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Predicted and Measured Seismic Velocities and Densities in Rock Samples from Chalk River, Ontario

by

C. Wright †

and

D.C. Kamineni *

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† Division of Seismology and Geothermal Studies
Earth Physics Branch
Department of Energy, Mines and Resources
Ottawa, Ontario K1A 0Y3

* Atomic Energy of Canada Limited
Whiteshell Nuclear Research Establishment
Pinawa, Manitoba.

Abstract

The intrinsic velocities of P waves in the rock samples of gneiss and monzonite taken from a borehole at Chalk River, Ontario, were calculated from the modal analyses and estimates of mineral velocities given in the literature. These velocities are systematically higher by about 0.5 km/s than the velocities measured in the laboratory at 2.5 kb by Simmons et al. (1978); they are probably better approximations to the intrinsic values than the laboratory measurements, because of the presence of residual porosity in rock cores at 2.5 kb. The calculated intrinsic velocities and the measured densities are well correlated and show a distinct layered structure that is barely resolvable in the velocities measured at 2.5 kb, and is absent in other velocity data. The degree of correlation between various sets of velocities and densities has been investigated and interpreted in terms of the relative contributions of mineralogical variations, crack parameters and observational errors. In spite of the large scatter of the borehole log velocities, approximate crack densities estimated from them show a weak but statistically significant correlation (0.39 ± 0.07) with the distribution of observed fractures in the drill cores.

Introduction

It is well known that the concentration and geometry of cracks, joints and other pore space are the main factors controlling the elastic wave velocities in near-surface crystalline rocks. Moreover, an understanding of the distribution of these parameters is essential for assessing the suitability of a particular rock body as a potential disposal site. It therefore seems natural and logical that some attempt should be made to use seismological techniques to measure crack parameters and related quantities. The application of mathematical theories of the elastic constants of cracked solids to the problem of quantifying cracks in rock cores from boreholes and in rocks in situ has been discussed in a complementary report by Wright and Langley (1979).

In order to uniquely determine the crack and saturation parameters of a rock, we require measurements of the P and S wave velocities and estimates of the intrinsic or 'crack free' P and S wave velocities. An important objective of this report is to determine whether or not the intrinsic P velocities in rocks can be reliably predicted from the modal analyses of borehole samples and the mineral velocities given in the literature. A similar analysis for S waves is not possible because of the absence of S wave velocity determinations in individual minerals. We have also compared the calculated 'crack free' velocities of samples from Chalk River with those measured at 2.5 kb by Simmons, Batzle and Cooper (1978). The P and S wave velocities in the rock body have been estimated by several methods; laboratory measurements at atmospheric pressure on cores from borehole CR1 have been tabulated by Simmons et al. (1978), a borehole log of the P wave velocities has been derived by

Neave (1978), and approximate P and S wave velocities in the vicinity of CR1 have been measured by Lam and Wright (1979) and Wright, Johnston and Lam (1979). We have examined and compared the uncertainties and limitations of several of these data sets.

We define the effective velocity of a mineral in a rock as the seismic P (or S) velocity estimated by regression analysis from modal analyses and rock velocity measurements on water-saturated core samples at room temperature and a pressure of 1 bar. Thus, effective mineral velocities reflect the concentration of cracks, grain boundaries, and the degree of alteration of a particular mineral averaged over a set of cores. The only estimates of effective velocities that we have found in the literature were for granulite facies rocks from Scotland (Hall and Al-Haddad, 1979). We have used these effective velocities and densities to estimate rock velocities and densities for Chalk River samples. The purpose of these calculations is to determine how closely the predicted velocities and densities agree with measured values at different depths in the borehole.

Finally, the degree of correlation between some of the sets of velocities and densities, the relationship of these data sets to petrological variations within the rock body and their utility in determining crack parameters have been considered. The synthesis of all these data sets has enabled us to recommend a procedure for making better use of both laboratory and field data in determining crack parameters, and to show how seismological field experiments and laboratory studies of seismic velocities complement one another.

Data Preparation

(a) Determination of Intrinsic Velocities

The intrinsic P wave velocities within a rock body are those velocities that would be measured if the rock were completely free of cracks and other pore space; they may be estimated from core samples in two ways: (i) from modal analyses and (ii) from laboratory measurements of velocities as a function of confining pressure. A literature search revealed no examples of the use of the first method. However, we have tried this first method because numerous modal analyses of the cores from Chalk River were available, whilst the number of laboratory determinations of velocities provided by Simmons et al. (1978) is rather limited.

(i) Seismic Velocities from Modal Analyses of Core Samples

The intrinsic P wave velocity V_p is given by

$$\frac{1}{V_P} = \sum_{i=1}^{N} \left(\frac{\beta_i}{V_i}\right) \qquad 1,$$

where β_i is the modal fraction of the ith constituent mineral, and V_i is the mineral velocity obtained from laboratory measurements or calculated as the average of the Voigt and Reuss values. Laboratory measurements have been utilized as much as possible; generally, laboratory data on aggregates of pure minerals at high pressures are unavailable, and results for rock types consisting largely of single minerals have been used. The mineral velocities are listed in Table 1, together with the type of measurement or calculation and the relevant literature citations.

Rock velocities have been calculated for pressures of 1 and 4 kb, and the average of these two values for each sample has been taken as the intrinsic value for a crack-free rock at atmospheric pressure. The results are

displayed in Fig. 1, and correspond to a pressure of about 2.5 kb, thus facilitating comparison with the results of Simmons et al. (1978) in which the maximum confining pressure on the samples was 2.5 kb.

The velocities obtained by calculation are systematically higher than the measured values by 0.52 $\pm$ 0.035 km/s. To illustrate this, we have plotted in Fig. 2 the differences between the calculated and measured velocities as a function of the calculated velocity. Whenever a modal analysis on a core used in the velocity measurements was not available, the calculated velocity for the nearest adjacent core has been used. The low scatter of the velocities within a layer of the same rock type justifies this procedure.

To explain the systematic differences between the velocities calculated by the two methods, we must check two possibilities: (i) that the laboratory samples still contain considerable pore space at pressures of 2.5 kb, and (ii) that one or more of the velocities used in Table 1 are systematically high. To investigate these alternatives, we have calculated velocities from modal analyses on five gneisses and a metagabbro from different regions, and the results are also plotted in Fig. 2. The data and the velocity measurements on the specimens have been taken from Christensen (1965).

The minerals present in Christensen's samples differ from those listed in Table 1 in the following respects. For four samples, the plagioclase present is andesine rather than oligoclase. Using the data of Birch (1961), the P wave velocities of andesine were calculated to be 6.52 and 6.60 km/s at 1 kb and 4 kb respectively. Three of the samples contained small quantities of muscovite ($<$1.1%); muscovite has slightly higher seismic velocities than biotite (Alexandrov and Ryzhova, 1961), but we have used the value for biotite in our calculations. One sample contained 1.3% of epidote. The epidote

velocity was arbitrarily taken as 7.0 km/s at 1 and 4 kb, since we know of no estimates of seismic velocities for this mineral. The velocities measured in an epidote amphibolite by Christensen (1965) suggest that the values we have used are a lower bound to the true value.

The systematic positive bias for the calculated velocities is slightly lower for Christensen's samples (0.36 $\pm$ 0.051 km/s) than for the Chalk River samples. This can be partly attributed to the use of an excessively high velocity for potash feldspar, since the four samples from Christensen (1965) that show the lowest bias also contain very small proportions of microcline. Christensen's data, however, show that microcline is not the only mineral for which the assumed velocity contributes a positive bias relative to the measured rock velocities.

Residual porosity is the most satisfactory explanation of most of the discrepancy between the calculated and measured velocities, since the difference between the crack porosity and the total porosity in crystalline rocks is well known (Nur and Simmons, 1969). Stiller, Wagner and Vollstadt (1977) have used the following empirical relationship between the intrinsic velocity V_F and the measured velocity V_θ, assuming that the crack volume decreases exponentially as a function of pressure:

$$V_\theta = V_F \left[1 + K \exp(-A\theta)\right]^{-1} \qquad 2,$$

where K is the fracture descriptive factor, θ is the pressure and A is a constant. For crystalline rocks, $K \simeq 0.3$ and $A \simeq 1\ \mathrm{kb}^{-1}$. If we put $\theta \simeq 2.5$ kb, $V_\theta \simeq 0.975\ V_F$, which is about 0.16 km/s less than the intrinsic value. The estimate of $V_F - V_\theta$ is only a rough approximation, and may be out by a factor of 2. Thus, if the velocities of Table 1 are reasonable estimates of the intrinsic values, we might expect the measured values for

rock cores to be 0.1 - 0.4 km/s less than the calculated values. Clearly, the relationship between calculated and measured velocities requires further investigation, preferably using velocities measured at higher pressures. We conclude that the calculated velocities are better estimates of the intrinsic velocities than the measured values at 2.5 kb.

(ii) Layering of Chalk River Rock Body

By visual inspection of the calculated intrinsic velocities, the rock body has been divided into seven major layers at depths down to 271 m, as shown in Fig. 1. These layers agree well with significant changes in mineralogy, as is evident from the display of the modal analyses on the right hand side of the diagram. Two additional thin layers, labelled T1 and T2, have been included where one or two velocities appeared anomalous with respect to the surrounding values. Between 90 and 170m the velocities are close to 6.9 km/s, corresponding to the monzonite. Gneiss is present above and below these depths; the velocities in the gneiss show much more scatter, and are typically about 0.3 km/s lower.

Changes in mineral content with depth give rise to both an apparent layering and a scatter in the calculated seismic velocities, especially within the gneiss. To quantify this scatter, we have assumed that the standard deviation on a single velocity estimate is independent of the layer concerned. This assumption enables the method of Jeffreys (1961, pp. 144-146) to be used to estimate the standard deviation on a single velocity measurement and the standard error on the mean velocity for each layer. The standard deviation is 0.069 km/s, and the standard errors for each layer are listed in Table 2. Thus, in the one instance when only one velocity estimate exists for

a particular layer, an error of 0.069 km/s has been assumed. The statistical treatment has been applied to the maximum measured velocities of Simmons et al. (1978), and the results are also listed in Table 2 and plotted in Fig. 1. Their data yield a standard deviation of 0.090 km/s on a single velocity measurement. The layer velocities of Simmons et al. (1978) follow a similar overall trend to the calculated velocities, but the measured velocities for the gneiss at depths below 210 m are as high as those for the monzonite.

Fig. 1 also shows the P velocities at atmospheric pressure measured by Simmons et al. (1978) on saturated cores. In addition, the velocities from the borehole log obtained by the Geological Survey of Canada have been plotted as a continuous line. The original borehole velocities have been smoothed by averaging five consecutive velocity estimates, and taking the depth to be the median value of the corresponding five consecutive depth values. Even the smoothed velocities show enormous scatter, and do not appear to be well-correlated with either the velocity measurements on saturated samples or with the calculated intrinsic velocitie-

Correlations between Various Velocity and Density Estimates and Their Interpretation.

(a) Qualitative Description of Parameters Used.

The calculated and measured P wave velocities at 2.5 kb depend primarily upon the mineral content of the rock. We have also examined other sets of velocities that depend primarily on the concentration of cracks, and have looked for correlations between them; these velocities, averaged by layer, are listed in Table 2.

The calculated velocities at 2.5 kb (Fig. 1 and Table 2) show a distinct layered structure, with the higher velocity monzonite at depths between 90 and 170 m surrounded by lower velocity gneiss. The upper region of gneiss, however, does yield velocities greater than those calculated for the monzonite at depths between 50 and 70 m. These higher intrinsic velocities both within the upper region of gneiss and the monzonite are due mainly to the relatively high proportions of potash feldspar and garnet in these layers. The less precise velocities determined from laboratory measurements at 2.5 kb do not show as distinct a layering. We believe that this near absence of layering is due to chemical alteration of the potash feldspar that results in a lower intrinsic velocity than the value of 7.0 km/s used in the calculations: an explanation that is supported by the results of Hall and Al-Haddad (1979).

In column 5 of Table 2 we have listed the P wave velocities calculated from the effective mineral velocities at 1 bar determined by Hall and Al-Haddad (1979). Because these effective velocities were determined by regression analysis on a large suite of rock samples, the calculated rock velocities reflect the average degree of cracking and alteration for the rocks of the region of Scotland sampled. The reason for calculating these velocities was to determine whether or not the effective mineral velocities for Scottish rocks accurately predict the observed velocities at Chalk River. The calculated P velocities of column 5 are systematically higher by more than 0.4 km/s than either the dry or saturated P velocities of Simmons et al. (1978). Thus, the samples from Chalk River are either more extensively fractured than the samples used by Hall and Al-Haddad, or large systematic discrepancies exist between the laboratories in which the velocities were measured. The P wave velocities measured at 1 bar on Chalk River samples by

Simmons et al. (1978) show no systematic changes with depth, though the limited data on S waves suggest slightly higher velocities below 90 m. This could be interpreted as a slight decrease in the density of cracks, although observations of fractures in the cores that will be discussed later suggest the opposite.

The P velocities determined from the borehole log at Chalk River are also systematically lower than the calculated velocities at 1 bar, but higher than the laboratory values, except in the uppermost 50 m. This provides some evidence that both higher crack densities and systematic differences between laboratories are responsible for the discrepancies between the calculated velocities and those measured by Simmons et al.. The borehole log velocities also seem to increase steadily in the top 70 m. If this effect is real, it can be interpreted as a progressive increase in the degree of saturation of the rock body, since there is no evidence for a significant decrease in fracture density at depths between 0 and 70 m.

Table 2 also lists three sets of densities for Chalk River samples: (i) measured, (ii) calculated from the modal analyses and the effective mineral densities estimated by regression analysis by Hall and Al-Haddad (1979), and (iii) calculated from the modal analyses and the expected densities of single crystals of the individual minerals. The densities of set (iii) are systematically higher than the other two, and the reasons for this will be discussed in the next section.

In Table 3, numerical information on all of the data sets used in the preparation of Table 2 has been listed. In particular, we have listed standard deviations on a single measurement that have been calculated (a) assuming that the rock body is layered and using the method of Jeffreys (1961, pp. 144-146),

and (b) assuming no layering. Both the range of the observations and the fractional increase in standard deviation when no layering is assumed give an indication of the reality of the layering for each parameter.

The parameters that are consistent with the layering hypothesis are V_1, V_4, V_6, ρ_1, ρ_2 and ρ_3 (Table 3). V_1 and V_6 are expected to show layering because they should depend primarily on mineral content. V_3 appears layered because of a steady increase in the top 70 m, possibly due to an increase in the degree of saturation of the cracks. The increase in the standard deviation of V_4 if no layering is assumed seems to be purely a sampling phenomenon caused by the small quantity of data. ρ_1, ρ_2 and ρ_3 exhibit layering presumably because of the dependence of density upon mineralogy. It is interesting that ρ_4 shows no evidence of layering. Since the effective grain densities obtained by Hall and Al-Haddad (1979) are in almost perfect agreement with the expected densities of single crystals of the major minerals (quartz, feldspars, amphibole and pyroxene), the apparent layering of the rock body is removed by using a density of 2.77 g cm^{-3} for the remaining minerals. The layering with respect to density therefore seems to be associated with changes in the amount of garnet present, for which the expected densities lie in the range 3.7 - 4.2 g cm^{-3}.

(b) Correlations between Parameters

The correlations between various pairs of the parameters listed in Table 3 have been studied, and slopes of the least-squares regression lines using the method of York (1966) have been calculated. The numerical results are listed in Table 4, in which ω_x/ω_y is the inverse ratio of the variance of each y-value to that of each x-value. For the regression of a velocity set upon

another velocity set or a density set upon another density set, it has initially been assumed that $\omega_x/\omega_y = 1$. The calculation was then repeated with ω_x/ω_y estimated from the standard deviations of Table 3. If there was no significant correlation between the parameters, the second calculation of the slope was omitted. A similar procedure was adopted in examining the correlation between velocities and densities; the slope of the initial least-squares line, however, was determined by regression of velocity upon density ($\omega_x/\omega_y = \infty$). Figures 3 - 7 are plots of those sets of data that have a correlation coefficient of at least 0.5; they are numbered in the order in which they appear in Table 4.

(i) Correlations between Various sets of Velocities

The small but significant correlation between V_1 and V_2 results from the fact that minerals with higher intrinsic P wave velocities tend to have higher effective velocities. A really high correlation, however, does not occur because the ratio of effective to intrinsic velocities or 'crack factor' for the major minerals varies between 0.80 and 0.94 (Hall and Al-Haddad, 1979). V_3 and V_1 show no significant correlation showing that the variability of the borehole log velocities is in no way related to the intrinsic rock velocities. The absence of any correlation between V_2 and V_3, even though they have similar averages, shows either that the P wave velocities from the borehole log are too scattered to provide useful information on crack parameters, or the crack factors of the individual minerals at Chalk River are not related to those of Hall and Al-Haddad (1979). The first of these explanations seems most probable, and we suggested earlier that variations in the saturation conditions in the borehole may be part of the cause of the variability of the borehole log velocities.

The slopes of the regression lines for the first two sets of data change dramatically as ω_x/ω_y changes, and are of not much significance, because of the low correlation coefficient and the small ranges of V_2 and V_3 compared with their standard deviations.

(ii) Correlations between Various Sets of Densities.

ρ_1 and ρ_3 are well-correlated, and the slope of the regression line taking ρ_1 as x - coordinate is about 2 for both of the values of ω_x/ω_y (Fig. 3). ρ_3 is also systematically higher on average than ρ_1 by 0.14 g cm^{-3}. These two results alone suggest that the effective densities of the high density minerals are much lower than the values assumed in calculating ρ_3. This interpretation is supported by the regression lines for ρ_4 and ρ_1 and for ρ_3 and ρ_4 (Fig. 4). ρ_4 and ρ_1 have averages that differ by only 0.05 g cm^{-3}, but they are not well correlated; the slightly better correlation between ρ_3 and ρ_4 suggests the explanation outlined below.

The effective mineral densities for the major minerals (quartz, feldspars, amphibole and pyroxene) used to calculate ρ_4 show only minor and statistically insignificant differences from those used in calculating ρ_3. The differences lie in the remaining minerals which are biotite, garnet, chlorite, opaques and calcite. The regression results of Hall and Al-Haddad (1979) necessitate the use of $\rho = 2.77$ g cm^{-3} for all these additional minerals in calculating ρ_4. The samples from Chalk River contain appreciable quantities of biotite, garnet and chlorite that have expected densities in the ranges 2.7 - 3.1, 3.7 - 4.2 and 2.65 - 2.94 g cm^{-3} respectively; the opaques and calcite are present in quantities that are far

too small to have a significant effect on the densities. Thus, the use of the low regression value of 2.77 g cm^{-3} for biotite, garnet and chlorite accounts for the larger slope (~ 5) of the ρ_3 versus ρ_4 line and the low slope (~ 0.2) of the ρ_1 versus ρ_4 line. The high average for ρ_3 compared with ρ_1 and ρ_4 seems to result largely from the use of higher densities for garnet and perhaps biotite than their effective grain densities within rock samples. There are sufficient density measurements with associated modal analyses for Chalk River cores to enable reliable effective density values to be calculated by regression analysis for the major minerals listed earlier, garnet and biotite. It is recommended that such a calculation should be undertaken.

(iii) Velocity - Density Correlations

The calculated velocities at 2.5 kb (V_1) show a high correlation with the measured densities (ρ_1). If we neglect the errors in ρ_1, the slope of the regression line is 2.0 (Fig. 5). When we use the standard deviations of Table 3, however, the slope increases to 4.5. The low value of ω_x/ω_y reflects the fact that the variability of measured density is much greater than the variability of the calculated velocities.

The foregoing results illustrate an important point. Previously published empirical velocity-density relationships appear to have been derived by regression of velocity upon density ($\omega_x/\omega_y = \infty$). It is true that density measurements can be made with higher precision than seismic velocity measurements, but the errors are not negligible. Consider the following example taken from Table 3. The errors in measured velocities at 1 bar and 2.5 kb are about 2% and 5% respectively; the errors in density measurements are about 2%,

giving ω_x/ω_y in the range 1 - 6. The error estimates represent the scatter within a rock unit that is petrologically fairly homogeneous, and therefore account for both the experimental errors in the measurement process and real fluctuations in velocity or density within the rock unit. The real variations in density are much greater than the measurement errors, and we suggest that allowance be made for these variations in deriving velocity-density relationships.

Rock velocities and densities calculated from effective mineral velocities and densities (V_2 and ρ_4) show a significant correlation of 0.50 (Fig. 5). The variability of crack factors between minerals therefore does not upset the general rule that higher density minerals will have higher seismic velocities. Thus a significant correlation between near-surface seismic velocity measurements or laboratory measurements at 1 bar and measured densities is expected, provided the density of cracks and the degree of saturation remain fairly uniform. V_3 and ρ_1 show no significant correlation because of the large errors in V_3. V_1 and ρ_3 also show a significant correlation of 0.68. The low slope of the regression line in comparison with the slope of the $V_1 - \rho_1$ relationship again illustrates that the densities used for the higher density minerals in calculating ρ_3 are systematically high. The P velocities measured in the laboratory at 2.5 kb show a marginally significant correlation with the measured densities. The low correlation coefficient in comparison with the good correlation between V_1 and ρ_1 reflects the large scatter in V_6, which is probably due to the variable crack densities of the samples and incomplete closure of pore space at 2.5 kb, rather than inaccuracies in the velocity measurements.

(c) Calculation of Aspect Ratios from Crack Density Parameters and Estimated Porosities.

Simmons et al. (1978) have estimated the crack porosities of the core samples on which velocity measurements were made. Using equation 4 of Wright and Langley (1979), these porosities and crack density parameters ε have been used to calculate the average aspect ratio. For each sample, we have used the average of the values of ε obtained from the dry and saturated samples, and the results are listed in Table 5. The set of values of ε has been calculated for the 'saturated isolated' case discussed by Wright and Langley (1979). Repeating the calculation for the 'saturated isobaric' case would not change the results significantly.

Since the crack density parameters obtained from the calculated intrinsic velocities are larger than those derived from the laboratory data, it is reasonable to postulate that the differences represent residual porosity due to cracks or voids of relatively large aspect ratios. This hypothesis is supported by the porosity estimates obtained from density measurements on wet and dry cores, which tend to be about an order of magnitude larger than crack porosity values obtained by Simmons et al.. The differences between the two porosity estimates and the two crack density parameter values for each sample have been used to calculate an aspect ratio for the residual porosity, and the results are also listed in Table 5. The residual porosities yield aspect ratios that are generally at least an order of magnitude higher than those estimated from the crack porosities. Thus, the two sets of aspect ratios provide a consistent explanation of why both the crack porosities and overall porosities of Table 5 differ considerably, and why the calculated seismic velocities at 2.5 kb are so much larger than the measured values.

(d) Correlation between the Crack Density Parameters and the Numbers of Fractures within a Given Depth Interval.

Figure 8 shows a histogram of the number of microscopic fractures observed in each interval of 5 m; the crack density parameters for the 'saturated isolated' case were estimated from the calculated intrinsic P velocities, the borehole log velocities and assumed S wave velocities, and are plotted in the lower portion of the diagram. To determine whether or not there is a significant correlation between these two parameters, the correlation coefficient r was calculated, and the data are displayed in Figure 9. $r = 0.39 \pm 0.07$, so that there is some slight correlation between the estimated crack density parameters and the number of observable fractures. The relatively low correlation coefficient may result from the large scatter in velocity estimates from the borehole log. Alternatively, a spurious correlation could occur if the real intrinsic P velocities for quartz monzonite are not significantly higher than the real values for gneiss.

Discussion

The results suggest that the intrinsic P velocities of crystalline rocks are best determined from the modal analyses of core samples and the mineral velocities obtained from the literature. The use of laboratory measurements at pressures of about 2.5 kb is unsatisfactory because of the presence of residual porosity in the samples. However, the value of laboratory data in determining crack-free P velocities could be improved by the use of an empirical formula such as equation 2 (Stiller et al., 1977); the intrinsic S wave velocities can only be estimated by this method.

A study of the correlations between various sets of velocities and densities has demonstrated the need for more accurate P wave borehole logs, and the inadequacy of the numerical and statistical techniques that have previously been used in the interpretation of such data. In particular, empirical velocity - density relationships should be determined using the least-squares line fitting technique of York (1966). Both the results of our estimates of intrinsic velocities and the calculations undertaken using the effective mineral velocities and densities of Hall and Al-Haddad (1979) suggest that linear regression techniques should be used to obtain a set of reference effective mineral velocities and densities against which the velocities and densities for a particular locality can be compared and interpreted in terms of anomalies in fracture content and mineral chemistry.

Recommendations for Future Studies

The work of Hall and Al-Haddad (1979) on effective mineral velocities and densities can be extended in a way that is of value to the 'RADWASTE' program. The essential ingredient is a large set of laboratory measurements of P and perhaps S wave velocities on crystalline rock samples for which modal analyses of the samples are available. The rocks should be of widely varying mineralogy, from well-dispersed localities, and the measurements should be from many different laboratories at several pressures up to at least 2.5 kb. Multiple linear regression techniques could then be used to derive effective velocities for all of the major minerals as a function of pressure. These mineral velocities could be used as a reference against which the velocities for a particular locality and a particular laboratory could be compared. There is an obvious analogy here with seismic travel times and

reference earth models. Actual rock velocity measurements would reflect both anomalies that are peculiar to a specific location and systematic differences between laboratories. Data could be further analysed to derive both site and laboratory anomalies in exactly the same way as one estimates source and station anomalies in seismology. The useful information that would come from such a study would be the site anomalies defined as the differences between the measured velocities corrected for systematic discrepancies between laboratories and the velocities calculated from the modal analyses and the reference mineral velocities. A comparison of these site anomalies for different rock types would determine whether or not the rocks of one locality were more cracked and altered than those of another.

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FIGURE CAPTIONS

Figure 1 P wave velocities for the borehole at Chalk River.

(i) Calculated velocities at 2.5 kb, V_1.

(ii) Maximum measured velocities of Simmons et al (1978), usually at 2.5 kb, V_6.

(iii) Smoothed velocities from Geological Survey of Canada borehole log, V_3.

(iv) Velocities measured by Simmons et al (1978) on saturated specimens, V_5.

A plot of the volume percentage of minerals as a function of depth from Dence and Kamineni (1978) is shown on the right. The symbols represent the following minerals: BI = biotite; CHL = chlorite; CPX = clinopyroxene; GA = garnet; HB = hornblende; KF = potash feldspar; OP = opaques; PL = plagioclase; QZ = quartz.

Figure 2 Differences between calculated and measured P wave velocities at 2.5 kb for data from Chalk River and from Christensen (1965), plotted as a function of the calculated velocity.

Figure 3 Plot of calculated densities, ρ_3, against measured densities, ρ_1.

Figure 4 Plot of calculated densities, ρ_3, against densities calculated from the regression results of Hall and Al-Haddad (1979), ρ_4.

Figure 5 Plot of calculated intrinsic P velocities, V_1, against measured densities, ρ_1.

Figure 6 Plot of velocities and densities, V_2 and ρ_4, calculated from the effective mineral velocities and densities of Hall and Al-Haddad (1979).

Figure 7 Plot of calculated intrinsic P velocities, V_1, against calculated densities, ρ_3.

Figure 8 Number of observed fractures within each five metre interval and the calculated crack density parameter plotted as a function of depth below the surface.

Figure 9 Diagram showing crack density parameters derived from borehole velocity logs plotted against the number of fractures observed in each interval of 5 m.

Table 1: P-Wave Velocities in Minerals Used to Determine Intrinsic Velocities in 63 Rock Samples

Mineral	P-Wave Velocity (Km/s) (a) 1kb	(b) 4kb	Maximum Modal Fraction	No. of Cores Containing the Mineral	Type of Measurement or Estimate	Reference
Quartz	6.05	6.19	0.396	63	Laboratory measurements on quartzite	Christensen (1965)
Potash Feldspar	6.95	7.06	0.589	63	Mean values of laboratory measurements on different orientations of microcline crystals	Simmons (1964) Press (1966)
Oligoclase	6.39	6.59	0.459	63	Calculations from laboratory measurements on anorthite and albite	Birch (1961)
Biotite*	6.04	6.04	0.319	46	Average of calculated values in 001, 100 directions	Alexandrov and Rhyzova (1961b) Christensen (1965)
Hornblende	7.17	7.27	0.168	34	Laboratory measurements on amphibolite	Press (1966)
Clino-* pyroxene	7.27	7.27	0.124	16	Average of calculated values for two pyroxenes	Alexandrov and Rhyzova (1961a), Alexandrov et al. (1964) Christensen (1965)
Garnet	7.81	7.99	0.247	5[illegible]	Laboratory measurements on almandite-pyrope	Press (1966)
Chlorite	6.75	6.92	0.216	39	Laboratory measurements on chlorite schist	Press (1966)
Opaques*	6.99	6.99	0.065	63	Laboratory measurements on magnetite at 4kb	Birch (1961)
Others‡	6.68	6.72	0.028	46	Somewhat arbitrarily, laboratory measurements on marble	Birch (1960)

* For biotite, pyroxenes and the opaques, the same values have been used for both 1 and 4kb. Only biotite and pyroxene, however, will tend to bias slightly the velocity differences between the two pressures, since the opaques always form less than 6.5% of the rock volume.

‡ The velocity used for the other minerals does not matter, since they form less than 3% of the rock volume. These other minerals are carbonates, sericite, zircon, apatite, sphene and epidote.

Table 2 : Complete List of P and S Wave Velocities and Densities Averaged by Layer

Layer No.	Depth Range	No. of Modal Analyses in each Layer	Calculated P-Wave Velocity at 2.5 kb km/s	Calculated P-Wave Velocity fr Regression Analysis km/s	P-Wave Velocity from Borehole Log km/s	Measured Density (wet) (M.I.T.) g/cm^3	Density Calculated from Modal Analysis g/cm^3	Density Calculated from Regression Analysis g/cm^3
1	0-50	7	6.530 ± 0.026	5.919 ± 0.064	5.243 ± 0.119	2.740 ± 0.020 2.777 ± 0.023	2.856 ± 0.030	2.722 ± 0.013
2	50-70	3	6.970 ± 0.040	5.853 ± 0.097	5.447 ± 0.145	2.806 ± 0.028 2.770 ± 0.028	3.023 ± 0.046	2.724 ± 0.020
3	70-90	3	6.713 ± 0.040	5.767 ± 0.097	5.697 ± 0.168	2.739 ± 0.034 2.67 ± 0.040	2.823 ± 0.046	2.711 ± 0.020
4	90-170	24	6.922 ± 0.014	5.956 ± 0.034	5.719 ± 0.055	2.806 ± 0.010 2.794 ± 0.018	2.987 ± 0.016	2.719 ± 0.007
5	170-210	9	6.529 ± 0.023	5.816 ± 0.056	5.665 ± 0.084	2.696 ± 0.016 2.727 ± 0.023	2.813 ± 0.026	2.721 ± 0.012
T1	193-194	1	6.97 ± 0.069	5.750 ± 0.168	5.640 ± 0.291	2.701 ± 0.049	2.780 ± 0.079	2.777 ± 0.035
6	210-247	8	6.664 ± 0.024	[illegible] ± 0.060	6.014 ± 0.103	2.729 ± 0.022 2.710 ± 0.028	2.843 ± 0.028	2.707 ± 0.012
7	247-265	6	6.547 ± 0.028	[illegible] ± 0.069	5.270 ± 0.206	2.707 ± 0.022 2.69 ± 0.040	2.803 ± 0.032	2.689 ± 0.014
T2	265-271	2	6.775 ± 0.049	6.010 ± 0.119		2.718 ± 0.034 2.72 ± 0.040	2.895 ± 0.056	2.711 ± 0.025

Layer No.	Depth Range	No. of Laboratory Measurements in Each Layer	Measured P Wave Velocity in Each Layer km/s (a) Sat. 1 bar	(b) Dry 1 bar	(c) Dry 2.5 kb	Measured S Wave Velocity in Each Layer at 1 bar km/s
1	0-50	3	5.397 ± 0.110	5.247 ± 0.152	6.113 ± 0.055	3.193 ± 0.097
2	50-70	2	5.015 ± 0.135	4.965 ± 0.186	6.285 ± 0.067	3.205 ± 0.119
3	70-90	1	5.63 ± 0.190	5.52 ± 0.263	6.26 ± 0.095	3.21 ± 0.168
4	90-170	5	5.576 ± 0.085	5.276 ± 0.118	6.280 ± 0.048	3.284 ± 0.075
5	170-210	3	5.453 ± 0.110	5.337 ± 0.152	5.973 ± 0.055	3.287 ± 0.097
T1	193-194	-	-	-	-	-
6	210-247	2	5.710 ± 0.135	5.29 ± 0.263	6.310 ± 0.067	3.380 ± 0.119
7	247-265	1	5.36 ± 0.190	5.21 ± 0.263	6.00 ± 0.095	3.44 ± 0.168
T2	265-271	1	5.67 ± 0.190	5.63 ± [illegible]	6.41 ± 0.095	3.62 ± 0.168

Table 3 : Statistical Data on Velocities and Densities

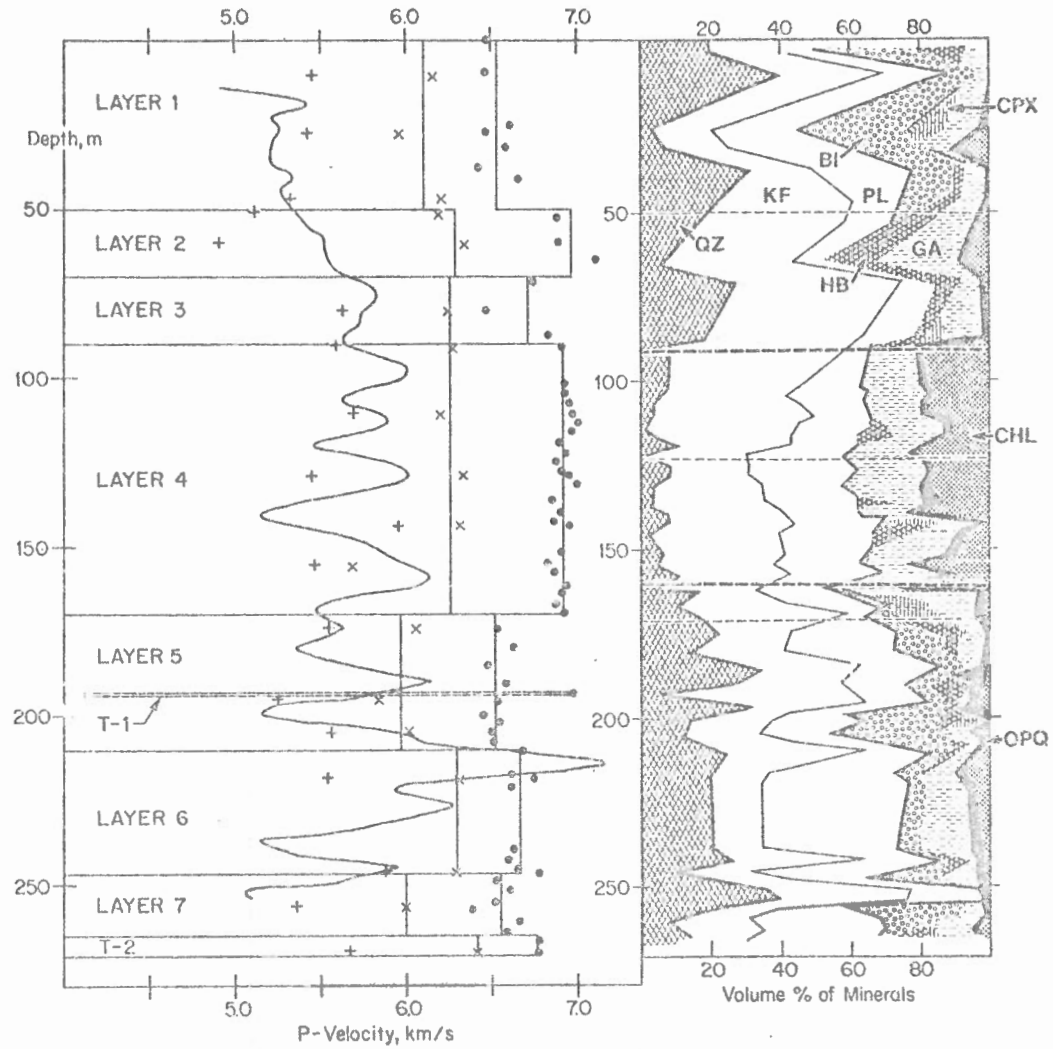
Data Set	Symbol	Units	Standard Deviation Assuming Layered Structure	No. of Data Points	No. of Degrees of Freedom	Range of Observations	Standard Deviation Assuming No Layering
Calculated P Wave Velocities at 2.5 kb	V_1	km/s	0.069	62	54	6.39 - 7.12 = 0.73	0.190
Calculated P Wave Velocities from Regression Analysis	V_2	km/s	0.168	62	54	5.44 - 6.20 = 0.76	0.173
P Wave Velocities from Borehole Log	V_3	km/s	0.291	63	56	5.05 - 6.80 = 1.75	0.345
Laboratory P Wave Velocities (Dry) at 1 bar	V_4	km/s	0.190	15	10	4.91 - 5.88 = 0.97	0.247
Laboratory P Wave Velocities (Saturated) at 1 bar	V_5	km/s	0.263	13	9	4.87 - 5.70 = 0.83	0.252
Laboratory P Wave Velocities (Dry) at 2.5 kb	V_6	km/s	0.095	14	9	5.84 - 6.41 = 0.57	0.160
Laboratory S Wave Velocities at 1 bar	V_7	km/s	0.168	15	10	2.97 - 3.62 = 0.65	0.168
Measured Density (Wet)	$\rho 1$	$g\ cm^{-3}$	0.049	54	46	2.620 - 2.865 = 0.245	0.065
Measured Density (MIT)	$\rho 2$	$g\ cm^{-3}$	0.040	15	10	2.66 - 2.86 = 0.20	0.052
Density Calculated from Model Analysis	$\rho 3$	$g\ cm^{-3}$	0.079	62	54	2.65 - 3.17 = 0.52	0.111
Density Calculated Using Data from Regression Analysis	$\rho 4$	$g\ cm^{-3}$	0.035	62	54	2.648 - 2.806 = 0.158	0.035

Table 4 : Correlation between Various Sets of Velocity and Density Data

Variable X	Variable Y	No. of Points	Correlation Coefficient	Slope of Regression Line	$\frac{\omega_x}{\omega_y}$	Data Set Averages $\bar{X}$	Data Set Averages $\bar{Y}$	Figure
V_2	V_1	63	0.24 ± 0.12	1.5 4.0	1.0 0.17	5.89	6.74	
V_1	V_3	55	0.13 ± 0.13	11.0 0.28	1.0 25.4	6.77	5.67	
V_2	V_3	55	0.00 ± 0.14	- -	1.0 -	5.89	5.67	
ρ_1	ρ_3	55	0.77 ± 0.09	1.92 2.05	1.00 0.43	2.76	2.90	3
ρ_1	ρ_4	55	0.31 ± 0.13	0.19 0.24	1.0 0.53	2.76	2.71	
ρ_4	ρ_3	63	0.50 ± 0.11	5.8 4.4	1.0 4.5	2.72	2.90	4
ρ_1	V_1	55	0.65 ± 0.10	2.0 4.5	∞ 0.33			5
ρ_4	V_2	63	0.50 ± 0.11	2.5 8.4	∞ 4.8			6
ρ_1	V_3	47	-0.13 ± 0.14	- -	7.0 -			
ρ_3	V_1	63	0.68 ± 0.09	1.17 2.47	∞ 0.14			7
ρ_2	V_6	17	0.27 ± 0.23	0.8 10.1	∞ 1.1			

Table 5: Estimates of Aspect Ratios for Crack Porosity and Residual Porosity

Layer No.	Depth Range, m	Ave. Crack Porosity from Simmons et al.	Ave. ε for Measured Velocities	Aspect Ratio for Cracks	Ave. Porosity from Density Measurements	Ave. ε for Intrinsic Velocities	Residual Porosity	Residual Crack Density Parameter	Residual Aspect Ratio
1	0-50	0.00030	0.135	5.2×10^{-4}	0.0038	0.269	0.0035	0.134	6.1×10^{-3}
2	50-70	0.00061	0.233	6.2×10^{-4}	0.0026	0.266	0.0020	0.033	1.4×10^{-2}
3	70-90	0.00030	0.130	5.3×10^{-4}	0.0023	0.258	0.0020	0.128	3.7×10^{-3}
4	90-170	0.00094	0.120	1.8×10^{-3}	0.0072	0.235	0.0063	0.115	1.3×10^{-2}
5	170-210	0.00070	0.105	1.6×10^{-3}	0.0036	0.219	0.0029	0.114	6.0×10^{-3}
6	210-247	0.00028	0.137	4.8×10^{-4}	0.0025	0.253	0.0022	0.116	4.5×10^{-3}
7	247-265	0.00046	0.140	7.7×10^{-4}	0.0023	0.256	0.0018	0.116	3.7×10^{-3}



• Calculated velocities at 2.5 kb
× Maximum measured velocities of Simmons et al. (1978), usually at 2.5 kb
⟋Smoothed velocities from borehole log
+ Velocities measured on saturated cores at atmospheric pressure

FIG. 1

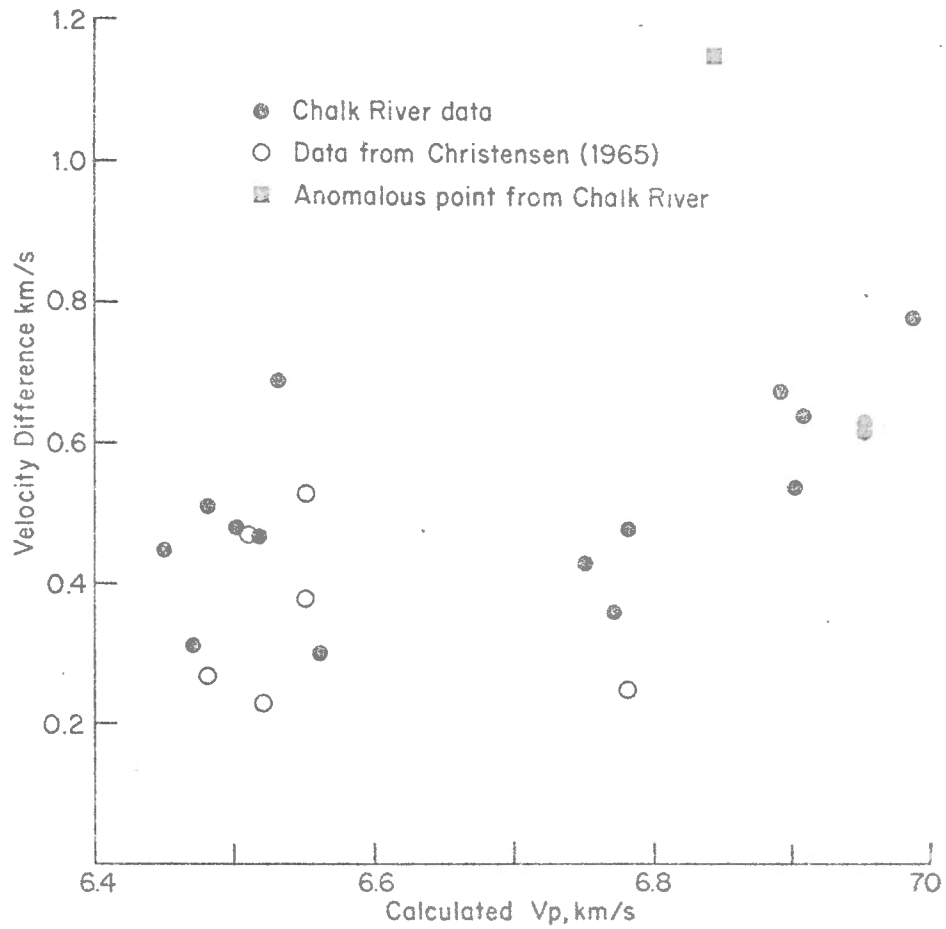

Chalk River data
Data from Christensen (1965)
Anomalous point from Chalk River
Velocity Difference km/s
1.2
1.0
0.8
0.6
0.4
0.2
6.4
6.6
6.8
70
Calculated Vp, km/s

FIG. 2

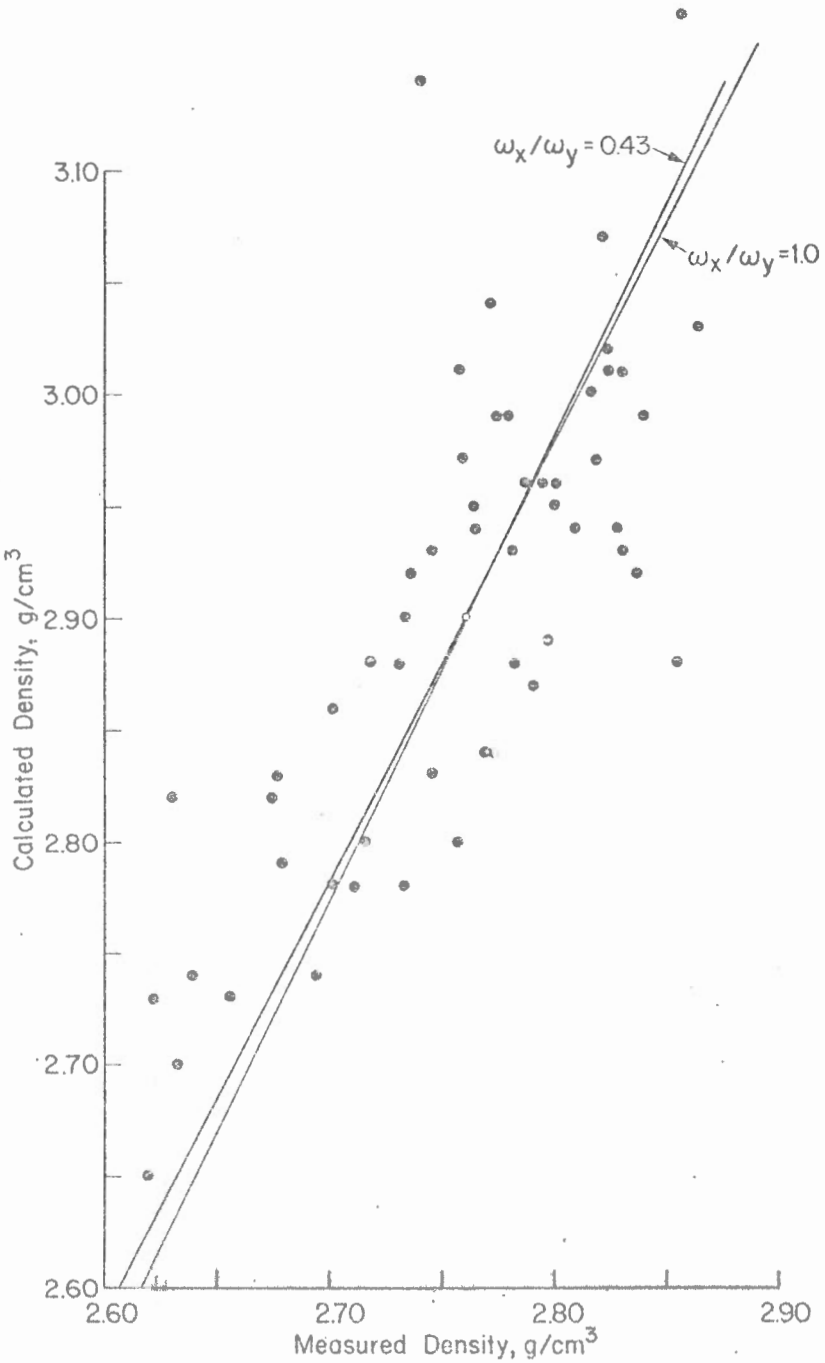

$\omega_x/\omega_y = 0.43$
$\omega_x/\omega_y = 1.0$
3.10
3.00
2.90
2.80
2.70
2.60
Calculated Density, g/cm^3
2.60
2.70
2.80
2.90
Measured Density, g/cm^3

FIG. 3

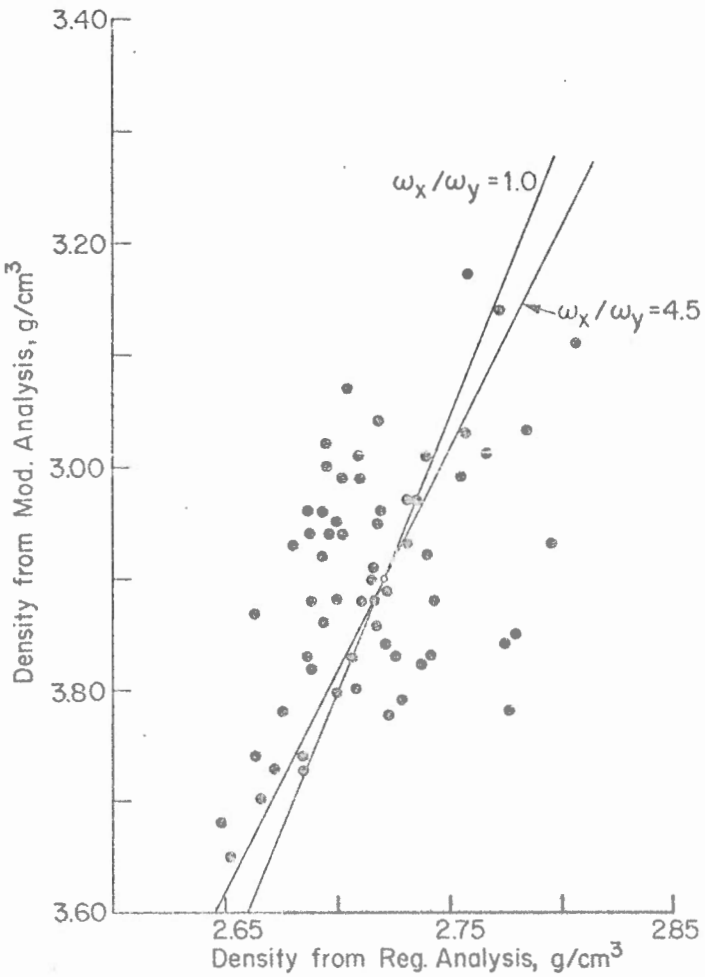

3.40
3.20
3.00
3.80
3.60
Density from Mod. Analysis, g/cm^3
$\omega_x/\omega_y = 1.0$
$\omega_x/\omega_y = 4.5$
2.65
2.75
2.85
Density from Reg. Analysis, g/cm^3

FIG. 4

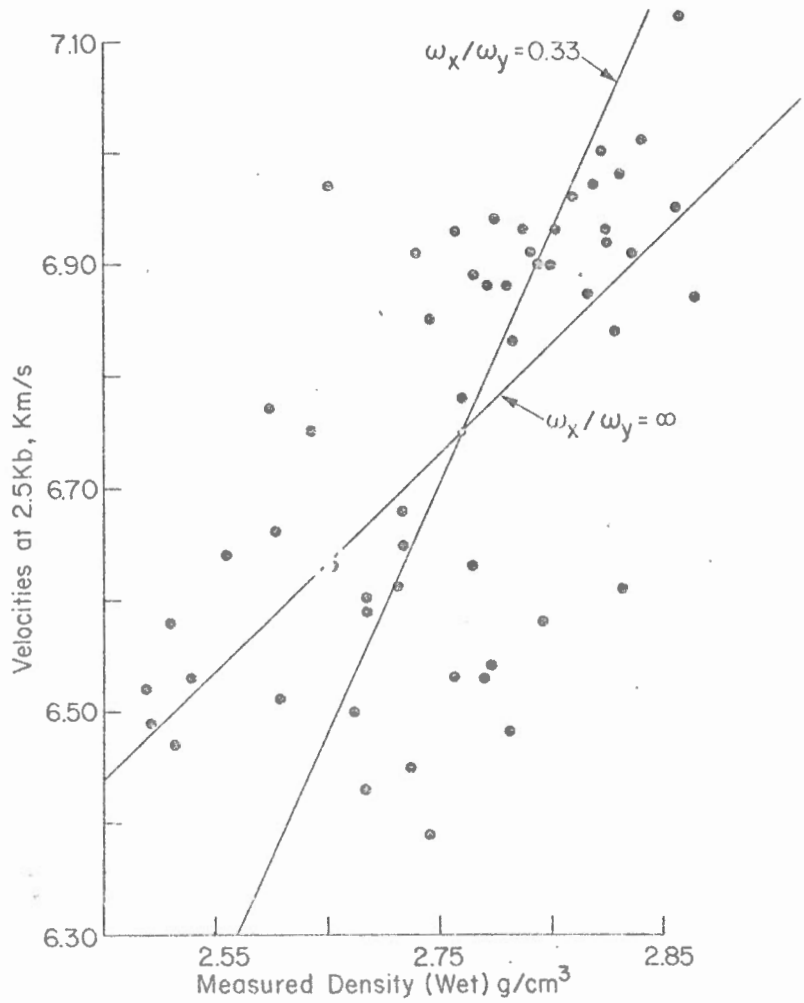

7.10
6.90
6.70
6.50
6.30
Velocities at 2.5Kb, Km/s
2.55
2.75
2.85
Measured Density (Wet) g/cm3
ωx/ωy=0.33
ωx/ωy=∞

Fig 5

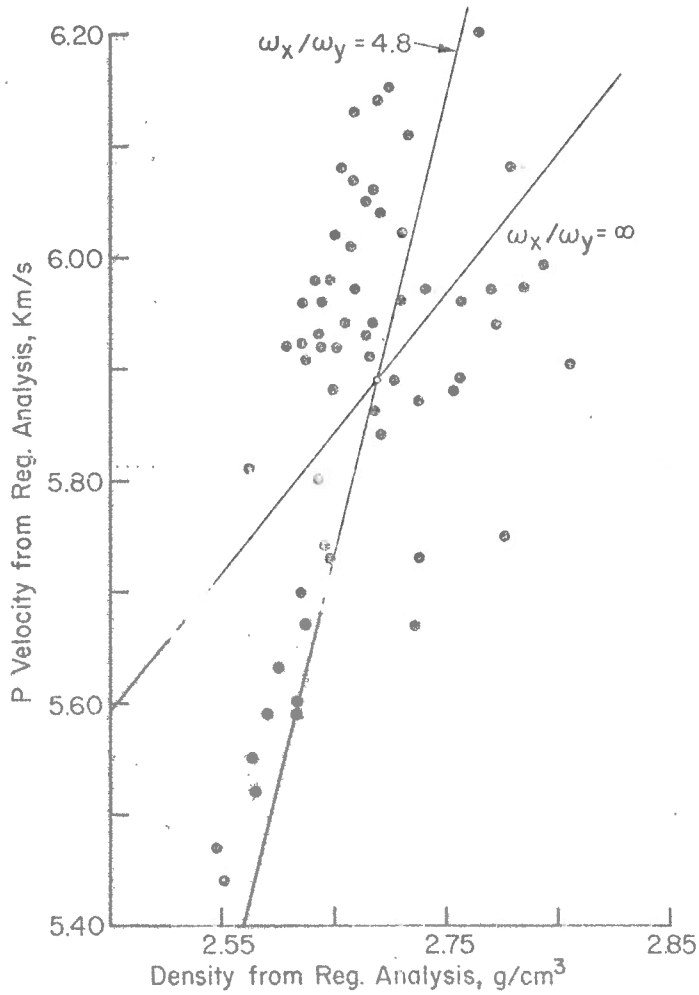

6.20
6.00
5.80
5.60
5.40
P Velocity from Reg. Analysis, Km/s
$\omega_x/\omega_y = 4.8$
$\omega_x/\omega_y = \infty$
2.55
2.75
2.85
Density from Reg. Analysis, g/cm^3

FIG. 6

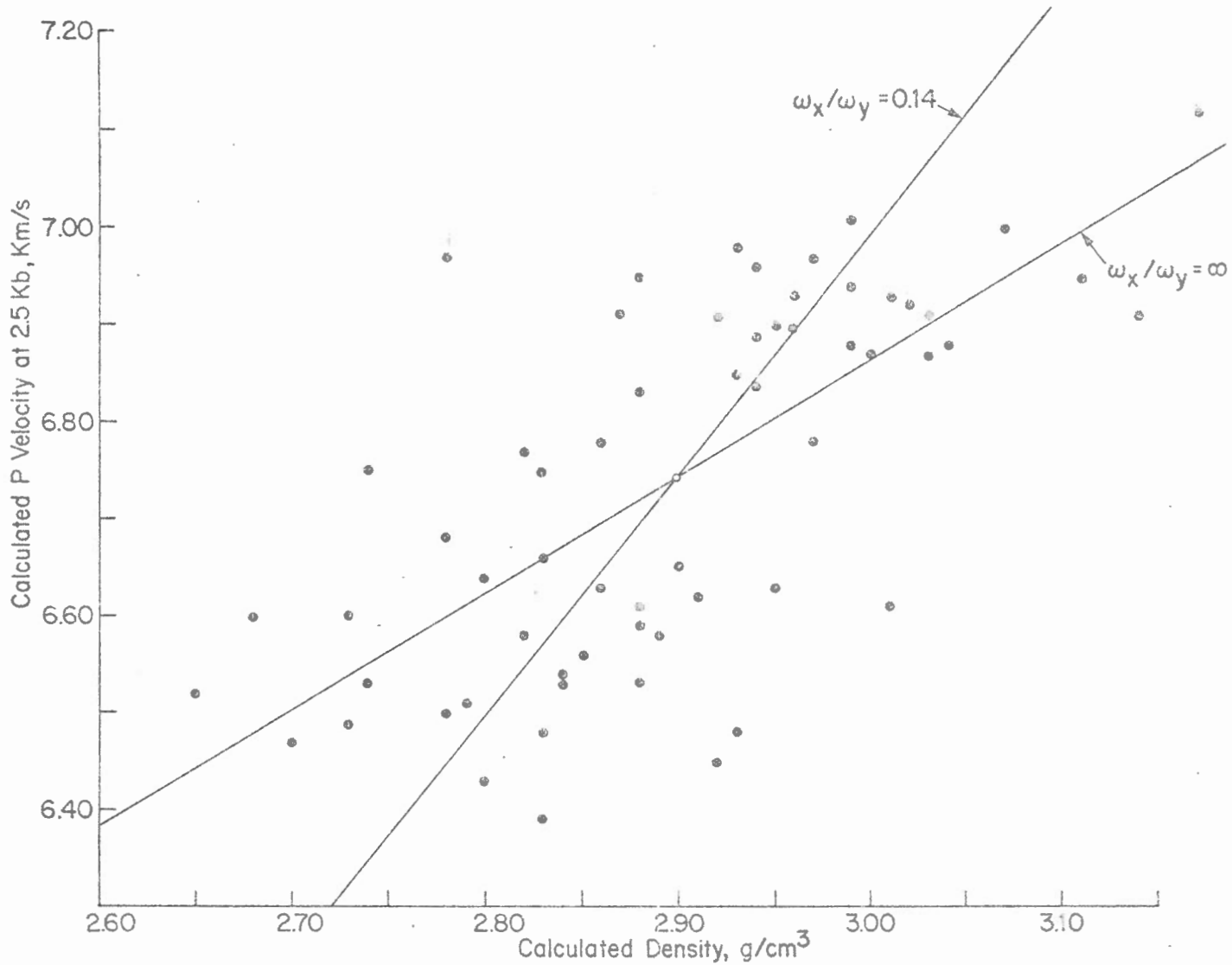

7.20
7.00
6.80
6.60
6.40
Calculated P Velocity at 2.5 Kb, Km/s
2.60
2.70
2.80
2.90
3.00
3.10
Calculated Density, g/cm3
ωx/ωy = 0.14
ωx/ωy = ∞

Fig. 7

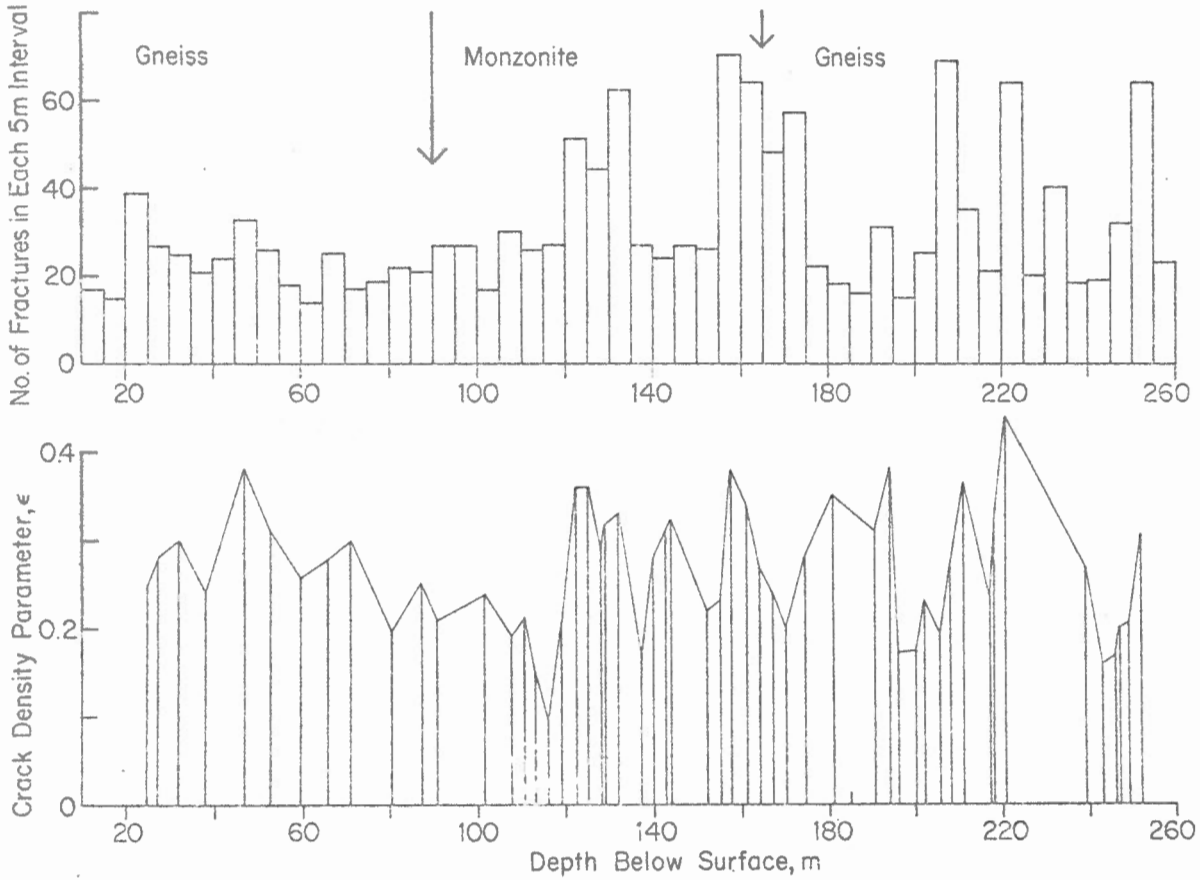

Gneiss
Monzonite
Gneiss
No. of Fractures in Each 5m Interval
0
20
40
60
20
60
100
140
180
220
260
Crack Density Parameter, ε
0
0.2
0.4
20
60
100
140
180
220
260
Depth Below Surface, m

FIG. 8

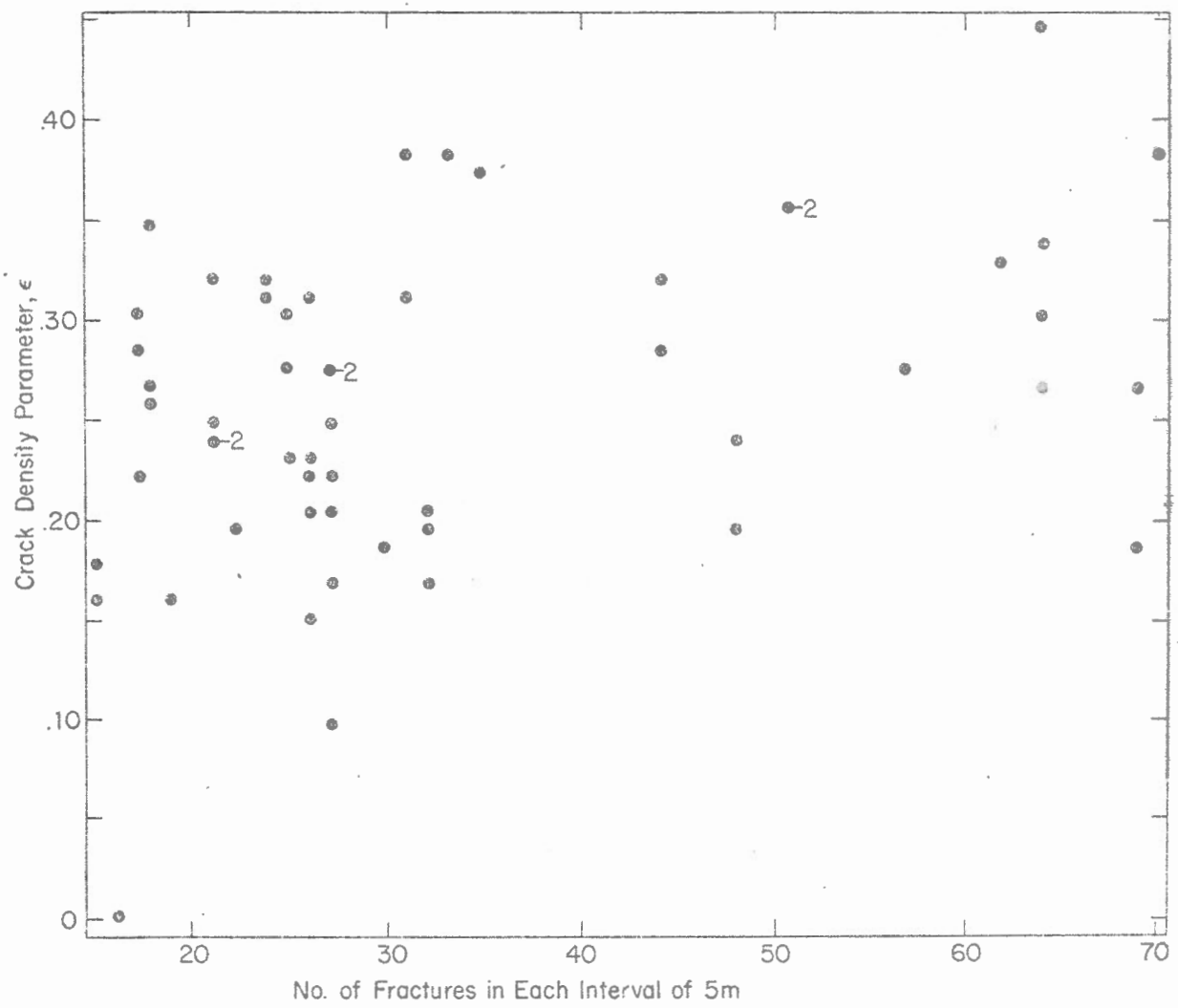

Crack Density Parameter, ε
.40
.30
.20
.10
0
20
30
40
50
60
70
No. of Fractures in Each Interval of 5m
-2
-2
-2

Fig. 9