Compressional Wave Velocities in Rocks at High Temperatures and Pressures, Critical Thermal Gradients, and Crustal Low-Velocity Zones

NIKOLAS I. CHRISTENSEN

Department of Geological Sciences and Graduate Program in Geophysics, University of Washington, Seattle, Washington 98195

The velocities of compressional waves have been determined for several igneous and metamorphic rocks to temperatures of 500°C at elevated confining pressures. At 2 kbar and between 25° and 300°C, temperature coefficients $(\partial V_p/\partial T)_P$ for many of the rocks range between -0.5×10^{-3} and -0.6×10^{-3} km s^{-1} °C^{-1}. At higher temperatures and 2 kbar the temperature coefficients show extreme variability, which is related to the opening of grain boundary cracks caused by anisotropic thermal expansion of the mineral components. Critical thermal gradients $(dT/dZ)_c$ for a low-velocity layer in the continental crust at pressures of 5–8 kbar are between 10° and 14°C/km. These values are probably high because of porosity; however, they are still lower than estimated crustal temperature gradients in normal and high heat flow provinces. Thus it is concluded that crustal velocity inversions produced by high temperature are likely to be common within the crust. It is significant, however, that the velocity decreases reported for some crustal low-velocity layers are much greater than the decrease which can be accounted for by temperature alone. At higher pressures (10–30 kbar), $(dT/dZ)_c$ = 6.3°C/km for dunite and eclogite, in excellent agreement with estimates from single-crystal data. The velocities in dunite as a function of temperature further support the conclusion that the observed increase in upper oceanic mantle velocity with age is a consequence of decreasing temperature.

INTRODUCTION

Seismology has provided valuable information on the nature of the earth's crust and upper mantle in the form of velocity distributions with depth. Crustal structure models often are based on the analyses of first-arrival time data, and derivations of velocity-depth relations result in homogeneous layered models. Recently, more realistic crustal models are being obtained which, in addition to first arrivals, satisfy secondary arrival as well as amplitude constraints. An important feature often found in these newer models is a velocity inversion in the crustal velocity-depth curve.

The geological interpretation of seismic velocities requires laboratory data on elastic wave propagation under in situ conditions for a wide variety of earth materials. In comparison with measurements at elevated pressures, the influence of temperature on seismic velocities has received little attention. The measurements reported in this study were undertaken to provide basic temperature data for a wide variety of igneous and metamorphic rocks believed to be major constituents of the earth's crust and upper mantle. Critical thermal gradients, which, when exceeded, produce velocity reversals, are calculated for the samples, and these results are used to evaluate the possibility that low-velocity zones within the crust originate from the decrease in velocity produced by increasing temperature.

EXPERIMENTAL DETAILS

Compressional wave velocity measurements were made in 13 samples as a function of hydrostatic pressure and temperature. The containing vessel consisted of a 46-cm-o.d. cylinder with a working cavity 18.5 cm long and 9.5 cm in diameter. The vessel was designed to accommodate a three-zone internal furnace (Figure 1) which allowed rapid heating and minimized temperature gradients within the rock samples. Pressure was generated by a standard two-stage intensifier using nitrogen as a pressure medium and was measured from the change in resistance of a calibrated manganin coil. The electronics were similar to those described in several previous velocity studies [e.g., *Birch*, 1960; *Christensen*, 1965; *Christensen and Shaw*, 1970]. A rectangular pulse of approximately 50 V was applied to a 2-MHz lithium niobate transducer on one end of a cylindrical sample. The time required for a compressional pulse to arrive at a second transducer on the opposite end of the sample was obtained with a dual-trace oscilloscope and calibrated variable length mercury delay line. The velocities were simply calculated from transit lines and sample lengths.

To exclude nitrogen from the pore spaces, the samples were jacketed with stainless steel tubes with wall thicknesses of 0.028 cm, machined from solid rod. Brass screens, 0.0254 cm thick and 100 mesh, were placed between the samples and the jackets (Figure 2), and machined stainless steel caps with wall thicknesses of 0.058 cm were placed within the jacket ends. This construction permitted welding the caps in place without excessive heating of the sample and assured no leakage of the pressure medium into the rock pore spaces. The transducers were held onto the end caps with spring pressure.

The rock cores were 1.626 cm in diameter and 3.208–3.315 cm long. The furnace windings were constructed of 0.05-cm-diameter nichrome V wire with resistance of 9.5 ohm for the center zone and 7 ohm for end zones. At 2-kbar confining pressure, approximately 500 W were required for heating to 500°C.

During each run the distribution of temperature along the length of the sample was investigated with three iron constantan thermocouples, and power to each of the three furnaces were adjusted to obtain a minimum temperature gradient. The difference in temperature between the two end thermocouples increased with mean temperature but did not exceed 1%. The accuracy of the temperature measurements at the junction of each thermocouple was about 1%.

Asbestos insulation was packed between the furnace and inner vessel wall, which allowed temperature equilibrium to be obtained rapidly and eliminated the need for external cooling of the pressure vessel. Confining pressures were first increased to 2 kbar, and temperature was then elevated at 25°C inter-

Paper number 9B0997.
0148-0227/79/009B-0997$01.00

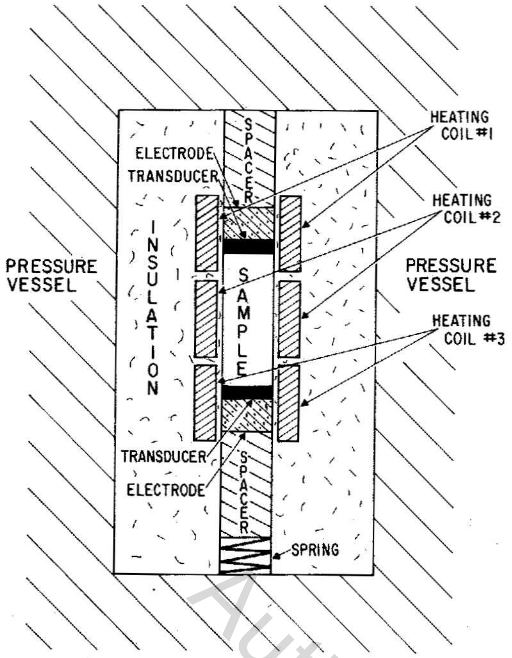

Fig. 1. Furnace and sample arrangement.

vals. Approximately 30 min was allowed for the rock specimens to reach the furnace temperature, and the pressure was adjusted to 2 kbar after each temperature change. During several runs the specimens were held at constant elevated pressures and temperatures for over 24 hours, and no changes in velocities were observed.

SAMPLE DESCRIPTIONS

The samples selected for this study represent a wide variety of rocks believed to be abundant constituents of the crust and upper mantle. Samples which were dredged from the ocean floor include basalt, serpentinite, gabbro, and amphibolite. Quartzite, granite, anorthosite, and several granulites with differing compositions were selected as representative continental crustal rocks. Possible upper mantle rocks selected for measurement include dunite and eclogite.

Velocities as a function of pressure, usually to 10 kbar and for selected samples to 30 kbar, have been reported in several previous investigations for cores from many of the same rocks of this study [*Christensen*, 1972, 1974; *Christensen and Fountain*, 1975]. The samples were carefully selected for (1) a minimum anisotropy, (2) a tight grain boundary structure, and (3) minimum alteration. Anisotropy was investigated in each sample by petrographic observations, and velocity measurements to 10 kbar in three perpendicular cores taken from the same specimens in which the cores were cut for the temperature measurements. For all samples, anisotropy was less than 5% and averaged about 1%. Petrographic and hand specimen examinations of the samples showed no visible weathering or fracturing. This is supported by the bulk densities of the cores from which the velocities were measured (Table 1).

DATA

Velocities measured at 2-kbar confining pressure as a function of temperature are shown in Figures 3 and 4. Although data were recorded at 25°C intervals at increasing and decreasing pressures, velocities are only shown for 50°C increments. Each core was first seasoned at 2 kbar by slowly cycling the temperature from 25° to 500°C and then down to 25°C. For many of the samples, velocities recorded during the first cycle did not differ appreciably from the velocities shown in Figures 3 and 4. However, slight increases in velocities were observed after initial cyclings of the granite, anorthosite, and granulites which are likely a result of dehydration of clay minerals present in minor amounts along grain boundaries and in fractures. Similar increases in velocities with heating have been reported by *Birch and Bancroft* [1938], who attributed the velocity increases to changes in internal stress state, and *Hughes and Jones* [1950], who suggested that heating produces permanent consolidation of rocks. The screens, which are wrapped between the samples and jackets, serve to trap released water and at the same time prevent pore pressure buildup within the samples.

Corrections for changes in length of the specimens due to thermal expansion have been made for the velocities in Figures 3 and 4. On the basis of data for the linear expansions of common rock forming minerals [*Skinner*, 1966] this correction averages 0.03 km/s at 500°C and is only significant in obtaining temperature derivatives of velocities.

DISCUSSION

Early attempts to measure velocities at elevated temperatures and atmospheric pressure showed large hysteresis effects. For example, *Ide* [1937] reported permanent decreases in velocities for rocks heated to 200°C and attributed the irreversible velocity decreases to the formation of cracks formed by thermal expansion of anisotropic minerals. Later studies [e.g., *Birch*, 1943; *Hughes and Jones*, 1950; *Hughes and Cross*,

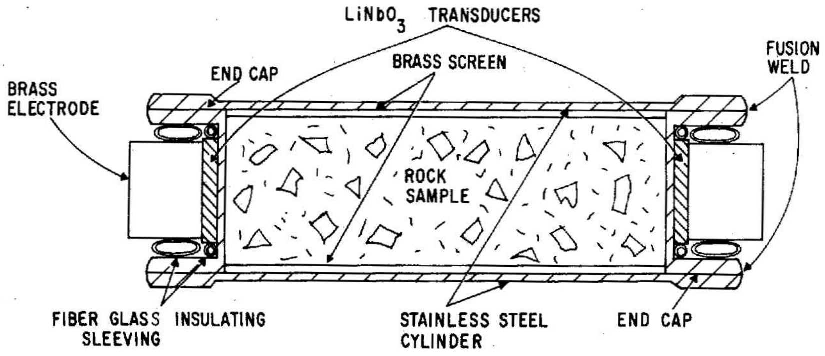

Fig. 2. Sample-jacket-transducer assembly.

TABLE 1. Sample Descriptions

Rock	Petrography and Density
Serpentinite, Mid-Atlantic Ridge	Massive with mesh structure and bastite pseudomorphs; 98% chrysotile and lizardite, 2% opaque; $\rho = 2.509$ g/cm^3.
Granite, Cape Ann, Massachusetts	Holocrystalline, coarse grained, inequigranular; 30% quartz, 40% perthite, 20% plagioclase (An_{12}), 10% biotite intergrown with opaque mineral, traces of zircon and tourmaline; $\rho = 2.621$ g/cm^3.
Quartzite, Baraboo, Wisconsin	Granulose, granoblastic, xenoblastic; 95% quartz with undulatory extinction, 4% muscovite, 1% graphite; $\rho = 2.647$ g/cm^3.
Granulite, New Jersey Highlands	Granulose, hypidioblastic, inequigranular; 30% quartz, 50% plagioclase (An_{30}), 17% hypersthene, 2% opaque mineral, 1% hornblende, trace of biotite; $\rho = 2.680$ g/cm^3.
Granulite, Saranac Lake, New York	Granulose, coarse grained, xenoblastic inequigranular; 35% perthite, 15% myrmekite, 2% plagioclase, 3% quartz, 15% garnet, 13% clinopyroxene (altered), 5% biotite, 4% hornblende, 4% opaque, 3% apatite, 1% zircon; $\rho = 2.848$ g/cm^3.
Basalt, East Pacific Rise	Intergranular, variolitic; 60% plagioclase (subradiating laths), 20% pyroxene, 20% groundmass; $\rho = 2.882$ g/cm^3.
Gabbro, Mid-Atlantic Ridge	Holocrystalline, medium grained; 50% plagioclase (An_{50}), 35% clinopyroxene, 15% olivine partially replaced by serpentinite, tremolite, and chlorite; $\rho = 2.901$ g/cm^3.
Granulite, Adirondack Mountains, New York	Gneissose, medium grained, xenoblastic; 32% plagioclase (An_{32}), 30% perthite, 38% pyroxene (altered), trace of opaque and sphene; $\rho = 2.911$ g/cm^3.
Amphibolite, Indian Ocean, Central Ridge	Holocrystalline, medium grained, inequigranular; 50% hornblende, 40% plagioclase (An_{30}-An_{50}), 5% enstatite, 5% actinolite and chlorite; $\rho = 2.930$ g/cm^3.
Anorthosite, Lake St. John, Quebec	Holocrystalline, coarse grained; 70% plagioclase (An_{57}), 7% hypersthene, 15% ilmenite, 5% hornblende, 2% biotite, 1% green spinel; $\rho = 2.971$ g/cm^3.
Granulite, Valle d'Ossola, Italy	Gneissose, medium to coarse grained; 60% plagioclase (An_{40}), 26% pyroxene, 9% garnet, 4% opaque, 1% microcline; $\rho = 3.125$ g/cm^3.
Dunite, Twin Sisters, Washington	Inequigranular, medium grained; 99% olivine, 1% opaque, trace of secondary alteration along fractures; $\rho = 3.306$ g/cm^3.
Eclogite, Nove' Dvory, Czechoslovakia	Granulose, medium to fine grained, symplectite intergrowths, xenoblastic; 49% garnet, 45% omphacite, 5% intergranular opaque mineral and rutile, 1% biotite; $\rho = 3.559$ g/cm^3.

1951; *Hughes and Maurette*, 1956] found that the cracking could be prevented by the application of high confining pressures to the rocks. The low hystereses common to the data obtained in the present study also demonstrate that fractures have not been produced by temperature cycles.

Recent studies of the effect of temperature on compressional wave velocities in igneous and metamorphic rocks at elevated pressure have found that below approximately 300°C velocity is approximately a linear function of temperature [e.g., *Stewart and Peselnick*, 1977; *Meissner and Fakhimi*, 1977], whereas at higher temperatures, velocity-temperature plots often show a rapid decrease in velocity [e.g., *Fielitz*, 1971; *Kern and Fakhimi*, 1975]. This more rapid lowering of velocity at temperatures above 300°C is pronounced in the granite, amphibolite, quartzite, anorthosite, and the granulites and has been interpreted as resulting from the opening of grain boundary cracks caused by anisotropic thermal expansion and differences in volume expansion of individual mineral components. *Kern* [1978], using a cubic pressure apparatus with a solid pressure medium, reported similar velocity-temperature relations at confining pressures up to 6 kbar. Kern attributed these changes to the opening of grain boundary cracks and supported this with thermal volumetric strain measurements, which showed correlations between velocity decreases with increasing temperature and volume changes. *Ramananantoandro and Manghnani* [1978] measured velocities in three cores of Twin Sisters dunite to 500°C at 10 kbar with a gas apparatus and found that the measured values of $(\partial V_p/\partial T)_P$ were larger than values calculated from olivine single-crystal data by 17–52%, depending upon the propagation direction. Ramananantoandro and Manghnani attributed this discrepancy to the widening of grain boundary cracks during heating and, on the basis of low hysteresis in their velocity-temperature plots, suggested that this effect is reversible. In addition, compressional wave velocities for quartz-bearing rocks show rapid decreases as temperatures approach the α-β quartz inversion temperature [*Fielitz*, 1971; *Kern*, 1978].

Petrographic examinations of samples included in the present study reinforce the earlier interpretations that grain boundary cracks open at elevated temperatures and suggest that rock texture is important in controlling the opening of grain boundaries. For example, the Twin Sisters dunite, which does not show a rapid decrease in velocity with increasing temperature above 300°C, has a grain boundary structure characterized by anhedral olivine crystals with irregular interlocking boundaries. This texture apparently inhibits the formation of large cracks between grain boundaries at high temperatures. Rocks with euhedral crystals and relatively simple igneous textures, on the other hand, have grain boundaries which open readily at elevated temperatures.

The slope of the velocity-temperature curve of the basalt is low in comparison to the gabbro. A similar low change in velocity with increasing temperature has been reported for another volcanic rock, an andesite from Colorado [*Hughes and Jones*, 1950]. Hughes and Jones suggested that the presence of

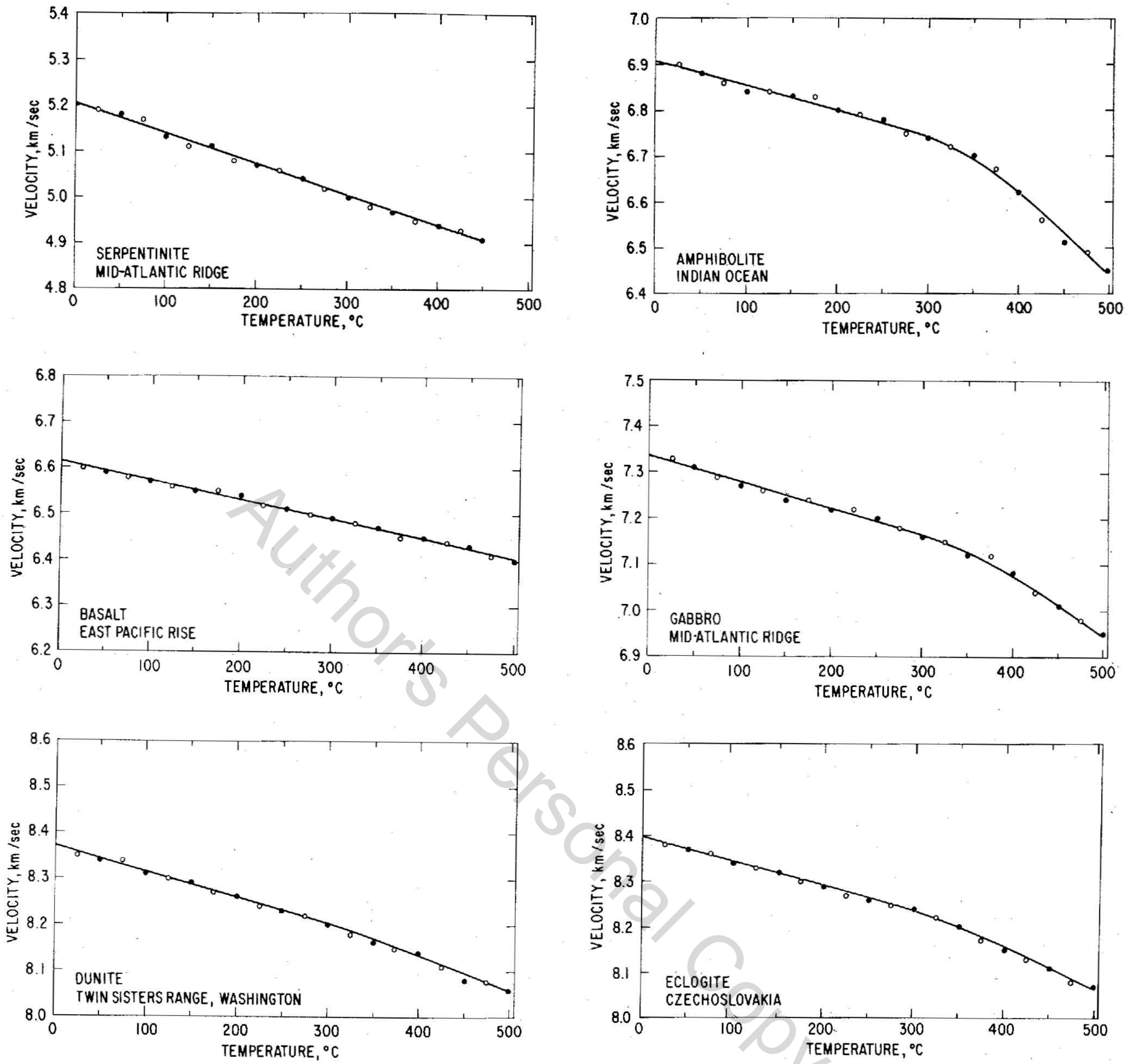

Fig. 3. Compressional wave velocities as a function of temperature at 2-kbar pressure for oceanic crustal rocks and possible upper mantle rocks. Dots indicate measurements with increasing temperature, and circles those with decreasing temperature.

glass in volcanic rocks would reduce the slope of the velocity-temperature curve and cited the early work of *Ide* [1937], who reported an increase in velocity with an increase in temperature for glass. An additional factor influencing the temperature derivative of velocity is suggested by the linear nature to 500°C of the basalt velocity-temperature curve. In contrast with the other rocks included in this study, the basalt crystallized at relatively low pressures. After formation, grain boundary cracks originated by thermal contraction associated with cooling. Because mineral volume changes are quite different for compression and thermal expansion, an application of pressure to volcanic rocks will not close grain boundaries as effectively as in plutonic rocks. Application of temperature, on the other hand, reduces grain boundary cracks up to temperatures much greater than 500°C. The combination of increased velocity resulting from decreased grain boundary porosity and decreased velocity due to the effect of temperature on the crystal lattice results in relatively low changes in velocity with increasing pressure in volcanics and a linear relationship between these variables. In contrast, the coarser-grained rocks show at first a characteristic linear decrease in velocity with increasing temperature and then a rapid decrease in velocity related to opening of grain boundaries.

Comparisons of the data presented here with previously reported velocities at 2–3 kbar in similar rocks (Figures 5, 6, and 7) show only fair agreement for the data sets from various laboratories. It is likely that some of the differences in the temperature derivatives are related to texture, as discussed above, or slightly differing mineralogies for a given rock type. It is also probable that experimental error is significant, since a

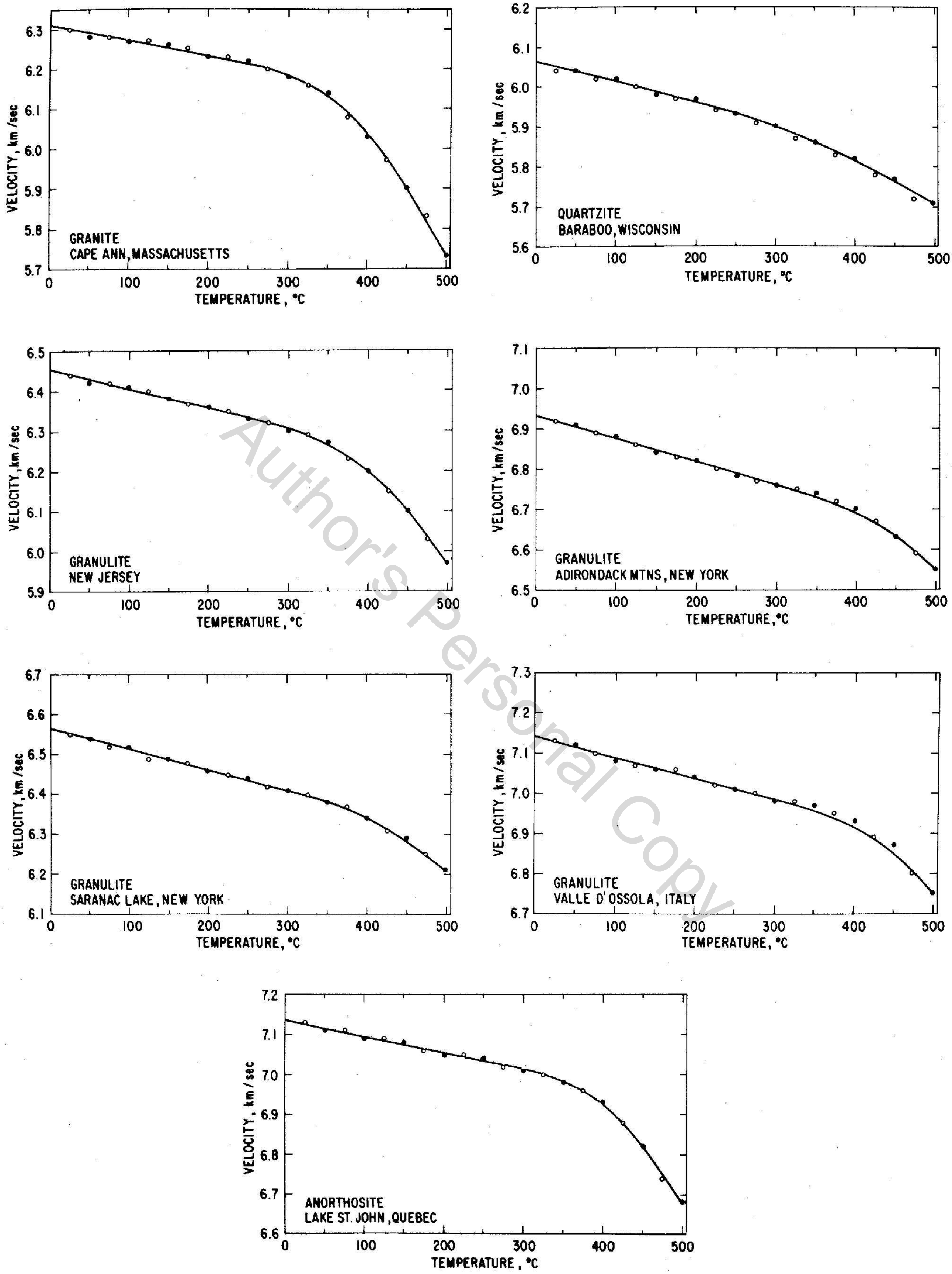

Fig. 4. Compressional wave velocities as a function of temperature at 2-kbar pressure for several continental crustal rocks. Dots indicate measurements with increasing temperature, and circles those with decreasing temperature.

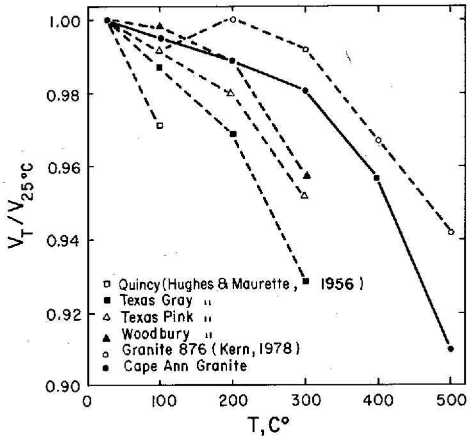

Fig. 5. A comparison of Cape Ann granite velocities with previously reported granite velocities at 2 kbar.

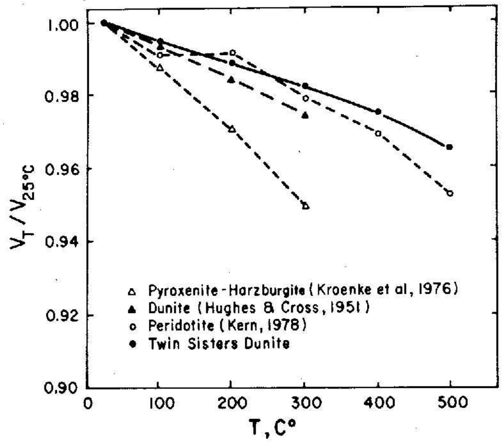

Fig. 7. A comparison of ultramafic rock velocities as a function of temperature at 2–3 kbar.

wide variety of furnace designs, jacketing techniques, transducer configurations, and methods of generating pressure have been employed.

Critical Thermal Gradients

A principal conclusion from the preceding discussion is that for many rocks the observed temperature coefficients at temperatures between 300° and 500°C are high owing to the opening of grain boundary cracks. On the other hand, the temperature coefficient obtained for the basalt is probably low at all temperatures. Examination of Figures 3 and 4 shows that the decreases in velocity with increasing temperature are close to linear for all samples at temperatures below 300°C. In Table 2, $(\partial V_p/\partial T)_P$ determined by linear least squares solutions is given for each sample for several temperature ranges. Of significance are the following: (1) Many samples have temperature coefficients between -0.5×10^{-3} and -0.6×10^{-3} km s^{-1} °C^{-1} over the temperature range 25°–300°C. Exceptions include the serpentinite with a relatively high temperature coefficient and the basalt, granite, and anorthosite. The basalt temperature coefficient is believed to be influenced by porosity, whereas the low values for granite and anorthosite may reflect an intrinsic low $(\partial V_p/\partial T)_P$ for feldspars. (2) The temperature coefficients at relatively high temperatures show extreme variability, which reflects the opening of grain boundary pore space, controlled by the rock texture. Since within the framework of geologic time, metamorphic reactions and flow at elevated crustal temperatures and pressures tend to eliminate grain boundary openings, it is unlikely that grain boundary pore space remains open within the earth's crust at depth. It is concluded that the data above 300°C are, in general, of little

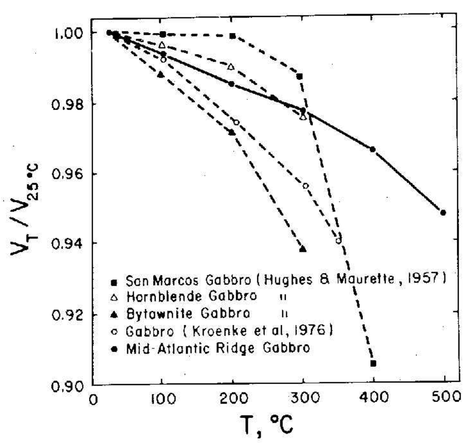

Fig. 6. Mid-Atlantic ridge gabbro velocities compared with other gabbro velocities at 2–3 kbar.

TABLE 2. Temperature Coefficients

Rock	Temperature Range, °C	$\partial V_p/\partial T$, km s^{-1} °C^{-1}
Serpentinite,	25–300	-0.68×10^{-3}
Mid-Atlantic Ridge	300–450	-0.57×10^{-3}
Granite, Cape Ann,	25–300	-0.39×10^{-3}
Massachusetts	300–400	-1.52×10^{-3}
	400–500	-2.96×10^{-3}
Quartzite,	25–300	-0.54×10^{-3}
Baraboo, Wisconsin	300–400	-0.80×10^{-3}
	400–500	-1.12×10^{-3}
Granulite,	25–300	-0.49×10^{-3}
New Jersey Highlands	300–400	-1.04×10^{-3}
	400–500	-2.32×10^{-3}
Granulite, Saranac	25–300	-0.51×10^{-3}
Lake, New York	300–400	-0.68×10^{-3}
	400–500	-1.28×10^{-3}
Basalt,	25–300	-0.39×10^{-3}
East Pacific Rise	300–400	-0.44×10^{-3}
	400–500	-0.52×10^{-3}
Gabbro,	25–300	-0.57×10^{-3}
Mid-Atlantic Ridge	300–400	-0.76×10^{-3}
	400–500	-1.28×10^{-3}
Granulite, Adirondack	25–300	-0.60×10^{-3}
Mountains, New York	300–400	-0.60×10^{-3}
	400–500	-1.52×10^{-3}
Amphibolite, Indian	25–300	-0.55×10^{-3}
Ocean, Central Ridge	300–400	-1.16×10^{-3}
	400–500	-1.64×10^{-3}
Anorthosite, Lake	25–300	-0.41×10^{-3}
St. John, Quebec	300–400	-0.80×10^{-3}
	400–500	-2.56×10^{-3}
Granulite, Valle	25–300	-0.52×10^{-3}
d'Ossola, Italy	300–400	-0.52×10^{-3}
	400–500	-1.80×10^{-3}
Dunite, Twin Sisters,	25–300	-0.56×10^{-3}
Washington	300–400	-0.60×10^{-3}
	400–500	-0.76×10^{-3}
Eclogite, Nove' Dvory,	25–300	-0.53×10^{-3}
Czechoslovakia	300–400	-0.84×10^{-3}
	400–500	-0.92×10^{-3}

TABLE 3. Critical Gradients

Rock	Pressure Range, kbar	$\partial V_p/\partial T$, km s^{-1} °C^{-1}	$\partial V_p/\partial P$, km s^{-1} kbar^{-1}	$(dT/dZ)_c$, °C km^{-1}
Serpentinite,	3–5	-0.68×10^{-3}	0.065	23.5
Mid-Atlantic Ridge	5–8	-0.68×10^{-3}	0.045	16.3
Granite, Cape Ann,	3–5	-0.39×10^{-3}	0.036	23.7
Massachusetts	5–8	-0.39×10^{-3}	0.025	16.4
Quartzite, Baraboo,	3–5	-0.54×10^{-3}	0.030	14.4
Wisconsin	5–8	-0.54×10^{-3}	0.030	14.4
Granulite, New	3–5	-0.49×10^{-3}	0.028	15.0
Jersey Highlands	5–8	-0.49×10^{-3}	0.020	10.9
Granulite, Saranac	3–5	-0.51×10^{-3}	0.024	13.1
Lake, New York	5–8	-0.51×10^{-3}	0.021	11.5
Basalt, East	3–5	-0.39×10^{-3}	0.035	25.4
Pacific Rise	5–8	-0.39×10^{-3}	0.016	11.7
Gabbro,	3–5	-0.57×10^{-3}	0.023	11.5
Mid-Atlantic Ridge	5–8	-0.57×10^{-3}	0.020	10.2
Granulite, Adirondack	3–5	-0.60×10^{-3}	0.033	15.7
Mountains, New York	5–8	-0.60×10^{-3}	0.028	13.2
Amphibolite, Indian	3–5	-0.55×10^{-3}	0.029	15.2
Ocean, Central Ridge	5–8	-0.55×10^{-3}	0.028	14.7
Anorthosite, Lake	3–5	-0.41×10^{-3}	0.021	14.9
St. John, Quebec	5–8	-0.41×10^{-3}	0.020	14.1
Granulite, Valle	3–5	-0.52×10^{-3}	0.035	20.6
d'Ossola, Italy	5–8	-0.52×10^{-3}	0.025	14.5
Dunite, Twin Sisters,	3–5	-0.56×10^{-3}	0.029	16.8
Washington	5–8	-0.56×10^{-3}	0.023	13.4
	10–30	-0.56×10^{-3}	0.011	6.3
Eclogite, Nove' Dvory,	3–5	-0.53×10^{-3}	0.020	13.2
Czechoslovakia	5–8	-0.53×10^{-3}	0.019	12.3
	10–30	-0.53×10^{-3}	0.010	6.3

value in estimating critical temperature gradients within the crust and upper mantle.

The influence of confining pressure on compressional wave velocities has been reported for several rocks to 10 kbar [e.g., *Birch*, 1960]. *Christensen* [1974] found that $(\partial V_p/\partial P)_T$ for igneous and metamorphic rocks below 10 kbar are influenced by grain boundary porosity as predicted by *Birch* [1969]. It was shown, on the other hand, that the pressure coefficient for dunite determined from velocity data between 10 and 30 kbar agrees well with extrapolated data from single-crystal olivine, which suggests that the effects of grain boundary porosity on rock velocities above 10 kbar is minimal.

The pressure coefficients for two of the rocks included in this study (dunite and eclogite) have been reported to 30 kbar. These are given in Table 3 with values at 3–5 kbar and 5–8 kbar for all of the rock specimens. The critical thermal gradients for a low-velocity layer $(dT/dZ)_c$ obtained from

$$\left(\frac{dT}{dZ}\right)_c = \rho g \left[-\frac{(\partial V_p/\partial P)_T}{(\partial V_p/\partial T)_P}\right]$$

(where Z is depth, ρ is the rock density, and g is the acceleration of gravity) are also given in Table 3 at various pressure ranges. For all calculations, values of $(\partial V_p/\partial T)_P$ were selected from the 25°–300°C temperature range.

Values of $(dT/dZ)_c$ in Table 3 are actually conditions for constant velocity; within the earth the condition $dV/dr = 0$ results in upward refraction [e.g., *Gutenberg*, 1959; *Birch*, 1969] and corrections for the earth's curvature at mantle depth can be substantial. Since we are primarily concerned with crustal low-velocity zones, the curvature of the earth is neglected in the following discussion.

The critical gradients for the eclogite and dunite in the pressure range of 10–30 kbar agree well with those calculated from single-crystal data [*Birch*, 1969; *Liebermann and Schreiber*, 1969]. In particular, $(dT/dZ)_c$ for olivine with a composition similar to the Twin Sisters dunite [*Kumazawa and Anderson*, 1969] is 6.9°C/km compared to 6.3°C/km for the rock (Table 3). As expected, critical thermal gradients are greater at lower pressures because of the rapid increase of velocity associated with closing of grain boundary cracks.

INTERPRETATION OF SEISMOLOGICAL DATA

Solutions to seismic data of the earth's crust are now undergoing substantial revisions, particularly with respect to velocity distributions with depth. Simple two or more layer solutions are now being replaced with more realistic models showing velocity gradients as a function of depth, many of which show well-substantiated reversals.

Evidence for low-velocity zones within the continental crust was first presented by *Gutenberg* [1951, 1955], based on obser-

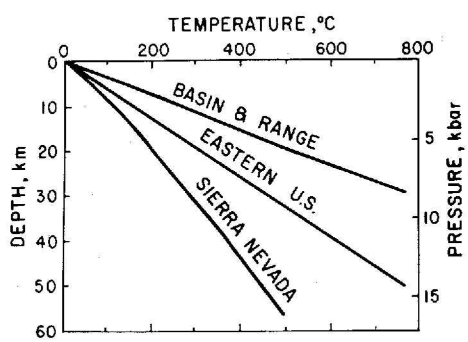

Fig. 8. Temperature distribution within the crust and upper mantle for three heat flow provinces [after *Blackwell*, 1971].

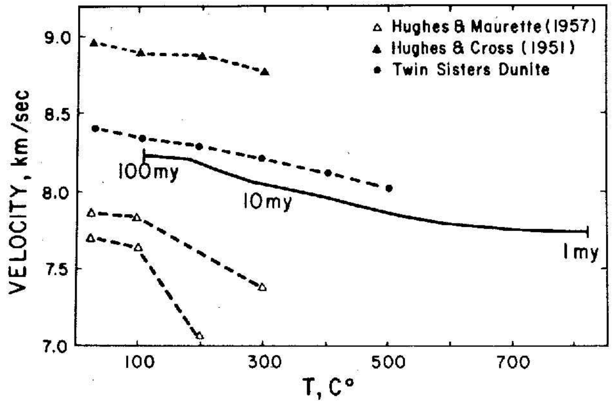

Fig. 9. Predicted oceanic upper mantle velocity-temperature relation (solid line) of *Bibee and Shor* [1976] compared with the Twin Sisters dunite velocity data of this paper and earlier velocity data for ultramafic rocks. The approximate mantle age is also shown in the diagram.

vations of explosions and earthquakes in southern California. More recent observations from crustal near-vertical reflection and refraction experiments, surface wave dispersion studies, and earthquakes further support the presence of velocity reversals throughout most of the world's continents [e.g., *Mueller and Landisman*, 1966; *Pavlenkova*, 1968; *Lukk et al.*, 1970; *Landisman et al.*, 1971; *Mitchell and Landisman*, 1971; *Berry and Fuchs*, 1973; *Bennett et al.*, 1975; *Keller et al.*, 1975; *Crosson*, 1976]. Although there is no consensus on the nature and extent of crustal low-velocity zones, many crustal models show a well-developed low-velocity zone at approximately 10-km depth. Velocities above the reversal are fairly constant, and velocities below the reversal show a steady increase with increasing depth. In some continental areas a second low-velocity zone has been proposed for depths of the order of 20 km [e.g., *Landisman et al.*, 1971].

The origin of low-velocity regions within the crust can be explained in a variety of ways. In addition to high-temperature gradients, proposals which have been advanced include (1) the transition from α to β quartz [*Gutenberg*, 1951], (2) velocity reversals originating from increased fluid pressure in rock pore spaces [*Gordon and Davis*, 1968], and (3) a region of relatively low velocity granitic plutons within the earth's crust [*Landisman et al.*, 1971]. Several other possibilities including the presence of partially molten rocks within the crust, particularly in high heat flow regions, and crustal zones of extensive hydration (i.e., serpentinization), could produce low-velocity channels.

All of these hypotheses are physically acceptable and therefore likely at least locally to produce velocity inversions within the crust. However, if details of the structure of the earth's crust reveal that a velocity reversal is a common crustal feature in many regions, especially of contrasting tectonic history, it appears that the simplest physical interpretation for crustal low-velocity zones is that they originate from high temperatures. With the experimental data presented in this paper it is possible to examine in detail this proposed origin of crustal velocity inversions.

On the basis of heat flow studies it is known that temperature-depth distributions within the continental crust show considerable variability [e.g., *Roy et al.*, 1968]. *Blackwell* [1971] has presented temperature-depth calculations for three heat flow provinces, the eastern United States, Basin and Range, and Sierra Nevada (Figure 8). The temperature distribution within the eastern United States is considered characteristic of 'normal' continental crust [*Roy et al.*, 1978], whereas the Basin and Range represents a region of high crustal temperatures, and temperatures underlying the Sierra Nevada province are relatively low. The oceanic geotherm of *Ringwood* [1969] closely approximates the temperature-depth curve for the eastern United States.

Blackwell [1971] presents for each province four models which differ from one another in assumed mantle heat flow, crustal structure, and lower crustal heat generation; however, the calculated temperature gradients, especially in the upper 20 km of crust, vary only slightly within a given province. Temperature gradients (dT/dZ) at depths between 10 and 20 km for the preferred model (model 3) are 8.4°C/km for the Sierra Nevada province, 15.2°C/km for the eastern United States, and 23.8°C/km for the Basin and Range province. A comparison of these gradients with the critical gradients $(dT/dZ)_c$ of Table 2 shows that for many of the rock types studied, the critical gradients are exceeded. For the Basin and Range province in particular a decrease of velocity with depth would occur in crustal regions of uniform lithology. Since the critical temperature gradients given in Table 2 for the pressure ranges 3–5 kbar and 5–8 kbar are influenced by rock porosity and thus are maximum values, the actual critical gradients at crustal pressures could be as low as the 8.4°C/km temperature gradient of the Sierra Nevada province. Thus regions with normal heat flow such as the eastern United States, which may be typical of stable portions of continents, are likely to have crustal low-velocity zones at depths where the crust is homogeneous.

The magnitude of the velocity decreases produced in a homogeneous crustal region by high temperature will depend upon many factors, one of the most important being the thickness of the region. Reported thicknesses of crustal low-velocity zones are usually between 2 and 6 km. Simple calculations based on the data presented in this paper and the temperature gradients of *Blackwell* [1971] show that the maximum decrease in velocity over a 4-km depth interval produced by temperature is unlikely to be greater than 0.05 km/s. In many continental regions the reported velocity reversals are much greater than 0.05 km/s, often reaching 0.6 km/s [e.g., *Landisman et al.*, 1971]. It should be emphasized that velocity reversals of this magnitude are unlikely to be produced by temperature alone, and other mechanisms must be invoked to explain their presence. However, since indistinguishable synthetic seismograms can be produced for models with wide ranges of negative velocity gradients and thicknesses, the actual magnitudes of many crustal velocity reversals may be lower than reported, and vertical extents may be greater. Thus temperature alone may be responsible for many crustal velocity inversions.

Unknown at present is whether low-velocity zones produced by temperature are common in the oceanic crust. The critical gradients observed for the gabbro from the Mid-Atlantic Ridge and the amphibolite from the Central Indian Ridge indicate that a velocity reversal within oceanic layer 3 is a distinct possibility, especially in regions near ridge crests. The existence of a low-velocity region near the base of the crust in the vicinity of the Explorer Ridge has recently been well documented by *Malecek and Clowes* [1978], the origin of which they attribute to a zone of partial melt.

Finally, *Bibee and Shor* [1976] have found that oceanic P_n

velocities corrected for anisotropy show a high correlation with age and attribute this to cooling away from ridge crests. Their velocity-age relationship was combined with the *Parker and Oldenburg* [1973] temperature-age model to produce a velocity-temperature plot for the upper oceanic mantle (Figure 9). Comparisons of this relationship with earlier velocity-temperature measurements for possible mantle rocks suggested that the increase in mantle velocity with age is consistent with decreasing temperature; however, *Bibee and Shor* [1976] emphasized the need for better experimental data before this finding would be used effectively. A comparison of the temperature-velocity data for the Twin Sisters dunite (Figure 3) with the velocity-temperature model of Bibee and Shor demonstrates without question the importance of this relationship (Figure 9).

Acknowledgments. This investigation was supported by the Office of Naval Research under contract N-0014-75-C-0502. I am indebted to R. McConaghy and V. Kiefer for assistance in constructing the furnace for the temperature-pressure vessel and obtaining the velocity data. Robert Crosson critically read the manuscript and offered helpful suggestions.

REFERENCES

Bennett, G. T., R. M. Clowes, and R. M. Ellis, A seismic refraction survey along the southern Rocky Mountain Trench, Canada, *Bull. Seismol. Soc. Amer., 65*, 37–54, 1975.

Berry, M. J., and K. Fuchs, Crustal structure of the Superior and Grenville provinces of the northeastern Canadian shield, *Bull. Seismol. Soc. Amer., 63*, 1393–1432, 1973.

Bibee, L. D., and G. G. Shor, Jr., Compressional wave anisotropy in the crust and upper mantle, *Geophys. Res. Lett., 3*, 639–642, 1976.

Birch, F., Elasticity of igneous rocks at high temperatures and pressures, *Geol. Soc. Amer. Bull., 54*, 263–286, 1943.

Birch, F., The velocity of compressional waves in rocks to 10 kilobars, 1, *J. Geophys. Res., 65*, 1083–1102, 1960.

Birch, F., Density and composition of the upper mantle: First approximation as an olivine layer, in *The Earth's Crust and Upper Mantle, Geophys. Monogr. Ser.*, vol. 13, edited by P. J. Hart, pp. 18–38, AGU, Washington, D. C., 1969.

Birch, F., and D. Bancroft, The effect of pressure on the rigidity of rocks, *J. Geol., 46*, 59–87, 113–141, 1938.

Blackwell, D. D., The thermal structure of the continental crust, in *The Structure and Physical Properties of the Earth's Crust, Geophys. Monogr. Ser.*, vol. 14, edited by J. G. Heacock, pp. 169–184, AGU, Washington, D. C., 1971.

Christensen, N. I., Compressional wave velocities in metamorphic rocks at pressures to 10 kbar, *J. Geophys. Res., 79*, 6147–6164, 1965.

Christensen, N. I., Compressional and shear wave velocities at pressures to 10 kilobars for basalts from the East Pacific rise, *Geophys. J. Roy Astron. Soc., 28*, 425–429, 1972.

Christensen, N. I., Compressional wave velocities in possible mantle rocks to pressures of 30 kbar, *J. Geophys. Res., 79*, 407–412, 1974.

Christensen, N. I., and D. M. Fountain, Constitution of the lower continental crust based on experimental studies of seismic velocities in granulite, *Geol. Soc. Amer. Bull., 86*, 227–236, 1975.

Christensen, N. I., and G. H. Shaw, Elasticity of mafic rocks from the mid-Atlantic ridge, *Geophys. J. Roy. Astron. Soc., 20*, 271–284, 1970.

Crosson, R. S., Crustal structure modeling of earthquake data, 2, Velocity structure of the Puget Sound region, Washington, *J. Geophys. Res., 81*, 3047–3054, 1976.

Fielitz, K., Elastic wave velocities in different rocks at high pressure and temperature up to 750°C, *Z. Geophys., 37*, 943–956, 1971.

Gordon, R., and L. Davis, Velocity and attenuation of seismic waves in imperfectly elastic rocks, *J. Geophys. Res., 73*, 3917–3935, 1968.

Gutenberg, B., Crustal layers of the continents and oceans, *Geol. Soc. Amer. Bull., 62*, 427–440, 1951.

Gutenberg, B., Channel waves in the Earth's crust, *Geophysics, 20*, 283–291, 1955.

Gutenberg, B., *Physics of the Earth's Interior*, 240 pp., Academic, New York, 1959.

Hughes, D. S., and J. H. Cross, Elastic wave velocities at high pressures and temperatures, *Geophysics, 16*, 577–593, 1951.

Hughes, D. S., and H. J. Jones, Variation of elastic moduli of igneous rocks with pressures and temperatures, *Geol. Soc. Amer. Bull., 61*, 843–856, 1950.

Hughes, D. S., and C. Maurette, Variation of elastic wave velocities in granites with pressure and temperature, *Geophysics, 21*, 277–284, 1956.

Hughes, D. S., and C. Maurette, Variation of elastic wave velocities in basic igneous rocks with pressure and temperature, *Geophysics, 22*, 23–31, 1957.

Ide, J. M., The velocity of sound in rocks and glasses as a function of temperature, *J. Geol., 45*, 689–716, 1937.

Keller, G. R., R. B. Smith, and L. W. Braile, Crustal structure along the Great Basin–Colorado plateau transition from seismic refraction studies, *J. Geophys. Res., 80*, 1093–1098, 1975.

Kern, H., The effect of high temperature and high confining pressure on compressional wave velocities in quartz-bearing and quartz-free igneous and metamorphic rocks, *Tectonophysics, 44*, 185–203, 1978.

Kern, H., and M. Fakhimi, Effect of fabric anisotropy on compressional wave propagation in various metamorphic rocks in the range of 20°C–700°C at 2 kb, *Tectonophysics, 28*, 227–244, 1975.

Kroenke, L. W., M. H. Manghnani, C. S. Rai, and R. Ramananantoandro, Elastic properties of selected ophiolitic rocks from Papua New Guinea: Nature and composition of lower crust and upper mantle, in *The Geophysics of the Pacific Basin and Its Margin, Geophys. Monogr. Ser.*, vol. 19, edited by G. H. Sutton, M. H. Manghnani, and R. Moberly, pp. 407–421, AGU, Washington, D. C., 1976.

Kumazawa, M., and O. L. Anderson, The elastic moduli, pressure derivatives, and temperature derivatives of single-crustal olivine and single-crystal forsterite, *J. Geophys. Res., 74*, 5961–5972, 1969.

Landisman, M., S. Mueller, and B. J. Mitchell, Review of evidence for velocity inversions in the continental crust, in *The Structure and Physical Properties of the Earth's Crust, Geophys. Monogr. Ser.*, vol. 14, edited by J. G. Heacock, pp. 11–34, AGU, Washington, D. C., 1971.

Liebermann, R. C., and E. Schreiber, Critical thermal gradients in the mantle, *Earth Planet. Sci. Lett., 7*, 77–81, 1969.

Lukk, A., I. Nersesov, and L. Chepkunas, Procedure for identification of the low velocity layer in the Earth's crust and mantle, *Izv. Acad. Sci. USSR, Phys. Solid Earth, 2*, 77–80, 1970.

Malecek, S. J., and R. M. Clowes, Crustal structure near Explorer Ridge from a marine deep seismic sounding survey, *J. Geophys. Res., 83*, 5899–5912, 1978.

Meissner, R., and M. Fakhimi, Seismic anisotropy as measured under high-pressure, high-temperature conditions, *Geophys. J. Roy. Astron. Soc., 49*, 133–143, 1977.

Mitchell, B., and M. Landisman, Geophysical measurements in the southern Great Plains, in *The Structure and Physical Properties of the Earth's Crust, Geophys. Monogr. Ser.*, vol. 14, edited by J. G. Heacock, pp. 77–93, AGU, Washington, D. C., 1971.

Mueller, S., and M. Landisman, Seismic studies of the Earth's crust in continents, I, Evidence for a low velocity zone in the upper part of the lithosphere, *Geophys. J., 10*, 525–538, 1966.

Parker, R. L., and D. W. Oldenburg, Thermal model of ocean ridges, *Nature, 242*, 137–139, 1973.

Pavlenkova, N., Methods of velocity determination from seismic crustal studies, report, 13 pp., Inst. Geophys. Acad. Sci. Ukranian SSR, 1968.

Ramananantoandro, R., and M. H. Manghnani, Temperature dependence of the compressional wave velocity in an isotropic dunite: Measures to 500°C at 10 kbar, *Tectonophysics, 47*, 73–84, 1978.

Ringwood, A. E., Composition and evolution of the upper mantle, in *The Earth's Crust and Upper Mantle, Geophys. Monogr. Ser.*, vol. 13, edited by P. J. Hart, pp. 1–17, AGU, Washington, D. C., 1969.

Roy, R. F., E. R. Decker, D. D. Blackwell, and F. Birch, Heat flow in the United States, *J. Geophys. Res., 73*, 5207–5222, 1968.

Skinner, G. J., Thermal expansion, Handbook of Physical Constants, *Geol. Soc. Amer. Mem., 97*, 75–96, 1966.

Stewart, R., and L. Peselnick, Velocity of compressional waves in dry Franciscan rocks to 8 kbar and 300°C, *J. Geophys. Res., 82*, 2027–2039, 1977.

(Received March 5, 1979;
revised June 18, 1979;
accepted June 27, 1979.)