATION

2.1 OBS AS A TOOL

Although many OBS experiments have been conducted so far, it is not yet routine work and the number of research groups at successful OBS work is still not many. This is not merely because of the costliness to prepare many OBSs and hire a ship for weeks, but because OBS work requires many lines of know-how of such as building deep-sea instruments, settling and recovering them, shooting explosives at sea, and processing vast amount of off-line data. This know-how is acquired through repeated experiments. For example, since sophisticated OBSs are not commercially available, usually a research group must build its own instruments whose specifications would limit the time and place of observation, the quantity and quality of records, and therefore, the scientific results to be obtained. Any research through a seismographic observation at sea includes above-mentioned scientific, technological and engineering efforts as integral parts.

The OBSs used in the present study record incoming signals continuously for several weeks. Noise characteristics of OBS sites can not be known beforehand and it is often experienced that sudden noise bursts of unknown cause (probably due to biological activities) have been frequently recorded by some OBSs. With a limited data storage capacity, it is difficult for an OBS of the event trigger type to record all the desired seismic events (earthquakes and/or shots) among noise. To record during preprogrammed time windows (timed trigger) is also possible in controlled source studies, but marine experiments would seldom go on schedule because of bad weather or some accident.

In the continuous recording mode, ground motion signals during weeks of observation period have been wholly copied on a magnetic tape with a limited frequency band width and a limited dynamic range. Therefore the processing of recovered data tape includes event detection, waveform filing and visualization, which would be performed on-line in case of land seismographic networks. Besides, recent improvements in OBS, especially the development of small-size, free-fall-pop-up OBSs (e.g. Johnson *et al.*, 1977; Kasahara *et al.*, 1979; Yamada *et al.*, 1981), have enabled the use of 10 - 20 OBSs in one experiment by a single ship. Such an experiment supplies larger amount of data (corresponding amount on the order of 10^{11} bits) in comparison with the conventional fewer OBS experiments. Processing of such amount of data is hardly conceivable without some automated tools. A system of such tools which greatly facilitates the processing of a large amount of OBS data will be described in Part 3.

2.2 RELEASING MODES

OBS TYPES

Basically two types of OBS were used in this study; a timed-release type and an acoustic-release type. Both are free-fall pop-up types. The timed-release type is compact and low-cost in comparison with the acoustic-release one, whereas the time schedule of recovery must be determined before launch and it constrains the experiment. To the contrary, although the acoustic-release type is large in size and expensive, it has the big advantage that it can be recalled at any time. The specifications of the sensor, recorder and clock are common to both types.

TIMED-RELEASE OBS

The timed-release type is that described by Yamada *et al.* (1981). It has a glass sphere of 43 cm diameter as a housing and buoyancy device, which contains geophones with gimbal mechanism, recording instruments and batteries. The glass sphere is armed with a plastic hard-hat, to the outside of which a radio beacon, a flashing light, a release timer (Urabe and Kanazawa, 1984) and anchors with release mechanism are attached. The size of the whole unit is 70 cm cubic and it weighs 80 kg in air. The timer releases the anchor by electrolyzing stainless-steel connecting plates to pop up the OBS.

ACOUSTIC-RELEASE OBS

There are several different versions of the acoustic-release type OBSs which were used in the experiment (Shimamura and Yamada, 1982). Basically they consist of a transponder and a seismometer units connected each other with a 100 m-long rope. The seismometer unit contains geophones with gimbals, recording instruments and batteries in an aluminum cylindrical pressure case which is 120 cm long and 20 cm in diameter. The seismometer unit weighs 50 kg in air. The transponder unit consists of a floatation system (glass spheres), finding instruments (a radio beacon and a flashing light), an acoustic transponder and an anchor. The transponder (Sonatech Inc.) is capable of answering to calls from sea surface unit and releasing a hook that holds the anchor on command. The floating system, when the anchor is released, can surface the transponder unit as well as the sensor unit.

2.3 SENSOR AND RECORDING SYSTEM

All OBSs are equipped with a vertical and a horizontal geophones which are mounted on gimbal mechanisms to keep level. The natural frequencies of the geophones are 2 or 3 Hz for the vertical and 4.5 Hz for the horizontal components.

Seismic signals are amplified and recorded on a Philips-type cassette tape by a ultra-slow-speed direct analog tape recorder (Shimamura and Asada, 1974; Yamada *et al.*, 1976; Yamada, 1980). In a chrome-type C-90 tape for high-fidelity audio recording running at a speed of 0.05 - 0.13 mm/s, seismic signals are recorded continuously for 11 - 25 days. The frequency response of the recorder is flat up to 30 Hz (-3 dB point) in case of 11-day recording, and it is inversely proportional to the recording period and is proportional to the tape speed. The recorder has four tracks on a tape, therefore it has four recording channels. One of them is assigned to record time code signals from a clock which is described later. The other three channels are used to record ground motion signals. The vertical component signals are recorded into two channels with amplifications normally different by 34 dB. The horizontal component signal is recorded in the remaining channel. The dynamic range of each recording channel is 40 dB.

Exceptionally for three OBSs used in 1985 (OBSs PR3, PR5 and PR7), vertical and horizontal components had two individual tape recorders in order to widen the dynamic range as large as 86 dB by using three recording channels with different amplifications for each component (Urabe and Ohmi, 1985). This configuration realized the dynamic range of the horizontal component as wide as that of the vertical component (see **Fig. 4.15**).

In 1985 experiment, sine waves and the time code signals were recorded in all seismic signal channels of the recorders before and after the operations at the sea bottom. These signals were used in calibrating amplitudes and correcting time for channel-to-channel skew.

2.4 CLOCK

The OBS clock is a time code generator (TCG) designed by Inatani and Furuya (1980) or its modified version (Urabe and Ohmi, 1985). The TCG generates a BCD (binary coded decimal)-coded clock signal in which day, hour, minute and system identification number are encoded in each one-minute-long frame by modulating the widths of second pulses. Its stability over a temperature range of 0 - 50 ^{o}C is 5 x 10^{-6} for the 1985 OBSs PR2, PR4 and PR6, and 5 x 10^{-7} for the others. The clocks were calibrated immediately before and after the OBS operations with the standard time (JJY radio signal) by a master clock (Shimamura, 1977). In consideration of the small temperature variation at the deep sea bottom and the stability of the tape path, the timing accuracy of seismic signals at playback is estimated to be better than 0.05 s throughout the observations.

2.5 POSITIONING

Positionings of ships were made by using the Navy Navigation Satellite System (NNSS) and Loran-C. In 1985 experiment, the positioning data from the navigation system were registered into floppy disks every minute together with water depth data from a depth sounder. The relative and absolute accuracies of the positionings are estimated to be 0.2 and 1.0 km, respectively. The ship position where each OBS was launched was adopted as the corresponding OBS position.

For the explosion OBSs in 1985 experiment, their positions were so corrected by using records from nearby shots that arrival times of direct water waves were consistent with the relative positions of the shots and the OBS on the profile.

3 DATA PLAYBACK SYSTEM

3.1 INTRODUCTION

REQUIREMENTS FOR AUTOMATIC DATA PLAYBACK SYSTEM

As stated in the previous part, we have adopted the continuous recording mode for our OBSs. To record ground motion with a frequency band of microearthquake and explosion signals (typically 3 - 20 Hz) continuously for more than a few days, the direct analog recording (DAR) on magnetic tape (e.g. Sacks, 1966; Shimamura and Asada, 1974) is the only possible method, although data quality is generally higher in digital recording or frequency modulation method than in the DAR method.

There are two special points in the playback procedure of the continuously recorded DAR data. One is that desired seismic events (earthquakes of interest or shots) must be cut out from the continuous record of the whole observation period, and the other is that the playback speed is several hundred times faster than the recording speed. The former is due to the inherency of the continuous recording, and the latter is due to an ultra slow tape speed at recording. Therefore a playback system for DAR data must be capable of detecting, cutting out and filing seismic events from continuous signal coming at a very fast rate.

CHARACTERISTICS OF DAR DATA AT PLAYBACK

Data on a cassette tape recorded by OBS is firstly copied to a reel-to-reel tape (1/4-inch wide), where the speed of the cassette tape on playback is 4.8 cm/s and the recording tape speed is 19 cm/s. This duplication is carried out in order to protect the original tape from the deterioration through repeated playbacks and to halve the time base by playing back the copy tape at 9.5 cm/s speed. Since the reel-to-reel tape recorder has a larger dynamic range and wider frequency response than the cassette tape recorder, the degradation of signal quality due to the duplication is negligibly small. In case that thus a duplicated 25-day-recorded tape is reproduced at 9.5 cm/s speed, the data rate is 400 times faster than that at recording time, so that signal frequencies are multiplied by 400.

FORMERLY USED PROCEDURE

Formerly used playback systems of DAR tape were based on analog data processing. In such systems (e.g. Moriya and Takeda, 1979; Yamada, 1980; Kasahara, 1981), earthquakes were detected by hearing seismic signals which is audible for its fast speed, and shots were searched according to time read with a time code reader. Records of events thus found were visualized on paper by manually triggering a fast strip-chart recorder such as an optical or ink-jet oscillograph. These playback methods, which had been the only possible ones until digital computers with a high speed A/D (analog-to-digital) converter became available, have two weak points. One is that they are primarily time-consuming manual labors so that they do not suit for data from long-term observations utilizing more OBSs and shots (see Part 2). The other is that they are incompatible with further data processings using digital computers.

NEW PROCEDURE

Urabe and Hirata (1984) developed a computer-aided digital processing system for DAR data from microearthquake OBS networks, which utilizes a high-speed (100 kHz) A/D converter and is free from above-mentioned weak points. The system automatically searches for seismic events from continuous OBS records and then creates digital waveform files of the events.

In this study, Urabe and Hirata (1984)'s system was used in the playback of the first OBS experiment in 1982. For the second experiment in 1985, the system was improved with respect to event discrimination and visualization and then used.

3.2 URABE AND HIRATA (1984)'S SYSTEM

GENERAL

In this section we review Urabe and Hirata (1984)'s DAR data playback system. This system has two special points. One is that detection of earthquakes and filing of their waveform data in digital form are performed through two respective passes to save data storage capacity; whole continuous data digitized at a low sampling rate are scanned at first, and then seismic events are digitized at a high sampling rate according to a time schedule table. The other special point is that station-multiplexed records in which records from all stations are arranged on a common time base are finally obtained; we refer to this record format as the multi-station format hereafter. The use of this format greatly facilitates the work of picking seismic phases and reduces mistakes in it because in that format seismic phases can be identified and picked in relation to other stations' records (e.g. Lee and Stewart, 1981).

By this system, digital and paper-transcribed multi-station records of earthquakes are produced from DAR tapes of an OBS array automatically and very rapidly with a little intervention by an operator. Although it is primarily for an microearthquake network data, filing of waveform data from controlled source experiments is also possible with a part of this system.

HARDWARE AND PROGRAMMING ENVIRONMENT

For data processing, a minicomputer system (S/140; Data General Corp.) with fast analog input/output facilities is mainly used and the computer center of the University of Tokyo (HITAC M280H) is partly used.

The block diagram of the minicomputer system is shown in Fig. 3.1. In this figure, the 147 Mbyte disk, the 1280 x 1024 display and the X-Y plotter are added for the improved system described in the next section. The 16-bit CPU has a floating point co-processor and 256 Kbyte read/write memory. The 73 Mbyte disk drive and the tape unit with a 1600 bits/inch recording density and a 1200 ft maximum tape length are external data storage devices. The tape unit reads or writes 20 Mbytes of data in 10 minutes. The CRT character display and a keyboard is used as a system console. The serial printer has a speed of 180 characters/s.

The A/D converter is capable of digitizing up to 16 channels of analog signals with a 12 bit resolution at a total sampling rate of 100 kHz, and at this rate the data can be stored into the disk without omission. The four-channel analog tape deck runs a 1/4 inch-wide reel-to-reel tape for playing back analog data. The tape deck has two selectable tape speeds (9.5 and 19 cm/s). The signal conditioning unit intervenes between the tape deck and the A/D converter. This unit has the selectable functions of amplifying signals, taking signal's single-side envelope by rectification for a seismic channel, and restoring waveform of the clock signal into its original form (pulse-width-modulated square waves).

The D/A (digital-to-analog) converter outputs up to 10 channels of data with a resolution of 12 bits. The output signals are transcribed on paper by the eight-channel thermal strip-chart recorder. The frequency response of the recorder is DC to 120 Hz, and it has an additional marker channel.

The computer center is used in the process of station multiplexing which requires larger disk storage capacity than the minicomputer. In that case, data are transferred by magnetic tapes.

All programs are written in FORTRAN. The area for waveform data in the disk consists of 15 contiguously-organized files with a total capacity of 64 Mbytes. In order to realize a data transfer throughput of 100 KHz, these files are so created that they contain no bad blocks. All these files organize one large virtual file, in which any waveform data created in the course of data processing are stored and ordered by 8 Kbyte-long block numbers.

FLOW OF DATA PROCESSING

Data from a microearthquake network of OBSs are processed as follows. In quotation marks are names of programs hereafter.

(1) The first A/D conversion for event detection: Signals of one seismic channel (normally the vertical low-gain) and clock from analog tape are wholly digitized at about 15 Hz in seismic time and stored into disk by "ADCBULK". The low sampling rate is chosen in order to store as long data as possible at one time; at this rate, 12-day-long data can be stored into a 65 Mbyte data file. To accommodate to the low sampling rate, the seismic signal is rectified and the clock signal is restored into square waves by the signal conditioning unit before the A/D conversion (see **Figs. 3.2 and 3.3**).

(2) Event detection: The rectified seismic signal in disk is scanned to detect seismic events by "EVDET". Ratio of short term average (*STA*) to long term average (*LTA*) of signal amplitude (e.g. Ambuter and Solomon, 1974) is used to judge seismic events. Detected events are listed in a disk file as shown in **Fig. 3.4A**.

After procedures (1) and (2) are applied to data from all stations;

(3) Scheduling for waveform filing: a time schedule table according to which waveform data will be digitized is made by "CAUTO" by compiling the event lists obtained for all stations of the OBS array. The table, hereafter we call the filing schedule table, is commonly applicable to all stations (**Fig. 3.4B**). Method of compiling event lists are illustrated in **Fig. 3.5**.

(4) **The second A/D conversion for waveform filing:** Waveform data are digitized from a continuously running analog tape by "AUTO" according to the filing schedule table. The sampling rate is 25 KHz in real-time for each of four channels at 9.5 cm/s playback speed. This rate corresponds to 60 - 150 Hz in seismic time. A record header is attached to each digitized record (**Fig. 3.4C**).

(5) **Resampling**: Since sampling rate of data digitized by "AUTO" fluctuates normally by a few percent due to fluctuation of tape speeds at recording and playback, the rate is accurately unified by "RESAMP" by numerical interpolation with reference to the clock signal.

After procedures (4) and (5) are applied for all stations;

(6) **Stations multiplexing:** Sampling-rate-unified records of all stations in the network are compiled into the multi-station format by "RAG" in the computer center. Time corrections for clock drifts and for inter-channel skew due to misalignment of recording and playback heads are applied at the same time.

(7) **Visualization:** Multi-station records are transcribed on paper by "DAMULT" normally with a 10 mm/s time scale by the eight-channel strip-chart recorder via the D/A converter (**Fig. 4.3**).

Seismic phases are then read manually from the multi-station paper records by means of a tablet digitizer.

PERFORMANCE OF THE SYSTEM

This processing system greatly facilitates and quickens playback of a large amount of DAR seismic data. For example, if this system is applied for data from a 25-day microearthquake observation with five OBSs, time required to produce multi-station records, which are unobtainable by conventional analog processing methods, is on the order of only 50 hours.

3.3 THE IMPROVED VERSION

GENERAL

The main improvement of the system is addition of an event discrimination procedure after the event detection and before the waveform filing. Efficient processing of local earthquakes is realized by inspecting multi-station time-compressed records and reducing the number of final multi-station records. Other improvements are in visualizations of multi-station records and refraction experiment records. By the use of an X-Y plotter or a laser printer, multi-station records of up to 27 channels and record sections of refraction experiments can be transcribed on paper.

For the improved system, a 147 Mbyte disk, a color graphic display with a 1280 x 1024 resolution connected to the computer via a fast line (307 Kbits/s) and an X-Y plotter with automatic paper feeder and maximum paper size of 420 mm x 300 mm were added. Total storage capacity of disk became 220 Mbytes, out of which 190 Mbytes are allocated to store waveform data in 51 files (23,900 of 8 Kbyte-long blocks). The computer center of Kyushu

University (FACOM M380) is also used to utilize its strip-chart X-Y plotters and laser printers.

Flow of the improved processing system is shown in **Fig. 3.6**.

ADDITIONAL DISCRIMINATION OF LOCAL EARTHQUAKES

In typical microearthquake observations, the total duration time of records of local earthquakes amounts on the order of one percent of the observation period. It corresponds to about 80 Mbytes of multi-station records for a 25-day observation with six stations, and this amount is near the upper-limit below which data are efficiently dealt with by the use of the minicomputer system for its limitation in data storage capacity. Unfortunately, it is difficult to select out only local events of interest by Urabe and Hirata (1984)'s method for the following cause. Some additional operation of event discrimination and data reduction is necessary to process local events efficiently.

Urabe and Hirata (1984)'s method of event discrimination by program "CAUTO" (**Fig. 3.5**) works well in omitting local noise bursts from earthquake events by setting a large N_{min} value (see **Fig. 3.5**). However, the events thus listed up contain not only local earthquakes of interest, but also many distant quakes and their *T* phases. Their total duration time often amounts several times that of the local quakes. Besides, if we want to analyze smaller local earthquakes which are detected at less stations, for example only one or two, a corresponding less N_{min} value must be set to list up those quakes in the filing schedule table. In such cases, many local noise bursts are also included in the table, and their total duration may be some ten times that of desired local quakes (about 800 MBytes, for example). It is unrealistic to store all these records in digital form with a full sampling rate.

In the improved system, a filing schedule table is made by setting a small N_{min} value such as one or two. According to the table, multi-station monitor records with a compressed time axis (1 mm/s) is transcribed on paper. Events of interest are discriminated from noise by inspecting the paper records manually.

It is often difficult to discriminate local events from other events or noise bursts by inspecting a single-station OBS record, even if it is done manually. For example; records' amplitude sometimes saturates so that *S* phases can not be identified; *T* phases are apt to be mistaken for local events; and distinction between local events and noise bursts is ambiguous if signal-to-noise ratio is small. As mentioned in Section 3.2, multi-station-format is useful in identifying seismic events because in that record form the waveforms of other stations can be compared with and apparent velocities of phases are also considered.

The procedure is as follows (see **Fig. 3.6**):

First, signals from the time code and a ground motion (normally vertical low-gain) channels are digitized at a full sampling rate (typically 50 KHz in real time at a 19 cm/s tape speed) by "AUTO" according to the filing schedule table. The table has been made by setting small N_{min} value in "CAUTO".

Secondly, "ENVEL2" numerically rectifies the seismic signals digitized by "AUTO", and reduces sampling rate to 5 Hz in seismic time by representing each 0.2 s interval by the maximum amplitude of that period. An accurate 5 Hz sampling rate is kept on the basis of the time code signal. Corrections for drift of the time code signal are applied by "TCGENV". Since data made by "ENVEL2" have a reduced and unified sampling rate and have no time code signal, they amount only several tenths of their original form.

Thirdly, after the rectified and resampled records are created by "ENVEL2" and "TCGENV" for every station in the network, records are multiplexed for each event into the multi-station form by "ERAG". In this record form, a total 24 hours of data for six-OBS array amounts only 5 Mbytes, which corresponds to about two percents of the original form.

Fourthly, The program "DAERAG" visualizes the station-multiplexed records on the eight-channel strip-chart recorder through D/A conversion which is performed at a speed ten times faster than the seismic time. An example of a rectified and time-compressed multi-station record is shown in **Fig. 4.14**. As shown in the figure, local events are easily discriminated from distant ones, *T* phases or other noises by visual inspection of wave signature, *S-P* times, apparent velocities etc. Then a schedule table for waveform data filing which contains only local earthquakes is made.

PLOT OF MULTI-STATION RECORDS

In Urabe and Hirata (1984)'s system, the use of an eight-channel strip-chart recorder limited the number of seismic channels on multi-station paper records to eight. Whereas, since each OBS has three seismic channels, in case of a six-OBS array for example, 18 channels are necessary to be plotted on one chart. To solve this shortage of plotting channels, in the improved version, stations multiplexing and plotting of multi-station records are carried out by a program "RAGPLOT" using a plotter in the computer center of Kyushu University.

"RAGPLOT" plots station-multiplexed records of up to nine stations (27 channels) with a strip chart X-Y plotter. In these records, time scale is normally 10 mm/s and amplitudes are normalized by the maximum of each trace (**Fig. 4.15**). Since a typical record is 100-second long and it becomes 1-meter long on paper, cut-sheet plotters or laser printers can not be used. It takes about 10 minutes to plot such a 27-channel record.

RECORD SECTIONS OF REFRACTION DATA

Records of explosion refraction experiments are digitized by "AUTO" according to schedule tables made from shot times and epicentral distances. The digitized data are then resampled by "RESAMP" to unify the sampling rate. Data thus obtained are plotted in the form of record sections by a program "PASTUP" which runs on both the computer center and the minicomputer. In minicomputer, record sections are plotted on the graphic display or with the X-Y plotter. In the computer center, they are plotted with laser printers. High-pass, low-pass, band-pass or band-stop filters (Saito, 1978) can be applied to the traces (**Figs. 4.5 - 4.12**). The graphic display and the laser printer plot a typical record section within a few minutes, whereas the X-Y plotter takes more than one hour to do it.

3.4 CONCLUSION

By utilizing computers and a fast analog input and output facility, digital processing of DAR data from OBSs have been realized. The new system is used as a set of tools to produce multi-station records of earthquake networks and record sections for refraction experiments. Not only it reduces time and labor needed for data playback, but also it produces station-multiplexed records which are unobtainable with the conventional analog data processing. The use of this record format greatly facilitates the work of identifying and picking seismic phases. By the use of the final version of this system, for example, the period required to produce multi-station records of local earthquakes from a 25-day earthquake observation with a six-OBS array would be a total of about 100 hours except for the manual discrimination of seismic events on paper records.

4 OBSERVATION AND DATA

4.1 EARTHQUAKE OBSERVATION IN 1982

4.1.1 OBSERVATION

A five-OBS array for seismicity study was deployed on July 14 and 15, 1982, with a chartered M/V Tokyo Maru (Ichikawa Offshore Co.) on the landward slope of the middle Japan Trench off Fukushima prefecture (Fukushima-Oki), 36.9 - 37.6 ^{o}N, at 0.5 - 4 km water depth. The OBS positions are shown in **Fig. 4.1.** The observation (= recording) period of the network is 25 days (July 15 - August 10). All OBSs were the acoustic-release type and were successfully recovered on September 2 and 3, in the second cruise of the same ship. The specifications of the OBSs are listed in **Table 4.1.**

4.1.2 DATA REDUCTION

The vertical component records of OBSs S17 and S18 were partly bad probably for some faults of the recording amplifiers. Therefore for these part only the horizontal component was used. The data were processed by using Urabe and Hirata (1984)'s system described in Section 3.2. The seismic events in which *STA/LTA* exceeded four were detected from recovered data tapes. The daily number of events detected at each OBS is shown in **Fig. 4.2.** About 600 earthquakes which were detected at more than two OBSs were selected out for further analysis. Since a large earthquake (the Ibaraki-Oki earthquake; $M = 7.0$) occurred on July 23 at 100 km south of the array, approximately 80 % of the 600 events were its foreshocks and aftershocks. The OBS observation covered the period of the fore-, main- and after-shock activity of the Ibaraki-Oki earthquake. This subject will be reported elsewhere so we do not deal with it here. Station-multiplexed digital waveform files of the selected events were created and visualized on paper through D/A conversion. An example of the record is shown in **Fig. 4.3**. P and S arrival times were read from the paper records with a tablet digitizer. The readings were classified into two ranks according to their quality, which were used to estimate errors in hypocenter determination. Excluding the Ibaraki-Oki activity by the use of $S-P$ times, 188 earthquakes near the OBS array were selected for further analyses.

4.2 EARTHQUAKE OBSERVATION AND REFRACTION EXPERIMENT IN 1985

4.2.1 OBSERVATION

GENERAL

A combined experiment of refraction and earthquake observations was conducted in the autumn of 1985, in the area of the inner trench slope of Fukushima-Oki. The experiment was carried out through two cruises, in both of which a 500-ton ship No. 3 Kaiko-Maru of Tokai Offshore Service Co. was chartered to use as a research vessel. The first cruise, October 2 - 18, surveyed an explosion refraction profile and deployed six OBSs for an earthquake observation. In the second cruise, November 2 - 6, the OBSs for earthquakes were retrieved. The positions of OBSs are shown in **Fig. 4.1** and their specifications are tabulated in **Table 4.2.** **Figure 4.4** shows the observation periods of the OBSs.

EXPLOSION

The explosion profile is 200 km long and parallels the Japan Trench at the area of 3 - 4 km water depth on the landward trench slope. Distance from the trench axis is 60 - 70 km. Ten OBSs were set on the profile. Since a storm delayed the operation of the explosions, three timed-release OBSs P3, P5 and P7 were replaced with the same type OBSs PR3, PR5 and PR7, respectively, at the same positions. The other OBSs (S1, S2, S4, S6, S8, S9 and S10) were of the acoustic-release type. All OBSs but S1, which was lost for unknown cause, were successfully retrieved.

Total of 5 tons of dynamite was detonated as 76 shots (4 x 400 kg, 20 x 100 kg, 1 x 75 kg, 2 x 50 kg and 49 x 25 kg) at the average spacing of 2.7 km on the profile, on October 4 and 8. The charge sizes, positions etc. of explosions are tabulated in **Table 4.3.** Each explosive was suspended at 50 - 100 m (depending on the optimum depth for charge size) from a surface buoy with a rope, and then electrically fired. Each of the 400 kg charges was partitioned into two 200 kg ones and fired simultaneously in order to enhance downgoing seismic signals

(Kanazawa and Shimamura, 1986). Since the shots were triggered manually for a technical reason, shot timings were obtained by recording detonator currents on a strip chart recorder. The positions where charges were launched were adopted as the shot positions. Since the sounder was faulty, the water depths along the profile were obtained from the travel times of bottom reflections of shots recorded with hydrophones suspended into the sea from the shipboard. A 9-liter airgun was also shot along the profile.

EARTHQUAKE

Earthquakes were observed with a network of six OBS stations which covered the southern half of the explosion profile, as shown in **Fig. 4.1.** The observation area is shifted by 50 km trenchward from that of the 1982 OBS array. The water depths of the OBS sites are 1.5 - 6 km. Three of the six stations are at the same positions as the refraction experiment. For these three stations, timed-release type OBSs PR2, PR4 and PR6 succeeded OBSs S2, S4 and S6, respectively, at the same position. Off-the-profile OBSs, P11, P12 and P13, were also the timed-release type. The observation period of the OBS array was from October 12 to November 4 (23 days). All OBSs were retrieved successfully on schedule.

4.2.2 DATA REDUCTION

EXPLOSION

The data were processed by using the system described in

Sections 3.2 and 3.3. The total amount of the digitized waveform data was about 60 Mbytes. The vertical component signal of OBS S8 had not been recorded, probably because of a fault of the amplifier. Thus the records of S8 were not used in the analyses. The horizontal components of OBSs PR3, S4 and PR5 also could not be used because the trimmings of the gimbals were improper.

Record sections with various parameter settings were made. They are shown in **Figs. 4.5 - 4.12.** In most OBSs, first arrivals are clearly seen for 25 kg shots at epicentral distances more than 100 km. First-arrival times at short epicentral distances (approximately less than 60 km) were digitized from the low-gain vertical component record sections to which a relatively broad (3 - 20 Hz) band-pass filter was applied in order to avoid errors due to phase delay. At the longer epicentral distances first arrival times were read from the high-gain record sections instead of the low-gain ones. The arrival times of later phases were digitized from those record sections to which a relatively narrow band-pass filter (3 - 10 Hz) was applied in order to enhance the signal-to-noise ratio. A record section and plots of the picked travel times of each OBS are shown in **Fig. 5.6 - 5.13**, together with the result of the ray-tracing which will be described later on.

EARTHQUAKE

Earthquake data were also processed with the system described in Sections 3.2 and 3.3. The vertical high-gain channels of OBSs P11, P12 and P13 had signal amplitudes as low as that of the low-gain channel probably because of bad temperature characteristics of some parts in the amplifiers. It was, however, not a serious problem. The horizontal component of OBS P11 had not been recorded for an unknown cause.

Seismic events were detected from records of the vertical low-gain channel or middle-gain channel (for PR2, PR4 and PR6; they had three different gains). The trigger levels are *STA/LTA* = 3 for OBSs P11, P12 and P13 and 5 for the other OBSs. At these trigger levels, almost all earthquake events that can be recognized manually on visual records would be detected.

The daily number of triggers at each station is shown in **Fig. 4.13.** These events include noise bursts as well. The high trigger rates of OBS P11 in the first half of the observation period is due to a high frequency of noise bursts. Except for these, the average trigger rates are 40 - 80 events/day, showing higher rates at the northwestern stations and lower at the southeastern ones.

To avoid omitting very small earthquakes which are detected by only one or two OBSs, it was decided that all the events detected at any OBS except for noisy P11 were to be visually inspected on time-compressed multi-station records (see Section 3.3). A filing schedule table was constructed from the event lists except for P11 by setting N_{min} = 1. The number of events in the schedule is about 1900 and the total event time is 72 hours (13 % of the observation period). Time-compressed station-multiplexed paper records were made according to the schedule table. An example of the record is shown in

Fig. 4.14. The signal channels used are the same as those used in the event detection.

The paper records thus obtained were used to discriminate local earthquakes (i.e. those occurred in and near the array) from other events, such as distant earthquakes, noise bursts, and *T* phases from distant earthquakes, through visual inspection. The earthquakes which had, or which were possible to have, *S-P* times less than 20 s were selected out for further analyses.

The number of the events selected out was about 230 and the total event time was 5.5 hours (1 % of the observation period). These events were digitized again from the analog tapes to create the waveform data files, which amounted to 80 Mbytes with a unified sampling rate of 60 Hz. The records were then stations-multiplexed and plotted on paper with X-Y plotters. All seismic channels (i.e. other than the clock signal channels) were plotted on the paper records. The numbers of seismic data channels of the multi-station file ranges 18 - 27; because three OBSs PR2, PR4 and PR6, which replaced OBSs S2, S4, and S6, respectively, have six seismic channels, while the others have three. An example of the paper record is shown in **Fig. 4.15.**

P and S arrival times of 218 earthquakes were read from the multi-station records by using a tablet digitizer. Accuracies were evaluated for each arrival time data, as ranks A, B, C, and D, which correspond to 0.1, 0.3, 0.5, 1.0 s, and more, respectively for P arrivals, and 0.3, 0.5, 1.0, 2.0 s, and more, respectively for S arrivals. These estimations of accuracies were used to evaluate the errors of the calculated hypocenters.

In practice, few events among those detected by only one or two OBSs were recognized on the final paper records as nearby earthquakes.

5 ANALYSIS AND RESULT

5.1 REFRACTION EXPERIMENT

5.1.1 METHOD

At first, to obtain a simple and objective first approximation of the velocity structure along the profile, a one-dimensional model was constructed for each single-sided profile of each OBS. This was performed by inverting intercept time and ray parameter (or apparent velocity) data of first arrivals by the slope-intercept method. The data were derived by fitting straight lines for the crustal phases and by a time-term analysis for the P_n phases. These models were used to set up the starting model for a ray-tracing in the next step.

The final model was obtained through two-dimensional ray-tracings. The model was repeatedly revised to fit the calculated travel times to the observed ones for all the OBSs. Later phases as well as first arrivals were considered, so that the layers which had been masked in the previous slope-intercept analysis were also introduced.

5.1.2 ONE-DIMENSIONAL MODELS

DATA OF CRUSTAL PHASES

Since breaks were not prominent in the arrival time curves of the crustal phases between the direct water waves and P_n

waves, data sets of slope and intercept (i.e. ray parameter and intercept time) of the first arrival time curves of crustal phases were measured by the following method:

Straight lines with intercept times equally spaced by step of 0.25 s were fitted to the smoothed first arrival time curves which were traced on the water-depth corrected record sections (i.e. travel times in seconds were reduced by the water depth in meters divided by 1500 m/s). Since the refracted waves from the sedimentary layer is hardly observed as first arrivals, the P-wave velocity of the sedimentary layer was assumed as 2.0 km/s.

In this method, each pair of slope and intercept data which represents a straight line fitted to the travel time curve introduces a corresponding homogeneous layer into the resultant structure model. The stability of this method can be testified by checking the dependency of the result on the selection of straight lines fitted to the curve. In **Fig. 5.1**, four structure models obtained by this method from a complete data set with the 0.25 s-spaced intercepts and from its three different subsets are shown. The first of the three subsets is with the eliminated first (uppermost) crustal slope-intercept pair from the complete data set, the second is with the eliminated last (lowermost) crustal pair, and the third is with eliminated every second pair. As seen in the figure, the complete and reduced data sets give similar models with velocities increasing from about 5.0 to 6.2 km/s in the depth range of 5 - 13 km, indicating high stability of this method.

DATA OF P_n PHASE

As for the mantle phases (P_n waves), the velocity and its time-terms (half the intercept time) were obtained by the time-term method (Scheidegger and Willmore, 1957). In this analysis, data are the water-depth corrected P_n travel times between OBS pairs, in which the shot nearest an OBS position (within 2 km) is used as the source at the OBS. The OBS pairs used are those in which the distances between the two are greater than 70 km beyond which P_n phase appears as first arrivals. The travel time data between two stations A and B is expressed as

$$T_{\mathrm{AB}} = X_{\mathrm{AB}}/ V + e_{\mathrm{A}} + e_{\mathrm{B}},$$

where X_{AB} is the horizontal distance between A and B, V is the wave velocity of the underlying layer (uppermost mantle in this case), and e_{A} and e_{B} are the time-terms of the stations A and B, respectively. Here, T_{AB} and X_{AB} are data, and V, e_{A} and e_{B} are unknowns which are to be determined under an over-determined condition with by least-squares method. The data set consists of 25 pairs of travel time and epicentral distance data from seven OBSs. Since OBS P5/PR5 yielded insufficient data and S8 had no vertical component record, these OBSs were omitted from the calculation. The results are tabulated in **Table 5.1**. The P_n velocity was determined as 8.26 $\pm$0.08 km/s.

SLOPE-INTERCEPT ANALYSIS

The intercept times and ray parameters thus obtained for crustal and P_n first arrivals are tabulated in **Table 5.2**. The time-term for OBS P5/PR5 was assumed by interpolation of the neighboring OBSs. These data were used to construct a one-dimensional P-wave velocity structure for each single-sided profile by the slope-intercept method using a formulation by Diebold and Stoffa (1981) (the τ-sum traveltime inversion). Every pair of a ray parameter and an intercept time is connected to a particular layer of a homogeneous velocity, and then a one-dimensional model which explains the data set is obtained.

The intercept time of the ray with a ray parameter p_k is expressed as

$$T(p_k) = 2\sum_{j=1}^{k-1} Z_j (u_j^2 - p_k^2)^{1/2} .$$

where, Z_j and u_j are the thickness and the slowness of the j-th layer, respectively. Then, by using a set of ray parameters p_i and intercept times T_i with i=1 to k+1, Z_k is obtained from the recurrence formula

$$Z_k = \frac{T(p_{k+1})/2 - \sum_{j=1}^{k-1} Z_j (u_j^2 - p_{k+1}^2)^{1/2}}{(u_k^2 - p_{k+1}^2)^{1/2}} .$$

The data and result of the calculation are tabulated in **Table 5.2** and shown in **Fig. 5.2**. The common features of the models are that below a 1 - 2 km thick sedimentary layer, the velocity increases from 5 to 6 km/s in the thickness range of 7 - 8 km. The Moho is at depths of 20 - 21 km.

5.1.3 RAY-TRACING

PROCEDURES

Naturally, one two-dimensionally varying structure must explain all the data, i.e. both first and later arrivals. To incorporate that in the model, a final P-wave velocity structure model was obtained with two-dimensional ray-tracings through the following procedures:

First, as many the picked phases as possible were identified and then the number of interfaces for the two-dimensional structure model was determined. Interfaces within the crust (i.e. below the bottom of the sedimentary layer and above the Moho) were introduced by the use of the later arrivals which were interpreted as reflected or refracted phases from the corresponding interfaces.

Secondly, an initial two-dimensional model having the obtained number of interfaces and the velocities as suggested by the previously obtained one-dimensional models was constructed.

Thirdly, the model parameters (i.e. velocities and shape of interfaces) were successively improved to fit the calculated travel times to the observed ones using a simple ray-tracing program (RT) in which the model is approximated as a stack of homogeneous layers with piece-wise interfaces. The model was constructed basically in a top-down manner.

Lastly, the model was finalized with another ray-tracing program (SEISOBS) in which a linear velocity gradient could be introduced to each layer and interfaces were defined by bicubic spline functions.

RAY-TRACING PROGRAMS

The firstly used ray-tracing program RT deals with a two-dimensionally layered structure in which layers have homogeneous velocity distributions. It generates direct, primary reflected, and critically refracted rays which travel along an interface. The program runs on a personal computer (NEC PC-9801) and utilizes interactive graphics.

The secondly used program SEISOBS (Hirata and Shinjo, 1986) is a modified version of SEIS83 (Cerveny and Psencik, 1983) for OBS data, in which the receiver is placed on the sea bottom instead of on the free surface. SEISOBS (and SEIS83 also) incorporates velocity gradients and no rays traveling along an interface. Direct, primary reflected and refracted waves were calculated. It also generates synthetic seismograms.

5.1.4 INTERPRETATION AND RESULT

GENERAL

The finally obtained P-wave velocity structure model is shown in **Fig. 5.3.** **Figure 5.4** shows the ray diagrams of three representative stations S2, S6 and S10. In **Figs. 5.5 - 5.12**, a record section and the comparisons of the theoretical travel times with the observed ones are shown for each OBS. The former times fit the latter ones within 0.1 s for the first arrivals which are picked with confidence. The synthetic record sections are also presented in these figures, however, they have not been used in constructing the structure model.

Except that the Moho deepens to the south, the structural change along the profile is rather small as has been assumed from the relatively simple time-distance curves. The data can be explained by a velocity structure model with isovelocity interfaces. The model is specified by the velocities immediately above and below curved interfaces between which the velocity is linearly interpolated between them.

SEDIMENTARY LAYER

The velocity of the sedimentary layer is assumed as 1.8 km/s at the top and 2.2 km/s at the bottom on the basis of the preliminary result of the airgun-OBS survey (Kaiho *et al.*, 1987; Kaiho, personal communication). The velocity of the layer underlying the sedimentary layer is set as 5.0 km/s throughout the model after the one-dimensional models, and it gives satisfactory fits for data also in the ray-tracing. From these velocities, the thickness of the sedimentary layer is determined as 1.3 - 1.9 km, being thinnest at around D = 90 km (position along the profile, D, is measured from OBS S10 and is positive southward) where the 5.0 km/s layer swells.

5.0 KM LAYER

The 5.0 km/s layer has a velocity gradient of about 0.1 s^{-1}, although the first-arrival time data can not resolve it. The velocity gradient is introduced because the travel time data over an epicentral distance range of about 25 - 65 km are successfully explained by a layer of 5.7 km/s for all the OBSs whereas the diving waves from the 5.0 km/s layer or the waves reflected by its bottom (i.e. 5.0 km/s branch) are not traced beyond about 30 km. If the 5.0 km/s layer has negligible velocity gradient, 5.0 km/s branch must be traced as secondary arrivals beyond that distance where the 5.7 km/s branch is observed as the first arrivals. A velocity gradient no less than about 0.1 s^{-1} is necessary to explain it. The bottom of the 5.0 km/s layer is somewhat ambiguous for the reflection from it has not been identified, so that it is not necessarily a clear velocity discontinuity. With that velocity gradient above, the 5.0 km/s layer has a thickness of 2.2 - 2.7 km.

5.7 KM/S LAYER

The 5.7 km/s layer is recognized from its refraction which forms the major part of the crustal first arrivals as mentioned above. This layer has little velocity gradient because in some OBSs the diving waves or bottom reflections of this layer are prominent for epicentral distances more than 100 km; they are observed in S6 (at D = 170 - 205 km in **Fig. 5.9**), P7/PR7 (at D = 140 - 205 km in **Fig. 5.10**) and in S10 (at D = 90 - 110 km in **Fig. 5.12**). They are traceable to an epicentral distance of 130 km in the south of OBS P7/PR7.

6.0 KM/S LAYER

The crustal structure below, including the bottom of the 5.7 km/s layer (i.e. lower half of the crust), is rather difficult to determine with a good resolution because they do not appear as first-arrivals.

The bottom of the 5.7 km/s layer, which is supposed to be the top of a 6.0 km/s layer, is rather ambiguously determined. It is introduced to explain the secondary arrivals observed at the north of S2 (D = 130 - 150 km in **Fig. 5.5**), the south of S4 (D = 170 - 190 km in **Fig. 5.7**), and the south of P7/PR7 (D = 95 - 115 km in **Fig. 5.10**), as the waves reflected by it. These phases are clearly observed only at the latter station, while they are somewhat vague at the former two stations. Therefore the velocity discontinuity (i.e. the top of this layer) is not necessarily required other than near the middle of the profile (D = 80 - 100 km). The velocity of the 6.0 km/s layer can not be resolved from the arrival time curves. However, it must be less than 6.2 km/s because refracted (or diving) waves from this layer would otherwise appear as first arrivals, and it can not be so close to 5.7 km/s to generate the reflected waves observed.

<u>6.8 KM/S LAYER</u>

Underlying the 6.0 km/s layer is a 6.8 km/s layer. This layer is suggested by interpretative sub-critically reflected waves in OBSs P3/PR3 (D = 105 - 125 km in **Fig. 5.6**), S6 (D = 60 - 80 km in **Fig. 5.9**), P7/PR7 (D = 115 - 125 km in **Fig. 5.10**), S9 (D = 65 - 75 km in **Fig. 5.11**) and S10 (D = 35 - 55 km in **Fig. 5.12**), and refracted or diving waves in S6 (D = 170 - 205 km in **Fig. 5.9**), P7/PR7 (D = 190 - 200 km in **Fig. 5.10**) and S10 (D = 80 - 120 km in **Fig. 5.12**). Although the appearance of these waves are relatively weak, being observed at many stations convinces us of this layer's existence. The estimation of velocity for this layer has ambiguity of $\pm$0.2 km/s, as much as that of the 6.0 km/s layer.

<u>UPPERMOST MANTLE</u>

The above crustal construction leads to the Moho depth of 21.2 km in the northern end of the profile and 23.5 km in the southern end, dipping southward. The dip is somewhat steeper in the middle than in the ends. The velocity of the uppermost mantle is 8.26 ±0.08 km/s as obtained with the time-term analysis. Calculated travel times satisfactorily fit to the observed ones for the P_n waves as shown in **Fig. 5.5 - 5.12.** *PmP* phases are observed at S2 (D = 105 - 125 km in **Fig. 5.5**), S4 (D = 90 - 110 km in **Fig. 5.7**), P5/PR5 (D = 90 - 100 km and 175 - 180 km in **Fig. 5.8**), S6 (D = 45 - 55 km and 125 - 155 km in **Fig. 5.9**), and P7/PR7 (D = 20 - 40 km and 115 - 130 km in **Fig. 5.10**).

The possible change of the structure above the Moho, which would be mostly within the lower crust, would cause the change in Moho depth on the order of 0.5 km. Replacement of the 6.8 km/s velocity with 6.0 km/s, though which is unrealistic, would deepen the Moho throughout by 1 km.

SUMMARY OF STRUCTURE

The characteristics of structure obtained for the refraction profile in Fukushima-Oki is summarized as follows:

Lateral variation of the constitution of the crust along the profile paralleling the trench is small. The sedimentary layer is 1.3 - 1.9 km thick with an assumed mean P-wave velocity of 2.0 km/s. The underlying 5.0 and 5.7 km/s layers are clearly defined throughout the profile from the first arrival data. The 6.0 and 6.8 km/s layers below are somewhat ambiguous as they are masked layers, and they have velocity ambiguity of ± 0.2 km/s. The existence of the 6.8 km/s layer is convincing, but that of the 6.0 km/s layer is certified only for the small part of the profile. The P_n velocity of 8.26 ± 0.08 km/s and the southward-dipping Moho which varies from 21 to 23.5 km (± 0.5 km) along the profile are definite from the clearly observed P_n waves.

5.2 EARTHQUAKES

5.2.1 HYPOCENTER DISTRIBUTION

COMPUTER PROGRAM

Hypocenters were calculated for 406 events (188 from 1982 and 218 from 1985) by Hirata and Matsu'ura (1987)'s computer program. This program estimates a maximum-likelihood solution from a Bayesian point of view, in which hypocentral parameters are estimated from both a priori information (i.e. an initial hypocenter and its reliability) and observed travel time data. The program works for velocity structure models having a piece-wise constant velocity gradient with depth without velocity discontinuity. Both *P* and *S* phases are equally used in the calculation in which each datum is weighted according to the estimated errors of the individual readings. The S-wave velocity structure is derived from a given P-wave velocity structure model by assuming a V_p-to-V_s ratio of 1.73. Errors involved in resultant hypocenters are calculated from the imperfect resolution due to the arrangement of stations and also from the assumed inaccuracy of the travel time data and the initial hypocenter. Since we had no a priori information about the hypocentral locations except that they were local events, the initial hypocenter was set at the center of the OBS arrays; 37.25 ^{o}N, 142.50 ^{o}E and a depth of 40 km. Large inaccuracies of 100 km in horizontal and 40 km in vertical coordinates were assumed for the initial hypocenter. With this setting, calculated hypocenters hardly depend on the initial hypocenter.

SEDIMENTARY LAYER

In the hypocenter determination, the effect of the sedimentary layer was took into account as static station corrections instead of being included in a horizontally layered velocity structure model (Urabe *et al.*, 1985). This is possible because the travel time within and the thickness of the sedimentary layer can be estimated using a commonly observed converted phase with some assumption.

It is ordinary in OBS records that an *S* phase converted from P wave at the bottom of the sedimentary layer is observed (e.g. Kasahara *et al.*, 1982; Hirata *et al.*, 1983). The observed time interval between *P* phase and the *P* to *S* converted phase (*PS-P* delay time; t_{ps-p}) was nearly constant for each station. The obtained mean t_{ps-p}'s are 2.2, 1.7, 2.0, 1.9, 2.0, 1.9, 1.9, 2.3, 1.5, 1.3 and 2.0 for OBSs S16, S17, S18, S19, S20, S2/PR2, S4/PR4, S6/PR6, P11, P12 and P13, respectively. Examples of the *P* to *S* converted phase are shown in **Figs. 4.3 and 4.15**.

Since the seismic ray in the sedimentary layer is near vertical for its slow wavespeed in comparison with those of the layers below, the effect of the sedimentary layer can be included in a static (fixed) station correction. This is performed by reducing the station's altitude by the sedimentary thickness and the arrival time by that within the sedimentary layer. Then the station is virtually placed on the top of the layer which underlies the sedimentary layer (i.e. the 5.0 km/s layer) and hypocenters are calculated for a structure model with the sedimentary layer eliminated.

Assuming a vertical incidence, t_{ps-p} is written as

$$t_{ps-p} = t_s - t_s$$
$$= t_p(V_p/V_s - 1),$$

here t_p and t_s are P- and S-wave travel times in the sedimentary layer, respectively, and V_p and V_s are P- and S-wave velocities of the sedimentary layer, respectively. We have assumed the mean P-wave velocity of the sedimentary layer as 2.0 km/s for the refraction profile on the basis of the airgun data (see Section 5.1.4). Its thickness has been obtained to be 1.8 km for each of the OBSs S2, S4 and S6, at which sites earthquakes also have been observed. The t_{ps-p} data, and the velocity and thickness of the sedimentary layer yield V_p/V_s's of 3.1, 3.1 and 2.6 for OBSs S2/PR2, S4/PR4 and S6/PR6, respectively. On the basis of this estimation, we assume a V_p/V_s ratio of 3.0 (Poisson's ratio of 0.44) which are to be applied to the sediments below all OBS sites in common.

The t_p, t_s and the thickness of the sedimentary layer estimated from the t_{ps-p} data and $V_p/V_s = 3.0$ for each OBS site are tabulated in **Table 5.3**. The t_p and t_s range from 0.65 to 1.15 s and from 1.95 to 3.45 s, respectively, and the sedimentary thickness ranges from 1.3 to 2.3 km.

Although the above estimation is based on the common and approximated V_p/V_s ratio obtained from the three OBSs, the error in resultant hypocenters due to improper estimation of V_p/V_s is probably small. This is because the measured t_{ps-p} itself corresponds to the station correction for the *S-P* time, so that no error is introduced to the *S-P* time. Besides, a 20 % error in assumed V_s would cause less than 0.1 s of relative changes in both the P- and S-wave station corrections and about 0.2 km of relative change in station altitude.

VELOCITY STRUCTURE MODELING

The velocity structure model used in the hypocenter determination was constructed on the basis of the result of the refraction experiment, with reference to the existing structural cross section of Sanriku-Oki region.

The earthquakes located in this study, in the result, are distributed almost at 50 - 170 km landward from the trench axis as will be shown later on. The two OBS arrays in 1982 and 1985 covered the ranges of 70 - 150 km and 30 - 100 km, respectively, landward the trench. Since we use a one-dimensional velocity structure model in hypocenter determination, the model must have a structure averaged over the whole range of 30 - 170 km landward the trench, which is the most part of the landward trench slope in Fukushima-Oki.

Since the penetration of the past studies on the crustal structure of Fukushima-Oki (Murauchi and Asanuma, 1970; Sakurai *et al.*, 1981; Suyehiro *et al.*, 1984) are no more than several kilometers below the sea bottom and the refraction profile in this study parallels the trench at just 60 - 70 km landward the trench axis, the postulated structural change across the trench can not be constructed from these data alone. The crust and upper mantle cross section of a neighboring region in the Japan Trench, Sanriku-Oki, is considered a good model to be referred to in constructing that of Fukushima-Oki region. As stated in Section 1.2, the structure of northern Sanriku-Oki has already been studied extensively.

Results of the studies on the structure of northern Sanriku-Oki (Ludwig *et al.*, 1966; Research Group for Explosion Seismology, 1977; Asano *et al.*, 1979; Asano *et al.*, 1981; Kanazawa *et al.*, 1985) compiled by Kanazawa *et al.* (1985) (**Fig. 1.5**) give the following characteristics which may be similar to the adjacent regions along the Japan Trench: The lithosphere beneath the continental slope consists of a sedimentary, 4.8 - 5.1 km/s, 5.7 - 6.0 km/s and 6.8 km/s layers, and the uppermost mantle with 8.1 km/s, from the top downward. The Moho deepens landward from 14 km at the trench axis to 27 km at the coast. The Moho depth in the distance range of 30 - 170 km from the trench axis is 16 - 26 km, and that at the middle of this range is 23 km. This deepening of the Moho is due to thickening of the 5.7 - 6.0 km/s layer. The Moho in Sanriku-Oki at the position corresponding to the refraction profile of Fukushima-Oki is 19 km deep. This is about 4 km shallower than that obtained around OBS S4 in Fukushima-Oki.

The constitution of the crust and uppermantle velocity structure of Fukushima-Oki obtained in this study's profile is a sedimentary, 5.0, 5.7, 6.0 and 6.8 layers and the Moho from top downward, and is very similar to that of Sanriku-Oki. Thus we naturally assume that the Moho in Fukushima-Oki likewise deepens landward due to a thickening of the 6.0 km/s layer. Considering the 4 km difference in depth, the Moho depth in the distance range of 30 - 170 km landward the trench is supposed to be about 20 - 30 km and that at the middle of this range to be about 27 km. Hypocenters were calculated for the seven different models which were basically obtained beneath OBS S4 but with thickened 6.0 km/s layers to place the Moho at depths of 23 - 29 km by steps of 1 km (**Fig. 5.13**). The averaged travel-time residuals for these models are plotted in **Fig. 5.14**. Although the residuals are monotonously smaller for the thicker crust models, we choose one with a Moho depth of 27 km as an optimum model for hypocenter determination because the deeper Moho seems rather unrealistic.

The effect of the ambiguity of structure model on a hypocenter located was took into account by adding one percent of the travel time to the estimated reading errors.

<u>RESULT</u>

Out of the 406 earthquakes (188 and 218 from the 1982 and 1985 observations, respectively) for which hypocenters were determined, the number of those located in the area of Fukushima-Oki (area of **Fig. 5.15**; approximately 36.5 - 38.0 $^{\circ}$N and 141.0 - 144.0 $^{\circ}$E) was 303 (147 from 1982 and 156 from 1985).

Figure 5.15 shows the epicenters of these events, in which those with estimated standard errors in location (square roots of the diagonal elements of the asymptotic covariance matrix for estimation errors) up to 50 km are included.

Among these, the number of those with standard errors less than 10 km are 195 (123 from 1982 and 72 from 1985). **Figs. 5.16 - 5.19** show the distributions of these hypocenters. In **Figs. 5.16 and 5.18**, the hypocentral locations for the 1982 and the 1985 observations, respectively, are plotted. **Figs. 5.17** and **5.19** indicate the estimated location errors of each hypocenter in **Figs. 5.16 and 5.18**, respectively, by error ellipses which represent one standard error (Hirata and Matsu'ura, 1987). The error ellipses make it possible to appreciate the resolution of the hypocenter location on a individual basis.

Characteristics of the hypocenter distributions during the periods of two observations in 1982 and 1985 are described as follows:

In both observations, there are small numbers of earthquakes located seaward the 3 km isobath, and most events are concentrated landward the 2 km isobath, which corresponds to the area more than about 90 km distant from the trench axis. In the 1985 observation, many events are located in the northern part. In comparison with that, in 1982, many events are found in the southern part. Besides, many events landward the 0.5 km isobath are located in 1982, while they are not in 1985. Shallow foci seaward the 4 km isobath (i.e. within about 70 km from the trench axis) are located only in the 1982 observation.

In view of the vertical sections across the trench, more events are located in the shallower (less than 25 - 30 km) zone than in the deeper zone, and almost all the deeper events are located at more than about 80 km landward from the trench axis; only several crustal events are located at farther seaward. Within about 50 km from the trench axis, no focus is located except for a deep (70 km) one in 1985.

Some of the patterns of hypocenter distribution described above may be artifacts due to the detectability and geometry of the OBS networks. As stated in the following sections, the patterns are reliable within approximately 50 km from the margin of each network.

5.2.2 RELIABILITY OF RESULT

MAGNITUDE

In order to evaluate the earthquake detectability of the network, it is necessary to determine the magnitudes of the events. Generally, magnitudes of small earthquakes are determined from their amplitudes or *F-P* times (t_{f-p}), and epicentral distances (e.g. Tsumura, 1967; Watanabe, 1971). We used the method of t_{f-p} because it does not use wave's amplitude which may be saturated, and within the epicentral distance range of 200 km the magnitudes thus obtained hardly depend on the epicentral distances (Tsumura, 1967).

We determined the magnitudes with reference to those determined by JMA (Japan Meteorological Agency). The t_{f-p}'s have been measured in the automatic event detection. In this procedure, "*F*" is the time at which a trigger threshold is no longer satisfied, and it is generally (and systematically) earlier than that read manually. This is because in the *STA/LTA* method, an *LTA* level, which normally represents the background noise level, rises during an event. This means that the trigger level itself also rises, so that the "*F*" is declared earlier than the apparent die-off of the oscillation. Since the time constant of *LTA* was set longer in the 1982 observation than in the 1985 observation (1000 s and 300 s, respectively), t_{f-p}'s are expected to be systematically longer in the former than in the latter.

Among the events located, the number of those which are listed on the seismological bulletin of JMA with a magnitude value (M_{JMA}) in an area between 36.5 - 38.0 ^{o}N and 141.0 - 144.0 ^{o}E, are 7 and 10 for the 1982 and the 1985 observations, respectively. The M_{JMA}'s of these events range 2.7 - 4.8 and 2.8 - 5.0, respectively. The relations between the *F-P* time and the M_{JMA} were established from these events in the least-squares sense (**Fig. 5.20**) as;

$$M_{JMA} = 1.73\ \overline{\log t_{f-p}} - 3.8 \quad \text{for 1982}$$

and

$$M_{JMA} = 2.50\ \overline{\log t_{f-p}} - 6.5 \quad \text{for 1985,}$$

where $\overline{\log t_{f-p}}$ is the stations average of $\log t_{f-p}$. The difference in coefficients between these two relations is due to the above-mentioned systematic difference of t_{f-p}'s between the data sets of the two observations.

DETECTABILITY

Using the relations above, the magnitudes of the about 300 events located in Fukushima-Oki area (36.5 - 38.0 ^{o}N and 141.0 - 144.0 ^{o}E) were determined to be 0.2 - 5.0 (0.4 - 4.8 for the 1982 observation and 0.2 - 5.0 for the 1985 observation). Although the smallest limit of the located earthquakes is M = 0.2 - 0.4 as observed from the magnitude-frequency relations (**Fig. 5.21**), some of the small events occurred in the OBS arrays above that limit may have been missed. Assuming the Gutenberg-Richter relation to hold, in which the b value is generally on the order of one, the events approximately larger than M = 2.5 are supposed to have been detected and located rather uniformly in the OBS arrays.

This magnitude level of uniform detectability for the OBS arrays of this area is not so low in comparison with that by the land based network (e.g. Yamamoto and Kono, 1983) but the same order as it. This is supposed to be due to relatively high noise levels at the OBS stations, however, accuracy of the location, especially the depth, is far better by the OBS arrays than by the land network because of the appropriate geometrical coverage.

AREA OF UNIFORM DETECTABILITY

The limitation of the area of uniform detection due to the geometry and sensitivity of the network can be estimated by comparing the epicenter distributions (**Figs. 5.15 and 5.19**) with those obtained with the land-based microearthquake network (**Fig. 5.22**; Faculty of Science, Tohoku University, 1983 and

1986). Patterns of the latter ones have the similar characteristics as those obtained with the OBS networks for the corresponding periods; near the OBS arrays, epicenters are restricted to the west of 142.5 ^{o}E, where the southern area is relatively active in 1982, whereas the northern area is active in 1985.

The pattern of epicentral distribution obtained with each OBS network is similar to that obtained by the larger land-based network for the area within some 50 km from the OBS network. Therefore, within that area, the hypocentral distribution is estimated to have been mapped rather uniformly by the OBS array.

With this in mind, since the OBS arrays are situated in the water depth ranges of 0.5 - 4 km and 1.5 - 6 km for the 1982 and the 1985 observations, respectively (or the landward distance ranges of 60 - 150 km and 30 - 100 km, respectively, from the trench axis), seaward aseismicity found in both 1982 (**Fig. 5.16**) and 1985 (**Fig. 5.18**) are true, but the landward low seismicity seen in 1985 is probably an artifact.

EFFECT OF STRUCTURE MODEL

Selection of the velocity structure model (**Fig. 5.13**) has little effect on the horizontal and vertical locations of the sub-crustal events and the horizontal locations of the crustal ones. The depths of the crustal events systematically change in nearly proportional to the Moho depth by the same order. For example, shallow foci located with the model with a 24 km deep Moho deepen by 2 - 8 km if one with a 29 km deep Moho is used.

It would be desirable to use in hypocenter location a laterally varying velocity structure. Such a structure will be modeled on the structural section across the trench, in which the crust thickens landward throughout the inner trench slope by about 10 km (see Section 5.2.1). Nevertheless, it was not undertaken for a technical reason. If it was, the hypocentral locations would be so affected that crustal events, especially those outside the OBS arrays, would change their depth by about the same order as the change of the crustal thickness there (i.e. several kilometers). Sub-crustal events would be affected little. These are because if there is no observation point in the vicinity of the epicenter, the focal depth of a crustal event is mainly constrained by the travel times of P_n waves which shot downward from the focus. Therefore, the crustal events would become deeper in the landward half and shallower in the trenchward half by several kilometers than those determined for the horizontal structure. Nevertheless, the crustal events would remain in the crust.

6 DISCUSSION

LATERAL VARIATION OF STRUCTURE ALONG THE JAPAN TRENCH

Here we compare the structure obtained for Fukushima-Oki with those for the neighboring regions, Sanriku-Oki and Boso-Oki, to discuss the lateral variation of the upper-lithospheric structure along the Japan Trench. Kanazawa *et al.* (1985) have conducted refraction profilings of the crust and uppermantle in the two regions adjacent to Fukushima-Oki, as was referred in Section 1.2 and shown in **Fig. 1.4**. The results of these profiles (denoted as OBS81 and OBS84 in **Fig. 1.4**) are comparable with this study's one because their water depth ranges and positions relative to the trench axis are much the same; i.e. 3 - 4 km and 60 - 70 km landward, respectively. The results appeared so far (Kanazawa *et al.*, 1985) are preliminary ones and further refinement remains to be done, however, there can not be major revisions in the structure models (Kanazawa, personal communication).

Figure 6.1a shows the projection of the P-wave velocity structure of the crust and upper mantle along the 3 - 4 km isobath landward the Japan Trench, projected on the north-south-striking vertical plane. We can see from this figure the lateral variation of the structure along the trench. The three regions have similar velocity structures except that the layers' thicknesses vary; in consideration of the resolving power of the experiments, there is no significant difference in the velocities of the corresponding layers among them. The 6.0 km/s layer exists in Fukushima-Oki and not in other regions, however, because of the vague appearance of this layer in the Fukushima-Oki records, it is inappropriate to discuss this layer in comparison with other regions.

The major lateral change in structure is that the crust 19 km thick in Sanriku-Oki thickens southward through Fukushima-Oki from 21 to 23 km, and then it connects smoothly with Boso-Oki. The southward dip of the Moho is larger in the middle to the north of the Fukushima-Oki profile, i.e. around the latitude of 38 ^{o}N (see also **Figs. 1.4, 4.1 and 5.3**).

The crustal thickening is due to the 5.7 (and 6.0) km/s layer, whereas the 5 and 6.8 km/s layers show relatively small changes in their thickness; the top of the 6.8 km/s layer nearly parallels the Moho. This observation supports the idea that the 6.8 km/s layer is the oceanic layer 3 of the descending Pacific plate; this idea has originally been reached by inspecting the structural cross section shown in **Fig. 1.5**. It also suggests that the lateral deepening of the top of this layer (and that of the Moho) along the trench represents the lateral deepening of the subducting plate itself; the 5.7 km/s layer fills up the space by its thickening instead. In other words, the dip angle of the subducting lithospheric slab at the Japan Trench is somewhat steeper in the south than in the north. The actual dip angle of the slab beneath the inner trench wall estimated from the seismic refraction studies is 4 - 5 ^{o} for Sanriku-Oki (from **Fig. 1.5**) and 8 - 9 ^{o} for southern Fukushima-Oki and Boso-Oki. Multi-channel seismic reflection profilings have also yielded the structural sections across these areas, in which sections a dipping oceanic layer is traced up to several tens of kilometers landward the trench axis (Nasu *et al.*, 1980; Sakurai *et al.*, 1981). The accurate dip of that layer, however, is difficult to estimate from these sections because of the ambiguity in the wave velocity.

The southward thickening of the crust along the trench is also supported by gravity data. **Figure 6.1b** shows the variation of the free-air gravity anomaly just along the 3.5 km isobath, which is approximately along the three refraction profiles and parallels the trench axis. From the north to the south, the free-air anomaly obviously begins trending towards negative at the middle of the Fukushima-Oki section. The intensity of the anomaly throughout these profiles is on the order of 100 mgal, which is reasonable for the seismic velocity structure obtained.

The origin of the southward-thickening 5.7 km/s layer (and the 6.0 km/s layer) is not clear. Since no evidence of accretion of the oceanic materials has been found in the Japan Trench region (KAIKO II Research Group, 1986), the 5.7 km/s layer in the models may be underlain by the oceanic layer 2 and possibly by the oceanic sediments as their original formation. However, it could not be resolved with these refraction experiments; they are, if they exist in such manner, expected to be slow and thin (1 - 2 km; see **Fig. 1.5**) in comparison with the overlying layers. It is also possible that they are mixed, perturbed and/or assimilated with the overriding layer.

The southward-deepening of the oceanic lithosphere observed beneath the inner slope along the Japan Trench seems to exist before the subduction. The bathymetry map of the Northwest Pacific (**Fig. 6.2**) shows that the ocean floor of the area between the Japan Trench and the Shatsky Rise is lower (i.e. deeper) in the south than in the north, as is represented by the 5.5 and 6.0 km isobaths in the figure. This becomes distinct when water depth profiles paralleling the Japan Trench axis on its ocean side are plotted (**Fig. 6.3**). The Northwest Pacific Ocean floor, which is estimated to be drifting west-northwestward relative to the Eurasia Plate (Minster and Jordan, 1978), obviously deepens southward by 0.3 - 0.4 km between the latitudes of 37 and 39 $^{\circ}$N. This tendency is extending eastward as far as about 1000 km from the trench to the foot of the Shatsky Rise.

It is natural to consider the southward-deepening of the subducting lithosphere along the Japan Trench in connection with the above extensive north-south depth contrast of the ocean floor. The simplest and most probable model accounting for these observations would be that the Pacific plate at subduction along the Japan Trench is inherently heavier (i.e. thicker or denser) in the south of 37 - 39 $^{\circ}$N than in the north of it by the extent causing several hundred meters of ocean floor deepening; such a heaviness contrast may so accelerate the downward motion of the subducting plate more in the southern region than in the northern region that a depth variation by about 4 km is formed beneath the inner trench wall 60 - 70 km landward the trench axis.

The lithospheric age can offer a qualitative explanation for the water depth contrast in the Northwest Pacific. The seafloor between the Japan Trench and the Shatsky Rise has been mapped for geomagnetic reversals giving its ages generally older in the south than in the north (Hilde *et al*, 1976; Isezaki, 1987; **Fig. 6.4**). And in the farther south at the Izu-Bonin Trench, the seafloor older than at the Japan trench is subducting. Since there is a global tendency that a seafloor deepens with its age (e.g. Sclater *et al.*, 1975), the above age data are in accordance with the fact that the seafloor is deeper and the subduction is steeper in the south including the Izu-Bonin Trench, where the water is generally deeper and the deep seismic zone is steeper than those in the Japan Trench.

Quantitatively, however, its age off the Japan Trench area ranges roughly 140 - 120 m.y.B.P. in the late Mesozoic, for which no more than about 100 m of water depth contrast is expected from the typical age-depth relation (Sclater *et al.*, 1975). Thus, the extent of the southward-deepening of the seafloor seemed to be somewhat more than expected from its age. Besides, the observed southward deepening of the Pacific plate is specifically at latitude of 38 - 39 ^{O}N (Figs. 5.3 and 6.3), which can not be explained from the age maps (Fig. 6.4).

Recently, Umino *et al.* (1984) and Mogi (1985) independently proposed a NW-SE trending "tectonic line" in the northern Honshu on the basis mainly of linear distribution of epicenters. The proposed line runs across the northern Honshu and passes the Fukushima-Oki connecting with one of the ancient fracture zones presented in the geomagnetic anomaly map by Hilde *et al.* (1977;Fig. 6.4a) (Umino *et al.*, 1984). This "tectonic line" may be some boundary related to the southward deepening of the

Pacific plate, however, detailed discussion is difficult for the lack of data. It should be noted that a recently revised geomagnetic anomaly map (Isezaki, 1987; Fig. 6.4b) does not have the fracture zone which was presented in the previous map at the corresponding position. Therefore the relationship between the proposed "tectonic line" and the "fracture zone" is not definite.

COMPARISON OF SEISMICITY; FUKUSHIMA-OKI AND SANRIKU-OKI

Although the low seismicity near the trench in Fukushima-Oki in comparison with Sanriku-Oki had been recognized from the land-based observations (see Section 1.2), the earthquake observations in this study offer a more accurate picture of hypocenter distribution in Fukushima-Oki, which enables its direct comparison with that accurately obtained with OBSs in Sanriku-Oki. Accurate hypocenter distribution can not be obtained without a correct velocity structure model and appropriate coverage of the seismic network. Such a distribution can be used to discuss the seismo-tectonics with reference to the structure.

Figure 6.5 shows vertical sections of hypocenter distribution across the Japan Trench for Sanriku-Oki and Fukushima-Oki regions. The former was made by compiling the results from the past three OBS observations conducted by Yamada *et al.* (1980) in southern Sanriku-Oki (or Miyagi-Oki; 38 - 39 ^{o}N) and Hirata *et al.* (1983 and 1985) in northern Sanriku-Oki (39 - 41 ^{o}N); and the latter shows the hypocenters presented in **Figs. 5.16 - 5.19**. Although the observation periods of the OBS arrays are just several weeks, consistency of the pattern of epicentral distribution with those obtained from the land-based long term observations (e.g. **Figs. 1.3 and 5.21**) supports that the general characteristics of hypocenter distributions in the two regions are correctly represented by these OBS data.

The hypocenter distributions of the two regions are obviously different. In Sanriku-Oki, hypocenters are rather uniformly distributed below the both inner and outer trench slopes except for a narrow seismicity gap at 20 - 50 km landward from the trench axis (Hirata *et al.*, 1983 and 1985). On the other hand, in Fukushima-Oki, crustal and sub-crustal seismicities are lower within 50 km and 80 km, respectively, landward from the trench axis. Here, some "crustal" events may be interplate ones occurred near the top of the 6.8 km/s layer or the Moho, however, they are only a small part of the whole crustal events. Concentration of hypocenters as deep as 10 - 30 km in a zone 100 - 150 km landward from the trench axis is also prominent in Fukushima-Oki. The hypocenter distributions farther landward show little difference between the two regions; the deeper (> 40 km) events seems to be connecting with the deep seismic zone.

The above characteristics are summarized as follows; in Fukushima-Oki, besides that an aseismic zone beneath the inner trench wall like one observed in Sanriku-Oki is present, the descending oceanic plate itself is almost aseismic as far as 80 km landward the trench axis, while the seismicity within the upper plate is significantly high. In striking contrast with it, hypocenters are distributed rather uniformly in Sanriku-Oki.

The aftershock activity of the great Sanriku earthquake of 1933 (M_S = 8.5), which was a great lithospheric normal faulting near the trench axis (Kanamori, 1971), may partly account for the seismicity around the trench in Sanriku-Oki. However, the presence of large and great thrust events in the shallow thrust zone in Sanriku-Oki (Kawakatsu and Seno, 1983; see **Fig. 1.2**, Section 1.2) and their absence in Fukushima-Oki can not be explained by it alone. All the known large earthquakes in Fukushima-Oki such as the five M = 7.1 - 7.7 events in 1938 (Abe, 1977) and the recent sequence of M = 6.0 - 6.7 events in 1987 (Faculty of Science, Tohoku University, 1987) have occurred in the landward area where both the upper and lower plates are found to be seismically active through the OBS observations. This zone corresponds to the deep thrust zone proposed by Kawakatsu and Seno (1983) (see Section 1.2). No major event has occurred farther seaward (shallow thrust zone) in Fukushima-Oki.

MODES OF SUBDUCTION ALONG THE JAPAN TRENCH

In the context of the modes of subduction, there are two typical modes; Chilean-type and Mariana-type (e.g. Uyeda and Kanamori, 1979; Uyeda, 1982). The former is associated with a compressional- or high-stress regime and the latter with a tensional- or low-stress regime. The subduction mode has been discussed with respect to the stress state, interplate earthquakes, topography, dip of the deep seismic zone, volcanism etc. Uyeda (1982) states that the Japan Trench region (or the Northern Honshu arc) had been of Mariana-type and was changed to Chilean type only several million years ago. This indicates that the Japan Trench is near the middle of the two typical types rather than either extremes, and easily transforms its subduction mode.

The previous discussions on structure and seismicity lead to the recognition that within the subduction zone of the Japan Trench, Fukushima-Oki has seismologically more characteristics of Mariana-type than Sanriku-Oki. In contrast to Sanriku-Oki where large thrust earthquakes occur in the shallow thrust zone, Fukushima-Oki (and probably Boso-Oki, too) has the deeper trench, steeper dip of the subducting slab (at least in the shallow part) and lower seismicity in the shallow thrust zone. A deeper trench and a steeper subducting slab are regarded as the indications of mechanically easier subduction or less mechanical coupling between the two plates (Uyeda and Kanamori, 1979). Accordingly the low seismicity in the shallow thrust zone of Fukushima-Oki can be explained as an indication of the weak mechanical coupling between the two plates, which is the result of the observed steepening of the subducting slab.

7 CONCLUSIONS

Seismological regionality of Fukushima-Oki in the Japan Trench area was investigated through two earthquake observations and a seismic refraction profiling conducted on the landward slope of that region by using OBS arrays. Data from the total of about 50 days of microearthquake observations in 36.5 - 38 ^{o}N and the 200 km long explosion-OBS survey paralleling the trench at 37 - 38.5 ^{o}N were analyzed by the aid of an improved digital data processing system. The results were discussed by comparing them with those of the adjacent regions, i.e. Sanriku-Oki and Boso-Oki. The following conclusions have been obtained through the investigations:

(1) The seismic velocity structure along a zone of 3 - 4 km water depth (60 - 70 km landward from the Japan Trench axis) in Fukushima-Oki is characterized by the southward-thickening crust, which consists of a sedimentary layer, a 5.0 km/s layer, a 5.7 km/s layer, possibly a 6 km/s layer and a 6.8 km/s layer, from the top downward. The Moho deepens from 21 to 23.5 km.

(2) About 200 local earthquakes were located with errors less than 10 km. Seismicity in the descending oceanic plate is significantly low as far as 80 km landward from the trench axis, and that in the upper plate is also low within 50 km from the trench axis while it is high farther landward. This forms a striking contrast to Sanriku-Oki region where foci are distributed rather uniformly.

(3) The structure obtained for Fukushima-Oki smoothly connects

with those of the corresponding profiles of Sanriku-Oki and Boso-Oki. Along these profiles which span the length of the Japan Trench, the subducting Pacific plate deepens southward at around 38 ^{o}N, causing a steeper dip of the descending slab in the south of Fukushima-Oki than in Sanriku-Oki by several degrees. The 6.8 km/s layer is regarded as the subducting oceanic layer 3.

(4) This depth contrast at subduction along the trench seems to have been prepared before the subduction, as indicated by several hundred meters of water-depth contrast lasting as long as 1000 km eastward on the Northwest Pacific ocean floor probably due to a north-south contrast of the lithospheric density or thickness.

(5) The trenchward low seismicity and the southward-steepening of the dip of the subducting oceanic plate along the trench at Fukushima-Oki is indicative of a weak mechanical interaction (i.e. coupling) between the two plates in Fukushima-Oki in comparison with Sanriku-Oki.

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Table 4.1 Specifications of OBSs in 1982 experiment

OBS name	position lat. (°N)	long. (°E)	water depth (m)	geophone fo (Hz) vert.	horiz.	recorder type (day)	Acoustic or Timed
S16	37.5840	141.9978	500	3.0	4.5	25	A
S17	37.167	141.75	460	3.0	4.5	25	A
S18	36.9147	142.3998	2840	3.0	4.5	25	A
S19	37.2502	142.8305	3980	3.0	4.5	25	A
S20	37.5005	142.4992	1210	3.0	4.5	25	A

Table 4.2 Specifications of OBSs in 1985 experiment

OBS name	position lat. (°N)	long. (°E)	water depth (m)	geophone fo (Hz) vert.	horiz.	recorder type (day)	Acoustic or Timed
S1	36.7668	142.5283	4599	3.0	4.5	15	A
S2	36.9973	142.6650	3915	3.0	4.5	15	A
PR2	36.9962	142.6610	4090	3.0	4.5	20	T
P3	37.1677	142.7607	3945	3.0	4.5	15	T
PR3	37.1682	142.7612	3945	3.0	4.5	15	T
S4	37.2842	142.8347	3785	3.0	4.5	15	A
PR4	37.2847	142.8350	3822	3.0	4.5	20	T
P5	37.3992	142.8987	3555	3.0	4.5	15	T
PR5	37.4002	142.8985	3555	3.0	4.5	11	T
S6	37.5913	143.0162	2950	3.0	4.5	15	A
PR6	37.5912	143.0137	2939	3.0	4.5	20	T
P7	37.8077	143.1330	2735	3.0	4.5	11	T
PR7	37.8065	143.1338	2735	3.0	4.5	11	T
S8	38.0003	143.2530	3010	3.0	4.5	15	A
S9	38.1923	143.3698	3200	3.0	4.5	15	A
S10	38.4185	143.5032	3100	3.0	4.5	15	A
P11	37.2325	143.2660	5849	2.0	4.5	25	T
P12	37.0015	143.0835	5374	2.0	4.5	25	T
P13	37.5003	142.5525	1360	2.0	4.5	25	T

Table 4.3 List of explosions

shot name	charge size (kg)	position lat. (°N)	position long. (°E)	shot depth (m)	bottom depth (m)
K06	25	38.2102	143.3790	50	3228
H05	100	37.9577	143.2290	100	2645
K09	25	38.1177	143.3230	50	3107
E01	400	38.3678	143.4740	130	3142
K00	25	38.4183	143.5000	50	3097
K01	25	38.3930	143.4870	50	3084
K02	25	38.3473	143.4600	50	3170
K03	25	38.3240	143.4460	50	3172
H01	100	38.3047	143.4330	100	3205
K04	25	38.2782	143.4180	50	3060
K05	25	38.2558	143.4060	50	3222
E02	400	38.2307	143.3910	130	3185
K07	25	38.1850	143.3630	50	3153
H02	100	38.1628	143.3520	100	3138
K08	25	38.1388	143.3360	50	3134
H03	100	38.0967	143.3090	100	3032
K10	25	38.0753	143.2980	50	2996
K11	25	38.0497	143.2840	50	2956
H04	100	38.0275	143.2710	100	2933
K12	25	38.0035	143.2540	50	3009
K13	25	37.9815	143.2410	50	2791
K14	25	37.9345	143.2140	50	2551
K15	25	37.9127	143.2010	50	2527
H06	100	37.8877	143.1880	100	2516
E04	400	36.8132	142.5550	130	4245
E03	400	36.9268	142.6210	120	3885
K48	25	36.7708	142.5260	50	4425
K47	25	36.7907	142.5360	50	4403
K46	25	36.8357	142.5670	50	4170
H20	100	36.8602	142.5760	90	4050
K45	25	36.8798	142.5910	50	3923
K44	25	36.9040	142.6060	50	3893
K43	25	36.9517	142.6340	50	3913
K42	25	36.9738	142.6470	50	3941
H19	100	36.9983	142.6610	90	3913
K41	25	37.0208	142.6780	50	3970
K40	25	37.0425	142.6890	50	3938
H18	100	37.0643	142.7020	90	3847
K39	25	37.0895	142.7170	50	3846
K38	25	37.1102	142.7290	50	3895
H17	100	37.1335	142.7430	90	3925
K37	25	37.1562	142.7570	50	3987
K36	25	37.1798	142.7710	50	3907
H16	100	37.2017	142.7840	90	4058
K35	25	37.2252	142.7980	50	3948

K34	25	37.2485	142.8100	50	3807
H15	100	37.2713	142.8250	90	3777
K33	25	37.2940	142.8370	50	3798
K32	25	37.3177	142.8490	50	3627
H14	100	37.3395	142.8660	90	3610
K31	25	37.3625	142.8800	50	3615
K30	25	37.3860	142.8900	50	3563
H13	100	37.4077	142.9060	90	3534
K29	25	37.4322	142.9200	50	3507
K28	25	37.4572	142.9330	50	3425
H12	100	37.4775	142.9450	90	3288
K27	50	37.4987	142.9570	50	3193
K26	25	37.5230	142.9680	50	2974
H11	100	37.5462	142.9840	90	3022
K25	25	37.5695	142.9970	50	3011
K24	25	37.5907	143.0020	50	2914
H10	100	37.6155	143.0250	90	2870
K23	50	37.6397	143.0400	50	2776
K22	25	37.6602	143.0530	50	2686
H09	100	37.6835	143.0680	70	2780
K21	25	37.7068	143.0810	50	2692
K20	75	37.7295	143.0950	50	2759
H08	100	37.7502	143.1090	90	2805
K19	25	37.7742	143.1240	50	2844
K18	25	37.7970	143.1360	50	2750
H07	100	37.8200	143.1520	90	2700
K17	25	37.8427	143.1640	50	2664
K16	25	37.8670	143.1780	50	2615
K49	25	37.8927	143.1890	50	2524

Table 5.1 Result of time-term analysis of P_n data

OBS name	:	S2	P3	S4	S6	P7	S9	S10
time term (s)	:	2.90	2.77	2.71	2.75	2.69	2.76	2.73
error (s)	:	0.10	0.09	0.07	0.02	0.02	0.07	0.10

velocity : 8.26 $\pm$0.08 (km/s)

Table 5.2 Data and result of slope-intercept analysis

layer No.	intercept time (s)	ray parameter (s/km)	Vp (km)	thickness (km)	top depth (km)
(S2 north side)					
0	-	-	1.50	3.92	0.00
1	0.00	0.5000	2.00	1.90	3.92
2	1.75	0.1950	5.13	1.80	5.82
3	2.00	0.1840	5.43	2.31	7.62
4	2.25	0.1790	5.59	1.96	9.92
5	2.50	0.1750	5.71	1.29	11.88
6	2.74	0.1710	5.85	7.97	13.17
7	5.80	0.1211	8.26	-	21.14
(P3/PR3 north side)					
0	-	-	1.50	3.95	0.00
1	0.00	0.5000	2.00	1.64	3.95
2	1.50	0.2040	4.90	1.39	5.59
3	1.75	0.1870	5.35	1.80	6.98
4	2.00	0.1790	5.59	2.00	8.77
5	2.25	0.1740	5.75	9.81	10.77
6	5.54	0.1211	8.26	-	20.58
(P3/PR3 south side)					
0	-	-	1.50	3.95	0.00
1	0.00	0.5000	2.00	1.62	3.95
2	1.50	0.1920	5.21	1.94	5.57
3	1.75	0.1820	5.49	1.72	7.51
4	2.00	0.1750	5.71	11.55	9.23
5	5.54	0.1211	8.26	-	20.78
(S4 north side)					
0	-	-	1.50	3.79	0.00
1	0.00	0.5000	2.00	1.40	3.79
2	1.25	0.2260	4.42	1.05	5.19
3	1.50	0.2010	4.98	1.32	6.24
4	1.75	0.1880	5.32	1.56	7.56
5	2.00	0.1800	5.56	1.31	9.12
6	2.25	0.1730	5.78	2.16	10.42
7	2.50	0.1690	5.92	8.08	12.58
8	5.42	0.1211	8.26	-	20.67

(S4 south side)

0	-	-	1.50	3.79	0.00
1	0.00	0.5000	2.00	1.38	3.79
2	1.25	0.2090	4.78	1.39	5.16
3	1.50	0.1920	5.21	1.38	6.55
4	1.75	0.1810	5.52	1.90	7.93
5	2.00	0.1750	5.71	1.27	9.83
6	2.25	0.1690	5.92	1.06	11.10
7	2.50	0.1630	6.13	1.89	12.16
8	2.75	0.1590	6.29	8.28	14.05
9	5.42	0.1211	8.26	-	22.33

(P5/PR5 north side)

0	-	-	1.50	3.56	0.00
1	0.00	0.5000	2.00	1.64	3.56
2	1.50	0.2000	5.00	1.51	5.19
3	1.75	0.1850	5.41	1.59	6.70
4	2.00	0.1760	5.68	2.53	8.29
5	2.25	0.1720	5.81	9.38	10.82
6	5.45	0.1211	8.26	-	20.20

(P5/PR5 south side)

0	-	-	1.50	3.56	0.00
1	0.00	0.5000	2.00	1.39	3.56
2	1.25	0.2210	4.52	1.22	4.95
3	1.50	0.2010	4.98	1.16	6.17
4	1.75	0.1870	5.35	1.42	7.32
5	2.00	0.1780	5.62	2.02	8.74
6	2.25	0.1730	5.78	9.49	10.76
7	5.45	0.1211	8.26	-	20.25

(S6 north side)

0	-	-	1.50	2.95	0.00
1	0.00	0.5000	2.00	1.63	2.95
2	1.50	0.1930	5.18	2.04	4.58
3	1.75	0.1840	5.43	1.12	6.62
4	2.00	0.1740	5.75	4.16	7.74
5	2.25	0.1720	5.81	0.48	11.90
6	2.50	0.1680	5.95	1.07	12.38
7	2.75	0.1640	6.10	3.11	13.45
8	3.00	0.1620	6.17	4.20	16.56
9	5.50	0.1211	8.26	-	20.76

(S6 south side)

0	-	-	1.50	2.95	0.00
1	0.00	0.5000	2.00	1.62	2.95
2	1.50	0.1860	5.38	2.59	4.57
3	1.75	0.1800	5.56	1.57	7.15
4	2.00	0.1740	5.75	2.07	8.73
5	2.25	0.1700	5.88	3.35	10.80
6	2.50	0.1680	5.95	0.00	14.14
7	2.75	0.1630	6.13	0.82	14.14
8	3.00	0.1600	6.25	6.16	14.96
9	5.50	0.1211	8.26	-	21.12

(P7/PR7 north side)

0	-	-	1.50	2.74	0.00
1	0.00	0.5000	2.00	1.36	2.74
2	1.25	0.1930	5.18	1.95	4.09
3	1.50	0.1830	5.46	2.27	6.04
4	1.75	0.1780	5.62	1.96	8.31
5	2.00	0.1740	5.75	1.42	10.27
6	2.25	0.1700	5.88	8.34	11.69
7	5.38	0.1211	8.26	-	20.03

(P7/PR7 south side)

0	-	-	1.50	2.74	0.00
1	0.00	0.5000	2.00	1.10	2.74
2	1.00	0.2070	4.83	1.64	3.83
3	1.25	0.1940	5.15	1.83	5.48
4	1.50	0.1870	5.35	1.00	7.31
5	1.75	0.1790	5.59	2.88	8.31
6	2.00	0.1760	5.68	8.47	11.19
7	5.38	0.1211	8.26	-	19.66

(S9 south side)

0	-	-	1.50	3.20	0.00
1	0.00	0.5000	2.00	1.37	3.20
2	1.25	0.2080	4.81	1.44	4.57
3	1.50	0.1920	5.21	1.35	6.02
4	1.75	0.1810	5.52	1.43	7.37
5	2.00	0.1730	5.78	1.56	8.80
6	2.25	0.1670	5.99	2.03	10.36
7	2.50	0.1630	6.13	9.43	12.39
8	5.52	0.1211	8.26	-	21.82

(S10 south side)

0	-	-	1.50	3.10	0.00
1	0.00	0.5000	2.00	1.38	3.10
2	1.25	0.2120	4.72	1.48	4.48
3	1.50	0.1970	5.08	1.54	5.96
4	1.75	0.1880	5.32	1.71	7.50
5	2.00	0.1820	5.49	1.19	9.20
6	2.25	0.1760	5.68	1.89	10.40
7	2.50	0.1720	5.81	5.64	12.29
8	2.75	0.1710	5.85	0.00	17.92
9	3.00	0.1640	6.10	2.05	17.92
10	5.46	0.1211	8.26	-	19.97

Table 5.3 Parameters of sedimentary layer

(V_p/V_s = 3.0 and V_p = 2.0 km/s)

OBS name	t_{ps-p} (s)	t_p (s)	t_s (s)	thickness (km)	water depth (m)	5.0 km/s depth (m)
S16	2.2	1.10	3.30	2.2	500	2700
S17	1.7	0.85	2.55	1.7	460	2160
S18	2.0	1.00	3.00	2.0	2840	4840
S19	1.9	0.95	2.85	1.9	3980	5880
S20	2.0	1.00	3.00	2.0	1210	3210
S2	1.9	0.95	2.85	1.9	3915	5815
PR2	1.9	0.95	2.85	1.9	4090	5990
S4	1.9	0.95	2.85	1.9	3785	5685
PR4	1.9	0.95	2.85	1.9	3822	5722
S6	2.3	1.15	3.45	2.3	2950	5250
PR6	2.3	1.15	3.45	2.3	2939	5239
P11	1.5	0.75	2.25	1.5	5849	7349
P12	1.3	0.65	1.95	1.3	5374	6674
P13	2.0	1.00	3.00	2.0	1360	3360