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Pore pressure and oceanic crustal seismic structure

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Summary. Marine refraction studies during the past decade have found considerable lateral variability in the seismic properties of the upper basaltic regions of the oceanic crust. In many localities, compressional and shear-wave velocities are quite low at the top of the basalt section and velocities increase rapidly with depth. It is concluded that pore pressure may be at least in part responsible for the low upper crustal velocities and contribute significantly to the lateral variability. This is supported by compressional and shear-wave velocity measurements as functions of confining pressure and pore pressure for basalt from the Juan de Fuca ridge and dolerite from the Samail ophiolite, Oman. Within the oceanic crust, regions of overpressure and underpressure will possess anomalous velocities, the magnitude of which will depend upon the porosity and the deviation of the pore pressure from hydrostatic. The influence of pore pressure on velocities is expected to diminish with depth and is unlikely to be significant at lower crustal depths where porosity is extremely low. Of significance, Poisson's ratio is shown to be dependent on pore pressure as well as confining pressure. At constant confining pressure, Poisson's ratio increases with increasing pore pressure. Thus, overpressured regions within the upper oceanic crust are likely to have relatively high Poisson's ratios as well as low compressional- and shear-wave velocities.

1 Introduction

Beginning in the early 1950s, the seismic structure of the oceanic crust has been the subject of numerous investigations. The results of early refraction studies at sea often showed a simple three-layered crustal model with relatively uniform velocities and thickness of individual layers (Raitt 1963). More recent seismic experiments using multichannel seismic techniques have determined detailed velocity–depth structures through the oceanic crust at several locations. Of particular significance, it has been found that the uppermost basalt of the oceanic crust has relatively low velocities, usually between 3.6 and 4.8 km s^{-1} (Houtz & Ewing 1976), which increase rapidly with depth giving rise to compressional wave velocity gradients in the upper 2 km of the crust averaging between 0.5 and 2.0 s^{-1} (Kennett 1977;

Whitmarsh 1978; Houtz 1980; Spudich & Orcutt 1980). In addition, recent refraction profiles recorded on ocean bottom seismometers by Au & Clowes (1984) confirm similar low shear-wave velocities in the upper basaltic crust followed by a rapid increase in velocity with depth. The low velocities and their variability in the upper oceanic crust immediately underlying deep sea sediments are generally attributed to the presence of high porosity, rubbly pillow basalt and breccia zones at shallow depths (e.g. Hyndman & Drury 1976) with the rapid increases in velocities with depth originating from decreasing porosity.

Marine heat flow observations, electrical soundings and geophysical logging have established that the pores and cracks are water saturated and often likely to be interconnected (Lister 1972; Kirkpatrick 1979; Hermance, Nur & Bjornsson 1972). This presence of pore water within the upper oceanic crust influences a wide range of properties including fracture strength, electrical conductivity and seismic velocities (Raleigh & Paterson 1965; Brace & Orange 1968; Christensen 1970). These properties are likely to depend not only on the external confining pressure at depth, but also on the fluid pressure within the rock cavities.

The question arises as to what extent pore pressure may be responsible for lowering velocities within the upper oceanic crust, as well as producing the observed variability in upper crustal seismic velocities. Laboratory velocity measurements on high porosity sedimentary rocks (Wyllie, Gregory & Gardner 1958; King 1966; Domenico 1977) and granite (Todd & Simmons 1972) under both controlled confining pressure and pore pressure have found that increasing pore pressure at constant confining pressure significantly lowers compressional wave velocity. This also has been experimentally verified for shear-wave velocities in sedimentary rocks (Banthia, King & Fatt 1965). Clearly similar results for oceanic basalt would demonstrate the importance of pore pressure as a parameter influencing oceanic crustal seismic properties.

In this study, compressional- and shear-wave velocities have been measured under controlled pore and external pressures for oceanic basalt and dolerite. It is concluded, as shown from these measurements, that any interpretation of crustal velocities in terms of composition, crack porosity or density must take into account pore pressure.

2 Experimental details and data collection

The basalt sample selected for the velocity measurements, a tholeiitic pillow basalt, was dredged from the western edge of the median valley of the Juan de Fuca ridge at a water depth of 2470 m. A modal analysis from a single thin section is as follows: 40 per cent plagioclase (An_{58-65}), 41 per cent clinopyroxene, 6 per cent opaque and 13 per cent basaltic glass and alteration products. Compressional-wave velocities measured from cores cut from the same pillow (Christensen 1970) demonstrate no significant anisotropy, but a marked dependence of velocity on the degree of water saturation within the rock.

A virgin sample approximately 2.5 cm in diameter and 5 cm long was cored from the basalt pillow. The ends were trimmed normal to the axis and parallel to 0.02 mm. The wet bulk density of the sample calculated from its dimensions and water saturated weight is 2.91 g cm^{-3}. Porosity occurring as connected microcracks and small vesicles calculated from wet and dry bulk densities is 4.0 per cent.

The dolerite was collected from the central portion of the sheeted dike section of the Samail ophiolite, Oman near Wadi Jizi at an estimated depth of 2.4 km beneath the sediment–basalt contact. The stratigraphic position of the sample within the ophiolite complex, as well as the laboratory measured velocities, place the sample within the lower portion of layer 2 of the oceanic crust (Christensen & Smewing 1981). Thus the velocity data should provide information on the influence of pore pressure on elastic properties of rocks from intermediate crustal depths.

The dolerite sample has been metamorphosed to the greenschist facies through the circulation of sea water at a mid-ocean ridge (Gregory & Taylor 1981). Mineral percentages by volume are: 48 per cent plagioclase (altered), 25 per cent clinopyroxene, 5 per cent opaque and 22 per cent alteration products consisting of albite, chlorite, epidote, actinolite, sphene and quartz. The wet bulk density of the sample is 2.8 g cm^{-3} and its porosity is 1.1 per cent.

The samples were encircled by thin shell aluminium containment cylinders (Fig. 1) containing 16 shallow longitudinal slots on their inside surface which are ported to the pore pressure pumping system, thereby exposing the circumferential surface of the sample to a controlled pore pressure. Each containment cylinder has a full length narrow opening filled with epoxy which allows closure with increasing confining pressure and assures unrestricted hydrostatic pressure on the sample. The electrodes and transducers are placed on the sleeved

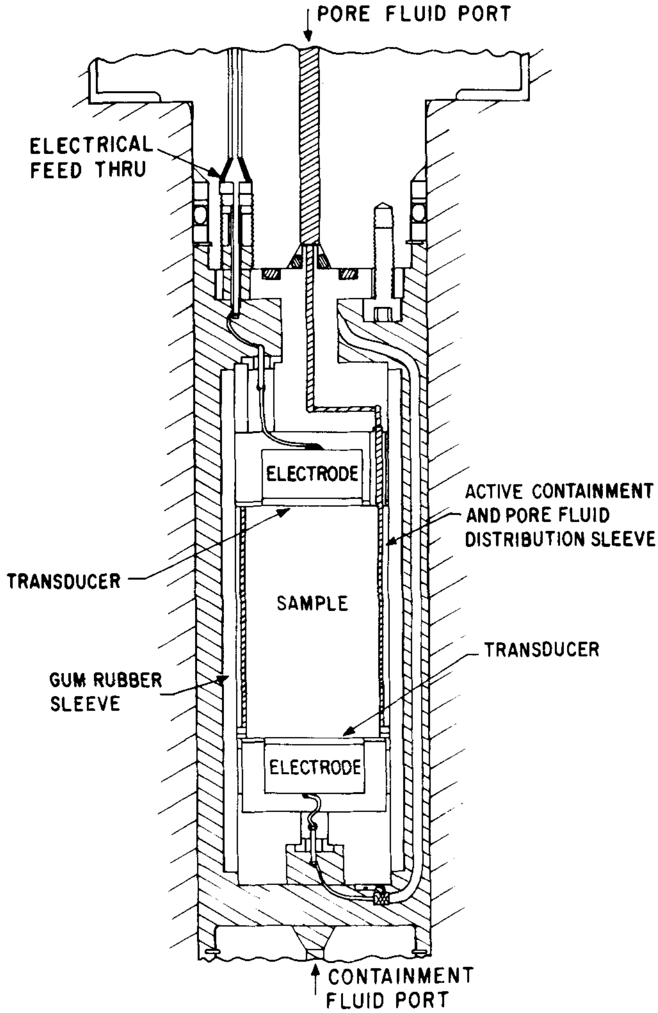

Figure 1. Diagram of sample assembly.

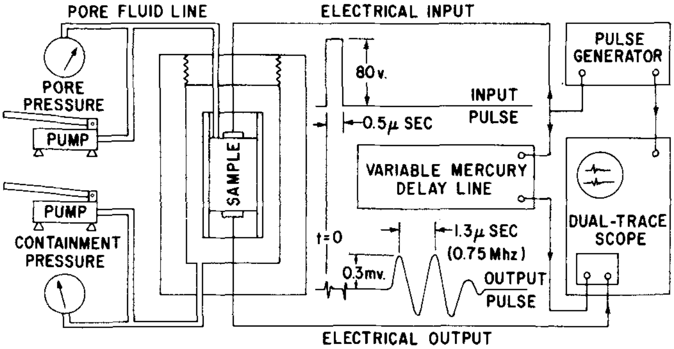

Figure 2. Schematic diagram of pressure system and electronics.

samples and jacketed with gum rubber tubing to prevent interaction of the containment and pore fluids. The mechanical configuration of the transducer electrodes, the electrical feed-throughs and the porting for pore and confining pressures are shown in Fig. 1.

A schematic diagram of the electrical and hydraulic systems designed for the velocity measurements is shown in Fig. 2. A rectangular pulse of approximately 80 V and 0.5 μs in width drives a sending transducer on one end of the rock core Lead-zirconate trans-

Table 1. Compressional wave velocities (V_p), shear wave velocities (V_s), and Poisson's ratios (σ) at various confining pressures (P_c) and pore pressures (P_p) for Juan de Fuca ridge basalt.

P_c(kbar)	P_p(kbar)	V_p(km/s)	V_s(km/s)	V_p/V_s	σ
0	0	4.601	2.110	2.181	0.367
0.1	0	4.859	2.299	2.114	0.356
0.2	0	5.034	2.441	2.062	0.346
0.4	0	5.258	2.682	1.960	0.324
0.4	0.2	5.074	2.444	2.076	0.349
0.6	0	5.396	2.814	1.918	0.314
0.6	0.2	5.293	2.689	1.968	0.326
0.6	0.4	5.103	2.452	2.081	0.350
0.8	0	5.556	2.928	1.898	0.308
0.8	0.2	5.430	2.822	1.924	0.315
0.8	0.4	5.307	2.704	1.963	0.325
0.8	0.6	5.147	2.450	2.101	0.353
1.0	0	5.660	3.012	1.879	0.303
1.0	0.2	5.598	2.936	1.907	0.310
1.0	0.4	5.458	2.822	1.934	0.318
1.0	0.6	5.359	2.704	1.982	0.329
1.0	0.8	5.184	2.458	2.109	0.355
1.0	1.0	4.810	-	-	-
1.2	0.4	5.603	-	-	-
1.5	0	5.888	3.200	1.840	0.290
1.5	0.5	5.723	3.041	1.882	0.303
1.5	0.7	5.648	2.953	1.913	0.313
1.5	0.9	5.529	2.844	1.944	0.320
1.5	1.1	5.430	2.721	2.000	0.333
1.5	1.3	5.280	2.469	2.139	0.360
1.6	1.6	4.950	-	-	-
2.0	0	5.980	3.305	1.809	0.280
2.0	1.0	5.773	3.057	1.888	0.306

Table 2. Compressional wave velocities (V_p), shear wave velocities (V_s), and Poisson's ratios (σ) at various confining pressures (P_c) and pore pressures (P_p) for Oman dolerite.

P_c(kbar)	P_p(kbar)	V_p(km/s)	V_s(km/s)	V_p/V_s	σ
0	0	5.960	3.230	1.845	0.292
0.05	0	5.979	3.243	1.844	0.292
0.2	0	6.031	3.290	1.833	0.288
0.2	0.15	5.986	3.251	1.841	0.290
0.4	0	6.078	3.333	1.824	0.285
0.4	0.35	5.991	3.253	1.842	0.291
0.6	0	6.121	3.369	1.817	0.283
0.6	0.2	6.088	3.340	1.823	0.285
0.6	0.55	5.998	3.253	1.844	0.292
0.8	0.4	6.100	3.344	1.824	0.285
0.8	0.75	6.002	3.261	1.841	0.290
1.0	0	6.185	3.409	1.814	0.282
1.0	0.6	6.108	3.353	1.822	0.284
1.0	0.95	6.017	3.263	1.844	0.292
1.3	0	6.232	3.448	1.807	0.279
1.3	0.3	6.201	3.413	1.817	0.283
1.3	0.9	6.118	3.355	1.824	0.285
1.3	1.25	6.026	3.276	1.839	0.290
1.5	0	6.252	3.452	1.811	0.280
1.5	0.5	6.210	3.418	1.817	0.283
1.5	1.1	6.127	3.360	1.824	0.285
1.5	1.45	6.036	3.281	1.840	0.290
1.7	0.7	6.224	3.422	1.819	0.284

ducers are used for transmitting and receiving compressional waves, whereas AC-cut quartz and lead-zirconate transducers generate and receive the shear waves, respectively. The electrical output from the receiving transducer is displayed on one trace of a dual trace oscilloscope. The transit time of the pulse through a sample is measured by superimposing the signals from the sample and a calibrated variable length mercury delay line (Birch 1960).

The containment and pore pressure pumps have operating pressure ranges of atmospheric to 2.7 kbar (270 MPa). Pressures are each monitored with identical Heise gauges with operating pressure ranges of 0–2.5 kbar and accuracies of 0.1 per cent of full scale.

Distilled water, which is used as the pore pressure medium, is introduced after the sample and the pore pressure plumbing have been evacuated for several hours. Data points are taken while holding a constant differential pressure (confining pressure minus pore pressure). The confining pressure and the corresponding pore pressure, required to maintain a constant differential pressure, are increased and decreased at random for each constant differential pressure data set. The average time required to reach equilibrium at each data point is approximately 12 hr. The repeatability is 0.5 per cent of the measured velocity. The accuracy of a given velocity measurement is 1 per cent.

Compressional- and shear-wave velocities, the ratio of compressional- to shear-wave velocity, and Poisson's ratios calculated from the velocities are given in Tables 1 and 2 for the basalt and dolerite. In Figs 3–6, the velocities are plotted as a function of confining

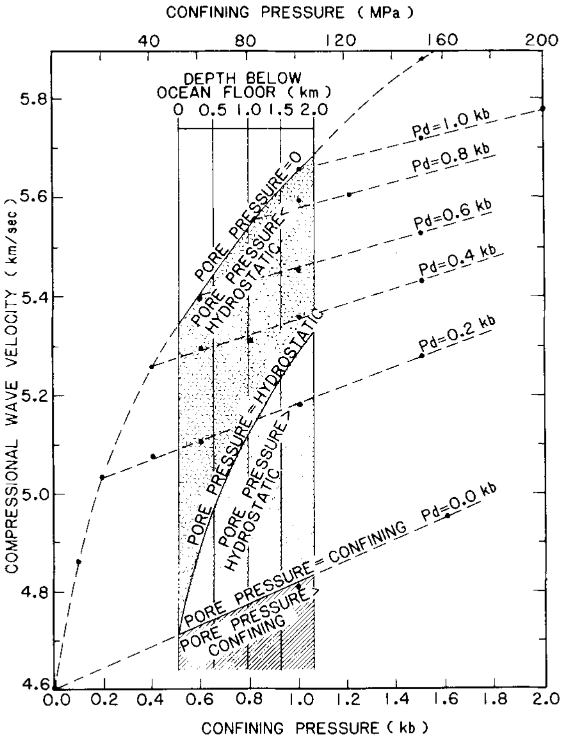

Figure 3. Basalt compressional-wave velocity as a function of confining pressure. Lines are shown for constant differential pressure (P_d).

pressures. The solid curves in Figs 5 and 6 show the variation of velocity with confining pressure at zero pore pressure and the data points beneath these curves represent velocities measured at elevated pore pressures. The dashed lines are lines of constant differential pressure, defined as confining pressure minus pore pressure. At a constant confining pressure, velocities decrease with increasing pore pressure along a vertical path in the figures (at any given point, the pore pressure is simply obtained by subtracting the differential pressure from the confining pressure).

3 Differential versus effective pressure

In addition to illustrating the dramatic decreases in velocities which accompany increasing pore pressure in oceanic rocks at fixed confining pressures, Figs 3–6 show that compressional- and shear-wave velocities are not constant at constant differential pressures. Over the pressure ranges of the measurements presented here, velocities increase at constant differential pressure as confining pressure is increased.

Several theoretical papers on acoustic wave propagation in fluid saturated porous elastic solids (Brandt 1955; Biot 1956, 1962; Biot & Willis 1957; Geertsma 1957; Fatt 1958) have concluded that seismic velocities are a function of an effective pressure $P_c - nP_p$ rather simple differential pressure $P_c - P_p$, where P_c is confining pressure, P_p is pore pressure and $n < 1$. Thus to maintain a constant velocity when confining pressure is increased, it is necessary to increase internal pore pressure an amount greater than the confining pressure.

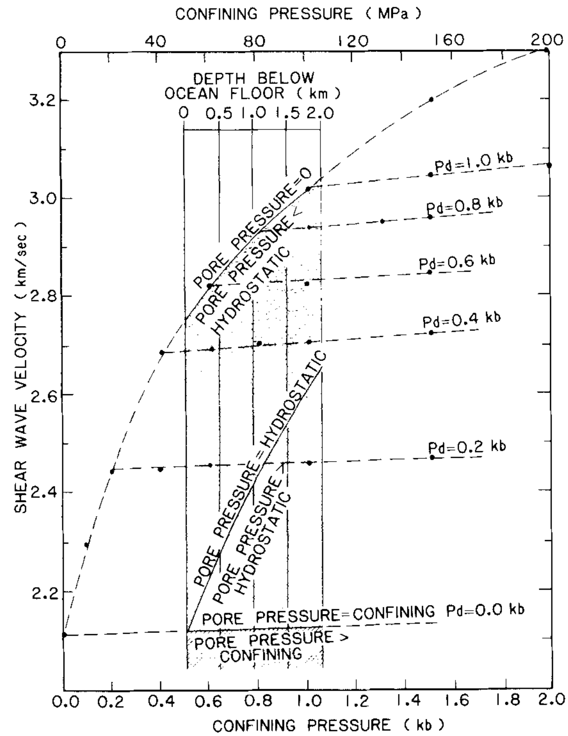

Figure 4. Basalt shear-wave velocity as a function of confining pressure. Lines are shown for constant differential pressure (P_d).

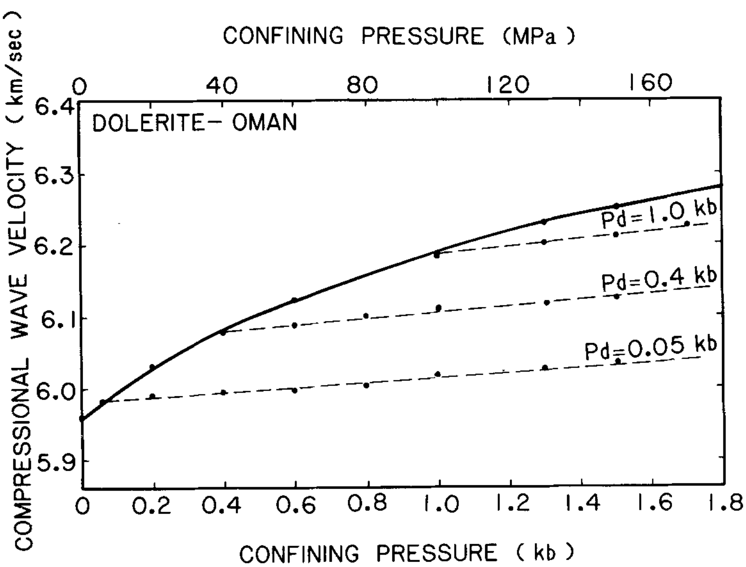

Figure 5. Dolerite compressional-wave velocity as functions of confining pressure and differential pressure (P_d).

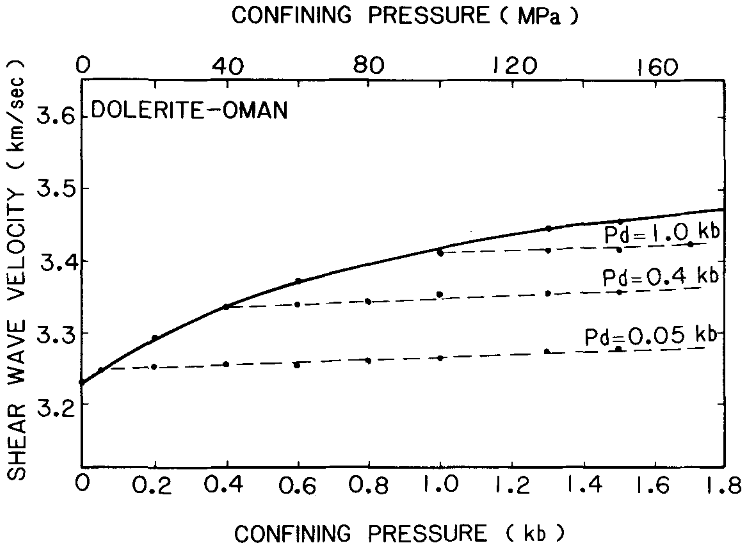

Figure 6. Dolerite shear-wave velocity as functions of confining pressure and differential pressure (P_d).

Early laboratory measurements of compressional-wave velocities in cores of sandstone (Wyllie *et al.* 1958; King 1966) suggested that velocity depends only on differential pressure; that is, $n = 1$. However, Banthia *et al.* (1965) found that for shear velocities in sandstones n is less than unity. Similar results were reported for compressional-wave velocities in low porosity samples of Chelmsford granite and Trigg limestone (Todd & Simmons 1972). More recently, Domenico (1977) has shown experimentally that velocities in unconsolidated Ottawa sand also depend on effective pressure. Of significance, n was found to be less than unity and greater for shear than compressional-wave velocities.

If we define differential pressure (P_d) as $P_c - P_p$ and effective pressure (P_e) as $P_e = P_c - nP_p$, values of n from our velocity studies can be calculated from the following relationship at constant velocity:

$$\Delta V = \left(\frac{\partial V}{\partial P_d}\right)_{P_p} \Delta P_d + \left(\frac{\partial V}{\partial P_p}\right)_{P_d} \Delta P_p = 0.$$

Since

$$\frac{\Delta P_d}{\Delta P_p} = -\frac{(\partial V/\partial P_p)_{P_d}}{(\partial V/\partial P_d)_{P_p}}$$

and for

$$\Delta P_e = 0$$

$$\Delta P_c - n\Delta P_p = 0,$$

it follows that

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

Values of n calculated from the above relationship are given in Tables 3 and 4. The results show that both compressional- and shear-wave velocities depend on effective pressure rather than differential pressure and n is greater for shear than compressional-wave velocities. At constant differential pressure n remains fairly constant over a wide range of pore pressure, whereas at constant pore pressure n decreases with increasing differential pressure.

4 Seismic velocities in the upper oceanic crust

The application of the data presented here to the interpretation of oceanic crustal velocities is significant. It is clear that if oceanic rocks have pore pressures approximating the hydrostatic pressure produced by the overlying column of seawater, their velocities and related elastic properties will be much different from laboratory measured properties in which pore pressures are zero.

Near ridge crests and other oceanic regions devoid or nearly devoid of sediment cover where cracks extend to the surface, the pore pressure at a given depth is likely to equate to the weight of a column of water extending to sea-level. In some regions of the upper oceanic crust, the pore pressure may be less than hydrostatic, as has recently been reported in the Costa Rica rift region by *Glomar Challenger* drilling (Anderson & Zoback 1982). In many regions it is also probable that pore pressure is in excess of hydrostatic and thus velocities are likely to be significantly depressed. It has been well established that overpressuring in sedimentary sections can have many origins, some of which may be applicable to basaltic regions of the oceanic crust. For example, tectonic processes (Fertl 1976; Gretener 1976), the smectite to illite transformation, which releases water during diagenesis (Burst 1969),

Table 3. Values of n for compressional (V_p) and shear-(V_s) wave velocities as a function of pore pressure (P_p) and differential pressure (P_d) for Juan de Fuca ridge basalt.

P_p(kb)	P_d = 0–0.2 kb		P_d = 0.4–0.6 kb		P_d = 08–1.0 kb	
	V_p	V_s	V_p	V_s	V_p	V_s
0	0.90	0.99	0.80	0.96	0.79	0.91
0.4	0.90	0.99	0.77	0.96	0.79	0.92
0.8	0.89	0.99	0.77	0.95		
1.2	0.89	0.99	0.75	0.95		

Table 4. Values of n for compressional-(V_p) and shear-(V_s) wave velocities as a function of pore pressure (P_p) and differential pressure (P_d) for Oman dolerite.

P_p(kb)	P_d = 0–0.2 kb		P_d = 0.4–0.6 kb		P_d = 08–1.0 kb	
	V_p	V_s	V_p	V_s	V_p	V_s
0	0.90	0.93	0.77	0.87	0.64	0.76
0.4	0.91	0.93	0.78	0.86	0.64	0.76
0.8	0.92	0.93	0.83	0.85		
1.2	0.92	0.93	0.85	0.85		

and rapid accumulation of low permeability shale, which hinders dewatering of underlying sediments (Gretener 1976), can all produce excess pore pressures. In tectonically active regions, such as subduction zones, high pore pressure may be developed in water saturated basalt by active folding and faulting. Likewise, the release of water accompanying relatively low grade metamorphic reactions in basalt may result in excess pore pressure. Regions in which impermeable sediments are interlayered with basalt may constitute seals, which would facilitate the development of high pore pressure.

The changes in compressional- and shear-wave velocities with increasing pore pressure are substantially greater for the basalt sample than the dolerite. This is presumably related to the higher porosity of the basalt and complicated by the presence of vesicles in addition to microcracks in the basalt. The importance of pore shape on elastic moduli and seismic velocities has been well demonstrated (Walsh 1965; O'Connell & Budiansky 1974; Toksoz, Cheng & Timur 1976). The following discussion will be primarily concerned with the effects of pore pressure on velocities in the upper basaltic regions of the oceanic crust and the data from the Juan de Fuca ridge sample will be used for illustrative purposes. It is assumed that the cracks and pores in the sample formed during cooling and were open *in situ*, as has been found for Iceland basalt (Kowallis *et al.* 1982).

In Figs 3 and 4, the basalt velocity data are shown as functions of confining pressure and differential pressure. In the upper portions of the figures, confining pressures have been equated to depth below the sea floor assuming a model consisting of a 5 km seawater column ($\rho = 1.03\,\mathrm{g\,cm^{-3}}$) overlying fresh basalt ($\rho = 2.85\,\mathrm{g\,cm^{-3}}$). The variations of velocities with depth are illustrated for zero pore pressure, as well as for pore pressure equal to hydrostatic in which the pore pressure is assumed to be a function of the weight of a column of free seawater extending to sea-level. The regions between these curves give compressional- and shear-wave velocities for oceanic crustal underpressure (i.e. pore pressure < hydrostatic pressure). Likewise, velocities below the pore pressure equal to hydrostatic curves represent crustal conditions of overpressure (pore pressure > hydrostatic). Variations of velocity with depth for pore pressures equivalent to confining pressures (differential pressure = 0) are also shown in these figures. This latter line for shear velocity at zero differential pressure was obtained using the zero confining pressure data and assuming a slope parallel to the 0.2 kb differential pressure line. Finally, Figs 3 and 4 predict extremely low velocities in oceanic crustal regions in which pore pressures exceed confining pressures.

Since increasing pore pressure at constant confining pressure has a greater effect on shear velocities than compressional velocities, Poisson's ratio (σ) calculated from the relation

$$\sigma = \frac{1}{2}\,\frac{(V_\mathrm{p}/V_\mathrm{s})^2 - 2}{(V_\mathrm{p}/V_\mathrm{s})^2 - 1}$$

increases for the basalt sample with increasing pore pressure. This is illustrated in Fig. 7. Examination of this figure also shows that if pore pressure is equivalent to hydrostatic in the upper 2 km of the oceanic crust, Poisson's ratio will decrease with increasing depth. Also, Poisson's ratio depends on an effective pressure, $P_\mathrm{e} = P_\mathrm{c} - nP_\mathrm{p}$, where $n > 1$.

It should be emphasized that the behaviour of velocities with confining and pore pressure obtained in this study is expected to vary for different basalts. Critical to this will be the percentage of connected pore space in a given sample and the pore geometry. Also, the measurements do not take into account the presence of large fractures and rubble zones which have been shown to be abundant within the upper basaltic regions of the oceanic crust. For a highly fractured basaltic region, the effects of pore pressure on seismic velocities may well be of greater importance than observed in this study.

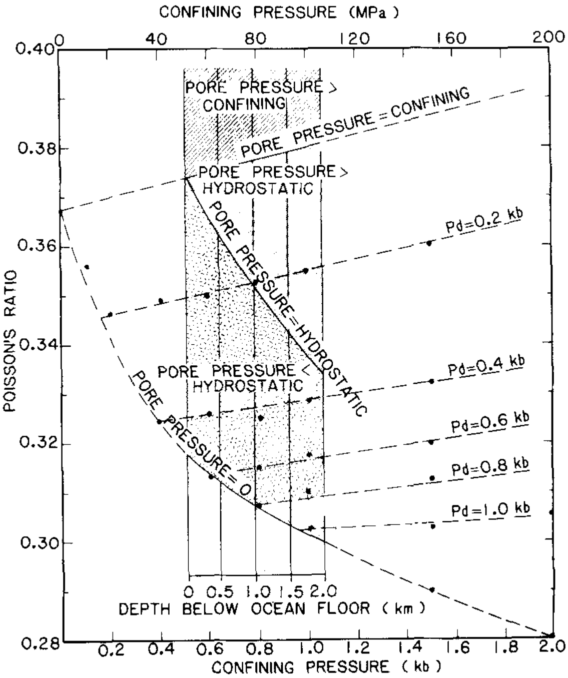

Figure 7. Poisson's ratio calculated from V_p and V_s for oceanic basalt as a function of depth within the crust and pore pressure.

5 Conclusions

Pore pressure is concluded to play an important role in influencing the velocities and elastic moduli of the oceanic crust. At constant confining pressure, compressional- and shear-wave velocities are significantly lowered by increasing pore pressure. Regions of overpressure and underpressure will possess anomalous velocities, which are dependent on depth, porosity, pore geometry and the deviation of the pore pressure from hydrostatic. Since porosity is likely to be extremely low in the lower oceanic crust, pore pressure is expected to have a greater effect on velocities in the upper few kilometres of crust.

Since pore pressure has little effect on bulk rock density, but can change velocities significantly, it becomes apparent that velocity–density relationships obtained from rocks under conditions of low pore pressure must be used with caution when applied to upper oceanic crustal velocities. For a given velocity, density is likely to be underestimated and porosity overestimated.

For oceanic basalt and dolerite, velocities are found to depend upon an effective pressure rather than simple differential pressure. When both confining pressure and pore pressure are varied, velocity increases with increasing confining pressure at constant differential pressure. Thus, the relationship $P_e = P_c - nP_p$ holds and $n < 1$. The value of n is less for compressional-wave velocities than shear-wave velocities and is not constant for a given rock.

The experimental results also show that Poisson's ratio is dependent on pore pressure as well as confining pressure. At constant confining pressure, Poisson's ratio increases with increasing pore pressure. This observation may prove significant in detecting overpressured and underpressured regions within the upper oceanic crust.

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