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Determination of Terrestrial Heat Flow in Southeastern North Dakota

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DETERMINATION OF TERRESTRIAL HEAT FLOW

IN SOUTHEASTERN NORTH DAKOTA

by
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This Thesis submitted by Dean A. Zabel in partial fulfillment of the requirements for the Degree of Master of Science from the University of North Dakota is hereby approved by the Faculty Advisory Committee under whom the work has been done.

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This Thesis meets the standards for appearance and conforms to the style and format requirements of the Graduate School of the University of North Dakota, and is hereby approved.

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Signature Dean A. Zabel

Date July 26, 1979

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ABSTRACT

Terrestrial heat flow measurements have been carried out at three sites in Southeastern North Dakota. The heat flow values were calculated from temperature gradients measured in three wells and thermal conductivities measured in the lab using samples from these wells.

At two of the sites values were obtained in Precambrian layers. Near Lidgerwood, North Dakota measurements in a layer of weathered Precambrian yielded a value of 1.21 HFU. At a site near Blanchard, North Dakota measurements in a Precambrian greenstone yielded a value of 0.76 HFU.

At a third site near Wheatland, North Dakota, no Precambrian layer was accessible for temperature gradient measurement. Temperature gradients (42.10 and 31.56^{o}C/km) measured in two Cretaceous sedimentary layers at this site were found to be in the same range as the gradients (45.06, 49.97 and 23.51^{o}C/km) measured in three corresponding Cretaceous sedimentary layers at the Lidgerwood site. These contrast with the gradients (12.58, 14.22 and 13.93^{o}C/km) measured in Cretaceous and Ordovician sedimentary layers at the Blanchard site. This contrast in these temperature gradients is reflected in the calculated heat flows.

Differences in the radiogenic heat productions of the underlying Precambrian rocks is a likely explanation for the difference between the two heat flow values in the Precambrian materials at the Blanchard and Lidgerwood sites. The observed variations in the temperature gradients, and hence in the heat flows, in the sedimentary layers are probably a result of ground water movement in the different aquifers present at the three sites.

INTRODUCTION

Terrestrial heat flow is an artifact of the very structure of the earth. The heat flow measured in the upper portions of the crust has two sources, heat production from radioactive elements in the crust itself and heat conduction from the mantle. On continental masses at least 66% of the heat flow is felt to originate in the crust itself (Stacey 1977, p. 186), making the distribution of radioactive elements an extremely important factor in any land measurement (Birch, Roy and Decker 1968; Roy, Decker, Blackwell and Birch 1968). When heat flow and heat production data are present for the same location, it is possible to make estimates of the heat conduction from the mantle into the crust (Lachenbruch 1970). Since heat flow is dependent upon the earth's structure, the earth's structure can be probed with the use of heat flow data. This is especially true when the heat flow data can be combined with those from other geophysical methods such as gravity and magnetic anomaly studies (Simmons 1967).

The United States has been divided into heat flow provinces which are based on the regional variations of the reduced heat flow (often linked to mantle heat flow) observed across the country (Roy, Blackwell and Decker 1972). The eastern half of North Dakota is usually considered as having heat flow similar to that of the Eastern United States (Scattolini 1978). Reduced heat flow in the Eastern United States averages 0.8 HFU (Roy, Blackwell and Decker 1972), where one HFU (Heat Flow Unit) is equal to 10^{-6} cal/s cm^2.

In a layer of earth material free of perturbing influences, the steady state heat flow is given by

$$Q = K(\partial T/\partial Z) \qquad \text{Eq. 1}$$

where K is the thermal conductivity and $(\partial T/\partial Z)$ is the vertical temperature gradient. It is this steady state heat flow that is of interest. To achieve reliable terrestrial heat flow values it is therefore necessary to obtain reliable values of both the temperature gradient (here after understood to be the vertical component of the temperature gradient) and the thermal conductivity of a stratigraphic layer. On land this requires a borehole to give access to stratigraphic layers for temperature measurements to determine the temperature gradients in them and to obtain samples of the chosen layers for laboratory thermal conductivity measurements.

Ideally, any layer used should be deep enough to be free of surface effects such as water movement and annual temperature variations (Roy, Decker, Blackwell and Birch 1968; Sass, Munroe and Lachenbruch 1968). Any layer used should also be sufficiently thick and uniform to exhibit a linear temperature gradient and yield a number of samples for reproducible thermal conductivity measurements. It is further necessary to allow the hole to return to thermal equilibrium after it is disturbed by the drilling process (Lee 1965, pp. 17, 18, 44).

The most reliable method of thermal conductivity measurement is the divided bar method with hard rock core samples (Birch 1950; Roy and others 1968). Hard rock core samples are preferred as they sustain the least alteration of physical and thermal properties of any of the commonly used sample forms. It also is possible to make measurements on rock chips

(Sass, Lachenbruch and Munroe 1971) and on unconsolidated sediments (Von Hersen and Maxwell 1959). However, randomness of orientation and the need for in situ porosity information in the first case and thermal conductivity dependence on water content (Baver and others 1972, pp. 272-274) in the second case make these latter methods less accurate and hence less desirable.

Heat flow data exist for some portions of the Williston Basin of North Dakota (Scattolini 1978). However, on the eastern edge of the basin, near the North Dakota-Minnesota border, little data have been collected to this date. The heat flow values previously obtained on the eastern edge of the basin (Scattolini 1978) are not of high precision for several reasons. Only shallow, uncemented wells (allowing vertical water flow between aquifers) which terminate in aquifers were available. Also, poor thermal conductivity data were obtained due to the lack of good samples. It is the purpose of this thesis research to more reliably determine heat flow values for this area.

In 1977 a series of wells were drilled for stratigraphic studies by the Bendix Field Engineering Company under subcontract to the Energy Research and Development Administration. Permission was granted to use three of these wells for this study. The three wells used are identified as RRVD #2, RRVD #8A and RRVD #10. RRVD #2 is located in Richland County, North Dakota at T 130N, R 51W, Sec. 19. RRVD #8A is in Cass County, North Dakota at T 140N, R 53W, Sec. 33. RRVD #10 is in Traill County, North Dakota at T 145N, R 52W, Sec. 27. Figure 1 locates these wells on a map of North Dakota. After drilling and geophysical logging of each chosen well was completed for the original project, it was cased and

MAP OF WELL LOCATIONS

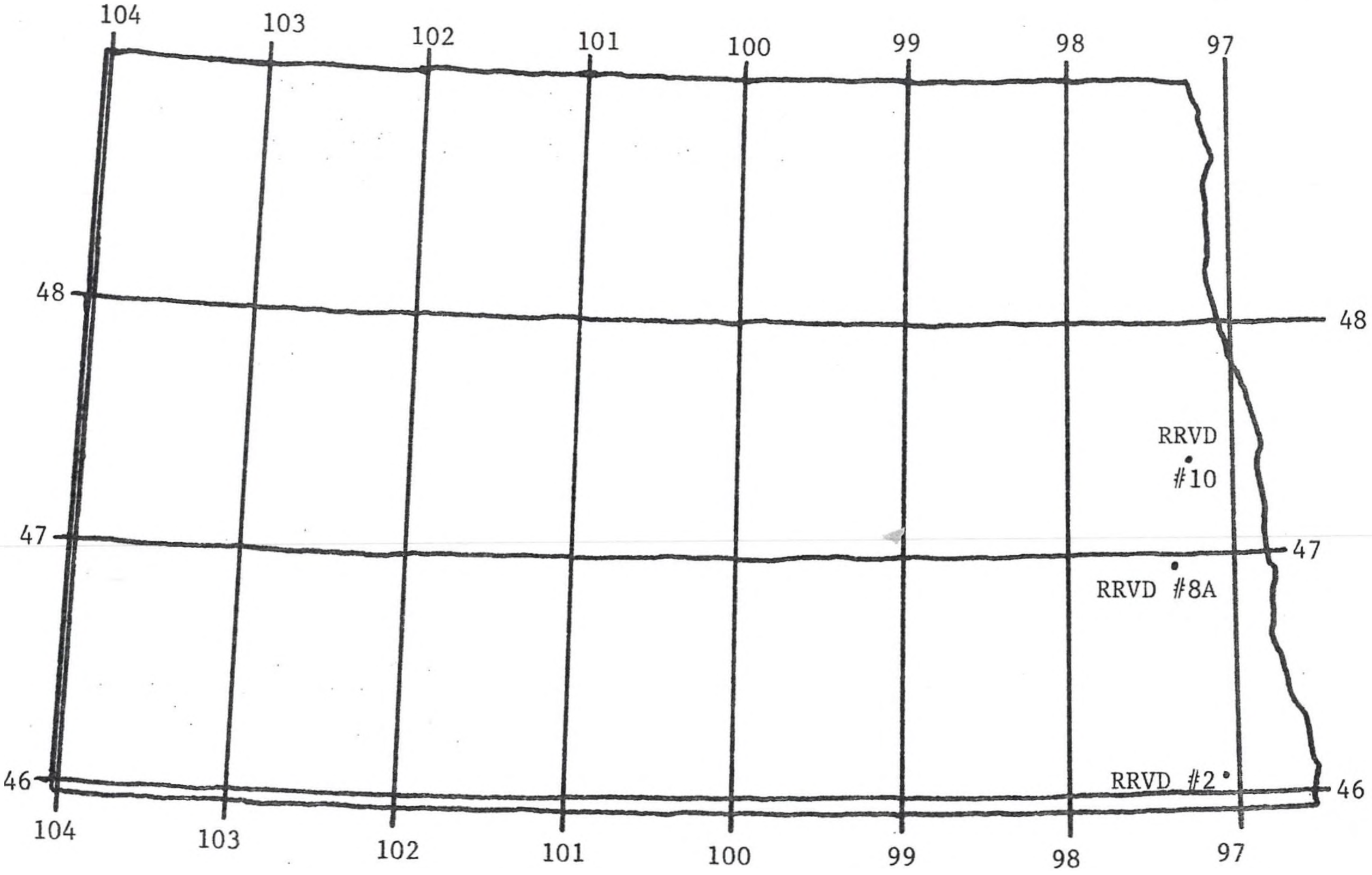


Figure 1

cemented to prevent ground water movement through the well and to provide good thermal contact with the neighboring earth material for this study.

Similar stratigraphies were found in the three wells. Two major stratigraphic differences were noted in the deeper layers. The Ordovician Winnipeg Group which is present in RRVD #10 and RRVD #8A was not found in RRVD #2. A member of this group, the Winnipeg Sand, is known to be an aquifer. Not present in RRVD #10 is a weathered Precambrian layer which is present in RRVD #8A and thickens as it extends south beyond RRVD #2. The stratigraphy used is given by Moore (1978).

Precambrian basement rock was penetrated in all three of the wells. Core samples of the Precambrian rock were recovered at each of the sites. The Precambrian rock was to be the lithologic unit of primary interest for heat flow determination. However, when the initial temperature measurements were made it was discovered that the cementing process left the Precambrian rock inaccessible in RRVD #2 and RRVD #8A. Therefore, only unconsolidated layers could be utilized in those wells. About three meters of Precambrian rock were accessible in RRVD #10. With the observed temperature gradient in the well, this length proved to be near the minimum for which an accurate measurement could be made.

EXPERIMENTAL

A. WELL SITE CONDITIONS

Upon completion of drilling and geophysical logging for the original project, each well was cased with 2 inch diameter black iron pipe cemented in place. The pipe was left water filled. Good thermal contact between the temperature sensing device and the surrounding earth material is provided in this way. The cement prevents water movement in the annulus between the pipe and the earth which could set up an artificial convective heat flow.

The wells were rotary-drilled, which disturbs the normal thermal equilibrium of the well site (Lee 1965, p. 17). It is estimated that reestablishment of the normal thermal equilibrium takes on the order of 20 times the amount of time required for drilling (Lee 1965, p.44). Drilling time for the deepest well was about 14 days so thermal equilibrium should have been reestablished in all of the wells within 300 days after drilling completion. No temperature data were collected in any of the wells prior to one year following completion of drilling. Temperature measurements were also made up to eight months after the initial measurements in order to check for temperature drift in wells RRVD #2 and RRVD #10. Remeasurement was impossible to accomplish for well RRVD #8A due to local conditions. Reasonable agreement between the repeated temperature gradients measured in RRVD #2 and RRVD #10 was observed (see the results section of this thesis).

B. WELL TEMPERATURE MEASUREMENTS

Temperature measurements in the wells were carried out using a Fenwall K212E thermistor connected to the surface with a four-lead cable. The thermistor resistance was measured with a Data Precision model 2540 A2 digital multimeter with four-lead connection compensation for the lead resistance.

This measurement system was calibrated in the laboratory against a Leeds and Northrup platinum resistance thermometer. The platinum resistance thermometer was last calibrated in 1975 by Leeds and Northrup against a standard traceable to the National Bureau of Standards. The thermistor calibration points were approximately 3°C apart and spanned the temperature range observed in the wells. Repeated measurements of the calibration points gave agreement to better than 0.02°C.

At the well sites the thermistor probe was lowered down the hole with a sinker bar to provide sufficient tension on the cable to accurately measure depth. Depth was measured by running the cable over a pulley of one foot circumference with a revolution counter attached on the pulley's axle. Measurement to the nearest 6 inches (0.15 meters) was possible in this way.

The sinker bar consisted of approximately 5.5 kilograms of lead in the shape of a slotted cylinder. It was clamped to the cable approximatedly 60 cm above the thermistor so as to minimize its affect as a heat sink on the water temperature at the thermistor's position. The probe was lowered slowly so as to induce as little turbulence as possible in the water which would cause a temperature mixing effect. Under these conditions the thermistor would come to equilibrium within 3 to 5 minutes.

C. THERMAL CONDUCTIVITY MEASUREMENTS

1. Divided Bar Technique

The thermal conductivity of the Precambrian rock core from RRVD #10 was measured employing a divided bar apparatus similar to that described by Birch (1950) and Roy and others (1968). A complete description of the apparatus used is given by Scattolini (1978) or Weispfenning (1977). The bottom of the stack was held at a constant temperature with circulating fluid from a constant temperature bath. The top of the stack was heated electrically to hold it at an elevated temperature. Fused quartz standards were placed in the stack above and below the sample to calibrate the system. Copper-constantan thermocouples were mounted in copper disks to measure the temperature on each side of the standards and the sample. The thermocouple potentials were measured with a Rubicon potentiometer with a precision of ±0.002 volts, which translates to a temperature precision of ±0.01°C.

After the sample was in place, the stack was allowed at least one hour before measurements were taken to establish equilibrium conditions. Measurements were then repeated over a period of at least 3 hours on each sample.

The thermal conductivity of the sample is calculated by comparison to the fused quartz standards according to the relation

$$K_r = K_q \left[\frac{(\partial T_u/\partial Z_u) + (\partial T_L/\partial Z_L)}{2(\partial T_r/\partial Z_r)} \right] \qquad \text{Eq.2}$$

where

K_r = the rock thermal conductivity

K_q = thermal conductivity of fused quartz (3.30 mcal/cm s °C)

$(\partial T_u/\partial Z_u)$ = temperature gradient in upper quartz standard

$(\partial T_L/\partial Z_L)$ = temperature gradient in lower quartz standard

$(\partial T_r/\partial Z_r)$ = temperature gradient in rock sample.

In effect this equates the heat flow through the sample to the average of the heat flows through the two fused quartz standards.

The samples were slices of core with thicknesses of from 1.26 to 1.30 cm. Both faces of each sample were lapped smooth. Thickness variations of each sample were no more than ±0.04 cm from the value used in the calculations. Immediately before being placed in the stack, each sample was placed in a chamber which could be evacuated and filled with distilled water. The chamber was evacuated for 45 minutes. Water was allowed to stand on the sample for 24 hours. In this fashion the moisture condition of the in situ rock was simulated as nearly as possible. When placed in the stack the faces of the sample were dried of water and a light coating of high thermal conductivity oil applied to insure good thermal contact between the sample and the thermocouple containing copper disks. An axial pressure was exerted on the stack by means of a compressed spring to further insure good thermal contact between elements of the stack.

2. Needle Probe Technique

A needle probe device of the type described by Von Herzen and Maxwell (1959) was used to measure the thermal conductivities of the unconsolidated samples. The probe used was constructed at the University of North Dakota Physics Department as part of the work in another thesis research (Weispfenning 1977). The probe was constructed of a 20 gauge hypodermic needle with nichrome wire as a heater and a thermistor as the temperature sensing device. The thermistor was calibrated against the same platinum resistance thermometer used to calibrate the well temperature probe. In this case the calibration points were approximately 5°C apart covering the needed temperature range. An absolute accuracy of ±0.02°C was observed from the calibration procedure. A relative accuracy of ±0.01°C was reasonably assumed.

To make a measurement, the probe was inserted into the center of a cylindrical sample, such as a core of unconsolidated sediments. At time t = 0, the heater was turned on by supplying a D. C. current to the nichrome wire. The thermistor resistance was recorded at 30 second intervals to provide probe temperature as a function of time. Each measurement required 10 minutes with a power input of just under 2 watts.

The temperature rise of the probe during heating becomes asymptotically linear with the logarithm of time as t approaches infinity. Weispfenning (1977) gives a full discussion of this problem as an appendix. The slope of this asymptote is related to the thermal conductivity of the sample by the relation

$$K = Q/(4\pi \text{ slope}) \qquad \text{Eq. 3}$$

where Q is the power input of the heater per unit length in mcal/cm s and the slope is in units of °C.

The samples used were of two types: (1) sedimentary core and (2) sedimentary drill cuttings. In these samples the thermal conductivity is dependent upon both the water content and the bulk density of the sample (Baver and others 1972). Both core and drill cutting samples were in a dehydrated state from being exposed to the atmosphere for approximately a year and a half before measurements were made on them. The density of the core samples is subject to decrease due to expansion of the core once the pressure from the overburden is removed. In the case of the drill cuttings, sample integrity does not exist.

The core samples were prepared by saturating them with distilled water, the assumption being that the layers represented are saturated in situ. Since the cores came from relatively shallow depths (all were less than 240 meters from the surface) it was felt that any density corrections would be relatively small and so were not attempted.

A somewhat similar procedure was chosen for use on the drill cuttings. First, samples of the cuttings were dry packed into a cylindrical mold having a diameter sufficient to meet the theoretical assumption that the sample boundary be at infinity compared to the probe radius. The samples were then saturated with distilled water and further packed until an appearance similar to that of the core samples of like material was obtained. The procedure is analogous to that used by Horai (1971) on powdered samples of pure minerals. From this point these samples were handled in the same way as the core samples, even though reestablishment of the in situ conditions was probably not attained.

RESULTS

Sample depth versus temperature plots for each well are shown in figures 2, 3 and 4. All of the temperature logs for the wells can be found in Appendix 1. The graphs of these data were used to identify layers exhibiting uniform temperature gradients. These layers, labeled on the graphs, were the ones chosen for heat flow measurements.

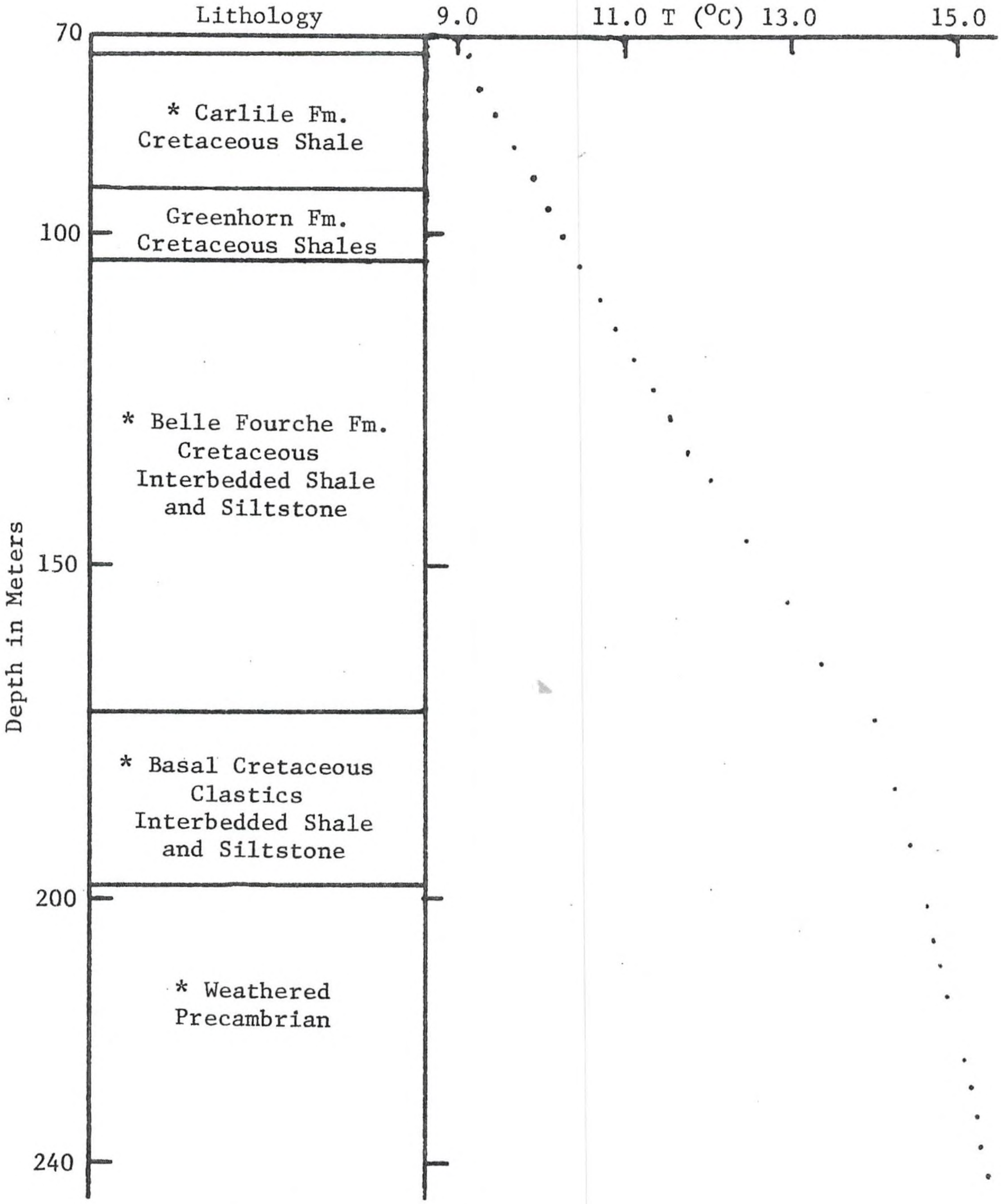
Once the layers were identified, the temperature gradients were calculated by least squares fitting. The temperature gradients listed in tables 1, 2 and 3 are representative of each layer as a whole and are averages from the multiple loggings. As an estimate of the uncertainty in these values the statistical standard deviation, σ_g, is reported as a percentage.

In the reporting of the average temperature gradients there were two individual values that were not used. In the first logging of well RRVD #10, temperature instabilities were noticed in the bottom 3 meters. This made accurate temperature determinations impossible for the Precambrian greenstone layer. A similar occurrence was observed in the fourth logging of RRVD #10 on the 173.13 meter reading, which justified the exclusion of this data point from the gradient calculation. In the second logging of well RRVD #2 the probe was lowered directly to the 73.15 meter level without pause. In retrospect it was felt that insufficient time was allowed for the probe and sinker bar to come to equilibrium with the well water before data collection was begun.

TEMPERATURE LOG

RRVD #2

17 May, 1979



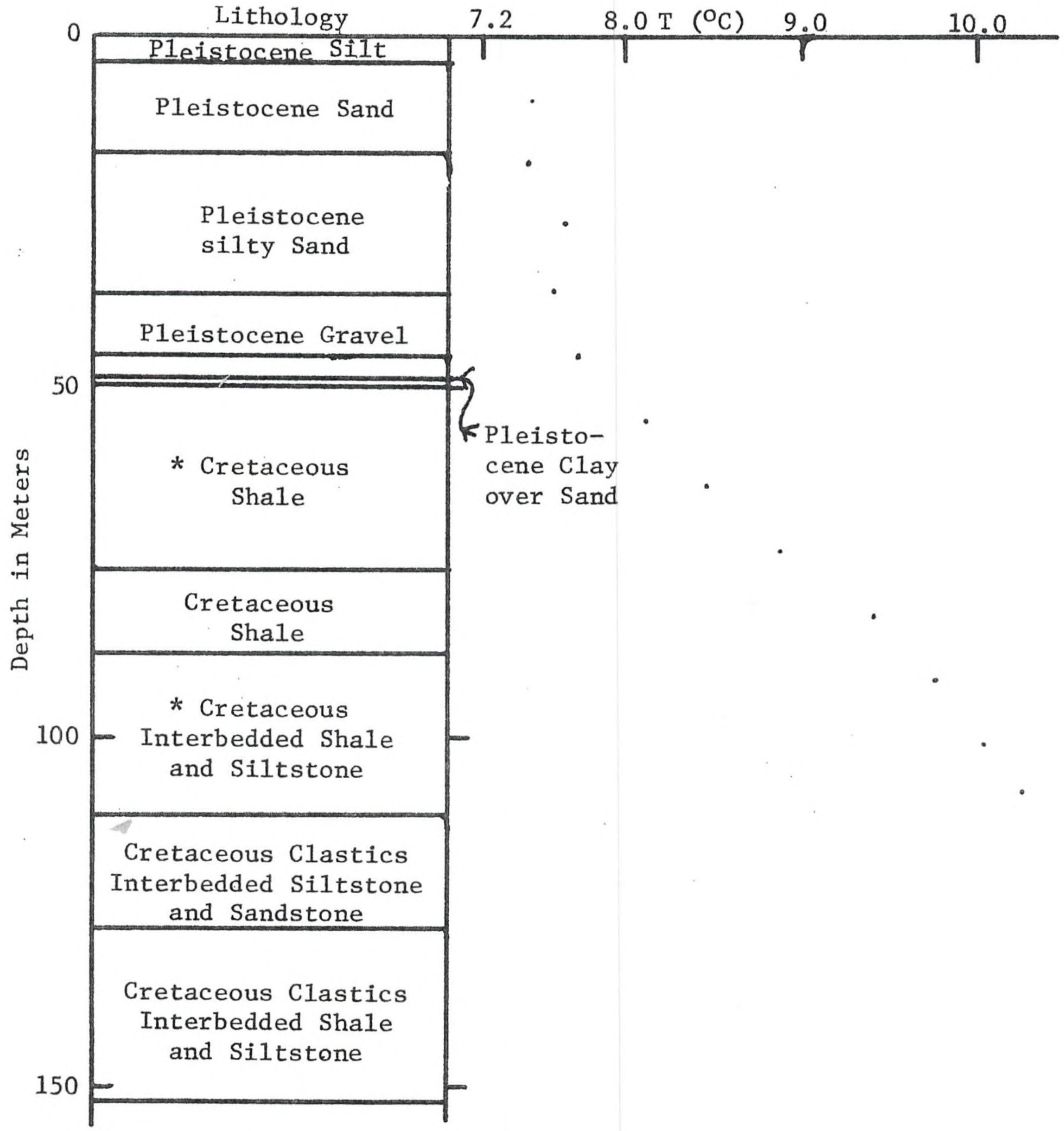
* Layers used for heat flow measurement

Figure 2

TEMPERATURE LOG

RRVD #8A

3 Aug. 1978



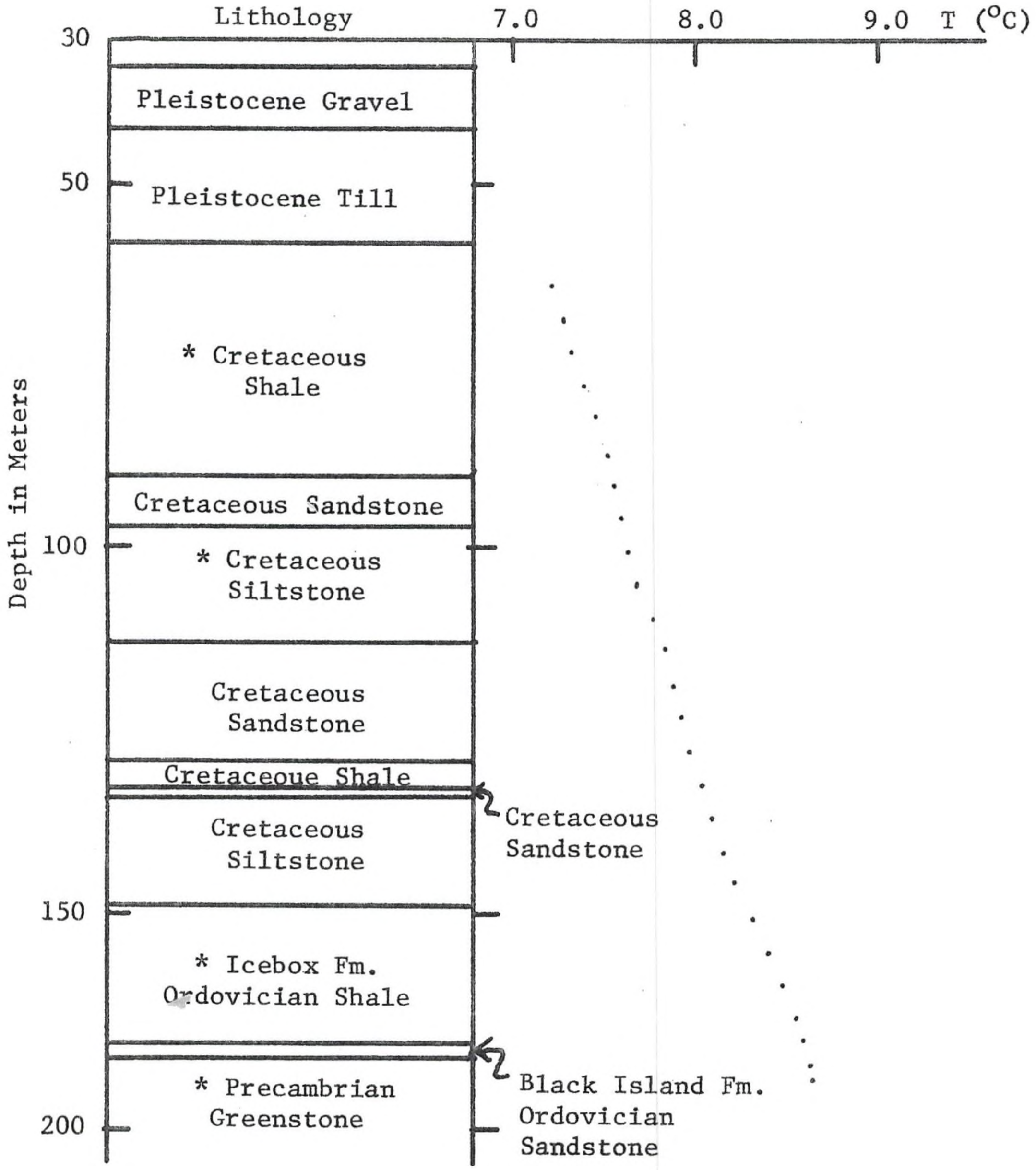
* Layers used for heat flow measurement

Figure 3

TEMPERATURE LOG

RRVD #10

15 May, 1979



* Layers used for heat flow measurement

Figure 4

Greater caution was used to prevent the reoccurrence of this problem in all subsequent logs.

Thermal conductivity values for the Precambrian greenstone from RRVD #10 found in table 3 are averages of six measurements made on the divided bar apparatus. All other thermal conductivities listed are averages of four measurements made with the needle probe. The statistical standard deviation, σ_K, is given as a percentage for each sample.

Figure 5 is a sample plot of temperature versus logarithm of time for the needle probe. The value of the slope used in equation 3 was in each case calculated by least squares fitting the data points beginning with the one at 90 seconds. Even though Eq. 3 is sufficiently accurate after 30 seconds (Weispfenning 1977), the 30 and 60 second data points were not used because of the difficulty of accurately reading the rapidly changing thermistor resistance in that time interval.

The heat flow for each layer listed in tables 1, 2 and 3 is calculated using the average thermal conductivity, $\bar{K}$, of the samples from that layer. A statistical standard deviation, $\sigma_{\bar{K}}$, for this average is given as a percentage of the uncertainty in the heat flow values. The value of σ_H is calculated with

$$\sigma_H = \sqrt{\sigma_K^{\,2} + \sigma_g^{\,2}} \qquad \text{Eq. 4}$$

(Sass, Munroe and Lachenbruch 1968). In this case σ_K is taken to be the largest value of $\sigma_{\bar{K}}$ or σ_K for that layer.

The final heat flow results with the thermal conductivities, temperature gradients and lithologies are also represented graphically in figures 6, 7 and 8.

NEEDLE PROBE PLOT

RRVD #10

Sample Depth 73.2 - 77.7 Meters

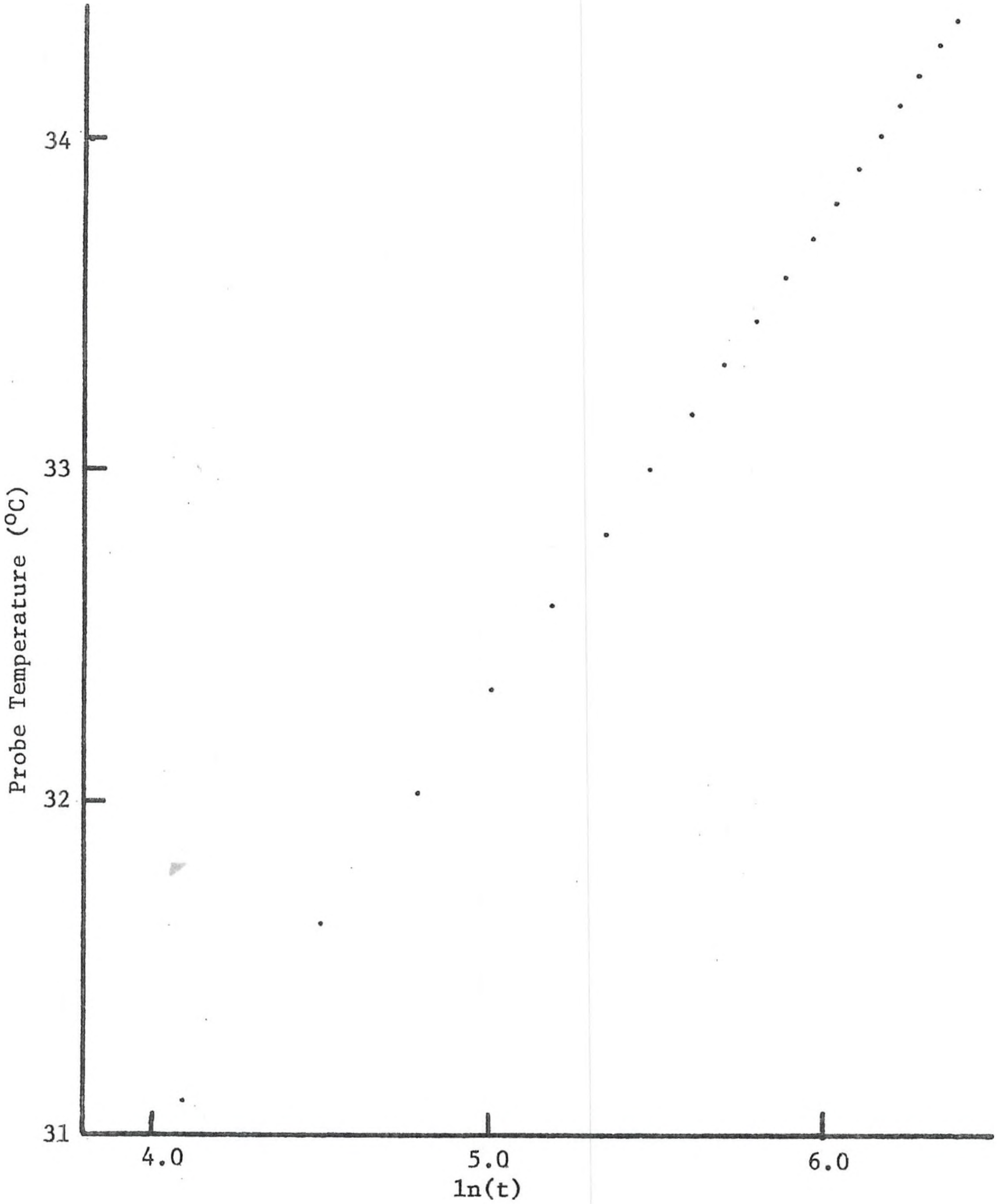


Figure 5

TABLE 1

RESULTS OF WELL RRVD #2

Sample Depth (meters)	K (mcal/cm s ^{o}C)	σ_K (%)	$\overline{K}$ (mcal/cm s ^{o}C)	$\sigma_{\overline{K}}$ (%)	$\partial T/\partial Z$ (^{o}C/km)	σ_g (%)	Heat Flow (HFU)	σ_H (%)
77.1- 85.3	3.29	5.5						
85.3- 89.9	3.34	1.5	3.35	1.8	45.06	2.7	1.51	6.1
89.9- 93.0	3.41	2.3						
106.7-111.3	3.43	1.2						
111.3-118.9	3.60	1.4	3.65	6.8	49.97	3.0	1.82	7.4
118.9-128.0	3.92	2.3						
176.8-182.9	4.16	2.4						
192.0-198.1	4.26	1.4	4.21	1.7	23.51	8.0	0.99	8.4
*201.8	7.02	3.8						
*202.7	5.93	9.4	6.22	11.3	19.49	9.4	1.21	13.3
*204.5	5.70	3.5						

* Weathered Precambrian core samples

TABLE 2

RESULTS OF WELL RRVD #8A

Sample Depth (meters)	K (mcal/cm s ^{o}C)	σ_K (%)	$\bar{K}$ (mcal/cm s ^{o}C)	$\sigma_{\bar{K}}$ (%)	$\partial T/\partial Z$ (^{o}C/km)	σ_g (%)	Heat Flow (HFU)	σ_H (%)
53.3- 65.5	4.48	10.5						
65.5- 67.1	4.73	3.8	4.41	8.3	42.10	3*	1.86	10.9
68.6- 74.7	4.01	3.7						
88.4- 94.5	4.51	6.9						
94.5-100.6	3.87	2.8	4.07	9.3	31.56	3*	1.28	9.8
100.6-106.7	3.84	2.1						

* Based on temperature and depth precision, see p. 25.

TABLE 3

RESULTS OF WELL RRVD #10

Sample Depth (meters)	K (mcal/cm s °C)	σ_K (%)	$\bar{K}$ (mcal/cm s °C)	$\sigma_{\bar{K}}$ (%)	$\partial T/\partial Z$ (°C/km)	σ_g (%)	Heat Flow (HFU)	σ_H (%)
57.9- 67.1	4.12	5.1						
67.1- 73.2	3.88	2.1	3.91	4.9	12.58	8.9	0.49	10.3
73.2- 77.7	3.74	0.8						
* 98.1	5.84	2.4						
* 98.8	6.42	3.3	6.45	9.8	14.22	14.0	0.92	17.1
*100.9	7.10	5.2						
150.9-153.9	5.58	3.0						
153.9-158.5	5.55	4.3	5.33	7.6	13.93	12.7	0.74	14.8
158.5-164.6	4.86	5.1						
+171.3	9.86	5.8						
+172.2	9.17	2.6	9.31	6.4	8.20	0.0	0.76	6.4
+173.1	9.68	4.1						
+174.0	8.53	0.9						

*Sedimentary core samples.
+Greenstone core samples, divided bar apparatus used.

HEAT FLOW RESULTS

RRVD #2

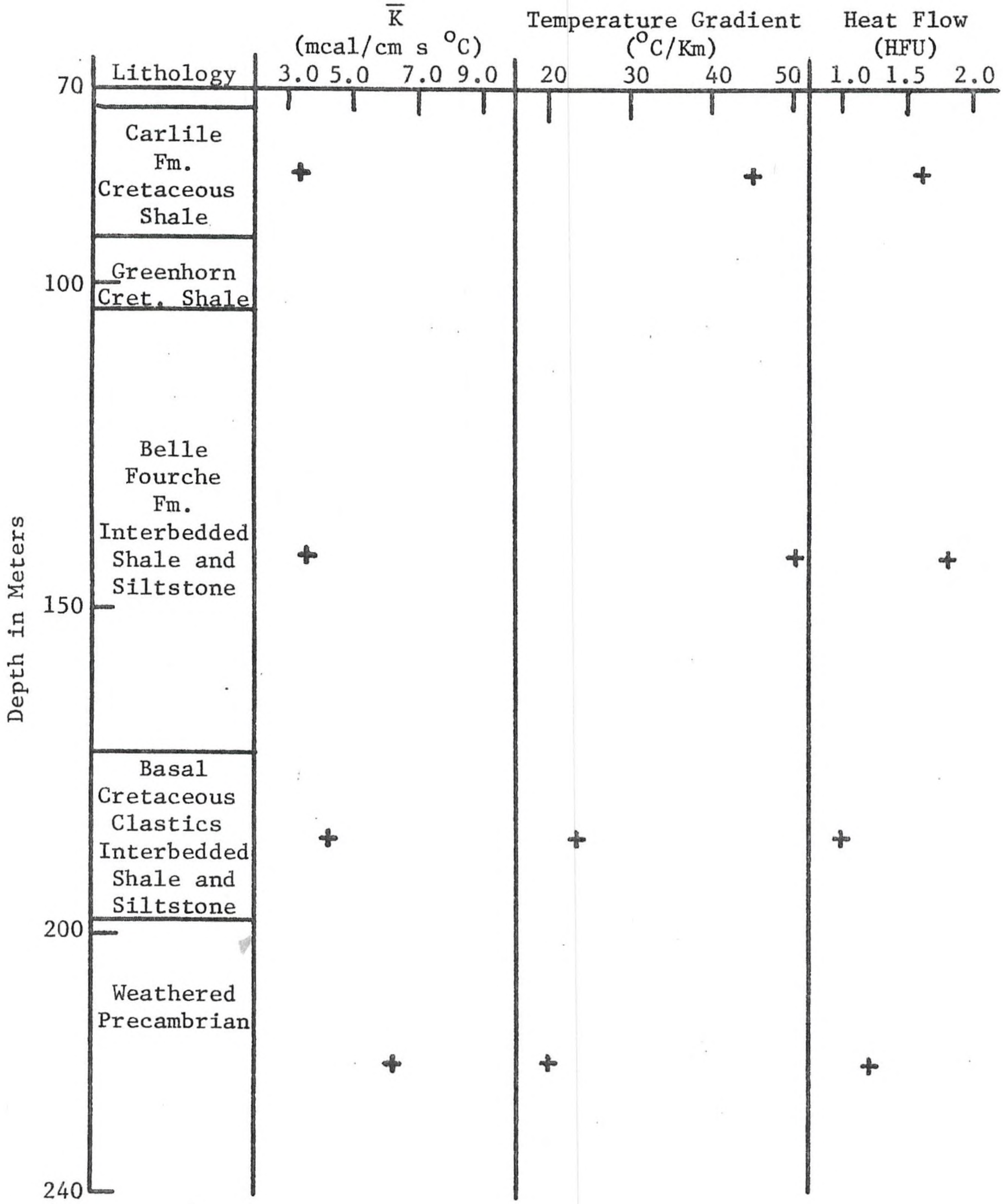
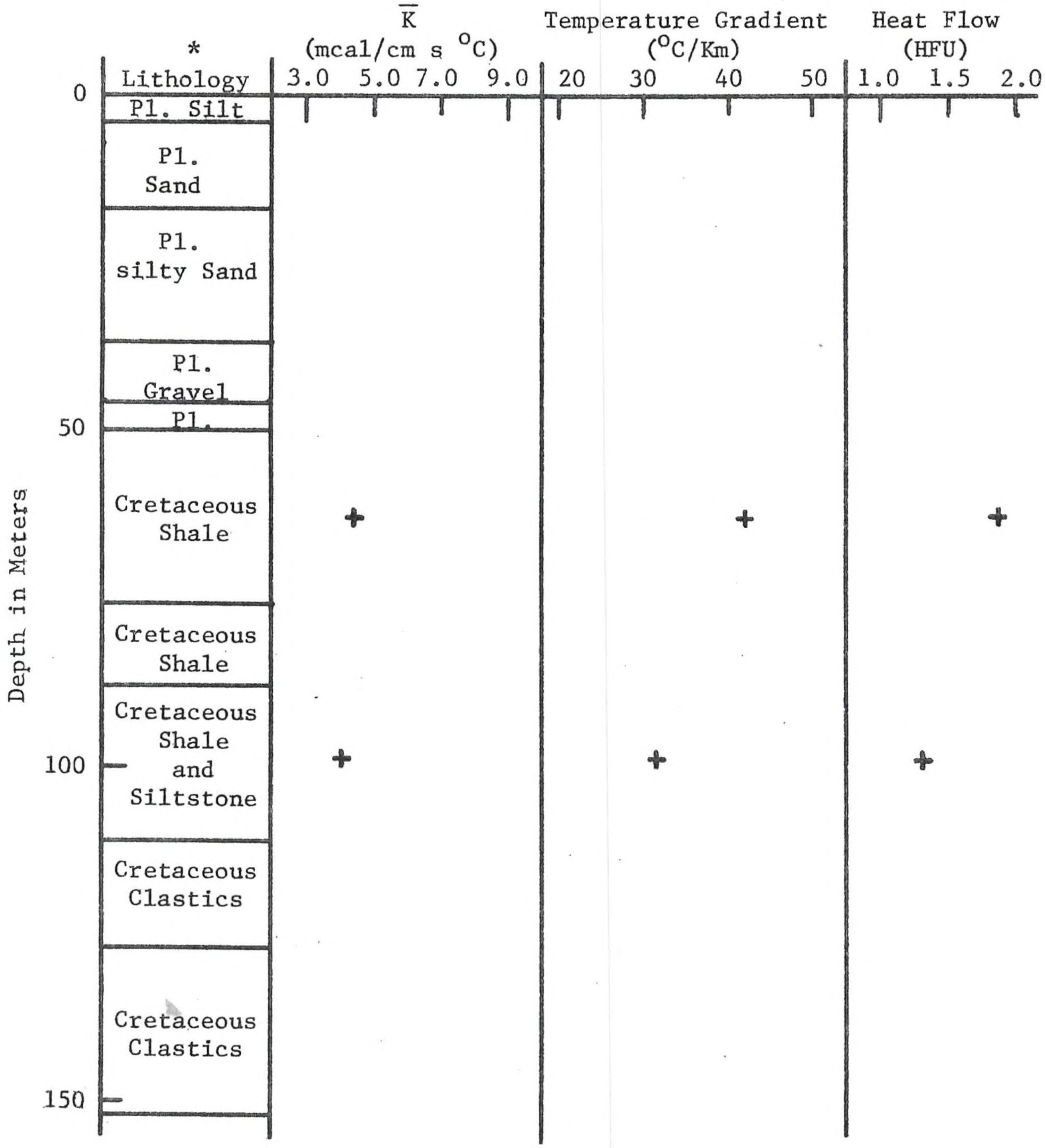


Figure 6

HEAT FLOW RESULTS

RRVD #8A

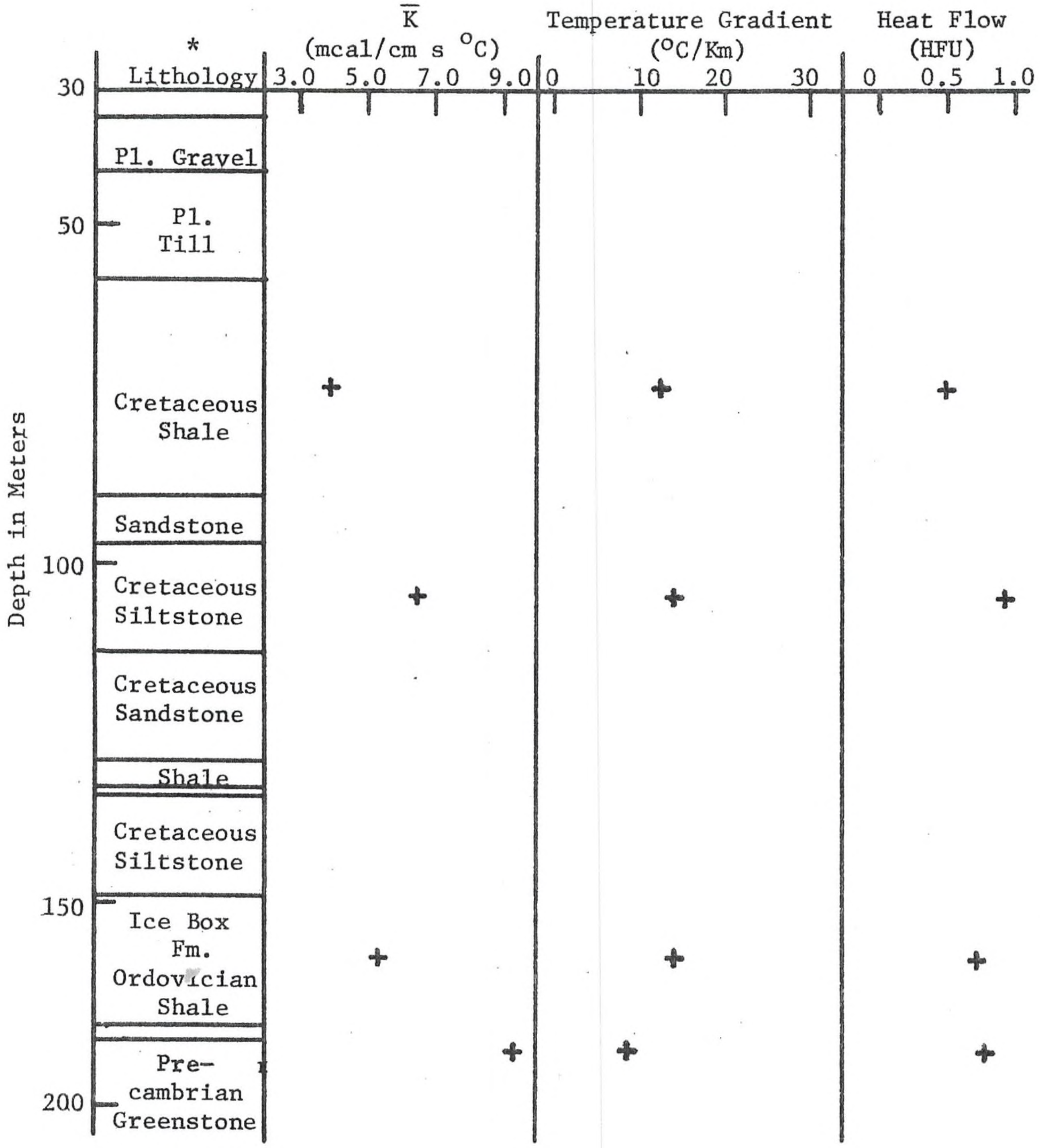


* See also figure 3 for better detail.

Figure 7

HEAT FLOW RESULTS

RRVD #10



* See also figure 4 for better detail.

Figure 8

DISCUSSION

The observed results are not of the quality originally hoped for due to the near total loss of access to the Precambrian rock for measurement. Under the experimental conditions observed, it is likely that both the temperature gradients and the thermal conductivities are not as reliable as would have been possible with the use of thick layers of hard rock.

Under ideal conditions the needle probe apparatus has an absolute accuracy of 3 to 4% (Von Hersen and Maxwell 1959), with the divided bar method slightly better than this (Roy, Decker, Blackwell and Birch 1968). A napthalene sample was run as a check on the needle probe. The average value obtained, 0.78 mcal/cm s ^{o}C, differs by only 2.8% from the recognized value of 0.804 mcal/cm s ^{o}C (International Critical Tables 1929). A third fused quartz standard was used as a sample to check the divided bar apparatus. Values of 3.26, 3.34 and 3.30 mcal/cm s ^{o}C were obtained in excellent agreement with the value of 3.30 mcal/cm s ^{o}C known for fused quartz.

Since the sedimentary and weathered Precambrian layers used for measurement are all deep enough to be water saturated, saturating the samples with distilled water before measurement should approximate the in situ water content. The layers from which drill cuttings were used were all high in silt content with traces of fine sands present (Moore 1978). These samples, when wet, packed together very well as the individual chunks of material disintegrated to form a single solid sample very similar in appearance to the siltstone core used from RRVD #10. The thermal conductivities measured on the drill cutting samples fall within the range of 3 to 5 mcal/cm s ^{o}C which is quoted by Sass, Munroe and Lachenbruch (1968) for similar materials. Fairly good consistency was observed for the samples. With the apparatus checks mentioned, the largest observed standard deviation in a layer's thermal conductivity measurements was assumed to be a reasonable measure of the uncertainty.

The temperature gradients are assumed to have uncertainties on the order of the statistical standard deviations, except for the values from RRVD #8A which are based on only one temperature log. Based on the depth and temperature measurement precision, and the layer thicknesses and temperature changes observed in RRVD #8A, a minimum uncertainty of 3% is set for those values. It may be noted that this is close to the statistical standard deviations observed in the two upper Cretaceous shale layers used for measurement in RRVD #2.

The heat flow values listed in tables 1, 2 and 3 are felt to represent the actual conductive heat flows present with one possible exception, the basal Cretaceous clastics layer in RRVD #2. It seems possible that some convective heat flow due to water movement exists within this layer.

Only the values for the Precambrian greenstone of RRVD #10 and the weathered Precambrian of RRVD #2 can realistically be said to represent the conductive heat flow in the upper portion of the Precambrian rocks of the area. The difference between these two values (0.76 and 1.21 HFU respectively) is believed representative of a real difference in the heat flows at these two locations. Both are in layers of nearly the same depth and are not separated by a very great land distance (about 160 km). This would indicate that any correction for past climatic history (Lee 1965, p.12) would be almost identical for the pair.

The value of 0.76 HFU for the Precambrian greenstone in RRVD #10 is close to the uncorrected value of 0.70 HFU observed in the Precambrian rock in a well in Winnipeg, Manitoba (Jessop and Judge 1970). In both wells the Precambrian rock has the same aquifer lying directly on top of it. This aquifer, commonly called the Winnipeg sand, is fairly thick (57 meters) in the Winnipeg well and fairly thin (2 meters) in RRVD #10. Jessop and Judge mention nonequilibrium water motion in the Winnipeg sand as possibly affecting their heat flow determination in the upper portion of the Precambrian gneisses in the Winnipeg well.

It is conceivable that even with only 2 meters of this sand above the Precambrian greenstone in RRVD #10, water movement in this sand could affect the observed heat flow since only the top 3 meters of the greenstone were accessible for temperature gradient measurement. Insufficient information is available to make a positive statement one way or the other in this matter. However, the agreement with the Winnipeg value, which was obtained in an interval of from 70 to 390 meters below the top surface of the Precambrian gneiss, indicates that the RRVD #10 greenstone value is probably not greatly in error.

The weathered Precambrian in RRVD #2 is an in situ weathered layer that grades downward into the underlying Precambrian chlorite schist. This weathered layer is believed to be impermeable, while the layer of basal Cretaceous clastics directly on top of it is probably permeable to some extent (Moore 1979, personal communication). While water movement in the Cretaceous clastics could affect the heat flow in the top of the weathered Precambrian, the temperature gradient in the upper portion of the weathered layer is not noticeably different from that at the bottom of the layer. Thus, if there is an effect, it is not observed in the data. The heat flow value of 1.21 HFU observed in this layer is close to values obtained by Scattolini (1978) farther to the west in North Dakota.

Thus, these two values from RRVD #2 and RRVD #10 (1.21 and 0.76 HFU respectively) represent the heat flows at these sites. Their difference is indicative of a substantial change in heat flow over the separation distance of 160 km. The reduced heat flow (mantle heat flow) in the Eastern United States is approximately 0.8±0.1 HFU (Roy, Blackwell and Decker 1972). Measureable mantle heat flow variations over a distance of 160 km are unlikely. The most probable explanation for the heat flow change is a difference in the heat production of the underlying rock materials. As of yet, no heat production measurements have been carried out for these sites. A more complete explanation of the heat flow data reported here is dependent upon the completion of such measurements.

The local hydrology seems to be a factor in all of the other heat flow values reported in this thesis. A likely aquifer is near each stratum from which these values come.

The Ordovician shale, the Cretaceous siltstone and the Cretaceous shale used for measurement in RRVD #10 are all in contact with sandstone directly beneath them. In each case, if cold water is moving horizontally through the sandstone heat from below would be convectively carried off, thus lowering the observed heat flow in the layer above the sandstone relative to the layer below it. If warm water is present, the effect would be reversed. On the basis of available information, this seems to be the most likely cause for the heat flow variations in RRVD #10.

Even though no effect of water motion in the basal Cretaceous clastics is noticed in the weathered Precambrian of RRVD #2, the presence of warm water movement would help to explain the heat flow profile of the well as a whole. Convection vertically in the clastics layer would have the effect of decreasing the observed temperature gradient relative to that which would be observed if no convection were present. This would cause a decrease in the observed heat flow in the layer. At the same time, the heat brought in by the water would cause an increase in the heat flow in the layers directly above. This is in fact the heat flow trend observed.

Similar reasoning could be used to help explain the relatively high heat flow values in RRVD #8A. The Cretaceous shale and interbedded shale and siltstone layers in which heat flows were determined are each in contact with a third shale layer. The interbedded shale and siltstone layer is also in contact with a basal Cretaceous clastics layer below it. In this well a good case for the existence of water movement in the clastics layer can be made from the occurrence of an artesian flow during drilling (Moore 1978). While sufficient evidence does not exist to prove this possible link between the heat flows in RRVD #2 and RRVD #8A, the simplicity of the explanation makes it an attractive one.

Regardless of this last speculation, it is quite clear that conditions are substantially different between RRVD #8A and RRVD #10. The drastic difference (42.10 and 12.58°C/km respectively) of the gradients in similar Cretaceous shales at nearly the same depths (50.3 and 57.9 meters respectively) is quite striking. Again, it is not known if the presence of different aquifers having different recharge areas is entirely responsible for the difference, but it seems a reasonable explanation for at least part of it.

To obtain more reliable conductive heat flow values than are reported in this thesis for this area, it will be necessary to utilize wells that penetrate the Precambrian rock more deeply than were available for this work.

The results of this research do seem to indicate that heat flow measurements in sedimentary layers could prove helpful in identifying and mapping water movement in aquifers.

APPENDIX

TEMPERATURE LOGS
WELL: RRVD #2

DEPTH (meters)	20 Aug. 78 T(^oC)	21 Oct. 78 T(^oC)	17 May 79 T(^oC)
73.15	9.44	9.61	9.09
77.72		9.67	9.23
82.30	9.78	9.89	9.43
86.87		9.88	9.67
91.44	10.28	10.13	9.88
96.01		10.31	10.09
100.58	10.65	10.48	10.25
105.16		10.67	10.43
109.73	11.03		10.67
114.30		11.12	10.86
118.87	11.51		11.09
123.44		11.55	11.33
128.02	11.90		11.53
132.59		12.01	11.77
137.16	12.33		12.01
141.73		12.49	
146.30	12.82		12.45
150.88		12.87	
155.45	13.28		12.92
160.02		13.37	
164.59	13.83		13.35
169.16		14.00	
173.74	14.17		13.99
178.31		14.23	
182.88	14.40		14.25
187.45		14.44	
192.02	14.58		14.40
196.60		14.70	
201.17	14.80		14.61
205.74		14.91	14.66
210.31	15.05		14.79
214.88		15.10	14.86
219.46	15.29	15.19	
224.03		15.33	15.06
228.60	15.44	15.37	15.12
233.17	15.50	15.44	15.19
237.74	15.60	15.51	15.25
242.32	15.69	15.53	15.34
246.89	15.85		
248.41	15.86		

TEMPERATURE LOG
WELL: RRVD #8A

DEPTH (meters)	3 Aug. 78 T(^{o}C)
9.14	7.49
18.29	7.45
27.43	7.66
36.58	7.60
45.72	7.73
54.86	8.11
64.01	8.46
73.15	8.88
82.30	9.41
91.44	9.75
100.58	10.04
107.29	10.25

TEMPERATURE LOGS
WELL: RRVD #10

DEPTH (meters)	8 Aug. 78 T(^oC)	4 Nov. 78 T(^oC)	15 May 79 T(^oC)	23 May 79 T(^oC)
9.14	6.73			
18.29	7.19			
27.43	7.29			
36.58	7.27			
45.72	7.39			
54.86	7.47			
64.01	7.59	7.39	7.22	7.20
68.58		7.40	7.28	7.25
73.15	7.75	7.46	7.33	7.32
77.72		7.54	7.40	7.38
82.30	7.79	7.60	7.46	7.45
86.87		7.67	7.51	7.50
91.44	8.04	7.68	7.56	7.54
96.01		7.70	7.60	7.61
100.58	8.10	7.76	7.64	7.63
105.16		7.80	7.69	7.68
109.73	8.24	7.87	7.79	7.75
114.30		7.92	7.84	7.81
118.87	8.34	7.95	7.88	7.88
123.44		8.00	7.93	7.90
128.02	8.40		7.97	7.95
132.59			8.03	8.02
137.16	8.55		8.10	8.07
141.73			8.16	8.16
146.30	8.66		8.22	8.20
150.88			8.31	8.28
152.40	8.79			
155.45			8.41	8.39
156.97	8.85			
160.02			8.48	8.43
161.54	8.90			
164.59			8.55	8.51
166.12	8.87			
167.64			8.60	8.56
170.69	8.85	8.84	8.63	8.61
171.60	8.85			
171.91				8.62
172.52	8.86			
173.13		8.86	8.65	8.65
173.43	8.88			

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