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**Geothermal Heat Flux at the COST B-2 and B-3 Wells,
U. S. Atlantic Continental Margin**

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ABSTRACT

Heat flow estimates at two sites on the U.S. Atlantic continental margin are presented. An estimate of the heat flowing from the basement also has been obtained. About 4.8 km of sediments penetrated at the COST B-2 and 4.0 km at the COST B-3 were deposited since the Upper Jurassic. Well logs were used to evaluate thermal gradients and sedimentation rates, whereas thermal conductivities and radiogenic heat productions were measured on drill cuttings samples.

A procedure to estimate in-situ thermal conductivity from drill cuttings and well logs is described. A substantial set of samples, in the form of drill cuttings, were sorted in four major lithologies: sandstones, siltstones, shales and limestones. Laboratory measurements of density, porosity, thermal conductivity, quartz (%), potassium (%), uranium (ppm) and thorium (ppm) were performed on 128 reorganized and pulverized samples. A significant correlation of the matrix thermal conductivity to quartz and potassium content was found. In situ porosity and volume fraction of each lithology, determined mainly from well logs, were used to calculate in situ mean thermal conductivity. Finally the mean in situ vertical component of the thermal conductivity, as required for heat flow values, has been estimated from a correction factor for the anisotropy of each lithology. The in-situ temperature and anisotropy effects substantially decrease estimates of thermal conductivity at depth.

Below the uppermost 1 km in both wells the best estimate of the thermal gradient is $26.3^{\circ}\text{C km}^{-1}$ at COST B-2 and $26.1^{\circ}\text{C km}^{-1}$ at COST B-3, whereas in situ mean thermal conductivities range between about 1.8 and $1.9 \text{ W m}^{-1} \text{ K}^{-1}$ (4.3 – 4.5 T.C.U.). The average heat flow is estimated as about 45 mWm^{-2} (1.07 H.F.U.) at COST B-2 and 44 mWm^{-2} (1.06 H.F.U.) at COST B-3, with an uncertainty of about 20–25%. The mean radiogenic production in sediments at the two sites has been estimated as 1.83 (COST B-2) and 1.44 (COST B-3) 10^{-6} Wm^{-3} . With a 12–14 km thick sedimentary sequence a radioactive contribution of 20 – 25 mWm^{-2} can be expected.

The effects of sediment deposition, compaction, pore water advection and radiogenic heat production have been combined in a numerical model (Hutchinson, 1985) to estimate the undisturbed basement heat flux. Although the sedimentation depresses the basement heat flux by 15–20%, this effect is more than compensated by radioactive heat production in the sediments, so that the surface flux is estimated to be higher than that from the basement. The latter is calculated at about 33 – 39 mWm^{-2} (0.8 – 0.9 H.F.U.), a relatively low value. The overall uncertainty is about ± 20 – 25% , and other estimates on continental margins with thick sediments (e.g. Reiter and Jessop, 1985) probably have at least a similar uncertainty.

1. INTRODUCTION

The evaluation of heat flow at the earth's surface is a complex problem which requires data on thermal conditions at depth through time. Important information about the nature, age and evolution of the crust can be obtained if a reliable estimate of the two main components of surface heat flow, i.e. radiogenic heat production in the lithosphere and heat flow transferred from the asthenosphere, is available (Keen and Lewis, 1982). The geological, geophysical and deep exploratory drilling data existing along the U.S. Atlantic continental margin give us an opportunity both for evaluation of these heat flow components through this province, and to propose a general procedure for the determination of heat flow starting from the usually available data from deep exploratory and/or oil wells.

An extensive regional study on terrestrial heat flow across the Canadian Atlantic continental margin has been recently published by Reiter and Jessop (1985) starting from bottom-hole-temperatures and thermal conductivities, estimated from lithologic descriptions, of 72 sites on the eastern Canadian shelf. This paper will analyse in detail only 2 deep wells drilled offshore New Jersey on the same north Atlantic margin of the U.S.A. Data from the COST B-2 and COST B-3 wells, drilled in the Baltimore Canyon Trough (Fig. 1), have been analyzed.

The evaluation of heat flow and thermal structure of the COST wells will contribute to a better understanding of the thermal history following continental margin formation. In particular, accurate measurements of heat flow in deep wells could be extremely useful in testing some of the thermal models (McKenzie, 1978; Royden et al., 1980; Hellinger and Sclater, 1983; and others) developed to explain thermally driven subsidence.

Continental rifting and crustal thinning are accompanied by extensive heating and vertical movements (uplift and initial subsidence) of the margins, evidenced most directly by listric faulting and volcanic intrusions; the rifting process is completed when a mid-ocean ridge is created. Following the onset of sea floor spreading, heat is lost from the margin both vertically and horizontally (Steckler, 1981) making the long term thermal subsidence (McKenzie, 1978; Hellinger and Sclater, 1983) more rapid and variable.

The amount of heating has important implications for the sediments that accumulate on the subsiding continental margin (Steckler and Watts, 1982); they will be heated during and after rifting, possibly creating the conditions for, or accelerating the process of, oil and gas maturation. Thermal models can predict, within restricted assumptions (initial lithosphere in isostatic and thermal equilibrium, heat transfer by conduction only, uniqueness of rifting event, etc.), the temperatures and heat flowing out of the cooling lithosphere for different times during continental margin formation. More accurate heat flow estimates in deep wells will better constrain such models, verifying the reliability of the initial assumptions.

In dealing with the heat flow determinations from well data, a major problem is the estimate of the thermal conductivity of the sedimentary formations penetrated, particularly when core sampling is inadequate or nonexistent. Many different approaches have been proposed to evaluate the thermal conductivity of sediments based only on standard logs (Goss et al., 1975;

Fig. 1. Geological map showing location of COST B-2 and B-3 wells and other exploratory wells off New Jersey (from Watts, 1981).

Evans, 1977; Houbolt and Wells, 1980; Drury and Jessop, 1983), although a general empirical and simple relation has not yet been determined. Radioactivity in sediments may contribute significantly to surface heat flux wherever the sediments are more than a few km thick. Radiogenic heat production has been approached by measuring directly the U, Th and ^{40}K concentrations in 122 samples of various sediment lithologies sorted from the drill cuttings originally available. In the opposite sense, the rapid accumulation of sediments at a margin suppresses the heat flux from the subsiding basement, depending primarily the sedimentation rate (Von Herzen and Uyeda, 1963; Hutchison, 1985).

The general approach followed in analyzing the COST B-2 and COST B-3 data is summarized in Fig. 2. Basically, a few hundred drill cuttings samples are separated into various lithologies, ground up, and the porosity, density, and thermal conductivity are measured in the saturated samples. The in situ thermal conductivity is estimated from well log parameters taking into account the anisotropy correction.

The surface heat flow is then computed by combining these data with the thermal gradients obtained from temperature logs. The thermal blanketing effect induced by the sedimentation is also considered, and radiogenic heat production in the sediments is estimated from U, Th and ^{40}K concentration measurements.

2. GEOLOGICAL SETTING

The U.S. Atlantic passive continental margin with its substantial thickness of sedimentary rocks contains a well preserved record of the evolutionary history of the area since the early opening of the Atlantic Ocean. During the past two decades it has been the site of extensive geological and geophysical investigation by academic, government and industry groups both for scientific and economic interest.

Gravity, magnetic, seismic and deep exploratory drilling data are available in many areas along the margin from Nova Scotia to Florida (Rabinowitz, 1974; Taylor et al., 1968; Schouten and Klitgord, 1977; Klitgord and Behrendt, 1979; Schlee et al., 1976; Sheridan, 1976; Grow and Markl, 1977; Grow et al., 1979; Buffler et al., 1979; Grow, 1980, 1981; Schlee, 1981; Scholle et al., 1977, 1980; Amato and Bebout, 1978). Based on these data, the general understanding of the origin and evolution of the Atlantic-type continental margin has been significantly improved (Grow and Klitgord, 1980; Watts, 1981; Watts and Thorne, 1984).

The multichannel seismic reflection profiles have shown considerable thicknesses of sediments beneath the shelf (up to 14 km) and slope (up to 8 km) regions in a sequence of basins separated by structural highs (Fig. 3). The COST test wells and more than 20 exploratory drillings (Fig. 1) provide information on stratigraphy, age and physical parameters of the seismic reflectors to depths of 5 km underlying the outer part of the shelf. These wells show a sedimentary section composed mainly of sandstones and shales with lesser amounts of limestone and subordinate amounts of coal and lignite (Scholle et al., 1977, 1980). Biostratigraphic studies indicate that these sediments are Jurassic to Quarternary in age and were deposited in continental or shallow marine environments.

Fig. 2. Block diagram of the sequence of procedures followed in the COST B-2 and B-3 data analysis for evaluation of the heat flow.

Fig. 3. Complete stratigraphic model cross-section of the U.S. Atlantic continental margin (after Watts and Thorne, 1984) including 2-layer stretching, flexure, variable compaction, sea level changes and slow erosion. Note location COST B-2/B-3 wells.

The margin formed by rifting and subsequent separation of the North American and African plates in Late Triassic-Early Jurassic time. The igneous and/or metamorphic basement (of Precambrian to Paleozoic age) involved in the rifting process started subsiding in response to the isostatic and thermal readjustments of the deeper structures of both lithosphere and asthenosphere. The transitional area between stretched continental and new forming oceanic lithosphere was exposed to significant vertical movements which allowed a large accumulation of sedimentary rocks.

The sediment column contains valuable information of the subsidence history of the underlying crust which is transitional in nature between continental and oceanic (Watts, 1981). The first marine transgression occurred in Late Triassic to Early Jurassic times on a rapidly subsiding shelf. In near-shore regions sands and shales were deposited, while in offshore regions limestones accumulated; they have been sampled in deep wells off Nova Scotia and on Georges Bank (Jansa and Wade, 1975; Grow and Sheridan, 1981).

Reefs formed in some parts of the slope regions, acting as barriers to terrigenous sediments infilling the newly formed basins. Growth of reefs may have continued until Cretaceous time when they became buried by prograding sands and shales. The margin continued to prograde during Late Cretaceous and Tertiary with sediment accumulation on both shelf and slope marine environments (Schlee, 1981; Watts, 1981). Erosion of the outer margin by bottom currents began in Middle Tertiary; further erosion occurred in a Late Pliocene and Pleistocene lowering of sea level.

A simple crustal cross-section off New Jersey with computed gravity and geoid effect (Fig. 4) shows little correlation with the extent of the trough indicating that the region is in isostatic equilibrium. The largest gravity anomalies are more likely related to the transition between oceanic and continental crust. In the same way the East coast magnetic anomaly (ECMA) could be correlated approximately with the boundary between oceanic and transitional crust (Rabinowitz, 1982).

A good observational data set along the margin (Watts, 1981) supports crustal extension during the early rifting history. Moreover a substantial amount of thinning must have occurred at the margin in order to explain the total tectonic subsidence at the COST wells.

Simple thermal models applied to the evolution of the margin can predict subsidence and heat flow through time, even if the large thickness of syn-rift and post-rift sediments makes it difficult to distinguish between the different models. Watts and Steckler (1981) have shown that the tectonic subsidence of the Atlantic margin can be equally well explained by a cooling plate model (Parsons and Sclater, 1977) and the stretching model of McKenzie (1978) with the stretching factor $\beta=6$. The heat flow predicted by these models mainly differs in the first 60 M.y. after rifting and thereafter tends asymptotically to the same equilibrium value. A complete stratigraphic model across the U.S. Atlantic continental margin (close to the COST wells) was proposed by Watts and Thorne (1984) (Fig. 3). This model includes: 2-layer stretching, flexure, variable compaction, sea level changes and slow erosion.

Fig. 4. Computed gravity and geoid effect (heavy curve) of a simple model for the crustal structure off New Jersey compared to observed (thin curve) free-air gravity anomaly and GEOS-3 satellite altimeter data. The numbers on the section are densities, and the dashed lines indicate the estimate of crustal thickness at the COST wells based on thermal models (from Watts, 1981, Fig. 29).

3. COST B-2 AND COST B-3 DATA SET

The COST B-2 and COST B-3 wells are two major deep stratigraphic tests located respectively on the U.S. Atlantic outer continental shelf and slope which have a continuing interest for both the petroleum industry and the scientific and academic community. Although the drilling of both wells was financed by the oil industry, the data from the COST holes were made generally available shortly after drilling for scientific and other purposes. This data includes considerable basic operational, lithologic, stratigraphic and geochemical information, most of it derived from industry reports (Smith et al., 1976; Amato and Simonis, 1979). Preliminary results have been published in Scholle et al. (1977) for the COST B-2 well and for the COST B-3 in Scholle et al. (1980). Drill cutting samples collected at discrete depth intervals during drilling were made available to us by ARCO and Phillips Petroleum oil companies (Tables 1 and 2).

4. LABORATORY MEASUREMENTS AND DATA PROCESSING

Our approach in analyzing the COST B-2 and COST B-3 data was to establish a general procedure to evaluate the heat flow of a given area starting from the usually available information from deep exploratory and/or oil wells. In particular, reliable estimates of thermal conductivity of the sedimentary materials penetrated by such wells are needed, even when (in most cases) core sampling is inadequate or non-existent.

A block diagram of the main steps followed in analyzing COST B-2 and B-3 data is summarized in Fig. 2. Basically, the drill cuttings samples are separated into various lithologies and their thermal conductivity measured (Sass et al., 1971). The in-situ thermal conductivity is estimated from well log parameters, and combined with thermal gradients measured in the hole to give heat flow values, following the usual methods. Radioactive elements in the sediments are determined quantitatively so that the contribution of heat generation within the sediments may be estimated.

The laboratory samples treatment, described below in detail, allowed us to measure the thermal conductivity of the saturated samples and to calculate their porosities and densities. With these two parameters it was possible to derive the solid thermal conductivity at zero porosity (Φ). By measuring the quartz content (weight percent) with x-ray diffraction, we could verify the strong positive correlation between quartz content and thermal conductivity of the solid component. The uranium and thorium contents of samples were measured with Neutron Activation methods and potassium by X-ray spectrometry and atomic absorption spectrophotometry to evaluate the radioactive heat production in the sediments.

The geophysical well logs were used to calculate the in situ porosity and the equilibrium temperature at discrete points. From these data the temperature gradients and the vertical component of in situ thermal conductivity K_z were estimated, assuming an anisotropy correction factor of conductivity for the primary lithologies from the literature. The surface heat flow was computed and the heat flow from the lithosphere was estimated taking into account the thermal blanketing effect of the sedimentation and the radiogenic heat production, which have overall contributions of opposite sign.

Table 1. Depth Intervals of COST B-2 Drill Cuttings Samples
(ft. below K.B.)

4110-4140	7980- 7990	12960-12970
4140-4170	8000- 8010	13160-13170
4170-4200	8220- 8240	13200-13210
4200-4230	8250- 8260	13220-13230
4230-4260	8450- 8460	13240-13250
4620-4630	8470- 8480	13260-13270
4640-4650	8530- 8540	13300-13310
4660-4670	8550- 8560	13340-13350
4880-4890	9290- 9300	13380-13390
4900-4910	9310- 9320	13410-13420
4920-4930	9370- 9380	13630-13640
5090-5100	9390- 9400	13690-13700
5110-5140	9620- 9630	13710-13720
5140-5170	9640- 9650	13870-13880
5200-5230	9660- 9670	14010-14020
6280-6290	9680- 9690	14030-14040
6300-6310	9700- 9710	15260-15270
6320-6330	10060-10070	15280-15290
6530-6540	10080-10090	15300-15310
6560-6560	10100-10110	15320-15330
6770-6780	10270-10300	15340-15350
6790-6800	11050-11080	15530-15540
6850-6860	11260-11290	15770-15780
6870-6880	11350-11380	15780-15790
6890-6900	11380-11410	15790-15800
7070-7080	11530-11560	15820-15830
7140-7150	11650-11680	15840-15850
7160-7170	11710-11740	15910-15920
7180-7190	11740-11770	
7220-7230	11770-11800	
7280-7290	11800-11830	
7320-7330	11980-12000	
7400-7410	12000-12020	
7540-7550	12370-12400	
7580-7590	12460-12490	
7720-7730	12550-12580	
7740-7750	12640-12650	
7760-7770	12680-12690	
7920-7930	12720-12730	
7940-7950	12740-12750	
7960-7970	12860-12870	
	12940-12950	

Table 2. Depth Intervals of COST B-3 Drill Cuttings Samples
(feet below K.B.)

3810-3840	6300-6360	9700- 9720	12090-12100
3840-3870	6400-6430	9800- 9850	12210-12220
3930-3960	6430-6460	9870- 9890	12300-12310
3960-3990	6500-6520	9900- 9930	12310-12320
4020-4050	6520-6550	9930- 9950	12390-12400
4080-4110	6600-6630	10000-10030	12420-12430
4290-4320	6630-6650	10100-10130	12490-12500
4380-4310	6700-6720	10130-10160	12580-12590
4440-4470	6730-6750	10200-10240	12620-12630
4560-4590	6760-6770	10300-10330	12660-12670
4590-4620	6800-6820	10400-10450	12690-12700
4650-4680	6830-6840	10450-10480	12750-12760
4800-4830	6900-6930	10600-10640	12780-12790
4890-4920	7000-7020	10700-10720	12820-12830
4950-4980	7030-7040	10720-10750	12900-12910
5010-5040	7100-7140	10790-10820	13040-13050
5100-5130	7200-7220	10920-10950	13190-13200
5130-5160	7230-7250	11030-11050	13240-13250
5220-5250	7300-7330	11060-11070	13350-13360
5310-5340	7330-7350	11120-11150	13440-13450
5400-5430	7400-7430	11170-11200	13550-13560
5490-5520	7500-7540	11210-11220	13680-13690
5610-5640	7600-7640	11220-11230	13750-13760
5730-5760	7700-7720	11250-11260	13870-13880
5790-5800	7720-7760	11310-11320	13970-13980
5830-5840	7810-7840	11320-11330	14020-14030
5870-5880	7900-7930	11390-11400	14080-14090
5900-5910	8000-8030	11490-11500	14200-14210
5940-5950	9000-9030	11560-11570	14310-14320
5980-5990	9100-9130	11650-11660	14380-14390
6020-6030	9200-9230	11700-11710	14440-14450
6050-6060	9300-9330	11800-11810	14550-14560
6180-6190	9400-9430	11910-11920	14620-14630
6260-6270	9500-9530	12000-12010	14710-14720

14860-14870
14950-14960
15070-15080
15190-15200
15250-15260
15360-15370
15470-15480
15550-15560
15680-15690
15750-15760
15790-15800

4.1 SAMPLE PREPARATION

The sample preparation and the measurements of porosity, density and thermal conductivity consumed most of the time and effort of this project; for the most part they were completed by four summer students between 1979 and 1982. A summary of the data distribution is presented in Table 3.

Table 3 - Summary of Cuttings Samples Used for Thermal Conductivity Measurements

	Original no. of samples	Depth Interval (ft. below K.B.)	Operator	No. Recombined samples	No. Reliable measurements
COST B-2	116	4,110-12,970	G. Moritz	38(?)	38
		11,980-15,930	L. Hobbie	47(?)	47
COST B-3	151	3,810-11,200	P. Borga	29	24
		11,210-14,630	K. Lange	16	16

*NB: the difference between the number of recombined samples and the number of final reliable values is due to the loss of 3 samples in the centrifuge and to an insufficient amount of material in another 2 samples.

Each operator used slightly different procedures in the treatment of samples mainly because the sorting of the samples by lithology and the accuracy of procedures (grinding, saturation, weighing) are somewhat subjective and therefore operator dependent. For the detailed description of the procedures used, refer to the original reports (Moritz, 1979; Hobbie, 1980; Borga, 1981; Lange, 1982). In any case, the main steps generally followed can be briefly described as follows:

- a) Washing: the rock chips were washed through a seive with a 1 mm opening size to remove drilling mud and miscellaneous fine particles.
- b) Drying: they were dried in an oven at 50-60°C for a few hours.
- c) Sorting: the rock fragments were sorted in 5 lithology groups. Sandstones, siltstones, shales, limestones and coal. The criteria used were mainly based on visible characteristics such as: grain size, texture, habit, shape of breaking surface, hardness, color, hydrochloric acid fizzing test.
- d) Recombining and labeling: a minimum amount of 40-45g of each lithotype from each depth interval is required for a thermal conductivity measurement and for a γ -ray spectrometry analysis. When insufficient material was available from a sample (depth interval), identical lithologies from nearby depth intervals had to be combined to obtain the minimum weight. These sorted and recombined samples were properly labelled.

- e) Grinding: The material must be reduced to a uniform grain size to correctly measure the thermal conductivity. This was accomplished by hand-grinding the rock chips with mortar and pestle. For the quartz analysis on the x-ray diffraction machine all the samples were reground again to a grain size of approximately 62 microns ($\mu = 10^{-6}\text{m}$).
- f) Drying: The pulverized samples were dried for at least 12 hours in an oven.

4.2 SAMPLES POROSITY AND DENSITY COMPUTATION

All the powdered samples weighed to an accuracy of ± 0.01 g are saturated with water in a moderate vacuum (~ 20 torr) and agitated to eliminate air bubbles. The samples are generally left for 12-36 hrs. to settle. A few COST B-3 samples (Borga, 1981) were run in a centrifuge for 3-5 hrs. at 200-500 rpm to enhance settling, but no significant effect was noticed. A complete settling is almost impossible to reach and the quality of the results is directly related to the grain size of the samples. The excess water above the upper sample interface is then removed either by a filter paper or a dropper and the amount of suspended sediment lost in this operation is measured after evaporation of the water in a drying oven.

The calculation of porosity (Φ) and the density (ρ) of the samples is simply calculated from the measured weights and volumes, e.g., Goldberg et al. (1980). The final relations are:

$$\Phi = \frac{V_w}{V_w + V_s} \quad (1)$$

$$\rho = \frac{W_s}{V_s} \quad (2)$$

where

$V_w = W_w =$ volume or weight of interstitial water (assuming $\rho_w = 1$)

$W_s =$ weight of the sediment

$V_s = W_w^1 - W_w =$ volume of sediment, where $W_w^1 =$ weight of equivalent volume of water

4.3 THERMAL CONDUCTIVITY MEASUREMENTS

The thermal conductivity of the pulverized and saturated samples was measured with the needle-probe technique (Von Herzen and Maxwell, 1959) following the procedure illustrated in Goldberg et al. (1980). The results are presented in Table 4.

Among the models proposed to calculate the thermal conductivity of a two-phase aggregate knowing the conductivities and volume fractions of the components (Zierfuss and Van der Vliet, 1956; Woodside and Messmer, 1961a, 1961b; Brailsford and Major, 1964; Budiansky, 1970; Adler et al., 1973;

Table 4.
LIST OF COST B-2 SAMPLE MEASUREMENTS

Samp. No.	Depth Interval (feet)	Lithology	Sample Porosity	Sample Density	K _{MEASURED} (TCU)	K matrix (TCU)	Quartz (%)	Potassium (%)	⁴⁰ K γ-ray (ppm)	U (ppm)	Th (ppm)	Th/U	A (μkm ⁻³)
46	4110-4200	*LS.	.675	2.60	2.50	7.42	7.4	0.55		1.84	3.2		0.72
44	~4150	LS.	.642	(1.72)2.4	2.16	4.31	4.8	0.71		3.99	4.3		(0.89) 1.23
45	4200-4900	"	.633	2.38	(2.63)	7.12	5.5	0.74		2.55	3.1		0.83
45	"	"	(.419)	2.43	(2.61)	(3.93)	2.5	0.08	0.09	5.34	3.5		1.46
42	5090-5140	Sh.	.640	2.34	2.32	5.36	16.6	1.96		4.93	14.3	2.89	2.11
42a	5090-5140	*Sl.		2.5**		?	18.7	1.38		8.54	4.4		2.43**
40	5110-6890	Ss.	(.370)	2.83	(4.61)	9.02	16.3	0.78	0.97	2.32	5.10	2.20	1.07
33	5140-7990	*LS.	(.649)	2.49	(2.53)	(6.83)	6.1	0.61		1.98	5.2	2.60	0.85
43	5200-6900	*LS.	.656	2.54	2.52	7.20	7.6	0.70		2.35	3.0		0.82
42	6530-6560	Ss.	(.438)	2.98	(4.17)	9.38	13.6	0.93		3.26	5.0	1.53	1.40
39	6770-7930	Ss.	(.401)	2.95	(4.49)	9.43	31.9	1.02	1.31	2.42	8.10	3.34	1.39
1/35	7070-7990	Sh.	.551	2.65	2.74	6.00	12.6	1.85		3.24	14.8	4.59	1.99
27	7540-8460	Ss.	(.413)	2.71	4.57	10.15	40.9	1.27		2.90	11.3	3.91	1.65
25	8000-8610	Sh.	.501	(1.74)2.6	3.00	6.21	12.2	2.07		3.36	15.3	4.54	(1.36) 2.03
21	8000-8730	Sh.	(.550)	(1.27)	(2.76)	(5.93)							
28	8220-8730	Ss.	(.475)	2.82	(3.85)	9.25	16.2	1.73		4.47	11.1	2.48	2.17
26	8470-8730	Ss.	(.421)	2.66	(4.69)	10.86	26.1	1.13		3.31	11.4	3.44	1.72
16	9290-9300	(Sl.)	(.664)	2.45	(2.40)	(6.41)	5.3	1.99		2.41	4.2	1.76	0.99
19	9290-9400	Ss.	.385	2.66	5.02	10.82	29.8	0.83		2.16	9.19	4.25	1.25
6/12	9290-10300	Sh.	.387	2.58	3.35	5.70	9.5	1.78	2.73	4.88	20.2	4.13	2.69
18	9310-9320	*LS.	.572	2.51	3.00	7.90	12.2	0.75		4.93	18.6	3.78	2.44
8	9370-9670	Sl.	.464	2.74	3.52	7.56							
3	9370-9850	Ss.	.405	2.73	4.85	10.92	43.3	1.01		3.18	16.3	5.14	2.06
15	9370-9850	Sh.	(.373)	2.40	(3.26)	(5.23)							
14	9620-9670	Sl.	.524	2.57	3.36	8.29	15.7	1.79		3.95	15.6	3.94	2.15
4	9640-9850	Ss.	.414	2.72	4.49	9.87	32.3	1.21		3.32	15.6	4.70	2.06
9	9660-9850	Ss.	.423	2.82	4.65	10.79	21.2	1.07		2.48	10.7	4.32	1.54
7	9660-9850	Sh.	(.501)	2.53	(3.01)	(6.14)							
11	10080-10090	Sh.	.556	2.54	2.97	7.31	11.6	2.13		4.33	15.4	3.55	2.23
57	10080-10110	*LS.	(.669)	2.63	(2.37)	(6.21)	4.5	0.63		2.22	2.9		0.81
53	10100-11560	Sh.	.508	2.54	3.17	7.09	11.8	2.10	3.05	4.42	15.9	3.59	2.29
5	10270-10300	Sh.	(.445)	2.27	3.23	6.22	12.4	1.69		4.93	16.9	3.42	2.18
51	11050-12750	Sh.	.404	2.76	2.94	4.75	9.1	1.89		5.04	14.4	2.86	2.52
52	11260-12970	Sl.	.493	2.43	3.45	7.88	15.8	1.99		4.78	14.4	3.00	2.17
50	11530-11830	Ss.	.432	2.58	4.58	10.88	41.1	1.42		5.34	19.5	3.66	2.72
58	11680-12970	Ss.	.430	2.55	4.51	10.49	16.5	1.27		2.92	10.2	3.49	1.49
1	11980-12580	Ss.		(2.65)		8.19	33.1	1.16	1.41	2.86	8.90	3.12	1.44
11	11980-12580	Sl.		(2.65)		6.49	11.1	1.86	2.34	3.98	15.6	3.93	2.23
59	12640-12870	Sh.	.429	2.47	3.08	5.42	8.6	2.11		4.57	15.0	3.29	2.20
54	12720-12750	Sh.	.423	2.56	2.89	4.78	7.8	2.24		4.45	14.0	3.15	2.20
55	12720-12750	Sh.	.441	2.58	3.04	5.46	9.1	2.11		4.51	14.3	3.17	2.24
12	12940-13350	Sl.		(2.65)		5.66	11.4	2.16		3.92	16.5	4.21	2.31
2	12940-13420	Ss.		(2.65)		10.30	30.1	1.14		3.27	13.5	4.12	1.84

Table 4. (cont.)

LIST OF COST B-2 SAMPLE MEASUREMENTS

Samp. No.	Depth Interval (feet)	Lithology	Sample Porosity	Sample Density	K _{MEASURED} (TCU)	K matrix (TCU)	Quartz (%)	Potassium (%)	⁴⁰ K γ-ray (ppm)	U (ppm)	Th (ppm)	Th/U	A (μMn ⁻³)
24	13340-13350	Sh.		(2.65)		5.40	8.0	2.02		4.89	15.0	3.07	2.44
25	13380-13390	Sh.		(2.65)		(6.44)	6.6	2.13		4.70	14.9	3.17	2.39
13	13380-13700	S1.		(2.65)		6.62	11.8	2.20	2.97	4.27	17.5	4.11	2.46
3	13630-13640	Ss.		(2.65)		11.16	28.4	1.10		2.75	8.49	3.09	1.37
26	13630-13640	Sh.		(2.65)		4.56	17.7	2.34		4.73	15.9	3.37	2.49
27	13690-13700	Sh.		(2.65)		(4.88)	7.2	2.25		5.22	14.4	2.77	2.50
5	13690-14040	Ss.		(2.65)		11.46	30.1	1.16	1.36	2.80	11.5	4.10	1.59
28	13710-13720	Sh.		(2.65)		4.83	6.7	2.28		5.15	11.8	2.28	2.31
16	13710-14040	S1.		(2.65)		6.00	13.4	2.29		4.44	16.3	3.67	2.43
?	13790-13800	Ss.		(2.65)		9.25							
14	13790-13800	S1.		(2.65)		6.71	22.4	2.27		4.47	14.6	3.27	2.33
29	13790-13800	Sh.		(2.65)		5.82	12.0	2.46		5.41	13.0	2.40	2.47
15	13870-13880	S1.		(2.65)		5.76	9.8	2.40	3.51	4.29	16.9	3.94	2.45
30	13870-13880	Sh.		(2.65)		(3.42)	9.1	2.43		5.26	14.0	2.66	2.50
4	13930-13940	Ss.		(2.65)		8.85	28.9	0.89		1.53	8.52	5.57	1.04
31	13930-13940	Sh.		(2.65)		5.03	10.3	2.48		5.61	15.7	2.81	2.71
32	14010-14020	Sh.		(2.65)		5.06	9.8	2.45		5.41	15.0	2.78	2.61
33	14030-14040	Sh.		(2.65)		(5.13)	10.3	2.45		5.64	14.7	2.61	2.64
34	15260-15270	Sh.		(2.65)		5.88	13.1	3.22	4.60	5.00	12.5	2.50	2.40
6	15260-15310	Ss.		(2.65)		10.55	38.6	0.70		2.46	9.72	3.95	1.34
17	15260-15330	S1.		(2.65)		8.29	15.5	2.63		4.38	14.9	3.40	2.50
35	15280-15290	Sh.		(2.65)		5.04	11.7	3.31		5.15	12.2	2.37	2.43
36	15300-15310	Sh.		(2.65)		5.28	14.1	3.20		4.80	12.0	2.49	2.32
37	15320-15330	Sh.		(2.65)		5.63	11.5	3.27		4.96	12.2	2.46	2.38
7	15320-15540	Ss.		(2.65)		9.56	30.3	0.48	0.75	1.62	7.43	4.58	0.96
18	15340-15540	S1.		(2.65)		7.07	20.7	2.62	3.74	4.06	16.1	3.97	2.36
38	15520-15530	Sh.		(2.65)		6.18	12.0	3.28		4.90	11.1	2.27	2.29
39	15530-15540	Sh.		(2.65)		5.18	12.8	3.28		4.62	14.3	3.10	2.44
8	15540-15550	Ss.		(2.65)		12.02	39.1	0.38		1.56	7.11	4.57	0.91
40	15540-15550	Sh.		(2.65)		5.11	17.8	3.25		4.89	13.5	2.76	2.45
49	15540-15780	S1.		(2.65)		7.07	18.9	2.61	3.63	4.33	15.4	3.55	2.38
41	15770-15780	Sh.		(2.65)		5.47	13.7	3.53		4.59	14.6	3.17	2.47
9	15770-15830	Ss.		(2.65)		10.51	33.2	1.27		2.97	12.7	4.29	1.73
42	15780-15790	Sh.		(2.65)		5.26	16.0	3.40	4.87	4.45	12.8	2.87	2.30
21	15780-15830	S1.		(2.65)		7.97	19.9	2.78		4.45	15.0	3.56	2.33
20	15790-15800	S1.		(2.65)		6.39	26.5	2.87		4.00	16.5	4.13	2.39
43	15790-15800	Sh.		(2.65)		(3.97)	14.1	3.46	4.87	4.59	13.2	2.87	2.37
44	15820-15830	Sh.		(2.65)		5.44	13.9	3.43		5.18	10.7	2.06	2.35
?	15830-15840	Sh.		(2.65)		5.50							
22	15830-15920	S1.		(2.65)		6.58	18.1	2.71		4.63	13.7	2.97	2.34
45	15840-15850	Sh.		(2.65)		(4.06)	15.0	3.29		4.13	15.5	3.74	2.39
10	15840-15930	Ss.		(2.65)		12.51	40.2	0.55	0.94	1.44	7.45	5.16	0.92
46	15910-15920	Sh.		(2.65)		(6.26)	17.6	3.34		3.79	14.3	3.77	2.23
23	15920-15930	S1.		(2.65)		8.02	17.2	2.90		4.06	14.0	3.44	2.24
47	15920-15930	Sh.		(2.65)		4.71	14.6	3.24		3.96	16.5	4.16	2.41

Table 4. (cont.)

LIST OF COST B-3 SAMPLE MEASUREMENTS

Samp. No.	Depth Interval (feet)	Lithology	Sample Porosity	Sample Density	K _{MEASURED} (TCU)	K matrix (TCU)	Quartz (%)	Potassium (%)	⁴⁰ K γ -ray (ppm)	U (ppm)	Th (ppm)	Th/U	A (μkm^{-3})
19	3810-3870	Sl.	.74	(3.34)2.5	2.13	6.56	15.3	1.28		3.71	8.17	2.20	1.52
20	4020-4050	Sl.	.73	(4.46)2.5	2.14	6.32	36.6	1.35		3.07	9.80	3.19	1.47
21	4560-4590	Sl.	.66	2.40	2.69	4.30	18.9	1.99		9.99	6.8	---	2.87
22	5010-5040	Sl.	.72	(3.03)2.5	2.13	5.65	18.2	1.59		5.71	4.6	---	1.79
23	5490-5520	*LS.	.64	2.70	2.14	4.22	2.5	0.41		1.78	2.9	---	0.70
24	6020-6030	*LS.	.61	2.41	2.04	3.52	5.1	0.65		1.06	3.2	3.03	0.49
1	6300-6650	*LS.	.68	2.32	2.33	6.25	5.3	0.92		1.50	4.4	2.91	0.67
2	7000-7350	Sl.	.72	2.45	2.15	5.94	10.1	1.36		3.16	7.43	2.35	1.31
3	7400-7760	Sl.	.72	2.47	2.22	6.57	10.9	1.50		3.25	9.29	2.85	1.48
5	7400-7760	*LS.	.64	2.55	2.43	6.20	3.8	0.70		1.37	4.7	3.45	0.70
6	7700-8030	*LS.	.74	(3.35)2.6	2.10	5.85	3.8	0.74		1.53	4.6	2.99	0.75
11	7810-8030	Sl.	.80	2.86	2.11	9.13	12.7	1.67		3.18	12.7	4.01	1.96
8	7810-9530	Sl.	.75	2.64	2.21	7.83	11.8	1.61		3.06	12.3	4.02	1.75
12	9000-9950	*LS.	.48	2.91	3.70	8.96	38.1	0.72	0.88	2.11	4.1	1.94	0.96
7	9000-9950	*LS.	.68	2.32	2.33	6.37	5.8	0.96		1.46	5.73	3.91	0.74
15	9000-10330	Sh.	.52	2.80	2.65	5.09	9.7	1.92		3.68	15.5	4.20	2.28
9	9700-9720	Sl.	.67	2.83	2.31	5.99	12.1	1.55		3.11	9.07	2.91	1.65
10	9800-10330	Sl.	.63	2.68	2.64	7.24	10.6	1.64		3.16	12.9	4.08	1.84
13	9800-10480	Sh.	.42	2.98	4.25	9.20	11.9	0.45	0.42	1.34	3.1	2.28	0.66
16	10000-10480	Sh.	.51	2.83	2.81	4.74	14.2	2.00		4.03	15.3	3.80	2.39
11	10400-11200	Sl.	.61	2.77	2.80	7.99	14.5	1.86		3.19	14.1	4.41	2.02
17	10400-10950	Sh.	.53	2.81	2.60	4.96	9.0	2.07		3.66	15.0	4.10	2.26
14	10600-11200	Sh.	.47	(3.27)2.7	3.79	9.03	12.9	0.53	0.90	1.81	4.0	2.23	0.79
18	10600-11200	Sh.	.54	(3.05)2.7	2.69	5.57	8.1	2.09		3.59	15.5	4.33	2.19
2	11210-11400	Sh.	.37	2.76	4.50	8.72	39.9	1.19		3.74	14.1	3.78	2.09
1	11210-11400	Sl.	.48	2.30	3.00	5.83	12.6	2.05	3.17	3.72	11.7	3.14	1.67
10	11650-11710	Sh.	.43	2.83	4.74	11.54	48.1	1.03		2.61	7.28	2.79	1.33
9	11650-11710	Sl.	.48	2.70	3.02	5.91	10.7	2.30		5.15	11.5	2.23	2.33
13	12090-12310	Sh.	.46	(3.20)2.7	4.90	13.72	18.4	1.09		2.05	6.64	3.24	1.09
14	12090-12310	Sl.	.46	2.55	3.21	6.28	12.9	2.13		3.88	14.9	3.84	2.10
16	12210-12400	Sl.	.48	2.75	2.42	5.98	9.6	2.40		4.96	17.5	3.52	2.76
6	12490-12580	*LS.	.38	2.53	4.14	7.84	11.8	0.37		1.53	4.6	2.99	0.70
5	12490-12580	Sl.	.50	2.56	2.74	5.14	10.0	1.67		1.37	4.7	3.45	0.79
8	12690-12900	Sl.	.59	(3.63)2.7	3.19	(9.82)	10.4	1.42		3.06	12.3	4.02	1.77
7	12690-12900	LS.	.61	2.67	2.08	3.61	9.8	1.69		1.46	5.73	3.91	0.92
11	13440-13560	Sh.	.46	(3.07)2.7	4.52	11.86	34.7	0.79		3.19	14.1	4.41	1.87
12	13440-13560	Sl.	.56	(3.07)2.7	3.02	7.63	9.4	2.02		2.11	4.1	1.94	1.01
3	14020-14310	*LS.	.51	2.97	2.06	2.94	2.3	0.06		3.25	9.29	2.85	1.63
4	14020-14310	Sh.	.46	2.36	2.75	4.71	6.3	2.05		3.18	12.7	4.01	1.65
5	14550-14630	Sl.	.53	(3.05)2.7	3.12	7.30	11.2	1.60		3.68	15.5	4.20	2.17

* Lithology designated following determination of quartz and potassium contents.

() = approximate or assumed value, with preferred estimated value determined from nearby samples, where available, presented following parentheses.

Robertson and Peck, 1974; Drury and Jessop, 1983), the Woodside-Messmer (1961a) weighted geometric mean model (3) was chosen because of its good fit with experimental data in the low porosity range (0-0.5), as shown by Drury and Jessop (1983), and Sekiguchi (1984):

$$K_m = K_f^{(\phi)} \cdot K_s^{(1-\phi)} \quad (3)$$

K_m = measured thermal conductivity of the aggregate

$K_f = K_w$ = fluid (water) thermal conductivity

K_s = solid fraction thermal conductivity

The procedures to calculate the solid conductivity K_s from the measured sample conductivity K_m and finally the in situ aggregate mean thermal conductivity K is shown in the block diagram of Fig. 5. The calculated values of K_s at room temperature assuming $K_w = 1.45$ T.C.U. (Clark, 1966) are reported in Table 4. The mean thermal conductivity of each lithology (K_s) at room temperature is presented in Table 5.

Table 5. Summary of thermal conductivities of various lithologies at zero porosity (K_s).

<u>Well</u>	<u>Sandstone</u>	<u>Siltstone</u>	<u>Shale</u>	<u>Limestone</u>
COST B-2	10.27 $\pm$.22 (4.30 $\pm$.09) n=23	7.03 $\pm$.22 (2.94 $\pm$.09) n=16	6.27 $\pm$.26 (2.62 $\pm$.11), n=7 (upper interval) 5.26 $\pm$.09 (2.20 $\pm$.04), n=22 (lower interval)	6.57 $\pm$.45 (2.75 $\pm$.19) n=8
COST B-3	10.43 $\pm$.73 (4.36 $\pm$.31) n=7	6.36 $\pm$.24 (2.66 $\pm$.10) n=17	5.01 $\pm$.16 (2.10 $\pm$.07) n=5	5.53 $\pm$.56 (2.31 $\pm$.23) n=9

Values given are mean + standard error, with n measurements for each lithology. Units in TCU, ($Wm^{-1} K^{-1}$).

Measurements at room temperature (25 $\pm$ 2 °C).

4.4 QUARTZ CONTENT EVALUATION

The weight percent of quartz in the samples has been determined by the x-ray powder diffraction technique. A description of the analytical methods used in defining the mineralogy of the DSDP cores is reported in Bader et al. (1970). An important part of determining accurate quartz content values is having a good standard; our technique for standard preparation and evaluation of quartz content in the samples is given in Appendix A. In general, the results we obtained show a reasonable estimate of the quartz content with an accuracy of about $\pm 5\%$.

Fig. 5. Block diagram summarizing the complete procedure used to determine in situ thermal conductivity from measurements on pulverized samples and estimates of in situ porosity, anisotropy and volume fraction.

The quartz content data are reported in Table 4 and the plots of sample quartz content vs. depth in COST B-2 and COST B-3 are shown in Fig. 6. The quartz content versus the solid thermal conductivity (Fig. 7) shows a strong positive correlation which can be represented by:

$$K_s = 4.53 + .151 \text{ Qtz}\% \quad (4)$$

where K_s = thermal conductivity of the solid fraction.

The scatter around the best fit straight line is large; however, errors in K_m and porosity (Φ) are greatly amplified by the Woodside-Messmer (1961) relation used in calculating K_s . The correlation coefficient is only 0.71, emphasizing that (eqn. 4) should not be used for purposes of accurate calculations.

The different lithologies considered in our study are reasonably well separated on such a plot:

- Limestones, clays and shales primarily occupy the region from 0 to 15% quartz content;
 - samples with quartz contents higher than 25% are mainly sandstones;
 - siltstones fill the intermediate range between 10 and 30% quartz content.
- As shown by Langseth (1984) for Arctic Ocean sediments, the high conductivity minerals, such as quartz, calcite and dolomite, are concentrated in the coarser fraction ($>62\mu$); K_m is positively correlated to the percentage of coarse fraction.

The lower K_m associated with fine grain materials, e.g., shales, clays and limestones, might be affected more than the others by larger errors related to the incomplete removal of air bubbles during the saturation process. However, the K_s intercept at 0% quartz content indicates a thermal conductivity of the matrix of 4.5 TCU which is close to the thermal conductivity of several common minerals: K-feldspar, plagioclase, mica (Horai, 1971), clay (Morin and Silva, 1984), common shale (Sekiguchi, 1984) and/or amorphous material (Fry, 1982). If we extrapolate the best fit linear relationship to 100% quartz, a value of 19.6 TCU is obtained. Horai (1971) measured 18.4 TCU ($7.7 \text{ Wm}^{-1}\text{K}^{-1}$) on pulverized quartz. Sekiguchi (1984) studied the relationship of conductivity of pure quartzose sandstone as a function of porosity applying different models; at zero porosity the thermal conductivity turns out to be 19.0 TCU. Thus it is reasonably clear that the conductivity of marine sediments and sedimentary rocks is strongly influenced by the abundance of quartz, as would be expected. Calcite and dolomite also have a high intrinsic contrast in thermal conductivity (8.5 and 13 T.C.U., respectively; Clark, 1966). Thus their abundances should also be correlated with thermal conductivity, but we did not analyze these possible relationships.

The relationship of K_s to quartz content can be incorporated in the simple geometric mean model (Woodside and Messmer, 1961a) to check the agreement between measured conductivities and calculated values using:

$$K_{calc.} = K_w^{(\Phi)} \cdot (4.53 + .151 \text{ Qtz } \%)^{(1-\Phi)} \quad (5)$$

where K_w = water conductivity

The utility of this model is limited since the derivation of equation (5) is in part circular. However this analysis is still useful because the mean

Fig. 6. Plot of quartz content (%) in the analyzed samples, paleobathymetry, and sediment accumulation rate versus depth in the COST B-2 (solid curve) and B-3 (dash-dot curve) wells (after Poag, 1980, Fig. 29). The paleobathymetry and sedimentation rate depth scales for the COST B-3 have been shifted to correspond to the geological time scale of the COST B-2. Paleobathymetry scale symbols are: U = upper slope, O = outer shelf, M = middle shelf, I = inner shelf, N = non-marine, in order of decreasing depth.

Fig. 7. Correlation of the quartz content with the thermal conductivity of the solid fraction K_s of the samples. Solid line is the best fitting linear relationship using all data, with the equation as shown and correlation coefficient of 0.71. Note that the various lithologies are generally separated on the plot.

conductivities of randomly oriented quartz and free quartz matrix are known independently from the model. The plot of Fig. 8 illustrates the results of the comparison: 34 samples from COST B-2 and 40 from COST B-3 are shown with different symbols. Most of the calculated values are within ± 0.3 T.C.U. ($\pm 0.13 \text{ Wm}^{-1} \text{ K}^{-1}$) of the measured ones. Thirteen samples have measured conductivities higher than calculated ones (11 sandstones, 1 limestone and 1 siltstone); these are probably caused by some other high conductivity minerals (perhaps calcite and/or dolomite) in the samples in addition to quartz, which increase the average thermal conductivity of the aggregate. Such minerals are more likely included in coarse sandstones or in limestones.

On the other hand there are 3 limestones, 1 sandstone and 1 siltstone with measured conductivities lower than calculated ones; this could be explained by the presence of significant amounts of low conductivity constituents such as amorphous material and non-crystalline carbonates. Errors, either in the measured or in the calculated conductivities (as a function of the quartz content) must also be considered. The general agreement of measured and calculated values appears to support the reliability of the approach.

4.5 LITHOLOGY DISTRIBUTION AND IN SITU POROSITY ESTIMATE

To obtain the mean in situ thermal conductivity over depth intervals, the volumetric fraction and the mean conductivity must be estimated for each lithology present. The mean conductivities for the various lithologies measured in our samples is given in Table 5. The lithology assignment for depth intervals of the COST B-2 well is based on Rhodehamel (1977) and on a reevaluation of the lithologic log to distinguish between clays/shales and siltstones (Fig. 9). For the COST B-3 well it is based on a subjective interpretation of the lithologic log (Table 6) by assigning for each 25' depth interval a certain proportion of each lithology.

Determination of the depth distribution of porosity is a major aim in explorative hydrocarbon drilling; it is measured directly on cores or estimated from well geophysical logs. The COST B-2 porosity data set is summarized over 1000 ft. depth intervals in Table 7. Sandstone porosities have been derived from density (FDC) and sonic (BHC) logs (Rhodehamel, 1977), measured on side wall cores (Smith et al., 1976; Core Laboratories, Inc., 1976), or evaluated from combinations of various well logs (Coriband, unpublished software by Schlumberger, Inc.). Shale porosities are estimated by the Coriband model or liquid pycnometry on chips. Siltstone and limestone porosities are derived from the Coriband model only. A few of these porosity distributions are shown in Fig. 10a-c.

Porosity data from the COST B-3 well were evaluated from Wu and Nichols (1980) LOGCALC models, utilizing the stratigraphic sequence of the lithologic log. The two well-log-analysis models of Wu and Nichols (1980) which we used are:

- 1) a sand-shale log-analysis model from 8200' to 12400' depth.
- 2) a carbonate-sandstone model from 12688' to 15810' depth.

Outside these depth intervals they were estimated from the COST B-2 well porosity distribution and from core analyses (Core Laboratories, Inc., 1979). This causes the shale and siltstone porosities for the deeper intervals of the COST B-3 well (Table 9) to be relatively poorly constrained.

Fig. 8. Plot of the K_m measured on the samples vs. K_m calculated from equation (5). Line gives relationship for ratio of 1 between axis variables.

Fig. 9. Principal lithologies vs. depth in the COST B-2 well averaged over 1000 ft (304.8m) intervals (from Rhodehamel, 1977, Fig. 7).

Fig. 10. Summary of the porosity distribution vs. depth for different lithologies in the COST B-2 well, averaged over 1000 ft (304.8m) intervals.

a) Sandstone, b) Siltstones and limestones, c) Shales.

Table 6. COST B-3 Lithology in 1000-ft. (304.8 m) depth intervals

<u>Depth Interval</u>	<u>Sand and Sandstone</u>	<u>Claystone Mudstone Shale</u>	<u>Siltstone</u>	<u>Limestone Dolomite</u>	<u>Total</u>
0-1000					
1000-2000	(Seafloor at 2728 ft.; No samples above 4000 ft.)				
2000-3000					
3000-4000					
4000-5000	12.0	80.0	0	8.0	100.0
5000-6000	6.0	67.5	7.5	19.0	100.0
6000-7000	13.5	48.0	26.0	12.5	100.0
7000-8000	12.0	53.0	17.0	18.0	100.0
8000-9000	35.5	38.0	9.5	17.0	100.0
9000-10000	11.0	57.0	12.0	19.5	99.5
10000-11000	40.0	30.0	1.0	29.0	100.0
11000-12000	43.0	39.5	12.0	5.5	100.0
12000-13000	12.5	47.5	6.0	34.0	100.0
13000-14000	10.0	36.5	3.0	50.5	100.0
14000-15000	21.0	35.9	1.5	39.0	97.4
15000-15810	34.2	37.9	8.7	16.1	96.9

Values (in volume percent) estimated from drill cuttings sample log (B.M. Pollack, in Scholle, 1980, pp. 20-23). For depth intervals where total is less than 100.0%, the remaining constituent lithology is primarily coal.

Table 7. Mean porosities of COST B-2 lithologies at 1000 ft. (304.8 m) depth intervals.

Depth Interval (ft)	Sandstone			Siltstone Calculated ³	Limestone Calculated ³	Shale Calculated ³	Shale (liquid pycnometry) ^{2,4}
	Logs ¹	Sidewall Cores ²	Calculated				
0-1100	(42)						
1100-2100	53.6+9.2	28.8+1.3					40.0+1.4
2100-3100	48.0+3.9	32.6+2.0					31.5+2.8
3100-4100	52.0+4.8	27.6+3.2					24.4+4.6
4100-5100	33.0+1.4	29.3+4.3					23.2+0.5
5100-6100	32.4+6.2	30.7+2.8					21.9+2.0
6100-7100	26.5+7.8	27.9+4.0	15.9+2.2	(10.3+0.9)	9.2+0.6	8.1+0.8	14.3+2.9
7100-8100	23.0+8.5	24.8+0.8	(18.5+1.4)	9.6+1.3		(9.2+0.6)	13.1+4.8
8100-9100	16.7+9.2	27.4+2.4	19.8+2.4	10.9+2.0	7.8+1.3	9.4+1.4	9.7+1.3
9100-10100	24+1.4	26.0+3.1	16.8+1.9	(14.2+1.9)	8.7+2.1	8.5+1.7	5.0+1.2
10100-11100	17.2+2.8	22.3+2.3	19.4+1.4	(9.4+5.1)	5.7+2.1	6.8+0.7	4.9+1.0
11100-12100	11+2.5	20.1+3.1	16.3+2.1	(7.0+3.1)	3.6+0.9	5.4+1.6	3.8+0.3
12000-13100	10+2.8	17.3+3.2	13.5+1.0		1.3+0.9	4.1+1.9	4.1+0.4
13100-14100	10.6+1.5	19.5+3.0	12.7+1.7	4.8+1.6	3.9+2.6	3.4+1.6	4.0+0.4
14100-15100	8+1.7	14.2+4.2	10.2+0.5	4.8+1.2	0.8+0.4	0.8+0.2	4.7+0.5
15100-16000	6.4+1.0	14.2+2.7	9.1+1.3	2.7+1.0		0.4+0.4	

All values in percent of total volume of specific lithology.

¹ FDC and BHC logs, from Rhodehamel (1977, Table 6)

² Smith et al. (1976)

³ Calculated from logs by CORIBAND^R, weighted by layer thickness

⁴ Unpublished data on picked drill cuttings by AMOCO

4.6 MEAN IN SITU THERMAL CONDUCTIVITY AND ANISOTROPY CORRECTION

According to our procedures (Figure 5) the mean thermal conductivity of the lithologies measured at laboratory conditions should be corrected for the in-situ pressures and temperatures before calculating the true K (with the formation porosities) by a second application of the Woodside and Messmer (1961a) geometric mean model (eqn. 3). The corrections for these in-situ conditions will be different for the solid matrix and fluid components of the lithologic types. The thermal conductivity of water increases directly with both pressure and temperature (Clark, 1966). Hydrostatic pressure and measured well temperature data (see below) were considered in correcting the pore water thermal conductivity as a function of depth on COST B-2 and COST B-3 wells (Tables 8, 9).

The thermal conductivities of the solid phases generally increase with pressure (Kappelmeyer and Haenel, 1974; Lubimova et al., 1979); the increase in conductivity is around a few percent per 100 MPa (1000 kg/cm²) for hard rocks or minerals, and may reach 10% or more for softer rocks such as claystone, siltstone and marlstone (Sekiguchi, 1984). This is related to the variation with pressure of the pore space more than variation in the conductivity of the grains, so that the effect of pressure for soft rocks is probably a function of the porosity.

On the other hand, the effect of temperature on thermal conductivity of rocks is significant and generally inversely correlated (Birch and Clark, 1940; Clark, 1966; Winkler, 1952; Schatz and Simmons, 1972; Kappelmeyer and Haenel, 1974; Lubimova et al., 1979; Sekiguchi, 1984). On the basis of these studies, the following correction factors were chosen for the different lithologies:

Lithologies Temp. coefficient

sandstone	-1%/5°C
siltstone	-1%/10°C
limestone	-1%/10°C
shale	-1%/20°C

The published data on shales are somewhat ambiguous. For example: Birch and Clark (1940) showed a slight decrease of conductivity with increasing temperature, whereas Winkler (1952) published an opposite trend. Sekiguchi (1984) suggests that the conductivity of poor conductors (4-5 T.C.U.), such as shale with less than 25% porosity, and basalt, is nearly independent of temperature. These different results for shales are probably related to the porosity of the samples; at high porosities the pore fluid becomes relatively more important and the conductivity might increase with temperature (Winkler, 1952).

The pressure and temperature correction of K_w and K_s (for each lithology) were calculated for the mid-depth of the interval for which the porosity and volume fraction were available. Application of the Woodside-Messmer relation (eqn. 3) to the in-situ values of K_w , K_s , and Φ gives the mean conductivity K .

At this point we have reconstructed the mean random thermal conductivity (K) of each lithology at depth without considering any anisotropy.

TABLE 8: COST B-2 THERMAL CONDUCTIVITY VS. DEPTH

Depth Interval [ft.]	Mean T [°C]	K _w	SANDSTONE K _s	φ	K _{pr}	SILTSTONE K _s	φ	K _{pr}	K _s	SHALE φ	K _{pr}	K _s	LIMESTONE φ	K _{pr}	SANDSTONE Vol.	K _s	SILTSTONE Vol.	K _s	SHALE Vol.	K _s	LIMESTONE Vol.	K _s	MEAN INTERVAL K _s
1.08-1.1	9	1.36	10.60	(55)	3.80	-	-	-	6.32	(60)	2.51	-	-	-	.986	3.80	0	-	.014	2.51	0	-	3.77
1.1-2.1	17	1.40	10.43	53.6	3.55	-	-	-	6.30	(50)	2.97	-	-	-	.775	3.55	0	-	.245	2.97	0	-	3.32
2.1-3.1	25	1.44	10.27*	48	4.00	7.03*	(30)	4.37	6.27*	40.0	3.48	6.57*	-	-	.425	4.00	.050	4.37	.525	3.48	0	-	3.72
3.1-4.1	34	1.48	10.09	52	3.72	6.97	(25)	4.73	6.24	31.5	3.97	6.51	(15)	5.21	.355	3.50	.179	3.97	.436	3.01	.030	5.4	3.37
4.1-5.1	43	1.51	9.90	33	5.32	6.90	(20)	5.09	6.21	24.4	4.40	6.45	(12)	5.42	.100	5.00	0	4.27	.885	3.33	.015	5.4	3.47
5.1-6.1	51	1.53	9.74	32.4	5.35	6.85	(15)	5.47	6.19	23.2	4.48	6.40	(10)	5.55	.515	5.03	.219	4.59	.246	3.48	.020	5.3	4.45
6.1-7.1	59	1.55	9.57	26.5	5.91	6.79	10.3	5.83	6.16	21.9	4.55	6.35	9.2	5.58	.655	5.56	.053	4.90	.280	3.45	0.12	5.3	4.71
7.1-8.1	66	1.57	9.43	23	6.24	6.74	9.6	5.86	6.14	14.3	5.05	6.30	(8.5)	5.60	.203	5.87	.436	4.92	.306	3.83	0.55	5.3	4.68
8.1-9.1	74	1.59	9.26	16.7	6.90	6.69	10.9	5.72	6.12	13.1	5.13	6.25	7.8	5.62	.286	6.49	.197	4.80	.419	3.89	.098	5.3	4.73
9.1-10.1	82	1.61	9.10	24	6.00	6.63	14.2	5.42	6.09	9.7	5.35	6.20	8.7	5.51	.503	5.64	.013	4.55	.309	4.06	.175	5.3	4.96
10.1-11.1	90	1.62	8.93	17.2	6.66	6.57	9.4	5.76	6.07	5.0	5.68	6.14	5.7	5.69	.415	6.29	.006	4.84	.519	4.31	.060	5.0	5.02
11.1-12.1	98	1.64	8.77	11	7.29	6.52	7.0	5.92	5.07	4.9	4.79	6.09	3.6	5.81	.245	6.85	.042	4.97	.353	3.63	.360	5.0	4.72
12.1-13.1	106	1.66	8.61	10	7.30	6.46	(6)	5.95	5.05	3.8	4.84	6.04	1.3	5.94	.082	6.86	0	5.00	.700	3.67	.218	5.0	4.07
13.1-14.1	114	1.67	8.44	10.6	7.11	6.40	4.8	6.00	5.03	4.1	4.81	5.99	3.9	5.70	.418	6.68	.110	5.03	.385	3.65	.087	5.0	4.82
14.1-15.1	123	1.70	8.26	8	7.28	6.34	4.8	5.95	5.00	4.0	4.79	5.93	0.8	5.87	.335	6.84	.346	5.00	.389	3.63	.030	4.0	4.61
15.1-16.043	128	1.72	8.15	6.4	7.38	6.31	2.7	6.09	4.99	4.7	4.75	5.89	(1.0)	5.82	.240	6.93	.340	5.11	.400	3.60	.020	4.0	4.62

T = equilibrium temperature from logs
 K_w = thermal conductivity of water (Clark, 1966), TCU
 K_s = thermal conductivity of solid component (TCU), corrected for temperature (see text)
 φ = porosity (from Table 6)
 K_{pr} = in-situ thermal conductivity, calculated from Woodside-Mesmer eqn., TCU
 K_z = vertical component of thermal conductivity, TCU

TABLE 9: COST B-3 THERMAL CONDUCTIVITY VS. DEPTH

Depth Interval [ft.]	Mean T [°C]	K _w	SANDSTONE K _s	φ	K _{pr}	SILTSTONE K _s	φ	K _{pr}	K _s	SHALE φ	K _{pr}	K _s	LIMESTONE φ	K _{pr}	SANDSTONE Vol.	K _s	SILTSTONE Vol.	K _s	SHALE Vol.	K _s	LIMESTONE Vol.	K _s	MEAN INTERVAL K _s
2.728-3.0	(6)	1.34	10.53	(55)	3.39	-	-	-	5.05	(55)	2.43	-	-	-	(.200)	3.39	0	-	(.800)	2.43	0	-	2.58
3.0-4.0	13	1.38	10.38	(50)	3.78	-	-	-	5.04	(45)	2.81	-	-	-	(.200)	3.78	0	-	(.800)	2.81	0	-	2.96
4.0-5.0	23	1.43	*10.18	(40)	4.64	*6.36	(30)	4.06	*5.01	(35)	3.23	5.53*	(15)	4.51	.120	4.36	0	3.41	.800	2.71	.080	3.7	2.91
5.0-6.0	34	1.48	9.96	(30)	5.62	6.29	(25)	4.38	4.98	(25)	3.68	5.47	(12)	4.68	.060	5.28	.075	3.68	.675	2.79	.190	3.9	3.11
6.0-7.0	42	1.51	9.79	(25)	6.14	6.24	(20)	4.70	4.96	(15)	4.15	5.42	(9)	4.83	.135	5.77	.260	3.95	.480	3.15	.125	4.0	3.67
7.0-8.0	50	1.53	9.63	(18.0)	6.92	6.19	(15)	5.02	4.94	(10)	4.39	5.38	(7)	4.93	.120	6.51	.170	4.21	.530	3.33	.180	4.1	3.82
8.0-9.0	58	1.55	9.47	12.3	7.58	6.14	(10)	5.35	4.92	5.1	4.64	5.34	(6.0)	4.96	.355	7.13	.095	4.49	.380	3.52	.170	4.1	4.61
9.0-10.0	66	1.57	9.30	14.9	7.13	6.09	(10)	5.32	4.90	(5)	4.63	5.29	3.6	5.06	.110	6.70	.120	4.47	.570	3.51	.195	4.2	3.98
10.0-11.0	73	1.59	9.16	13.2	7.27	6.04	(7)	5.50	4.88	(5)	4.61	5.25	3.9	5.01	.400	6.83	.010	4.62	.300	3.50	.290	4.1	4.64
11.0-12.0	81	1.61	9.00	13.1	7.18	5.99	(7)	5.46	4.86	(4)	4.65	5.21	(6.0)	4.86	.430	6.75	.120	4.58	.395	3.53	.055	4.1	4.64
12.0-13.0	89	1.62	8.84	10.6	7.38	5.94	(5)	5.57	4.84	(4)	4.63	5.17	8.3	4.69	.125	6.94	.060	4.68	.475	3.51	.340	3.9	3.96
13.0-14.0	98	1.64	8.65	7.7	7.61	5.88	(5)	5.52	4.82	(4)	4.62	5.12	7.3	4.71	.100	7.15	.030	4.63	.365	3.50	.505	3.9	3.96
14.0-15.0	106	1.66	8.49	7.7	7.49	5.83	(4)	5.54	4.80	(3)	4.65	5.07	4.1	4.84	.210	7.04	.015	4.65	.359	3.53	.390	4.0	4.33
15.0-15.81	114	1.67	8.33	7.6	7.37	5.78	(4)	5.50	4.78	(3)	4.63	5.03	5.4	4.74	.342	6.93	.087	4.62	.379	3.51	.161	3.9	4.62

T = equilibrium temperature from logs
 K_w = thermal conductivity of water (Clark, 1966), TCU
 K_s = thermal conductivity of solid component (TCU), corrected for temperature (see text)
 φ = porosity, estimated value in parentheses
 K_{pr} = in-situ thermal conductivity, calculated from Woodside-Mesmer eqn., TCU
 K_z = vertical component of thermal conductivity, TCU
 Vol. = volume fraction of respective lithologies

Sedimentary formations, such as those drilled at COST wells, show well defined bedding planes, layering, lamination, or foliation, particularly for shales and limestones. In many cases, the thermal conductivity is significantly less perpendicular to the bedding (vertical direction), than parallel to the bedding in the horizontal (x-y) plane (Birch and Clark, 1940; Simmons, 1961; King and Simmons, 1972; Kappelmeyer and Haenel, 1974). The anisotropy effect may influence the final computation of heat flow by 30% or more (King and Simmons, 1972).

However, in pulverizing the samples any anisotropy due to bedding or texture was destroyed. Assuming a horizontal geometry of stratification (not exactly true in the case of COST B-3 well drilled on the slope) it is possible to calculate the vertical component of the thermal conductivity if the anisotropy factor $A(z)$ is known, where

$$A = \frac{K_{x,y}}{K_z} \quad (6)$$

and $K_{x,y}$ is the conductivity parallel to bedding and K_z is perpendicular to it. The conductivity in an arbitrary direction K_ϕ is calculated (Kappelmeyer and Haenel, 1974, p. 57, eqn. 3.5) as

$$K_\phi = K_z \cos^2 \phi + K_{x,y} \sin^2 \phi \quad (7)$$

where ϕ is the angle between the direction and the perpendicular to bedding. Since the conductivity $\bar{K}$ measured on pulverized rock is the result of random orientations of anisotropic material, we have

$$\bar{K} = \frac{\int_0^{\pi/2} K_\phi d\phi}{\int_0^{\pi/2} d\phi} = \frac{K_z + K_{x,y}}{2} = \frac{K_z(1+A)}{2} \quad (8)$$

The anisotropy correction was applied below 3000 ft. K.B. (857 m below sea floor) for the COST B-2 and below 5000 ft. K.B. (693m below sea floor) for the COST B-3, considering that well-defined bedding and layering was significant at lithification depth (from lithologic logs) and below. Anisotropy is probably related to ambient pressure and temperature, as well as to diagenetic and metamorphic processes which occur over time, but with no further information we assumed a constant factor with depth.

Reasonable values of anisotropy for each lithology have been assigned from the literature (previously cited papers), i.e., $A = 1.1$ for sandstone, 1.3 for limestone, and 1.5 for shale. We have calculated the average conductivity $\bar{K}_z$ in each depth interval bounded with in-situ temperature measurements by considering the volume-averaged fraction of all the contributing lithologies:

$$\bar{K}_z = Z / [\Delta Z_1 V_1 / K_{z1} + \dots + \Delta Z_n V_n / K_{zn}] \quad (9)$$

with: $V_1, \dots, V_n$ = volume fraction of each lithology in the interval ΔZ_n
 $\Delta Z_1, \dots, \Delta Z_n$ = thicknesses of units 1 through n,
 $K_{z1}, \dots, K_{zn}$ = in situ K_z for each lithology.

4.7 ESTIMATION OF THERMAL CONDUCTIVITY FROM WELL LOGS

We tried to find a general and reliable relationship between thermal conductivity and some of the rock physical properties to estimate the thermal characteristics of sedimentary rocks from standard geophysical well logs. Several previous studies of this type have attempted to address the problem; however, the relations obtained are applicable only for particular lithologies, mainly sandstones (Zierfuss and Van der Vliet, 1956, Tikhomirov, 1968, Anand et al., 1973; Goss et al., 1975) or siliciclastic and carbonate rocks (Houbolt and Wells, 1980). An empirical linear relationship between the ratio of sound velocity (V_p) to thermal conductivity (K) and formation temperature (T_z) is deduced in the latter study. The relation obtained from 10 North Sea wells has the form:

$$K_z = \frac{V_p \cdot 77}{1.039 (80.31 + T_z)} \quad (10)$$

where K_z = vertical component of in situ thermal conductivity, in $\text{Wm}^{-1}\text{K}^{-1}$.

A comparison of the average interval K_z with the model of Houbolt and Wells (1980) for the COST B-2 well shows a reasonable agreement (Fig. 11). The values of the compressional velocity V_p for each lithology and their variation with depth were estimated from the sonic logs.

4.8 EQUILIBRIUM TEMPERATURES AND THERMAL GRADIENTS

Multiple logging runs were made to several depths during drilling of the COST B-2 and COST B-3 wells, and temperature measurements were recorded on most of these logs. In the COST B-2 a High Resolution Temperature log (HRT) was run as well.

Therefore, near-bottom hole temperatures (BHT) measured several times after cessation of drilling and circulation are available. True formation temperatures can be estimated with good accuracy (within 1°C) if the total perturbing time (t_1) (time elapsed during drilling of last 30 feet of well and subsequent circulation), and the time elapsed between end of circulation and the temperature measurement (t_2) are known (Dowdle and Cobb, 1975; Fertl and Wichmann, 1977; Middleton, 1979):

$$\text{BHT} = T - C \ln (1 + t_1/t_2) \quad (11)$$

where T is the true temperature and C is a constant determined by least squares fit of equation (11). It is assumed that the disturbance due to circulation is of constant magnitude over its duration, and eqn. (11) is a good approximation when the in-situ temperatures are measured over at least several thermal time constants of the hole (a^2/κ , where a is the hole radius and κ is thermal diffusivity of surrounding material) after cessation of circulation (Bullard, 1947). The information contained in the well logs and daily drilling reports is not always sufficient to accurately determine t_1 and t_2 , particularly when the logging operations are interrupted by mud conditioning intervals, fishing attempts, reaming, etc.

Fig. 11. Mean acoustic velocities (V_p , determined from logs, left) and vertical component of thermal conductivity (K_z , right) over 1000 ft. (304.8m) intervals of COST B-2 well. Vertical bars are calculated conductivity values (Table 8), and circles are those predicted from acoustic velocities following Houbolt and Wells (1980).

In the COST B-2 well, 4 BHTs were measured at various depths in the hole (Robbins, 1977) (Table 10, Fig. 12). An accurate analysis of all the information available gave the same results as those obtained by Von Herzen and Helwig (1984), at least for the first three logging depths (Table 10). The analysis of the deepest BHT is problematical because of the many trips, conditioning intervals, reaming, etc., which indicate that the calculated value of 127.9°C is near the lower bound of the possible range (127°-143°C). The HRT value at the same depth was 284°F (140°C), which suggests a higher equilibrium temperature than that deduced from the BHTs, although the HRT calibration was not confirmed and is therefore uncertain itself.

The mean sea floor temperature (at 118.3 meters depth below K.B.) is not well constrained because of the annual and possible longer period temperature variations. The value of 6°C was determined from a mean of nearby hydrographic data (unpublished) over the past several decades.

In the COST B-3 well temperatures were measured only at 3 depths which were useful for estimating equilibrium in-situ temperatures. At 5731 ft. depth the temperature recorded on various logs (Dual Induction SFL, FDC-CNL, FDC, LSS, SRS, HRD) was uniform at 97°F (36.1°C). At the next measured depth (12655 ft.) increasing temperatures were measured on successive log runs. A formation temperature of 195°F (90.6°C) was calculated from 4 points, with a correlation coefficient of 0.996 (Fig. 13). The calculation of the deepest undisturbed temperature was less accurate because of less certain knowledge of t_1 and t_2 ; an average value of 241°F (116.1°C) was found. Like the COST B-2, the sea floor temperature of COST B-3 is not accurately known. A value of ~40°F (~4.4°C) was estimated at the sea floor (818.7m below M.S.L.) based on existing (unpublished) hydrographic data. Formation temperatures and interval gradients are summarized in Table 11.

5. CALCULATION OF HEAT FLUX

The calculated interval heat flow at the COST B-2 and COST B-3 wells is reported in Table 12 together with interval K_z and gradient. A summary of the calculated equilibrium temperatures, interval gradients, mean interval K_z and interval heat fluxes is presented in Fig. 14 for the COST B-2 well and in Fig. 15 for the COST B-3 well. The broken lines represent estimated values, or values for which the computations only provide an estimate of the relevant parameters, such as formation porosity, volume fraction of a certain lithology, sea floor mean temperature. Therefore the values in the shallower intervals are generally affected by larger errors caused by lack of information and poor control on the variability of the upper boundary conditions (i.e., temperature), or of in-situ physical properties (i.e., porosity).

The heat flux may be affected by several factors, the most important of which for the COST B2/B3 wells are heat production by radioactivity and the effects of sedimentation. We have analyzed these factors and combined them in a model to determine the unperturbed heat flux from the basement (crystalline) rocks.

5.1 RADIOACTIVE HEAT PRODUCTION IN THE SEDIMENTS

Radioactive heat production is a scalar and isotropic petrophysical property independent of in situ temperature and pressure. In radioactive decay

TABLE 10. Summary of Temperature Measurements COST B-2

Depth below K.B. ft (m)	t ₁ (hrs)	t ₂ (hrs)	Recorded Temp., T (°F)	In-situ Temp., °F (°C)	Temperature Gradient, °C/km
388 (118.3)				~43(6.0) [†]	
4015 (1223.7)	0.23	4.0	99.5		29.0
	0.23	7.0	100	100.5(38.1)	
	0.23	14.0	100		
10,064 (3067.5)	3.94	7.0	150		25.9
	3.94	12.8	164		
	3.94	20.0	170	186.7(85.9)	
	3.94	22.5	170		
	3.94	23.5	178		27.2
	3.94	43	180		
13,465 (4104.4)	6.5	7	204		
	6.5	13	216	237.4(114.1)	
	6.5	17.8	222		17.8
16,005 (4878.3)	4.62	7.3	230		
	4.62	14.5	245	262.2(127.9)	
	4.62	18.8	247		

† estimated from hydrographic data

COST B-2 DOWNHOLE TEMPERATURES

Fig. 12. Plot of recorded bottom hole temperatures vs. $\ln (1 + t_1/t_2)$ for multiple log runs at three depths in the COST B-2 well. t_1 is total disturbance (circulation) time at each depth, and t_2 is the time since cessation of the disturbance. The temperature intercept is the undisturbed formation temperature (Table 10).

TABLE 11. Summary of Temperature Measurements COST B-3

Depth below K.B. ft (m)	t_1 (hrs)	t_2 (hrs)	Recorded Temp., T(°F)	Calculated in-situ† Temp., T_o , °F(°C)	Temperature Gradient, °C/km
(831.5)(SF)	----	----	~40	~40(4.4)	34.6
5731 (1746.8)	0.365	9.0	96	97(36.1)	
	0.365	5.0	97		
	0.365	11.0	97		
	0.365	13.0	97		25.8
12,655 (3857.2)	6.57	7.75	138	195(90.6)	
	6.57	14.0	156		
	15.5?	80	178		
	15.5?	101.5	(158)*		
	15.5?	103.5	183		26.8
15,779 (4809.4)	5.4	11.25	210	241(116.1)	
	5.4	17.25	217		
	5.4	24.75	225		
	5.4	41.25	232		
	5.4	51.75	239		
	6.9	123.75	232		

* Doubtful value, omitted from extrapolation to in-situ temperature

† Calculated by least-squares fitting of $T = T_o + A \cdot \ln(1+t_1/t_2)$

t_1 = total time of drilling/circulation disturbance at depth being considered.

t_2 = time from end of drilling/circulation disturbance to measurement.

COST B-3 DOWNHOLE TEMPERATURES

Fig. 13. Plot of recorded bottom hole temperatures vs. $\ln(1 + t_1/t_2)$ for multiple log runs at two depths in the COST B-3 well.

COST B-2

Fig. 14. Summary of COST B-2 equilibrium temperatures, interval gradients, mean vertical thermal conductivity (K_z) and interval heat flow vs. depth (dashed lines and values in parentheses are estimated).

COST B-3

Fig. 15. Summary of COST B-3 equilibrium temperatures, interval gradients, mean vertical thermal conductivity ($\bar{K}_z$) and interval heat flow vs. depth (dashed lines and values in parentheses are estimated).

Table 12. Summary of geothermal data averaged over vertical intervals.

-----C O S T B-2-----				-----C O S T B-3-----			
Depth below Seafloor(m)	* $\bar{K}_z$ [TCU] $\frac{\text{Wm}^{-1}}{\text{K}^{-1}}$	Thermal Gradient $\frac{^{\circ}\text{C km}^{-1}}{\text{km}^{-1}}$	Heat Flow [HFU], $\frac{\text{mWm}^{-2}}{\text{km}^{-2}}$	Depth below Seafloor (m)	$\bar{K}_z$ [TCU] $\frac{\text{Wm}^{-1}}{\text{K}^{-1}}$	Thermal Gradient $\frac{^{\circ}\text{C km}^{-1}}{\text{km}^{-1}}$	Heat Flow [HFU], $\frac{\text{mWm}^{-2}}{\text{km}^{-2}}$
0				0			
	[3.52] 1.47	29.0	[1.02] 43		[2.94] 1.23	34.6	[1.02] 43
1105.4				915.3			
	[4.41] 1.85	25.9	[1.14] 48		[4.41] 1.85	25.8	[1.06] 44
2949.2				3025.7			
	[4.60] 1.92	27.2	[1.25] 52		[4.23] 1.77	26.8	[1.13] 47
3986.1				3977.9			
	[4.67] 1.95	(17.8)	[(.83)] (35)				
4760.0							

* $\bar{K}_z$ is mean vertical component of thermal conductivity.

almost the entire decay energy is converted to heat. All natural radioactive isotopes generate heat, but only the contributions of the U, Th and ^{40}K are significant (Rybach, 1976). Radioactive heat production is important in sedimentary sequences because it may provide an appreciable contribution to their thermal history (Richardson and Oxburgh, 1978). Its effect is to increase the heat flow and the temperature within the sediments. For example, Keen and Lewis (1982) have shown that the radiogenic heat production increases the heat flow by up to 15% and the present temperatures within the sediments by 17°C at 5 km depth for the continental margin off Nova Scotia.

All the available samples (82 for COST B-2 plus 40 for COST B-3) were measured for U, Th and K: neutron activation techniques were used to determine the concentrations of U and Th for various lithologies. Total potassium content was determined by atomic absorption spectrophotometry (Appendix B). A more restricted set of 21 samples were counted for ^{40}K peaks on a γ -ray spectrometer.

a) U and Th Determination By Neutron Activation

U and Th analyses on the samples were performed at the U.S.G.S. Laboratories in Denver, CO, with a delayed neutron activation method (e.g., Amiel, 1962). A small amount of sample (less than 10 g) is irradiated in a nuclear reactor and the induced U and Th decay energy peaks are measured. The measuring time required for each sample is about 1 minute. A computerized routine calculates U and Th concentration in ppm with the relative coefficient of variation and the ratio Th/U (Table 4). The coefficient of variation is expressed as percentage of concentration and is equal to 1 standard deviation. The range of U and Th concentration in our samples is bounded between 1-10 ppm and 2.7-20 ppm respectively.

b) Potassium Analysis Through Atomic Absorption Spectrophotometry

Atomic absorption spectrophotometry is an accurate analytical technique to measure the total potassium content of a sample (Appendix B). The potassium content in % weight of each sample is reported on Table 4. The accuracy of the determinations is better than 0.1%. The range of variability was found to fall between 0.06 and 3.53%.

Since potassium in sedimentary rocks is mainly contained in potassium-rich feldspars, micas, illite, glauconite and other secondary minerals, its relative concentration is qualitatively related to the lithology of the sedimentary rocks. Thus a correlation should exist between potassium content and thermal conductivity of the solid fraction, K_s . A plot of K_s as a function of the potassium content (Fig. 16) shows a significant inverse relationship:

$$K_s = 9.21 - 1.24 \cdot \text{POT\%} \quad (12)$$

The correlation coefficient is low, 0.48, reflecting the data scatter particularly at low potassium content. This scatter appears to decrease as the potassium content increases. Similar to the plot of quartz content vs. K_s (Fig. 7), the different lithologies are also separated on this plot of potassium content vs. K_s . At low potassium content two different coexisting lithologies cause a maximum range in K_s : sandstones at the upper K_s range

Fig. 16. Correlation of potassium content and thermal conductivity of the solid fraction K_s of samples. Solid line and equation give the best linear fit to all data. Note the general separation of lithologies.

and limestones at the lower. Potassium concentrations higher than 3% are observed only in the deepest COST B-2 shales (from 15,000 to 16,000 feet).

These correlations have provided a useful tool in reassigning the lithology of the original samples, which were somewhat subjectively estimated by the original operators. In particular, 13 samples have been moved to limestone from lithologies previously identified differently, as follows:

<u>ID #</u>	<u>Depth (feet)</u>	<u>Previously Identified lithology</u>
<u>COST B2</u>		
M46	4110-4200	Shale
M33	5140-7990	Shale
M43	5200-6900	Shale
M18	9310-9320	Shale
M57	10080-10110	Sandstone
<u>COST B3</u>		
B23	5490-5520	Siltstone
B24	6020-6030	Siltstone
B1	6300-6650	Siltstone
B5	7400-7760	Siltstone
B6	7700-8030	Siltstone
B7	9000-9950	Siltstone
L6	12490-12580	Sandstone
L3	14020-14310	Shale

The new mean K_s for each lithology in the 2 wells is given in Table 5.

In summary, the quartz % and K% variability ranges for most of the COST B-2 and COST B-3 lithologies are given approximately in Table 13. The "theoretical" quartz values for sandstones and shales are deduced from the ranges of K_s for those lithologies, using the parameter curves of Robertson (1979). They are obviously consistent with the observed values.

Table 13. Ranges of matrix thermal conductivity (K_s), quartz content, and potassium content for COST B-2/B-3 lithologies.

<u>Lithology</u>	<u>K_s (T.C.U.)</u>	<u>Quartz % content</u>		<u>K% content</u>
		<u>Theoretical*</u>	<u>Observed (Fig.)</u>	
Sandstone	8.0-13.5	29-63	>20	<1.5
Siltstone	5.0- 9.0	--	10 - 25	1.3 - 2.8
Shale	4.0- 6.5	3-18	<18	>1.7
Limestone	3.5- 8.0	--	<10	<1.0

*from Robertson, 1979

c) ^{40}K γ -Ray Spectrometer Analysis

The direct determination of ^{40}K concentration through natural γ -ray spectrometry is a relatively time consuming technique and has only moderate sensitivity (Lewis, 1974; Rybach, 1976). The natural γ -radiation intensities are low; therefore sample size and weight should be at least a few hundred grams (Rybach, 1976; Keen and Lewis, 1982). Since the COST B-2/B-3 samples were not more than 40-50 g each, only a limited set of samples was measured to check the results obtained with two independent methods.

Twenty-one samples were analyzed at the W.H.O.I. γ -spectrometer (Redfield laboratory, H. Livingston and D. Schneider). The mean detecting time was 14 hours each sample. Two Ge-Li detectors were used alternatively. The output counts per sec. were corrected for the background level, and the net signal was ratioed to the NBS SRM4350 ^{40}K standard of 0.54 disintegrations per second per gram. Calculating the number of ^{40}K atoms and knowing the weight of a mole of ^{40}K , the weight in g of ^{40}K per gram of sample can be obtained.

The ^{40}K concentration in ppm is shown in Table 4. The total K concentration was calculated, assuming a ratio $^{40}\text{K}/\text{K}$ equal to 1.167×10^{-4} (Dohrenwend, et al., 1984). The comparison of these results with the corresponding atomic absorption values gives an acceptable agreement (Fig. 17). A systematic enhancement of the potassium content of about 17% determined by the γ -spectrometer vs. the absorption spectrophotometric technique is attributed to an absolute calibration problem of the γ -spectrometer.

The U, Th and K contents of COST B-2 and COST B-3 samples, summarized on Tables 4 and 14, show that a significant correlation exists between their concentration and lithologies. The U (ppm) vs Th (ppm) plot (Fig. 18a)

TABLE 14. U, Th and Potassium Content and Heat Production of primary lithologies

Lithology	K(%) Content	U (ppm) Content	Th (ppm) Content	Heat Production	(μWm^{-3})*
				COST B-2	COST B-3
Limestone	< 1.0	< 5	3 - 6(?)	$1.15 \pm .21(8)$	$0.81 \pm .11(9)$
Sandstone	< 1.5	< 5	3 - 20	$1.53 \pm .10(22)$	$1.26 \pm .20(7)$
Siltstone	1.3 - 2.8	2 - 10	4 - 18	$2.34 \pm .03(16)$	$1.80 \pm .12(19)$
Shale	> 1.7	3 - 6	11 - 20	$2.37 \pm .03(36)$	$2.15 \pm .13(5)$

* Values of heat production given are mean $\pm$ standard error, number of measurements in parentheses.

shows different relationships for each lithology. Shales exhibit a nearly constant high U (3-6 ppm) and high Th (11-20 ppm) content, siltstones show a more scattered distribution with U between 2 and 10 ppm and Th between 4 and

Fig. 17. Plot of the 21 samples analyzed for potassium (K) with two techniques: γ -ray spectrometry and atomic absorption spectrophotometry. The heavy solid line is the best fit, and the thin line is the relationship of equivalence between both methods.

Fig. 18. Plot of a) Uranium (ppm) vs. Thorium (ppm), b) U (ppm) vs. K (%), and c) Th (ppm) vs. K (%) of COST B-2 and COST B-3 samples. Different symbols adopted for each lithology given in legend.

18 ppm, sandstones are characterized by a roughly constant Th/U ratio (less than 5 ppm U, with Th ranging between 4 and 20 ppm), whereas limestones are characterized by low Th content (3-6 ppm) and no clear Th/U ratio. Nevertheless, the four lithologies taken altogether show a constant Th/U ratio of about 3.3; this value is in good agreement with the ratio in sedimentary rocks, ranging from 2 to 5 (Adams and Weaver, 1958; Adams et al., 1959; Pliler and Adams, 1962; Adams et al., 1964). Figs. 18b and c show plots of U and Th (ppm) vs. total K (%), respectively. A general observation is that limestones have appreciable U, Th and K contents (pure limestones would be expected to contain only minor traces of these elements), and that shales show the highest concentrations. Sandstones and siltstones exhibit variable contents of U and Th depending on their mineralogical composition. In fact U and Th are present in rocks in two different fractions (Rybach and Buntebarth, 1982): incorporated in the mineral structures and occupying lattice positions and/or adsorbed in the intergranular spaces. The latter fraction is highly mobile, and thereby may be redistributed by water circulation; it also increases with the total U content. COST B-2 lithologies include higher concentrations of U, Th and K than those of COST B-3. This difference may be related to the more abundant carbonate fraction present in the COST B-3 well (Table 6).

d) Heat Production Evaluation

Radioactive heat production in a given rock can be calculated from the heat production constants (Rybach, 1976; Rybach and Buntebarth, 1982), which give the heat generated per unit time if the rock density (ρ) and its U, Th and K contents (C_U , C_{Th} and C_K) are known. The following relation, taken from Rybach and Buntebarth (1982), has been used:

$$A(\mu W m^{-3}) = 0.1325\rho(0.718 C_U + 0.193 C_{Th} + 0.262 C_K) \quad (13)$$

where U and Th contents are in ppm, K content is in wt. % and density (ρ) is in $g\ cm^{-3}$. Heat productions calculated from U, Th and K contents of each sample are listed in Table 4 and summarized by lithologies for the two COST wells in Table 14. Values of sample density and porosity for 49 COST B-2 samples (11 Ss., 13 Sl. and 25 Sh.) were unfortunately not available and therefore an average grain density of $2.65\ g/cm^3$ was assumed, approximately the average density of siltstones, sandstones and shales from the same depth interval of the COST B-3 well. From Table 14 it can be seen that the COST B-3 lithologies have mean heat productions systematically lower (by about 10-30%) than those of COST B-2. We surmise that this difference may result from more abundant carbonate deposition at the COST B-3 site (on the continental slope) than at the COST B-2 (on the outer continental shelf).

Rybach and Buntebarth (1982) have shown clear relationships between the petrophysical properties: density, seismic velocity, heat generation and mineralogical constituents for crystalline rocks. Previously Rybach (1975) demonstrated, from 110 measurements on Alpine rocks (mainly igneous and metamorphic, also including 12 carbonates), that no correlation exists between the average thermal conductivity and heat production neither in general nor within individual rock types. On the other hand he suggests that in sand-shale alternating sequences a most probable inverse correlation should exist: thermal conductivity decreases with increasing heat production. On Fig. 19 heat production versus thermal conductivity of the solid fraction is plotted for the

Fig. 19. Radiogenic heat production correlated with thermal conductivity of the solid fraction, K_s . The solid line and equation give the best fit linear relationship to the data, excluding limestones.

122 COST B-2 and COST B-3 lithologies. Excluding limestones, a weak inverse correlation can be observed for the other three remaining lithologies, as suggested by Rybach (1976). The linear fit is:

$$A(\mu\text{W m}^{-3}) = -0.33 K_s (\text{W m}^{-1} \text{K}^{-1}) + 3.12 \quad (14)$$

Table 15 summarizes the calculated radioactive heat production over the vertical intervals at COST B-2/B-3 wells. The overall average heat production within the drilled sediments at the two sites is $1.83 \mu\text{Wm}^{-3}$ and $1.46 \mu\text{Wm}^{-3}$ respectively. Such heat production rates generate radiogenic heat flows of about 1.8 and 1.5 mWm^{-2} per thousand meters of sediments. The total steady-state contribution to surface heat flux of the radiogenic heat production of the drilled sediments is 8.6 mWm^{-2} at COST B-2, and 5.8 mWm^{-2} at COST B-3. If we assume a total sediment thickness of about 12 to 14 km in the Baltimore Trough (Grow, 1980; Watts, 1981) and the same concentration of radiogenic elements in the deep section as in the drilled one, then a total steady-state radiogenic heat flow of 25 and 20 mWm^{-2} can be expected from the sediments at COST B-2 and COST B-3 sites, respectively. This is a substantial fraction (40-50%) of the total heat flux deduced at these sites (see Sec. 5.2) and considerably greater than the radiogenic heat production determined for the sediments of the Nova Scotia margin (Keen and Lewis, 1982).

We note that the sediment heat production should cause a decrease in heat flow with depth over the vertical intervals measured, by about 9 mWm^{-2} over the uppermost 5 km of the COST B-2, and 6 mWm^{-2} over the uppermost 4 km of the COST B-3 (Table 15). Instead, except for the lowermost interval (below 4 km) of the COST B-2 with uncertain data, the heat flux appears to increase somewhat with depth in both wells (Table 12). Probably other data uncertainties are overwhelming the measurement of any systematic decrease of heat flow with depth caused by radioactive heat production.

5.2 THERMAL BLANKETING EFFECT AND HEAT FLUX FROM THE LITHOSPHERE

The sediments which have accumulated at the North American eastern continental margin since the rifting of the North Atlantic Ocean constitute a substantial thickness (14 km at the COST B-2 and perhaps 12 km at the COST B-3 site), affecting the thermal regime of the upper crust. The heat from the "undisturbed" deep crust flowing in through the base of the sedimentary sequence is substantially altered by a number of geological processes which interact with each other through time. These include sedimentary processes - sedimentation, erosion, compaction and water advection; physical and chemical processes, which contribute to the internal heat generation such as the radioactive decay; and volcanic and tectonic processes such as volcanic intrusions, hydrothermal circulation, lateral inhomogeneities, etc. The physical parameters, geometry and timing of some of these processes, particularly erosion, volcanic intrusions, hydrothermal circulation and lateral inhomogeneities may have been important at times and in some settings, but for which the necessary detailed data are lacking; therefore they will be not considered here. In the following the effects of sedimentary processes and radiogenic heat production are evaluated quantitatively.

The 4.8 km of sediments drilled at the COST B-2 and 4.0 km at the COST B-3 were deposited since the Upper Jurassic (~ 140-150 Ma), leaving an additional 8 to 9 km or so of older Mesozoic sediments above the crystalline

TABLE 15. Calculated Radioactive Heat Production
over vertical intervals.

Depth Interval below K.B. (m)	$\bar{A}$ (μWm^{-3}) [*]	Interval heat flow Contribution (mWm^{-2})
<u>COST B-2</u>		
118.2(SF)-1223.7	1.76	1.9
1223.7-3067.5	1.86	3.4
3067.5-4104.4	1.75	1.8
4104.4-4878.3	1.97	1.5
	Mean†: 1.83	Total 8.6
<u>COST B-3</u>		
831.5(SF)-1746.8	1.76	1.6
1746.8-3857.2	1.43	3.0
3857.2-4809.4	1.24	1.2
	Mean†: 1.46	Total 5.8

* Calculated as $\bar{A} = 0.1325p (0.718 C_u + 0.193 C_{th} + 0.262 C_K)$, from
Rybach and Buntebarth (1982)

where C_u , C_{th} , and C_K are respective concentrations of U, Th
(both in ppm), and K(%), using the values measured in Table 14
considering the relative amounts of lithologies over depth
intervals in Tables 8 and 9.

† Mean value weighted according to vertical intervals.

basement. The total sediment accumulation in the Baltimore Trough (Fig. 20), mainly taken from Watts and Thorne (1984), shows two major distinct patterns: a very high deposition rate from the beginning of the rifting (some 190 Ma) until Neocomian times (about 130 Ma), followed by a much slower deposition up to the present; in this latter interval a significant increase in sedimentation during the Serravallian (15-11 Ma) can be observed (COST B-2). Applying the simple Hamilton (1959) formula:

$$H = [(1+e_o)/(1+e_f)](H-\delta H) \quad (15)$$

where: H = original thickness of unconsolidated sediments
 δH = amount of consolidation
 $H-\delta H$ = present thickness
 e_o = original void ratio = original porosity/(1-porosity)
 e_f = final void ratio = final porosity/(1-porosity)

and assuming a reasonable mean porosity value of 3-5% for the older undrilled section and a mean porosity of about 10-20% from the COST B-2 and B-3 porosity/depth distributions for the drilled section (Fig. 10), a rough estimate of these two sedimentation rates can be obtained: about 300-350 m/M.y. during pre-Neocomian times and 50-70 m/M.y. during the last 130 M.y. This latter value is in good agreement with Poag's (1980) estimate (Fig. 6).

Assuming an average constant sedimentation of 150 m/M.y. over 190 M.y. and using Von Herzen and Uyeda's (1963) model, the surface non-radiogenic heat flow will be about the 70-75% of the value if no sedimentation would have taken place. The $\sim 45 \text{ mWm}^{-2}$ measured, reduced by the radiogenic contribution of $20\text{-}25 \text{ mWm}^{-2}$, will result in a deep undisturbed heat flow of only $25\text{-}35 \text{ mWm}^{-2}$ for a constant sedimentation rate of 150 m/M.y. for 190 M.y.

Recently, more realistic numerical models of the sedimentation processes, accounting for compaction, pore water movement and surface temperature changes have been developed (Hutchison, 1985). The physical one-dimensional model consists of 2 main layers: the lower one is assumed to be non-porous basement, whereas in the upper layer the proportions of sediment grains and pore water change as the sediment column compacts. As the deposition proceeds and compaction take place, pore water is assumed to move upward (one-dimensionally) as a function of depth with respect to the subsiding basement surface. Radiogenic heat production within the sediments is also included in the model.

Appropriate physical and thermal constants for the sediments and interstitial water have been assumed in the model, and the heat flux at the basement interface is assumed constant. The mean matrix density, specific heat, and thermal conductivity of sediment grains are taken as 2650 kg m^{-3} , $1045 \text{ J Kg}^{-1} \text{ K}^{-1}$, and $2.5 \text{ W m}^{-1} \text{ K}^{-1}$, respectively; the same parameters for water are 1020 kg m^{-3} , $4180 \text{ J Kg}^{-1} \text{ K}^{-1}$, and $0.67 \text{ W m}^{-1} \text{ K}^{-1}$. The dependence of these on temperature and pressure was not included. Compaction of sediments is assumed to have an exponential dependence with depth(z), so that porosity (Φ) is given by

$$\Phi(z) = \Phi_o e^{-z/\lambda} \quad (16)$$

Fig. 20. Estimated sediment accumulation over time at the COST B-2 and COST B-3 wells (top), taken mainly from Watts and Thorne (1984). Dashed lines represent the earliest (constant) deposition, beginning about 190 Ma. The relative heat flow over time (bottom) gives the effect of the sedimentation, not including internal heat production. Model calculations assume two different sedimentation rates, one for the earliest 60 M.y. and another for the remaining 130 M.y. (see text).

where Φ_0 is the surface porosity, λ is the compaction constant. Φ_0 and λ , as well as the radioactive heat production, were varied in model calculations. For a given Φ_0 and λ , a two-layer approximation is made to estimate the sedimentation rates for the COST B-2 and B-3 sites: 9 and 8 km deposited between 190 and 130 Ma, and 5 and 4 km between 130 Ma and present, respectively (Fig. 20).

Some experiments have been run with the numerical model to investigate the sensitivity of results to the measured or assumed parameters. For a given total thickness of non-radioactive sediments (14 km for COST B-2, 12 km for COST B-3), the (assumed constant) basement heat flux is reduced by factors of about 0.81 and 0.84, respectively (somewhat higher sediment rates at COST B-2 vs. COST B-3). With or without sediment radioactivity, for a given sediment thickness the resultant surface heat flux is almost independent of the surface porosity or compaction constant. This can be understood if the sediments are viewed as falling through the pore water column as they compact, such that the transient surface heat flux is primarily a function of the total mass of (solid) sediment and its rate of deposition, and insensitive to the initial porosity or rate of compaction of the sediments with depth. This result is fortunate since the total sediment thickness is one of the better known parameters, whereas the physical properties of the sediments over time are less certain.

The surface heat flux is very dependent on the mean radioactive heat production in sediments when they are as thick as those at the COST B2/B3 sites, as noted by Keen and Lewis (1982). For the sedimentation histories at the COST B2/B3 sites, the numerical model experiments show that the additional surface heat flux caused by uniform sediment heat production is reduced by about 12-15% in comparison to the steady-state. The reduction is somewhat greater for the COST B-2, because of the greater sediment thickness, than for the B-3. The contributions to surface heat flux are proportional to the mean heat production, and for the respective total sediment thicknesses at the COST B-2 and B-3 are 11.2 and 9.7 mWm^{-2} per 1 μWm^{-3} mean heat production. Hence, the surface heat flux produced by the sediment column at these COST sites is nearly half the total measured if the sediment has a mean heat generation of $2\mu\text{Wm}^{-3}$. Lacking samples, we do not know the radioactive contributions from over half of either the COST B-2 or B-3 sediments, but we have no reason for expecting that the mean values below the sections drilled are substantially different from the samples measured.

Table 16 summarizes the heat flow and corrections. If we use the measured mean sediment heat production values (Table 15), the contributions to the present surface heat flux of the sediments at the COST B-2 and B-3 sites are about 21 and 14 mWm^{-2} , respectively. Combined with the uncorrected heat flow estimates (Table 12), and the transient suppression of the surface heat flux by the sedimentation, the estimated basement heat flux at both the COST B-2 and B-3 wells is in the range of 33-39 mWm^{-2} . The basement flux does not appear significantly different at the two sites within the uncertainties of the measurements and assumptions.

6. DISCUSSION AND CONCLUSIONS

Heat flow measurements at the Earth surface are particularly meaningful in geological and geophysical interpretation if they are accurately

Table 16. Heat Flow Summary

Site	Uncorrected* Heat Flow (mWm ⁻²)	Mean radioactive† Heat production (μWm ⁻³)	Sediment Radioactive Correction (mWm ⁻²)	Transient Sedimentation Factor	Estimated Basement Flux (mWm ⁻²)
COST B-2	48 - 52	1.83	21	0.81	33 - 38
COST B-3	44 - 47	1.46	14	0.84	36 - 39

* from Table 12

† " Table 15

estimated in deep wells. In most cases, thermal gradients estimated from logs in oil and gas wells are accurate, on average, within about ± 10 -15% (Reiter and Jessop, 1985). These accuracies can be improved to perhaps ± 4 -5% if multiple temperature measurements are available at the same depth at different times and if the errors produced by instrumental and environmental disturbances are known.

On the other hand, thermal conductivity evaluation is generally more elusive and difficult to calculate in situ within known bounds. This parameter depends on the mineral constituency of the geological materials (Horai and Simmons, 1969; Horai, 1971; Horai and Baldrige, 1972), geometry of mineral grains (King and Simmons, 1972; Kappelmeyer and Haenel, 1974), and in situ conditions (Birch and Clark, 1940; Kappelmeyer and Haenel, 1974; Sekiguchi, 1984). In Section 4.6 we have shown that temperature, together with the porosity and anisotropy of minerals, all generally a function of depth, substantially influence the in situ thermal conductivity of the sedimentary formations.

Thermal conductivity measurements on well samples are scarce; their dependencies on temperature, pressure-porosity and direction are not always adequately constrained. More often, thermal conductivity values are estimated from lithological composition, assuming mean values of the thermal conductivity of constituents from the literature. In this case any rock type has a spread of values that can exceed a factor of 2 (Kappelmeyer and Haenel, 1974; Reiter and Jessop, 1985). The total uncertainty introduced by these errors and/or assumptions on thermal conductivity evaluation could be reasonably estimated as ± 30 -50%. Thus in estimating the heat flow from deep wells, the thermal conductivity is at least 2-5 times more uncertain than the respective thermal gradient.

Five out of the twelve data groups identified by Reiter and Jessop (1985) along offshore eastern Canada show a mean thermal gradient almost coincident with those of COST B-2 and COST B-3 wells. However estimated mean thermal conductivities are substantially higher (about 50%) in the Canadian province, even though age, nature and evolution of the sedimentary sequences are comparable. Rather than there being real differences in the thermal conductivity of the rocks in the two provinces of this order, it seems likely that the methods of estimation used are responsible for much of the discrepancy.

Although direct measurements of thermal conductivity are tedious, they may be required whenever the most accurate heat flow values are desired.

6.1 SAMPLES TREATMENT AND LABORATORY MEASUREMENTS

Sorting of the original samples by lithology was somewhat subjective, but it was significantly improved by a posteriori quartz and potassium content determinations. Density and porosity determinations on pulverized samples are probably the most uncertain among all the laboratory measurements (Table 4) mainly because of the difficulty in saturation of the samples (air bubbles included in the mixture) and thereby in weighing the pore water. This quantity is directly proportional to grain density and porosity of the sample, therefore an eventual underestimate of pore water content leads to an underestimate of matrix conductivity (K_s) when the Woodside and Messmer (1961a) geometric mean model (eqn. 3) is applied to conductivity measured (K_m) on saturated laboratory samples. In the same way an underestimate of the density lowers the radioactive heat production of a rock with fixed U, Th and potassium contents. Of course these errors propagate into the final heat flow evaluation. The porosities and densities of many COST B-2 and a few COST B-3 samples (Table 4) are quite approximate (values in parentheses). The mean K_s for each lithology (Table 5) was calculated on a selected set of values measured on samples with both reliable porosities and K_m . For the determination of radioactive heat production (Tables 4 and 14) we used all the reliable density estimates, assigning values from nearby samples for unreliable determinations or assuming a mean density of 2.65 g/cm^3 for missing values. Uncertainties in porosity and density of the samples are probably about $\pm 10\%$ and $\pm 5\%$ respectively.

The thermal conductivity of the saturated sample (K_m) has been measured with the needle probe technique (Von Herzen and Maxwell, 1959) within about $\pm 1\text{--}2\%$ uncertainty (Goldberg et al., 1980). The uncertainty in thermal conductivity of the matrix (K_s) is obtained by combining those associated with porosity and K_m . At low porosity (0–10%) the uncertainty is between $\pm 2\%$ and $\pm 5\%$, whereas at high porosity (40–50%) it ranges around $\pm 15\text{--}18\%$. Averaging all the K_s determinations by lithology (Table 6) decreases the total uncertainty. A mean uncertainty of $\pm 8\text{--}10\%$ can reasonably be estimated for the K_s of each lithology. The uncertainty of quartz content determined by x-ray powder diffraction has been estimated as $\pm 5\%$. A moderate but significant correlation between quartz content and thermal conductivity of the solid fraction has been observed (Fig. 7). A plot of measured vs. calculated thermal conductivities based on quartz content (Fig. 8) shows that quartz is the most abundant high conductivity mineral in the samples, followed by a subordinate percentage of probable dolomite and calcite.

Potassium analysis has been done mainly to evaluate the amount of ^{40}K of the samples (Table 4). An interesting correlation was found to exist between potassium content and thermal conductivity of the matrix K_s (Fig. 16); this observation proved to be useful in lithological assignments and as an indicator of shale content of the samples. The accuracy of the atomic absorption spectrophotometry technique is better than 0.1%.

Uranium and thorium contents in the samples were measured by neutron activation methodology. These measurements constitute a substantial data set (122 samples) for sedimentary rocks. The accuracy of these determinations is estimated to average ± 5 to $\pm 10\%$ for U, depending on the concentration, and $\pm 2-3\%$ for Th.

6.2 WELL LOG ANALYSIS

The volumetric fraction of each lithology vs. depth was averaged over 1000 feet (304.8 m) intervals (Fig. 9 for COST B-2 and Table 6 for COST B-3); this estimate is a subjective interpretation of the lithological log and therefore a $\pm 10\%$ uncertainty is reasonable.

Porosities were estimated from: sidewall cores, drill cuttings (liquid pycnometry) and well geophysical and lithological logs. COST B-2 porosities (Table 7) are based on more data than those of the COST B-3 (Table 9). The porosity of sandstones shows a large scatter and an appreciable difference for the various methods used (Fig. 10a), whereas siltstones, limestones and shales (Fig. 10b,c) seem to have less variability. The uncertainties of situ porosities should be within the variability for a lithology at a given depth interval, and generally is of the order of $\pm 10-20\%$.

Sonic and lithological logs of COST B-2 were analyzed to verify the relationship between sound velocity (V_p), thermal conductivity (K_z) and formation temperature (Houbolt and Wells, 1980). The agreement between average interval K_z and the values predicted independently by the model (Fig. 11) is reasonable. A systematic bias of about 20% which seems to increase with depth, could be explained by an underestimate of the anisotropy factor, which also probably increases with depth, and/or from the temperature effect.

The anisotropy of the thermal conductivity is more prominent for layered formations such as shales limestones and siltstones (King and Simmons, 1972; Kappelmeyer and Haenel, 1974). Insufficient work has been done to fully understand the behaviour of thermal anisotropy of rocks under varying conditions of temperature, pressure and diagenetic processes. Anisotropy must be considered when thermal conductivity measurements on oriented and sufficiently large samples are not possible.

The final evaluation of the average in situ vertical component of the thermal conductivity K_z (Figs. 14, 15), (with $\pm 15\%$ uncertainty estimated for in situ porosity, $\pm 10\%$ for matrix thermal conductivity, $\pm 10\%$ for the anisotropy factor, and $\pm 10\%$ for the volume fraction estimate) has an accuracy of the order of $\pm 20\%$.

The calculation of equilibrium temperatures at logging depth for the two sites was determined from extrapolation of temperatures measured on multiple logging runs at certain depths (Eqn. 11). A well-established extrapolation procedure (Bullard, 1947; Middleton, 1979) allowed us to estimate the true formation temperature within an accuracy of about $\pm 1-2\%$ (Fig. 12 and 13) when reliable notes on operations during logging were available. The deepest COST B-2 and COST B-3 temperatures were less accurate and problematic. The estimate of the appropriate sea floor temperatures has a much larger uncertainty (perhaps $\pm 10-20\%$) and therefore is only approximate. The mean

deep thermal gradients obtained at the two sites are, on average, accurate to $\pm 5-8\%$.

6.3 HEAT FLOW ESTIMATE

Temperatures, thermal gradients, conductivities and heat flows of COST B-2 and COST B-3 wells are summarized on Figs. 14 and 15. The most reliable COST B-2 thermal gradient should range between the mean value calculated from the two intermediate intervals ($26.3\text{ }^{\circ}\text{C km}^{-1}$) and the mean of all the intervals excluding the uppermost ($23.0\text{ }^{\circ}\text{C km}^{-1}$). The deepest COST B-2 interval gradient is anomalous with respect to the other interval gradients; it is probably caused by a poor estimate of the deepest temperature. The COST B-3 average thermal gradient of $26.1\text{ }^{\circ}\text{C km}^{-1}$ is better constrained and agrees with the value for the intermediate depth intervals of COST B-2. The gradients may increase slightly with depth in both the COST B-2 and B-3, although the increase in both cases is probably not significant nor does it correlate with decreased thermal conductivity with depth, as would be expected if heat flow were constant with depth. The gradients are much more uniform than found for boreholes in offshore eastern Canada by Reiter and Jessop (1985).

Thermal conductivities from the two sites are comparable; the mean values for the COST B-2 are systematically higher than those for the COST B-3 by about 10%. This is caused mainly by the different fraction of the various lithologies present in the wells; sandstones are more abundant in COST B-2, whereas more limestones occur in COST B-3. Excluding the uppermost interval, the average values are $1.89\text{ W m}^{-1}\text{ K}^{-1}$ (4.52 T.C.U.) for COST B-2 and $1.82\text{ W m}^{-1}\text{ K}^{-1}$ (4.35 T.C.U.) for COST B-3.

The average heat flow calculated at the two sites is 45 mW m^{-2} (1.07 HFU) for the COST B-2 and 44 mW m^{-2} (1.06 HFU) for the COST B-3. Excluding the poorly controlled top intervals from both wells does not significantly change these estimates. If the bottom interval gradient is also excluded, the COST B-2 value would be about 49 mW m^{-2} (1.18 HFU). The total uncertainty on these estimates is probably of the order of 20-25%.

6.4 INTERNAL HEAT PRODUCTION AND THERMAL BLANKETING EFFECT

Uranium, thorium and ^{40}K concentrations in COST B-2 and B-3 samples are within normal ranges for sediments and sedimentary rocks (Adams and Weaver, 1958; Adams et al., 1964; Rybach, 1976). Significant amounts of radioactive isotopes are found in limestones, indicating most likely the presence of other lithologies or possibly a redistribution of the isotopes by interstitial water circulation.

A weak inverse correlation has been observed between internal heat production of sandstones, siltstones and shales and their matrix thermal conductivity K_s (Fig. 19b), whereas limestones show nearly a constant heat production rate independent of K_s (Fig. 19a).

A significant difference in mean radiogenic heat production is found between COST B-2 and COST B-3 vertical intervals (Table 15); on average the COST B-3 values are about 20% less than those of the COST B-2. Moreover, in the COST B-3 well the heat production rate decreases with depth as the limestone content increases, i.e., approaching the prograding carbonate bank or

reef (Grow, 1980; 1981). Extrapolating the overall average heat production rate (Table 15) over the 14- to 12- km-thick sedimentary sequence at the COST B-2 and B-3 sites, gives an estimated total contribution from the sediments to the surface heat flow of about 25 and 20 mWm^{-2} , respectively.

In evaluating the sedimentation rates on the margin since the early rifting, formation thicknesses and ages are needed. For the undrilled section no data are available, and therefore the sedimentary interval of about 8 to 9 km between the basement and the bottom of the wells was considered produced by uniform sedimentation over 60 M.y. time span. For the COST wells we used the chronostratigraphic table of Watts and Thorne (1984) even though the age of the deepest drilled horizon in the two wells is somewhat uncertain between Early Cretaceous and Late Jurassic times (some 130-150 Ma). The total sediment thickness (Fig. 20) is to a good approximation the result of sedimentation (combined with compaction) at two substantially different rates.

The sedimentation correction program of Hutchison (1985) includes radiogenic heat production in the sediments. Considering the effect of sedimentation only (with no internal heat production), the constant sedimentation model (Von Herzen and Uyeda, 1963) indicates a reduction of the original heat flow to a value (70-75%) which is comparable with that (83%) obtained by Hutchison's (1985) model. The estimated deep undisturbed heat flow through the basement is 33 to 39 mWm^{-2} . The quite surprising result of a surface heat flow higher than that flowing through the basement, primarily because of the radioactivity in the 12-14 km thick sediment column, should be investigated and verified elsewhere.

The slight increase (5-10%) of heat flow with depth over the most reliable intervals of the COST B-2 and B-3 wells is probably not significant since the uncertainties of the data, especially thermal conductivity estimates, are much larger. However, we note that a majority (41 of 71) of the offshore Canadian wells studied by Reiter and Jessop (1985) also show this type of variation, frequently of larger magnitude. Factors which may cause such effects include rapid (transient) sedimentation or water circulation within the measurement intervals. A transient pulse of sedimentation would cause a difference in heat flow between the surface and depth z which would substantially decay over a time given by z^2/κ (κ = sediment thermal diffusivity $\sim 6 \times 10^{-3} \text{ cm}^2 \text{ s}^{-1}$), which is about 1 million years for a depth of 5 km. Although there may have been a higher sedimentation rate for the COST B-2 (not B-3) about 10-15 Ma (Fig. 20), its effects on heat flow variation with depth would have dissipated long ago, and apparently there are no more recent significant perturbations. A steady sedimentation rate could also create a decrease of heat flow with depth. Numerical model results show that the decrease could have been marginally significant during the time of initial high sedimentation rate at the COST B-2/B-3 sites (130-150 Ma, $\sim 4\%$ increase between the surface and 5 km depth), but that the present difference due to this effect over the same depth range is insignificant ($< 1\%$).

Reiter and Jessop (1985) suggested that slow water circulation, at rates between 2 to 13 mm/year, could explain heat flow variations vertically at any site as well as laterally between sites, on the Canadian Atlantic margin. Similarly it is possible that a very slow downward component of water movement ($\sim 1 \text{ mm/year}$) could account for the vertical differences in heat flux at the COST B-2/B-3 sites. Like Reiter and Jessop, we have no independent evidence

to establish the existence of such water circulation, and it seems probable that other sources of data error are responsible. The uniformity of heat flow between the COST B-2 and B-3 sites, about 60 km apart, argues against any significant water circulation patterns on this scale which could affect the heat flow measurements.

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Appendix A. ANALYSIS OF SAMPLES QUARTZ CONTENT

Quartz standards preparation.

Since the amount of quartz requested for the standards preparation and that of α - Al_2O_3 , necessary for both standards and samples, was significant, second order standards have been acquired.

Selected pure quartz from COST B-2 samples was ground and its x-ray pattern compared with the standard α -quartz furnished from the manufacturer of the x-ray machine; α - Al_2O_3 99% (#23,474-5) was acquired from Aldrich Chemical Co. and compared with the NBS α -alumina standard #674. Both the standard sets have been run 3 times on two different machines (USGS and McLean); the secondary standards were found coherent with the first order standards.

The quartz standards and all the samples were x-rayed using a Phillips Norelco x-ray diffraction unit with monochrometer and under the following specifications:

- Radiation - $\text{CuK}\alpha$ monochrometized by means of a curved graphite crystal diffracted-beam monochrometer
- Tube voltage = 45 Kv
- Tube current = 25 mA
- Divergence slit = 1°
- Receiving slit = 0.1 mm
- Scintillation voltage = .8716 Kv
- Gain = 128
- Baseline = 202
- Window = 913
- Mode = Differential
- Time constant = 2
- % signal suppression = 0
- Goniometer speed = $1^\circ/\text{min}$.
- Chart speed = 1 inch/min.

The grain size to which standards and later samples have been reduced by hand grinding was checked on a few samples, but not always controlled. On average, it was that material which would pass through a 100 mesh sieve (less than 149μ). The background sediment chosen to constitute the matrix of the standards and close enough to the real samples matrix was a marine sediment from Panama Basin (obtained from G. Jones, WHOI). This material was not completely quartz free as we would have preferred, but in any case the small amount present (about 2-3%) was taken into account in the standards.

Four samples with a mean range quartz content: 40% quartz, 60% Panama Basin sediment and different amounts of internal standard mixed (i.e. 5%, 10%, 20%, 30% α -alumina) were run to select the best combination in comparing quartz peaks and α -alumina peaks. The 30% internal standard mixture was adopted (Ellis, 1972, chose 33.33% in practice because it is equivalent to half of the sample weight).

Mixtures of standard quartz and Panama Basin sediment were x-rayed with α -alumina as an internal standard. The 3.343 Angstrom quartz peak (101) was ratioed to the 3.479 Angstrom α - Al_2O_3 peak (012) after subtraction of background noise.

A preliminary attempt was made to use the peak height ratio; but since the height, width and shape of the peaks is a function of the abundance as well as grain size uniformity and distribution, the peak does not give by itself as good a fit as shown in Fig. A-1. Therefore the areas under the peaks were measured three times each to smooth the reading errors. The results show that the scatter is significantly reduced (Fig. A-2).

The best fitting straight line equation is:

$$R = -.4098 + .2433 \text{ Qtz\%} \quad (\text{A-1})$$

where R = Quartz peak area/internal standard peak area.

The correlation coefficient is .996, N=21. The fit is not significantly improved by imposing a quadratic relationship. We find that the quartz content can be estimated to an accuracy of $\pm 5\%$ in absolute amount at the 99% confidence level from the area ratio (quartz peak area/ α -alumina peak area), over the range from 0-50% quartz content.

Quartz content evaluation in the samples.

In analyzing the 82 COST B-2 and 40 COST B-3 samples a problem arose of interference of the quartz peak (101) with a small illite (003) peak (A=3.33) and/or mica. In 31 cases the quartz peak was uncontaminated (Fig. A-3, parts A and B), in 63 cases it showed a "shoulder" on one side close to the base of the peak (Fig. A-3, parts C and D), and in the remainder 28 cases a distinct small peak was located on the basal area of the quartz peak (Fig. A-3, part E). In a few samples the α -alumina peak also interfered with distinct clay minerals peaks, but in these cases it was usually quite easy to separate the respective contributions to the patterns (Fig. A-3, part D).

For deep-sea sediments, Heath and Pisias (1979) suggested solving the overlapping of quartz (101) and illite (003) by measuring the main illite peak (001) (shown on the right side of each diffractogram in Fig. 8) and subtract from the total quartz peak area 12.5% of the main illite peak. This percentage was experimentally determined for their samples. This technique was not adopted for the following reasons:

- The COST B-2/B-3 samples are not deep-ocean sediments.
- The overlapping illite peak in the deep-ocean sediments sits just on top of the quartz peak and therefore is not seen.
- Since the continental margin sedimentary sequence underwent diagenetic processes with some of the clay minerals altered to micas, it is not reliable to assume the experimental percentage found by Heath and Pisias (1979) for deep-ocean sediment.
- A comparison of the illite (001) peak with the supposed interfering illite (003) where both are clearly visible (Fig. A-3, part E) indicates that the proportion found by Heath and Pisias (1979) does not apply to our samples.

The interference was graphically eliminated by assuming symmetry of the quartz peak area (Fig. A-3).

Fig. A-1. Quartz standard best linear fitting: the ordinate gives the height ratio between 3.343A quartz peak (101) and 3.479A α - Al_2O_3 peak (012), and the abscissa shows the percentage of quartz in the standards. The correlation coefficient is 0.975.

Fig. A-2. Quartz standard best linear fitting: the ordinate shows the area ratio between quartz and internal standard. Correlation coefficient is 0.996.

Fig. A-3. X-ray diffractograms showing the main different kinds of patterns observed on various lithologies:

- A) COST B-3 limestone (3.8% quartz). Clear pattern; the major illite peak occurring at about 9°, where present, is shown on the right of each pattern.
- B) COST B-3 sandstone (34.7% quartz). Clear pattern, a feldspar peak is observed on the left of quartz peak.
- C) COST B-2 sandstone (30.1% quartz). The "shoulder" (shaded) at the base of the quartz peak is produced by a small interfering peak of illite and/or mica.
- D) COST B-2 siltstone (11.1% quartz). The same interference as in C) is observed on the quartz peak and a kaolinite peak falls nearby the α -alumina peak at 24.9°.
- E) COST B-2 shale (14.1% quartz). A distinct illite and/or mica peak is obvious near the base of the quartz peak pattern.

Appendix B. POTASSIUM ANALYSIS IN SAMPLES BY ATOMIC ABSORPTION SPECTROPHOTOMETRY

A Perkin-Elmer Mod. 403 instrument was used following the appropriate specifications suggested by the manufacturer in preparing the potassium standards and in fusing and running the samples.

Potassium Standards

The potassium standards have to be treated in the same way as the samples. These are supposed to be mixed with lithium metaborate in 1:7 proportion (100 mg of sample and 700 mg of LiBO_2), fused at 1000°C for 15 minutes, dissolved in 4% nitric acid solution and then diluted to a certain volume of distilled water (to be experimentally determined) in order to keep the maximum expected value of K content in the range of linear response of the machine. In successive attempts on a selected set of samples, a dilution factor of 1,000 (dissolved sample diluted to 1000 ml) was verified to be appropriate.

A mixture of LiBO_2 and HNO_3 solutions were prepared to receive the potassium standard exactly as for the samples. KNO_3 was used as a source of potassium because it does not introduce new anions in the solution. To have 1 g/l (or 1000 ppm weight per unit volume (W/V)) of potassium, 2.588 g of KNO_3 should be dissolved and diluted to 1 liter in solution. An intermediate solution with 100 ppm K was prepared taking 10 ml of the 1000 ppm K solution and diluting them to 100 ml with diluting solution. Eight K standards were prepared from 0 to 3.5 ppm at even 0.5 ppm intervals. The complete set of standards was analyzed 6 times during the sample measurements (each 30-40 samples) to check any temporal variation of the absorbance measured with the instrumentation. An intermediate K standard (1.5 ppm K) was run each 5 samples. No significant variation of the absorbance with time was observed.

A straight line best fit of the absorbance as a function of the K content for the 6 different runs gives (Fig. B-1):

$$A = .0936 \text{ ppm K} + .0037 \quad \text{corr. coeff.} = 0.9977$$

$$A = .0947 \text{ ppm K} + .0041 \quad \text{corr. coeff.} = 0.9976$$

$$A = .0949 \text{ ppm K} + .0044 \quad \text{corr. coeff.} = 0.9976$$

$$A = .0936 \text{ ppm K} + .0040 \quad \text{corr. coeff.} = 0.9982$$

$$A = .0914 \text{ ppm K} + .0101 \quad \text{corr. coeff.} = 0.9974$$

$$A = .0884 \text{ ppm K} + .0085 \quad \text{corr. coeff.} = 0.9979$$

A = Instrumentation absorbance measurement

Samples Treatment

Pulverized samples and LiBO_2 were dried in an oven at 60°C for 6-8 hrs., put in a drier to cool down for 2 hrs. and carefully weighed: 100 mg

Fig. B-1. Potassium concentration calibration of atomic absorption instrumentation. Absorbance measurement units are plotted vs. potassium (ppm) in the standard solution. The set of points associated with each line (least square fit) is a separate calibration, taken approximately at 30-40 min. intervals during measurement of samples. Successive calibrations are shifted along the potassium axis by 0.5 ppm for clarity.

of sample to be mixed with 700 mg of LiBO_2 . To obtain a good fusion a complete dry mixing of the samples was required (1 minute each vial on a vortex mixer). Each sample was dropped in a pure graphite crucible and fused in a furnace at 1000°C for 15 minutes. The molten sample was poured in a small amount of solution (HNO_3 , 4%) causing its immediate solidification in glassy chips, which are subsequently dissolved by agitation with magnetic stirrers. The solution was then diluted to the final volume of 1000 ml. All the samples together with the potassium standards were analyzed in the same day.

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