33. ELECTRICAL RESISTIVITY, SOUND VELOCITY, THERMAL CONDUCTIVITY, DENSITY-POROSITY, AND TEMPERATURE, OBTAINED BY LABORATORY TECHNIQUES AND WELL LOGS: SITE 462 IN THE NAURU BASIN OF THE PACIFIC OCEAN¹

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ABSTRACT

At Deep Sea Drilling Project (DSDP) Site 462, from mudline to 447 meters below the sea floor, Cenozoic nannofossil oozes and chalks have acoustic anisotropies such that horizontal sonic velocities are 0 to 2.5% faster than those in the vertical direction. In laminated chalk, anisotropy of 5% is typical, and in limestones, radiolarian oozes, porcellanites, and cherts, the anisotropies range from 4 to 13%. Middle Maestrichtian volcaniclastics from 447 meters to 560 meters below the sea floor have an acoustic anisotropy of 2 to 32%; 4 to 13% is typical. Basalt flows and sills occur between 560 meters and 1068 meters, and have no apparent anisotropy, but minor interbedded volcaniclastics have anisotropies from 0 to 20% (5% is typical). These volcaniclastics frequently have very small anisotropies, however, compared with the volcaniclastic sequence above the basalt section.

Data in cross-plots of the laboratory-measured compressional sound velocity versus wet-bulk density, wet-water content, and porosity of sediments and sedimentary rock typically lie between the equations derived by Wyllie et al. (1956) and Wood (1941); the Wyllie et al. (1956) equation has a fair fit with similar basalt velocity cross-plots. Cross-plots of thermal conductivity and sound velocity indicate only a fair correlation. Cross-plots of thermal conductivity versus porosity, wet-water content, and wet-bulk density correlate well with equations derived by Maxwell (1904), Ratcliff (1960), Parasnis (1960), and Bullard and Day (1961). Electrical formation factor versus porosity for sediments and sedimentary rock agrees with the Archie (1942) equation, with *m* values of 2.6; for basalt, an *m* of about 2.1 is typical. Basalt pore-water resistivities do not appear to be greatly different from sea water. Formation factors are greater than those derived from equations in Maxwell (1904), Winsauer et al. (1952), Boyce (1968), and Kermabon et al. (1969). An "apparent interstitial water resistivity" (R_{wa}) curve was derived from the density and induction logs. This R_{wa} curve indicated an anomaly, at 393.5 to 396.5 meters, which could be interpreted as (1) 76% hydrocarbons, (2) relatively fresh pore water (1.8%₀ salinity), or (3) low-grain-density (2.2 g/cm³) semi-lithified porcellanite-chert. Porcellanite-chert is the most plausible interpretation.

In situ temperatures measured by the Uyeda temperature probe were about 2 to 5°C (50%) higher than the equilibrium temperature (Lachenbruch and Brewer, 1959) extrapolated from two Gearhart-Owen continuous temperature logs; this discrepancy probably arises because the hole was washed out in this depth interval, so these extrapolated temperatures are probably not reliable. If one ignores all precautions as to temperature artifacts, then the equilibrium temperatures of the Gearhart-Owen temperature logs suggest that hydrothermal circulation is occurring in at least the upper 40 meters of the basalt section and heat is transferred by convection and not conduction. Hydrothermal circulation is probably *not* indicated, however, and the temperature anomalies probably result from excessive artificial cooling of the fractured basalt zones by circulation of water during drilling.

INTRODUCTION

Purpose

This paper studies certain physical property relationships, using samples and well logs obtained at DSDP drill Site 462 in the Nauru Basin of the Pacific Ocean (Fig. 1). Its principal aims may be summarized as follows:

1) To introduce additional systematic studies of compressional sound velocity and acoustic anisotropy for basalt and pelagic and volcaniclastic sediments and rocks, and their relationships to wet-bulk density, wetwater content, and porosity. These relationships are important in the proper interpretation of gravity, seismic reflection, seismic refraction, and sonobuoy data.

2) To provide additional systematic studies of thermal conductivity and its relationship to wet-bulk density, wet-water content, porosity, and sound velocity. These studies are important in proper interpretation of heat flow studies, hydrothermal circulation, and prediction of temperature distribution within the crust.

3) To study the temperature distribution in the crust, which is important in understanding hydrothermal circulation, deposition of vein material, or potential ore deposits. (Heat flow calculations are in the Site Summary, this volume.)

4) To study electrical formation factor and porosity relationships (empirically derived) for sediments, sedimentary rock, and basalt, which are essential in the proper interpretation of electric logs.

5) To study the density and electric logs, which are important because if the porosities derived from the density and electric logs do not match, within certain limits of error, assuming the logging data are accurate, one or more of the following is indicated: (a) conductive metallic minerals, (b) anomalies in the salinities of interstitial water, (c) an anomalous temperature, (d) a large amount of minerals with very high or low grain density, or (e) the presence of hydrocarbons.

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Figure 1. Location of DSDP drill Site 462 in the Pacific Ocean. Temperature at time of measurement with logging device (T_m) versus ln [t/(t - s)], where t = time since the bit reached the depth in question and s = duration of drilling or circulation after bit reached that depth.

DATA, DEFINITIONS, METHODS

The sediment and basalt classification is discussed in the Explanatory Notes (this volume). Wet-bulk density is the ratio of weight of the wet-saturated sediment or rock sample to its volume, expressed in g/cm^3 . Wet-water content is the ratio of the weight of sea water in the sample to the weight of the wet saturated sample, and is expressed as a percentage. Porosity is the ratio of the pore volume in a sample to the volume of the wet saturated sample, and is expressed as a percentage in some cases and as a fraction in others. All of these equations, derivations, and techniques are discussed in detail in Appendix I (Boyce) of this volume.

The following technique was used for sedimentary samples: Generally, in the *Glomar Challenger* laboratories, an undisturbed (visibly undistorted bedding), wet-saturated sample, about 2.5 cm thick and 5 cm long, was cut and removed from a split core liner after

waiting about 2 hours after the core was on deck to allow it to approach room temperature. Then the thermal conductivity was measured using the Quick Thermal Conductivity Meter (QTM) described in Appendix I of this volume. The sample was then carefully cut, if necessary, with a diamond saw, and smoothed with a sharp knife or file to a D-shaped sample 2.5 cm thick and with a 2.5-cm radius. Compressional sound velocities $(\pm 2\%)$ perpendicular and parallel to bedding were measured with the Hamilton Frame velocimeter (Boyce, 1976a, and Appendix I, this volume). Immediately afterward, wetbulk density was measured within ± 2 or 3% using special two-minute gamma-ray counts with the Gamma-Ray Attenuation Porosity Evaluator (GRAPE) (Evans, 1965) as modified by Boyce (1976a, and Appendix I, this volume). Between various measurements, the sample was then wrapped in plastic and stored in a sealed plastic box with a wet sponge so that it would not dry out. The wet-water content, wetbulk density, and porosity of a sub-sample were then determined by weighing the water-saturated sample in water and after drying for 24 hours at 110°C. The weight of evaporated water was corrected for salt content (35%) to give the weight of sea water (Boyce, 1976a; Boyce, Appendix I, this volume; Hamilton, 1971b). The estimated precision of wet-bulk density is ± 0.01 g/cm³ (absolute), and the precision of wet-water content and porosity is $\pm 0.5\%$ absolute units. The acoustic impedance, in units of $(g \cdot 10^5)/(cm^2 \cdot s)$, is obtained from the product of the vertical (if possible) velocity and the gravimetric (if possible) wet-bulk density. The laboratory results are reported in tables in the Site Summary (this volume).

Basalt velocity samples were processed in a slightly different manner from the sediments. Some velocities were measured when the basalt first arrived in the laboratory; this allowed us to be certain that the sample was saturated with water. Others were measured on minicores (small cylinders drilled out of the rock). The velocities and the corresponding technique for each velocity are reported in tables in the Site Summary (this volume). Detailed methods are discussed in Appendix I (this volume). There was an attempt to keep these samples water-saturated. Questionable (i.e., all) mini-core samples from Hole 462 have been deleted from major cross-plots.

In general, all heat conductivity measurements on basalt were performed in the same manner as that described for sedimentary material. All basalt GRAPE 2-minute wet-bulk densities, and gravimetric wet-bulk densities, wet-water contents, and porosities were determined on mini-cores, using techniques identical to those employed for the sedimentary material.

In situ wet-bulk density and electrical resistivity were obtained from Gearhart-Owen well log combinations: (1) compensated density, caliper, and gamma-ray logs; and (2) induction, 16-inch normal, and gamma-ray logs. These are presented in the Site Summary (this volume), and the tools and precautions regarding the data are discussed in Appendix I (this volume).

With respect to the accuracy of logging data, I do not have absolute techniques available—e.g., *in situ* standards or *in situ* beds with precisely known *in situ* physical property values—to check empirically the validity of the logging data. Therefore, I have cross-plotted these data to see if they appear to have reasonable interrelationships. Crossplots of density and electrical resistivity, where hole conditions were good, are within preconceived limits, as would be expected on the basis of lithology of the cores.

The laboratory (two-minute GRAPE) density data were superimposed on the well log density plots, and appear basically to agree (within ± 15 m vertical depth shifts, without porosity adjustments to *in situ* overburden pressures) with the compensated density log data, where hole conditions were good. Therefore, the density log appears to be valid. There are thin beds, however, such as chert beds, where high or low density values are not represented on the density log. These beds are probably thinner than the logging tool's resolution.

Because we do not have any absolute method to evaluate the logging data, therefore, any log-derived relationships between electrical resistivity and density-porosity are subject to bias if the logging tools are not working properly.

In situ temperature measurements were obtained using two Gearhart-Owen continuous temperature logs. Spot temperature measurements were obtained using the Uyeda temperature probe; the resulting data are discussed in the Site Summary (this volume), and the equipment and technique employed are discussed in Appendix I (this volume).

Electrical Resistivity

The electrical resistivity of any material is defined as the resistance, in ohms, between opposite faces of a unit cube of that material. If the resistance of a conducting cube with length L and cross-sectional area A is r, then the resistivity R_0 is

$$R_0 = rA/L = \text{ohm-m} \tag{1}$$

Electrical conduction through saturated sediment is complicated by a framework that generally consists of nonconducting mineral grains. If the sediment consists of nonconducting minerals, electrical conduction is primarily through the interstitial water, whose conductivity varies with temperature, and salinity (Horne, 1965; Horne and Courant, 1964; Horne and Frysinger, 1963; Thomas et al., 1934). Conduction through the fluid can be modified significantly, however, if there are present metallic minerals with appreciable conductivity or claytype minerals that exchange or withdraw ions from the interstitial water (de Witte, 1950a, b; Patnode and Wyllie, 1950; Keller, 1951; Berg, 1952; Winsauer and McCardell, 1953; Wyllie, 1955). Charged colloidal particles and exchanged ions are not necessarily removed from the sediment when the interstital water is sampled, so they do not contribute to what is normally thought of as the water salinity (Keller, 1951; Howell, 1953).

The formation factor, F, is the ratio of the electrical resistivity of the saturated sediment, R_0 , to the resistivity of the interstitial water, R_w , at the same temperature and pressure (Archie, 1942):

$$F = R_0 / R_w \tag{2}$$

The formation factor has been related to porosity and fluid salinity of rocks or sediments by Archie (1942; 1947), Winsauer et al. (1952), and others (Appendix A).

If the mineral composition of the sediment forms a nonconductive matrix, and if the interstitial water conductivity is high, then this ratio is considered to be the "true" formation factor. With increasing salinity of the interstitial water, this "true" formation factor approaches a constant value for a given porosity and rock sample (Patnode and Wyllie, 1950; Keller and Frischknecht, 1966).

If sediments contain minerals which are conductors, then this ratio is considered to be an "apparent" formation factor, and is less than the "true" formation factor of sediments for a given set of porosity, textural, and cementation characteristics. The "apparent" formation factor approaches a constant value with different salinities, at given porosity, only if the conductivity of the interstitial water is much greater than that of the conducting minerals (Berg, 1952; Howell, 1953; Wyllie and Southwick, 1954; Wyllie, 1955).

The variation of apparent formation factor with interstitial water resistivity may be related in part to the distribution of conducting grains in a sample. Wyllie and Southwick (1954) developed a model showing that the connected conducting grains are conductors in parallel with, and isolated conducting grains are conductors in series with, the interstitial fluid. If the interstitial fluid is a good conductor, all the conducting grains will contribute to the overall conduction. If the interstitial fluid is a moderate or poor conductor, the conducting grains in series with interstitial water will contribute a reduced proportion of the overall conduction of the rock-matrix; thus, the formation factor appears to increase as the resistivity of the fluid increases.

Clay-type minerals with varying exchange capacities may act as resistors or conductors relative to different interstitial water resistivities. Because of the clay-type minerals and other possible conducting minerals, the "formation factor" (for a given sample) may not be constant for different interstitial water resistivities (Keller, 1951; Wyllie, 1955; Berg, 1952; Wyllie and Gregory, 1953; Winsauer et al., 1952; Winsauer and McCardell, 1953; Wyllie and Southwick, 1954; Keller and Frischknecht, 1966).

The resistivity of interstitial water may be estimated by measuring the resistivity of the water squeezed from the geologic sample, or by taking it to be equal to the resistivity of sea water. However, interstitial-water sampling may not remove ions that are filtered or trapped by clay-type minerals (Scholl, 1963), and the natural sediment compaction from overburden pressure may trap or filter various ions as the fluid migrates; thus, the interstitial fluid may have a chemical composition different from that of the original interstitial sea water (Siever et al., 1961; Siever et al., 1965). The electrical resistivity of the interstitial water determined, for example, by using the data of Thomas et al. (1934), may therefore be in error, because Thomas et al. assumed a chemical composition identical to that of sea water.

Electrical resistivity through fresh sediment may be isotropic (Bedcher, 1965), but consolidated sediments and rock have anisotropic resistivities. Resistivity parallel to bedding is typically less than the resistivity perpendicular to bedding (Keller, 1966; Keller and Frischknecht, 1966).

Textures of the individual mineral grains affect the electrical resistivity. The more angular textures create a longer path length through the sediment and thus a higher resistivity and a higher formation factor for given porosity (Wyllie and Gregory, 1953). The resistivity is also affected by grain-size distribution, particularly for clay-type minerals. A finer grain size gives a greater surface area with ionic exchange capacity, and so increases the number of ionic cloud conductors in a given sample. This is also true, to a lesser degree, of non-clay-type minerals, such as quartz and feldspar (Keller and Frischknecht, 1966).

Study of the relationship between electrical formation factor and porosity is especially important in interpreting the electrical well logs in terms of porosity. The "formation factor-porosity" relationship is of particular interest because there are only a few published studies of modern unconsolidated marine sediments (Boyce, 1968, 1980; Kermabon et al., 1969; and Donnelly et al. 1980). For a summary of various types of lithology and their F-porosity relations, see Keller (1966) and Keller and Frischknecht (1966).

Maxwell's (1904) equation is a theoretical approach for spheres in suspension, which should provide an upper limit of porosity. Most sediments or rocks are not accurately defined, however, by this simple model. Sediments or rocks generally have irregularly shaped grains, so that conducting ions must travel a longer average path than would be required between spherical particles, and sediments or rocks will therefore have greater resistivities and formation factors than those derived by Maxwell's (1904) equation.

Among DSDP cruises, the following have resulted in published electrical resistivities of water-saturated basalt: Leg 26 (Hyndman and Ade-Hall, 1974); Leg 34 (Drury, 1976); Leg 37 (Hyndman and Drury, 1977; Katsube et al., 1977); Leg 46 (Christensen et al., 1979; Kirkpatrick, 1979); and Legs 51, 52, 53 (Donnelly et al., 1980; Hamano, 1980; Salisbury et al., 1980).

We will interpret the DSDP Hole 462 density and electrical logs by using a technique developed by the petroleum industry (Schlumberger, Ltd., 1972) called the "apparent electrical resistivity of the interstitial water" (R_{wa} curve). This technique will here involve calculating the porosity from the wet-bulk density log, assuming grain densities based on the lithology of the cores:

$$\phi = \frac{\varrho_g - \varrho_b}{\varrho_g - \varrho_w} \tag{3}$$

where

 ϕ = fractional porosity;

- $\varrho_g = \text{grain density, g/cm}^3;$
- ϱ_w = pore fluid density, g/cm³;

 ϱ_b = wet-bulk density, g/cm³, from the density log.

Then, by using a simplified form of Archie's (1942) equation for the Site 462 data:

$$F = R_a / R_{wa} = \phi^{-m} = \frac{1.0}{\phi^2}$$
(4)

$$R_{wa} = R_a \phi^2 \tag{5}$$

By substituting in this equation (5) the "apparent formation resistivity" (R_a) (not corrected for bore-hole diameter, bore-hole fluids, or the thicknesses of beds with contrasting resistivity) from the induction log and the ϕ derived from the density log, we can then solve for R_{wa} .

If the formation is homogeneous calcareous ooze with a uniform pore-water salinity and a uniform and normal temperature gradient, the " R_{wa} versus depth" plot will theoretically be a straight line, but R_{wa} will decrease slightly because of increasing temperature with increasing depth. The method is useful because R_{wa} will be anomalously high if there are any unexpected zones (which can sometimes be very distinct) of (1) hydrocarbons, (2) low grain density, (3) relatively fresh water in the pores, or (4) negative-temperature anomalies. The R_{wa} curve will give anomalously low values if there are any unexpected zones of (1) high grain density, (2) electrical conductors (metallic deposits), (3) relatively saltier pore waters, or (4) high-temperature anomalies. Since the composition of the pore fluids is known from samples of the sedimentary rocks collected on the *Challenger*, and we know the temperature of the formation, then we know what range of R_{wa} to expect, and should thus be able to identify the anomalies. If hydrocarbons are present, the pore-water saturation $S_w = (R_{wa} \text{ ex-}pected/R_{wa} \text{ anomalous})^{1/2}$ when using Equation 4.

Sound Velocity

Compressional sound velocity in isotropic material has been defined (Wood, 1941; Bullen, 1947; Birch, 1961; Hamilton, 1971a) as

$$V = \left(\frac{i+4\mu/3}{\varrho_b}\right)^{1/2} \tag{6}$$

where

V is the compressional velocity;

 ϱ_b is the wet-bulk density in g/cm³ and $\varrho_b = \varrho_w \phi$ + $(1 - \phi) \varrho_g$ (here ϕ is the fractional porosity of the sediment or rock and the subscripts *b*, *g*, and *w* represent the wet-bulk density, grain density, and water density, respectively);

i is the incompressibility or bulk modulus; and μ is the shear (rigidity) modulus.

Where samples are anisotropic, i and μ may have unique values for the corresponding vertical and horizontal directions. See Laughton (1957); Carlson and Christensen (1977); and Bachman (1979) for discussions of anisotropy.

Compressional velocity of sediments and rocks has been related to the sediment components by Wood (1941), Wyllie et al. (1956), and Nafe and Drake (1957), whose equations are listed in Appendix B. These will be discussed later. Velocity is related to mineralogical composition, fluid content, water saturation of pores, temperature, pressure, grain size, texture, cementation, direction with respect to bedding or foliation, and alteration, as summarized by Press (1966). Recently, Hamilton (1978) has summarized velocity-density relationships of sediment and rock of the sea floor. Christensen et al. (1975) have summarized velocity-density relationships of basalt under pressure.

Basalt velocities at zero pressure have been published for cores recovered on Leg 37 (Hyndman, 1977); Leg 46 (Matthews, 1979), and Legs 51, 52, 53 (Salisbury et al., 1980; Donnelly et al., 1980; Hamano, 1980). Sonic logs taken in sea-floor basalt have been reported by Kirkpatrick (1979) for Leg 46 and by Salisbury et al. (1980) for Legs 51, 52, 53.

Thermal Conductivity

Green (1962) defines thermal conductivity as follows: q is the quantity of heat which flows in j seconds through a body which has a cross-sectional area A and a length L. When the temperature difference between the ends of a path is T_d , and the material has a bulk thermal conductivity coefficient k_b , then

$$q = k_b \frac{A}{L} j T_d \tag{7}$$

where q is in calories $\times 10^{-3}$; A is in cm²; L is in cm; and T is in °C.

Therefore, the units for Equation 7 are

calorie ×
$$10^{-3} = k_b \frac{\text{cm}^2}{\text{cm}} \circ \text{C} \cdot \text{s}$$

Solving for the bulk thermal conductivity coefficient,

$$k_b = \frac{\text{calorie} \times 10^{-3}}{\text{cm} \cdot {}^{\circ}\text{C} \cdot \text{s}}$$
(8)

Thermal conductivity is the quantity of heat per second that will flow between opposite faces of a cube, when there is a 1°C temperature difference between opposite faces of the cube. For a 1-cm cube, heat conductivity is a coefficient in units of (calorie $\times 10^{-3}$)/(cm °C s).

Theory of thermal conduction in earth material is discussed in detail and in theoretical terms by Jaeger (1965), Beck (1965), and Langseth (1965). Many thermal conductivity measurements and parameters affecting thermal conductivity are summarized by Clark (1966). Bulk thermal conductivity varies with temperature, pressure, different minerals with correspondingly different thermal conductivities, direction of heat flow along bedding planes (anisotropy), mineral shape and texture, porosity, and pore shape, pore fluids, and degree of pore saturation. Because of the different conductivity values of many minerals and pore fluids, the wet-saturated bulk sediment-rock heat conductivity can be analogous to some of the electrical conductivity models discussed earlier, especially the model by Wyllie and Southwick (1954).

Many scientists have attempted to correlate thermal conductivity with porosity, assuming that mineral contents and their corresponding grain conductivities remain relatively constant compared with the thermal conductivity of the pore fluid. These comparisons assume a simple mineralogy or a single mineral such as calcite. Thermal conductivity has been correlated with porosity-like parameters by Maxwell (1904), Parasnis (1960), Ratcliffe (1960), Bullard and Day (1961), and others. Some of these authors' equations are listed in Appendix C. (None is given for Ratcliff (1960), who made empirical correlations.) These investigators use end-member systems where the bulk thermal conductivity can be calculated from (1) a given porosity parameter, (2) the grain conductivity and the fluid conductivity, and (3) shape of grains. The Maxwell (1904) equation was obtained from the work of Bullard et al. (1956), whose empirical data agree with the Maxwell equations. Bullard (1963) and Nafe and Drake (1963) have summarized some of these various correlation equations, but they use different end-member values for mineral and pore-water thermal conductivities, so comparison is difficult.

There have been many thermal conductivity measurements on DSDP sediments, but few with associated porosity parameters. Recently, thermal conductivity of sea-floor basalt has been measured on Leg 37 (Hyndman et al., 1977), Leg 46 (Erickson and Hyndman, 1979), and Legs 51, 52, 53 (Hamano, 1980).

THERMAL CONDUCTIVITY METHOD

In this study, thermal conductivity was measured using a Quick Thermal Conductivity Meter (QTM). This device employs a rectangular pad with a heater and thermocouple, which is placed on a flat rock sample, and the thermal conductivity is "automatically" measured, calculated, and displayed on a panel. The device has known standards whose heat conductivities are measured with those of the samples. Deviation from these standards is slight, and correction factors (Appendix I, this volume) were applied to all thermal conductivities after they were adjusted to 21 °C. The QTM's principle is the same as that described by Maxwell and Von Herzen (1969), except that the heater and thermocouple are between a pad and the rock (on the flat surface of a split core); in the Maxwell and Von Herzen (1969) technique, a needle is centered in a complete round sediment core. Because the QTM's needle is placed on the sample surface, the formula used by Maxwell and Von Herzen (1969) is divided in half and a new empirical correction factor is inserted to calibrate the data for the probe, etc.

To be certain of obtaining good data, measurements were obtained on as many standards as possible and a correction factor taking into account the results of these measurements was applied to the sample results. One of the reasons for doing this was that the only needle pad for wet samples was damaged, necessitating use of a non-waterproof needle-pad and insertion of a thin film of plastic between the needle pad and the flat rock surface. See Appendix I (this volume) for details on equipment standards and assumptions.

The thermal conductivity standards are accurate only within $\pm 5\%$, and the precision of the QTM data is $\pm 10\%$. All raw thermal conductivity data are presented in Table 4 of Appendix I (this volume).

Temperature

When drilling a hole, if the temperature of the circulating water is greater or less than the *in situ* temperatures of the formation, then the formation will be either heated or cooled, respectively (Jaeger, 1965; Raymond, 1969). The heating or cooling can affect the surrounding formation as far as about 3 meters away from the borehole. Therefore, when a temperature measuring device is placed in the hole fluid, the temperature measured is not the true undisturbed *in situ* formation temperature, and one must use procedures such as those given by Timko and Fertl (1972) or Keller et al. (1979) to obtain an estimate of equilibrium temperature (temperature before drilling).

I will present here a study of the two Gearhart-Owen continuous temperature logs obtained from Hole 462 and attempt to extrapolate these data to theoretical equilibrium temperatures. I will employ the graphic techniques of Keller et al. (1979), who use an equation derived by Lachenbruch and Brewer (1959); this equation defines borehole fluid temperatures in terms of duration of "drilling-circulation" at a given depth and the time interval between the last borehole circulation and the time of the temperature measurement on the log at that depth. Jaeger (1965) discusses the Lachenbruch and Brewer (1959) equation, and Jaeger (1965) and Raymond (1969) discuss formation temperature disturbances caused by borehole drilling activities, so I will not discuss these matters in detail here.

The following is Lachenbruch's and Brewer's (1959) equation, which attempts to quantify temperature equilibrium in the borehole:

$$T_m - T_0 = \frac{q}{4\pi k_b} \ln\left(\frac{t}{t-s}\right) \tag{9}$$

where

- T_m = temperature at the time of measurement with the logging device;
- T_0 = equilibrium temperature or temperature before drilling the hole;
- t = time since the drill bit reached the depth in question;
- s = duration of drilling or circulation after the drill bit reached the depth in question;
- q = heat source;
- k_b = thermal conductivity.

The $(q/4\pi k_b)$ portion of the equation can be the slope in the diagram (Fig. 2) if we plot T_m versus $\ln [t/(t - s)]$, as in Keller (1979). For example, (1) plot the T_m of the first temperature logging run at a given depth versus its corresponding $\ln [t/(t - s)]$; (2) plot the T_m of the second temperature log at the same given depth versus its corresponding $\ln [t/(t - s)]$; then (3) draw a straight line through these two points on Figure 2 and extrapolate this line until it intersects the zero axis of $\ln [t/(t - s)]$. The temperature where the zero axis of $\ln [t/(t - s)]$ and the extrapolated line intersect is the equilibrium temperature.

These temperature-extrapolation techniques will be reliable only where the diameter of the bore is not extremely variable or washed out, and where no permeable fractures occur in the formation. Fractured formations allow the drill-circulation water to either heat or cool the formation much more than at intervals where the formation is not fractured. Therefore, if temperature anomalies occur in the basalt portions of the hole which are fractured, then these anomalies are always suspect of being artifacts of the drilling operation. In addition, the theoretical aspects of the Lachenbuch and Brewer (1959) equation may not be directly applicable to the bottom of the hole (Langseth, personal communication). Also, the hole may not have had enough time to come to equilibrium, so that the equation may not properly apply.

Keller et al. (1979) used the foregoing technique on data from the borehole on Kilauea Volcano, where they identified zones of intense hydrothermal circulation.



Figure 2. Equilibrium temperature can be estimated from the graphic solution. See text for explanation and discussion.

Oceanic basaltic crust drilled on Leg 37 at Sites 332, 334, 335, and 337 (Hyndman et al., 1977) and on Leg 46 at Site 396 (Erickson and Hyndman, 1979) also reportedly contain zones in the basalt where convection or hydrothermal circulation appears to be operating. This type of transfer is important to a better understanding of heat flow, genesis of veins, and metallic mineral deposition.

RESULTS

Scatter Diagrams

The scatter diagrams which follow are presented in order to provide empirical relationships, for comparison with previous or future studies, and to help develop predictive relationships.

The first scatter diagram (Fig. 3) shows the gravimetrically determined wet-bulk density versus gravimetrically determined porosity. The data are for Cenozoic nannofossil ooze, chalk, limestone, radiolarian ooze, porcellanite, and chert, from 0 to 447 meters below the sea floor, and for Cretaceous volcaniclastics, basalt flows, and (some doleritic) sills, from 447 to 1068 meters below the sea floor at Site 462. On this diagram, the grain density of each sample may be estimated by a line from "1.025 g/cm³ (for 35% salinity) density at 100% porosity" through "the given datum point" to



Figure 3. Gravimetrically determined wet-bulk density versus gravimetrically determined porosity, Site 462.

the "0% porosity axis." The grain density is the bulk density value at 0% porosity. This grain density determination is subject to great uncertainty, especially at high porosity, but at least it allows identification of sample data of questionable accuracy. Unusual grain density values could result from laboratory mistakes or from gas in the samples.

Figure 4 shows gravimetrically determined wet-bulk density versus wet-bulk density determined by the GRAPE 2-minute count. In general, the GRAPE data appear to be slightly higher. Considering all the assumptions of grain densities and attenuation coefficients, as discussed in Boyce (1976a), however, the correlation of the data is good.

Acoustic anisotropy (Fig. 5) is important for estimating vertical velocities (for seismic reflection profiles) from (1) the horizontal velocities determined by refraction techniques, and from (2) oblique velocities determined by sonobuoy techniques. Acoustic anisotropy in sedimentary rock may be created by some combination of the following variables, as summarized by Press (1966), Carlson and Christensen (1977), and Bachman (1979): (1) alternating layers with high- or low-velocity materials; (2) tabular minerals aligned with bedding, which create fewer gaps (containing pore water) in a direction parallel to bedding; (3) acoustically anisotropic minerals whose high-velocity axis may be aligned with the bedding plane; and the (4) foliation parallel to bedding.

From 0 to 447 meters below the sea floor, Cenozoic nannofossil ooze and chalk have acoustic anisotropies of 0 to 2.5% (sonic velocity faster horizontally than vertically), and 5% is typical in laminated chalk; limestones



Figure 4. Gravimetrically determined wet-bulk density versus special two-minute-GRAPE wet-bulk density, Site 462.



Figure 5. Vertical versus horizontal compressional velocity, Site 462.

and radiolarian oozes, porcellanites, and cherts have anisotropies of 4 to 13%. Middle Maestrichtian volcaniclastics from 447 to 560 meters have an acoustic anisotropy of 2 to 32%, and 4 to 13% is typical. From 560 to 1068 meters, basalt flows and basalt-dolerite sills occur which show no apparent anisotropy, but minor interbedded volcaniclastics have anisotropies from 0 to 20%and 5% is typical. However, the volcaniclastics frequently have very small anisotropies compared with the volcaniclastic sequence above the basalt section.

Scatter diagrams of horizontal and vertical velocity versus porosity (Fig. 6), wet-water content (Fig. 7), and wet-bulk density (Fig. 8) represent the Cenozoic oozes, chalk, limestone, chert, and porcellanites and the Cretaceous volcanics from Site 462, from 0 to 1068 meters below the sea floor. These figures illustrate the Wood (1941), Wyllie et al. (1956), and Nafe and Drake (1957) theoretical equations (listed in Appendix B), which utilize here, for simplicity, a calcium carbonate matrix (6.45 km/s; 2.72 g/cm³) saturated with sea water (1.53 km/s; 1.025 g/cm³). Wood's (1941) equation assumes a suspension of spheres without rigidity, and theoretically best applies to soft unconsolidated sediment. This equation would tend to give the lower velocity limit. The Wyllie et al. (1956) equation assumes complete rigidity of the carbonate matrix, and should theoretically give the upper velocity limit. The Nafe and Drake equation is shown for n values of 4, 6, and 9. No single value of nfits all the data. For some values of n, the velocities obtained from the Nafe and Drake (1957) equation may be too high (greater than the velocities from the Wyllie et al. equation) or too low (lower than the velocities from the Wood equation).

Basalt velocity measurements are particularly sensitive to whether the sample is fully saturated with water (Christensen and Salibury, 1975). When we first cored



Figure 6. Horizontal and vertical compressional velocity versus gravimetrically determined porosity for sedimentary rock at Site 462. See text for discussion.



Figure 7. Horizonal and vertical compressional velocity versus gravimetrically determined wet-water content for sedimentary rock at Site 462. See text for discussion.

basalt at Hole 462, we drilled mini-cores (about 2 cm in diameter and 2.5 cm long) and measured the velocity on these, but to do so required a long waiting period for official DSDP curatorial procedures to be completed, so there was a possibility of the samples drying. To test this, we measured the velocity across the 6.61-cm diameter of the basalt cores, as they first arrived in the laboratory, and then compared them with the velocities measured across the same interval in mini-cores (Fig. 9). Velocities measured on the whole core (6.61 cm in diameter) are typically 8% faster than velocities measured on the mini-cores. Therefore, from Core 66 downward in Hole 462, basalt velocities were measured on the whole core. Basalt mini-core velocities for Hole 462 are not included in any other cross-plots. The temperature of the whole core was 15°C, versus 21°C for the mini-core; this difference creates only a small bias (less than 0.1%, Press, 1966; Boyce, 1976b) in the basalt velocities. After Core 74 was recovered from Hole 462A, I left the ship and the new scientist (Dr. Fujii) went back to a minicore system which appears to be valid; his velocities match the acoustic well log data and do not appear to be biased in plots of basalt velocity versus porosity (Fig. 10), wet-water content (Fig. 11), and wet-bulk density (Fig. 12).

The cross-plots of basalt velocity versus porosity, wet-water content, and wet-bulk density include the



Figure 8. Horizonal and vertical compressional velocity versus gravimetrically determined wet-bulk density for sedimentary rock at Site 462. See text for discussion.

Wyllie et al. (1956) equation (Appendix B), using a grain velocity of 6.45 km/s, a pore-fluid velocity of 1.53 km/s, and grain and fluid densities of 3.0 g/cm^3 and 1.025 g/cm^3 , respectively.

The plot of basalt velocity against porosity agrees with data from Leg 46 (Matthews, 1979; Kirkpatrick, 1979) and Legs 51, 52, 53 (Salisbury et al., 1980; Donnelly et al., 1980; Hamano, 1980). But the data from Leg 46 and 51, 52, 53 also lie in a zone between the 5 km/s-10% porosity data point and the 6.25 km/s-2% porosity data point, in addition to the zone shown in Figure 10.

The plot of basalt velocity versus wet-bulk density in Figure 12 does not match similar plots from Legs 37, 46, and 51, 52, 53, because velocity is primarily controlled by variation in porosity and not wet-bulk density when the grain density fluctuates between 2.8 and 3.0 g/cm³. Leg 61 basalt had a grain density of about 3.0 and that from Legs 37, 46, and 51, 52, 53 typically had lesser grain densities.

Acoustic impedance is plotted against vertical velocity in Figure 13 for the Cenozoic to Cretaceous sediments, sedimentary rock, and basalt from Site 462. The plot approximates a linear relationship and normally segregates different mineralogies, such as basalt, clastics, limestone, and chert, into lines representing dif-



Figure 9. Comparison of basalt compressional velocity measured on the mini-core (2 cm in diameter) and the whole core (6.6 cm in diameter), Core 66, Hole 462. Line A indicates theoretical equality of the two samples. Line B shows the average of these data. Whole-core velocities average 8% (0.426 km/s) faster than minicore velocities. This apparently resulted from air getting into the mini-core sample (after two hours of processing, core-splitting, and drilling the mini-core). Whole-core velocities were measured immediately when the core came into the laboratory.

ferent bulk elasticities (Boyce, 1976b). In Figure 13, however, a single line is developed for the soft clastic sediments and calcareous oozes, but cherts, basalt, and some limestone give distinct trends.

Figures 14 and 15 show horizontal velocity and vertical velocity, respectively, versus thermal conductivity. The correlation between velocity and thermal conductivity appears to be fair.

Correlations of thermal conductivity with porosity (Fig. 16), wet-water content (Fig. 17), and wet-bulk density (Fig. 18) include equations from Maxwell (1904), Parasnis (1960), and Bullard and Day (1961) (these equations are given in Appendix C). Also shown in these figures is an empirical relationship by Ratcliff (1960). The equations of Maxwell (1904) and Parasnis (1960) use a calcite (crystal) grain thermal conductivity (1 atm and 21°C) of 9.0 \times 10⁻³ cal/(cm s°C) and a pore-fluid thermal conductivity (1 atm and 21°C) of 1.4×10^{-3} cal/(cm s°C) (Clark, 1966; Nafe and Drake, 1963). The equation of Bullard and Day (1961), which is for a linear relationship between the grain and pore-water thermal resistivities, was also adjusted for the grain and fluid thermal conductivities, just mentioned, by using the inverse of each of these thermal conductivity values. Site 462 data agree well with these equations and their corresponding grain shapes and direction of flow. Of course, the basalt data will not match the equations derived, because the equation lines calculated were for limestone, but would match if the equation lines were also calculated for basalt grains.

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Figure 10. Compressional velocity (parallel) of Site 462 basalt at laboratory conditions versus gravimetrically determined porosity. See text for discussion.



Figure 11. Compressional velocity of basalt (parallel) at laboratory conditions versus gravimetrically determined wet-water content Site 462. See text for discussion.

The basalt thermal conductivities are in agreement with data from Leg 46 (Erickson and Hyndman, 1979) and Legs 51, 52, 53 (Hamano, 1980); the authors cited have published density values with their basalt thermal conductivity data.



Figure 12. Compressional velocity of basalt (parallel) at laboratory conditions, versus gravimetrically determined wet-bulk density, Site 462. See text for discussion.

Figure 19 shows electrical resistivity plotted against wet-bulk density, and Figure 20 shows formation factor versus porosity. The data are from the in situ Gearhart-Owen induction log and compensated density log for Hole 462 (Fig. 21). The data have been selected only for the portions of the hole where the hole diameter was not washed out, between 335 and 600 meters below the sea floor. Specific data points and raw data are listed in Table 1, along with interstitial pore-water salinities from the Site Summary (this volume). In situ pore-water electrical resistivities are calculated from these salinities and the in situ temperatures by using charts in Thomas et al. (1934) and by then correcting these data for hydrostatic pressure with techniques given by Horne and Frysinger (1963). The induction log electrical resistivities are corrected for borehole diameter, for the electrical resistivity of the borehole, and for the thicknesses of beds with contrasting resistivities. The induction log electrical resistivity is parallel to bedding.

Figure 19, electrical resistivity versus wet-bulk density, indicates that the data selected appear to be valid, with no unusual lithologies or pore-water resistivities represented. The resistivity corresponding to a density value of one should be the resistivity of the pore waters. The other lines drawn in indicate where the samples with grain densities of 3.0, 2.7, and 2.4 g/cm³ should appear. The Site 462 basalt data tend to fall on the 3.0 g/cm³ line; and the cherty porcellanite rocks fall between the lines for 2.7 and 2.4 g/cm³, which agrees with Figure 3, where corresponding sedimentary rocks from the same interval below the sea floor have low-grain-density solutions.



Figure 13. Compressional velocity (vertical where possible) versus acoustic impedance at laboratory conditions Site 462. Basalt velocities measured on mini-cores from Hole 462 have been deleted. Acoustic impedance is the product of the laboratory-determined velocity and the gravimetrically determined wet-bulk density.

Since Figure 19 is applicable mainly where there are not any significant pore-water salinity changes, the plots of formation factor versus porosity (Fig. 20) remove any variation related to pore-water salinity or grain density, and allow comparison of texture and porosity data with data of other investigators.

As shown in Figure 20, the Archie (1942) equation, $F = \phi^{-m}$, would fit the sedimentary data if it had an *m* value of about 2.6; and this equation would fit the data for basalt (fractured in some zones) if it had an *m* value of about 2.1. In general, basalt resistivities range from 18 to 100 ohm-m, with (unfractured basalt) porosities of 4 to 15%.

The basalt's formation-porosity relationships agree fairly well with those for material collected on Legs 46 and 51, 52, 53, and, to a lesser degree, Leg 37, if the pertinent data are replotted and a comparison is made using only the $F = \phi^{-m}$ relationship: for Leg 37 (Hyndman and Drury, 1977), typical *m* is about 2.5; for Leg 46 (Christensen et al., 1979), typical *m* is about 1.9 (Kirkpatrick's 1979 logging data appear to be anomalous, perhaps because his density-porosity data are not quantitative enough); and for Legs 51, 52, 53 (Salisbury et al., 1980; Hamano, 1980), typical *m* is about 1.9.



Figure 14. Horizontal compressional sound velocity versus thermal conductivity at laboratory conditions, Site 462. The basalt minicores from Hole 462, Cores 61 to 66, are omitted. This figure is not directly comparable to Figure 15, because there is not necessarily a corresponding vertical velocity for each horizontal velocity. For homogeneous basalt units, the scatter of velocity versus thermal conductivity is significantly less.



Figure 15. Vertical compressional sound velocity versus thermal conductivity at laboratory conditions, Site 462. The basalt mini-cores from Hole 462, Cores 61 to 66, are omitted. This figure is not directly comparable to Figure 14, because there is not necessarily a corresponding horizontal velocity for each vertical velocity. The plot of basalt velocity versus conductivity here is for the more homogeneous basalts from Cores 75 to 91, Hole 462A, so the scatter is less than that among data in Figure 14, which is for a great range of basalt types.



Figure 16. Thermal conductivity versus gravimetrically determined porosity at laboratory pressure and temperature (21 °C), Site 462. A = Maxwell (1904) and Parasnis (1960): calcite spheres; k_g and $k_w = 9.0 \times 10^{-3}$ and 1.4×10^{-3} cal/(cm °C s). B = Ratcliff (1960): empirical. C = Bullard and Day (1961): modified for calcite; k_g and $k_w = 9.0 \times 10^{-3}$ and 1.4×10^{-3} cal/(cm °C s). D = Parasnis (1960): calcite; flow perpendicular to axis of thin plates; k_g and $k_w = 9.0 \times 10^{-3}$ and 1.4×10^{-3} cal/(cm °C s). E = Parasnis (1960): calcite; flow parallel to axis of thin plates; k_g and $k_w = 9.0 \times 10^{-3}$ and 1.4×10^{-3} cal/(cm °C s). F = Parasnis (1960): calcite; flow perpendicular to axis of long cylindrical grains; k_g and $k_w = 1.9 \times 10^{-3}$ and 1.4×10^{-3} cal/(cm °C s).

Therefore, it appears that Site 462 basalt data do not disagree significantly with data for other DSDP deep holes in basalt, except perhaps those of Leg 37. Leg 37 holes may have penetrated basalts with some combination of the following: (1) a more tortuous texture, (2) lack of fractures, (3) a lower pore-water salinity than assumed, (4) less conductive solids. Other possible reasons for the difference in Leg 37 data are that (1) the induction log measured resistivity in a direction parallel to bedding; (2) different techniques were used in measuring the resistivity—Leg 37 data were laboratory-determined; (3) logs were incorrect; or (4) Site 462 basalts have greater pore-water salinities than one would expect. In general, it appears that the resistivities meas-



Figure 17. Thermal conductivity versus gravimetrically determined wet-water content at laboratory pressure and temperature (21 °C), Site 462. See caption of Figure 16 for further explanation.

ured by well logs tend to be lower in absolute value than the laboratory-measured resistivities; this seems to result from the logs including cracks and fractures in their sampling, as suggested by Christensen et al. (1979), Kirkpatrick (1979), and Salisbury et al. (1980), but also, in part, from the induction log resistivity being measured parallel to bedding.

In Figure 20, formation factor versus porosity, the Archie (1942) equations agree with the data. Equations derived by Maxwell (1904), Winsauer et al. (1952), Boyce (1968), Kermabon et al. (1969) (see Appendix A for these equations) gave formation factors which were too low. These equations, however, are theoretically more applicable to modern unconsolidated formations (Boyce, 1980).

Interpretations of Density and Electrical Logs versus Depth

The R_{wa} curve (Fig. 21) is calculated by rearranging the Archie (1942) equation: R_{wa} = (resistivity induction log) (ϕ^2), where ϕ is the fractional porosity derived from the density log, assuming a 2.7 g/cm³ grain density



Figure 18. Thermal conductivity versus gravimetrically determined wet-bulk density at laboratory pressure and temperature (21°C), Site 462. See caption of Figure 16 for further explanation.

above the basalt (561 m) and 3.0 g/cm³ below 561 meters. This is a continuous plot versus depth, and is used here mainly as a tool to identify unusual geological zones, such as (1) metallic deposits, (2) temperature anomalies, (3) interstitial water salinity anomalies, (4) heavy mineral deposits, (5) very light mineral deposits, or (6) hydrocarbons. It is not designed to calculate the R_{wa} accurately, but only to indicate anomalous lithologic zones.

In Figure 21, the R_{wa} curve shows a real anomaly, at 393.5 to 396.5 meters, which could be (1) 76% hydrocarbon (water saturation in pores = $S_w = [R_{wa} \text{ nor-} mal/R_{wa} \text{ maximum anomaly}]^{1/2} = [0.25 \text{ ohm-m/4.4}]^{1/2}$ ohm-m]^{1/2} = 25%]); (2) low-grain-density (2.2 g/cm³) opal silica chert or porcellanite; or (3) relatively fresh pore water (1.8% salinity). The low-grain-density chert appears to be the true cause of the anomaly, since (1) the last occurrence of opal radiolarian ooze was recovered at 380 meters, and below 380 meters radiolarians begin to be replaced by opal-CT and clinoptilolite (which are also low in grain density: 2 to 2.4 g/cm³); and (2) low grain densities are indicated by cross-plots of gravimetrically determined porosity and wet-bulk density (Fig. 3), using data for cores recovered in this interval below the sea floor, which encompasses mainly chert, porcellanite, and limestone; softer siliceous or calcareous oozes could have been washed away in the drilling process, however.



Figure 19. In situ formation electrical resistivity versus in situ wetbulk density, as measured on the Gearhart-Owen single Induction Log and Compensated Density Log. The Induction Log data were corrected for borehole diameter, hole fluid resistivity, and for beds of contrasting resistivity and bed thickness. Numbers on the data points correspond to column one in Table 1.

In the R_{wa} curve, the data from above 390 meters are not good; they are not quantitative, because the hole is either washed out or of anomalously small diameter. The sediments between the basalt layers give slightly high R_{wa} anomalies, but this appears to be because their true grain densities are 2.7 instead of the 3.0 g/cm³ assumed in the R_{wa} calculation. All other anomalies are artifacts of (1) slight misalignment of the density and induction logs; (2) thin beds with contrasting resistivities, since the induction log resistivity was not adjusted for borehole conditions; and (3) a 1.2-meter vertical resolution of the induction log relative to the 23-cm vertical resolution of the density log. The R_{wa} curve, in general, decreases slightly with increasing depth, because of increasing temperature. Anomalies in the temperature log in Figure 21 are artifacts, and occur where the tool hung up in the hole on its way down and required several minutes before continuing.

Interpretation of Temperature Logs versus Depth

Shown in Figure 22 are the Gearhart-Owen continuous temperature logs versus depth, the equilibrium



Figure 20. Electrical formation factor versus *in situ* porosity from Density Logs, Hole 462, assuming grain density of 2.4 g/cm³ for points 1 and 2; 2.7 g/cm³ for points 3–7, 11, 13, and 14; and 3.0 g/cm³ for points 8, 10, and 12. Number on data points correspond to raw data in Table 1. Shown in the figure are lines representing Archie's (1942) formula $F = a \phi^{-m}$, where a = 1.0 and different values of *m* result in different slopes. These data agree with Boyce's (1968) discussion of Archie's (1942) formula where *m* is between 2.0 and 2.5, but the other equations derived by Boyce (1968), Maxwell (1904), Kermabon et al. (1969), and Winsauer et al. (1952) do not give large enough formation factors. Thus, the data agree with the summary and discussions in Keller (1966) and Keller and Frischknecht (1966), where cementation and more consolidated sediments have typically higher values of *m* (e.g., 2.5).

temperature calculated from the two sets of Gearhart-Owen log data, and the Uyeda temperature probe data from Site Summary, this volume, where heat-flow calculations are reported); the density log is also shown in Figure 22 to aid as an indicator of lithology and thermal conductivity changes which would also cause changes in temperature gradient with depth. The methods of calculating temperature are discussed in the methods section, and the technique will not be discussed further. Measurements in the upper 100 meters of the hole were taken with the drill pipe and bottom-hole assembly present.

Anomalous data occur where the drill string was present, because the temperature tools were pumped down the pipe to prevent the tool from floating in the pipe. Thus, this water circulation invalidates the temperature data in this zone.

The Gearhart-Owen data and the corresponding equilibrium temperatures show many anomalies versus depth, and these were caused by the temperature sensor getting hung up in the hole (on the way down) for a considerable time before it could be worked past any blockage or shelves in the hole.

In addition, equilibrium temperatures and any temperature anomalies are suspect as artifacts in any portion of the hole where its diameter is extremely variable, washed out, or where the formation is fractured. Fractures in the basalts, for example, allow the drill-circulation water to cool the formation more than at intervals where the formation is not fractured. The negative equilibrium temperature anomalies in the basalt zone may therefore be artifacts of the drilling operation.

The calculated equilibrium temperatures are very sensitive to small variations in the raw data, so most of the variation in the final equilibrium temperature is probably only "noisy" data.

If one ignores all the above precautions, however, then in the basalt section (below 561 m) the equilibrium temperatures suggest a vertical section where the temperature does not increase with depth, and even decreases with depth. This decrease with depth probably reflects artifacts in the data; but if the temperature variation over this section is in fact representative of in situ conditions, then the data indicate convection of water, at least in the upper portion of the basalt section at Site 462. Heat would in this case be transferred primarily by convection, and conduction would be insignificant. Such a set of circumstances seems to be indicated by the data for basalt drilled on Legs 37, 46, and 51, 52, 53. If these indications are valid, then it is of particular interest with respect to the older crust at Site 462, where reduced permeability has been predicted because of vein material precipitating in the fractures.

SUMMARY AND CONCLUSIONS

1. From 0 to 447 meters below the sea floor, Cenozoic nannofossil oozes and chalks have acoustic anisotropies of 0 to 2.5% (sonic velocity faster horizontally), and 5% is typical in laminated chalk; limestones and radiolarian oozes, porcellanites, and cherts have anisotropies of 4 to 13%. Middle Maestrichtian volcaniclastics from 447 to 560 meters have an acoustic anisotropy of 2 to 32%, and 4 to 13% is typical. Basalt flows and sills with no apparent anisotropy occur from 560 to 1068 meters, but minor interbedded volcaniclastics have anisotropies of 0 to 20% and 5% is typical. These volcaniclastics frequently have very small anisotropies, however, compared with the volcaniclastic sequences above the basalt section.



Figure 21. Selected Gearhart-Owen logging data versus depth in Hole 462, which are required to calculate (or interpret) the "apparent interstitial water electrical resistivity" (Rwa) curve. See text for discussion.

Table 1. Selected data derived from compensated density log and induction log.

Plot No.	Depth below	Depth below Sea Floor ^a (m)	Hydrostatic Pressure ^b (kg/cm ²)	Interstitial Water Salinity ^c (‰)	Temperature ^d (°C)	In situ Interstitial Water Resistivity, ^e R _w (ohm-m)	Hole Diameter (cm)	Resistivity of Hole Fluid ^f (ohm-m)	True Formation Resistivity, ^g <i>R_I</i> (ohm-m)	Formation Factor, $F = (R_I/R_W)$	Porosity $\phi = \sqrt[m]{1/F}$ Sed. $m = 2.6$ Bas. $m = 2.1$	Wet-Bulk Density (g/cm ³)		Porosity ^h			Gamma- Rav	
	Floor (m)												$Q_g = 3.0$ (%)	$\varrho_{g} = 2.7$ (%)	$\varrho_{g} = 2.6$	$Q_g = 2.4$	Log (API)	Probable Lithology Recovered in Core
1	5583	394	577	35.6	7.8	0.255	27.9	0.259	8.77	34.39	0.256	2.19		(30.5)	(26.1)	15.3	10	Silicified limestone; chert.
2	5605	416	579	35.6	8.5	0.251	28.7	0.256	2.15	8.57	0.438	1.89		(48.5)	(45.2)	37.2	25	Silicified limestone; chert.
3	5627	438	581	35.6	11.8	0.232	29.9	0.237	4.00	17.24	0.335	2.20		29.9			12	Silicified limestone; chert.
4	5652	463	584	35.7	13.0	0.226	29.2	0.230	1.05	4.62	0.555	1.70		60.0			30	Volcanic siltstone; limestone; nannofossil marl; nannofossil chalk.
5	5683	494	587	35.7	14.3	0.219	33.8	0.224	2.78	12.68	0.376	2.23		28.1			20	Claystone; calcareous claystone.
6	5715	526	591	35.6	15.3	0.215	35.6 + ?	0.219	0.90?	4.17?	0.577	1.72?		58.7			48	Zeolitic claystone.
7	5734	545	592	34.2	15.7	0.221	33.5	0.217	1.08	4.89	0.543	1.75		56.9			22	Zeolitic claystone; zeolitic mudstone; nannofossil marlstone.
8	5753	564	594	34.2	16.3	0.219	25.4	0.214	17.86	81.54	0.123	2.75	12.7				20	Basalt, altered.
9	5755	566	595	34.2	16.3	0.219	24.9	0.214	30.30	138.37	0.096	2.68	16.2				20	Basalt, altered.
10	5757	568	595	34.2	16.3	0.219	24.1	0.214	100.00	456.62	0.054	2.88	6.1				7	Basalt.
11	5760	571	595	34.2	16.3	0.219	27.4	0.214	1.92	8.78	0.434	1.85		50.9			49	Volcaniclastics(?).
12	\$762	573	595	34.2	16.4	0.219	23.9	0.214	46.00	210.05	0.078	2.92	4.1				8	Basalt.
13	5764.5	\$75.5	596	34.2	16.5	0.218	31.8	0.214	1.63	7.46	0.462	1.83	6265	52.4			31	Volcaniclastics(?).
14	5768	579	596	34.2	17.0	0.216	23.4	0.212	3.28	15.18	0.351	2.08		37.1			42	Volcaniclastics(?).

^a Depth below sea floor = depth below the rig floor minus water depth (5179 m) minus 10 meters.

^b Hydrostatic pressure = depth below sea level × 1.035 g/cm³

^c Interstitial water salinity is from shipboard determination on sedimentary samples, and these data are assumed to apply to the basalt section.

^d Temperature is from second temperature log.

⁶ Electrical resistivity is calculated from temperature and salinity based on data of Thomas et al. (1934) and then corrected for hydrostatic pressure by techniques formulated by Horne and Frysinger (1963).
⁷ The hole fluid was sea water with a salinity of 35%, and its *in situ* resistivity was calculated by the same method as that for interstitial water resistivity in footnote "e."

8 This R_i is corrected for borehole diameters and bed thickness and adjacent bed resistivity. ^h Porosity = $(\varrho_g - \varrho_b)/(\varrho_g - \varrho_w)$, where ϱ_g is 2.7 and 2.4 g/cm³ for sedimentary material and 3.0 g/cm³ for basalt, $\varrho_w = 1.03$ g/cm³, and ϱ_b = wet bulk density from density log. ϱ_g estimated from laboratory porosity-density measurements.



Figure 22. Temperature versus depth below the sea floor at Hole 462. Temperatures shown are the two Gearhart-Owen continuous Temperature Logs and the theoretical equilibrium temperatures calculated from these two Gearhart-Owen Logs, and temperatures from the Uyeda Temperature Probe. The wet-bulk Density Log is also shown to indicate lithology and to provide an index to thermal conductivity.

2. Cross-plots of the laboratory-measured compressional sound velocity versus wet-bulk density, wet-water content, and porosity of sediments and sedimentary rock typically agree with equations derived by Wyllie et al. (1956) and Wood (1941); the equation of Wyllie et al. (1956) fits well with similar basalt cross-plots. Crossplots of basalt velocity versus porosity agree well with the data from Legs 37, 46, and 51, 52, 53. Basalt velocity versus wet-bulk density did not agree well, because of varying grain density.

3. Cross-plots of thermal conductivity and sound velocity indicate only a fair correlation. Cross-plots of thermal conductivity versus porosity, wet-water content, and wet-bulk density correlate well with equations derived by Maxwell (1904), Ratcliff (1960), Parasnis (1960), and Bullard and Day (1961). Thermal conductivity of basalt agrees with thermal conductivity data for Legs 46 and 51, 52, 53, published also with their corresponding density data.

4. Electrical formation factor versus porosity for sediments and sedimentary rock agrees with the Archie (1942) equation with m values of 2.6; for (in part fractured), basalt an m of about 2.1 is typical. Formation factors are greater than those derived from equations in Maxwell (1904), Winsauer et al. (1952), Boyce (1968), and Kermabon et al. (1969). These equations should be more applicable for unconsolidated sediments, for which they were derived. The basalt formation factor-porosity relationship, $F = \phi^{-m}$, has an m which agrees with the value of 1.9 for Legs 46 and 51, 52, 53. Leg 37 material, however, appears to have a slightly higher m of 2.5, suggesting some combination of the following: (1) Leg 37 material has a different texture (a more tortuous path for the ions to travel); (2) Leg 37

material is not as fractured; (3) Leg 37 material has a lower pore-water salinity than expected; (4) our Leg 61 pore-water salinity is higher than we suspect; (5) Leg 37 material has less conductive solids; (6) the induction log measured parallel to bedding; (7) perhaps the higher mreflects the different techniques used on Leg 37. For Hole 462, the pore-water salinity of the basalt does not seem to differ greatly from that in the overlying sedimentary section.

5. An "apparent interstitial water resistivity" (R_{wa}) curve was derived from the density and induction logs. This R_{wa} curve indicated an anomaly between 393.5 and 396.5 meters, which could be interpreted as (1) 76% hydrocarbons, (2) relatively fresh pore water (1.8% salinity), or (3) low-grain-density (2.2 g/cm³) semilithified porcellanite-chert. The low-grain-density porcellanite-chert appears to be the most plausible interpretation.

6. In situ temperatures measured by the Uyeda temperature probe were about (2 to 5° C) 50% higher than the equilibrium temperatures (Lachenbruch and Brewer, 1956) extrapolated from two Gearhart-Owen continuous temperature logs. This difference probably results from washing out of the hole in this interval, so equilibrium temperatures are not reliable over this depth interval.

7. Ignoring all precautions about temperature artifacts, the equilibrium temperatures of the Gearhart-Owen temperature logs suggest that in at least the upper 40 meters of the basalt section hydrothermal circulation is occurring and heat is transferred by convection, not conduction. This is consonant with the apparent water convection discovered on Legs 37, 46, and 51, 52, 53 in the deep basalt holes. Such hydrothermal circulation/convection is also of interest with reference to older basalt crust: it has been thought that the older basalt would be less permeable, because its fractures have had more time to be filled by veins. The permeability at Site 462 may still be less(?), but at least the basalt section still appears to be permeable. The anomalous equilibrium temperature is probably an artifact, however, of (1) varying hole diameter and fractured formation (so that the formation may have been cooled excessively during drilling operations), and (2) inappropriate application of the Lachenbuch and Brewer (1959) equation. Obviously, this whole question needs further investigation.

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APPENDIX A Electrical Formation Factor-Porosity Relations of Wet-Saturated Sedimentary Rock and Sediment¹

Maxwell (1904) theoretically derived the following relationship for a suspension of spheres:

$$F = \frac{3-\phi}{2\phi}$$

where

 $F = R_0/R_w =$ formation factor;

 R_0 = the electrical resistivity of the saturated formation;

 R_w = the resistivity of the interstitial water;

 ϕ = the porosity expressed as a fraction or decimal.

Archie's (1942) equation was derived for consolidated sandstone without clay material:

$$F = \phi^{-m} = \phi^{-2}$$

where m is a variable depending on consolidation, texture, and cementation.

Winsauer et al. (1952) derived a slightly different empirical formula for various sandstone formations:

$$F = a\phi^{-m} = 0.62 \theta^{-2.15}$$

where a and m are variables depending on cementation, textures, and mineralogy of the formation.

Boyce (1968) derived the following empirical equation for modern marine diatomaceous silty clay to silty sand (this equation is of the same form as that of Winsauer et al., 1952):

$$F = 1.30 \theta^{-1.45}$$

Kermabon et al. (1969) derived the following empirical equation (one of three), also for modern marine clays and turbidite sands:

$$F = \left(\frac{1.45}{\phi}\right)^{1.46} - 0.719$$

APPENDIX B

Theoretical Equations Relating Compressional Velocity of the Wet-Saturated Rock to the Velocities and Densities of the Fluid and Solid Grain End-Member Constituents

The Wood (1941) equation applies to velocities through suspensions without rigidity:

$$V_b = \left(\frac{1}{[\phi \beta_w + (1 - \phi) \beta_g] [\phi \varrho_w + (1 - \phi) \varrho_g]} \right)^{1/2}$$

where

V = compressional velocity;

 ρ = density, β = compressibility, and subscripts g, w, and b represent the solid grains, interstitial water, and wet-bulk rock or sediment, respectively;

¹ Keller (1966) and Keller and Frischknecht (1966) summarize and discuss similar equations derived for continental formations.

ϕ = fractional porosity, where $\rho_b = \phi \rho_w + (1 - \phi) \rho_e$.

The Wyllie et al. (1956) equation applies in rocks with rigidity:

$$\frac{1}{V_b} = \frac{\phi}{V_w} + \frac{(1-\phi)}{V_g}$$

The Nafe and Drake (1957) equation applies to rock with varying degrees of rigidity, which is controlled in the equation by the value of n:

$$V_b^2 = \phi V_w^2 \left[1 + \left(\frac{\varrho_w}{\varrho_b} \right) (1 - \phi) \right] + \frac{\varrho_g}{\varrho_b} (1 - \phi)^n V_g^2$$

APPENDIX C Relationships of Thermal Conductivity to Porosity, Wet-Water Content, and Wet-Bulk Density

Maxwell (1904); grains are spherical:

$$k_b = k_w \left[\frac{3 - 2\phi \left(1 - \frac{k_w}{k_g}\right)}{3k_w/k_g + \phi \left(1 - \frac{k_w}{k_g}\right)} \right]$$

where

 k_b = thermal conductivity of the bulk water saturated sediment;

 k_w = thermal conductivity of sea water;

 k_g = thermal conductivity of the grains; ϕ = porosity

 $= w/[w + (1 - w) \varrho_w/\varrho_g]$

in which

w = wet-water content;

 $\varrho_w = \text{density of water;}$

 ϱ_g = density of solid grains.

Parasnis (1960):

Model 1: Grains are long thin plates oriented perpendicular to the direction of heat flow:

$$\frac{k_{w}}{k_{g}} = \frac{\left(\frac{k_{w}}{k_{g}}\right) + e\left(\frac{k_{w}}{k_{g}}\right)}{\left(\frac{k_{w}}{k_{g}}\right) + e}$$

where

 $e = \text{void ratio} = (\text{weight water})/(\text{weight of dry solids}) = (\varrho \cdot \phi)/(\psi \cdot \phi)$ $\varrho (1 - \phi)$

Model 2: Grains are long cylindrical bodies with axes of cylinders perpendicular to direction of heat flow:

$$\frac{k_b}{k_g} = \frac{2\left(\frac{k_w}{k_g}\right) + e\left(\frac{k_w}{k_g}\right)\left[1 + \left(\frac{k_w}{k_g}\right)\right]}{2\left(\frac{k_w}{k_g}\right) + e\left(1 + \frac{k_w}{k_g}\right)}$$

Model 3: Grains are spherical:

$$\frac{k_b}{k_g} = \frac{3\left(\frac{k_w}{k_g}\right) + e\left(\frac{k_w}{k_g}\right)\left[1 + 2\left(\frac{k_w}{k_g}\right)\right]}{3\left(\frac{k_w}{k_g}\right) + e\left[1 + 2\left(\frac{k_w}{k_g}\right)\right]}$$

Model 4: Grains are long thin plates with their lengths parallel to the direction of heat flow:

$$\frac{k_g}{k_g} = \frac{1 + e\left(\frac{k_w}{k_g}\right)}{1 + e}$$

Bullard and Day (1961): Empirical deviation for thermal conductivity of sediments:

$$\frac{1}{k_b} = 161 + 651 w$$

The above is true for $k_w = 1.23 \cdot 10^{-3}$ (cal-cm)(cm² s°C) and $k_g = 6.21 \cdot 10^{-3}$ (cal-cm)/(cm² s°C).

$$\frac{1}{k_b} = 111 + 603 w$$

The above is modified by Boyce for $k_w = 1.40 \cdot 10^{-3} \text{ (cal-cm)/(cm^2 s^{\circ}C)}$ and $k_g = 9.0 \cdot 10^{-3} \text{ (cal-cm)/(cm^2 s^{\circ}C)}$.