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Pore pressure, seismic velocities, and crustal structure

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ABSTRACT

The seismic velocity structure of the Earth's crust is examined in relation to the role played by high pore pressures. Compressional and shear-wave velocities measured at carefully controlled confining and pore pressures show significant decreases with increasing pore pressure. This behavior is important in sedimentary rocks and in crystalline rocks likely to occur at deep crustal levels. Velocities are shown to be a function of effective pressure rather than differential pressure. Similar results are found for crystalline rocks in which gas or water is the pore fluid. Thus, pervasive CO_2 at elevated pore pressures will lower crustal velocities. Low-velocity regions originating from high pore pressures have high Poisson's ratios. Within the crust, dehydration accompanying progressive metamorphism can produce high pore pressures and regions of low velocities if pore fluids are contained. A conceptual model is presented for the continental crust in which midcrustal discontinuities such as the Conrad, where present, separate an overpressured upper region of igneous and metamorphic rocks from underlying dry rocks. Fluids released by lower crustal dehydration are trapped above the Conrad discontinuity. In midcrustal regions, where pore pressures are low due to loss of fluids, progressive metamorphism with depth will be detected as a gradual increase in velocity rather than a sharp discontinuity.

INTRODUCTION

It is well known that many physical properties of rocks depend not only on their mineral constituents but also on the nature of contained pore fluids. Of particular significance in geology and geophysics is the pressure of the interstitial pore fluid. For more than half a century, workers have recognized the importance of pore pressure in rock mechanics. For example, wide acceptance of the role of high pore pressure in overthrusting originated from the theory of Hubbert and Rubey (1959), which showed that fracture strength is proportional to the difference between the total normal stress across a fault and the pore fluid pressure within the fault zone. More recently, pore pressure has been shown to have important tectonic implications in the control of earthquakes (e.g., Raleigh, 1971) and in evaluating nuclear waste burial sites (e.g., Trimmer and others, 1980). In refraction seismology, the interpretation of crustal composition relies to a large extent on comparisons of laboratory-measured velocities with velocities determined at various depths. Knowledge of the extent to which pore pressure will influence velocities is critical for such comparisons. Crustal bright spots and deep laminated reflectors (Fuchs, 1969) may also relate to regions of alternating pore pressures.

Compared with standard measurements of velocity as a function of confining pressure, measurements of velocities under controlled pore pressure are complex. A jacket must be constructed that will isolate the rock from the external confining pressure fluid and contain a port so that internal pore pressure can be varied independently. Such experiments have been undertaken for a limited number of rock types, most of which have been sandstones (Wyllie and others, 1958; Banthia and others, 1965; King, 1966; Christensen and Wang, 1985). Nur and Simmons (1969), by comparing jacketed and unjacketed measurements, showed that high pore pressures significantly lower compressional wave velocities in low-porosity rocks such as granite. Other pore-pressure studies of interest include shear-wave velocity measurements on chalk (Banthia and others, 1965), compressional-wave velocity measurements in limestone and granite (Todd and

Christensen, N. I., 1989, Pore pressure, seismic velocities, and crustal structure, *in* Pakiser, L. C., and Mooney, W. D., Geophysical framework of the continental United States: Boulder, Colorado, Geological Society of America Memoir 172.

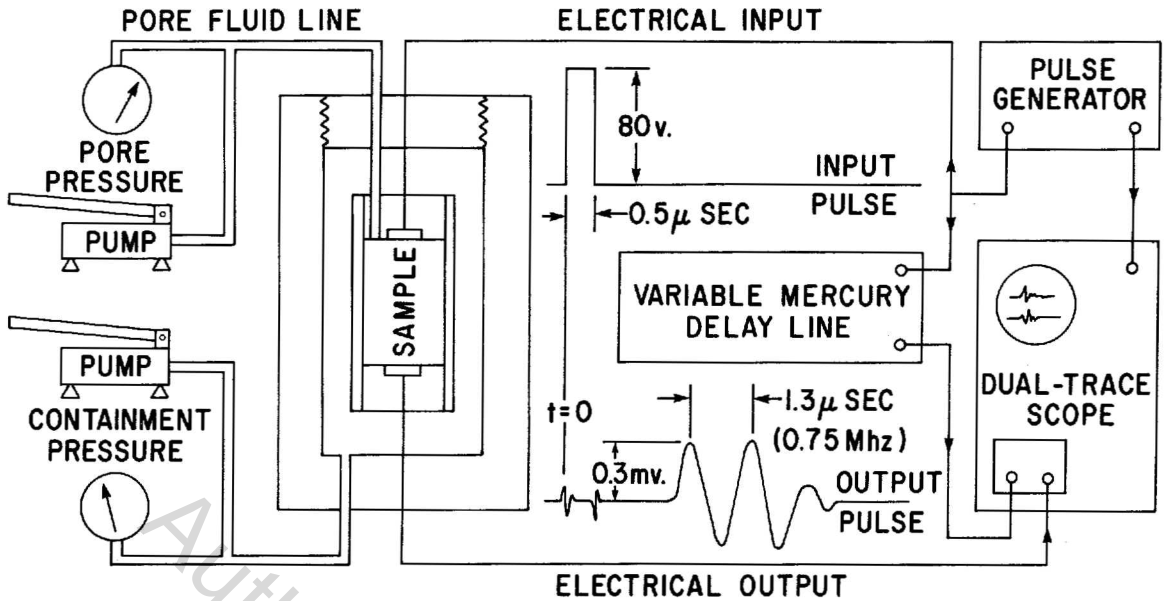


Figure 1. Schematic diagram of electronics and pressure system for velocity measurements at elevated pore and containment pressures.

Simmons, 1972) and both shear- and compressional-wave velocity studies on oceanic basalt and dolerite (Christensen, 1984).

The few papers that have been concerned either directly or indirectly with the influence of pore pressure on seismic velocities and their geophysical consequences differ widely in experimental observations and in interpretation of the data. For example, early velocity measurements by Wyllie and others (1958) for Berea sandstone show that an increase in compressional-wave velocities produced by an increase in confining pressure (P_c) is cancelled by an increase in pore pressure (P_p). Thus velocity was found to be constant for a given differential pressure (defined as $P_c - P_p$). This is in direct conflict with theoretical analyses (Brandt, 1955) and later measurements (King, 1966; Christensen and Wang, 1985), which concluded that an incremental change in pore pressure does not entirely cancel changes in compressional wave velocities produced by a similar change in confining pressure. This and other differences in opinion on this important subject originate, in part, from the lack of detailed experimental data.

An objective of this study has been to obtain detailed velocity data for a wide variety of rocks at carefully controlled confining and pore pressures. Several different rock types, both sedimentary and crystalline, originating from a range of crustal depths, have been chosen for investigation. The measurements provide an understanding of many important aspects of the influence of pore pressure on seismic velocities, thereby demonstrating the significant role played by pore pressure in the seismic velocity structure of the Earth's crust.

EXPERIMENTAL TECHNIQUES

The combined effects of confining and pore pressure on velocities in several principal rock types have been investigated to pressures to 200 MPa (2 kbar). Samples selected were as follows: (1) and (2) relatively pure quartz-rich sandstones with differing porosities, (3) dolomite, (4) a low-porosity basalt, (5) andesite, (6) granite, and (7) lherzolite. In addition, data reported by Christensen (1984) and Christensen and Wang (1985) for (8) a relatively high-porosity oceanic basalt, (9) a greenschist facies meladolerite, and (10) a quartz sandstone with abundant clay clasts and clay cement are included for comparisons with the new measurements. Descriptions of the rocks are given in Table 1.

A schematic diagram illustrating the electrical and hydraulic systems used for the velocity measurements is shown in Figure 1. A 0.5-μsec-wide rectangular pulse of up to 500 V drives the sending transducer on one end of the rock core. Lead-zirconate-titanate (PZT) transducers with 1-MHz resonant frequencies are used for transmitting and receiving compressional waves, whereas similar frequency AC-cut quartz and PZT transducers generate and receive the shear waves, respectively. The elastic wave produced by the sending transducer is transmitted through the sample, detected by the receiving transducer and amplified and displayed on one trace of a dual-trace oscilloscope. The transit times of the elastic waves are measured by superimposing arrivals from the samples and from a calibrated variable-length mercury delay line.

The pore-pressure hydraulic system used distilled water as the pore-pressure medium and operates independent of the confining-pressure hydraulic system. The confining-pressure medium is hydraulic oil. Pressure is generated with hand pumps and measured with Heise bourdon gauges with accuracies of 0.1 percent to 250 MPa (2.5 kbar).

The cylindrical samples are jacketed with aluminum that has longitudinal slots on its inside surface, which are ported to the pore-pressure pumping system, thus exposing the circumferential

TABLE 1. SAMPLE DESCRIPTIONS

Rock	Density (g/cc)	Porosity (%)	Mineralogy (%)	Comments
Mt. Simon Sandstone 1970.2	2.430	9.2	84 quartz, 14 clay, 2 opaque	Obtained from drill core at 1,970.2-ft depth in northern Illinois
Mt. Simon Sandstone 2008.6	2.523	8.0	91 quartz, 8 clay, 1 opaque	Obtained from drill core at 2,008.6-ft depth in northern Illinois
Berea Sandstone	2.253	19.3	70 quartz, 10 clay, 20 lithic fragments, feldspar, and carbonate	Clay-filled pore spaces
Oneota Dolomite	2.755	4.0	93 dolomite, 6 quartz	Obtained from drill core at 539.1-ft depth in northern Illinois
St. Helens Andesite	2.488	3.2	55 plagioclase, 25 amphibole, 3 opaque, 17 groundmass	From massive flow, slightly vesicular
Juan de Fuca Ridge Basalt	2.912	3.1	40 plagioclase, 41 pyroxene, 6 opaque, 13 groundmass	Core of pillow, vesicular
Iceland Basalt	2.788	0.4	55 plagioclase, 37 pyroxene, 5 alteration, 3 opaque	Massive dike, cored at 972.5-ft depth
Oman Dolerite	2.801	1.5	48 plagioclase, 25 orthopyroxene, 5 opaque, 22 alteration	Sheeted dike of Samail ophiolite
Illinois "Granite"	2.648	1.6	35 quartz, 31 K-feldspar, 30 plagioclase, 4 biotite	Coarse grained, from drill core at 4,061-ft depth
Kilborne Hole Lherzolite	3.256	3.9	80 olivine, 12 orthopyroxene, 5 clinopyroxene, 1 opaque, 2 serpentine	Peridotite xenolith, very loose, porous texture typical of many xenoliths

surface of the sample to the controlled pore pressure. Each jacket has a narrow slot parallel to the core axis filled with epoxy that allows for jacket adjustment with changes in confining pressure and assures hydrostatic confining pressure. Electrodes and transducers are placed on the jacketed sample ends and covered with gum rubber tubing to prevent interaction of the confining and pore fluids.

The velocity measurements at elevated temperatures and pressures are made using a two-stage double gas compressor. Argon is the confining pressure medium. Temperatures are produced with a three-zone furnace described by Christensen (1979) and measured with thermocouples located directly on the sample. Temperatures are estimated to be accurate to 1 percent.

VELOCITY DATA

The velocity data as functions of confining and differential pressures are shown in Figure 2. The velocities were measured along paths of increasing and decreasing confining pressure with atmospheric pore pressure, and at constant differential pressure. Data points for each constant differential pressure data set were obtained at random confining pressures. This procedure eliminates possible bias in the determination of the slope of a line of constant differential pressure due to possible hysteresis effects.

The procedure used to obtain velocities along lines of constant differential pressure is illustrated in Figure 3 for the dolomite at P_d = 5 MPa (0.05 kbar). After the velocity was measured

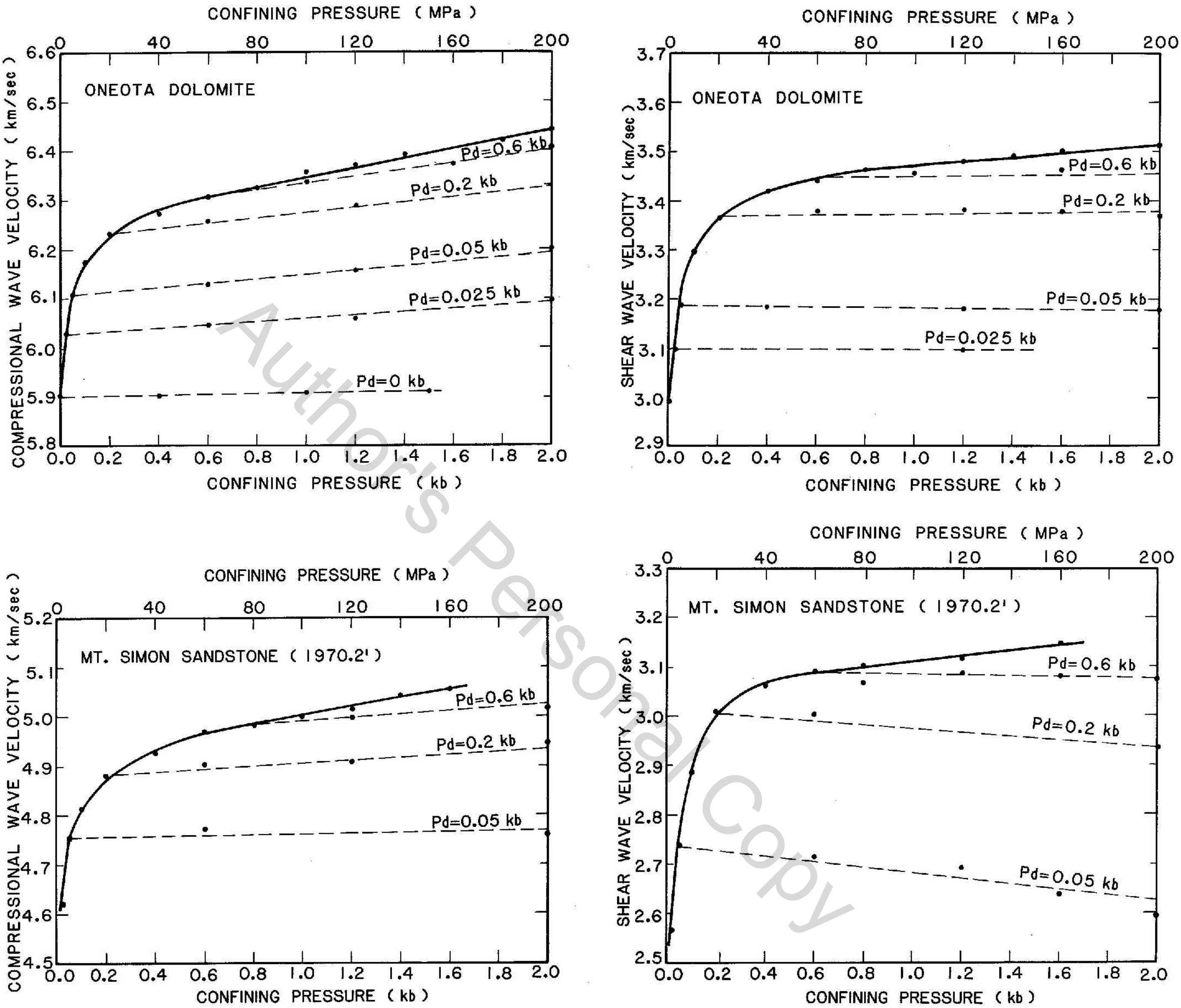
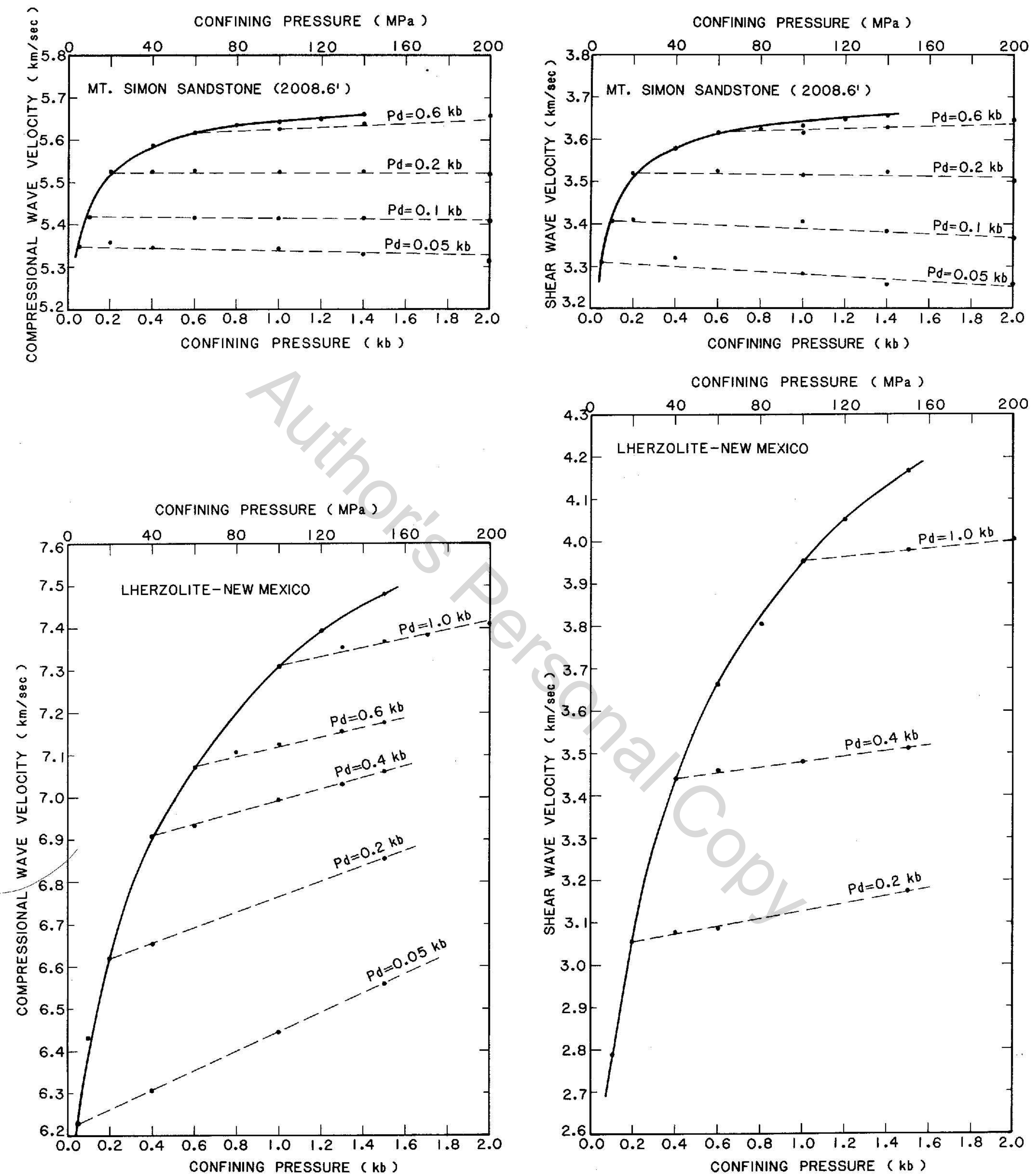


Figure 2a–e. Velocities as a function of confining pressure and differential pressure (P_d).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SHEAR WAVE VELOCITY (km/sec)
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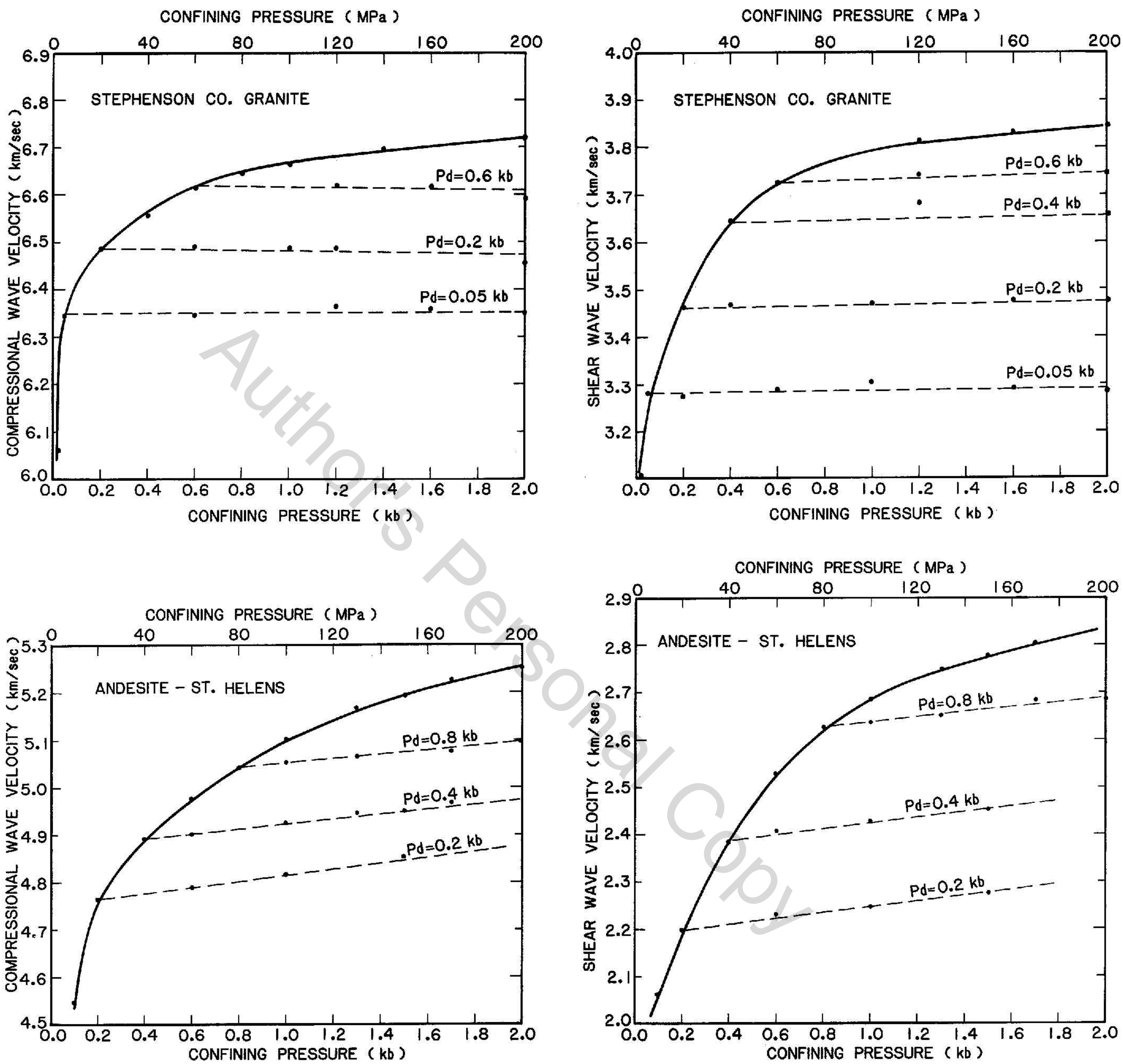
Figure 2b.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STEPHENSON CO. GRANITE
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ANDESITE – ST. HELENS
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ANDESITE – ST. HELENS
SHEAR WAVE VELOCITY (km/sec)
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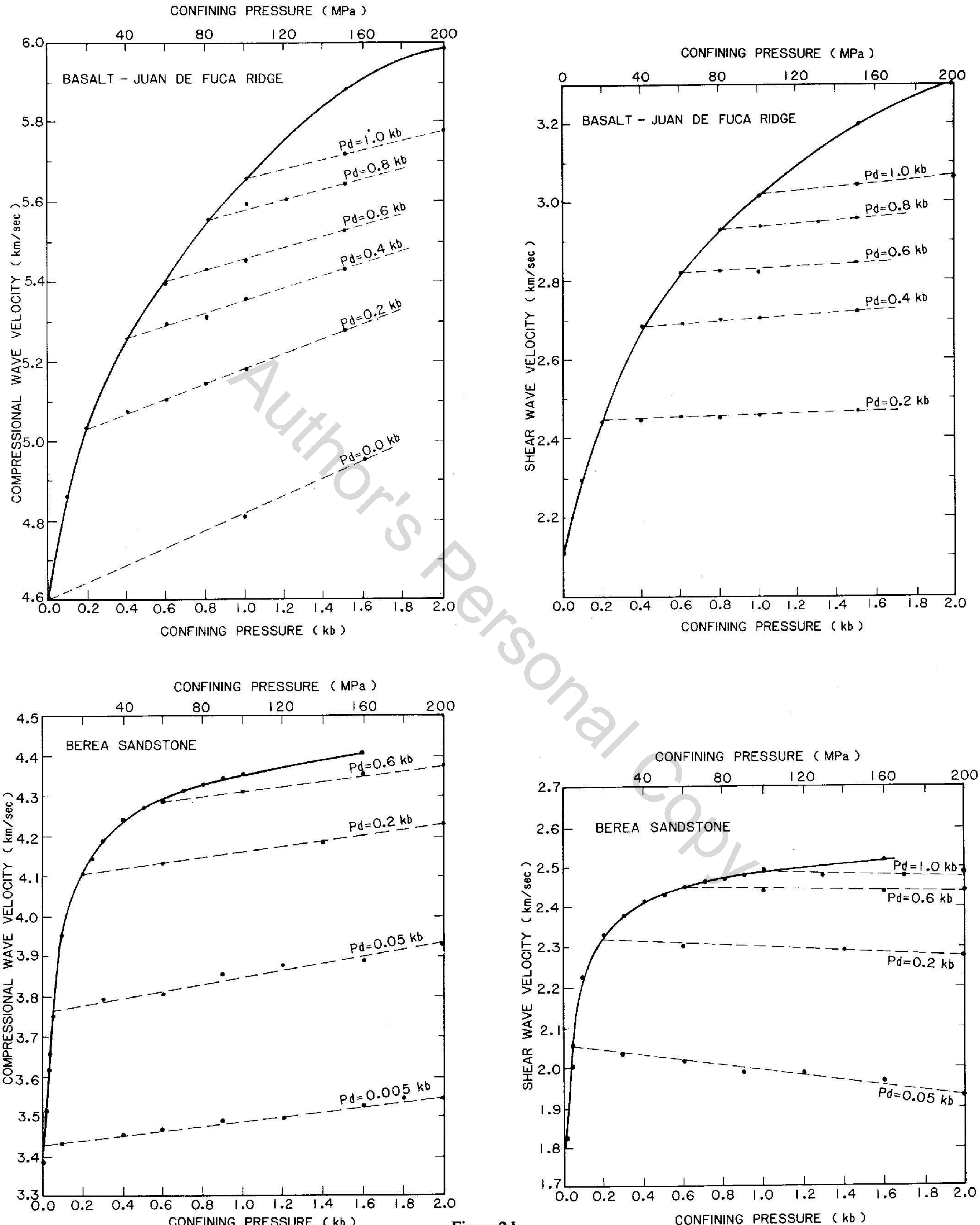
Figure 2c.


CONFINING PRESSURE (MPa)
BASALT – JUAN DE FUCA RIDGE
COMPRESSIONAL WAVE VELOCITY (km/sec)
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BASALT – JUAN DE FUCA RIDGE
SHEAR WAVE VELOCITY (km/sec)
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BEREA SANDSTONE
COMPRESSIONAL WAVE VELOCITY (km/sec)
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SHEAR WAVE VELOCITY (km/sec)
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Figure 2d.

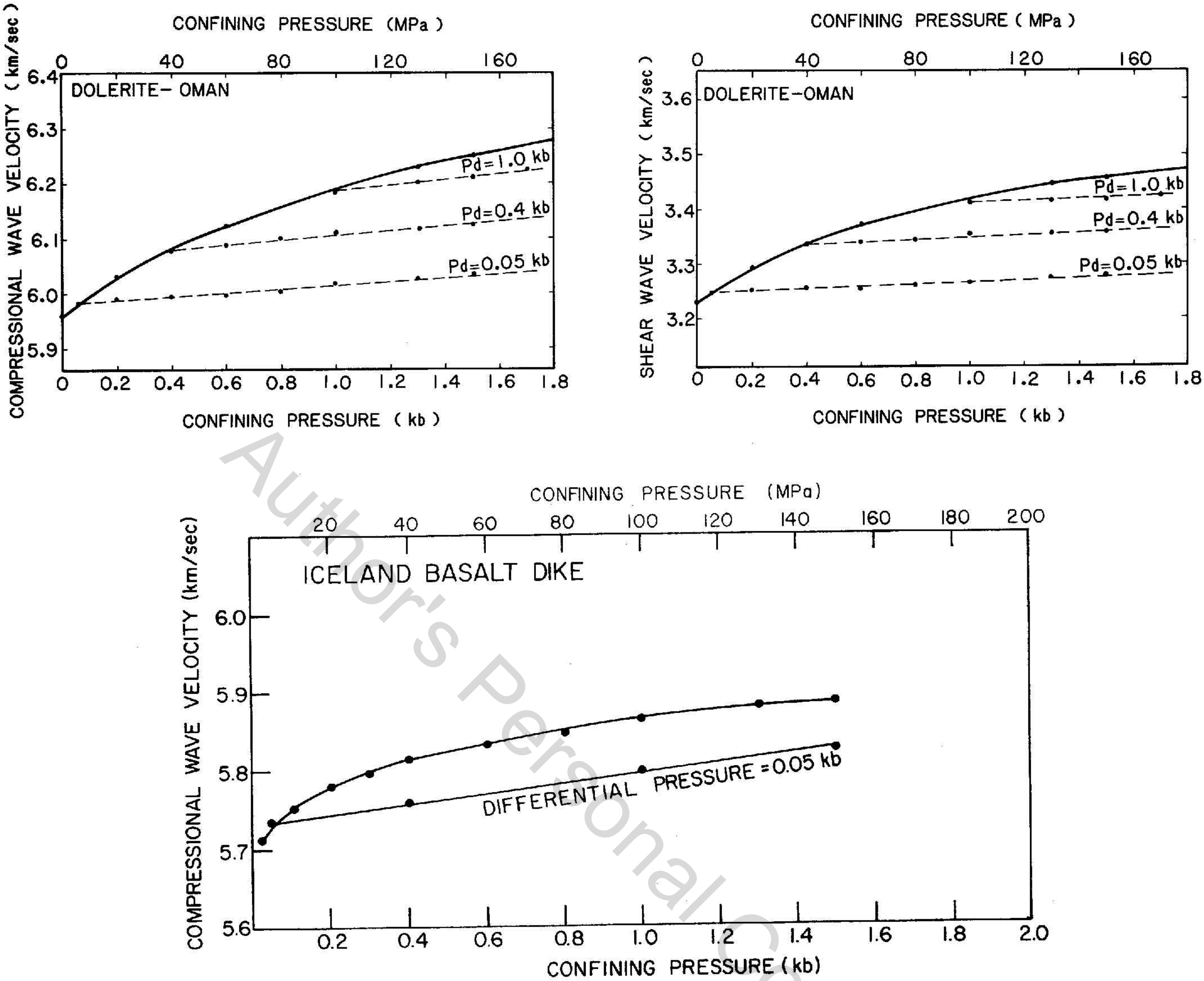

Figure 2e.

at $P_c = 200$ MPa (point 1, Fig. 3), pressures were lowered to $P_c = 100$ MPa and $P_p = 95$ MPa. The initial measured velocity was 5.98 km/s (point 2) and at equilibrium the velocity increased to 6.10 km/sec. Velocity behavior in going to point 5 and back to point 3 is also illustrated.

Compressional and shear-wave velocities were measured on the same samples during separate runs. Runs for a single rock lasted between 20 and 48 days, and pressures were maintained for an average of approximately 24 hr prior to each velocity measurement. The repeatability of a measured velocity is better than 0.5 percent; the accuracy is 1 percent.

VELOCITIES IN TERMS OF EFFECTIVE STRESS

To a first approximation, confining pressure (P_c) and pore pressure (P_p) are expected to have opposite, but roughly equal, effects on velocities. That is, the increase in velocity produced by an increase in confining pressure should be similar to the decrease in velocity produced by an equivalent increase in pore pressure. For this to hold, the lines of constant differential pressure (P_d) in Figure 2 will have zero slopes, i.e., $(\partial V_p/\partial P_c)P_d = 0$.

Several theoretical papers on wave propagation in fluid-saturated rocks (Brandt, 1955; Biot, 1962; Biot and Willis, 1957; Geertsma, 1957; Fatt, 1958) concluded that rather than differential pressure, velocities should be functions of an effective pressure, P_e, defined as $P_e = P_c - nP_p$, where n is an empirical factor usually less than 1. A value of n less than 1 implies that a pore pressure increment does not entirely cancel a confining-pressure increment. Experimental studies usually found the effective pressure law to hold, with values of n less than 1 common for a variety of rocks (Banthia and others, 1965; Todd and Simmons,

TABLE 2. VALUES OF n FOR $P_p = 0$

Rock		*Pd* (MPa)							
		0.5	5	10	20	40	60	80	100
Berea Sandstone	Vp	0.99	0.98	–	0.95	–	0.70	–	–
	VS	–	1.02	–	1.03	–	0.04	–	1.16
Mt. Simon Sandstone (1920)	Vp	–	1.00	–	0.92	–	0.76	–	–
	Vs	–	1.03	–	1.06	–	1.07	–	–
Mt. Simon Sandstone (2008)	Vp	–	1.02	1.02	1.02	–	1.02	–	–
	Vs	–	1.02	1.02	1.02	–	1.02	–	–
Oneota Dolomite	Vp	1.00	0.98	–	0.87	–	0.68	–	–
	Vs	–	1.00	–	0.99	–	0.88	–	–
Mt. St. Helens Andesite	Vp	–	0.98	–	0.94	0.90	–	0.85	–
	Vs	–	–	–	0.96	0.93	–	0.85	–
Juan de Fuca Basalt	Vp	0.98	–	–	0.86	0.84	0.81	0.81	–
	Vs	–	–	–	0.98	0.96	0.95	0.93	–
Iceland Basalt	Vp	–	0.87	–	–	–	–	–	–
	Vs	–	–	–	–	–	–	–	–
Illinois Granite	Vp	–	1.00	–	0.99	–	0.99	–	–
	Vs	–	1.00	–	0.99	0.99	0.95	–	–
Oman Dolerite	Vp	–	0.91	–	–	0.81	–	–	0.63
	Vsq–	–	0.95	–	–	0.87	–	–	0.81
Kilborne Lherzolite	Vp	–	0.94	–	0.89	0.87	0.86	–	0.79
	Vs	–	–	–	0.96	0.95	–	–	0.91

1972; Dominico, 1977; Christensen, 1984). Christensen and Wang (1985) found that, for equal increments of increased confining and pore pressure in Berea sandstone, V_p increased and V_s decreased. Thus, n is less than 1 for compressional waves but greater than 1 for shear waves. It was postulated that this behavior originated from the presence of high-compressibility clay that lines pores within the quartz framework of the sandstone.

Values of n can be calculated from the expression (Todd and Simmons, 1972; Christensen, 1984):

$$n = 1 - \frac{(\partial V/\partial P_p)_{P_d}}{(\partial V/\partial P_d)_{P_p}}.$$

Values calculated along the curve $P_p = 0$ are given in Table 2. The slopes $(\partial V/\partial P_c)_{P_d}$ of the dashed lines in Figure 2 were used for the $(\partial V/\partial P_p)_{P_d}$ term, since $\partial P_p = \partial P_c$ for constant P_d. The $(\partial V/\partial P_d)_{P_p}$ term was approximated from a curve fit to the P_p = 0 data.

The n values presented in Table 2 show that the effective pressure law holds for the rocks included in this study. For compressional waves, n is usually less than unity and decreases with increasing differential pressure. The three sandstones (Fig. 4), as well as the dolomite, show relatively large decreases in n with increasing confining pressure, which is likely related to the nature of their porosity. Similarly, shear-wave values of n for the crystalline rocks are less than unity and decrease with increasing differential pressure. The three sandstones again are anomalous, with shear-wave n values greater than 1, which remain constant or increase with increasing differential pressure.

APPLICATIONS TO CRUSTAL SEISMIC STRUCTURE

As interest in crustal fluids has increased, a large number of papers have appeared dealing either directly or indirectly with the distribution of water in the crust and the geochemical and geophysical consequences (e.g., O'Neil and Hanks, 1980; Etheridge and others, 1984; Angevine and Turcotte, 1983; Walder and Nur, 1984; Christensen, 1984; Newmark and others, 1985). Several studies dealing primarily with the sedimentary portion of the

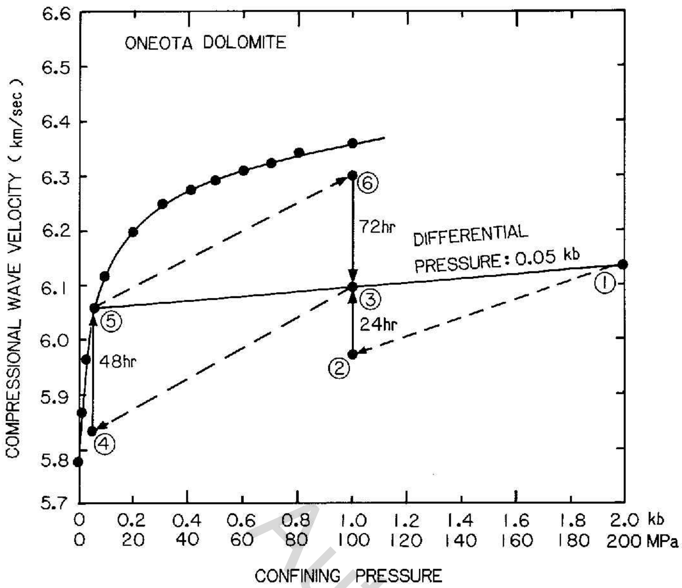

Figure 3. Velocity measurements along a line of constant differential pressure illustrating equilibrium time (see text).

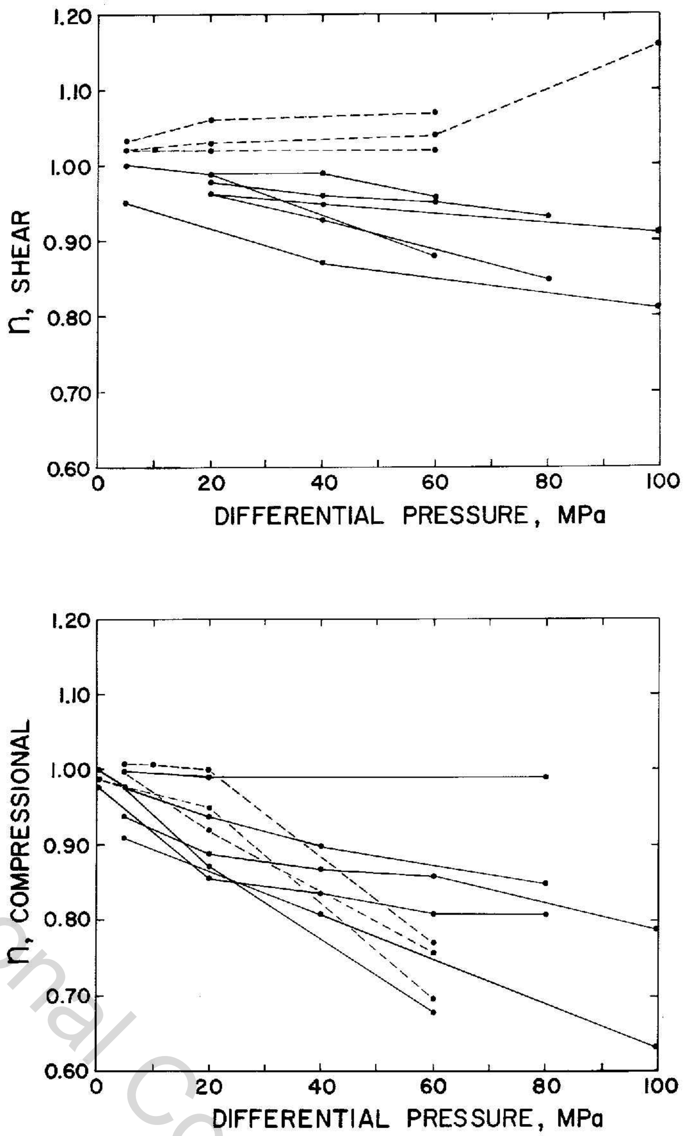

Figure 4. Values of *n* at various differential pressures (see text for definition of *n*). Sandstones are shown as dashed lines.

Earth's crust have suggested possible mechanisms that could produce high pore pressures in fluid-saturated rocks. For example, tectonic compression (Hubbert and Rubey, 1959; Fertl, 1976), porosity reduction by precipitation of minerals (Levorsen, 1954), and mineral reactions that release water such as the smectite to illite or gypsum to anhydrite transformations (Burst, 1969; Heard and Rubey, 1966) are all likely to produce excess pore pressures in sedimentary sections. It is apparent that the creation of high pore pressures in sedimentary formations will produce anomalously low seismic velocities. Thus seismic field observations of low velocities or atypically low rates of increase in velocity with depth could easily be attributed to high pore pressure. However, consideration of temperature and changes in mineralogy with depth will complicate this interpretation. If related to pore pressure, the velocities will depend on the magnitudes of the confining and pore pressures based on the effective-pressure law. The response will also be related to porosity: velocities of higher porosity rocks will show larger changes than low-porosity rocks for a given change in pore pressure (Fig. 5).

Many theoretical studies dealing with the influence of porosity on the physical properties of rocks have found that pore aspect ratios are critical parameters (e.g., Walsh, 1965). Crystalline rocks have porosities consisting primarily of microcracks with high aspect ratios, whereas the sandstones and dolomite porosities are in the form of low-aspect ratio pores. Thus it is not surprising that the influence of pore pressure on velocities in igneous and metamorphic rocks is quite different from sedimentary rocks. The results plotted in Figure 5 show that for a given porosity, the velocity response to a change in pore pressure is much greater for crystalline rocks. Even for small porosities, pore pressure markedly affects velocities of the igneous and metamorphic rocks.

As pore pressure is increased to confining pressure, changes in velocity are not linear functions of pore pressure. Initial increases in pore pressure lower velocities only slightly. Maximum decreases in velocity occur as pore pressure approaches confining pressure. This is illustrated in Figure 6 for four crystalline rocks with a range of porosities.

Since increasing pore pressure at constant confining pressure has a greater effect on shear velocities than compressional velocities (Fig. 2), V_p and V_s ratios decrease and Poisson's ratios increase (Fig. 7). Thus, V_p and V_s measurements in overpressured sedimentary strata can give estimates of the magnitudes of the pore pressures, provided reasonable constraints are placed on porosities and mineralogies. For example, the two Mt. Simon sandstones, because of their high quartz contents, have extremely low Poisson's ratios (0.14 and 0.18) at elevated confining pressures and low pore pressures (Fig. 7). However, if pore pressures approach confining pressures, Poisson's ratios increase to values common for many quartz-free rocks at low pore pressures (0.22 and 0.28). Similarly, the 0.29 Poisson's ratio of the dolomite at elevated confining pressure and low pore pressure increases to 0.33 as pore pressures become equivalent to confining pressures.

The applications of these measurements to interpreting seismic velocities in the oceanic crust have been discussed by Christensen (1984). Within the upper few hundred meters of the volcanics that constitute oceanic layer 2, velocities are likely to be strongly influenced by pore pressure. For fractures connected to the surface, pore pressures will be hydrostatic. A major question, still unanswered, is to what depths porosity and water exist within the oceanic crust. Recent deep drilling in the Pacific at DSDP Site 504B found that porosity is extremely low at about 1-km depth below the sea floor (e.g., Salisbury and others, 1985). Deep drilling in Icelandic volcanics, however, recovered cores with large vugs lined with quartz, epidote, and carbonate at depths of 1,900 m (Mehegan and others, 1982).

Several major questions arise concerning pore fluids in the continental crust. What are the vertical and lateral distributions of pore fluid? To what depths are pore fluids no longer connected through crack systems to the surface? What geologic processes are responsible for trapping pore fluids at depth? Are aqueous fluids or CO_2 more important as lower continental crustal pore fluids? The answers to these questions have important implications for the seismic properties of the crust as well as other physical properties.

Researchers concerned with the electrical conductivity structure of the continental crust for many years have been concerned with the abundance and nature of crustal pore fluids. The role of pore water in producing high electrical conductivity is well documented from laboratory studies (e.g., Brace, 1972; Olhoeft, 1981). Thus, aqueous fluids are generally considered to be the principal source of high crustal conductivity (e.g., Connerney and others, 1980; Olhoeft, 1981; Jiracek and others, 1983).

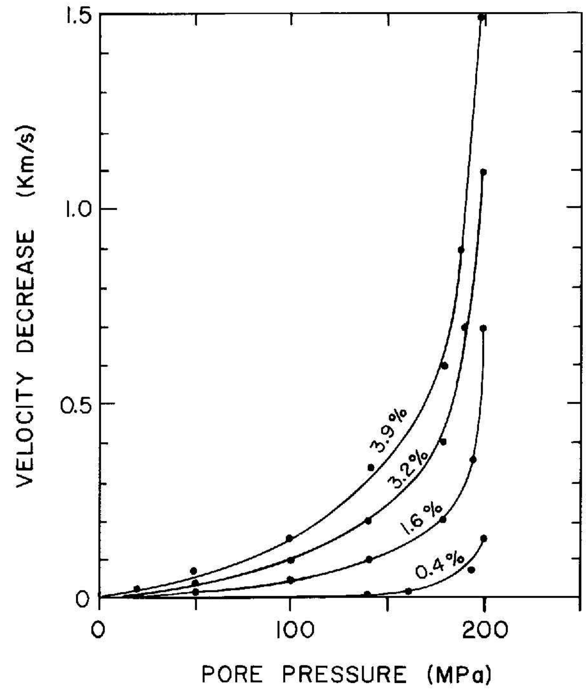

Figure 6. Velocity decreases at 200 MPa confining pressure with increasing pore pressure for basalt (porosity = 0.4 percent), dolerite (porosity = 1.6 percent), andesite (porosity = 3.2 percent), and lherzolite (porosity = 3.9 percent).

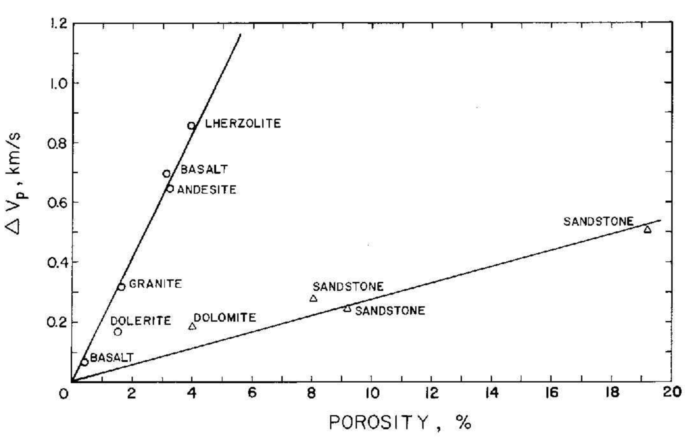

Figure 5. Compressional wave velocity decreases at 100 MPa confining pressure produced by an increase in pore pressure of 95 MPa.

Several authors have cited evidence for high electrical conductivity of the lower continental crust (e.g., Hyndman and Hyndman, 1968; Shankland and Ander, 1983; Jones, 1987). Meissner (1986), on the other hand, has cited regions where the lower crust apparently has low electrical conductivity; he emphasized problems in resolution of vertical structure using electromagnetic investigations. Critical to the present study are the observations of Olhoeft (1981), which indicate that hydrous minerals such as amphibole and mica have relatively low conductivities. Thus, lower crustal high-conductivity regions likely contain aqueous fluids that may occur at elevated pore pressures. Fyfe and others (1978) suggested that, under lower crustal conditions, fluid pressures prevail, often approaching lithostatic pressure.

Within the continental crust, progressive metamorphism proceeds by a sequence of reactions involving dehydration (Bowen, 1940; Thompson, 1955; Turner and Weiss, 1963). Common examples include the metamorphism of pelitic rocks, where shale is converted to slate, phyllite, mica schist, and granulite, and the reactions of mafic greenschist through amphibolite to mafic granulite. Reactions involving hydrous silicates such as micas, chlorites, and amphiboles are thus likely to be important in providing H_2O pore fluid at significant crustal depths (e.g., Burst, 1969; Meissner, 1986). In these regions the assumption that the pore pressure of

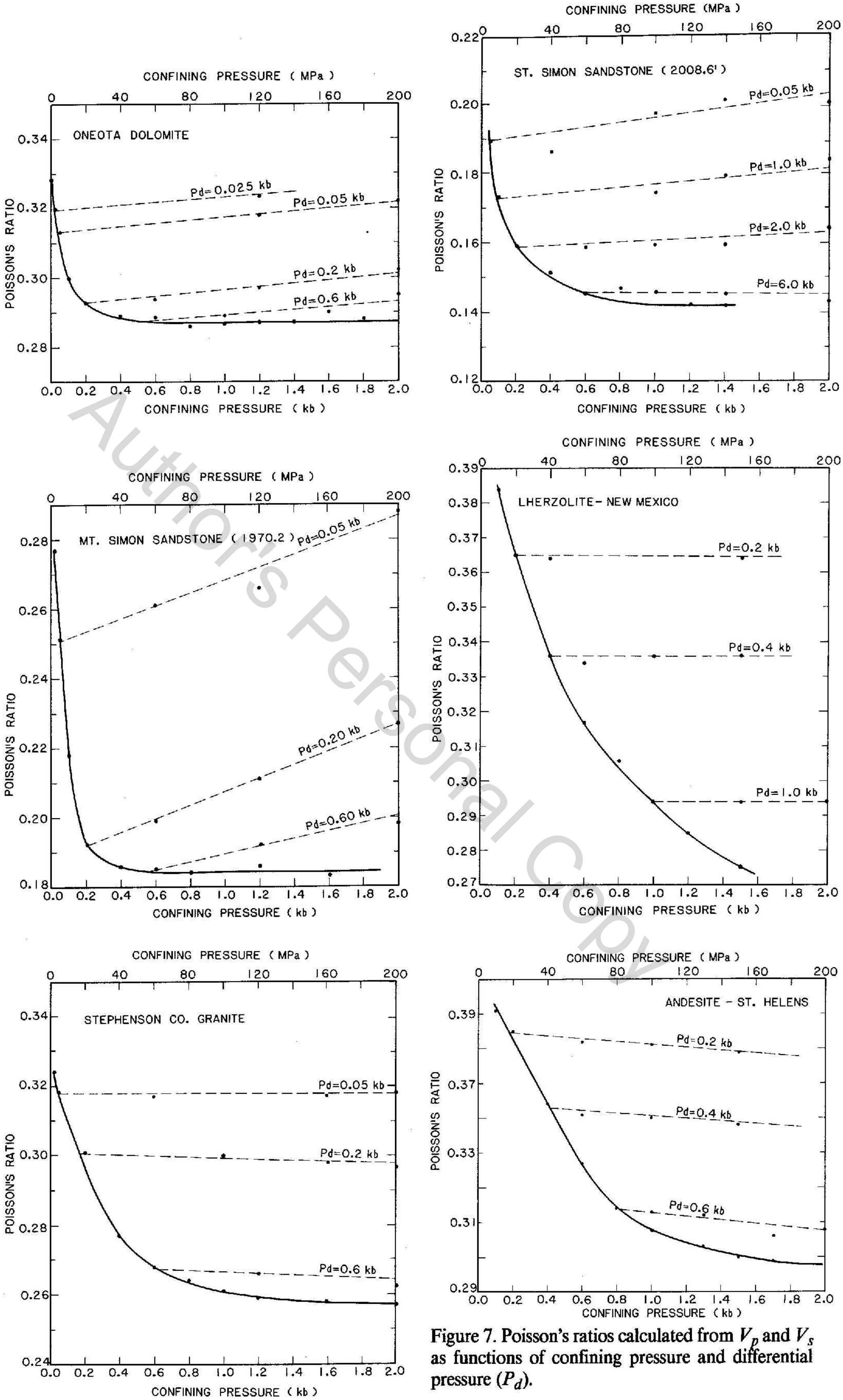

Figure 7. Poisson's ratios calculated from V_p and V_s as functions of confining pressure and differential pressure (P_d).

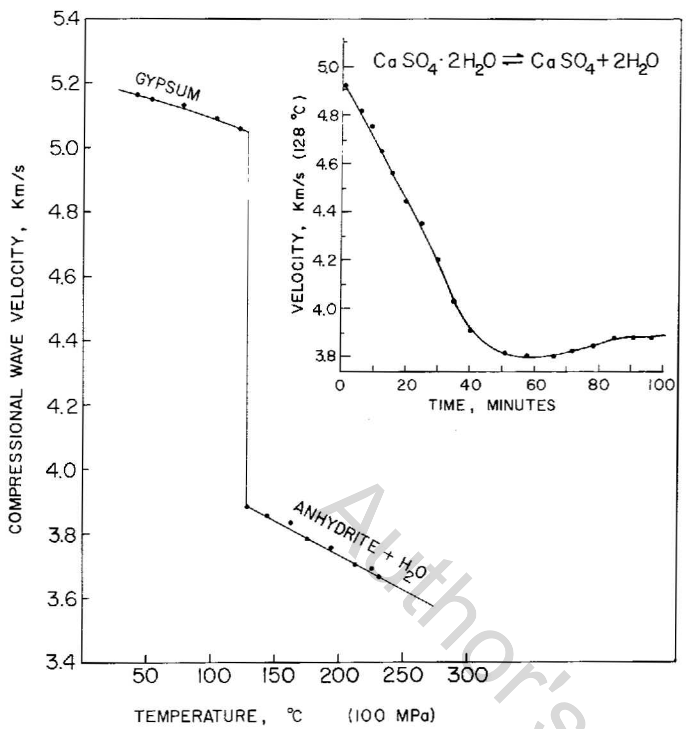

Figure 8. Changes in velocity associated with the gypsum-anhydrite transition. Inset shows velocity as a function of time at 128°C.

H_2O is equivalent to load pressure must, in many instances, be incorrect, since overpressures are likely where permeabilities are low.

The gypsum-anhydrite reaction:

$$\underset{\text{gypsum}}{CaSO_4 \cdot 2H_2O} \rightleftharpoons \underset{\text{anhydrite}}{CaSO_4} + \underset{\text{water}}{2H_2O}$$

illustrates the lowering of seismic velocities that accompanies crustal metamorphic reactions involving pore-pressure buildup during dehydration. The transition temperature is complicated by many factors, including the presence of impurities, the original gypsum porosity, and grain size, pressure, and often the formation of the semihydrate bassanite (Sonnenfeld, 1984). Gypsum has a compressional-wave velocity of approximately 5.2 km/sec at confining pressures high enough to eliminate most crack effects on velocities, whereas anhydrite has velocities of approximately 6.0 km/sec (Birch, 1960). Thus if H_2O is allowed to escape during the dehydration process, velocities will increase rapidly once the rotation proceeds to anhydrite. The anhydrite and water together occupy about 11 percent more volume than the original gypsum. However, if the released pore fluid is contained, much different results are observed due to fluid-pressure buildup (Fig. 8). In this experiment, a monomineralic aggregate of gypsum was completely enclosed in a copper jacket. The sample was placed under a hydrostatic confining pressure of 100 MPa, and velocities were measured with increasing temperature. Velocities first decreased slightly due to the effect of increasing temperature on the gypsum aggregate. At approximately 128°C, velocities decrease significantly due to pore pressure associated with the dehydration process. During the reaction, which occurred over a time interval of 60 min (Fig. 8), the signal from the receiving transducer remained on the oscilloscope screen, and a steady decrease in velocity was observed. After completion of the reaction, a further increase in temperature provided velocity measurements in the assemblage anhydrite + H_2O. A puncture in the copper jacket terminated the experiment. The recovered sample consisted of anhydrite; it is not known if the semihydrate formed during the experiment.

Experiments showing major velocity decreases associated with metamorphic dehydration are particularly important in view of recent findings from deep continental drilling. In 1970 the Soviet Union began a major drilling project in the northeast part of the Baltic shield on the Kola Peninsula above the Arctic Circle (Fig. 9). In 1984 a book published in Russian (Kozlovskiy, 1984) summarized in detail the drilling results to a depth of 11.6 km, nearly one-third of the total continental crustal thickness in this region. The upper 6,847 m of the hole penetrated Proterozoic tholeiitic metabasalt, metapyroxenite, metaperidotite, metatuffs, metasandstones, metaarkoses, and carbonates. Below 6,847 m, the drilling recovered Archean metamorphics consisting of mica-plagioclase gneisses in the upper section and biotite-plagioclase gneisses and amphibolites in the lower section (Fig. 10). Of particular significance was the finding of porous, water-saturated fractured rocks in the lower section of the drillhole that are not connected hydraulically with the overlying zone. At all depths, fractures are coated with various mineral fillings, indicating that they were not formed by the drilling process. In general, porosity was found to *increase* with depth, and logging velocities *decreased* with depth.

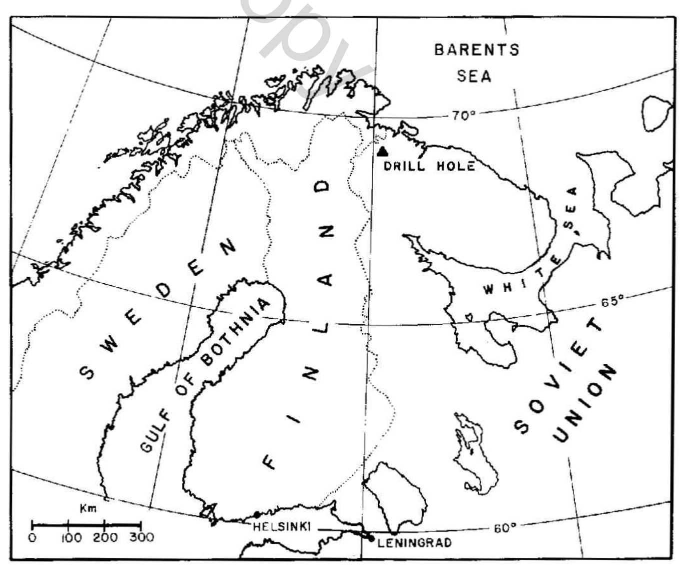

Figure 9. Location of Kola drillhole.

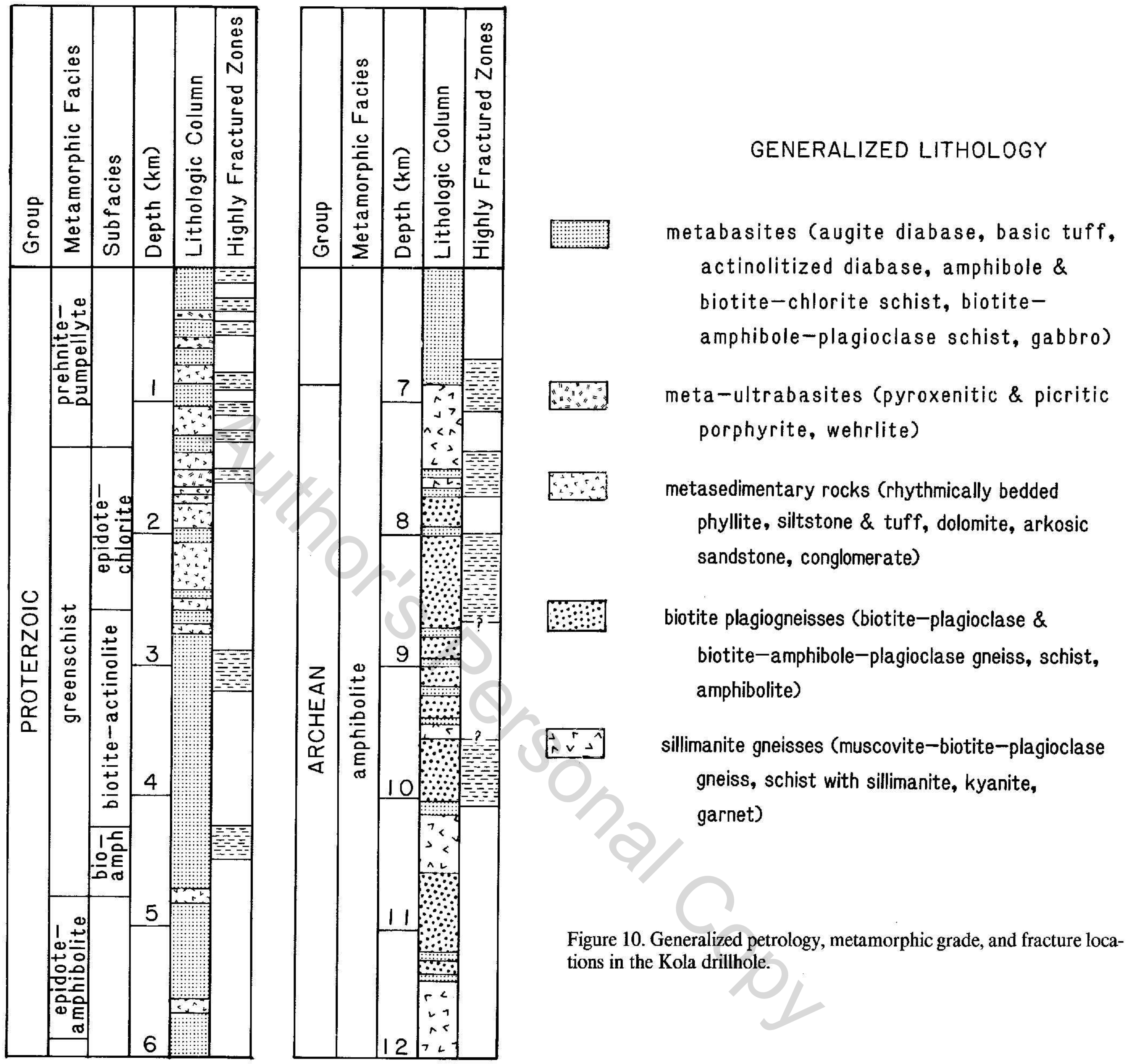

Figure 10. Generalized petrology, metamorphic grade, and fracture locations in the Kola drillhole.

The zones of increased fracturing are nonuniform in depth and separated by nearly impermeable regions. Chemical analyses and textural observations suggest that the origin of water at depths below 4.5 km was mineral dehydration formed during metamorphism. Calculations based on dehydration reactions associated with the greenschist to epidote-amphibolite facies transition show that, in the depth interval of 4.0 to 6.8 km, total water losses were 5.5×10^7 g/m^3, equivalent to 6.7 percent of the original rock volume (Kozlovskiy, 1984). This water loss, if confined, would produce pore pressures of approximately 3 GPa, equivalent to expected mantle pressures at a 100-km depth.

Laboratory experiments on sediments have shown that high pore pressures in rocks containing nonaqueous fluid also lower velocities (Banthia and others, 1965). A preliminary study of compressional-wave velocities in basalt with helium as a pore-pressure saturant shows that crystalline rocks behave in a similar manner (Fig. 11). Note that at equivalent confining and pore pressures, velocities are higher for the water-saturated sample. Also, lines of constant differential pressure for He saturation have lower slopes. Thus, n values are closer to unity with He as a pore-pressure medium. Calculated values of n along the $P_p = 0$ curve range from 0.792 at $P_d = 30$ MPa to 0.940 at $P_d = 80$ MPa.

Studies of mineral equilibria in rocks from exposed lower crustal granulite facies terranes often find evidence for low H_2O activities. One theory to explain these low activities hypothesizes that granulites form from a major influx of CO_2 that dilutes H_2O (Schuiling and Kreulen, 1979; Newton and others, 1980). This is supported by observations of CO_2–rich fluid inclusions in granulites. However, other probable causes of low H_2O activity at lower crustal depths include preferential partitioning of H_2O into partial melt and recrystallization of dry rocks (Valley and O'Neil, 1984). The possibility of the presence of CO_2, at least locally, in the lower continental crust raises speculation as to whether elevated CO_2 pore pressure could play an important role in lowering seismic velocities. Of significance, abundant lower crustal CO_2, because of its nonpolar nature, would not produce a region of high electrical conductivity (Kay and Kay, 1981).

CONCLUSIONS

It is becoming apparent that porosity containing aqueous fluids occurs at considerable depths within the crystalline rocks of the Earth's crust. The interpretation of seismic velocities in these regions requires knowledge of the in-situ pore pressure in addition to confining pressure and temperature. Laboratory studies demonstrate that the application of pore pressure decreases velocities in igneous and metamorphic rocks to values usually attributed to high crustal temperatures, extensive fracturing, or even partial melting.

Velocities in crustal sedimentary sections will be anomalously low if pore pressures and porosities are high. High Poisson's ratios are characteristic of these overpressured regions. Within the oceanic crust, pore pressures within cracks connected to the overlying sea-water column cause lower velocities and may be responsible for some observations of horizontal variability in the velocities of seismic layer 2.

If the findings of the Soviet Union Kola drilling project are typical of continental crustal regions, it is likely that much of the velocity structure of the upper continental crust is directly affected by pore pressure. If connected to the surface, water-saturated regions in the upper few kilometers of the crust are likely to have lowered velocities due to hydrostatic pore pressures. At greater depths, where porous fractured metamorphic rocks are not connected hydraulically with near-surface fractures, pore pressures generated by metamorphic processes and possibly tectonic forces are likely to produce a dramatic lowering of seismic velocities. Dehydration with accompanying high pore pressure resulting from progressive metamorphism is likely to produce hydraulic fracturing. In some deeper crustal regions, high-grade metamorphism to the granulite facies has produced relatively anhydrous coarse-grained rocks, and fluids rich in CO_2 at elevated pore pressure may be important.

Refraction crustal studies have reported a variety of crustal structures at intermediate depths (e.g., Meissner, 1986). Velocity inversions are sometimes observed (e.g., Berry and Fuchs, 1973); a major first-order discontinuity, often referred to as the Conrad (e.g., Steinhart and Meyer, 1961), is common in many regions; and strong positive velocity gradients are frequently reported (e.g., Pavlenkova, 1979). One of many explanations for the origin of low-velocity zones is high pore pressure that arises from dehydration reactions likely involving mica and amphibole. As was discussed earlier, the low-velocity zones will be characterized by high Poisson's ratios and high electrical conductivities. Rather than originating from a major lithologic change, the Conrad discontinuity may in many regions separate an overpressured upper crust from a "dry" lower crustal region. Velocities will increase abruptly across the discontinuity, and electrical conductivity will decrease. In midcrustal regions, where pore pressures are low due to loss of fluids, progressive metamorphism with increasing depth, if isochemical, will be observed as a positive velocity gradient.

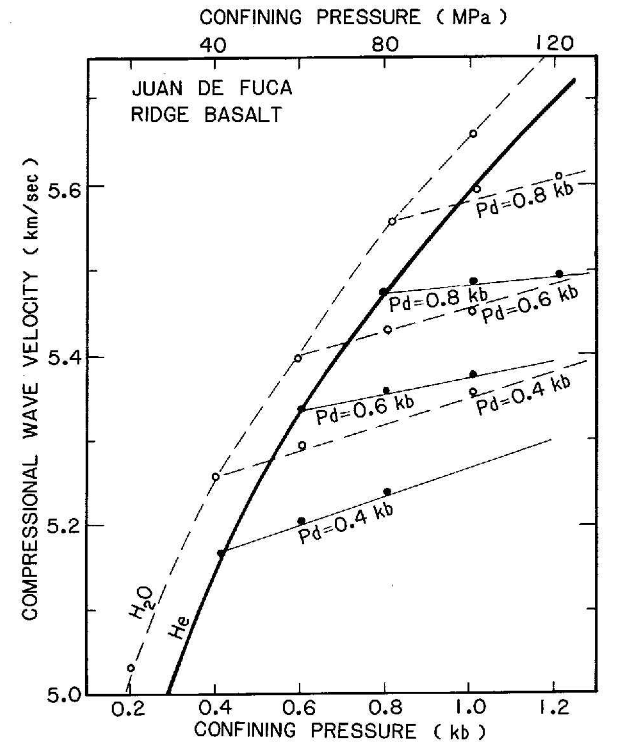

Figure 11. Compressional wave velocities in basalt with water (dashed curves) and helium (solid curves) as pore pressure media.

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