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# Velocity and Porosity of Sediments from CRP-1 Drillhole, Ross Sea, Antarctica

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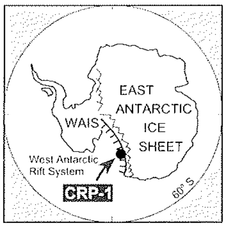


**Abstract** - The relationship between whole-core compressional wave velocities and gamma-ray attenuation porosities of sediments cored at CRP-1 is examined and compared with results from core-plug samples and global models. Both core-plug and whole-core velocities show a strong dependence on porosity; this relationship appears to be independent of lithology. In the range from 0.1 to 0.4 of fractional porosity (Miocene strata), plug velocities are generally 0.2 - 0.5 km s$^{-1}$ higher than whole-core velocities. Possible reasons include decreased rigidity in the whole core and diagenetic changes in the plugs. Possibly both velocity measurements are correct but neither is fully representative for *in situ* conditions. It appears that the core-plug results are more compatible with data from other regions than the whole-core data. After removing first-order compaction control from the whole-core porosity record, a second-order control by clay content can be quantified as a simple positive linear regression (R=0.6). In contrast, after correction for first-order control, porosity and velocity are not significantly influenced by lonestone abundance except for rare, very large lonestones.

## INTRODUCTION

The Cape Roberts Project (CRP) is investigating the Cenozoic and Cretaceous climate history of the Antarctic by coring scientific drillholes offshore Cape Roberts, Ross Sea. The first drillhole (CRP-1) penetrated 148 metres of Quaternary and Miocene sediments. P-wave velocities of these sediment cores provide a bridge between core depths and regional seismic profiles (Cape Roberts Science Team, 1998).

Relationships of P-wave velocity and porosity can be diagnostic tools for the interpretation of a complex imprint on the petrophysics of sediments and sedimentary rocks. For example, porosity-velocity relationships by Wood (1941) and Wyllie et al. (1956) were long used to describe the petrophysical characteristics of high porosity sediments and low porosity sedimentary rocks, respectively. The Wood equation simplified the theoretical Hookean elastic equations (*e.g.*, Gassmann, 1951a, 1951b), by assuming a sediment suspension in which shear modulus and frame bulk modulus are zero. Consequently, the relationship is, at best, appropriate only at very high porosities of >0.7. In fact, a multitude of variables affects the elastic moduli (and, therefore, the compressional-wave velocities) of siliciclastic, sand-shale sediments and sedimentary rocks. In a classic series of papers (*e.g.*, Biot, 1962), Biot developed a general theoretical description for the viscoelastic responses of fluid and mineral framework of a porous medium. Biot theory, though rigorous and powerful for sensitivity studies, requires specification of 13 parameters, a formidable hurdle for most geophysical studies of sound propagation. For the special case in which frequency approaches zero, and therefore internal friction, attenuation, and frequency dependence can be neglected, compressional velocity can be specified largely in terms of dynamic elastic moduli (Gassmann, 1951a, 1951b). However, these moduli are difficult to predict for high-porosity sediments such as those cored at CRP-1.

An alternative approach to sediment velocities is based on empirical, rather than theoretical, relationships. For example, Wyllie et al. (1956) recognised that porosity dominates most variables controlling velocity, and he therefore introduced an empirical relationship between porosity and compressional velocity. His time average equation uses grain and fluid velocities, rather than bulk moduli, to describe the velocity/porosity relationship. This equation most closely approximates the behaviour of lithified sedimentary rocks. Its validity is usually restricted to sediments with porosities <0.3. Later investigators extended this empirical approach to include variations in sand/shale content (Castagna et al., 1985; Han et al., 1986) and higher porosities (Raymer et al., 1980). Erickson & Jarrard (in press) proposed "global" empirical relationships for predicting velocity based on porosity, sand/shale content, and consolidation history.

Physical properties of CRP-1 core, including porosity and P-wave velocity, were determined in two ways: by measuring gamma-ray attenuation and P-wave travel time on the whole core prior to core cutting at the drill site (Cape Roberts Science Team ,1998), and by using core-plug samples (Brink & Jarrard, this volume). Down-core variations in the whole-core data are plotted and their implications are discussed by Niessen et al. (this volume). The aim of this study is to examine the relationship between velocities and porosities of sediments cored at CRP-1, and to compare these velocities and porosities with other parameters measured in the core such as grain size (Ehrmann, this volume; Woolfe et al., this volume) and lonestone volume (Brink et al., this volume).

CRP-1 coring recovered a broad range of lithologies including diamict, sandstone, mudstone, and claystone; all have distinctive physical properties responses. For example, diamicts contain lonestones which significantly decrease porosity and increase velocity. Opposite trends may be visible in the porosity data for units showing significant increase in clay content. Other factors that may have left imprints on the physical properties include diagenesis, fracturing and brecciation by tectonic stress (Wilson & Paulsen, this volume), overcompaction by glacial loading, and rebound due to both deglaciation and erosion of overburden strata. Thus, if the variables affecting velocity and porosity in these sediments can be identified and their influences quantified, velocity and porosity patterns may provide clues to the history of these causal variables. Furthermore, this isolation of relevant variables provides a foundation for interpretation of seismic reflection profiles beyond the site (*e.g.*, Bücker et al., this volume).

## METHODS

During the CRP-1 coring campaign, the drillsite laboratory work included non-destructive, almost continuous determinations of wet bulk density (WBD) and P-wave velocity with 2-cm spacings. A Multi Sensor Core Logger (MSCL, Geotek Ltd., UK) was used to measure core temperature, core diameter, P-wave travel time, gamma-ray attenuation, and magnetic susceptibility. The technical specifications of the MSCL system are tabulated in the CRP-1 Initial Report (Cape Roberts Science Team, 1998). Data acquisition and processing are described in detail by Niessen et al. (this volume) and briefly summarised below.

P-wave velocities were calculated from the core diameter and travel time after subtraction of the travel time through the transducer caps (Cape Roberts Science Team, 1998). The latter was determined empirically by putting together transmitter and receiver transducers, including caps. The arrival time of the P-wave pulse is detected using the second zero-crossing of the received waveform. Resulting P-wave velocities are normalised to 20°C using the core temperature logs. For temperature logging, an infrared sensor was used which was adjusted to detect temperature on the core surface. Displacement (core diameter) and infrared (temperature) sensors were calibrated at the beginning of the CRP-1 campaign.

Wet bulk density (WBD) was determined from attenuation of a gamma-ray beam transmitted from a radioactive source ($^{137}Cs$). The gamma-ray detector was calibrated using aluminum, carbon and nylon of known densities and specific gamma-ray attenuation coefficients. Quantification of wet bulk densities was carried out according to the following formula:

$$WBD = a + b * (1/-\mu * d) * \ln (I/Io)$$

where a and b are instrument-specific variables to correct for count-rate dependent errors as described by Weber et al. (1997), d is core diameter, $\mu$ is specific attenuation coefficient for gamma rays, and ln (I/Io) is natural logarithm of the ratio of attenuated (sample) over non-attenuated (air) gamma counts per second.

Porosity ($\Phi$) is calculated from wet bulk density as follows:

$$\Phi = (dg - WBD) / (dg - dw)$$

where dg is grain density and dw is pore-water density. Grain density is assumed to be 2.70 Mg m$^{-3}$ based on the core plug measurements of Brink & Jarrard (this volume), and pore-water density is assumed to be 1.02 Mg m$^{-3}$.

## RESULTS AND DISCUSSION

### VELOCITY/POROSITY RELATIONSHIP

For all stratigraphic units of the CRP-1 core in which velocity (Vp) and porosity ($\Phi$) were measured (whole core), there is a clear positive correlation between the two parameters (Fig. 1). For the entire data-set including both Quaternary and Miocene units, the best fit was observed for a 3rd-order polynomial function (Fig. 1):

$$Vp = 3.9788 - 12.321*\Phi + 25.065*\Phi^2 - 17.891*\Phi^3$$
$$(R = 0.93)$$

For most velocities above 4.5 km s$^{-1}$ the equivalent porosity is negative. The reason is that porosity was calculated from bulk density (Niessen et al., this volume) assuming a constant grain density of 2.7 Mg m$^{-3}$. Most of the observed bulk densities are lower than 2.7 Mg m$^{-3}$ and

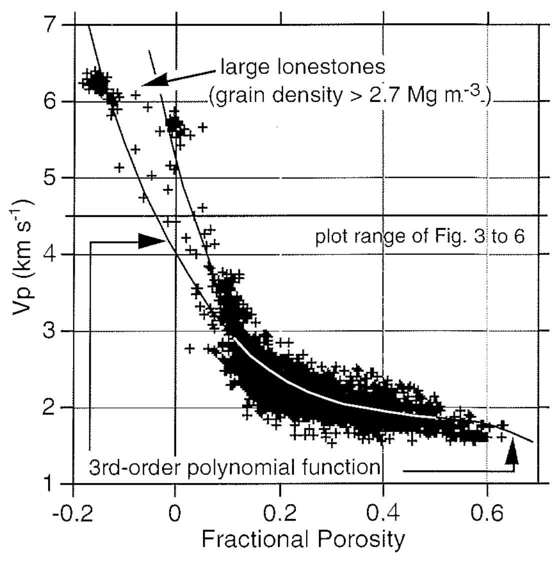


*Fig. 1* - Comparison of P-wave velocity and porosity for all CRP-1 units, along with a 3rd-order polynomial fit.

*Tab. 1* - Core intervals of "negative porosities" of CRP-1, clast description, measured P-wave velocities (Vp), and wet bulk densities (WBD), compared to reference values from Schön (1996).

| depth interval (mbsf) | lithology | measured Vp (km s$^{-1}$) | measured WBD (Mg m$^{-3}$) | reference Vp (km s$^{-1}$) | reference density (Mg m$^{-3}$) |
|---|---|---|---|---|---|
| 15.60-16.30 | lonestone, dolerite | 5.9-6.4 | 2.8-3.0 | 6.3-6.5 | 2.9-3.2 |
| 32.86-32.88 | lonestone, undetermined | 6.0-6.2 | 2.90 | | |
| 63.12-63.14 | diamict with lonestones | 4.4-4.9 | 2.73 | | |
| 80.59-80.99 | lonestone, dolerite | 6.1-6.4 | 2.95-3.00 | 6.3-6.5 | 2.9-3.2 |
| 107.07-107.30 | lonestone, granite | 5.6-5.9 | 2.69-2.72 | 5.1-5.6 | 2.6-2.7 |
| 114.80-114.88 | sandstone, cemented | 4.4-5.6 | 2.70-2.78 | 1.0-5.3 | 1.7-2.8 |
| 117.00-117.08 | lonestone, dolerite | 5.1-5.9 | 2.72-2.92 | 6.3-6.5 | 2.9-3.2 |
| 123.12-123.18 | lonestone, dolerite | 6.1-6.2 | 2.80-2.95 | 6.3-6.5 | 2.9-3.2 |

calculated fractional porosities range between 0.0 and 0.7, as one would expect. In some intervals, however, bulk densities above 2.7 Mg m$^{-3}$ were measured (Niessen et al., this volume) which results in negative porosities. Almost all of these measurements were in depth intervals where the entire core consists of a large lonestone (Tab. 1). Most of these clasts are derived from granites, granitoids, granodiorites and dolerites (Cape Roberts Science Team, 1998). These rock types are normally characterised by relatively high densities from 2.6 to 3.0 Mg m$^{-3}$ and P-wave velocities from well above 5 to more than 6 km s$^{-1}$ (*e.g.*, Schön, 1996).

The effect of large clasts on high P-wave velocity can be tested by plotting Vp *versus* magnetic susceptibility, which was measured in the same depth intervals (Niessen et al., this volume). Dolerites are expected to have the highest velocities of all basement rocks of the hinterland. Also, their magnetic susceptibilities should be strongly increased compared with other clasts or sediment matrix of the CRP-1 core. Our data are consistent with these expectations (Fig. 2). Except for 3 measurements, all other CRP-1 velocities around or above 6 km s$^{-1}$ are associated with magnetic susceptibilities above 400 ($10^{-5}$ SI), and are thus indicative of dolerite clasts. In contrast, velocities between 5 and 6 km s$^{-1}$ are associated with relatively low susceptibilities. This can be explained by granite clasts because granite has velocities in this range but is low in magnetic minerals. The individual relationships of velocity to porosity for different clast lithologies is probably the reason for the relatively large scatter in the velocity-porosity plot above 4.5 km s$^{-1}$.

Whole-core data from the porous majority of the core (porosity from 0 to 0.7 and velocity range from 1.5 to 4.5 km s$^{-1}$, respectively) are compared with results from plug measurements and different models in figures 3 to 6. Despite a relatively large scatter in the whole-core data, a significant difference between plug and whole-core data is evident. This difference, which is most distinct for data measured in Miocene strata (Fig. 4), appears to be largely independent of lithology (Figs. 5 & 6). For a given porosity, core-plug velocities are generally 0.2-0.5 km s$^{-1}$ higher than whole-core velocities (Fig. 4). This pattern is particularly evident for porosities between 0.1 and 0.4. Theoretically, this could be caused by actual grain densities being higher than the assumed constant 2.7 Mg m$^{-3}$ which was used for calculating whole-core porosities. Higher grain densities would particularly increase porosities at

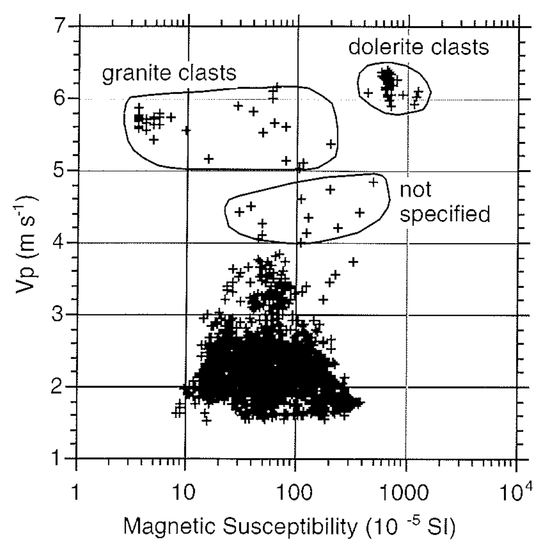


*Fig. 2* - Comparison of P-wave velocity and magnetic susceptibility for all CRP-1 units.

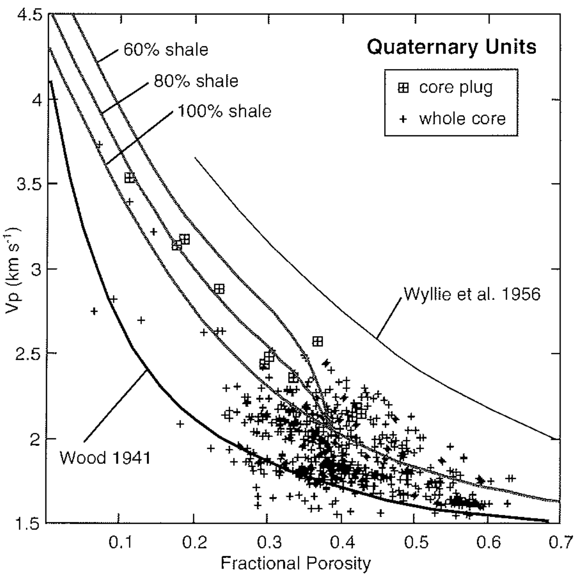


*Fig. 3* - Comparison of P-wave velocity and porosity for all Quaternary CRP-1 units, for both whole-core and core-plug measurements. Model curves for suspensions (Wood, 1941), sandstone (Wyllie et al., 1956), and 60%, 80%, and 100% shale (Erickson & Jarrard, in press) are discussed in text.

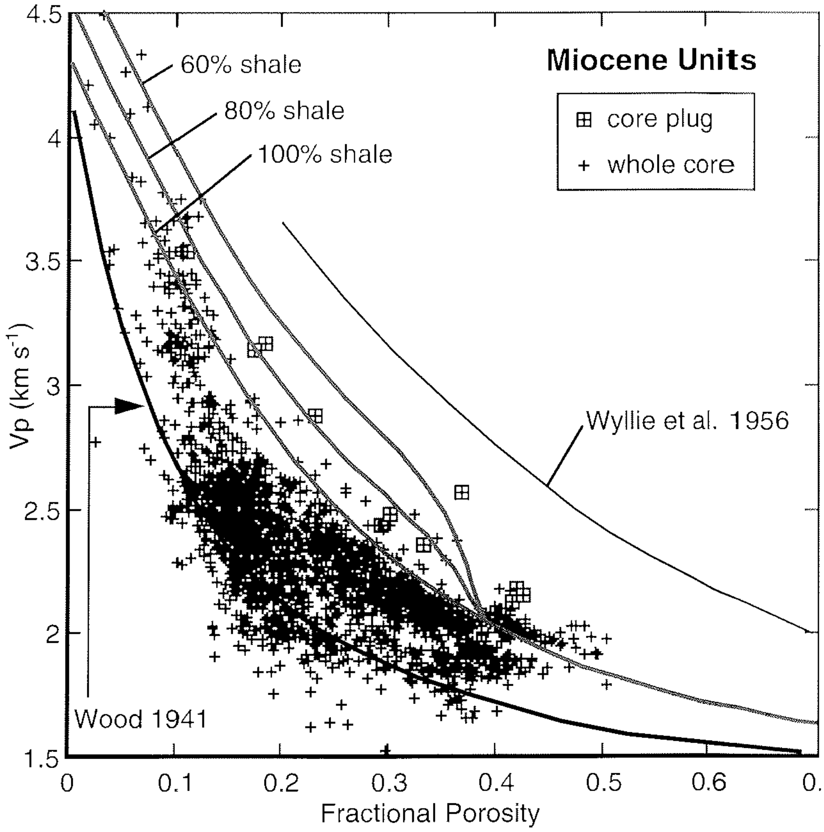


*Fig. 4* - Comparison of P-wave velocity and porosity for all Miocene CRP-1 units, for both whole-core and core-plug measurements. Model curves for suspensions (Wood, 1941), sandstone (Wyllie et al., 1956), and 60%, 80%, and 100% shale (Erickson & Jarrard, in press) are discussed in text.

low porosity levels and thus move the whole-core and plug data closer together. However, grain densities cannot be the reason for the discrepancy, because, except for three samples, plug porosities and whole-core porosities exhibit very good agreement, including multiple measurements in units of very low porosity (Brink & Jarrard, this volume). Also, grain densities around 2.7 Mg m$^{-3}$ are very similar to those observed in the nearby CIROS-1 core (2.67 Mg m$^{-3}$, Bücker et al., this volume). Thus, the difference in the velocity-porosity pattern of whole-core and plug data is attributable to velocity, rather than porosity.

Neither frequency-dependent velocity nor anisotropy can account for the difference between plug and whole-core velocity measurements; measurement frequencies are similar, and both measurements are perpendicular to the core axis. The difference could be caused by undetected bias in either the core-plug or whole-core velocity measurements, but both measurement suites included standards. Very large clasts, which were measured in various levels of the whole core, show very realistic velocities consistent with data from various references (Tab. 1). This suggests correct detection of travel times by the MSCL-system for different depth levels of the core. On the other hand, there may be some errors in the whole-core velocity data because the Geotek system has difficulties in detecting the arrival time of the P-wave pulse correctly when amplitude levels on the receiver side are very low (Cape Roberts Science Team, 1998). This error may be up to + 20% for some travel times if the detection is affected by an offset of one wavelength on the received wavelet. This could result in lower velocities for some of the measurements but can hardly account for the discrepancy seen in the entire porosity range between 0.1 and 0.4 (Figs. 4 to 6).

Alternatively, it is possible that both sets of measurements are accurate, but neither is fully representative of *in situ* conditions. This leaves us with the following suggestions to explain the discrepancy:

1) most of the whole-core velocities are surprisingly low, compared to global models (*e.g.*, Wyllie et al., 1956; Castagna et al., 1985; Erickson & Jarrard, in press). Some velocities are even as low as the Wood (1941) model, implying no rigidity (Figs. 4 to 6) although the relationship of Wyllie et al. (1956) is more realistic at these low porosities. Loss of rigidity could have occurred due to either *in situ* brecciation (Passchier et al., this volume), *in situ* exhumation (Jarrard & Erickson, 1997), core rebound (Hamilton, 1976) or, at least in some sands, core disturbance. It is interesting to note that the discrepancy is largely restricted to Miocene units (Fig. 4), which are generally more strongly fractured than Quaternary units (Wilson & Paulsen, this volume) and which are mostly uncemented (Cape Roberts Science Team, 1998). Some small depth intervals between 114 to 116 metres below sea floor (mbsf) of sandstone Unit 6.2, however, are cemented, which should result in an increase in rigidity and thus higher velocities (Tab. 1). Indeed, these intervals are characterised by velocities of about 3.5 km s$^{-1}$ and porosities slightly above 0.1, which agree quite well with the results from the plug samples (Fig. 5). Exhumation can decrease velocities by as much as 1 km s$^{-1}$ due to microcrack opening (Jarrard & Erickson, 1997). Seismic profiles across CRP-1 demonstrate that some exhumation has occurred (Cape Roberts Science Team, 1998). The CRP-1

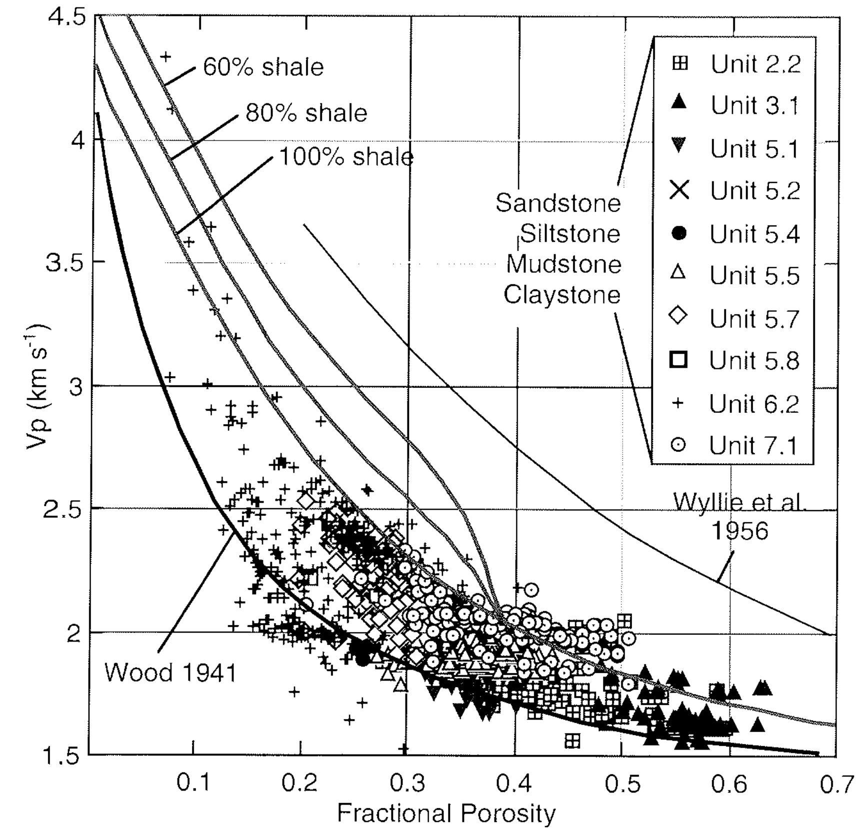


*Fig. 5* - Comparison of P-wave velocity and porosity for the following CRP-1 lithologies: sandstone (Units 5.4, 6.2), sandstone dominant (Units 2.2, 3.1, 5.1, 5.5), siltstone dominant (Units 5.2, 5.7), mudstone (Unit 5.8) and claystone (Unit 7.1).

compaction trends suggest that its magnitude may be between 300 and 650 m (Niessen et al., this volume) which is less than estimated for the adjacent CIROS-1 site (800 to 1 000 m, Bücker et al., this volume);

2) the core-plugs could have undergone diagenetic change associated with sample drying. Drying could possibly cause precipitation of dissolved calcium carbonate, and it certainly causes precipitation of sea salt, and neither would be dissolved by the kerosene saturation used for velocity measurements. Although such precipitates would be volumetrically small, they might increase the frame bulk modulus and shear modulus significantly. If so, then

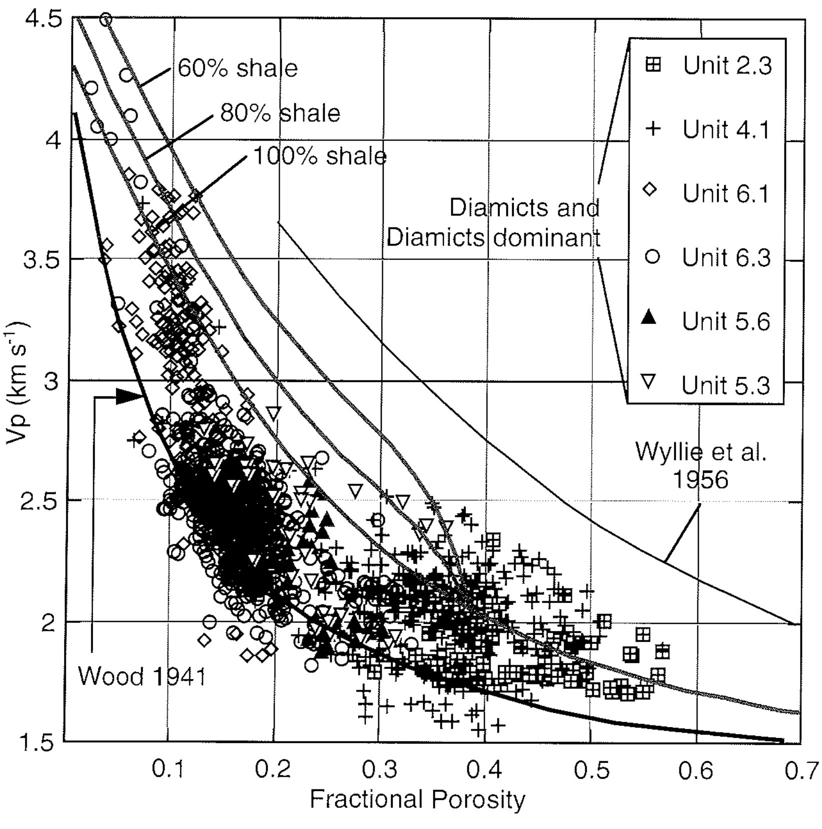


*Fig. 6* - Comparison of P-wave velocity and porosity for the following CRP-1 lithologies: diamicts (Units 2.3, 4.1, 5.3, 6.1, 6.3) and diamicts dominant (Unit 5.6).

the evidence of late-stage diagenetic "annealing", lack of microcracks, and typical Vp/Vs ratios (see Brink & Jarrard, this volume) may not be indicative of *in situ* or initial whole-core conditions.

Compressional velocity of high-porosity sediments is controlled by different variables than those affecting low-porosity sedimentary rocks (Erickson & Jarrard, in press). Velocities of siliciclastic sedimentary rocks decrease rapidly with both increasing porosity and increasing clay content. At fractional porosities higher than about 0.4, however, velocity exhibits a subtle dependence on porosity, and clay content has no direct influence on velocity. Both clay content and sorting do indirectly affect velocity, through their control of porosity. Erickson & Jarrard (in press) found that burial affects velocity by both compaction-related porosity decrease and pressure-induced increase of intergrain coupling. At a critical porosity (Marion et al., 1992) of about 0.38 for highly consolidated sediments and 0.31 for normally consolidated sediments, velocities are expected to start increasing rapidly due to increasing influence of frame bulk modulus and shear modulus on velocity (Erickson & Jarrard, in press). However, critical porosity is very sensitive to early consolidation and diagenesis and is therefore poorly predictable (Vernik, 1997; Erickson & Jarrard, in press).

In figures 3 and 4 whole core data and plug samples are compared with trends empirically determined for highly compacted shaly sediments (Erickson & Jarrard, in press). Core plug results correlate with trends observed for shale contents of 60 to 80%. In this relationship the critical porosity, a kink where velocities increase rapidly, is located at about 0.38. Such a point of rapidly increasing velocity is also observed for whole-core data but at the much lower porosity level of about 0.15 (Fig. 1), far lower than predicted by any of the empirically derived global models discussed above. Therefore, based on their velocity/porosity patterns, the core plug results are more compatible with data from other regions than are the whole-core results. This is particularly true for the units between 60 and 110 mbsf (Units 5.3 to 6.2, Figs. 5 & 6), which experienced relatively strong overcompaction (Niessen et al., this volume): these sediments should have increased velocities below fractional porosities of about 0.3, which is not observed. Moreover, the less compacted Quaternary diamictons seem to be more compatible with highly compacted shaly sediments than are the more strongly compacted Miocene diamictites (Fig. 6, Niessen et al., this volume).

At present, the cause of the discrepancy in velocities remains uncertain. Based on CRP-1 data alone, we cannot distinguish among various possibilities discussed above. We hope that CRP-2 will provide not only whole-core and core-plug data but also downhole logs, and this combination is likely to resolve the present uncertainty concerning velocities and the velocity/porosity relationship for Cape Roberts sediments.

## EFFECT OF CLAY CONTENT

Porosities of most siliciclastic sediments depend on grain size and compaction history. Analyses of very shallow (mostly <10 mbsf) marine sediment core samples show that initial porosity depends strongly on average grain size and sorting: well-sorted sands have porosities of only about 0.4, whereas clays have porosities of up to 0.8 (*e.g.*, Shumway, 1960a, 1960b; Hamilton, 1976). Initial porosities are subsequently reduced by both mechanical compaction and chemical diagenesis. Niessen et al. (this volume) demonstrate that most of the fluctuation of porosity that is superimposed on the compaction-induced downcore trend of porosity seems to correlate with the downcore pattern of variation in clay content (Ehrmann, this volume; Woolfe et al., this volume). The question is whether there is a way of quantitatively describing the effect of clay content on porosity.

Grain-size and porosity data for the CRP-1 core cannot be compared statistically in a direct way because the down-core trend of porosity caused by compaction would overprint any possible correlation. Niessen et al. (this volume) determine empirically that the steep downcore decrease in porosity observed from the top of Unit 2.1 to the bottom of Unit 6.3 can be described by a simple linear regression of porosities from non-diamict units:

$$\Phi = 0.5574 - 0.002914 * Z \qquad (R = 0.7455)$$

where $\Phi$ is fractional porosity and Z is depth in mbsf. We have removed this first-order control from the entire porosity data set, resulting in porosity residuals, here defined as the differences between observed porosities and those predicted from depth. Because grain-size data are relatively rare (about one sample per metre) and sometimes obtained from crumbled or fractured core intervals that are unsuitable for whole-core porosity determinations, we calculated residuals using a smoothed data set of porosities (running mean with a window size of 30, which is equivalent to 0.6 m) (Niessen et al., this volume). Also, negative porosities (Fig. 1) were removed from the dataset prior to smoothing, because it is inappropriate to compare porosity data from large clasts to grain-size analyses of matrix sediments. Porosity residuals are plotted *versus* clay content in figure 7. There is a clear positive correlation between porosity residuals and percent clay (<2 µm), which can be described by the following linear regression:

$$\Phi\ \text{res} = -0.06946 + 0.0097413 * c \qquad (R = 0.6)$$

where $\Phi$ res is the residual of fractional porosity and c is the clay content in % <2µm as published by Ehrmann (this volume). A similar regression analysis for non-smoothed porosity data yields a very similar result (-0.068 + 0.0092 * c, R=0.6).

The scatter in these data can have various causes. Grain-size and whole core measurements represent different volumes. Most plausible, however, is that the linear regression used to remove first-order control on porosity by compaction is probably too simple. Because there are several diamicts in the record which may represent basal till and thus glacial loading (Cape Roberts Science Team, 1998), different units may have undergone different compaction. This local differential compaction cannot be

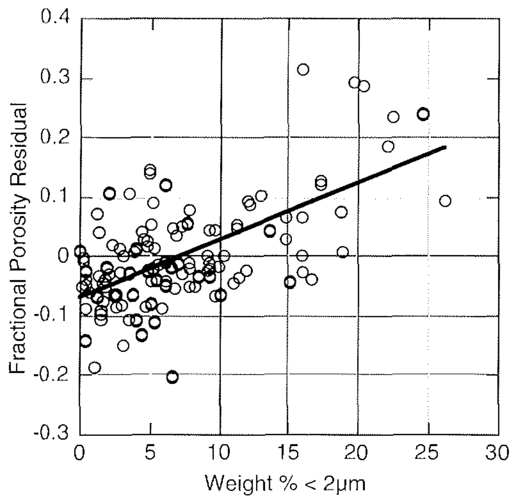


*Fig. 7* - Comparison of percent clay content (Ehrmann, this volume) to porosity residuals.

removed correctly from the porosity trend because we do not know the exact compaction history of the individual units (Niessen et al., this volume). We conclude, however, that grain size, in particular clay content, had a relatively strong influence on the porosity of unconsolidated CRP-1 sediments. This grain-size effect has remained to some extent because the sediments are largely uncemented. We therefore suggest that the above regression can be used to remove grain-size effects on the CRP-1 core in order to further study the effect of overconsolidation as discussed in Niessen et al. (this volume).

## EFFECT OF LONESTONES

The effect of large lonestones on CRP-1 velocities and porosities is obvious: isolated points have real porosity of near zero (or negative whole-core porosities as shown in Fig.1) and velocity of about 6 km $s^{-1}$. But do variations in lonestone abundance cause a detectable effect on either velocity or porosity of diamicts in general? We can test for possible second-order effects of lonestones in a similar way to the effect of grain size by removing the first-order controls on porosity and velocity, then examining residuals. Porosity residuals are here calculated from the non-smoothed 2-cm interval porosities (including large clasts), using the same down-core compaction trend as for grain sizes.

Porosity residuals are plotted *vs* lonestone volume (Brink et al., this volume) for all units in figure 8. There is a weak positive correlation (R = 0.3) of lonestone volume and porosity residual, mostly defined by the very large clasts as previously noted. Almost all samples with more than 25% lonestones have positive residuals. Zooming in on those with less than 10% lonestones (Fig. 8), it appears that porosity residuals are systematically positive all the way down to about 2% of lonestone volume. But does this correlation necessarily imply that lonestone abundance affects porosity even at low lonestone abundances? The causal relations may be that many diamicts have lower porosities than other sediments, and diamicts contain generally more lonestones than other sediments. For example, diamicts often fall below the downcore trend of porosities defined by sandstones, siltstones, and mudstones (Niessen et al., this volume), probably because over-compaction is most common in diamicts, and because diamicts have poorer sorting than sandstones and siltstones. Removing non-diamict units from figure 8 would not change the pattern because most lonestones are from diamicts (Brink et al., this volume). However, the large scatter in the data below the 10% lonestone level in figure 8 and the weak positive correlation suggest that the effect of lonestones on porosity is rather small or negligible. Moreover, other measures of lonestone abundance (*e.g.*, number of lonestones per 10 cm interval, volume of lonestones ≤16 mm) exhibit no correlation with porosity residual.

The main factor accounting for CRP-1 velocities is porosity (Fig. 1). Velocity residuals, here defined as the difference between observed velocity and that predicted from porosity according to the 3rd order polynomial function (Fig. 1), show no correlation with lonestone volume. Other measures of lonestone abundance (*e.g.*,

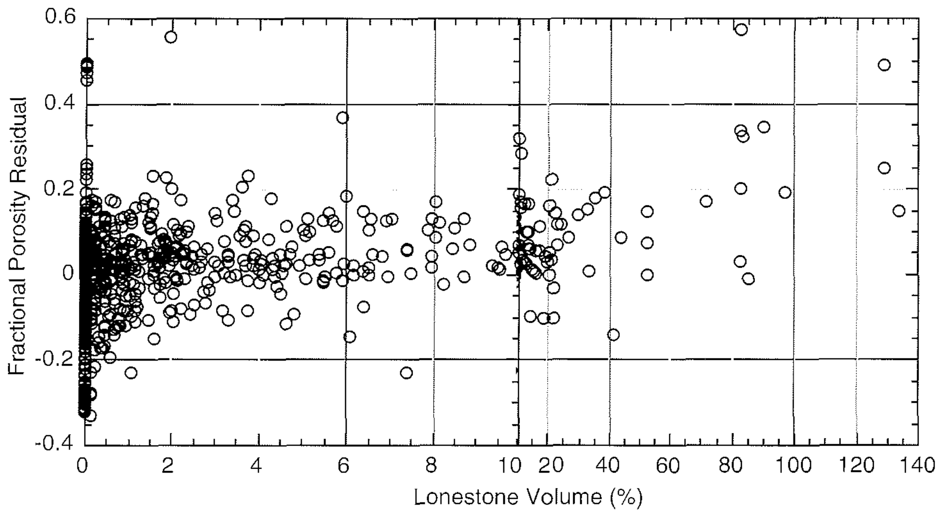


*Fig. 8* - Comparison of lonestone volume (Brink et al., this volume) to porosity residuals. Note the change in lonestone volume scale at 10%.

number of lonestones per 10 cm interval, volume of lonestones ≤16 mm) also exhibit no correlation with porosity residual.

## CONCLUSIONS

Both core-plug and whole-core velocities exhibit a very strong dependence on porosity. We conclude that a single velocity/porosity trend for CRP-1 is capable of accounting for the first-order variations in velocity. This trend can be used to predict velocity based on porosity, thereby taking advantage of the nearly continuous whole-core porosity record to produce a much more continuous velocity record than was possible with the multisensor track.

Furthermore, both core-plug and whole-core data show velocity/porosity relationships that appear to be independent of lithology. However, the two data-sets indicate significantly different velocity/porosity patterns. For a given porosity, core-plug velocities are generally 0.2-0.5 km $s^{-1}$ faster than whole-core velocities. This difference is in the velocity observations rather than in porosities. The cause of this discrepancy is uncertain. Based on CRP-1 data alone, we cannot distinguish among various possibilities such as undetected bias in the velocity measurements, non-representation of *in situ* conditions in the core or plugs, lack of rigidity in the core, and diagenetic change in the plugs. Fortunately, CRP-2 is expected to provide not only whole-core and core-plug data but also downhole logs, and this combination is likely to resolve the present uncertainty concerning velocities and the velocity/porosity relationship for Cape Roberts sediments.

From comparison of porosity and velocity with other data measured in the CRP-1 core, we conclude that variations in lonestone abundance have no direct influence on CRP-1 velocities and porosities, except for rare, very large lonestones. On the other hand, there is a significant dependence of compaction-corrected porosities on clay content. We suggest a relationship between porosity and clay content which can be used to correct the down-core porosity trend for grain-size effects in order to detect overcompaction in some of the CRP-1 units.

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