

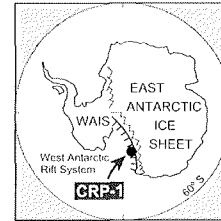
Petrophysics of Core Plugs from CRP-1 Drillhole, Victoria Land Basin, Antarctica

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Abstract - This paper reports measurements of velocity vs pressure and of bulk density, porosity, matrix density, and magnetic susceptibility in 18 core plugs from CRP-1. Comparison of our bulk densities with continuous whole-core density records shows very good agreement. Core-plug measurements of matrix density permit conversion of the whole-core density record to porosity. Agreement between our magnetic susceptibility measurements and the continuous, whole-core data is excellent. In contrast, our atmospheric-pressure measurements of P-wave velocity are ~10% faster than whole-core data obtained at the same pressure. Our measurements of velocity *versus* pressure indicate that *in situ* P-wave velocities are probably only 1-3% higher than those measured at atmospheric pressure. Although the Miocene section has undergone significant exhumation, we do not observe typical exhumation signatures of anomalously low initial velocities followed by microcrack closing as pressure is increased. Instead, velocity response to pressure appears to be dominated by a small amount of post-exhumation cementation.



INTRODUCTION

The Cape Roberts Project (CRP) is an international drilling project whose aim is to reconstruct Neogene to Palaeogene palaeoclimate by obtaining continuous core and well-logs from a site near Cape Roberts, Antarctica. The first CRP drillhole, CRP-1, obtained 148 metres of Quaternary and Miocene sediments. Lithologies sampled for this petrophysical study include diamictites, sandstones, and mudstones (Cape Roberts Science Team, 1998; Woolfe et al., this volume).

Velocity, density, and porosity of sediments drilled by the CRP provide insights into controls on compaction (Niessen et al., this volume) and velocity (Niessen & Jarrard, this volume) at the drillsite, as well as a link between drillhole depth and regional seismic reflection profiles (Cape Roberts Science Team, 1998; Bucker et al., this volume). These three parameters can be determined in three ways: by laboratory measurements on core plugs, by whole-core measurements, or by downhole logging. CRP-1 had no downhole logging. Continuous whole-core measurements of bulk density, velocity, and magnetic susceptibility were made at the rig-site. Whole-core results were published in the CRP-1 Initial Reports volume (Cape Roberts Science Team, 1998), and revised results are presented by Niessen et al. (this volume) and Niessen & Jarrard (this volume).

This study provides a complementary dataset: laboratory measurements of velocity vs pressure and of bulk density, porosity, matrix density, and magnetic susceptibility for core plugs. All major lithologies present in CRP-1 were sampled for this study. Petrophysical measurements were made on a small subset of the available cores (only 18 samples), in contrast to whole-core measurements of nearly all cores. They can, however,

address questions critical to the analysis and interpretation of the whole-core measurements:

- what is the matrix density of the CRP-1 sediments? Matrix density is needed for conversion of whole-core densities to porosities;
- do core-plug measurements confirm the accuracy of continuous whole-core velocity and density data, or is a recalibration of the latter needed?
- are velocity measurements made at atmospheric pressure representative of *in situ* values? Can measurements of velocity vs pressure provide a correction factor for the whole-core velocity measurements, permitting their extrapolation to *in situ* velocities?

METHODS

At McMurdo Station, Antarctica, 26 cylindrical samples were drilled from the working halves of the CRP-1 cores; the circulating "fluid" used to remove cuttings was air, rather than water, to minimise core damage. Sample diameters were 2.5 cm. Volumes of most samples were 10-11 cm³, but three samples were only 3-7 cm³. Quaternary samples were too unconsolidated to drill, so 8 samples were obtained with a cylindrical plastic sampling sleeve. These Quaternary samples fell apart during drying, however, and therefore their properties are not reported here. All samples analysed were Miocene in age and were stored in sealed bags to retain moisture. Shortly after sampling, palaeomagnetic measurements were undertaken on most samples. These measurements included remanence, alternating field demagnetisation, and magnetic susceptibility (Cape Roberts Science Team, 1998). The magnetic susceptibility of the samples was

later remeasured at the University of Utah, using a KLY-2 Kappa bridge rather than the less sensitive Bartington instrument used for the original measurements. During the several months between sampling and our measurements, samples were dried for computerised tomography.

In the University of Utah laboratory, two samples were exposed to seawater in an attempt to resaturate them. The integrity of both samples was degraded by this resaturation: one sample completely disintegrated and the other underwent substantial surface spalling. Most CRP-1 sediments contain some smectite (Ehrmann, this volume), and exposure of smectite-bearing sediments to water often causes clay swelling and associated spalling. Consequently, we evacuated the remaining water from all core-plug samples and used kerosene as the saturating fluid, rather than seawater. Kerosene is often used in the petroleum industry for core-plug drilling and petrophysical measurements, because it does not adversely affect the integrity of shale-rich sediments.

Porosity, bulk density, and matrix density of the core plugs were determined using a simple weight-and-volume technique. Samples were evacuated for about three days to remove pore water, with a final vacuum pressure of 9-11 Pa, then dry weight was measured. Samples were then evacuated again at 9-11 Pa for one day. Samples were flooded with kerosene while still under vacuum. Next, external pressure was changed to atmospheric, permitting the high vacuum within each sample's pores to suck kerosene into its pores. Wet volume and wet weight of each sample were then measured. Accuracy of this technique was confirmed by measuring a suite of standard samples. These standards are Ferron sandstones that had previously been measured by Amoco, using a helium porosimeter and mercury immersion, as described by Sondergeld & Rai (1993).

The density of kerosene (780 kg/m^3) is much lower than that of seawater (1020 kg/m^3). To assure that core-plug bulk densities are representative of *in situ* conditions and to permit direct comparison of these bulk densities to continuous whole-core data, we converted kerosene-saturated bulk densities to water-saturated values.

Velocities of kerosene-saturated samples were measured in a New England Research velocimeter, at confining pressures of 0-17.2 MPa. Pore pressures were atmospheric, so confining pressure was equal to differential pressure. Velocimeter accuracy was confirmed by replication of Amoco results on Ferron sandstone samples, for both compressional velocity (V_p) and shear velocity (V_s) and for both saturated and dry states.

Because the P-wave velocity of seawater (1500 m/s) is higher than that of kerosene (1300 m/s), the fluid bulk modulus of seawater (2.4 GPa) is almost double that of kerosene (1.3 GPa). Therefore, the Gassmann (1951) equation was used to convert measured velocities of kerosene-saturated samples to those of seawater-saturated samples. For this conversion, mineral bulk modulus was assumed to be that for silty clay (50 GPa) (Hamilton, 1971); an alternative assumption of quartz (39 GPa) reduces velocity by only 10-30 m/s. This conversion has been applied to the low-pressure data shown in table 1 and later used in comparisons with whole-core measurements. This conversion indicates that water-saturated velocities are 6-18%, or 150-340 km/s, higher than kerosene-saturated velocities. Patterns within velocity runs, such as percentage differences between different pressures, are not significantly affected by the difference in saturating-fluid, and we therefore retain original kerosene-saturated data for such comparisons.

RESULTS

POROSITY, BULK DENSITY, AND MATRIX DENSITY

Table 1 presents the results of the core-plug measurements of porosity, water-wet bulk density, and matrix density.

Figure 1 compares bulk densities of these core plugs with the continuous whole-core density records of Niessen et al. (this volume). Because of the larger sampled volume for whole-core measurements compared with core plugs, exact comparison of samples at the same depth is not

Tab. 1- Petrophysical measurements on CRP-1 core plug samples.

Depth (mbsf)	Density (kg/m^3)	Porosity (fractional)	Matrix Density (kg/m^3)	Vol. Mag. Susc. ($\times 10^{-6}$ SI)	Pressure (MPa)	V_p (m/s)	V_s (m/s)	V_p/V_s
62.88-62.91	2293	0.233	2680	562	0.00	2865	1418	2.02
66.63-66.67	2176	0.295	2660	220	0.69	2425	1202	2.02
70.51-70.55	2200	0.337	2800	139				
75.46-75.48	1968	0.415	2640	211	0.00	2105	1088	1.94
86.21-86.25	2193	0.302	2700	1768	0.00	2462	1216	2.02
86.96-86.99	2131	0.331	2680					
98.15-98.18	2112	0.334	2660	2344	0.00	2338	1276	1.83
104.20-104.05	2538	0.112	2730	577	0.00	3533	1696	2.08
107.62-107.66	2396	0.186	2710	517	0.00	3156	1601	1.97
112.90-112.94	2166	0.301	2660	1000				
115.38-115.42	2388	0.176	2680	157	0.69	3126	1659	1.88
123.01-123.05	2333	0.25	2770	324				
124.86-124.89	2357	0.257	2820	886				
131.11-131.14	2300	0.238	2700	1089				
135.90-135.93	2341	0.241	2760	1371				
142.03-142.07	2080	0.368	-	1654	0.00	2550	1297	1.97
144.72-144.76	1993	0.421	2700	825	0.00	2154	1144	1.88
146.42-146.46	1971	0.427	2680	807	0.00	2124	1160	1.83

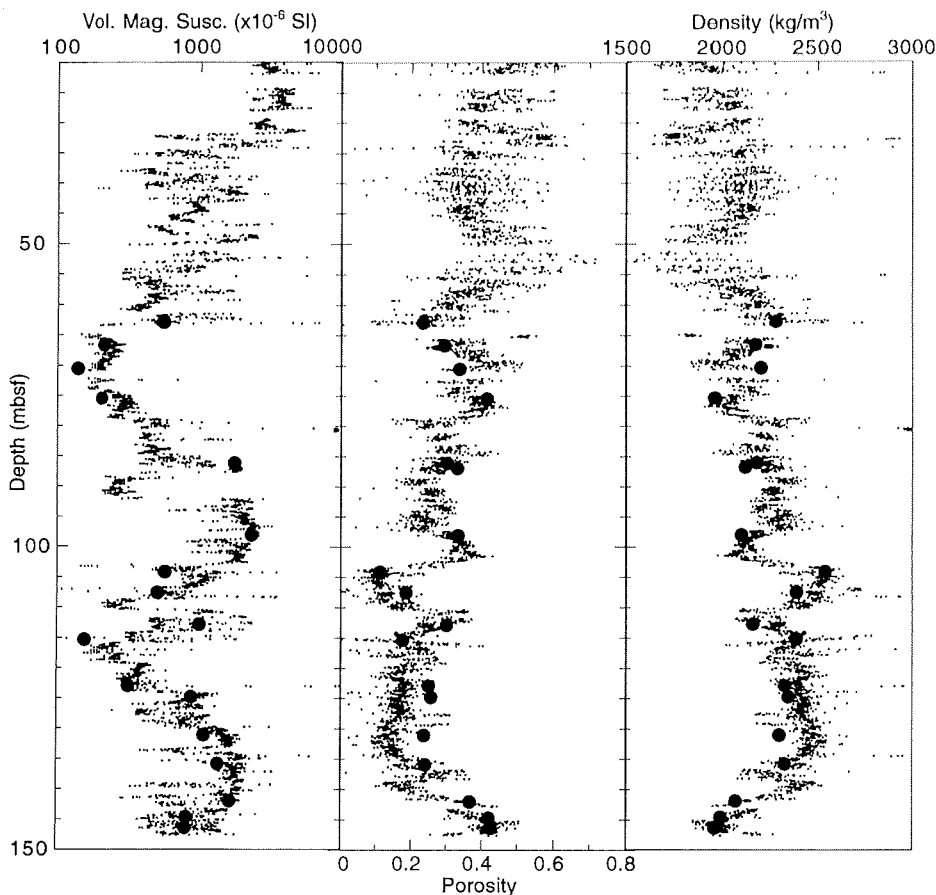


Fig. 1 - Comparison of core-plug (this study) and continuous whole-core (Niessen et al., this volume) measurements of density, porosity, and magnetic susceptibility. Small dots: whole-core data; large dots: core-plug data. Whole-core porosities are based on converting density measurements to porosity, using a matrix density of 2700 kg/m^3 determined from core plugs. Agreement of the two data sources is generally good to excellent.

possible. Nevertheless, the overall pattern is clearly one of very good agreement between the two measurement techniques. This consistency confirms the general accuracy of the whole-core dataset. *In situ* densities may be slightly higher than both core-plug and whole-core densities due to rebound, the expansion that cores undergo when removed from *in situ* lithostatic pressures to atmospheric pressure. Porosity and density rebound are, however, usually less than 1-2% for cores from <150 metres below sea floor (mbsf) (Hamilton, 1976).

To convert whole-core densities to porosities, it is necessary to assume a constant matrix density. Our core-plug measurements (Tab. 1) show that matrix density is quite uniform within the Miocene section of CRP-1, with a mean value of 2700 kg/m^3 . Accordingly, this value has been used by Niessen et al. (this volume) for conversion of whole-core densities to porosity. These data indicate that the assumption of uniform matrix density introduces only minor errors into the conversion from density to porosity. We note, however, that calculated matrix densities can be biased when smectite is abundant, because sample drying depletes smectite interlayer water. Smectite is present in small amounts throughout the hole, and is locally abundant within the interval 70-45 mbsf (Ehrmann, this volume).

Figure 1 also compares core-plug and whole-core porosities. Overall agreement is quite good. Three core-plug porosities for the interval 123-131 mbsf are higher than whole-core porosities, possibly suggesting that matrix density for this short interval is higher than the 2700 kg/m^3 used for conversion of whole-core densities to porosity.

MAGNETIC SUSCEPTIBILITY

Volume magnetic susceptibilities of these core plugs were measured both on a Bartington bridge (Cape Roberts Science Team, 1998) and on a Kappa bridge (this study). The two sets of measurements are generally consistent. The Kappa bridge data are systematically higher by about 20%, but this difference is small in comparison to intersample variations of more than an order of magnitude (Fig. 1). Agreement between the Kappa bridge data and the continuous, whole-core measurements of Niessen et al. (this volume) is excellent.

In some sedimentary environments, magnetic susceptibility can be highly correlated with clay content, because magnetic minerals often are most abundant in finest-grained sediments. Niessen et al. (this volume) find that CRP-1 magnetic susceptibilities tend to be higher in muds than in sands, but the overall correlation between magnetic susceptibility and clay content is weak. In contrast, they observe a good correlation between porosity and clay content. Accordingly, for the core-plug dataset of table 1, no correlation is observed between magnetic susceptibility and either porosity, density, velocity, or matrix density.

VELOCITY VERSUS PRESSURE

Table 1 lists P-wave velocity, S-wave velocity, and V_p/V_s for the lowest-pressure steps of all samples that exhibited adequate coupling for useful measurements.

This section concentrates on P-wave velocities. We note, however, that atmospheric-pressure V_p/V_s is remarkably consistent among all samples (1.8-2.1) and agrees well with ratios predicted (Castagna et al., 1985; Han et al., 1986) for siliciclastic rocks of similar porosity.

Background

Measurements of velocity at atmospheric pressure are usually not representative of *in situ* velocities, for two reasons: reduced interparticle coupling and microcrack opening.

Increased overburden pressure increases the number and area of interparticle contacts, thereby increasing shear modulus and frame bulk modulus, and this increased framework stiffness increases velocity (Stoll, 1989). This effect is present even if microcracks are absent. Hydrostatic fluid pressure cannot accomplish these skeletal changes, and consequently the velocities of sea-floor sediments are independent of water depth. For elastic moduli, the relevant skeletal pressure is the differential pressure, or difference between overburden (or lithostatic) pressure and pore pressure (Wyllie et al., 1958). In relatively permeable sedimentary sequences, differential pressure increases with depth because lithostatic pressure increases at about double the rate of hydrostatic pressure increase. The expected magnitude of pressure-induced velocity influence is, however, both model-dependent and sensitive to assumptions concerning soft sediment elastic moduli (Stoll, 1989; Dvorkin & Nur, 1995).

Stress relaxation, whether *in situ* or coring-induced, can generate and open microcracks. Virtually all core samples, regardless of lithology, exhibit patterns of increasing P-wave velocity with increasing pressure attributable to closing of microcracks (e.g., Nur, 1971; Bourbié et al., 1987). Initial microcrack porosities of <0.005 are sufficient to cause pressure-dependent velocity variations of 5-50%, indicating that the primary effect of this pressure on velocity is through its impact on frame bulk modulus, not on porosity or density (Walsh, 1965; Nur & Murphy, 1981; Bourbié et al., 1987).

Microcrack opening is not confined to cores removed from *in situ* conditions; it can also occur in response to *in situ* changes in stress state. Microcracks may affect the velocity-porosity relationship of any sediment that has undergone a large decrease in overburden stress. For example, Jarrard & Erickson (1997) found that exhumation changes the velocity-porosity pattern for both well-logs and core measurements from the Ferron Sandstone. They found that the velocity-porosity relationship seen for low-pressure velocity measurements on outcrop cores agreed with shallow log data, whereas that for high-pressure velocity measurements agreed with deep log data. This pattern was consistent with the hypothesis of a pressure-release effect due to exhumation. The Miocene section at CRP-1 has been exhumed, based on seismic evidence for an angular unconformity (Henry et al., this volume). Therefore, the possibility of an associated velocity decrease must be investigated.

Velocity Response to Pressure Change

To permit extrapolation of atmospheric-pressure velocity measurements to *in situ* conditions, the effects of stress relaxation and microcrack opening on core samples must be reversed by measuring the samples at *in situ* pressures. Modern *in situ* lithostatic pressures for these shallowly buried (63-147 mbsf) sediments are only 0.6-1.5 MPa. However, lithostatic pressure at maximum burial was much higher: roughly 10.3 MPa if maximum burial was 1 000 m. Niessen and Jarrard (this volume) suggest that the degree of compaction of CRP-1 sediments is comparable to sediments from much less than 1 000 m burial. However, in view of uncertainties in maximum burial depth, our velocity experiments extend to 17.2 MPa.

A standard suite of velocity measurements typically consists of 5-8 pressure steps between atmospheric pressure and maximum pressure, usually beginning at maximum pressure and then decreasing pressure until coupling is lost at about 3.45 MPa. Both V_p and V_s are measured at each pressure step. Such a pattern is appropriate for lithified rocks, but not necessarily for CRP-1 sediments.

Unlike well-lithified sedimentary rocks, unlithified sediments such as those from CRP-1 do not necessarily respond elastically to variations in confining pressure. They may deform viscoelastically, or they may break down, at fairly modest pressures. Therefore, we measured velocity on both upgoing and downcoming pressure cycles, beginning with the upgoing cycle. Ideally, upgoing and downcoming cycles would overlap, but actual runs often show slightly lower velocities on the upgoing cycle than on the downcoming cycle. This difference is attributable to the very short time (<10 minutes) between changing pressure and measuring. Following a change in confining pressure, fluid moves into or out of the sample pores, and this equilibration of the pore fluid to the pressure change can take a few minutes, particularly for relatively impermeable samples. From the perspective of differential pressure rather than measured confining pressure, therefore, measurements made during the upgoing cycle are slightly overpressured, and measurements during the downcoming cycle are slightly underpressured. If the highest-pressure (17.2 MPa) measurement is repeated after 20 minutes, its velocity rises by 1-3% (Fig. 2) due to this equilibration. An improved experimental design would take into account the full amount of time necessary for pressure equilibrium to take place within the samples at individual pressure increments.

The first five sample runs consisted (approximately) of the following pressure steps: 0, 0.69, 1.38, 3.45, 5.17, 6.90, 10.3, 13.8, 17.2, 13.8, 10.3, 6.90, 5.17, 3.45, 1.38, 0.69, and 0 MPa. As is often noted for more lithified rocks, it was not always possible to detect useful P or S arrivals at the 0 MPa and 0.69 MPa steps, due to insufficient coupling of sample to transducer. Two of these samples exhibited much lower lowest-pressure velocities for the decreasing-pressure cycle than for the increasing-pressure cycle, possibly indicating that the high-pressure steps had damaged sample strength. No visual breakdown or other change was noted for the samples, but stress-induced modification of intergrain contacts may decrease

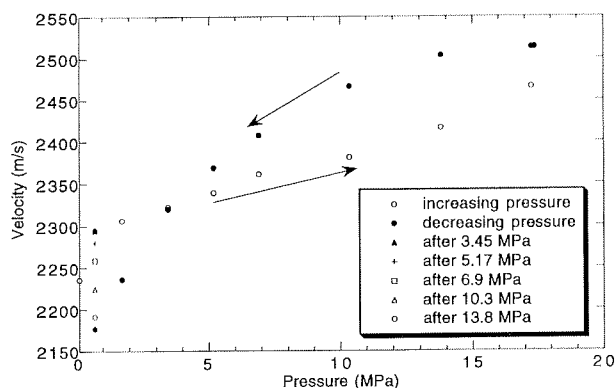


Fig. 2 - Example of compressional-wave velocity measured as a function of pressure. At high pressures, velocities measured during the decreasing-pressure cycle are a few percent higher than those made during the increasing-pressure cycle. Repeat 0.69 MPa measurements made after each increasing-pressure increment demonstrate a systematic decrease, caused by pressure-induced incipient breakdown of the sample.

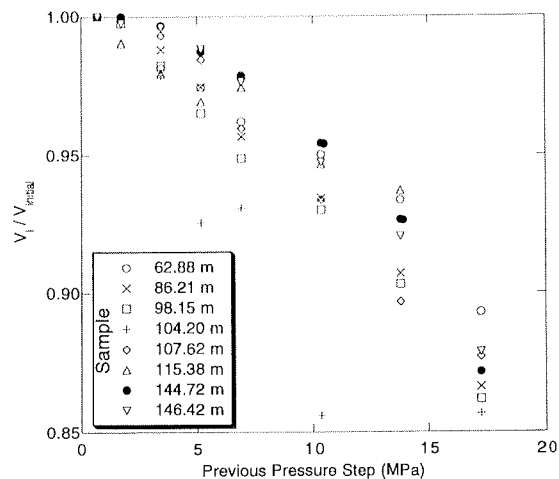


Fig. 3 - Effect of sample cycling to increasing pressures on 0.69 MPa compressional-wave velocity, for 8 samples. Exposure to pressures of 5.2-6.9 MPa decreases velocities subsequently measured at 0.69 MPa by 2-7%, due to incipient breakdown of the sample. Exposure to pressures of 13.8-17.2 MPa increases this breakdown effect to 7-14%.

framework bulk modulus and shear modulus without causing macrofractures. Two other samples in this first batch had already been subjected to 17.2 MPa pressures during prior runs that had been unsuccessful because of poor coupling; though they showed similar velocities for increasing- and decreasing-pressure runs, both runs could have been adversely affected by incipient breakdown, so these data are excluded from table 1 and the figures.

To document and isolate any breakdown effect of the experiment on the samples, runs on a subsequent sample followed each increased-pressure increment with a repeat of earlier, lower-pressure steps. This experimental suite shows two major features. First, the initial effect of pressure reduction after high pressures is to produce velocities that are higher on the decreasing-pressure cycle than on the increasing-pressure cycle. This hysteresis effect, as discussed above, is attributable to measurement times that are rapid in comparison with establishment of equilibrium pore pressures within the sample. Second, repeated measurements at the lowest pressure (0.69 MPa) are affected by cycling to progressively higher pressures: initial low-pressure steps have no effect on the 0.69 MPa results, but exposure to pressures of 10.3-17.2 MPa causes a substantial reduction in subsequent 0.69 MPa measurements. This result confirms that pressures of 10.3-17.2 MPa are sufficient to cause incipient breakdown of sample strength (frame and shear moduli).

For all samples subsequently measured, each increased-pressure increment was alternated with a return to 0.69 MPa for remeasurement. A plot of these repeated 0.69 MPa measurements for each sample (Fig. 3) shows that most samples exhibit incipient breakdown at pressures of 3.45-6.90 MPa, followed by substantial breakdown and associated 5-14% velocity reduction at pressures of 10.3-17.2 MPa. Separate consolidation tests would have provided a useful perspective on this pressure-dependent behaviour. Consolidation tests had been planned for some CRP-1 whole-round samples, but the samples were considered to be unsuitable.

This pattern of incipient breakdown at modest pressures is incompatible with the behaviour expected for rocks whose strength was established by consolidation and cementation at a maximum burial much greater than one kilometre. Nor do many of the CRP-1 samples exhibit the strong initial pressure dependence of velocity that is the diagnostic signature of microcrack closing. Apparently, burial was accompanied by little or no cementation, and the subsequent exhumation did not induce pervasive microcrack opening. In a study of sediments from Nankai prism, Karig (1993) found that sample breakdown may be caused by repeated pressure cycling in addition to sample exposure to high pressures. In this study, however, replicate velocity measurements at the same pressure indicate little or no breakdown related to cycling alone.

An alternative hypothesis, more compatible with the velocity results, is that a diagenetic "annealing" episode responsible for the present rock strength occurred following exhumation. Petrographic and oxygen isotope studies indicate that the final stage of diagenesis of CRP-1 sediments was an episode of carbonate cementation, probably from mixed fresh water and seawater, occurring after brittle fracturing (Baker & Fielding, this volume). The degree of carbonate cementation was small, with carbonate contents that are generally only about 1% (Ehrmann, this volume; Baker & Fielding, this volume) and consisting of thin rims of calcite or siderite (Claps & Aghib, this volume; Baker & Fielding, this volume). A relatively shallow, post-erosional timing for diagenesis best fits both these sedimentological constraints and our petrophysical evidence that the incipient cementation responsible for the framework and shear moduli occurred at relatively shallow burial. Consequently, this strength begins to break down at pressures of only 3.45 to 6.90 MPa.

An unfortunate consequence of this apparent late-stage stabilisation of acoustic properties is that the samples have lost any fingerprint of ancient deeper burial. Thus,

the high-pressure portions of these velocity runs are not representative of more deeply buried portions of the same V3 formation, eastward of CRP-1.

Implications for *In Situ* Velocities

When considered in conjunction with the whole-core velocity measurements of Niessen et al. (this volume), these core-plug velocity data have two significant implications for *in situ* velocities. First, the core-plug measurements provide an independent confirmation of the reliability of the whole-core measurements. Second, the core-plug data indicate how different *in situ* velocities are likely to be from whole-core measurements made at laboratory pressure.

Figure 4 overlays our lowest-pressure core-plug measurements on a plot of whole-core velocities *versus* depth. The core-plug measurements shown are generally atmospheric-pressure velocities measured before the upgoing pressure cycle, and therefore before any possible breakdown. For two samples, atmospheric-pressure velocity could not be measured because of inadequate coupling, so 0.69 MPa velocity is shown instead. Exact correspondence of individual measurements in figure 4 is not expected, because of the different volumes measured, but overall patterns of agreement or disagreement are useful. A few whole-core velocities of $>4\,000$ m/s (not shown) are associated with large limestones (Niessen et

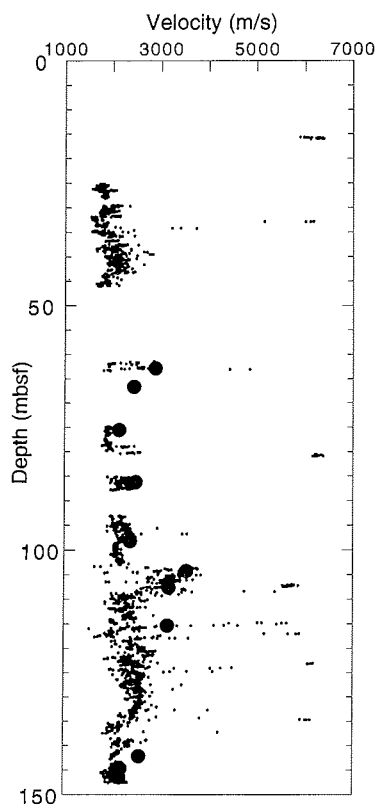


Fig. 4 - Comparison of compressional wave velocities measured on core-plugs (at 0 or 0.69 MPa) with those measured at atmospheric pressure on whole cores (Niessen et al., this volume). Some core-plug velocities appear to be systematically higher than whole-core velocities. Small dots: whole-core data; large dots: core-plug data.

al., this volume), avoided by the core-plug sampling. Small limestones are pervasive in CRP-1 (Cape Roberts Science Team, 1998; Brink et al., this volume), but they appear to have little effect on porosities and velocities (Niessen & Jarrard, this volume).

Although both whole-core and initial core-plug velocity measurements were at atmospheric pressure, the two may not be identical because sediment rebound, due to removal from *in situ* pressures to atmospheric pressure, is gradual, not instantaneous. The whole-core measurements were made within hours of core retrieval, whereas the core-plug measurements were made 8 months later. If residual rebound occurred after the whole-core measurements were made, then core-plug velocities would be lowered with respect to whole-core velocities. Instead, a systematic bias may be present that is of opposite sign to rebound effect.

An additional consistency check, which avoids the problem of comparing velocity data from different volumes and depths, is comparison of the velocity/porosity relationships of the two datasets. As shown in figure 5, both core-plug and whole-core measurements indicate a strong effect of porosity on velocity. For a given porosity, core-plug velocities are systematically about 10%, or 200-500 m/s, faster than whole-core velocities. The cause of this difference is uncertain. The difference could be caused by undetected bias in either the core-plug or whole-core velocity measurements, but both measurement suites included standards. Alternatively, both datasets may be accurate but not fully representative of *in situ* conditions. The low whole-core velocities may be related to a loss of rigidity caused by *in situ* brecciation, *in situ* exhumation, core rebound or core disturbance. Another possibility is that drying of the core-plugs may have induced diagenetic change such as salt precipitation or smectite dehydration.

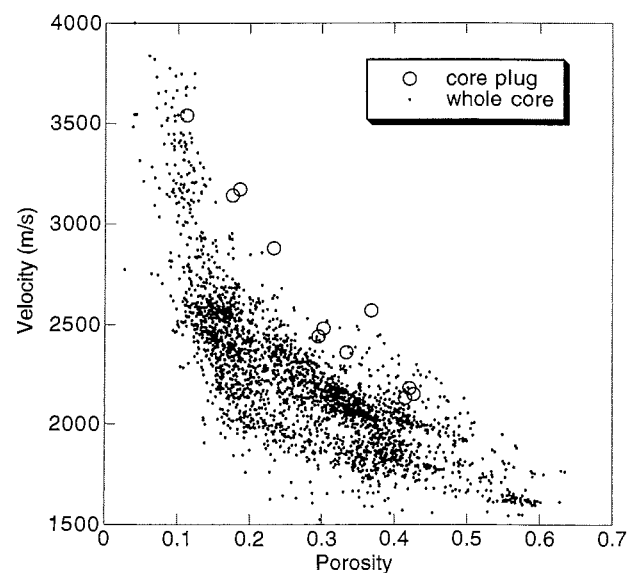


Fig. 5 - Velocity-positivity relationship for CRP-1, based on both core-plug data of this study and whole-core data of Niessen et al. (this volume). Both datasets demonstrate the strong effect of porosity on velocity, but core-plug velocities are systematically a few percent higher than whole-core velocities.

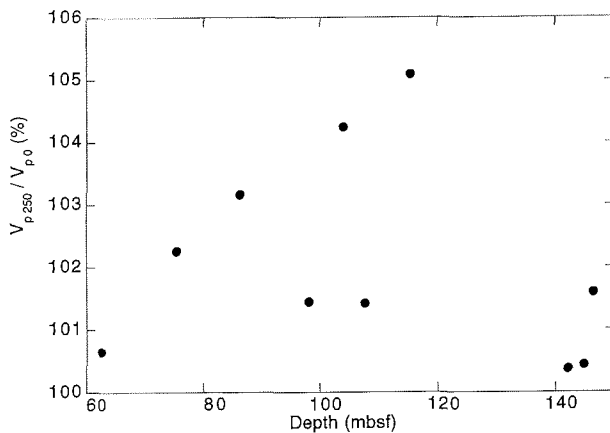


Fig. 6 - Percentage difference between 1.38 MPa measurements and atmospheric-pressure measurements of P-wave velocity, plotted versus depth. Any CRP-1 velocity measurements made at atmospheric pressure should be increased by 1-5% to be representative of velocities at *in situ* pressures.

However, the discrepancy is similar in smectite-rich and smectite-poor intervals. Niessen & Jarrard (this volume) discuss in detail the possible mechanisms for this difference, in the context of the CRP-1 velocity/porosity relationship.

Our measurements of velocity *versus* pressure provide an indication of the likely differences between *in situ* velocities and those measured on continuous cores at laboratory pressure. Figure 6 plots the percentage difference between 1.38 MPa measurements and atmospheric-pressure measurements *versus* depth; in both cases upgoing-cycle measurements are used. *In situ* differential pressures are about 0.62-1.51 MPa, so the values shown in figure 6 slightly overestimate the likely difference between *in situ* and laboratory measurements. On average, *in situ* velocities are probably 1-3% higher than those measured at atmospheric pressure. Consequently, the depth of the V3-V4 seismic reflector is probably 1-3% deeper, or only 2-5 m deeper, than is estimated with atmospheric-pressure velocities (*e.g.*, Cape Roberts Science Team, 1998; Bucker et al., this volume).

ACKNOWLEDGEMENTS

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REFERENCES

- Bourbié T., Coussy O. & Zinszner B., 1987. *Acoustics of Porous Media*. Ed. Tech., Paris, 334 p.
- Cape Roberts Science Team, 1998. Initial Report on CRP-1, Cape Roberts Project, Antarctica. *Terra Antarctica*, **5**(1), 187 p.
- Castagna J.P., Batzle M.L. & Eastwood R.L., 1985. Relationships between compressional-wave and shear-wave velocities in elastic silicate rocks. *Geophysics*, **50**, 571-581.
- Dvorkin J.P., & Nur A.M., 1995. Elasticity of high-porosity sandstones: theory for two North Sea datasets. *Expanded Abstracts, 65th Ann. Int. SEG Meeting, Houston*, 890-893.
- Gassmann F., 1951. Elastic waves through a packing of spheres. *Geophysics*, **16**, 673-685.
- Hamilton E.L., 1971. Elastic properties of marine sediments. *J. Geophys. Res.*, **76**, 579-604.
- Hamilton E.L., 1976. Variations of density and porosity with depth in deep-sea sediments. *J. Sediment. Petrol.*, **46**, 280-300.
- Han D., Nur A. & Morgan D., 1986. Effects of porosity and clay content on wave velocities in sandstones. *Geophysics*, **51**, 2093-2107.
- Jarrard R.D. & Erickson S.N., 1997. Sandstone exhumation effects on velocity and porosity: perspectives from the Ferron Sandstone. *Abstracts AAPG Annual Meeting, Dallas*.
- Karig D.E., 1993. Reconsolidation tests and sonic velocity measurements of clay-rich sediments from the Nankai Trough. *Proc. Ocean Drilling Program Sci. Results*, **131**, 247-260.
- Nur A., 1971. Effects of stress on velocity anisotropy in rocks with cracks. *J. Geophys. Res.*, **76**, 2022-2034.
- Nur A. & Murphy W., 1981. Wave velocities and attenuation in porous media with fluids. *Proc. 4th Int. Conf. on Continuum Models of Discrete Systems*, Stockholm, 311-327.
- Sondergeld C.H. & Rai C.S., 1993. A new exploration tool: quantitative core characterization. *PAGEOPH*, **141**, 249-268.
- Stoll R.D., 1989. *Sediment Acoustics*. Springer-Verlag, Berlin, 153 p.
- Walsh J.B., 1965. The effect of cracks on the compressibility of rock. *J. Geophys. Res.*, **70**, 381-389.
- Wyllie M.R.J., Gregory A.R. & Gardner G.H.F., 1958. An experimental investigation of factors affecting elastic wave velocities in porous media. *Geophysics*, **23**, 459-493.