VELOCITIES OF *P* AND *S* WAVES FOR DRILLING CORE ROCKS AT SYOWA STATION, ANTARCTICA

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Abstract: Velocities of compressional and shear waves for the drilling core samples at Syowa Station, East Antarctica are measured at confining pressure up to 0.5 GPa (5 kbar). The samples consist predominantly of granitic complex or gneiss being older than 350–533 m.y. in the shield region. The *P*-wave velocities for three orthogonal directions and the *S*-wave velocities for two orthogonal directions are measured to reveal anisotropy of the rocks. The velocities increase rapidly with increasing pressure at low pressures below 50–120 MPa due to crack closure. Whereas, the velocities increase linearly with increasing pressure at higher pressures than 50–120 MPa, and the velocities for each direction are nearly the same.

A velocity model in the upper 20 km of the crust is estimated from the laboratory data, assuming an appropriate temperature profile. The proposed model of *P*-wave velocity increasing gradually with depth is consistent with the upper crustal structure obtained by the explosion seismic experiments in the Ongul Islands and the Mizuho Plateau, East Antarctica.

1. Introduction

Many studies of rocks for geology, geochemistry and geomagnetism have been made by the exertion of many Antarctic researchers. Only a few studies, however, have been performed on the elastic properties of rocks in the vicinity of Syowa Station since 1978. The explosion seismic experiments in the Ongul Islands, Prince Olav Coast and Mizuho Plateau have been conducted by the 20th and 21st Japanese Antarctic Research Expedition (JARE-20, -21) in 1979–1981, and the velocity structures of the crust have been reported by IKAMI *et al.* (1981, 1984) and ITO *et al.* (1984). The core samples, on the other hand, were obtained from the core drilling at Syowa Station for the installation of a borehole tiltmeter and for the under-ground temperature observations by JARE-21 and -22 during 1980–1982.

Presence of open cracks in rocks on the ground surface is important to geophysics such as seismic velocity, gravity, heat-flow, tectonic stress, etc. Many studies of crack behaviors have been attempted to solve the problems by differential strain analysis (SIMMONS *et al.*, 1974), scanning electron microscopy (PADOVANI *et al.*, 1982), and crack density with elastic wave velocity (O'CONNELL and BUDIANSKY, 1974).

We measured apparent porosity of the seven core samples from the drillings, and measured *P*- and *S*-wave velocities of two or three orthogonal directions for the selected four samples at confining pressure up to 0.5 GPa. The laboratory data are converted into the velocity-depth profile of the upper crust on the assumption of a suitable pressure and temperature variation with depth. The proposed model is compared with the results of the explosion seismic experiments by JARE and by the research expeditions of other countries.

2. Experimental Method

Oil was used as a pressure transmitting medium for a 0.5 GPa high pressure vessel. Confining pressure was measured with a pressure transducer of strain-gage type within uncertainties of 0.5%. The working space in the cylindrical pressure vessel is 140 mm long and 70 mm in diameter.

The samples prepared from the drilling cores were cut out to cylinders of 65 mm long and 29 mm in diameter, which were dried for more than 24 hours at 110°C. The accuracy of length measured by a dial caliper is within 0.01 mm. The small correction (0.6%/GPa) of length with pressure was made by typical compressibility of granite listed by CLARK (1966). The sample was jacketed with a 0.08 mm thick Bytac film made of alminium foil protected by overlay of teflon, besides shielded with synthetic rubber.

The velocity measurements were made by the pulse transmitting method. Lead-zirconate-titanate transducers (PZT) of compressional and shear modes used were 6 mm square, and those resonance frequencies were 5 and 2 MHz, respectively. Eight transducers were fitted to the surface of the sample with stainless steel attachments. A VHF-switch selected pairs of transmitting and corresponding receiving transducers in sequence, and the velocities of four paths as shown in Fig. 1b were measured.

Schematic diagram of the measuring system of velocity is shown in Fig. 1a. All instruments were controlled by a microcomputer (Hewlett-Pacard 9836) connected

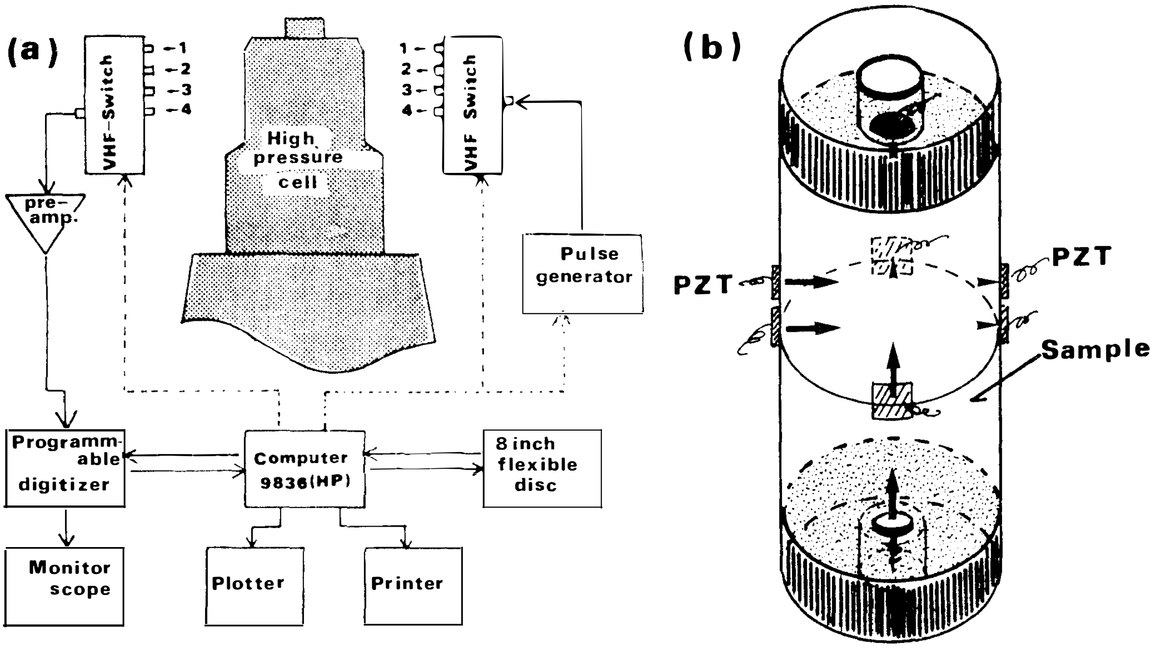

Fig. 1. (a) Schematic diagram of the experimental system. (b) Arrangement of transmitting and receiving transducers of sample assembly in the high pressure vessel.

with GP-IB interface. A pulse of 0.7 μs wide and 10 V high emitted by the pulse generator was applied to one side of the transducer to excite longitudinal or transverse vibration. The elastic waves were picked up by another transducer mounted on the other side of the sample, and then converted into digital signals by a programmable digitizer (Sony-Tektronix 390-AD) with a sampling frequency of 60 MHz. The elastic wave signals averaged over five times shots were stored in an 8-inches flexible disc. The precision of velocity measurements is better than 0.5%.

3. Samples and Geology

The location of the drilling holes is shown in Fig. 2. A hole of 30 m deep and 50 mm in diameter was drilled at site I for the underground temperature observation (KAMINUMA, 1983). The seven samples of SP-B to SP-H as shown in Fig. 3 were selected from a 30-m core. The sample of SP-V was also obtained from the other core at site II, where a borehole tiltmeter was set in a hole of 140 mm in diameter and 3 m deep.

The Ongul Islands situated in the northeast of Lützow-Holm Bay are composed

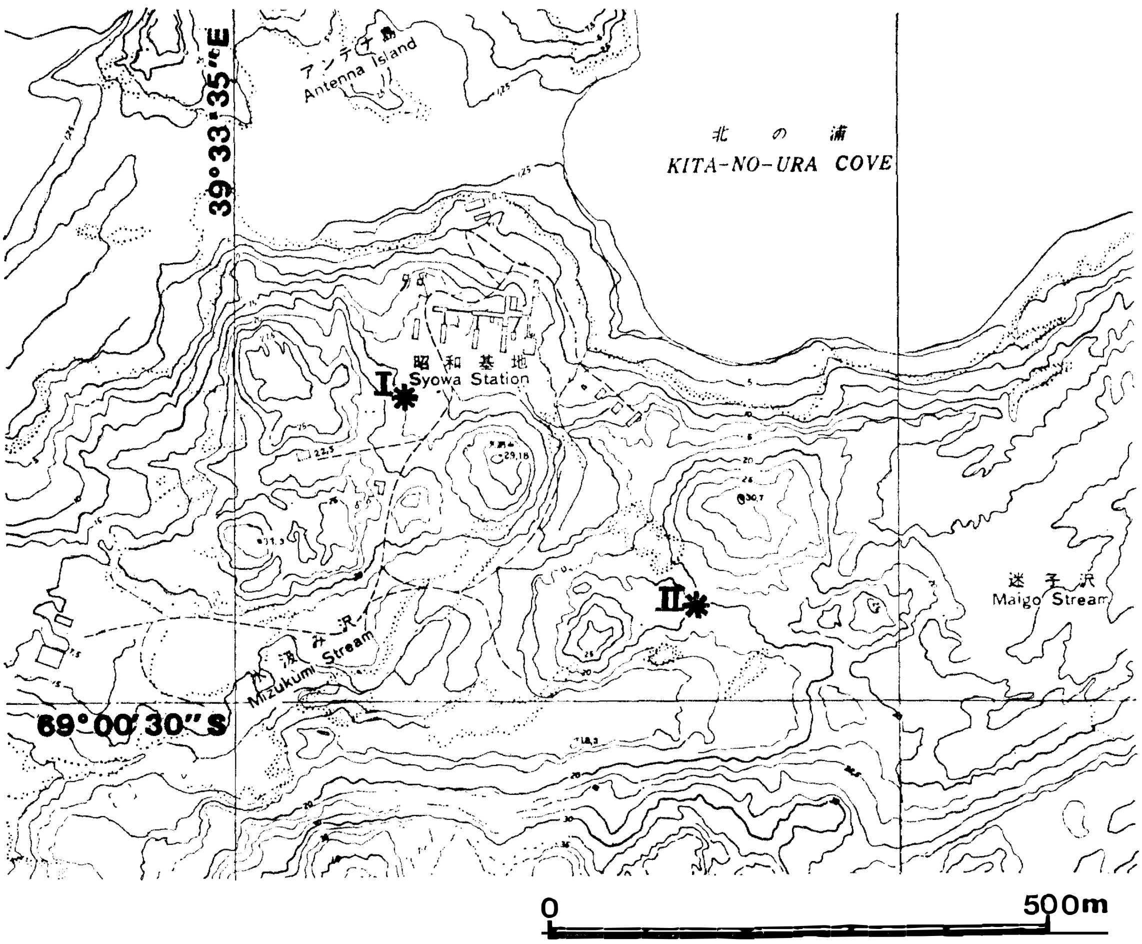

Fig. 2. Sample locations in East Ongul Island. Site I, a hole of 30 m deep was drilled for the underground temperature measurements. Site II, a hole of 3 m deep and 140 mm in diameter was drilled for the installation of a borehole tiltmeter.

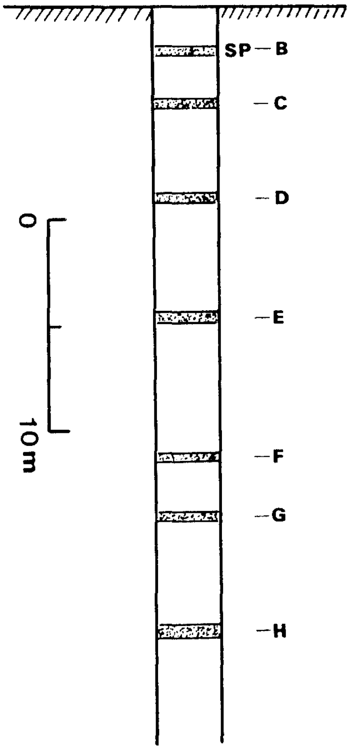

Fig. 3. Depth of samples selected from the 30-m core at site I in Fig. 2.

Table 1. Density, porosity and major mineralogical constituents.

Symbol of sample	Depth of sample (m)	Bulk density (g/cm^3)	Apparent porosity (%)	Major mineralogical constituents
SP-B	2.1	2.686	0.3	Quartz (rich), feldspar, biotite
SP-C	4.4	2.612	1.0	Quartz (rich), feldspar, biotite
SP-D	9.1	2.822	0.6	Quartz, feldspar, magnetite, hornblende (rich)
SP-E	14.6	2.775	0.6	Quartz, feldspar (rich), biotite
SP-G	23.8	2.618	0.7	Quartz, feldspar, hornblende
SP-H	29.1	2.668	0.8	Quartz, feldspar, biotite, hornblende
SP-V	2	2.664	1.8	Quartz, feldspar, garnet

mainly by garnet gneiss, hornblende gneiss and pyroxene gneiss, and patches of feldspathic gneiss and hornblende biotite gneissose granite also occur in this region (YANAI *et al.*, 1974). Ages of the rocks determined by K-Ar method ranged from 350 to 533 m.y. (KANEOKA *et al.*, 1968; YANAI and UEDA, 1974). The appearance of all samples is granitic complex or gneiss, but each sample differs in details. The mineralogical constituents of the samples observed by a polarizing microscope are listed in Table 1.

Porosity and density of the samples are also listed in Table 1. Apparent pore volume was measured by an immersion technique. Bulk volume was obtained from the cylindrical sample, and apparent porosity was calculated from the bulk volume and the pore volume. This technique underestimates the pore volume, because of no intruding water into occluded pores. The fair scatter of density as discerned in Table 1 depends on the difference in content of heavy minerals such as garnet, hornblende and magnetite.

4. Experimental Results

The results with velocities are summarized in Table 2. Z-V_p is P-wave velocity

Table 2. Elastic velocities at hydrostatic pressure and pressure derivatives.

		SP-C	SP-E	SP-G	SP-V
Velocity at pressure of 1 MPa interpolated	Z-V_p	4.6	4.8	4.8	5.1
	$R1$-V_p	3.8	4.4	3.9	4.4
	$R2$-V_p	3.3			4.4
	$R1$-V_s		2.8	3.0	
	$R2$-V_s	2.6	2.8	(/)*	3.0
Velocity at pressure of 500 MPa extrapolated	Z-V_p	6.00	6.64	6.28	6.25
	$R1$-V_p	6.06	6.75	6.38	6.20
	$R2$-V_p	5.93			6.28
	$R1$-V_s		(/)*	(/)*	
	$R2$-V_s	3.63	3.60	3.51	3.62
Pressure derivative (km/s/GPa)	Z-V_p	0.40	0.76	0.39	0.39
	$R1$-V_p	0.63	0.68	0.56	0.37
	$R2$-V_p	0.47			0.44
	$R1$-V_s		(/)*	(/)*	
	$R2$-V_s	0.47	0.54	0.35	0.29

Velocity is in km/s. * Data were not obtained owing to accidents in the experiment.

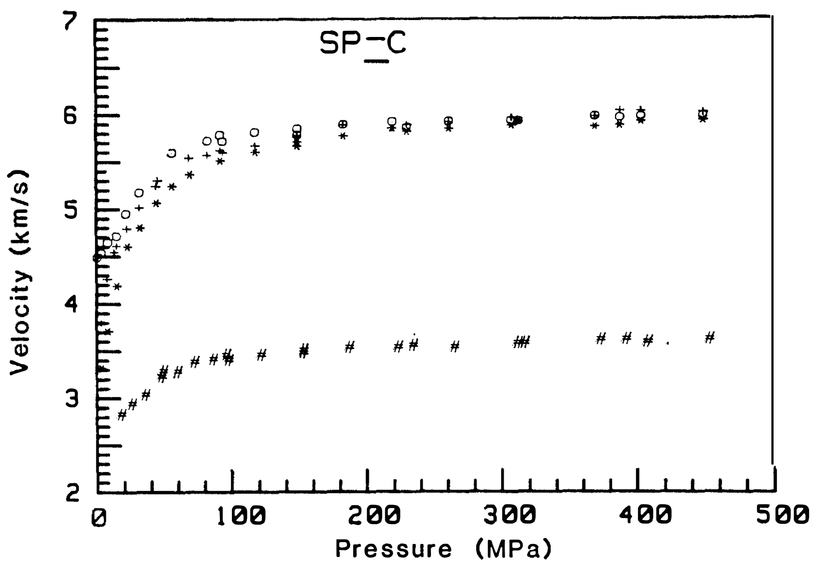

Fig. 4. P-wave and S-wave velocities for sample SP-C as a function of pressure. Symbol (⊙) is P-wave velocity for axial direction of cylinderical sample (Z-V_p), symbols (+) and (∗) P-wave velocity for two orthogonal radial directions ($R1$-V_p, $R2$-V_p), respectively, and symbol (#) S-wave velocity for radial direction ($R2$-V_s).

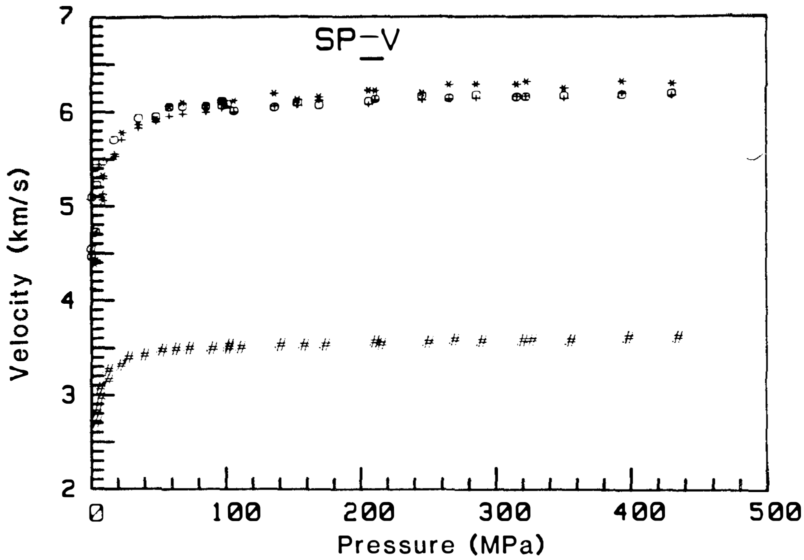

Fig. 5. P- and S-wave velocities for sample SP-V as a function of pressure. For explanation of notation, see Fig. 4.

for axial direction of the cylinderical sample, and $R1$-V_p and $R2$-V_p are P-wave velocities for two orthogonal radial directions, respectively. $R1$-V_s and $R2$-V_s are S-wave velocities for the same radial directions, respectively. All runs of the experiments completely came to an end on samples SP-C and SP-V. These velocities for each path as a function of pressure are shown in Figs. 4 and 5, respectively. The velocities for SP-V increase rapidly for all paths up to the pressure of 30 MPa, and those for SP-C up to the pressure of 100 MPa, because of the rapid crack closure with increasing pressure. The velocities at the pressure of 1 MPa as the lowest pressure of the present experiment are listed in Table 2 for the reason of completing a surface of contact between the sample and the transducer. The velocities at low pressures fairly differ in three directions. These differences are caused by anisotropic distributions of very thin ellipsoidal cracks, which is more or less seen in such an experiment as, for example, by SIEGFRIED and SIMMONS (1978).

The pressure derivative is obtained by fitting a straight line to the data above pressures of 50 to 120 MPa. The extrapolated values at the pressure of 500 MPa for three directions of a cylinderical sample are almost the same. Probably, a large anisotropy may not exist in the upper crust. As the velocities at high pressures are no more affected by crack closure, the velocity difference with the samples at a high pressure reflects the difference in mineral composition as shown in Table 1.

5. Discussion

A linear relation between the density and the P-wave velocity is widely used for calculating a gravity anomaly. KURININ and GRIKUROV (1982) used a correlation between the density and the P-wave velocity as shown in Fig. 6 to evaluate the gravity anomaly for the crustal structures revealed in the explosion seismology in East Antarctica. The apparent density calculated from the pore volume and the P-wave

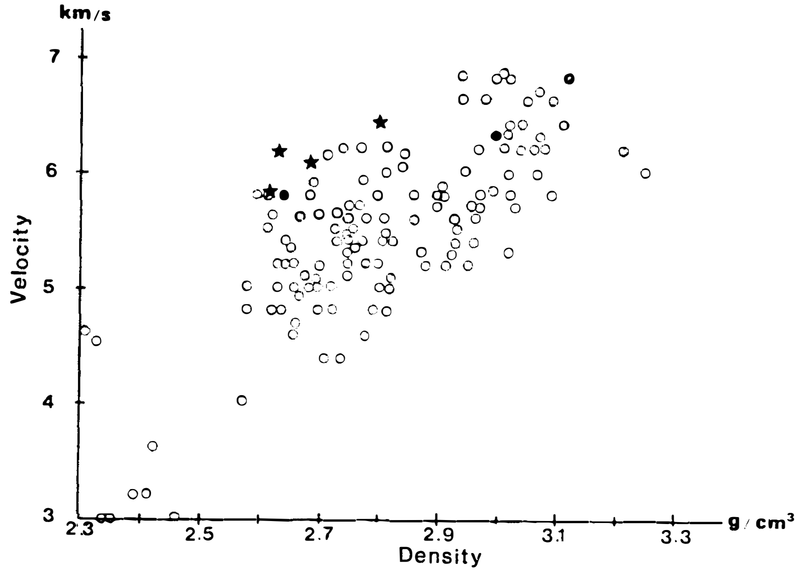

Fig. 6. Correlation between the density and P-wave velocity of rocks in East Antarctica after KURININ *and* GRIKUROV *(1982). The present data are indicated by asterisks.*

velocity extrapolated at porosity 0% is plotted in Fig. 6 with asterisks. Our velocities are higher than those by KURININ and GRIKUROV at the same density, because the latter values were obtained mainly at the atmospheric pressure and at 60 MPa at most.

If the pressure (P) and temperature (T) profiles in the upper crust are given, the P-wave velocity as a function of depth (h), $V_p(h)$ can be calculated from the laboratory data. On the assumption of a constant mineral composition,

$$V_p(h)=V_p(0)+\int_0^h\left[\left(\frac{\partial V_p}{\partial P}\right)_T\cdot\frac{dP}{dh}+\left(\frac{\partial V_p}{\partial T}\right)_P\cdot\frac{dT}{dh}\right]\cdot dh\,. \tag{1}$$

If the pressure profile in the crust is assumed to be hydrostatic, $P(h)$ is represented by the following equation,

$$P(h)=\int_0^h \rho g dh\,, \tag{2}$$

where ρ is density and g is gravitational constant. The pressure dependence of velocity with depth can be calculated from the pressure derivatives in Table 2 and the pressure variation with depth in the upper crust estimated from eq. (2). The pressure profile at a low pressure is deduced from the slope of the smoothed curve which is fitted to the velocity-pressure data as shown in Figs. 4 and 5.

The temperature profile in the crust of the Baltic Shield by LUBIMOVA (1967) is appropriately used as the geothermal gradient of the shield region in East Antarctica. We also referred to the values (6.12–8.89×10^{-4} km/s/°C) of quartz monzonite and hornblende gabbro by LIN and WANG (1980) as the approximate temperature derivative. The calculated $V_p(h)$ and $V_s(h)$ and the used temperature profile (12.5°C/km) are shown in Fig. 9a. The stippled regions in Fig. 9a indicate the velocity range obtained from the experiments of samples SP-E, SP-G and SP-V for two or three orthogonal directions, except the lowest curve of the sample SP-C including no heavy minerals.

The small values of the velocity near the surface, 1–2 km in depth, are caused by cracks of the dry sample as mentioned in the previous section. Such a small value of velocity may not occur *in situ*, because cracks would be filled with water of ice. The *P*-wave velocity near the surface of an ice-free area in the Ongul Islands was determined from the explosion seismic experiments with the very short line of 1 km and the line of 5 km as shown in Fig. 7a (ITO *et al.*, 1984). The results indicate that the *P*-wave velocity from the surface to a depth of 1–2 km depth is nearly 6.0 kms. The values are fairly well consistent with the velocities in the laboratory data extrapolated to porosity 0%.

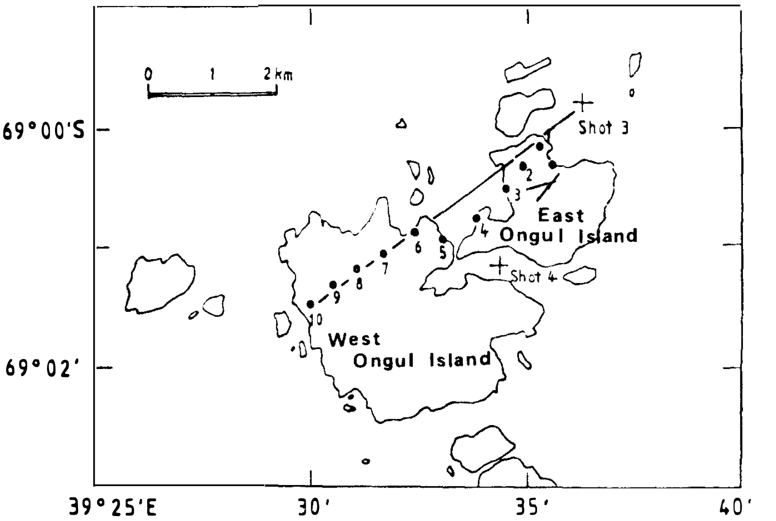

Fig. 7a. Location of profile lines of seismic experiments conducted in the ice-free area of the Ongul Islands near Syowa Station, East Antarctica.

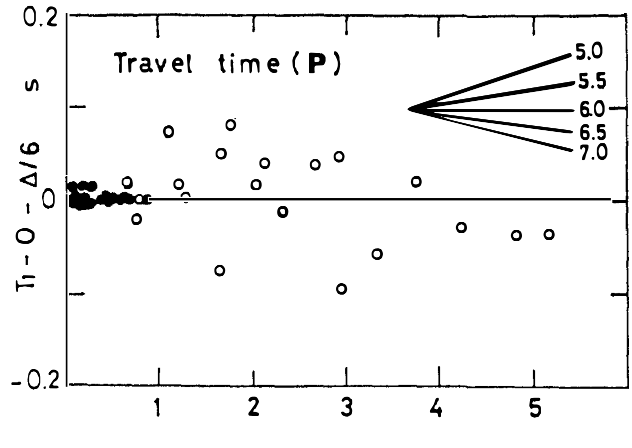

Fig. 7b. Reduced travel times of the seismic experiments in the Ongul Islands. Solid circles are the travel times of short profiles and open circles are those of a 5 km long profile after ITO *et al. (1984). The P-wave velocity in the ice-free area is about 6 km/s.*

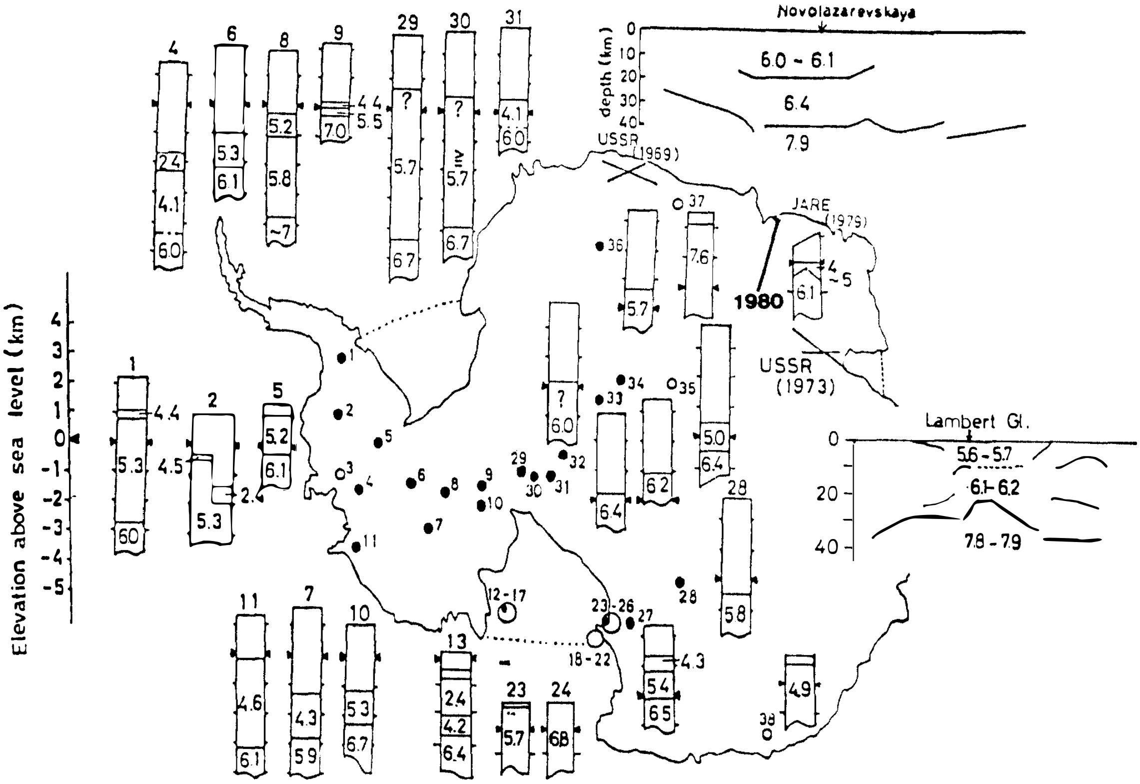

Fig. 8. Location of the seismic profiles and the P-wave velocity in km/s, in Antarctica, by Antarctic research expeditions of USA, USSR and Japan. Notice that a layer with velocities 5.7–6.4 km/s exists at most in East Antarctica.

The P-wave velocity of the crust was inferred from the explosion seismic experiment (IKAMI *et al.*, 1984). The travel times of the experiments are shown in Fig. 9b. The apparent velocities are 6.2–6.3 km/s at short distance from the shot points and become larger with increasing distances, which indicates that the velocity of the upper crust increases with depth as illustrated in Fig. 9a.

Seismic velocities in Fig. 8 are after BENTLEY and CLOUGH (1972), the results of USSR (KOGAN, 1972; KURININ and GRIKUROV, 1982) and Japan (IKAMI *et al.*, 1984). The P-wave velocity of 5.7–6.4 km/s in the upper crust is widely distributed, and the layer of velocity less than 5.7 km/s does not exist in East Antarctica. The model calculated from the laboratory data is passable to the upper crust of East Antarctica.

The velocity increases rapidly with increasing pressure up to 30 MPa for sample SP-V and up to 100 MPa for sample SP-C due to crack closure. Sample SP-C consists mainly of quartz and feldspar, and its mineral composition is nearly granite. YUKUTAKE and SHIMADA (1982) reported that cracks did not close completely up to a pressure of 200 MPa for the granite of 60–100 m.y. from Setonaikai, Japan. Crack closure for the present samples of 350–533 m.y. in the shield region occurs at much lower pressures than those for younger rocks in Japan. It is interesting to study in further details the response of crack closure with pressure, for the process of reaction of igneous rocks, for the age of rocks and for the grade of metamorphism by use of the rocks in Antarctica.

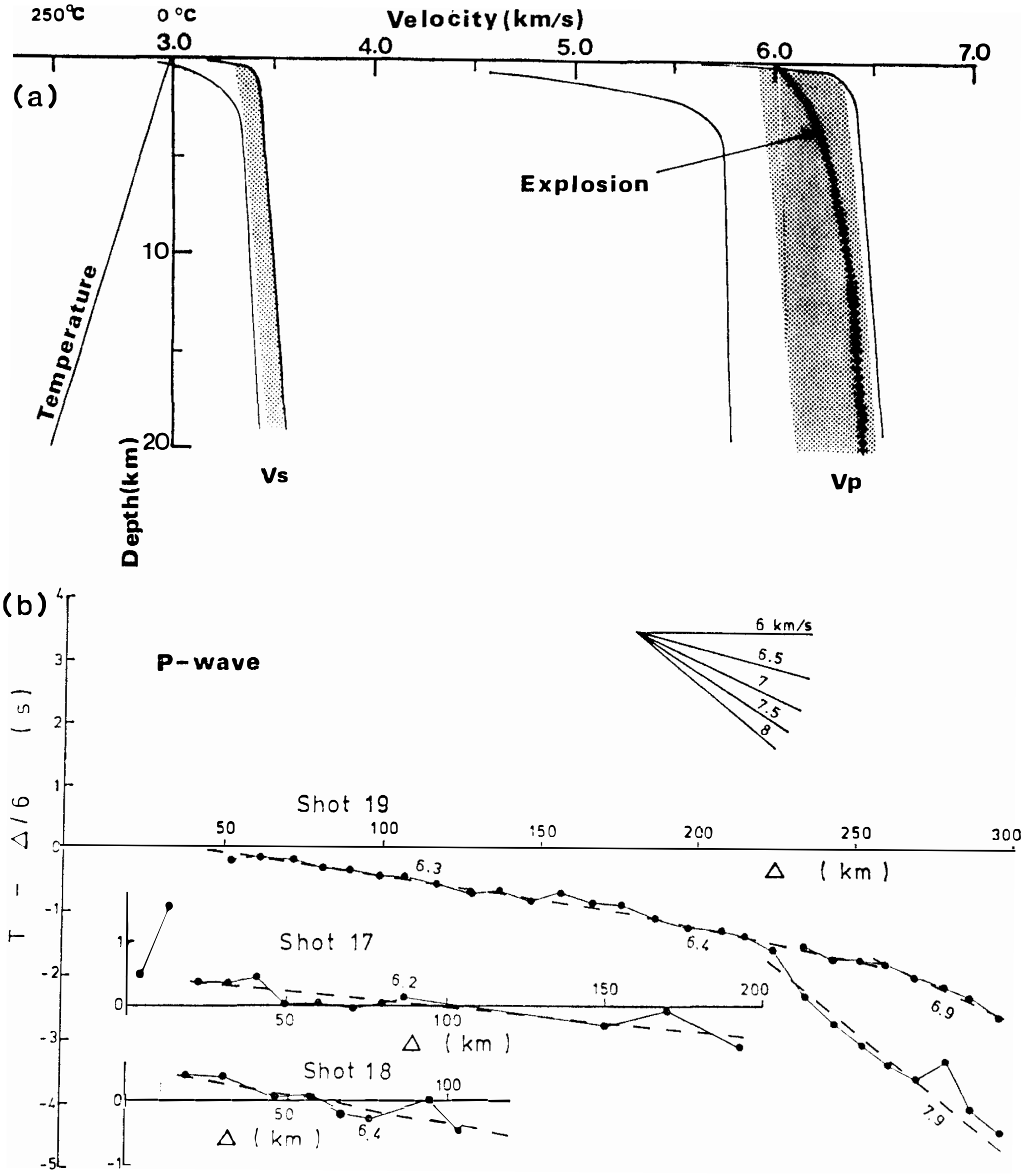

Fig. 9. (*a*) *Assumed temperature profile, and P- and S-wave velocities of the upper crust calculated from the laboratory data. The P-wave velocity from the explosion seismic experiments in the Mizuho Plateau is also given in the figure.* (*b*) *Reduced travel time plots for large explosion shots 17, 18 and 19 after* IKAMI *et al.* (*1984*). *Numerals attached to dashed lines are the apparent velocities in km/s.*

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