Revista Brasileira de Geofísica; 1987, Vol. 5, 117-126

doi: 10.22564/brjg.v5i2.2232

ESTIMATION OF FORMATION TEMPERATURE AND HEAT FLOW FROM MEASUREMENTS MADE IN SHALLOW WATER WELLS

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Temperature Logs made in water wells are generally unsuitable for the determination of geothermal gradientes because of thermal disturbances by water flows between aquifers through the well. However, bottom parts of the well are usually free of such induced water flows and results obtained in the present work show that temperature measurements at the well bottom can be of use in estimating geothermal heat flux. This method, denominated BWT (bottom well temperature) is essentially similar to the BHT method used in petroleum industry, the main difference being that the thermal probe is placed at the bottom of the well and not some meters above it as is the case in the latter. Furthermore, precision and accuracy of the temperature data is much higher than those in routine BHT measurements and in many cases facilities exist for making measurements long after drilling disturbances have died out. Even in cases where measurements have to be made soon after cessation of drilling the quality and quantity of temperature data that can be obtained permit determination of formation temperatures with a high degree of confidence. Simple theoretical models were developed for correcting drilling disturbances in such cases. The results of modelling studies show that the bottom of the well returns to thermal equilibrium more rapidly than parts situated three or more well diameters above it. An example of application of the method, with measurements in a shallow water well is also presented.

Perfis de temperatura feitos em poços perfurados para produção de água são, em muitos casos, inadequados para a determinação de gradientes geotérmicos, devido à presença de movimento de água no poço. Tal movimento pode ser induzido quando dois aquíferos isolados são interligados pelo poço. A parte mais profunda dessas perfurações, no entanto, não é na maioria dos casos afetada por esse fluxo de água e, o presente trabalho mostra que medidas feitas no fundo dos poços podem ser utilizadas na determinação do fluxo geotérmico. O método desenvolvido é semelhante ao método das temperaturas de fundo de poço (BHT) usado na indústria do petróleo, sendo uma das principais diferenças a posição em que as medidas de temperatura são feitas. No método BHT essas medidas são feitas alguns metros acima do fundo do poço, ao passo que neste caso o termômetro é posto no fundo. Por outro lado, a precisão das medidas é, neste caso, muito maior do que a das temperaturas de fundo de poço e além disso, em alguns casos é possível fazer-se essas medidas com o poço em equilíbrio térmico com a rocha. No caso em que as medidas de temperatura são feitas fora do equilíbrio térmico, a quantidade e a qualidade dos dados permite que se determine a temperatura de rocha com boa precisão desde que se disponha de modelos para extrapolar essa temperatura a partir das medidas. Neste trabalho foi desenvolvido um modelo simples para a determinação da temperatura da rocha. Os resultados obtidos com esse modelo mostram que no fundo do poço o retorno ao equilíbrio térmico é mais rápido do que a uma distância acima do fundo equivalente a três ou mais diâmetros do poço. Um exemplo de aplicação do método é apresentado.

(Traduzido pela Revista)

INTRODUCTION

The large number of water wells normally found in sedimentary basins makes attractive their use in the estimation of terrestrial heat flow densities. There are however, several problems in the use of water wells drilled in sedimentary basins in geothermal studies.

The presence of water flows in aquifers for instance, constitutes a heat transfer mechanism which is much more effective than conduction and, redistributes the heat flow in the basin. Donaldson (1962), has considered the effect of water movement in the redistribution of heat flow. Majorowicz et al. (1985) and Majorowicz et al. (1986) have interpreted the variability of heat flow densities in Canadian sedimentary basins in terms of groundwater flow effects. Since transport constitutes an important heat transfer mechanism, a correct evaluation of heat flow in sedimentary basins would only be possible if water flow were completely known. The condition stated above is rarely found and in general only the conductive component of the heat flow (q) can be evaluated.

Waters wells in sedimentary basins are in general shallow. This fact constitutes an additional difficulty since heat flow analysis made in these wells can provide information only of the upper part of the sedimentary basin. Although it is possible in some cases to have full temperature logs made with the well in thermal equilibrium conditions with the formation, the use of these logs are difficult for two reasons. First if the well cuts permeable formations, which is always the case in water wells, the temperature profile is affected by water flow in the formation. The presence of formation water flow distorts the temperature distribution with depth and the conductive heat flow varies from point to point in the well. Mansure & Reiter (1979) presented a correction for heat flow in the case that there exists a constant vertical groundwater movement. This model however is very simple and may not correspond to most cases that can be found in sedimentary basins. The second problem occurs when the well cuts two or more permeable layers separated by impervious strata. When this situation is found, it is possible that the two aquifers exchange water since the hydraulic head in them is, in general, different. This situation is schematically shown in Fig. 1. Becker et al. (1983a) and Becker et al. (1983b) analysing deep oceanic crustal geothermal measurements, have found a similar situation in a well that crossed impervious sedimentary layers and reached a highly permeable basement. They have used the temperature profile to evaluate the water flow into the permeable basement. Some water wells of the Taubaté Sedimentary Basin, South-east Brazil present thermal equilibrium profiles (Fig. 2) that are probably affected by both disturbing effects of water flow.

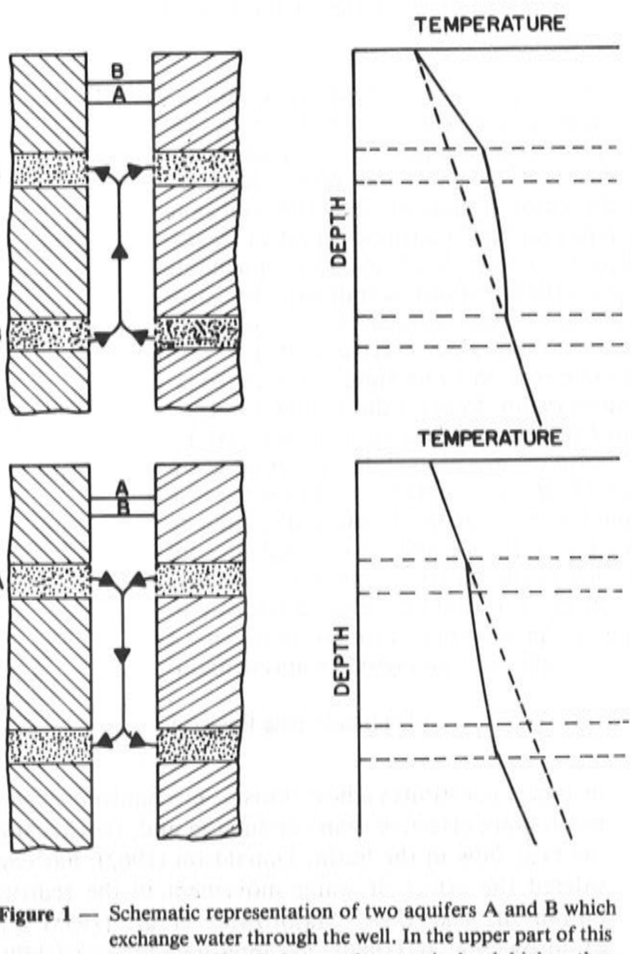

Figure 1 — Schematic representation of two aquifers A and B which exchange water through the well. In the upper part of this figure, aquifer B has a piezometric level higher than aquifer B has a piezometric level higher than aquifer A. In the bottom part the opposite is true.

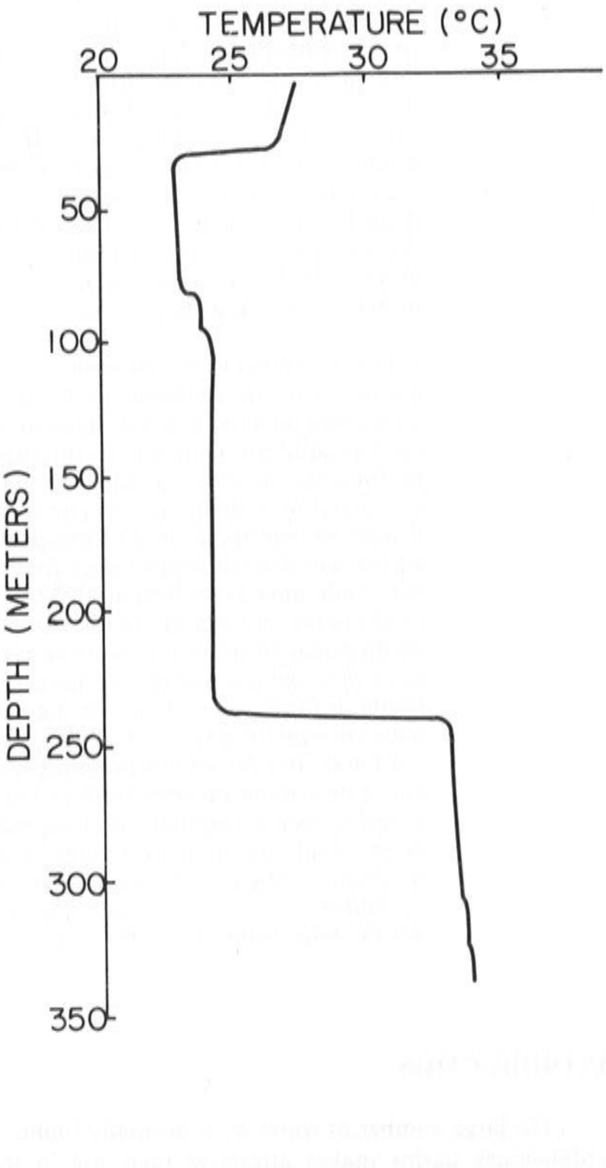

Figure 2 Temperature profile in a water well at the country of Lorena, Taubaté Sedimentary Basin — Southeast Brazil. The aspect of this temperature profile, with almost isothermic intervals, suggests that water exchange is occuring through the well.

In the present paper, a method for estimating the conductive component of heat flow density using temperature measurements in shallow water wells is presented. The method is essentially the classical thermal resistance method, put forward by Bullard (1939) but uses a single temperature measurement in the bottom of the well. The thermal conductivities of the sediments are measured on the drilling cuttings, recovered during drilling in closely spaced intervals.

If the temperature measurements in the bottom of the well are made soon after drilling, and consequently far from being the equilibrium values, it is necessary to correct the obtained values for drilling disturbances. In this paper a simple model to correct temperature measurements made in the bottom of the well is presented. The method is similar to Middleton's (1979) method for correcting BHT measurements but, it takes in consideration the presence of the bottom of the well. It is necessary to consider the end effect of the well bottom since the thermal probe in this technique is placed at the bottom and not some meters above it as is the case in BHT measurements (Fig. 3).

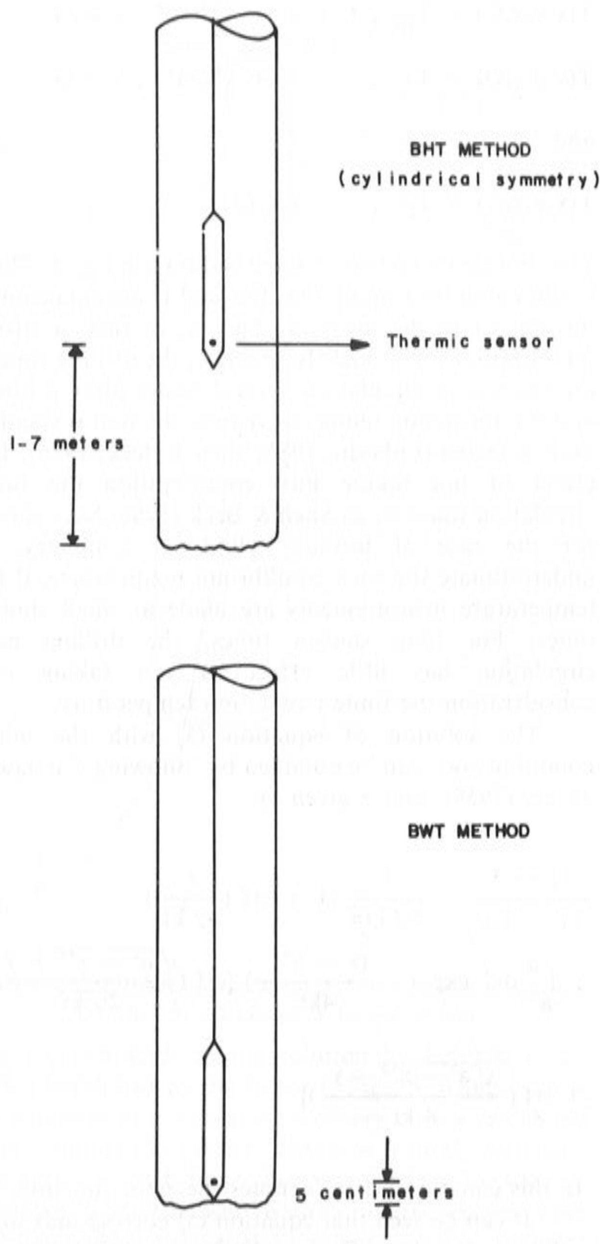

Figure 3— Difference in the location of the thermal probe between the BHT method and the method proposed in this paper.

The necessity to take into consideration end effects limits the possibility of obtaining analytical models that give a better representation of the real situation in the bottom of the well, like those presented by Shen & Beck (1986) for correcting BHT data. On the other hand, a greater number of temperature measurements can be made in the case of water wells and the measurement precision is greater if a thermistor probe is used.

DESCRIPTION OF THE METHOD

This method of heat flow density determination in water wells is essentially an application of the thermal resistance method developed by Bullard (1939). It utilizes only one temperature measurement in the bottom of the well, in a similar way that BHT measurements are made in oil wells, but the temperature determination is much more precise. In water wells, it is sometimes possible to obtain a complete temperature profile made with the well in thermal equilibrium with the rock. In this case although the temperature profile can be distorted by hydraulic factors, the temperature at the well bottom is well determined. At the well bottom, the temperature can be affected by thermal disturbances due to heat transport in the formation but, in general, it is free of disturbances induced by water flow through the well.

In the case of water well which is in thermal equilibrium with the rock, the apparent conductive heat flow density can be simply estimated by the relation

$$q = \frac{T_B - T_o}{R} \quad (1)$$

where T_B is the temperature at the well bottom, T_o is the mean surface temperature. R is the total thermal resistance defined by

$$R = \sum_{i=1}^{n} \frac{l_i}{K_i} \quad (2)$$

l_i, K_i, being the thicknesses and thermal conductivities of the sedimentary layers between surface and the well bottom. The term "apparent heat flow density" is used here to stress that it is calculated under the assumption of negligible water movement in the formation. In the presence of water movement the conductive heat flow varies as a function of space. Only the total heat flow, which is the sum of conduction and transport components, is constant when the system is in steady state and heat production is negligible. In spite of this limitation, equation 1 can be used as an estimation of the conductive component of the heat flow density.

Although it is sometimes possible to make temperature measurements in water wells which are in thermal equilibrium with the rock, it is frequent that such measurements are only possible soon after the end of drilling mud circulation, that means immediately after the drilling activity. In this situation, it is necessary to extrapolate the equilibrium formation temperature from measurements made at the bottom of the well. This is the same situation found in temperature measurements made in deep oil wells. In our case, however, it is possible to make a series of temperature measurements with a close time interval for several hours, without having to remove the thermal probe from the well. Furthermore, the temperature measurements are made with a thermistor thermal probe which allows a greater precision. Another difference between the conventional BHT method and this technique, which for the purpose of differentiating from the previous one will be called bottom well temperature (BWT) method, is that the thermal probe is placed at the well bottom and not one to ten meters above it (Fig. 3). As a consequence, the thermal equilibrium recover is, as will be soon demonstrated, faster than in the first case.

The location of the thermal probe at the well bottom has, however, an inconvenience. The techniques used for extrapolation of the equilibrium temperature from BHT assume an infinite cylindrical symmetry, since in that situation, the heat flow is essentially radial. In the case of BWT method, the proximity of the well bottom invalidat s this assumption.

A quite simple model for extrapolation of equilibrium rock temperature from temperature measurements made near the well bottom was developed following Middleton (1979). This method is based on the solution of the heat conduction equation

$$\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2} = \frac{1}{k}\frac{\partial T}{\partial t} \qquad (3)$$

for a semi-infinite cylindrical well emplaced in a homogeneous medium. It is supposed that there is no contrast between the thermal properties of the well fluid and the sourrounding medium. Leblanc et al. (1981) argued that this model limitation, in the case of a well with infinite cylindrical symmetry, has no practical importance if a thermal diffusivity proper to the drilling fluid is used. Luheshi (1983) and Shen Beck (1986) however, have shown that the thermal diffusivity contrast has a significant influence on the recovery of the rock equilibrium temperature. Blackwell (1953, 1954, 1956) presented analytical solutions of equation (3), in the presence of thermal properties contrast between the cylinder and the surrounding medium, in order to evaluate the axial heat flow error in measurements made by the conductivity probe method. His results can also be applied in the evaluation of the distance from the hole bottom beyond which, its temperature envolves as in an infinite cylinder. However, the geometrical models adopted by Blackwell can not be used in evaluating the hole bottom effect over the equilibrium recovery in its proximities. Also the numerical solutions presented by Luheshi (1983) can not be applied to our problem. This author approximated the vertical temperature gradient by assuming that the horizontal plan through the hole bottom is held fixed at the equilibrium temperature.

In this paper, the homogeneous model is used since the inclusion of thermal properties contrast would lead to a complex analytical solution. We have chosen to sacrifice the model precision in order to keep its simplicity.

The initial conditions applied to equation (3) are that at the end of drilling mud circulation ($t = O$) the temperature inside the well (T_m) is constant. The temperature in the sourrounding medium, at $t = O$, is homogeneous that equal to T_f. Mathematically, this can be expressed by

$$T(x,y,z,O) = T_m \;, \quad O \leqslant x^2 + y^2 \leqslant a^2 \;, \; z \geqslant O$$

$$T(x,y,z,O) = T_f \;, \quad x^2 + y^2 > a^2 \;, \; z \geqslant O$$

and (4)

$$T(x,y,z,O) = T_f \;, \quad z < O$$

The coordinate system adopted is shown in Fig. 4. These initial conditions imply that the well is instantaneously introduced in the medium. This is, in fact, a strong limitation of the model. In practice, the drilling fluid is maintained in circulation several hours after drilling, and the formation temperature near the well is significantly affected (Luheshi, 1983; Shen & Beck, 1986). The effect of not taking into consideration the finite circulation times is, as Shen & Beck (1986) have shown for the case of infinite cylindrical symmetry, to underestimate the rock equilibrium temperature, if the temperature measurements are made at small shut-in times. For long shut-in times, the drilling mud circulation has little effect of not taking into consideration the finite circulation temperature.

The solution of equation (3) with the initial conditions (4), can be obtained by following Carslaw & Jaeger (1959), and is given by

$$\frac{T_f - T}{T_f - T_m} = \frac{1}{8\sqrt{kt\pi}} \left(1 + \operatorname{erf}\left(\frac{z}{2\sqrt{kt}}\right)\right) \cdot$$

$$\cdot \int_{-a}^{a} dx' \exp\left(-\frac{(x - x')^2}{4kt}\right) \left[\operatorname{erf}\left(\frac{\sqrt{a^2 - x'^2} + y}{2\sqrt{kt}}\right) + \right.$$

$$\left. + \operatorname{erf}\left(\frac{\sqrt{a^2 - x'^2} - y}{4\,kt}\right)\right] \qquad (5)$$

In this equation erf (x) denotes the error function.

It can be seen that equation (5) corresponds to the solution for an infinite cylinder emplaced instantaneously in a homogeneous medium multiplied by a factor

$$f = \frac{1}{2}\left(1 + \operatorname{erf}\left(\frac{z}{2\sqrt{kt}}\right)\right)$$

This implies that the recovery of equilibrium temperature is faster near the well bottom by a maximum factor of two. On the cylinder axis equation (5) reduces to

$$\frac{T_f - T}{T_f - T_m} = \frac{1}{2}\left(1 + \operatorname{erf}\left(\frac{z}{2\sqrt{kt}}\right)\right)\left(1 - \exp\left(-\frac{a^2}{4kt}\right)\right)$$

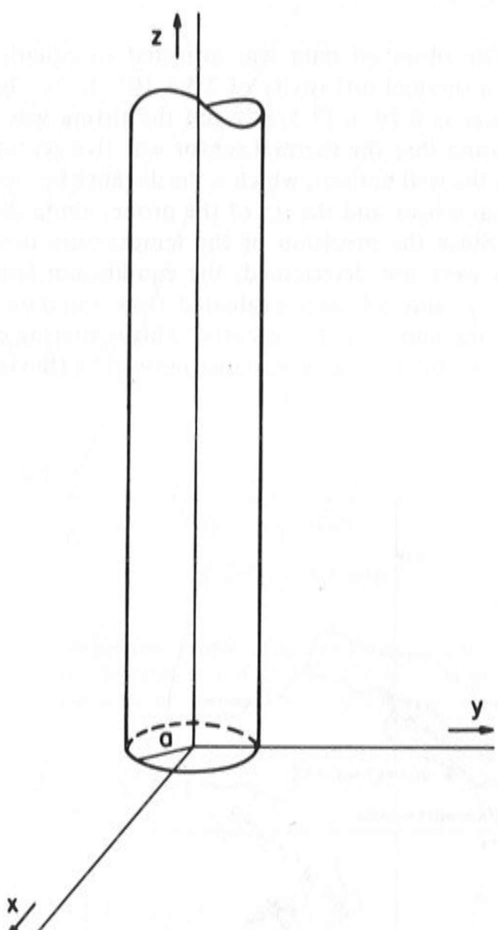

Figure 4— Coordinate systems adopted for the solution of equation (3). The origin is located at the well bottom.

which corresponds to the solution by Leblanc et al. (1981) multiplied by the factor f. Figs. 5, 6 and 7 show the equilibrium temperature recovery curves calculated from equation (5) for three distances from the well axis. The curves in each figure are parametrized for various distances from the well bottom. These curves were calculated using a thermal diffusivity of 2.5 x 10^{-7} m^2/s which seems to be a good value for the drilling mud (Leblanc et al., 1981).

It is apparent from these figures that near the well bottom is, the return to thermal equilibrium occurs faster. Furthermore, comparison between these figures shows that the recovery of equilibrium temperature is faster near the well wall.

To extrapolate the equilibrium rock temperature from a series of temperature measurements made at the well bottom, the measured data BWT (t_i) is adjusted to the relation

$$BWT(t_i) = T_f + \Delta \varphi_i \qquad (7)$$

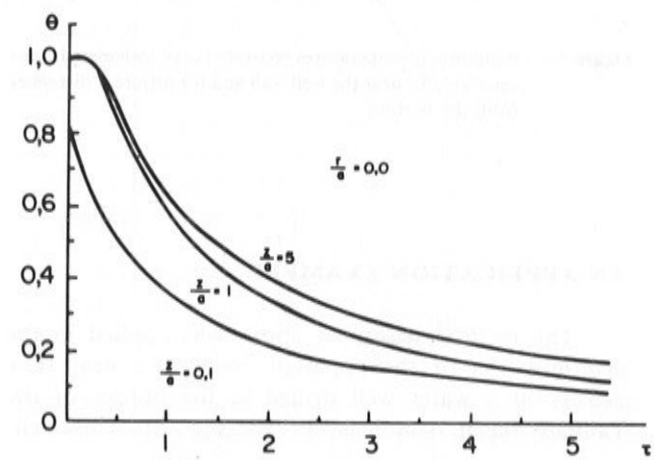

Figure 5 — Equilibrium temperature recovery curve, calculated from equation (5), at the center of the well and for different distances (z) from the well bottom. Θ is the relative temperature difference $(T - T_f)/(T_m - T_f)$ and , is the adimensional time (kt/as^2).

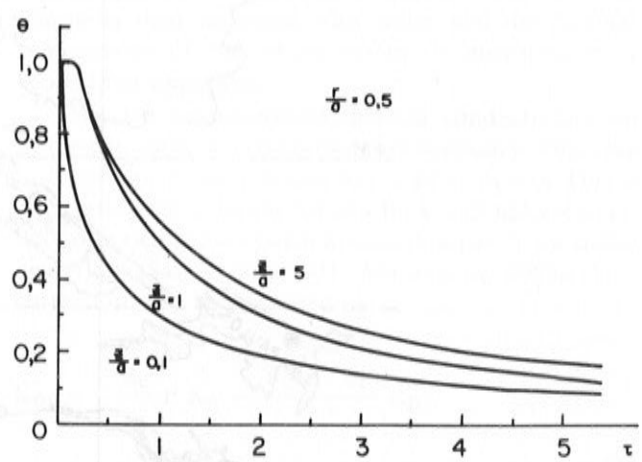

Figure 6— Equilibrium temperature recovery curve, calculated from equation (5), half way between the center and the well wall and for different distances from the bottom.

where φ_i is the right side of equations (5) or (6) calculated at the time ti and ΔT is the difference between the initial mud temperature (T_m) and the initial formation temperature (T_f).

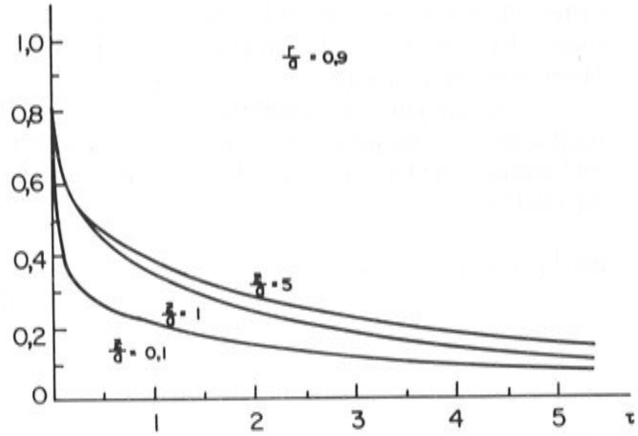

Figure 7— Equilibrium temperatures recovery curve, calculated from equation (5), near the well wall and for different distances from the bottom.

AN APPLICATION EXAMPLE

The method described above was applied in the determination of the apparent conductive heat flow density in a water well drilled at the border of the Taubaté Basin, southeast Brazil (Fig. 8). This well, 152m in depth, was drilled for water supply to the county of Guaratinguetá. Although this well was not in thermal equilibrium, a full temperature profile (Fig. 9) was made. The temperature at the well bottom was recorded in intervals of half an hour, between forty and forty seven hours of shut-in time (Fig. 10).

The instrument utilized for the measurements was a thermal probe with a thermistor sensor. During the field operation the electrical resistance of the sensor was measured with a Wheatstone bridge and the corresponding temperature was determined by the calibration curve of the instrument, furnished by the Instituto de Pesquisas Tecnológicas de São Paulo S/A. The individual precisions of the temperature determinations were not evaluated from the calibration curve.

The observed data was adjusted to equation (7) using a thermal diffusivity of 2.5 x 10^{-7} m^2/s. The well diameter is 0.19 m (7 5/8'') and the fitting was made supposing that the thermal sensor was five centimeters above the well bottom, which is the distance between the thermal sensor and the tip of the probe, along the well axis. Since the precision of the temperature measurements were not determined, the equilibrium temperature (T_f) and ΔT were evaluated from the data point scattering about the fitted curve. This scattering can be measured by the sample variance defined by (Bevington, 1969).

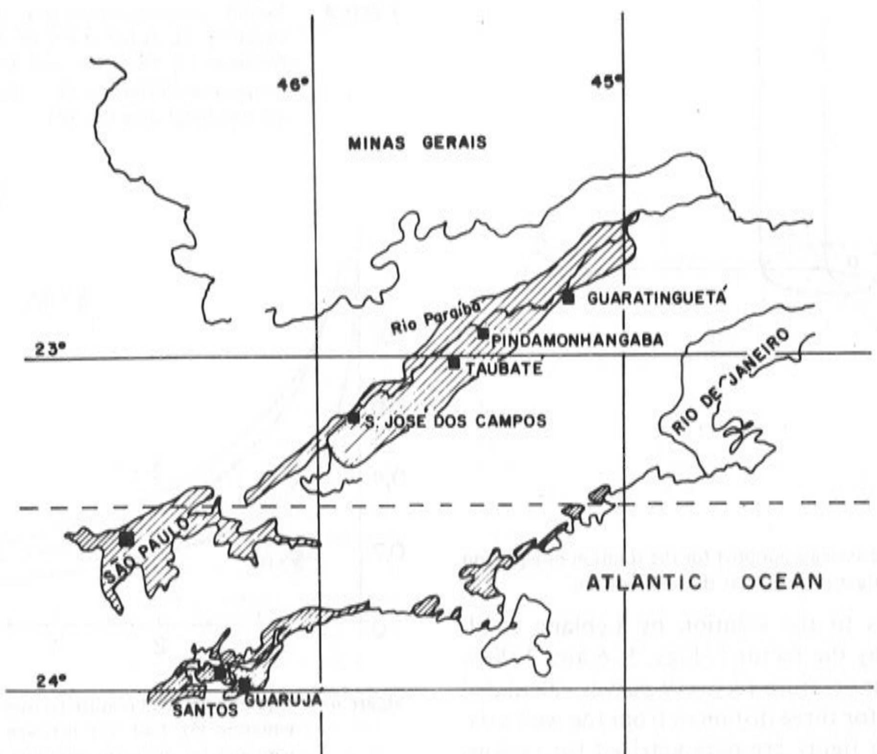

Figure 8— Location map of the Taubaté Sedimentary basin and the city of Guaratinguetá, state of São Paulo, Brazil.

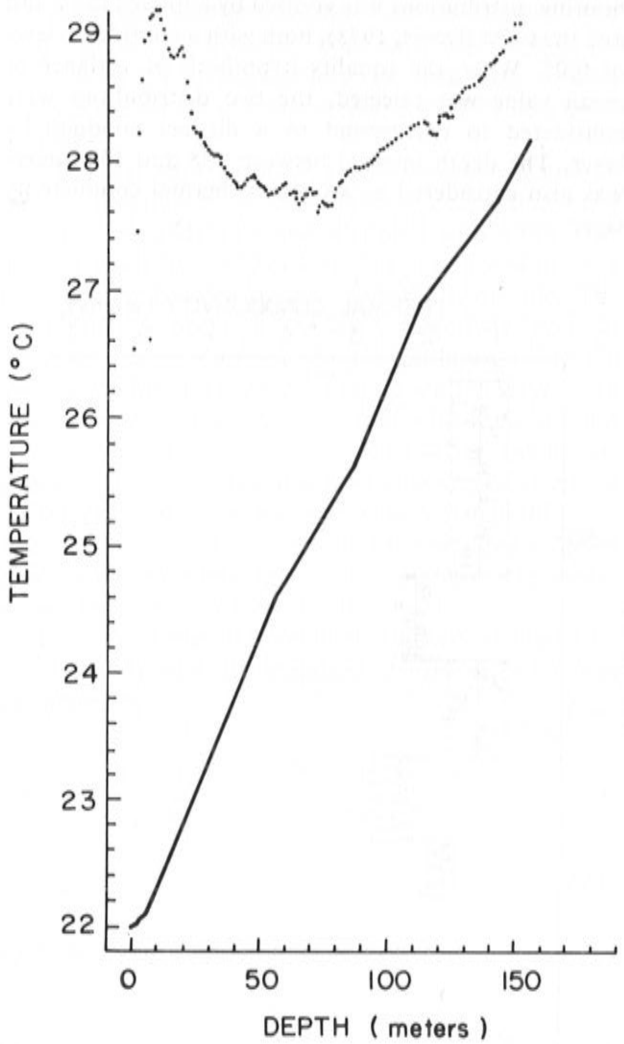

Figure 9— Temperature profile of the well, represented by dots, recorded after end of mud circulation. Continuous line represents the calculated conductive temperature profile.

$$s^2 = \frac{1}{N-2} \sum_{i=1}^{N} (T_i - T_{c_i})^2 \tag{8}$$

where T_i and T_{c_i} are the corresponding measured and calculated temperatures and N is the number of measurements. The calculated sample variance was 3 x 10^{-6} (°C)².

The extrapolated formation temperature (T_f) obtained with this procedure is (28.1 ± 0.4) °C and the difference between the formation temperature and the initial drilling mud temperature (T_m) is (— 6.57 ± 0.07) °C.

For the determination of the total thermal resistance (equation 2), it is necessary to measure the thermal conductivity of rocks that constitute the sedimentary layers crossed by the well. Since the drilling technique used in this well does not permit the recovery of sample cores, the thermal conductivities were measured in drilling cuttings recovered continuously during drilling for depth intervals of two meters. The divided bar method (Beck, 1965) adapted for thermal conductivity measurements in drilling cuttings (Sass et al., 1971; Marangoni, 1986) was used in this work.

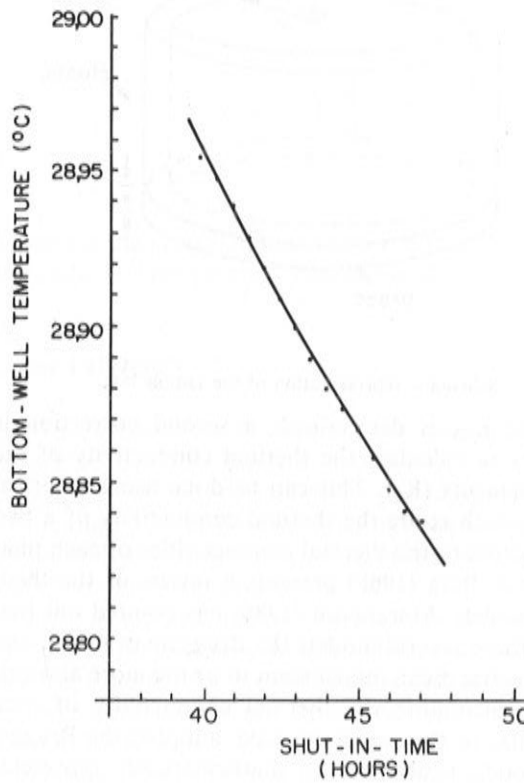

Figure 10— Temperature recovery curve at the well bottom. Measured data is represented by dots. Continuous line represents the adjusted temperature recover curve.

The drilling cuttings were dried at 50°C and packed in a sample box which is schematized in Fig. 11. The sample is then saturated with water and the thermal conductivity of the whole system is measured in a divided bar apparatus.

For the determination thermal conductivities on drilling cuttings two corrections are necessary. First, the influence of the sample box has to be removed. This is done using the relation for the bulk thermal conductivity of a two phase system arranged in parallel with the heat flow (Sass et al., 1971; Marangoni, 1986). This correction may be expressed as

$$K_m = \left(\frac{d_e}{d_i}\right)^2 K_a - \frac{d_e^2 - d_i^2}{d_i^2} K_p \tag{9}$$

where d_e and d_i are the external and internal diameters of the sample box, K_a is the measured thermal conductivity of the whole system (sample box and the saturated sample). K_p is the thermal conductivity of the sample box wall and K_m is the thermal conductivity of the saturated sample.

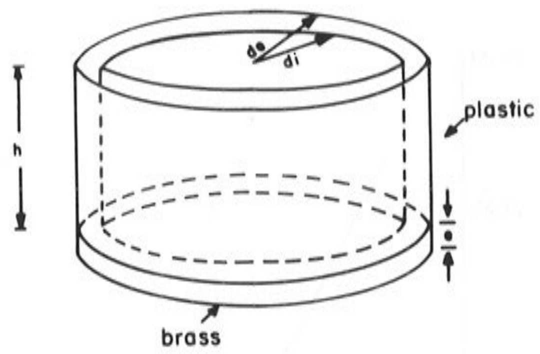

Figure 11— Schematic representation of the sample box.

Once K_m is determined, a second correction is necessary to calculate the thermal conductivity of the rock fragments (K_S). This can be done using theorical models which relate the thermal conductivity of a two phase system to the thermal conductivities of each phase. Hutt & Berg (1968) presents a review of the theoretical models. Marangoni (1986) has pointed out that among these several models the Bruggeman model and the geometric mean model seem to be the more apropriate for calculating the thermal conductivity of rock fragments. In this paper we have adopted the Bruggeman model, which can be mathematically represented by

$$\varphi = \frac{(K - K_S)}{(\frac{K}{K_f})^{1/3}(K_f - K_S)} \qquad (10)$$

where φ and K_f are the water volumetric fraction and thermal conductivity respectively. Equations (8) and (9) can be combined to furnish directly the thermal conductivity of the rock fragments (Marangoni, 1986).

$$K_S = K_f \frac{CR - \varphi\,(CR)^{1/3}}{1 - \varphi\,(CR)^{1/3}} \qquad (11)$$

where CR is K_m/K_f (K_m given by equation (9)).

The obtained thermal conductivities for each 2m deph interval were combined to give the conductivity profile shown in Fig. 12. The conductivity values of the profile in Table 1 were grouped in depth intervals which Table 1

seemed, in a visual inspection, to represent a distribution of thermal conductivities around a mean value. The normality of these distributions were verified using the χ^2 test with four probability bands (Davis, 1973). The hypothesis of normality for all distributions, except that corresponding to the depth interval between 138 and 152 meters, could not be rejected at a confidence level of 0.05. In the depth interval (138, 152m) the eight individual values are either 1.6 or 1.8 W/mK, with one exception of 1.9 W/mK. The hypothesis of equality of variance and mean value between each pair of neighbouring distributions was verified by applying the F-test and the t-test (Davis, 1973), both with a confidence level of 0.05. When the equality hypothesis of variance or mean value was rejected, the two distributions were considered to correspond to a distinct conductivity layer. The depth interval between 138 and 152 meters was also considered as a distinct thermal conductivity layer.

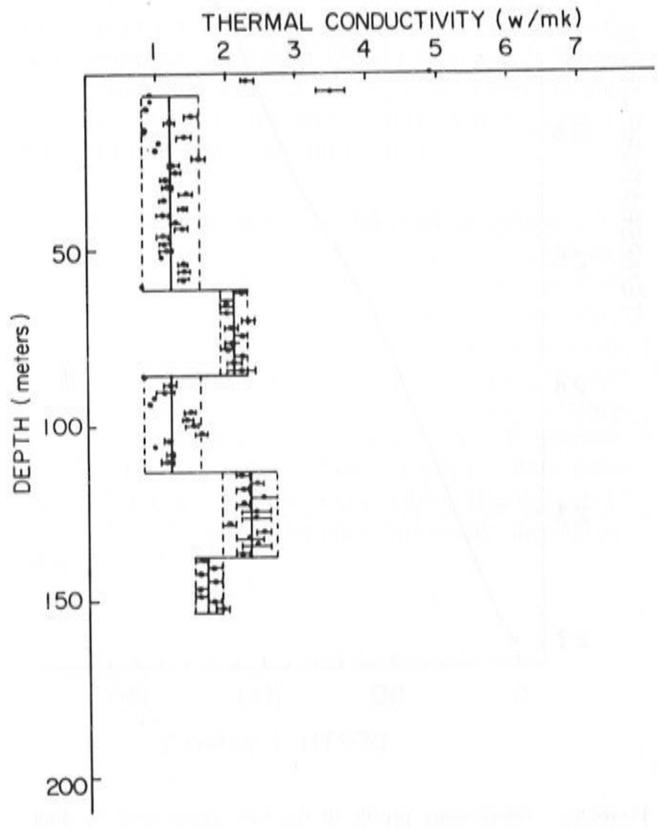

Figure 12— Thermal conductivity profile. Dots represent the measured data. Continuous line represents the mean conductivity of each layer and the interrupted lines limit a variation of 2 τ around the mean.

Table 1 — Thermal conductivities and layer thicknesses of the thermal conductive profile represented in Figure 12.

Layer	Depth interval (m)	Layer thickness (m)	Thermal conductivity (W/mk)
1	0 – 2	2.0 ± 0.2	4.9 ± 0.3
2	2 – 4	2.0 ± 0.2	2.3 ± 0.3
3	4 – 6	2.0 ± 0.2	3.5 ± 0.2
4	6 – 62	56 ± 2	1.2 ± 0.2
5	62 – 66	24 ± 2	2.1 ± 0.1
6	66 – 112	26 ± 3	1.2 ± 0.2
7	112 – 138	26 ± 3	2.3 ± 0.2
8	138 – 152	14 ± 2	1.7 ± 0.1

The thermal conductivity profile constructed in the way described above has a reasonable correlation with the lithologic profile furnished by the driller. The sampling process, however, does not permit a great resolution of the stratigraphic levels, since the drilling cuttings are mixed to some extend by the drilling mud circulation.

The thermal conductivity layer between 62 and 86 meters corresponds, in the lithologic description, to a sequence of sandstones and siltstones with a thin clay layer between 70 and 72 meters. The presence of the clay layer cannot be seen in the conductivity profile. The same situation occurs in the last conductivity layer. In the depth interval corresponding to this layer, 138 to 150 meters, the lithologic profile describes a first layer with fine to medium grain size sandstone with some amount of clay. A sandstone layer between 144 and 148 meters appeared in a quite clear manner in the SP and electrical resistance (single point electrode) logs made in this well. The bottom part of this section was described as green clay. None of these lithologic variations are cleary shown in the conductivity profile.

The total thermal conductivity was calculated by equation (2) and its standard deviation (ΔR) was calculated by

$$\Delta R^2 = \left(\sum_{i=1}^{N} \frac{\partial R}{\partial l_i} dl_i + \sum_{i=1}^{N} \frac{\partial R}{\partial K_i} dK_i \right)^2 \qquad (12)$$

with the constraint that

$$dl = \sum_{i=1}^{N} dl_i \qquad (13)$$

where dl is the error in the well depth, dl_i the error in the thickness of the layers, and dK_i the standard deviations of the corresponding thermal conductivities. Since the depth to a layer boundary corresponds to the depth where a conductivity change occurs and since the sampling is made continuously for depth intervals of two meters, the error in depth was assumed to be half of this interval. The first three thin layers are exceptions in this rule since for small depths the mixing of the drilling cuttings is much smaller.

The obtained total thermal resistance is (101 ± 12) m^2 K/W and the apparent conductive heat flow density in this well is (60 ± 8) mW/m^2.

The conduction steady state temperature profile can be obtained by integrating the Fourier law

$$q = -k \frac{dT}{dz}$$

where q is the measured heat flow. In Fig. 9 the calculated conductive temperature profile for this well is shown.

CONCLUSIONS

In this paper, a method for the determination of apparent heat flow densities in shallow water wells is presented. The method was developed to permit the determination of heat flow densities in wells with substancial hydrologic disturbances, which make it difficult the use of complete temperature profiles.

In this method, temperatures are measured only at the well bottom and a detailed determination of the thermal conductivity profile is made. When the temperature measurements are made with the well out of thermal equilibrium, the rock temperature needs to be extrapolated from the measured temperatures. However, since the measurements are made at the well bottom, the extrapolation models based on the solution of the heat transfer equation with infinite cylindrical symmetry must be modified to take into consideration the presence of the well bottom. A very simple model is presented in this paper.

There are some aspects in this method that call for further investigation. The removal of the influence of surface temperature variations on the heat flow density determination is an important point. On the other hand, the thermal conductivities are measured on drilling cuttings and, for this reason, they may not represent the real conductivities at depth. Rock porosity and saturation for instance, represent rock characteristics that are altered by drilling. The combination of thermal conductivity measurements made on drilling cuttings, and the porosity and saturation information obtained from well logging may lead to a better determination of the thermal conductivity profile.

The author is indebted to Dr. Valiya M. Hamza for suggesting the theme of this work and for several usefull discussions on it. The author is also indebted to Miss Julia Paes Leme for the carefull thermal conductivity measurements. Eduardo Becker and Pedro Viana dos Santos from IPT are thanked for their help in obtaining the well data.

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