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USE OF GAMMA RAY ATTENUATION; ACOUSTIC VELOCITY;
AND ELECTRICAL RESISTIVITY IN THE EXTRACTION
OF SEDIMENT PHYSICAL PROPERTIES

BY
VEITH ALTMANN

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
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IN
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UNIVERSITY OF RHODE ISLAND

2003

MASTER OF SCIENCE
OF
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ABSTRACT

Sediment physical properties, such as porosity, shear strength, shear modulus, and grain size, from six different geographical marine environments (Central Arctic Ocean, Central Scotian Shelf, Gulf of Mexico, Caribbean Sea, North-Eastern Pacific, Western Equatorial Pacific and Mid-Atlantic Ridge) were analyzed and correlated with non-destructive measurements of compressional wave velocity, bulk density and electrical resistivity.

In coarse grained sediments (median grain diameter $>4 \mu\text{m}$) with a sand fraction greater than 15%, larger grain size is associated with higher velocity. In fine grained sediments (median grain diameter $< 4 \mu\text{m}$) velocity does not correlate with median grain size. However, a pronounced linear relationship exists between compressional wave velocity and percent clay fraction (grain diameter $< 2 \mu\text{m}$) regardless of sediment grain size.

The use of electrical resistivity is limited for predicting porosity in unconsolidated sediments. In general, the trend of decreasing sedimentary porosity with increasing electrical resistivity, is consistent with previous observations (e.g., Archie, 1942). However, the trends are sediment-type dependent and therefore resistivity measurements cannot be used as sole predictor of porosity.

Compressional wave velocity and bulk density, expressed as the elastic parameter $\Xi (V_p^2 * \rho_B)$ correlate well with miniature vane shear data. These non-destructive measurements can be used to predict undrained shear strength.

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PREFACE

This thesis was prepared and submitted in the manuscript style and includes three manuscripts and one appendix. The manuscripts are formatted for publication in a scholarly journal. Each manuscript describes the evaluation of physical properties by non-destructive measurements such as gamma-ray attenuation, compressional wave velocity, and electrical resistivity. These non-destructive measurements are compared to direct measurements of porosity, grain size and shear strength. Relationships were found between sediment properties and non-destructive measurements from a wide range of marine environments to increase the boarder applicability of the findings.

Manuscript I focuses on the correlation of compressional wave velocity and bulk density with grain size. The prediction of porosity by means of electrical resistivity, compressional wave velocity and bulk density comprises Manuscript II. The last manuscript describes the prediction of shear parameters such as shear strength and shear modulus using compressional wave velocity and bulk density.

Appendix A contains a comprehensive description of non-destructive measurement methods. A Multi-Sensor-Track (MST), compressional wave velocity logger (P-Wave Logger), Gamma-Ray Porosity Evaluator (GRAPE) and a resistivity logger are introduced, and a detailed description of principles and methods used to evaluate the respective properties is given.

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Manuscript I

Compressional Wave Velocity and Bulk Density in the Prediction of Grain Size

Introduction

Compressional wave (P-Wave) velocity has been used as a predictor of geotechnical and physical properties of soils, sediments, and rocks since the 1950's. Compressional wave velocity strongly depends on mass physical properties such as porosity, bulk density, and elastic properties such as bulk modulus and shear modulus. Grain size also depends on with compressional wave velocity, although to a lesser extent. These parameters are of great interest for the geotechnical engineer.

Developing fully automated systems such as Multi-Sensor-Tracks (MST) to measure compressional wave velocity and other parameters (e.g., gamma-ray attenuation, electrical resistivity, and magnetic susceptibility), which also can be related to physical properties, enabled us to correlate between non-destructive measurements and physical properties.

The advantage of these new systems to evaluate physical properties is a reduced cost and time. In addition, non-destructive evaluation of physical properties provides continuous records of geotechnical properties down core, and measurements are performed on whole core samples encased in the original sample liner the cores are not disturbed by discrete sampling. Misinterpretations, due to the interpolation or extrapolation of information obtained by discrete measurements are minimized by

continuous measurements obtained by Multi-Sensor-Tracks.

Many studies concern the relationships between compressional wave velocity and physical properties such as bulk density, porosity, and grain size (Bachman, 1985; Hamilton, 1970; Schreiber, 1968). Impedance, the product of compressional wave velocity and bulk density has also been used as an acoustic property, to determine physical properties (Bachman, 1985; Schnack-Friedrichsen *et al.*, 2001).

The purpose of this paper is to determine the empirical correlation between compressional wave velocity and grain size. The aim is to present a relationship that is significant enough to allow for prediction of grain size. This prediction can be made in terms of median grain diameter or in terms of fractional contents of different particle sizes (Sand, Silt, Clay).

The empirical correlation was performed by using laboratory determined grain size values and non-destructive evaluated compressional wave velocity. Furthermore the incorporation of non-destructive derived bulk density (GRAPE density) is used as a factor to improve the correlation results.

For this study, sediment physical properties of five cores from three different geographic marine environments were analyzed and correlated: (1) three cores from the Central Arctic Ocean (Lomonosov Ridge); one core from the Central Scotian Shelf (Emerald Basin); and one core from the Western Equatorial Pacific (Ontong Java Plateau). The core analyses were performed for each respective scientific project and not by the author, however, a sufficient description of the testing conditions and physical properties is provided.

Background

In general, the velocity of a compressional wave is an expression of the elastic properties of the medium, through which it penetrates. The basic equation for the velocity of a compressional wave is:

$$V_p = \left(\frac{B + \frac{4}{3} * G}{\rho_B} \right)^{\frac{1}{2}} \quad (1)$$

where V_p is the compressional wave velocity (m/s), B is the bulk modulus (KN/m²), G is the shear modulus (KN/m²), and ρ_B is the bulk density (Mg/m³) (Hamilton and Bachman, 1982).

Other physical properties such as porosity and grain size affect the compressional wave velocity through its effects on the elasticity of the sediments (bulk modulus and shear modulus) and on the bulk density. A general trend is observed in the relationship between velocity and grain size: an increase in grain size is associated with an increase in velocity (Hamilton and Bachman, 1982; Morgan, 1969; Sutton *et al.*, 1957).

The grain size affects the velocity predominately by its influence on porosity. The porosity, in turn, highly influences the bulk density and the bulk modulus of the sediments and thus indirectly influences the velocity. The influence of porosity on the bulk modulus is straight forward. Marine sediments can be considered as two-phase mixtures and consequently the porosity is an expression of the water content since typically no gaseous phase is present (100% saturated). The relationship between porosity and bulk modulus is given by Equation (2):

$$B = \frac{B_F * B_S}{\phi * (B_S - B_F) + B_F} \quad (2)$$

where B_F is the pore fluid bulk modulus (KN/m²), B_S is the bulk modulus of the

mineral constituent (KN/m^2), and ϕ is the fractional porosity (-) (Hamilton, 1971). This equation is only valid for sediments that lack rigidity. However, to emphasize the influence of porosity on bulk modulus, this equation is appropriate. From Equation (2) it can be deduced that a decrease in porosity decreases the denominator and thus decreases the influence of the water bulk modulus on the overall bulk modulus. Although within the same order of magnitude, the bulk modulus of water is lower than the bulk modulus of the mineral solids. Therefore, a decrease in water increases the bulk modulus. An increase in bulk modulus is associated with an increase in compressional wave velocity (Equation (1)).

In comparison to the porosity-bulk modulus relationship, the influence of grain size on porosity is very complex. Numerous factors affect this relationship. The most important interrelated factors, in terms of the mineral grains and influence on porosity, are grain size, uniformity of grain size (sorting), grain shape, packing of grains, and mineralogy. In general, an increase in porosity is interdependent with a decrease in grain size (Hamilton and Bachman, 1982). Some general observations of the influences of mineral grain characteristics on porosity are:

- (1) well-sorted sediments (high uniformity) have higher porosities;
- (2) fine grained platy sediments have a more porous structure than coarser spherical sediments due to interparticle forces that cause the fine particles to stick together and do not allow a reorientation of the particles to form a more dense packing;
and
- (3) well rounded (spherical) grains are less porous than angular grains.

Also, grain size influences the sediment shear modulus and, therefore, influences velocity (Sutton *et al.*, 1957). Given the same velocity, less solid grain-to-grain contact occur in fine grained sediments with oriented, platy grains than in coarser sediments with more spherical grains. This grain-to-grain contact (packing) influences the shear modulus in terms of an increasing shear modulus with increase in grain-to-grain contacts. This, however, contributes, to a lesser extent, to the velocity-grain size relationship.

The grain size also affects bulk density. The relationship between bulk density and porosity was investigated in earlier studies (Hamilton, 1970). As expected, a decrease in porosity is linearly related to an increase in bulk density. A decrease in porosity increases the bulk modulus and thus increases the velocity. Since bulk density is in the denominator of Equation (1), an increase is linked to a decrease in velocity when other parameters remain constant. However, the influence of porosity (in terms of grain size) on bulk density and hence on velocity, when compared with its influence on the bulk modulus is negligible. This becomes evident when absolute values are compared. A decrease in porosity as a consequence of a variation in grain size changes bulk density by less than 10^{-1} , whereas the bulk modulus changes under the same conditions by at least 10^3 .

Methods

Compressional Wave Velocity. Compressional wave velocity was determined using a P-wave logger (PWL) on the Multi Sensor Track (GEOTEK). An ultrasonic pulse with a dominant frequency of 500 kHz was transmitted across the unopened core sample and the travel time was measured. The velocity was then calculated by

dividing the core diameter by the pulse travel time. Corrections for transducer and core liner time delays as well as for core diameter deviations were applied (APPENDIX A, Schultheiss and McPhail, 1989).

Bulk Density. Bulk density measurements were performed using a gamma-ray attenuation porosity evaluator (GRAPE) (Evans, 1965). The measurement of sediment bulk density using gamma-rays is based on the principles of Compton scattering and attenuation. A parallel, monoenergetic beam of gamma-rays (^{137}Cs) penetrates the core sample and is detected on the opposite side by a scintillation counter. When passing through the sample some of the gamma-rays are absorbed or scattered and lose energy and direction, respectively. The scintillation counter detects the gamma-rays that pass through the absorber without any loss of energy. The energy loss and the attenuation respectively are directly related to bulk density. A discrete value for bulk density is then derived by calibrating the attenuation of gamma-rays through the unopened core sample with the attenuation through standards of aluminum and water (APPENDIX A, Boyce, 1976).

Porosity. The porosity ϕ was determined from the bulk density using an estimated specific gravity of 2.75 with Eq. (17):

$$\phi = \frac{\rho_s - \rho_B}{\rho_s - \rho_F} \quad (3)$$

where ρ_s is the specific gravity (Mg/m^3), ρ_B is the bulk density (Mg/m^3), and ρ_F is the pore fluid density (Mg/m^3) (density of sea water, $\rho_{sw} = 1.025 \text{ Mg/m}^3$).

Grain Size Analysis: Grain size analyses were performed using two different methods. Sediments from the Central Arctic Ocean (Lomonosov Ridge) were analyzed using X-ray attenuation (SediGraph grain size analyzer) and the coulter counter

method (Coulter Counter) was used to determine the grain size of sediments from the Central Scotian Shelf (Emerald Basin) and the Western Equatorial Pacific (Ontong Java Plateau).

The Sedigraph grain size analyzer measures the attenuation of X-rays by particles that are suspended in a solution (Jones *et al.*, 1988). The SediGraph determines the concentration of particles remaining at decreasing depth within a suspension as a function of time. The principle of Stoke's Law of Settling is used to convert vertical profiles of suspension density to weight percentages of grain size.

The coulter counter method is based on a principle where particles suspended in an electrolyte pass through a small aperture with electrodes on both sides. The passing particles displace their own volume of electrolyte, whereby the resistance in the current is changed in proportion of the volumetric size of the particles. The number of changes per time reflects the number of particles per volume in suspension.

Before analysis, the sediments were chemically pretreated to remove organic matter and to disaggregate the particles. The suspension was washed on a 63 μm sieve to separate the sand-sized grains from silt and clay. The retained portion was oven dried and then dry sieved, following ASTM D421/422. The clay and silt fraction ($<63 \mu\text{m}$) was analyzed in the Coulter Counter and in the Sedigraph, respectively.

Finally, the results of both tests were combined and a cumulative semi-logarithmic frequency curve for each analyzed sample was developed. The median grain diameter which is the value that corresponds to the 50% mark (d_{50}) on the cumulative frequency curve was then determined.

Site Description and Sediment Properties

Five cores from three different sites were selected for this study. The major geographic environments are: (1) the Central Arctic Ocean (Lomonosov Ridge), (2) the Scotian Shelf (Emerald Basin), and (3) the Western Equatorial Pacific (Ontong Java Plateau). A summary of the cores used in this study and a detailed description of the locations are shown in Table 1.

Central Arctic Ocean (Lomonosov Ridge). The Lomonosov Ridge is located in the Central Arctic Ocean between the longitude 130° to 155° East and the latitude 85° to 90° North (Figure 1). The Lomonosov Ridge separates the Markarov Basin and the Amundsen Basin. The ridge crest is at its highest less than 1000 m below sea level and drops down on both sides into the adjacent basins with depths of more than 3000 meters below sea level. The sediments of the Lomonosov Ridge are predominantly hemipelagic with minor ice rafting components.

The sediments of the retrieved cores from the Lomonosov Ridge (96/09-1pc, 96/12-1pc, and 96/13-1pc) are described as clays and silty clays and the lithology was divided into three geotechnical units (Jakobsson *et al.*, 2001). Unit I consists of a thin layer of dark brown clay at the surface and is underlain by a layer of yellowish brown to dark gray silty clay. The second Unit II is composed of olive gray clay and silty clay and unit III is an indurated dark olive gray silty clay (Figure 2 to Figure 4). The physical properties of all cores follow identical trends with depth. The bulk density of Unit I and Unit III increases with depth. Values range from 1.80 Mg/m³ in Unit I to 2.06 Mg/m³ in Unit III. The bulk density of Unit II is lower than the bulk density of the other two units. Bulk density of Unit II is alternating between maxima of 1.85 Mg/m³ and minima of 1.60 Mg/m³.

The compressional wave velocity follows the trend of the bulk density. The velocity in Unit I increases with depth and does not have much scatter. The velocity trend of Unit I is continued in Unit III. Values of compressional velocity range from 1450 m/s to 1590 m/s. The velocity of Unit II is much lower and is nearly constant with depth. The average velocity value for Unit II is about 1460 m/s.

The median grain size of Unit I is increasing with depth. The observed values vary from 2 μm and 60 μm . Scatter in the trend is caused by thin lenses of fine grained sand. At the unit break (Unit I and II), the median grain sizes decrease to a value of approximately 3 μm that continues almost constantly with depth (Unit II and III).

Central Scotian Shelf (Emerald Basin). The Emerald Basin is a 2430 km^2 depression located on the central Scotian Shelf approximately 40 km off the coast of Nova Scotia and reaches its maximum depth at 290 m below sea level (Figure 5). The Basin is filled with glacial till that is overlain by Quaternary fine grained glacio-marine and marine sediments and underlain by firm bedrock. In general the sediment stratigraphy can be divided into three major geotechnical units (Figure 6). However, the sediments retrieved with core 87003-02 are only composed of the upper two units. Unit I consists predominantly of marine silty clay and clayey silt, olive gray in color. Unit II is mainly dark gray glacio-marine silty clay (Moran *et al.*, 1991). Bulk density increases uniformly with depth and no distinct variation in bulk density due to the differences in composition between Unit I and II are observed. Density ranges from 1.40 Mg/m^3 at the top of the core to 1.6 Mg/m^3 at a depth of 16.5 meters. The compressional wave velocity is relative constant with depth and has an average value of 1446 m/s.

Median grain size is constant with depth and is narrowed to a very small range with an average size of 2 μm . A peak median grain size of 20 μm appears to be anomalous at the unit break.

Western Equatorial Pacific (Ontong Java Plateau). The Ontong Java Plateau, with an area of 1.5 million km^2 , is the largest plateau found in the world oceans. It is located in the western equatorial Pacific north of the Solomon Islands (Figure 7). The Ocean Drilling Program Site 807A is located at the northern rim of the plateau in a water depth of 2804 meters. In general the sediments are divided into three major geotechnical units. However, grain size analyses were performed on the sediments of Unit I only and therefore this unit is emphasized in this discussion (Figure 8).

Unit I consists primarily of silts that range from sandy silts to clayey silts. Noticeably, the sediments are predominately composed of foraminifers and nannofossils. The nannofossil content in the sediments averages about 75% and the foraminifer abundance averages approximately 23%. The remaining fraction is composed of trace amounts of quartz and clay minerals.

Foraminifers and nannofossils are skeletal remains of pelagic organisms and predominantly composed of carbonate. The size of these particles range between 10 and 40 μm . Foraminifer and nannofossils are characterized by a hollow structure consisting of one or more affiliated chambers with wall thicknesses between 1.0 to 1.5 micrometers (Figure 9).

Bulk density is relatively uniform and increases with depth from 1.50 Mg/m^3 to 1.80 Mg/m^3 . Except for some scatter, the compressional wave velocity follows the same trend as the bulk density. A slight increase with depth is observed. Values for the

velocity range between 1450 m/s and 1500 m/s and average 1475 m/s. The average median grain size is approximately 28 μm (medium silt). Slight variations from average values are observed in the middle section of the unit where the median increases to 40 μm .

Results

Figure 10 and Figure 11 show plots of median grain size versus compressional wave velocity and impedance for all sediments. Two main data populations are observed. The hemipelagic sediments from the Central Arctic Ocean and the Central Scotian Shelf seem to form a different population than the calcareous sediments from the Western Equatorial Pacific. The grouping of the sediments is more pronounced in the correlation of median grain size and impedance (Figure 11). On this account the hemipelagic and calcareous sediments are separately investigated in the following.

Hemipelagic Silts and Clays: Plots of median grain size versus compressional wave velocity and impedance ($V_P \times \rho_B$) are shown in Figure 12 and Figure 13. The empirical relationships between these parameters published by Bachman (1985) and Hamilton (1982) are also presented. Two different trends are observed. Grain sizes between 1 μm and 4 μm seem to have no influence on compressional wave velocity. The observed trend of these fine grained sediments is parallel to the velocity axis and covers nearly the whole range of velocities used in this study (1420 m/s to 1600 m/s). At a median of 4 μm , the influence of the grain size on the velocity becomes more pronounced. An increase in median grain size is associated with an increase in velocity. A logarithmic regression line, fit to all data, results in a low interdependency, demonstrated by low correlation ($R^2 = 0.297$). An analysis of the standard error of

estimate yields a value of 27.88, i.e., the error in estimating the velocity from grain size, using the regression equation, results in an average velocity error of 27.88 m/s. The empirical relationships, proposed by Bachman (1985) and Hamilton (1982), and the obtained logarithmic regression line are similar with Hamilton's approximation showing a lower limit and Bachman's an upper limit.

The relationship of impedance and median grain size is similar to that of velocity and median grain size (Figure 13). Between 1 μm and 4 μm the trend is parallel to the impedance axis, indicating no distinct interdependency. Above a median grain size of 4 μm , a trend of increasing impedance with increasing median grain size is observed. The correlation coefficient ($R^2 = 0.290$) suggests considerable scatter around the logarithmic regression line. The analysis of the standard error of estimate results in 105.23 $\text{Mg/m}^2\text{s}$. The empirical relationship provided by Bachman (1985) falls below the logarithmic regression line.

Figure 14 shows the plot of compressional wave velocity versus the percentage of sand, silt, and clay constituents. The difference between fractions is adopted from the Udden-Wentworth size classification system for sediment grains. Therefore, the boundary between sand and silt is at 63 μm and all particles smaller than 2 μm are classified as clay. The boundary between clay versus silt may be physically better presented at 4 μm . However, data used here define less than the 2 μm size as the clay size. Linear regression lines fit to the populations of sand, silt, and clay, only show correlation with the clay fraction ($R^2 = 0.598$). The sand fraction shows small interdependency ($R^2 = 0.432$) while the silt fraction shows no correlation ($R^2 = 0.122$). The relationship is the same as the velocity correlations when described in terms of

impedance versus the percentage of sand, clay, and silt (Figure 15).

In Figure 16 and Figure 17, the sand and clay fraction are extracted from the data in Figure 14 and are plotted versus compressional wave velocity. In the plot of percentage clay fraction the earlier-derived empirical relationship (Hamilton and Bachman, 1982) is included. The slope of Hamilton's proposed relationship is lower than the slope of the regression line for the sediments used in this study. Noteworthy is the intersection of curves at 50% clay fraction. Also shown in Figure 16 are the prediction intervals (confidence 95%). The regression analysis for the data falling into this interval results in an improved regression coefficient of 0.881.

A closer look at the relationship between percent sand fraction and compressional wave velocity shows two trends (Figure 17). A sand fraction of approximately 15% appears to be the limit for the influence of the grain size on velocity. At lower sand contents, the data is scattered and shows no correlation with velocity. Above a sand fraction of 15% the velocity increases linear with the amount of sand in the sediment (solid line). However, the low correlation coefficient ($R^2 = 0.294$) suggest wide scatter around the linear regression. The regression coefficient is slightly improved where the data falls in the 95% confidence prediction interval.

Porosity is correlated with median grain size in Figure 18. In the fine-grained sediments (median grain size $< 4 \mu\text{m}$), porosity ranges between 0.4 and 0.8 and no distinct relationship is apparent. Above $4 \mu\text{m}$, an increase in median grain size is correlated with a decrease in porosity.

Calcareous Sediments: Plots of compressional wave velocity and impedance versus median grain size of the calcareous sediments from the Western Equatorial

Pacific (Ontong Java Plateau) are shown in Figure 19 and Figure 20. No distinct trend can be seen. The scatter around the regression line is too wide to predict any correlation. Contrary to the common trend of increasing velocity with increasing grain size, calcareous sediments seem to behave differently. The velocity is almost constant over the range of median grain sizes and a small decrease in velocity with increasing median grain size is indicated by the regression line. This trend is even more pronounced when comparing impedance instead of velocity with grain size.

Discussion

As the correlations show, the observed relationships between compressional wave velocity and grain size generally follow trends and prediction lines published by other authors (Bachman, 1985; Hamilton and Bachman, 1982). The discrepancy between the published regression lines (Bachman, 1985; Hamilton and Bachman, 1982) and those fit to the data (Figure 12, Figure 13) can be attributed to the difference in the parameter that describes the grain size characteristics, to the fact that the proposed relationships may not be applicable for a wide range of sediment types, or to measurement errors.

In this study, the grain size characteristic is expressed in terms of the median grain diameter which is defined as the 50% mark on the cumulative frequency curve (divides the normal frequency curve into two equal parts). The linear relationship, proposed by Hamilton and the polynomial relationship of Bachman, are based on the mean grain diameter after Folk and Ward (1957), which is defined as the mean of the average grain size of the coarsest fraction of the sediment, the finest fraction and the medium fraction and is expressed in ϕ -units. The difference in mean and median is

only of influence when the grain size distribution is strongly skewed or of bimodal character. However, since the evaluated regression curve lies between the proposed relationships it is assumed that the difference between mean and median grain size can be neglected.

Early efforts to determine the compressional wave velocity of sediments used small chunk samples. Methods such as the resonant chamber method (Toulis, 1956) suffered from the problem that the velocity could not be determined from undisturbed samples. Later, compressional wave velocity measurements were taken on chunk samples with an apparatus known as a Hamilton Frame (Boyce, 1976). These samples were visually undisturbed (e.g. undistorted bedding), but likely suffered from stress-relief and sample disturbance. Compressional wave velocity measurements on whole-core samples encased in the original sample liner are considered as relatively undisturbed.

Therefore, the offset of the three relationships may be attributed to a lack of general applicability or to differences in the measurement methods of velocity due to sample disturbance.

Comparing the porosity-grain size relationship (Figure 18) and the velocity-grain size relationship (Figure 12), interdependency is observed. In the fine-grained sediments (median grain size $< 4 \mu\text{m}$), the porosity and the grain size show no correlation. When exceeding a median grain size of $4 \mu\text{m}$ the porosity seems to be directly related to the grain size. The relationship between grain size and velocity shows the same behavior. In the fine-grained section, grain size changes have no influence on the velocity. Above $4 \mu\text{m}$, grain size shows a more distinct trend. Since

compressional wave velocity and porosity show similar trends when compared with grain size, it is apparent that grain size influences the velocity in terms of porosity. Porosity, in turn affects the bulk modulus and thus the velocity.

This behavior can probably be attributed to the mineral-grain structure of these sediments. Fine grained sediments (silt and clay) are apt to adhere together when deposited because of physico-chemical bonds (Mitchell, 1993). The structure formed is controlled by interparticle forces. In marine environments, the particles are in edge to face contact and form a three-dimensional porous structure (Figure 21d and f). When subjected to overburden pressure, the weak interparticle force can not retain this porous configuration. Therefore, the porosity change is predominantly caused by pressure and particle rearrangement. This behavior is also seen by an increase in bulk density with depth, linearly related to a decrease in porosity with depth (Figure 2 to Figure 8).

When the median grain size increases, the structure formed by the particles changes. Spaces are filled with larger grains (silt, sand) and mineral-grain structures such as in Figure 21c and e are formed. Since the grains have no grain-to-grain contact, the porosity is still dominated by pressure, but the influence of particle size slightly increases. Not until the particles have grain-to grain contact (Figure 21a and b), does the porosity form a strong dependency with grain size. An increase in pressure forces the particles to rearrange and the rearrangement or packing is dependent on the size of the grains. The amount of larger particles (sand) needed to influence the grain-to-grain structure, seems to be approximately 15% sand (Figure 17). At this percentage and above, an increasing velocity with percent sand is

observed. The correlation, observed in Figure 16 between velocity and percentage clay fraction affirms these findings. A decrease in the clay fraction is correlated with an increase in velocity.

The regression equations for the more distinct relationships between grain size and compressional velocity for the hemipelagic sediments are as follows:

- Velocity, V_p (m/s) versus median grain size, d_{50} (μm) (Figure 12):

$$V_p = 1494.38 + 26.72 * \ln(d_{50}) \quad (4)$$

- Velocity, V_p (m/s) versus clay fraction, C (%) (Figure 16):

$$V_p = 1653.38 - 2.69 * C \quad (5)$$

The lack of a relationship between median grain size and compressional wave velocity of the calcareous sediments from the Ontong Java Plateau may be explained by the nature of the calcareous particles. These particles have no physico-chemical bonds. The chamber-like structure of the foraminifers and the rigid frame of nannofossils (Figure 9), results in a very high porosity (50% - 70%). The high porosity results in low bulk moduli since water has a much lower bulk modulus (higher compressibility) than the mineral phase. The decrease in porosity due to an increase in the size of the grains that is shown to be an important factor for hemipelagic sediments seems to be of less influence when calcareous particles are abundant. The variations in porosity due to changes in grain size are limited and results in limited variations of velocity (Figure 19).

Conclusion and Recommendations

The observations made in this study on the relationships between compressional wave velocity and grain size generally follow findings of other investigators (Bachman, 1985; Hamilton and Bachman, 1982; Sutton *et al.*, 1957). The results are summarized as follows:

- (1) Grain size affects compressional wave velocity indirectly. Grain size is one of the controls of porosity. Porosity controls bulk modulus and thus compressional wave velocity.
- (2) In coarse-grained sediments, an increase in median grain size is directly associated with an increase in velocity. In fine-grained sediments, however, grain-size changes have little influence on the velocity. A median grain size of 4 μm and a sand content of 15% appear to be the limiting factors in the distinction between fine-grained behavior and coarse grained behavior.
- (3) The use of the percent clay (grain size $< 2 \mu\text{m}$) is a much better parameter for correlation with compressional wave velocity than the median diameter parameter.
- (4) When calcareous particles (nannofossils, foraminifers) are abundant in sediments, there is no relationship between velocity and grain size. The influence of the hollow structure of the particles on porosity and thus on bulk modulus and compressional wave velocity dominates. Therefore, the influence of grain size on the porosity, bulk modulus and velocity is too small to detect with standard methods.

In general, the influence of grain size on physical properties is highly complex. The graduations and classifications developed to characterize sediments in terms of the mineral grains (size, distribution) may not be adequate to be described by compressional wave velocity. On this account considerations should be made in further studies, to employ different characteristics or limits of grain size in the correlation with velocity (e.g. the fraction of particles smaller than 3 μm might be better approximated by velocity than the fraction of 2 μm particles).

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Core	Location	Water Depth (m)	Recovery (cm)
Central Arctic Ocean (96/09-1pc)	143°26'37"E, 86°24'52"N	927	270
Central Arctic Ocean (96/12-1pc)	144°46'22"E, 87°05'51"N	1003	1610
Central Arctic Ocean (96/13-1pc)	145°10'08"E, 87°09'12"N	978	1510
Western Equatorial Pacific (130/807A)	156°37'00"E, 3°36'00"N	2804	82290
Central Scotian Shelf (003-002)	63°02'02"W, 44°00'56"N	215	1610

Table 1: Core Locations.

Figure 1: Location and bathymetric map of the Central Arctic Ocean (Lomonosov Ridge).

Figure 2: Bulk density, compressional wave velocity and median grain size of Core 96/09-1pc (Central Arctic Ocean, Lomonosov Ridge).

Figure 3: Bulk density, compressional wave velocity and median grain size of Core 96/12-1pc (Central Arctic Ocean, Lomonosov Ridge).

Figure 4: Bulk density, compressional wave velocity and median grain size of Core 96/13-1pc (Central Arctic Ocean, Lomonosov Ridge).

Figure 5: Location and bathymetric map of the Central Scotian Shelf (Emerald Basin).

Figure 6: Bulk density, compressional wave velocity and median grain size of Core 87003-02 (Central Scotian Shelf, Emerald Basin).

Figure 7: Location and bathymetric map of the Western Equatorial Pacific (Ontong Java Plateau)

Figure 8: Bulk density, compressional wave velocity and median grain size of Core 807A (Western Equatorial Pacific, Ontong Java Plateau).

Figure 9: (a) Foraminifera (x120), (b) nannofossil (x400)

Figure 10: Plot of compressional wave velocity versus median grain size for sediments from the Central Arctic Ocean, the Central Scotian Shelf, and the Western Equatorial Pacific.

Figure 11: Plot of impedance versus median grain size for sediments from the Central Arctic Ocean, the Central Scotian Shelf, and the Western Equatorial Pacific.

Figure 12: Plot of compressional wave velocity versus median grain size for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 13: Plot of impedance versus median grain size for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 14: Plot of compressional wave velocity versus the sand, silt, and clay fraction for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 15: Plot of compressional wave velocity versus the sand, silt, and clay fraction for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 16: Plot of compressional wave velocity versus percent clay fraction for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 17: Plot of compressional wave velocity versus percent sand fraction for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 18: Plot fractional porosity versus median grain size for sediments from the Central Arctic Ocean and the Central Scotian Shelf.

Figure 19: Plot of compressional wave velocity versus median grain size for calcareous sediments of the Western Equatorial Pacific (Ontong Java).

Figure 20: Impedance versus median grain size for calcareous sediments of the Western Equatorial Pacific (Ontong Java).

Figure 21: Common mineral-grain structures of marine sediments. (a) single-grained structure (uniform sands); (b) mixed grain structure (sands, silty sands, sandy silts); (c) single-grained structure with platy minerals; (d) book-house structure (clays); (e) book-house structure, suspended with silt or sand (silty clays, clayey silts); (f) pelagic clay structure (Hamilton and Bachman, 1982).

Manuscript II

Resistivity-Porosity Relationships in Unconsolidated Marine Sediments

Introduction

Electrical resistivity measurements as a predictor of porosity are used extensively by the hydrocarbon exploration industry. Porosity is of fundamental interest in the exploration of hydrocarbon resources, since it affects the quality of both the source rock and the reservoir rock. In geotechnical engineering, the porosity is of primary interest in the prediction of soil and sediment behavior.

Archie (1942) successfully correlated porosities of reservoir rocks (sandstone) in sedimentary strata with electrical resistivity and developed an empirical relationship known as Archie's Law. Winsauer *et al.* (1952) extended Archie's Law to shale and developed the Humble Formula which is probably the most widely used equation in the hydrocarbon (well logging) industry.

Commonly, electrical resistivity is measured in situ by means of down-hole logging tools. With the development and introduction of smaller resistivity probes (Wenner spread, galvanic method) and the extension of investigation of the sea floor, marine scientists adopted the resistivity method to evaluate porosity of unconsolidated marine sediments (Boyce, 1968; Kermabon *et al.*, 1969).

However, large scatter is observed in the relationship of porosity and electrical resistivity of unconsolidated marine sediments. This suggests that Archie's Law, applied on unconsolidated marine sediments, may not be an effective predictor of

porosity.

The purpose of this work is to investigate the accuracy of using electrical resistivity to predict the porosity of unconsolidated marine sediments. Therefore, resistivity measurements of cores from three different geological marine environments were analyzed. Porosity was plotted versus resistivity to evaluate the coefficients used in Archie's Law. To improve predictions, other non-destructive measurements were incorporated: compressional wave velocity and bulk density. The results are compared with published relationships.

The core analyses were performed for each respective scientific project and not by the author, however, a description of the testing conditions and physical properties is provided.

Background

Electrical resistivity (reciprocal of electrical conductivity) is defined as the resistance between opposite faces of a sample of a given material to the flow of an electrical current. The resistance is a function of the material's resistivity and shape (length and cross-section). In porous media such as marine sediments the bulk resistivity is composed of the resistivity of the interstitial pore fluid and the resistivity of the mineral grains. In general, the mineral grains are assumed to be infinitely resistive. Since the resistance is also a function of the path of electrical flow around soil particles, the bulk resistivity of sediment samples yields information of the pore space occupied by the pore fluid.

Electrical resistivity measurements as a predictor for physical properties of sediments and rocks was first applied by Archie (1942). Archie introduced the

formation resistivity factor FF which is defined as the bulk resistivity R (ohm-m) divided by the resistivity of the interstitial pore fluid R_w (ohm-m) (Eq.(6)).

$$FF = \frac{R}{R_w} \quad (6)$$

From investigations on sandstone and clean sand, Archie also developed an empirical relationship that links the resistivity of the fluid saturated sandstone and the resistivity of the pore fluid (formation factor FF) to the fractional porosity. This relationship is referred to in the literature as Archie's Law:

$$FF = \phi^{-m} \quad (7)$$

with ϕ as fractional porosity (-) and m as a dimensionless factor that describes the nature of the sandstone. From his experimental work, Archie found m to be dependent on the degree of cementation of the individual grains and determined m to be in the range of 1.8 to 2.0 for sandstone and 1.3 for loosely consolidated sand.

Winsauer *et al.* (1952) introduced a more general form of the equation:

$$FF = a * \phi^{-m} \quad (8)$$

where a is a dimensionless coefficient between 0.6 and 2.0 depending on lithology and m is the cementation or tortuosity factor and ranges between 1.0 and 3.0. From Equation (8) the Humble Formula was developed which is probably the most widely used equation to evaluate porosity from resistivity. The a and m coefficient used in the Humble Formula are 0.62 and 2.15, respectively.

Atkins and Smith (1961) showed that systems of cohesionless particles obey Archie's Law, and that the magnitude of m depends on the shape of the individual particles, increasing as they become less spherical. They also demonstrated

theoretically that a combination of different particles (different shapes) leads to a relation in the form of Equation (8).

Even though Archie's Law and its derivations (Winsauer *et al.*, 1952) are based on the assumptions of infinite resistivity of the mineral grains, several other investigators have adapted the equations for sediments containing clay particles. Clay particles have negative charges and therefore the assumption of infinite resistivity of the mineral particles does not hold. However, Brace *et al.* (1965) showed in their research on fine-grained rocks of low porosity that the surface conductivity is negligible when the pore fluid resistivity has a magnitude similar to that of sea water.

Boyce (1968) investigated marine sediments of the Bering Sea and obtained values for the coefficients a and m of 1.30 and 1.45, respectively. Taylor-Smith (1971) used Archie's Law to describe his results for clay rich marines sediments and found m equal to 2.0 when the porosity is greater than 0.6. Kermabon *et al.* (1969) fitted a third degree polynomial curve to their data to approximate the porosity resistivity relationship.

A comprehensive review of parameters affecting the electrical resistivity of porous media was compiled by Dakhnov (1962). He summarized his findings in a general equation:

$$R = f_1(c) * f_2(\phi) * f_3(s) * f_4(T) * f_5(Q) * f_6(R_s) * f_7(R_w) \quad (9)$$

where

R = bulk resistivity of the sediment (ohm-m),

c = amount of clay and silt in the sediment,

ϕ = porosity of the sediment,

S = partial saturation of the sediment,

T = temperature,

Q = mineral cation exchange capacity,

R_s = sediment minerals resistivity, and

R_w =interstitial water resistivity.

Since some of these variables are sometimes omitted, the complexity of Equation (9) can be reduced. For example, when sand is used, Equation (9) can be simplified to

$$R = f_2(\phi) * f_7(R_w) \quad (10)$$

which is the general expression of Archie's Law.

Methods

Electrical Resistivity: Electrical resistivity was measured using a linear four-electrode-array (Wenner spread). The probes were pushed 2 mm into the split-core surface and an alternating current was applied to the two outer electrodes. The potential drop across the two inner probes was measured and converted to resistance by dividing by the instrument current. The resistivity was then obtained by multiplying the resistance by the instrument cell constant. The cell constant is defined as the cross-sectional area of the sediment divided by the distance of the two voltage electrodes and was determined by measuring the resistance of sea water (known resistivity) at a controlled temperature (Gerland *et al.*, 1992). The formation factor was calculated using Equation (6), with a pore fluid resistivity of 0.209 ohm-m (Dietrich *et al.*, 1989).

Compressional Wave Velocity. Compressional wave velocity was determined using a P-wave logger (PWL) on the Multi Sensor Track (GEOTEK). An ultrasonic pulse with a dominant frequency of 500 kHz was transmitted across the unopened core

sample and the travel time was measured. The velocity was then calculated by dividing the core diameter by the pulse travel time. Corrections for transducer and core liner time delays as well as for core diameter deviations were applied (APPENDIX A, Schultheiss and McPhail, 1989).

Bulk Density. Bulk density measurements were performed using a gamma-ray attenuation porosity evaluator (GRAPE) (Evans, 1965). The measurement of sediment bulk density using gamma-rays is based on the principles of Compton scattering and attenuation. A parallel, monoenergetic beam of gamma-rays (^{137}Cs) penetrates the core sample and is detected on the opposite side by a scintillation counter. When passing through the sample some of the gamma-rays are absorbed or scattered and lose energy and direction, respectively. The scintillation counter detects the gamma-rays that pass through the absorber without any loss of energy. The energy loss and the attenuation respectively are directly related to bulk density. A discrete value for bulk density is then derived by calibrating the attenuation of gamma-rays through the unopened core sample with the attenuation through standards of aluminum and water (APPENDIX A, Boyce, 1976).

Porosity: Porosity was determined from measurements of wet and dry sediment mass and wet sediment volume. Samples of approximately 10 cm^3 were taken from the sediment cores. Wet sediment mass was determined using an electronic balance. Wet sediment volume was calculated by means of a helium-displacement pycnometer. The dry sediment mass was obtained after oven-drying at 105° to 110°C for 12h to 24h and weighing on an electrical balance.

These measurements were used to calculate water content and bulk density

following the methods of the ASTM D 2216. All measurements were corrected for salt assuming a pore water salinity of 35‰. Porosity was calculated from bulk density and water content using the following equation:

$$\phi = \frac{(w * \rho_B)}{[(1 + w) * \rho_w]} \quad (11)$$

where ρ_B is bulk density (Mg/m^3), ρ_w is the density of pore fluid (Mg/m^3), and w is water content (-).

In cases where no discrete measurements of physical properties were performed, the porosity was calculated from GRAPE bulk density using an assumed specific gravity of 2.75:

$$\phi = \frac{\rho_S - \rho_B}{\rho_S - \rho_w} \quad (12)$$

where ρ_G is the specific gravity (Mg/m^3), ρ_B is the bulk density (Mg/m^3), and ρ_w is the pore fluid density (sea water $\rho_w = 1.024 \text{ Mg/m}^3$).

Site Descriptions and Sediment Properties

14 cores from three different sites were selected for this study. The major geographic environments are (1) the North Atlantic (Ceara Rise), (2) the Northeastern Pacific (Cascadia Margin) and (3) Caribbean Sea (Barbados Ridge). A summary of the cores used in this study and a detailed description of the location are given in Table 2.

North Atlantic (Ceara Rise): The Ceara Rise is located in the eastern equatorial North Atlantic between the longitude 42° to 45° West and the latitude 3° to 7° North (Figure 22). The Ceara Rise is a bathymetric high and reaches a minimum water depth of about 2600 m. The rise consists of a series of platform-shaped shoals oriented in a

northwest-southeast direction. It is bounded on the north-east and east by the Ceara Abyssal Plain and on the north, west, and southwest by the Amazon Fan with water depth of about 4500 m (Curry *et al.*, 1995).

The sediments recovered from the drilling sites along the Ceara Rise are predominately pelagic oozes, clays, chalks, claystones, and limestones. Calcareous nannofossils, foraminifers, and clay minerals are primary constituents. In general, the sediments can be divided into three major geotechnical units. Sediments of Unit I are dominated by grayish brown nannofossil clay with foraminifers alternating with light brownish gray, clayey nannofossil ooze. Unit II consists of light gray nannofossil ooze with varying amounts of clay alternating with grayish brown nannofossil ooze. The difference between Unit I and Unit II is based on the nearly 20% higher carbonate content in Unit II. Unit III consists of light greenish gray to greenish gray nannofossil chalk and limestone with variable amounts of foraminifers and clay minerals (Figure 23 to Figure 25).

Electrical resistivity increases with depth and is consistent with the index properties results of decreasing porosity with depth. Observed electrical resistivity values range from 0.1 Ωm to 0.4 Ωm with an average of 0.2 Ωm . Site 927 has the lowest resistivity values, followed by Site 925, and Site 926 exhibits the highest average resistivity. In the lower parts of core 154/925C, 154/925D, and 154/926D (Figure 23, Figure 24) a sharp increase in resistivity is observed. The resistivity increase is not associated with an increase in porosity and suggests changes in the microfabric of the sediments (cementation, tortuosity). The resistivity values of these intervals range between 0.4 Ωm and 1.3 Ωm . Porosity values range between 0.4 and

0.7 and except for minor scatter follow a decreasing linear trend (Figure 23 to Figure 25).

Northeastern Pacific (Cascadia Margin): The Cascadia Margin stretches along the western border of the North American continent for about 800 km from Vancouver Island, British Columbia in the North to Cape Blanco, Oregon in the South (Figure 26). The Margin breaks to the west into the Cascadia Basin with water depths of approximately 3000 m. Two large deep-sea fans (Nitinat and Astoria) are formed at the base of the Cascadia continental slope (Westbrook *et al.*, 1994).

The sediments of Cascadia Margin can be described as terrigenous and hemipelagic clays and silts. Three geotechnical units are used to distinguish the sediments. The recovered sediments of Unit I are predominately dark gray to dark olive gray silty clays and clayey silts with alternating thin layers of fine sand. Unit II is composed of firm clayey silts of dark gray to dark olive gray color. The sediments contain varying amounts of biogenic components (up to 10%). Unit II differs from Unit I in that Unit II has a lower abundance of sand and thus is more fine-grained than Unit I. Unit II and Unit III are identical in structure and composition. The difference in both is based on significant increases in glauconite in Unit III (Figure 28, Figure 29).

Electrical resistivity increases with depth and the observed values range from 0.5 Ωm to 2.4 Ωm . The increase in resistivity is associated with a decrease in porosity. Porosities range between 0.3 and 0.7 with most of the values occurring between 0.4 and 0.6.

Caribbean Sea (Barbados Ridge): The Barbados Ridge Accretionary Complex lies on the margin between the Caribbean Sea and the Atlantic Ocean, south and east of

Barbados Island. The complex narrows from approximately 300 km width in the south, east of Tobago, to less than 100 km width in the north, east of Barbuda (Figure 27).

The sediments recovered from the drilling sites along the Barbados Ridge are predominately calcareous oozes, calcareous clays, claystones, and limestones. Calcareous nannofossils, foraminifers, and clay minerals are the primary constituents. The sediments from the Barbados Ridge are divided into three geotechnical units. Unit I consists of homogeneous olive gray and brown calcareous clays and claystone interbedded with ash layers. The ash layers are dark gray in color and are highly bioturbated. Sediments of Unit II are dominated by olive gray claystones. Unit III is characterized by cyclic alterations of green to pale green, and olive gray calcareous claystone (Figure 29)

Electrical resistivity increases almost linear with depth. Values range from 0.3 Ωm to 1.5 Ωm . The increase in resistivity is consistent with a decrease in porosity which varies between 0.4 and 0.7.

Results

Plots of the formation factor versus measured values of porosity at each site are shown in Figure 30 to Figure 36. The empirical relationship between formation factor and porosity published by Boyce (1968) and Winsauer *et al.* (1952, Humble Formula) are also shown in the plots. Power law regression lines, following the form of Equation (8) (solid line), are fit to the data to obtain coefficients a (lithology) and m (cementation, tortuosity).

In general, the sediments of all sites follow the relationship of decreasing porosity

with increasing resistivity (Archie, 1942; Winsauer *et al.*, 1952). Exceptions to this general trend are Site 926 (Figure 31) and Site 891 (Figure 34). These sediments indicate no distinct increase in resistivity with decreasing porosity. The regression lines for these two are almost linear with a very small gradient and the regression line fit to the sediments from Site 926 actually show a decrease in resistivity with a decrease in porosity, indicated by the positive exponent. The observed values for the coefficients a and m range from 0.30 to 6.22 and -0.135 to 2.48, respectively (Table 3).

All data sets show wide scatter around the regression lines. Correlation, expressed in terms of the R^2 value, is best in sediments from Site 671 (Figure 36) and Site 925 (Figure 30) with values of 0.65 and 0.61, respectively. The lowest correlation coefficients are present in the sediments of Site 926 and 891 with values of 0.006 and 0.06. Values of the standard error, made by estimating the formation factor from porosity, vary between 0.138 for the sediments of Site 927 (Figure 32) and 1.478 for sediments of Site 891 (Figure 34). Reversing the relationship and calculating the error made by estimating porosity from the formation factor, the standard errors of estimate results in 0.16, 0.13 and 0.04 for sediments from the North Atlantic, Northeastern Pacific and Caribbean Sea, respectively.

Compared with the published relationship of Boyce (1968) and the Humble Formula, the sediments of the Ceara Rise (Site 925, 926, and 927) lie below these curves. While the sediments of the Northeastern Pacific (Site 889, 891, and 892) lie above, the sediments of the Caribbean Sea (Site 671) also lie above the published relationships but show the best correlation with these curves (Figure 36).

Figure 38 shows a plot of the formation factor evaluated from resistivity measurements versus a calculated formation factor using bulk density and compressional velocity ratio as a and m coefficients in Equation (8), and the measured porosity for sediments of the North Atlantic (Ceara Rise). Due to lack in data quality and quantity of compressional wave velocity and bulk density, sediments from the Caribbean Sea (Barbados Ridge) and Northeastern Pacific (Cascadia Margin) were excluded. The compressional wave velocity ratio is the ratio between the sediment and pore fluid (sea water) velocity. For this study a sea water velocity of 1522 m/s is assumed, which corresponds to a temperature of 20°C and a salinity of 35‰ (Chen and Millero, 1977).

In Figure 38, the coefficient a is approximated by bulk density and the coefficient m by the velocity ratio. A linear regression line, fit to the data, indicates a very low correlation ($R^2 = 0.079$).

Discussion

The general trend observed in the resistivity-porosity relationship of unconsolidated marine sediments is similar to published relationships. The a and m coefficients range between 0.30 to 6.22 and -0.135 to 2.48, respectively (Table 3). With the exception of the negative m value obtained from the sediments at Site 926, and a very high a value from Site 891 the values are generally in agreement with published data that range between 0.6 and 1.30 for a and 1.2 to 3.0 for m (Erchul, 1972; Jackson *et al.*, 1978).

The wide scatter around the regression lines indicates high variations in a (lithology) and m (tortuosity, cementation). Even in sediments of adjacent sites (Figure

30 to Figure 32) and with identical lithologies, major differences in these coefficients are apparent. In sands and sandstones where electrical resistivity is usually used to extract porosity, no such extensive scatter is observed (Jackson, 1975; Jackson *et al.*, 1978).

Reasons for the extensive scatter and the high variation in the coefficients a and m in unconsolidated marine sediments may be due to the textural properties of the sediments. When fine particles (clay, silt) are deposited on the sea floor, they adhere to each other by means of physico-chemical bonds (Mitchell, 1993). The mineral-grain structure that is formed is thereby highly random and irregular (high tortuosity, Figure 39d, e, f), even in intervals of similar sediment content. Electrical current is constrained to follow complex meandering paths whose lengths increase with tortuosity and whose cross-sectional areas (and hence resistance) vary erratically between the pores and fine interconnecting capillaries. Furthermore, clay particles can not be considered as infinitely resistive due to the negative charges on the particle surfaces. The path of the electrical current is therefore not only dependent on tortuosity, but also dependent on the conductivity of the particles.

Previous investigators show significant relationships between dielectric constant and porosity (Arulanandan, 1991; Smith and Arulanandan, 1981). However, these studies predominantly use uniform soils that are not representative of natural deposits.

When coarser particles are deposited, the mineral-grain structure is controlled by gravity and the structure is more regular than in fine grained sediments (Figure 39a, b, c). The tortuosity is therefore more dependent on the arrangement of the particles. Tortuosity variations within similar sediments are less for sands. Since coarser grains

are predominantly composed of quartz, they can be considered non-conductive and therefore, variations in resistivity due to the conductivity of particles are negligible.

Comparing the results of this study (e.g. a and m values) with published relationships, the porosity is not predicted well. The closest relationship is obtained between the equation published by Boyce (1968) and the sediments from Site 671. These Site 671 sediments (silty marine clays) are very similar to the sediments Boyce studied. In general, applying published relationships to sediments types that are different from those used to develop the relationships does not result in a good prediction of porosity. Even in very identical sediments (Site 671 and Boyce (1968)) the variations are too large for an adequate prediction of porosity (Figure 37).

The use of bulk density and compressional wave velocity (velocity ratio) as a first approximation of the coefficients a and m results in weak correlations (Figure 38) as indicated by the low R^2 values of 0.079.

Yet, the approach seems to be reasonable if one considers the characteristics of a and m . The coefficient a reflects the lithology and a change in lithology is usually associated with a change in bulk density. The coefficient m incorporates cementation and tortuosity (increase in cementation and tortuosity, increases the factor m). These two parameters also influence the compressional wave velocity. An increase in cementation increases the bulk modulus and the shear modulus of sediments. Since these two parameters are interdependent with velocity, the velocity would also increase. However, the relationship in Figure 38 indicates no interdependency between a , m , bulk density and velocity ratio.

Conclusion

The applicability of electrical resistivity to extract porosity of clean sands, sandstones, and limestones is well demonstrated (Archie, 1942; Winsauer *et al.*, 1952). It is also common practice in the hydrocarbon exploration industry to use published charts of formation factor to predict porosity of source and reservoir rocks (Figure 40). However, the results of this study indicate that the resistivity–porosity relationship in unconsolidated fine-grained sediments is much more complex. It may not be possible to develop similar chart-type relationships to predict porosity for these sediment types. The wide scatter of data around regression lines results in poor approximations of porosity, even when the a and m coefficients are known (i.e. evaluated by means of discrete porosity measurements). However, as a starting point, the high porosity fine-grained data presented in this study are presented in chart form (Figure 41).

Compressional wave velocity and bulk density show no relationship with either coefficient a or m . However, changes in velocity and density may be used to identify changes in lithology, cementation and tortuosity. This could be used to group data of similar texture and lithology to apply appropriate a or m values and to refine charts that can be used to predict porosity.

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Core	Location	Water Depth (m)	Recovery (m)
North Atlantic (Leg 154, 925C)	43°29'21"W, 4°12'15"N	3052	369
North Atlantic (Leg 154, 925D)	43°29'21"W, 4°12'15"N	3052	364
North Atlantic (Leg 154, 925E)	43°29'20"W, 4°12'15"N	3053	55.7
North Atlantic (Leg 154, 926A)	42°54'29"W, 3°43'09"N	3610	335
North Atlantic (Leg 154, 926B)	42°54'30"W, 3°43'09"N	3610	593
North Atlantic (Leg 154, 926C)	42°54'30"W, 3°43'08"N	3610	394
North Atlantic (Leg 154, 927A)	44°28'50"W, 5°27'46"N	3325	316
North Atlantic (Leg 154, 927B)	44°28'50"W, 5°27'46"N	3327	268
North Atlantic (Leg 154, 927C)	44°28'50"W, 5°27'46"N	3328	263
North-Eastern Pacific (Leg 146, 889A)	126°52'05"W, 48°41'57"N	1322	223
North-Eastern Pacific (Leg 146, 891B)	125°19'33"W, 44°38'38"N	2674	54
North-Eastern Pacific (Leg 146, 892A)	125°07'08"W, 44°40'26"N	686	64
North-Eastern Pacific (Leg 146, 892D)	125°07'08"W, 44°40'26"N	686	76
Caribbean Sea 110, 671B)	58°43'57"W, 15°31'33"N	4915	562

Table 2: Core Locations

Figure 22: Location and bathymetric map of the North Atlantic (Ceara Rise).

Figure 23: Plots of electrical resistivity and porosity versus depth for sediments from Site 925 (North Atlantic, Ceara Rise)

Figure 24: Plots of electrical resistivity and porosity versus depth for sediments from Site 926 (North Atlantic, Ceara Rise)

Figure 25: Plots of electrical resistivity and porosity versus depth for sediments from Site 927 (North Atlantic, Ceara Rise)

Figure 26: Location and bathymetric map of the Northeastern Pacific (Cascadia Margin).

Figure 27: Location and bathymetric map of the Caribbean Sea (Barbados Ridge).

Figure 28: Plots of electrical resistivity and porosity versus depth for sediments from Site 889 and 891 (Northeastern Pacific, Cascadia Margin)

Figure 29: Plots of electrical resistivity and porosity versus depth for sediments from Site 892 (Northeastern Pacific, Cascadia Margin) and Site 671 (Caribbean Sea, Barbados Ridge).

Figure 30: Plot of formation factor versus porosity (Site 925, North Atlantic, Ceara Rise).

Figure 31: Plot of formation factor versus porosity (Site 926, North Atlantic, Ceara Rise).

Figure 32: Plot of formation factor versus porosity (Site 927, North Atlantic, Ceara Rise).

Figure 33: Plot of formation factor versus porosity (Site 889, Northeastern Pacific, Cascadia Margin).

Figure 34: Plot of formation factor versus porosity (Site 891, Northeastern Pacific, Cascadia Margin).

Figure 35: Plot of formation factor versus porosity (Site 892, Northeastern Pacific, Cascadia Margin).

Figure 36: Plot of formation factor versus porosity (Site 671, Caribbean Sea, Barbados Ridge).

Site	Coefficient a	Coefficient m	R^2
North Atlantic, Site 925 (Figure 30)	0.30	2.48	0.61
North Atlantic, Site 926 (Figure 31)	1.48	-0.14	0.01
North Atlantic, Site 927 (Figure 32)	0.30	1.61	0.30
North Atlantic, All Sites	0.45	1.61	0.17
Northeastern Pacific, Site 889 (Figure 33)	1.71	2.05	0.50
Northeastern Pacific, Site 891 (Figure 34)	6.22	2.29	0.02
Northeastern Pacific, Site 892 (Figure 35)	2.31	1.21	0.52
Northeastern Pacific	3.48	0.90	0.37
Caribbean Sea, Site 671 (Figure 36)	0.94	2.41	0.65
Boyce (1968)	1.30	1.45	--
Humble Formula	0.62	2.15	--

Table 3: Summary of the coefficients a (lithology) and m (cementation, tortuosity).

Figure 37: Plot of measured porosity and calculated porosity (Boyce, 1968) versus depth for (Caribbean Sea, Barbados Ridge)

Figure 38: Plot of formation factor versus calculated formation factor (a = bulk density, m = velocity ratio; North Atlantic, Ceara Rise).

Figure 39: Common mineral-grain structures of marine sediments. (a) single-grained structure (uniform sands); (b) mixed grain structure (sands, silty sands, sandy silts); (c) single-grained structure with platy minerals; (d) book-house structure (clays); (e) book-house structure, suspended with silt or sand (silty clays, clayey silts); (f) pelagic clay structure.

Figure 40: Plot of formation factor versus porosity commonly used in the hydrocarbon industry to predict porosity (courtesy Schlumberger©). The proper choice is best determined by laboratory measurements or experiences in the area. In the absence of this knowledge, for hard formations $a \neq 1.0$ is recommended and for soft formations $a = 1.0$ with appropriate cementation factor, m .

Figure 41: Plot of formation factor versus porosity of the data presented in this study. Each site represents a different sediment type: pelagic clays with high carbonate content (North Atlantic, Ceara Rise), hemipelagic and terrigenous clays and silts with low carbonate content (Northeastern Pacific, Cascadia Margin), and pelagic clay with low carbonate content (Caribbean Sea, Barbados Ridge).

Manuscript III

Compressional Wave Velocity and Bulk Density in the Extraction of Shear Strength Characteristics

Introduction

The shear strength and the shear modulus of soils and sediments are two important parameters used in geotechnical engineering. The bearing capacity of foundations, slope stability, and soil retaining structures are all affected by the shear strength of the soil in the immediate vicinity of the structure. The initial shear modulus is also a key parameter for small strain dynamic analyses and can be used as a predictor of soil behavior during cyclic loading caused by wind or waves, earthquakes, explosions, and machine or traffic vibrations.

The proper evaluation of these properties in the laboratory is labor and cost intensive. The quantity of information obtained is limited because of the effort required to acquire data information on how shear strength and shear modulus vary with depth and sediment type. For regional studies this limited amount of information makes it necessary to interpolate or extrapolate between data points to get an area wide profile that can be used to make assumptions for design. This extrapolation of information can lead to extensive errors and misinterpretations, since soils and sediments are inhomogeneous, non-linear, and non-uniform. In geotechnical engineering, uncertainties and errors caused by lack of information and extrapolation are compensated by the use of the factor of safety approach and consequently, an

underestimation of the capacity of a structure. Methods to improve the quantity and quality of such data would reduce uncertainty and risk. Thus, this study's focus is the investigation of non-destructive methods for assessment of shear strength properties, to improve data quantity and quality at a relative low cost. The ability to extract sediment shear characteristics, such as the shear strength and the shear modulus non-destructively, by measurements of compressional wave (P-wave) velocity and bulk density would lead to high resolution profiles and ultimately to a better estimation of design parameters.

Compressional wave velocity and bulk density can readily be measured by means of a P-wave logger (PWL) and gamma-ray attenuation porosity evaluator (GRAPE), respectively. Both sensors are incorporated into the Multi-Sensor-Track (MST) and are part of a standard instrument configuration (APPENDIX A). This state-of-the-art device is fully automated and able to measure these parameters at high resolutions (less than 1 cm intervals) along a sediment core. Providing a relationship between nondestructive measurements such as these and shear strength could potentially improve the design and the safety of structures. Furthermore, the influence of sample disturbance on the parameters is reduced, since the MST measures relatively undisturbed whole-core samples encased in the original sample liner.

The main objective of this study is to develop empirical relationships between non-destructive measurements of P-wave velocity, bulk density and discrete measurements of undrained shear strength or shear modulus evaluated by miniature vane shear tests. Although the provided relationship is completely based on statistical and empirical approaches, a discussion of the theoretical significance of the results is also included.

The sediments used for the empirical correlation in this study are from three different geographic marine environments: (1) Central Arctic Ocean (Lomonosov Ridge; (2) Central Scotian Shelf (Emerald Basin); and (3) Gulf of Mexico (Texas-Louisiana Slope and Rise). The core analyses were performed for each respective scientific project and not by the author, however, a sufficient description of the testing conditions and physical properties is provided.

Background

Several empirical studies have been published, investigating the relationship between compressional wave velocity of marine sediments and undrained shear strength (Hamilton, 1970; Horn *et al.*, 1968). Horn *et al.* (1968) analyzed marine sediments of 22 cores from the Norwegian Basin and the Mediterranean Sea and correlated compressional wave velocity to undrained shear strength measurements evaluated by means of a fall-cone penetrometer and concluded that the undrained shear strength is not a reliable predictor for compressional wave velocity. The analyzed data fell into distinct groups of sediments with similar grain size and sediment genesis (deep sea mud and clay, turbidities and ash) and an over-all increase in velocity with increasing shear strength was observed in all sediment groups, except for the fine-grained pyroclastic sediments (ash) where the velocity seemed to be independent of shear strength.

Hamilton (1970) made a similar correlation with sediments from major physiographic provinces in the North Pacific Ocean and came to a similar conclusion. Even though the sediments used in the study covered a wide range of different types and shear strength values, no distinct correlation was observed. Hamilton based the

correlation on the assumption of increasing velocity with increasing rigidity (i.e. shear modulus), and since the cohesion (i.e. undrained shear strength) is a measure of rigidity, an increase of velocity should result in an increase in shear strength. As an explanation for the high scatter of the data, Hamilton suggested that the monotonic undrained shear strength cannot be compared with the dynamic rigidity and that the influence of the rigidity on the velocity is marginal in marine sediments.

Despite these results (Hamilton, 1970; Horn *et al.*, 1968; Schreiber, 1968), an evaluation of shear strength or shear modulus from miniature vane shear tests (MV) and non-destructively derived compressional wave velocity and bulk density was conducted by following a different approach. The basis of this study differs from former studies, in that the velocity is not directly compared with shear strength parameters (peak shear strength and shear modulus) but rather with velocity in conjunction with bulk density because both parameters are related to moduli. In addition, these earlier studies were not non-destructive; probes were used or samples taken to measure velocity and density.

Theoretically the relationships among shear strength, compressional wave velocity and bulk density are based on the basic equation for the velocity of an acoustic compressional wave:

$$V_p^2 * \rho_B = B + \frac{4}{3} * G \quad (13)$$

where V_p is the compressional wave velocity (m/s), B is the bulk modulus (KN/m²), G is the shear modulus (KN/m²), and ρ_B is the bulk density (Mg/m³). Since shear strength depends on shear modulus, these are directly related. Cadling *et al.* (1950) developed a relationship to estimate the shear modulus, G , from the initial slope of a

miniature vane stress versus strain (rotation) data plot as follows:

$$G = \frac{\tau_0}{2 * \Theta_0} \quad (14)$$

where τ_0 is the shear stress (KN/m²) and Θ_0 is the angle of vane rotation (°). Since Equations (13) and (14) are functions of the shear modulus, it is reasonable to assume a relation between the two. The shear modulus derived from Eq. (13) is a dynamic shear modulus at very small strains (< 0.001%) whereas the shear modulus using Eq. (14) is large strain caused by the rotation of a four bladed vane in an assumed linear elastic medium. The assumptions used to evaluate the shear modulus from Eq. (14) are:

- (1) the failure surface is fully mobilized and in the same shape and size as the cylinder formed by the vane rotation;
- (2) the disturbance of the stress distribution due to insertion of the vane is negligible;
- (3) drainage and progressive failure do not occur in significant magnitude during the test;
- (4) the material is isotropic and infinite in all directions; and
- (5) Hooke's law is valid.

Although, the vane shear test is a "destructive" one, resulting in a measure of large strain behavior, it is assumed that there is a relationship between small strain and large strain behavior. The initial slope of the vane shear stress - rotation curve expresses the shear properties of the sediment in terms of large strain (static) with a different absolute value than the small strain (dynamic) shear modulus that is obtained using , for example, resonant column (Kramer, 1996) or shear wave velocity measurements

(Hamilton *et al.*, 1970). Therefore, the relationship between dynamic shear modulus in terms of the factor $V_p^2 * \rho_B$ (Eq.(13)) and the vane derived shear properties was investigated. For convenience, the $V_p^2 * \rho_B$, an elastic parameter, is defined here as Ξ , as an expression for the elastic behavior of sediments in terms of bulk modulus and shear modulus. To verify the assumption that Ξ can be used to express G , Ξ is correlated with an empirically derived dynamic shear modulus (Jamiolkowski *et al.*, 1991).

Methods

Compressional Wave Velocity. Compressional wave velocity was determined using a P-wave logger (PWL) on the Multi Sensor Track (GEOTEK). An ultrasonic pulse with a dominant frequency of 500 kHz was transmitted across the unopened core sample and the travel time was measured. The velocity was then calculated by dividing the core diameter by the pulse travel time. Corrections for transducer and core liner time delays as well as for core diameter deviations were applied (APPENDIX A, Schultheiss and McPhail, 1989).

Bulk Density. Bulk density measurements were performed using a gamma-ray attenuation porosity evaluator (GRAPE) (Evans, 1965). The measurement of sediment bulk density using gamma-rays is based on the principles of Compton scattering and attenuation. A parallel, monoenergetic beam of gamma-rays (^{137}Cs) penetrates the core sample and is detected on the opposite side by a scintillation counter. When passing through the sample some of the gamma-rays are absorbed or scattered and lose energy and direction, respectively. The scintillation counter detects the gamma-rays that pass through the absorber without any loss of energy. The energy loss and the attenuation

respectively are directly related to bulk density. A discrete value for bulk density is then derived by calibrating the attenuation of gamma-rays through the unopened core sample with the attenuation through standards of aluminum and water (APPENDIX A, Boyce, 1976).

Undrained Shear Strength. The undrained shear strength of the sediments was determined at selected intervals using a motorized miniature vane shear device (MV). The vane blade torque was measured either using an electronic torque transducer or linear springs. A four bladed vane with a geometry ratio of 1 (the height is equal to the diameter) was used and the measurements were performed in accordance with ASTM D 4648-94.

Undrained Shear Modulus (G_{MV}). The maximum undrained shear modulus G_{MV} was evaluated using the initial slope of the shear stress - rotation plot derived from miniature vane shear measurements. The maximum shear modulus occurs at strains lower than 0.001% and decreases after exceeding this limit (Kramer, 1996). Conversion from angle of vane rotation was calculated following Equation (14);

$$G = \frac{\tau_0}{2 * \Theta_0} = \frac{\tau_0}{\gamma_0} \Leftrightarrow 2 * \Theta_0 = \gamma_0 \quad (15)$$

where τ_0 is the shear stress (KN/m²), Θ_0 is the angle of vane rotation (°), and γ_0 is the shear strain (%). The angle of rotation that is equivalent to a shear strain of 0.001 % results in a 0.0005° vane rotation. Due to the low resolution of the shear stress – rotation plot of approximately one degree, a best fit non-linear regression line, based on the natural logarithm with two parameters, was fit to the data estimate the shear strength at 0.0005° (Figure 42). The shear modulus G_{MV} was then calculated from the equation of Cadling *et al.* (1950).

Dynamic Shear Modulus (G_{MAX}). The dynamic shear modulus G_{MAX} was approximated using the empirical equation proposed by Jamiolkowski *et al.* (1991);

$$G_{MAX} = \frac{198}{e^{1.3}} * OCR^k * \sqrt{p'} \quad (16)$$

where G_{MAX} is the maximum shear modulus (initial slope of the shear stress – strain curve, MN/m²), e is the void ratio (-), OCR is the over-consolidation ratio (-), k is the over-consolidation exponent as a function of plasticity index (Hardin and Drnevich, 1972), and p' is the mean effective pressure (MN/m²). The void ratio e was determined from the bulk density using an estimated specific gravity of 2.75 with Eq. (17) and Eq. (18):

$$\phi = \frac{\rho_s - \rho_B}{\rho_s - \rho_F} \quad (17)$$

$$e = \frac{\phi}{1 - \phi} \quad (18)$$

where ϕ is the porosity (-), ρ_s is the grain density (Mg/m³), ρ_B is the bulk density (Mg/m³), and ρ_F is the pore fluid density (Mg/m³) (density of sea water, $\rho_{sw} = 1.025$ Mg/m³). The over-consolidation ratio was assumed to be 1.0 for normally consolidated sediments. For over-consolidated sediments the normalized vane shear strength was used in conjunction with published charts that approximate the over-consolidation ratio (Bradshaw, 1999). The plasticity index was derived from normalized vane shear strength (Lambe and Whitman, 1969). Mean effective pressure is calculated using Equation (19);

$$p' = \frac{\sigma'_1 + \sigma'_2 + \sigma'_3}{3} \quad \text{with} \quad \sigma'_2 = K_0 * \sigma'_1 = \sigma'_3 \quad (19)$$

σ' are the principle effective effective stresses in each direction (MN/m²) and K_0 is the

earth pressure coefficient at rest. An assumption that $K_0 = 0.5$ simplifies Eq.(20) to;

$$p' = \frac{2 * \sigma_1'}{3} \quad (20)$$

where the major principle stress (overburden) was derived from the bulk density.

Site Description and Sediment Properties

Four cores from three different sites were selected for this study. The major geographic environments are: (1) the Central Arctic Ocean (Lomonosov Ridge), (2) the Scotian Shelf (Emerald Basin), and (3) the Gulf of Mexico (Texas-Louisiana Slope and Rise). A summary of the cores used in this study and a detailed description of the locations are shown in Table 4.

Central Arctic Ocean (Lomonosov Ridge). The Lomonosov Ridge is located in the Central Arctic Ocean between the longitude 130° to 155° East and the latitude 85° to 90° North (Figure 43). The Lomonosov Ridge separates the Markarov Basin and the Amundsen Basin. The ridge crest is at its highest less than 1000 m below sea level and drops down on both sides into the adjacent basins with depth of more than 3000 meters below sea level. The sediments of the Lomonosov Ridge are predominantly hemipelagic with minor ice rafting components.

The sediments of Core 96/09-1pc are described as clays and silty clays and the lithology was divided into three geotechnical units (Jakobsson *et al.*, 2001). Unit I consists of a thin layer of dark brown clay at the surface and is underlain by a layer of yellowish brown to dark gray silty clay. The second Unit II is composed of olive gray clay and silty clay and unit III is an indurated dark olive gray silty clay (Figure 44). The bulk density of Unit I and Unit III increases with depth. Values range from 1.80

Mg/m³ in Unit I to 2.06 Mg/m³ in Unit III. The bulk density of Unit II is lower than the bulk density of the other two units. Bulk density of Unit II is alternating between maxima of 1.85 Mg/m³ and minima of 1.60 Mg/m³.

The compressional wave velocity follows the trend of the bulk density. The velocity in Unit I is increasing with depth and does not have much scatter. The velocity trend of Unit I is continued in Unit III. Values of compressional velocity range from 1450 m/s to 1590 m/s. The velocity of Unit II is much lower and is nearly constant with depth. The average velocity value for Unit II is about 1460 m/s.

The shear strength of Unit I and Unit II is relatively low. Values range between 7.4 kPa and 21.3 kPa and are increasing with depth. The sediments in Unit I and II can be considered as normally consolidated sediments. In comparison to that, Unit III is characterized by a steep increase in shear strength from 16.6 to 95.0 kPa and then increases with depth, and with little scatter to a maximum of 170.0 kPa. The sediments of Unit III are interpreted as over-consolidated due to the high increase in shear strength and the quasi-constant bulk density.

Central Scotian Shelf (Emerald Basin). The Emerald Basin is a 2430 km² depression located on the central Scotian Shelf approximately 40 km off the coast of Nova Scotia and reaches its maximum depth at 290 m below sea level (Figure 45). The Basin is filled with glacial till that is overlain by Quaternary fine grained glacio-marine and marine sediments and underlain by firm bedrock. In general the sediment stratigraphy can be divided into three major geotechnical units (Figure 46). However, the sediments retrieved with core 87003-02 are only composed of the upper two units. Unit I consists predominantly of marine silty clay and clayey silt, olive gray in color.

Unit II is mainly dark gray glacio-marine silty clay (Moran *et al.*, 1991). The marine silty clay and clayey silts have relatively low shear strength and a high plasticity index. Shear strength increases with depth except for several anomalous peaks that are probably measurement errors. Bulk density is uniformly increasing with depth and no distinct variation in density due to the differences in composition between Unit I and II can be observed. Density ranges from 1.40 Mg/m³ at the top of the core to 1.6 Mg/m³ at a depth of 16.5 meters. The compressional wave velocity is relative constant with depth and has an average value of 1446 m/s.

Gulf of Mexico (Texas-Louisiana Slope and Rise). The Texas-Louisiana slope and rise encompasses approximately 120,000 km² (Figure 47). In general the sediments can be divided into four major geotechnical units. Core JPC 32 contains sediments of all four units whereas Unit IV in core JPC 41 is missing (Figure 48). Unit I consist of brown to light brownish gray silty clays with some minor amounts of carbonate sands. Unit II is composed of olive gray, grayish brown and dark gray silty clays. Unit III is brownish gray and light olive gray silty clays with significant amounts of foraminifera and unit IV is a stratified deposit consisting of thick layers of reddish silty clays and dark gray silty clays (Bradshaw, 1999).

Unit I of core JPC 32 has a relatively constant bulk density with depth of about 1.48 Mg/m³. At the unit break the density drops to 1.60 Mg/m³ and increases to 1.68 at the break to Unit III. Between Unit II and III a density drop occurs, again. The trend of the density in Unit III and IV is identical and increases with depth from 1.48 Mg/m³ to 1.68 Mg/m³. The velocity of core JPC 32 is uniformly increasing with depth from 1470 m/s to 1520 m/s. The shear strength of Unit I, II, and III is linear increasing with

depth from 2.8 kPa to 16.8 kPa. At the break between Unit III and Unit IV the shear strength drops to a value of 11 kPa and then increases linearly with depth to a value of 19.0 kPa.

Properties of Core JPC 41 differ from that of Core JPC 32. The density in Unit I and II is almost constant at 1.4 Mg/m^3 . At the unit break between II and III the density increases to a peak value of 1.58 Mg/m^3 and drops just after that to a minimum value of 1.32 Mg/m^3 . Below this interval, the density increases uniformly with depth to a maximum of 1.8 Mg/m^3 . The velocity profile follows the trend of the bulk density with values ranging from 1480 m/s at the top of the core to 1550 m/s at a depth of 7.1 meter. The shear strength is slightly increasing with depth in Unit I, II, and parts of Unit III. At the middle of Unit III, a steep increase in shear strength is observed and below this depth the trend of the shear strength is similar to the upper part of the core. Values for the shear strength range in the upper part of the core between 2.5 kPa and 7.0 kPa and after the step-wise increase from 29.3 kPa to 40.75 kPa (Figure 49). The sediments of the upper part of the core are interpreted as normally consolidated and in the lower part (depth $> 5.20 \text{ m}$) overconsolidated.

Results

Large Strain and Small Strain Shear Modulus. The large strain shear modulus G_{MV} derived from the initial slope of the shear stress – rotation plot is compared with the small strain shear modulus G_{MAX} , calculated using Equation (16) and the elastic parameter Ξ (Figure 50 and Figure 51). Both plots show scatter in the observed values. Scatter in the large strain shear modulus is wider for the elastic parameter Ξ than for the small strain shear modulus G_{MAX} . However, the difference in both plots is, if

expressed in terms of the correlation coefficient, very small ($R^2 = 0.461$ and 0.338). In general, the vane derived small strain shear modulus (G_{MV}) is much smaller than the calculated small strain shear modulus (G_{MAX}), by two orders of magnitude. Values in the higher shear modulus range seem to be more scattered than in the lower range.

Figure 52 shows plots of large strain and small strain shear modulus (a), small strain shear modulus and elastic parameter (b), and large strain shear modulus and elastic parameter (c) versus depth. A strong relationship between the G_{MAX} and Ξ is indicated by a nearly parallel depth trend (Figure 52b). The figures on the left and on the right (Figure 52a and c) represent the same data as Figure 50 and Figure 51. While the scatter plots show low R^2 values, the plots versus depth of these values suggest an interdependency, except, at a depth of 1.80 m (high shear modulus range) where there is increasing scatter of G_{MV} .

Undrained Shear Strength and Small Strain Shear Modulus. A strong relation is apparent between the undrained shear strength S_u and the small strain shear modulus G_{MAX} (Figure 53). The values of the sediments from Core JPC 41, JPC 32 (Gulf of Mexico) and 87003-02 (Emerald Basin) follow a linear trend where an increase in shear strength is associated with an increase in shear modulus. However, the sediments of Core 96/09-1pc (Lomonosov Ridge) seem to follow a different trend. In the Lomonosov Ridge sediments, the normally consolidated samples (low shear strength) follow the trend of the other sites, while the over-consolidated sediments (high shear strength) cause a steep increase of the regression line, i.e. the increase in shear strength is not associated with an appropriate increase in shear modulus. Apparently, the over-consolidated sediments from the Arctic behave differently from

the normally consolidated sediments or the offset can be ascribed to measurement derived errors. Since the over-consolidated values of Core JPC 41 (Gulf of Mexico) are within the limits of the general trend, it is more likely an artifact of errors in data acquisition. This is also confirmed by the unusual high ratio of undrained shear strength over the effective overburden pressure that exceeds absolute values of 5.0. This is much higher than values published for similar sediments (Bradshaw, 1999; Lambe and Whitman, 1969). On this account, values exceeding shear strengths of 50 kPa are considered carefully and additional correlations are provided that excludes these values. Figure 54 shows the correlation without the anomalous shear strength values of the over-consolidated Lomonosov Ridge sediments. The values follow a strong linear trend, with a correlation coefficient of 0.733. Also shown in this figure are the 95% confidence prediction intervals. A regression analysis for the data falling in this interval yields a correlation coefficient of 0.769. Noteworthy are the values in the upper part of the plot representing the over-consolidated sediments of Core JPC 41 that diverge from the linear regression line.

Undrained Shear Strength and Elastic Parameter Ξ The relationship between undrained shear strength S_u and the elastic parameter Ξ is almost identical to the relationship between undrained shear strength and small strain shear modulus (Figure 55). However, the correlation in terms of coefficient R^2 is not as pronounced. In Figure 56 the over-consolidated sediments of the Arctic are removed. The scatter around the regression line is much wider than in previous correlation with the shear modulus. Two data populations are noticeable. Above the regression line the over-consolidated data from Core JPC 41 and below the regression curve on the right side

the normally consolidated sediment from the Arctic seem to follow different trends.

Small Strain Shear Modulus and Elastic Parameter Ξ : Figure 57 shows the plot of the elastic parameter Ξ and the calculated small strain shear modulus G_{MAX} . Since Ξ is an expression of the elastic properties of the sediments, in terms of bulk modulus and shear modulus, a good correlation between these parameters is obvious. Sediments of the different cores are all following a linear relationship (with slightly varying slopes and intercepts) with increased shear modulus associated with an increase in the elastic parameter Ξ . Noteworthy is that the regression lines of Core JPC 41 and 96/09-1pc follow the same trend, even though they are slightly offset. Core JPC 32 and 87003-02 also follow this trend, but offset from the other.

Discussion

The correlation between the large strain shear modulus G_{MV} and the small strain shear modulus G_{MAX} and the elastic parameter Ξ , respectively, yield a low coefficient correlation relationship, but show a trend of interdependence. Scatter is too wide to be able to use the regressions equation as predictors of large strain shear behavior. Reasons for the data scatter could range from variable stress history to small scale sedimentological differences to measurement errors.

The most probable factor is the low measurement resolution of the shear stress – rotation (strain) plots. A resolution of 1° vane rotation is too large for evaluation of the initial slope (shear modulus) of the curve. To obtain a value for the shear modulus, the initial part of the shear stress rotation curve was approximated by a best fit nonlinear regression line. This procedure is based on a subjective extraction of points on the curve, used to fit the regression line into the shear stress – rotation plot. Small changes

to the input parameter slightly causes an overall change of the regression line and thus a change in the initial slope of the curve. Additionally, minor problems with the vane shear device during data acquisition, such as slipping of the driving belt, resulted in a non-uniform curve and complicated the approximation of shear modulus (Figure 42).

In general, the vane-derived shear modulus is much lower than the small strain shear modulus. In addition to the low resolution shear stress – rotation plots, reasons for this difference could be attributed to the vane shear procedure. Assumptions made with regard to the absence of sediment disturbance (see assumption (2) in the background chapter), to evaluate the shear modulus and shear strength from vane shear tests (Cadling and Odenstad, 1950), are not met with the first insertion of the vane into the sediment. The vane insertion forces the soil to displace, and the strains caused by the insertion exceed 0.001%. This error affects the shear modulus to a much higher degree than the shear strength, since the shear modulus is a measure at very small strains ($< 0.001\%$) whereas the strength is measured at high strain.

In spite of the scatter and associated errors, general trends exist. The general trend of increasing large strain shear modulus G_{MV} with increasing small strain shear modulus is observed, despite the data quality. Another indication of a relationship between large strain and small strain shear modulus or the elastic parameter is seen in Figure 52. All three parameters, compared with each other, show an identical trend with depth.

The correlation between the undrained shear strength and the small strain shear modulus results in a very distinct linear relationship. The resulting regression coefficient R^2 of 0.733 is good, particularly given the large variation in sediments

studied. The scatter of the observed values around the regression might be explained by errors in measurements of the shear strength or by errors in the calculation of the shear modulus. However, it is more probable that the error is made in the evaluation of the shear modulus, since the equation that is employed is empirical and factors such as the over consolidation ratio and the plasticity index are based on empirical approaches (Bradshaw, 1999; Lambe and Whitman, 1969).

The influence of over-consolidation is demonstrated by the wide scatter around the regression line that can be observed for the over-consolidated sediments of Core JPC 41 (Figure 54). The over-consolidation ratio of normally consolidated sediment is one and, therefore, the plasticity index that is represented in the over-consolidation ratio exponent, does not influence the equation. This is seen in the small data scatter of normally consolidated sediments at moderate shear strength (< 25 KPa). The error that is made in the empirical determination of the over-consolidation ratio for over-consolidated sediments and in the determination of the plasticity index and over-consolidation ratio exponent k is expressed by the shift in the data for the over-consolidated sediments. An analysis of the data results in a standard error of estimate of 4.27, that is, the values of shear strength obtained using the regression equation with the shear modulus G_{MAX} varies about ± 4.27 KPa from the true value. With regard to the range of shear strength values of 2.0 KPa to 40 KPa in which the sediments of this study predominately fell, this error is acceptable.

The relationship that is found between the undrained shear strength and the small strain shear modulus is consistent with the results found in the correlation between undrained shear strength and small strain peak shear modulus (Hardin and Drnevich,

1972). Shear stress versus shear strain data for different strain amplitudes from resonant column tests (Figure 58) are similar to large strain stress - rotation curve from vane shear tests. The end points of each hysteresis loop can roughly be approximated as a simple hyperbolic function and is dependent on the shear modulus and the peak shear strength.

The correlation between undrained shear strength and the elastic parameter results in an identical relationship as observed by the shear strength and the shear modulus correlation (Figure 56). The scatter around the linear regression line is wide (correlation coefficient of 0.423). However, an analysis of the standard error of estimate results in a value 6.43 KPa which is similar to the error derived from the correlation of S_U and G_{MAX} . The scatter of the observed values can be explained by means of Figure 57. Since, the elastic parameter Ξ is an expression of the bulk modulus K and the shear modulus G , the parallel shifts of the regression lines can be attributed to differences in bulk modulus. These differences are reflected in Figure 56 in terms of the scatter of the data populations. The over-consolidated sediments from core JPC 41 have a higher bulk modulus, as do some of the deeper test from the Arctic sediments.

Conclusion and Recommendations

Correlations between parameters derived from miniature vane shear tests, including large strain shear modulus and undrained shear strength, and non-destructive measurements of compressional wave velocity and bulk density are presented. An empirically derived small strain shear modulus was compared with undrained shear strength and shear modulus acquired from miniature vane shear tests. Data from three different geographical environments were used to establish an empirical relationship that allows for a prediction of shear parameters from non-destructive measurements.

The study has shown that small strain and large strain shear characteristics are interdependent. This is shown by a strong linear relationship between the undrained shear strength and small strain shear modulus ($R^2 = 0.733$, $\sigma_{est} = 4.27$ KPa).

The existing relationship between compressional wave velocity, bulk density and small strain shear modulus (Eq.(13)) can be used to approximate undrained shear strength. However, numerous other factors complicate the prediction and have to be taken into account when deriving an empirical equation. Furthermore, the relationship between S_u and Ξ is not as strong as S_u and G_{MAX} because Ξ is representative of the bulk modulus and shear modulus. Variability in bulk modulus among sites (Figure 57) introduces more scatter into the S_u versus Ξ plot (Figure 56) because G_{MAX} appears to be a more sensitive indicator of S_u , since it does not rely on the determination and incorporation of bulk modulus.

A definitive relationship between the large strain shear modulus of the initial slope of the vane shear – rotation plot and small strain shear modulus or Ξ can not be made. The resolution of the vane shear stress – rotation plot (1° vane rotation) is not high

enough to evaluate a sufficient shear modulus. Improving the resolution of the vane shear stress – rotations plots, holds promise for developing such a correlation.

This study suggests that a general equation can be developed to predict shear strength or shear modulus from non-destructive measurements for normally consolidated fine grained sediments. More experimental research is needed to refine the relationships presented here. The promising results of this study provide a basis on which further studies can be built.

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Figure 42: Plot of shear stress versus vane rotation (miniature vane shear test) and the conversion from vane rotation to strain after Cadling et al. (1950).

Core	Location	Water Depth (m)	Recovery (cm)
Central Arctic Ocean (96/09-1pc)	143°26'37"E, 86°24'52"N	927	270
Central Scotian Shelf (87003-002)	63°02'02"W, 44°00'56"N	215	1610
Gulf of Mexico (JPC 32)	92°19'38"W, 26°43'16"N	1915	1510
Gulf of Mexico (JPC 41)	92°13'7"W, 26°34'49"N	2383	1220

Table 4: Core Locations.

Figure 43: Location and bathymetric map of the Central Arctic Ocean (Lomonosov Ridge).

Figure 44: Bulk density, compressional wave velocity and undrained shear strength of Core 96/09-1pc (Central Arctic Ocean, Lomonosov Ridge).

Figure 45: Location and bathymetric map of the Central Scotian Shelf (Emerald Basin).

Figure 46: Bulk density, compressional wave velocity and undrained shear strength of Core 87003-02 (Central Scotian Shelf, Emerald Basin).

Figure 47: Location and bathymetric map of the Gulf of Mexico (Texas-Louisiana Slope and Rise).

Figure 48: Bulk density, compressional wave velocity and undrained shear strength of Core JPC 32 (Gulf of Mexico, Texas Louisiana Slope and Rise).

Figure 49: Bulk density, compressional wave velocity and undrained shear strength of Core JPC 41 (Gulf of Mexico, Texas Louisiana Slope and Rise).

Figure 50: Plot of the initial shear modulus G_{MV} from miniature vane shear test versus the maximum shear modulus G_{MAX} after Jamiolkowski (1991).

Figure 51: Plot of the initial shear modulus G_{MV} from miniature vane shear test versus the elastic parameter Ξ ($V_p^2 \cdot \rho_B$).

Figure 52: Plot of shear modulus G_{MV} , G_{MAX} and elastic parameter Ξ versus depth (Core 96/09-1pc, Central Arctic Ocean).

Figure 53: Plot of undrained shear strength versus small strain shear modulus (after Jamiolkowski, 1991), with linear regression lines and correlation coefficients for each core.

Figure 54: Plot of undrained shear strength versus small strain shear modulus (after Jamiolkowski, 1991) and without the over-consolidated values from Core 96/09-1pc.

Figure 55: Plot of elastic parameter Ξ versus undrained shear strength with linear regression lines and correlation coefficients for each core.

Figure 56: Plot of elastic parameter Ξ versus undrained shear strength and without the over-consolidated values from Core 96/09-1pc.

Figure 57: Plot of elastic parameter Ξ versus large strain shear modulus (after Jamiolkowski, 1991) with linear regression lines for each core.

Figure 58 (a) Results of a resonant column tests at different stress amplitudes. (b) Approximation of the stress strain curve base on results of resonant column tests (Hardin et al., 1972)

APPENDIX A: Non-Destructive Laboratory Testing (MSCL)

Introduction

The determination and evaluation of the physical properties of soils, rocks and sediments from land and sea is an essential part of geotechnical engineering. These properties control the strength, compressibility, and permeability characteristics of sediments and rock that are important for safe design of buildings, offshore structures, transportation, and environmental protection systems. Furthermore, physical property data, especially of marine sediments and rocks, are of fundamental interest for paleoclimatic, paleoenvironmental and stratigraphic studies. The Multi-Sensor Core Logger (also referred to as MSCL) represents an automated, nondestructive and rapid method for the continuous measurement of the physical properties of soil, sediment, and rock cores.

The history of nondestructive logging devices began in the 1960's with the development of the Gamma-Ray Attenuation Porosity Evaluator (GRAPE) to measure bulk density. Over the years, the devices became more and more sophisticated and automated and other equipment such as full waveform logging systems and magnetic susceptibility sensors were added. The availability of appropriate data processors and technology contributed to the fast development of automated, nondestructive devices and eventually resulted in the present day multi-sensor core logging systems.

The multi sensor core logging system is a versatile instrument because of its modular design. This design allows the user to adapt the system to individual testing requirements. The adaptability affects not only the different sensor configurations but

also the adaptability to samples. MSCL devices are able to process whole cores and split cores encased in plastic liners as well as exposed hard rock samples. The basis of the system configuration is formed by a horizontal motorized core conveyer system that moves the cores past the different stationary sensors using a core pusher (Figure A 1). The conveyer is driven by a stepper motor, which controls forward motion to intervals as low as 0.5 mm. The sensors, or measurement devices, are either placed within or around the conveyer system on a center stand. The range of sensors available is multifaceted and currently consists of measurement devices for core diameter and temperature, compressional wave velocity (P-Wave), gamma-ray attenuation, magnetic susceptibility, core imaging, natural gamma ray and electrical resistivity. All integral sensors are connected to a central electronic rack that controls the sensor settings and provides the interface between the sensors and a computer. Customized software enables full automated core logging and recording of the data. The whole MSCL unit is at least 4 meters long (depending on the configuration) and is capable of logging core units with a length of up to 1.50 meters.

The following chapters focus on the gamma-ray attenuation porosity evaluator, the compressional wave velocity device and on the electrical resistivity device. The individual components, their configuration and the theory that underpins each are described in detail. Methods for calibration and data processing are also presented.

The different MSCL devices shown in the figures are fabricated by GEOTEK Ltd. and only depict examples of the different sensors available. However, since the described mechanisms and principles are standards, they are applicable to all other devices available and are not only limited to GEOTEK systems.

Displacement Transducers

The measurement of the sediment thickness and outside core diameter, is essential for almost all sensors mounted on the MSCL. The thickness is a fundamental value for calibrating the instrument response when calculating bulk density, compressional wave velocity, magnetic susceptibility and electrical resistivity.

Apparatus. MSCL systems are equipped with an integral thickness measuring device. In most instances, the determination of the core or sediment thickness results from a displacement measurement. A common approach is to couple rectilinear displacement transducers to the P-Wave transducers (Figure A 2).

Due to the fact that the P-Wave transducers must have close contact to the sample they must be able to compensate for core thickness discontinuities during measurement. This is achieved by imbedding the transducers into a plastic housing using a spring mechanism and horizontal or vertical housing slides perpendicular to the core sample. The coupling of the P-Wave transducer with the displacement transducer results in a simultaneous movement during the logging process. Since this layout yields only a displacement deviation of the core being logged, the displacement transducers have to be adjusted to a reference thickness.

Calibration. A cylindrical reference sample with a known thickness (diameter) is placed between the P-Wave transducer faces. The transducer housing is then adjusted to the reference sample in such a way that the displacement of the spring is approximately midway, (ie. the transducer is able to move in and out of the housing by the same distance). The position of the displacement transducers is then set at zero.

The reference sample does not necessarily need to be of the same size as the core being logged, however the thickness should be equivalent because once the vertical slides of the transducer housings are fixed and the zero position is set, the core sample has to fit between the P-Wave transducers in the same manner as the reference sample.

The sediment thickness d_s can be derived from the core thickness deviation measured by the displacement transducers and follows the relationship:

$$d_s = d_{ref} - d_w + \frac{CTD}{1000} \quad (A 1)$$

In this equation the d_{ref} is the reference sample core or diameter (in m), d_w is the total wall thickness of the liner of the core sample being logged (in m) and CTD is the core thickness deviation (in mm) provided by the transducers.

Gamma-Ray Attenuation

The Gamma-Ray Attenuation Porosity Evaluator (GRAPE) provides continuous, non-destructive determination of sediment and rock bulk density. These data can also be used to calculate porosity and water content.

Apparatus. A single GRAPE apparatus consists of a gamma-ray source, a scintillation detector and a caliper (Figure A 3). The gamma-ray source and the scintillation counter (detector) can be aligned vertically or horizontally to measure across split or whole cores, respectively. The design of the gamma-ray source is simple. The radio active element is housed in a capsule which, in turn, is inside a lead filled steel container (Figure A 4).

The size of the lead container in comparison with the source capsule is very large to provide for user radiation protection. Using a 10 mC ^{137}Cs source, the maximum radiation at the surface of the steel container is less than 7.5 $\mu\text{Sv/h}$ (GEOTEK), subject to the dimensions for enclosure given in Figure A 4. This radiation dose is, in comparison with the current regulatory occupational exposure limit of 25 $\mu\text{Sv/h}$ (0.05 μSv per year, 2000 working hours per year), a safe level (U.S. Nuclear Regulatory Commission (10 CFR 20), U.S. Department of Energy (10 CFR 835)).

The isotope ^{137}Cs (Cesium) is commonly used as the radioactive source in gamma ray attenuation devices, however other source materials like ^{133}Ba (Barium) and ^{60}Co (Cobalt) can be used. The choice of the radiation material is based on the need for a certain level of energy as well as practical considerations. The main reason for choosing ^{137}Cs as the source is its half-life period of 30.2 years which allows for a good balance between energy level (0.662 MeV) and service life. In comparison to ^{137}Cs , ^{133}Ba (Barium) has a half-life of only 10.51 years and ^{60}Co (Cobalt) only 5.26 years (Table A 1).

Slits placed in front of the gamma-ray source collimation facilitate nearly parallel beams and can also be used to alter the beam diameter. For general applications, collimator slit diameters of 5 mm are adequate. Collimator slits with a diameter less than 5 mm (e.g. 2.5 mm) are utilized only to obtain measuring data of very high resolution.

The gamma-ray detector looks similar to the source from its external appearance, but internally is configured differently (Figure A 5). The detector is also encased in steel and lead housing. The detection element, a NaI crystal with a photo-multiplier

tube, is located in the center of the housing.

Principle. The measurement of sediment and rock bulk density using gamma-rays is based on the principles of Compton scattering and attenuation (Figure A 6). A parallel, monoenergetic beam of gamma-rays penetrates a sample (also referred to as absorber). When passing through the absorber some of the gamma-rays are absorbed or scattered and lose energy and direction, respectively. The scintillation counter only detects the gamma-rays that pass through the absorber without any loss of energy. The energy loss and the attenuation are directly related to bulk density.

Three main mechanisms affect the attenuation or alteration of the gamma-rays: Photoelectric absorption, pair production and Compton scattering. The effectiveness of the different attenuation processes depends on the energy level of the gamma-rays and on the nature of the absorber. At an energy level below 0.05 MeV the photoelectric absorption (total energy of the gamma-ray that is imparted to an electron of the absorber) is dominant, whereas the pair production process becomes important at energies above 5 MeV. This high level of energy changes the charge of the hit electron resulting in different charged electrons. If the energy level ranges between 0.2 MeV and 4.0 MeV Compton scattering dominates. Compton scattering is a well understood process. The influence of the other two mechanisms is negligible at this energy range.

Compton scattering, also known as the Compton Effect, is the scattering of photons at quasi-free electrons. If a photon hits a quasi-free electron (electron of the outermost orbit of the atom), the energy and impulse transmission results in the scattering of the photons, and can be described as a classic particle impulse. The photon has a higher wave-length λ' ($\lambda' > \lambda$) after it hits the electron and, consequently,

a lower energy. The scattering process, and thus attenuation, is highly dependent on the number of electrons in the absorber and the number of photons in the gamma-ray beam. The equation is:

$$I = I_0 * e^{-\mu * \rho_B * d} \quad (A 2)$$

Solving Equation (A 2) for bulk density, the following expression is obtained:

$$\rho_B = \frac{1}{\mu * d} * \ln \left[\frac{I_0}{I} \right] \quad (A 3)$$

where,

I = the intensity of the detected gamma-rays with the same,
energy as the source (counts/sec),

I_0 = the intensity of the gamma-ray source (counts/sec),

μ = the specific Compton mass absorption coefficient (m^2/Mg),

d = the thickness (m),

ρ_B = the true bulk density (Mg/m^3).

Attenuation Coefficient. In the determination of bulk density, the parameters I , I_0 , and d are straightforward and easy to measure. However, the determination of the Compton mass absorption coefficient μ is more difficult because it depends on the nature of the absorber. The equation for the coefficient μ can be written as:

$$\mu = \sigma_e * \frac{Z}{A} * N \quad (A 4)$$

where,

μ = the specific Compton mass absorption coefficient (m^2/Mg),

σ_e = the average collision cross section ($m^2/electron$),

Z = the number of electrons per atom (atomic number),

A = the atomic weight of the absorber (Mg/mole), and

N = the number of atoms per mole in the absorber.

The influence that the energy level of the bombarding photons has on the attenuation coefficient was studied and results show that the coefficient does not vary much for common materials at particular gamma-ray energies (Berry, 1961; Chappell, 1956). This energy range is between 0.2 and 3.0 MeV. For example the attenuation coefficient of quartz, using ^{137}Cs (0.662 MeV), has a μ of 7.4 m²/Mg, and, the value of μ is 10.0 m²/Mg, using ^{133}Ba (0.30 to 0.36 MeV) as the radiation source (Table A 2).

Similarly, for common minerals found in sediments and rock, the influence ratio of the atomic number Z to the atomic weight of the absorber A is almost constant. Both, the energy level and the ratio Z/A results in a constant attenuation coefficient for compounds of C, O, Na, Al, Mg, Ca and Si.

The mass attenuation coefficient can vary for other reasons. Problems occur when the coefficient μ diverges from the mean value. That is, the Equation (A 3) is only accurate for sediments and rocks composed of minerals which have similar coefficients. For example, the ratio of Z/A of hydrogen is not the same as many common materials. Water for instance has a 10 % higher μ value then the most common minerals ($\mu_{\text{water}} = 8.56$ m²/Mg and $\mu_{\text{Si}} = 7.4$ m²/Mg by using ^{137}Cs as radiation material). Therefore, the attenuation coefficient changes when sediments have a large percentage of hydrogen (e.g. water) in their pore space.

Calibration. The most common core samples the Multi-Sensor Core Logger operates on are sediments contained in plastic liner or rocks. If the coefficient μ is known with sufficient accuracy, it is possible to calculate the bulk density directly by using Equation (A 3).

Sediments and rocks, for example, are naturally composed of different minerals and therefore the specific mass attenuation coefficient of the sample is dependant on the composition. Furthermore, sediment samples are mostly saturated with water, therefore the determination of the mass attenuation coefficient is difficult due to the attenuation coefficient being 10 percent higher. Finally, attenuation through the liner wall as well as the spreading of the emitted gamma-ray beam results in imprecise data. Considering this, the computed bulk density from Equation (A 3) under an assumed or estimated average coefficient would not lead to valid results.

It is more practical and accurate to evaluate the bulk density by using an empirical approach. The basic principle of this approach is based on a comparison of the measured gamma-ray counts of a sample with a standard curve, evaluated by means of measuring bulk density standards. The core sample is looked upon as a two phase system, wherein phase one is minerals and phase two is the pore or interstitial water. This system can be replicated by means of density standard. As the bulk density standard for the mineral phase, aluminum is used due to its similar mass attenuation coefficient to the most frequently occurring minerals (e.g. quartz). The liquid phase is replaced by distilled water. The assumption is now that the attenuation of the standard, consisting of aluminum and distilled water, is equivalent to the attenuation of the core sample, consisting of minerals and water, at the same density level. Therefore, only

the attenuation has to be determined to evaluate the bulk density of the core.

In practice the calibration curve is derived by using predefined standards. Actual devices are illustrated Figure A 7 and Figure A 8 and consist of a cylindrical or half cylindrical aluminum rod of varying diameter, contained in a liner which is completely filled with distilled water. To obtain an accurate calibration curve the standard should have at least 5 diameter gradations. Besides the use of prefabricated standards, another method to evaluate the calibration curve is measuring discrete plates of aluminum. This approach is based on the same principle as previously described (Figure A 9).

The advantage of this method is that for each measurement the specific liner of the core sample can be used for calibration, that is, the liner attenuation is taken into account. However, the difficult handling of this method (the plates must be aligned perpendicular to the gamma-ray beam) is a disadvantage which eventually resulted in the development and application of the prefabricated aluminum rod standards.

The average density of a standard can easily be calculated because the densities of aluminum and water are known as well as the geometry. The density at each section (each different diameter of the rod) is then:

$$\rho_i = \frac{d_i}{D} * \rho_{Al} + \frac{D - d_i}{D} * \rho_F \quad (A\ 5)$$

where ρ_i = the average density at section I (Mg/m^3),

d_i = the aluminum rod thickness at section I (m),

D = the maximum aluminum rod thickness or inside diameter of
the liner (m),

ρ_{Al} = the density of aluminum (Mg/m^3),

ρ_F = the density of distilled water (Mg/m^3).

The standard is then placed between the sensors of the GRAPE device and gamma counts are measured at each section. The minimum count period (t_{cal}) should not fall below 100 seconds per section. The detected counts have to be normalized, that is dividing the counts by the time. The calibration curve is acquired by plotting the average density (A) times the internal diameter of the liner ($d_{int,liner}$) over the natural logarithm of normalized gamma counts. By multiplying the average density times the internal diameter of the liner the density becomes normalized, too. Through the calibration points ($\ln(counts/s)_i, \rho_i$) a second order polynomial, fit to the calibration data ($y = Ax^2 + Bx + C$) yields the calibration coefficients A , B , and C . An example of a calibration graph is illustrated in Figure A 10. The density of the core sample is then determined as:

$$\ln\left(\frac{\text{counts}}{t_s}\right) = A * (\rho_{BC} * d_s)^2 + B * (\rho_{BC} * d_s) + C \quad (A\ 6)$$

where

ρ_{BC} = the bulk density calculated from gamma counts (Mg/m^3),

A, B, C = the calibration coefficients,

t_s = the sampling period (sec),

counts = the actual measured counts on the sample,

d_s = the actual sediment thickness (m).

The sediment thickness used in Equation (A 6) is determined from the core thickness deviation measurement (Eq.(A 1)).

In the densiometry method, described above, the attenuation coefficient is assumed

to be constant for the mineral phase, that is, the minerals have nearly the same μ as the aluminum standard. Yet, if mineralogical analysis indicates significant differences in coefficients, corrections for the bulk density must be applied.

$$\rho_B = \rho_{BC} * \frac{\mu}{\mu_s} \quad (\text{A } 7)$$

where

ρ_B = the true bulk density (Mg/m³),

μ = the standard coefficient of aluminum (m²/Mg),

μ_s = the true mass attenuation coefficient (m²/Mg).

Prior to the introduction of the calibration method, described above, calibration of the GRAPE device was performed with two aluminum cylinders of different thickness. In this procedure, different attenuation behavior of water was not taken into account and therefore the calculated density was about 10 % too high. However, the over estimated density data can be corrected by using Equation (A 9), which was developed by Boyce (1976) and accounts for the lower Compton mass attenuation coefficient of water.

$$\rho_B = \frac{(\rho_{BC} - \rho_{FC}) * (\rho_s - \rho_F)}{\rho_{SC} - \rho_{FC}} + \rho_F \quad (\text{A } 9)$$

where

ρ_B = the true bulk density (Mg/m³),

ρ_{BC} = the bulk density calculated from gamma counts (Mg/m³),

ρ_F = the true fluid density (Mg/m³),

ρ_{FC} = the fluid density calculated from gamma counts (Mg/m^3),

ρ_S = the true grain density (Mg/m^3), and

ρ_{SC} = the grain density calculated from gamma counts (Mg/m^3).

In Equation (A 9), ρ_{BC} as well as ρ_{FC} are known and calculated, respectively. The values for true grain density, grain density calculated from gamma counts, and true fluid density must be estimated. To simplify Equation (A 9), ρ_G is equated with ρ_{GC} . Values for grain density of different geologic minerals and fluid densities are given in Table A 3.

Today, the use of fluid correction is not necessary because the new calibration procedure accounts for the influence of fluids (water). However, the procedure is still needed to readjust older log data in which the influence of fluids is disregarded.

Porosity and Water Content. In porous materials bulk density is related to the matrix density and grain density, respectively, as well as to the fluid density and can be described by the following expression:

$$\rho_B = \rho_S * (1 - \phi) + \rho_F * \phi \quad (\text{A } 10)$$

where

ρ_B = the bulk density (Mg/m^3),

ρ_S = the grain or matrix density (Mg/m^3),

ρ_F = the fluid density (Mg/m^3), and

ϕ = the porosity (-).

Reconverting Equation (A 10), porosity is determined by:

$$\phi = \frac{\rho_s - \rho_B}{\rho_s - \rho_F} \quad (\text{A } 11)$$

The unknown values in (A 11) are the grain density ρ_G and the density of the interstitial fluid ρ_F . For common geologic minerals, estimation of these values can be made with sufficient accuracy. Table A 3 displays grain densities of common minerals as well as densities for frequently occurring interstitial fluids.

Compressional Wave Velocity

The compressional (P) wave velocity system can be used to differentiate and quantify physical properties such as elastic modulus, density, porosity, homogeneity and grain structure of solids and liquids.

Apparatus. The compressional (P) wave velocity unit is mounted on the center stand of the MSCL device. The assembly consists of a pair of P-Wave transducers (PWT) and the rectilinear displacement transducers (DT) that are located diametrically across the core sample (Figure A 11).

Two different types of PWT's are available in present devices; a) stainless steel piston transducers (old style) and b) oil filled acoustic rolling contact (ARC) transducers (new style). The heart of the stainless steel piston transducer is a piezo-electric element that is embedded in epoxy resin and encased in a stainless steel container. A piezo-electric element is a crystal plate of natural or synthetic material that is able to deform mechanically by means of an applied electric field and thereby

generates ultrasonic vibrations. The performance of the piezo-electric element is invertible, that is the element also produces an electrical pulse while deforming. Thus, the transducers can either operate as a transmitter or as a receiver. The epoxy resin attenuates and scatters the back-transmission and thus shields back of the source. The front face of the piston consists of a thin plastic window that allows for the forward energy transmission (Figure A 12).

The oil filled acoustic rolling contact (ARC) transducer has a similar design as the piston transducer. The active element is a piezo-electric crystal embedded in a shielding epoxy resin. However, the crystal is mounted stationary on a center spindle and is surrounded by a rotating, soft and deformable diaphragm. The space between the active element and the diaphragm is filled with castor-oil (Figure A 13).

For whole core logging the alignment of the transducers is horizontal. To ensure proper coupling between the sample and the transducers, the active element is spring-loaded and mounted within a plastic housing. By using piston transducers, the ultrasonic coupling can vastly be improved if the liner surface or the face of the transducer is moistened with water or other lubricants. Since, the ARC transducer already has an excellent ultrasonic coupling due to the soft oil-filled diaphragm it becomes redundant to moisten the surfaces during the logging process.

While logging split cores, the transducers are vertically orientated. The receiver is usually an ARC transducer and is mounted stationary within the conveyer track. The receiver has permanent contact to the sample whereas the upper PWT (piston type) is mounted on a vertical slide and can be raised or lowered during measurement intervals by means of a stepper motor. To protect the upper PWT from contamination and to

improve coupling the exposed core surface has to be covered with a thin plastic wrap and moistened.

Principle. The velocity of the P-Wave, traveling through a medium (sample), is controlled by the physical properties of the medium and the ambient temperature. The elasticity and density are the basic physical properties that govern the velocity of the wave propagation. The compressional (P) wave is a longitudinal wave in which the displacement of the particles through which the wave travels is parallel to the propagation of the wave. The displacement is expressed as an elastic back and forth oscillation about an equilibrium position, rather than as an absolute change of position.

The ignition of a P-Wave is a vibration (caused by the active element) that sets the adjacent medium in motion (Figure A 14) The particles in front of the wave are forced together creating a zone of higher density (compression) and behind the wave an area of lower density (rarefaction). The propagation of the P-Wave through the medium can be described as a chain reaction. The first particles that are set into an oscillating motion transfer the energy to the adjacent particles which in turn begin to oscillate and transmit the energy to the next particles and so on. The velocity with which the energy transmission occurs strongly depends on the inertia of the medium. Denser mediums possess a higher inertia and the time needed to set the particles in oscillating motion increases. Consequently the transmission of the P-Wave (energy transfer from particle to particle) is slowed down and results in a decreasing velocity. The elasticity (restoring forces) affects the P-Wave velocity in the opposite way. The higher the elasticity of the medium, the better is the ability of the particles to oscillate, and

transfer energy to the adjacent particles.

The basic equation for the relationship between density, elasticity and velocity is given by:

$$V_p = \left(\frac{B + \frac{4}{3} * G}{\rho_B} \right)^{\frac{1}{2}} \quad (A 13)$$

where

V_p = the P-Wave velocity in the sediment (in m/s),

ρ_B = the bulk modulus (Mg/m²),

G = the shear or (rigidity) modulus (Mg/m²),

B = the bulk modulus (in MG/m³).

Solving Equation (A 13) for the bulk modulus the following equation is obtained;

$$B = V_p^2 * \rho_B - \frac{4}{3} * G \quad (A 14)$$

The bulk density is determined by discrete measurements (destructive) or by the gamma-ray attenuation porosity evaluator (nondestructive) and the shear (rigidity) modulus can be estimated from charts or empirical correlations.

The most essential processes to evaluate accurate P-Wave velocity data are accurate pulse travel times, and the setting of the pulse timing circuitry. An ultrasonic pulse transmitter produces a 500 kHz pulse (corresponding to a wave period T of 2 μ s; $T=1/f$) with a voltage spike of 120 V and a repetition rate of 1 kHz. The pulse is sent to the transmitter transducer and excites the piezo-electric element that in turn launches the P-Wave at approximately 500 kHz. The frequency of the emitted pulse or the P-Wave can vary between a few Hz and a GHz because the pulse travel time

through a medium is not affected by the frequency. However, shorter wavelengths are more responsive to changes in the medium through which they pass. For geotechnical applications it turned out that a frequency between 250 kHz and 500 kHz is sufficient. The P-Wave then propagates through the sample, detected on the other side by the receiver transducer and is amplified by an automatic gain control (AGC). The time period or the travel time of the wave which is needed to propagate the sample is recorded by the system. The accuracy with which the travel time is measured depends strongly on the first arrival detection of the wave at the receiver side (oscilloscope). In manually operated P-Wave logging systems and for sporadic measurements a subjective, visual first arrival pick (amplitude) can be performed. However, in full-automated systems it becomes necessary to incorporate a more reliable and electronically definitive technique to obtain continuous high resolution data. Most of the present devices use zero-crossings to determine the first arrival and the pulse travel time.

This detection procedure is insensitive to amplitude variations and noise, and is therefore reliable for weak signals. A short time after the ultrasonic pulse transducer has generated the transmit pulse (500 kHz, 120V), a delay pulse is triggered. The time set of the delay pulse is just a few microseconds less than the transmit pulse needed to propagate through the sample. The end of the delay time pulse is a signal for the oscilloscope to start recording the P-Wave and to set the gate time pulse that in turn describes the duration of the oscilloscope recording. The delay time pulse works like a shield for the receiver in a way that it insulates the receiver from background noise before the objective wave arrives. This ensures that there is no interference or pre-

stimulation that could alter the pulse time measurement. At the end of the gate pulse, which is normally set to at least three frequencies, the gate pulse terminates the oscilloscope recording. While the oscilloscope is recording the incoming pulse a threshold detector, integrated in the oscilloscope, detects all negative deflections of the received pulse that are crossing a preset threshold level. The magnitude of the threshold level depends on the intensity of the incoming signal amplitude and consequently on the sample characteristics and is set to approximately 10 to 20 % of the maximum signal amplitude voltage. Parallel to the threshold detection a zero-crossing detector (also integrated into the oscilloscope) monitors all crossings of the pulse signal with the abscissa. The count time pulse that is triggered at the same time as the ultrasonic pulse is then terminated and recorded after the first zero-crossing following the first threshold level detection (Figure A 15).

When the pulse travel time is determined the velocity is derived using:

$$V_p = \frac{d_s}{t_s} \quad (\text{A } 15)$$

where

V_p = the P-Wave velocity in the sediment (in m/s),

d_s = the sediment thickness (in m; see Eq.(3.1)),

t_s = the measured travel time through the sample (in s).

Besides an accurate pulse time measurement, environmental and non-physical influences have to be taken into account for sufficient velocity data acquisition. The P-Wave velocity is very sensitive to temperature. This property mainly affects the interstitial fluid. A temperature difference of one degree can cause velocity variation

higher than between different sediment strata. Increasing temperature results in decreasing density and this in turn results in increasing P-Wave velocity (Figure A 16).

Core quality is regarded as a non-physical influence which can influence the P-Wave velocity to a great extent. Insufficient coupling between the core liner and the sample due to sampling effects as well as embedded gas voids and bubbles in the sample may result in variations of the velocity. This is mainly attributed to the different P-Wave velocities of gases. The velocity of gases is about four times less than that of solid or liquids and due to the fact that the size of the voids is not known sufficiently the correction of the velocity is impractical. Table A 4 displays compressional (P) wave velocities of common sediment types, elements and fluids.

Calibration. The measured total pulse travel time (count pulse, see Figure A 15) has to be calibrated before calculating the P-Wave velocity. Four time delays have to be taken into account for the purpose of evaluating the absolute pulse travel time through the core sample; (1) pulse delay, (2) transducer delay, (3) circuitry delay, and (4) core liner delay.

The pulse delay is an error in time measurement due to the first arrival detection method (zero crossing). The pulse delay is a time constant and depends on the wave length and the system wiring. Figure A 17 illustrates the dependency of the wiring on the pulse delay. If the first peak is positive, the pulse delay is one times the wave length because the threshold detection is only triggered for incoming negative signal peaks. In case of an inverse wiring the first signal peak is negative and therefore the

pulse delay is one and a half times the wave length. Since the pulse delay is a constant it is already integrated into the edit routine of the PWL system and usually requires no adjustment while logging. However, if hardware is replaced or samples with a different geometry are measured the pulse delay has to be adjusted.

The calibration for circuitry delay (time needed by the emitted pulse of the pulse transmitter to reach the active element and vice versa from the receiver to the oscilloscope as well as electrical delays), transducer delay (time needed by the P-Wave to propagate the distance between the active elements and the transducer faces) and liner delay (time needed by the P-Wave to travel through the liner walls) is performed simultaneously by means of a calibration standard. A small segment of liner filled with distilled water is placed between the transducer faces. The most suitable calibration liner is a liner segment of the same type and size as the core being logged. After placing the standard in the core track, an ultrasonic pulse is triggered (same pulse as if logging a normal core) and the total pulse travel time is recorded, including all delays described above. The absolute travel time through the sediment (distilled water) can be calculated due to the fact that the P-Wave velocity of distilled water at a particular temperature is known (Figure A 18). The equation to calculate the absolute travel time is given by:

$$t_s = \frac{(d_c - d_w)}{V_{p,w}} = \frac{d_s}{V_{p,w}} \quad (\text{A } 16)$$

where t_s is the pulse travel time in sediment (distilled water), d_c is the distance between the transducer faces, d_w is the total wall thickness of the liner of the calibration standard, d_s is the sediment thickness obtained from the displacement measurement (see 3.1) and $V_{p,w}$ is the P-Wave velocity of distilled water.

The total delay time of the pulse (t_{delay}) due to circuitry delay (t_{circ}), transducer delay (t_{trans}), liner delay (t_w) and pulse delay (t_{pulse}) results in:

$$t_{\text{delay}} = t_{\text{tot}} - t_{\text{pulse}} - t_s = t_{\text{tot}} - t_{\text{pulse}} - \frac{(d_c - 2 * d_w)}{V_{p,w}} \quad (\text{A } 17)$$

where t_{tot} is the recorded travel time of the measurement.

The total delay time is a constant of the measurement device in conjunction with the calibration standard liner (liner of the core being logged) and therefore has only to be inserted into the PWL system once before logging.

To obtain the P-Wave velocity through the sediment inside the core liner (corrected velocity) Equation (A 18) is used:

$$V_p = \frac{d_s}{t_{\text{tot}} - t_{\text{delay}}} \quad (\text{A } 18)$$

Data Processing. Post-processing of the logged P-Wave velocity is inevitable to provide universal and comparable data sets that can be used for investigation and research. The major corrections of the measured laboratory P-Wave velocity are for temperature (laboratory and in situ correction), salinity (in situ correction) and pressure (in situ correction).

The temperature of the interstitial fluid effect on the P-Wave velocity is described previously. However, differences in temperature between different logging environments (sites) have to be addressed to obtain compatible data sets. The standard or set temperature for velocity data is 20°C. An equation for the approximation of the ambient or environmental temperature to the standard temperature is provided by:

$$\frac{1}{V_{p,T}} = \frac{1}{V_p} + \left(\frac{\phi}{V_{p,T,W}} - \frac{\phi}{V_{p,T,L}} \right) \quad (\text{A } 19)$$

In this equation $V_{p,T}$ (in m/s) is the corrected P-Wave velocity at temperature T ($^{\circ}\text{C}$), V_p the logged P-Wave velocity (in m/s) $V_{p,T,W}$ is the velocity of the pore water at desired temperature (m/s) and $V_{p,T,L}$ is the temperature of the pore water at the moment of logging.

In some cases it is desirable to compare the laboratory measurement with in situ down-hole measurements. However, the environment in respect of temperature, salinity and pressure (depth) is obviously very different to provide sufficient statements. To account for these differences a polynomial function (Chen and Millero, 1977) together with the ratio method (Hamilton, 1971) can be used;

$$V_{p,z} = V_{p,20} * \left(\frac{1449.2 + 4.6T - 0.55T^2 + 0.00029T^3 + (1.34 - 0.01T)(S - 35) + 0.016z}{V_{p,s,w}} \right) \quad (\text{A } 20)$$

where $V_{p,z}$ is the in situ P-Wave velocity (in m/s), $V_{p,20}$ is the logged P-Wave velocity at 20°C (in m/s), z is the depth (in m), T is the temperature (in $^{\circ}\text{C}$), S is the salinity (in ‰) and $V_{p,sw}$ is P-Wave velocity of sea water at 20°C .

Electrical Resistivity

The electrical resistivity has been used as a fundamental tool for physical property evaluation for over 60 years. Commonly it is measured in situ by means of down hole probes. Recently, non-contacting electrical resistivity (NCR) measurements were

performed with core logging devices using a methodology called the eddy current technique. The most important property that can be derived by resistivity is porosity.

Apparatus. The NCR device is integrated at the end of the sensor track. It is placed below the conveyer track and just between the set of rails (Figure A 19). The distance between the sensor and the moving core sample is limited to a minimum, due to physical restrictions of the methodology of the resistivity measurement. The external appearance of the sensor is box shaped and about 10 by 15 cm. The internal configuration of the sensor and the electrical units as well as the specifications remains unknown for the user due to the fact that most of these devices are legally patent protected. However, the main components are briefly described and are similar for a wide range of devices.

The NCR device uses eddy current techniques that consist of two sets of coils. The first set of coils (measuring coils) is located in the upper part of the sensor box and is aligned vertically and perpendicular to the core sample axis. One of these coils acts as the transmitter coil and is connected to an alternating current source at the electronic rack. The other one acts as the receiver coil and is connected to an impedance analyzer (eddyscope) at the electronic rack. The second set of coils is identical but located in the lower part of the sensor and is insulated from the sphere of influence of the measuring coils. The connections to the electronic rack are similar to the first set.

Principle. The main principle, on which the non-contact resistivity (NCR) is based, is the eddy current technique. The eddy current technique takes advantage of the

ability of conductive materials to generate a magnetic field while under the influence of an electric current that can be measured and converted into resistivity. The transmitter coil is connected to an alternating current (AC). If the current is passed through the coil, an alternating magnetic field is generated in and around the coil. If the coil is placed close to the sample the alternating magnetic field of the coil penetrates into the sample. The penetration depth of the magnetic field depends on the frequency of the excitation current, the design of the coil (number of loops), and the conductivity and magnetic permeability of the sample. The depth increases with decreasing frequency, increasing number of loops (larger magnetic field) and increasing conductivity and magnetic permeability (Figure A 20).

Due to the conductivity and the magnetic permeability of the sample, the applied magnetic field initiates current flow within the sediment sample by means of induction. The direction of movement is orbital and perpendicular to the magnetic flux. This induced current is called the eddy current because of its similar appearance to swirling water currents. The eddy currents in the sediment sample, in turn, generate their own magnetic field which can be detected and converted into a voltage by the receiver coil (re-induction).

The magnetic field generated by the eddy currents is very small. To be able to measure this small magnetic field with as much accuracy as possible, a reference set of coils operating in air is used. This methodology minimizes the effect and influence of natural or artificial (electronic devices) magnetic fields in the immediate vicinity.

The measured voltage, as result of the eddy current magnetic field, is a function of the conductivity of the sample and, consequently, its resistivity which is the reciprocal

of the conductivity.

Calibration. As described previously, the non-contact resistivity (NCR) device measures only the voltage of the receiver coil, induced by the eddy currents. Developing a resistivity versus NCR response curve establishes a convenient way to convert the measured voltage into resistivity. To develop the response plot a series of saline solutions at different concentrations are used. The principle behind this procedure is straight forward. The resistivity of saline solutions at specific concentrations is well established and can be gathered from the literature or calculated by using the following relationship:

$$R_w = 5.3394 * C^{-0.8987} \quad (A\ 21)$$

where

R_w = the electrical resistivity (ohm-m),

C = the salt concentration (g/l).

Placing the saline solution contained in a core liner over the electrical resistivity sensor and measuring the sensor response (voltage) for each case, a plot of resistivity over sensor response and voltage, is evaluated. The core liner used for the calibration should have the same liner specifications as the liner of the core being logged (size, shape and material) because the resistivity sensor is very sensitive to different measurement geometries.

Once the sensor is calibrated (i.e. Table A 5 and Figure A 21), the measured voltage of the sample is converted to resistivity using of the equation obtained from the calibration curve in Figure A 21 (power law curve). The resistivity of the sample

follows Equation (A 22):

$$R = A * V^B \quad (A\ 22)$$

where

A = coefficient of the calibration curve,

B = coefficient of the calibration curve,

V = sensor response (mV), and

R = electrical resistivity of the sample (ohm-m).

The resistivity of materials is very sensitive to variations in temperature. The resistivity decreases as temperature increases. To be able to provide comparable resistivity data it is essential to apply a temperature correction for the logging environment. This can be done by using the charts given in the literature or by using an approximation (Arps, 1964):

$$R_{T,2} = R_{T,1} * \left[\frac{T_1 + 21.5}{T_2 + 21.5} \right] \quad (A\ 23)$$

In Equation (A 23), $R_{T,2}$, is the electrical resistivity corrected for temperature (ohm-m), $R_{T,1}$ is the measured electrical resistivity, T_1 is the temperature at which the sample was measured ($^{\circ}\text{C}$) and T_2 is the desired temperature ($^{\circ}\text{C}$). It should be mentioned that Equation (A 23) is an approximation.

Data Processing. For the relationship between the electrical resistivity and the porosity no universal model covers the wide range of sediment types and mixtures. Several empirical and theoretical models that yield sufficient results for the porosity evaluation are described in the literature. The resistivity of a sample is influenced by

several factors such as type of sediment (sand, silt, clay, etc), size, shape, structure, distribution and cementation of particles, and tortuosity that have to be taken into account by selecting an appropriate model.

In 1942, Archie was the first who linked the electrical resistivity of fluid saturated sandstone to the porosity and to the resistivity of the interstitial pore fluid. He introduced the formation factor FF which is the ratio of formation resistivity (resistivity of the whole sample) and the resistivity of the interstitial pore fluid and is unique for each soil or sediment composition:

$$FF = \frac{R}{R_w} \quad (A\ 24)$$

where

FF = the formation factor (non-dimensional),

R = electrical resistivity of the sample (ohm-m), and

R_w = electrical resistivity of the interstitial pore fluid (ohm-m).

The evaluation of the electrical resistivity of the sample is described above. The resistivity of the interstitial pore fluid can be calculated using Equation (A 21).

Furthermore, Archie provided an empirical relationship between the formation factor and the porosity that he developed from investigations of sandstones:

$$FF = \phi^{-m} \quad (A\ 25)$$

In this equation m is a dimensionless constant and allows for tortuosity of the pore sediments. Archie's equation (also referred to as Archie's Law) is based on the assumption that the conductivity of a fluid saturated sediment sample is due to the connectivity of conductive fluid-filled pores. The solid phase (i.e. particles) is

considered to have infinitive resistance ($R \approx \infty$). However this assumption is only valid for sands with little or no clay-mineral content. If the samples contain clay particles, the conductivity of the sample is not only due to the conductivity of the interstitial pore-fluid but also due to the conductivity of the clay particles. Clay particles are charged particles and therefore are able to conduct currents.

The enhancement of Archie's Law (Winsauer *et al.*, 1952) includes the additional dimensionless constant a that takes the lithology of the sediments into account:

$$FF = a * \phi^{-m} \quad (A\ 26)$$

Equation (A 26) is the basic approach for almost all models and concepts developed for different sediment compositions. The constant a varies between 0.6 and 2.0 and the constant m between 1 and 3. Table A 6 shows a small assortment of a and m values derived by different authors together with specifications on the type of sediment for which they can be applied.

A partial empirical method to evaluate the constants a and m is to generate a log-log scale plot of formation factor FF versus porosity. Porosity data are determined discretely in the laboratory, from well-logs (neutron density combination) or by density-porosity relationships (GRAPE porosity). A straight line fit through the points in the log-log scale plot yields the a and m constants (Figure A 22), as follows:

$$m = \frac{\log(FF_2) - \log(FF_1)}{\phi_2 - \phi_1} \quad (A\ 27)$$

$$a = FF_1 * \phi_1^m = FF_2 * \phi_2^m$$

In these equations F_1 and ϕ_1 as well as F_2 and ϕ_2 are any two points of the straight line.

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Figure A 1: Typical MSCL system configuration for split and whole core logging (GEOTEK); (1) active conveyer track, (2) passive conveyer track, (3) core pusher, (4) controlling computer and sensor electronics, (5) linescan camera, (6) gamma-ray attenuation porosity evaluator (GRAPE), (7) compressional (P) wave sensor, (8) magnetic susceptibility point sensor, (9) magnetic susceptibility loop sensor, (10) natural gamma sensor, (11) non-contact electric resistivity sensor.

Figure A 2: Example configuration of a displacement transducer mounted on a spring loaded P-Wave transducer and their coupling.

Figure A 3: Standard configuration of the GRAPE apparatus on the MSCL. Shown in this image is one of the GEOTEK systems.

Figure A 4: Gamma-ray source location within its lead-shielded housing (all denoted dimensions are minimum values to ensure the radiation shielding).

Source Element		Energy Level		Half-Life	
⁶⁰ Co	(Cobalt)	1.17 to 1.33	MeV	5.26	years
¹³³ Ba	(Barium)	0.30 to 0.36	MeV	10.51	years
¹³⁷ Cs	(Cesium)	0.662	MeV	30.20	years

Table A 1: Energy level and half-life of capable radiation materials.

Figure A 5: Scintillation counter (detector) configuration schematic.

Figure A 6: Schematic illustration of the Compton scattering process. The quasi-free electron (e) is hit by a gamma wave (γ) with the specific wave length (λ). After the collision the gamma wave (γ') possesses a higher wave length (λ').

Type	$\mu (^{133}\text{Ba})$ [m ² /Mg]	$\mu (^{137}\text{Cs})$ [m ² /Mg]
Sandstone	10.10	7.80
Limestone	10.20	7.70
Dolomite	10.20	7.70
Shale	10.10	7.60
Anhydrite	10.20	7.70
Feldspar	9.90	7.40
Quarz	10.00	7.40
Aluminum	9.80	7.40
Water	1.08	8.56

Table A 2: Mass attenuation coefficient μ for various minerals, aluminum, and water at different energy levels.

Figure A 7: Bulk density standard for whole core calibration (telescope rod)

Figure A 8: Bulk density standard for split core calibration (telescope rod)

Figure A 9: Bulk density standard for whole core calibration (plates)

Figure A 10: Calibration graph

Mineral Type	Grain Density ρ_G [Mg/m ³]	Mineral Type	Grain Density ρ_G [Mg/m ³]
Albite	2.61	Magnesite	2.97
Almandine	4.08	Magnetite	5.18
Anorthite	2.75	Matrolite	2.25
Apatite	3.23	Muscovite	2.78
Aragonite	2.99	Nornblende	3.15
Augite	3.32	Olivine	3.31
Barite	4.50	Phlogo	2.80
Biotite	3.00	Plagioclase	2.64
Calcite	2.71	Pyrite	5.02
Clinoperthite	2.54	Quartz	2.65
Diopside	3.31	Rhodochrosite	3.57
Dolomite	2.85	Rutile	4.20
Epiote	3.40	Siderite	3.75
Fluorite	3.18	Sillikanite	3.19
Forsterite	3.22	Sphalerite	4.00
Galena	7.50	Staurolite	3.78
Garnet	3.60	Sulfur	2.07
Halite	2.16	Sylvite	1.99
Hematite	5.26	Topaz	3.50
Hornblende	3.12	Tourmaline	3.23
Distilled Water	1.000	Seawater (S=15‰)	1.010
Pure Water	0.9981	Seawater (S=35‰)	1.025

Table A 3: Matrix density ρ_G of common minerals and densities for interstitial fluids ρ_F (the densities of the fluids are valid at a temperature of 20°C).

Figure A 11: (a) Cross-section of a stainless steel piston transducer (PT), (b) cross-section of an oil filled acoustic rolling contact (ARC) transducer

Figure A 12: Compressional (P) wave velocity unit with stainless steel piston transducers and rectilinear displacement transducers for whole core logging (GEOTEK System)

Figure A 13: Acoustic Rolling Contact (ARC) transducer in a vertical configuration for split core logging (GEOTEK)

Figure A 14: Visualization of a longitudinal wave. Because of the compressional and dilatational forces also called pressure or compressional wave.

Figure A 15: Signal and pulse timing diagram for PWL system using the zero-crossing method.

Figure A 16: P-Wave velocity in distilled water at atmospheric pressure (NAVORD REPORT 6747)

Sediment Type	Bulk Density ρ_B [Mg/m ³]	Velocity V_p [m/s]	Temperature [°C]
Sand (Coarse)	2.03	1836	23
Sand (Fine)	1.98	1742	23
Sand (Very Fine)	1.91	1711	23
Silty Sand	1.83	1677	23
Sandy Silt	1.56	1552	23
Sand-Silt-Clay	1.58	1578	23
Clayey Silt	1.43	1535	23
Clayey Silt	1.38	1535	23
Clayey Silt	1.41	1531	23
Silty Clay	1.42	1519	23
Silty Clay	1.24	1521	23
Silty Clay	1.37	1507	23
Clay	1.26	1505	23
Clay	1.42	1491	23
Distilled Water	1.000	1497	25
Seawater (S=35‰)	1.025	1533	25
Aluminum	2.700	~5100	--
Lead	11.35	~1322	--
Air	0.001275	343	20

Table A 4: Average sound velocities of sediments and reference materials (Hamilton, 1971).

Figure A 17: Influence of inverse wiring on the pulse delay. On the left hand side the first incoming peak is positive and results in a $t_{\text{Pulse}} = 1.0 \times \lambda$. The right side indicates a negative incoming first peak which results in a $t_{\text{Pulse}} = 1.5 \times \lambda$.

Figure A 18: Pulse delay times

Figure A 19: Electrical resistivity sensor (GEOTEK system)

Figure A 20: The figure on the left side illustrates the eddy currents which are induced by the magnetic field of the coil. The figure on the right side illustrates the magnetic field generated by the eddy currents.

Concentration C [g/l]	Resistivity R _w [ohm-m]	Sensor Respons V [mV]
35.00	0.209	405.790
17.50	0.390	210.477
3.50	1.730	43.765
1.75	3.321	22.008
0.35	15.481	4.341

Table A 5: Example for a calibration table; saline concentration, calculated resistivity and measured sensor voltage.

Figure A 21: Example for a calibration curve of calculated resistivity versus sensor response.

Author	<i>a</i> Value	<i>m</i> Value	Sediment Type
Archie (1942)	1.00	1.80 – 2.00	clean consolidated sandstone
	1.00	1.30	clean unconsolidated sand
Winsauer (1952, 1953)	0.62	2.15	sandstone, detrital quartz (with intergranular porosity; Humble Formula)
	1.00	$1.87+0.019\phi^{-1}$	non-fissured carbonates of low porosity (Shell formula)
Atkins (1961)	1.00	3.28	sodium montmorillonite
	1.00	2.70	calcium montmorillonite
	1.00	2.11	illite
	1.00	1.87	kaolinite
	1.00	1.60	sand
Boyce (1968)	1.30	1.45	silty, sandy sediments with a clay fracture of 5% – 26% (kaolinite, montmorillonite, illite and chlorite)
Kermabon (1969)	1440	1.46	clay, sand, silt (Note: $FF = (a * \phi^{-m}) - 0.7193$)
Taylor-Smith (1971)	1.35	1.20	sand, silt and clay
	1.00	2.00	only silts and clays ($\phi > 60\%$)
	1.00	1.50	only sands and coarse silts ($\phi < 60\%$)

Table A 6: Values for the constants *a* and *m* of different authors; Archie (1942), Winsauer (1952, 1953), Atkins (1961), Boyce (1968), Kermabon (1969) and Taylor-Smith (1971).

Figure A 22: Plot of formation factor versus porosity for the determination of '*a*' and '*m*' (Serra, 1984).

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