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GEOTHERMAL PROCESSES
at the
GALAPAGOS SPREADING CENTER
by
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BIOGRAPHICAL SKETCH

I came into the world in the usual way in Conway, Arkansas, on June 28, 1951, the second of three boys born in my family. My education proceeded in parochial schools through the sixth grade, thence to public junior high and high school in Shawnee Mission, Kansas. The American Field Service provided me with a year in the Gymnasium Buergdorf, Bern, Switzerland, which greatly broadened my appreciation of the finer things in life. My undergraduate years were spent at Massachusetts Institute of Technology where Drs. Aviva Brecher and John Sclater exposed me to the variety of research open in the field of paleomagic. I received the Bachelor of Science in Geophysics in June, 1974, and immediately entered the Joint Program in Oceanography of M.I.T./W.H.O.I. During the first four years of my tenure at W.H.O.I. I was priveleged to spend eight months at sea, almost all of this seafaring was aboard the R/V KNORR with Captain Emerson Hiller and crew. I am especially indebted to the skilled assistance of the captain and crew for the success of the data gathering portion of this thesis.

ACKNOWLEDGMENT

This thesis is one product of a multi-institutional investigation of the Galapagos Spreading Center, and consists in part of papers co-authored by scientists from many institutions. I am especially grateful to Dave Williams for his observations and ideas which form the heart of Chapter II. The work reported here benefited greatly from the constant encouragement of R.P. Von Herzen, whose experience in oceanic heat flow measurement and interpretation provided much needed guidance. The ideas I have developed here originated in discussions with Dick Von Herzen and Dave Williams. Whereever I have lead these ideas astray, any errors in the following, are mine alone. The officers and crew of the KNORR and LULU cheerfully assisted in the station work and provided me with a great respect for their work. The support staff at W.H.O.I. has been very cooperative, especially the ACNAV group and Chris Wooding, for endless computer assistance. Paul Murray, Skip Pelletier and Jim Akens have provided me with constantly improving instrumentation, and instruction on much of Life in general. The other graduate students have enriched my education greatly. John Crowe has shared much of the work involved here. Sandy Pelletier was a great help in the final flail.

Special thanks are necessary for Nancy and Sandy Shor, and Caleb, who were always willing to take in a poor student. The Almeida and Hudson bus lines made this a real trip.

This work is dedicated to my lonely, loving wife, Beth, whose long-suffering devotion supported me through it all.

Preface

Dear Reader,

The following chapters deal with my investigations over the past five and one half years of the thermal history of the Galapagos Spreading Center. The first two chapters have been written to stand alone in publication and hence some repetition of background material occurs. At this point I will outline the general thrust of the thesis and major conclusions. Please bear in mind that in many cases I refer to the oceanic ridge system as if the GSC were representative of processes at all ridges. This is patently incorrect, but because it becomes boring to continually qualify my remarks to apply to only the detailed study area within which I have sufficient data to draw conclusions, I simply ask the reader to mentally add the necessary qualifiers.

The objectives of my thesis were firstly the characterization of the thermal environment at the GSC in sufficient areal detail to verify or deny the existence of spatially coherent heat flow variations, and secondly the modeling of these variations as the observable effects of hydrothermal circulation. To achieve the first objective, I assisted in the development of a digitally recording heat flow instrument with extended capabilities over previous analog instrumentation. This resulted in a significant improvement in the quality of the data currently obtained at WHOI, LDGO, and SIO (all of

whom now are sharing this instrument design, thanks to the efforts of R. Von Herzen at improving the world heat flow data base). However, the data obtained for the GSC were partially degraded by developmental "bugs in the system." Removal of those bugs has protracted the appearance in the literature of the data and this thesis. Herewith the thesis.

It is shown in Chapter I that the large variance in oceanic heat flow near spreading axes may well be the result of averaging a spatially smoothly varying heat flow distribution. This spatial variation appears correlated with the topographic elements of the GSC; consistently higher heat flow is found on the elevated sides of escarpments. Chapter II summarizes the extensive detailed observations of localized venting in the field of hydrothermal mounds found south of the axis. The speculations in this chapter preceded the numerical modeling experiments of Chapter III, and represent the views of several of my coworkers on the Galapagos Hydrothermal Project. In the true spirit of geological investigation, we failed to reach agreement on the current status or evolutionary history of the mounds. The mounds are the surface expression of the deeper hydrothermal system. The variety of opinion on their detailed internal structure does not contradict the basic observation, that seawater-derived fluids are venting from the oceanic crust more than 18 km from the axis of spreading. A significant hydrothermal circulation system is therefore required.

With the verification of a coherent heat flow field observed over the south flank of the GSC, I felt justified in comparing the results to thermally driven convection in a porous layer. Chapter III explains the physical system modeled, the boundary conditions applicable for the hydrothermal system of the oceanic crust near the spreading axis, and the results of varying the depth and permeability distribution of a porous medium analog to the oceanic crust. The thermal input from below to the hydrothermal layer was fixed by analogy with the conduction solution, and at the spreading axis we simulated continuous addition of hot material. Significant differences were found from the two previous studies of Fehn and Cathles (1979) and Ribando et al. (1973) which lead us to reinterpret our data in terms of a time-varying circulation system. The additional uncertainty this time dependence lends to the data, together with the fact that we are already simplifying the system considerably in our modeling, leaves the choice of a particular model of the upper oceanic crust open for further speculation. Specific implications of the parameter range investigated are given at the end of Chapter III.

ABSTRACT

The measurements of the heat flow field of the Galapagos Spreading Center in an area of about 570 km^2 reveal the planform of the conductive flux and permit the first truly areal estimate of the near-axis heat flux for comparison with theoretical plate cooling models. The intrusion process and associated hydrothermal circulation dominate the surface heat flow pattern, with circulation apparently continuing beyond the limits of our survey. The areal average of the conductive heat flux is $7.1 \pm .8 \text{ HFU}$ ($295 \pm 33 \text{ mW/m}^2$), about one-third the heat flux predicted by plate models. The remaining heat is apparently removed by venting of hydrothermal waters at the spreading axis and through basalt outcrops and hydrothermal mounds off-axis. The pattern of surface heat flux is lineated parallel to the axis and the strongly lineated topography. Sharp lateral gradients in heat flow, greater than 10 HFU/km near escarpments and commonly expressed as high heat flow at the tops of scarps and lower heat flow in the valleys, may indicate a local concentration of the circulation by surface fault systems and/or variable sediment thickness.

The mounds of the Galapagos Rift are spectacular hydrothermal features. Their internal temperatures have been measured at up to 13°C above the bottom water temperature and total heat flow (conducted plus convected) can be several hundred times the normal oceanic value. Fluids, when they discharge from the mound, do so at a very slow rate and at temperatures probably quite near the bottom water temperature. The mounds are principally composed of iron silicates intermixed and encrusted with lesser amounts of manganese

oxides. They are generally found in rows, in a uniformly sedimented area above faults or fractures in the crustal rocks which permit fluids to escape from a deep hydrothermal aquifer. The mounds field, covering an area of at least 200 square kilometers and consisting of thousands of individual mounds, is probably less than 300,000 years old; and many of the mounds may be only a few tens of thousands of years old or less.

Numerical modeling of two-dimensional convection within the oceanic crust is constrained by these observations of the detailed heat flow field. By comparing the estimated heat and mass flow rates with model flows, the maximum permeability-depth distribution is limited to no greater than .1 mD (10^{-12} cm^2) at depths greater than 2 km. The bulk permeability of the upper 2 km appears constrained to greater than 1.0 mD (10^{-11} cm^2); bulk permeabilities in excess of 5. mD are limited to the uppermost km.

One major consequence of the convection is the reduction of the overall temperature of the upper crust relative to conductive models. We found that convection with variable fluid properties responds to thermal forcing in a predictable fashion based on the dimensionless Rayleigh number evaluated at elevated temperature. Correlation of model fluxes and comparison to results for slow spreading rate models indicates a depth of fluid penetration shallower for the Galapagos than other ridges. The heat input at the ridge axis forces the wavelength of the resulting circulation to reflect the depth of strong circulation at the axis, and indicates 60 % or more of the "missing" heat flux near mid-ocean ridges is removed by hydrothermal upwelling at the axis of spreading.

CHAPTER I.

THE GALAPAGOS SPREADING CENTER AT 86°W:

A DETAILED GEOTHERMAL FIELD STUDY

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ABSTRACT

We report here measurements of the heat flow field of the Galapagos Spreading Center on crust of age less than 1.0 My. The 443 measurements in an area of about 570 km^2 reveal the planform of the conductive flux and permit the first truly areal estimate of the near-axis heat flux for comparison with theoretical plate cooling models. The intrusion process and associated hydrothermal circulation dominate the surface heat flow pattern, with circulation apparently continuing beyond the limits of our survey. The areal average of the conductive heat flux is $7.1 \pm .8 \text{ HFU}$ ($295 \pm 33 \text{ mW/m}^2$), about one-third the heat flux predicted by plate models. The remaining heat is apparently removed by venting of hydrothermal waters at the spreading axis and through basalt outcrops and hydrothermal mounds off-axis. The pattern of surface heat flux is lineated parallel to the axis and the strongly lineated topography. Sharp lateral gradients in heat flow, greater than 10 HFU/km near escarpments and commonly expressed as high heat flow at the tops of scarps and lower heat flow in the valleys, may indicate a local concentration of the circulation by surface fault systems and/or variable sediment thickness.

INTRODUCTION

The detailed mechanisms of heat transfer at mid-ocean ridges have been of interest since the general acceptance of plate tectonics and sea floor spreading. Elder (1965) realized that the heat transfer near the volcanic axis would be complicated by the

intermittency of the intrusion process and by hydrothermal circulation. The nature of circulation in the oceanic crust is not yet well understood, but many recent studies have emphasized the role of hydrothermal heat transfer in the tectonic evolution of plates (e.g. Anderson et al., 1977; Williams and Von Herzen, 1974), and the effects of exchanged hydrothermal fluids on oceanic chemistry (Wolery and Sleep, 1976; Corliss et al., 1979).

The reliable measurement of heat flow in near-axis environments has proved difficult due to rugged terrain, uneven sediment cover, and unstable shallow hydrographic environment (Talwani et al., 1971; Sclater et al, 1976; Davis and Lister, 1977; Williams et al., 1977). Oceanic heat flow measurements on crust of age less than 5 My are highly variable and are characterized by a wide variance as spreading axes are approached (Parsons and Sclater, 1977). These young crust heat-flow observations average well below the theoretical mean heat flow predicted by plate tectonic models but approach the theoretical value with increasing age, at different ages for each oceanic plate (Anderson et al., 1977). Lister (1972) suggested that the discrepancy between measured and predicted heat flow arose from the redistribution of heat flux by the mechanism of hydrothermal circulation. A limited number of measurements on the Juan de Fuca Ridge showed relatively higher heat flux on topographic highs, which Lister suggested indicated a "chimney-like" effect on the interstitial water forcing flow up within the basaltic crust and possibly out through thin sediments.

Williams et al. (1974) discovered a spatial heat-flow pattern in a survey of the Galapagos Spreading Center (GSC), varying from about 1 to over 15 HFU ($\text{HFU} = 10^{-6} \text{ cal/cm}^2/\text{sec} = 41.83 \text{ mW/m}^2$) on a profile normal to the axis over a distance of about 30 km. Several oscillations in a cyclic pattern of heat flow with distance from the axis were measured, with a rising mean value approaching their inferred theoretical heat flux at the maximum survey age of 0.9 My. The site was revisited in 1976 during the Pleiades expedition of the Scripps Institution of Oceanography, and in 1977 a major diving and surface ship program was conducted with R/V KNORR, R/V LULU, and DSR/V ALVIN submersible of the Woods Hole Oceanographic Institution (Corliss et al., 1979). We report here the extension of the single profile of heat flow measurements from Williams et al. (1974), to an areal survey comprising 443 surface ship measurements. Together with detailed studies from the deep submersible ALVIN of the hydrothermal mounds area (Williams et al., in press), these measurements enable us to make a much better estimate of the total heat flow from crust of less than 1 My age.

TECTONIC SETTING

The GSC is situated in the Panama Basin between the Cocos and Nazca plates, extending east from a triple junction with the East Pacific Rise at 2°N latitude, 102°W longitude to the Panama fracture zone at 82°W longitude (inset, Figure I-1). The tectonic evolution of this area has been described most recently by Lonsdale and Klitgord (1978) and by Hey (1977). Our survey area is located at

Figure I- 1. Detailed heat-flow survey area on the south flank of the Galapagos Spreading Center (inset shows the location in the Panama Basin). Volcanic axis shown with the associated hydrothermal vent areas, including those discovered on the most recent DSRV ALVIN, R/V GILLIS diving expedition in spring 1979 (K. Crane, personal communication). Heat-flow measurements are indicated by dots and contoured at 2 HFU in the interval 0-10 HFU, with additional contours at 14 and 22 HFU. Thirteen stations marked with open circles were omitted in contouring to emphasize the trend. Age (My) presuming uniform spreading at 35 Km/My indicated normal to axis. The area within the dashed outline was used in the average flux calculation shown in Figure 5 and Table 1. Documented mounds areas (Lonsdale, 1977; K. Crane, personal communication) are indicated by stippling pattern south of about 0.5 My age.

the eastern end of the high amplitude magnetic anomalies associated with the Galapagos "hot spot," (Trehu, 1975). The tectonic history of the past 20 My has been complicated by a number of ridge jumps (Hey, 1977). Surface ship magnetic tracks have been interpreted to show a southerly ridge jump near 86°W around anomaly 2' to 3 (3 My)(Anderson et al., 1976).

The detailed near-bottom Deep Tow magnetics show continuous, nearly symmetric spreading at the rate of 34.7 to 35.2 km/My over at least the last 2 My, with four magnetic reversal boundaries located symmetrically about the present axis (Klitgord and Mudie, 1974). Using these boundaries, Klitgord and Mudie inferred a maximum half-width of volcanic activity, i.e. the region within which volcanic intrusion and extrusion create the magnetic source layer, of between 2 and 4 km. The detailed chart from a multi-beam bathymetric survey of the area (Allmendinger and Riis, 1979) shows lineated topography on either side of the ridge crest, slowly varying for over 70 km along strike. Profiles based on this chart (Figure I-2) show the regular, block faulted pattern described by Klitgord and Mudie (1974), superimposed on the increasing depth with age. Between 5 and 35 km south of the axial high, there is a slow increase in depth across the area from east to west. The regularity in the sea-floor spreading process, the presence of uniform sediment cover and the relatively subdued topography create a favorable environment for a systematic and detailed geothermal study.

Figure I- 2. North-south profiles of bathymetry taken from the detailed narrow beam sonar chart of the survey area. Vertical exaggeration of 56X emphasizes the regular increase of depth with distance from the axis. Fine pattern, shallower than 1400 fm, coarse pattern, 1400-1450 fm.

OBSERVATION TECHNIQUE

The 443 surface ship measurements of heat flow listed in the appendix I-1 include 81 values previously published by Williams et al. (1974) and by Sclater and Klitgord (1973). Several reevaluations of location and depth of SOUTHTOW stations (Williams et al., 1974) resulted in minor changes to the previously published data. The new measurements were made on PLEIADES II and KNORR 64 during 1976 and 1977. The large number of measurements was possible in part due to the use of a self-contained, digitally recording heat-flow instrument, with telemetered data and extended battery life allowing stations to continue up to 18 hours (Green et al., in prep.). Most of the measurements were made during such multipenetration "pogo" stations, with the instrument and/or the ship acoustically navigated within bottom transponder nets. Less reliably navigated stations were made by radar-ranging on another surface ship which was acoustically navigating.

Sediment temperature gradients were measured with three outrigger thermistors mounted on a 2.5-m lance as described in Von Herzen and Anderson (1972). Extrapolation of the gradients to the bottom water temperature measured on the instrument 1 m above the base generally showed that the lower probe was about 2.8 m below the sediment water interface. We have corrected these penetration depths for measured tilts of up to 30° from vertical, and for partial penetration.

Thermal conductivity was measured on four gravity cores on PLEIADES II and two piston cores on KNORR 64 with the needle probe method of Von Herzen and Maxwell (1959). In contrast with cores further north (Sclater et al., 1974), there was no appreciable difference either down core or between cores, with a mean and standard deviation of $1.78 \pm .07$ TCU (TCU = 10^{-3} cal/cm sec $^{\circ}\text{C}$ = 0.418 W/m-K). Two piston cores reported by Williams et al. (1974) yielded thermal conductivities within this range, although they preferred a 5 % increase over the upper 4-m interval of sediment. Due to this uniformity in conductivity, and the small scale of the area, we have used the mean conductivity in converting the measured thermal gradients to heat flow. The uncertainty of a few percent is insignificant for the conclusions of this chapter. Cores raised within the hydrothermal mounds field sometimes recovered very coarse grained, rubbly material unsuited to the needle probe technique due to the coarse particle size (Chapter II; Williams, et al., in press). We measured thermal conductivity only on those portions of cores containing no hydrothermal phases.

PERTURBATIONS TO HEAT FLUX

Our survey area, shown in Figure I-1, is located entirely over crust of less than about 1.2 My. The sediment thickness increases from zero near the axis of spreading to more than 50 m at the southern boundary. Records from both the Deep Tow 4 kHz penetrating seismic system (Klitgord and Mudie, 1974) and our 12-kHz profiling telemetry pinger show uniformly increasing sediment cover over most

of the area more than 4 km from the axis, with one or more internal reflectors draping the basalt as evidence of pelagic sedimentation over the relatively smooth basement.

Small-scale basement topography (order 10-100 m) and sediment thickness variations, as well as sediment slumping, may redistribute the conducted heat flux (Von Herzen and Uyeda, 1963). In general, such conductive anomalies will act to reduce the measured thermal gradient and result in concentrating the heat loss near elevated basalt outcrops which are difficult for conventional measurement techniques to sample. The widespread occurrence of thermal gradients less than one-tenth the values predicted from simple plate models would require such effects to be significant over about half of the 570-km² study area. Because of the uniformity of sediment distribution and relatively subdued topography the observed heat flow pattern cannot be explained by conductive anomalies alone.

DISCUSSION OF THE HEAT FLOW FIELD

The primary purposes of the heat flow study were to establish the areal distribution of conductive heat flux near an actively spreading ridge, and to relate the mean flux to theoretical models of plate creation including hydrothermal fluid circulation. There are only 13 measurements that are not compatible with the remarkably smooth heat flow contours of Figure I-1. The regularity in the heat flow measurements at this scale, which allowed contouring of the data with relatively little gerrymandering, contrasts with the heterogeneous variation shown by detailed surveys with wider

station spacing over broader areas (e.g. Davis and Lister, 1977). The areal distribution of measurements shows significant lineation in the heat flow pattern parallel to the axis of spreading and parallel to the strongly lineated topography. In general, the distribution enabled us to predict higher or lower relative heat flow across the region on the basis of the previous data (Williams et al., 1974).

The most reasonable explanation of the spatially varying heat flow field and its amplitude is hydrothermal circulation of seawater driven by both the intrusion of new crustal material at the ridge axis and the convective instability of the seawater saturated oceanic crust (Elder, 1965; Bodvarson and Lowell, 1972; and Lister, 1972). The venting of circulating seawater-derived fluids directly to the sea accounts for the large discrepancy between the conducted heat flow and the calculated heat loss from a theoretical model. By analogy with land hydrothermal systems, high and variable heat flux is expected in a hydrothermally active crust due to variations in the near-surface plumbing.

Figure I-3 illustrates the uniformity of the low heat flow values measured in several stations along the valleys between the scarps crossing our area, and the sharp change that occurs as some of the scarps are crossed. The strong correlation between this subdued, block-faulted topography and the original SOUTHTOW measurements, located along the left side of the figure, can be seen to persist across the region.

Figure I- 3. Large-scale escarpments from topographic base chart, with heat-flow measurements in symbols corresponding to the contour intervals of Figure I-1. Slopes facing axis are generally steeper, and from Deep Tow profiles (Klitgord and Mudie, 1974) appear to be primarily exposed basalt fault scarps. Slopes facing away from the axis are not as steep locally, but total offset exceeds inward-facing fault offset. 13- and 17-km scarps indicated and discussed in text. Profile A A' shown in Figure I-4.

Lister (1972) and Williams et al. (1974) showed predominantly higher heat flux on elevated regions near scarps with a heat flow anomaly amplitude greater than that calculated from conduction theory. Venting of hydrothermal fluids could account for this increased heat flux. We attempted to investigate the relationship between escarpments and heat flow by making closely spaced, well-navigated measurements near two prominent scarps in the region. A sensitive depth sensor on our instrument allowed us to closely determine the positions of measurements with respect to topographic features.

The survey shown in Figure I-3 includes detailed measurements along the scarp at 17 km south of the axis, across to the region where the vertical offset of this scarp dies out, and across a scarp 13 km south of the axis. We note the very sharp transition to high, variable heat flow in crossing the 13 km scarp, and an irregular decrease in the total heat flux to both sides of the scarp (Figure I-4). A similar decrease was ascribed by Williams et al. (1974) to the wavelength of a convecting layer. Again, the heat flux is higher on the up side of the scarp at 17 km although by a smaller amount, with a decreasing difference east along strike. Where the surface offset has decreased below 10 fm, the heat flow is high and uniform.

These closely spaced measurements suggest that localized upwellings along scarps contribute to the heat loss in young crust. Because we do not always find elevated heat flow on scarp tops, the circulation cannot be solely forced by the topography. Of 37 scarp crossings shown in Figure I-3, the heat flux measured at the top

Figure I- 4. Profile A-A' (from Figure I-3) showing the rapid increase in conductive heat flux at the top of the 13-km scarp. The dashed line is drawn between contour values from Figure I-1, and the solid curve from the detailed topographic base chart, along the profile. Dots show actual values of heat flow and pressure-derived depths determined at stations. Distance along the profile is marked in kilometers. The obliquity of the profile with respect to the trend of the topography results in some scatter in the plotted points both in depth and heat flux. The vertical line indicates the sharp offset of the scarp which is smoothed out on the contour chart, and shows the sharp division in the heat flow to either side of the scarp.

of the scarp relative to that measured near the base increased definitely 16 times, showed a slight increase 9 times, showed no change or slight decrease 11 times, and definitely decreased only once (in the generally high heat flow area near $0^{\circ}39'N$, $86^{\circ}1'W$). This predominance of increased heat flux at the high sides of scarps cannot result from uniform conduction models. Circulation here may be along strike of the fault, in a permeable zone embedded in less permeable surrounding crust (Bodvarsson and Lowell, 1972). Our data show that this convection is more complex than the simple roll cell convection of perturbation theory (Lapwood, 1948), and that significant three-dimensional variations exist.

As pointed out by Sleep and Wolery (1978), in the immediate vicinity of venting hydrothermal fluids, the conductive heat flux can be very large, and will contribute to the heat missing in comparing the surface ship data to theoretical models. Such high gradients are found within a field of hydrothermal mounds identified initially on the basis of Deep Tow side-scan sonar observations (Klitgord and Mudie, 1973; Lonsdale, 1977). Indeed, with the ALVIN submersible we found portions of individual mounds where the temperature gradient exceeded $5^{\circ}C/m$ (Williams et al, in press; and ALVIN data in Appendix I-1). These mounds and mound ridges are less than about 50 m wide, and are estimated by Williams et al. (in press) to occupy less than 2.5 % of our study area. They concluded

that these hydrothermal mounds contribute a heat loss of 7 cal/sec/cm ridge maximum, or less than 15% of the overall heat budget for crust less than 1 My (Table 1.).

COMPARISON OF HEAT FLOW FIELD TO THEORETICAL HEAT FLUX

Figures I-1 and I-3 show a fairly smooth spatial variation of the conductive heat flow within the region of closely spaced measurements. This data encourages us to believe that we have adequately sampled the heat flow for the computation of an areal average. We can only put a lower bound on the total heat flux because some heat may be escaping undetected near the mounds or other undersampled areas of rapidly changing conductive heat loss near sites of fluid discharge as just discussed.

The total heat flux was determined by planimetering the area within the contour intervals indicated in Figure 1 for each of three independently contoured heat flow charts; variations between these three determinations were used as an error estimate of the accuracy of the power out of the area (Figure I-5 and Table I-1). The overall power output is quite well constrained at $(40.8 \pm 4.7) \times 10^6$ cal/sec (169 ± 20 MW). Thus the total area of 573 km^2 over which the contours were integrated, shown in Figure I-1, has an average heat flux of 7.1 ± 0.8 HFU ($295 \pm 33 \text{ mW/m}^2$). This estimate is somewhat less than the preliminary estimate in Corliss et al. (1979), reflecting a slight change in the region considered and in the contours resulting from improved station data.

Figure I- 5. Histogram of number of observations (dotted line) and planimetered contour area (solid line) versus heat-flow class intervals. Observations falling outside planimetered study area (dashed outline of Figure I-1) but listed in Appendix I.1 are not included in the total number of observations. Errors in the contour areas discussed in text.

Table I-1: Calculation of Heat Balance of GSC

Total area (dashed outline, Figure I-1)	573 km ²
Age range	0.10 to 0.85 my
Distance range(from axis)	3.50 to 29.8 km
Equivalent width (area/distance)	21.8 km
Integrated heat flux	
Observed	40.8 x 10 ⁶ cal/sec
Divided by width	18.7 cal/cm ridge/sec
(Age) ^{-1/2} theoretical	50.9 cal/cm ridge/sec

Simple plate tectonic models predict singular heat flux on axis, and a high, monotonically decreasing heat flow over young crust. We would like to establish the limits to which conductive models of heat loss can be used in comparison to the field data. Parsons and Sclater (1977) showed that for distances greater than the advective length scale d (km) based on the plate velocity v (here 35 km/My), and the thermal diffusivity κ (for shallow crust, about $19 \text{ km}^2/\text{My}$), the one-dimensional heat flow model is a good approximation to the thermal structure of a conductively cooling plate model. In the Galapagos study, we have $d = \kappa/v = 0.54 \text{ km}$. Using Parsons and Sclater's (1977) best-fit value of diffusivity of $8.04 \times 10^{-3} \text{ cm}^2/\text{sec}$ ($25.3 \text{ km}^2/\text{My}$), obtained empirically from older crust, we obtain a length scale of $d = 0.72 \text{ km}$. If the plate were cooling solely by conduction, then for distances much greater than about a kilometer, the plate advection would dominate horizontal conduction, and the heat loss would be asymptotically similar to the one-dimensional model. A more severe limitation at this scale is the width of the accretion zone, marking the formation of the rigid crust. Convective and time-dependent effects obviously will limit the extent to which the theoretical heat flux applies, both within the hydrothermal layer and below in the plastically flowing mantle. However, due to the moderate plate velocity and the rapid thickening of the plate with age, we can use the $(\text{age})^{1/2}$ model with care over the surveyed area, which begins 3.5 km off axis.

The heat flux from older oceanic crust has been shown to follow the relationship of $12/(\text{age(my)})^{1/2}$ (Lister, 1977). Integration over the age range of the area from .1 to .85 My (Figure I-1) yields an average heat flow of about 19.4 HFU. Our mean flux of 7.1 HFU is thus about 37% of this empirically predicted value. Most of the heat flow from young crust is derived from cooling of material in the shallow levels directly below the surface. Thus the discrepancy between our mean value and the predicted heat flow may reflect crustal material that is significantly cooler than predicted from conduction models.

The quantity of heat removed at the axis by venting of hydrothermal fluids will depend on the depth of circulation and the temperature at the base of the layer. The model of Sclater and Klitgord (1973), which included horizontal conduction but not off-axis hydrothermal activity , illustrated that quenched temperature profiles assigned at the axis to arbitrary depth will decay back to the predicted plate temperatures monotonically on a time scale fixed by the depth of the initial cooling.

Circulation of hydrothermal fluids provides a mechanism for cooling the crust and causing local surface heat-flow anomalies. The detailed measurements reported here provide evidence for this process close to a spreading center. Ranges of some relevant parameters have been investigated with numerical techniques and will be reported in Chapter III.

One possible model is that the spatial variation of surface heat flow reflects the intensity of cooling at the spreading axis which varies over geological time. The surface heat flow measurements do not put tight constraints on the axial cooling depth, but with increasing age the effects of conduction would be to reduce the amplitudes and broaden the wavelength of the anomalies. Instead, we observe undiminished, narrow anomalies to the limits of our survey region ($0^{\circ} 25' N$, $86^{\circ} 06' W$, Figure I-3), about 1.2 My age. Thus we infer that circulation continues to be active over most of our survey area.

The heat flow field appears to change systematically in a direction parallel to the axis. The dominant peaks and valleys in the heat flow pattern deduced from the SOUTHTOW data continue east-west across the area about 10 km before dying out in the gradual increase in the mean flux to the east, a transition that may not be complete in the 30 km width of our survey. Such a large-scale variation may arise from a variety of causes. The tectonic and emplacement processes within 4 km around the axis where we have no heat flow data could result in irregular horizontal and vertical permeability to seawater flow within the crust, or change the overall thickness of the crust. Such crustal thickness changes are suggested by recent sonobuoy and refraction studies (Houtz and Ewing, 1976), and may well influence the convective pattern. By analogy with laboratory studies which show more effective heat transfer in thicker layers (Elder, 1965), the increase in the mean conductive heat flux from 6 HFU in the west to 12 HFU in the east

across our area may imply a thinner circulation layer in the west.

A large fracture zone 40 km further east may influence the local tectonics, and hence the thermal input to the plate across our area. It seems unlikely that the increase in the mean heat flux to the east is related directly to this fracture zone, because it is too distant to have a conductive effect from the age differential across the boundary. In any case, older (and hence cooler) crust on the east side of the boundary would have an effect opposite to that observed (increase towards boundary).

The gradual topographic slope along strike between 5 to 35 km south of the axis (Figure I-2), if mimicked by the base of the water-saturated crust, will result in a slow flow upslope at the base of the hydrothermal layer and parallel to the ridge crest (Bories and Combarrous, 1973). The broad high observed in the east may be the diffuse zone over which such a forced convection would rise. For example, horizontal mass flow of 10^{-8} g/cm²/sec at 5 km depth, at a mean temperature of 300°C, would transport more than 3 HFU.

Of course, due to the relatively more widely spaced stations in the east, there may be an aliased higher frequency variation which we have simply undersampled. More measurements to the east are necessary before this broad-scale variation can be confirmed. From the continuity in the heat flow observations over most of our area parallel to the spreading center, the distance over which this pattern remains coherent is at least 25 km.

CONCLUSIONS

The variation of conducted heat flux has been well determined for 0.1 to 1.0 My age crust at the GSC at 86°W longitude. The dense net of stations has provided a detailed picture of the variation of heat flow on young crust. The surface data will constrain future estimates of how the thermal regime of the plate forms and evolves and the relation of the tectonics of emplacement to the surrounding plate. We summarize these observations as follows:

1. The mean conducted flux is about one-third of that expected on the basis of the one-dimensional conductive model which fits heat flow on older crust, with the missing heat lost by hydrothermal venting.
2. The systematic variations in heat flux reflect seawater-derived fluids circulating in the crust throughout the area, partially maintaining the relatively cool crustal temperatures established at the axis.
3. Corresponding lineations of both the topography and heat-flow pattern suggest continued circulation influenced by the topographic escarpments.
4. Large local variation in conducted heat flux across hydrothermal mounds and topographic escarpments suggests local venting of low-temperature fluids extending more than 35 km off-axis to crust older than 1 My.

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Appendix I-1: Heat-flow measurements on the Galapagos Spreading Center
near 86° W, 1° N.

The table lists heat flow measurements taken from surface ships and from the submersible ALVIN during joint WHOI-SIO cruises SOUTHTOW, PLEIADES, and KNORR 64. The SOUTHTOW, SCAN, and PIQUERO data have been previously published by Williams, et al. (1974), and by Sclater and Klitgord (1973); here we list the data in a format consistent with our data, and to include changes in published locations and depths on those Southtow stations marked with an asterisk. The table entries are:

Cruise Cruise name

Station Heat Flow Station Number.

No. Penetration number during station.

0° N Latitude in decimal minutes north of $0^{\circ} 0'$ N.

min. Note: Stations north of 1° N are listed as greater than $60'$.

86° W Longitude in decimal minutes west of $86^{\circ} 0'$ W.

min. Note: Stations east of 86° W are listed as negative minutes.

Navigation

Q: code referring to navigation method

A--acoustically navigated fish (147 penetrations).

B--acoustically navigated ship (153 penetrations).

C--single transponder fish only (69 penetrations).

D--single range ship only, either radar ranging on the ALVIN tender LULU or acoustic ranging (18 penetrations).

E--satellite navigation of ship (50 penetrations).

F--dead-reckoned (16 penetrations).

max min brg:

Position error ellipse - maximum and minimum uncertainty in meters relative to the KNORR 64 acoustic navigation net together with the strike of the major uncertainty axis, in degrees.

Depth

fm: depth of instrument at penetration, in uncorrected fathoms.

S : Depth source: 1.--from instrument pressure sensor.
2.--from combined ship fathometer and topographic base chart.
3.--from ship fathometer.

E : Environment--local slope (within 500 m) of sea floor based on combined analysis of ship fathometer and instrument 12 kHz records, and base topographic chart.
1.--Flat bottom, less than 5 fm/100 fm relief.
2.--Flat slope, 5 to 10 fm/100 fm.
3.--Slope, 10 to 15 fm/100 fm.
4.--Steep slope, 15 to 25 fm/100 fm.
5.--Scarp, greater than 25 fm/100 fm.
6.--Ridge, greater than 10-fm decrease in two opposite directions.
7.--Valley, greater than 10-fm rise in two opposite directions.

N : Number of thermistors used in gradient measurement.

Sed: Depth, in meters, of deepest reflector visible on telemetry record. The pinger within our instrument allowed us to verify the continuity of shallow internal reflectors between many penetrations, and yielded minimum sediment thicknesses at most stations. Where no reflector was visible, the thickness is given as 0.

Pen: Depth, in meters, of lowest thermistor below sediment surface (inferred).

T: Tilt, in degrees from vertical, measured by instrument.

HFU Heat flow value, in HFU ($\text{HFU} = 10^{-6} \text{ cal/cm}^2/\text{sec} = 41.83 \text{ mW/m}^2$).

All heat-flow measurements during multiple penetration stations must assume a sediment thermal conductivity in converting the measured temperature gradients to heat fluxes. We found no significant change in the thermal conductivity measured on piston and gravity cores throughout the area. Therefore we converted all thermal gradients to heat flow using a conductivity value of $1.77 \text{ mcal/cm/sec/}^{\circ}$ during PLEIADES II and KNORR 64. The small change in conductivity for reported stations from SOUTHTOW, PEQUIRO, and SCAN does not contribute significantly to the observed heat-flux changes.

The measurements from the DSRV ALVIN have the last seven columns replaced with measurement parameters used in correcting temperature-gradient measurements for the effects of water percolation, as described in Chapter II:

- q_{cond} The heat flux (HFU) obtained by multiplying the mean gradient for the three thermistors by the mean sediment conductivity of 1.77 TCU.
- w_o The mass flux (micrograms/cm²/sec) derived from a model calculation fit to the observed temperature probe-data.
- q_{tot} The total heat flux (HFU) including both conduction and advection.

Where the simple model does not fit the data, the mean gradient and heat flow are given without a water flux.

Cruise	Station	No.	0°N 86°W		Q	Navigation			Depth			Sed.			Pen.		HFU
			min.	min.		max	min	brg	fm	S	E	N	m	m	T		
KNORR 64	1	1	44.99	10.56	A	50	50	0	1520	1	7	3	9.6	2.8	0	6.3	
KNORR 64	1	2	44.91	10.52	A	50	50	0	1521	1	7	3	9.1	2.8	0	2.1	
KNORR 64	1	3	44.84	10.29	A	50	50	0	1524	1	7	3	9.8	2.8	0	1.1	
KNORR 64	1	4	44.74	9.83	A	50	50	0	1519	1	7	3	9.6	2.7	15	0.7	
KNORR 64	1	5	44.71	9.60	B	150	150	0	1520	1	7	3	6.7	2.7	15	1.7	
KNORR 64	1	7	44.65	9.13	B	150	150	0	1509	1	7	3	0.0	2.7	15	0.1	
KNORR 64	2	1	38.38	5.61	A	25	10	90	1476	1	3	3	14.1	2.8	0	3.9	
KNORR 64	2	2	38.32	5.38	A	20	10	90	1464	1	3	3	14.1	2.4	30	5.0	
KNORR 64	2	3	38.22	5.13	A	20	20	0	1462	1	3	3	11.8	2.7	15	6.4	
KNORR 64	3	1	38.20	4.92	A	20	10	90	1450	1	3	3	7.2	2.8	0	11.5	
KNORR 64	3	2	38.22	4.54	A	20	10	90	1454	1	3	2	12.6	2.4	30	6.1	
KNORR 64	3	3	38.10	4.06	A	20	20	0	1445	1	3	3	9.3	2.7	15	5.9	
KNORR 64	3	4	37.95	3.65	A	25	25	0	1450	1	3	3	12.0	2.7	15	9.7	
KNORR 64	3	5	37.87	3.22	A	25	25	0	1448	1	3	3	12.2	2.7	15	9.8	
KNORR 64	3	6	37.80	2.85	B	50	50	104	1446	1	3	3	12.4	2.7	15	11.3	
KNORR 64	3	7	37.70	2.38	F	150	100	104	1432	1	3	3	11.1	2.7	15	14.8	
KNORR 64	3	8	37.54	1.96	F	200	150	104	1427	1	3	3	20.9	2.8	0	14.3	
KNORR 64	3	9	37.43	1.53	F	200	150	104	1428	1	3	3	12.6	2.7	15	18.1	
KNORR 64	4	1	37.59	7.53	A	20	20	0	1446	1	5	3	14.4	2.8	0	4.9	
KNORR 64	4	2	37.66	7.35	A	20	20	0	1441	1	5	3	12.0	2.8	0	4.6	
KNORR 64	4	3	37.74	6.74	A	100	100	0	1439	1	5	3	12.8	2.4	30	2.6	
KNORR 64	4	4	37.80	6.74	A	10	10	0	1441	1	5	3	9.8	2.7	15	3.4	
KNORR 64	4	5	37.73	6.37	A	50	50	0	1436	1	5	3	12.2	2.7	15	4.6	
KNORR 64	4	6	37.64	5.91	A	100	100	0	1436	1	5	3	6.3	2.8	0	7.8	
KNORR 64	4	7	37.68	5.54	A	100	100	0	1443	1	5	3	21.8	2.8	0	12.8	
KNORR 64	4	8	37.63	5.25	A	50	20	90	1439	1	5	3	13.1	2.8	0	8.6	
KNORR 64	5	1	60.63	11.10	E	800	400	95	1483	1	2	3	6.1	2.8	0	6.4	
KNORR 64	5	2	58.98	10.63	E	400	200	95	1510	1	6	3	0.0	2.7	15	3.8	
KNORR 64	5	3	57.88	10.63	E	400	200	95	1446	1	3	3	16.5	2.8	0	3.2	
KNORR 64	5	4	56.98	10.63	E	400	300	95	1432	1	2	3	16.1	2.8	0	7.1	
KNORR 64	5	5	55.75	11.00	C	300	300	0	1448	1	1	3	8.9	2.8	0	11.0	
KNORR 64	5	6	55.10	11.24	C	300	200	95	1453	1	3	3	8.0	2.8	0	8.6	
KNORR 64	5	7	54.35	11.00	C	300	200	95	1430	1	1	3	9.1	2.8	0	13.7	
KNORR 64	5	8	53.94	11.00	C	300	200	95	1449	1	5	3	7.0	2.8	0	1.5	
KNORR 64	5	9	52.93	10.68	C	200	200	0	1448	1	3	3	10.0	2.8	0	5.0	
KNORR 64	5	10	52.07	11.25	C	200	150	95	1407	1	2	3	10.0	2.7	15	11.8	
KNORR 64	5	11	51.31	11.11	E	150	150	95	1406	1	2	3	7.6	2.4	30	7.9	
KNORR 64	6	1	38.29	7.67	A	40	40	0	1476	1	2	3	8.5	2.8	0	2.0	
KNORR 64	6	2	38.18	7.41	A	40	40	0	1471	1	2	3	16.5	2.8	0	2.6	
KNORR 64	6	3	38.03	6.60	A	20	20	0	1489	1	7	3	13.3	2.8	0	3.0	
KNORR 64	6	4	38.18	6.13	A	20	20	0	1472	1	4	3	12.4	2.8	0	2.7	
KNORR 64	6	5	37.92	5.57	A	20	20	0	1475	1	7	3	10.9	2.8	0	3.8	
KNORR 64	7	1	30.12	-0.62	C	200	100	45	1528	1	1	3	13.1	2.8	0	5.2	
KNORR 64	7	2	30.70	-0.52	C	200	100	45	1527	1	2	3	12.9	2.8	0	4.8	
KNORR 64	7	3	31.21	-0.32	C	200	150	45	1521	1	1	3	11.5	2.8	0	5.9	
KNORR 64	7	4	32.05	0.02	C	300	250	45	1509	1	3	3	10.2	2.8	0	6.0	
KNORR 64	7	5	32.90	0.00	C	300	250	45	1486	1	1	3	14.4	2.8	0	9.3	
KNORR 64	7	6	33.20	0.18	C	300	250	30	1476	1	5	3	8.9	2.8	0	8.0	
KNORR 64	7	7	34.20	0.18	C	250	200	90	1500	1	4	3	12.9	2.8	0	4.4	

Cruise	Station No.	0°N		86°W		Q	Navigation			Depth			Sed.			Pen. m	HFU
		min.	min.	max	min		brg	fm	S	E	N	m	T				
KNORR 64	8	1	36.08	0.73	C	200	100	90	1469	1	2	3	11.3	2.8	0	19.6	
KNORR 64	8	2	36.43	0.73	C	300	150	90	1454	1	2	3	10.5	2.8	0	16.1	
KNORR 64	8	3	36.95	0.73	C	300	150	90	1433	1	2	3	4.6	2.8	0	20.3	
KNORR 64	8	4	37.95	0.85	C	300	150	90	1448	1	1	3	11.8	2.8	0	13.9	
KNORR 64	8	5	38.36	1.02	C	300	150	90	1448	1	1	3	17.0	2.7	15	13.5	
KNORR 64	8	6	39.13	1.20	C	300	150	90	1439	1	3	3	15.0	2.7	15	13.5	
KNORR 64	8	7	39.52	1.18	C	300	150	90	1435	1	3	3	6.8	2.8	0	26.9	
KNORR 64	8	8	40.00	1.20	C	300	150	90	1444	1	3	3	13.5	2.8	0	17.4	
KNORR 64	9	1	40.50	0.18	C	150	150	0	1419	1	2	3	8.5	2.8	0	14.4	
KNORR 64	9	2	41.40	-0.02	C	250	150	90	1451	1	3	3	10.9	2.8	0	0.9	
KNORR 64	9	3	42.48	0.03	C	250	150	90	1416	1	3	3	10.0	2.8	0	8.2	
KNORR 64	9	4	43.10	0.08	C	250	150	90	1420	1	3	3	7.6	2.8	0	6.1	
KNORR 64	9	6	44.12	0.18	C	250	150	90	1360	1	4	2	6.7	2.4	30	0.4	
KNORR 64	9	7	45.83	0.65	C	350	200	90	1375	1	4	1	0.0	2.4	30	0.6	
KNORR 64	9	9	44.25	1.69	C	150	150	0	1400	1	3	3	5.2	2.8	0	1.5	
KNORR 64	9	10	43.91	2.38	C	250	150	90	1395	1	2	3	5.6	2.8	0	5.9	
KNORR 64	9	11	43.90	3.37	C	250	150	90	1395	1	6	3	5.7	2.8	0	8.4	
KNORR 64	9	12	43.95	3.70	C	250	150	90	1415	1	3	3	10.0	2.8	0	5.5	
KNORR 64	11	1	35.54	5.73	A	20	20	0	1467	2	1	3	4.3	8.7	0	15.3	
KNORR 64	12	1	43.52	2.32	D	200	200	0	1415	3	2	3	7.2	2.8	0	5.7	
KNORR 64	12	2	43.34	2.66	D	200	200	0	1415	3	2	3	6.5	2.8	0	5.8	
KNORR 64	12	3	43.10	2.92	D	200	200	0	1420	3	7	3	6.5	2.8	0	4.3	
KNORR 64	12	4	43.12	3.36	D	200	200	0	1420	3	7	3	6.8	2.8	0	5.2	
KNORR 64	12	5	43.15	3.96	D	200	200	0	1450	3	7	3	5.9	2.8	0	3.4	
KNORR 64	12	6	43.00	4.40	D	200	200	0	1440	3	7	3	5.4	2.8	0	4.9	
KNORR 64	12	8	43.10	5.02	D	300	200	90	1410	3	5	1	5.2	2.4	30	2.6	
KNORR 64	12	9	43.78	4.98	D	300	200	90	1440	3	2	3	6.3	2.4	30	3.6	
KNORR 64	12	10	43.80	4.60	D	300	200	90	1430	3	2	3	6.5	2.8	0	4.7	
KNORR 64	12	11	44.38	4.40	D	300	200	90	1430	3	2	3	6.5	2.8	0	2.0	
KNORR 64	12	12	44.68	4.42	D	300	200	90	1430	3	2	3	3.9	2.8	0	0.9	
KNORR 64	13	1	36.62	5.86	A	20	10	0	1461	1	1	4	5.9	5.7	0	12.3	
KNORR 64	14	1	41.84	4.48	C	250	100	90	1473	1	2	4	9.4	7.8	0	1.8	
KNORR 64	15	1	40.30	5.03	A	40	20	90	1459	1	7	3	5.6	2.8	0	3.6	
KNORR 64	15	2	40.21	4.97	A	60	40	0	1454	1	7	3	2.8	2.7	15	5.0	
KNORR 64	15	3	40.05	4.85	A	20	10	0	1465	1	5	3	4.3	2.8	0	7.8	
KNORR 64	15	4	39.95	4.88	A	20	20	0	1423	1	5	3	8.1	2.8	0	15.2	
KNORR 64	15	5	39.87	4.88	A	20	20	0	1421	1	5	3	10.9	2.8	0	11.7	
KNORR 64	15	6	39.74	4.77	A	20	20	0	1420	1	5	3	3.9	2.8	0	20.3	
KNORR 64	15	7	39.72	4.69	A	20	20	0	1423	1	5	3	12.6	2.8	0	11.4	
KNORR 64	15	8	39.71	4.67	A	20	20	0	1424	1	5	3	5.2	2.4	30	23.9	
KNORR 64	15	9	39.73	4.53	A	20	20	0	1422	1	5	3	5.2	2.8	0	18.8	
KNORR 64	15	10	39.59	4.53	A	50	20	0	1429	1	4	3	0.0	2.4	30	22.8	
KNORR 64	15	11	39.51	4.51	A	30	20	0	1432	1	3	3	2.8	2.7	15	7.9	
KNORR 64	15	12	39.46	4.48	A	30	20	90	1433	1	3	3	2.8	2.7	15	9.2	
KNORR 64	15	13	39.38	4.40	A	20	20	0	1440	1	3	3	16.8	2.8	0	10.9	
KNORR 64	15	14	39.32	4.35	A	40	40	0	1445	1	2	3	6.8	2.8	0	7.2	
KNORR 64	15	15	39.16	4.23	A	40	40	0	1446	1	2	3	5.9	2.8	0	8.7	
KNORR 64	15	16	39.12	4.08	A	20	20	0	1456	1	2	3	12.8	2.8	0	7.0	
KNORR 64	15	17	39.04	4.01	A	30	20	0	1458	1	2	3	16.8	2.8	0	6.1	

Cruise	Station	No.	0°N	86°W	Q	Navigation			Depth				Sed.	Pen.	HFU	
			min.	min.		max	min	brg	fm	S	E	N	m	m	T	
KNORR 64	15	18	39.03	4.00	A	20	20	0	1457	1	2	3	5.9	2.8	0	4.1
KNORR 64	15	19	38.93	4.01	A	20	20	0	1457	1	3	3	16.3	2.8	0	7.0
KNORR 64	15	20	38.68	3.95	A	50	50	0	1459	1	1	3	17.9	2.8	0	6.3
KNORR 64	15	21	38.46	3.76	A	50	30	90	1456	1	1	3	9.3	2.8	0	15.8
KNORR 64	15	22	38.53	3.55	A	50	50	0	1456	1	1	3	13.9	2.8	0	6.5
KNORR 64	15	23	38.44	3.42	A	20	20	0	1454	1	1	3	12.4	2.8	0	11.6
KNORR 64	15	24	38.55	3.09	A	100	100	0	1456	1	1	3	0.0	2.8	0	6.5
KNORR 64	16	1	36.98	8.65	A	10	10	0	1439	1	4	3	7.0	2.8	0	7.0
KNORR 64	16	2	37.08	8.68	A	20	10	0	1429	1	4	3	7.6	2.8	0	9.1
KNORR 64	16	3	37.20	8.72	A	80	40	0	1420	1	4	3	0.0	2.8	0	8.5
KNORR 64	16	4	37.38	8.76	A	40	40	0	1420	1	3	3	6.1	2.8	0	11.3
KNORR 64	16	5	37.45	8.76	A	40	20	0	1414	1	3	3	11.3	2.8	0	9.8
KNORR 64	16	6	37.55	8.73	A	30	10	0	1411	1	3	3	5.2	2.8	0	7.8
KNORR 64	16	7	37.75	8.71	A	30	10	0	1408	1	3	3	5.4	2.4	30	9.1
KNORR 64	16	8	38.00	8.64	A	40	20	0	1403	1	3	3	5.4	2.4	30	5.6
KNORR 64	16	9	38.60	8.05	A	10	10	0	1452	1	3	3	6.1	2.8	0	1.6
KNORR 64	16	10	38.55	7.96	A	40	20	0	1450	1	3	3	2.0	2.8	0	1.9
KNORR 64	16	11	38.39	7.96	A	10	10	0	1454	1	3	3	10.9	2.8	0	1.7
KNORR 64	16	12	38.20	7.93	A	20	10	0	1467	1	5	3	5.9	2.8	0	2.5
KNORR 64	16	15	38.16	7.97	A	10	10	0	1463	1	5	3	16.5	2.4	30	1.7
KNORR 64	17	1	40.25	4.95	B	200	100	90	1437	1	7	3	14.4	2.8	0	5.0
KNORR 64	17	2	40.45	5.00	B	200	100	90	1424	1	3	3	0.0	2.8	0	5.8
KNORR 64	17	3	40.70	4.95	C	300	200	90	1436	1	7	3	11.8	2.8	0	4.3
KNORR 64	17	6	41.95	4.50	C	400	200	90	1439	1	3	2	6.7	2.8	0	2.4
KNORR 64	17	7	42.28	4.27	C	400	200	90	1403	1	4	2	8.0	2.8	0	1.9
KNORR 64	17	8	42.60	4.25	C	400	200	90	1387	1	4	2	7.2	2.8	0	5.0
KNORR 64	18	1	42.62	4.05	D	200	150	90	1395	3	4	3	7.6	2.8	0	3.7
KNORR 64	18	2	42.72	4.05	D	300	150	90	1395	3	4	3	8.0	2.8	0	3.9
KNORR 64	18	3	42.88	3.80	D	400	200	90	1425	3	5	3	6.1	2.8	0	2.0
KNORR 64	18	4	43.10	3.62	D	500	200	90	1428	3	5	3	0.0	2.8	0	8.0
KNORR 64	18	5	43.12	3.60	D	500	200	90	1428	3	5	3	11.3	2.8	0	6.7
KNORR 64	18	6	43.14	3.58	D	500	200	90	1428	3	5	3	7.2	2.8	0	4.7
KNORR 64	19	1	35.66	5.62	A	20	20	0	1459	2	1	1	3.0	3.2	0	7.2
KNORR 64	20	1	35.63	8.31	A	10	10	0	1439	1	2	4	0.0	3.8	0	14.5
KNORR 64	21	1	36.10	8.27	A	10	10	0	1450	1	2	3	2.0	2.8	0	8.1
KNORR 64	21	2	36.30	8.23	A	20	10	0	1461	1	1	3	13.1	2.8	0	8.4
KNORR 64	21	3	36.50	8.21	A	20	10	0	1468	1	1	3	6.3	2.8	0	8.6
KNORR 64	21	5	36.96	8.10	A	20	10	0	1467	2	5	3	0.0	2.8	0	5.2
KNORR 64	21	6	37.25	8.04	A	20	20	0	1434	2	1	3	3.5	2.8	0	5.2
KNORR 64	21	7	37.31	7.97	A	20	20	0	1432	2	1	3	6.7	2.8	0	2.3
KNORR 64	21	8	37.55	7.82	A	30	20	0	1440	2	6	3	8.3	2.8	0	3.0
KNORR 64	22	1	33.20	9.43	D	150	100	90	1485	3	2	3	12.8	2.8	0	5.3
KNORR 64	23	1	33.64	10.00	C	150	150	0	1448	1	2	3	5.4	2.8	0	5.7
KNORR 64	23	2	33.42	9.13	C	200	150	90	1473	1	4	3	5.6	2.8	0	5.3
KNORR 64	23	3	33.25	8.60	C	200	150	90	1487	1	4	3	4.4	2.7	15	7.6
KNORR 64	23	4	33.00	8.38	C	200	200	0	1493	1	4	3	5.7	2.7	15	4.3
KNORR 64	23	5	33.22	7.90	C	300	100	90	1474	1	4	3	10.4	2.8	0	4.4
KNORR 64	23	6	32.90	7.75	C	200	200	0	1488	1	4	3	12.9	2.8	0	3.1
KNORR 64	23	7	32.87	7.52	C	200	200	0	1489	1	4	3	13.9	2.7	15	3.0

Cruise	Station	No.	0°N	86°W	Q	Navigation			Depth				Sed.	Pen.	HFU	
			min.	min.		max	min	brg	fm	S	E	N	m		m	T
KNORR 64	23	8	32.87	7.42	C	200	200	0	1491	1	4	3	13.7	2.7	15	3.0
KNORR 64	24	1	35.01	8.25	A	10	10	0	1445	2	1	5	0.0	6.2	0	29.4
KNORR 64	25	1	35.58	5.87	A	10	10	0	1466	2	1	5	0.0	6.9	0	26.1
KNORR 64	26	1	42.00	-1.50	F	500	300	90	1407	1	2	3	6.8	2.8	0	9.8
KNORR 64	26	2	41.00	-1.50	F	500	300	90	1428	1	2	3	9.6	2.8	0	6.0
KNORR 64	26	3	40.35	-1.50	F	500	300	90	1451	1	2	3	10.9	2.8	0	3.9
KNORR 64	27	1	39.81	0.21	F	500	300	90	1440	3	2	3	15.0	2.8	0	12.9
KNORR 64	27	2	38.81	0.21	F	500	300	90	1452	3	2	3	15.2	2.8	0	10.9
PLEIADES II	3	1	38.72	6.15	B	200	100	90	1457	1	2	3	13.1	2.8	0	6.8
PLEIADES II	3	2	38.73	6.13	B	200	100	90	1463	1	2	3	13.1	2.8	0	5.0
PLEIADES II	3	3	38.45	6.22	B	200	100	90	1474	1	7	3	15.2	2.4	30	3.7
PLEIADES II	3	4	38.30	6.35	B	200	100	90	1475	1	7	3	0.0	2.4	30	3.2
PLEIADES II	5	1	37.23	5.63	B	100	50	90	1434	1	1	3	16.5	2.8	0	9.3
PLEIADES II	5	2	36.95	5.48	B	100	50	90	1454	1	2	3	16.5	2.8	0	6.2
PLEIADES II	5	3	36.55	5.53	B	100	50	90	1457	1	2	3	16.5	2.8	0	10.0
PLEIADES II	5	4	36.29	5.79	B	200	100	90	1466	1	1	3	8.0	2.4	30	10.9
PLEIADES II	5	5	36.04	5.75	B	200	100	90	1464	1	1	3	14.6	2.4	30	12.7
PLEIADES II	5	6	35.74	5.77	B	200	100	90	1463	1	2	3	18.5	2.8	0	13.5
PLEIADES II	5	7	35.46	5.80	B	200	100	90	1466	1	2	3	16.5	2.8	0	11.5
PLEIADES II	5	8	35.26	5.72	B	200	100	90	1467	1	1	3	16.5	2.8	0	11.6
PLEIADES II	5	9	35.01	5.65	B	100	50	90	1467	1	1	3	16.5	2.8	0	13.4
PLEIADES II	5	10	34.79	5.65	B	200	100	90	1467	1	1	3	13.1	2.4	30	12.1
PLEIADES II	5	11	34.38	5.73	B	200	100	90	1466	1	2	3	12.6	2.4	30	11.3
PLEIADES II	5	12	33.96	5.91	B	200	100	90	1456	1	1	3	6.7	2.8	0	8.1
PLEIADES II	5	13	33.46	6.05	B	200	100	90	1461	1	3	3	14.6	2.8	0	9.0
PLEIADES II	5	14	33.20	6.07	B	200	100	90	1474	1	4	3	15.2	2.8	0	6.6
PLEIADES II	5	15	32.82	6.17	B	200	100	90	1488	1	3	3	15.2	2.8	0	2.1
PLEIADES II	5	16	32.29	6.26	B	300	300	0	1492	1	2	3	15.2	2.8	0	1.4
PLEIADES II	5	17	32.00	6.30	B	300	300	0	1494	1	1	3	14.6	2.8	0	3.8
PLEIADES II	5	18	31.62	6.48	B	300	300	0	1494	1	1	3	15.2	2.8	0	7.4
PLEIADES II	5	19	31.33	6.55	B	150	150	0	1494	1	3	3	15.2	2.8	0	7.7
PLEIADES II	5	20	30.91	6.63	B	150	150	0	1515	1	2	3	15.9	2.8	0	6.9
PLEIADES II	6	1	40.55	9.75	C	200	100	45	1418	1	4	3	17.8	2.8	0	1.4
PLEIADES II	6	2	40.34	9.66	C	200	100	90	1442	1	4	3	17.8	2.4	30	2.5
PLEIADES II	6	3	40.10	9.60	C	150	150	0	1449	1	6	3	16.5	2.8	0	2.7
PLEIADES II	6	4	39.84	9.48	C	200	100	90	1453	1	3	3	16.5	2.4	30	3.0
PLEIADES II	6	5	39.60	9.29	C	200	100	90	1466	1	3	3	17.8	2.8	0	2.1
PLEIADES II	6	6	39.19	9.12	C	200	100	90	1480	1	5	3	8.0	2.4	30	1.9
PLEIADES II	6	7	38.81	9.03	C	200	100	90	1472	1	3	3	13.1	2.8	0	2.3
PLEIADES II	6	8	38.61	8.97	C	150	150	0	1474	1	2	3	13.9	2.8	0	1.9
PLEIADES II	6	9	38.45	8.82	A	150	150	0	1435	1	5	3	11.3	2.8	0	3.6
PLEIADES II	7	1	38.09	8.78	A	100	100	0	1418	1	1	3	15.9	2.8	0	5.1
PLEIADES II	7	2	37.84	8.59	A	100	100	0	1420	1	1	3	10.0	2.8	0	3.5
PLEIADES II	7	3	37.46	8.38	A	100	100	0	1431	1	2	3	6.7	2.8	0	6.1
PLEIADES II	7	4	36.64	8.55	A	100	100	0	1461	1	2	3	0.0	2.4	30	6.2
PLEIADES II	7	5	36.63	8.50	A	100	100	0	1462	1	2	3	5.9	2.7	15	4.9
PLEIADES II	7	6	36.53	8.70	B	250	100	90	1461	1	1	3	15.2	2.8	0	5.8
PLEIADES II	7	7	36.43	8.84	B	250	100	90	1442	1	3	3	15.9	2.8	0	8.2
PLEIADES II	7	8	35.69	8.83	B	250	100	90	1438	1	1	3	16.5	2.8	0	8.4

Cruise	Station	No.	0°N 86°W		Q	Navigation			Depth				Sed.			Pen.	HFU	
			min.	min.		max	min	brg	fm	S	E	N	m	m	T			
PLEIADES	II	7	9	35.30	8.93	B	250	100	90	1440	1	3	3	14.6	2.8	0	12.5	
PLEIADES	II	7	10	35.12	8.81	B	250	100	90	1448	1	1	3	13.1	2.8	0	12.5	
PLEIADES	II	7	11	34.73	8.88	B	250	100	90	1458	1	2	3	15.2	2.8	0	13.6	
PLEIADES	II	7	12	34.22	8.88	B	250	100	90	1455	1	2	3	15.2	2.8	0	13.9	
PLEIADES	II	8	1	40.58	2.50	C	200	100	95	1424	1	6	3	8.0	2.8	0	2.7	
PLEIADES	II	8	2	40.38	2.48	C	100	100	0	1423	1	6	3	12.2	2.8	0	1.7	
PLEIADES	II	8	3	40.20	2.50	C	200	200	0	1433	1	3	3	13.7	2.8	0	5.3	
PLEIADES	II	8	4	39.98	2.45	B	250	100	95	1438	1	6	3	14.8	2.8	0	6.5	
PLEIADES	II	8	5	39.43	2.40	B	200	200	0	1435	1	5	2	6.3	2.8	0	18.8	
PLEIADES	II	8	6	39.27	2.40	B	200	200	0	1430	1	3	3	16.5	2.8	0	9.6	
PLEIADES	II	8	7	38.87	2.40	B	200	200	0	1446	1	2	3	8.0	2.8	0	7.5	
PLEIADES	II	8	8	38.53	2.40	B	200	200	0	1451	1	3	3	5.4	2.8	0	6.6	
PLEIADES	II	8	9	38.20	2.42	B	250	100	0	1455	1	2	3	13.1	2.8	0	10.1	
PLEIADES	II	8	10	37.95	2.40	B	200	200	0	1453	1	5	3	11.1	2.8	0	10.6	
PLEIADES	II	8	11	37.65	2.48	B	200	200	0	1441	1	2	3	5.4	2.8	0	12.0	
PLEIADES	II	8	12	37.32	2.42	B	250	100	0	1450	1	2	3	10.5	2.8	0	14.1	
PLEIADES	II	8	13	36.70	2.55	B	250	100	95	1440	1	6	3	12.2	2.8	0	14.5	
PLEIADES	II	8	14	36.33	2.95	B	300	100	95	1455	1	3	3	11.7	2.8	0	9.5	
PLEIADES	II	8	15	35.88	2.95	B	200	200	0	1472	1	4	3	12.2	2.8	0	8.3	
PLEIADES	II	8	16	35.60	3.07	B	200	100	90	1471	1	1	3	13.7	2.8	0	7.5	
PLEIADES	II	8	17	35.29	3.04	B	200	100	90	1469	1	1	3	11.7	2.8	0	10.9	
PLEIADES	II	8	18	35.05	3.03	B	200	100	90	1468	1	1	3	13.1	2.8	0	6.7	
PLEIADES	II	8	19	34.70	3.06	B	200	100	90	1467	1	1	3	12.2	2.8	0	7.6	
PLEIADES	II	8	20	34.23	3.00	B	200	100	90	1468	1	1	3	12.2	2.8	0	7.8	
PLEIADES	II	8	21	33.84	2.97	B	200	100	90	1471	1	1	3	10.5	2.8	0	5.5	
PLEIADES	II	8	22	33.28	2.80	B	200	100	100	1476	1	2	3	5.4	2.8	0	13.9	
PLEIADES	II	8	23	33.02	2.74	B	200	200	0	1444	1	6	3	9.4	2.8	0	9.7	
PLEIADES	II	8	24	32.48	2.72	B	200	100	80	1453	1	4	3	10.5	2.8	0	4.8	
PLEIADES	II	8	25	32.17	2.88	B	200	100	90	1471	1	2	3	12.2	2.8	0	8.6	
PLEIADES	II	9	1	37.25	6.15	B	100	100	0	1449	1	3	3	12.2	2.8	0	6.7	
PLEIADES	II	9	2	37.07	5.98	B	100	100	0	1456	1	2	3	12.2	2.8	0	6.2	
PLEIADES	II	9	3	36.97	5.86	B	100	100	0	1457	1	2	3	12.8	2.8	0	9.8	
PLEIADES	II	9	4	36.84	5.74	B	100	100	0	1455	1	2	3	12.2	2.8	0	8.0	
PLEIADES	II	9	5	36.79	5.68	B	100	100	0	1452	1	1	3	12.2	2.8	0	10.1	
PLEIADES	II	9	6	36.75	5.61	B	100	100	0	1453	1	1	3	13.7	2.8	0	11.9	
PLEIADES	II	9	7	36.67	5.52	B	100	100	0	1454	1	6	3	6.3	2.8	0	10.4	
PLEIADES	II	9	8	36.57	5.45	B	100	100	0	1452	1	1	3	6.8	2.8	0	9.8	
PLEIADES	II	9	9	36.47	5.35	B	100	100	0	1454	1	1	3	13.7	2.8	0	10.0	
PLEIADES	II	9	10	36.35	5.25	B	100	100	0	1455	1	2	3	0.0	2.4	30	37.0	
PLEIADES	II	9	12	36.26	5.16	B	100	100	0	1463	1	7	3	2.6	2.4	30	6.4	
PLEIADES	II	9	13	36.23	5.13	B	100	100	0	1461	1	7	2	0.0	2.4	30	37.0	
PLEIADES	II	10	1	35.88	5.59	B	100	100	0	1464	1	2	3	12.2	2.8	0	11.5	
PLEIADES	II	10	2	35.93	5.65	B	100	100	0	1464	1	2	3	13.7	2.4	30	13.1	
PLEIADES	II	10	3	36.00	5.73	B	100	100	0	1462	1	1	2	13.7	2.4	30	13.3	
PLEIADES	II	10	4	36.06	5.83	B	100	100	0	1463	1	1	3	13.7	2.4	30	15.2	
PLEIADES	II	11	1	29.00	5.30	C	400	100	60	1508	1	2	3	15.9	2.8	0	6.8	
PLEIADES	II	11	2	28.70	5.33	C	300	125	60	1511	1	1	3	13.1	2.8	0	7.0	
PLEIADES	II	11	3	28.45	5.35	C	300	100	60	1508	1	1	3	13.7	2.8	0	6.8	
PLEIADES	II	11	4	28.18	5.40	C	250	100	90	1506	1	1	3	13.1	2.8	0	7.0	

Cruise	Station	No.	0°N 86°W		Q	Navigation			Depth			Sed.			Pen.		HFU
			min.	min.		max	min	brg	fm	S	E	N	m		m	T	
PLEIADES	II	11	5	27.70	5.50	C	150	100	90	1507	1	1	3	13.1	2.8	0	4.8
PLEIADES	II	11	6	27.20	5.52	C	150	150	0	1470	1	1	3	13.7	2.8	0	5.2
PLEIADES	II	11	7	26.78	5.55	C	250	75	100	1475	1	1	3	8.5	2.8	0	9.2
PLEIADES	II	11	8	26.48	5.60	C	225	100	120	1463	1	4	3	12.8	2.8	0	11.1
PLEIADES	II	11	9	26.23	5.65	C	275	75	90	1465	1	1	3	4.3	2.8	0	4.4
PLEIADES	II	11	10	25.83	5.70	E	400	400	90	1459	1	2	3	11.7	2.8	0	4.8
PLEIADES	II	11	11	25.40	5.80	E	400	400	0	1478	1	3	3	0.0	2.8	0	10.1
PLEIADES	II	12	1	24.30	5.50	F	300	100	90	1525	1	7	3	4.8	2.8	0	2.3
PLEIADES	II	13	1	35.78	7.85	A	100	100	0	1448	1	1	3	13.1	2.8	0	12.0
PLEIADES	II	13	2	35.78	7.86	A	100	100	0	1448	1	1	2	12.2	2.8	0	10.2
PLEIADES	II	13	3	35.76	8.01	A	100	100	0	1445	1	1	2	12.8	2.8	0	10.1
PLEIADES	II	13	4	35.82	8.10	A	100	100	0	1444	1	1	2	12.8	2.8	0	9.2
PLEIADES	II	13	5	35.80	8.22	A	100	100	0	1442	1	1	2	5.4	2.8	0	8.2
PLEIADES	II	13	7	35.63	8.46	A	100	100	0	1438	1	1	2	11.7	2.8	0	10.8
PLEIADES	II	13	8	35.57	8.60	A	100	100	0	1438	1	1	2	12.2	2.7	15	15.1
PLEIADES	II	13	9	35.56	8.70	A	100	100	0	1438	1	1	2	11.1	2.7	15	12.7
PLEIADES	II	13	10	35.55	8.81	A	100	100	0	1438	1	1	2	10.5	2.4	30	16.7
PLEIADES	II	13	11	35.49	8.74	A	100	100	0	1438	1	1	2	11.7	2.8	0	12.5
PLEIADES	II	13	12	35.46	8.61	B	100	100	0	1438	1	1	2	11.7	2.4	30	13.6
PLEIADES	II	13	13	35.54	8.56	B	100	100	0	1438	1	1	2	11.7	2.8	0	13.4
PLEIADES	II	13	14	35.79	8.62	B	100	100	0	1438	1	1	2	10.5	2.8	0	12.1
PLEIADES	II	13	15	36.21	8.90	B	100	100	0	1462	1	1	2	10.5	2.8	0	10.9
PLEIADES	II	13	16	35.85	8.48	A	100	100	0	1438	1	1	2	10.5	2.8	0	9.3
PLEIADES	II	13	17	35.63	8.45	A	100	100	0	1438	1	1	2	11.1	2.7	15	10.4
PLEIADES	II	13	18	35.52	8.62	A	100	100	0	1438	1	1	1	5.4	2.4	30	37.0
PLEIADES	II	13	19	35.46	8.68	B	100	100	0	1438	1	1	2	11.1	2.4	30	16.3
PLEIADES	II	14	1	39.00	5.79	B	250	100	0	1450	1	6	3	11.7	2.8	0	3.6
PLEIADES	II	14	2	39.28	5.59	B	200	200	0	1460	1	3	3	12.8	2.8	0	4.7
PLEIADES	II	14	3	39.61	5.55	B	250	100	90	1445	1	3	3	0.0	2.8	0	2.7
PLEIADES	II	14	4	39.91	5.50	B	250	100	90	1445	1	5	3	14.2	2.8	0	7.6
PLEIADES	II	14	5	40.49	5.41	B	200	200	0	1418	1	6	3	0.0	2.8	0	4.3
PLEIADES	II	14	6	40.87	5.46	B	100	100	0	1440	1	5	1	8.0	2.8	0	16.2
PLEIADES	II	14	8	41.30	5.51	B	250	100	90	1455	1	5	3	8.0	2.8	0	1.1
PLEIADES	II	14	9	41.98	5.68	B	250	100	90	1426	1	4	3	10.5	2.8	0	1.7
PLEIADES	II	14	11	42.88	5.58	B	250	100	90	1435	1	5	3	8.0	2.7	15	7.1
PLEIADES	II	14	14	43.45	5.50	B	200	200	0	1452	1	3	3	4.3	2.4	30	4.2
PLEIADES	II	14	15	43.70	5.51	B	200	200	0	1450	1	5	3	9.1	2.7	15	3.6
PLEIADES	II	14	16	44.10	5.35	B	250	100	100	1464	1	7	3	5.4	2.4	30	5.7
PLEIADES	II	15	1	40.50	3.10	C	250	100	90	1446	1	3	3	4.3	2.8	0	4.0
PLEIADES	II	15	2	40.70	3.05	C	250	100	90	1450	1	6	3	9.1	2.8	0	3.6
PLEIADES	II	15	3	41.40	3.02	C	200	200	0	1400	1	5	3	8.0	2.8	0	1.2
PLEIADES	II	15	4	41.80	2.95	C	250	100	90	1380	1	5	3	4.3	2.8	0	3.4
PLEIADES	II	16	1	44.97	2.41	B	200	100	90	1390	1	5	3	6.3	2.8	0	3.6
PLEIADES	II	16	2	44.91	2.54	B	200	100	90	1404	1	4	3	8.0	2.8	0	1.7
PLEIADES	II	16	3	44.71	2.71	B	200	100	90	1425	1	3	3	11.7	2.4	30	0.9
PLEIADES	II	17	1	43.15	8.24	B	200	100	90	1393	1	2	2	6.3	2.8	0	11.7
PLEIADES	II	17	2	42.85	8.10	B	200	100	90	1405	1	5	2	5.4	2.8	0	11.8
PLEIADES	II	17	3	42.61	7.66	B	250	100	100	1458	1	4	2	8.0	2.8	0	13.8
PLEIADES	II	17	5	42.41	7.74	B	250	100	100	1463	1	1	2	5.4	2.4	30	8.5

Cruise	Station	No.	0°N	86°W		Navigation			Depth				Sed.	Pen.	HFU	
			min.	min.	Q	max	min	brg	fm	S	E	N	m	m	T	
PLEIADES	II 18	1	42.48	7.02	B	250	100	100	1455	1	4	2	6.8	2.8	0	9.8
PLEIADES	II 18	2	42.33	6.88	B	250	100	100	1460	1	3	2	13.1	2.8	0	5.8
PLEIADES	II 18	3	41.83	7.03	B	200	100	90	1464	1	1	2	11.1	2.8	0	2.9
PLEIADES	II 18	4	41.58	6.92	B	200	100	90	1466	1	1	2	10.5	2.8	0	0.8
PLEIADES	II 18	5	41.38	6.79	C	250	100	90	1414	1	3	2	8.5	2.8	0	1.8
PLEIADES	II 18	6	40.79	6.51	B	200	100	90	1417	1	1	2	7.4	2.4	30	9.1
PLEIADES	II 18	7	40.48	6.20	B	200	200	0	1443	1	3	2	5.4	2.7	15	5.8
PLEIADES	II 18	8	39.95	6.20	B	250	100	90	1433	1	5	2	5.4	2.4	30	10.7
PLEIADES	II 20	1	39.41	8.51	B	200	100	90	1465	1	3	3	19.6	2.8	0	1.4
PLEIADES	II 20	2	39.14	8.38	A	200	100	90	1443	1	4	3	14.8	2.8	0	2.2
PLEIADES	II 20	3	38.72	8.24	A	200	100	90	1480	1	7	3	13.7	2.8	0	2.5
PLEIADES	II 20	4	38.15	7.95	A	200	100	90	1465	1	5	3	8.0	2.7	15	1.7
PLEIADES	II 20	5	37.72	7.67	A	200	100	90	1437	1	3	3	6.3	2.7	15	1.1
PLEIADES	II 20	6	37.35	7.36	A	200	100	90	1442	1	3	3	5.4	2.7	15	5.0
PLEIADES	II 20	7	37.01	6.73	A	200	100	90	1458	1	3	3	11.7	2.7	15	5.6
PLEIADES	II 20	8	36.85	6.41	A	200	100	90	1462	1	1	3	11.1	2.4	30	3.2
PLEIADES	II 20	9	36.71	5.93	A	200	100	90	1463	1	2	3	13.1	2.7	15	7.2
PLEIADES	II 20	10	36.60	5.45	A	200	100	90	1453	1	1	3	12.2	2.7	15	13.1
PLEIADES	II 20	11	36.70	5.11	A	200	100	90	1454	1	1	3	5.4	2.8	0	10.5
PLEIADES	II 20	12	36.80	4.64	A	200	100	90	1457	1	1	3	12.2	2.7	15	9.4
PLEIADES	II 20	13	36.89	4.13	A	200	100	90	1454	1	1	3	8.5	2.4	30	13.3
PLEIADES	II 20	15	36.95	3.60	B	200	100	90	1448	1	1	3	0.0	2.4	30	13.0
PLEIADES	II 21	1	36.41	4.67	A	100	100	0	1462	1	1	3	13.1	2.8	0	9.9
PLEIADES	II 21	2	36.15	4.90	A	100	100	0	1462	1	1	3	11.7	2.8	0	11.5
PLEIADES	II 21	3	35.97	5.13	A	100	100	0	1466	1	2	3	11.7	2.8	0	12.1
PLEIADES	II 21	4	35.80	5.42	A	100	100	0	1466	1	2	3	5.7	2.8	0	15.5
PLEIADES	II 21	5	35.71	5.60	A	100	100	0	1460	1	2	1	0.0	2.4	30	37.0
PLEIADES	II 21	6	35.67	5.70	A	100	100	0	1463	1	2	3	0.0	2.7	15	17.8
PLEIADES	II 21	7	35.60	5.85	A	100	100	0	1463	1	2	3	0.0	2.4	30	12.8
PLEIADES	II 21	8	35.53	6.03	A	100	100	0	1462	1	2	3	0.0	2.7	15	10.2
PLEIADES	II 21	9	35.33	6.19	A	100	100	0	1465	1	1	3	0.0	2.8	0	8.6
PLEIADES	II 21	10	35.13	6.30	A	100	100	0	1463	1	1	3	0.0	2.4	30	13.1
PLEIADES	II 21	11	34.79	6.34	A	100	100	0	1465	1	1	3	4.6	2.8	0	8.8
PLEIADES	II 21	12	34.30	6.64	A	100	100	0	1462	1	2	3	0.0	2.4	30	8.8
PLEIADES	II 21	13	33.98	7.09	A	100	100	0	1461	1	2	3	0.0	2.4	30	8.2
PLEIADES	II 21	14	33.62	7.37	A	100	100	0	1462	1	1	3	0.0	2.8	0	5.3
PLEIADES	II 21	15	33.28	7.65	A	100	100	0	1469	1	3	3	0.0	2.8	0	4.5
PLEIADES	II 21	16	32.97	7.97	A	100	100	0	1484	1	3	3	0.0	2.7	15	2.7
PLEIADES	II 21	17	32.66	8.22	A	100	100	90	1495	1	3	3	0.0	2.7	15	3.8
PLEIADES	II 22	1	39.25	2.77	B	200	200	0	1433	1	3	3	0.0	2.8	0	6.6
PLEIADES	II 22	2	39.24	2.95	B	200	200	0	1434	1	3	3	0.0	2.8	0	6.0
PLEIADES	II 22	4	38.97	3.39	B	200	200	0	1468	1	7	3	0.0	2.8	0	2.5
PLEIADES	II 22	5	38.88	4.09	B	200	200	0	1455	1	1	3	0.0	2.8	0	6.5
PLEIADES	II 22	6	38.98	4.81	A	200	200	0	1467	1	2	3	0.0	2.8	0	8.0
PLEIADES	II 22	7	39.09	5.36	A	200	200	0	1462	1	1	3	0.0	2.7	15	2.8
PLEIADES	II 22	8	39.17	5.98	A	200	200	0	1464	1	3	3	0.0	2.8	0	2.6
PLEIADES	II 22	9	39.23	6.92	A	200	200	0	1448	1	3	3	0.0	2.8	0	3.6
PLEIADES	II 22	10	39.19	7.41	A	200	200	0	1458	1	3	3	0.0	2.8	0	2.6
PLEIADES	II 22	13	40.18	7.86	B	200	200	0	1448	1	3	3	0.0	2.4	30	4.6

Cruise	Station	No.	0°N 86°W		Q	Navigation			Depth				Sed. m	Pen. m	HFU	
			min.	min.		max	min	brg	fm	S	E	N			T	
PLEIADES	II 23	1	41.65	3.17	B	200	200	0	1452	1	3	3	0.0	2.8	0	0.6
PLEIADES	II 23	4	41.34	3.66	B	200	100	110	1468	1	1	3	0.0	2.8	0	1.6
PLEIADES	II 23	5	41.86	4.55	B	200	200	0	1464	1	1	3	0.0	2.8	0	1.1
PLEIADES	II 23	6	41.68	4.99	B	200	200	0	1450	1	5	3	0.0	2.8	0	1.1
PLEIADES	II 23	7	41.69	5.38	B	200	200	0	1456	1	1	3	0.0	2.8	0	1.9
PLEIADES	II 24	1	41.34	6.57	B	200	200	0	1462	1	1	3	0.0	2.8	0	3.2
PLEIADES	II 24	2	41.49	7.04	B	200	200	0	1467	1	1	3	0.0	2.8	0	3.5
PLEIADES	II 24	3	41.75	7.47	B	200	200	0	1464	1	1	3	0.0	2.8	0	1.4
PLEIADES	II 25	1	41.62	7.55	B	100	100	0	1466	1	1	3	0.0	2.8	0	1.1
PLEIADES	II 25	2	41.62	7.77	B	100	100	0	1462	1	1	3	0.0	2.8	0	1.7
PLEIADES	II 25	3	41.82	8.04	B	100	100	0	1459	1	3	3	0.0	2.8	0	1.3
PLEIADES	II 25	4	42.02	8.56	B	100	100	0	1456	1	1	3	0.0	2.8	0	1.4
PLEIADES	II 25	5	42.10	9.01	B	100	100	0	1452	1	1	3	0.0	2.8	0	1.2
PLEIADES	II 26	1	45.12	10.53	B	200	200	0	1512	1	7	3	0.0	2.8	0	1.5
PLEIADES	II 26	2	45.21	10.66	B	200	200	0	1505	1	7	3	0.0	2.8	0	1.2
PLEIADES	II 27	1	39.44	5.83	B	200	200	0	1445	1	2	3	0.0	2.8	0	4.6
PLEIADES	II 27	2	39.13	5.79	B	200	200	0	1455	1	3	3	0.0	2.8	0	3.0
PLEIADES	II 27	3	38.87	5.69	B	200	200	0	1452	1	1	3	0.0	2.8	0	5.6
PLEIADES	II 27	4	38.66	5.44	B	200	200	0	1457	1	1	3	0.0	2.8	0	8.0
PIQUERO	8 6	1	10.80	3.70	F	2000	2000	90	1570	3	8	3	0.0	2.4	0	0.1
PIQUERO	8 7	1	18.20	4.10	F	2000	2000	90	1464	3	8	3	0.0	2.4	0	3.9
PIQUERO	8 8	1	26.30	4.90	F	2000	2000	90	1477	3	1	3	0.0	2.4	0	6.7
PIQUERO	8 9	1	39.10	5.70	F	2000	2000	90	1448	3	6	3	0.0	2.4	0	10.3
PIQUERO	8 10	1	42.50	6.00	F	2000	2000	90	1408	3	3	3	0.0	2.4	0	3.4
PIQUERO	8 11	3	44.30	6.00	F	2000	2000	90	1455	3	4	3	0.0	2.4	0	5.5
SCAN	10 17	1	40.00	8.70	E	2000	2000	90	1456	3	3	3	0.0	2.4	0	2.3
SCAN	10 18	1	37.60	8.10	E	2000	2000	90	1429	3	1	3	0.0	2.4	0	4.0
SCAN	10 19	1	35.30	6.80	E	2000	2000	90	1460	3	1	3	0.0	2.4	0	5.4
SCAN	10 20	1	33.70	6.50	E	2000	2000	90	1449	3	6	3	0.0	2.4	0	9.4
SOUTHTOW	43	1	50.30	8.90	B	200	200	90	1445	1	7	3	0.0	2.4	0	0.3
SOUTHTOW	45	1	50.90	10.10	B	200	200	90	1390	1	3	3	0.0	2.4	0	4.4
SOUTHTOW	45	2	50.60	9.90	B	200	200	90	1425	1	7	1	0.0	0.7	30	2.2
SOUTHTOW	46	1	51.70	8.80	B	400	400	90	1400	1	2	3	0.0	2.4	0	10.4
SOUTHTOW	50	1	45.00	7.40	B	400	400	90	1441	1	2	3	0.0	2.4	0	0.0
SOUTHTOW	50	2	44.80	7.10	B	400	400	90	1443	1	1	3	0.0	2.4	0	1.3
SOUTHTOW	51	1	43.00	6.70	E	1000	1000	90	1384	1	2	3	0.0	2.4	0	12.1
SOUTHTOW	52	1	44.50	8.20	B	400	400	90	1475	1	5	3	0.0	2.4	0	2.4
SOUTHTOW *	53	1	45.70	9.00	A	200	200	90	1370	1	6	3	6.7	2.1	30	4.3
SOUTHTOW	53	2	44.90	8.80	A	200	200	90	1460	1	4	3	5.7	2.2	30	2.1
SOUTHTOW	54	1	45.10	7.80	E	1000	1000	90	1440	1	2	3	11.7	2.5	0	0.8
SOUTHTOW	54	2	44.90	7.80	E	1000	1000	90	1442	1	1	3	8.5	2.5	0	1.1
SOUTHTOW	54	3	44.80	7.80	E	1000	1000	90	1443	1	1	3	9.1	2.9	0	2.2
SOUTHTOW *	54	4	44.50	7.80	E	1000	400	90	1449	1	5	1	15.4	2.5	30	19.0
SOUTHTOW	54	5	44.30	7.90	E	1000	400	90	1485	1	7	3	11.1	2.9	0	5.8
SOUTHTOW	54	6	43.80	8.00	E	1000	1000	90	1465	1	5	3	6.5	2.4	0	6.2
SOUTHTOW	54	7	43.50	8.00	E	1000	1000	90	1394	1	5	2	8.1	2.5	0	17.0
SOUTHTOW	54	8	43.10	8.10	E	1000	1000	90	1392	1	2	2	8.9	2.7	30	8.8
SOUTHTOW *	55	1	42.20	11.50	E	1000	200	90	1462	1	1	3	0.0	2.4	0	1.9
SOUTHTOW *	55	2	42.00	11.60	E	1000	200	90	1445	1	4	3	0.0	2.4	0	3.2

Cruise	Station	No.	0°N 86°W		Navigation			Depth				Sed.	Pen.	HFU			
			min.	min.	Q	max	min	brg	fm	S	E	N	m	m	T		
SOUTHTOW	*	55	3	41.70	11.60	E	1000	200	90	1428	1	6	3	0.0	2.4	0	1.0
SOUTHTOW	*	56	1	40.70	11.50	E	1000	200	90	1475	1	5	3	0.0	2.4	0	2.7
SOUTHTOW	*	56	2	40.50	11.50	E	1000	200	90	1445	1	6	3	0.0	2.4	0	3.0
SOUTHTOW		57	1	36.60	11.90	B	400	400	90	1464	1	1	3	0.0	2.4	0	3.4
SOUTHTOW		57	2	37.00	12.10	B	400	400	90	1464	1	1	3	0.0	2.4	0	4.1
SOUTHTOW		58	1	44.30	7.30	E	1000	1000	90	1466	1	4	3	0.0	2.7	0	2.2
SOUTHTOW		59	1	41.70	8.40	E	1000	500	90	1453	1	1	2	0.0	3.0	0	1.2
SOUTHTOW		59	2	41.50	8.30	E	1000	500	90	1450	1	5	2	13.7	2.4	0	1.3
SOUTHTOW		59	3	41.20	8.40	E	1000	1000	90	1423	1	1	2	10.2	2.4	30	5.9
SOUTHTOW		59	4	40.90	8.70	E	1000	1000	90	1423	1	1	2	10.4	3.0	0	2.7
SOUTHTOW	*	59	5	40.60	8.90	E	1000	1000	90	1438	1	4	2	13.5	3.3	30	2.8
SOUTHTOW		59	6	40.30	9.20	E	1000	1000	90	1448	1	6	2	13.7	3.0	30	2.1
SOUTHTOW	*	60	1	38.30	10.60	E	1000	1000	90	1455	1	5	3	0.0	2.4	0	9.2
SOUTHTOW		61	1	40.30	10.20	E	1000	500	90	1454	1	5	3	0.0	2.4	0	1.8
SOUTHTOW	*	61	2	40.50	9.90	E	1000	500	90	1475	1	3	3	0.0	2.4	0	1.3
SOUTHTOW	*	61	3	40.70	9.60	E	1000	500	90	1495	1	1	3	0.0	2.4	0	1.1
SOUTHTOW		62	1	35.10	11.10	B	400	400	90	1461	1	2	2	12.2	2.4	0	4.6
SOUTHTOW		63	1	39.10	12.00	B	400	400	90	1459	1	3	3	13.5	2.5	0	0.0
SOUTHTOW		63	2	38.80	12.00	B	400	400	90	1457	1	1	3	6.8	2.9	0	5.6
SOUTHTOW		63	3	38.60	11.90	A	200	200	90	1456	1	1	3	8.3	3.0	30	6.1
SOUTHTOW		63	4	38.20	11.90	B	400	400	90	1418	1	1	3	11.1	2.9	0	7.0
SOUTHTOW		63	5	37.90	11.80	B	400	400	90	1420	1	2	2	9.8	3.0	0	11.9
SOUTHTOW		63	6	37.40	11.70	B	500	400	90	1448	1	4	3	11.1	2.9	0	9.5
SOUTHTOW		63	7	37.00	11.50	B	500	400	90	1466	1	1	3	8.9	2.7	0	5.3
SOUTHTOW		63	8	36.30	11.30	B	500	400	90	1463	1	1	3	4.4	2.6	0	2.0
SOUTHTOW		63	9	36.40	11.30	B	600	500	90	1463	1	1	3	0.0	2.7	0	1.9
SOUTHTOW		63	10	36.00	11.10	B	600	500	90	1465	1	1	3	5.4	2.4	30	8.5
SOUTHTOW		63	11	35.80	11.00	B	600	500	90	1454	1	2	3	8.0	2.4	30	7.7
SOUTHTOW		64	1	35.70	10.90	B	400	400	90	1456	1	2	3	0.0	2.4	0	12.7
SOUTHTOW		64	2	35.60	10.90	B	400	400	90	1458	1	2	3	0.0	2.4	0	19.0
SOUTHTOW		64	3	35.40	10.90	E	400	400	90	1461	1	2	3	0.0	2.4	0	8.6
SOUTHTOW		65	1	34.20	10.60	E	1000	1000	90	1463	1	2	3	0.0	2.4	0	9.0
SOUTHTOW		65	2	34.10	10.50	E	1000	1000	90	1463	1	2	3	0.0	2.4	0	10.6
SOUTHTOW		66	1	33.10	11.20	A	200	200	90	1502	1	2	3	0.0	2.4	0	2.3
SOUTHTOW		66	2	32.80	11.10	A	200	200	90	1497	1	1	3	0.0	2.4	0	1.9
SOUTHTOW		67	1	34.50	8.70	E	1000	500	90	1456	1	2	2	14.8	2.7	0	13.0
SOUTHTOW		67	2	34.10	8.70	E	1000	500	90	1462	1	2	2	9.3	2.9	0	10.1
SOUTHTOW		67	3	33.70	8.80	E	1000	300	90	1461	1	3	3	6.1	2.9	0	7.1
SOUTHTOW		67	4	33.40	8.80	A	200	200	90	1470	1	4	3	11.1	2.4	0	5.2
SOUTHTOW		67	5	33.00	8.90	A	200	200	90	1493	1	2	3	8.3	2.4	0	3.6
SOUTHTOW		67	6	32.80	9.00	A	200	200	90	1496	1	2	3	7.6	2.7	0	4.4
SOUTHTOW		67	7	32.40	9.00	A	200	200	90	1494	1	2	3	7.6	2.4	0	8.4
SOUTHTOW		68	1	34.00	13.80	B	1000	1000	90	1465	1	2	5	0.0	11.2	0	6.9
SOUTHTOW		69	1	32.00	10.70	E	1000	1000	90	1478	1	2	3	8.3	2.9	0	6.3
SOUTHTOW		69	2	31.80	10.70	E	1000	1000	90	1474	1	1	3	3.7	2.9	0	7.7
SOUTHTOW		69	3	31.60	10.70	E	1000	1000	90	1473	1	1	3	0.0	2.9	30	9.3
SOUTHTOW		69	4	31.10	10.70	E	1000	1000	90	1452	1	4	2	7.6	2.1	30	30.3
SOUTHTOW		69	5	30.80	10.70	E	1000	1000	90	1477	1	5	3	0.0	2.9	30	8.5
SOUTHTOW		69	6	30.50	10.60	E	1000	1000	90	1486	1	4	3	0.0	2.9	0	8.0

Cruise	Station No.	0°N 86°W		Q	Navigation			Depth				Sed. m	Pen. m	HFU	
		min.	min.		max	min	brg	fm	S	E	N			T	
SOUTHTOW	69	7	30.20	10.50	E 1000	1000	90	1485	1	3	3	11.3	2.4	0	8.0
SOUTHTOW	69	8	29.90	10.40	E 1000	1000	90	1490	1	3	3	0.0	2.9	0	7.6

Galapagos Rift Dive Data	No.	0°N 86°W		Q	Navigation			Depth fm	q _o	w _o	Pen	q _{tot}
		min.	min.		max	min	brg					
ALVIN DIVE 721	1	36.657	5.988		10	10	0	1450	36.7	4.3	0.5	38.7
ALVIN DIVE 721	2	36.593	5.855		10	10	0	1456	9.6	2.7	0.5	9.9
ALVIN DIVE 721	3	36.611	5.851		10	10	0	1458	6.4		0.5	6.4
ALVIN DIVE 721	4	36.618	5.855		10	10	0	1458	21.7	-7.1	0.5	20.0
ALVIN DIVE 721	5	36.634	5.851		10	10	0	1452	15.0	-25.0	0.5	12.2
ALVIN DIVE 721	6	36.642	5.850		10	10	0	1447	33.4	16.1	0.5	43.1
ALVIN DIVE 721	7	36.642	5.848		10	10	0	1444	292.0	70.6	0.5	450.9
ALVIN DIVE 721	8	36.648	5.878		10	10	0	1443	76.0	159.1	0.5	759.1
ALVIN DIVE 721	9	36.650	5.898		10	10	0	1450	13.3	-6.0	0.5	12.2
ALVIN DIVE 721	10	36.646	5.891		10	10	0	1448	39.5	-33.4	0.5	15.8
ALVIN DIVE 721	11	36.642	5.875		10	10	0	1443	9.6		0.5	9.6
ALVIN DIVE 721	12	36.667	5.882		10	10	0	1455	8.1	-40.7	0.5	3.7
ALVIN DIVE 721	13	36.658	5.871		10	10	0	1452	8.4	-33.7	0.5	4.2
ALVIN DIVE 721	14	36.655	5.875		10	10	0	1450	33.4	44.4	0.5	59.1
ALVIN DIVE 721	15	36.647	5.882		10	10	0	1443	99.2	-13.1	0.5	63.9
ALVIN DIVE 732	1	38.604	7.887		10	10	0	1458	2.1	6.7	0.5	2.2
ALVIN DIVE 732	2	38.382	7.901		10	10	0	1464	2.2	3.0	0.5	2.3
ALVIN DIVE 732	3	38.273	7.914		10	10	0	1476	4.2	3.1	0.5	4.3
ALVIN DIVE 732	4	38.159	7.936		10	10	0	1477	2.7		0.5	2.7
ALVIN DIVE 732	5	38.097	7.930		10	10	0	1426	6.0	4.4	0.5	6.3
ALVIN DIVE 732	6	37.926	7.941		10	10	0	1424	2.2	15.7	0.5	2.5
ALVIN DIVE 732	7	37.82	8.035		10	10	0	1427	4.6		0.5	4.6
ALVIN DIVE 732	8	37.771	8.108		10	10	0	1424	24.6	16.8	0.5	34.3
ALVIN DIVE 732	9	37.679	8.194		10	10	0	1425	0.8		0.5	0.8
ALVIN DIVE 732	10	36.993	8.275		10	10	0	1443	41.2	33.2	0.5	53.2
ALVIN DIVE 734	1	34.175	5.641		10	10	0	1451	25.0	1.8	0.5	25.5
ALVIN DIVE 734	2	34.311	5.603		10	10	0	1458	27.4	37.9	0.5	44.5
ALVIN DIVE 734	3	34.307	5.621		10	10	0	1458	23.6	11.7	0.5	27.9
ALVIN DIVE 734	4	34.309	5.660		10	10	0	1459	47.2	88.6	0.5	119.2
ALVIN DIVE 734	5	34.298	5.683		10	10	0	1464	0.64		0.5	0.6
ALVIN DIVE 734	6	34.439	5.605		10	10	0	1458	-0.4		0.5	-0.4
ALVIN DIVE 734	7	34.512	5.921		10	10	0	1457	108.0	168.2	0.5	510.0
ALVIN DIVE 734	8	34.503	6.001		10	10	0	1458	-3.0		0.5	-3.0
ALVIN DIVE 734	9	34.918	5.973		10	10	0	1456	5.7	-35.8	0.5	3.8
ALVIN DIVE 734	10	34.916	5.967		10	10	0	1455	197.6	61.2	0.5	372.3

APPENDIX I-2. Count reduction

The procedure for determining the resistance corresponding to the count data from the DHF instrument is very simple. The amplifier and digitizer in the instrument are assumed to be linear with constant offset and gain during the course of one or more record cycles. The only nonlinearity arises from the Wheatstone bridge, whose excitation voltage is assumed to be fixed. In that case, the simple voltage divider relationship yields the following transfer function:

$$N = a / (1 + (R_o / R_x)) + b$$

where: N is the output count

R_o is the resistance in the upper arm of the bridge,

R_x is the unknown thermistor (or known calibration) resistance

a is the slope of the line, involving the bridge excitation, amplifier and digitizer gains and

b is the intercept at zero counts involving the offsets of the digitizer and amplifier, and the balance point of the bridge.

Solving for the thermistor resistance, we obtain:

$$R_x = R_o / (a / (N - b) - 1)$$

As this is linear in the coefficients, the computation of the two constants a and b from the calibration data is straightforward:

$$a = (F - Z) / (C - D)$$

$$b = (D * F - C * Z) / (D - C)$$

where:

F = counts of upper calibration point

Z = counts of lower calibration point

$$C = (1 + R_O/R_F)^{-1}$$

$$D = (1 + R_O/R_Z)^{-1}$$

R_F = resistance of upper calibration point

R_Z = resistance of lower calibration point

Assuming the following setup for the bridge:

R_O = upper arm of the bridge = 15000. ohms

R_F = "full scale" or upper calibration point = 4000 ohms

R_Z = "zero" (misnomer) or lower calibration = 5000 ohms

This 1000 ohm span corresponds to roughly 5°C ; we find the bridge constants

$$C = (1 + 15000./4000.)^{-1} = 0.21053$$

$$D = (1 + 15000./5000.)^{-1} = 0.25000$$

Table I-2 contains a typical section of the digital record which will serve to illustrate the procedures for data reduction. This data actually came from the recent cruise to Crater Lake, using DHF 5, a close descendent of the instruments used on the Galapagos cruise.

Then before penetration, at record 70,

Z = lower calibration count = 856

F = upper calibration count = $3658 + n*4096$

where n is the number of rollover spans which are observed from the telemetered data PGR record.

TABLE I 2.

Time	Tilt	LoCal	Tilt	HiCal	Tw	Tw	Tl	Tl	T2	T2	T3	T3
70	660	856	659	3658	794	795	819	822	802	803	843	844
71	658	856	660	3656	795	797	820	821	802	803	842	844
72	656	856	658	3656	794	794	819	821	806	807	845	847
73	657	856	658	3658	798	798	824	825	807	807	847	849
74	656	856	656	3657	801	804	1118	1126	1016	1016	909	911
75	656	854	657	3656	803	804	1099	1101	980	979	910	912
76	656	853	656	3657	803	804	1093	1095	971	977	910	912
77	655	852	655	3656	803	804	1089	1093	973	975	911	911
78	654	852	653	3657	803	804	1090	134	1944	1948	1822	0
79	655	852	556	3656	802	804	1089	1091	971	972	911	911
80	653	851	653	3656	803	803	1087	1089	972	973	910	912
81	653	852	653	3655	803	804	1088	1089	973	971	911	911
82	653	851	654	3654	803	804	1086	1089	971	971	910	911
83	652	852	652	3654	803	804	1087	1089	970	972	910	911
84	652	851	650	3656	803	804	1085	1087	971	971	910	911
85	650	852	651	3654	803	804	1086	1087	969	971	910	911
86	650	851	651	3657	804	804	1085	1086	971	970	910	911
87	651	851	650	3657	805	805	1086	1161	862	818	853	853
88	651	852	649	3656	798	799	824	827	811	808	848	849
89	648	853	649	3656	798	799	823	825	813	809	847	848
90	648	854	649	3656	796	797	821	823	804	806	844	846

Let us assume that you observe one rollover from the PGR setup. Then F equals 7754, and the scale is roughly $5^{\circ}\text{C}/(7754-856)\text{counts} = 1 \text{ m}^{\circ}\text{C}/\text{count}$. The transfer coefficients become

$$a = (F-Z)/(C-D) = -1.747 * 10^5$$

$$b = (D*F-C*Z)/(D-C) = 4.4543 * 10^4$$

Putting $N=856$, we obtain $R_x = 15000. / (a/(856-b) - 1) = 5000.$ ohms as a check. The null point of the bridge is calculated to verify that the number of rollovers assigned to the upper calibration point was correct. The value of the resistance will correspond to the comparison resistance in the reference side of the bridge, and should be known from the setup of the instrument. In this case, 0 counts equals 5131.5 ohms, which sounds reasonable. Continuing to calculate the resistances for other variables:

Variable	Record 70		Record 85	
	Counts	Resistance	Counts	Resistance
T_W	794	5009.47	803	5008.09
T_1	819	5005.65	1096	4964.96
T_2	802	5008.24	969	4982.77
T_3	843	5001.98	910	4991.76

In view of the small change apparent in T_3 , it appears no rollovers occurred. This can only be verified with the PGR record. If the upper calibration point is not in the second range (i.e. there was not one rollover) the calculation would yield significantly different resistances.

By way of comparison, the change in resistance of the lowest thermistor, T_1 , is calculated as 40.69 ohms. Were the scale assumed to be linear, with a value of 1000. ohms/ 6898 counts, the calculated resistance change would be 38.71 ohms, or nearly 5 per cent lower than the correct value. For this reason the nonlinear nature of the bridge should be included in calculations using widely spaced calibration data.

We have found that the regression of counts versus $1/\text{time}$ is not significantly different from the regression of temperatures versus $1/\text{time}$, hence in extrapolating the penetration data to steady state it is only necessary to plot the counts, and convert only the extrapolated count to temperature. Extrapolation was carried out for a few of the data reported here, which showed no significant change in the gradients. Therefore the data are all treated as mean gradients and multiplied by a standard conductivity in recognition of the uniformity of sediment type and measurement operations.

Chapter II.
The Hydrothermal Mounds of the Galapagos Rift: Observations
with DSRV ALVIN and Detailed Heat Flow Studies

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ABSTRACT

Ranging in height from gentle hills of less than a meter to steep-sided giants of more than 20 meters, the mounds of the Galapagos Rift are spectacular hydrothermal features. Their internal temperatures have been measured at up to 13°C above the bottom water temperature and total heat flow (conducted plus convected) can be several hundred to several thousand times the normal oceanic values. Fluids, when they discharge from the mound, do so at a very slow rate and at temperatures probably quite near the bottom water temperature. The mounds are principally composed of iron silicates intermixed and encrusted with lesser amounts of manganese oxides. They are generally found in rows, in a uniformly sedimented area above faults or fractures in the crustal rocks which permit fluids to escape from a deep hydrothermal aquifer. The sediment blanket in some way alters the chemistry of the ascending thermal fluids and leads to the development of mounds. The mounds field, covering an area of at least 200 square kilometers and consisting of thousands of individual mounds, is probably less than 300,000 years old; and many of the mounds may be only a few tens of thousands of years old or less.

INTRODUCTION

Hot magma rising beneath sea floor spreading centers provides one of the principal means of cooling of the Earth. As the newly formed crust solidifies and moves away, it loses heat to the overlying ocean water by three separate mechanisms of heat transfer: thermal conduction through the sea floor, the quenching and cooling of extruded lavas, and the discharge of thermal water from cooling dikes and intrusives. Until recently, this last process had not been directly observed, but in early 1977 a series of thermal springs on the Galapagos Rift in the eastern equatorial Pacific was studied in detail with the submersible ALVIN of the Woods Hole Oceanographic Institution. The results of the study of thermal springs on the Galapagos Rift axis have been published elsewhere (Corliss et al., 1979; van Andel and Ballard, 1979; Crane and Ballard, 1979). Here we describe our visual observations of a series of hydrothermal mounds on the south flank of the Galapagos Rift, discuss the results of detailed heat flow studies of these mounds, and venture some initial hypotheses regarding their formation.

The mounds field, first noted by Klitgord and Mudie (1974) and described in some detail by Lonsdale (1977) on the basis of observations with deep-towed geophysical instruments, occurs in a region 18 to 32 km south of the Galapagos rift axis at 86° W (Figure II-1). The field extends at least 27 km in an east-west direction and may be appreciably longer. Given the spreading rate of the Galapagos Rift of 3.4 cm/yr (Klitgord and Mudie, 1974), the basement age in the mounds area ranges from 0.5 m.y. in the north to 0.9 m.y. at the

Figure II-1. Location of area of investigation in the Panama Basin, eastern equatorial Pacific, Bathymetric contours are in uncorrected meters at a 100-m contour interval and are from Van Andel et al.(1971).

southern edge. The sedimentation rate in the deep Panama Basin is high (about 50 m/m.y.) and the sediment thickness in the mounds area increases southward from 20 to 50 m. Numerous north-facing normal faults, with displacements up to 100 m, divide the flank into 2-4-km-wide fault blocks with a steep north slope and a gentle south flank. These principal structural elements appear to have been fully developed within 10 km from the axis (Allmendinger and Riis, 1979). A finer scale of faulting with displacements of less than 10 m is evident within the fault blocks; in the mounds area minor relief resulting from this fault is completely buried.

The mounds field lies within an area of very high heat flow from which Williams et al. (1974) deduced that it might have a hydrothermal origin. On the basis of these studies, carried out in 1972, a multi-institutional study of hydrothermal processes on the Galapagos Rift was developed, beginning with expedition Pleiades in 1976. During this expedition detailed deep-towed geophysical, photographic, heat flow, and coring studies were carried out on the Galapagos Rift crest and in the mounds field (Corliss et al., 1978). This laid the foundation for a major study in early 1977 with the DSRV ALVIN, supported by the research vessels KNORR and LULU of the Woods Hole Oceanographic Institution, Woods Hole, Mass. This expedition included seven ALVIN dives in the mounds area as well as more heat flow, photographic and coring work from the R. V. KNORR. During the dives ALVIN visited approximately 80 individual mounds, collected samples of hydrothermal waters and sediments, made measurements of heat flow and

water temperature, and took thousands of photographs. Additionally, the drilling ship GLOMAR CHALLENGER drilled four holes in the mounds area, some of which penetrated individual mounds (Hekinian and Rosendahl, 1977; Hekinian et al., 1978).

FIELD OPERATIONS AND DATA SETS

The initial studies in 1972 and 1976 (Klitgord and Mudie, 1974; Williams et al., 1974; Lonsdale, 1977) clearly delineated the prospective target areas in the mounds field. In addition, a detailed precision bathymetric survey with the multi-narrowbeam system of the U.S. Navy Oceanographic Office supplied a topographic base that displays details of the mound area geography. The ALVIN dives were precisely navigated with anchored acoustic transponders, and a precision pressure depth measuring system coupled with an acoustic altimeter provided a fine scale microtopography along the dive tracks. The microtopographic data, recorded digitally together with geochemical data, are not yet available. This has not only prevented the final editing of ALVIN dive tracks but also the reconciliation of the ALVIN navigation grid with the precision bathymetric chart; consequently our discussion of the morphology of the mounds remains to some extent preliminary and is not yet backed by the precise microtopographic detail of which the system is capable.

The field operations of ALVIN, with the exception of the use of the heat flow probe, were conventional and have been described in detail by Ballard and van Andel (1977). Considerable use was made of ALVIN's

side-scanning sonar which, when near the bottom, displayed with great clarity the various rows of mounds in the vicinity (Figure II-2) and allowed the crew to select various mounds for close inspection. In addition to the standard external camera system equipped for color photography, and hand-held cameras, ALVIN was also equipped with a variety of temperature sensing and recording equipment, including a heat flow probe, water sampling devices, a temperature, salinity, oxygen and conductivity monitoring apparatus, and sediment sampling equipment. The following discussion of the mounds relies primarily on visual observations, analysis of the ALVIN external photography, surface ship and ALVIN heat flow data. The diving program was limited to seven dives on a total of fewer than 80 mounds, whereas the DEEP TOW map shows more than 180 chains of mounds. Thus our discussion relies on observations of only a small portion of the total field.

DESCRIPTION OF THE MOUNDS

The mounds field contains a very large number of mounds which present a spectacular manifestation of the discharge of thermal water. Most mounds occur in rows (Figure II-3), aligned approximately east-west, parallel to the axial trend of the Galapagos Rift. Some rows, however, deviate more than 30° from this direction, a spread of azimuths similar to that of the fault patterns of the rift axis and upper flanks where individual faults also deviate up to that amount from the regional trend (Allmendinger and Riis, 1979 ; van Andel and Ballard, 1979). Some mounds are isolated rather than in rows. The mounds range in height from less than 1 m to more than 20 m (Figure II-4).

Figure II-2. ALVIN scanning sonar display of rows of mounds. ALVIN is in the center of the screen; top of screen is 0° relative bearing. (Left) A row of mounds 30 m ahead of ALVIN; sonar is on 50-m range setting. (Right) ALVIN is sitting on a row of mounds; a second row is visible 100 m off ALVIN's port beam. Sonar is on 150-m setting.

Figure II-3. Mounds field. Bathymetry in uncorrected fathoms (1 fathom = 1.83 m) with a 5-fathom contour interval. Heavy dashed lines are rows of large mounds as located by Scripps Deep Tow. Dashes do not indicate individual mounds but only general location of rows. Rows of small mounds are not shown. Solid lines are ALVIN tracks for seven numbered dives. Light dashed lines indicate approximate boundaries of mounds areas.

Smaller mounds (< 4 m high) tend to have gentler slopes (less than 30°) than higher ones (above 4 m), which exhibit slopes up to vertical. In plane view, mounds vary from roughly circular, to very elongate and ridgelike. Craggy knobs or pinnacles, ranging in height from 0.5 to 6 m, occur on the tops of many mounds (Figure II-5). These knobs and pinnacles consist of an earthy, dark-green to brown material that is relatively soft. Dark, ribbon-like bands, often in intricate arabesque patterns, mark the upper portions of all mounds where the slope is gentle (Figure II-6). The bands are flush with the surface or slightly (up to 1 cm) elevated. The visible width is usually 10 cm, and upon excavation the thickness of the bands appears to vary from 1 to more than 6 cm. The material ranges from soft and crumbly to fairly hard. These dark arabesques are ubiquitous and are, on small mounds, the only feature that distinguishes them from the normal abyssal sediment surface.

On the steep middle and upper slopes of large mounds, massive blocky outcrops of dark-brown, obviously relatively coherent material occur quite commonly (Figure II-7). Often these outcrops show distinct layers parallel to the mound surface implying erosion or collapse. Slump scars and zones of debris of hydrothermal crust material give further evidence of extensive erosion or collapse over the entire upper part of the mounds including near the knobs.

In at least one case, a series of several major knobs was seen to arise from a set of subparallel outcropping vertical slabs extending along the crest of an elongate mound.

Figure II-4. Profile sketches of typical mounds as observed from the ALVIN . See text for explanation.

Figure II-5. Composite photographs of a pinnacle. The photograph has been put together from four photographs taken as ALVIN was ascending to the top of a pinnacle. Note the dense biological community attracted to the pinnacle. These animals are typical representatives of the usual deep sea benthic community and apart from the unusual density of life are not uncommon species.

Figure II-6. Photograph of dark arabesque bands. These bands are outcropping ridges of manganese oxide crust found only on the hydrothermally altered sea floor. The presence of these bands enabled diving scientists to differentiate between small sediment swells and small or incipient mounds. High thermal gradients were only found in the presence of this crust.

Figure II-7. Photograph of a blocky outcrop on the steep side of a large mound with a sponge (20-30 cm tall) attached.

The seven ALVIN dives traversed an aggregate of several kilometers of mound surface of all types. Nowhere along these tracks did the continuous water temperature monitor reveal the presence of bottom water temperature anomalies like those invariably associated with the thermal springs of the crustal rift valley 18 km to the north (Corliss et al., 1979). However, a slow and temporary flow of warm water was often observed in skidmarks and shallow excavations made by ALVIN, suggesting a slight overpressure of the interstitial water sealed in, or at least strongly impeded by, the surface material of the mounds. Also, the presence of hydrothermal deposits at and just below the surface, and the results of the heat flow studies demand a significant upward flow of thermal water that must eventually be released through the surface. Such a release may be accomplished by means of slow diffusion through the entire mound surface, or may be more concentrated in specific "vent" areas. In either case the flow must very slow or temporary because no measurable temperature anomalies are created. The three types of outcrops of hydrothermal deposits, the black bands or arabesques, the blocky outcrops, and the knobs and pinnacles are candidates for such localized areas of thermal flow. In fact these features are difficult to explain without calling on hydrothermal flow. On several pinnacles, clusters of small tubes were observed (Figure II-8), 2-4 mm in diameter and 1-6 cm long, which protruded from vertical or overhanging sides of a pinnacle. The tubes consisted of fragile, earthy, yellow material and frequently were observed to shed a dusting of that material on the solid surface below them. These tubes

Figure II-8. Photograph of clusters of yellow tubes on the side of a pinnacle. The brittle starfish has a diameter of about 15 cm. Note the light (yellow) stain on sediment beneath tube. These tubes may be small vents.

possibly represent small vents, but, unfortunately, this cannot be confirmed, since at the time they were observed, the temperature monitor was not functioning.

The exposure of knobs, blocky outcrops, and arabesque bands, despite the rapid sedimentation in this area, requires that these features be either very young, continuously renewed, or exposed by slumping and erosion of the overlying pelagic sediments by slow bottom currents. Since there is no evidence of excessive thickness of pelagic sediments piled around the mounds, we favor the continued growth of the mounds by hydrothermal input.

The mounds support a dense and active biological community, mainly concentrated on the hard surface of blocky outcrops and pinnacles (Figure II-5). The composition of these communities is restricted to species normally found in these depths (about 2700 m), and their location on the hard substrate afforded by the mounds appears entirely opportunistic; hence, they are not comparable to the very unusual communities of the thermal springs in the Galapagos inner valley (Corliss et al., 1979).

SURFACE SEDIMENTS AND INTERNAL STRUCTURE

Morphological variations of the mound deposits reflect the mechanical degradation of competent crusts and semilithified (Fe-hydroxides) clays, while color variations are due to the differences in the mineralogical and chemical composition of the hydrothermal ferromanganese phases, the oxidation state of iron and the

degree of admixture of the normal pelagic siliceous-calcareous oozes of the region. The analysis of the samples collected by ALVIN is not complete, but the results on material dredged in 1976 by the Pleiades expedition (Corliss et al., 1978) can be correlated with dive observations and descriptions of samples collected by ALVIN. Corliss et al. (1978) identified three principal phases in their samples: (1) relatively pure Mn minerals, todorokite and birnessite, of exceptional crystallinity, (2) well-crystallized nontronite, and (3) amorphous Fe hydroxides. The todorokite and birnessite samples were black, sometimes porous, sometimes in the form of dense plates and crusts varying between 1 and 6 cm thick. The nontronite samples consisted of dark-green to yellow-green irregular crusts and lumps which commonly had Mn-rich crusts. The amorphous Fe-hydroxide samples primarily occurred as orange coatings on some of the birnessite-todorokite crusts.

The mounds visited by ALVIN were generally covered with an unlithified sediment which, judging from ALVIN cores, appears to be the normal biogenic sediment of the region with a strong admixture of ferromanganese oxides. The dark arabesque bands are comparable to the Mn-rich birnessite and todorokite crusts. In some cases, these crusts occur in the form of thin, scalloped layers barely visible through the partial cover of overlying pelagic sediment, suggesting that this birnessite and todorokite crust is a nearly ubiquitous phenomenon. The blocky outcrops appear to consist of layers of this crust and semilithified Fe-rich nontronite. The pinnacles are nontronite coated with the black crust (Figure II-9). The nontronite ranges in color

Figure II-9. Photograph of a small excavation on a knob (about 1 m tall) Note the layering of the deposits. The bulk of the knob is the dark green nontronite mud.

from bright green and yellow to chocolate brown. Cores and ALVIN tracks sometimes show the dark birnessite and todorokite crust separated from the nontronite by yellow intermediate zones. Particularly on mounds with well-defined pinnacles, bright-yellow layers occur of what is most probably amorphous Fe-hydroxide. On several occasions, on both large and small mounds, cavities were found under a fairly competent crust of dark Mn-oxides.

The strong fractionation of Fe and Mn represented by the three main phases must result from Eh and possibly pH gradients set up by mixing of oxygenated bottom water and migrating, low-oxygen hydrothermal fluids within the mound. Under the slightly reducing conditions just below the crust surface, Fe-rich nontronite is likely to be the stable phase removing iron and silica from solution. Manganese will precipitate at somewhat higher Eh and form relatively pure crusts near the mound-water interface. With time and continuing deposition of crusts, the stability boundary for the Mn- and Fe-rich phases migrates upward, causing the dissolution of previously deposited Mn and redeposition near the surface. Thus it seems likely that mounds which have the green, reduced nontronite close to the surface have the most active hydrothermal flow. If the flow were to diminish, the more highly oxidized layer would thicken and Mn deposition could take place at greater depth. The lower Eh indicated for the interior of the mounds limits the amount of mixing of bottom water into the mound, and suggests that all of the pore water in the mounds is of hydrothermal origin--probably ultimately sea water, but of deeper circulation than just within the mounds.

All of this concerns only the upper surface of the mounds to a depth of a few tens of centimeters. The nature of the deposits at greater depth and the position of the base of the mounds are far more uncertain.

Near-bottom acoustic reflection records show that the volcanic basement is continuous under the mounds, although most mounds appear to be located above small (a few meters) vertical basement offsets, presumably faults (Lonsdale, 1977; Klitgord and Mudie, 1974). In the regional sediment cover, two prominent reflectors occur near depths of about 7 and 17 m. These have been interpreted as ash layers (Lonsdale, 1977) or a single thick layer of green hydrothermal mud (Hekinian & Rosendahl, 1977, see below) but may equally well represent the normal banding of equatorial pelagic calcareous deposits that is due to changes in carbonate dissolution (Berger & Mayer, 1978). In some cases where basement reflections continue beneath the mounds, these reflectors are interrupted or bowed upward more than can reasonably be ascribed to velocity changes within the mound deposits. This could occur either through disruption of the reflectors at the time they formed regionally, or through more recent growth of mound deposits within or below the reflector horizons. Thus the acoustic reflection data provide little information regarding the level and timing of initial mound formation.

Four holes were drilled by the Deep Sea Drilling Project in the mounds area (Hekinian et al., 1978). Two of these (sites 424B and 424C) were capped by several meters of normal biogenic sediment and are

probably on a lower slope or between mounds. The other two (424 and 424A) begin in a mixture of Fe-Mn crust fragments and green hydrothermal mud very similar to that recovered in two piston cores of the mounds (Figure II-10). This material does not occur in the upper few tens of centimeters of the mounds. This material may have been sucked up in the coring and drilling process but alternatively could form the body of the structures, either in the form of layers or as mixtures, or they may be slump deposits beneath the mound flanks. In the two drill sites (424 and 424A) this lithology extends approximately 30 m below the sea floor, where it overlies about 15 m of normal pelagic ooze above unaltered basement of fragmented pillow basalts and massive sheet flows. The other two DSDP drill holes (424B and 424C) encountered green hydrothermal mud beneath several meters of normal pelagic ooze. They were reportedly drilled between mounds. However, they are located in an area that contains numerous small mounds; hence, their position relative to the small mounds cannot be determined precisely.

HEAT FLOW

A total of one hundred and forty-four successful heat flow measurements have been made in the mounds area (Figure II-11). One hundred and nine were made with a 2 1/2-m multipenetration probe with three outrigger thermistors. Six were made with four to five outrigger thermistors mounted on piston cores from 6 to 9 m long. Twenty-nine were made using a 55-cm-long, three thermistor probe deployed by ALVIN's mechanical arm (Figure II-12). In addition six measurements were made by ALVIN just north of the mounds area on dive 732. The

Figure II-10. Photograph of a core of a mixture of fragments of manganese oxide and nontronite from piston core on or near a mound.

Figure II-11. Map of heat flow measured in the mounds field and adjacent areas. Numbers are heat flow values in heat flow units. ALVIN heat flow measurements are not shown; only the conductive component of heat flow is given. Bathymetry is in uncorrected fathoms with a 5-fathom contour interval.

Figure II -12 . Photograph of the ALVIN heat flow probe being pushed into the side of a pinnacle by ALVIN's mechanical arm. The photograph shows about 1.5 m of the vertical extent of the pinnacle.

older data are published in Williams et al. (1974), the ALVIN and piston core data are in Table II-1, and the remainder are found in Appendix I-1.

We deal with the ALVIN measurements first. As can be seen in Figure II-13 many of the thermal gradients are strikingly nonlinear. The most probable explanation for the nonlinear gradients in materials lacking large thermal conductivity contrasts is the movement of interstitial fluids. In the following paragraphs we will attempt to determine the total heat flow conducted plus convected from the mounds. To do so, we need to consider the complete physical system and impose constraints consistent with our geologic knowledge to yield a simple interpretational model. The observations summarized in other sections make it reasonable to assume there is a source of hydrothermal fluids in the basaltic crust beneath the mounds providing the materials from which the mounds grow. Since this process has been going on for at least a few thousand years, the upper part of the mound may be close to thermal steady state. Because of the standing water interface, the pressure gradient within the upper tens of cm can only vary vertically and the vertical flow must be constant as shown by Lapwood (1948). Isotherms and isobars within the upper layer closely parallel the topographic surface of the mound. The fluid flow then forms a thermal boundary layer whose thickness is proportional to the flow rate w .

Figure II-13. Examples of thermal gradients measured by ALVIN.

The equations governing fluid flow and heat transfer within this layer are given by Wooding (1960). The full equations are given in Chapter III, with the numerical modeling results. In the simplified case considered here, the thermal energy conservation equation becomes,

$$(II-1) \quad 0 = \nabla \cdot \underline{q} = \kappa_m \frac{d^2 T}{dz^2} + w \frac{dT}{dz}$$

with $\kappa_m = \frac{K}{(\rho c_p)_f}$ and subject to the boundary condition: $T = 0$ at $z = 0$ and $q = \text{constant}, q_0$, everywhere,

where

z the vertical distance (positive downwards),

K the wet matrix thermal conductivity,

w the vertical volume flux per unit area (positive upwards),
the fluid density,

c_p the fluid heat capacity,

q the vertical heat (flux positive upwards),

the solution with a constant heat flux q_0 and zero anomalous temperature at $z = 0$ is

$$(II-2) \quad T(z) = \frac{q_0}{(\rho c_p)_f w} (1 - e^{-wz/\kappa_m})$$

This simple exponential profile can be fit to our measurements to provide an estimate of the fluid velocity and total heat transfer across the boundary where we made our observations. The stability of the vertical flow for such a model was studied by Wooding (1960). In the case $w > 0$, the temperature approaches $q_0/(\rho c_p)_f w$ far from the surface, where the heat flux is entirely convective. In the case $w < 0$, curvature is positive and the physical situation pertains to a layer of

thickness L , with a basal temperature

$$(II-3) \quad T(z=L) = \frac{q_0}{(\rho c_p) f w} (1 - e^{-wL/\kappa})$$

If we have three temperatures, T_u , T_m , and T_L equally spaced at Δz , we can find an expression for

$$(II-4) \quad \frac{q_0}{(\rho c_p) f w} = \frac{T_m^2 - T_u \cdot T_L}{2T_m - T_u - T_L}$$

We can separately find the velocity as

$$(II-5) \quad w = \frac{\kappa}{\Delta z} \ln \frac{T_m - T_u}{T_L - T_m}$$

Thus our estimate of w depends linearly on the modified thermal diffusivity and inversely on the spacing of the observations. The heat flux, q_0 , is calculated from equation 4. Hence q_0 is always positive, and depends linearly on the assumed κ_m . From the mean thermal conductivity values on the piston cores of 1.78×10^{-3} cal/cm-sec- $^{\circ}$ C, we estimate a thermal diffusivity for the admixture of Mn oxides, nontronite, and biogenic ooze of the mounds of 2×10^{-3} cm²/sec. Using this value, Table II-1 was constructed. It can be seen that small nonlinearities (e.g. measurements 721-5,-9) yield small corrections to the conductive heat flux calculated solely on the basis of the upper gradient, whereas large nonlinearities (e.g. 721-7,-15) yield large corrections.

Table II-1. Alvin and Piston Core heat flow data

Galapagos Rift Dive Data	No.	Depth fm	q _{cond}	w _o	q _{tot}
ALVIN DIVE 721	1	1450	36.7	4.3	38.7
ALVIN DIVE 721	2	1456	9.6	2.7	9.9
ALVIN DIVE 721	3	1458	6.4		6.4
ALVIN DIVE 721	4	1458	21.7	-7.1	20.0
ALVIN DIVE 721	5	1452	15.0	-25.0	12.2
ALVIN DIVE 721	6	1447	33.4	16.1	43.1
ALVIN DIVE 721	7	1444	292.0	70.6	450.9
ALVIN DIVE 721	8	1443	76.0	159.1	759.1
ALVIN DIVE 721	9	1450	13.3	-6.0	12.2
ALVIN DIVE 721	10	1448	39.5	-33.4	15.8
ALVIN DIVE 721	11	1443	9.6		9.6
ALVIN DIVE 721	12	1455	8.1	-40.7	3.7
ALVIN DIVE 721	13	1452	8.4	-33.7	4.2
ALVIN DIVE 721	14	1450	33.4	44.4	59.1
ALVIN DIVE 721	15	1443	99.2	-13.1	63.9
ALVIN DIVE 732	1	1458	2.1	6.7	2.2
ALVIN DIVE 732	2	1464	2.2	3.0	2.3
ALVIN DIVE 732	3	1476	4.2	3.1	4.3
ALVIN DIVE 732	4	1477	2.7		2.7
ALVIN DIVE 732	5	1426	6.0	4.4	6.3
ALVIN DIVE 732	6	1424	2.2	15.7	2.5
ALVIN DIVE 732	7	1427	4.6		4.6
ALVIN DIVE 732	8	1424	24.6	16.8	34.3
ALVIN DIVE 732	9	1425	0.8		0.8
ALVIN DIVE 732	10	1443	41.2	33.2	53.2
ALVIN DIVE 734	1	1451	25.0	1.8	25.5
ALVIN DIVE 734	2	1458	27.4	37.9	44.5
ALVIN DIVE 734	3	1458	23.6	11.7	27.9
ALVIN DIVE 734	4	1459	47.2	88.6	119.2
ALVIN DIVE 734	5	1464	0.64		0.6
ALVIN DIVE 734	6	1458	-0.4		-0.4
ALVIN DIVE 734	7	1457	108.0	168.2	510.0
ALVIN DIVE 734	8	1458	-3.0		-3.0
ALVIN DIVE 734	9	1456	5.7	-35.8	3.8
ALVIN DIVE 734	10	1455	197.6	61.2	372.3
KNORR 64 PC7		1467	27.5	2.3	34.9
KNORR 64 PC8		1459	12.3	...	12.3
KNORR 64 PC12		1439	16.6	0.6	18.1
KNORR 64 PC15		1445	29.4	2.8	32.0
KNORR 64 PC16		1466	26.1	4.4	43.2
SOUTH TOW PC12		1465	4.6	14.0	64.0

q_{cond} the heat flux (HFU) obtained by multiplying the mean gradient for the three thermistors by the mean sediment conductivity of 1.78 TCU.

w_o the mass flux (micrograms/cm²/sec) derived from a model calculation fit to the observed temperature probe data

q_{tot} the total heat flux (HFU) including both conduction and advection.

Where the simple model does not fit the data, the mean gradient and heat flow are given without a water flux.

Eight of the thirty-five measurements yielded gradients which do not fit this simple model. Some of these showed obvious heating of the upper thermistor, apparently from water flowing up the hole made by the probe, while others showed maximum temperatures at the middle sensor, and hence could not fit a monotonic one-dimensional model. Errors in temperature measurements and probable thermal conductivity variations can account for nonlinearities interpreted as velocities in the ALVIN measurements below about 10^{-5} cm/sec. Thus only eighteen of the ALVIN measurements can be considered to require vertical flow. Of these, twelve indicate an upward flow of fluids and add to the conducted heat flow value and six yield downward flow subtracting from the conducted value. We are not positive of the downward flows implied by these measurements. They tend to occur on mound flanks and could represent the entrainment of bottom waters in the final stages of hydrothermal discharge. The inferred total heat flow values (q_0), for all the measurements we applied this model to, vary between 15 and 759 HFU, and imply fluid velocity at the axis of mound chains to be as high as 1.6×10^{-4} cm/sec.

The six piston core measurements shown in Figure II-15 and listed in Table II-1 have been handled with the same one-dimensional theory. Since they cover a greater depth range they are sensitive to lower velocities. We do not have adequate measurements for the thermal conductivity of the hydrothermal phases.

Figure II-14. Curves and geometry for one-dimensional flow model.

Explanation in the text. The slow velocities of interstitial water motion consistent with the observations imply that all of the heat transfer is taking place conductively exactly at the sediment/water interface. Thus the thermal gradient increases or decreases with depth depending on the sign of the water velocity and its contribution to total heat transfer.

Figure II-15. Thermal gradients from piston core heat flow stations. Numbers refer to piston core stations of KNORR cruise 64, except for number 62, which is from South Tow expedition (Williams, et al. 1974).

The remaining 109 measurements were made with the 2 1/2-meter probe. Although several of these showed nonlinear gradients, we do not feel that the quality of the data is sufficient to rule out errors resulting from experimental difficulties such as a bent probe or thermal conductivity variations as an explanation. Therefore, only the conducted heat flow values are presented in Figure II-11. A numerical average of these values would be somewhat misleading. Besides omitting the convected component of heat, they are not evenly distributed. Also, because of excessive tilt of the heat flow instrument many are only minimum values. Instead of averaging the values we have attempted to determine if the measurements could be grouped in some way. Figure II-16 is a histogram of all 138 mounds area measurements that yielded acceptable thermal gradients. There are 33 values above 14 HFU, ranging up to values of several hundred HFU. If these were plotted on a histogram in increments of 2 HFU they would reduce its usefulness; therefore, we have grouped these together, but the reader should realize this when looking at Figure II-16.

The location of ALVIN measurements relative to a mound or group of mounds is known to a fair degree of accuracy. Values above about 10 HFU seem to occur only on or immediately adjacent to mounds. Therefore, we have divided the histogram in Figure II-16 into two groups. The first are those 79 measurements above 10 HFU which we call "mound high heat flow." We speculate that nearly all of these measurements are on or immediately adjacent to mounds. Several of the measurements not plotted on the histogram would also be in this group. These were stations on top of mounds in which rapid upward flow of fluids along the probe is indicated.

Figure II-16. Histogram of mounds heat flow data. Class interval above 14 HFU includes all values >14 HFU, some of which range up to over 700 HFU. All values taken from within the areas outlined in Figure II-3.

The second group are those values that are clearly in the mounds area but apparently not adjacent to mounds. This is a group containing 59 measurements, which we call "mounds background." Its upper limit is 10 HFU, and it contains values down to 3.2 HFU. Most of the lower values are clustered near the northern boundary of the mounds area and quite possibly could be outside this roughly defined boundary.

The boundaries between these two groups are by no means distinct. We have drawn it at what we believe to be the best point, but by nature it is speculative. Nevertheless, it does provide a framework for understanding the thermal regime of the mounds area.

The "mounds background" heat flow presumably represents heat being transferred from a thermal aquifer through a cap rock, primarily by conduction. Using the heat conduction equation, $q = K \Delta T / \Delta Z$ we can do a parametric study of the subseafloor thermal regime. The "mound background" heat flow is q , typically 8 HFU. The thermal conductivity, K , is about 2×10^{-3} cal/cm-sec- $^{\circ}\text{C}$ for sediments and about 4.5×10^{-3} cal/cm-sec- $^{\circ}\text{C}$ for basalt lavas. The temperature difference between the seafloor and the top of the thermal aquifer feeding the mounds is ΔT , and the distance over which this ΔT occurs is ΔZ . If the top of the thermal aquifer is at the sediment-basalt interface (35 meters beneath the seafloor) the temperature there is approximately 16°C . If the top were 100 meters deep it would be 24°C or 183°C at 1 kilometer. We have measured temperatures of 15°C at less than 1 m beneath the mound surface. Our one-dimensional flow model suggests that fluids entering a mound at between 15° and 25°C would be sufficient to explain our observations.

Fluids rising from a deep circulation system along a fissured zone will lose heat laterally to the rock mass surrounding them and cause elevated heat flux around the vent. This problem was addressed by Sleep and Wolery (1978) and the near vent heat flux for a variety of idealized geometries was calculated. We note that a particularly simple form of the heat flow distribution arises for a conductive slab bounded by a flowing fissure with a linear gradient and by the isothermal heat reservoirs of the ocean above ($T = 0$) and the circulation at depth below ($T = T_1$). Fluids enter the fracture at $T = T_1$ and $z = 0$ and exit at $T = T_0$ and $z = L$. An anomalous temperature field is caused by the existence of the flowing fracture and is superimposed upon a normal temperature gradient within the slab of $T(z) = T_1(1-z/L)$. The problem has a simple analytical solution when the temperature within the fracture T_f can be approximated by $T_f(z) = \frac{z}{L}(f - 1) + T_1$ where $f = T_0/T_1$. Solving the steady conduction equation $\nabla^2 T = 0$, we find

$$\begin{aligned}
 \text{(II-6)} \quad T(y, z) &= \frac{2T_1b}{\pi} \sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n} e^{-n\pi y/L} \sin \frac{n\pi z}{L} + T_1(1-\frac{z}{L}) \\
 &= T_1 \left(1-\frac{z}{L} + \frac{2b}{\pi} \tan^{-1} \frac{e^{-\pi y/L} \sin(\pi z/L)}{1+e^{-\pi y/L} \cos(\pi z/L)}\right)
 \end{aligned}$$

Heat flow at the upper surface of the slab, as a function of distance y from the fracture, is then given by

$$\text{(II-7)} \quad q(y, z=L) = \frac{KT_1}{L} \left(1 + \frac{2f}{(e^{\pi y/L} - 1)}\right)$$

In Figure II-17 we have plotted the quantities of steady state anomalous heat flow (q) versus distance from the fracture (y), for $f = 1$ where $q_0 = 8$ HFU, the heat flow in a slab without a fracture. From this we can see that as the thickness of the slab increases so does the width of the heat flow anomaly. In Figure II-17 we have plotted three examples using equation 7. They do not correspond exactly with the simple parametric study we gave earlier in the section because the flowing fracture model uses a single thermal conductivity instead of the more realistic two-conductivity model. Nevertheless, the results are a good approximation. The broad anomaly derived from a 1000-m-thick lid, and $f = 1$ is probably an extreme. Water would enter the fracture at over 180°C , and it apparently enters the mound around 20°C , yielding $f \sim 0.1$. Also the assumption of steady state thermal conditions begins to break down for the 1000-m-thick lid. The middle curve in Figure II-17 is probably the broadest anomaly we would expect. This size anomaly, especially with the close fracture spacing we see in the mounds area, can lead to the large scatter in observed heat flow and implies that 8 HFU may be an artificially high "mounds background" heat flow value. The heat flow values in the "mounds high heat flow" group, as stated earlier, are presumably from locations near mounds and within the heat flow anomaly caused by the flowing fractures.

How does the observed total heat flow in the mounds area compare with the predicted heat flow? The age of the mounds varies from 0.5 to 0.9 m.y. Using the Parsons and Sclater (1977) model, the predicted heat flow is $q = 11.3\sqrt{t}$, where t is the crustal age in million years.

Figure II-17. Three examples of heat flow near a flowing fracture using (II-7). The assumed mounds background heat flow is 8 HFU. The three curves are explained in the text.

The heat flow in the mounds area should then vary between 12 and 16 HFU. The mounds background heat flow is observed to be 8 to 10 HFU. The high heat flow of the mounds may occur on 5% of the mound area. If their average heat flow is 100 HFU, the average total heat flow for the mounds area would be 13 to 15 HFU, or in other words, approximately the predicted value. This coincidence is probably accidental, since in light of the overall pattern, part of the heat flux in the mounds is derived from lateral flow.

DISCUSSION

The 1977 investigations have added greatly to our knowledge of the morphology, constructional materials, and heat flow of the mounds. They have also made clear that the processes operating in the mounds area are more complicated than was originally envisaged. In the following pages we present a series of discussions, models, and hypotheses dealing with the principal questions that emerge from our data. Such questions concern the age and mode of origin of the mounds, the reasons for the morphological differences between various mounds, the fluid temperatures and heat and mass flow rates in the hydrothermal system, and the reason for the location of the mounds field as a whole. Unfortunately, the present data are still too sparse for definitive answers to any of these questions. We shall present several alternate models, each favored by different combinations of the authors, without being able to select, in most cases, a preferred model.

The Age and Mode of Origin of the Mounds

The lineated pattern of the mounds, with a spread of about 30° in azimuth, is parallel to the fault patterns originating in the rift valley and transferred down the flanks. Lonsdale (1977) further demonstrated that most mounds are located above small faults with throws of a few meters. These faults are likely candidates for the parts of the plumbing system that furnish access for the hydrothermal waters to the sediments, although the sediments themselves may not be competent enough to fracture and may provide transport channels mainly through the natural permeability, perhaps enlarged by overpressure. The only evidence we have for a possible role of a fissure is the vertical slab on the crest of one mound that forms the base of several pinnacles, as discussed earlier.

The only direct evidence for the time of initiation of mound building comes from DSDP sites 424 and 424A. At these sites, hydrothermal muds overlie about 10 to 15 meters of normal pelagic sediment. Assuming a long-term sedimentation rate in the area of 45 m/m.y. based on site 425 (Hekinian and Rosendahl, 1977), the first hydrothermal deposits formed about 210,000 to 330,000 years after the formation of the crust or about 300,000 to 420,000 years ago, an age that agrees with the biostratigraphy of the cores. At that time, the mounds area must have been located about 7 to 11 km south of the ridge axis, well beyond the zone where the structural development of the ridge crest is completed (Allmendinger and Riis, 1979). Since beyond that zone the only structural activity is a gradual reduction in relief by relaxation on normal faults (Klitgord and Mudie, 1974), there

apparently is more involved in the initiation of mound building, which we will discuss later.

Somewhat indirect evidence of the age of the mounds comes from the location of mounds today. The mound field begins just south of the major scarp 17 km south of the spreading axis. Besides some isolated small deposits very near the spreading center, no mounds or rows of mounds were observed on younger crust. The intermediate area was covered in both the DEEP-TOW Surveys, and a mound field containing large mounds (>4 m high) would not have been overlooked.

The area north of the 17-km scarp is one of low heat flow today, although it may not have been when the crust under the mounds was that distance from the spreading center. Also, the area north of the 17-km scarp is marked by numerous basement outcrops and exposed "scarps". South of the 17-km scarp the sediment has achieved sufficient thickness to bury most of these outcrops. The nearly complete sediment cover apparently is an important ingredient in the initiation of mound building, as we discuss in a later section. Therefore, we have some evidence that suggests that the mounds did not begin to develop until the underlying crust was 17 km from the ridge crest or correspondingly, was 500,000 years old. This implies that the mounds drilled at DSDP sites 424 and 424A are less than 130,000 years old.

The apparent discrepancy between the 330,000-year age derived from drilling data and the 130,000-year age, based on the youngest crust on which mounds are found today, could be due to the way in which mounds develop. If mound building can incorporate some of preexisting sediment material into the hydrothermal muds, then the unaltered

underlying sediment layer may have been thinned by the mound development. This would mean that the mounds are much younger than our interpretation of the DSDP drilling data indicates. We favor the younger age for the mounds.

The mounds themselves are the basic surface expression of hydrothermal flow through a sediment layer. This is clearly illustrated by the occurrence of very small, incipient mounds ("minimounds") which we observed in the inner rift valley. In general, the axial floor is free of sediment and the numerous springs issue directly from crevices in the basalt. Toward the edges of the active zone, however, patches of thicker (perhaps 1 m) and more continuous sediment cover occur. In such areas, on the outer limits of the active volcanic zone, small conical mounds several meters in diameter and 1-2m high have been observed on bottom photographs. They consist of reddish-brown sediment with a central zone of dark-brown to black crust and resemble the mounds we describe here--but on a very reduced scale.

The uniformly high heat flow of the entire mounds area suggests that it is underlain by thermal waters recharged from the low heat flow regions to the north and south (Williams et al., 1974 and Corliss et al., 1979). If this is the case, then the mounds mark the locations of breaks in an otherwise impermeable lid over the hydrothermal aquifer. This lid may consist solely of sediments, or it may include some basement rocks sealed by hydrothermal precipitates.

One of the most perplexing problems is that of the growth of the mounds. The DSDP data indicate that in at least two mounds about 15 m of hydrothermal sediment is deposited that is essentially free of

normal pelagic material. Laterally, this deposit corresponds to an only slightly smaller thickness of pelagic sediments free of hydrothermal material. What happens to the pelagic calcareous siliceous material that must, initially, be deposited on the surface of the mounds? The ALVIN observations indicate that, at most, several centimeters of pelagic sediment blanket the mound surface, overlying purely hydrothermal material. Is the pelagic rain, after incorporation in the surface layers of the mounds, dissolved under the prevailing physico-chemical conditions, the calcium being released to the outside and the silica perhaps incorporated in the nontronite? Or does the thin surface layer of pelagic sediment continuously slough off, under the influence of gravity, currents and bioturbation, to be redeposited at the foot of the mound? On the steep-sided older mounds this would seem to be a feasible solution, especially in view of the ubiquitous evidence for slumping and erosion and the likelihood of sediment instability due to excess pore water pressure. However, the smaller mounds do not generally have steep slopes, and show little evidence of erosion. Moreover, no evidence exists for significant accumulation of pelagic sediments around the mounds; on the contrary, many of the larger ones are surrounded by a moat 1-2 m deep and up to 20 m wide (Lonsdale, 1977).

As regards the accumulation of hydrothermal sediment within the mounds themselves, the available evidence strongly indicates that a large part of the deposition occurs internally. The dark surface crust of Mn-minerals forms in the zone of mixing between oxygenated bottom water and upward-migrating, low-oxygen hydrothermal fluids. As the

mound grows, this zone also migrates upward, resulting in the dissolution of the lower part of the Mn-crust and reprecipitation on top. Internally, under reducing conditions, nontronite is the main phase that is precipitated. Thus, in a sense, the mound swells like a balloon, continuously adding material internally and stretching and thickening its skin. A highly generalized diagram of the evolution of mounds is given in Figure II-18.

Morphogenesis of the Mounds

The mounds come in a considerable variety of sizes and shapes. In a general way, we can distinguish small mounds without knobs or blocky outcrops, large mounds with knobs and blocky outcrops, large mounds with only blocky outcrops, and small mounds with knobs. At least three variables may be important in determining the size of a mound: its age, the mineral content of the hydrothermal fluids, and the average rate of fluid supply. In addition, the shape of the mound might further be influenced by an internal structure surrounding a plumbing system (if water is not diffused evenly), and by erosion of the mound surface.

It seems probable that the hydrothermal fluids come from a large reservoir in the crust that underlies the entire area. Thus, the properties of the reservoir and of the fluids are not likely to vary over a short distance. Consequently, supply rates rather than the nature of the fluid must account for size differences between mounds in a given row or adjacent rows, if we assume that there are no significant age variations in a row of mounds associated with a single fault. On the other hand, a systematic size difference between rows

Figure 18. Simplified diagram of the possible development of a mound from the premound condition (Figure II-18-a) to the mature mound (Figure II-18-d).

(a)

(b)

(c)

(d)

may be the result of an age difference.

In the mounds area, the number of observed small mounds, presumably identical with what Lonsdale (1977) described as "stripes and patches of manganese crust," is greater than that of large mounds. The small mounds are elevated a mere few meters above the sea floor and have gentle slopes, marked only with the dark band arabesques. We regard these small mounds as either young stages in the mound-building process or the product of a lower fluid supply rate.

Two alternative hypotheses are possible to account for the knobs, pinnacles, and perhaps also for the blocky outcrops. In the first model, the knobs and pinnacles are regarded as the protruding ends of active vents, of the hydrothermal plumbing system, and the main channels of hydrothermal flow. These channels are filled, in this view, with a permeable mass of Fe-Mn-minerals, indicating that conditions in the channels are not highly reducing. Old mounds would have highly developed plumbing systems, frequently but not always extending to the surface or, probably as a result of limited erosion, well above the surface.

In the second model, the knobs and pinnacles are also regarded as infillings of a plumbing system, but not an active one. In this view, a mound contains an internal system of tubes and channels that is constantly being clogged by precipitates, leading to abandonment and the opening of new channels. From this internal plumbing system, in much the same manner as blood transport from main vessels to a pervasive system of small vessels under the skin, solutions diffuse through permeability pathways in the upper parts of the entire mound

and reach the surface ubiquitously. As the mound swells with age through internal deposition of hydrothermal materials, it oversteepens; slumping and erosion, aided by overpressured pore water, set in and expose the ends of the older, abandoned channels. In this view, the mounds with pinnacles and knobs are the oldest ones and the small mounds without any knobs the very young ones.

The occurrence of knobs may be a function of the rate of fluid flow rather than of age. Large mounds exist that do not have knobs and yet are warm inside; such mounds may be the result of old age and very slow fluid supply, in contrast to small mounds without knobs that are simply immature. From these observations one could conclude that a well-developed plumbing system resulting in knobs and pinnacles is evidence of a rapid rate of fluid supply, whether the knobs are active vents or erosional remnants of old ones.

The available data do not allow us to choose between these various models. Such a choice must rest on a more comprehensive understanding of the internal structure of the mounds--and that kind of information will probably be difficult to obtain. Geophysical methods have trouble penetrating the steep and hard surfaces of the mounds. The same is true of coring and drilling efforts. Almost any operation attempted from a surface vessel results in a large uncertainty in the exact location of the drill hole relative to a mound.

Mass and Flow Constraints

With the data we have it is possible to make a crude mass budget calculation. We can put bounds on the mass flux by applying the

results of Seyfried and Bischoff (1977). Upon reacting seawater with basaltic glass at 260°C and 500 bars, they obtained a solution containing 45 ppm Fe. The nontronite clay which forms the bulk of the mound deposits contains 25% Fe, according to Corliss et al. (1978). The mounds have a porosity of at least 50% and possibly more. The mass of the nontronite in a 20-m column of 1 cm^2 area will then be about 2000 grams of clay, 500 grams of which is Fe. If all of this Fe came from a hydrothermal solution of 45 ppm, then about 11×10^6 grams of fluid would be required. The ALVIN heat flow data show a range of temperature gradients consistent with flow velocities of from 10^{-4} to less than 10^{-5} cm/sec. If we use an average velocity of 10^{-5} cm/sec, then about 35,000 years would be required to build a mound 20 m high.

Despite the severe limitations imposed by the one-dimensional flow model, we feel that the data are consistent with maximum fluid supply temperatures of $15^{\circ}\text{--}25^{\circ}\text{C}$ and flow velocities of 10^{-4} to 10^{-6} cm/sec. The residence time of the fluid in the 20-m column would be 3.2 years. If, instead, flow in the mounds takes place in more discrete channels, so long as the channels are not more than a few meters apart, the porous flow calculations will still apply.

The Location of the Mounds Hydrothermal Field

Deposits similar to those of the mounds have been reported from various spreading centers. In transform fault A in the FAMOUS area on the Mid-Atlantic Ridge, two inactive vents are surrounded by black, yellow, and green hydrothermal deposits (ARCYANA, 1975). Farther south

on the same ridge, at 26° N, hydrothermal activity has produced manganese deposits in the rift valley (Rona et al., 1976), whereas Cann et al. (1977) obtained a dredge haul of manganese crust associated with green nontronite and amorphous iron oxide from the floor of the Gulf of Aden. Manganese crusts have also been found in the Galapagos Rift valley (Corliss et al., 1979). Mounds, however, have not been observed in any of these areas, nor on other parts of the Galapagos Rift than the area described here (Klitgord and Mudie, 1974; Anderson et al., 1976). Without near bottom observations they could have easily been overlooked. Therefore, although the evidence suggests they are uncommon features, we cannot say that they are unique to the Galapagos Rift at 86° W long. It seems more likely that they are simply peculiar to this specific type of geologic setting and since the geologic setting is rare so is the occurrence of mounds.

On the Galapagos Rift, hydrothermal vents have been identified in two areas, the axial part of the rift valley, and the mounds field. In the rift valley, the hydrothermal venting is concentrated in the innermost zone of active volcanism and is apparently quite short-lived; there is clear evidence that the discharge is greatly reduced by the time the newly formed crust becomes part of the marginal zones of the rift valley, where only inactive vents have been found. This limits the duration of the bulk of the venting to a few thousand years, in striking contrast with perhaps more than a hundred thousand years of venting in the mounds area. It is thus possible to attribute the axial hydrothermal activity to the cooling of the underlying magma body with

only superficial sea water convection.

Presumably, between the ridge crest vents and the mounds area vents there are two additional vent areas coincident with observed heat flow highs (Corliss et al., 1979; see Figure I-1). No serious effort was made to locate any of these vents. We neither attempted a survey of bottom water temperatures close to the seafloor (within a few meters) in these areas--nor did we dive on them with ALVIN. We do know that the DEEP TOW did not observe any mounds north of the mounds area--but it may not be able to detect small mounds (less than a few meters high).

These regularly spaced vent areas are apparently a result of an extensive regional convective system extending to appreciable depth in the lithosphere. The model given by Williams et al. (1974) proposes a semi-regular set of sea water convection cells beneath the ridge flanks that are somewhat influenced by the rather regular pattern of major fault systems. This model could yield the pattern of heat flow reported in Williams et al. (1974) and Corliss et al. (1979). The observed heat flow highs would mark the location of upwelling thermal fluids and the lows regions where recharge of cold sea water is occurring.

An alternative to this model is to have the entire convection system driven by recent intrusives. In this case the mounds would simply be above such intrusives. This alternative was discussed by Williams et al. (1974). Since there are numerous areas of high heat flow and many have large areal extent, substantial intrusive activity would be indicated. Given the apparent age of the mounds this activity

would have to be quite enduring. We favor the regional convective circulation system, but we cannot rule out the existence of a widespread and enduring area of off-axis intrusive activity.

If the high heat flow areas north of the mounds are vent areas but do not contain mounds, why are mounds found 18 to 25 km from the Galapagos ridge axis on 0.5- to 0.8-m.y.-old crust? We note that south of the 17-km scarp the sea floor is smooth and the sediment cover is generally ubiquitous. The fluid permeability of sediments is probably several orders of magnitude smaller than for much of the young ocean crust (Anderson et al., 1977). The mean conducted heat flow in the vicinity of the mounds is also higher than that a few kilometers to the south or north of the mounds region.

The heat flow is summarized in Williams et al. (1974) and Corliss et al. (1979). The conductive heat flux oscillates systematically with age of sea floor, presumably due to cellular patterns of hydrothermal circulation (Williams et al., 1974). Due to open exchange of sea water with the hydrothermal system, the measured conductive flux is generally less than the total geothermal flux from the spreading axis to the vicinity of the mounds. The effective hydraulic impedance increases only slowly with age due to increasing sediment cover, until in the vicinity of the mounds the complete blanketing of relatively permeable basement by sediment causes a dramatic increase in the hydraulic impedance. The physical situation is analogous to the large increase in electrical conductance in rock when a certain fraction and geometry of conducting interstitial fluid is attained (Shankland and Waff, 1974).

The location where sealing by sediments becomes complete is a function of spreading rate, sedimentation rate, and topographic roughness. The dramatic increase it causes in hydraulic impedance causes a similar reduction in the rate of exchange with sea water, which in turn leads to increases in fluid temperatures and changes in fluid chemistry. The fluids discharging from the ridge crest thermal springs are very low in iron, yet the mounds contain a substantial quantity of hydrothermal iron. Both of these fluid types have probably achieved temperature levels and chemical properties adequate to have permitted the dissolution of iron. Our explanation is that the ridge crest vents encounter oxygen-rich sea water as they ascend and the iron is precipitated beneath the seafloor. For the mound vents, the dissolved oxygen is not encountered until the fluids reach the base of the mound's manganese crust, because the sediments restrict the penetration of oxygen-rich seawater.

As normal pelagic sediment thickness continues to increase with time, the hydraulic impedance eventually becomes sufficient to effectively stop flow, and/or the total heat flux drops to a value insufficient to drive a significant amount of fluid through the low-permeability sediments (i.e., lower Rayleigh number), and the mounds become inactive and are buried. Circulation may still continue in the more permeable crust below, however.

If this explanation is correct, it may provide an approximate measure of the ocean crust permeability. The effect of a surface layer of sediment becomes important when its effective hydraulic impedance,

given approximately by h/k_s (k_s = permeability, h = layer thickness), is large compared to the impedance of the underlying crust (Skilbeck and Anderson, 1978). In the mounds area the sediment thickness is about 30 m, with a permeability estimated at about 10^{-12} cm^2 (Anderson et al., 1977). If the depth of effective fluid permeability in the crust is about 3 km (Williams et al., 1974), then the minimum effective crustal permeability, k_c , is given as $k_c = k_s (3000/30) 10^{-10} \text{ cm}^2$. This is approximately the same value obtained by Anderson et al. (1977) by another method, and is here estimated to be uncertain by ~ 1 order of magnitude. It is a minimum value since we do not know the lower limit on the thickness of continuous sediment cover that would effectively inhibit fluid flow.

It might be asked whether mounds comparable to those found in the Galapagos will be found near other spreading ridges, which question will only be answered by additional field investigations elsewhere. However, if our reasoning about the origin of mounds is correct, then they may not be ubiquitously associated with other spreading centers. The subdued topography and rapid sedimentation at the Galapagos spreading center may be particularly conducive for their formation. The conductive heat flow approaches the theoretical mean in 1 m.y. or less here, and for the Galapagos overall it appears to be several m.y. (Anderson and Hobart, 1976). For other ridges, the ages are much greater, perhaps several tens of m.y. for the Pacific and Indian Ocean ridges, and perhaps as long as 80 m.y. for the Atlantic (Anderson et al., 1977). By those times, the intensity of heat loss may be low

enough to inhibit formation of mounds.

CONCLUSION

As is so often the case, our detailed study of the mounds has raised as many questions as it has answered. The highlights of what we have learned are summarized in the abstract. Here we will summarize the important problems that presently lack adequate solutions.

The internal structure and plumbing of the mounds remain a mystery as does the composition and temperature of the hydrothermal fluids that enter and discharge from a mound. The age of individual mounds and the age of the mounds field can only be roughly estimated. The path taken by the fluids and the heat that are discharged from the mounds is only vaguely understood, other than to state that the water is predominantly sea water and the heat is from molten basalt from the mantle. Some of these questions can and presumably will be answered in additional diving and deep drilling that are scheduled for the mounds in 1979.

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Chapter III.
Numerical Modeling of Hydrothermal Systems
at the
Galapagos Spreading Center

ABSTRACT

Numerical modeling of two-dimensional convection within the oceanic crust is constrained by observations of the detailed heat flow field within 35 km of the Galapagos Spreading Center (on crust less than 1 Ma). By comparison of the observed heat and mass flow rates previously described in Chapters I and II with model flows, the maximum permeability-depth distribution is limited to no greater than .1 mD (10^{-12} cm²) at depths greater than 2 km. The bulk permeability of the upper 2 km appears constrained to greater than 1.0 mD (10^{-11} cm²); bulk permeabilities in excess of 5. mD are limited to the uppermost km. One major consequence of the convection is the reduction of the overall temperature of the upper crust relative to conductive models. We found that convection with variable fluid properties responds to thermal forcing in a predictable fashion based on the dimensionless Rayleigh number evaluated at elevated temperature. Correlation of model fluxes with Rayleigh number and comparison to results for slow spreading rate models, as well as comparison to the oceanic heat flow data, indicates a depth of fluid penetration shallower for the Galapagos than other ridges. The heat input at the ridge axis as a boundary for our models forces the wavelength of the resulting circulation to reflect the depth of strong circulation at the axis, and indicates 60 % or more of the "missing" heat flux near mid-ocean ridges is removed by hydrothermal upwelling at the axis of spreading.

INTRODUCTION

Recent direct observations with the Alvin submersible have provided spectacular evidence of hydrothermal activity at oceanic spreading centers (Corliss et al., 1979, Ballard et al., in prep.). The volcanic intrusive and extrusive activity localized at the spreading axis has long been assumed to drive convective systems within the oceanic crust comparable in size and effect on surface heat flux to subaerial geothermal systems (Bodvarsson and Lowell, 1972; Elder, 1965). Field studies near oceanic spreading centers generally show scattered data averaging well below the theoretical mean heat loss. The difference between observed and predicted heat flow versus age, based on observations of topographic subsidence and measured thermal properties of rocks thought to make up the oceanic plate, as well as plate thermal models, is usually attributed to heat loss by interstitial seawater circulation. Abnormally low conductive heat flow from oceanic plates apparently subsiding thermally provides evidence for the continuing removal of heat by water motion far beyond the direct influence of active spreading centers (Anderson et al., 1977).

Active venting has been discovered at two areas to date, the GSC (Corliss et al., 1979), and the East Pacific Rise (EPR) at 21°N (Ballard et al., in prep.). The nature of venting at these areas differs significantly in the total volume of fluid flow, exit temperatures of various vents, and spatial distribution of vent fields (Edmond, personal communication, Crane and Ballard, in press). Despite these differences, the inferred temperature of

chemical equilibration for most chemical species is found to be similar for all axial vents, in the range of 300-350°C (Edmond et al., 1979). While the physical reason for the uniformity of the reaction temperature is not understood, the likely candidates are reactions greatly reducing the permeability of the rock substrate to further water penetration, or some limitation inherent in the physical properties of the circulating fluids themselves.

THE GEOLOGICAL MODEL

We have a fair understanding of the materials that comprise the uppermost few hundred meters of the oceanic crust, based on extensive dredging and limited drilling results. The deeper regions of the crust have been investigated geophysically with marine seismics, gravity and magnetics, and petrologically with studies of ophiolites and dredge samples from fracture zones. The high porosity of portions of the oceanic crust drilled thus far implies a high permeability at least locally (Kirkpatrick, 1979). The seawater saturated crust can be treated as a uniform porous medium if there is sufficient porosity connected for fluid flow. The alternative first order model of the oceanic crust would be a fissure or vein flow model. The response of the crustal hydrothermal system modeled here assumes that discrete porous elements are closely spaced (100 m) so that conduction brings the impermeable rock into equilibrium with the surrounding fluid within a small fraction of the fluid flow time scale.

In this chapter, we investigate the consequences of various assumed permeabilities for fluid motion within the oceanic crust

under the influence of the thermal input due to the spreading process. We will adopt a model of fluid circulation based on averaged crustal properties resulting in an equivalent porous medium. The permeability of the oceanic crust is expected to decrease with depth due to a combination of changing rock type and increasing overburden effective stress. The functional form of this decrease is not known; in this study, for mathematical convenience an exponential decrease of permeability with depth was used, $k(z) = k_0 \exp(-z/L)$, with k_0 the surface permeability, L a length scale, and z depth in plate. This simple form of $k(z)$ yields a rapid decrease in permeability near the surface and a smooth transition to effectively impermeable crust at depth. Since we are dealing with crust from 0 to 1 my in age, no change in permeability with age was used. We specifically included heat input due to the spreading process in our calculations as one boundary condition. The spreading center is commonly referred to as the spreading axis, and we will refer to the first circulation cell, downwelling plus upwelling area, as the "axial cell." The intense circulation associated with the axial cell greatly influenced the spacing of fluid upwellings.

FIELD OBSERVATIONS SUMMARY

The most detailed survey of the thermal environment of young oceanic crust has been obtained on the Galapagos Spreading Center as reported in the previous chapters. We will use the observed heat flow data to constrain the choice of model parameters for the hydrothermal system. The average heat flow south of the Galapagos

Spreading Center is shown in Figure III- 1. The measurements south of the axis were averaged over 2 km ($\approx$ 57 kyr) intervals parallel to the ridge crest (i.e. along isochrons) and the mean $\pm$ variance is plotted on a log scale against distance.

Observations are limited to the region more than approximately 2 km away from the axis by the lack of sufficient sediment cover to obtain a gradient measurement. The increase in variance in each of the peaks in the heat flow distribution (somewhat masked by the log scale) is due to the averaging process smearing out individual very high measurements. For example, the variance of measurements within the hydrothermal mound deposits around 20 km south derives from a coherent, narrow ($\leq$ 50 m) high heat flow zone around each mound ridge, as described in Chapter II.

The characteristics of the heat flow field reported in Chapters I and II which are relevant to the numerical modeling are summarized here for reference:

1. The areal pattern revealed by 443 surface gradients is roughly two dimensional, with the heat flow near the axis depressed generally below 2 HFU along the entire 22 km width of the survey.
2. The peak values of heat flow around 20 km south are associated with fields of hydrothermally deposited "mounds" presently located in discontinuous regions from 18-32 km south of the ridge crest on crust from .5 to .9 Ma. old (Figure II-2).
3. Available detailed observations with the submersible ALVIN show extremely narrow high heat flow patterns around the mounds, with heat flow higher than the background level of 8 HFU only within 50 m of the crest of the mounds.
4. The mean heat flow through the 22 x 26 km survey area is 7.31 HFU as compared with the predicted heat loss from a simple plate cooling model of 19 HFU.

Figure III- 1. Heat flow distribution on the south flank of the Galapagos Spreading Center from Appendix I- 1. The mean $\pm$ one standard deviation for the measurements falling within non-overlapping 2 km strips parallel to the axis are shown by the stippled area. The dashed curve is the empirical $12(\text{age})^{-1/2}$ heat flow correlation from older crust (Lister, 1977). The heavy curve is the heat flux calculated numerically from a conductive model, M3c (3 km depth). The difference between the numerical model and empirical curves arises from several causes. First, the material parameters in the numerical model are appropriate for the upper crust, while the heat released in older crust arises from mantle material. Second, the heat flow from below the model is limited by the boundary condition to be asymptotic to $8(\text{age})^{-1/2}$. Note the logarithmic heat flow scale which masks the increased arithmetic variance in high heat flow areas.

Model M3c is shown in cross section with the conductive equilibrium temperature distribution in the lower part of the figure. The isotherms are labeled in hundreds of degrees Celsius. The geometry of all model cross sections will be similar in the following figures. The axis of spreading is on the left. Model depths vary from 2 to 5 kilometers. The width is always 23.2 km (0.4 km/horizontal grid). Vertical exaggeration is 2x. Spreading velocity is 35 km/myr, thus the models represent crust from 0 to .67 myr age.

5. The heat flow varies relatively smoothly over the area, with peaks at 6, 12, 20 and 28 km shown in Figure III- 1 yielding a wavelength of around 7 km normal to the spreading axis. The mean conductive heat flow in the region increased significantly from 2 to 10 km from the ridge crest and rose only slightly in going from west to east.

6. Locally elevated heat flow occurred consistently on the elevated side of scarps paralleling the axis of spreading, with depressed heat flux at the base of scarps, contrary to expected conductive cooling behavior (see discussion in Chapter I).

Observations further south show continuing heat flow variation with the mean rising up to the empirical $(\text{age})^{-1/2}$ model at about 3 Ma, or 100+ km distance (Figure III-2, Figure I-1, Von Herzen and Anderson 1972, Sclater and Klitgord 1973, Anderson and Hobart, 1976).

As emphasized in previous chapters, the major features of the heat flow field which can only be accounted for by convective heat transfer are the depressed heat flow near axis, the large magnitude of the short wavelength heat flow anomalies, and the elevated heat flow over topographic highs. This topographic anomaly is not modeled in these convection models. Topographic forcing of fluid flow within the crust will interact with the hydrothermal system established by the strong forcing from the volcanic activity at the axis of spreading. This interaction should be studied separately, once the basic response of the hydrothermal layer is known. We assume in this modeling that topographically forced motions are small in comparison to the overall circulation. The fact that the heat flow is only relatively correlated with topography suggests no dominance exists overall.

Figure III-2. Heat flow versus age for the Galapagos Spreading Center, averaged over 2.0 myr overlapping intervals, from Anderson and Hobart (1976). The mean $\pm$ standard deviation for the data are indicated by the solid lines. Similarly averaged data from the East Pacific Rise, with only the means given, is also plotted, with data east of the EPR given by the dashed line, west of the EPR by the dotted line. The data show a consistent nearly linear trend parallel and below the $12(\text{age})^{-1/2}$ curve. The mean lies just below the $6(\text{age})^{-1/2}$ curve until a sharp rise at about 3 myr on the GSC, and a somewhat less abrupt rise in the EPR data at 8-12 myr. The theoretical curves are explained in the discussion section.

THE NUMERICAL MODEL

We based our numerical model on the approximation of porous medium fluid flow within the oceanic crust limited by a variable depth distribution of isotropic permeability (Cathles, 1977, Fehn and Cathles, 1979). The equations of motion incorporate Darcy's Law to balance gravitational body force and pressure gradients (Bear, 1972; see Cheng, 1978, for a recent summary of the derivation):

$$(III-1) \quad (\nu/k) \underline{q} = - \underline{\nabla} p + \rho_f \underline{g}$$

Balances of mass and thermal energy are:

$$(III-2) \quad \underline{\nabla} \cdot \underline{q} = 0$$

$$(III-3) \quad \rho_m c_m T_t = - \underline{\nabla} \cdot (c_f T \underline{q}) + \underline{\nabla} \cdot (K_m \underline{\nabla} T) + \rho_m c_m \nu_s \underline{\nabla} T$$

The fluid equation of state is expressed as:

$$(III-4) \quad \rho_f = \rho_f(T, p)$$

In the above equations the temperature of both matrix and fluid is T ; fluid pore pressure p ; the fluid specific heat c_f , kinematic viscosity ν , and density ρ_f are tabulated as a function of temperature and pressure from the data of Keenen et al. (1969) for pure water, (see Tables III-1); fluid-saturated matrix specific

heat $c_m (= 0.2 \text{ cal/cm}^3/^{\circ}\text{C})$, thermal conductivity $K_m (= 0.006 \text{ cal/cm/sec/^{\circ}\text{C}})$, permeability $k(z)$, density $\rho_m (= 2.70 \text{ g/cm}^3)$ and spreading velocity, $v_s (= 35 \text{ km/Ma})$; g = gravitational acceleration $(= 980 \text{ cm/sec}^2)$ and fluid mass flux, q ($\text{g/cm}^2/\text{sec}$). Subscripts x, z , and t denote spatial and temporal derivatives.

Average pore fluid velocities can be derived from the fluid mass flux (also known as Darcy flux) by dividing by the porosity, ϵ , open to fluid flow. Typically this flow porosity is taken to be .001, considerably less than the intrinsic rock porosity (Kirkpatrick, 1979). The coordinate system is oriented with the x -axis normal to the spreading center and the z -axis vertically up from the top of the hydrothermal layer.

The use of a single temperature assumes local thermodynamic equilibrium of both fluid and rock matrix; this is consistent with the slow flows calculated under these assumptions. This system of equations assumes further that inertial forces in fluid motion can be neglected, which will be true for all flows calculated here. The matrix is isotropic, although not necessarily homogeneous.

We will assume two-dimensional fluid flow normal to the axis. The presence of the spreading axis as a natural center of symmetry is thought to force the geometry of at least the initial circulation of hydrothermal fluids. This assumption seems justified by the heat flow observations at the GSC which show a large area of depressed heat flow adjacent to the axis, with elevated measurements occurring at about the same distance from the ridge in each profile (Figure I-1.) In addition, theoretical and experimental investigations of the

Table III-1. Fluid property variation tables used in the modeling of hydrothermal circulation.

The specific volume in cc/g, enthalpy in joules/g, and viscosity in millistokes were taken from the tables of Keenen et al. (1969). The fluid specific heat is derived by dividing the enthalpy by the temperature. The sharp variation of parameters around the extension of the critical curve must be taken with a grain of salt, so to speak, since the true hydrothermal fluids will contain significant dissolved solids. This will extend the critical curve to higher temperatures and pressures than for the pure water phase, and modify the transport properties of the fluid (Sourirajan and Kennedy, 1962; Helgeson and Kirkham, 1974). We do not consider the effect of dissolved salts in the models reported here. The effect of dissolved solids on the minimum stability problem has been considered by Nield (1968). The possibility of double diffusive phenomena and chemical alteration of permeability patterns can only be studied with a more detailed numerical model. If the solute effect on the density of the fluid is less than the thermal forcing, these models should be adequate.

Table III-1a. Specific volume, cm³/gram fluid

<u>T°C</u>	Pressure (bars)								
	200.	250.	300.	400.	500.	700.	900.	1000.	1500.
0.	0.99	0.99	0.99	0.98	0.98	0.97	0.97	0.97	0.97
20.	0.99	0.99	0.98	0.98	0.98	0.97	0.97	0.96	0.96
50.	1.00	1.00	0.99	0.99	0.99	0.98	0.98	0.97	0.97
100.	1.03	1.03	1.02	1.02	1.02	1.01	1.00	1.00	1.00
150.	1.08	1.08	1.07	1.06	1.06	1.05	1.04	1.04	1.03
200.	1.13	1.13	1.13	1.12	1.11	1.10	1.09	1.08	1.08
250.	1.22	1.22	1.21	1.19	1.18	1.16	1.15	1.14	1.13
300.	1.35	1.34	1.33	1.30	1.28	1.25	1.22	1.21	1.20
350.	1.66	1.59	1.55	1.48	1.44	1.37	1.35	1.31	1.29
400.	6.0	6.0	2.79	1.90	1.73	1.56	1.48	1.45	1.41
450.	10.0	9.16	6.73	3.69	2.48	1.89	1.69	1.63	1.57
500.	12.0	11.12	8.67	5.62	3.89	2.46	2.02	1.89	1.80
550.	13.0	12.72	10.16	6.98	5.11	3.22	2.46	2.25	2.10
600.	15.0	14.13	11.44	8.09	6.11	3.97	2.96	2.67	2.45
650.	16.0	15.43	12.59	9.06	6.96	4.65	3.47	3.11	2.83
700.	17.0	16.64	13.66	9.94	7.72	5.25	3.96	3.54	3.22
800.	19.0	19.90	15.60	11.50	9.07	6.32	4.84	4.34	3.94
900.	21.0	21.04	17.45	12.96	10.28	7.25	5.64	5.09	4.62
1000.	23.0	23.10	19.19	14.32	11.41	8.11	6.25	5.68	5.17
1100.	25.0	25.12	20.90	15.64	12.49	8.93	6.75	6.13	5.60
1200.	27.0	27.11	22.59	16.94	13.56	9.72	7.15	6.50	5.95
1500.	36.0	29.10	24.26	18.23	14.62	10.51	7.45	6.80	6.25

Table III-1b. Enthalpy, joules/gram.

	Pressure (bars)									
$T^{\circ}\text{C}$	200.	250.	300.	400.	500.	700.	900.	1000.	1500.	
0.	20.01	24.95	29.82	39.51	49.03	67.71	70.	70.	70.	
20.	102.	107.	111.	120.	130.	147.	166.	174.	183.	
50.	226.	230.	235.	243.	252.	268.	286.	294.	302.	
100.	434.	437.	410.	449.	456.	472.	488.	496.	503.	
150.	644.	647.	651.	657.	664.	677.	691.	698.	705.	
200.	860.	862.	865.	870.	875.	886.	899.	904.	910.	
250.	1086.	1087.	1088.	1090.	1093.	1100.	1110.	1115.	1120.	
300.	1333.	1330.	1327.	1324.	1323.	1323.	1327.	1329.	1332.	
350.	1645.	1623.	1608.	1588.	1575.	1565.	1556.	1554.	1553.	
400.	2600.	2580.	2151.	1930.	1874.	1822.	1806.	1799.	1793.	
450.	3000.	2949.	2821.	2512.	2284.	2122.	2067.	2050.	2036.	
500.	3200.	3162.	3081.	2903.	2720.	2463.	2355.	2321.	2295.	
550.	3400.	3335.	3275.	3149.	3019.	2789.	2646.	2598.	2560.	
600.	3500.	3491.	3443.	3346.	3247.	3061.	2918.	2862.	2817.	
650.	3700.	3637.	3598.	3520.	3441.	3290.	3159.	3104.	3058.	
700.	3800.	3777.	3745.	3681.	3616.	3492.	3375.	3326.	3282.	
800.	4100.	4047.	4024.	3978.	3933.	3845.	3758.	3720.	3685.	
900.	4326.	4309.	4291.	4257.	4224.	4159.	4110.	4071.	4046.	
1000.	4582.	4568.	4554.	4527.	4501.	4449.	4410.	4379.	4361.	
1100.	4840.	4828.	4816.	4793.	4770.	4727.	4670.	4630.	4610.	
1200.	5101.	5090.	5079.	5057.	5037.	4998.	4865.	4830.	4815.	
1500.	5365.	5354.	5344.	5323.	5303.	5266.	5010.	4980.	4970.	

Table III-lc. Viscosity, millistokes

<u>T°C</u>	<u>Pressure (bars)</u>								
	200.	250.	300.	400.	500.	700.	900.	1000.	1500.
0.	17.20	17.20	17.10	17.00	16.80	16.50	16.50	16.50	16.50
20.	10.02	10.02	10.02	10.02	10.02	10.02	10.02	10.02	10.02
50.	5.48	5.47	5.47	5.45	5.44	5.44	5.44	5.44	5.44
100.	2.93	2.93	2.93	2.94	2.95	2.95	2.95	2.95	2.95
150.	2.00	2.01	2.01	2.03	2.04	2.04	2.04	2.04	2.04
200.	1.58	1.58	1.59	1.61	1.62	1.62	1.62	1.62	1.62
250.	1.36	1.37	1.37	1.39	1.40	1.40	1.40	1.40	1.40
300.	1.27	1.27	1.27	1.28	1.30	1.30	1.30	1.30	1.30
350.	1.22	1.22	1.22	1.22	1.23	1.23	1.23	1.23	1.23
400.	2.00	1.93	1.28	1.20	1.20	1.20	1.20	1.20	1.20
450.	2.84	2.84	2.23	1.52	1.30	1.30	1.30	1.30	1.30
500.	3.58	3.58	2.90	2.08	1.64	1.64	1.64	1.64	1.64
550.	4.28	4.28	3.52	2.58	2.05	2.05	2.05	2.05	2.05
600.	4.99	4.99	4.13	3.07	2.45	2.45	2.45	2.45	2.45
650.	5.73	5.73	4.75	3.55	2.86	2.86	2.86	2.86	2.86
700.	6.48	6.48	5.40	4.06	3.27	3.27	3.27	3.27	3.27
800.	7.78	7.78	6.70	5.08	4.09	4.09	4.09	4.09	4.09
900.	8.00	8.48	7.40	6.06	4.90	4.90	4.90	4.90	4.90
1000.	9.00	9.48	8.40	7.06	5.72	5.72	5.72	5.72	5.72
1100.	10.00	10.48	9.40	8.06	6.53	6.53	6.53	6.53	6.53
1200.	11.00	11.48	10.40	9.06	7.34	7.34	7.34	7.34	7.34
1500.	12.00	12.48	11.40	10.06	8.15	8.15	8.15	8.15	8.15

geometry of circulation indicate roll-type convection occurs for low to moderately unstable systems (Lapwood, 1948, Nield, 1968, Combarnous and Bories, 1975). The system of equations is solved by introducing the stream function, ψ , which is related to the mass flux by $q_z = \psi_x$, $q_x = -\psi_z$. After eliminating pressure from the primitive equations by taking the curl of equation (1) we have:

$$(III-5) \quad T_t = -((c_f T \psi_z)_x - (c_f T \psi_x)_z) + (K_m T_x)_x + (K_m T_z)_z + \rho_m c_m v_s T_x$$

$$(III-6) \quad 0 = ((\nu/k) \psi_z)_z + ((\nu/k) \psi_x)_x + g(\rho_f)_x$$

where subscripts x, z, and t denote partial derivatives with respect to space and time.

Since the momentum equation (III-5) is a steady state, instantaneous dynamic balance of fluid flow rate and horizontal density variations, it is solved iteratively, the time advance made via the energy equation (III-6) with advection of both fluid and rock matrix. These partial differential equations are cast as centered finite differences evaluated at grid points spaced at regular x and z spacings, as derived in standard numerical methods texts (Forsythe and Wasow, 1960, Carnahan et al., 1969, Roache, 1976). The horizontal grid spacing of all the models was 0.4 km. Fourteen vertical grid spacings gave adequate resolution of the temperature field. Further details of the modeling procedures and parameters are given in the Appendix III-1.

Stability and convergence were studied by reducing the grid spacing and time increments. We found the computed global quantities affecting heat and mass transfer were not significantly altered and feel most results are accurate to $\pm 15\%$. Most models exhibit unsteady axial temperature and velocity structures which may be a direct result of the interaction of the axial boundary condition and the varying fluid properties and thus total fluxes usually varied cyclically by more than 10% after "equilibrium" was reached. The vertical flux at the axis frequently was rapid enough to cause a small oscillation in the axial temperature and velocity. We adjusted the time steps from 0.1 kyr at the start to 1.0 kyr steps in the steady state.

In order to observe the interaction of the axial cell with convection further off-axis, we ran the models with a width of 23.2 km, resulting in aspect ratios from 4.6 to 11.6 for depths from 5 to 2 km. The cell geometry is not distorted by the far boundary, which was assumed insulating with no through flow of fluid, or this aspect ratio variation.

Rayleigh Number

Equations III-4, III-5, and III-6 allow an arbitrary variation of fluid transport properties to be used in the numerical model. For the purpose of comparing our results with previous studies and to further elucidate the basic physics of the convection process, we will temporarily simplify the fluid properties. The numerical model uses linear interpolation in both temperature and pressure of the data in

Table III-1 instead of constant property values. If we assume for the moment that density is expressed by $\rho_f = \rho_0(1 - \alpha T)$, then the parameter α (thermal expansion coefficient, $\alpha = (1/\rho) d\rho/dT$) joins the other fluid and medium properties in specifying the problem. Taking characteristic values of these parameters, we can nondimensionalize the system of partial differential equations to obtain :

$$(III-7a) \quad 0 = \psi_{zz} + A^2 \psi_{xx} + Ra T_x$$

$$(III-7b) \quad (\rho_m c_m / \rho_f c_f) T_t = (A v_s / \chi_m) T_x - (T \psi_z)_x + (T \psi_x)_z + T_{xx} + T_{zz}$$

where the dimensionless quantities

$$Ra = \text{Rayleigh number} = k \alpha g \Delta T H^3 / \chi_m = (H^2 / \chi_m) / (H / (k \alpha g \Delta T))$$

$$A = \text{aspect ratio} = H/W,$$

allow the solution to be studied and rescaled to any particular parameter range desired.

The Rayleigh number can be defined with a temperature difference, ΔT , based on either the base heating, Q_{base} , or the side heating, Q_{axis} , for fixed heat flux boundaries. We found the heat input at the axis was relatively constant within the range of permeabilities studied. We used the average heat flow into the base of the model to define the Ra.

The Rayleigh number arising in the nondimensionalized equations expresses the relative importance of convection and conduction. The

Ra can be seen to consist of the ratio of fluid transit time to thermal transit time, where the characteristic fluid velocity is given by

$$(III-8) \quad q_r = \epsilon k \alpha_f g \Delta T / \nu$$

This scale velocity is useful in comparing fluid flows between models and in predicting the magnitude and time scale of advection.

If the fluid parameters are considered independent of the temperature and pressure, Ra is the only parameter which determines the solution of the coupled partial differential equations III-5a, III-6a for a given set of boundary conditions. When the functions specifying fluid properties are variable, as is the case here, the relative importance of each term of the governing differential equations will vary in a nonlinear manner within the solution domain. This variation can be partially studied by examining the local variation of the layer and fluid properties in the Ra, namely $Ra' = \alpha_f \rho_f c g / \nu K$, as a function of temperature and pressure. These parameters are plotted separately and in combination in Figure III-3, evaluated from the tabulated data for pure water used in the numerical model (Tables III-1). The pressures of 300 and 600 bars correspond to the static reference fluid pressures at 3 and 6 km depth. It is apparent from the Table III-1 that the effect of pressure is most important at temperatures above that of the critical point (374°C, 218 bars). The common feature of a maximum in this instability parameter Ra' at around 400°C irrespective of the depth (pressure)

indicates a maximum effectiveness of convective transport. This arises primarily from the change in volume and enthalpy of the pure water substance at temperatures near the critical point. The tradeoff between rapid fluid transport of heat and viscous dissipation in the porous medium thus is optimized for fluids near 400°C . It is possible that this maximum in transport favors the development of relatively constant maximum fluid temperatures found in the field.

We anticipate that the models will retain some dependence of volume flux and steady state characteristics on a "mean" Ra of some type. From Figure III-3 it appears that the Ra computed at any temperature below about 400°C is relatively independent of pressure, while undergoing an increase of almost two orders of magnitude with temperature. The choice of a constant temperature for referencing the model Ra results in fixing the ordinal sequence of the models, while varying the reference temperature changes the cardinal values of the series. Plotting the results on a log scale will then reveal any systematic trends even if a constant multiplier must be applied to match other investigations. We choose to define the Ra with the (fixed) basal heat flux and fluid parameters evaluated at 175°C , a temperature near the mean of the bulk of the circulating fluids in most models. This parameter correlates with much of the variation in model results.

Figure III- 3. Variation of the fluid transport properties and their effect on the Rayleigh number, $Ra = Ra' k \Delta T H$, with $Ra' = \frac{\alpha \rho g \beta}{\eta}$. The model Ra_{base} used in the text is evaluated at 175°C . The curves for thermal expansion coefficient, heat capacity c_p , and viscosity, are evaluated from the Table III-1 at 300 bars (solid) and 600 bars (dashed). The maximum in thermal transport properties with the broad minimum in viscosity favors convective motions at temperatures near 400°C . The principal component influencing the Rayleigh number appears to be the thermal expansion coefficient, which has a sharp peak near the critical curve for water/steam equilibrium. Boiling is inhibited by pressures in excess of the critical point, however, a large volume expansion is still observed. Hydrothermal fluids derived from seawater will have a slightly higher critical point in temperature and pressure, but similar behavior to pure water is expected (Sourirajan and Kennedy, 1962).

BOUNDARY AND INITIAL CONDITIONS

We will summarize the boundary conditions used in the model and then expand the justification in the following section. The fluid flow was assumed to have free discharge at the upper boundary only, the other three sides being impermeable. This fixes the velocity at the two vertical boundaries to be purely vertical. The temperature of the upper boundary was assumed isothermal. The far boundary from the axis (to the right in the figures) was assumed insulating, and the spreading plane and bottom of the model had specified heat flux as derived in the following section.

Initial Conditions

Most models reported here were started from a conductive model near equilibrium, except in the case of very high surface permeability (50 mD) where the circulation was first allowed to proceed at 5 mD and then the permeability was increased once the heat content of the system had decreased. Tests were made with starting temperature distributions both cold ($T = 0$) and linear in depth ($T = a \cdot z$). While the peak mass flux and temperature history varied markedly, the initial conditions were not found to influence the final flow and temperature fields by more than the variance found within each "steady" solution. We are not interested in the initial flow of large magnitude resulting from the initially large heat content of the model, because this flow would only occur if circulation suddenly became possible by a general increase in permeability, or if a jumping spreading center were to force flow in older crust.

Conductive heat input

The heat input from below into the region is fixed by a simplified conductive model. The general conservation equation for heat energy III-5 is simplified by assuming no fluid flow beneath the hydrothermal layer and steady state temperature distribution with respect to the ridge crest frame of reference. The temperature as a function of distance, x , from the axis and depth, z , within the plate resulting from a balance of horizontal advection and vertical diffusion of heat to an isothermal upper surface at $T = 0$, is given by (Lister, 1972)

$$(III-9) \quad T(x,z) = T_M \operatorname{erf}\left(z \left(4x\alpha/v_s \right)^{-1/2} \right)$$

where T_M is the initial temperature and other terms previously defined. The derivative of this function evaluated at the base of the hydrothermal layer was used as the heat flux into the hydrothermal layer:

$$(III-10) \quad Q(x,H) = (KT_m/(\pi\alpha x/v_s)^{1/2}) \exp(-H^2/(4\alpha x/v_s))$$

This heat flow distribution is compared to that from the Davis and Lister (1974) model, which incorporated lateral conduction, in Figure III- 4. The model thermal parameters given earlier were used in evaluating both the lower boundary flux and the Davis and Lister series. This reduces the total heat flow to $8(\text{age})^{-1/2}$ relative to the empirical curve of $12(\text{age})^{-1/2}$ (Figure III-1) and enters as a constant factor of 1.50 in comparing the model results with global flux estimates (e.g. Edmond et al., 1980).

Figure III- 4. Solid curves: heat flow into the base of the models vs distance. Dashed lines: the Davis and Lister (1974) series solution heat flux, evaluated at the depth of the models using the same thermal parameters as in the numerical model ($K=.006$ TCU, $c_p = 0.2$, $T_{\text{intrusion}}=1200^{\circ}$). The vertical heat flux in the one dimensional approximation is somewhat higher than the Davis and Lister solution because of the neglect of horizontal conduction.

The heat flux into the base of the layer is zero at the origin because the injected material is initially isothermal. The vertical heat flow peaks very quickly with distance for the shallow 2 km depth, thereafter falling asymptotically as the square root of distance, while the basal heat flux for a 5 km deep model shows only a broad maximum. The average heat input at the base is tabulated in Table III-2, with the other model parameters. Shown in Figure III-5 is the temperature distribution calculated by the finite difference model together with the Davis and Lister model for the same plate thermal parameters and plate thickness of 80 km. The lower finite difference calculation includes a latent heat of 100 cal/gram which is neglected in the Davis and Lister solution. For conduction acting alone, temperatures within the upper two km exceeding 1000°C result from the plate velocity used here for comparison to the Galapagos spreading center (35 km/Ma). This has been inferred to allow the existence of shallow magma chambers at moderate spreading rates (Sleep, 1975).

Spreading axis

The spreading axis thermal boundary condition was investigated by Fehn and Cathles (1979) as part of their study of hydrothermal convection for slow spreading rates (10 km/myr). The heat input at the axis is governed by the spreading rate, latent heat of intruded material, temperature of intrusion, and heat of reaction of any metamorphic products produced during metasomatism. For slow spreading ridges, circulation driven by periodic intrusions of 200m width was compared to circulation driven by heat input with the

Figure III- 5. Equilibrium temperature distribution for conduction acting alone.

(a.) Davis and Lister (1974) series solution using an 80 km plate, with spreading velocity of 35 km/my, evaluated over the upper 3 km. The region at a temperature in excess of about 900°C would be partially molten.

(b.) Numerical model solution with no latent heat.

(c.) Numerical model solution with 100 cal/g latent heat advected in with the plate.

These cross sections illustrate that the boundary conditions and numerical approximations used in the model duplicate the analytic solution in the pure condition case.

TABLE III-2 Model Descriptions

- M2A Standard 2 km depth, 23.2 km width model with exponentially decreasing permeability from 5 to .05 mD (with an e-folding depth of 434 m). Spreading and lower boundary flux conditions as described in the text. Conductive equilibrium temperature distribution as shown in Figure III-5 (model M2c) was used for initial condition. Average basal heat flux is 11.42 HFU.
- M2B As for M2A except lower boundary required to be isothermal at 333°C, corresponding to a background heat flux of 10 HFU.
- M2C As for M2A except permeability increased everywhere by a factor of 10 so that the range is 50-.5 mD. Starting temperature distribution from M2A.
- M2D As for M2A except permeability constant at 5.0 mD.
- M2E As for M2D except permeability reduced to 0.5 mD.
- M2F Initial and boundary conditions as for M2A, but with fixed water thermal properties and linear expansion coefficient. Permeability decreasing exponentially with depth from 50-.5 mD.
- M3A As for M2A except depth changed to 3 km. The permeability decreases from 5.0 mD with the 434m length scale and thus reduces to .005 mD at 3 km depth. Conductive equilibrium temperature distribution used for initial conditions (model M3c). Average basal heat flux 9.19 HFU.
- M5A As for M3A except depth increased to 5 km. The flow in M3A indicated that reducing the permeability below .05 mD produced essentially conductive heat transfer, therefore the permeability range was kept at 5.0-.005 mD and the length scale was changed to 724 m. Conductive equilibrium initial temperature distribution (M5c). Average basal heat flux of 5.71 HFU.

continuous boundary condition suggested by Oldenburg (1975):

$$(III-11) \quad -K_m T_x = v_s (L + (T_m - T(t,z)) c_m)$$

where L = latent heat of fusion ($= 100 \text{ cal/cm}^3$) and T_f is the temperature at the end of the previous time step. Fehn and Cathles (1979) concluded from the similarity of the average and maximum temperatures and fluid flows that the continuous boundary condition was a good representation of the spreading process. Since the intrusion process is not directly modeled here but only its average effect, especially on the conditions off-axis, we used this continuous condition. At the medium spreading rate simulated here of 35 km/my, the more continuous volcanic activity inferred at the Galapagos Spreading Center (Van Andel and Ballard, 1979) may be approximated more closely by the above boundary condition than at the 10 km/my spreading rate studied by Fehn and Cathles (1979).

The axial temperature will in reality be highly transitory, since injection must take place at T_m while convection will quickly cool the intrusion, within a time of order kiloyears (Lister, 1974; Lowell, 1975; Cheng and Lau, 1977; Cheng and Minkowycz, 1977; Torrance and Sheu, 1978). We compute the mean wall temperature and net mass flux during an essentially continuous spreading process. We will examine models which would have essentially no flow without the axial forcing, as well as models unstable with the basal heating. Heat released within the hydrothermal layer at a depth z is subject to a pre-existing flow

caused by intrusions below z , plus the overall cellular overturn as a function of heating below the hydrothermal layer. The effect on the cooling history of pre-existing overturning (in a constant fluid properties porous layer) has been studied by Torrance and Sheu (1978), and found to significantly shorten the time response of the convection system. Since we are concerned with equilibrium conditions the time-average wall temperature will be lowered by a more rapid response.

Effect of Sediment on Upper Boundary Condition

Previous studies have emphasized that the isothermal boundary of the ocean/sediment-basalt interface will prevent hot fluids from exiting the crust at the low porous flow velocities computed for free convection (Ribando et al., 1976, Sleep and Wolery, 1978). Ribando et al. (1976), in comparing the Williams et al. (1974) GSC heat flow observations with the heat flow from a convecting fluid in a uniform porous layer heated from below, simplified the problem by neglecting two crucial geologic features: (1) the heat input associated with the ridge axis and (2) the venting of hot fluids from active upwelling areas. They approximated the heat loss at the surface with a conduction and Newtonian "radiation cooling" boundary condition,

$$(III-12) \quad T + h \, dT/dz = 0, \quad \text{at } z = 0.$$

where h is the thermal resistance of the surface, derived below.

This "radiation cooling" boundary condition arises from implicitly including the sediment cover in the following manner. By neglecting horizontal conduction and the heat capacity of the thin sediment layer (less than 50m of sediment, and less than 100m

vertical relief) and assuming that the fluid motion through the sediment is very slow (less than about 10^{-5} cm/sec (Lubimova et al., 1965)) and controlled by the underlying basalt permeability, the temperature profile in the sediments becomes

$$(III-13) \quad T(z) = (K_b/K_s) G z, \quad 0 \leq z \leq L,$$

where G is the (locally fixed) temperature gradient at the basalt sediment interface at depth L ; K_b and K_s are basalt and sediment thermal conductivities. The temperature at the sediment/basalt interface is $T(L) = (K_b/K_s)GL$; this temperature must be matched in the basalt. We can also neglect horizontal motion of fluid in the basalt just below the sediment due to the proximity of the free isobaric surface; then, since the convective heat transfer of any fluid motion across the interface will be the same in the basalt and sediment, the heat transferred by the gradient in temperature must be matched, just as the pure conduction case. Thus the condition that the temperature at the top of the sediment be fixed becomes a condition on both the temperature and the gradient at the sediment/basalt interface; since these are linearly dependent they can be combined to yield the homogeneous boundary condition:

$$(III-14) \quad T(z) - L(K_b/K_s) dT/dz = 0, \quad z = L.$$

If we nondimensionalize this equation with the scale of the hydrothermal layer, H , and the heat flow into the layer at depth H , Q_b we obtain

$$(III-15) \quad T'(z') - (L/H)(K_b/K_s) dT'/dz' = 0, \quad z' = L/H.$$

We thus find the heat transfer coefficient: $h = K_b L / K_s H$.

Ribando et al. (1976) found this boundary condition allowed the exiting of hot fluids only under very unrealistic assumptions about

the thermal resistance afforded by the sedimented interface. In the oceanic environment, sediment thickness rarely exceeds 25 m within the first half million years, and specifically for the GSC the sediment thickness increases at 50 m/my within our area (Klitgord and Mudie, 1974). Thus for the shallowest circulation depth considered here of 2 km, and with $K_b/K_s=3$, the heat transfer coefficient h is less than 0.04. The numerical model of Ribando et al. (1973) investigated the effects of varying the heat transfer coefficient on the upper boundary temperature. They found that a significant quantity of heat was carried out by convection of hot water across the upper boundary only for h greater than 0.3. Thus in our modeling we took the condition $T=0$. The effect of a finite value of h would be to lower the conductive heat flux and increase the convective flux at the level where these fluxes are evaluated in the models described here.

Model Calculations

The problem of circulation as posed above now depends on two further conditions: the depth of fluid circulation and the permeability distribution. We will first examine the results of models run at 2, 3 and 5 km with similar permeability variations, and then discuss a suite of models at a fixed 2 km depth.

The computed stream function and isotherms are shown in Figure III- 6 A, C, and D for three models with permeability decreasing exponentially with depth from 5.0 mD ($= 5 \times 10^{-11} \text{ cm}^2$) from the surface to the base of the domain. Figure III-6 C shows a model with permeability decreasing exponentially with depth from a surface permeability of 50. mD.

The models are shown in cross-section normal to the ridge with a vertical exaggeration of 2x. Injection of new material takes place at the ridge axis at the left of the figures. Dimensionless streamlines are shown as dashed lines over the isotherms. The stream function interval is variable in order to show the structure of the flow in regions of widely varying intensity. The intervals increase by factors of two from .125. The volume transported across the area between the streamlines (separation in the x-z plane per cm ridge crest) is equal to the difference in stream function values times the thermal diffusivity divided by the model depth. Thus the total mass flux across a surface between adjacent streamlines increases by a factor of two for each succeeding pair. The temperature isotherms are at 100°C intervals with the 500°C and 1000°C isotherms emphasized.

By comparison with the conduction temperature field shown in Figure III-5, we note that all the models are cooled to less than half the mean temperature appropriate to conduction acting alone. Steady state conditions, i.e. the balance between the heat inputs and output, were defined by observing the leveling off of the mean temperature of the domain as shown in Figure III- 7. For models M3A and M5A the excess heat content initially is much greater than for the 2 km models. In addition, the lower portions of these models (permeability below .10 mD) seem almost to respond conductively as seen by comparing the axial temperature distributions to the conductive cases (Figure III-8). This delays the approach to equilibrium of the deeper models.

Figure III- 6. Model streamfunction (dashed lines) and isotherm (solid lines) crossections just prior to formation of new upwellings near axis of spreading. Model permeabilities as marked. The models are shown in cross-section normal to the ridge. Injection of new material takes place at the ridge axis at the left of the figures. The dimensionless stream function interval is variable in order to show the structure of the flow in regions of widely varying intensity. The intervals increase by factors of two from .125. The volume transported across the area between the streamlines (separation in the x-z plane per cm ridge crest) is proportional to the difference in streamfunction values. Thus the total mass flux across a surface between adjacent streamlines increases by a factor of two for each succeeding pair. The local fluid velocity is inversely proportional to the spacing of the streamlines. The temperature isotherms are at 100°C intervals with the 500°C and 1000°C isotherms emphasized. The vertical exaggeration is 2 x.

Figure III- 7. Temperature averaged over the domain, T_{ave} , as a function of time, showing the approach of the models shown in Figure III- 6 to equilibrium convection. Temperatures for the conductive equilibrium case are given for each model depth.

The streamlines and isotherms in the upwelling region at the axis on the left of Figure III-6 are quasi-steady-state, with regular fluctuation of the maximum stream function in time. This fluctuation appears to arise from the numerical approximation of the fluid properties used (see Tables III-1), however the time step interval or an unstable thermal boundary layer may contribute to the variation. The strong upwelling at the axis produces a nearly exponential drop in the temperature as the upper isothermal boundary is approached. This thermal boundary layer is poorly resolved by our vertical grid spacing (Figure III-8). A relatively large fluctuation in the temperature of the first grid point at the axis results. This fluctuation is taking place at a mean temperature near the critical temperature for water where the properties of water are changing most rapidly. As a result, axial fluxes fluctuate regularly about the mean in most models. This fluctuation contributes to variance in maximum stream function ($\psi_{\max}$), mass flux and heat flux, listed in Table III-3.

Model M2F was run with a linear expansion coefficient and fixed fluid properties. The results show none of the fluctuations seen in the variable fluid properties cases. This suggests that the primary cause of the cyclic axial flux is the piecewise linear interpolation of fluid properties within the tables. The equivalence of upwelling and downwelling for this model with linear thermal properties resulted in equidimensional cellular convection.

Table III-3 Model Results

Model	$\psi_{\max} \pm \text{s.d.}$	$T_{\max} \pm \text{s.d.}$ ($^{\circ}\text{C}$)	$T_{\text{ave}} \pm \text{s.d.}$ ($^{\circ}\text{C}$)	Mass Flux $\pm \text{s.d.}$ 10^6 g/yr/cm	Heat Flux $\pm \text{s.d.}$ cal/sec/cm
M2A	10.09 ± 3.59	529.00 ± 23.3	171.50 ± 0.50	1.896 ± 0.501	38.08 ± 6.91
M2B	8.91 ± 3.40	458.80 ± 19.68	153.59 ± 0.25	1.398 ± 0.433	34.42 ± 6.71
M2C	23.13 ± 2.84	327.89 ± 1.36	91.92 ± 0.02	8.041 ± 0.362	39.32 ± 2.89
M2D	16.90 ± 0.19	303.57 ± 6.811	75.42 ± 2.75	6.261 ± 0.381	40.89 ± 1.83
M2E*	8.09 ± 0.36	487.56 ± 38.34	153.80 ± 0.16	1.922 ± 0.070	41.56 ± 1.29
M2F	14.50 ± 0.00	638.80 ± 0.00	132.36 ± 0.05	7.560 ± 0.052	40.25 ± 0.08
M3A	10.57 ± 2.91	(1200.)	256.20 ± 0.20	2.138 ± 0.564	44.54 ± 8.60
M5A	16.24 ± 2.66	(1200.)	262.06 ± 0.29	3.484 ± 0.482	51.09 ± 7.99

* Model M2E had only 11 grid points vertically, leading to largest variation in $T_{\max}$.

Figure III- 8. Vertical temperature profiles at the axis of spreading for the models in Figure III-6 at the time step shown in Figure III-6. The depth of the .1 mD permeability is indicated for models M3A and M5A. Regions of the model with such low permeabilities were dominated by thermal conduction. The vertical velocity at the axis in all of the models is high enough ($1-100 \times 10^{-6}$ cm/sec) to produce a sharp thermal boundary layer at the upper interface. This thermal boundary layer would be tens of meters thick and is not resolved by the vertical grid spacing. Experiments with a finer grid showed similar total volume flux and vertical average temperature.

Alternatively, it is possible that this fluctuation we observe is also the result of thermal lag in the boundary layer next to the axis of spreading. Horne and O'Sullivan (1974) observed varying intensity of circulation in a model with heating along one portion of the lower boundary with constant fluid properties. The instability of the boundary layer structure to variation in the supply of heat at the wall is found to depend on the spatial wavelength of the disturbance (Cheng, 1978). The free interface will affect the flow by providing a constant pressure surface hence a downflow at the edge of the vertical boundary layer. Temperature and velocity disturbances thus will be bounded in the vertical direction, and will produce pulsations in the axial flow, appearing as billowing isotherms. The temperature inversions found in some time steps appear to arise in this unsteady manner.

We found that the cells off-axis migrate with the plate, and thus represent only temporary locations of high fluid and heat flux relative to the ridge axis. The cell migration took two forms: in the first, with uniform permeability the downwelling adjacent to and feeding the upwelling at the axis broadened and eventually split to form a new cell. In cases with high surface permeability, 50. mD decreasing with depth, the first upwelling off-axis remains at a fixed distance from the spreading center. Figure III-9 shows the first type of migration of upwellings, in M2D (5.0 mD). The dashed line indicates the spreading plate motion. Each dot shows the location of the center of a zone of upwelling at a particular time. The axial upwelling is fixed at 0 km is not indicated. The cells

Figure III- 9. M2D migration of cells with time. The locations of centers of upwelling are shown through the birth of a new cell. The dashed line marks the plate advection due to the spreading process. The upwelling forced by the volcanic intrusion axis is fixed by the thermal input at the axis (0 distance) and is not shown. The off-axis upwellings track closely with the plate motion, maintaining the same relative separation. The open triangle at 120 kyr indicates the growth of a new upwelling. However, the leading upwelling stops migrating under the influence of the new upwelling, and they merge at 175 kyr. A new upwelling may form at 200 kyr, since the cell spacing is almost the same as at 120 kyr.

migrate initially with the plate velocity, although the insulating boundary at 23.2 km causes an approaching upwelling to jog slightly. The migration of the first upwelling off-axis produces a broadened zone of downwelling. Downwelling near the axis decreases in intensity as the first upwelling recedes, producing an increasingly broad stagnant region of fluid near the bottom of the model. This fluid is heated from below and from the axis; on a time scale dependent on the plate velocity, lower boundary heat input, and permeability, a recirculation eventually develops which rises and finally reaches the surface as a newly spawned upwelling. In Figure III-9, this is marked by the triangle at 120 kyr and 5 km distance. This upwelling is however too close to the leading upwelling for stability. The two cells merge at 145 kyr as the leading cell actually moves against the spreading motion. A new upwelling is about to form at 200 kyr elapsed time. As a result, a maximum distance off-axis exists at which the first upwelling is found (in this case about 10 km).

Figure III-10 shows the evolution of the 2 km, 50-.5 mD exponentially decreasing permeability model M2C in 20 kyr time frames during one complete cycle of cell birth. The high permeability at the shallow levels allowed this model to reach thermal equilibrium 20 kyr after the permeability was increased. Here the second upwelling as well as the first remains fixed with respect to the ridge axis.

As the downwelling between two merging cells decreases, conduction warms the region reducing downflow even more, so that a

Figure III-10. Time evolution of M2C, after "equilibrium"

circulation was obtained. The cycle of cell growth appears to take some 120 kyr. The new cell formed near to the old one, and their separation increases during migration. This is the only time observed in these experiments when adjacent cells moved apart; the cells are moving but not at the plate velocity immediately. Some settling time appears necessary for the velocities to adjust. The cell upwellings were almost fixed with respect to the surrounding rock mass at 23 km off-axis.

positive feedback system is set up. The rapid merger of cell upwellings produces a temporary imbalance in the conducted and convected heat flux at the edge of the upwelling. This heat flow asymmetry is a result of the time evolution of the fluid motion (hysteresis), and might be expected to arise in the field as well. Thus finding a region of high heat flow does not imply fluids are rising currently in the exact center. This adjustment of the streamlines to changing temperature distribution could be more pronounced when the permeability of the uppermost crust increases even more than in this simulation.

Conductive models of plate thermal structure such as McKenzie (1967) or Davis and Lister (1973) generally solve for a steady state temperature distribution in the frame of reference fixed to the ridge crest. The cell migration found here results in an unsteady temperature field near the axis, as a broadening downwelling eventually leads to sufficient stagnation to form a new upwelling cell. In the case of decreasing permeability with depth, with average permeability of 1 mD, the first cell geometry is found to be fixed, while the less vigorous second cell widened and then split (Figure III-10). The constant 5 mD model M2D did not stabilize the first upwelling, however, when the axial cell split, the new upwelling formed too close to the old cell, now the second off-axis, and they merged, as shown in Figure III-9.

The time required for the splitting of the cells is a function of the depth of circulation, the spreading rate, permeability, and the separation of the cells. As the model depth increases, the

distance of the first upwelling off-axis also increases. Minimum cell widths just as the cell splits, and maximum cell widths just before the cell splits are dependent on the depth of the permeable layer. We must caution that not all conceivable permeability distributions will have upwellings off-axis fixed with respect to the surrounding rock mass. As the permeability increases, especially at depth, the cells will most likely feel the stabilizing influence of the forced axial motion to increasing distances.

Model Heat Flow Results

The heat flow field from these models is not steady, as illustrated for model M2C in Figure III-11. For purposes of comparison, the models are shown approximately at the time a new upwelling would begin forming. The heat flow resulting from each of the models shown in Figure III-6 is shown in Figure III-12. Each of the convective models has greater heat flux from the axial upwelling region than from the conductive model, but immediately the heat flow, both total and conductive, falls below the conductive curve as downwelling occurs. As the depth of circulation increases, the width of this depressed heat flow zone broadens (cf. Figure III-12-A and D). The peak values of heat flow in the off-axis regions are observed to decrease with distance. The mean level of the off-axis heat flow depends critically on the permeability of the layer, and less so on the maximum depth of the model. The heat flux at the upper surface depends on the relative velocity of the cell with respect to the moving plate and the heat input from below.

Relative to the conductive equilibrium temperature structure, convective heat transfer lowers crustal temperatures everywhere

Figure III-11. Total (solid line) and conducted (dashed line) heat flow from model M2C at (A) 0 kyr and (B) 120 kyr time steps shown in Figure III-10. The heat flow is evaluated at the first grid point below the upper surface (141 m depth) in order to estimate the contribution of fluid flow to the total heat flux. The variation in elevated heat flow spacing exhibited in this simple cell migration model complicates the interpretation of high heat flow zone spacings as direct indications of depth of fluid circulation.

Figure III-12. Heat flow evaluated at the first grid point below the surface for each model shown in Figure III-6. Total heat flow (solid lines) consists of the conductive heat flow (dashed lines) plus the heat transported by the water motion. The conductive heat flow profile calculated in model M3c is shown for reference.

except in the upper part of the upwelling regions. This is also true at the axis, as seen in Figure III-8. Because of our boundary condition balancing the heat input and the temperature difference between the newly injected basalt and the in situ temperature, more heat is input to the model by the spreading process. In addition, the heat advected out of the model by plate motion decreases as the temperature of the advected crust is lowered. The total heat flow out the upper surface is elevated relative to the conductive case.

Figure III-13 is a comparison of the integrated heat flux (with respect to distance, not time) for the upper surface of the conductive and convective models, M2c and M2C. The dashed curve is the cumulative heat flux out the upper surface, that is, the integral of the heat flow shown in Figure III-11. The solid curve is the integral of the conductive model M2c surface heat flow. The dotted line is the integrated heat flux into the base of the model. As remarked earlier, the heat flow into the base is set to zero immediately at the spreading center. The slope of the integrated heat flux out of the upper surface is asymptotic to the slope of the heat into the bottom, while the mean separation of the two curves equals the total heat input at the axis. The mean separation for the first cell is larger than the separation at greater distances. The mean separation decreases because the mean temperature of the second cell increases over that of the first cell, absorbing some of the basal heat flux. The conductive flux at the spreading axis is only 170 HFU for M2c, which is not as high as the convective case, M2C. The displacement between the two curves is the extra heat flux

Figure III-13. Integrated heat flux in the lower boundary (Q_{base}) and out the upper boundary for models M2C (50 mD surface permeability) and M2c (conduction only). The integrated heat flux for the vigorously convecting M2C shows discontinuous increases as each upwelling is crossed. All three curves should become parallel at sufficiently great distances if the cells migrate completely fixed with respect to the plate.

removed by the convection process.

The cyclic nature of the heat flow field suggests the use of minimum stability theory to estimate the depth of the fluid circulation. The horizontal wavelength of elevated heat flow measurements of the GSC was equated by Williams et al. (1974) to the wavelength of porous media convection. This wavelength is a function of the boundary conditions and layer parameters. For constant fluid properties, between isothermal impermeable boundaries, this wavelength is found by minimum stability analysis and experiment to equal twice the layer depth (Lapwood, 1948; Katto and Matsuado, 1964). This wavelength was used by Williams et al. (1974) to estimate the depth of penetration of the hydrothermal system at 3.5 km. For other boundary conditions, the minimum stability eigenvalue problem was solved by Nield (1968) with the results given in Table III-3. The case analogous to the model boundaries of fixed heat flux, impermeable lower boundary and fixed temperature, permeable upper surface has a wavelength of 3.59 times the layer thickness. The input of heat from the side in our model imposes motion and removes the critical Ra from the system, as does the variable heating from below. However, the fundamental response of the system should favor modes which are composed of the natural system response.

Direct comparison of models M2A and M3A (Figure III-6) shows the effect of the addition of a nearly impermeable layer to the

Table III-4
 Values of the Critical Rayleigh Number, Ra_c ,
 and the corresponding wavenumber, k_c ,
 and wavelength, λ_c , for
 various boundary conditions (after Nield, 1968)

<u>Flow Boundaries</u>		<u>Thermal boundaries</u>		<u>Ra_c</u>	<u>k_c</u>	<u>λ_c</u>
<u>Bottom</u>	<u>Top</u>	<u>Bottom</u>	<u>Top</u>			
Rigid	Rigid	T_H	T_0	$4\pi^2$	π	2.00 H
Rigid	Rigid	T_H	Q_0	27.10	2.33	2.70 H
Rigid	Rigid	Q_0	Q_0	12.	0.00	∞
Rigid	Free	T_H	T_0	27.10	2.33	2.00 H
Rigid	Free	Q_0	T_0	17.65	1.75	3.59 H
Rigid	Free	T_H	Q_0	9.87	1.57	4.00 H
Rigid	Free	Q_0	Q_0	3.	0.00	∞
Free	Free	any		00.00	0.00	∞

Rigid = impermeable boundary
 Free = permeable boundary, in contact with free standing fluid
 T_H = Temperature fixed at heated bottom boundary
 T_0 = Temperature fixed at cooled top boundary
 Q_0 = Heat flux fixed at boundary, i.e. temperature gradient fixed at boundary.

bottom of the circulating system. Forced flow at the axis extends somewhat deeper, only increasing the magnitude of the overall fluid flux by about 10 %. This cools the axial cell relative to the 2 km model, lowering the isotherms throughout the model. Heat flux in the downwelling areas is decreased below .2 HFU even in the farthest downwelling. With a sharply defined lower limit to the circulation in model M2A, the cell spacing remains the same, and fluid flux remains nearly constant, but the effect on the mean heat flow is reduced. In both of these models, the circulation is dying out with distance too rapidly to match the observations. As the circulation decays, the surface heat flux decays to the conductive value, indicating as hydrothermal circulation slows the crust retains a linear temperature gradient appropriate to the basal heat flux.

The depth of the .25 streamline in models M2A, M3A, and M5A coincides roughly with a permeability of 0.10 mD. The axial temperature gradients shown in Figure III- 8 also show a similar gradient below the depth of the 0.10 mD in each model to the conductive equilibrium models. The gradients are somewhat less than in the conductive case and decrease upward due to the increasing heat transfer by the fluid motion.

Comparison of Models and Observed Heat Flow

Various permeability distributions were run for 2 km depth of circulation . Primary parameters of the results are listed in Tables III-2 and III-3. The basic model M2A produced convection only in the axial region where the heat input at the vertical boundary forced fluid motion. This forced convection died out within the 23 km width of the

model. Heat flow through the upper surface decayed about the mean of the conductive solution, M2c (Figure III-12 A). The conductive heat flux is reestablished more rapidly than in the conductive case discussed below.

The use of constant thermal properties for the circulating fluid in model M2F produces uniform areas of upwelling and downwelling shown in Figure III-14. The tabulated fluid properties used for all other models allowed the more effective heat transport at higher temperatures to produce narrowed zones of upwelling. Maximum temperatures in the upwelling fluid do not exceed 300°C except in the axial cell. These narrow upwellings arise without extra fissure permeability, and suggest that narrow high heat flow regions can arise from the effects of variations in water properties as well as spatial variations in crustal permeability.

Figure III-15 shows the observed heat flow distribution superposed on the predicted heat flow for the model M5A which brackets but does not match in detail the observed heat flow. The heat flow is depressed off-axis over too wide an area for 5 km circulation systems. In addition, the width of the high heat flow zones and the ratio of elevated heat flow area to depressed heat flow increases. Thus the observed high variance, six or seven kilometer wavelength heat flow field indicates a shallow depth of circulation. The time dependence of the models discussed here indicates that refinements of the model parameters in an attempt to duplicate the exact wavelength and amplitude of heat flow observations would be meaningless. The variation across our area from east to west which we initially

Figure III-14. Heat flow and model cross-section for M2F.

Constant thermal expansion coefficient of $10^{-3} (^{\circ}\text{C})^{-1}$, heat capacity $1.0 \text{ cal cm}^{-3} (^{\circ}\text{C})^{-1}$, and viscosity $10^{-3} \text{ cm}^2 \text{ sec}^{-1}$ were used to verify that the narrow upwellings observed in other models arose from the fluid properties and not from the numerical method. The constant properties make upwelling and downwelling closely symmetric.

Figure III-15. Observed variation of heat flow compared to the 5 km model M5A at three time steps. The maximum heat flux matches the range observed near the axis, but not the overall pattern. The observed heat flow highs near the axis suggest a slightly shallower depth of circulation than this model, but similar bulk permeability.

attributed to tectonic causes may actually arise from an unsteady hydrothermal system with many parameter variations.

Correlation of Results

The models illustrated have common features summarized as follows:

(1) axial upwelling forced by the addition of hot, new material is always the strongest motion of the model; this can stabilize the location of the first upwelling off-axis for some permeability distributions (e.g. M2C),

(2) upwellings off-axis migrate with the plate for most permeability distributions, so that at these spreading velocities the thermal history of a piece of crust should reflect initially deuteritic cooling to a state near 300°C followed by less rapid cooling, and then reheating to either a moderate or low temperature as the off-axis cells become established (Figure III-6),

(3) fluctuating heat and mass flux throughout the evolution of the model due to coupled heat input and temperature at the axis, and cell migration,

(4) upwelling spacing fixed roughly by depth of major circulation at the axis,

(5) narrower upwelling regions than downwelling due to the increased heat transport properties of the fluid at higher temperatures (Figure III-12).

It is apparent from the varying vigor of circulation that some influence of the basic Rayleigh number persists even in these models which have temperature and pressure dependent fluid properties.

Total mass flux is plotted versus Ra_{base} in Figure III-16 from the data in Table III-3, as well as the results of Fehn and Cathles (1979) and additional models from this study not described in detail here. The generally positive correlation of fluid flux with increasing Ra_{base} arises partly from the increased axial circulation, which constitutes up to 100% of the mass flow (Fehn and Cathles 10 km model). Also indicated in Figure III-16 is the inferred mass flux estimate of Sleep and Wolery (1978). They estimated the global heat flux anomaly associated with hydrothermal circulation and calculated the flow volume necessary at 140°C of $3.8 \text{ kg/km}^2/\text{sec}$ on crust less than 1 Ma, which converts to $2.78 \times 10^7 \text{ g/yr/cm}$ ridge over the first 23.2 km. We observe that our models can produce depressions in the heat flow adequate to explain the heat flow anomaly with an order of magnitude less mass flow. The difference may partially arise from the boundary conditions chosen here resulting in somewhat lower total heat flow; they may have also underestimated the effectiveness of fluid flow in removing the heat at the axis.

Discussion

The discovery of time-varying cellular convection in the range of parameters investigated allows further speculation about the significance of the birth of cells for the oceans. Experimentalists (e.g. Seyfried and Bischoff, 1975, and Bischoff and Dickson, 1975) have characterized the reaction of basalt and seawater as distinctly different for different water-rock ratios. Most of the circulation cells are entirely open to the ocean during their course. The axial

Figure III-16. Total mass flux plotted versus the $Ra_{\text{base}}(175^{\circ}\text{C})$, illustrating the close correspondence to a linear increase, despite variety of basal depths and heat fluxes. The 10 km models of Fehn and Cathles (1979) had essentially all of the flow taking place at the axis, and may not be comparable to the other models. For a given depth of fluid penetration, there is some curvature suggesting the mass flux is increasing less rapidly with increasing basal Ra .

cell is completely open, with mean fluid residence times ranging from less than 1 to over 10 kyr. However, the spawning of a new cell takes place at the bottom of the domain. Recirculation for several thousand years of seawater at elevated temperature will produce significantly enriched fluids relative to the open system at the axis. Volumetrically, this is not a particularly major component of the circulation, the maximum absolute value of the streamfunction in the recirculation being around 1.0, as compared to 16. in the axial cell. However, the altered rocks may appear quite different, and will occur whenever the new cells form, which is to say around 10 % of the basalt layer will be exposed to this "birth" process.

The previous investigation of Ribando et al., (1976) had concluded that the form of circulation in the exponentially varying case would be the same as the uniform case, except that the length scale of the circulation would be fixed by the e-folding depth of the permeability. We do not impose the wavelength of the flow. The narrow upwellings found here seem forced by the axial heat input over more than one cell, unlike the symmetric cells suggested by Ribando et al. (1976). They imposed that no heat be available to drive the circulation except that beneath the half-wavelength they modeled. Due to the unsteady cell configuration we find here, plate motion allows interchange of energy between the upwellings. The circulation cells also communicate with the axis by virtue of the forced down/up motion necessary to supply successive cells.

We have attempted to use the observable spatial distribution of maximum and minimum heat fluxes at the ocean floor in selecting the

parameters of the model hydrothermal system. The ratio of maximum to minimum surface heat flux has been used by Ribando, et al. (1976) to estimate the Rayleigh number (Ra) for a uniform property model of the oceanic crust. Their results show this ratio is strongly dependent on the Rayleigh number when the fluid properties are held fixed and the upper boundary is permeable, but somewhat less informative for impermeable upper boundary or decreasing mobility (k/η) with depth. In the case of decreasing permeability with depth, for fully varying fluid properties, this ratio is not uniquely defined. Under the assumptions of their model, while the ratio of $q_{\max}/q_{\min}$ was useful, the variation in area covered by heat flow above the mean was only from 50-40%. The area of elevated heat flux must be much less than 40% to match observations (Figure III- 1 and Figure I-5).

While Ribando et al. (1976) picked $q_{\max}/q_{\min} = 6$ as representative of the GSC on the basis of the Williams et al. (1974) survey, such a ratio test is notoriously unreliable for a poorly sampled population, especially for positive definite quantities with variance as high as oceanic heat flow. Thus, with our much more extensive sample, we would now pick limits, based on 1% levels of the population of interpolated heat flow data, of 22 HFU maximum and .5 HFU minimum, or a ratio of 40 ± 10 . Use of such a ratio is not valid for the apparently unsteady circulation and heat flow fields, and thus even with much more data, we cannot constrain the system on the basis of this ratio test.

Instead, the overall trend of rising mean heat flow normal to the ridge can be taken, with the increasing minimum heat flux, to indicate a general increase of the mean temperature of the crust. Such a general rise can be seen in several of the numerical models, and arises due to the over-efficient cooling forced at the axis by the heat flux from the side. The intense downwelling cools the crust below the level which free cellular convection would maintain thus resulting in a rising mean temperature away from the axis. Mean temperatures within the upwelling and downwelling regions of M2E and M2A are shown in Fig. III-17. For the higher permeability cases, the rising regions had temperatures higher than any downwelling regions; both up- and downwelling regions maintained nearly linear gradients because of the relatively slow flows involved (except at the axis). The variation in bottom boundary temperature was surprisingly small. Convection seems to maintain both axial and basal temperatures in the range 300 to 400°C, presumably due to the maximum in property variations around that temperature, i.e. the very efficient heat transfer at that temperature under high pressure conditions (Figure III-3).

The relatively uniform temperatures maintained at the base of the hydrothermal layer allow us to consider the conductively cooling underlying mantle in a new light. We previously have viewed the plate as conductively cooling from an injection temperature of around 1200°C through the isothermal (0°C) sea floor. We now have the suggestion that instead of cooling directly into the ocean heat sink, an intermediate layer has been introduced, which

Figure III-17. Average temperatures in the upwelling and downwelling portions of M2A and M2E. The cells are numbered sequentially from the axial upwelling, with the arrows indicating the direction of fluid flow. The small variation in these almost linear gradients is an indication of the fact the models are not at very high Ra , with the axial heating forcing the flows.

initially maintains the top of the mantle at a temperature of around 300-500°C, and which subsequently transfers approximately as much heat as a layer with thermal diffusivity about that of the mantle. Such a situation can be mathematically modeled again by balancing horizontal advection and vertical conduction as in traditional plate thermal models, but with an additional upper quenched layer of thickness L , similarly to the models of Sclater and Klitgord (1973) and Lubimova et al. (1975).

We can compare our results to a one-dimensional model quenched to depth Z . This comparison is valid since the axial cell removes most of the heat put in at the axis. The time behaviour of the heat flux in this case is shown in Figure III-18. The two layer problem was solved by the method of images as described in Carslaw and Jaeger (1959) yielding the solution as a sum of two error functions. The heat flow crossing the base of the quenched layer is infinite at the origin but has only half the value of the full plate case, $12/\sqrt{t}$ (my) since the temperature is free to rise at the lower interface. The heat flow at the upper surface is initially zero, but rapidly rises to $12/(e \cdot t)^{1/2}$, before following the $12/\sqrt{t}$ curve asymptotically. The flat heat flow at times near the peak, that is near H^2/α , suggests that the generally observed pattern of low heat flow coming slowly onto the $12/\sqrt{t}$ curve may not be a valid way to look at the time of cessation of the hydrothermal circulation. Circulation may continue only until the mean heat flow is $1/\sqrt{e}$ of the predicted heat flow. Then the heat flow slowly increases as the

Figure III-18. Heat flow versus time (distance/velocity) for quenched layer model. Surface heat flux recovers in time dependent on the thickness of the quenched layer, while heat flux from the underlying conducting halfspace initially has an asymptote of half the value appropriate to later times, once the surface $T=0$ is felt. The flat approach of the heat flux to the asymptote may be difficult to observe in oceanic data.

underlying plate rewarms the layer. Sleep and Wolery (1978) treated this point, but were not considering that the mean heat flow would reflect the depth of circulation and the maximum temperature of hydrothermal fluids at which it effectively caps the plate. This variation in the depth and temperature may account for some of the observed elevation differences between the ridge crests in the various world ocean ridges.

Summary

The basic porous medium model adopted here used lower boundary heating derived from the conduction solution, Parker and Oldenburg (1971) heat input at the ridge crest, isothermal sea floor, and an insulating, impermeable far boundary. Model depths of 2, 3, and 5 km with exponentially decreasing permeability were compared for effective steady wavelength of the convection cells for comparison with the heat flow data. While each model remains time-dependent, the distance off-axis of the first upwelling cell and the mean separation of the next upwellings increased with the model depth, as shown in Figure III- 12 . The length scale of the permeability distribution was found to influence only the initial cellular instability which advected out the initial (conductive equilibrium) temperature distribution. However, with continued growth of the cells and plate advection, the upwellings merged or died out, leaving fewer upwellings spaced out proportional to the model depth scale. For this reason, we feel the circulation responds strongly to the axial depth of fluid circulation, and therefore would limit most of the fluid flow at the GSC to the upper 2 km of oceanic crust.

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APPENDIX III: Numerical Methods

The basic idea of numerical modeling of complex physical systems is to recast the partial differential equations representing the physical interactions of the continuum in reality into the form of simple arithmetic differences of functions evaluated at discrete points. The assumptions of continuity of derivatives of all orders for each of temperature, pressure, and velocity functions allow the representation of the function derivatives by discrete differences of the function at neighboring grid points.

The partial differential equations describing our system have been given in Chapter III (III-5, III-6):

$$(AIII-1) \quad T_t = -((c_f T \psi_z)_x - (c_f T \psi_x)_z) + (K_m T_x)_x + (K_m T_z)_z + \rho_m c_m v_s T_x$$

$$(AIII-2) \quad 0 = ((\nu/k) \psi_z)_z + ((\nu/k) \psi_x)_x + g(\rho_f)_x$$

where subscripts x, z , and t denote differentiation with respect to x, z , and time respectively; T and ψ are temperature and stream function; ν , ρ_f , and c_f are fluid kinematic viscosity, density and specific heat; and K_m, c_m , and k are the porous matrix conductivity, specific heat and permeability.

At each grid point, we have material constants specified as initial conditions. While the coding allows a time or temperature dependence for each of these parameters, all of the models reported

maintained constant plate properties. The spatial variation of permeability was always decreasing or constant with depth. At each time step, fluid properties appropriate to the pressure and temperature at that time were derived from tables.

The solution sought is the temperature and flow field through time, $T(x,z;t)$ and $(x,z;t)$. The similarity of the above equations is increased if we reintroduce the time derivative of the stream function which was dropped as negligible in comparison to the other terms in the momentum equation:

$$(AIII-3) \quad \Psi_t = ((\nu/k)\Psi_z)_z + ((\nu/k)\Psi_x)_x + g\rho_x$$

Equations AIII-1 and AIII-3 are now of the form:

$$(AIII-4) \quad ET_t = (1/FG)((FT_x)_x + (FT_z)_z) - ((UCT)_x + (WCT)_z) + Q$$

A general solution subroutine for this equation was written by L. Cathles (1977) and used alternately for the solution of the stream function field and the temperature field. The solution algorithm begins with a centered difference approximation to the space derivatives and a forward time difference. Thus, spatial derivatives in the above equations are replaced as:

$$(AIII-4) \quad f_{x(z)} = (f((n+1)dx(z)) - f((n-1)dx(z))) / 2dx(z)$$

$$f_{xx(z)} = (f((n+1)dx(z)) - 2f((n)dx(z)) + f((n-1)dx(z))) / 2dx(z)$$

where the increments of x (or z) may be a function of z (or x).

Time derivatives are replaced with:

$$(AIII-5) \quad f_t = (f(n \cdot dt) - f((n-1) \cdot dt)) / dt$$

Conservative differencing requires that the material coefficients arising in the difference equations be evaluated at the same grid points as in the differencing equation. Conservative differencing is employed as discussed in Roache (1976, p.28 ff.) so that the variation in material properties across the domain is stably treated. In the above difference equations the time intervals are denoted by $n \cdot dt$ and the space intervals by $n \cdot dx(z)$.

Boundary conditions are given in Chapter III, equations III-10 and III-11:

$$(AIII-6) \quad A \cdot f + B \cdot f_{x(z)} = G$$

with $A=1$, $B=0$ for specified boundary value, $A=0$, $B=1$ for specified boundary gradient in either stream function or temperature with respect to x (or z). Substitution of the finite difference estimate of the gradient results in an equation for the boundary grid value which depends on the neighboring grid points and the input value at the boundary. Where symmetry is assumed, as at the axis, the gradient at the boundary must be zero.

The rectangular model domain results in a tridiagonal matrix of coefficients derived from the local properties arrays, whose inversion is achieved by the alternating direction implicit iteration scheme described in Carnahan, et al. (1969). This method has been shown to be convergent under arbitrary time steps for

linear coefficients, however in the current scheme a limitation on the time steps results from the nonlinear advection terms. As discussed previously, the very large vertical velocities at the axis resulting from the enforced vertical boundary heating causes the temperature at the axis to fluctuate between time steps. In advancing to the steady convective state, time steps were increased for each model from 0.5 kyr during the initial transient to 1 kyr during the late stages of the transient cooling. ^{Doubling} Halving the time steps during the later time period did not affect the heat or mass flux calculations. A simplified model flow chart is given in Figure A-III-1. Additional model parameters are given in Table A-III-1. The comparison of the current model results at higher spreading velocity with those of Fehn and Cathles (1979) shows that a regular increase of convection follows the increasing basal Rayleigh number, as discussed in Chapter III.

Figure A-1. Simplified Model Flowchart.

TABLE A-III-1. Comparison of Model Results

$k(z)^1$ (mD)	H_p (km)	Ra_b^2	Mass Flux Total ($\times 10^6$ g/yr)	Flux Axial (g/yr)	T_{ave}^3 °C	Q_{eff}	Q_{base}	Q_{axis}	Q_{top}^4	Nusselt Number ⁵	γ_{max}^6
						(HFU)					
.1	10.	80.	.70	.70	490.	5.88	6.0	0.	0.	1.02	0.0
.25	10.	200.	2.02	1.60	225.	2.70	6.0	0.	0.	2.22	0.0
.25	5.	50.	1.81	1.46	485.	5.82	6.0	0.	0.	1.02	0.0
.25	3.	18.	0.82	0.80	610.	7.32	6.0	0.	0.	0.82	0.0
.50	10.	400.	3.21	2.22	190.	2.28	6.0	0.	0.	2.63	0.0
.50	5.	100.	3.02	2.20	435.	5.22	6.0	0.0	0.0	1.15	0.0
.50	3.	36.	1.75	1.53	560.	6.72	6.0	0.	0.	0.89	0.0
1.e2	10.	174.	3.95	2.51	400.	4.80	6.0	0.	0.	1.25	0.0
.65	10.	113.	3.71	2.29	420.	5.04	6.0	0.	0.	1.19	0.0
1.0	5.	200.	3.90	2.55	410.	4.92	6.0	0.	0.	1.22	0.0
1.0	3.	72.	2.75	2.26	520.	6.24	6.0	0.0	0.	0.96	0.0
2.5e2	10.	434.	4.81	2.87	300.	3.60	6.0	0.	0.	1.67	0.0
2.5	5.	500.	4.72	3.01	390.	4.68	6.0	0.	0.	1.28	0.0
2.5	3.	180.	3.56	2.34	500.	6.00	6.0	0.	0.	1.00	0.0
50.e2	2.0	651.	8.18	3.28	91.1	5.47	11.42	84.5	17.2	2.08	-22.5
5.0	2.0	304.	6.12	2.90	75.3	4.51	11.42	85.2	17.3	2.53	-17.0
5.e2	2.0	65.1	2.41	1.96	170.4	10.22	11.42	74.8	18.7	1.12	-13.3
0.5	2.0	30.4	2.03	1.51	153.8	9.23	11.42	72.7	18.3	1.24	-8.5
5.e3	3.0	79.9	2.30	1.66	261.4	10.46	9.19	51.78	13.0	1.19	9.8
5.e3	5.0	137.2	3.53	0.0	264.5	6.34	5.71	52.30	10.4	1.11	0.0

1. $k(z)$ permeability distribution indicated by surface permeability in units of milliDarcy (1 mD = 10^{-11} cm²), followed by "e" if permeability decreases exponentially with depth and 2 or 3 for two or three orders of magnitude of permeability reduction across model depth.

2. Ra_b is the Rayleigh number based on the average heat input from below and the layer parameters evaluated at 175°C.

3. T_{ave} average temperature derived from summation over the solution domain.

4. $Q_{eff} = K T_{ave} / (H/2)$

5. Nusselt Number = Q_{base} / Q_{eff}

6. Fehn and Cathles (1979) data first 14 model results, remainder reported here. Q_{axis} , Q_{base} , and γ_{max} were not available from Fehn and Cathles.

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