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Publication Date
1984-10-01

LBL-18564
Preprint c.1

Lawrence Berkeley Laboratory
UNIVERSITY OF CALIFORNIA

EARTH SCIENCES DIVISION

Submitted to the Journal of Geophysical Research

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October 1984

Prepared for the U.S. Department of Energy under Contract DE-AC03-76SF00098

DISCLAIMER

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This work was supported by the Director, Office of Energy Research Office of Basic Energy Sciences, Division of Engineering, Mathematics and Geosciences, U. S. Department of Energy under Contract No. DE-AC03-76SF00098.

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Convection in the Oceanic Crust: Numerical Simulation of Observations from DSDP Hole 504B, Costa Rica Rift

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ABSTRACT

Three dimensional modeling of convection in the oceanic crust at DSDP site 504B using an Integral Finite Difference Model closely duplicates underpressures, surface heat flow, downhole temperature profiles and fluid drawdown rates observed by in situ measurements in the borehole. The major constraint to produce such good fits is the permeability versus depth function which was actually measured in the borehole. Pronounced "underpressuring" (fluid pressures less than the hydrostat) occur throughout the convection cell if and only if a tight, impermeable cap rock (chert) exists over the cell. The computed flow rates of water from the ocean through the borehole into the basalt agree closely with the measurements carried out 65, 720 and 1280 days after drilling. The model predicts a inflow rate of 50 to 60 litres per hour during the expected next occupation of the site by mid 1986. Numerical modeling has confirmed that an active hydrothermal convection cell exists in the region of Hole 504B and that a very-low-permeability cap rock is necessary for the existence of the convection cell. Convective hydrothermal cells in the oceanic crust can explain observed periodic heat flow variations in sub-oceanic crust.

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INTRODUCTION

Surface heat flow measurements made near the Costa Rica Rift in the Eastern Equatorial Pacific during the past 14 years have shown substantial deviations from the heat flow distributions predicted by conductive plate models of the cooling lithosphere (Figure 1) (Williams et al., 1974; Anderson and Hobart, 1976; Sclater et al., 1976; Anderson and Skilbeck, 1981; Hobart et al., in press). One explanation of these heat flow anomalies invokes the continued convection of sea water within the oceanic crust at considerable distances from the midocean ridge axis (Anderson et al, 1977).

As part of the Deep Sea Drilling Project (DSDP), extensive measurements have been made of the physical, chemical, and thermal state of the oceanic crust surrounding Hole 504B, which penetrates 275 m of sediment and 1076 m of basement in 6-m.y.-old crust on the flank of the Costa Rica Rift (Figure 2) (cf. Anderson, Honnorez, et al., 1982; Cann, Langseth, Honnorez, Von Herzen, White et al., 1983) Data collected included temperature logs, permeability and porosity measurements, thermal conductivity measurements, and an observation of significant under-pressuring of 8 bars or 2% of hydrostatic pressure at that depth at a penetration depth of 450 m into the formation (Anderson and Zoback, 1982; Becker et al., 1982, 1983a, 1983b, 1984). Anderson and Zoback (1982) suggested that this under-pressuring could best be explained by the presence of an active convection cell within the basement near Hole 504B, despite the observation that surface heat flows near the hole vary closely about the values predicted for purely conductive plate heat transfer (figure 1) (Hobart et al., in press).

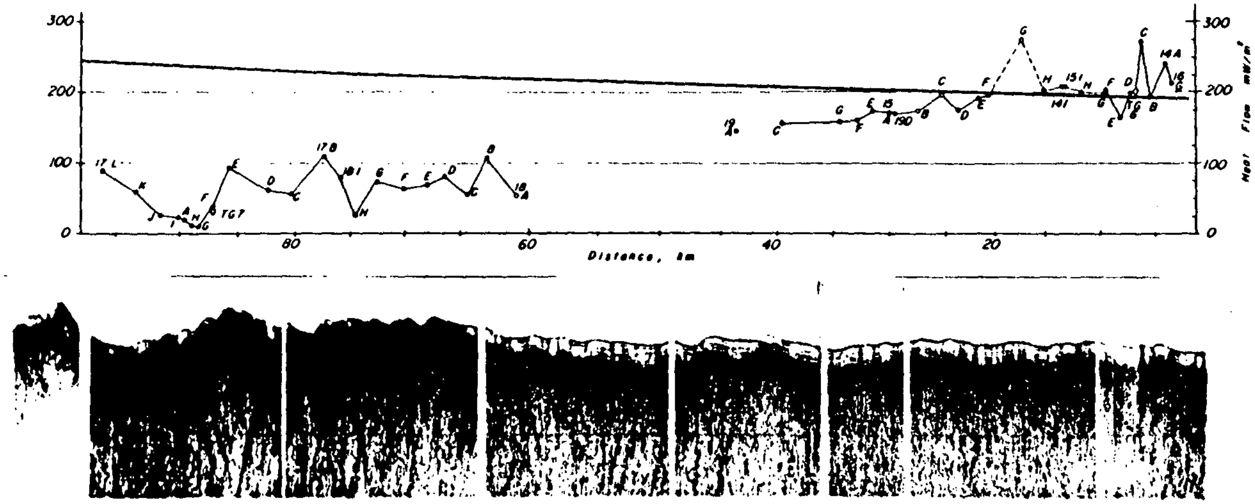

Figure 1: Heat flow profile on the South flank of the Costa Rica Rift (see map, Fig. 2 from Anderson and Zoback, 1982).

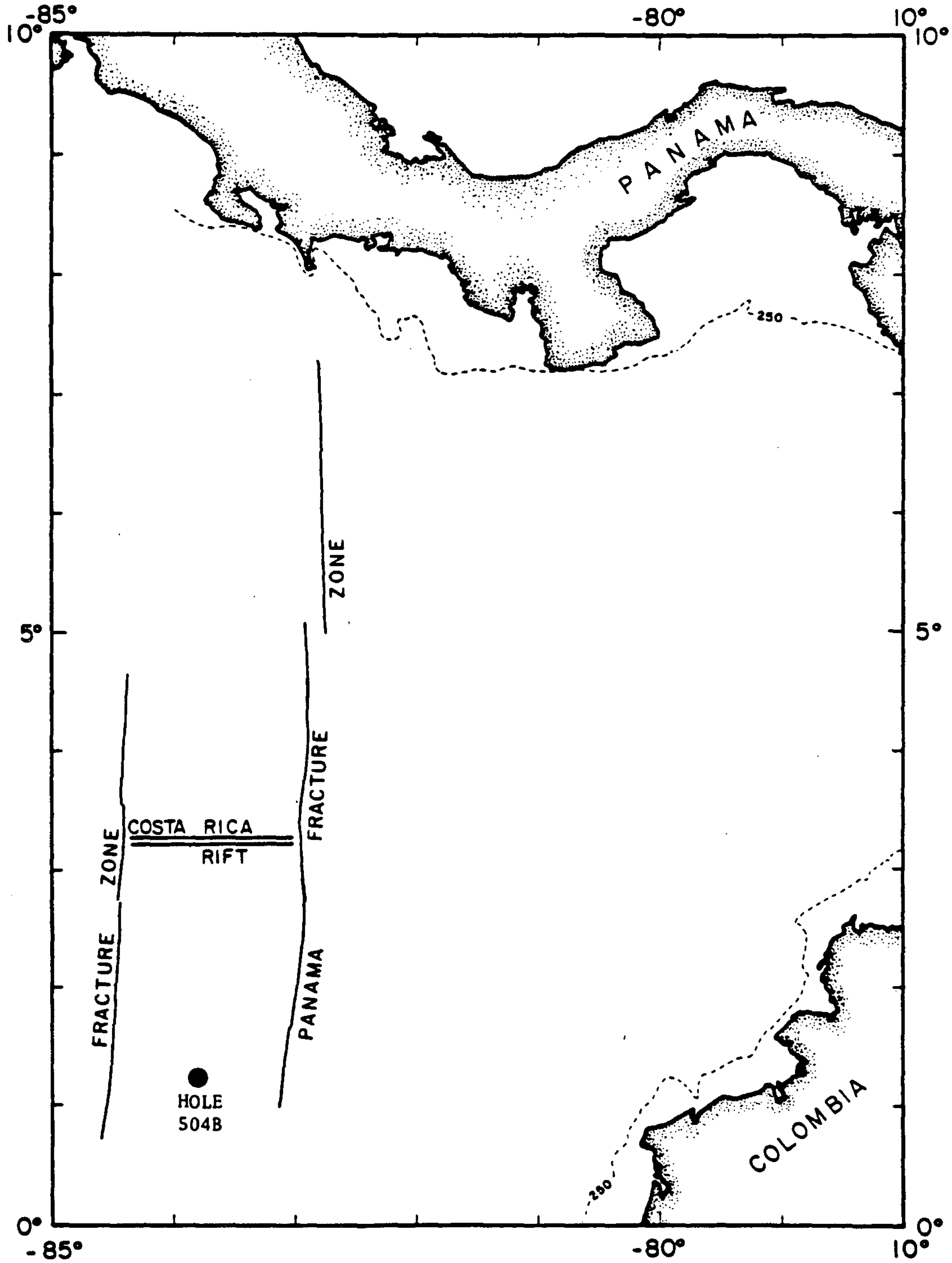

Figure 2: Location of DSDP Hole 504B (from Anderson and Zoback, 1982).

The purpose of the present study is to test this convection cell hypothesis with a three-dimensional mathematical model of the oceanic crust near 504B, using the permeability, thermal conductivity, and other measurements obtained through the drilling programs at that site. Physical and thermal conditions at other locations on the rift flank have been estimated by varying some of the structural parameters of the system. In addition, the effect of the wellbore itself on observations of the temperature and pressure distributions in the natural system has also been investigated. Although other studies have used numerical methods to model convection that occurs near the ridge axis (e.g., Ribando, et al., 1976; Patterson and Lowell 1982; Gartling, 1978) no previous studies have been conducted which use actual formation data in the convection cell analysis, because prior to Hole 504B no such measurements were available.

FIELD DATA

Hole 504B was occupied by four different DSDP legs between 1979 and 1983, first drilled during Leg 69 in 1979, and deepened during legs 70 and 83 to 1350 m below sea floor or 1076 m into basement, twice the basement penetration of any DSDP hole. A large array of in situ geophysical and geochemical experiments were completed in the borehole during Legs 69, 70, 83, and 92. Hole 504B is the only drillhole in the oceans for which permeability and pore pressure have been measured over enough depth to describe their variation with depth. At the same time, flow rate into the formation through the drill hole was estimated over the four years from temperature profiles measured during legs 70, 83 and 92. The most extensive suite of geophysical logs from the Deep Sea Drilling Project were also run in Hole 504B during Leg 83. One final unique feature of Hole 504B is that it remains

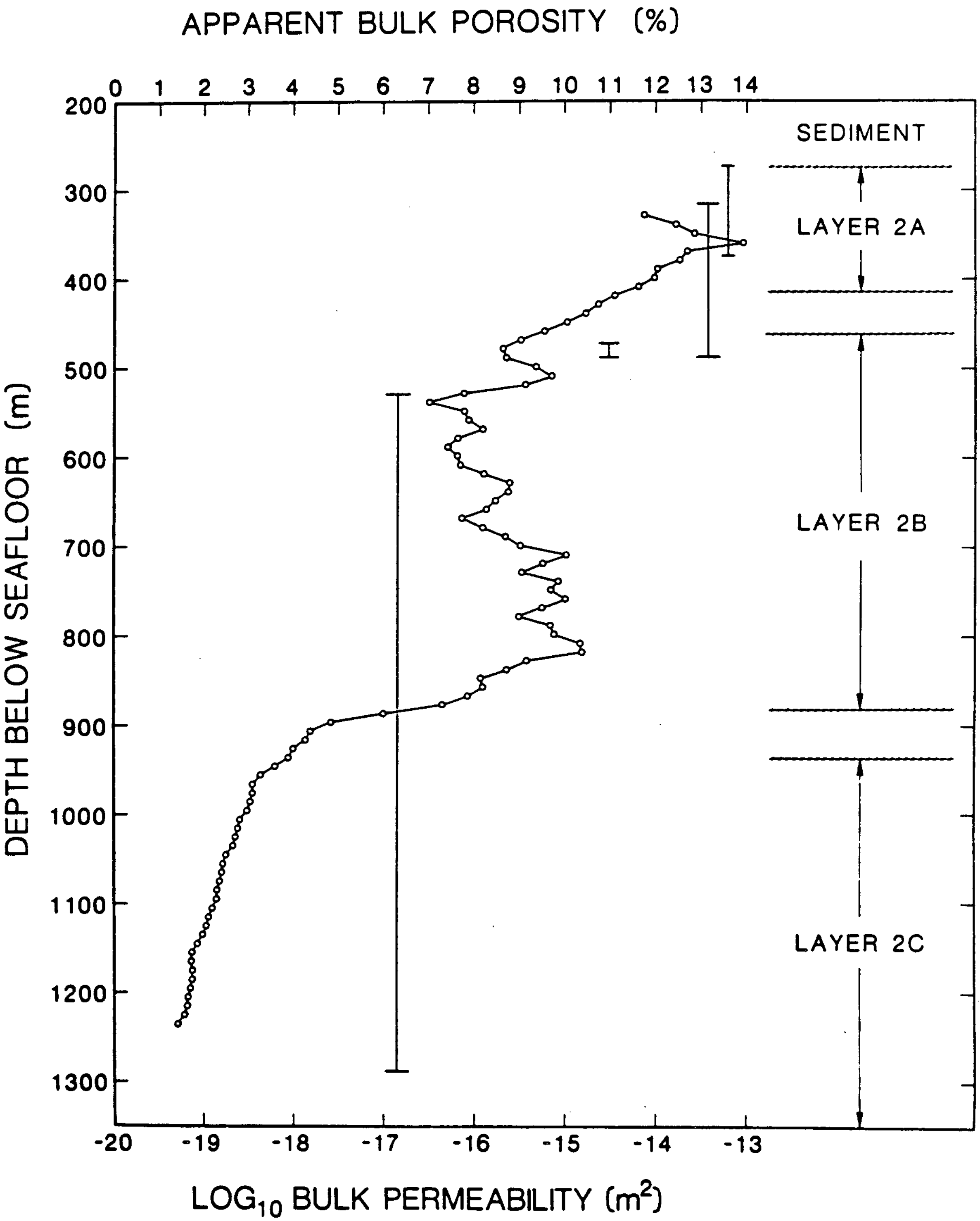

Figure 3: Vertical distribution of porosity and permeability, DSDP Hole 504B. Porosities were determined from large scale electrical resistivity measurements (Becker et al., 1982). Permeabilities are average values over the indicated vertical intervals (Anderson and Zoback, 1982).

open to reoccupation during the upcoming decade of the new Ocean Drilling Project. Therefore, predictions made by our modeling study can be tested in the future.

The stratigraphy of Hole 504B is very similar to that of ophiolites on land (Anderson, et al., 1982). The upper 275 meters of sediments are equatorial carbonate oozes grading downward to chalks, and a basal 50 meters of chert. The basement (below a depth of about 300 m from the sea floor) is basalt of remarkably constant chemistry. Structurally, the upper crust is gradationally layered from an upper 150 meters of rubbly pillow basalts (Layer 2A), to 450 meters of smaller, more coherent and more highly altered pillow basalts (Layer 2B), and finally the hole bottoms in 500 meters of sheeted dikes (top of Layer 2C). Although core recovery averaged only 20%, this lithostratigraphy is known so precisely because the entire basement section was imaged with an ultrasonic borehole televiewer (Zoback and Anderson, 1982; Newmark et al., 1984), which substantially aided interpretations of cuttings and logs.

Permeability Versus Depth

Anderson and Zoback (1982) and Anderson et al. (1984) used a drill stem packer to conduct flow tests over three intervals of Hole 504B. They measured a decrease in permeability from 10^{-13} to 10^{-14} m^2 in Layer 2A, to 10^{-15} to 10^{-17} m^2 in Layer 2B, and to 10^{-17} m^2 and lower in Layer 2C. They estimated a permeability (k) versus depth (z, in meters) function to be approximately:

$$k(z) = 0.11\ e^{-z/50} \times 10^{-12}\ m^2. \qquad (1)$$

The accuracy of such a function is somewhat verified from the porosity versus depth function measured independently using ultra long spacing electrical resistivity by Becker et al. (1982, 1984), (Figure 3). Also the high permeabilities measured in Layer 2A were supported by the values required to model temperatures and flow rates measured in the layer by Becker et al. (1983a, 1983b, 1984).

TABLE 1

Summary of Mesh Materials Properties

Material	Permeability, md*	Porosity (%)	Thermal Conductivity (W/m-K)
Chert	0	1	1.26
Basalt 1	100	11	1.67
Basalt 2	0.4	8	1.67
Basalt 3	0.2	3	2.20

* 1 md $\approx 10^{-15}$ m^2

TABLE 2

Basalt Permeability Variations

Material	Permeability (md)* Case 1	Case 2	Case 3
Basalt 1	200	200	500
Basalt 2	0.8	0.8	2.0
Basalt 3	0.4	0.8	1.0

* 1 md $\approx 10^{-15}$ m^2

Pore Pressures less than the Hydrostatic

In addition to permeability, the packer flow tests yielded measurements of in situ pore pressure in the formation. Anderson and Zoback (1982) reported underpressures in Layer 2A of 8 bars below hydrostatic at a depth of 450 m below sea floor. Again this observation was corroborated by temperature and flow rate measurements which demonstrate conclusively that oceanic bottom water has been continuously flowing into the formation through the borehole ever since the drill bit first penetrated this underpressure zone (Becker et al., 1984).

Temperature and Flow Rate versus Time

Becker et al. (1983a, 1983b, 1984) present temperature versus depth measurements at Hole 504B one month, two years, and 3.5 years after the hole penetrated basement during Leg 69. These profiles show a nearly conductive distribution of heat in Layers 2B and 2C with the upper most 100 m of basement acting as an aquifer drawing cold ocean bottom water into the upper oceanic crust through the borehole itself (Figure 4). According to the authors, flow rates down the casing and into the formation are required to decreased from 6,000 to 7,000 ℓ/hr. (about 90 m/hr) 50 days after drilling, to 1,700 ℓ/hr (25 m/hr) after 772 days, and 2.4 m/hr after 1272 days (Figures 5) in order to match the observed temperature profiles.

THE SIMULATION PROBLEM

The task of any numerical simulation of convection in the upper oceanic crust is to describe the physics of the system well enough to replicate all of the observations reported above. We will describe a model below that does this, and predicts the temperature and flow-rate observations to be made in Hole 504B during its next reoccupation planned for 1986.

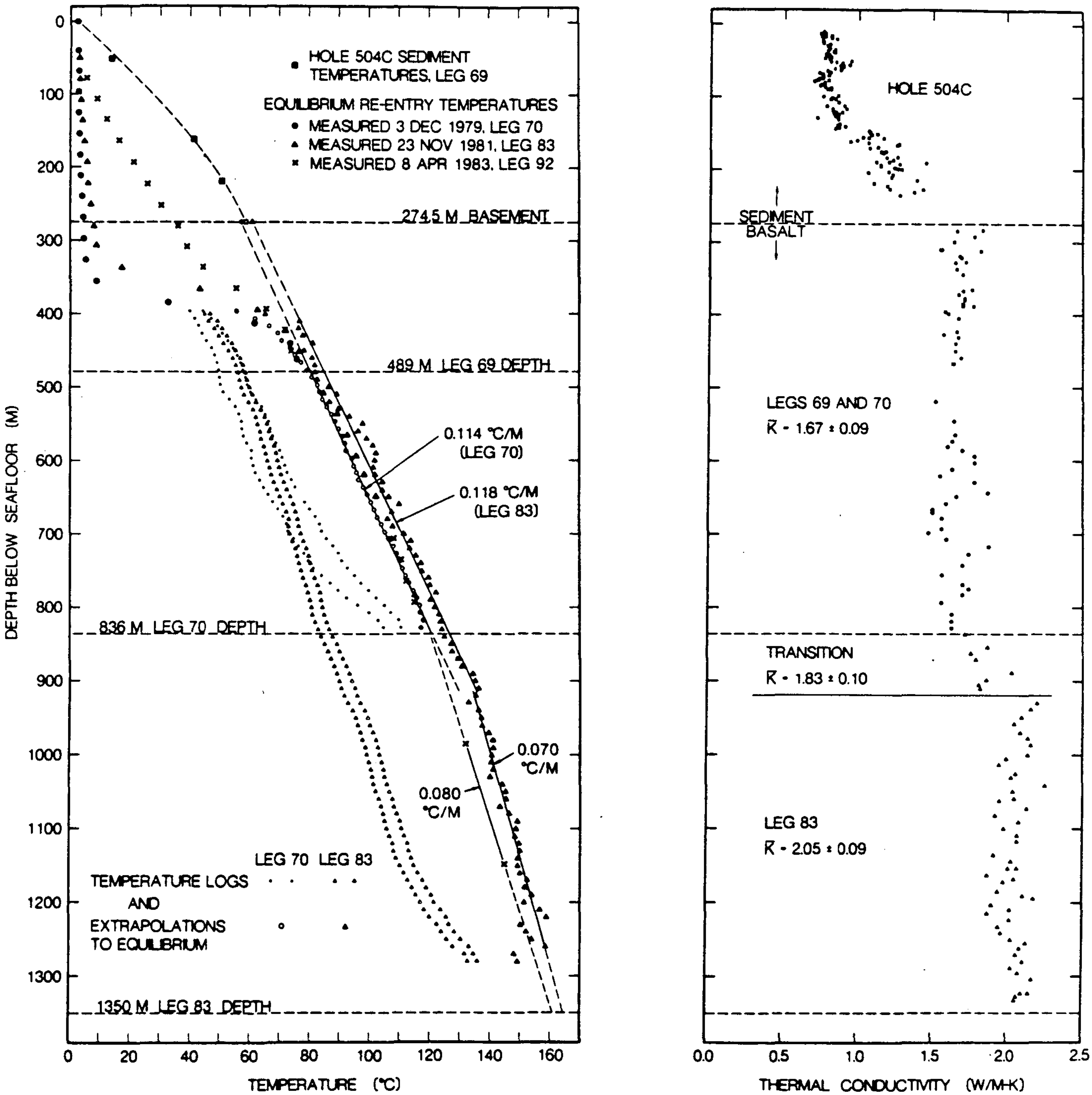

Figure 4: Observed temperature profiles and thermal conductivities of recovered samples, DSDP Hole 504B. Equilibrium temperatures were measured in the sediments in Hole 504C offset about 60 m form Hole 504B.

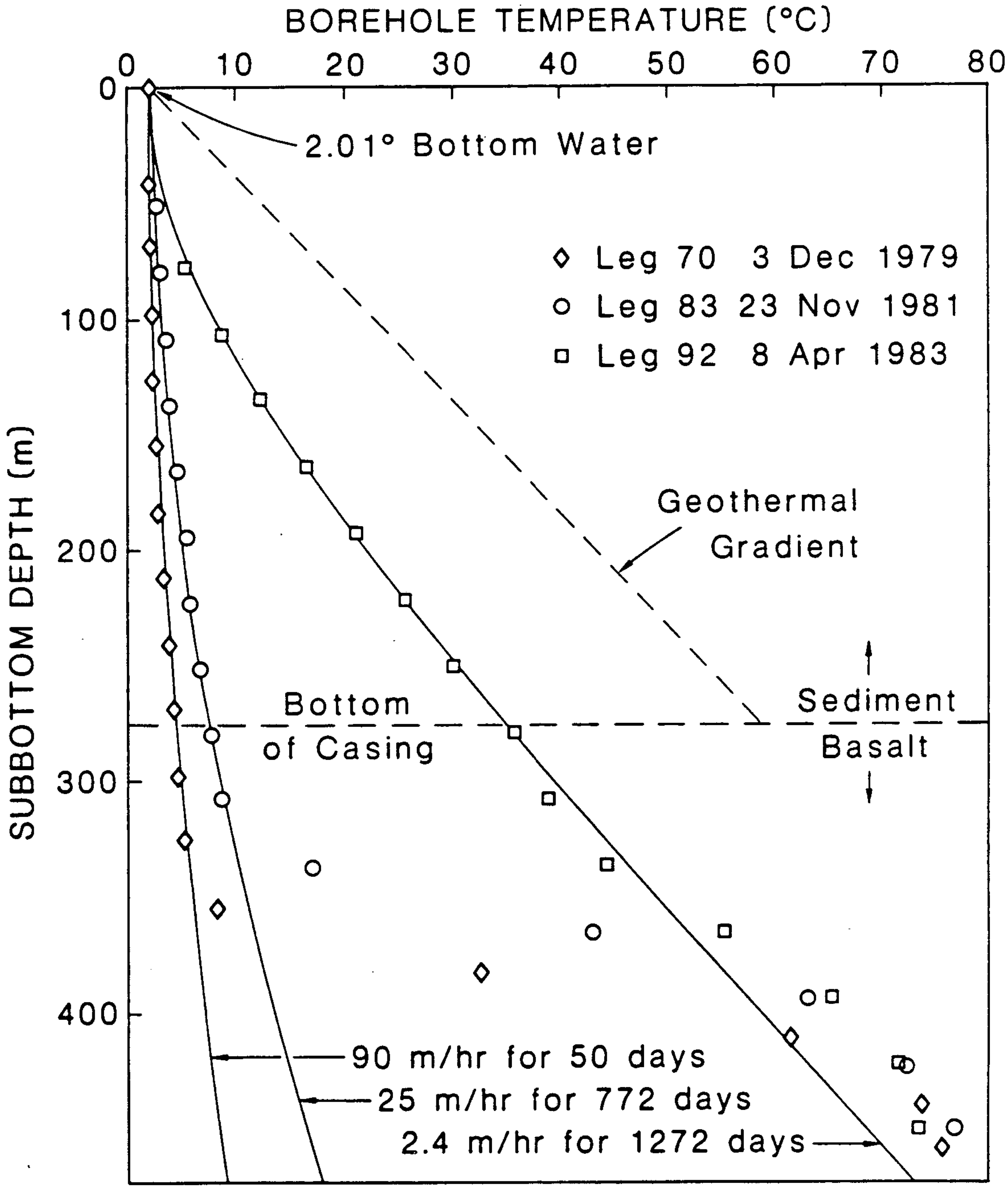

Figure 5. Equilibrium borehole temperatures measured in Hole 504B (symbols) and temperature profiles (curved lines) predicted with a constant flow rate model (Becker et al., 1983).

The Mathematical Model

The governing equations employed in this model are the mass and energy balance laws for a liquid-saturated, non-isothermal porous medium. For any mass flow equation can be written as

$$\frac{\partial}{\partial t} \int_V \varphi \rho \, dV = - \int_A \rho \, \vec{V}_d \cdot \vec{n} \, dA + \int_V G_f \, dV \tag{2}$$

and the integral form of the energy equation as

$$\frac{\partial}{\partial t} \int_V \rho e_f \, dV = \int_A \lambda \vec{\nabla} T . \vec{n} \, dA - \int_A \rho c_f \, \delta T \, \vec{V}_d . \vec{n} \, dA + \int_V G_n dV \tag{3}$$

where ρ is the density, ϕ is the porosity, λ is the thermal conductivity of the material, c_f is the heat capacity of the fluid, e_f is the internal energy of the fluid, δT is the interface temperature, G_f is a mass source or sink, G_n is a heat source or sink, $\vec{n}$ is an outward pointing unit vector, and t is time.

In both equations the left-hand side represents an accumulation term which is balanced on the right-hand side by a combination of flux and source terms. The fluid fluxes are calculated using Darcy's Law,

$$\vec{V}_d = - \frac{k}{\mu} (\vec{\nabla} p - \rho \vec{g}) \tag{4}$$

where k is the absolute permeability, μ is the dynamic viscosity, and $\vec{g}$ is acceleration due to gravity.

The model uses the Integrated Finite Difference Method (IFDM) (Narasimhan and Witherspoon, 1976) to evaluate the integrals in (1) and (2) by discretizing the space and time domains. The mesh generated for the IFDM scheme is used in the computer program PT (pressure-temperature), in which the fully coupled governing equations are solved simultaneously with an efficient sparse solver (Bodvarsson, 1981).

The Three-Dimensional Model

For the perceived heat flow situation near the Costa Rica Rift, the following symmetry considerations were used in constructing the nodal mesh. First, we assumed that the areas of high heat flow near 504B could be idealized and arranged in a regular square grid such that a small symmetrical section would provide sufficient information for the analysis of the entire flow field (Figure 6). Second, from observations of the spacings of peak heat flow values, we estimated that the approximate horizontal dimension of the convection cells was 4,000 m (Hobart et. al., 1984) and thus scaled the mesh accordingly. Third, we assumed that the variations of basal heat flow across the bottom of the cell was too small to interfere with the symmetry arguments mentioned above. We later tested both the scaling and basal heat flow hypotheses through expansion of the horizontal dimensions of the mesh.

The mesh used in the simulation is shown in Figure 7. The dimensions were 4000 meters by 4000 meters for the horizontal directions and 1300 meters for the vertical. The choice of the vertical dimension was based on the downhole permeability and porosity measurements (Figure 3), which indicated the decrease of porosity to a negligible level beyond the 1300 meter depth. The mesh consisted of 234 block-centered volume elements and 546 interfaces between the elements. Distances between nodal points varied from 83 meters for the closest to 1008 meters for the farthest, while element volumes ranged from 3.9×10^5 to 9.3×10^7 m^3.

For the material properties employed in the mesh we took the measurements made in the study of Hole 504B (Figures 3 and 4) and incorporated them into the four layers shown in Fig. 7. A low permeability chert/sediment layer occupied the first 300 meters of the mesh, with three different types

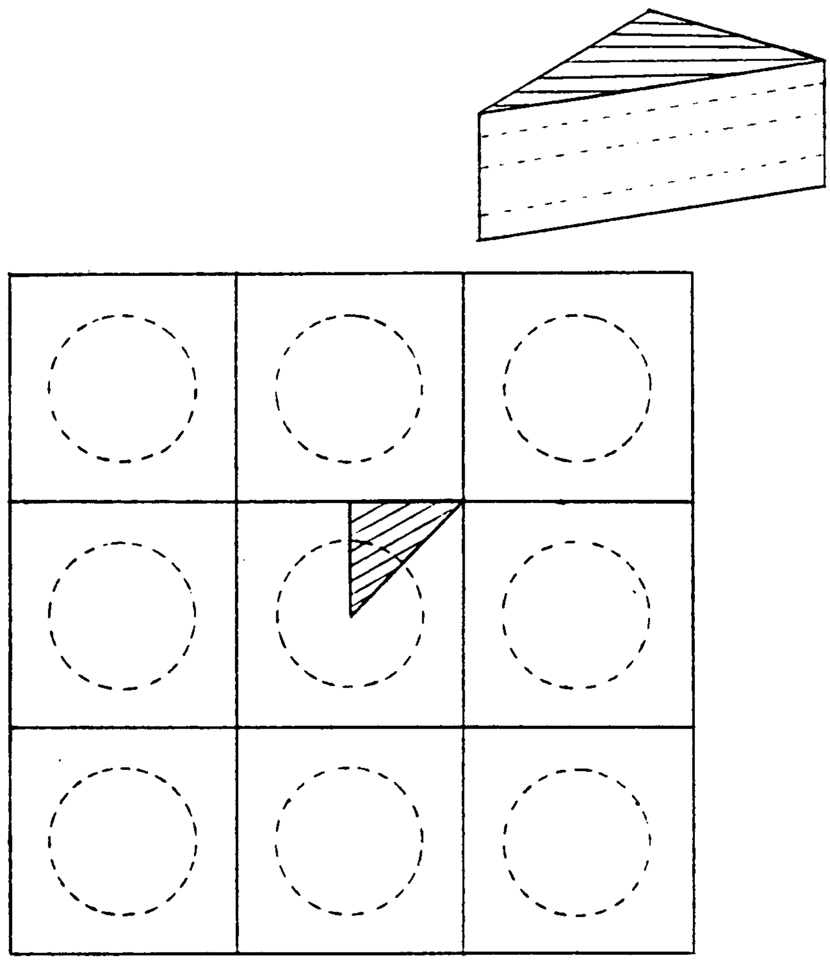

Figure 6: Idealized plan view of periodic heat flow highs observed on the ocean floor.

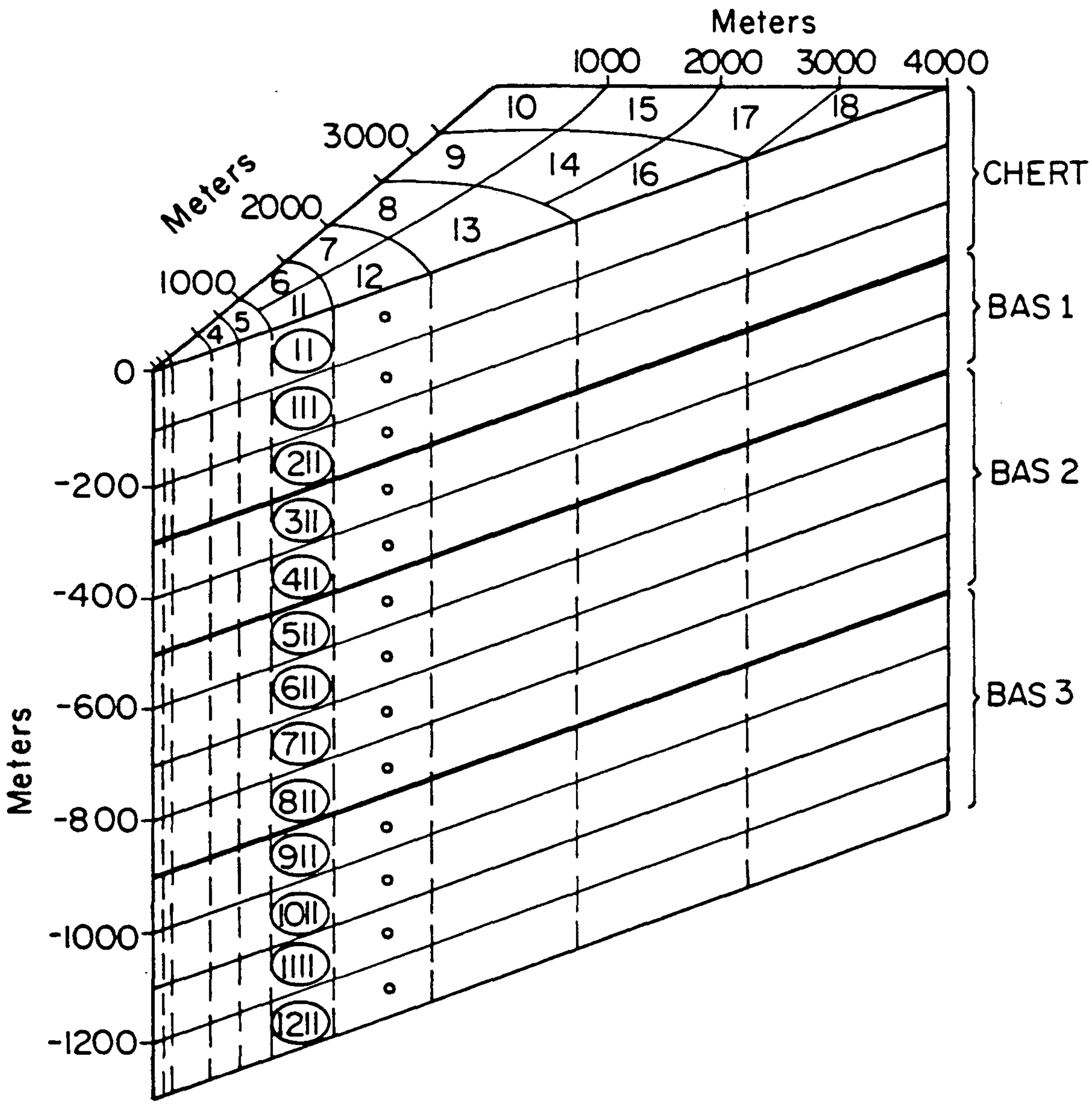

Figure 7: Three dimensional nodal mesh used for the mathematical model.

of basaltic rock filling out the remaining 1000 meters. The permeabilities, porosities, and thermal conductivities of these rocks are presented in Table 1.

The flow region was subjected to the following boundary conditions. The vertical boundaries were taken to be impermeable and adiabatic (streamlines of the flow field), while the top boundary was specified as a chert/ocean interface, with the ocean temperature set at a constant 0°C with pressure set constant at the ocean bottom value. For the bottom face of the mesh, a constant heat flux, varying from 215 mW/m^2 at the first node to 209 mW/m^2 at the farthest, was maintained in order to simulate the basal heat flow present near Hole 504B. The initial pressure distribution in the mesh was hydrostatic with the initial fluid temperature profile corresponding to that established by the basal heat flow when conduction was the only mode of heat transfer. The basic simulation run set a permeability for the chert layer and then ran the calculations out until the system reached equilibrium, usually after approximately 1 million years of simulation time, with time steps starting at 0.1 seconds and increasing up to 30,000 years.

SIMULATION OF FIELD OBSERVATIONS

The thrust of the calculations was to simulate the observed temperature profiles, fluid pressure profiles as well the heat flux at the ocean bottom using the measured systems parameters. Also, the calculations were aimed at investigating the sensitivity of the afore mentioned profiles to the distribution of permeability within the system. In addition, a second part of the calculations addressed the question of the interaction between the ocean and the convecting cell through Hole 504B. The different cases studied are as follows. In all the following cases, the measured values of thermal conductivity (Table 1) were used.

(i) Purely conductive regime: Permeabilities of the chert as well the basalt layers set everywhere to zero

(ii) Measured values: Chert permeability set to zero and the permeabilities of the basalt layer set to observed values (Table 1)

(iii) Permeable chert: Same as (ii) but with a finite chert permeability of 0.1 md

(IV) Permeable Basalt: Chert permeability set to zero but with the permeability of the basaltic layers increased by a factor of two to five

(V) Stretched flow region: Same as (ii) but with the horizontal dimensions of the systems enhanced by a factor of two.

The purely conductive case not only provided the initial temperature distribution for all the other runs, but also provided a base for heat flux to which results from all the other runs could be compared.

The relation between depth and underpressuring for the various cases is shown in Fig. 8. In this figure the values of underpressure denote the departure of the computed fluid pressure from the expected value of the hydrostat at the depth of interest. Note from the figure that the actually observed undepressure value at a depth of 450 m agrees closely with the computed profile for the Measured values case. Also, assigning even a small value of permeability to the chert layer dramatically suppresses underpressuring due to the establishment of a noticeable hydraulic connection between the ocean water and the convecting cell. On the contrary, increasing the permeability of the basalt layers (Case 3, Table 2) causes a more pronounced underpressuring. It is thus evident from Fig. 8 that an impermeable chert cap is an essential requirement for the existence of underpressuring in the oceanic crust. Also, convective motion appears to accentuate the magnitude of underpressuring. Detailed examination of the computed results showed that except in the most permeable case (Case 3, Table 2), the horizontal gradients in temperature as well as fluid pressure throughout the system were small.

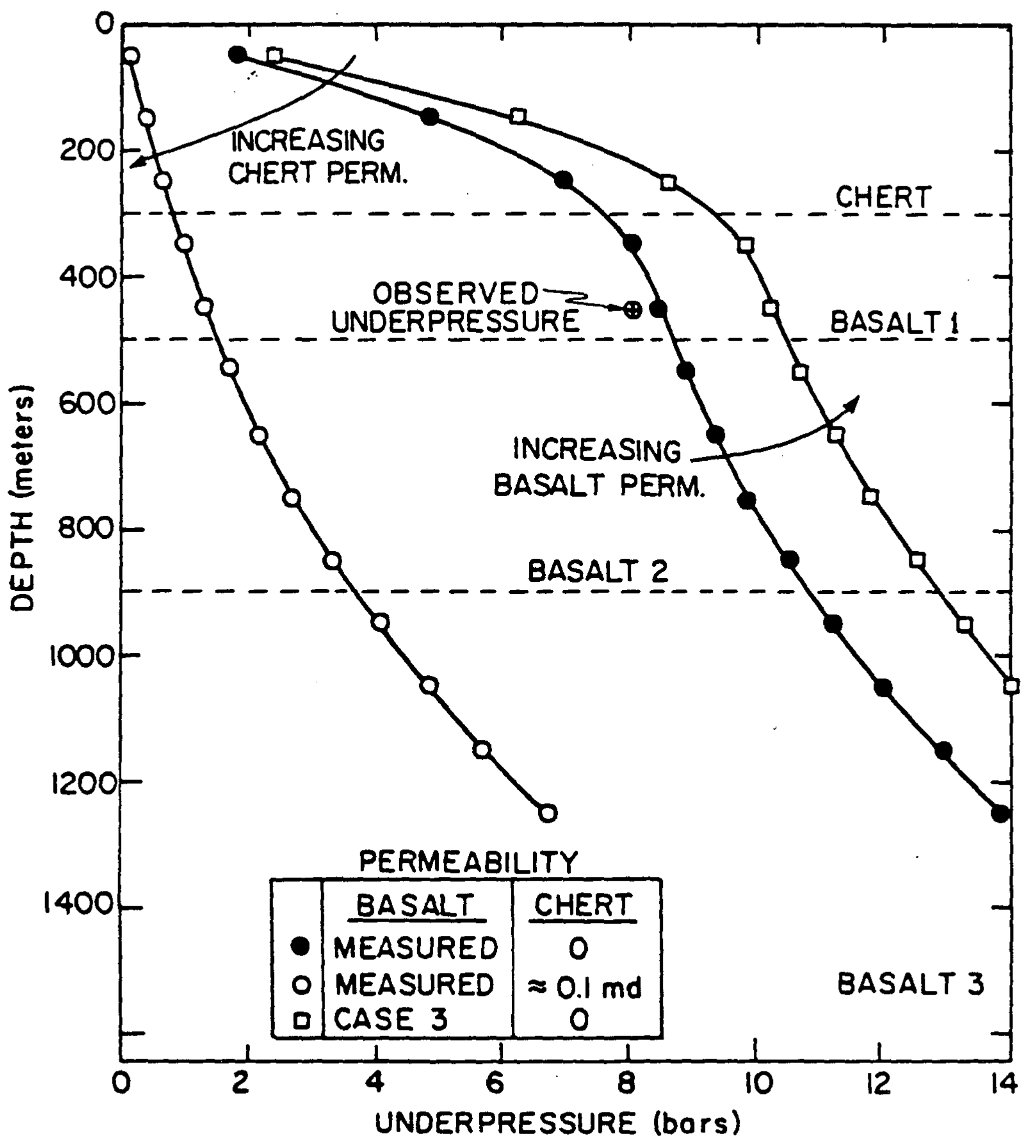

Figure 8: System underpressure as a function of depth for various permeability variations. Permeable caprock attenuates underpressuring while permeable basalt accentuates it.

These results clearly show that hole 504B could, in principle, be anywhere in the convecting system, and that the observations made on this hole do not necessarily suggest a down going limb of a convection cell as was originally speculated by Anderson and Zoback (1982).

A comparison of the observed temperature profile with the simulated values for the Measured Values case is given in Fig. 9. As can be seen, the agreement is good. In as much as the temperature gradients in the horizontal direction within the system are very small, the simulated profile shown in Fig. 9 would be more or less characteristic of any vertical line within the system.

Plots of the surface heat flux at the ocean bottom for the various cases are shown compared in Fig. 10. Note that the computed heat fluxes using measured values closely approximate the pure convection case which agrees with the heat flow measurements near Hole 504B (Fig.1) and with Hobart et al., (1984). It can be seen that the surface heat flux pattern tends to become more and more peaked as the permeability of the basalt layer is increased (Case 1, 2 and 3, Fig. 10). The range of heat flux variations seen in Fig. 10 is consistent with the extreme heat flux variations observed away from the ridge axis (Fig.1). Thus, one possible explanation for the variation in heat flux away from the ridge axis is that the effective formation permeability varies across the ridge flank, yielding widely divergent convective heat flow patterns.

The velocity distributions for the Measured Values Case and for Case 3 and shown compared in Figs. 11a and 11b. As can be seen, the single large convection cell of the Measured Values Case breaks up into two cells as the magnitude of convection increases. It was found that for the Measured Values Case, the fluid velocities (Darcy velocities) varied from 2.5 cm/year within the upper most basalt to less than 0.25 cm/year elsewhere.

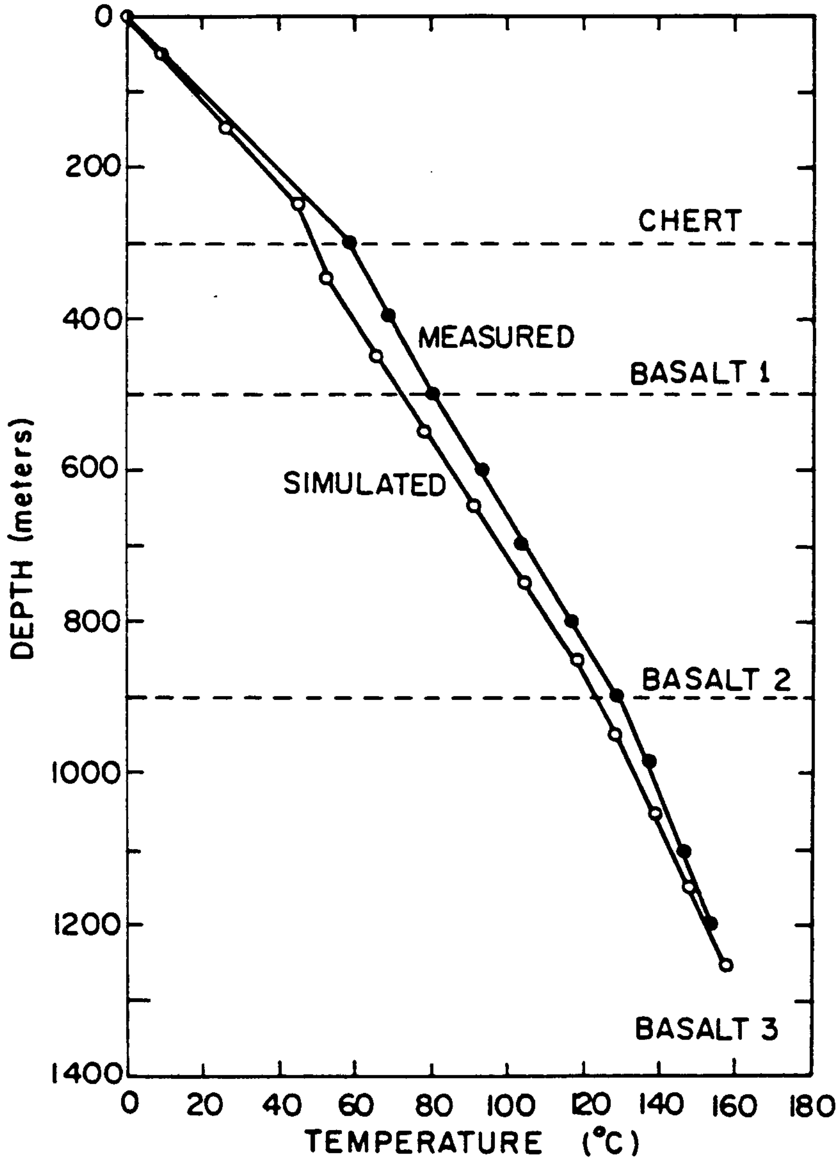

Figure 9: Comparison of down hole temperature profiles for simulated and measured equilibrium conditions.

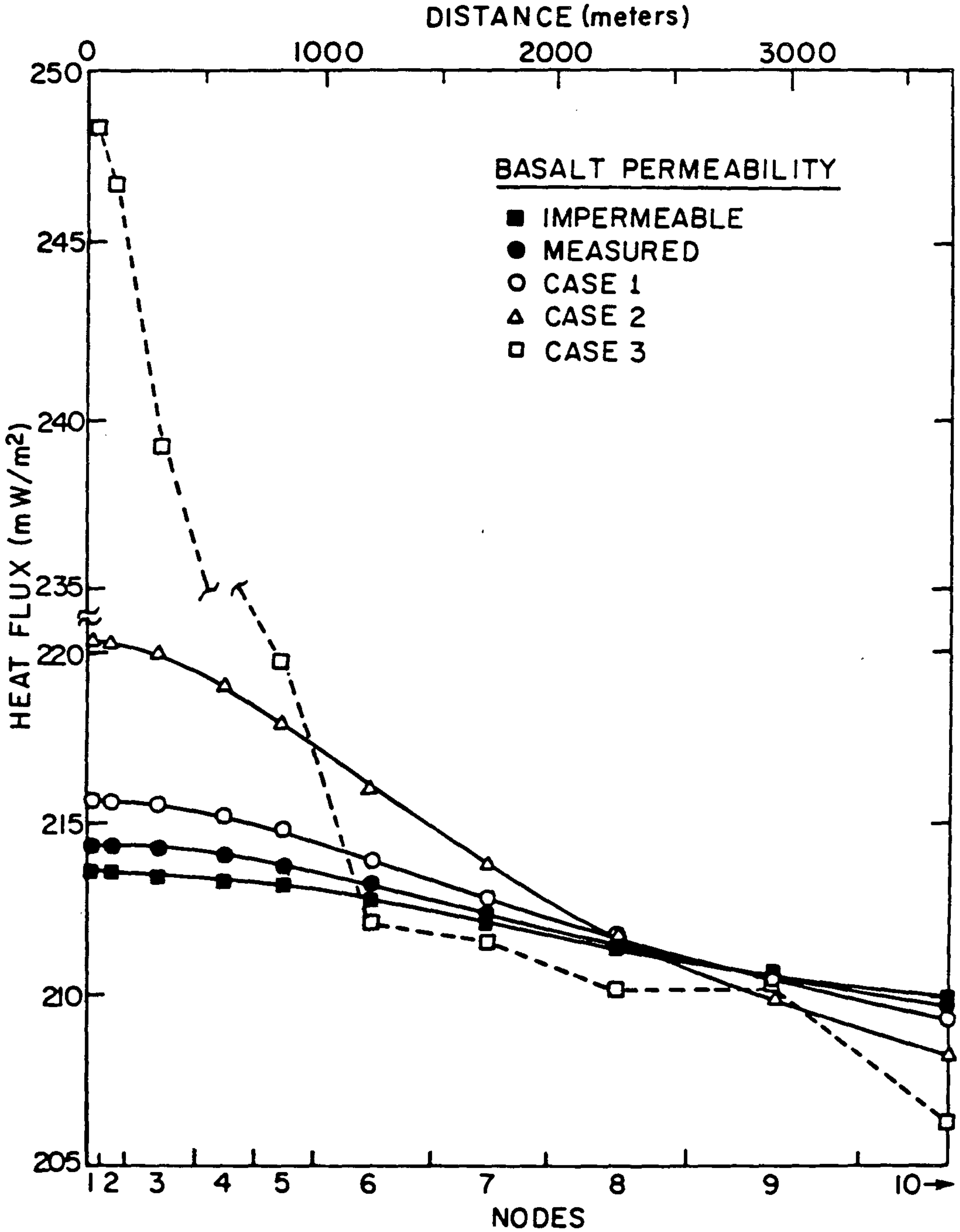

Figure 10: Plot of surface heat flux for five permeability variations.

As mentioned previously, scaling considerations used in constructing the nodal mesh were based on assumptions concerning the geometry of the convection cells simulated by the model. In order to test these assumptions, we gradually expanded the size of the mesh, looking for evidence of the point at which a second convection cell would appear and the exact form it would take, either symmetrical or assymmetrical. With the nodal mesh expanded to twice the horizontal dimensions of the original mesh, a second convection cell limb had formed, starting at approximately 3800 meters from the first nodal point and extending through the rest of the mesh. Thus, the bottom heat flux variation did not seem to affect the symmetry of the system to any unreasonable degree, and the estimate of 4000 meters for the horizontal dimension seems to be approximately correct.

WELL BORE INTERACTION STUDIES

As a final step in the investigations, we examined the possible role that the well bore itself may play in perturbing the pressure and temperature observations of the natural convecting system. Becker et. al. (1983a, 1983b, 1984), in performing heat flow measurements at Hole 504B, logged down hole temperatures for all three drilling legs, both at times when drilling disturbances were present and when equilibrium had been regained. They observed that the presence of a layer of high permeability basalt in the uppermost 100 m of basement could be inferred from a nearly isothermal region that appeared in the temperature logs, indicating substantial flow of ocean bottom water down the hole and into this layer. In order to model this well bore effect we refined the nodes at the origin of the mesh with dimensions appropriate to a penetrating well. The permeability of the well bore nodes varied according to the particular drilling leg we wished to model. Temperatures present at the early stages of well bore presence were then compared to the measured disturbance values.

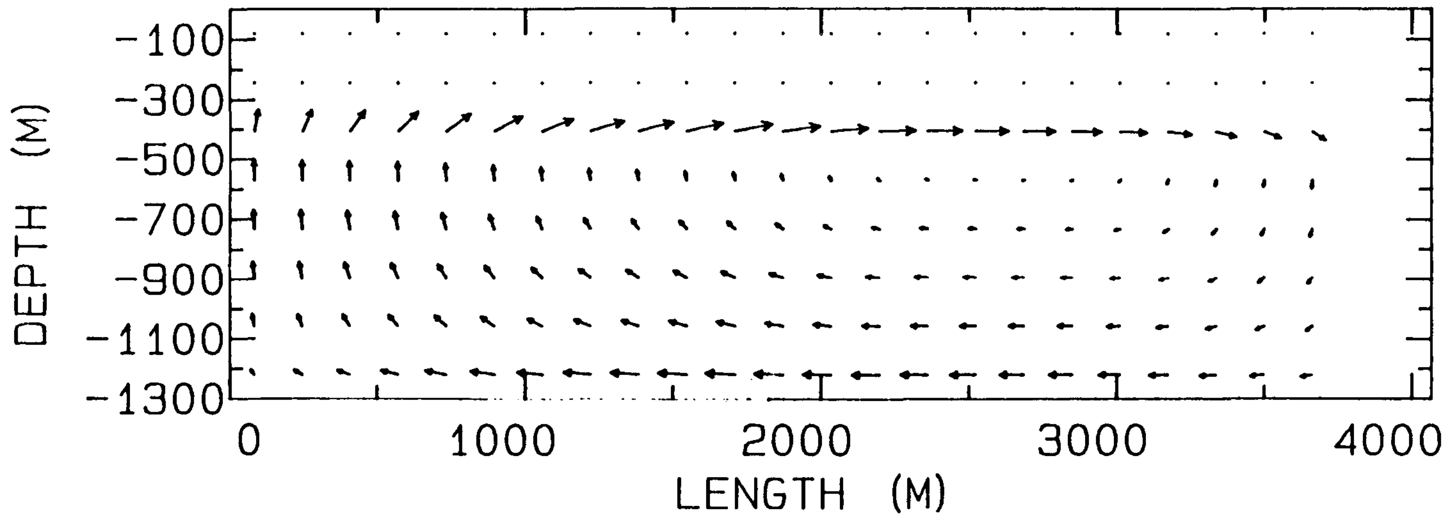

Figure 11a: Flow field plot for simulation case based on system parameters.

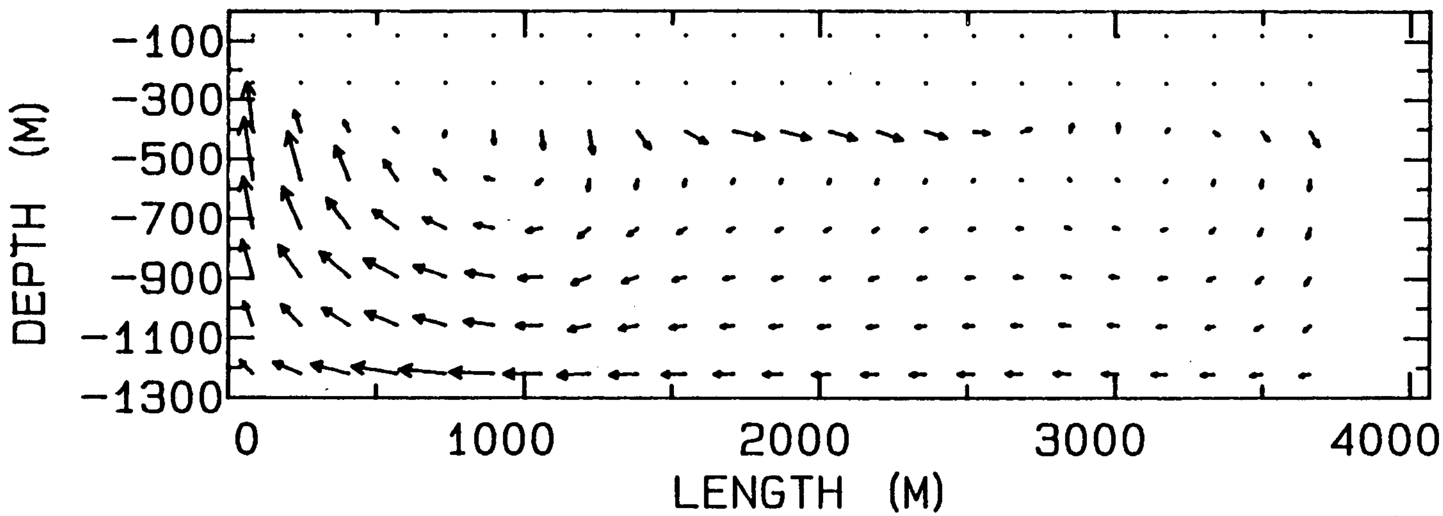

Figure 11b: Flow field plot for simulation Case 3.

The results obtained through this modeling approach were plotted along with measured disturbance values for Leg 69 (489 meter penetration), Leg 70 (836 meter penetration), and Leg 83 (1350 meter penetration) (Figures 12, 13, and 14). The plots obtained agree quite well in general shape, but the isothermal section goes to a greater depth in the simulation plots than the temperature observations indicate. This suggests that the basalt layer of high permeability is not as thick as indicated by the downhole permeability measurements and as modeled here, but is, instead, about 100 m-thick as indicated by the downhole temperature and porosity measurements. The misplacement of this layer does not seem to have significantly affected the results presented in Figs. 8-10, however.

Becker et. al. (1983a, 1983b, 1984) used the temperature profiles obtained to estimate the downhole flow rate as it changed with time. At the beginning of Leg 70; they estimated an initial flow rate of between 6000 and 7000 l/hr (90 m/hr). Two years later, just before the drilling of Leg 83, they fould a flow rate of approximately 1700 l/hr (25 m/hr) and, one and one half years after that, at the drilling of Leg 92, a final flow rate of 170 l/hr (2.4 m/hr) was measured (Figure 5). These measured rates were compared with values calculated by our mathematical model (Figure 15), and there was fair agreement between rates at which the downhole flow slowed, although the simulated flow rate remained at a slightly higher value than the measured flow rate. This could be due to the fact that a 200 m thick highly permeabile basalt layer was used in the model, while the actual thickness may be less (≈100 m) as suggested by the temperature logs.

An important application for our model will be its predictive capability considering the wide array of hydrogeological experiments planned for the next ten years of the Ocean Drilling Project. Investigation of the crustal hydrothermal convection system is one of the major priorities of that

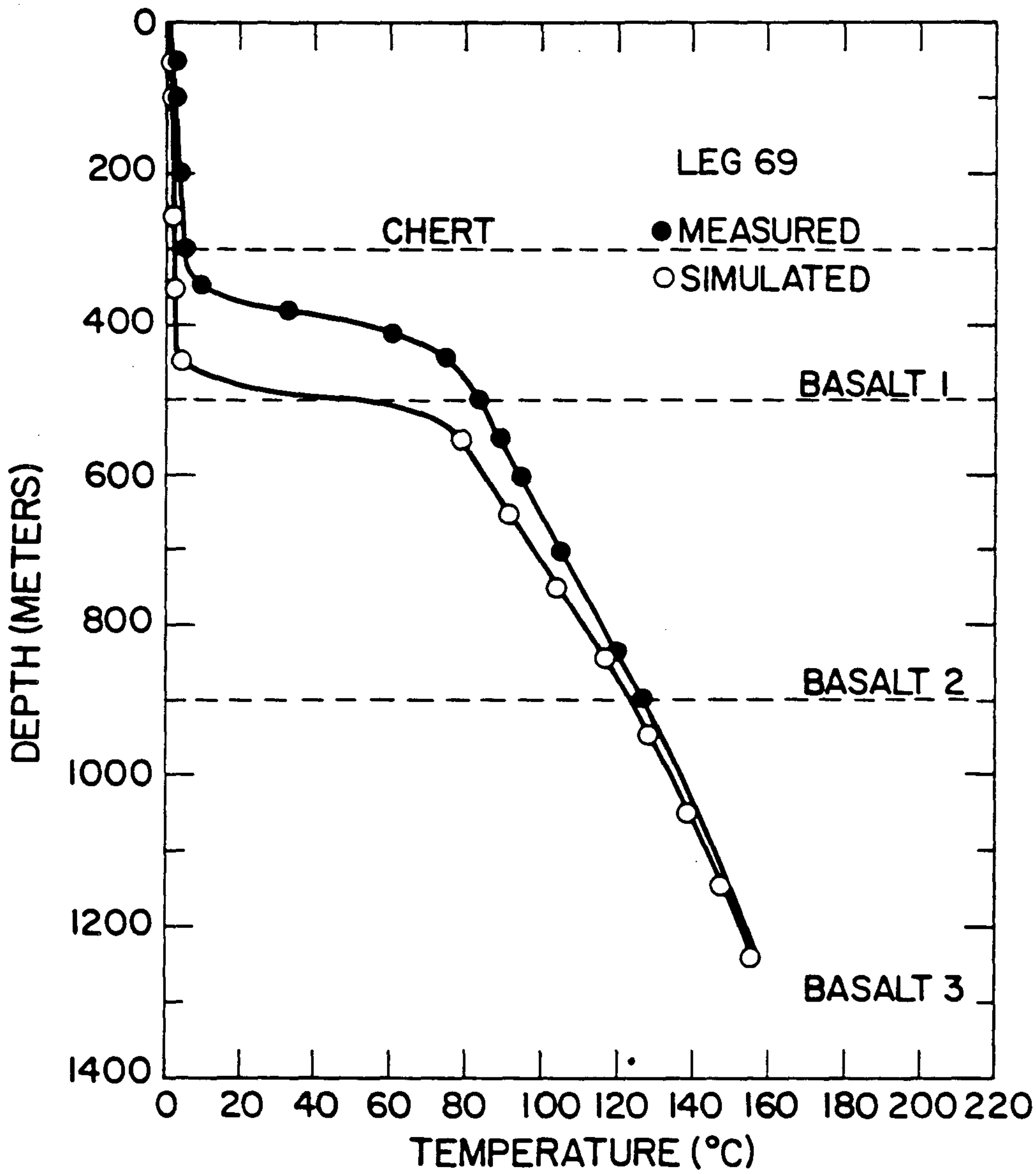

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Figure 12: Measured and simulated disturbance temperature profiles for Leg 69, the first drilling program at 504B.

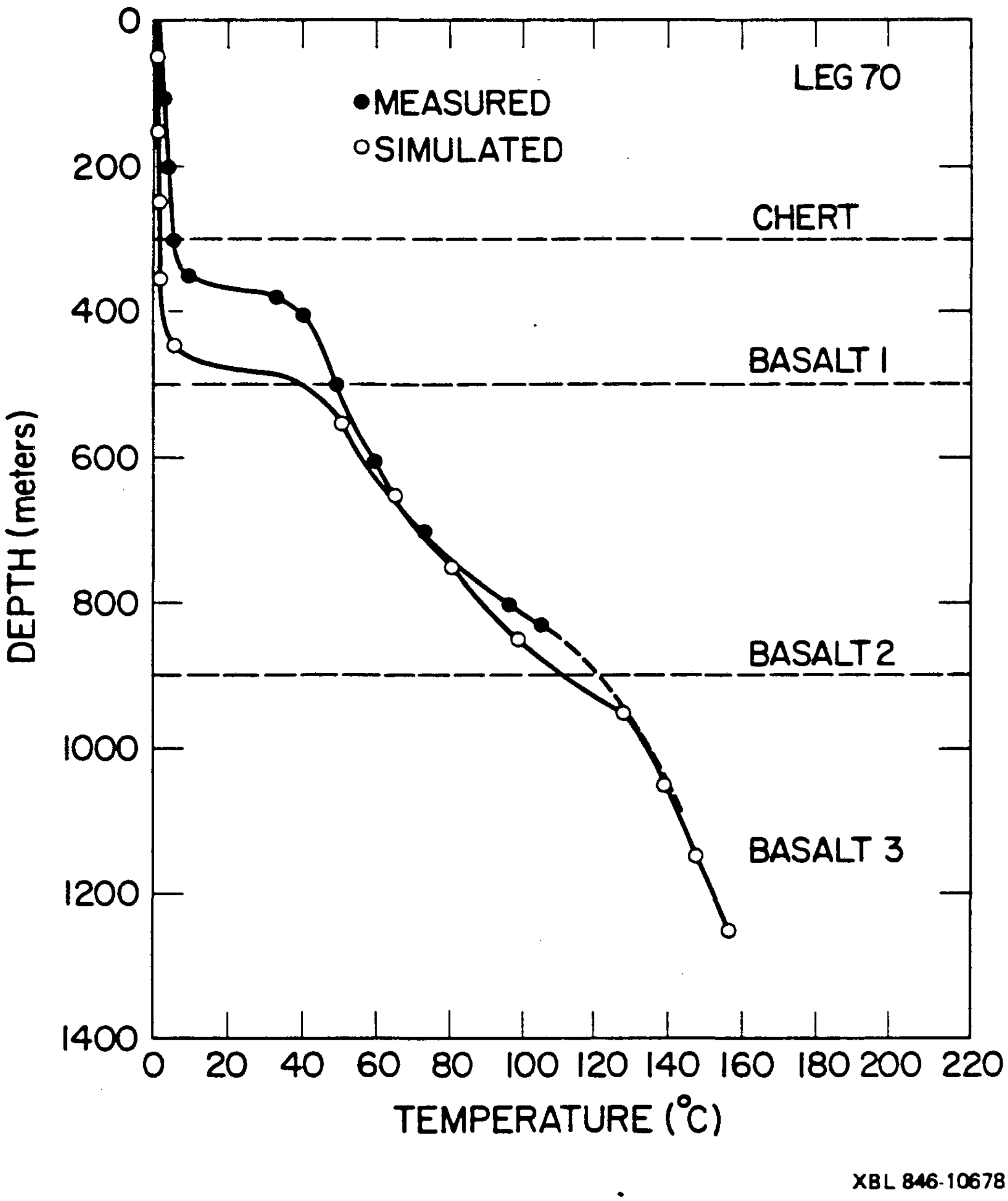

Figure 13: Measured and simulated disturbance temperature profiles for Leg 70, the second drilling program at 504B.

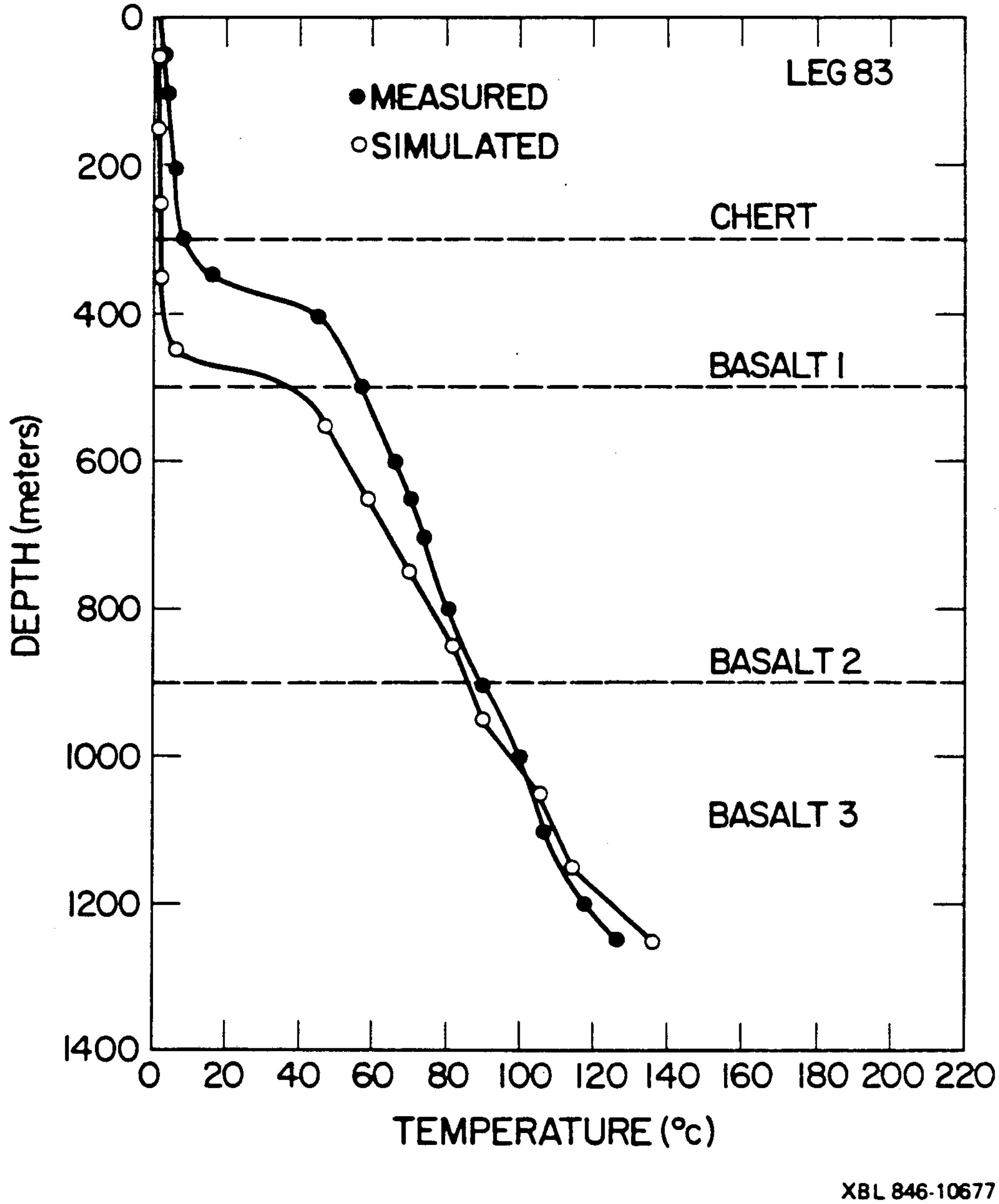

Figure 14: Measured and simulated disturbance temperature profiles for Leg 83, the third drilling program at 504B.

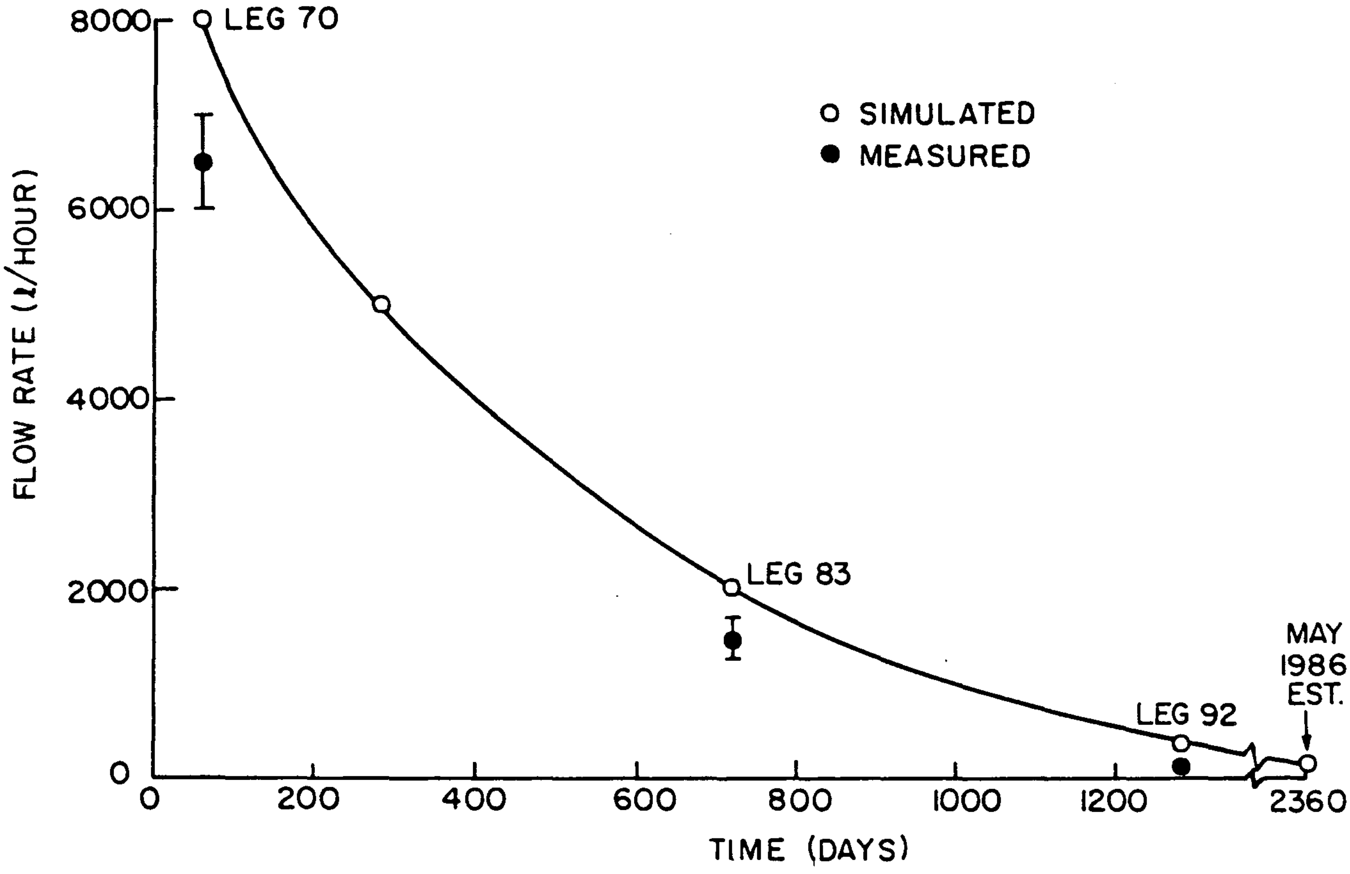

Figure 15: Inflow of ocean-bottom water into the basalt aquifer: comparison of computed results with the estimates of Becker et al. (1984)

program. As an example, consider the fact that the Ocean Drilling Project plans to return to Hole 504B in about May, 1986, some 2400 days after the initial drilling program. According to the numerical model, the downhole flow rate at this time should be approximately 50 to 60 l/hr. Whether the prediction is credible or not remains to be seen.

At this point it is worth interjecting a note of caution against overly adjusting the system parameters to obtain very refined matching of the field observations. Considering the remoteness of the system and the difficulties inherent in making even rudimentary physical measurements within the system, it is encouraging that the use of measured parameters has led to results that agree closely with field observations in respect of fluid pressure, temperature, fluid flow and heat flow. Thus the merit of the present exercise is that it has helped is our obtaining a good understanding of the physics of the hydrothermal system in the vicinity of DSDP Hole 504B.

CONCLUSION

The model for convection in a porous medium presented here satisfactorally reproduces all the hydrodynamic observations from DSDP Hole 504B. Specifically, the permeability versus depth function measured in the hole, coupled with an impermeable chert cap produces underpressures within the entire convection cell in the oceanic crust. The values of underpressures calculated by the model are the same as those measured in the 504B borehole. Removal of the cap results in loss of the underpressures. The natural evolution of a ridge axis hydrothermal system through its sea floor spreading history will carry the oceanic crust from an open system at the ridge crest to an insulated, capped system on the flank. The same kind of physical system that produces the spectacular "black smoker" hot springs on the axis will produce the underpressured, convection system observed on the Costa Rica Rift if the sediments form a sufficiently impermeable cap on the basalt crust.

Underpressures may not be observed universally on ridge flanks, however, as the existence of chert or some other low permeability seal above the crust plays a crucial role. Though the chert found at Hole 504B rests on the youngest ocean crust yet found, it is far from anomalous. Virtually every western Pacific plate drill hole has encountered chert. Chert is also common in old Atlantic holes, and is characteristic of the basal sedimentary layer of ophiolites found on land (Anderson et al., 1982). The requirement that sufficiently well capped convection cells develop underpressures throughout their volume may explain why so many DSDP drillholes have shown evidence for drawdown (Becker et al., 1983b, 1984).

The capped convection cells established by the model also explain the the anomalous heat flow pattern observed on the sea floor near crustal spreading centers. Apparently, a number of convection cells develop in the new crust material, and the values of permeability and porosity present in each cell dictate the amplitude of the periodic heat flow highs. Thus, the presence of hydrothermal circulation in the oceanic crust brings together all of the various field observations in a single, realistic model. Another important result of our modeling study is the realization that nearly conductive temperatures toward the bottom of a cell are the natural result of exponentially decreasing permeability with depth. While slow convection does occur within these zones, the most vigorous flow is confined to the upper oceanic crust. It is important to determine in the next few years exactly how representative the permeability versus depth profile is at Hole 504B. The only other hole in the oceans in which permeability measurements were made (Hole 395A in the Atlantic) also had low permeability 500 meters into the crust (Hickman et al., 1984). Temperature profiles also indicated a zone of high permeability in the uppermost crust at Hole 395A as well and flow rates predicted in theat layer were calculated to be upto 1m/year by Langseth et al. (1984, Becker et al., 1984).

Concerning the continued decrease in flow rate with time since drilling was completed at Hole 504B, our model clearly shows that the crustal convection cells have been disrupted by the intrusion of cold ocean bottom water through a hole in the impermeable cap. This accounts for the decrease in flow rate downward into the crust with time. The model can make specific predictions of the rate of flow to be encountered upon reoccupation of the site in the near future. Such verification of the model is extremely important, for it implies that no regional bounds to the reservoir need exist in order to account for the drop in downhole flow rate (drawdown) with time.

ACKNOWLEDGMENTS

We would like to thank Sally Benson and Marcelo Lippman for constructive criticisms of the manuscript. This work was partly supported by the Director, Office of Energy Research, Office of Basic Energy Sciences, Division of Engineering, Mathematics and Geosciences, U.S. Department of Energy under Contract No. DE-AC03-76SF00098.

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