

Bangor University

DOCTOR OF PHILOSOPHY

Resistivity, thermal conductivity, porosity relationships for marine sediments

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Award date: 2019

Awarding institution: University College of North Wales, Bangor

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RESISTIVITY - THERMAL CONDUCTIVITY - POROSITY RELATIONSHIPS FOR MARINE SEDIMENTS

by Michael Anthony Lovell

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A thesis presented for the degree of Philosophiae Doctor University of Wales

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U.K.

December 1983



SUMMARY

A laboratory examination of the electrical resistivity formation factor, thermal conductivity, and porosity properties of natural and artificial marine sediments, both in terms of their own interdependences and of their interrelationships with other geophysical and geotechnical quantities has been carried out. Marine sediments have been considered as two separate systems, sands and clays. For the sands, measurements of electrical formation factor, thermal conductivity, porosity, and permeability have been made using three cells. For the deep sea clays, an oedometer cell has been modified to enable resistivity, thermal conductivity, compressional wave and shear wave velocity measurements to be made, permeability and porosity values being derived from uniaxial consolidation theory.

Formation factor (and compressional wave velocity in the clays) exhibit close interrelationships with permeability, and the capability of predicting permeability to within an order of magnitude both empirically and theoretically (using Biot's equations) is shown. Electrical formation factor is shown to help in defining a mass coupling factor for this theoretical model.

Formation factor is related to porosity for both sands and clays. While any one sample is best represented by Archie's (sands) or Winsauer's (clays) empirical law, the overall trend is a 3rd degree polynomial; particle shape appears to dominate both porosity and permeability relationships with electrical resistivity.

Thermal conductivity exhibits a dependence on porosity for saturated sediments. The successful prediction of thermal conductivity using a geometrical model requiring volume and thermal conductivity values has been demonstrated for up to 3 components, and for a variety of particle shapes. Using the common relation, porosity, thermal conductivity may be related to formation factor for both sands and clays.

Anisotropy is shown to exist in sands and clays for some of the energy transfer processes measured. The recognition and subsequent measurement of these phenomena may improve the already good correlations described above. Things derive their being and nature by mutual dependence and are nothing in themselves.

Nagarjuna.

For Hélène, Matthew, Zoë, and Natasha.

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The author is indebted to his supervisor, Mr.D.Taylor Smith, for his constant support and encouragement throughout this study, and for critically appraising the original manuscript.

An examination of this nature, involving the design and construction of laboratory cells and geophysical measuring equipment, inevitably calls on the resources of a large number of colleagues and support staff. The many contributions to the project, largely by members of the Marine Science Laboratories and the School of Electronic Engineering Science within the UCNW are innumerable and varied in extent, but the total help afforded is considerable and is gratefully acknowledged by the author.

Amongst the more important contributions the following are particularly thanked;

Mr.F.C.Dewes for his help in designing the cells, and particularly for his enduring practical support during the deep sea clays testing programme. Thanks are also due to him for drawing the diagrams for this thesis.

Mr.P.Ogden for designing and constructing the electronics associated with the deep sea clays testing programme (see Appendix C), particularly the needle probe, and for practical support during the data collection. Mr.A.R.Owens for designing and constructing the electronics for the needle probe used on the sand samples.

Messrs. D.Gill and F.Jacques, together with the late Mr. J. Byng, for constructing and modifying many cells, many times.

Dr. R.Penny (Oxford University heat flow group) and Dr. J.R.Bloomer (British Petroleum heat flow group) for use of the Oxford microprocessor needle probe and help in calibrating the sands samples needle probe.

Dr. P.D.Jackson (Institute of Geological Sciences) for informative discussions on the use of his porosity cell and the design of laboratory cells for making electrical resistivity measurements.

Dr. P.J.Schultheiss (Institute of Oceanographic Sciences) for supplying the deep sea sediment samples, and for help in constructing the shear wave transducers in the modified oedometer cell.

This research has been financially supported by a NERC studentship, a student grant from British Petroleum, and a joint research contract from the European Atomic Energy Commission and the Department of the Environment. All of these sources are gratefully acknowledged.

Finally, the author wishes to thank his family for their constant support over a long period.

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1. INTRODUCTION

The search for a suitable repository in which high level radioactive waste can be safely confined over long periods of time has led to renewed interest in the geotechnical properties of the deep ocean floor. Meanwhile, the continued search for economic mineral resources has directed attention towards increasing the extraction efficiency in the exploitation of hydrocarbons, together with the modelling of hydrocarbon formation processes; these currently focus on continental shelf deposits and porous media generally. The main body of research associated with the geotechnical state of the seafloor has concentrated over the past twenty years or so on identifying the nature of the material for foundation purposes in association with the exploitation of mineral resources; the majority of this data pertains to the oil industry located on the continental shelf areas of the oceans.

In attempting to identify a suitable location for the disposal (or storage) of high level radioactive waste, where the body is firstly heat generating as the radioactive decay continues, and secondly the body's casing can only be regarded as a temporary shield for the containment of the waste, two aspects of the geotechnical state of the sediment mass are of special importance, the thermal character and pore space nature. These same two features are also of importance in the renewed interest in improving hydrocarbon exploitation, as well as with the proposed laying of sensitive fibre optic communication cables on the seafloor in 1984.

The thermal nature of a medium is characterised by the thermal diffusivity, being the ratio of the thermal conductivity to the volumetric heat capacity. It is thus possible to determine the thermal diffusivity through the individual measurement of these two parameters; the volumetric heat capacity is easily assessed from density and water content (or porosity in a saturated sediment) determinations (Bullard, 1954; Hillel, 1980), while the thermal conductivity may be measured using the transient needle probe technique (Von Herzen and Maxwell, 1959) either in the laboratory or in-Since the volumetric heat capacity is comparatively situ. easy to assess if the porosity is known, attention is focussed on the measurement of the thermal conductivity . All direct thermal measurements suffer from problems of ambient temperature fluctuations, while the same problems of sampling applicable to the assessment of the pore space apply equally here; these problems are discussed in detail below.

A simple approach to a marine sediment is to consider the medium as an assemblage of grains, the pore spaces between which are filled with a pore fluid. For marine sediments generally this pore fluid consists largely of seawater, although, under certain circumstances, quantities of gas may also present themselves. The proportion of space taken up by the pores is referred to as the porosity, being the ratio of the volume of voids to the total volume. This quantity defines in simple terms a facet of the physical state of the material and is of prime importance in many geotechnical and geophysical analyses. The relative quantity of space

contained in a sediment can affect the consolidation characteristics, and indeed the stability of a material, while both the propagation of compressional waves and electrical currents are dramatically affected by the nature of the medium contained therein. The major drawback with porosity, if any, is its static, scalar nature. The porosity of two sediments, one isotropic, one anisotropic, may be equal, although many of the geotechnical and geophysical parameters will exhibit variations according to the direction or orientation of the measurements within that framework. A parameter which extends the concept of the pore space in terms of its distribution and interconnections is the permeability of the medium.

Permeability refers to the ease with which a fluid flows through a medium under an applied pressure gradient. Its accurate assessment is complicated by many factors. In-situ, costs are exceptionally expensive, and results may be prone to bias if there is fissuring or faulting in an area. Laboratory measurements suffer from requiring a representative sample of the in-situ condition of the material. It is difficult to obtain either a sufficient sampling density to reproduce the in-situ variation of material, or to obtain an undisturbed sample of the sea floor (Bailey, 1971). This latter point is particularly poignant where removal involves dramatic variation in temperature and pressure (Noorany, 1973). Pertinent to both areas of measuring fluid flow is the additional controversy regarding the magnitude of the hydraulic gradient used in determining permeability. Pane et

<u>al</u>, (1983) in summarising the situation, state that Darcy's law has been challenged extensively, particularly with reference to soft clays of low permeabilities. They point to Olsen (1966) as identifying the problem of applying Darcy's law to certain clay sediments, whilst noting its applicability generally, and to Mitchell and Younger (1967) who severely question the applicability of measurements using hydraulic gradients far in excess of field gradients, especially since deviations from Darcy's law are most severe at low gradients typical of those encountered in-situ.

Electrical flow in marine sediments has been considered both theoretically and experimentally, and relationships between porosity and electrical formation factor shown to exist (Schopper, 1966; Boyce, 1968; Kermabon<u>et al</u>, 1969; Jackson, 1975a,b; Jackson <u>et al</u>, 1978; Mendelson and Cohen, 1983). The extension of electrical flow as being analagous to fluid flow (permeability) has been noted frequently and various attempts at relating the two exist (Archie, 1942; Schopper, 1966; Brace, 1977). Much controversy reigns over the role of the pore space in defining the flows of fluid and electric currents (Dullien 1979), for while a dependency is ubiquitously proposed, the precise scale and nature is illdefined.

A major aim of the current research into the disposal of high level radioactive waste is that of initial containment, and to this end it is necessary to identify the geotechnical condition of the sea floor not only at its surface boundary but also with depth. Conventionally, depths in excess of a

few metres are only accessible by drilling, and whilst the past decade has witnessed the success of Glomar Challenger in doing so, the areal coverage so accomplished is strictly limited. In the continental shelf environment, there has been considerable success in developing geophysical techniques capable of identifying the geotechnical state of a material. Bennell et al (1982) have demonstrated the usefulness of sea floor probes in measuring electrical resistivity, compressional wave speed and shear wave velocity, while Whitmarsh and Lilwall (1982) have shown the feasibility of measuring transverse waves over larger distances on the deep The relationships of these geophysical ocean floor. parameters with the geotechnical state of the material are well documented elsewhere (Taylor Smith, 1974; Hamdi and Taylor Smith, 1982; Schultheiss, 1981). With the exception of Hamdi and Taylor Smith (1982) whose work is referred to in detail later and to an extent forms the basis of some of this research programme, none of the attempts at interrelating the geotechnical and geophysical properties of sea floor sediments have been directed towards identifying the permeability or thermal conductivity of a sample through geophysical There is, however, a certain amount of work on measurements. permeability relating to the assessment of lithified sediments in aquifers where limited success has been found using either electrical resistivity formation factor or compressional wave speed (Worthington, 1973; Barker and Worthington, 1973). On thermal conductivity assessment various relationships have

been shown to exist with the pore space volume and the nature and proportion of pore fluid present, although the only attempt at relating thermal and electrical conductivity for marine sediments (Hutt and Berg, 1968) used a two-electrode system with little success.

Since the problem of containment initially requires an assessment of the sea bed both laterally and vertically, preferably using in-situ techniques which create as little disturbance as possible to the material, the potential role of geophysics as a non-destructive tool is apparent. The success enjoyed previously in various in-situ shallow water environments and laboratory simulations lends to its application in the definition of the geotechnical state of the deep sea floor. However, for this to be realised it is firstly necessary to define in the controlled confines of the laboratory some useful relationships between geophysical parameters, and permeability and/or thermal conductivity; these relationships may be empirical or theoretical providing they enable a valid working prediction of the relevant parameter to be made.

The aim of this research project, carried out under a variety of contracts and grants, is to examine the interrelationships, if any, that exist between electrical formation factor, porosity, and thermal conductivity for marine sediments. The use of any such relationships, as noted earlier, may be far-reaching, and could possibly form the basis for predicting thermal conductivity or permeability by remote geophysical means. Permeability, it should be noted,

creeps into this study under the auspices of its similarities with porosity, being itself dependent on the nature of the pore space, and also its analogy with electrical flow in porous media. It is, however, present only as an extension to the porosity concept and thus only as a means to an end, and not as a subject in its own right. To achieve this study, marine sediments have been grouped into clean sands and clays, in many ways two extremes of the sediment spectrum. A series of laboratory cells have been modified or designed to enable the simultaneous measurement of a range of geophysical and geotechnical parameters in an attempt at elucidating some interrelatioships between the three stated parameters. For the clean sands a suite of thirteen artificially prepared samples and a suite of four natural samples have been tested, while ten deep sea surface clays and four deep sea drilling project specimens from depth within the sediment column have also been tested. Although in attempting to define the role of the pore space in terms of the electrical and thermal character of marine sediments, sound speed and shear wave velocity have been measured, these parameters are only presented as necessary adjuncts to the main purpose of the study.

2. BACKGROUND

2.1 <u>Permeable Porous Media</u>

A marine sediment may in its simplest form be considered an assemblage of grains, the pore spaces between which are filled with a pore fluid. The grains range from discrete physical particles with little chemical or electronic bonding between them (e.g. sands) to complex lattice arrangements of particles where chemical and electronic bonding may be important (e.g. clays). The pore fluid may vary from saline water to mixtures of saline water and gaseous phases; these gaseous phases may be in or out of solution.

Generally, in an unconsolidated sediment which shows little sign of any lithification processes the pore spaces are considered to be interconnected. The presence of voids separated from other neighbouring voids in such sediments is considered unlikely and hence all of the voids may be assumed as playing some relative role in controlling fluid flow (permeability) through the sediment. Additionally, the absence of any lithification processes denotes the lack of any physical cementation between the individual grains; while this is particularly true for sand-like sediments, problems arise with fine clays especially in the deep sea where 'rigid' bonding due to the long time constants associated with the very slow rates of deposition may produce a characteristic response similar to that expected with cementation.

2.2 <u>Geophysical Measurements in Marine Sediments</u>

Geophysical measurements on marine sediment samples, both in-situ and in laboratatory studies, have enabled various relationships to be evaluated with a range of geotechnical properties (Shumway, 1960; Boyce, 1968; Buchan <u>et al</u>, 1967; Hamilton, 1970; Jackson, 1975a). Of the wide range of geophysical parameters measured there are, perhaps, three which have proved both of use and of general applicability.

(i) Compressional wave (P-wave) velocity, or sound speed is by far the most commonly encountered, its extensive use as a search tool in the underwater and hydrocarbon industry lending emphasis to its application. The general consensus of experimental results shows the magnitude of the P-wave is primarily a function of the nature and quantity of the pore fluid present. The introduction of a very small proportion of gas into the pore volume dramatically reduces both the amplitude and velocity of propagation (Anderson and Hampton, 1980).

(ii) Shear wave velocity measurements in unconsolidated sediments have proved to be both difficult and controversial, with much discussion relating to the precise nature of the received waveform. The past five years, however, has seen dramatic improvements. On land, in-situ measurements have been developed as a routine site investigation tool (Davis and Schultheiss, 1980) while measurements using pulse techniques applied to 'bimorph' bender elements enable shear waves to be monitored in the laboratory (Shirley and Anderson, 1975;



Schultheiss, 1981; Hamdi and Taylor Smith, 1982). In the marine environment Schwarz and Conwell (1974) provided a similar in -situ arrangement but with apparently limited success, whilst in-situ probes enabling measurements over short distances have been successfully deployed in a variety of sediment types (Bennell <u>et al</u>, 1982). The overwhelming evidence from these various measurements is that the shear wave is in practice independent of the pore fluid, being primarily a function of the rigidity of the mineral framework. Any liquid present seems to act as a lubricant, slightly reducing the magnitude of the velocity.

(iii) Electrical resistivity measurements have been used ubiquitously in the oil industry for predicting the porosity of a reservoir rock (Archie, 1942,1950; Guyod, 1944; Schlumberger, 1951) and for determining the nature of the pore fluid. This success is based on the premise that electrical conduction in rocks and sediments takes place through the pore fluid, the grains themselves being insulators in the absence of any ore minerals (metallic conductor). The measured electrical resistivity is thus a function of the nature and distribution of the pore fluid; for marine sands this dependence on pore shape (dictated by particle shape) has been clearly demonstrated (Jackson <u>et al</u>, 1978).

Although for cohesionless sediments the assumption of non-conducting particles is generally valid, the same cannot be said of clays. The charged particles inherent on the surface of clay particles can provide solid conduction paths within the sediment mass. The effect of this phenomena is to

reduce the overall resistance of the sample and hence reduce the value of the formation factor; this reduced value is known as an apparent formation factor, as opposed to the true formation factor. Additionally, the charges present on the surfaces can lead to the formation of complex lattice arrangements of the particles; one of the features of these structures is the ability of the charged particles to adsorb water on to their surfaces, effectively increasing their size and hence reducing the size of the pore channels. The ability to adsorb water and the magnitude of the surface charge are directly related, though while the former reduces the hydraulic conduction path the latter increases the electrical conduction path. Consolidated sands containing noticeable proportions of clay occur in the hydrogeological industry, and the technique there is to resaturate the rock in the laboratory with solutions of varying salinity so as to define the true formation factor (Worthington & Barker, 1972). With unconsolidated clays, however, resaturation with a different salinity solution will affect the fabric and nature of the sample so this technique is not directly applicable. However, the effect of the conducting matrix decreases with increasing pore fluid salinity, and for pore fluid salinities similar to those of standard seawater (35%) the effect is generally attributed to be negligible (Brace et al, 1965; Kermabon, et al, 1969; Barker, pers. comm. 1982).

Since electrical resistivity measurements depend on both the nature and distribution of the pore fluid, there appeared

potential in using the measurement to define the pore structure; this necessitates the removal of the dependence of the electrical resistivity on the nature of the pore fluid. To accomplish this the electrical resistivity of the sample is normalised with respect to the electrical resistivity of the pore fluid, and a Formation Factor (FF) is defined:

The formation factor represents the resistance to electrical flow through a sample; where the grains are insulators relative to the conducting properties of the pore fluid, the formation factor is independent of the nature of the pore fluid.

2.3 <u>Permeability</u>

The basic law concerning laminar fluid flow in homogeneous porous media is Darcy's law:

$$Q = \frac{\emptyset A (h_2 - h_1)}{L}$$
 (2.2)

where \emptyset is a constant (permeability) Q is the total volume of fluid percolating in unit time L is the length of sample of cross-sectional area A h_2-h_1 is the hydraulic head

The constant Ø depends on the properties of the fluid and of the porous medium and has the units of velocity. To remove

the effect of the fluid properties such that the coefficient is simply a function of the porous medium we may define the specific or absolute permeability K' as;

$$K' = \frac{\partial \eta}{\gamma}$$
(2.3)

 η = viscosity of pore fluid γ = unit weight of pore fluid The units of this specific permeability are in area (cm²).

The validity of Darcy's law has been widely tested and is generally accepted for the majority of natural soil bodies. Taylor (1948) shows that exceptions do occur for uniform particle diameters greater than 0.5 mm, while Scheiddeger (1960) notes discrepancies occurring at high velocities. Since deviations from Darcy's law are most severe at low gradients, such as those in nature, the validity of applying the results of a laboratory analysis where gradients tend to be excessively high to the field situation is queried by Mitchell and Younger (1967). They also query the use of permeability values from consolidation tests where the developed gradients are very high. This raises the discrepancies evident when comparing the permeability values measured directly (see Akroyd, 1964; British Standard 1377 for methodology) and those derived from oedometer theory (see oedometer section 5.1; Taylor, 1948; Lambe and Whitman, 1979). Generally those derived from oedometer theory are between one and two orders of magnitude less than those measured directly. (Bryant et al, 1981; Hamdi, 1981; Schultheiss, pers. comm.

1983). Whilst there is no definitive explanation of this phenomenon, various experimenters have pointed to the difference in the state of the samples being tested, particularly the strain imposed in the oedometer. This strain, while creating a large hydraulic gradient which would tend to produce a high value of permeability, is countered by the particle rearrangement which may tend to reduce the flow channels resulting in a decrease in permeability.

2.3.1 Predicting the coefficient of permeability

Many attempts at elucidating some quantitative relationship between the permeability of a material and its geotechnical and geometrical properties have been made, both experimentally and theoretically. Loudon (1952) provided a good review of much of the earlier work; Hazen's formula empirically related the "effective" grain size to the permeability for filter sands of grain size 0.1 - 3.0 mm.

$$\emptyset = C(d_{10})^2$$
 (2.4)

where d_{10} is defined as the size such that 10% of the material is of smaller grains. C = constant (approx. 100). Slichter (1899) used the geometry of the voids for uniform spheres to derive a formula;

$$\emptyset = \frac{771 \ d^2}{C}$$
 (2.5)

where C is inversely related to porosity.

Terzaghi (1925) extended Slichter's formula to cover sand of non-uniform grain size and variable grain shape:

$$\emptyset = \frac{C}{\eta} \left(\frac{n - 0.13}{\sqrt[3]{1 - n}} \right)^{2} \cdot d_{10}^{2} \quad (2.6)$$

 η = viscosity of water n = porosity

Krumbein and Monk (1942) devised an empirical equation for sands depending only on the mean grain size and the spread of sizes present;

 $\emptyset = 760 d_m^2 e^{-1.31\sigma_{\phi}}$ (2.7)

Loudon (1952) experimented with a variety of quartz particles (glass beads, crushed glass, and natural sands), defining a linear relationship between log Ø and n;

 $\log_{10}(\emptyset S^2) = b.n + constant \qquad (2.8)$

S = specific surface area b = 5.15 (constant for all particle types)
In doing so, he also emphasised the error margin in permeability predictions, since a 0.01 change in fractional porosity for one sample represented a 13% change in permeability.

Various theoretical models have also been proposed in attempts to predict the permeability of porous media. Detailed accounts of these mathematical and physical analogies are given by Scheidegger (1960), Bear <u>et al</u> (1968) and Dullien (1979). Generally such models are based on the capillaric equation of Poiseuille; the equation is subsequently modified to account for the deviation of the pore channels from straight circular pipes, and then combined with Darcy's law to evaluate a coefficient of permeability. Scheidegger (1960), whilst providing a comprehensive review of such models, emphasises that in general, natural porous media are extremely disordered, so that it seems a rather poor procedure to represent them by something which is intrinsically ordered.

A simple version may be briefly considered whereby an average pore channel is represented as having a cross sectional area a and length L' where L' is greater than the length L of the sample (Carman, 1948; Wyllie and Spangler, 1952). The result of such an analysis is the Kozeny - Carman equation;

$$\emptyset = \frac{m^2 n}{k_0 \eta} \left(\frac{L}{L'}\right)^2$$
(2.9)

where

k_o is a constant (2.0 - 2.5)
(L/L') is the hydraulic tortuosity
 m is the hydraulic radius
(m² is a measure of pore channel area)

Wyllie and Gregory (1953) expressed the equation in the form;

$$\emptyset = \frac{n^3}{(1-n)^2} \cdot \frac{1}{s^2 k_0} \cdot \left(\frac{L}{L}\right)^2$$
(2.10)

where S is the specific surface area. The value of $k_0(L/L')^2$ being the Kozeny - Carman constant, dependent on the shape of the particles.

Many other models have been defined, generally of increasing complexity, and reference to Scheidegger (1960) and Dullien (1979) should be made for a thorough analysis of these. The main characteristic of the models and empirical relationships, however, which consistenly recurs is the role of the pore space in somehow controlling the permeability. Garcia - Bengochea et al (1979) used pore size distribution to predict the permeability of compacted mixtures of silt and kaolin. They point out that empirical predictive equations between permeability and pore volume, whilst useful in sands, are of limited value with cohesive materials. Pore shape was used in association with electrical formation factor measurements to predict the permeability of granite (Brace, 1977). Graton and Fraser (1935), whilst acknowledging the role of the pore space, emphasise that permeability depends on absolute dimensions rather than simply angular relationships, and whilst vitally dependent on the percentage of voids, is governed quantitatively by their size, shape and continuity. Identification of the role of the pore space is perhaps an obvious criteria; its quantitative measurement, however, poses

many problems. Dullien (1975a) uses a network model incorporating two pore size distributions to predict the permeability of a suite of sandstones, measuring the distributions by metal porosimetry techniques, while Garcia-Bengochea <u>et al</u> (1979) use two forms of a capillary model, again using mercury intrusion porosimetry to determine the pore size distribution. These statistical parameters require the use of auxiliary apparatus beyond the scope of this study, and thus render many of the more complex and perhaps rational models unusable within the present terms of reference.

The majority of published data relate to clean cohesionless materials, there being a sparcity of physical property and permeability data combined for cohesive Studies of the effects of consolidation on materials. permeability measurements for soft clays have been carried out by Pane et al (1983). They confirm and re-emphasise the critical role played by permeability in the consolidation process and note that while this role has long been recognised, the current technology and practice for obtaining and applying information on the permeability of soils is in a primitive state. The controversy surrounding permeability measurements in clays and the adherence of fluid flow to Darcy's law have already been mentioned. Bryant et al (1981) consolidated deep sea clay samples in the laboratory and provide clay fabric data to accompany the consolidation results. Their clay fabric data reveal that for an unconsolidated sediment the microstructure is a loose, open,

random arrangement of particles. At 800 kPa load, the structure is a much denser packing of particles with a tendency for the particles to be orientated in packets, but without a high degree of preferred orientation. At 1500 kPa load a high degree of preferred orientation occurs. Similar orientations have been shown to occur naturally at equal insitu pressures by two of the co-authors of that study (Bennett <u>et al</u>, 1977). This suggests that an unconsolidated sample of deep sea clay should have a reasonably open structure, and even at the maximum pressures encountered in this study, the recent sediments will still have a comparatively open structure with little gross alignment of the particles.

2.4 <u>An Analogy between Fluid Flow and Electrical Conduction in</u> <u>Porous Media</u>?

Fluid flow in a porous medium is controlled by the pore channel configuration; it is thus a function of the interconnection of the pore spaces, which in an unconsolidated sediment are themselves dictated by the shape, size and distribution of the particles present. Electrical conduction generally occurs through the fluid present and similarly is a function of the arrangement and shape of the pore structure (electrical relationships with porosity are but one expression of this). It seems therefore that an analogy does exist between the two and hence a relationship between electrical formation factor and permeability may be expected.

This analogy has long been considered as theoretically

applicable to soils, although the application in practice has generally been disappointing. Archie (1942) showed a correlation between the two properties in the form of;

$$FF = b \emptyset^{-m}$$
 (2.11)

where b and m are empirical constants peculiar to the particular rock under study. Whilst similar to Archie's equation relating electrical formation factor and porosity, he did note that his equation represented a trend or average line and that significant errors in quantitive applications must be tolerated.

Archie's law relating electrical formation factor and porosity

$$FF = n^{-m}$$
 (Archie 1942) (2.12)

was later modified by Winsauer et al (1952);

$$FF = Cn^{-m}$$
(2.13)

where C is an empirical constant. This empirical equation produced a better fit to their data for brine saturated sands although it violates the boundary condition of FF = 1 at n = 100% for values of C \neq 0. Although they made no direct measurement of permeability, they did examine the tortuosity of the samples, concluding that the values obtained for electrical and fluid flow were not in agreement with each

other.

Wyllie and Spangler (1952) took the simple capillary model described earlier and applied it to the flow of electric currents in porous media, producing;

$$FF = \frac{1 L'}{n L}$$
(2.14)

where (L'/L) is the electric tortuosity. Combining this equation with Archie's law produces;

$$\frac{L'}{L} = n^{-(m-1)}$$
(2.15)

suggesting that the electric tortuosity depends on the value of m, and the porosity of the sample. Recent studies of the variation of the constant m for sand samples deposited in a laboratory cell have shown a clear relationship with particle shape; m varies from 1.4 to 2.0 for spherical to plate-like particles with an average of 1.5 for the majority of sands (Jackson et al, 1978).

Schopper (1966), in supposing that there must evidently be relations between porosity, permeability, and formation factor, considered both the hydraulic and electric tortuosities (T,X). Whilst he noted certain similarities between electric conductance and hydraulic flow (the dependency on the duct length being the same for both processes), he also identified an essential difference in the dependency on the pore channel cross-section, there being a

linear dependency with the electric process, a quadratic dependency with the hydrodynamic process.

Dullien (1975b) also considers the two tortuosities, concluding that the electric parameter, X, is incorrectly termed and relates to additional effects not necessarily connected with the tortuous nature of the conduction path. By expressing the idea in terms of a simple model of a porous medium consisting of identical parallel capillaries with periodic step changes in their diameter, he shows how the hydraulic tortuosity remains constant at unity, and the electric tortuosity increases as a function of the ratio of the large to small diameters of the capillary segments. Thus while a narrow constriction will affect the porosity or effective path length only minimally, the electric tortuosity (X) will increase. X therefore measures effects unconnected with the tortuous nature of the path and should not be called tortuosity. Dullien refers to Schopper (1966) as an example of the typical confusion concerning the two tortuosities, where the reduction in area is equated to the presence of a matrix (particles) as opposed to 100% porosity conditions, rather than the reduction in channel area due to constrictions. Effectively, the true area of channel cross section may be much less than that defined by porosity due to the presence of constrictions.

Jackson (1975a) conducted a brief analysis of the flow of electric currents and liquids in simple pore shapes. In comparing the electric and hydraulic tortuosities (X,T) for four simple channel geometries, representative in part of real

porous media, results showed that the hydraulic tortuosity (T) is generally greater than the electric tortuosity (X). No theoretical comparison could really be made because a complete analysis is impossible, due to the effects and processes which cannot be satisfactorily assessed (pore geometry, turbulence and distortion of the electric field). The results showed that even for the simplest of shapes no universal FF - Ørelationship existed. This is emphasised experimentally in that shelly sands and clays may both exhibit formation factors of 2 but in terms of permeability values are at opposite extremities $(10^{-3}$ to 10^{-9} m/sec). Jackson's study did show, however, that both Ø and FF are modified in the same way, suggesting that Ø should be single-valued for any particular value of FF for any one sample; thus empirical equations should be useful, and indeed preliminary tests on artificially mixed sands bear agreement (Fig. 2.1, reproduced from Hamdi and Taylor Smith, 1982).

Worthington (1973) established relationships between electrical formation factor and both porosity and permeability for a sandstone aquifer. The porosity relationship followed that of Winsauer (eq. 2.13), while the permeability relationship with formation factor was along the lines of that proposed by Archie (eq. 2.11). The accuracy of the permeability prediction from the measurement of the formation factor was considered to be within an order of magnitude.

Brace (1977) showed that for granite a reversed form of Archie's equation could be applied;



(Lovell 1981, reproduced from Hamdi and Taylor Smith, 1982).
$$\emptyset = \left(\frac{h^2}{k_0}\right)_{FF} - 1.5$$

(2.16)

where k_0 is a constant and h is the hydraulic radius. The main problem identified in this study concerned the identification of the pore space configuration, particularly with depth. Brace showed that for similar values of electrical resistivity rocks may exhibit differences in permeability by a factor of 100.

Brown (1980) in considering the fluid-solid coupling for acoustic waves in fluid-filled porous media drew an analogy between fluid flow and electrical flow in comparing the differential equations for each. He concluded that the two kinds of flow satisfied the same differential equation with the same boundary conditions, giving the same "formation factor" to describe the effect of the geometry of the porous medium on the macroscopic fluid or electrical flow.

A theoretical study of the effect of grain anisotropy on the electrical properties of rocks by Mendelson and Cohen (1982) considered the state of Archie's law in defining electrical conduction in porous media. They suggested that while for randomly orientated grains the conductivity is a scalar quantity (eq. 2.17), for orientated (i.e. anisotropic grain distribution) the quantity is a tensor (eq. 2.18).

$$\sigma = \sigma_{u} n^{m} \qquad (2.17)$$

 $\sigma_j = a_j(n) \sigma_w n^m$

(2.18)

More importantly, perhaps, they show that different distributions of grain orientation can lead to the same effective conductivity; they thus conclude that measurements of conductivity are not sufficient to determine grain distribution. The effect of this conclusion is to reject any unique $FF - \emptyset$ relationship, since \emptyset appears to be primarily a function of pore shape and size distribution which is effectively governed by the grain distribution. This may well explain the disappointing results from investigations into $FF - \emptyset$ relationships, though it is obviously in contrast to the work and results of others.

Support for Mendelson and Cohen comes indirectly from Dullien (1979) who cites experimental and theoretical work by El-Sayed (1978). El-Sayed quotes differences in computed and measured electrical formation factors due to the difference in scale between the two phenomena of electrical flow (measured) and fluid flow (used to compute the formation factor). The medium tested was a sandstone where the macrostructure primarily contributed to the fluid flow with the microstructure (particularly small pore spaces) effect negligible; the reverse proved true for electrical conduction. For this result to be extended to other types of porous media an initial constraint is the presence of a very wide pore This difference in scale between the size distribution. hydraulic flow and electric conductance is raised earlier by Schopper (1966) who concludes that one of the three quantities defining the two tortuosities is the staistical distribution of channels and suggests that this may affect the two by

different amounts. A similar conclusion was reached by Jackson (1975a) in his brief study of tortuosities.

2.5 <u>Permeability and Compressional Wave Velocity</u>

Compressional wave measurements, both in-situ and in the laboratory, have shown the dependence of the speed of propagation in porous media on the nature and distribution of the pore fluid. At a given porosity the speed of propagation falls considerably if a small quantity of the pore fluid is replaced by gas, whilst for a saturated sediment an inverse relationship between the speed of propagation and porosity is well documented (Nafe and Drake, 1957; Buchan <u>et al</u>, 1971; Boyce, 1976). Electrical flow which is also dependent on the pore fluid phase of a sediment has been shown in the previous section to exhibit some form of analogy with fluid flow, and sound speed is known in foliated clay materials to exhibit anisotropy (Fig. 2.2), though whether this is due to a difference in the fluid flow arrangement of the sediment structure or its elasticity is difficult to determine.

2.5.1 <u>Theoretical model</u>

In describing the static and dynamic response of porous materials under varying conditions, Biot in a series of papers (1941, 1956, 1962a,b) set out in terms of a sediment framework and the interrelationships between it and the fluid contained within, the effect of propagating a compressional wave through



a sediment. The resulting differential equations suggest that the compressional wave velocity increases with frequency and that the increase depends upon the permeability of the medium. To study the role played by permeability Hamdi and Taylor Smith (1982) developed a marine sediment model along the lines first proposed by Gassman (1951) and Biot (1956) and tested this experimentally using an earlier modified consolidation cell. A brief summary of the theoretical model is presented in Appendix A, although the original reference is best consulted if a thorough examination and discussion of the model is required.

The solution to the theoretical model, derived by Hamdi and Taylor Smith (1982), may be expressed as;

$$\mathcal{O} = \frac{ng}{\omega} \left(\frac{\rho}{b\rho - n\rho_{W}} \right) \frac{V_{o}^{2}}{V_{o}^{2}} \left[\frac{V_{p}^{2}}{V_{o}} - 1 - \frac{1}{1 - \left(\frac{V_{p}^{2}}{V_{o}} \right)} \right]$$
(2.19)

where	Ø	= coefficient of permeability in cm/sec
	n	= fractional porosity
	ω	= angular frequency of propagation
	ρ	= bulk or total density
	Pw	= pore fluid density
	Ь	= mass coupling factor
	Vp	measured P wave velocity
v _o	, V 🖷	= theoretical velocities at "zero" and "infinite"
		frequencies
	S i	<pre>= acceleration due to gravity</pre>

In defining the model account is taken of the fact that not all of the fluid moves in the general direction of the pressure gradient by introducing a parameter m; this has the dimensions of density and is a consequence of the shape of the

grains, their interconnection and the form of the cavities. The parameter may be written;

$$m = \frac{b \rho_w}{n}$$
(2.20)

where b is a mass coupling factor which must have a numerical value greater than 1; for b = 1 there is no fluid - solid coupling. A mass coupling factor of 1 thus forms a limiting boundary condition in the absence of an accurate value for b. The results in Fig. 2.3 (dotted lines) are taken from Hamdi and Taylor Smith using a value of 1 for the mass coupling factor, and show that the model predicts the permeability reasonably well. Brown (1980) has suggested that through an analogy between electrical and fluid flow it may be possible to derive a value for b, pertinent to each sample, where

$$b = FF.n \qquad (2.21)$$

The value so derived would provide a lower limit to the value of b, and whilst the magnitude does not vary greatly the model is reasonably sensitive to it. To evaluate the effect of changing the value of b by this technique, formation factor values have been computed for the data presented in Fig. 2.3 by substituting the porosity values into Archie's law (eq. 2.12). The values used for the exponent are generally in agreement with the work on formation factor - porosity relationships (Taylor Smith, 1971; Jackson <u>et al</u>, 1978; and this study); these are listed in Table 2.1. The modified

Measured values.
Predicted values with b = Fn.
:----: Predicted values by Hamdi & Taylor Smith.



Fig. 2.3. Comparison of measured permeability values with computed permeability values from the theoretical model, with b=1 and b=Fn. (After Hamdi and Taylor Smith, 1982).

TABLE 2.1

DATA FOR FIGURE 2.3

Sample No.	e Measured	Permeabili Computed (b=1)	ty Computed (b=FFn)	Porosity n	Formation Factor	Archie Slope
	10 ⁻⁷ m/s	10 ⁻⁷ m/s	10 ⁻⁷ m/s	8	•	•
Fl	3.0	4.5	3.7	0.600	2.50	-1.8
F2	4.6	5.4	4.8	0.549	2.94	-1.8
F3	14.0	6.7	12.0	0.445	4.00	-1.7
F4	20.0	8.9	22.0	0.430	3.34	-1.4
F5	40.0	12.0	42.5	0.393	3.12	-1.3

and the second second

results are also plotted on Fig. 2.3 (solid lines) and show an improved fit, although neither of the set of predictions is far from the measured values.

It thus appears that it is possible to predict the permeability of a marine sediment to within an accuracy of at worst an order of magnitude using the theoretical model; the use of electrical formation factor in defining the mass coupling factor appears to help in precisely defining the magnitude of the permeability.

2.5.2 <u>Empirical relationship between permeability and</u> <u>compressional wave velocity</u>

In the field of hydrogeology the need for some method of quantitatively assessing the yield of an aquifer has prompted the search for simple geophysical techniques of assessing the in-situ permeability. Empirical relationships exist between the permeability of a consolidated sandstone and its compressional wave velocity (Barker and Worthington, 1973), where

$$V_{\rm p} = C - d \log \emptyset \qquad (2.22)$$

C and d are empirical constants.

As with the electrical formation factor - permeability relationships the 'fit' of the curve to the data is but a trend, and considerable errors are likely in any quantative assessment. For deep sea clays the relationship between

sound speed and the pore space is effectively concerned with the porosity; since a small change in porosity can create a large change in permeability (Loudon, 1952) and a small change in porosity can have little effect on sound speed for high values of porosity (Buchan <u>et al</u>, 1971) this error source is likely to be considerable.

2.6 Thermal Diffusivity

In attempting to characterise the thermal properties of a medium the thermal diffusivity is perhaps the most useful and important parameter;

$$D = \frac{k}{C}$$

where $C = \rho c$

(2.24)

(2.23)

D	=	thermal diffusivity
k	×	thermal conductivity
ρ	=	density
С	=	specific heat capacity
С	-	volumetric heat capacity
		—

The thermal dffusivity may be measured directly (Jackson and Taylor, 1965; Jaeger, 1958) or indirectly by determining independently both the thermal conductivity and volumetric heat capacity (Von Herzen and Maxwell, 1959; Bullard, 1954; Hillel, 1980). The volumetric heat capacity of a porous medium can be accurately defined in terms of the relative proportions of the constituents;

 $C = nC_w + (1-n)C_s$ (Hillel, 1980) (2.25) $c = wc_w + (1-w)c_s$ (Von Herzen & Maxwell, 1959)

(2.26)

where C =
$$\rho$$
c, n = porosity
w = water content (by weight)

or

If the porosity of a sample is known, then the thermal diffusivity may be determined by measuring the thermal conductivity of the sample.

2.6.1 <u>Thermal conductivity measurement</u>

The classical technique for measuring the thermal conductivity of a material is the divided bar apparatus whereby a sample is subjected to a one dimensional steady state flow of heat. The thermal conductivity is deduced from the temperature gradient across, and the heat flux through, the sample. There are problems, however, in achieving unidimensional heat flow, and the special preparation of samples for insertion into the apparatus does not lend the technique to marine sediments. A technique more applicable to loose unconsolidated sediments is the transient needle probe, firstly developed for use on industrial insulating materials (D'Eustachio and Schreiner, 1952) and since modified for use in soils (Lachenbruch, 1957) and deep sea sediments (Von Herzen and Maxwell, 1959). The principle of the transient

needle probe technique involves the heating effect of an infinite line source, represented by a hypodermic needle, inserted into the sample. A plot of temperature rise against the logarithm of time enables a value for the thermal conductivity to be attained within minutes.

In-situ, the needle probe technique has been adapted for use (Sclater <u>et al</u>, 1969; Erickson <u>et al</u>, 1975; Sass <u>et al</u>, 1981), and a more rapid pulse technique developed (Lister, 1979). In the laboratory, measurements with a needle probe can produce satisfactory results (Maximum error 5%) in short times (<100 seconds), thus enabling rapid collection and checking of results with little loss of moisture from saturated samples.

For granular soils, De Vries <u>et al</u> (1975) have shown that the thermal conductivity may be predicted by a knowledge of the relative proportions of the constituent minerals and fluids, their thermal properties, and the mineral shapes;

$$k = \frac{x_0 k_0 + b x_1 k_1}{x_0 + b x_1}$$
 (2.27)

where k = thermal conductivity x = volume fraction subscripts o, l refer to fluid, solid respectively

The quantity b is the ratio of the average temperature gradient in the granules and the corresponding quantity in the fluid, and is shown to be a function of the shape and orientation of the particles and the ratio of the

conductivities. While theoretically this model only holds for particles ellipsoidal in shape and sufficiently dispersed so as not to influence each other, de Vries extends it to a variety of soils by applying simple corrections.

Bullard et al (1956) and Ratcliffe (1960) have shown a clear dependence of thermal conductivity on water content, while Sass et al (1971) provide a simple geometrical relationship (eq. 4.4) which allows the computation of the bulk conductivity from the volumes and conductivities of the constituents. In a series of experiments on clean cohesionless sands, Lovell (1981) obtained data which has shown this geometrical relationship to hold (Bloomer, 1981 pers. comm.), and an association, as then undefined, to exist with electrical formation factor (Figs. 2.4 & 2.5). Both of these results form part of this study and are described in detail later. Bruggeman (cited Hutt and Berg, 1968) also provided a formula for the normalised bulk conductivity in terms of the normalised grain conductivity and the porosity of the system (see Table 6.1), while Von Herzen and Maxwell (1959) showed a linear relationship to exist between thermal conductivity and thermal diffusivity (computed from k/C) for deep sea clays.



Fig. 2.4. Thermal conductivity plotted against porosity for sands of different particle sizes and shapes.

Needle horizontal

× Needle vertical



Fig. 2.5. Thermal conductivity plotted against electrical formation factor for sands of different particle sizes and shapes.

3. EXPERIMENTS WITH CLEAN SANDS

The cohesionless samples studied consist of two groups of clean sands. The artificial suite of thirteen samples was constituted to provide a range of particle shapes and sizes, and size distributions, using glass ballotini, commercially available quartz sand, and crushed shells. The natural suite of four samples was selected to provide a range of particles covering some of the features exhibited by the artificial suite. Descriptions of all the artificial and natural sands are presented in Table 3.1

3.1 <u>Sample Descriptions</u> (sands)

Artificial Group

A total of thirteen samples comprising three suites, exhibit variations in grain size distribution or spread of sizes (samples a - d), mean grain size (samples e - h), and grain shape (samples i - n). Samples (a - h) are quartz sands, while samples (i - m) represent a range of carbonate shell / quartz sand mixtures from pure shell to pure quartz sand. Sample (n) is a glass ballotini specimen, introduced to represent spherical particles. These samples have been deposited in the porosity cell and the square permeameter. Grain size histograms for the major components of the group are presented in Fig. 3.1, while the grain shapes are demonstrated in Plate A.

TABLE 3.1

SAND SAMPLES

Artificially Prepared Samples

Sample No.			Mø	σø
Quartz	Sands			
a b c d	increasing spread of sizes		0.90 1.07 0.97 1.10	0.21 0.50 0.61 1.34
e f g h	similar spreads different means	a a a a a a a a a a	-0.29 0.75 1.72 2.74	0.13 0.27 0.35 0.14
Quartz	/ Shell Mixtures			
i j k l m	decreasing shell content	70%i 30%f 50%i 50%f 30%i 70%f 0%i 100%f	0.33	0.26
Glass	Ballotini			
n	spherical particles		0.72	0.45

Natural Samples

Sample No.	Location	Carbonate Content	^м ø	σø
0	Newborough	6.5	2.44	0.30
p	Trearddur Bay	51.4	1.63	0.88
đ	Casha Bay West	78.8	1.36	0.49
r	Casha Bay S.E.	90.8	-0.21	0.50



Fig. 3.1. Grain size histograms for suite of 9 artificial sands.

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Natural Group

Four samples from different locations were chosen to briefly exhibit the natural representation of the variations in grain shape. They range from a rounded quartz sand (Newborough), a shelly quartz sand (Trearddur Bay) through to a guartz - shelly carbonate sand (Casha Bay West) and an almost pure shelly carbonate sand of irregular plates/spindles (Casha Bay South East). The Newborough and Trearddur Bay sands are from the Isle of Anglesey, situated off the north west coast of Wales, while the two Casha Bay sands are from Galway, Eire. All four have been deposited in each of the three cells. Where possible the sampled sands have been used in their natural state, although a small percentage of very fine particles was removed from the Newborough sand since it is not possible to adequately redeposit such a range of particles in the present experimental arrangement due to the difference in their settling times. The Trearddur Bay sand and the Casha Bay South East sand were both sieved through a 4mm mesh to remove whole shells from the sample such that any major inhomogeneities could be removed from the deposited sands. Obviously, both of these procedures could alter the structure of the sands and hence their physical charcteristics and the results obtained, though under the experimental techniques adopted in this study such measures were considered necessary. Fig. 3.2 presents the grain size data, while Plate B shows the grain shapes, for these four sands.









Fig. 3, 2. Grain size histograms for natural sands.



3.2 Laboratory Cells

The packing of sands by various depositional techniques in the laboratory was closely studied by Kolbuszewski (1948a,b) who noted that the value of maximum porosity would be increased by up to 10% when using containers smaller than 76 mm in diameter. He further showed that by pouring a sand in fluid without any entrapped air the sample achieved a loose packing state whereby the exhibited porosity tended to a maximum value. Using such a depositional technique Jackson (1975a,b) developed a novel porosity cell which enabled a deposited sand to be compacted through a known series of porosity values from its maximum towards its minimum. The cell contained electrical resistivity electrodes and differed from previous cells (Kermabon et al, 1969; Erchul and Nacci, 1972) in that the staged settling of the sample could be monitored to produce a sequence of porosity - electrical data as the sand became increasingly dense. This cell is referred to here as Jackson's Porosity Cell and forms the basis of the laboratory measurements on sands, and the development of two permeameter cells as detailed below. It should be noted that both permeameter cells may be filled such that the sample exhibits an initial loose packing which may be compacted in stages, as in the porosity cell.

3.2.1 Jackson's porosity cell

The cell is machined out of solid perspex (Plate C) and is shown diagrammatically in Fig. 3.3. It consists of a

PLATE C

The Cylindrical Permeameter

The Square Permeameter Cell

Jackson's Porosity Cell

封書











Fig. 3. 4. The electrode plate.

rectangular chamber some 100 mm square and 75 mm tall; the volume then tapers upwards at an angle of 45° to a cylinder 40 mm in diameter and 250 mm tall. This cylinder serves as both a sediment reservoir and a volume measuring device; since the cross-sectional area of the cylinder is approximately one eighth that of the cell chamber, a volume change in the chamber is amplified about eight times in the cylinder. This assumes that there is no measurable variation in packing between the chamber and the cylinder; to minimise the effect of any discrepancies the cylinder is filled to the minimum height sufficient for a compaction run. However, because of the difference in cross-sectional area between the two, piping may occur as the electrolyte is displaced by sediment particles; to prevent this a by-pass system is incorporated in the design of the cell which allows the electrolyte to pass from the sloping roof of the chamber to the top of the cylinder, any sediment being contained in the chamber by 0.063 mm sieve mesh. The cylinder is graduated every 10 cm^3 which enables a resolution of fractional porosity between 0.0015 and 0.0035 depending on the type of sediment.

The porosity of the sample at any stage during its compaction may be determined from its dry weight and specific gravity, together with the volume occupied.

The electrical resistivity of the sample is determined using a four-electrode array mounted in two electrode plates on opposite sides of the cell chamber. Each plate contains a potential electrode and a current electrode. The plates are

composed of a matrix of stainless steel screws (figure 3.4), each electrode constituting the alternate series of screws which are wired together in series to form a plate electrode; the pairs of opposing current and potential electrodes are in turn connected in parallel and the complete arrangement is thus equivalent to a standard four-electrode array.

To enable the measurement of thermal conductivity during a test the cell is slightly modified from its original design to accommodate a needle probe. The probe, which may be inserted either vertically or horizontally into the cell, is described in detail later.

3.2.2 <u>Cylindrical permeameter cell</u>

This cell (Plate C) is modified from a standard soils laboratory constant head permeameter and is illustrated in Fig. 3.5. The modification consists of the introduction of six circular electrodes into the wall of the cell such that any four may be used to measure the electrical resistivity or formation factor. The positions of the six electrodes enables selection of four adjacent electrodes such that the measured electrical parameter approximates to the same region as the permeability measurement between two adjacent sets of manometer outlets.

The electrical measurement obviously relates more to the vertical direction, just as in Jackson's porosity cell the measurement relates more to the horizontal direction. A possible problem here may concern the flow of current at the



Not to scale

Fig. 3.5. Schematic section through cylindrical permeameter.

interface between the sample and the cell wall, a feature not encountered in the previous porosity cell or the following square permeameter. This would tend to lower the resistance measured as a narrow layer of electrolyte presented an easier path for current transfer. The electrical measurement is accomplished using similar equipment and technique to the porosity cell, and is described later.

Measurement of the permeability of the sample is accomplished using the arrangement outlined in Fig. 3.6. A constant head supply of de-aired saline water is passed through the chamber, and the hydraulic gradient and volume of electrolyte passing monitored over a fixed time interval. The permeability (which relates to the vertical direction) is then computed from Darcy's law (eq. 2.2), (see Lambe, 1964; Akroyd, 1964 for method).

3.2.3 <u>Square permeameter cell</u>

The square constant head permeameter cell (Plate C) was purposefully designed for this study to combine the effectiveness of the square electrodes, originally designed by Jackson for his porosity cell, with the ability to measure the permeability directly. It effectively constitutes a hybrid of the two previous cells in that the permeability is still measured vertically, but the electrical measurements are horizontally orientated. The cell cross-section is of similar size and shape to the porosity cell, as are the electrode plates. Additionally it is possible to introduce a thermal



Fig. 3. 6. Block diagram of the permeameter experiment.

conductivity needle probe into the cell horizontally. Fig. 3.7 shows the cell schematically, while Fig. 3.8 shows a plan of one of the electrode plates. The electrode plate is similar in operation to the matrix electrodes of the porosity cell (Fig. 3.4); the potential electrode still consists of a number of screws, wired in series, across the face of the plate, although the current electrode is the background surface through which the potential screws protrude. This surface is electrically conducting silver paint and is insulated from the potential electrode. This modified design follows that used by Jackson in a triaxial cell (Schultheiss, 1982).

All measurements (electrical, thermal, and permeability) are made as for the previous cells.

3.3 <u>Geophysical Measuring Systems</u>

3.3.1 <u>Electrical resistivity</u>

The electrical resistivity measuring system consists of a four electrode array; two current electrodes and two potential electrodes. To eliminate the problem of polarisation at the current electrodes a constant current of known magnitude is electronically swiched to produce a low frequency alternating current of 4 Hz (ABEM AC Terrameter) or 0.4 Hz (ABEM SAS Terrameter). This system avoids the problem of any contact impedance at the current electrodes and eliminates any D.C. offset at the potential electrodes.

The electrical formation factor of the sample under test



Not to scale

Fig. 3. 7. Schematic section through square permeameter.



Fig. 3.8. Electrode plate (square permeameter).

neres la contriend inside, a maintain diest hypersons candle of assestal diseaser 1.0 mc. Thailes the shole interaction these the shole interaction with runs the shole interact and a base thereining is interact at the sid-parts. A interact outrept severes of 110 ms provides the memory size alls a consecut and inter while the provides to machteral strengt may be simply computed, being the ratio of the resistance of the sample to the resistance of the electrolyte for the same cell geometry. Should the electrical resistivity of the sediment or electrolyte need to be computed in preference to the formation factor, the geometric constant must be determined. This procedure is described in detail by Jackson (1975a).

3.3. Thermal conductivity

The thermal conductivity of the sample may be determined at any pause in the compaction process using a needle probe inserted through the side wall (porosity cell and square permeameter) or base (porosity cell).

The needle probe (Fig. 3.9) is designed to approximate an infinite line source with a constant heat flux. The theory and limitations of the technique are given in section 5.6 along with a description of a more refined and sensitive probe used in the measurements on clays. For the measurements on sands the probe is more robust, and, since there is no stringent limitation to the amount of heat supplied and therefore the temperature rise incurred, less sensitive. This probe is contained inside a stainless steel hypodermic needle of external diameter 1.5 mm. Inside the tube, (internal diameter 1.0 mm), a heater wire runs the whole length and a bead thermistor is located at its mid-point. A constant current source of 120 mA provides the heater wire with a constant heat flux, while the temperature is monitored through



Fig. 3. 9. Thermal conductivity needle probe (sand samples).
the bead thermistor, the response of which is linearised and amplified before being displayed on the Y-axis of an X - Yplotter. The X-axis of the plotter is supplied with a logarithmic time base over the period 100 seconds such that the output to the plotter is a graph of temperature rise against logarithm of time. The slope is thus indicative of the thermal conductivity of the sample. Fig. 3.10 presents a block diagram of the thermal conductivity measuring system.

As with the needle probe used for the clay measurements the heat output is not calibrated in absolute terms. Instead the slope $(q/4\pi k)$ for any unknown sample is compared with the slope for a ceramic glass of known thermal conductivity; this method follows that of Bloomer and Ward (1979) and allows computation of the unknown thermal conductivity to within an accuracy of 7%. The needle probe technique is based on the assumption that the heating element approximates an infinite line source, supplied by a constant heat flux. Together with the additional constraints concerning the uniformity of the temperature of the sample and the edge effects due to the sample container, certain time limitations are imposed on the needle probe results. These are explained in detail in section 5.6 together with the theory of the needle probe technique. They are computed in a similar manner here for this probe.



Fig. 3.10. Block diagram of the thermal conductivity measuring system. (Sand samples).

Time limitations

Lower time limit

Equation (5.12) holds only for small values of r and large values of t;

$$E_{L} = \frac{r^2}{4D^+} << 1$$

r = probe radius D = thermal diffusivity t = time

The term E_L will thus be a maximum for sediment of low thermal diffusivity. Thermal diffusivity shows a clear relationship with water content (or porosity in a saturated sediment). Two limiting cases may be considered.

Minimum porosity = (Quartz sand)	35%	Maximum porosity = 70% (Carbonate sand)
$D = 1.26 \ 10^{-6} \ m^2 s^{-1}$ r = 0.00075m		$D = 2.83 \ 10^{-7} \ m^2 s^{-1}$ r = 0.00075m
EL	t	EL
	(seconds)	
0.1116 0.0558 0.0223 0.0112 0.0074	1 2 5 10 15	0.4969 0.2485 0.0994 0.0497 0.0331
0.0030	20	0.0249

These results suggest that the equation should hold for times in excess of 5 seconds.

Upper time limit

This time limit is dictated by the edge effects of the sample; these are in turn a function of the thermal diffusivity and size of the sample. DeVries and Peck (1958) provide an equation defining a limit on the dimension of the sample:

$$E_u = exp - \left(\frac{R^2}{4Dt}\right) << 1$$

D = thermal diffusivity
R = radius of sample centred on probe
t = time

Again two limiting cases of extreme sediment type may be considered;

Minimum porosity = : (Quartz sand)	35 %	Maximum porosity = 70% (Carbonate sand)
$D = 1.26 \ 10^{-6} \ m^2 s^{-1}$	L = 100 seconds	$D = 2.83 \ 10^{-7} \ m^2 s^{-1}$
Eu	R	Eu
0.0070 0.0418 0.1677	0.05m 0.04m 0.03m	2.56 E-10 7.27 E-7 3.53 E-4

This suggests that for a thermal conductivity measurement over a 100 seconds period a sample of radius 4 cm would be acceptable.

Summary of time limits

The time interval taken for computing the thermal conductivity measurements is between 20 and 80 seconds. Using this time interval, a block of ceramic glass was calibrated using the needle probe against another block of ceramic glass of known absolute thermal conductivity. A comparison of the result obtained with that from an independent laboratory (Oxford University, Department of Geology and Mineralogy) shows a reproducibility of within 3%. The values for the ceramics have in turn been calibrated against a standard of quartz glass and checked against divided bar measurements for one of the ceramics.

3.4 <u>Temperature Effects</u>

Electrical Resistivity / Formation Factor

Jackson (1975) and Stanford (1975) in using the porosity cell both show the variation of the resistance of the electrolyte at the potential electrodes with change in temperature. They record an almost linear variation of 2.4% per O C with the resistance falling as the electrolyte conductivity increases with temperature, over the range of temperature encountered in a daily variation. Similar results pertain for this study, and it is therefore necessary to monitor the temperature of the sediment body throughout any one run. This is achieved by use of a thermistor located in a stainless steel tube within the main chamber of the cell. The formation factor is then simply computed by;

$$FF = \frac{R_{sample}(T^{O}C)}{R_{electrolyte}(T^{O}C)}$$

Permeability

Permeability measurements are particularly sensitive to changes in temperature due to the associated variation in the viscosity of the pore fluid. It is essential that the temperature of the permeant be closely monitored to within 0.1° C so that the computed permeability values can be corrected to a standard temperature of 20° C;

$$\emptyset_{20^{\circ}C} = \frac{\mu_{T^{\circ}C}}{\mu_{20^{\circ}C}} \emptyset_{T^{\circ}C}$$

where μ = viscosity of electrolyte

Thermal Conductivity

Although thermal conductivity does vary with temperature, the typical room temperature variation during a run is sufficiently small as to be considered negligible. An additional problem, however, can arise in that with each successive thermal conductivity measurement, heat is added to the system and effectively an anomalous temperature regime established. It is important therefore to allow sufficient time to elapse between measurements for this heat to dissipate and a reasonably stable situation to redevelop. This is particularly important in the porosity cell where there is no flow of water which can be suitably annexed to aid this recovery. The thermal conductivity measurements presented in this study relate to a temperature range of approximately 19°C - 23°C.

3.5 <u>Experimental Procedure</u>

This is largely similar for both the porosity cell and the permeameter cells, although there are obvious peculiarities confined to each type of cell.

3.5.1 <u>Sample preparation</u>

Each sample is washed carefully to remove any salt deposits and then dried overnight in an oven at 110°C. The required amount of sample is then weighed (between 900 g for a shelly carbonate and 1700 g for a quartz sand) and placed in a vessel containing electrolyte. The electrolyte and sample are thoroughly mixed and allowed to remain overnight to achieve equilibrium; the electrolyte is a dilute seawater solution, diluted to enable increased resolution in the electrical resistivity measurements.

The cells are filled with sand in a manner shown by Kolbuszewski (1948a,b) to produce a very loose packing state. A polythene funnel is introduced into the top of the cell and the saturated sand allowed to enter through this route, under water. The rate of deposition should be chosen to avoid any turbulent effects as the grains tumble through the water column. If any turbulent effects are observed these may be reduced by using a narrower funnel and adjacent variable flow valve (Schultheiss, 1982). Care must be taken during the filling of the cells not to produce any accidental knocks or vibrations which would reduce the porosity of the sample as it settled into a more compact structure. Once the cell contains

all of the weighed sample, measurements of the necessary parameters are made on a routine basis, the sample being gently compacted in stages. The compaction is accomplished by vertical impacts imparted to the cell by use of a variable amplitude sieve shaker. As the porosity of the sample decreases and the packing becomes closer, longer periods and higher amplitudes are required to reduce the porosity further. Each run is repeated twice for each sample to enable an accurate check on the repeatability of the measurements.

The peculiarities of the permeameter cells in terms of the experimental procedure concern the necessity for a 'filter bed' of gravel at the lower end to prevent the sample being washed out by the permeant, thus blocking the outlet pipe. This must be laid for each test prior to depositing the sample in the cell. Additionally, care is required to allow a constant hydraulic gradient to be achieved before commencing the test; this is important since any intervening thermal conductivity measurements require the permeant flow to be temporarily suspended.

4. SANDS RESULTS : ANALYSIS AND INTERPRETATION

4.1 <u>Experiment Summary</u>

4.1.1 Artificial sands

The artificial group of sands have been deposited in the porosity cell and the square permeameter. In the porosity cell, measurements of porosity - horizontal formation factor were made, both without a needle probe present, and with the needle probe inserted vertically through the base of the cell. Lovell (1979) showed the introduction of the needle probe to effect a slight increase in the formation factor for a volume determined porosity; the difference was difficult to quantify since it was inseparable from experimental error (Fig.4.1). . In the square permeameter the permeability was measured in a vertical orientation, along with the horizontal formation factor, and thermal conductivity with the needle probe inserted horizontally through the wall of the cell. Again no detrimental effects on the permeability measurements were observed by the introduction of the needle probe. Measurements using the cylindrical permeameter (vertical permeability and vertical formation factor) are restricted in this group to a brief analysis of the magnitude of any anisotropy of the formation factor for the same sand at a similar permeability state.

4.1.2 <u>Natural sands</u>

The four natural sands studied have been deposited in all three cells. Additionally, in the porosity cell the thermal



Fig. 4.1. Comparison of formation factor results for the porosity cell with and without needle probe. (Lovell, 1979).

conductivity has been measured with the needle probe both horizontally orientated and vertically orientated.

Table 4.1 summarises the parameters and their orientations for each of the two groups. It should be noted that whilst the orientation of the electrical and permeability measurements generally refers to the direction of flow, for the needle probe the thermal conductivity measurement approximates the value for the material perpendicular to the orientation of the axis of the probe (see section 4.5.1 on anisotropy).

4.1.3 <u>Matrix thermal conductivity</u>

Various equations exist expressing the bulk thermal conductivity of a porous system in terms of its components (Sass <u>et al</u>, 1971; DeVries <u>et al</u>, 1975; Bruggeman, cited Hutt and Berg, 1968). In an earlier attempt to briefly determine the appropriateness of any, or all, of these, a series of simple experiments were carried out on a variety of particles (nylon ballotini, quartz sand, broken shell sand) using two different pore fluids (distilled water and alcohol). Each combination was deposited in a beaker in a manner which gave a fairly close packing state, and the mean value of 20 thermal conductivity measurements taken.

4.2 <u>Electrical Formation Factor - Porosity</u>

Jackson's porosity cell is designed primarily to produce electrical formation factor - porosity data for a single sample over a range of packing states from its loosest (high

TABLE 4.1

Measurements on Sands

Artificial Sands

Porosity cell:		FF_h	-	n		
		FFh	-	n	-	k _v
Square permeameter:		FFh	-	ø _v		
	Аларана Аларана	FFh	-	ø _v	-	k _h
Natural Sands	v					
Porosity cell:		FFh		n		
		FFh	-	n	-	k _v
、		FFh		n	-	k _h
Square permeameter		^{FF} h	-	øv		
		FFh	•••	ø _v	-	^k h
Cylindrical permeameter		FF_v	-	Øv		

Note:

FF = electrical formation factor n = porosity Ø = permeability k = thermal conductivity

Subscripts h, v refer to horizontal, vertical directions. For electrical and permeability measurements these are directions of flow, though for thermal conductivity measurements they refer to the orientation of the needle probe. porosity) to its densest (low porosity). Initial results showed the success of the design in producing such information (Jackson, 1975a,b) and a more detailed study of the ranges of data produced by a variety of sand shapes and sizes has followed (Stanford. 1975; Jackson et al, 1978). The essence of the detailed work shows that all sands obey Archie's law (eq.2.12) to a reasonable degree, with the exponent m decreasing with increasing sphericity of the particles. The value of m varies from close to 2 for shelly sands to 1.5 for natural quartz sands. Spherical particles (glass ballotini) can exhibit an m value of 1.4 to 1.3. For a single sand sample undergoing compaction from its loosest to densest states the formation factor will vary by 20 - 25 %, corresponding to a fractional porosity change of between 0.05 (shelly sands) to 0.10 (natural quartz sands). In comparing sand samples the position of the formation factor - porosity plot may be indicative of a difference in particle shape, but could also be due to a difference in the size distribution of the particles. As the particle sphericity increases, the slope of the plot will decrease, and the formation factor increases corresponding to lower porosity values. An increase in the particle size distribution produces a decrease in porosity as voids are partially filled by smaller particles corresponding to an increase in formation factor, although the slope of the plot will remain unchanged as long as the particle shapes do not vary.

The results obtained for this study are presented in Fig.



Fig. 4.2. Best fit Archie lines for artificially constructed range of quartz and shell sands.

4.2 for the artificial samples, and in Fig. 4.3 for the natural samples. They confirm the earlier work already described, with each sample reasonably represented by Archie's law (eq. 2.12). For some of the samples an equation of the form used by Winsauer (eq. 2.13) correlates better, though within the bounds of experimental error both adequately describe the behaviour. An expression describing all of the data in the form of a 3rd degree polynomial is;

 $n = 1.4154 - 0.4799(FF) + 0.0687(FF)^2 - 0.0033(FF)^3$

More recently, Schultheiss (1982) has carried out collaborative work with Jackson on the cell for the purpose of measuring shear waves in addition to the electrical - porosity data. A feature of their study concerned an examination of the various techniques possible in filling the cell; this was aided by the inclusion of a smaller pair of potential electrodes positioned towards the centre of the cell chamber to enable an evaluation of any variation in the packing of the sand across the cell volume. They showed that during a compaction run the porosity in the centre of the cell is initially lower than at the edges, but becomes higher as the vibration proceeds before returning to a lower value at the end of the test. Furthermore, experiments at inverting the cell to enable a rerun of the test led to the porosity at the centre of the cell being lower than at the edges, although the difference generally decreased, as the test proceeded. They report the maximum difference in fractional porosity between the centre and periphery of the cell was approximately 0.015,



Fig. 4.3. Best fit Archie lines for natural sand samples.

based on a change in formation factor of approximately 0.12. In an attempt to control these packing irregularities all of the measurements in this study have been made using primary deposition samples (i.e. no inversions of the cell to redeposit the sample loosely have been made).

4.3 <u>Electrical Formation Factor - Permeability</u>

Plotting formation factor - permeability as a log - log relationship shows a linear trend for each sample (Figs. 4.4 & 4.5), suggesting there exists some relationship between the two similar to that of Brace (1977, eq.2.16)

 $\emptyset = C(FF)^{-X}$

where $a = h^2/k_0$

h = hydraulic radius k = shape factor (2 - 3)

The results for such plots are tabulated in Table 4.2 and show both C and x increase with decreasing sphericity of the particle shape. For an increasing spread of sizes C decreases and x increases. The results for the artificial sand / shell mixtures are slightly confused by the mixing of two sands of similar spreads but different means, although the results for the natural samples, however, are more positive. The coefficient C appears to be a function of the pore size, both in terms of the equation of Brace (and earlier, Wyllie and Spangler, 1952) and these results; with increasing spread of sizes C decreases as the pore space volume decreases







Fig. 4.5. Permeability plotted against Apparent Formation Factor for natural sand samples.

TABLE 4.2

Permeability - Formation Factor Regression Lines for Sand Samples

$\emptyset = C (FF)^{-x}$

Sample No. С

х

R

Vertical Ø - Horizontal FF

Artificial Sands

a b c d f & m g h i j k l n	0.1069 0.0851 0.0892 0.0695 0.8347 0.0552 0.0457 0.0056 0.8483 0.0966 	2.9632 2.9860 3.2423 3.2947 3.2160 2.2344 3.2341 2.8141 6.3746 3.5308 3.9109 3.5810	0.98 0.99 0.98 0.98 0.98 0.97 0.98 0.99 0.94 0.95 0.99 0.99
--	--	--	--

Natural Sands

0	0.0105	3.1598	0.98
p	0.0221	3.3622	0.98
q	0.0876	3.9623	0.98
r	3.9104	6.7278	0.98

Vertical Ø - Vertical FF

Natural Sands

0	0.1823	3.1819	0.99
P	0.1039	3.5623	0.99
q	0.2076	3.9263	0.99
r	75.1688	7.8880	0.96

(decreasing n, increasing FF), while a decrease in sphericity of the particles towards plate-like grains opens out the structure (increasing n, decreasing FF) and hence C increases. The exponent x, meanwhile, increases with both decreasing sphericity and increasing spread of sizes. In developing his model from that of Wyllie and Spangler (1952) through Archie's law (eq. 2.12), Brace attaches a dependence of x on the exponent in Archie's law. This would suppose a certain dependence of x on the shape of the particles, x increasing with decreasing sphericity: this is upheld by the results presented here.

Archie (1942) showed a clear relationship to exist between porosity and permeability of the form;

$$n = C \emptyset Y$$

Combining this with his relationship for electrical formation factor and porosity;

$$FF = n^{-m}$$

an equation relating electrical formation factor to permeability is arrived at;

$$FF = \frac{1}{C^m} \phi^{-my}$$

i.e. $FF = b \emptyset^{-x}$

where

$$b = 1/C^m \qquad x = my$$

This is similar to the equation of Worthington (1973) and shows a clear dependence of both the coefficient and exponent in the formation factor - permeability plot on the Archie exponent m. As already shown (Jackson <u>et al</u>, 1978; and this study) this exponent in saturated clean sands appears to be primarily influenced by particle shape.

The coefficient C may be expressed in terms of the hydraulic radius h (the ratio of the volume of the interconnecting pores to their surface area) and a shape factor k. Stanford suggested that a shape factor was almost numerically equal to the tortuosity, and defined it as the product FF.n. This is equivalent to the electrical tortuosity $n^{-(m-1)}$ of Wyllie and Spangler (1952). Stanford further developed the concept of a "Formation Sphericity" derived from the cosine of the angle of the gradient of the straight line relationship between his shape factor and porosity. This cosine treatment gave values between 0 and 1. He tentatively concluded that the formation sphericity could provide an alternative method of quantifying grain shape and provided comparative data of measured sphericity and the cosine-derived sphericity. The data match could apparently be improved by taking the square root of the cosine. Perhaps, the best aspect of Stanford's sphericity is that it refers to the pore space rather than the particles individually. However, it is noteworthy that for his data the m exponent of Archie's law bears a reverse correlation with measured sphericity, and the square root of its inverse bears as close agreement as the

square root cosine sphericity. The problem with both techniques is that, while they may adequately assess the shape of the grains, neither accounts for the variation of the pore shape with changes in porosity; for this purpose the individual shape factor FF.n (or electric tortuosity of Wyllie and Spangler, 1952) is preferable. It would appear that the shape factor and the hydraulic radius of equation 2.16 may thus be expressible as one function describing the pore space.

4.3.1 Anisotropy

For the natural suite of samples measurements have been made using both permeameter cells with similar depositional techniques for each. Data sets are thus available for measurements of vertical permeability corresponding to both horizontal and vertical electrical formation factors. The plot in Fig. 4.5 shows both sets of data, and reference to Table 4.2 provides the least squares linear fits to these data. While the vertical formation factor is far in excess of the horizontal, for the same vertical permeability, the slopes of the fits remain remarkably similar for any of the four samples. This raises the problem of applying laboratory conclusions to in-situ conditions and the correlation of electrical and hydraulic parameters. All of the measurements for the artificial suite are for electrical and hydraulic flows in mutually perpendicular orientations. It would seem a reasonable suggestion that, on the basis of these results and since hydraulic anisotropy is recorded in the literature (e.g.

Maasland, 1957), hydraulic anisotropy would also be evident where electrical anisotropy is so pronounced. Extreme values for the electrical resistivity anisotropy (λ_e) are 1.08 and 1.20 (after Keller and Frischknecht, 1966: $\lambda_e = (FF_v/FF_h)^{0.5}$). It is therefore necessary to define the orientation of a permeability and/or electrical measurement for the purposes of predicting one from the other. In this connection the formation factor - porosity curves already depicted should be noted as being in terms of horizontal electrical measurements; attempts at measuring vertical electrical formation factors in the porosity cell produced inconclusive results as to the nature of the exponent m in Archie's law for different orientations of current flow.

4.4 Thermal Conductivity - Porosity

Measurements on saturated deep sea sediments (Bullard, 1954) and also on a variety of powders (Ratcliffe, 1960) have shown the thermal conductivity to depend largely on the water content. Similar results pertain in this study for measurements on deep sea clays (see sections 6.4.4 and 6.6).

Thermal conductivity values have been obtained for both suites of sand samples using the transient needle probe technique; measurements were made on each sample with the needle orientated vertically, and also horizontally.

For the artificial suite the measurements with the needle vertical are for the porosity cell, while the horizontal needle measurements were carried out in the square

permeameter. Fig. 4.6 shows a log - arithmetic plot of thermal conductivity against porosity. This shows a general adherence of the data to the geometric equation (Sass <u>et al</u>, 1971) relating the total or bulk thermal conductvity to the thermal conductivities of the components;

$$k_{\rm b} = k_{\rm s}^{(1 - n)} k_{\rm w}^{\rm n}$$

For a sand composing more than one solid fraction the equation may be extended to;

$$k_{b} = k_{s_{1}}^{n_{1}} k_{s_{2}}^{n_{2}} k_{w}^{n}$$

where $n_1 + n_2 = 1 - n$ and n_1, n_2 refer to volume proportions of k_{s_1}, k_{s_2} Thus, the majority of the data relate to quartz sands and are adequately described by the equation;

$$k_{\rm h} = 8.58^{(1 - n)} 0.64^{(n)}$$

(least squares fit)

where $k_{quartz} = 8.58 \text{ W/mK}$ $k_{water} = 0.64 \text{ W/mK}$

The value of quartz compares very well with the range of values quoted by Clark (1966), although the value for water represents a 7% error at most (compared with 0.60 - 0.61 W/mK over the temperature range of the measurements).

The shelly sand mixtures, however, require use of the three component model. This introduces another variable whose value cannot be accurately established (the thermal conductivity of the shelly material). Using the two values already established for quartz and water, the use of a value



Fig. 4.6. Thermal conductivity plotted against porosity for sands of different particle sizes and shapes.

for carbonate of 3.32 W/mK gives a reasonable fit to the data (Fig. 4.6) for the pure shell sand, with the mixtures falling between the two.

A similar approach to the data for the natural suite of sands (Fig.4.7) yields a similar range of matrix (solid) conductivities between the quartz and carbonate extremes, although the value for guartz of 8.58 W/mK obtained from the arificial suite appears perhaps slightly too high. The changeover appears to be less distinct than in the artificial suite of samples, and the four models fitted are based on the carbonate content determinations for each of the natural samples, listed in Table 4.3. Each of the four models appears to put an upper bound on the corresponding data set. This feature may be due to a difference in the conductivity value assigned to the quartz particles for the artificial and natural suites. Since there is only sufficient data relating to comparatively pure quartz sand in the artificial suite, the conductivity value for guartz determined using the geometric model on that data has been used throughout. It is easily seen, however, that a slight adjustment to this conductivity value could shift the model to a position which correlates better with the natural sands data.

An alternative approach to predicting the thermal conductivty of a particulate material is that of DeVries (1975). In addition to defining the volume fractions and thermal conductivities of the system components, the ratio of the average temperature gradient in the granules to that in



Fig. 4.7. Thermal conductivity plotted against porosity for natural sands with 3 component geometric model fits.

TABLE 4.3

CARBONATE DETERMINATIONS FOR NATURAL SANDS

Sample No.	<pre>% Carbonate</pre>
0	6.5
p	51.4
q	78.8
r	90.8

the fluid is required. This parameter is shown to be a function of the shape and orientation of the particles and the ratio of the conductivities. Two limiting cases may be applied to the present study; that of a high conductivity equidimensional particulate structure (e.g. a quartz sand) and that of a low conductivity plate-like or lamellae structure (a carbonate shell sand). The boundaries set by these two conditions are plotted in Fig 4.8.

4.5 <u>Thermal Conductivity - Electrical Formation Factor</u>

Attempts at relating electrical and thermal conductivities have generally met with little success (Hutt and Berg, 1968) although individual relationships between each parameter and porosity have been shown both experimentally and theoretically (Archie, 1942; Ratcliffe, 1960). Fig. 4.9 is a plot of thermal conductivity against electrical formation factor for both suites of samples. Two features are apparent; firstly some relationship between the two parameters would appear to exist, and secondly there is a tendency for the two sets of measurements relating to different orientations of the needle probe to separate out. In terms of the relationship between the two, a predictor equation based on combining the two separate porosity relationships should be feasible.

The relationship between thermal conductivity and porosity may be described using the geometric equation, already noted, or by an equation proposed by Bruggeman (cited Hutt and Berg, 1968);



Fig. 4.8. Thermal conductivity plotted against porosity for all sands data. Model fit after deVries (eq. 2.27).

- Needle horizontal
- × Needle vertical



Fig. 4.9. Thermal conductivity plotted against electrical formation factor for artificial and natural sands.

$$n = \frac{k_{b}' - k_{s}'}{k_{b}'^{1/3} (1 - k_{s}')}$$

where $k_b' k_s'$ are the conductivities of the bulk and particles respectively, normalised with respect to the conductivity of the pore fluid. The disadvantage of this equation is that it does not readily avail itself for a system composing particles of more than one conductivity. The electrical formation factor variation with porosity is best described in terms of Archie's law (eq. 2.12), the exponent m taking a separate value for each sample, though a reasonable description may also be given by a polynomial of the form;

 $n = 1.4155 - 0.4799(FF) + 0.0687(FF)^2 - 0.0033(FF)^3$

These two pairs of equations may be combined such that the thermal and electrical conductivities may be related through porosity, which is common to each equation. Figs. 4.10 - 4.12 show the results of this approach and each combination used appears to provide a reasonable fit. The Bruggeman - Archie combination (Fig. 4.10) provides only boundary limits to the data for 100% pure carbonate or quartz; this is because modifications to the Bruggeman equation are required to accommodate more than one solid particle conductivity. Similarly the Geometric - Polynomial combination (Fig. 4.11) has been used only for the pure ends of the sample spectrum, since the polynomial equation does not describe the individual electrical - porosity data sets as well as Archie's law. Both of these combinations provide reasonable bounds to the data.

• Needle horizontal

× Needle vertical





• Needle horizontal

x Needle vertical





The combination of the Geometric and Archie equations (Fig. 4.12) provides for a wider area between the upper and lower limits than the Polynomial - Archie combination, and also allows for precise definition of each sample. This Geometric -Archie model fit, like that of the geometric model on the thermal conductivity - porosity graphs (Figs. 4.6 & 4.7), sets an upper bound to the data. As before this is considered to be a function of the use of the incorrect conductivity for the particles (the value for quartz of 8.58 W/mK is derived from the artificial suite data set), and can be corrected for by using a slightly lower value.

4.5.1 Thermal conductivity and anisotropy

The mass of data relating to measurements with the needle probe orientated in two mutually perpendicular directions, but with the formation factor measured horizontally in both cases, suggests a directional dependence of thermal conductivity in sands. The transient needle probe technique approximates an infinite line source, which in an homogeneous, isotropic material would produce circular isothermals based on the axis of the probe. The direction of heat conduction under such conditions is mutually perpendicular to these isotherms and relates to the planes radiating out through 360°, from the axis of the probe. These directions may in turn be related in terms of two perpendicular components, corresponding to axes of the thermal conducivity system (Fig. 4.13).

For an anisotropic medium Bloomer (1980) has suggested
- Needle horizontal
- × Needle vertical







Fig. 4.13. Needle orientation and measured thermal conductivity.





that the shape of the isothermals centred on the needle will be modified to an ellipse (Fig. 4.14, after Bloomer, 1980) where the two conductivities perpendicular to the needle probe orientation are not equal. The numerical solution to this problem is difficult and an intuitive approach is taken by Bloomer, assuming the temperature - time variation of a constant power needle probe remains unchanged, and the net heat flow perpendicular to the isothermals is constant, for the transformation to an elliptical shape. Thus, if the needle probe is orientated as in Fig. 4.14 (a) the measurement will produce a value for $k_x = k_y$. However, if the needle probe is orientated as in (b) the measurement k_m may be defined as;

$$k_{\rm m} = \frac{2}{((1/k_{\rm z}) + (1/k_{\rm x}))}$$

Thus if k_x can be defined, k_z may be computed from the measured value.

To test the result, measurements were made by Bloomer on a supposed homogeneous block of anisotropic slate for which divided bar measurements were available; the results appear to confirm the validity of the equation.

Applying these results to the present data set it should be possible to compute the vertical component of thermal conductivity k_z from the measurement with the needle horizontal, using the result for the horizontal component (with the needle vertical). This would provide a maximum and minimum limit of conductivity values over which the measured

value could be expected to occur. In defining the electrical formation factor relationship with thermal conductivity it is preferable to compare similar components. Since all of the electrical measurements made in connection with the thermal measurements are for the horizontal plane this restricts the specific definition to the horizontal thermal conductivity measurements (with the needle probe vertical). Such a consideration reduces Fig. 4.9 to Fig. 4.15. and may help to explain the upper bound fit of the models described earlier.

It should be possible to calculate a coefficient of anisotropy for thermal conductivity analagous to that for electrical resistivity ;

$$\lambda_{t} = (k_{x} / k_{z})^{0.5}$$

where k_{x} , k_{z} are the thermal conductivities horizontally and vertically; k_{x} is measured directly, while k_{z} is computed from two mutually perpendicular measurements. Attempts to do so for the natural sediment suite, however, produce inconclusive results with values of λ_{t} ranging between 1.08 and 1.17 for the four samples.

4.6 Matrix Thermal Conductivity Results

The results of an earlier attempt to examine the appropriateness of the Geometric and Bruggeman equations in phase describing the thermal conductivity of a two/porous medium are presented in Table 4.4. Three different types of particles were deposited, in turn, in distilled water, and in industrial



Fig. 4.15. Thermal conductivity plotted against electrical formation factor for sands of different particle sizes and shapes.

TABLE 4.4

MATRIX THERMAL CONDUCTIVITY RESULTS

Particle Type	Porosity n	Pore Fluid	Thermal C Measured	Conductivity k Geometric Br model	(W/mK) uggeman model
Nylon	0.385	water	0.34	0.35	0.37
Ballotini	0.402	alcohol	0.20	0.22	0.23
Quartz Sand	0.366	water	3.18	3.26 (1) 3.22 (2)	3.40 (1) 3.36 (2)
	0.350	alcohol	1.83	2.26 (1) 2.23 (2)	2.07 (1) 2.05 (2)
Broken	0.630	water	1.04	1.14	1.19
Shell	0.610	alcohol	0.60	0.58	0.57

Component Thermal Conductivity Values:

(1)	^k qua	artz	= 8.58	W/mK		both val	Lues empirio	cally	
(2)	^k qua	artz	= 8.49	W/mK		Geometri artifici	ic model on lal quartz s	sands.	
k _{shel}	.1 =	3.32	W/mK	(Arti	ficia	l sands	result)		
k _{nylc}	on =	0.25	W/mK	(Kaye	and I	aby, 196	56)		
k _{wate}	er =	0.61	W/mK	(Chall	oner	and Powe	211, 1957)		
k _{alco}	hol	= 0.	17 W/mH	(ethy	1),	0.19W/mH	(methyl),	(Weast,	1970)

Specific Gravity Values:

```
water = 1.00
alcohol = 0.80
nylon ballotini = 1.05
quartz sand = 2.65
broken shell = 2.78
```

alcohol. Measurement of the thermal conductivity of each system was made with the needle probe orientated vertically, for one porosity value; the mean of 20 measurements on each sample was taken as being characteristic of the sample. The results show that both models describe the thermal conductivity of a porous medium reasonably well. At worst the deviation between measured and computed values is 24%, and generally this difference is in single figures. Allowing for an error of up to 7% in the thermal conductivity measurement the two equations would seem to produce acceptable representations of the thermal conductivity of a porous medium. Additional errors in defining the conductivities of quartz and broken shell (carbonate), and the porosity of the system, can only enhance the validity of using either model. To choose between the two models, for sands containing water as the pore fluid, the Geometric Model appears to be slightly better, although a different choice of k_{solids} could feasibly reverse the situation. Additionally, the Geometric model has the enhanced capability of being extended to several solid phases of differing conductivities, and with sufficient data, to a detailed consideration of anisotropic media.

5. EXPERIMENTS ON CLAYS

The aim of this part of the study is to assess the interrelationships between electrical resistivity, porosity, and thermal conductivity for deep sea clays. The nature of such material allows for adequate sampling, but does not lend itself to redeposition in the laboratory. In order to obtain a variety of conditions for one sample therefore, an oedometer, or consolidometer, was chosen as the basis of the laboratory measurements; in normal use this allows for uniaxial stress and strain, mechanically similar to that achieved by normal overburden pressure. It is thus possible to reproduce, at least in part, the conditions at depth within the sediment column. Additionally, the use of this apparatus allows calculation of the porosity and permeability of the sample being tested. The model chosen for the study was a Wykeham Farrance Engineering, rear loading, dead weight, lever arm oedometer. This apparatus is as standard, although the cell used to contain the sample was substantially modified as detailed below.

It is worth noting at this point that the modified cell does not allow for backpressuring of the system to enable complete resaturation where gas has come out of solution, or direct permeability measurements. Whilst these features would be ideally suited to this study, both financial and, in particular, temporal constraints prevailed.

5.1 The Modified Oedometer Cell

The oedometer is a well tried machine, familiar in most soils engineering establishments, and fully detailed in the literature (Lambe, 1964; Akroyd, 1964; Taylor, 1948). Effectively, a dead weight is applied to the system, and through a system of lever(s) uniaxial stress is applied to a sample contained in a fixed ring; the resulting uniaxial strain is monitored with time.

The modifications carried out under the terms of this project have been based on a standard fixed ring cell where a sample is contained within a ring with a fixed porous disc below, and a porous disc above which is free to move inside the ring (see Fig. 5.1, Plate D). The main novelty of the modified cell is its construction, being manufactured almost entirely out of electrically non-conducting PVC. This is to effectively remove the conducting effect of the cell during electrical resistivity measurements on the sample. The only metal part is an upper cap of stainless steel to minimize distortion at high values of uniaxial stress. The dimensions of the cell are similar to the conventional apparatus; the sample ring has a diameter of 75 mm and height of 20 mm.

Within this basic design are contained piezoelectric transducers to enable measurement of the compressional wave velocity (1 MHz), and piezoelectric bimorph elements (Fig. 5.2) for measuring the shear wave velocity of the sample. Each of the pair of transducers, one transmitter and one receiver, is located in a perspex plate, which itself is

PLATE D

The modified oedometer.

The modified oedometer cell, with exposed top cap, sample ring, and needle probe.







Fig. 5.2. Operation of a piezoelectric bimorph element for transmitting and receiving shear waves.

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(Schultheiss, 1981)

located adjacent to the exterior faces of the porous discs, with respect to the sample (see Fig. 5.1). The inner surface of each of the perspex plates is then coated with electrically conducting paint to form two current electrodes. Potential electrodes, in the form of two single circular stainless steel wires, are located on the inner surfaces of the porous discs, the whole comprising a standard four electrode resistivity measuring system.

The sample ring itself is also constructed out of PVC. A hole in the side wall of the sample ring (approx. 0.8 mm diam.) enables the introduction of a thermal conductivity measuring needle probe. The precise nature and workings of this probe are detailed elsewhere in this report. The cell base is constructed with a slot cut away in the side wall to enable the introduction with ease; a single thermistor probe may be inserted, if no needle is to be present, to enable the temperature of the sample to be monitored throughout the test.

An additional pair of piezoelectric compressional wave transducers (1MHz) are introduced into the cell wall giving a horizontal propagation path. These enable more accurate monitoring of velocity variations over a fixed separation and also allow for a simple assessment of any sound speed anisotropy.

The initial results from test samples consolidated in this apparatus show that there is no appreciable loss of material due to introduction of the needle probe in the sample. Experiments with two contiguous samples of apparently homogeneous material from the same core, consolidated in a

conventional and modified oedometer show there to be negligible effect on the consolidation behaviour due to the presence of the geophysical elements, suggesting that there is minimal interference with the pore water drainage from the sample.

Data Analysis

The application of each incremental load provides a series of displacement values with time from which the 100% and 50% primary consolidation times can be computed using standard techniques (logarithm time, square root time fitting methods: Taylor, 1948). These data can then be used together with various mass determinations to compute the following parameters for each load; this type of analysis is thoroughly documented in any standard soil mechanics textbook (eg Lambe, 1964; Akroyd, 1964).

a) Void Ratio (e)

Appendix B provides equations for computing various other volume fraction parameters (porosity, bulk density) from this parameter (with the aid of the specific gravity of the particles).

b) Coefficient of Consolidation (C_v) The decrease in volume of the sediment sample with time may be

expressed by this coefficient. Defined here it represents an average coefficient of consolidation;

$$C_{v} = \frac{0.197 (H/2)^2}{t_{50}}$$
(5.2)

H/2 = half thickness of sample t₅₀ = time for 50% primary consolidation

c) Uniaxial compressibility coefficient (M_v)

Defined as the change in volume per unit volume per unit increment in pressure.

$$M_{v} = \frac{\Delta e}{\Delta P(1 + e)}$$
(5.3)

d) Coefficient of permeability (Ø)

$$\emptyset = C_{y}M_{y}Y_{u} \qquad (5.4)$$

 Y_w = unit weight of pore fluid

Calibration data

The displacement transducer (LVDT) is calibrated in a vertical jig providing an accuracy of +/- .005 mm.

5.2 <u>Temperature</u> Control

To enable accurate thermal conductivity measurements the sample must be maintained at a uniform and steady temperature.

The oedometer is thus positioned in a temperature controlled room in which the temperature is kept at approximately 24°C. Geophysical measurements are then corrected to 20°C as follows:

Thermal conductivity values are corrected using the technique adopted by Ratcliffe (1960) whereby the thermal conductivity decreased by 6% over the temperature range 25° C to 4° C.

Compressional wave velocities are corrected by applying the same correction as applicable to the compressional wave velocity in seawater (Shumway, 1958).

Electrical resistivity values are converted to electrical formation factors at the corresponding temperature using empirical calibration curves and electrical conductivity data for seawater (Thomas <u>et al</u>, 1934).

No corrections to the shear wave velocities are considered necessary over the temperature range studied.

5.3 The Compressional Wave Measuring System

The compressional wave velocity of the sediment sample is determined by comparing the travel time through the sample to the travel time through standard seawater. The electronic instrumentation for generating and receiving high frequency sine waves is shown schematically in Fig. 5.3. The transmitting 1MHz piezoelectric transducer is excited by an 'Arenberg' pulsed oscillator via a 100 ohm variable attenuator. A pulse generator triggers both the pulsed



Fig. 5.3. Block diagram of the acoustic electronics used with the modified consolidation cell.

oscillator and the oscilloscope used for observing the received signal from the 1 MHz piezoelectric receiver. The time measurement between the voltage applied to the transmitting crystal and the voltage across the receiving crystal was measured by using a digital counter with a resolution of 0.05 microseconds.

A novel feature of the modified cell is the location of the compressional wave piezoelectric transducers beyond each porous disc. This allows for less disturbance of the drainage paths of the sample and under certain circumstances can ease the identification of the onset of the pulse by its later arrival.

The measuring system for the additional compressional wave transducers with a horizontal propagation path is of a similar design.

Data Analysis

The compressional wave velocity is effectively computed by comparing the travel time through the sample to the travel time through standard seawater.

Since the propagated wave travels through two porous stones in addition to the sample thickness there is a considerable delay time which must be subtracted from the total travel time of the pulse propagating through the sample. This delay time is a cumulative delay incorporating any electronic effects. The delay time value is determined by calibrating the system using standard seawater as the propagating medium for a series of transducer separations.

Plotting the travel times against the separation distances, the intercept of the best fit line is a measure of the delay of the system.

The compressional wave velocity (v_p) for the sample is then calculated;

$$v_{p} = \frac{H}{t_{m} - t_{d}}$$
(5.5)

 t_m = measured travel time through sample t_d = delay time

Errors:

The relative errors involved in measuring V_p are +/- .05 mm and +/- .05 microseconds. Since the sample thickness decreased during the test, the error in the measurement consequently increased; the range encountered varies between +/- 6 m/s and +/- 20 m/s for the samples in this study.

An additional absolute error may be introduced due to the assumption that the sample thickness is initially 20.00 mm. All sample thicknesses during the test are computed relative to this assumption, and an error of 0.1 mm, whilst small initially, can be substantial at the end of the test when the sample thickness may be reduced by more than 50%. This error is particularly important since it is difficult to estimate its magnitude, but for an initial unrealised offset of 0.1 mm can be as high as +/- 30 m/s.

5.4 The Shear Wave Measuring System

The shear wave measuring system uses an arrangement originally developed and tested by Schultheiss (1981), involving piezoelectric bimorph crystals as transducers. The preliminary development of such shear wave transducers was carried out by Shirley and Anderson (1975) and Shirley and Hampton (1978) at the Applied Research Laboratories, University of Texas at Austin. A bimorph element (Fig. 5.2) is a composite sheet of two polarised piezoelectric ceramic plates bonded to a centre electrode. Voltages applied across the plates cause the element to bend producing a shearing motion in the surrounding medium. The electronic system for transmitting and receiving shear waves is shown schematically in Fig. 5.3. Two shear elements of 10.0 mm x 10.0 mm x 0.4 mm dimensions are mounted, one each on the perspex discs, and protrude through the porous discs to a depth of 3.0 mm into the sediment sample. A 10 volt positive-going D.C. step of a square wave generated by a function generator is used to drive the shear wave transmitting element. At the same time a 2.5 volt square wave is used to trigger the oscilloscope which displays the received signal, the travel time being measured by a digital timer with 0.1 microseconds resolution.

Data analysis

Calibration of the shear wave measuring system is rather more difficult than for the compressional wave since there is no standard reference material with a known shear wave velocity. An attempt at defining such a medium for this

purpose is to use a remoulded sample of commercially available potter's clay, prepared in such a way as to provide a homogenous quantity which may be trimmed in stages so that a travel time is obtained for successively smaller transducer separations. A plot of the travel time against the transducer separation (measured from the inner leading edges) shows the delay time of the system to be negligible.

The velocity of the shear wave (V_S) during consolidation is then determined;

$$V_{s} = \frac{H - TS}{t_{s}}$$
(5.6)

TS = length transducers protrude into sediment sample t_s = measured travel time

Errors:

The error situation is similar to that for the P-wave with a relative error of between +/- lm/s and +/- 3m/s. The maximum error attributable to an initial sample thickness offset from 20.00 mm of 0.1 mm is +/- 8 m/s.

5.5 The Electrical Resistivity Measuring System

A four-electrode array comprising two current and two potential electrodes is utilised within the modified cell. The two current electrodes are located as the inner surfaces of the perspex discs housing the ultrasonic transducers; they are coated with electrically conducting silver paint such that

current flow through the sample volume may be approximated as vertical and parallel. The two potential electrodes, each in the form of a single stainless steel wire, are located on the inner surfaces of the porous discs, adjacent to the sample. An alternating current at a nominal frequency of 0.4 Hz is supplied by an ABEM SAS300 terrameter, the voltage across the sample being measured at the potential electrodes by a potentiometer integral to the intrument. The measurement is designed to give the resistance of the sample; this may be converted to a resistivity through a geometrical constant which relates to the geometry of the electrode arrangement. During the consolidation of a sample the geometry will of course vary and hence this constant is computed for a continuous range of sample thicknesses. The electrical measurement is in turn computed as a formation factor (i.e. the resistivity of the sample is normalised with respect to the resistivity of the pore fluid).

In defining the electrical formation factor of the marine samples used in this study, the resistivity of the pore fluid is taken to be that of the bottom water at the location where the core was taken (Kullenberg, 1952; Siever <u>et al</u>, 1965).

Data analysis

Since the geometry of the cell continously changes as the sample thickness decreases, it is therefore necessary to calibrate the cell for a range of sample thicknesses using seawater of similar salinity and temperature to the sample. The formation factor of a sample may then be determined by

dividing the resistance of the sample by the resistance of the pore fluid (seawater) at those same conditions.



(for similar conditions of l, a β T(°C)

P =resistivity
R =resistance
a =area
1 =length
T =temperature

The nature of this determination is to assume that the formation factor does not dramatically vary with temperature; effectively the variation of the resistivity of the sample is assumed to be similar to that of the pore fluid. Since electrical conduction is largely through the pore fluid, and the range of temperature being considered is at most $3^{\circ}C$ (25°C - 22°C) and therefore unlikely to produce important chemical or mineralogical changes, this assumption appears reasonable.

The errors involved in the determination of the formation factor centre on the measurement of the electrcial resistance, thickness, and temperatue of the sample; a combination of the effects of these parameters on the final determination is difficult to examine precisely, but is estimated relatively at 3% with a maximum of 5%.

5.6 The Thermal Conductivity Measuring System

The thermal conductivity of the sample is measured using the transient needle probe technique originally applied to marine sediments by Von Herzen and Maxwell (1959).

The needle probe (Fig. 5.4) is designed to approximate an infinite line source with a constant heat flux. The thermal conductivity of an infinite sample may be approximated by;

 $\frac{\delta T}{\delta(\ln(t))} = \frac{Q}{4\pi k}$ (5,11)

k = thermal conductivity of sample

Q = rate of heat dissipation per unit length of an infinite line source

- T = temperatue
- t = time

However, because of the finite dimensions of the probe and of the sample, the time interval must be kept to within confined limits. If the rate of supply of heat Q is reduced, t may be increased; however, the temperature range measured



Fig. 5.4. Cut-away view of the thermal conductivity needle probe.

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will be small and the sample must be initially at a uniform temperature. To accomplish this the sample and measuring systems are contained inside a constant temperature room $(25^{\circ}C)$. A value for Q may be determined by calibrating the probe against a standard of known thermal conductivity (Bloomer and Ward, 1979).

The needle probe design incorporates a central heating element within a fine diameter copper tube and a resistance thermometer to monitor surface temperature. The copper tubing has an external diameter of 0.65 mm and a 0.25 mm bore, with 42 SWG enamelled constantan wire as the heater. The latter is soldered to the end of the tube thus allowing the tubing to be used as the heater current return. Any space remaining within the tube is filled with epoxy resin potting compound to give extra rigidity to the construction. The use of copper tubing ensures an even distribution of heat throughout the probe and also a fast heating response, therefore overcoming problems incurred with previous stainless steel probe designs.

A non - inductive winding of 49 SWG enamelled copper wire covers the surface of the tubing. This provides an average temperature reading at the probe surface. The probe is finally covered with epoxy resin, a fine protective coating being achieved by certain heating techniques. The final diameter of the probe is typically 0.8 mm, the length 75 mm.

All electrical connections are made within a perspex housing, encapsulated with potting compound, to provide strain relief. The measuring system is shown schematically in

Fig. 5.5.

The resistance thermometer is contained within a finely tuned bridge network with a four terminal connection at the probe. The potential developed across the resistance thermometer is digitised using the Intersil 7109, a 12 bit, dual slope, integrating Analogue to Digital Converter (ADC). The bridge network also provides a fixed voltage reference for the device.

The ADC is interfaced to a low power CMOS microprocessor system consisting of the Motorola 6805 processor with additional 5k Bytes of Erasable Programmable Read Only Memory (EPROM) and 1k Bytes of Random Access Memory (RAM). An audio cassette interface is also incorporated to allow data to be 'dumped' to a remote Apple computer. Instrument control is obtained through a 16 way keypad and data output through either a dual 4 digit display or the cassette interface.

Low power CMOS integrated circuits are used wherever possible to keep power consumption low, allowing easy conversion to battery operation if necessary; this also helps to reduce system noise levels.

A constant current source has been designed to provide the heater current; the output can be selected between 100 mA and 140 mA.

Calibration of the resistance thermometer is made against a precision thermistor located in a thermally insulated block of brass at a point half way along the needle probe. A thermal gradient is induced into the brass in a direction along the length of the probe. Thus the average temperature





of the probe could be accurately determined from the resistance of the thermistor. The digital output of the ADC is recorded in one degree intervals over the range 20° C to 40° C, for each probe. The resulting curve is then matched to a third order polynomial equation, effectively linearising the resistance thermometer and the ADC. The coefficients obtained for each probe are stored in the non-volatile memory of the microprocessor system, enabling their use in an iterative software routine to produce a temperature reading.

Mathematical conversion of the ADC output to a temperature reading was typically 570 ms allowing temperature to be recorded at one second intervals during a thermal conductivity measurement. To avoid any further inaccuracy in the temperature reading, all mathematical calculations are made using a numbering system whereby each number is stored in three consecutive bytes in a floating-point format. The system permits representation of zero, positive and negative numbers, in the dynamic range E-38 to E+38, with an accuracy of five digits.

For a 20°C temperature range the total ADC input voltage swing is typically 35 mV and by using the ten most significant bits of the output a resolution of 35 microvolts can be obtained. This relates to an increase of typically 0.01° C to 0.02° C per bit of the ADC. The system noise is effectively reduced such that overall accuracy of the resistance thermometer is governed by the accuracy of the temperature calibration. This is estimated to be around +/- 0.05° C over

the range 20° C to 40° C.

Typically, a single thermal conductivity measurement is monitored over 100 seconds, temperature readings being taken at one second intervals. The precise timing of the measurements are made by the 'on - board' timer of the 6805 microprocessor.

A visual inspection of this curve allows rejection of any temperature measurements before the probe achieves equilibrium with the sediment at the probe/sediment interface, and after the boundary effects of the sample size and finite probe length become important at increased time. The remaining data are plotted on a temperature - log time scale and by comparing the slope with that for a ceramic of known thermal conductivity (Bloomer and Ward, 1979) the sample thermal

<u>Data analysis</u>

The thermal conductivity is measured using a needle probe which effectively provides temperature data over a 100 second period for a needle inserted into the sample and supplied by a constant heat supply along its length. The needle approximates an infinite line source, the theory of which has been worked out in considerable detail (Jaeger, 1956, 1958). The temperature rise at the probe is given to a good approximation by;

$$T = \frac{q}{4\pi k} \cdot \ln\left(\frac{4Dt}{r^2}\right) + C \qquad (5.12)$$

(Carslaw and Jaeger, 1959)

where t = time T = temperature q = heat input per unit length per unit time k = thermal conductivity of sediment sample D = thermal diffusivity of sediment sample r = probe radius

If the temperature-time data is potted on a linear-logarithm scale the plot should asymptotically approach a straight line of slope $(q/4\pi k)$. If $(q/4\pi k)$ is a known quantity k can be determined.

The method used for determining $(q/4\pi k)$ is to use a calibration standard of known thermal conductivity and to compare this slope with the slope for any unknown sample (Bloomer and Ward, 1979);

(5.13)

The standards used for this study included a glass ceramic, and gelled water (Hyndman <u>et al</u>, 1979).

Typically, the thermal conductivity for a sample at a given load was taken as the mean of five individual measurements; the resulting accuracy was generally within a 7% margin. However, certain limitations to the needle probe must be considered in computing the results;

1 Lower time limit

A limitation to equation (5.12) is that the approximation only holds for small values of r and large values of t, i.e.

r²

probe radius r = 0.0004 mtypically, at start of test water content = 70% $D = 2.10^{-7} \text{ m}^2 \text{s}^{-1}$

end of test water content = 35% D = $3.5.10^{-7}$ m²s⁻¹

$E_1 = 0.2000$ t	t seconds	$E_1 = 0.1143$ t
0.020	10	0.011
0.013	15	0.008
0.010	20	0.006

From these results, the approximation to a linear slope is attained at 20 seconds and 12 seconds corresponding to samples tyically of 20 mm and 10 mm thickness. Generally a time interval of 20 seconds is sufficient for the approximation to hold.

2 Upper time limit

The upper time limit is dictated by the diameter of the sample, centred on the needle, which is effectively the sample

thickness. De Vries and Peck (1958) provide an equation defining a limit on the dimension of the sample;

$$E_{u} = \exp\left(\frac{-R^{2}}{4Dt}\right) \quad << 1 \qquad (5.15)$$

where	R = radius of $D = thermal d$ $t = time (sec$	sample iffusivity s)			•
For R	= 0.01 m	$D = 2.0.10^{-7} \text{m}^2 \text{s}^{-1}$	(water	content	70%)
t 20 50 65 80 100	Eu 0.0019 0.0821 0.1462 0.2096 0.2865				
ForR	= 0.006 m	$D = 3.5.10^{-7} \text{m}^2 \text{s}^{-1}$	(water	content	35%)
t 20 30 40 50	Eu 0.2765 0.4244 0.5258 0.5979				

These results suggest that at low thermal diffusivities corresponding to large sample thicknesses (> 20 mm) an upper limit of 60 seconds is acceptable. However, an upper limit of less than 20 seconds may be required for samples of high thermal diffusivity corresponding to small thicknesses; this limit is particularly constricting but is rarely approached in practice. (It should be noted that the dominant term is the thickness of the sample rather than the change in thermal diffusivity with consolidation).

Summary of limitations

The above limitations to the needle probe results when

applied to the present suite of samples suggest the following time intervals may be considered as sufficiently satisfying the conditions appropriate to equation (5.12).

20	mm	samples	20	-	60	seconds
12	mm	samples	12		32	seconds

These narrow time intervals obviously restrict the accuracy of the measurements, particularly the limitation imposed by the narrow sample thickness towards the end of the test. The error in the measurement is, however, considered as, at worst, 10%.

5.7 <u>Description of the Sediment Samples</u>

Two suites of samples have been considered in this study, although only the primary suite provided samples of sufficient size to enable complete testing. The secondary suite of samples from the Deep Sea Drilling Project are only 50 mm in diameter and this restricted size prevented measurement of either the electrical or thermal parameters.

The primary suite consists of nine samples from the N.E. Atlantic. The samples are from three separate 4 m long Kastenlot gravity cores, with a tenth specimen providing a commissioning trial test sample for the apparatus. Each core was subsampled in 8 cm sections from which a PVC sample ring was used to obtain an "undisturbed" specimen from the central section. Once prepared the samples were stored in seawater to prevent loss of moisture, and at 4° C to minimize any chemical

or bacteriological activity. Initial testing of the apparatus was accomplished using commercially available potter's clay.

The secondary suite consisted of four DSDP samples; two red clays and two carbonate oozes. These were of 20 mm height and 50 mm diameter, contained in a teflon coated stainless steel ring. Because of the narrower diameter, the cell had to be adapted and a new top cap constructed; the short length across the centre of the ring prevented thermal conductivity measurements with the needle probe, and the stainless steel ring eliminated the possibility of making meaningful electrical resistivity measurements. They are included here for completeness only. Table 5.1 summarises all of the clay samples tested.

5.8 <u>Experimental Procedure</u>

5.8.1 Sample preparation

Sample preparation is carried out at sea upon recovery of the cored section, hence little is required beyond setting the sample in the cell, and the cell in the oedometer.

Initially the base and top cap of the cell are placed under seawater and a vacuum applied to ensure complete saturation of the porous stones. The cell is then carefully retrieved and the sample placed in position with the top cap on, positioned centrally with locating spacers to prevent any transference of stress onto the sample; the whole is kept submerged in seawater to prevent the porous stones drying out.

TABLE 5.1

CLAY SAMPLES

-	· · · · · · · · · · · · · · · · · · ·					
Sample No.	Sediment Type	Мø]	Location	Sample Code	Depth
•	•					
Kastenl	ot Gravity Co	res		н. Т		
1 · ·	Pelagic Ooze	8.9	32 22	34.7'N 27.5'W	D10406/1	20cm
2	Turbidite	9.4		11	D10406/3	36cm
3	Turbidite	9.3		11	D10406/7	104cm
4	Pelagic Ooze	8.9		59	D10406/11	185cm
5	Pelagic Ooze	10.3		FT	D10406/20	304cm
6	Nanofossil Turbidite	9.7	31 24	33.2'N 50.5'W	S126/2-1	44 cm
7	Π	8.9		11	S126/2-4	70cm
8	Calcareous Pelagic Clay	9.2	30 23	22.0'N 35.0'W	S126/15-4	81cm
9	Π	10.0			s126/15-11	179cm

Deep Sea Drilling Project Samples

DSDP	4			
1	Red Clay	10.1	RC576A-2-3//	13.15m
2	Red Clay	10.9	RC576A-5-5//	44.65m
3	Carbonate Ooze	8.8	577A-3-4//	24.85m
4	Carbonate Ooze	7.2	577A-9-2//	78.85m
At this stage a thermal conductivity measurement may be made, following which the needle is removed and replaced with a single small temperature sensor which locates in the wall of the ring.

The cell is then placed on the beam of the oedometer. A dial gauge and linear variable displacement transducer (LVDT) enable monitoring of any displacement once the locating spacers supporting the top cap are removed. Measurements of the remaining geophysical parameters may be made without affecting the consolidation once the top cap is in contact with the sample.

5.8.2 <u>Consolidation testing with simultaneous geophysical</u> measurements.

Since the modifications to the oedometer cell do not alter the basic engineering of the oedometer design, the method of testing by which increments of uniaxial stress are applied to the sample and the resulting strain is monitored remains as documented elsewhere (Lambe, 1964; Akroyd, 1964).

A load of 5 kN on the beam hanger produces a pressure of 12.5 kPa (28 kPa) on a 75 mm (50 mm) diameter sample.

When a fully saturated sediment is loaded, the load is initially carried by the fluid in the pores; this gradually transfers to the soil skeleton as water drains out to the surrounding medium, producing a change in the water content corresponding to a change in the void ratio of the sediment. The transfer time of the application of each load depends on

the time for the load to become the total effective pressure when the excess pore water pressure is reduced to zero; this is called the 100% primary consolidation stage and depends on the soil thickness, compressibility, and permeability. A clean coarse grained sand will attain its 100% primary consolidation within a few minutes while a fine clay requires 24 hours or more to reach this stage.

The thickness of the sample, or its height, is monitored by a dial gauge and a linear variable transducer.

The compressional wave and shear wave velocities together with the electrical resistivity measurements are quoted throughout this report for the 100% primary consolidation stage of each loading; this standard is adopted simply because it represents an easily repeated condition, which enables comparisons of the data between samples. To achieve this it is necessary to monitor the geophysical parameters throughout the loading cycle. Additional thermal conductivity measurements during the loading cycle are made after the attainment of 100% primary consolidation, at some time prior to the application of the next load; this is to minimize the. effect of any temperature fluctuations on the consolidation and other geophysical measurements.

Following the attainment of 100% primary consolidation the sample continues to decrease in volume as the particles rearrange their relative positions; this phenomenon is referred to as secondary consolidation. The next loading is applied at some time after the primary consolidation is complete and generally involves a doubling of the total

applied pressure.

The sample is unloaded in similar increments to the loading process, following which the sample is removed, dried overnight, and weighed.

6. <u>CLAY RESULTS</u> : ANALYSIS AND INTERPRETATION

6.1 <u>Calculation of Equivalent Depths</u>

The use of an oedometer in this study enables the variation of geophysical and geotechnical parameters to be monitored with the application of incrementally increased axial loads. These uniaxial stresses may be equated to the stresses exerted by the material forming the overburden in a sedimentary column; thus any specific applied load can be equated to a depth in the sediment column beneath the sea floor.

$$D = \frac{\sigma_z}{\gamma_{sub}}$$
(4.1)

where

Ysub

$$\frac{\gamma_{W}(\rho_{3} - 1)g}{(1 + e)}$$
 (4.2)

D	22	depth
σz	-	vertical stress
YW.	=	unit weight of pore fluid
Ps	*	specific mass density of solid
g	33	acceleration due to gravity

The equivalent depths computed using this equation are evaluated on the basis of the characteristic nature of the test sample, and assume there is no gross variation of material with depth. The equivalent depth thus becomes a function of the porosity of the test sample at a given pressure; obviously the errors in these assumptions can be considerable.

6.2 <u>Laboratory Consolidation and its Relation to the In situ</u> <u>Sediment Column</u>

The laboratory oedometer test effectively consolidates a sample of material very rapidly in comparison with the very slow rates of deposition, and hence consolidation, encountered naturally in the deep ocean environment. Whilst it simulates the reduction in volume of the pore space, the sudden application of relatively large increments of vertical stress precludes any time-related effects developing. Thus the important field or in-situ processes of interparticle bonding, cementation, and rigid bonding (an effect produced in addition to normal interparticle bonding by the extremely long time constants involved - Bryant et al, 1981) are neither reproduced, nor can they be reasonably accounted for quantitatively. As a result of this severe limitation, the oedometer test - while providing useful information relating to the consolidation behaviour of a material - can only approximate the condition of a sample at depth. The interpretation of results, particularly with regard to establishing gradients of a parameter with increasing depth, must be viewed cautiously; such a gradient can only represent the variation of that particular material with depth, and cannot wholly account for in-situ conditions especially at increasing depths where time related effects may be considerable. Even with these constraints, however, the laboratory predicted compressional wave velocity gradients do compare reasonably well with those measured in-situ (Hamdi, 1981), suggesting that some credence at least can be

attributed to the oedometer simulation of overburden pressure.

6.3 <u>Previous Stress History : Preconsolidation Pressure</u>

The application of a load to a sample whereby the load effects grain movement (sliding, rolling) which is non-elastic and produces a decrease in the void ratio, followed by elastic deformation of the grains, considers the sample to be normally consolidated. The term 'normally consolidated' refers to a soil element that is at equilibrium under the maximum stress it has ever experienced; however, if the sample is at equilibrium under a stress which is less than that to which it was once consolidated, it is termed overconsolidated (Lambe & Whitman, 1979).

The state of a sample in these terms may be obtained by comparing the overburden and preconsolidation pressures. The overburden pressure is calculated as that due to the overburden present at the time of sampling. The preconsolidation pressure, which is the maximum previous stress applied to the sample, may be estimated from a plot of void ratio against pressure (e - log P) (see Schmertman, 1953 for details). A quantity known as the overconsolidation ratio (OCR) may be defined;

OCR =	Preconsolidation Pressure	(6.3)
001	Overburden Pressure	(0.07

Thus three states of samples may be identified;

OCR	<	1	underconsolidated
OCR	Ħ	1	normally consolidated
OCR	>	1	overconsolidated

The e-log P plots presented in Figs. 6.1 - 6.6 show all but one of the surficial sediments, Sample 6, to be moderately overconsolidated. The suite of DSDP samples all show overconsolidation although the amount varies, while the differential between the overburden and preconsolidation pressures is similar within each pair of samples from a single location; this suggests the sediment column between each pair of samples is continuous.

The overconsolidation shown by the majority of the samples is surprising in that the development of a previous vertical stress in excess of that produced by the present overburden would not be expected on the deep ocean floor, particularly for the primary suite (some of which represent comparatively recently deposited material). In the context of previous results for cored samples taken from the North Atlantic (Buchan et al, 1971) these results show similar overconsolidation phenomena. The overconsolidation apparent in the consolidation data of Buchan and colleagues was attributed to bonding after long exposure on the sea floor, or an alternative suggestion concerning the method of sampling using a free fall corer. Sample 6 in this study is anomalous in that it appears normally consolidated. This feature may be similar to that described by Hamilton (1964) where a disturbed sample may appear as an unconsolidated sample in terms of an e - log P plot.



Fig. 6. 1. e - log.p plots.



Fig. 6.2. e-log. p plots.



Fig.6.3. e-log.p plots.



Fig. 6.4. e - log.p plots.





Fig. 6.6. e-log.p plots.

6.3.1 Effect of preconsolidation on sample behaviour

The process of consolidation consists of two phases. Firstly, the non-elastic movement of grains relative to each other resulting in a decrease in void ratio; the sample will not return to its original state if the applied stress is removed. Secondly, elastic deformation of the skeleton and individual particles will occur; this phase is reversible and removal of the applied stress will allow for some recovery of the sample towards its initial state.

If, therefore, a sample has undergone consolidation at some previous maximum stress, and is then effectively unloaded to a lower stress level, the recovery towards its original state is only partial. Reloading the sample at pressures below the previous maximum stress means that any new strain will result purely from elastic deformation, the non-elastic part having occurred during the previous loading. Only at stress levels greater than the maximum previous stress will normal consolidation (elastic and non-elastic) occur, whilst at the transition from elastic to non-elastic behaviour at the preconsolidation stress the behaviour of the sample may be unpredictable.

6.3.2 Effect of preconsolidation on oedometer test results

The theory of unidimensional consolidation allows for the computation of the coefficient of consolidation and the uniaxial compressibility, from which a coefficient of permeability may be derived (eq. 5.4). The coefficient of

consolidation is theoretically a constant, but as Richards and Hamilton (1967) note it has unfortunately been found in practice to change with confining stress. During the reloading portion of an e-log P curve the coefficient is often either indeterminate or subject to large errors and hence consideration is restricted to values corresponding to pressures greater than that of the preconsolidation load i.e. the straight line portion of the e-log P curve. In computing the permeability of the sample from oedometer theory it is therefore necessary to restrict the results analysis to pressures both in excess of the preconsolidation pressure and over the linear portion of the e - log P plot.

6.3.3 Effect of preconsolidation on porosity, electrical formation factor and thermal conductivity measurements

The porosity of a saturated marine sediment is a measure of the proportion of voids present; it is related to both electrical formation factor and thermal conductivity. Since none of the parameters appears to be influenced by the elastic properties of a sediment, the only effect is liable to be in connection with the initial resistance of the sediment to consolidation. This feature should produce little initial change in void ratio (and hence parameters dependent upon it) with increasing stress, until the preconsolidation stress is exceeded.

This effect suggests that gradients of any parameters whether functions of the elastic properties of the sample or

its volumetric proportions - may only be reasonably computed for depths in excess of that represented by the preconsolidation pressure.

6.3.4 Effect of preconsolidation on velocity measurements

a) <u>Compressional wave propagation</u>

The effect of the previous stress history on the compressional wave velocity of a marine sediment was investigated by Hamdi (1981). By prestressing a sample to various states of artificially imposed overconsolidation he was able to monitor the variation of the compressional wave velocity with applied load from pressures less than the previous maximum stress to pressures in excess of it. He concluded that the expected increase in the compressional wave velocity with increasing consolidation pressure becomes disturbed as a function of the preconsolidation pressure. A similar result shows the non-linear effect on the constrained modulus which initially rises in magnitude with increasing applied pressure as strain results from the elastic deformation of the sample, the non elastic deformation having already occurred during the previous loading. The modulus thus reaches a maximum, but as the applied load becomes greater than the preconsolidation pressure, the grains start to slide relative to each other, the resistance to straining decreases and the constrained modulus reaches a minimum. Further increase in the applied pressure results in an increase in the modulus as the grains come into a more compact

arrangement and interparticle sliding becomes more difficult (Whitman <u>et al</u>, 1964, cited Hamdi, 1981). Hamdi explains the non-linear behaviour of the compressional wave velocity around the preconsolidation pressure as a reduction in porosity matching the opposite effect of an increase in the sediment bulk density on the elastic moduli of the sediment: the net result being that there is little or no increase of the compressional velocity with increasing loads.

b) <u>Shear wave propagation</u>

The effect of a previous stress history on the shear wave velocity is not documented at present, and since it may be reasonably expected to be affected by the time dependent effects of consolidation it is probably not ideal to use an artificially preconsolidated sample to test the effect. However, applying the results applicable to the compressional wave to the present case a decrease in the shear wave velocity gradient may be expected, although the absolute value would continue to increase with increased applied pressure.

Consideration of the DSDP suite of samples provides, perhaps, a better experimental approach than an artificial sample. A plot of shear wave velocity against strain (Fig. 6.7) shows an evident decrease in shear wave velocity gradient at a pressure similar to that of the preconsolidation value. This effect is somewhat masked by the initial drop in velocity following the application of each load. This latter effect is considered to be a function of the development of excess pore



pressure immediately following the loading, but appears to be a minimum as the sample passes from elastic to non-elastic deformation at the preconsolidation pressure.

6.4 <u>Variation of Geophysical / Geotechnical Properties with</u> Depth

6.4.1 <u>Compressional wave velocity gradients</u>

Compressional wave velocity measurements were made at a frequency of 1MHz and are corrected where necessary to a standard temperature of 20⁰C, and a salinity of 35 parts per thousand.

For the suite of surficial samples the preconsolidation pressure is comparatively low and close to the start of the loading cycle. In this region the compressional wave velocity is close to that of the pore fluid and the gradient is small; together with the difficulty of accurately reproducing the unknown preconsolidation pressure, this tends to make the detection of any subtle changes problematical.

The DSDP suite of samples, however, exhibit much larger preconsolidation pressures, and slightly greater gradients; the detection of the effect of the preconsolidation is easier, and since it is apparent, must be considered in computing gradients for these samples.

Thus for the surficial suite of samples the effect of preconsolidation is very limited and generally falls within the overall measurement error; gradients may therefore be computed from the whole data set. The DSDP suite, however, is

affected to a greater extent and gradients can only be computed for pressures exceeding that of the preconsolidation.

Ranges of gradient

A plot of compressional wave velocity with depth for the cored surficial sediments is presented in Fig. 6.8. Generally the velocity increases with increasing depth of overburden; this is primarily a function of the decrease in porosity. Two groups may be defined on the basis of initial void ratio;

```
Group 1: e < 5
Samples 1 and 3
```

the data is best represented by the polynominal

 $v_p = 1489.77 + 0.3455D - 0.00006035D^2$ R = .99

or by the linear equation

 $V_{\rm p} = 1489.95 + 0.3346D$ R = .99

Group 2: e < 5 Samples 4, 5, 6, 7, 8, 9.

 $V_p = 1486.75 + 1.3090D - 0.00476D^2$ R = 0.87

Sample 6 is anomalous to group 2 in that its initial void ratio is greater than 5. However, it rapidly falls below 4 at low pressures; it is also anomalous in appearing to be the



Fig. 6.8. Compressional wave velocity variation with effective depth for the surficial sediments.

only normally consolidated sample and may therefore be disturbed.

The equivalent linear gradients over the depth range to 150m for the two groups are;

Group 1 0.7 s^{-1} - 0.3 s^{-1} Group 2 1.3 s^{-1} - 0.3 s^{-1}

Compressional wave velocities for the DSDP suite of samples are plotted with depth in Fig. 6.9. The nature of the preconsolidation pressure prevents gradients being computed for stresses less than this value; the number of data points available is thus reduced and the gradients may thus be expressed in terms of either polynominal or linear functions with equally good fits;

DSDP 1 (red clay)	$V_{p} = 1514.03 + 0.5479D$	R = .98
DSDP 2 (red clay)	$V_p = 1501.56 + 0.7625D$	R = .94
DSDP 3 (carbonate ooze)	$V_p = 1564.77 + 0.5835D$	R = .99
DSDP 4 (carbonate ooze)	$V_p = 1680.08 + 0.5933D$	R = .98

These equations are for equivalent depths in excess of 39 m; they represent linear gradients of between 0.5 and 0.8 s⁻¹ for the red clay, and 0.6 s⁻¹ for the carbonate ooze.



6.4.2 Shear wave velocity gradients

Shear wave velocity measurements were made using bimorph 'bender' type crystals, the frequency of which varied with the nature of the material and the applied pressure. Typically for the surficial materials the frequency was initially 5 kHz, though for the DSDP carbonate ooze samples at high stress levels it was close to 100 kHz.

A plot of shear wave velocity with depth is presented in Fig. 6.10 for the surficial sediments. The general trend is for increasing velocity with increased depth, though changes in the gradient similar to that discussed in association with the preconsolidation pressure are apparent in many of the samples at low pressure.

Similar perturbations occur in Fig. 6.11 of the shear wave velocity - depth plot for the DSDP suite of samples. It is therefore necessary to compute gradients for both suites of samples using pressures in excess of that of preconsolidation.

Ranges of gradient

For the surficial suite, all but Sample 6 can be suitably expressed by a power equation of the form;

$$V_{\rm c} = 42.7064 \ {\rm D}^{0.3069} \ {\rm R} = 0.89$$

Sample 6 is again anomalous and is best described by;

 $V_{c} = 18.5183 D^{0.5166} R = 0.97$





though for a similar pressure range to the rest of the samples this equation reduces to

$$V_{c} = 29.5184 D^{0.3866} R = 0.98$$

which is close to that originally describing the majority of samples.

Sample 6 exhibits equivalent linear gradients varying in magnitude from 8 s⁻¹ to 1 s⁻¹ over the depth range 1 to 140 m, while for the remaining samples the gradients are 4 s⁻¹ to 1 s⁻¹ over the depth range 10 to 140 m. Extrapolation of this curve back to 1 m depth gives a gradient of 11 s⁻¹.

The DSDP suite of samples may be equally well described by power curves as by linear equations in terms of the shear wave velocity;

> DSDP 1 $V_s = 58.1127 D^{0.2484}$ R = 0.97DSDP 2 $V_s = 20.4994 D^{0.4496}$ R = 0.99DSDP 3 $V_s = 62.5223 D^{0.3537}$ R = 0.99DSDP 4 $V_s = 96.7977 D^{0.29326}$ R = 0.99

These are approximately equivalent to linear gradients of $0.45 \text{ s}^{-1} - 0.5 \text{ s}^{-1}$ for the red clays (DSDP1 and 2), and $1.0 \text{ s}^{-1} - 0.6 \text{ s}^{-1}$ for the carbonate oozes (DSDP3 and 4). They are for pressures equivalent to depths between 30 and 400 m.

6.4.3 <u>Electrical formation factor and porosity variations</u> with equivalent depth

Electrical formation factor measurements are not available for the DSDP suite of samples and therefore only the surficial sediments suite may be considered.

Each sample exhibited an increase in the formation factor with decreasing porosity (increasing vertical stress). The nature of this relationship is linear for each particular sample when plotted on a log-log scale (Figs. 6.12 & 6.13) and is best described after Winsauer <u>et al</u> (1952) by equation (2.13);

$$FF = Cn^{-m}$$

where C and m are empirical constants peculiar to each sample. Reasonable fit lines are also obtainable using Archie's equation (2.12), although with a lower degree of adherence;

$$FF = n^{-m}$$

where m = 2 (Taylor Smith, 1971).

Plotting all the individual linear relationships on a single graph (Fig. 6.14) the slope m of each 'best fit' is obviously a function of the range of porosity covered; m decreasing as n decreases, and C generally increasing simultaneously. Jackson <u>et al</u> (1978) showed that for sands the variation in m is a function of particle shape, m decreasing with increasing sphericity of the particles.







Fig. 6.13. Electrical formation factor plotted against porosity for a single sample during a consolidation test.



Fig. 6.14. Electrical formation factor – porosity data for all nine surficial sediment samples; each interpreted individually as FF = Cn-m.

A scatter plot of the individual measurements for samples 1 to 9 is presented in Fig. 6.15. It becomes apparent that while any individual sample may be expressed by an Archie or Winsauer type equation, the total solution is not a linear trend on a log-log plot.

A similar conclusion was made by Kermabon <u>et al</u> (1969) who produced a best fit line described by a 3rd degree polynominal;

 $n = -5.9021 (FF)^3 + 40.0416 (FF)^2$ -105.3899 (FF) + 171.2504

Boyce (1968) produced a Winsauer type equation for a small number (<50) of measurements on recent sediments from the Bering Sea.

 $FF = 1.3 n^{-1.45}$

More recently (Boyce, 1980) published data for samples from the DSDP show that in order of preference the following models agreed; Boyce (1968), Archie (1942), Kermabon <u>et al</u> (1969), Winsauer <u>et al</u> (1952), and Maxwell (1904).

The best fit obtained in this study for more than 250 points is an equation of the form;

n = 1.3861 - 0.4626 (FF) + 0.0833 (FF)² - 0.0073 (FF)³



Fig. 6.15. Electrical formation factor plotted against porosity for all nine surficial sediments.

This equation is subject to the boundary condition FF = 1 at 100%. A least squares fit to the data on a log - log plot, not subject to any boundary condition is;

$$FF = 1.29 n^{-1.42}$$

The similarity with that of Boyce, given above, is remarkable. Both of these fits, together with the functions defined by other workers in the field are plotted in Fig. 6.16.

It is interesting to compare the electrical formation factor results in the oedometer with the results obtained elsewhere for DSDP samples (Boyce, 1980). Fig. 6.17 shows the results for both sets of measurements plotted against depth, while Fig. 6.18 shows the functions describing both sets of data in terms of their porosity relationships. At high porosities the data agree reasonably well with published values for recent sediments. However, for the low porosity measurements where the oedometer attempts to artificially create the conditions at depth, the formation factor values fall on the low side for similar porosities in-situ; i.e. the electrical resistivity of the in-situ sample is greater than that of the equivalent laboratory consolidated sample. The difference may be explained in terms of the comparatively open structure of the laboratory sample, and the lack of any bonding or cementation which may occur in-situ. Additionally, if the sediment is anisotropic then this difference may exceed that shown here since the DSDP formation factors relate to the horizontal direction while the measurements for this study



Fig. 6.16. Electrical formation factor plotted against porosity for all nine surficial sediments interpreted as a single fraction.



Fig. 6.17. Electrical formation factor plotted against effective depth for the surficial sediments. Also plotted are DSDP data (Boyce, 1980).


Fig. 6.18. Electrical formation factor plotted against porosity for the surficial sediments. Also plotted are DSDP data (Boyce, 1980).

relate to the vertical direction. In a single vertically anistropic sample the horizontal formation factor would be expected to be less than the vertical.

The plot of formation factor with depth (Fig. 6.17) also shows the discrepancy between the DSDP values and those of this study; however, the scatter of points is such that the formation factor - depth relationship is difficult to precisely determine. This is primarily a function of the difficulty in defining the porosity - depth relationship. Bryant et al, (1981) suggest that there is no exact solution since the variables involved are too numerous. A laboratory predicted porosity gradient will differ from an in-situ gradient due to variations in temperature, grain characteristics, depositional history, physiochemical factors, tectonic stress, and numerous other factors to a lesser degree. Even with these limitations Stepheson (1977) proposes that it is still possible to gain a certain insight through porosity - depth predictions since the unknown effect of any one factor cannot possibly exceed the known combined effects of all the factors affecting the sediment porosity. The main obstacle then is in determining the relative importance of these individual factors for a given location.

In utilizing the formation factor measurements to predict porosity with depth, the comparison of laboratory and in-situ results suggests laboratory relationships will underestimate the in-situ porosity while in-situ measurements require greater definition particularly with regard to any effects of

anisotropy.

6.4.4 Thermal conductivity variations with depth

Measurements on saturated deep sea sediments by Bullard (1954) and Ratcliffe (1960) have shown clear relationships to exist between water content and thermal conductivity, the magnitude of the conductivity being little influenced by small changes in the mineralogy of the sample. For DSDP samples Erickson (1973) has shown clear but undefined relationships to exist for some samples between thermal conductivity and bulk density, porosity and natural gamma activity, although he emphasises that mechanical disturbance during coring may produce variations in physical properties as large or larger than those occuring naturally.

In this study measurements were only possible on the surficial sediment suite, and for each sample values of conductivity were determined both before loading and after loading. Where it was possible to push the needle into the sample under load measurements were also made, although only limited success was achieved. It was found to be unsatisfactory to leave the needle in position during the loading cycle since stressing and curvature of the probe resulted in calibration inaccuracies.

The thermal conductivity of an aggregate may be expressed as;

$$k_{b} = k_{1}^{n_{1}} k_{2}^{n_{2}} k_{i}^{n_{1}}$$

(Sass et al, 1971, eq. 6.4)

where the ith constituent occupies volume faction i. For an isotropic, homogeneous saturated sediment this expression reduces to

$$k_{\rm b} = k_{\rm s}^{(1-n)} k_{\rm w}^{\rm n}$$
 (6.5)

thus;

 $\log k_b = n (\log k_w - \log k_s) + \log k_s \qquad (6.6)$

Fig. 6.19 is a plot of log thermal conductivity - porosity; the linear trend with a correlation of 0.79 suggests the above equation reasonably describes the results of this study and provides conductivity values of 2.01 W/mK for the solids and 0.63 W/mK for the pore fluid. The value for the solids compares with 8.8 W/mK for quartz, 2.9 W/mK for other minerals and 0.25 W/mK for organic matter (Hillel, 1980), while the value for water of 0.60 W/mK (Challoner and Powell, 1957) represents a maximum consistent error of 5%.

Equation 6.23 may be adapted to materials of several mineral types by taking k_s as the geometric mean conductivity; similarly if the material is anisotropic then the measured conductivity may be estimated using this model by taking the geometric mean of the components.

Any analysis of the data with effective depth is made difficult by the sparcity of data points at pressure. From



Fig. 6.19. Measured thermal conductivity plotted against porosity for the surficial sediments.

the data available it is feasible, however to suggest that there is no measureable effect beyond the reduction in void ratio (porosity) and hence the water content. This appears to be in general agreement with the DSDP results of Erickson (1973) where no universal relationship between thermal conductivity and depth is apparent.

6.4.5 Permeability and its variation with depth

Permeability values are computed from oedometer theory for each of the samples tested over the range for which the coefficient of consolidation is constant (i.e. the e-log P plot is linear). Fig. 6.20 shows a plot of permeability against depth for the surficial samples; over the initial 140 m the permeability decreases by between one and two orders of magnitude. Fig. 6.21 shows a similar plot of permeability against axial load.

The relationship between permeability and axial load or equivalent depth may be expressed with reasonable correlation (generally R = 0.98) by power equations;

For axial load;

 $\emptyset = \emptyset_0 \mathbf{p}^{-m} \tag{6.7}$

where \emptyset_0 varies between 7.17 10^{-7} to 2.8 10^{-8} m/s and m varies between 0.71 and 0.99.

For the two data sets of DSDP samples, samples 1 and 2 provide values of $\emptyset_0 = 2.7 \ 10^{-4}$ m/s and m = 2.2, while for



Fig. 6. 20. Permeability variation with effective depth for the surficial sediments.



samples 3 and 4 $\emptyset_0 = 8.1 \ 10^{-3}$ m/s and m = 1.9.

For equivalent depth

$$\emptyset = \emptyset_0 D^{-m} \qquad (6.8)$$

 $Ø_{0}$ varies between 2.6 10^{-7} and 1.1 10^{-7} m/s, and m between 0.79 and 1.18 for the surficial sediments.

For DSDP 1 & 2 $Ø_0 = 5.25 \ 10^{-4} \text{ m/s}, \text{ m} = 3.0, \text{ and DSDP}$ 3 & 4 $Ø_0 = 5.02 \ 10^{-4} \text{ m/s} \text{ and m} = 2.13.$

 \emptyset_{0} approximates the coefficient of permeability at a load of lkPa and therefore varies with the physical characteristics of each sediment; for the DSDP samples their initial equilibrium pressure is greatly in excess of this value and the value of \emptyset_{0} is thus artificial and of little meaning.

Variations in the permeability of cohesive sediments may be due to a number of physical reasons, though the main factor appears to be grain size (Bryant <u>et al</u>, 1981). Theoretically permeability should be governed primarily by the shape, size and distribution (interconnection) of the pore spaces. In sands this is mainly an expression of the grain characteristics, but for clays the role of the chemical and electrical bonding in defining the nature of the layer of adsorbed water may be as important. Partical size analyses show insufficient variation in the samples at such fine sizes (< 9 phi) on which to base the permeability variations.

6.5 <u>Permeability Prediction and Electrical Formation Factor</u>

6.5.1 Which value of permeability?

Reference has already been made (p.13) to the controversy regarding the accuracy and validity of the various measurements of permeability. Although discrepancies do exist between the values obtained by direct measurement and those obtained from oedometer theory and the consolidation characteristics of a material, both sets of data relate in some manner to fluid flow in the test sample. It would seem likely therefore that there is probably some, as yet, undefined relationship between the two sets of data, even if that relationship involves defining a difference in the nature of the fluid flow. A relationship with one set of data therefore, may be taken as indicating the likelihood of a similar, though not necessarily the same, relationship with the alternate data set.

For the purpose of this study the permeability values are computed from the consolidation test data. These data are however briefly compared to direct permeability measurements and predicted permeability measurements in section (6.5.3) provided by Dr. P. Schultheiss at I.O.S. (Wormley) for similar samples of material.

Additionally, it should be noted that consideration of all measured parameters in this study for depth locations in the sediment column are based on equating the applied axial load to the depth of overburden. This technique is obviously

only valid for the mechanical loading of a sediment and does not include any of the time-related effects associated with the deposition of fine grained material, and is only valid for similar material (particle size and shape, mineralogy, void ratio etc.) at those depths.

6.5.2 <u>Empirical prediction</u>

Empirically, numerous authors have related permeability to void ratio through a variety of equations (see chapter 2). Fig. 6.22 shows such a plot, using e-logØ as the relationship. Each sample appears to show a linear trend, the slope of which is similar allowing for a reasonable error margin. The relative position of each sample trend is primarily a function of initial void ratio e_0 , although the overall range of permeability values shifts by less than half an order of magnitude over the total suite of samples, whilst the maximum range is two orders of magnitude. This suggests that while the volume proportions of the sample may assume similar values, the packing structure may differ for those samples, producing different values of permeability.

Considering the above suggestion that the initial void ratio (e_0) to some extent dictates the range of void ratios and permeabilities (i.e. the relative position of a line) it is possible on the basis of e_0 values to establish three sample groups:



Group	A	5<	e _o		1,3,6
Group	B	3<	e _o	<5	2,4,7
Group	C		e _o	<3	5,8,9

Since the variation in permeability may be a function of a different packing structure, this facet may be evident in the shear wave velocity profiles. Reference to Fig. 6.10 shows that there is a reasonable correlation with groups A,B,C having increasing shear wave velocities for similar equivalent depths. It may thus be possible to utilize the variation in shear wave velocity measurements in further delineating the permeability variation for a given e_0 ;

 $Vs = bD^{m}$

Sample	b	m
1	36.3528	0.3326
3	15.8098	0.5113
6	18.5183	0.5166
2	29.3015	0.4269
4	23.3945	0.4913
7	28.5810	0.4038
5	38.2470	0.3695
8	47.3505	0.3478
9	34.2016	0.4133

An interesting feature of the three groups above is that they do not strictly relate to the three cores used in this study. Particle size data for each sample show no distinguishing features, all samples being predominantly clay sized particles. In considering the particle size data, however, the values presented (Table 5.1) are for oven dried

samples. But in a deep sea sediment, the clay fabric detracts from individual discrete particles with matrices developed using chemical and electrical bonding, and more importantly perhaps, these bonds can cause a layer of water to adhere to the structure. This water may be effectively fixed as part of the framework of the sample, unlike the pore fluid in a cohesionless sand which is free to move relative to the framework. The result of this feature of the clay fabric is that the actual pore size may be less than that dictated by particle size measurements. How much less will depend on the samples ability to attract water to the surface of its particles, this ability is in turn a function of the mineralogy of the particles and the salinity of the pore A group of geotechnical properties indirectly linked fluid. to these phenomena are the Atterberg Limits, and evaluation of these may be of use in helping to define the variability in permeability with pore volume measurements (void ratio, electrical formation factor, compressional wave velocity). Within the limitations of this study this has not been feasible, partly due to insufficient sediment and the need to ovendry the sample provided in connection with the main phase of the programme; Grim (1962) points out the alteration in plasticity of a material upon drying.

The relationship defined in plotting e-logØ is similar to many empirical plots in the geotechnical literature (Taylor, 1948; Lambe & Whitman, 1979). The least square fit equations for the data form similar groups to those detailed;

Samples 1, 3, 6 $(5 < e_0)$ $e = 0.5910 \log_e \emptyset + 14.9846 R = 0.85$ Samples 2, 4, 7 $(3 < e_0 < 5)$ $e = 0.3266 \log_e \emptyset + 8.8626 R = 0.84$ Samples 5, 8, 9 $(e_0 < 3)$ $e = 0.2936 \log_e \emptyset + 7.7527 R = 0.76$ DSDP 1 & 2 (red clay) $e_0 = 5.19 e = 0.5381 \log_e \emptyset + 14.5473 R = 0.99$ $e_0 = 2.98 e = 0.3389 \log_e \emptyset + 10.0536 R = 0.99$ DSDP 3 & 4 (carbonate ooze) $e_0 = 1.95 e = 0.0946 \log_e \emptyset + 3.0695 R = 0.91$ $e_0 = 1.39 e = 0.2623 \log_e \emptyset + 6.2349 R = 0.99$

A comparison of the regression equations for the surficial and deep sea drilling project samples generally shows remarkable similarity in the coefficients for similar values of initial void ratio e_0 .

(a) <u>Electrical formation factor</u>

Plotting formation factor - permeability as a log-log relationship shows a reasonable linear trend for each sample, suggesting some practical analogy exists between the two parameters (Fig. 6.23).

Individual least squares fits may be fitted to each of the samples;

$$\emptyset = C FF^{-X}$$



Fig. 6.23. Electrical formation factor plotted against permeability for the surficial sediments.

C (10 ⁻⁶)	x	R
3.5141	9.5443	0.98
3.7139	9.0813 9.3383 9.2878	0.99
0.5609	10.8424	0.99
0.0254 0.1123	6.3903 7.5583	0.96
	C (10 ⁻⁶) 3.5141 1.5001 3.7139 9.4569 0.5609 3.7828 0.0254 0.1123	C (10^{-6}) x3.51419.54431.50019.68133.71399.33839.45699.28780.560910.84243.782812.34940.02546.39030.11237.5583

This type of fit is similar to that used by Brace (eq.2.16);

$$\emptyset = h^2/k_0 (FF)^{-1.5}$$

where k_0 is a constant and h is the hydraulic radius. The exponent value of 1.5 was determined by combining a model based on Poiseuille's law with the empirically derived Archie's law. The Archie exponent was taken as 2. The equation was successfully tested for a variety of granites, sandstones, and ceramics, over the range of formation factors from 4 to 16,000, representing a change in permeability of almost ten orders of magnitude. The maximum porosity considered was 0.37.

Worthington (1973) also produced a similar type of fit to data for a sandstone aquifer. His apparent formation factors varied from 2.6 to 9.9, which he proposed corresponded to true formation fators in the region 6.6 to 31.0. Permeability values ranged from 10^{-5} to 10^{-8} m/s, corresponding to porosity values between 0.11 and 0.29. The equation arrived at in Worthington's study, whilst similar in conception to that of

Brace, provides an exponent closer to those determined here for consolidated clays;

 $\emptyset = 1.1190 (FF)^{-5.88}$

(Worthington, 1973)

Similarly, if the results for the sand samples examined earlier in this study are considered (Table 4.2), the values of the exponents vary from being comparatively close to that of Brace (lowest values 2 to 3) for spherical particles, to values similar to those for clays (>7) for shelly, less spherical particles. This would suggest that the role of particle shape, and hence pore shape is important.

The dominant effect may not be associated with the initial physical state of the sample (e_o values being ungrouped by these divisions), but may rather be connected with the particle arrangements. This is supported in part by the physical descriptions of each core being different, and also by consideration of the results for the sand samples discussed earlier. For the sands both C and x increase with decreasing sphericity, while C decreases and x increases with increasing spread of sizes. The results for the cored samples presented here suggest the former effect due to grain shape is the more important and identify the three cores as being composed of increasingly spherical particles in the order;

sph	eric	ity	sphericity			
Core Sl26/2- Samples 6,7	<	Core Samples	D10406/ 1,2,3,4,	<	Core San	S126/15- mples 8,9

The segregation of the samples into the groups corresponding to each core allows least square fits to be determined for each group;

Samples 1,3,4,5 (Core D10406/) $\emptyset = 2.5334 \ 10^{-7} \ (FF)^{-6.0043}$ R = 0.87 Samples 6,7 (Core S126/2-) $\emptyset = 1.1739 \ 10^{-5} \ (FF)^{-11.4056}$ R = 0.93 Samples 8,9 (Core S126/15-) $\emptyset = 4.7645 \ 10^{-7} \ (FF)^{-6.8568}$ R = 0.94

The proposed variation in particle shape may be due to variations in mineralogy between the cores; if this is so, then the mineralogical differences may give rise to differences in the water adsorbtion qualities of the particle structure, together with variations in the conducting properties of the particle matrix. These two effects may respectively influence the hydraulic and electric flow, characteristics of the sediment. However, since the latter effect is generally considered as negligible for sediments with a saline pore fluid, the particle shape argument coupled with differences in mineralogy provides a reasonable explanation.

(b) <u>Compressional wave velocity</u>

The only documented example of an attempt at simply relating compressional wave velocity and permeability produced equation (2.22);

 $V_p = C - d \log_{10} \emptyset$ (Barker & Worthington, 1973)

Their data plot showed a large amount of scatter with a trend represented by the least squares line $(v_p = 1.4 - 0.38 \log_{10}0,$ for v_p in km/s and 0 in mm/s); field application of the equation would therefore be subject to considerable errors.

Least squares fits to each of the samples tested provide for good correlation of the two parameters, differentiation being made again on the basis of initial void ratio.

Group 1 Sample 3 $e_0 = 5.34$ (no P-wave visible) Group 2 Samples 1,4,6,7 $3 < e_0 < 5.5 V_p = 1100.2236 - 21.3639 \log_e \emptyset R = 0.90$ Group 3 Samples 5,8,9 $e_0 < 3 V_p = 910.4009 - 29.8139 \log_e \emptyset R = 0.69$

However, application of these equations in attempting to predict permeability must involve substantial errors, partly on account of the spread of data (Fig. 6.24), but particularly since the compressional wave velocity varies very little with changes in porosity for surficial deep sea sediments.

For the DSDP sediment samples, the high values of preconsolidation pressure limit the number of valid data points such that it is not possible to evaluate any clear empirical relationship with compressional wave velocity.



6.5.3 <u>Theoretical model</u>

The idea and background of creating a theoretical model is examined in section (2.5) and full details of the nature of the model used in this study are presented in Appendix A.

In defining the model a characteristic angular frequency, ω_c , for the material under consideration is introduced;

$$\omega_{c} = \frac{ng}{\vartheta} \left(\frac{\rho}{b\rho - n\rho_{W}} \right)$$
(6.9)

An approximate solution may then be realised by use of two theoretical velocities; the "zero frequency velocity" V_0 for frequencies much lower than the characteristic frequency, and the "infinite frequency velocity" V_{∞} for very high frequencies.

The solution to the model is then equation 2.19:

$$\delta = \frac{ng}{\omega} \left(\frac{\rho}{b\rho - n\rho_{W}} \right) \frac{V_{0}^{2}}{V_{0}^{2}} \left[\frac{\left(V_{p}^{2} \right) - 1}{1 - \left(V_{p}^{2} \right)} \right]^{0.5} (2.19)$$

For deep sea clays the characteristic frequency given by equation (6.9) is typically in excess of 600 MHz, in comparison with the propagating frequency of 1MHz for the compressional wave velocity measurements in this study. The measurements are thus made at a frequency very much less than

the characteristic frequency and the measured velocity V_p is represented by the zero frequency velocity V_o . It is suggested by Hamdi and Taylor Smith (1982) that ideally the sediment sample should be scanned at a wide range of frequencies to enable identification of the characteristic frequency and so permit an accurate evaluation of the permeability (equation 2.19). Whilst this is perhaps feasible for fine sands with a typical characteristic frequency around 100 KHz (although problems of scattering may arise) it is highly impracticable to attain propagation frequencies much in excess of 1MHz, and hence even the maximum frequency applied will be much lower than the frequency for clays.

(a) Input data

In order to compute the permeability on the basis of equation 2.19 it is first necessary to calculate the zero frequency and infinite frequency velocities V_0 and V_{∞} . To accomplish this the following parameters must be defined (see Appendix A for equation);

Porosity n Bulk density $\rho = \rho_{s}(n-1) + \rho_{w}n$ Solid density ρ_{s} Fluid density ρ_{w} Rigidity modulus G Compressibility of solids "fluid "frame

The measurement of electrical formation factor would allow for

the definition of a lower limit for the mass-coupling factor (Brown, 1980) although an estimate of this value is now possible through porosity (see section 6.4.3).

The porosity and densities may be defined in the laboratory situation by standard formulae and measurements · (see Appendix B).

The rigidity modulus and the frame modulus are interrelated through;

$$B_{f} = \frac{2 (1 + \mu) G}{3(1 - 2\mu)}$$

where μ = Poisson's ratio

Both parameters are directly dependent on the sediment structure, and have little association with the contained fluid. Taylor Smith (1983), in considering the situation, points to Stoll (1980) who suggests that since μ has a limited range from 0.1 to 0.2, Bf can be approximated as G. (B_f varies from approximately 0.9 G to 1.3 G). The rigidity modulus (pV $_{\rm s}^2$) may be assessed directly from shear wave velocity measurements, as may the frame compressibility Cf (the inverse of B_f). Hamdi and Taylor Smith (1982) established an empirical relationship between the modulus obtained from the unloading curve of the oedometer (B_f) and the shear wave velocity; this relationship is reinforced by data from this study (Fig. 6.25). A comparison of the frame modulus determined in this way against the shear modulus obtained from the velocity shows that G is always slightly greater than B_f (Taylor Smith, 1983); this would mean



Fig. 6.25. Frame compressibility, calculated from slope of unloading curve, plotted against shear wave velocity. (After Hamdi and Taylor Smith, 1982).

Poisson's ratio assuming a value of 0.1 or less. For the purposes of this brief study of the theoretical model, the shear wave velocity is used to assess both the rigidity modulus (using density values also) and the frame compressibility. The possible error introduced into the model in doing so is considered to be small in comparison with that inherent in the compressional wave velocity measurement, as discussed later.

McCann (1968) used the compressibility of phlogopite mica as being representative of the solids for a clay $(2.33.10^{-11}$ $m^2/N)$, while Shumway (1958) provided a curve for the compressibility of seawater with temperature based on Wood's emulsion equation (1940). (For DSDP samples 3 and 4 of carbonate ooze the compressibility of CaCO₃ (1.3.10⁻¹¹ m^2/N) is used.

Given the capability of measuring these parameters using the modified oedometer, it should be feasible to predict the permeability of a sample by measuring the compressional wave velocity.

(b) <u>Predicted Permeability values and their interpretation</u>

Using the model to predict a range of permeabilities corresponding to increments in applied axial stress in the oedometer (effective depths), it is possible to plot a diagram such as that in Fig. 6.26. Hamdi and Taylor Smith (1982) showed the relationship between the difference in V_p and V_o , and the permeability of the medium (see Fig. 2.3) which is an



Fig.6.26. Void ratio plotted against permeability (theoretical model) for a single sample during a consolidation test.

expression of the difference between the applied and characteristic frequencies (a function of the permeability); $(V_p - V_o)$ increases with increasing permeability for a fixed applied propagating frequency.

Since the permeability of the medium varies during the consolidation test, the magnitude of $(v_p - v_o)$ should also vary accordingly: both parameters \emptyset and $(v_p - v_o)$ would generally be expected to decrease with increasing strain. However, because $(v_p - v_o)$ tends towards zero anyway (i.e the applied frequency is far less than the characteristic frequency) and at increased strain the error inherent in the compressional wave measurement increases, it is difficult to acccurately monitor the variation. Furthermore, since $(v_p - v_o)$ approaches zero, and since the error in the compressional wave velocity may reasonably be larger than $(v_p - v_o)$, it is possible to define the ratio v_p/v_o in equation 2.19 as less than unity; under these circumstances the model breaks down since the equation demands the square root of a negative number.

A consequence of this feature of equation 2.19 is rather more important since for a value of V_p equal to V_o the model returns a null value of permeability; thus as V_p approaches V_o , the permeability Ø decreases rapidly, and a minimal error in V_p can produce a large change in the magnitude of Ø.

Fig. 6.27 is a modified form of Fig. 6.26 but includes error margins corresponding to the effect of the error in the compressional wave velocity measurement (solid lines). The dotted lines show the range where V_p is supposed equal to (V_o



Fig. 6.27. Reproduction of figure 5.5. with error bars on predicted permeabilities corresponding to errors in compressional wave velocity measurement. Dotted error bars correspond to V_p = V_o + 0·1 m/s.

+ 0.1m/s) and are produced since for most points $(v_p - v_o)$ is less than the error margin in v_p . The overall impression presented by Fig 6.27 is the limitation of the present laboratory set up whereby v_p is measured on a small sample with increasing error margin at increasing strains. These limitations apply to the present system for samples where the applied frequency is far less than the characteristic frequency. However, it does show the success in delineating the maximum value of permeability expected for the sample, particularly at low strain where the error in v_p is small.

Table 6.1 shows values of measured (oedometer theory) and predicted permeability for each of the samples tested. Fig. 6.28 is a plot of predicted permeability against measured permeability for each surficial sample at an axial load of 49.8 KPa.

The range of predicted permeability exhibited in Fig. 6.28 is between 10^{-6} and 10^{-7} m/s, while the oedometer permeability varies between 10^{-8} and 10^{-9} m/s. These compare with direct measurements on similar materials (Dr.P.Schultheiss, pers. comm.) which provide typical values of $3 \ 10^{-8}$ m/s for e = 3.6 and 5.5 $\cdot 10^{-7}$ m/s for e = 5.0 (core S126/4 which is similar to samples 6 and 7 in this study), and $3 \ 10^{-8}$ m/s for e = 2.5 (samples 1,4,5). This suggests the predicted values are close to those measured although it is impossible to conduct any detailed analysis without simultaneous measurements on single samples of material.

TABLE 6.1

PERMEABILITY VALUES

Kastenlot Gravity Cores

Sample No.	Oedometer 10 ⁻⁹ m/s	Theoretical 10 ⁻⁷ m/s	Oedometer 10 ⁻⁹ m/s	Theoretical 10 ⁻⁷ m/s
Applied	Pressure 49	.8 kPa	398.4	kPa
l	8.3	10.5	1.7	15.4
2	-	. –	-	-
3	11.8	10.1	1.2	6.8
4	7.3	10.7	0.9	5.7
5	3.4	2.5	1.3	4.0
6	4.6	5.2	0.6	4.3
7	8.9	8.0	1.7	10.2
8	2.5	3.3	0.6	1.6

Deep Sea Drilling Project Samples

No.	Applied Pressure (kPa)	Permeab	Permeability			
		Oedometer m/s	Theoretical m/s			
1	450	$6.1 \ 10^{-10}$	6.0 10 ⁻⁶			
2	281	1.2 10 ⁻¹⁰	4.0 10 ⁻⁶			
3	450	9.8 10 ⁻⁸	6.3 10 ⁻⁷			
4	900	1.6 10 ⁻⁸	15.0 10 ⁻⁷			



Fig. 6.28. Predicted permeability (theoretical model) plotted against measured permeability (oedometer) for the surficial sediments.

For the DSDP samples the number of data points is small due to the large values of preconsolidation pressure and the nature of the e-logP curve. It is possible to see a similar relationship to that for the surficial sediments in the carbonate ooze samples (DSDP 3 & 4 : see Table 6.1), although the red clay samples (DSDP 1 & 2) exhibit surprisingly very large discrepancies between the measured and predicted permeabilities. This may be due to an inaccurate choice of input values (for example the compressibility of the solids) or an undetected error in the original experimental data. The sparcity of data and the varied nature of the samples thus prevent any further analysis beyond that for the surficial sediments.

6.5.4 <u>Summary</u>

Empirical predictions of permeability utilising electrical formation factor or compressional wave velocity can be made although because both relationships describe general trends considerable errors may occur. Additional control data can help to reduce this scatter; initial void ratio in particular appears to segregate the permeability behaviour of different types of sample. The use of shear wave velocity in extracting some measure of the packing structure for a given void ratio is also suggested.

The theoretical model (Hamdi and Taylor Smith, 1982) enables an upper limit to be assigned to the permeability of a particular material providing sufficient input data is

available. The instability of the model with respect to fine grained material and the error margins dictated by the experimental arrangement prevent the monitoring of any subtle changes in permeability during the consolidation test. However, it does appear possible to predict the permeability of a deep sea sediment to within an order of magnitude.

6.6 <u>Thermal Conductivity Prediction and Electrical</u> <u>Formation Factor</u>

In Chapter 2 thermal diffusivity was shown to be defined in terms of the thermal conductivity and volume heat capacity; the latter being a function of the porosity of a saturated sediment.

6.6.1 <u>Thermal conductivity and porosity</u>

Bullard and Day (1961) found the thermal resistivity (the reciprocal of the conductivity) to be linearly related to the water content;

$$R = 1/k = (161 + - 14) + (651 + - 30)w$$
 (6.10)

This equation has been used to compute thermal conductivities corresponding to the porosities at which measurements were made; Fig. 6.29 is a plot of computed against measured conductivity with a line of slope = 1 also drawn. The agreement is reasonably good considering the errors in both the measured and computed values.



Fig.6.29. Predicted thermal conductivity plotted against measured thermal conductivity. (Prediction from Bullard and Day, 1961; eq.6.1.). Slope of line is 1.

Results for a series of artificially prepared sands exhibiting a wide variety of size and shape (Lovell, 1981) show a relationship of the form of equation (6.4);

 $k_b = k_s^{(1-n)} k_w^n$ (after Sass <u>et al</u>, 1981)

i.e. $\log k_b = n (\log k_w - \log k_s) + \log k_s$

(data reproduced in Fig. 2.4) based on the geometric equation (Sass <u>et al</u>, 1971). A similar plot of the data from this study is presented in Fig. 6.30 showing a similar dependence on porosity for deep sea sediments. Again the agreement is reasonably good.

The scatter in both plots is attributed to the strict boundary conditions applicable to the needle probe with regard to the sample size which effectively reduce the precision of the measurements. How well such a relationship would fit data for samples from depth in the sediment column where time effects, bonding, and cementation may occur rather than the simple mechanical loading of the oedometer, is difficult to estimate.

6.6.2 <u>Geophysical prediction of porosity and thermal</u> <u>conductivity</u>

Since both constituent parameters defining the thermal diffusivity of a saturated sediment appear to be related to the porosity, and since mineralogical variations have little


Fig. 6.30. Measured thermal conductivity plotted against porosity for the surficial sediments.

effect on either the heat capacity (Bullard, 1954; Hillel, 1980) or the thermal conductivity (Ratcliffe, 1960), the prediction of thermal diffusivity in marine sediments is reduced to the problem of accurately predicting the porosity.

Compressional wave velocity shows a clear dependence on the porosity of an unconsolidated marine sediment (Wood, 1940; Nafe and Drake, 1957) although empirically the relationship exhibits a scatter to which Wood's emulsion equation represents a lower bound. Particularly at high porosities typical of deep sea clays, the relationship is poorly defined with the P-wave velocity often exhibiting a value less than that for seawater, and a value that changes little with changes in porosity. This notable feature detracts from any expected relationship between V_p and the thermal parameters of the medium since whilst the latter may depend purely on the volume fractions of the constituents present, the compressional wave velocity is obviously dependent on additional properties of the medium, not expressed in terms of its porosity. A plot of thermal conductivity against compressional wave velocity shows a certain obvious lack of precision in interrelating the two properties (Fig. 6.31).

Electrical formation factor has been shown to exhibit an apparently unique relationship with porosity for each sample (Figs. 6.12 and 6.13), while a general equation may describe the suite of samples together (Fig. 6.16). By combining an equation relating formation factor with porosity and an equation relating thermal conductivity with porosity, it is possible to achieve equations relating formation factor to



Fig. 6.31. Measured thermal conductivity plotted against compressional wave velocity for the surficial sediments.

thermal conductivity; this analysis is similar to that of Hutt and Berg (1968).

Fig. 6.32 is a plot of formation factor-thermal conductivity with the graphs of three such equations also shown. The equations used are drawn from the four presented in Table 6.2. The best fit appears to be that of the polynominal - geometric; this is to be expected since both equations on their own showed good correlations with porosity.

A plot of the predicted thermal conductivity using this model against the measured thermal conductivity is presented in Fig. 6.33.

Thermal diffusivity values may be computed from the volume heat capacity and thermal conductivity of a sample (e.g. eqs. 2.23 -2.26). The resulting values are plotted in Fig. 6.34 against electrical formation factor and show a similar distribution to that of Fig. 6.32 for thermal conductivity - formation factor.

6.6.3 Prediction of sediment column porosity

In order to predict the variation of the thermal character of the sediment column it is necessary to predict the variation of porosity with depth. Hamdi and Taylor Smith (1981) have suggested a technique whereby this may be achieved from the in-situ measurement of the compressional wave velocity, void ratio and permeability, once the gradient of velocity with overburden pressure for the sediment in question can be defined. The particular behaviour of a sediment in-

- * POLYNOMIAL GEOMETRIC
- ARCHIE_e GEOMETRIC_t
- POLYNOMIALe BRUGGEMANt



Fig.6.32, Measured thermal conductivity plotted against electricat formation factor for the surficial sediments. The three models are based on the equations in table 1.

TABLE 6.2

ELECTRICAL FORMATION FACTOR - POROSITY RELATIONSHIPS

Archie (e)

 $FF = n^{-2}$ (eq. 2.12)

Polynominal (e)

n = 1.3861 - 0.4626 (FF) + 0.0833 (FF)² - 0.0073 (FF)³ (data from this study, see section 4.4.3).

THERMAL CONDUCTIVITY - POROSITY RELATIONSHIPS

Geometric (t)

 $K_{\rm m} = K_{\rm s}^{(1-n)} K_{\rm w}^{(n)}$ (eq. 4.4).

Bruggeman (t)

$$n = \frac{k'_{m} - k'_{s}}{k'_{m}^{1/3} (1-k'_{s})}$$

(Hutt & Berg, 1967)

k'; are normalised conductivities with respect to the pore fluid.



Fig. 6.33. Predicted thermal conductivity (Polynomial_e – Geometric_t) plotted against measured thermal conductivity for the surficial sediments. Slope of line is 1.



situ can be defined by establishing master curves relating velocity to overburden pressure and void ratio to overburden pressure through laboratory experiments. Although the study only considered normally consolidated sediments, the results successfully predicted the consolidation behaviour for a number of samples where the initial void ratio was less than 2. Extension of the method to overconsolidated samples, and to samples with an initial void ratio greater than 2 (for void ratios between 2 and 4 there is little change in compressional wave velocity) requires further study, together with an analysis of which value of permeability relates to settlement behaviour (see section 6.5).

6.6.4 <u>Summary</u>

The thermal diffusivity of a deep sea marine sediment may be computed from a knowledge of the porosity of the sample. This may be assessed mechanically (by oven drying) or by geophysical means utilising electrical formation factor or compressional wave velocity measurements. Given a value for porosity, equations 6.4 and 6.10 may be used to give a reasonable estimate of thermal conductivity, while the polynominal relating formation factor and porosity may be utilised with equation 6.4 to produce a formation factorthermal conductivity relationship. Evaluation of sediment column thermal conductivity requires knowledge of the variation of porosity with depth; this can be achieved within certain limitations by geophysical means though extension of

the technique to overconsolidated and high void ratio sediments is required.

The overall scatter apparent in the thermal conductivity measurements may be due to the experimental arrangement of using very small samples, though some of the deviation may be due to the effects of anisotropy or incomplete saturation of the sample. These effects, particularly that of anisotropy, require further study.

7. <u>CONCLUSIONS</u>

The aim of this research has been to examine the electrical resistivity - thermal conductivity - porosity properties of a marine sediment both in terms of their own interdependences and of their interrelationships with other geophysical and geotechnical quantities. In order to achieve this examination, it has been necessary to consider sands and clays as two separate sediment systems, which indeed they are. The results of this rather broad investigation are both considerable and varied, and whilst each individually identified facet may not have been completely explored, certain important interrelationships have been initially identified or confirmed.

Overwhelmingly what has emerged has been the close interrelationship between permeability and the geophysical parameters, electrical resistivity and compressional wave velocity, and the capability of predicting the one by measuring the other. This is particularly of importance in relation to Biot's prediction of compressional wave velocity.

In tandem to this is the interrelationship between thermal conductivity and the porosity of a saturated sediment, and the extension of this dependency to allow the prediction of thermal conductivity by measurement of electrical formation factor.

These major features of the investigation, together with some of the less dominant though important conclusions, are presented in more detail below.

1. Thermal conductivity in both clean saturated sands and deep sea clays is dependent on the water content of the sample. For a fully saturated sediment this dependency may be expressed in terms of the porosity of the sample.

2. In deep sea clays the mineralogical composition does not appear to affect the thermal conductivity, the dominant control being the amount of fluid present. For sands, however, the mineralogy can have some determinable effect (the thermal conductivity decreasing with increasing proportions of carbonate - as compared with quartz); this effect is only noticeable at high variations in mineralogical composition and is secondary to the water content dependency outlined in (1).

3. Thermal conductivity in deep sea clays may be adequately described, in terms of the water content of a sample, by the empirical equation of Bullard and Day (1961);

R = 1/k = (161 + - 14) + (651 + - 30)w

where w = water content

While for both the deep sea clays and clean saturated sands studied, two equations predict the thermal conductivity on the basis of porosity;

$$k_{b} = k_{s1}^{n1} k_{s2}^{n2} k_{w}^{n}$$

(Geometric model after Sass <u>et al</u>, 1971)

where n = porosity n1 + n2 = 1 - nsl,s2 are solid fractions, w pore fluid fraction.

$$n = \frac{k'_{m} - k'_{s}}{k'_{m}^{1/3} (1 - k'_{s})}$$

(Bruggeman, cited Hutt and Berg, 1968)

4. Employment of the geometrical model to the data obtained in this study yields the following thermal conductivity values;

Seawater	0.61 -0.64 W/mK
Carbonate shell particles	3.32 W/mK
Quartz sand particles	8.58 W/mK
Deep sea clay particles	2.01 W/mK

5. Thermal conductivity may be predicted using electrical formation factor measurements through the common relation, porosity (see conclusions 7 & 8). Good correlations are achieved by using two equations, one each relating thermal conductivity or electrical formation factor to porosity. For deep sea clays either of the Geometric or Bruggeman equations (thermal conductivity - porosity) may be combined with a 3rd degree polynomial relating formation factor to porosity. For clean saturated sands the best results appear to be obtained with the combination of Archie (electric) and Geometric (thermal) equations.

6. The saturated sand samples exhibit both thermal and electrical anisotropy. Recognition of these phenomena, and their subsequent measurement and analysis, may improve the already good correlation between the two. Since the deep sea clays exhibit some slight compressional wave velocity anisotropy, it is to be expected they would also exhibit anisotropy in terms of thermal conductivity and electrical formaton factor.

7. For deep sea clays, each sample produces a linear trend on a log-log plot of formation factor - porosity, representing the change in formation factor with mechanical loading (producing decreasing porosity values). Each of these trends may be described by a Winsauer type equation;

$$FF = Cn^{-m}$$
 (eq. 2.13)

where m varies between 1.36 and 3.50 and C varies between 0.95 and 1.25

Alternatively, all of the data may be crudely approximated by Archie's law (C =1 in the above equation) with an exponent equal to 2. A plot of all of the data points, however, suggests the overall relationship between formation factor and porosity is not linear but is better represented by a 3rd degree polynomial;

 $n = 1.3861 - 0.4626(FF) + 0.0833(FF)^2 - 0.0073(FF)^3$

& For clean saturated sands, porosity exhibits a linear relationship to electrical formation factor when plotted on a log - log scale. Each sample may be described by Archie's empirical law;

$$FF = n^{-m}$$
 (eq. 2.12)

where m varies between 1.9 for lamellae shaped particles to 1.4 for equidimensional quartz grains. A value as low as 1.3 may be returned for perfectly spherical particles. The absolute range of porosity - formation factor varies, increasing with increasing spread of sizes. The range of quartz and shell sands studied may also be expressed in terms of a 3rd degree polynomial;

 $n = 1.4154 - 0.4799(FF) + 0.0687(FF)^2 - 0.0033(FF)^3$

9. Permeability, like porosity, is a function of the proportion of pore space within a sediment. Unlike porosity, however, it is determined by the nature of the interconnection of the pores and is thus a directionally dependent quantity. While porosity cannot therefore exhibit anisotropy, permeability can.

10. Permeability shows a linear relationship with electrical formation factor, for clean saturated sands when plotted on a log-log basis;

$$\emptyset = C (FF)^{-X}$$

This empirical equation allows permeability to be predicted to within an order of magnitude. Both C and x increase with decreasing sphericity of the particles, while for increasing

spread of sizes C decreases and x increases. The coefficient C would seem to be a function of the pore size, while x can be shown to exhibit a certain dependence on the exponent m in Archie's empirical law relating formation factor to porosity. The coefficient x would thus become a function of the particle shape, or indirectly pore shape.

C varies from 0.0006 to 0.0889, x varies from 0.0022 to 0.0064, for Ø in m/s.

11. Permeability - electrical formation factor measurements on clean sands show a clear repeatability which suggests that the relationship between the two is single valued for any one sample.

12. Permeability values from direct measurements on clays are between one and two orders of magnitude greater than permeability values derived from the consolidation test. The former relates to the application of a hydraulic gradient across the stationary skeletal framework of the sample, the latter to the expulsion of pore fluid from a sample during the process of consolidation. Which of these values is pertinent to the present study, and precisely why the difference between them is so large, remains unanswered.

13. Empirical relationships between void ratio and oedometer-derived permeability for deep sea clays have been shown to exist;

 $e = b \log \emptyset + C$

Both coefficients b and C decrease with decreasing initial void ratio. For the surficial suite;

Samples	1,	3,	6 (5 < e ₀)		
	e	=	0.5910 log _e Ø +	14.9846	R = 0.85
Samples	2,	4,	7 (3 < e ₀ <5)		
	e	3	0.3266 log _e Ø +	8.8626	R = 0.84
Samples	5,	8,	9 (e ₀ < 3)		
	e	=	0.2936 log _e Ø +	7.7527	R = 0.76

The results for the DSDP samples show remarkable similarities to these equations for similar initial void ratios. The empirical equations allow the permeability to be assessed from the initial void ratio and void ratio at a given state to within an order of magnitude.

Similar groupings of the shear wave velocity data suggest that since permeability may be influenced by packing structure, it may be possible to further delineate the permeability variations for a given initial void ratio by utilising shear wave velocity measurements.

14. Oedometer derived permeability values for deep sea clays exhibit a linear relationship on a log - log basis with electrical formation factor. Individual least squares fits produce similar equations for each member of a core sequence, thus three groups are defined corresponding to each core;

Samples 1,3,4,5 (Core D10406/) $\emptyset = 2.5334 \ 10^{-7} \ (FF)^{-6.0043}$ R = 0.87 Samples 6,7 (Core S126/2-) $\emptyset = 1.1739 \ 10^{-5} \ (FF)^{-11.4056}$ R = 0.93 Samples 8,9 (Core S126/15-) $\emptyset = 4.7645 \ 10^{-7} \ (FF)^{-6.8568}$ R = 0.94

The notable absence of any correlation here with initial void ratio, or with particle size data (there being little to distinguish between any of the samples' particle size curves) suggests the dominant effect may be particle shape, effected by differences in mineralogy between the cores. Since differences in mineralogy may effect differences in the nature and quantity of the adsorbed water layer the important facet may be pore shape rather than mineral particle shape. Empirically, the equations allow the permeability to be established to within an order of magnitude.

15. Compressional wave velocity measurements in deep sea clays show a relationship with oedometer-derived permeability values of the form;

$$V_p = d - C \log_e \emptyset$$

While the correlation between the two parameters is theoretically good, changes in permeability (or porosity) of a high porosity sediment often do not effect measurable changes in the compressional wave velocity. Thus, while the equation may empirically predict the permeability of a sediment to

within 2 orders of magnitude, subtle differences in permeability within one sediment type may not necessarily be detected.

16. Compressional wave velocity measurements on deep sea clays, when used in a theoretical model based on Biot's equations, produce permeability values which bear a relationship with oedometer-derived permeability values, although there is a difference of two orders of magnitude between the two sets of results. The use of electrical formation factor in defining a mass coupling factor (FF.n) can i mprove this prediction under certain conditions; this term is often referred to in studies of fluid flow in sands as a shape factor. The permeability prediction is considered to be within an order of magnitude, and appears to bear some agreement with direct permeability measurements.

17. For the DSDP samples, the two carbonate oozes show similar results in terms of a theoretical model. The two red clay samples show a wider differential between the two sets of permeability values; the reason for this is unclear, but with one of the samples (DSDP 1) the compressional wave received pulse was ill-defined. The high preconsolidation pressures in all four DSDP samples allow for very few data points to be extracted for analysis in terms of permeability values.

8. FURTHER WORK

The obvious extension of any laboratory based study of the natural environment is to transfer the work into the real world and test whether the laboratory-defined relationships are supported by in-situ measurements. Before that is done, or perhaps in tandem with that transition, there are several areas of interest arising from this study which require further laboratory investigation.

1. <u>Anisotropy</u>

The role of anisotropy in defining both the thermal and hydraulic characters of sands has been identified. Similar effects may be considered likely in clay sediments. The relationships defined between electrical flow and these characteristics have shown initial potential, and since electrical flow appears to be also modified by anisotropic media, the potential for exploring and defining the three dimensional nature of marine sediments is apparent. In particular, porosity - formation factor requires further study to identify the effects of anisotropy on the relationship.

2. <u>Particle Shape - Pore Shape</u>

Electrical and hydraulic flow both appear to be modified by changes in particle shape. General assumptions throughout the literature are that this effect is a function of pore shape, and the results from this study lend further weight to

this. The range of porosity exhibited, together with the absolute magnitude, is influenced not only by particle shape, but by the spread of sizes present; both of these effects may alter the pore shape and size. Thermal conductivity for a saturated sediment appears to be primarily a function of porosity, although like electrical flow, it too is affected by anisotropy within the medium.

These effects suggest that a close examination of the role of the particle and its pore space in defining the physical nature of the sediment is required. In particular, precise control is sought after in analysing the range of sizes and shapes, and the variety of ways they may interact.

3. <u>Redeposition Techniques</u>

Conclusion (11) suggests that the permeability formation factor relationship, while not expressing a single unique equation for all sediments, is single-valued for any one sample. This is in contrast to the work of others. The single valued relationship, however, stems from results for similar packing techniques, or redeposition techniques. It may be possible that a different packing structure, arrived at by a different means of depositing the sample may affect the hydraulic character more than the electrical, or vice versa. Experiments with various packing techniques, particularly moving away from solely vertical compaction and movement of grains, would appear to be necessary. These could form an extension of the work required on pore shape, since both are

concerned primarily with packing structure.

4. In-situ Application

In transferring the laboratory derived relationships and techniques to the in-situ environment of the continental shelf or deep ocean floor, it is important to consider the limitations imposed in a controlled laboratory environment and hence the contrasting conditions to be found in-situ. Of particular note is the effective homogeneity, and in some respects isotropy of a laboratory sample. In-situ, both environments exhibit inhomogeneities in the form of fissures and fractures, variations in void ratio, and repetitive laminar beds. For example, with respect to deep sea clays and the disposal of high level radioactive waste, an artificially imposed inhomogeneity in the form of the remoulded material surrounding a buried canister is vitally important, since this material must in all probability form a weak link in any containment process.

To verify whether the laboratory relationships hold for in-situ conditions, tests are required at a specific site, where simultaneous measurements of all the necessary parameters are required to be made.

Initially these measurements could be made on land using an overconsolidated material (e.g. Oxford clay, London clay) where bore hole(s) could be used to assess the physical Properties required against the geophysical prediction. The Costs of such an exercise would be comparatively small (as

opposed to marine operations) and would enable thorough field testing of the predictor equations and models.

Measurements in a marine environment on near surface sediments could be readily accomplished using probe vehicles (e.g. Bennell <u>et al</u>, 1982). However, for assessing the usefulness of the relationship for predicting parameters at depth, advances in measurement techniques are required, particularly with electrical resistivity. Additional control may be provided through bore holes and DSDP samples.

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<u>APPENDIX A</u>

Theoretical Model

The theoretical model used in this study to evaluate permeability values for deep sea sediments through compressional wave velocity measurements is that developed by Hamdi and Taylor Smith (1982). The model is based on Biot's equations and relates the compressional wave velocity measured at a fixed frequency to computed velocities at zero and infinite frequencies in terms of sediment porosity and permeability. This appendix summarises the nature of the model, though for a full detailed analysis reference should be made to the original paper.

<u>Notation</u>

```
ρ
    bulk density
    solid density
ρ
    fluid density
Pw.
   fractional porosity
absolute dilatation of solid
nï
 e
Ę
    relative dilatation (n(e-E))
    absolute dilatation of fluid
XYZ
    sediment elastic moduli (units of pressure)
ь
    mass coupling factor
F
    electrical formation factor
R = C_{\pm}/C_{\pm}
    compressibility of solid particle
Cs .
Cw
    fluid compressibility
    frame (bulk) compressibility
C1
    rigidity (shear) modulus
G
    angular frequency
ω
t
    wave number
    measured compressional wave velocity
٧p
    "zero frequency" compressional wave velocity
Vo.
    "infinite frquency" compressional wave velocity
۷_
    acceleration due to gravity
 g
    permeability of sediment (units of area)
permeability of sediment (units of velocity)
 K
ø
    absolute viscosity of water
μ
```

In order to define a mathematical model which is relatively simple to handle, certain assumptions are necessary. The sediment consits of an elastic skeleton, itself somehow formed from elastic grains, the voids of which are totally filled with a compressible fluid which is free to flow in them. If a compressional wave propagates through a sediment the different compressibilities of the constituents provides for different behaviour of the pore space and the fluid contained there in. There is a greater reduction in the seawater volume than in the surrounding solid volume and the water tends to flow through the framework under the influence of an induced pressure gradient. Using this concept Biot (1962) has obtained two fundamental equations which describe this propagation, where the relative motion of the fluid takes place according to Darcy's law and where scattering effects can be ignored:

$$\nabla^{2}(Xe - Y\xi) = \frac{\delta^{2}}{\delta t^{2}} \left(\rho e - \rho_{\omega} \xi \right)$$
 (A1.1.)

and

$$\nabla^{2}(Ye - \vec{Z}\xi) = \frac{\delta^{2}}{\delta t^{2}} (\rho_{w}e - m\xi) - \frac{\mu}{\kappa} \cdot \frac{\delta\xi}{\delta t}$$
(A1.2.)

The parameter m has the dimensions of density and accounts for the fact that not all of the fluid moves in the general direction of the pressure gradient; this is a consequence of the shape of the grains, their interconnection and the form of the cavities. The parameter may be written

$$m = \frac{b\rho_w}{n}$$
(A1.3.)

where b is a mass coupling factor which must have a numerical

value greater than unity (b=1 indicates there is no fluid-solid coupling).

Brown (1980) suggests a lower limit for b may be obtained by an analogy with electrical conduction;

The parameters X, Y, & Z, are each a form of sediment elastic modulus and may be related to the compressibilities of the constituents;

$$X = \frac{(1-R)^2}{(1-n-R)C_s + nC_w} + \frac{1}{C_f} + \frac{4}{3}G$$
 (A1.5.)

$$Y = \frac{1 - R}{(1 - n - R)C_{s} + nC_{w}}$$
(A1.6.)

$$Z = \frac{1}{(1 - n - R)C_{s} + nC_{w}}$$
(A1.7.)

Biot's differential equations predict the existence of two plane dilatational waves with very different properties. The first kind of wave is a true wave analagous to the dilatational wave which propagates in ordinary elastic media, attenuating relatively slowly. The second kind of wave attenuates rapidly, is in the nature of a diffusion wave moving much more slowly than the first kind of wave, and is only observable in the immediate vicinity of the wave source. It would seem that the wave of the first kind is transmitted mainly through the fluid - the one usually measured as the compressional wave. and

$$\omega_{c} = \frac{\mu}{\kappa} \left(\frac{\rho}{m\rho - \rho_{w}^{2}} \right)$$
 (A1.11.)

Equation (Al.11) may be rewritten to include the mass-coupling factor b and to convert the permeability coefficient k from units of area to units of velocity (ϕ):

$$\omega_{c} = \frac{gn}{g} \left(\frac{\rho}{b\rho - n\rho_{w}} \right) \qquad (A1.12.)$$

with

$$K = \frac{g\rho_w}{g\rho_w}$$

From equations (A1.8) and (A1.12);

$$\varphi = \frac{ng}{\omega} \left(\frac{\rho}{b\rho - n\rho_w} \right) - \frac{V_o^2}{V_o^2} \sqrt{\left[\frac{\left[(V_p^2 / V_o^2) - 1 \right]}{\left[1 - (V_p^2 / V_o^2) \right]} \right]}$$

Hence if V_{∞} and V_{∞} can be calculated, a measured value of $V_{P_{1}}$ should enable a value of permeability (\emptyset) to be calculated for a sediment is known physical state (i.e. known porosity).

APPENDIX C

Modified Oedometer : Electronics Diagrams

1000 512 BYTES Y4	1200 512 BYTES ROM 7 13 FF ROM 7 14 00 512 BYTES ROM 6 15 FF ROM 6 16 00 512 BYTES ROM 5	1917 1915 1915 1915 1000 512 BYTES ROM 3 1015 ROM 3 1015 ROM 2 ROM 2 ROM 2 ROM 2 ROM 1 1005 ROM 1 ROM 1 ROM 1 ROM 1 ROM 1 ROM 1 ROM 2 ROM 1 ROM 3 ROM 1 ROM 3 ROM 3	IFFS IFFS THER INTERRUPT IFFS FROM WAIT STATE IFFS THER INTERRUPT IFFA EXTERNAL INT. IFFS SOFTWARE INT. IFFF RESET IFFF RESET
Y0 0000 UNUSED Y3 0011 HBYTE 709 RRD1	00000000000000000000000000000000000000	VILLU I.D. I.E. / IV. P.M.D.3 0C44E L. BYTE 7109 BRD3 0C45 L. BYTE 7109 BRD3 0C66 UNUSED 0C66 0C06 ADC GHOST 0C66 Y1 0C66 UNUSED 0C06 ADC GHOST 0C66 0C06 OUNUSED 0C66	TZ OFFF ADC GHOST OFDOME TER DATA LOGGER L6805) HEMORY MAP
00.80 8 BARS BAR DISP 00.81 DISP. GHOST 00.63 01SP. GHOST 00.64 8 BARS BAR DISP.	0085 DISP. GHOST 0087 DISP. GHOST 0088 6 BARS BAR DISP 0088 6 BARS BAR DISP 0088 DISP. GHOST 0088 DISP. DISP. 0099 DI 0099 DI 0099 DI 0099 DI 0099 DI 0099 DI	0092 D2 B011UN UISP 0093 D1 B0TT0H DISP 00995 DISP GH0ST 00998 UNUSED 00998 UNUSED 00990 DISP GH0ST 00997 DISP GH0ST 0080 OISP GH0ST 01400 OISP GH0ST	0155 03000 14 RAH 18 RAH
0000 PORT A	0002 EXT.MEM. SPACE 0003 EXT.MEM. SPACE 0004 DDR A 0005 DDR B 0006 EXT.MEM. SPACE 0006 EXT.MEM. SPACE 0007 EXT.MEM. SPACE 0007 EXT.MEM. SPACE 0007 EXT.MEM. SPACE 0008 EXT.MEM. SPACE 0006 EXT.MEM. SPACE 00010 EXT.MEM. SPACE 0010 EXT.MEM. SPACE	SCRATCH PAD RAM	00315 0010 0075 0075







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TEMP. PROBE DISPLAY BOARD BL





.

Sec. 1





APPENDIX D

Tables Of Results

- 1. Artificial Sand Samples
- 2. Natural Sand Samples
- 3. Surficial Deep Sea Clay Samples
- 4. Deep Sea Drilling Project Samples

1. Artificial Sand Samples

n = fractional porosity w = fractional water content FF = horizontal electrical formation factor Ø = permeability (m/s) k = thermal conductivity (W/mK)

All electrical values relate to a horizontal orientation.

All thermal conductivity values are measured with the needle orientated vertically.

Samples (a - d) : increasing spread of sizes
Samples (e - f) : similar spread of sizes, different means
Samples(i - m) : Shell / quartz sand mixtures
shell% decreasing
i = 100%

j = 70% k = 50% Shell l = 30% m = 0%

Sample n : Glass ballotini

(e) = estimated value from adjacent data

w

FF

k

Ø

SAM	?LE	:	a
-----	-----	---	---

0.4413 0.4336 0.4204 0.4009 0.3830 0.3641 0.3509 0.3440	0.2307 0.2252 0.2159 0.2026 0.1907 0.1786 0.1703 0.1660	3.28 3.36 3.51 3.73 4.00 4.29 4.47 4.58		2.57 2.59 2.82 3.00 3.18 3.16 3.52
		3.47 3.57 3.73 3.89 4.03 4.17 4.29 4.43	0.002565 0.002412 0.002303 0.001861 0.001811 0.001665 0.001356 0.001244	2.32 2.46 2.49 2.61 2.71 2.77 2.87
SAMPLE: b 0.4277 0.4244 0.4148 0.4059 0.4002 0.3914 0.3823 0.3761 0.3698 0.3633 0.3568 0.3501 0.3432 0.3362 0.3291 0.3232	0.2210 0.2187 0.2120 0.2059 0.2021 0.1962 0.1903 0.1862 0.1822 0.1780 0.1738 0.1655 0.1613 0.1570 0.1535	3.50 3.53 3.65 3.71 3.77 3.85 3.98 4.08 4.18 4.29 4.40 4.29 4.40 4.57 4.67 4.80 4.90 5.04		2.55 2.58 2.87 3.04 2.74 3.33 3.56 3.83 4.12 3.58 3.61 3.97 3.37 3.65 3.12 3.83
0.4272 0.4168 0.4026 0.3890 0.3710 0.3518 0.3302	0.2207 0.2134 0.2037 0.1947 0.1830 0.1708 0.1577	3.49 3.58 3.72 3.85 4.18 4.51 4.94		2.58 3.02 3.27 3.60 3.43 3.60 3.50

n	W	FF	Ø	k
SAMPLE: b	(cont.)			
		3.62 3.77 4.05 4.39 4.58 4.80	0.001733 0.001714 0.001312 0.001034 0.000895 0.000782	2.56 2.56 2.69 2.84 3.00 2.99
SAMPLE: C				
0.4173 0.4094 0.3986 0.3873 0.3696 0.3509 0.3309 0.3098	0.2138 0.2083 0.2010 0.1935 0.1821 0.1703 0.1581 0.1456	3.66 3.73 3.82 3.98 4.23 4.57 4.99 5.48		2.78 3.30 3.46 3.30 3.28 3.63 3.42 3.69
0.4173 0.4041 0.3873 0.3696 0.3509 0.3309 0.3098	0.2138 0.2047 0.1935 0.1821 0.1703 0.1581 0.1456	3.60 3.74 3.92 4.17 4.54 4.93 5.37		2.78 3.09 3.37 3.42 3.38 3.58 3.60
		3.72 3.86 4.01 4.25 4.42 4.63 4.90 5.07 5.30	0.001054 0.000967 0.000858 0.000757 0.000652 0.000521 0.000442 0.000388	2.49 3.29 2.92 2.97 3.07 3.52 3.49 3.56
SAMPLE: d				
0.3728 0.3642 0.3463 0.3339 0.3209 0.3074 0.2934 0.2800 0.2712	0.1841 0.1786 0.1674 0.1599 0.1521 0.1442 0.1362 0.1286 0.1238	4.26 4.37 4.51 4.83 5.12 5.43 5.93 6.44 6.82		2.48 2.59 2.89 2.94 2.81 3.09 3.14

SAMPLE: d (cont.)

4.53	0.000412	2.87
5.20	0.000362	2.94
5.63	0.000247	3.54
6.21	0.000178	4.12
6.60	0.000128	3.60
4.59	0.000409	2.97
4.81	0.000397	3.05
5.12	0.000344	3.25
5.55	0.000270	3.46
5.82	0.000215	3.30
6.07	0.000186	3.68
6.57	0.000137	3.65
6.74	0.000114	3.96

Ø

n	W	FF	Ø	k
		,		
SAMPLE: e				
0.4475 0.4375 0.4272 0.4165 0.4053 0.3938 0.3817 0.3692 0.3562 0.3474	0.2353 0.2281 0.2208 0.2134 0.2057 0.1980 0.1900 0.1819 0.1737 0.1682	3.17 3.25 3.35 3.45 3.57 3.69 3.88 4.09 4.29 4.45		2.62 2.86 3.02 3.05 3.09 3.03 3.30 3.30 3.35 3.49 3.54
		3.14 3.35 3.46 3.71 3.99 4.06	0.021913 0.015541 0.016782 0.011512 0.010090 0.009246 0.009120	2.49 2.45 2.62 2.69 2.85 3.03 3.36
SAMPLES: f	and m			
0.4444 0.4367 0.4206 0.4035 0.3854 0.3661	0.2328 0.2273 0.2160 0.2043 0.1922 0.1798	3.32 3.39 3.57 3.79 4.07 4.33		2.50 2.80 2.65 2.80 2.77 3.18
		3.23 3.46 3.68 3.83 3.96 4.28 4.42 4.60	0.003861 0.003504 0.003268 0.002767 0.002429 0.001936 0.002195 0.001798	2.83 2.81 2.90 2.87 3.04 3.14 3.10 3.27

SAMPLE: g				
0.4634	0.2468	3.11		2.47
0.4554	0.2409	3.14	an a	2.59
0.4457	0.2338	3.23		2.78
0.4392	0.2291	3.31		2.82
0.4347	0.2259	3.35		2.74
0.4305	0.2229	3.42		2.80
0.4242	0.2185	3.51		2.84
0.4199	0.2155	3.56		2.75

n	W	FF	Ø	k	
SAMPLE: g	(cont.)				
0.4122 0.4049 0.3974	0.2102 0.2052 0.2002	3.61 3.67 3.73		2.77 2.81 2.82	
0.4587 0.4412 0.4289 0.4215 0.3974 0.3837 0.3725	0.2433 0.2306 0.2218 0.2109 0.2002 0.1911 0.1839			2.41 2.56 2.58 2.61 2.96 3.25 3.31	•
		3.15 3.38 3.54 3.66 3.84 4.01 4.18	0.00104 0.00091 0.00078 0.00073 0.00065 0.00047 0.00043	2.55 2.64 2.67 2.67 2.75 2.89 2.98	
SAMPLE: h					
0.4841 0.4696 0.4595 0.4490 0.4381 0.4267 0.4208 0.4149 0.4088 0.4025 0.3961	0.2628 0.2517 0.2441 0.2364 0.2285 0.2204 0.2163 0.2122 0.2080 0.2080 0.2038 0.1995	3.07 3.10 3.19 3.26 3.40 3.52 3.57 3.64 3.75 3.78 3.88	•	2.41 2.66 2.69 2.60 2.95 3.20 3.17 3.14 3.39	
		2.98 3.07 3.29 3.57 3.57 3.69 3.80	0.000260 0.000233 0.000198 0.000157 0.000157 0.000140 0.000130		
		2.90(e) 3.06(e) 3.36(e) 3.54(e) 3.83(e) 4.06(e)	0.000284 0.000238 0.000182 0.000158 0.000126 0.000105	2.17 2.32 2.54 2.55 2.57 2.69	

م می باد. ۲۰۰۸ کاملا می بادی کاملا می میگرد کار این می میگرد این می این این می این می این می این می این می این ۲۰۰۸ کاملا می می بادی این می این م ۲۰۰۸ کاملا می این می

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SAMPLE: i

0.7140 0.7070 0.6980 0.6910 0.6820 0.6740 0.6650 0.6530	0.4746 0.4662 0.4554 0.4473 0.4370 0.4280 0.4180 0.4051			0.98 1.06 1.00 1.05 1.06 1.08 1.10 1.07
0.7053 0.7002 0.6973 0.6857 0.6716 0.6585 0.6396 0.6291	0.4641 0.4580 0.4546 0.4405 0.4253 0.4110 0.3911 0.3803			0.92 1.00 0.96 0.96 0.97 1.00 1.01
0.6935 0.6907 0.6879 0.6850 0.6820 0.6790 0.6760 0.6728 0.6664 0.6631 0.6597 0.6563 0.6527 0.6492 0.6418 0.6379 0.6340 0.6300		2.09 2.10 2.12 2.13 2.15 2.17 2.19 2.22 2.24 2.26 2.28 2.31 2.33 2.37 2.39 2.41 2.44		
		2.09 2.13 2.22 2.26	0.008571 0.007086 0.005497 0.004038	
	マ・コン・マート (1) 1972 (1) 1 (1) 10 83 (1) (1) 10 83 (1) (1) 10 84 85 (1)	2.09 2.15 2.25 2.32	0.006951 0.006325 0.004914 0.004268	

n	W	FF	Ø	k
SAMPLE: j				
0.6532 0.6440 0.6342 0.6240 0.6131 0.5976 0.5873 0.5764 0.5672 0.5528 0.5452	0.4088 0.3991 0.3889 0.3786 0.3678 0.3528 0.3432 0.3432 0.3331 0.3248 0.3122 0.3056	2.18 2.24 2.29 2.40 2.45 2.55 2.63 2.74 2.79 2.92 3.01		1.40 1.64 1.60 1.62 1.57 1.74 1.82 1.92 2.07 1.78 1.80
	:	2.27 2.36 2.53 2.68 2.82 2.87	0.005154 0.004210 0.004327 0.003322 0.002362 0.002125	1.29 1.26 1.31 1.41 1.51 1.55
SAMPLE: k	•			
0.6089 0.6005 0.5917 0.5881 0.5769 0.5650 0.5524 0.5391 0.5298 0.5200 0.5099	0.3659 0.3578 0.3494 0.3461 0.3357 0.3250 0.3139 0.3024 0.2946 0.2865 0.2783	2.35 2.42 2.42 2.47 2.60 2.67 2.89 2.99 3.01 3.08 3.05		1.72 1.90 1.94 2.09 1.98 2.12 2.34 2.18 2.49 2.46
SAMPLE: 1				•
0.5528 0.5452 0.5374 0.5251 0.5165 0.5076 0.4984 0.4888 0.4789 0.4685 0.4578 0.4466 0.4437	0.3163 0.3097 0.3030 0.2926 0.2856 0.2784 0.2710 0.2635 0.2559 0.2480 0.2401 0.2319 0.2298	2.67 2.69 2.77 2.87 2.98 3.06 3.16 3.25 3.33 3.44 3.57 3.71 3.71		2.16 2.22 2.20 2.32 2.59 2.64 2.49 2.61 2.54 2.84 3.04 2.87 2.91

SAMPLE: 1 (cont.)			
	2.83 2.94 3.10 3.24 3.36 3.46 3.54	0.004697 0.003813 0.003575 0.002801 0.002355 0.002068 0.001978	1.93 2.02 2.10 2.30 2.33 2.31 2.40
			•
SAMPLE: n	с С		
0.3850 0.3740 0.3625 0.3506 0.3445	3.46 3.56 3.68 3.83 3.89		0.87 0.82 0.87 0.84 0.88
	3.52 3.65 3.80 3.89 3.98 3.98 3.98 4.06	0.000873 0.000832 0.000701 0.000662 0.000581 0.000581 0.000515	

FF

W

n

Ø

k

SAMPLE: 0

2.89 2.90 2.94 k

2.10

2.42

2.42

2.49

2.54

2.42

2.83

2.92

2.09

2.82

3.43

2.17 2.08 2.08 2.08 2.08 2.06 2.10 2.01

2.11

(v)

(v)

(v)

(v)

(v)

(v)

(v)

(v)

(h)

(h)

(h)

(h)

0.4747	
0.4716	
0.4645	
0.4609	
0.4562	

0.4521 0.4473 0.4425 0.4375 0.4330 0.4274 0.4222 0.4169 0.4115 0.4061 0.4005 0.3948 0.3901 0.3890

0.4650

0.4558

0.4415

0.4264

0.4105

0.3936

0.3936

0.3878

0.4581

0.4238

0.3848

0.3503

3.39 3.54 3.78 3.92 4.04 4.12 4.16	2.87 2.99 3.09 3.24 3.38 3.50 3.72 3.72	3.07 3.28 3.66 4.09	3.65	2.87 2.91 3.01 3.14 3.28 3.50	2.95 2.98 3.01 3.05 3.07 3.12 3.14 3.20 3.25 3.32 3.38 3.49 3.57 3.60 3.63 3.64
(V) (V) (V) (V) (V) (V) (V)	(h) (h) (h) (h) (h) (h) (h)				
<u>к</u>					
0. 0. 0. 0. 0.	0. 0. 0. 0. 0.				
000 000 000 000 000 000					
357 334 278 243 215 196 190	369 319 284 296 226 181 173				

SAMPLE:	р
---------	---

0.4853 0.4832 0.4790 0.4747 0.4703 0.4659 0.4613 0.4567 0.4521 0.4521 0.4473 0.4425 0.4425 0.4375 0.4325 0.4274 0.4222 0.4169 0.4116	2.89 2.86 2.87 2.89 2.91 3.01 2.96 3.01 3.09 3.15 3.22 3.23 3.25 3.28 3.35 3.39 3.45		
0.4894	2.84		1.75 (v)
0.4790	2.89		2.15 (v)
0.4659	3.02		2.05 (v)
0.4521	3.18		1.86 (v)
0.4375	3.36		2.31 (v)
0.4222	3.68		2.45 (v)
0.4169	3.76		2.47 (v)
0.4832	2.94	· · · · · · · · · · · · · · · · · · ·	1.54 (h)
0.4703	2.99		1.72 (h)
0.4567	3.08		1.69 (h)
0.4425	3.23		1.81 (h)
0.4274	3.39		1.95 (h)
0.4116	3.60		2.05 (h)
0.4005	3.74		1.85 (h)
	2.79 (h) 2.87 (h) 2.93 (h) 3.09 (h) 3.20 (h) 3.29 (h) 3.37 (h) 3.43 (h) 3.43 (h) 3.49 (h) 3.55 (h)	0.000688 0.00053 0.000576 0.000513 0.000444 0.000401 0.000359 0.000398 0.000318 0.000318	1.53 1.49 1.57 1.63 1.69 1.70 1.73 1.76 1.78 1.80

k

Sample p (cont.)

4.12 4.24 4.41 4.52 4.65 4.78 4.89 5.02	(V) (V) (V) (V) (V) (V) (V)	0.000629 0.000624 0.000557 0.000474 0.000442 0.000397 0.000362 0.000323	V
5.02 5.10	(v) (v)	0.000323	

0

k

SAMPLE: q

0.5767 0.5739 0.5702 0.5683 0.5665 0.5628 0.5589 0.5550 0.5511 0.5470 0.5470 0.5429 0.5387 0.5345 0.5345 0.5301 0.5257 0.5212 0.5166 0.5119 0.5071	2.55 2.56 2.58 2.60 2.62 2.65 2.65 2.68 2.71 2.74 2.78 2.81 2.85 2.89 2.92 2.98 3.01 3.05 3.09 3.14			
0.5767 0.5713 0.5620 0.5542 0.5421 0.5293 0.5203 0.5110 0.5013	2.52 2.56 2.67 2.76 2.90 3.08 3.19 3.26 3.34		· · · · · ·	1.17 (v 1.48 (v 1.42 (v 1.33 (v 1.38 (v 1.25 (v 1.43 (v 1.43 (v
0.5820 0.5601 0.5501 0.5483 0.5483 0.5358 0.4938	2.50 2.59 2.61 2.67 2.67 2.76 3.15			1.14 (h 1.20 (h 1.32 (h 1.20 (h 1.33 (h 1.33 (h 1.44 (h
	2.66 2.74 2.84 2.96 3.02 3.13 3.17 3.23 3.23 3.23 3.29 3.33 3.35	<pre>(h) (h) (h) (h) (h) (h) (h) (h) (h) (h)</pre>	0.001688 0.001681 0.001429 0.001255 0.001213 0.000954 0.000881 0.000878 0.000814 0.000805 0.000696 0.000734	1.09 1.06 1.08 1.06 1.06 1.05 1.13 1.11 1.16 1.13 1.14

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SAMPLE: q (cont.)

3.47	(v)	0.001498
3.65	(v)	0.001318
3.76	(v)	0.001188
3.85	(v)	0.001070
3.95	(v)	0.000955
4.01	(v)	0.000873
4.09	(v)	0.000805
4.14	(v)	0.000775

SAMPLE:	r
---------	---

0.6982 0.6964 0.6917 0.6890 0.6862 0.6834 0.6806 0.6777 0.6748 0.6718 0.6656 0.6656 0.66592 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559 0.6559	1.98 1.99 2.00 2.01 2.03 2.05 2.06 2.08 2.10 2.11 2.13 2.15 2.16 2.18 2.20 2.22 2.23 2.25 2.27 2.29 2.32 2.38 2.38	
0.6968 0.6929 0.6876 0.6821 0.6763 0.6704 0.6611 0.6512 0.6407 0.6296 0.6218	1.83 1.84 1.85 1.91 1.97 2.00 2.08 2.13 2.20 2.26 2.30	
0.7018 0.6665 0.6567 0.6464 0.6354 0.6277	2.03 2.23 2.29 2.37 2.45 2.51	

1.09 1.27 1.36 1.29 1.16 1.31 1.29 1.21 1.21 1.21 1.20 1.23	 (V) (V)
0.91	(h)
1.05	(h)
1.05	(h)
1.25	(h)
1.22	(h)
1.32	(h)

SAMPLE: r (cont.)

2.00	(h)	0.0371	0.94
2.04	(h)	0.0286	0.94
2.13	(h)	0.0261	0.96
2.21	(h)	0.0215	0.99
2.25	(h)	0.0168	1.03
2.32	(h)	0.0136	1.02
2.35	(h)	0.0115	1.01
2.20	(h)	0.0103	1 03
2.30	(h)	0 0118	1 02
2.55	(11)	0.0110	1.02
2.56	(v)	0.0438	
2.61	(v)	0.0412	
2.67	(v)	0.0392	
2.75	(v)	0.0207	
2.84	(v)	0.0180	
2.90	(v)	0.0165	
2.96	(v)	0.0148	
2.99	(v)	0.0148	. *
	2.00 2.04 2.13 2.21 2.25 2.32 2.35 2.39 2.39 2.39 2.56 2.61 2.67 2.75 2.84 2.90 2.96 2.99	2.00 (h) 2.04 (h) 2.13 (h) 2.21 (h) 2.25 (h) 2.32 (h) 2.35 (h) 2.39 (h) 2.39 (h) 2.39 (h) 2.61 (v) 2.61 (v) 2.67 (v) 2.67 (v) 2.75 (v) 2.84 (v) 2.90 (v) 2.96 (v) 2.99 (v)	2.00 (h) 0.0371 2.04 (h) 0.0286 2.13 (h) 0.0261 2.21 (h) 0.0215 2.25 (h) 0.0168 2.32 (h) 0.0136 2.35 (h) 0.0115 2.39 (h) 0.0103 2.39 (h) 0.0103 2.39 (h) 0.0118 2.56 (v) 0.0438 2.61 (v) 0.0412 2.67 (v) 0.0207 2.84 (v) 0.0180 2.90 (v) 0.0148 2.96 (v) 0.0148

3. Deep Sea Clay Samples : Surficial Suite

P = applied vertical pressure (kPa) e = void ratio n = fractional porosity Vpv = compressional wave velocity in vertical direction Vph = compressional wave velocity in horizontal direction Vs = shear wave velocity in vertical direction (all velocities are in m/s) FF =vertical electrical formation factor Ø = permeability derived from consolidation results (× 10⁻¹¹ m/s) D = equivalent depth calculated from applied load (m)

Void ratio values are quoted for t100 times; i.e. for 100% primary consolidation. Exceptions are initial void ratios (e₀) and bracketed values (the latter correspond to the last reading prior to a change in applied load; t1440). Tabulated porosity values correspond to these conditions.

	P	е	n	v _{pv}	v_{ph}	Vs	FF	Ø	D
	SAMPLE	2 1							
	Specif	ic Gravi	ty = 2.6	8		e _o = 4	5.19		
•• .	1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 12.5	5.19 4.89 4.75 4.58 4.4.37 4.13 3.82 3.48 (3.37) (3.55)	0.8385 0.8302 0.8261 0.8208 0.8138 0.8051 0.7925 0.7768 0.7712 0.7802	1492 1493 1494 1496 1501 1510 1527 1558		31 60 82 99 106 134 160 232 230 140	1.77 1.87 1.91 1.94 1.98 2.04 2.20 2.41	2844.0 1207.0 831.9 471.6 342.2 167.6 89.0	0.56 4.3 8.3 16 31 59 111 206
	SAMPLE	2							
	Specif	ic Gravi	ty = 2.6	7		e ₀ = 4	4.60		
	1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 12.5	4.58 4.32 4.07 3.65 3.22 2.84 2.49 2.14 (2.09) (2.26)	0.8208 0.8120 0.8028 0.7849 0.763 0.7396 0.7135 0.6815 0.6815 0.6764 0.6933			32 45 67 122 153 183 245 263 146	1.64 1.65 1.70 1.78 1.89 1.96 1.97 2.15 1.97	4832.0 2737.0 2368.0 673.0 946.0 360.0 61.7	0.5 3.9 7.4 13 25 25 81 145
	SAMPLE	2 3	anna Martin Station Anna Anna Anna Anna Anna Anna Anna Ann				n Na san Na san		
	Specif	ic Gravi	ty = 2.6	6		e_ = !	5.34		
	1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 12.5	5.32 4.96 4.55 4.04 3.50 3.00 2.56 2.13 (2.07) (2.32)	0.8418 0.8322 0.8198 0.8016 0.7778 0.7500 0.7191 0.6805 0.6743 0.6988	1488 1490 1491 1494 1506 1518 1539 1535 1510		19 30 43 62 91 119 156 184 197 121	1.50 1.56 1.60 1.64 1.77 1.89 2.05 2.35 2.42 2.25	4542.0 2146.0 1176.0 633.1 271.2 120.0 49.9	0.57 4.4 8.1 15 26 46 83 145
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SAMPLE	4							
Specif	ic Gravi	1 = 2.6	6		e _o = 3	3.21		
1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 12.5	3.20 3.14 3.01 2.73 2.38 2.04 1.70 1.39 (1.33) (1.50)	0.7619 0.7585 0.7506 0.7319 0.7041 0.6711 0.6296 0.5816 0.5708 0.6000	1501 1501 1503 1516 1518 1534 1555 1582	1505 1507 1512 1520 1542 1552 1589 1642 1565	35 44 56 74 104 140 179 229 239 175	1.87 1.88 1.91 1.97 2.09 2.23 2.44 2.73	1330.0 973.8 731.8 353.8 161.4 91.2 35.0	0.4 3.1 6 11 20 35 63 110
SAMPLE Specif	5 ic Gravi	.ty = 2.6	6		e _o = 2	2.10		
1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 199.2 99.6 49.8 24.9 12.5	2.10 2.06 2.03 1.95 1.83 1.69 1.53 1.35 (1.32) (1.34) (1.36) (1.38) (1.41) (1.43) (1.48)	0.6774 0.6732 0.6700 0.6610 0.6466 0.6283 0.6047 0.5745 0.5690 0.5726 0.5763 0.5798 0.5851 0.5885 0.5968	1486 1490 1494 1520 1530 1547 1576 1551 1530 1528 1528 1528 1544 1511	1507 1508 1511 1514 1524 1533 1554 1577 1604 1590 1583 1573 1563 1557	50 63 68 85 108 143 173 213 216 194 158 135 115 86	2.30 2.32 2.34 2.37 2.44 2.53 2.66 2.84 3.02 2.98 2.95 2.90 2.86 2.84 2.73	819.3 1207.0 342.5 219.7 149.6 127.0 55.5	0.28 2.2 4.4 9 16 31 59 109
P	e	n	v _{pv}	v_{ph}	Vs	FF	Ø	D
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SAMPLE	6							
Specif	ic Gravi	.ty = 2.6	5		e _o = !	5.43		
1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 199.2 99.6 49.8 24.9 12.5 1.6	5.12 4.19 3.80 3.46 3.12 2.77 2.42 2.08 (2.00) (2.02) (2.04) (2.09) (2.14) (2.19) (2.31)	0.8366 0.8073 0.7917 0.7758 0.7573 0.7348 0.7076 0.6753 0.6667 0.6689 0.6711 0.6764 0.6815 0.6865 0.6979	1476 1478 1484 1493 1505 1525 1553 1568	1483 1489 1489 1507 1529 1554 1590 1554 1560 1551 1541 1538 1519	10 49 60 79 110 139 170 179 168 146 128 111 116 92	1.66 1.80 1.94 2.02 2.15 2.30 2.50	2411.0 639.7 460.1 262.5 130.3 64.3 29.0	0.6 3.8 7.0 14 26 44 80 143
SAMPLE	7							
Specif	ic Gravi	ty = 2.7	0		e _o = 3	3.44		
1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 1.6	3.41 3.36 3.26 3.01 2.77 2.52 2.22 2.06 (1.90) (2.03)	0.7732 0.7706 0.7653 0.7506 0.7347 0.7159 0.6894 0.6732 0.6552 0.6700	1492 1495 1499 1518 1528 1544	· · · · · · · · · · · · · · · · · · ·	31 40 57 80 101 125 160	1.86 1.88 1.90 1.97 2.03 2.13 2.25	1761.0 1990.0 892.5 573.7 347.4 167.0 52.9	0.4 3.2 6.1 12 22 40 7.3 133

P	е	n	v _{pv}	v_{ph}	Vs	FF	Ø	D
SAMPLE	8					-		
Specif	ic Gravi	ty = 2.6	8		e _o =	2.22		
1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 199.2 99.6 49.8 24.9 12.5 1.6	2.21 2.15 2.07 1.96 1.80 1.61 1.41 1.22 (1.20) (1.22) (1.24) (1.26) (1.27) (1.29) (1.35)	0.6885 0.6743 0.6622 0.6429 0.6169 0.5851 0.5495 0.5495 0.5536 0.5575 0.5595 0.5633 0.5745	1488 1495 1497 1506 1516 1539 1564 1596 1614 1590 1581 1564 1581 1567 1560	1503 1505 1514 1525 1535 1552 1587 1604 1654 1643 1624 1616 1612 1602 1591	47 67 76 103 131 155 189 234 243 230 197 168 154 135 96	2.10 2.14 2.19 2.27 2.37 2.51 2.75 3.01 3.03 2.99 2.90 2.89 2.83 2.75	233.7 322.5 251.5 179.7 81.1 55.7	0.29 2.3 4.5 8.6 16 30 56
SAMPLE	9							
Specif	ic Gravi	.ty = 2.6	9		e _o =	2.54		
 1.6 12.5 24.9 49.8 99.6 199.2 398.4 796.7 398.4 12.5 1.6	2.50 2.48 2.37 2.22 2.02 1.81 1.59 1.38 (1.32) (1.49) (1.55)	0.7143 0.7126 0.7033 0.6894 0.6689 0.6441 0.6139 0.5798 0.5690 0.5984 0.6078	1484 1485 1485 1487 1496 1503 1528 1564 1547 1518 1516	1485 1493 1493 1493 1512 1525	37 46 65 87 112 144 185 235 245 120 87	2.02 2.20 2.27 2.33 2.43 2.74 3.03	413.8 162.6 161.0 113.7 79.7 29.4 26.7	0.32 2.54 4.89 9.3 17 32 60 108

Thermal Conductivity Data Summary : Surficial Clay Samples

Sample No.	P	n	W	k	FF
1	Pre	0.8337	0.6581	0.83	1.77
1	Post	0.7702	0.5626	0.96	2.40
2	Pre	0.8169	0.6314	0.75	1.64
2	49.8	0.7798	0.5762	0.82	1.74
2	Post	0.6859	0.4560	0.95	1.97
3	Pre	0.8389	0.6666	0.70	1.50
4	Pre	0.7571	0.5448	0.86	1.87
5	Pre	0.6706	0.4387	0.87	2.30
5	Post	0.5888	0.3547	0.97	2.73
6	Pre	0.8435	0.6717	0.69	1.66
6	Post	0.6956	0.4645	0.82	2.37
7	Pre	0.7689	0.5581	0.81	1.86
7	Post	0.6615	0.4260	0.87	2.38
8	Pre	0.6839	0.4560	0.94	2.10
8	99.6	0.6300	0.3901	0.84	2.39
8	Post	0.5675	0.3511	1.00	2.75
9	Pre	0.7116	0.4837	1.00	2.21
9	796.7	0.5554	0.3217	1.21	3.03
9	Post	0.6000	0.3629	0.95	2.65
CS CS	Pre Post	0.7615 0.6336	0.5488	0.83	

= Applied vertical pressure (kPa). 'Pre', and 'Post' refer to Ρ before and after the oedometer test.

- = porosity. n
- water content. W =

= thermal conductivity (W/mK): needle inserted horizontally. = vertical electrical formation factor. k

FF

CS commissioning sample.

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4. Deep Sea Drilling Project Samples

1.41

Void ratio values are quoted for the end of each loading increment (e_{1440}) . Where it is possible to extract times for 100% primary consolidation e_{100} (and porosity n_{100}) values are also given.

Ρ

^e1440 ⁿ100 e100

vp

Vs

Ø

SAN	MPLE DSDP.	1					
Spe	ecific Gra	avity =	2.67	e _o	= 5.06		
1.5 56 112 224 450 900 1800 900	5 5 4.92 4.68 4.15 3.28 2.28 1.49 3	5.06 5.01 4.85 4.54 3.89 2.99 2.07 1.39 1.43 1.78	0.8350 0.8311 0.8239 0.8058 0.7664 0.6951 0.5984	1489 1497 1486 1489 1552 1565 1612 1652	98 123 138 149 65 180 215 231 208 156	1349.0 1079.0 672.0 295.0 57.0 7.0 0.71	0.5 10 20 37 65 107 165 256 130 4.6
SAI Spe	MPLE DSDP ecific Gr	2 avity =	2.67	eo	= 2.98		
1.: 28 56 84 11: 14: 16: 28: 450 900 180 360 450 28	5 2.95 2.88 2.81 2.73 2.67 9 2.59 1 2.36 0 2.05 0 1.61 00 1.20 00 0.88	2.98 2.93 2.79 2.72 2.65 2.57 2.30 2.01 1.55 1.16 0.84 1.09 1.52	0.7468 0.7423 0.7375 0.7319 0.7275 0.7214 0.7024 0.6721 0.6169 0.5455 0.4681	1529 1542 1557 1562 1555 1569 1558 1556 1585 1559 1685 1809 1772 1651	73 74 82 88 94 96 108 120 154 181 255 288 274 171	40.0 22.0 28.0 16.0 16.0 5.0 12.0 5.0 3.0 0.9 0.3	0.4 7 13 19 25 31 36 55 81 137 232 395 56 4.2

D

P

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ⁿ100

.

vp

vs

D

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Ø

SAMPL	E USDP.	2					
Speci	fic Gra	avity =	2.70	eo	= 1.95		
1.5		1.95		1535	91		0.3
28		1.90		1549	113		4.8
56		1.85		1554	128		9.0
84		1.82			161		14
112		1.80		1561	175		18
140		1.77		1574	140		23
168		1.75		1583	168		27
252	1.67	1.65	0.6255	1584	230	134.0	39
450	1.58	1.56	0.6124	1603	276	98.0	67
900	1.43	1.40	0.5885	1646	341	70.2	126
1800	1.26	1.22	0.5575	1698	435	2.7	234
900	· *.	1.23		1706	438	. A	5. 1 t
450		1.24		1666	384		
252		1.25		1654	349		н. 1. т. н.
168		1.26		1644	312		
140		1.27		1630	300		
112		1.27		1632	287		
84		1.27		1618	267		
56		1.28		1615	258		
28		1.29		1598	205		
1.5		1.34		1579	169		

P

e₁₀₀ e₁₄₄₀ n₁₀₀ V_p

vs

Ø

D

SAMPLE DSDP4								
Specific Gravity = 2.70 $e_0 = 1.39$								
1.5 28 56 112 224 336 448 560 672 900 1350 1800 1350 900 672 560 448 336 224 112 56 28	1.26 1.16 1.12 1.10 0.97	1.37 1.35 1.33 1.30 1.26 1.24 1.22 1.21 1.19 1.15 1.11 1.09 0.96 0.97 0.97 0.97 0.98 0.98 0.98 0.98 0.98 0.98 0.98 0.99 1.00 1.00	0.5575 0.5370 0.5283 0.5238 0.4924	1609 1608 1619 1642 1670 1684 1701 1722 1731 1735 1822 1925 1925 1906 1866 1849 1831 1813 1813 1776 1744 1731 1718	105 154 176 233 246 260 328 381 386 436 471 565 595 580 550 550 550 550 550 550 550 550 55	16.3 4.3 3.1 9.1 1.8	0.2 3.9 7.6 15 30 44 58 72 86 113 167 220 413	
1.5		1.02		1698	182			