Kendall (1985) state that the poor results are due to poor data quality and they maintain that good data give good results, although they do not clarify what is good in either context.

The philosophy of the approach I have adopted is described below, but most of the mathematics of the inversion procedure is given in Appendix 4. I assume that the subsidence and sedimentation histories and all relevant physical properties (thermal conductivity, porosity, heat production) are well known and that steady state thermal conditions exist at all times. As I will make all the same assumptions for forward and inverse modelling, any inconsistency between the two models should reflect the insensitivity of the maturation functions and their calibration to reflectance values.

7.4.2 Method

The relevant steady state energy equation for the sedimentary column is

\[ \frac{d}{dz} \left( k(z) \frac{dT}{dz} \right) + A(z) = 0 \]  \hspace{1cm} (7.4.1)

where \( k(z) \) and \( A(z) \) are the thermal conductivity and internal heat production of the sediments as a function of depth. The lower boundary condition at \( z_b \) is the basal heat flow, \( Q_b(t) \) (positive upwards) and the upper boundary condition at \( z = 0 \) is the surface temperature, \( T_s \). The temperature profile is

\[ T(z',t) = T_s + \int_{0}^{z'} \frac{1}{k(z)} Q_z(t) \, dz \]  \hspace{1cm} (7.4.2a)

with

\[ Q_z(t) = Q_b(t) + \int_{z'}^{z_b} A(z) \, dz \]  \hspace{1cm} (7.4.2b)

This is most easily solved numerically by assuming that the physical properties are constant between grid points \( z_{n-1} \) and \( z_n \) and for \( N \) grid points, with \( T(0,t) = T_s \) and \( z_N = z_b \).
\[ T(z_{n+1}) = T(z_{n-1}) + \frac{z_n - z_{n-1}}{k_n} \left[ Q_b(t) + \sum_{i=n}^{N} \xi_i A_i (z_i - z_{i-1}) \right] \] (7.4.3)

where \( \xi = 0.5 \) if \( i = n \), otherwise \( \xi = 1 \). The maturation function, eqn. 7.3.1, can be written as

\[ M = \int_{t_0}^{t_1} a(\alpha F(Q(t)) - \beta) \, dt \]

where the temperature, \( F(Q(t)) \), is a function of the heat flow and the burial histories. Therefore, given a discrete set of such functions or observations, \( M_i \), we want to find the best model of the heat flow function \( Q_b(t) \) such that the original data can be adequately predicted. This problem can be written in vector form as

\[ M = GQ \] (7.4.4a)

where \( G \) can be regarded as a response function or filter that maps \( Q \) into \( M \). \( Q \) is a \( (M \times 1) \) vector of heat flow parameters and \( M \) is a \( (N \times 1) \) vector of maturation indices, converted from the observable vitrinite reflectance values with one of the relationships given earlier. To solve for \( Q \) we simply write

\[ Q = G^{-1} M \] (7.4.4b)

In order to estimate the goodness of fit of between the observed and calculated values, it is necessary to define a misfit function, \( E = f(M^\text{obs}_i - M^\text{calc}_i) \) and it is the aim of the inversion procedure to minimize this objective function. The standard least squares method minimizes the \( L_2 \) norm, or more simply the length of the misfit vector, so that

\[ E = \sum_{i=1}^{N} (M^\text{obs}_i - M^\text{calc}_i)^2 = \sum_{i=1}^{N} \delta M_i^2 \] (7.4.5)

The root mean square error (RMS) is defined as \( \sqrt{E} \). A discrete inversion approach for \( M \) model parameters of the type
\[ Q_b(t) = \sum_{j=1}^{M} Q_j \]  

(7.4.6)

suggests two simple forms for the heat flow function. The first is in terms of block functions where the heat flow is constant over discrete time intervals, where \( t_m \) is the present day, and \( t_0 \) is the beginning of deposition (\( t_1 \) and \( t_0 \) in eqns. 7.2.1-3),

\[
Q_b(t) = \begin{cases} 
Q_j & t_j \leq t < t_{j-1}, \ j=1,M \\
0 & \text{otherwise}
\end{cases}
\]  

(7.4.7a)

Another reasonable parameterization is a polynomial in time,

\[
Q_b(t) = Q_1 + Q_2 t + \ldots + Q_M t^{M-1}
\]  

(7.4.7b)

and the linear function used by Lerche et al. (1984) is a special case of this form. The aim of the inversion is to obtain the best estimates for the parameters \( Q_j, j=1,M \). Here we adopt the block function form as this allows a more intuitive interpretation of the results as a function of time. This form is also more suited to model the type of heat flow history predicted from many basin formation models, usually in the form of an initially elevated heat flow gradually decaying over time (e.g. fig. 2.3.4).

The least squares approach requires the derivative of each maturation function, \( M_i, i=1,N \), with respect to the model parameters. As \( a, \alpha, \beta \) and the limits of the integral in equation are independent of \( Q_j \) then the first derivative is given by

\[
\frac{\partial M_i}{\partial Q_j} = \int_{t_{j-1}}^{t_j} \frac{\partial F(Q)}{\partial Q_j} a(\alpha F(Q(t)) \cdot \beta) \alpha \log a \ dt
\]  

(7.4.8a)

and the second derivative is

\[
\frac{\partial^2 M_i}{\partial Q_j \partial Q_k} = \int_{t_{j-1}}^{t_j} \frac{\partial F(Q)}{\partial Q_j} \frac{\partial F(Q)}{\partial Q_k} a(\alpha F(Q(t)) \cdot \beta) (\alpha \log a)^2 \ dt
\]  

(7.4.8b)
From 7.3.5a,b

\[
\frac{\partial F(Q)}{\partial Q_j} = \int_0^{z^*} \frac{1}{k(z)} \, dz \tag{7.4.9a}
\]

and

\[
\frac{\partial M_i}{\partial Q_j} \frac{M_j}{\partial Q_k} = 0 \quad j \neq k \tag{7.4.9b}
\]

\[= \left( \frac{\partial M_i}{\partial Q_j} \right)^2 \quad \text{otherwise}
\]

It can be seen from eqn. 7.4.9b that the second derivative matrix is diagonal. In practice, the data misfit (eqn. 7.4.5) is calculated after the maturation functions have been converted to reflectance values, for each of the three functions given in table 7.3.1. I have not included a data covariance matrix in the inversion procedure because we wish to examine the results obtained with synthetic data.

The forward problem is solved by evaluating each of the three maturation integrals (table 7.3.1) for a specified heat flow and depositional history, including compaction, over 250 m.y. at 1 m.y. intervals. Ages were specified for the deposition of sediment containing a piece of vitrinite, and these points were tracked through their burial histories. The temperatures were calculated using eqn. 7.4.3 with \( T_s = 20^\circ \text{C} \) and the maturation integrals were calculated numerically. To illustrate the method I used 4 simple heat flow histories, two with constant values,

(i) \( Q(t) = 50 \, \text{mW m}^{-2} \)

(ii) \( Q(t) = 100 \, \text{mW m}^{-2} \)

and two linear relationships, where \( t \) is age (Ma) and \( t = 0 \) is the present day,

(iii) \( Q(t) = 50 + \frac{(250 - t)}{5} \, \text{mW m}^{-2} \)

(iv) \( Q(t) = 100 - \frac{(250 - t)}{5} \, \text{mW m}^{-2} \)
The final values of the maturation integrals for the specified sediment ages were used as the input data for the iterative inversion procedure, as described in Appendix 4, to obtain a set of M model parameters \( Q_j, j=1, M \).

7.4.c Modelling approach and presentation of results

I found the inversion procedure was more stable using an initial guess for the solution lower than the expected result and I generally used 40 mWm\(^{-2}\), but I will discuss the significance of the initial model later. Unless stated otherwise, the inversion procedure was continued for 40 iterations, and by this stage all the solutions had RMS of less than 0.02 % reflectance.

The results for the four heat flow histories (i)-(iv) given in the last section are summarized graphically in the following form in Appendix 5 (figs. A5.i-iv), and see fig. 7.4.1 for immediate reference): The reflectance/depth profiles were obtained by converting the time-temperature integrals using the relationships given in table 7.3.1. The assumed heat flow model gives a reflectance profile which is shown as a solid line and the reflectance values obtained from the inversion derived heat flow models are plotted as '+' for each data point. The inversion derived heat flows are then given separately for each of the three maturation functions, together with the assumed model (solid line) and the initial starting model (dashed line). The age of each reflectance value, or really the age of the sediment in which the organic material would be found is shown by a 'x' at the top of the uppermost heat flow model. Obviously, the oldest 'x' corresponds to the deepest '+' on the reflectance/depth profiles. I used sediment thicknesses of 5 and 2 km and incorporated the physical properties of shale (section 5.2.b, Appendix 1). The maximum temperatures at the base of the sediments were approximately 150°C for 5 km and 80°C for 2 km with 50 mWm\(^{-2}\) and 280°C for 5 km and 140°C for 2 km with 100 mWm\(^{-2}\). For a given heat flow
Fig. 7.4.1 An example of the inversion results presented in appendix 5. The reflectance values for each of the three maturation functions (bottom panel) are shown as a solid line for the assumed heat flow model and as 'x' for the inversion generated model. The upper three panels show the assumed heat flow (solid line), starting model (dashed line) and the inversion generated result (step functions) for each maturation function. The age of each reflectance value is marked by a 'x' at the top of the upper panel.
history the results will be referred to as Ro2 and Ro5 for the Royden et al. (1980) integral with 2 or 5 km of sediment. Similarly, Wa2 and Wa5 refer to the Waples (1980) integral and Ri2 and Ri5 to the Ritter (1984) integral.

7.4.4 Results and discussion

What is apparent from the results of the inversion procedure is that the maturation integrals are most sensitive to the recent heat flow history and with this method, results can be obtained which adequately reproduce the reflectance values but the resolution of the early heat flow history is poor. The most consistent illustration of this is where the heat flow increases to the present day, e.g. fig. A5.iii.b, where 5 km of sediment was deposited. In these models the data fit is seen to be very good but the model heat flow is nearly constant or actually increases back in time. The apparently reasonable agreement of the earliest model parameter in Ro5(iii) and Ri5(iii) is merely due to the fact that I set any heat flow parameter which became greater than 300 mWm$^{-2}$, or negative, back to its initial value of 40 mWm$^{-2}$. However, if I had kept these high values the effect on the data fit is insignificant. The results shown in fig. A5.iii.a, where only 2 km of sediment was deposited, are more satisfactory in that they approximate the true solution reasonably well, except perhaps for Wa2(iii). This improved model resolution is because of the power law form of the maturation integrals. The shallower sediments do not reach temperatures high enough to totally reduce the sensitivity of the functions to the early heat flow history.

Generally then, it appears that the more recent heat flow parameters are reasonably well resolved, although some models do show the effect of the tradeoff between temperature and time e.g. Ri2(ii), Wa2(ii) (fig. A5.ii.a), Wa2(iv) (fig. A5.iv.a), Ri5(iv), Wa5(iv), Ro5(iv) (fig. A5.iv.b). These models have the most recent heat flow parameter either too high or too low and consequently, the next two or three parameters show
departures from the true model in the opposite sense to compensate for this. It is not surprising that such results are obtained. The recent heat flow history is constrained by a greater number of observations or data, and then is well resolved relative to the early period, which is only constrained by the older or presently deeper data. As the older data points become more deeply buried they experience progressively higher temperatures and, with the power law dependence of the maturation functions on temperature, this effectively swamps the low temperature, or shallow burial, part of the maturation integrals, restricting the potential for the resolution of the earlier heat flow parameters. This is a fundamental drawback to the use of such integrated time-temperature functions where one value represents the total thermal history. The non-uniqueness will become more acute when long (50-100 m.y.) periods of non-deposition occur as the constraints on the temperature history of a given data point afforded by the burial history are no longer available.

Some of the inversion results show good agreement with the true models, even when the heat flow has increased to the present day e.g. Ro2(i-iv), Ro5(i), Ri2(i-ii), Ri5(ii, iv). However, as there are no errors associated with the reflectance values it is expected that provided the inversion algorithm continues to minimize the objective function, that it is only a question of the number of iterations made before a 'good' answer is obtained. This number will of course vary according to the maturation integral, the initial model and the minimization path. Consequently the results mentioned above may be telling us little about the potential for actually resolving the model. The fact that none of the Waples models have reached equally satisfactory solutions, while still providing very good fits to the data, suggests this is likely to be the case.

To illustrate the point I used a starting model of 100 mWm\(^{-2}\) for the heat flow models (ii) and (iv) with 2 km of sediment and the Royden et al. (1980) maturation function. For model (ii) (fig. 7.4.2) a good result was obtained after the 40 iterations but as can be seen a reasonable data fit is achieved after 25 iterations although with a spurious model which would suggest that the heat flow had been nearly a factor of two higher in the past. The error in the present day heat flow is less than 5 mWm\(^{-2}\) and the maximum misfit
Fig. 7.4.2  Inversion results for the Royden maturation integral with $Q(t)$ (solid line) = 50 mWm$^{-2}$ for 2 km of sediment deposited over 250 m.y.. For the purposes of illustration, the reflectances for the middle and upper models have been successively offset by 0.1%. The starting model is shown by the dashed line.

Upper panel starting model $Q = 40$ mWm$^{-2}$, run for 40 iterations
Middle panel starting model $Q = 100$ mWm$^{-2}$, run for 40 iterations
Lower panel starting model $Q = 100$ mWm$^{-2}$, run for 25 iterations
in the predicted and 'observed' data is less than 0.005% reflectance. It is most unlikely that either of these would be resolvable with real data. For model (iv) it can be seen (fig. 7.4.3) that once again a satisfactory heat flow model is reached after 40 iterations (data misfit RMS = 0.0073) with and initial model of 100 mWm\(^{-2}\) but after 20 iterations the data is adequately predicted (RMS = 0.013), although the model heat flow is not. A similar result is obtained for the starting model of 40 mWm\(^{-2}\) after 9 iterations (RMS=0.011). It is obvious then that even with perfect data any conclusions regarding the early heat flow will be dependent on the inferred data resolution, or for the inversion procedure, the specified convergence criterion. With real data and associated errors, the resolution is expected to be very poor.

As a final example, I used a heat flow history more relevant to basin formation mechanisms. This form increases linearly from 50 to 100 mWm\(^{-2}\) over the first 50 Ma and then decays exponentially, with a time constant of 62.8 m.y. (e.g. McKenzie, 1978) towards a final value of 40 mWm\(^{-2}\). I also used two sedimentation histories such that 2 km of the final sediment thickness (5 km) was deposited linearly over the first 50 Ma and the rest was deposited either linearly or exponentially with a time constant of 62.8 m.y.. The distribution of data points over time can be seen at the top of the uppermost heat flow model in figs. 7.4.4 and 7.4.5. The results can be gauged in the light of the comments made above concerning the convergence criterion, but as can be seen after 40 iterations the models are not well resolved, except perhaps for the suggestion that heat flow was higher at some time in the past. Inversion of the Ritter maturation function has achieved the best representation of the true model but still shows a lack of sensitivity to the early heat flow.

In fig. 7.4.6 I show the reflectance profiles generated using the exponential heat flow function described above and the Ritter maturation function, both with and without internal heat production in the sediments (totaling about 10 mWm\(^{-2}\) in this case) and that obtained from a constant heat flow of 55 mWm\(^{-2}\) with internal heat production. The latter model predicts reflectance values as close or closer to the true data than the model where heat production has been neglected and similar discrepancies may arise when the thermal conductivity is not corrected for temperature. The influence of both of these factors is to
Fig. 7.4.3 Inversion results for the Royden maturation integral with \( Q(t) \) (solid line) increasing linearly from 50 to 100 mWm\(^{-2} \) for 2 km of sediment deposited over 250 m.y. For the purposes of illustration the reflectances for the middle and upper models have been successively offset by 0.1%. The starting models are shown by a dashed line.

Upper panel starting model \( Q = 100 \) mWm\(^{-2} \), run for 40 iterations
Middle panel starting model \( Q = 100 \) mWm\(^{-2} \), run for 20 iterations
Lower panel starting model \( Q = 40 \) mWm\(^{-2} \), run for 9 iterations
Fig. 7.4.4: Results of inversions with $Q(t)$ (solid line) linear from 50 to 100 mWm$^{-2}$ for 50 m.y., then decaying exponentially to 40 mWm$^{-2}$, with 5 km of sediment linearly over 250 m.y. (see $x$ at the top of the upper panel for the age of the reflectance samples). All have the same starting model (dashed line) and were run for 40 iterations.
Fig. 7.4.5 Results of inversions with $Q(t)$ (solid line) linear from 50 to 100 mWm$^{-2}$ for 50 m.y., then decaying exponentially to 40 mWm$^{-2}$, with a total of 2 km of sediment, deposited linearly for 50 m.y. and then exponentially for the next 200 m.y. (see x at the top of the upper panel for the age of the reflectance samples). All have the same starting model (dashed line) and were run for 40 iterations.
Fig. 7.4.6. A comparison of the predicted reflectances for the exponentially decaying heat flow as shown in figs. 7.4.4 and 7.4.5 with (solid boxes) and without (open boxes) internal heat production included in the sediments and also for a constant heat flow of 55 mWm$^{-2}$ with internal heat production (triangles) The upper panel is for exponential sedimentation from 200 Ma and the lower panel is for linear sedimentation.
increase the temperature at a given depth and it is clear that they should be considered in thermal modelling studies if only to quantify the magnitude of uncertainties associated with their subsequent neglect.

A brief concluding comment may be made regarding the inversion procedure. It is clear from some of the results that there is a degree of tradeoff between the parameters. This is a common feature of the discrete block function parameterization used, analogous to the simple layered velocity or resistivity models used in seismic refraction or electromagnetic inversions. An approach to overcome this type of problem has been suggested by Constable et al. (1987). They propose that the preferred model should be as smooth as possible. To this end, the inversion strategy is based on the minimization of a regularization function, which is simply a measure of the roughness of the model, this being defined as a function of the first or second derivative of the model parameters with respect to the time for the heat flow problem. The minimization of this function is subject to the constraint that the data misfit is also minimized. The method is straightforward for a 1-D problem. The approach would not alleviate the lack on sensitivity of the inversions to the early heat flow history, but would reduce the tradeoff between the more well constrained parameters. However, as I will discuss in the next section, the heat flow in the Eromanga Basin appears to have been elevated by about 20-25% in the last few million years, and in this situation a smooth model might not be warranted. The problem is also aggravated in the Eromanga Basin by the fact that sedimentation effectively ceased about 90 Ma, and as I mentioned earlier, this means that the constraints on the thermal evolution provided by the burial history are not available over this period.

7.5 Concluding remarks

Groundwater flow can have a fundamental role in the heat balance of a sedimentary basin and it is prudent to be aware of this when interpreting temperatures or heat flow estimates. The latter will be the more sensitive as the relative change in the
temperature gradient will be greater than the relative change in the temperature itself. The central region of the Eromanga Basin, where the major Jurassic aquifers are at depths between 1500 and 2300 m, can be considered to be in a state of conductive equilibrium, although I will expand on this point later.

Permeability is a major control on both convective and advective heat transfer, yet it is a difficult parameter to quantify. As I mentioned in section 7.2.a, measured permeabilities in the Eromanga Basin vary by several orders of magnitude, and the problem is to obtain a representative value even for a single formation. The presence of thin impermeable layers (shale, siltstone) within an aquifer will tend to inhibit free convection, but forced convection moving though a relatively permeable bed will be less affected by local obstructions. Therefore, it is probable that for most sedimentary basins, where the geothermal gradients are less than 60°C/km, advective heat transport by groundwater flow will be more important than thermally induced convection.

The thermal blanketing effect of sedimentation can cause significant departures from steady state equilibrium if sedimentation rates are high (e.g. over 100 m/m.y.). Once deposition has ceased, however, the geotherm relaxes towards a steady state equilibrium value. After the deposition of 5 km of sediment, the basin should be within about 10% of steady state after 40 m.y. or so, even for sedimentation rates of 1000 m/m.y.. Therefore the assumption of steady state during the evolution of a sedimentary basin will generally be a reasonable first order approximation for sedimentation rates of less than about 100 m/m.y. and where less than 5 km of sediment has been deposited. In these situations it is unlikely that the relevant thermal parameters could be constrained to the required accuracy to resolve the difference between a full transient solution and that obtained under the assumption of steady state.

Although fluid flow and sedimentation can perturb the steady state conductive thermal regime, overall these effects will be relatively unimportant in intracratonic basins, where the horizontal dimension is usually 3 orders of magnitude greater than the depth and
sedimentation rates are usually less than 50 m/m.y. Where fluid flow is dominantly vertical, such as may occur at the basin margins or along faults, then the local geotherm may be enhanced (for upward flow) or reduced (for downward flow). I address the question of horizontal flow over a localized heat source in more detail in section 8.3.f.

It has been shown that reflectance measurements, even when error free, will be of little use in constraining basin formation mechanisms which tend to predict a decreasing heat flow with time. The major problem is that the temperature dependence of the maturation functions results in an obliteration of the early history. Although methods such as fission track analysis are potentially more informative than reflectance measurements, I suspect that the interpretation of data would be subject to similar problems regarding the early temperature history as temperatures may be reached where the tracks will totally anneal on geological timescales. A combination of methods with a range of temperature sensitivity applied to the subsiding basin region and also to adjacent uplifted (?)source) regions, is likely to prove most satisfactory. Additional problems will arise if thermal parameters such as thermal conductivity and heat production are not reliably assessed, although it may be argued that relative variations in thermal histories could be inferred from reflectance values if it can be demonstrated that such thermal properties are reasonably constant within or between sedimentary basins. This assumption may be based on the similarity of lithological types and tectonic setting, for example.
8. THE THERMAL STATE OF THE CENTRAL EROMANGA BASIN REGION

8.1 Introduction

It was recognized early this century that artesian water from boreholes in the Eromanga Basin could flow to the surface at temperatures of up to 100°C (Gregory, 1906). Since then it has been demonstrated that this is due not to a primary igneous source for the water, as originally proposed by Gregory, but by heating of meteoric groundwater as it flows into an area with high geothermal gradients. Some of the high temperature gradients observed in shallow (<1000 m) artesian wells may be the result of vertical groundwater flow along faults (Habermehl, 1980) or simply attributable to hot water flowing up the borehole to the point of measurement. Over the last twenty years or so, exploration drilling in the Cooper Basin has shown the area to be of significant economic importance due to the presence of hydrocarbons. Temperature measurements made at depths of two to four kilometres during drilling suggest that the geothermal gradients are considerably higher than the commonly assumed shallow geotherm of 25°C/km appropriate for Palaeozoic terrains, and are between 30 and 60°C/km (Polak and Horsfall, 1977; Middleton, 1979a; Pitt, 1982, 1986; Kantsler et al., 1983, 1986; Cull and Conley, 1983; Piper, 1986).

8.2 Temperatures, thermal conductivity and heat flow

8.2.a Temperature observations and geothermal gradients

Temperature measurements made during exploration drilling should be regarded with suspicion as a matter of course. Two types of temperature measurement are considered in this study - those made during wireline logging runs and drill stem tests (DST).
8.2

Bottom hole temperatures (BHT) are recorded with a maximum temperature thermometer during successive logging runs but these are perturbed due to the presence of drilling mud which is generally not in equilibrium with the true formation temperature. The effect of mud circulation may be corrected for if the time since the cessation of drilling and/or mud circulation is known (e.g. Fertl and Wichmann, 1977; Middleton 1979b, 1982c; Leblanc et al., 1981; Lee, 1982; Catala, 1984). The drilling mud will have a cooling effect at depth and as such, the raw observations may be regarded as lower limits for the true formation temperature. A simple practical approach to obtain an estimate of the steady state temperature, when two or more logging temperatures are known, is to extrapolate from observed temperatures on a plot of temperature against the log of time. This has some physical basis, given the generally exponential time dependence for the decay of a thermal perturbation. Pitt (1986) suggests that this method will give values within 2% of estimates made with those which incorporate the correct equations of energy conservation. In the absence of information regarding the time since drilling ceased, Andrews-Speed et al. (1984) used an arbitrary correction of $0.15(T_z-T_s)$ which was added to the observed temperature, $T_z$, observed at depth $z$, $T_s$ being the assumed surface temperature. Although this correction is not quantitatively justified, it is likely that the raw data will be adjusted in the correct sense as the difference between the log derived temperatures and the true formation temperatures would be expected to be greater for higher formation temperatures. As a result of the correction the temperature gradient will be increased. The error associated with this approach is quoted as about 15% by Andrews-Speed et al. (1984) although the value will depend on the time since mud circulation ceased. Other secondary factors such as the borehole diameter will affect the magnitude of the difference between the true formation temperature and that measured during logging.

Drill stem testing involves mounting a tool at the end of the drill pipe and isolating the sedimentary section to be tested by inflatable packers. Formation fluids flow out in to the borehole and up the drill pipe. Subsequently, the tool is shut and the formation pressure is allowed to build up back to more or less the original value. The primary objectives of the
test are to obtain continuous measurements of fluid flow rate (during the flowing period) and formation pressure (during the shut-in period). Temperature measurements are commonly made during the testing. DST temperature data are generally thought to be reliable estimates of the true formation temperature provided sufficient formation fluid flows into the tool and the correct equilibrium temperature is obtained. The definition of sufficient fluid flow is vague and depends on the type of fluid. The data set that I have used was obtained from Delhi Petroleum (Piper, 1986) and corrections had been made to DST measurements when less than the equivalent of about $1 \times 10^6$ cubic feet/day ($\sim 3.3 \times 10^{-7}$ m$^3$ s$^{-1}$) of gas or 100 barrels of oil/day (1 barrel $\sim$ 160 litres, 0.19 m$^3$ s$^{-1}$) was measured during the test.

Perrier and Raiga-Clemenceau (1984) qualitatively discuss some of the factors influencing the temperature of flowing fluids during DST. They obtained continuous temperature recordings over a period of nearly 21 hours and suggested that the temperature generally increases during the flow period (10 hours) as a result of the inflow of formation fluid which has a higher temperature than the drilling mud and also frictional heating due to flow through the permeable formation. Frictional heating is interpreted as the heating associated with viscous dissipation, and is equal to the mechanical power needed to force the fluid through the permeable medium (Bejan, 1984). The amount of heat per unit time per unit volume is given as

$$Q_v = \frac{\mu}{K} (u)^2$$  \hspace{1cm} (8.2.1a)

where $\mu$ is the dynamic viscosity (Pa s), $K$ is the permeability (m$^2$) and $u$ is the average velocity vector, which from eqn. 7.2.2, is given by

$$u = - \frac{K}{\mu} \left[ \nabla P - \rho_0 (1 - \zeta(T)) g \right]$$  \hspace{1cm} (8.2.1b)

The temperature anomaly associated with this heat (for a unit volume) is given by

$$\Delta T = \frac{Q_v}{\rho_e c_e}$$  \hspace{1cm} (8.2.1c)
Perrier and Raiga-Clemenceau (1984) do not describe the pressure, flow rate or fluid volume measurements but they attribute a temperature increase of about 3°C to frictional heating, equivalent to $Q_v$ of $3.5 \times 10^2$ Wm$^{-3}$ for water ($\rho_f = 1000$ kg m$^{-3}$, $c_f = 4.2 \times 10^3$ J kg$^{-1}$ K$^{-1}$). Note that this value is heat required to linearly increase the temperature in one cubic metre of water by 3°C over 10 hours. Perrier and Raiga-Clemenceau (1984) witnessed an increase in temperature of about 4°C over 5 minutes or so immediately after the tool valve was shut which they attributed to adiabatic compressional heating. During the shut-in period with no flow (11 hours), the temperature gradually decreased towards the equilibrium value of the formation fluid, presumably because the system was no longer adiabatic. Therefore, the overall maximum temperatures obtained during DST are too high (~7°C or about 5%) in the example of Perrier and Raiga-Clemenceau. Of course, it would be informative to know during which period of the DST the recording was made - flowing or shut-in. However the data relevant to the time of temperature recording is generally not available. Perrier and Raiga-Clemenceau (1984) compared logging derived and DST derived temperatures and concluded that the correctly extrapolated DST data, if available, should be the most reliable estimate of the true formation temperature and even corrected logging temperatures may be too low. Other factors affecting the accuracy of the DST temperatures would be expansion and adiabatic cooling of the fluid (especially for gas), the amount of drilling mud present in the tool and the time since drilling ceased before the test. The temperature will also depend on where it was measured - at the formation being tested or further up the drill pipe, although it is usually the former. Corrections similar to those applied to logging temperatures may be used when it is suspected that the measured temperature is reflecting that of the drilling mud, rather than the true formation fluid, as would occur if the fluid flow was insufficient. This is probably the most common source of error associated with DST temperature data.

Fig. 8.2.1 illustrates the distribution of geothermal gradients in the Cooper and Simpson Desert Basin areas. These values are based on corrected logging and DST data taken from Piper (1986) and calculated with a surface temperature of 20°C. The corrections
Figure 8.2.1 Geothermal gradients (°C/km) estimated from BHT and DST data from the central Eromanga Basin region. The larger unfilled symbols are averages plotted where the well density prevented illustration. The consistently highest gradients (>50°C/km) occur in or near the Nappamerri Trough (Burley, Moomba, Strzelecki) and along its northern margin (Jackson), whereas the Patchawarra Trough (Tirrawarra) and the Simpson Desert Basin have lower (<45°C/km) gradients. Gradients to the northeast of this area in Queensland are between 50 and 45°C/km (Pitt 1982, 1986).
were made by Piper (1986) and involved estimating the equilibrium temperature by extrapolating from logging temperatures made at different times at the same depth in the same well. Where the observed DST data were considered unreliable as a result of insufficient flow, a similar correction was made by Piper by assuming that the time-temperature relationship deduced from logging runs was applicable to the DST data at similar depths in the same well. This of course requires some knowledge of the time since mud circulation ceased.

In fig. 8.2.2 I reproduce a map of the basement lithologies intersected by wells in the southern Cooper Basin area. In this region the highest gradients (50-55°C/km) are often associated with granite basement, for example Moomba-Big Lake area, Burley-Mcleod wells in the Nappamerri Trough. High gradients are also observed along the basement highs which form the northern and eastern margins of the Nappamerri Trough. These are made up of Palaeozoic/Proterozoic sediments and metasediments and the gradients are 40-55 °C/km and are highest at the eastern end (e.g. Jackson). The lowest values (about 30-35°C/km) occur in the Patchawarra Trough (e.g. Tirrawarra) where the basement is predominantly carbonates, sandstone or siltstone. In other wells the gradients are between about 35-55°C/km and the basement is typically represented by Palaeozoic or possibly Proterozoic sediments and metasediments. Clearly then, the geothermal gradient is higher than the average 25°C/km observed in Phanerozoic regions. However, before a meaningful interpretation of these observations can be made in terms of heat flow, the thermal conductivity of the sediments needs to be quantified.

8.2.b Thermal conductivity

Generally surface heat flow, $Q_s$, is estimated from Fourier's Law of heat conduction. This expresses heat flow, $Q$, as the product of the temperature gradient and thermal conductivity,
Fig. 8.2.2 Interpreted basement geology of the part of the southern Cooper Basin in South Australia (from Gatehouse, 1986) showing the location of some known granite intrusions in the basement.
\[ Q = -k \frac{dT}{dz} \quad (8.2.2) \]

For the earth, heat flow is defined to be positive upwards so at the surface \( z = 0 \), \( Q_s = -Q \).

To have confidence in any heat flow estimate it is necessary to have reliable control on the downhole thermal conductivity variations. Various models have been suggested for the prediction of the effective, or bulk, thermal conductivity \( (k_b) \) in porous rocks (see Appendix 1). Generally, these models are functions incorporating the fractional porosity, \( \phi \), the thermal conductivities of the solid grains or matrix, \( k_g \), and the fluid or gas phase in the porespace, \( k_f \), and possibly some assumptions concerning the geometry of the solid material and the porespace. No one model has proven unequivocal when compared to observations and often empirical or ad hoc adjustments are required to fit the data. However, most of the models fall within the Hashin-Strickman bounds derived by variational calculus (Hashin and Strickman, 1962) for a two phase material (e.g. Brailsford and Major, 1964; Horai, 1971).

Measurements of thermal conductivity were made on samples from the southern Cooper Basin and a complete description of the method and results is given in Appendix 1. Based on these results, we adopt the arithmetic mean of two formulations, the Maxwell model and the geometric mean model, as a suitable function for the sediments from the southern Cooper basin area. The Maxwell model is equal to the upper Hashin-Strickman limit and the geometric model is close to, but greater than, the lower limit. For these two models, the bulk thermal conductivity is given by:

**Geometric model**

\[ k_b = k_g (1 - \phi) k_f \phi \quad (8.2.3a) \]

**Maxwell model**

\[ k_b = k_g \frac{(2r+1) - 2\phi(r-1)}{(2r+1) + \phi(r-1))} \quad (8.2.3b) \]

where \( r = k_g/k_f \)
and $k_g$ is assumed to be the continuous phase, $k_f$ the dispersed phase (see Beck, 1976) and $\phi$ is the fractional porosity (0-1). I have assigned representative matrix thermal conductivities to the four lithologies, coarse grained sandstone, fine grained sandstone, siltstone and shale (table 8.2.1). The internal heat production values given were based on data from Taylor and McLennan (1985). Some of the formations in the sedimentary sequence contain 10-15% of coal and the matrix conductivity adopted for coal is based on values given by Cěrmak and Rybach (1982). The other parameters for coal (internal heat production, porosity/depth relationship) are taken to be the same as those for shale. The pore fluid is assumed to be water with a thermal conductivity of 0.61 Wm$^{-1}$K$^{-1}$. The porosity/depth relationships are discussed in section 6.2.b.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Matrix thermal conductivity ($k_g$) Wm$^{-1}$K$^{-1}$</th>
<th>Heat production (A) $\mu$Wm$^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse Sandstone</td>
<td>7.0</td>
<td>0.795</td>
</tr>
<tr>
<td>Fine Sandstone</td>
<td>4.5</td>
<td>0.795</td>
</tr>
<tr>
<td>Siltstone</td>
<td>3.5</td>
<td>1.375</td>
</tr>
<tr>
<td>Shale</td>
<td>2.5</td>
<td>2.291</td>
</tr>
<tr>
<td>Coal</td>
<td>0.3</td>
<td>2.291</td>
</tr>
</tbody>
</table>

Table 8.2.1 Matrix thermal conductivity and internal heat production values assigned to the lithologies in the Eromanga/Cooper Basin sediments.

Pressures of up to 1 kbar, apart from their role in compaction, have a negligible effect on thermal conductivity (Lubimova et al., 1977) but the effect of temperature is significant (Birch and Clark, 1940; Lubimova et al., 1977). The empirical temperature correction suggested by Sekiguchi (1984) is used in this study and this is given as

$$k = 358(k_0 - 1.84)(1/T - 0.00068) + 1.84$$  \hspace{1cm} (8.2.4)

$T$ is the absolute temperature (K), and $k_0$ is the thermal conductivity (Wm$^{-1}$K$^{-1}$) at 288 K ($15^\circ$C). This relationship is illustrated in fig. 8.2.3.
Fig. 8.2.3 The empirical temperature correction for thermal conductivity proposed by Sekiguchi (1984) as given in eqn. 8.2.4. Note how the higher thermal conductivities reflect the strong temperature dependence of quartz.
Fig. 8.2.4 shows the calculated downhole porosity (see section 5.2.b), thermal conductivity and temperature profiles for 4 wells in the southern Cooper Basin, together with observed values where available. The relative proportions of each lithology were obtained from downhole lithology logs. The location of these wells can be found in fig. 6.2.1a or Appendix 1, fig. 1. The thermal conductivity and the predicted temperature profiles are shown both with and without the temperature correction. The higher thermal conductivities associated with the quartz rich Jurassic sandstone reservoirs are apparent in fig. 8.2.4 at depths of about 1500-2300 m. The calculation of the temperature profiles will be discussed in the next section.

It should be noted here that errors cited for thermal conductivity measurements are of the order of 10-15%. Measurements made on small samples at the surface are often extrapolated over scales of hundreds of metres and it has been observed that heat flow calculated as a function of depth can be discontinuous implying that the thermal conductivity may be erroneous (Blackwell et al., 1981; Cull et al., 1988). There will also be some bias towards the more consolidated sediments as samples often cannot be obtained for the friable shaly lithologies and in these cases the thermal conductivity will be overestimated. It can be difficult, then, to obtain a reliable heat flow estimate but, as I will discuss later, if it can be assumed that the thermal conductivity is relatively constant for a given area (i.e. similar lithologies), the trend of geothermal gradients will directly reflect the heat flow variations, provided that conductive heat transfer is the dominant process.

8.2.c Thermal resistance and heat flow

Downhole temperatures may be used to estimate the heat flow through the sediments to the surface. The technique is known as the thermal resistance method (Bullard, 1939; Chapman et al., 1984) and is based on the assumption of 1-D steady state heat conduction. Fig. 8.2.5 illustrates how the heat flow may be estimated from downhole temperature measurements in a sedimentary sequence with depth dependent thermal conductivity. In
Fig. 8.2.4a  Observed and estimated porosity (upper panel), thermal conductivity (centre panel) and temperature (lower panel) for the Dullingari 1 and Innamincka 1 wells in the southern Cooper Basin (see fig. 6.2.1a and appendix 1, fig. 1 for the location of the wells). Corrected logging temperatures are given by a solid circle and the uncorrected values are shown by small triangles. The observed thermal conductivity is given as a range from the results given in appendix 1, but the diamond symbols shown on the thermal conductivity profile for Innamincka 1 are taken from Howard (1960). The predicted thermal conductivity and temperature profiles are shown where the thermal conductivity is both uncorrected and corrected for temperature.
Fig. 8.2.4b Observed and estimated porosity (upper panel), thermal conductivity (centre panel) and temperature (lower panel) for Mcleod 1 and Burley 2 wells in the southern Cooper Basin (see fig. 6.2.1a and appendix 1, fig. 1 for the location of the wells). Corrected logging temperatures are given by a solid circle, corrected DST by a solid square and the uncorrected values are shown by small triangles. The observed thermal conductivity is given as a range from the results given in appendix 1. The predicted thermal conductivity and temperature profiles are shown where the thermal conductivity is both uncorrected and corrected for temperature.
Fig. 8.2.5 The thermal resistance method of estimating heat flow (Bullard, 1939). The
method is based on the assumption of 1-D steady state with no internal heat production
and the temperature at a depth $z$ is given as

$$T_z = T_s + Q \sum_{i=1}^{N} \frac{\Delta z_i}{k_i}$$

where $T_s$ is the assumed surface temperature, $k_i$ is the interval thermal conductivity over
$\Delta z_i$ and $Q$ is the heat flow. $Q$ is then the slope of a temperature versus thermal resistance,
$\Sigma(\Delta z/k)$, plot. If only one temperature measurement, $T_z$, is available the method may still
be used (e.g. Chapman et al., 1984). The above equation can then be reformulated as

$$Q = \frac{-k(T_z - T_s)}{Z} \quad \text{with} \quad \frac{Z}{k} = \sum_{i=1}^{N} \frac{\Delta z_i}{k_i}$$

and $\frac{Z}{k}$ is simply the harmonic mean of the thermal conductivity weighted by the interval
thickness, referred to as the depth averaged thermal conductivity.
figs. 8.2.6a,b I show the results of applying the method to the Dullingari 1, Innamincka 1, Burley 2 and Mcleod 1 wells (see fig. 8.2.4). To obtain the best fit to the data, in a least squares sense, I used either a surface temperature \( T_s \) of 20°C or left \( T_s \) as a free parameter. The thermal resistance term, \( \Sigma(\Delta z/k) \), was calculated with the thermal conductivity both corrected and uncorrected for temperature. The preferred values for the heat flow are those with \( T_s = 20°C \) and the thermal conductivity corrected for temperature. The temperature profiles shown in fig. 8.2.4 were calculated with the heat flow estimated from the BHT in each well, \( T_s = 20°C \), the depth averaged thermal conductivity to the base of the well and including internal heat production in the sediments (see table 8.2.1). The downhole temperatures in Innamincka 1 and Dullingari 1 were not included in the Delhi Petroleum database, and so were corrected by using the method suggested by Andrews-Speed et al. (1984), i.e. \( 0.15(T_s - T) \) as no other information was available. The BHT in Innamincka 1 obtained with this method, 155°C, was higher than expected from the other downhole measurements, and also that cited by Pitt (1982) so a value of 151°C, as given by Pitt (1982), was used. Table 8.2.2 summarizes the heat flow estimates described above.

The depth averaged thermal conductivity, \( \bar{K} \), is evaluated as the weighted harmonic mean of the interval thermal conductivities. The relatively small variation in the thermal conductivities in these 4 wells is expected from the overall similarity of lithological types in the sedimentary units. As I mentioned in chapter 6, detailed lithology logs were not always available and in these cases what were considered to be representative values (Delhi Petroleum, pers. comm., 1985) were assigned for the proportions of each lithology in each well. However, thermal conductivity is considerably more sensitive than the backstripping procedure to the assumed lithology and porosity/depth function, especially for higher values of the matrix conductivity (quartz rich sediments). Using the representative lithology proportions in every well, estimates of the depth averaged thermal conductivity for a given well are plotted versus the total sediment thickness for each of the 40 wells used in the backstripping (chapter 6) are given in fig. 8.2.7.
Fig. 8.2.6a Estimates of heat flow made for Dullingari 1 and Innamincka 1 from the southern Cooper Basin. The thermal resistance is estimated both with and without temperature corrections applied to the thermal conductivity (empty and solid squares respectively). The least squares regression was made with the surface temperature ($T_s$) constrained to be 20°C and also left as a free parameter (See table 8.2.2). Both these wells have metasedimentary basement.
Fig. 8.2.6b As fig. 8.2.6a but for Mcleod 1 and Burley 2 from the Nappamerrri Trough, southern Cooper Basin. Both of these wells have granitic basement.
<table>
<thead>
<tr>
<th>Well</th>
<th>BHT</th>
<th>A</th>
<th>$\bar{k}$</th>
<th>Q (BHT)</th>
<th>$T_S$</th>
<th>Q</th>
<th>$T_S$</th>
<th>Q</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>°C</td>
<td>μWm$^{-3}$</td>
<td>Wm$^{-1}$K$^{-1}$</td>
<td>mWm$^{-2}$</td>
<td>°C</td>
<td>mWm$^{-2}$</td>
<td>°C</td>
<td>mWm$^{-2}$</td>
</tr>
<tr>
<td>Mcleod 1</td>
<td>227</td>
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<td>2.50</td>
<td>129</td>
<td>20</td>
<td>133</td>
<td>-16</td>
<td>161</td>
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<td></td>
<td></td>
<td>2.14</td>
<td>112</td>
<td></td>
<td>20</td>
<td>114</td>
<td>-8</td>
<td>132</td>
</tr>
<tr>
<td>Burley 2</td>
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<td>142</td>
<td>20</td>
<td>144</td>
<td>-2</td>
<td>163</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.14</td>
<td>119</td>
<td></td>
<td>20</td>
<td>123</td>
<td>12</td>
<td>129</td>
</tr>
<tr>
<td>Dullingari 1</td>
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<td>105</td>
<td>20</td>
<td>113</td>
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<td>104</td>
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<td>84</td>
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<tr>
<td>Innamincka 1</td>
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<td>2.47</td>
<td>79</td>
<td>20</td>
<td>84</td>
<td>0</td>
<td>99</td>
</tr>
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<td></td>
<td></td>
<td>2.16</td>
<td>69</td>
<td></td>
<td>20</td>
<td>75</td>
<td>12</td>
<td>80</td>
</tr>
</tbody>
</table>

Table 8.2.2 Estimates of heat flow for 4 wells from the southern Cooper Basin area, including heat production (A) in the sediments. The upper line for each well are the values without the depth averaged thermal conductivity ($\bar{k}$) being corrected for temperature and the lower line includes the temperature correction. Q(BHT) is the heat flow obtained from the BHT with $T_S=20^\circ$C after allowing for internal heat production in the sediments. The other two values are those shown in fig. 8.2.6 where all the downhole measurements were used and a best fit line was obtained with the thermal resistance method. The first value had $T_S$ constrained to be $20^\circ$C and the second value had $T_S$ as a free parameter.

The wells with a thicker sedimentary sequence tend to have higher $\bar{k}$ as a result of the lower mean porosity. The values, uncorrected for temperature fall, in a range of 2.2-3.0 Wm$^{-1}$K$^{-1}$ and, after a temperature correction is applied, the range is about 2.0-2.4 Wm$^{-1}$K$^{-1}$.

The thermal conductivities derived with the representative values proportions for each lithology can be compared to those in table 8.2.2. For Burley 2 and Mcleod 1, the depth averaged thermal conductivities derived from a detailed lithology log are 0.5 Wm$^{-1}$K$^{-1}$ less than those obtained with the generalized lithology distribution without the temperature correction being applied (open triangles, fig. 8.2.7). The maximum difference is reduced about 0.25 Wm$^{-1}$K$^{-1}$ with the temperature correction (solid triangles, fig. 8.2.7). The depth averaged thermal conductivity value given in table 8.2.2 for the Innamincka 1
Fig. 8.2.7 Depth-averaged thermal conductivities, derived by using the same representative values for the proportions of each lithology for each formation for the 40 wells used in backstripping (chapter 6) as a function of the sediment thickness. The solid triangles are corrected for temperature, and the open triangles are not. The best fit lines are given in the form $\bar{k} = a \cdot z^b$ with $a = 0.9397$, $b = 0.1168$ for the corrected values and $a = 0.4238$ and $b = 0.2374$ for the uncorrected values. These values can be slightly higher than those obtained for wells where detailed lithology logs were available to constrain the lithologies more accurately.
well includes the effect of nearly 2000 m of Palaeozoic metasediment below the Permian-Cretaceous sediments. If this is taken into consideration then the depth averaged thermal conductivity for the Permian-Cretaceous section is reduced to about 2.0 and 2.2 Wm\(^{-1}\)K\(^{-1}\), with and without a temperature correction respectively. These values are about 0.25 and 0.4 Wm\(^{-1}\)K\(^{-1}\) less than those estimated with the representative proportion of each lithology. Dullingari 1 was not used in the backstripping process and so was not included in fig. 8.2.7.

It is probable then that the depth averaged thermal conductivities as given in fig. 8.2.7 are too high, probably as a result of an overestimation of the amount of sandstone present. Overall, however, the spread of the values is low and, after a temperature correction is applied, the range is of a similar magnitude to the measurement uncertainty (~15 %). The differences between the geothermal gradients across the basin (fig. 8.2.1) are then most probably due to real differences in heat flow from the underlying basement.

The inclusion of heat production in the sediments as given in table 8.2.1 results in a maximum heat flow contribution of 5.5 mWm\(^{-2}\) and up to 6°C to the BHT for the Burley 2 and Mcleod 1 wells and the geothermal gradient, as defined by the predicted BHT, is increased by less than about 2°C/km. Middleton (1979a) has shown that some of the Lower Permian sediments in the Moomba-Big Lake region have relatively high concentrations of radiogenic elements (1-9 ppm uranium and 7-23 ppm thorium), which he attributes to their derivation from a granitic source. However these concentrations would make only a small additional contribution to the heat budget (~3 µWm\(^{-3}\)) and would increase the geothermal gradient by less than 1 °C/km and thus the statement above, concerning variations in the basement heat flow, is still valid.

Assuming that the depth averaged thermal conductivity, \(\bar{k}\), of the sediment is constant, and a value of about 2.0-2.2 Wm\(^{-1}\)K\(^{-1}\) is probably appropriate after applying a temperature correction, then the ratio of two geothermal gradients (\(G_1, G_2\)) will be equal to
\[ \Delta Q = Q_2 - Q_1 = Q_2 (1 - G_1/G_2) \]

This relationship is illustrated in fig. 8.2.8 for a range of values for \( Q_2 \). In the southern Cooper basin area the higher gradients (45-55°C/km) are commonly associated with granite basement (e.g. Moomba-Big Lake, Burley-Mcleod) or basement highs, the lower values occur in the Patchawarra Trough (<35°C/km) and in other wells the gradients are about 35-45°C/km suggesting regional gradient ratios of between 0.65 and 0.85 relative to the higher values (45-55°C/km). This implies that, relative to a background of 75 mWm\(^{-2}\) (gradient \( \approx 35°C/km \)), the heat flow in areas with granitic or shallow basement is enhanced by between 15 and 45 mWm\(^{-2}\) for \( Q_2 = 90-120 \) mWm\(^{-2}\). There would be an uncertainty of up to 20 mWm\(^{-2}\) in these heat flow values, assuming errors of \( \pm 5°C/km \) in the geothermal gradient and \( \pm 0.2 \) Wm\(^{-1}\)K\(^{-1}\) in the depth averaged thermal conductivity.

### 8.2.d Causes of heat flow variations

The most obvious source for the excess heat flow is from internal heat production within the basement. Under the assumption that the radiogenic elements are uniformly distributed vertically through a layer, fig. 8.2.9 illustrates the relationship between heat flow, heat production and the thickness of the body. Taking the average values of Middleton (1979a) for the concentration of radiogenic elements measured in unweathered samples of the Moomba and Big Lake granites (uranium = 15.1 ppm, thorium = 60 ppm and potassium = 5.6%), then the rate of heat production in a granite of density 2650 kg m\(^{-3}\) would be 8.7 \( \mu \)Wm\(^{-3}\) (using the rates of heat production for the radiogenic elements given by Turcotte and Schubert, 1982, pg. 140). In order to explain the excess heat flow of up to 40 mWm\(^{-2}\), the granite bodies would have to be 4-5 km thick.

Assuming a uniform vertical distribution of radiogenic elements may be justified on a local scale, but if not, there are two possible alternatives involving fractionation of the radiogenic elements. Either their concentration decreases with depth, as is commonly assumed in larger scale crustal heat production models (Lachenbruch, 1970), and then the
Fig. 8.2.8 Excess heat flow ($\Delta Q = Q_2 - Q_1$) in mWm$^{-2}$ as a function of the ratio of the geothermal gradients ($G_1/G_2$, $G_2 > G_1$) for values of $Q_2$. The parameter space relevant to the central Eromanga Basin is shaded.
Fig. 8.2.9 Contours of constant heat flow in mWm$^{-2}$ ($Q_{1-D}$) as a function of thickness and uniform heat production rate of a layer, assuming only vertical heat transfer.
inferred thickness will be too low. Alternatively, the amounts measured from the core samples by Middleton (1979a) are underestimates as a result of leaching. As the upper surface of the granite has been subjected to weathering so that the Lower Permian sediments unconformably overlies the intrusive, then this latter option is certainly possible. However, this would imply that concentration of of radiogenic elements away from the zone of leaching is abnormally high, given that the values of Middleton (1979a) are nearly twice as high as those observed in granites elsewhere (e.g. Taylor and McLennan, 1985; Sawka and Chappell, 1986; Lucazeau and Mailhe, 1986; Lee et al. 1987).

More importantly, the relationship between heat production and granite thickness shown in fig. 8.2.9 is based on the assumption that heat transfer occurs only in a vertical direction (no lateral heat flow). Middleton (1979a) made this assumption when modelling the thermal state in the Moomba area and, with this restriction, his 1-D steady state models provided an explanation for the total heat flow, as opposed to simply the excess heat flow as we have defined it, associated with the granite areas. Middleton's models incorporate a high heat production (8.4 μWm⁻³) body of between 7 and 10 km in thickness, although the mantle heat flow he uses is less than 19 mWm⁻², which is at the low end of the range cited by Sclater et al. (1980a) even for old stable continents (17-31 mWm⁻²). The thickness of his high heat production bodies could be reduced by about 1.5 km if the mantle contribution to the heat flow is 31 mWm⁻².

Pitt (1982, 1986) attributed the high gradients associated with the granites not only to internal heat production but he also suggested that the granite bodies may be acting as conductive pipes, channeling heat upwards into the overlying sediments. As granites are often quartz rich, they will tend to have a higher thermal conductivity than low grade metasediments or shales, for example. It is also possible this effect may also contribute to the higher gradients associated with basement ridges, e.g. in the Jackson area at the extreme northeastern end of the Nappamerrri Trough, as the consolidated basement would be expected to have a higher thermal conductivity than the adjacent, more porous, sediments. Only five thermal conductivity measurements were made on basement samples (Appendix
1, table 2a). The one granite sample from Moomba 1 has a thermal conductivity of 3.3-4.1 Wm\(^{-1}\)K\(^{-1}\) which is comparable to the two shale samples from Dullingari 1. The relatively high values (3.35-4.15 Wm\(^{-1}\)K\(^{-1}\)) obtained for these shales is attributed to the fact that they are dolomitized (cf. a value of 2.45 Wm\(^{-1}\)K\(^{-1}\) from Gidgealpa 1). The one other thermal conductivity measurement was made on a clean quartz sandstone from the basement of Tirrawarra 1 well. This sample had a predictably high value of 5.05 Wm\(^{-1}\)K\(^{-1}\). From these few measurements it is not possible to conclude that the granites will have a significantly higher thermal conductivity than the other basement lithologies, but overall, the metasedimentary basement may have a thermal conductivity up to about 1 Wm\(^{-1}\)K\(^{-1}\) greater than the sediments.

8.3 Numerical modelling of steady state heat flow with laterally varying thermal properties

8.3.a Introduction

Sclater et al. (1970) have numerically examined the influence of 2-D, or laterally varying, thermal conductivity structure on the surface heat flow and England et al. (1980) extended the formulation to include variable heat production. These two studies both concluded that 2-D variations in thermal properties can have a significant effect on the surface heat flow. In this section we address the effect of laterally varying thermal properties on heat flow and temperatures that would be observed over a depth range of 3-4 km, the depth of many of the wells used in this study. Some wells, for example in the Moomba-Big Lake area, while having high geothermal gradients do not encounter granitic basement. The elevated heat flow may be due to lateral heat flow from nearby granites or heat refraction as a result of lateral thermal conductivity contrasts. For the wells where
granitic basement is known, we want to know the sensitivity to lateral heat flow of the inferences made under the assumption of 1-D (vertical) heat flow regarding the size of, and internal heat production in, the granites required to maintain the high geothermal gradients. Indirectly then we may be able to infer the distribution of different basement lithologies from an appraisal of the geothermal gradient data.

8.3.b Apparent and interval heat flow

If the assumption of 1-D steady state is valid, and provided there is no internal heat generation, then the estimate of heat flow will be constant, irrespective of the depth of the temperature measurements. However, if these assumptions are not justified, then the thermal resistance method may not provide a reliable estimate of the true heat flow. Therefore, it is necessary to distinguish between the apparent heat flow, estimated using a downhole temperature and an assumed surface temperature, and an interval heat flow, estimated using two downhole temperatures. Departure from the 1-D case may be the result of vertical or horizontal fluid movement and variable internal heat generation and thermal conductivities. This will result in the apparent heat flow being lower than the interval heat flow at depth, but an overestimate of the true surface heat flow if the true geotherm is concave upwards relative to the geotherm implied by the bottom hole temperature, and greater than the interval heat flow if the converse is the case. Fig. 8.3.1 illustrates the first situation, where a concave upwards geotherm may be the result of downwards or horizontal fluid flow, or a high heat producing body in the basement and the associated lateral heat flow in the overlying sediments where the temperature measurements are made. One downhole temperature measurement, $T_{z2}$, is made in a well where the thermal conductivity is constant with depth. The gradient extrapolated from $T_{z2}$ to the surface would be correct if a 1-D steady state situation existed and the apparent heat flow, $Q_{z2}$ would be equal to the true heat flow, which would be depth independent. However, the additional observation, $T_{z1}$, and the apparent heat flow estimated from it, $Q_{z1}$, show that
\textbf{Fig. 8.3.1} An illustration of the difference between apparent heat flow defined at depths \( z_1 \) and \( z_2 \), \( Q_1 \) and \( Q_2 \), and the interval heat flow, \( Q_i \). Apparent heat flow is estimated from the temperature at a given depth and an assumed surface temperature whereas interval heat flow uses two downhole temperature observations. The depth averaged thermal conductivity between the two relevant temperature values is used to calculate the heat flow (see fig. 8.2.5). The three values should agree if the 1-D (vertical heat flow) and no internal heat production approximations are valid, although the interval heat flow will be more sensitive to errors in the temperature measurements. If there is a departure from the 1-D situation then obviously the value heat flow estimates (apparent and interval) will be a function of depth, as shown above for \( Q_i, Q_1 \) and \( Q_2 \).
the 1-D approximation is invalid. The two values of apparent heat flow are different and both are less than the interval heat flow, \( Q_i \) estimated from \( T_{z1} \) and \( T_{z2} \). \( Q_i \) would be fairly close to the true heat flow at these depths, provided the temperature measurements were as accurate as represented in fig. 8.3.1. The true surface heat flow, \( Q_s \), estimated from the near surface temperature gradient would be less than either of the apparent heat flow estimates.

Although apparent heat flow is not perhaps the most obvious property to use, the uncertainties in the temperature measurements effectively preclude using the interval heat flow, estimates of which can vary considerably (e.g. by a factor of 2-3) depending on the temperature gradients derived from the data over small depth ranges. When calculating the apparent heat flow, errors in the temperature measurements at depth are smoothed by extrapolating a gradient to the surface. However, it should be recognized that the apparent heat flow is, physically, a somewhat artificial measure and not necessarily representative of the true heat flow, either at depth or at the surface.

8.3.c Method

The steady state energy equation to be solved is

\[
\nabla \cdot (k_b \nabla T) + A = \rho_f c_f v \cdot \nabla T \quad (8.3.1)
\]

with \( k = \) thermal conductivity, \( c = \) heat capacity, \( b, f = \) bulk and fluid subscripts, \( v = \) pore fluid velocity. Eqn. 8.3.1 is expressed in two dimensions using cartesian coordinates with only horizontal fluid flow as,

\[
\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} = -\frac{1}{k_b} \left[ A + \frac{\partial T}{\partial x} \frac{\partial k_b}{\partial x} - \rho_f c_f v \frac{\partial T}{\partial x} + \frac{\partial T}{\partial z} \frac{\partial k_b}{\partial z} \right] \quad (8.3.2a)
\]

and in cylindrical coordinates, assuming radial symmetry and no fluid flow,

\[
\frac{\partial^2 T}{\partial r^2} + \frac{1}{r} \frac{\partial T}{\partial r} + \frac{\partial^2 T}{\partial z^2} = -\frac{1}{k_b} \left[ A + \frac{\partial T}{\partial r} \left( \frac{\partial k_b}{\partial r} + \frac{k_b}{r} \right) + \frac{\partial T}{\partial z} \frac{\partial k_b}{\partial z} \right] \quad (8.3.2b)
\]
The details of the numerical procedure used to model the 2-D steady state heat transfer are given in Appendix 6 and a schematic representation of the model is given in fig. 8.3.2. As England et al. (1980) illustrated, the enhancement of heat flow is dependent on the depth to the top of the anomalous body, its geometry or aspect ratio and the thermal properties. The sediment thickness was taken as either 3 or 4 km (cf. Moomba 1 and Burley 2 respectively) and I typically used heat producing or high conductivity bodies of 2, 5 and 10 km in width and aspect ratios (width/height or diameter/height) of 0.5, 1 and 2. Additional calculations were made with greater width, or larger aspect ratio, bodies where it was considered relevant. The apparent heat flow, as defined above, was calculated using the temperature \( T(z_b) \) at the base of the sediments \( (z_b) \). Thus

\[
Q_{app}(z_b) = \bar{k} \left( \frac{T(z_b) - T_s}{z_b} \right) \tag{8.3.3}
\]

where \( \bar{k} \) is the depth averaged thermal conductivity through the sediments. We then define the excess heat flow as \( Q_{app} - Q_b \), where \( Q_b \) is the prescribed background heat flow into the base of the model (80 mWm\(^{-2}\)). Unless specifically stated I always used the temperature at the base of the sediments to estimate the apparent heat flow. The results in the tables are given for the profile directly above the symmetry axis of the anomalous body. I will discuss the effect of fluid flow in an aquifer, as illustrated in fig. 8.3.2, later.

### 8.3.d The effect of lateral variations in thermal conductivity

Table 8.3.1 gives the results of calculations made with a thermal conductivity contrast \( (\Delta k) \) of 1 Wm\(^{-1}\)K\(^{-1}\) or a ratio of 1.33 between the body \( (k = 4 \text{ Wm}^{-1}\text{K}^{-1}) \) and the background \( (k = 3 \text{ Wm}^{-1}\text{K}^{-1}) \) for cylindrical and cartesian geometries. I also give the results for a body 5 km across with a contrast of 3 Wm\(^{-1}\)K\(^{-1}\) (i.e. \( k = 6 \text{ Wm}^{-1}\text{K}^{-1} \)). Fig. 8.3.3 illustrates the results for both a cylindrical and cartesian body with a thermal
\[ T = T_{\text{surface}} \]

**Fig. 8.3.2** Schematic representation of the model used to examine the effects of lateral heat transfer and fluid flow in an aquifer over a heat source (e.g. granite intrusion). The heat source may have a positive thermal conductivity contrast with the adjacent basement which has a thermal conductivity of 3 Wm\(^{-1}\)K\(^{-1}\). The thermal conductivity of the aquifer when flow is included in the models is 3 Wm\(^{-1}\)K\(^{-1}\) otherwise the sediments have a thermal conductivity of 2 Wm\(^{-1}\)K\(^{-1}\) throughout. The surface temperature and basal heat flow \((Q_b)\) are specified and there is no lateral temperature gradient at the sides of the model. The geometry and grid spacings used for the model are variable and are discussed in the text and in appendix 6.
conductivity contrast of 1 Wm\(^{-1}\)K\(^{-1}\) under 4 km of sediment cover. These calculations did not include any heat production in the body.

<table>
<thead>
<tr>
<th>Sediment Thickness</th>
<th>3 km</th>
<th>4 km</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Aspect ratio</td>
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</tr>
<tr>
<td></td>
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<td>cartesian</td>
</tr>
<tr>
<td>2 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\Delta k = 1.0 \text{ Wm}^{-1}\text{K}^{-1})</td>
<td>1.6</td>
<td>1.7</td>
</tr>
<tr>
<td>2.0</td>
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<td>3.7</td>
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<td></td>
</tr>
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<td>5.6</td>
</tr>
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<tr>
<td>0.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>20 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\Delta k = 1.0 \text{ Wm}^{-1}\text{K}^{-1})</td>
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<td>4.4</td>
</tr>
<tr>
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</tr>
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<tr>
<td>5 km</td>
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<td>12.1</td>
<td>12.3</td>
</tr>
<tr>
<td>1.0</td>
<td>14.7</td>
<td>17.0</td>
</tr>
</tbody>
</table>

Table 8.3.1 The excess heat flow, \(Q_{app} - Q_b\), in mWm\(^{-2}\) over the axis of a body with a positive thermal conductivity contrast for aspect ratios (width/height) of 2.0, 1.0, and 0.5 and widths (cartesian) or diameters (cylindrical) of 2, 5, 10 and 20 km. The body underlies either 3 or 4 km of sediment cover (k = 2 Wm\(^{-1}\)K\(^{-1}\)).

It is apparent from these calculations that lateral variations in the thermal conductivity alone are unable to account for the excess heat flow of 15-40 mWm\(^{-2}\) associated with the granites and basement highs with a reasonable thermal conductivity contrast of 1 Wm\(^{-1}\)K\(^{-1}\), although there are some points worth noting here. The role of the
Fig. 8.3.3 Excess heat flow, $Q_{\text{app}} - Q_{\text{b}}$, over the axis of a body 2, 5, 10 and 20 km in diameter with aspect ratios (width/height) of 0.5, 1.0 and 2.0. The body has no internal heat production, but has a thermal conductivity contrast of 1 Wm$^{-1}$K$^{-1}$ with the adjacent basement ($k = 4$ and 3 Wm$^{-1}$K$^{-1}$ respectively). The body underlies 4 km of sediment ($k = 2$ Wm$^{-1}$K$^{-1}$). The solid squares are for a cartesian geometry and the unfilled squares are for a cylindrical geometry.
geometry of the body (i.e. aspect ratio) is clear - a vertically elongated shape (small aspect ratio) results in a greater enhancement of heat flow as there is a greater length over which to refract heat laterally into the body. The enhancement of heat flow above a body due to heat refraction is an edge effect, arising as a result of the discontinuity in the thermal conductivity and, as a body widens, this heat flow above its central axis will tend to the background value.

This effect can be seen from table 8.3.1 where the cylindrical body shows a greater enhancement than the rectangular one as the aspect ratio decreases i.e. the body becomes vertically elongated (compare 5 and 10 km widths). As the width increases to 20 km the enhancement of the axial heat flow tends to decrease, although it is clear that the enhancement effect is still more significant with the cylindrical geometry for lower aspect ratios. The final feature to note is that the shallower the body, or less sediment cover, the greater the enhancement of heat flow. This is expected from the constant temperature upper boundary condition which results in the heat transfer being dominantly in a vertical direction. Therefore there is effectively less room for the heat refracted into the body, and subsequently conducted out of its upper surface, to spread laterally before reaching the surface.

8.3.e The effect of lateral variations in internal heat production

Table 8.3.2 gives the results of calculations made where internal heat production (A) of 3, 6 and 10 μWm⁻³ was included in the body. These results are for a cylindrical body with a positive thermal conductivity contrast of 1 Wm⁻¹K⁻¹. The difference between these results and those made without the lateral conductivity variation are always less than the values given in table 8.3.1 for the thermal conductivity contrast alone. I also include in table 8.3.2 the heat flow expected for purely vertical heat transfer as illustrated in fig. 8.2.9.
Table 8.3.2  Excess heat flow, $Q_{\text{app}} - Q_{b}$, in mWm$^{-2}$ over the axis of a cylindrical body with internal heat production ($A=3$, 6 and 10 $\mu$Wm$^{-3}$) for diameters of 2, 5 and 10 km. The body also has a thermal conductivity contrast of 1 Wm$^{-1}$K$^{-1}$ with the adjacent basement and underlies 3 or 4 km of sediment. The excess heat flow expected for 1-D (i.e. vertical) heat transfer is also given.

These calculations illustrate the importance of lateral heat flow as the excess apparent heat flow estimated above the centre of the body is between 10 and 50% of the value predicted with the 1-D model. In fig. 8.3.4 I illustrate the results of similar calculations made for a cylindrical body 5 km thick and up to 50 km in diameter underlying 4 km of sediment. The results of these calculations and those made with 3 km of sediment are summarized in table 8.3.3a. It is seen that the departure from the 1-D case is reduced for the wider bodies even though the heat flow may still be up to 20% less than that predicted from a 1-D model for a body 50 km wide (aspect ratio of 10). As with the lateral variation of thermal conductivity, the enhancement of heat flow over the axis of a body with a positive heat production contrast is greater for a shallower body, all other factors being equal and this is again a consequence of the upper boundary condition.
Fig. 8.3.4 Excess heat flow, $Q_{\text{app}} - Q_{\text{b}}$, over the axis of a cylindrical body 5 km thick and up to 50 km in diameter with a thermal conductivity ($k$) of 4 Wm$^{-1}$K$^{-1}$ and an internal heat production contrast ($A$) of 3, 6 and 10 $\mu$Wm$^{-3}$ with the adjacent basement ($k = 3$ Wm$^{-1}$K$^{-1}$). The body underlies 4 km of sediment ($k = 2$ Wm$^{-1}$K$^{-1}$).
<table>
<thead>
<tr>
<th>Sediment thickness</th>
<th>4 km</th>
<th>3 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>A (µWm⁻³)</td>
<td>3</td>
<td>6</td>
</tr>
<tr>
<td>Diameter (km)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5.0</td>
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<td>20.0</td>
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</tr>
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</tr>
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<td>40.0</td>
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</tr>
<tr>
<td>50.0</td>
<td>13.9</td>
<td>24.8</td>
</tr>
</tbody>
</table>

Table 8.3.3a. Excess heat flow, $Q_{app} - Q_b$, in mWm⁻² over the axis of a cylindrical body of up to 50 km in diameter and 5 km thick with internal heat production ($A = 3, 6$ and $10$ $\mu$Wm⁻³).

The difference between the surface heat flow and the apparent heat flow calculated with the temperature at the base of the sediments is shown in fig. 8.3.5 for a cylindrical body of 10 and 50 km in diameter. Although I refer to a surface heat flow, it is strictly the apparent heat flow calculated with the temperature at the shallowest grid point (500 m). As a result of lateral heat transfer, the apparent heat flow will be equal or slightly higher than the surface heat flow over the centre of heat producing body and slightly less over a short distance away from the margin of the body. An important feature of this diagram is that the influence of lateral heat transfer on the vertical heat flow adjacent to the body is only significantly affected within a distance of about 10% of the diameter of the body. Notice also how the isotherms are pulled up for the 50 km body compared to the 10 km body. Table 8.3.3b gives the temperature at the base of the sediments for these models and these can be compared to BHT values of 220-230°C in the Burley 1, Burley 2 and Mcleod 1 wells (about 4 km of sediment) and 180-200°C in the Moomba-Big Lake area (about 3 km of sediment).
Fig. 8.3.5  Surface and apparent heat flow and isotherms at 50°C intervals for cylindrical bodies of diameter 10 km (lower panel) and 50 km (upper panel) and 5 km thick. The body, located at the left hand side of the diagram, has an internal heat production rate of 10 μWm⁻³ and a thermal conductivity of 4 Wm⁻¹K⁻¹. The adjacent basement and the 4 km of overburden have no internal heat production and thermal conductivities of 3 and 2 Wm⁻¹K⁻¹ respectively.
Table 8.3.3b. Temperature (°C) at the base of the sediments over the axis of a cylindrical body as described for table 8.3.3a. The surface temperature is taken as 20°C

These results suggest that lateral heat production contrasts may be the explanation for the elevated heat flow and temperatures in these areas, provided that the bodies are ≥30 km in diameter and have a positive heat production contrast of 6-10 µWm⁻³. These heat production contrasts could, of course, always be reduced by increasing the thickness of the body. Finally, table 8.3.4 compares the results for a 5 km thick body with cartesian and cylindrical geometries for aspect ratios of 2, 1 and 0.5. These calculations were made both with and without internal heat production and thermal conductivity contrasts. The influence of the cylindrical geometry is such that, for these models, the lateral heat loss is up to 3 times that of the rectangular body. For a body of a given geometry, the two significant ratios are the width/thickness (aspect ratio) and the width/depth to top of the body. As these both increase then a 1-D approximation will become more valid over the centre of the body. The actual enhancement of heat flow will also depend, of course, on the thermal properties (thermal conductivity and heat production).
Table 8.3.4  Excess heat flow, \( Q_{\text{app}} - Q_b \), in mWm\(^{-2}\) over a body 5 km in width (or diameter) with no thermal conductivity contrast (2 columns on the left) and with \( \Delta k = 1 \) Wm\(^{-1}\)K\(^{-1}\) for various internal heat production rates (A = 0, 3, 6 and 10 \( \mu \)Wm\(^{-3}\)).

8.3.f    The effect of fluid flow in an aquifer over a heat source

As stated earlier, the Eromanga Basin is a major component of the Great Artesian Basin. The hydrological characteristics of the basin have been described by Habermehl (1980, 1986) and the directions of groundwater flow are shown in fig. 8.3.6. The aquifer system can be broadly subdivided into the major Jurassic and less important Cretaceous aquifers. It was described in section 7.2.b how hydrological flow can perturb the simple conductive geotherm especially in the recharge and discharge areas where considerable vertical water movement may occur. The regional dips of the aquifers in the Eromanga Basin are low and less than 5° (Senior et al., 1978) although of course groundwater may move vertically along faults. Generally, however, the vertical permeabilities are low, especially in the confining beds (Habermehl, 1980, 1986), and the major perturbations would be associated with horizontal fluid flow. Horizontal water velocities of up to
Fig. 8.3.6. Recharge and natural discharge (springs) areas, and directions of regional groundwater flow in the Eromanga Basin (from Habermehl, 1986).
1.5x10^{-7} \text{ m s}^{-1} (~5 \text{ m yr}^{-1}) are observed at the recharge and discharge margins of the basin (Habermehl, 1980, 1986). The advective transport of heat due to groundwater water flow may then be locally important in redistributing heat when flowing over a heat source, such as a granite body, and this effect may influence heat flow estimates downstream from the body.

To simulate the flow of water in an aquifer over a localized heat source the original 2-D cartesian model was adapted (see fig. 8.3.2) and the horizontal advective term was included in the numerical algorithm (Appendix 6). The aquifer is assumed to be 500 m thick and with its upper surface at a depth of 1500 m, which is a reasonable approximation to the Jurassic aquifer system as observed in this area. I used horizontal flow velocities of 0.1, 0.2, 0.5 and 1.0 x 10^{-7} \text{ m s}^{-1} (0.32, 0.64, 1.6 and 3.2 \text{ m yr}^{-1}). The heat producing body was 5 km thick and 40 km in width with 3 and 4 km of sediment cover. The body had no thermal conductivity contrast with the adjacent basement but the aquifer was assigned a thermal conductivity of 3 \text{ W m}^{-1}\text{K}^{-1} which is suggested by the downhole thermal conductivity profiles (fig. 8.2.4). The model can no longer be regarded as symmetrical and so the granite body is located in the centre of the mesh. The width of the mesh was increased so that the surface heat flow at the two ends was equal to the background value of 80 \text{ mW m}^{-2}.

The purpose of these calculations was to examine the variation with depth of the heat flow calculated from the 2-D model but under the assumption of vertical heat conduction only, and also the horizontal length scales over which fluid flow may perturb the heat flow. The interval heat flow, as defined in section 8.3.b, was calculated using the temperatures at vertically adjacent mesh points (500 m spacing) and the apparent heat flow (also see section 8.3.b) was estimated from the temperature at each mesh point and the surface temperature (eqn. 8.3.3). Figs. 8.3.7 a,b summarize the results for the temperature profile and heat flow profile directly above the centre of the anomalous body. It is seen that for all practical purposes that, for a given flow velocity, the apparent and interval heat flows are equal and constant above the aquifer and always less than the value without flow. For
Fig. 8.3.7a Temperature and apparent and interval heat flow profiles over the centre of a cartesian body 40 km wide and 5 km thick with an internal heat production rate of 10 $\mu$Wm$^{-3}$. The body has no thermal conductivity contrast with the adjacent basement ($k = 3$ Wm$^{-1}$K$^{-1}$). Four km of sediment overlies the body and includes an aquifer 500 m thick between 1500 and 2000 m. The aquifer has a thermal conductivity of 3 Wm$^{-1}$K$^{-1}$ and a value of 2 Wm$^{-1}$K$^{-1}$ is assigned to the rest of the sediment. The numbers next to the profiles are the flow velocities in the aquifer in m s$^{-1} \times 10^{-7}$ (1 m s$^{-1} \times 10^{-7}$ = 3.2 m yr$^{-1}$).
Fig. 8.3.7b  As 8.3.7a but with 3 km of sediment.
increasing flow velocities the interval heat flow below the aquifer is progressively greater than the no flow case whereas the opposite applies to the apparent heat flow and the interval heat flow above the aquifer. Examination of the geotherms (lower panel, figs. 8.3.7a,b) reveals why this is observed. Fluid flow advectively transports heat away from the region vertically above the source, reducing the vertical heat flow and the temperature at the base of the sediments. This latter effect is the reason why the apparent heat flow is reduced below the aquifer. The flow also causes a steepening of the thermal gradient below the aquifer and a shallowing above it which is the reason for the variation in the interval heat flow. Of course, heat flow above the aquifer will become relatively enhanced 'downstream' and the distance over which this occurs depends on the fluid flow velocity. This effect is illustrated in fig. 8.3.8, where the surface heat flow and the apparent heat flow calculated with the temperature at the base of the sediments (4 km). The fluid flow depresses the isotherms over the body and, although affected by the presence of the aquifer, the apparent heat flow would provide a better estimate of the size of the heat producing body than the surface heat flow.

The downhole temperature measurements from Burley 2 and Mcleod 1 show lower temperatures in the region of, and just below, the Jurassic aquifers at depths of about 1800-2500 m (fig. 8.2.4) than those calculated using the thermal resistance method. In fig. 8.3.9 I show the geotherms obtained for 4 km of sediment with a velocity of 0.5 x 10^{-7} m s^{-1} (1.6 m yr^{-1}) and also the geotherm that would be obtained by using the temperature at the base of the sediments with the thermal resistance method. Differences in temperature of up to nearly 20°C can arise due to the horizontal flow of groundwater over a heat source and this effect may be manifested in the downhole temperatures in Burley 2 and Mcleod 1. The latter well shows a greater deviation from the 1-D case which may be attributable to the fact that, although both wells directly overlie a granite intrusion (probably the same one), Mcleod 1 lies up the flow gradient (i.e. east) relative to Burley 2. However, considerably more accurate temperature data would be required to support this statement, and it is
Fig. 8.3.8  Surface and apparent heat flow and isotherms at 50°C intervals for the cartesian body 40 km wide and 5 km thick. The body, located from 200 - 240 km, has an internal heat production rate of 10 μWm⁻³ and no thermal conductivity contrast with the adjacent basement (k = 3 Wm⁻¹K⁻¹). 4 km of sediment overburden is present and an aquifer 500 m thick is situated at 1500-2000 m. Fluid flow occurs from left to right at velocities of 0, 2.5 x 10⁻⁷ and 5.0 x10⁻⁷ m s⁻¹ (lower, centre and upper panels respectively). The aquifer has a thermal conductivity of 3 Wm⁻¹K⁻¹ the rest of the sediment has value of 2 Wm⁻¹K⁻¹. The apparent and surface heat flows are very similar for the no flow case and cannot be distinguished on this figure.
Fig. 8.3.9 Calculated geotherms over the centre of the heat producing body as described in fig. 8.3.7a with fluid flow in a 500 m thick aquifer at 1500 m with a velocity of $0.5 \times 10^{-7}$ m s$^{-1}$ (1.6 m yr$^{-1}$). The profile labeled (a) is that obtained from the numerical solution to the steady state energy equation (8.3.2a) and the profile labeled (b) is that estimated by using the lowermost temperature value from (a) and assuming 1-D steady state, similar to the thermal resistance method, with no internal heat production or fluid flow. The thermal conductivity was not corrected for temperature. The corrected temperature values for Mcleod 1 and Burley 2 are also shown and the Jurassic aquifer system is between about 1800 and 2300 m in these two wells.
possible that the suggestion of lower temperatures at these depths arises due to errors in the temperature estimates themselves.

Although groundwater flow may perturb the local geotherm in areas of high heat flow, the results of the heat flow calculations above show that, in the presence of such advective heat transport, the apparent heat flow will be a minimum estimate of the true heat flow from the basement, at least over the centre of any high heat producing bodies. Even away from the central region of a body, the apparent heat flow estimates are within a few mWm\(^{-2}\) of the value calculated in the absence of fluid flow (see fig. 8.3.8). Therefore, the values of high apparent heat flow in the wells without granitic basement are unlikely to be the result of enhancement of heat flow by advection.

8.4 Discussion

8.4.a Implications for the central Eromanga Basin basement geology

Obviously, the relationship of estimates of depth scales of heat production or the vertical distribution of radiogenic elements to the actual situation will be dependent on the geometry of the bodies being considered. Lateral heat transfer acts to reduce the inferred thickness of a heat producing body and this heat loss will be more significant than enhancement by a positive thermal conductivity contrast for the values considered in the previous section. The departure from the 1-D situation is a function of the ratio of the horizontal and vertical wavelengths of the body and, as shown by Jaupart (1983), for crustal scale heat production heterogeneities, horizontal heat transfer results in the heat flow representing an average distribution over distances of 200-350 km.

The results of the numerical calculations described above suggest that the high heat flows and BHT associated with granite intrusions in the southern Cooper Basin area are the
result of shallow crustal batholiths, greater than about 30 km across, rather than smaller (<10 km) isolated bodies. With an average internal heat production contrast of 6-10 μWm\(^{-3}\), they would be at least 5 km or so in thickness. The similarity of this thickness to that obtained with the 1-D assumption is because the wider bodies approximate the 1-D case, at least in the axial region. The interpreted basement geology as shown in fig 8.3.2 shows a patchy distribution of granites and the implication is that they must extend laterally underneath the lower Palaeozoic/Proterozoic sediments and metasediments. The wells which do intersect granite are probably then encountering either cupolas, as appears to be the case in the southwest of England (Bott et al., 1958) or fault bounded subcrops. In either case, these areas were uplifted and eroded prior to the commencement of Permian sedimentation as these sediments are unconformably overlying the granites.

In the Nappamerri Trough, it may be inferred that the Moomba-Big Lake area is underlain by a granite at least 30 km across and some 5 km or more in thickness, and a similar situation is proposed for the Burley/Mcleod area to the northeast. There is no data to confirm whether the two intrusions are connected or not. As the average gradient in this area is over 40°C/km it is likely that granites are widespread. The Strzelecki and Jackson areas have equally high geothermal gradients but no obvious indication of granite in the basement. The easterly trending basement ridge, on which the Jackson field is situated, also correlates with a trend of high gradients. As it has been shown that thermal conductivity variations alone can account for less than 10 mWm\(^{-2}\) of excess heat flow, then it is probable that internal heat production within the elevated basement is responsible for the high gradients. The elongated nature of the ridge makes the cartesian geometry models more relevant and the results in table 8.3.4, for example, imply that the heat production contrast between the basement and the sediments need not be as high as that required for the granites and the adjacent basement. Also the basement comes to within 2 km of the surface and a heat production contrast of about 3 μWm\(^{-3}\) could explain the excess heat flow of 10-20 mWm\(^{-2}\) along the ridge which would be at least 20 km across. The geothermal gradients in the Jackson area at the eastern end of the ridge would additionally require a locally
increased heat production and there may be an intrusive body below the metasedimentary basement, and a similar statement may be made about the Strzelecki area. The role of the basement lithology on the heat flow is apparent in the Patchawarra Trough where the geothermal gradients are relatively low (30-40°C/km). The basement lithologies in this area include carbonates and quartzose sandstones, both of which would have low heat generating capabilities.

8.4.b Implications for the thermal history of the Eromanga Basin

It is possible, then, to explain the differences in heat flow in the central Eromanga/Cooper Basin area by steady state models involving primarily lateral variations in heat production within the upper 5-10 km of the basement rocks. However as stated at the beginning of this section even the lower heat flow estimates (~75 mWm⁻²) in this region are high when compared to global averages (< 60 mWm⁻²). There is a strong positive correlation between the geothermal gradient and vitrinite reflectance (Kantsler et al., 1982, 1986) and, for example, reflectances of over 6.0 % have been measured in the lower parts of the section in wells in the Burley region (Kantsler et al., 1978,1983; Delhi Petroleum, pers. comm., 1986). Considering that the oil window is assumed to correspond to reflectances of 0.5 to 1.3 %, or temperatures of about 70-150°C, it is not surprising that only traces of gas were found in the lower section of these wells. I briefly mentioned earlier that maturation modelling in this area has led some authors to conclude that the present day thermal regime is a relatively recent (< 10 Ma) phenomena at which time an elevation of the regional geotherm of 10-15°C/km occurred (Pitt, 1982, 1986; Kantsler et al., 1978, 1983), corresponding to an increase in heat flow of about 20-30 mWm⁻² (~ 20-25 %). Kantsler et al. (1986) suggest that the observed reflectances can be modeled with the present day geothermal gradients being in existence for 300 Ma and also are consistent with the gradients being higher in the Permo-Triassic. They hypothesize that the early high heat
flow could be the result of a thermally driven tectonic subsidence mechanism (e.g. section 2.3) and possibly this was enhanced by initially high levels of radioactive decay in the basement granites. The former statement is certainly a possibility but, taken literally, the latter is not given that the half lives of the radiogenic elements are $10^{8-10}$ years and the granites are only 300-350 m.y. old. It is possible that leaching of these elements has occurred, but as I described earlier this would imply elemental concentrations about 3 or 4 time higher than average measured values. The results of the inversion procedure described in section 6.4 would lead us to conclude that if a recent elevation of heat flow has occurred then maturation modelling just will not be sensitive to the early thermal history and it is clear that the variety of conclusions arrived at by modelling reflectances is the result of this lack of resolution. However, the interpretations made above that much of the Nappamerri Trough is underlain by granites implies that, prior to sedimentation, the thermal regime of the area was elevated sufficiently to result in widespread intrusion. The granites encountered in drill holes in the area yield radiometric dates of 305-360 Ma and although the younger ages may reflect cooling, these ages suggest that the elevated thermal regime may have existed for up to 50 m.y. before deposition commenced in the Late Carboniferous-Early Permian (~ 290 Ma). Unfortunately, it is unlikely that any current palaeogeothermometer would be sensitive to this early phase of the thermal history of the basin, especially if the relatively high thermal regime is a recent (< 10 Ma) phenomena.

Fig. 8.3.10 gives the result of a single $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum made on K-feldspar extracted from the basement granite encountered in the Moomba 1 well. Unfortunately other basement and sediment samples proved unsatisfactory for this method as either there were no appropriate minerals or the feldspars were too heavily altered. The bottom hole temperature in this well is nearly 200°C (Middleton, 1979a) and yet the diffusion profile is more or less flat past about 0.3 $^{39}\text{Ar}$ released. The result was not good enough to obtain reliable estimates of the diffusion parameters. A useful comparison may be made, however, with the study of Harrison et al. (1986) who used the $^{40}\text{Ar}/^{39}\text{Ar}$ technique to constrain the thermal history of the Fenton Hill caldera, New Mexico. They concluded that...
Fig. 8.3.10 $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum for feldspar from the granite intersected in the Moomba 1 well as pre-Permian basement, currently at a temperature of nearly 200°C. The plateau age is similar that estimated stratigraphically for the overlying sediments (~290 Ma) but the lack of argon loss (i.e. < 15%) is apparent, implying that the present day temperatures have been operative for less than 10 Ma.
gradients observed at the present day were established less than 1 Ma and they attribute the
temperature increase to geothermal flow associated with nearby volcanic activity. Their
diffusion profiles for temperatures at 110-197°C show more argon loss than that of the
Moomba 1 basement sample and a greater range of apparent age (i.e. a less well defined
plateau). The age spectrum obtained for the Moomba granite can therefore be qualitatively
interpreted as indicating that the present day thermal regime in this area is also recent and
represents an elevation of 10-20°C/km at possibly less than 1 Ma and certainly less than 10
Ma. Preliminary results from fission track analysis lead to the same conclusion (Duddy
1987) and indeed suggest that the gradient has been elevated over most of the central
Eromanga Basin. The fission track data give ages which are often in accord with the
stratigraphic ages although the observed temperatures, if operative for longer then 10-20
Ma, would have resulted in track annealing and a reduction in the fission track ages. Also,
the estimates made of the temperature at which the observed mean track length would be
produced over a 10 or 20 Ma interval, are below those currently observed by 20-30°C.
These results are consequently interpreted by Duddy (1987) as being indicative of an
increase in the geotherm within the last million years or so. Duddy (1987) describes similar
evidence for an increase in the geothermal gradient from one well as far west as the
Simpson Desert area and recent results suggest that the increase may also be observed in
western Queensland (Duddy, pers. comm., 1987).

There is obviously strong geochemical evidence for a recent (<10 Ma)
elevation of the geothermal gradient in the central Eromanga Basin. This cannot be
attributed to depositional processes as significant sedimentation effectively ceased with the
Winton Formation at about 91 Ma. As shown in section 7.2.c, with sedimentation rates of
250 m/m.y. for 5 m.y., the temperature at the base of 3-4 km of sediment will be within at
least 5 % of the steady state value at all times and in any case lithospheric thermal relaxation
time scales are too long to facilitate the proposed recent elevation. The 100 m or so of
irregularly distributed Tertiary sediments are too thin, and also too old, to account for an
increase in the gradient of 10-20°C/km in the last 10 Ma. Although variations in the near
surface (top 10 km) distribution of heat production can explain the *differences* in the heat flow estimates around the southern Cooper Basin region, in the light of the above statements it is also not possible to appeal to a model of deeper crustal or mantle heat production by radiogenic decay to explain the high background heat flow of about 75-80 mWm$^{-3}$. This is because the basement is necessarily older than the oldest sediment (~290 m.y.) and the high basal temperatures would have been operative from more or less the time sedimentation ceased (~91 Ma).

In their maturation studies of the southern Cooper Basin, Kanttsler et al. (1978, 1983) considered the possibility of a mantle hot spot and/or the presence of magma in the crust under the area as an explanation for the recent elevation of the geothermal gradients. The major objection to a mantle plume model is the lack of topographic expression for such a feature. As a result of thermal expansion when excess heat is conducted upwards, we would expect to see a few hundred metres of topography in the form of a broad swell (Mareschal, 1981; Lambeck and Nakiboglu, 1985; McMullen and Mohraz, 1987) or possibly even some dynamic uplift component (Houseman and England, 1986). In contrast, the topography in the Eromanga Basin shows an overall decrease from east to west, being about 400 m in the central Queensland and at or just below sea level in the Lake Eyre region in South Australia. Mareschal (1983) has shown how a similar topographic expression would be expected if large scale rapid intrusion of magma into the lower lithosphere took place. Thus it seems that the cause of the thermal anomaly cannot be too deep and not in the lower lithosphere. However, if heat is advectively transported to a shallow region of the crust, either by magma migration or fluid movement, then widespread uplift will not necessarily result.

Carbon dioxide isotope analyses of natural gases from the Cooper/Eromanga Basin sediments and the observation that CO$_2$ content increases with temperature (contrary to what is expected during organic maturation) are interpreted to indicate an igneous (low $^{13}$C/$^{12}$C) component (Rigby and Smith, 1981) and possibly related to magmatic fluids moving upwards. Unfortunately, it is not possible to date the timing of the gas formation
and the gases may be older than the 1-10 Ma inferred for the elevation of the geotherm. Changes in the flow regime within the aquifer system may locally alter the geothermal gradient but it is difficult to envisage how the thermal regime could be enhanced by a factor of 20-25% relatively uniformly over an area possibly more than 600 km across. A similar criticism may be made regarding the possibility of deep fluid circulation or intrusive processes. Thus, at this stage, it is not possible to make any statements less than purely speculative as to the mechanism responsible for the recent elevation of the geothermal gradients in the central and western Eromanga Basin.

8.5 Concluding remarks

Geothermal gradients in the central Eromanga Basin region estimated from BHT and DST measurements are between 30 and 60°C/km. Measurements of the thermal conductivity of the sediments (appendix 1) reflect the role of lithological type and porosity, with low porosity quartz rich sandstones having the highest values (5-6 Wm⁻¹K⁻¹). Incorporating what is considered to be representative lithological representation of each sedimentary formation, estimates of the depth averaged thermal conductivity range from 2.2 and 3.0 Wm⁻¹K⁻¹, and after correcting for temperature, the range is 2.0-2.4 Wm⁻¹K⁻¹, although these may be high (5-10%) when compared to values obtained with detailed lithology logs and a range of 2.0-2.2 Wm⁻¹K⁻¹ is considered reasonable. Combining the geothermal gradients with the thermal conductivity data, heat flow in the region is estimated to be between 75 and 120 mWm⁻². Using steady state models, the differences in heat flow over and above a background value of about 75-80 mWm⁻² are attributed to lateral variations in heat production in the basement underlying the Permian-Cretaceous sediments. Some wells have intersected granites, and these contain high concentrations of radiogenic heat producing elements. Some other wells are located on basement highs, and it is considered that lateral contrasts in heat production between these ridges and the sedimentary sequence
may be responsible for the enhanced heat flow observed in these regions. Lateral variations in thermal conductivity are considered relatively unimportant for the models examined here. The few measurements made on basement lithologies are inconclusive regarding the possibility of the thermal conductivity of granitic basement being significantly higher than the metasedimentary basement. The distribution of granites under the sedimentary sequence in the Nappamerri Trough must be more widespread than would be interpreted from well data alone, as lateral heat transfer precludes the intrusives from being less than 30 km in diameter. In areas such as Moomba-Big Lake, where the heat flow is 100-120 mWm$^{-2}$, yet not all the wells encounter granitic basement, the inference is that some wells have intersected the upper surface of granitic cupolas, or fault bounded blocks, and the intrusives are laterally more extensive underneath the other basement lithologies. The inferred widespread presence of granites, probably less than 50 m.y. older than the earliest sediments in the southern Cooper Basin, implies that the thermal regime of this region was elevated prior to, and possibly during, the fault controlled phase of subsidence in the Permian. The granites were uplifted and eroded before sedimentation commenced. This uplift and erosion may have been the result of regional compression acting on eastern Australia during the Late Devonian-Carboniferous. Given the irregular lithological nature of the basement intersected in drill holes, it is probable that faulting played an active role in the uplift. Subsequently, the same faults may have be reactivated under a more oblique stress regime as the southern Cooper Basin began to develop in the Permian. The thermal regime may have remained relatively high during the Permian and the Triassic to Cretaceous subsidence may have been driven by the subsequent decay of the elevated thermal regime.

The thermal history of the Eromanga Basin appears to have been complicated by a recent (< 10 Ma) elevation of the geothermal gradient in the Cooper Basin and possibly the Simpson Desert Basin areas. The cause of this elevation is unknown but is likely to be related to processes acting in the crust rather than the lower lithosphere as there is no topographic expression of a deep heat source.
9. CONCLUSIONS

Geodynamic models for the formation of sedimentary basins allow predictions to be made which can subsequently be tested against observations. Two of the primary predictions made by many models are the subsidence and heat flow histories.

Tectonic subsidence is defined as that component of the overall subsidence attributed to one or more physical mechanisms acting during the evolution of a sedimentary basin. The tectonic subsidence is generally obtained by the backstripping method where the effect of sedimentary loading is removed and additional corrections may be made for palaeobathymetry and sea level variations. Usually, an attempt is made to restore the sedimentary column to its decompacted thickness at the time of deposition of a given formation or unit. The overall shape of the subsidence curve is relatively insensitive to the porosity/depth function assumed in the decompaction procedure. Errors of about 10-20% may arise in the absolute values of the subsidence curve as a result of the use of an unrepresentative porosity function. It is considered that, in many cases, incorrect assumptions regarding palaeobathymetry, sea level variations and isostatic models may have greater uncertainties which are directly mapped into the tectonic subsidence estimates.

Application of the backstripping method to the intracratonic Eromanga and Cooper Basin sequences reveals some consistent features in the subsidence history. The Permian was a period of fault controlled subsidence in the southern Cooper Basin with rapid variations in thickness and lithology and the development of local unconformities, reflecting the dominant influence of the regional stress field on subsidence. The pattern of faulting during this time was complicated but appears to have a major strike-slip component. During the Triassic, sedimentation became more widespread, extending over the whole of the Cooper Basin and the Simpson Desert Basin. A Late Triassic unconformity in the area is attributed to large scale tectonics associated with activity at the eastern plate margin, rather than a fundamental change in the subsidence mechanism. The Early Jurassic sediments
were deposited over a similar geographical area to the Late Triassic formations in what was, by that time, probably a lowland area. By the Middle-Late Jurassic the area of deposition had expanded to coalesce with the Surat and Carpentaria Basins to the east and north respectively. However, the major depocentre was still the Cooper-Simpson Desert Basin region. The subsidence from the Triassic to the Earliest Cretaceous (Cadna-Owie Formation) is consistent with an exponential form with a time constant of between 50 and 200 Ma. This subsidence is attributed to the decay of a thermal perturbation in the lithosphere and the driving force is implicitly attributed to density changes as a result of cooling and/or phase changes. By the end of the deposition of the Cadna-Owie Formation, marine conditions existed over the Eromanga Basin. Also about this time, the subsidence rate increased significantly, and this feature is observed in all the wells used in the subsidence analysis. The short timescale (~20 m.y.) over which this rapid subsidence phase occurred and the abrupt cessation of deposition tends to preclude a large scale thermal process in the lithosphere. The increase in subsidence rate corresponds to a marked increase in the amount of volcanogenic detritus derived from a volcanic arc, assumed to be situated off the present day eastern coast of Queensland. The phase of subsidence is attributed to a two stage increase in the sediment influx by a factor of 10-40 greater than that expected to keep the basin in equilibrium with the thermally driven subsidence. The maximum rates of sediment influx do not exceed 170 m/m.y..

This model can provide an explanation for the apparent contradiction with global sea level curves which predict a continuously rising sea level during the Cretaceous while the sediments in the Eromanga Basin show two transitions from marine to paralic/non-marine conditions. These facies variations have previously been interpreted as global sea level transgression/regression cycles, but in the model proposed in this thesis the basin was twice filled near to, or over and above, the contemporary sea level. The sea level variations which have been proposed for the Canning and Officer Basins in Western Australia in the Early Cretaceous, but which apparently are not seen in the Eucla Basin, are also attributed to a more local mechanism (uplift) rather than a global sea level change. The overfilling of the Eromanga Basin results in a topography which is 100-200 m above that seen in the
central region, and some additional unknown mechanism, possibly related to rifting at the southern and eastern margins or pressure induced phase changes below the basin, needs to be invoked to explain the topographic discrepancy. A phase change is probably the more reasonable of the two mechanisms, but it is difficult to demonstrate convincingly that the transformations would proceed under the relatively small pressure perturbation due to the excess sediment loading.

When modelling thermal histories in evolving sedimentary basins, simplifying assumptions are commonly made, either as a result of lack of information or to facilitate the solution of the problem. Results from simple analytical models suggest that for intracratonic platform basins, where the width of the basin is usually two or three orders of magnitude greater than the depth, a constant background heat flow is unlikely to be perturbed by fluid flow except possibly at the margins where vertical movement may be significant. Also, in these broad basins rates of deposition are generally too low (< 50 m/m.y.) to cause more than a 5-10% departure from the steady state approximation for the temperatures in the sediments.

Thermal history analysis is not only of economic importance but may also provide useful constraints on tectonic processes involved in sedimentary basin formation. The methods available are likely to be somewhat insensitive to the early heat flow history in a fully developed basin spanning 200 m.y. or so. Additionally, the low temperature geothermometers need to be quantified in terms of reaction kinetics and other factors which may influence the results, for example the chlorine content of apatites used for fission track analysis. The calibration of experimental results needs refinement but, as more advanced understanding of the methods is achieved, the potential exists for combining the different approaches to improve the resolution of the thermal history in sedimentary basins. These techniques need to be further tested in the geological environment, most probably in developing Tertiary or younger sedimentary basins, e.g. the intra-Carpathian basins, (Sclater et al., 1980b) or recent continental margin basins, to assess their suitability for constraining geodynamic models.
In the study of the central Eromanga Basin region, application of the $^{40}\text{Ar}^{39}\text{Ar}$ method was restricted to only one granite sample from the basement of the Moomba 1 well because of the unsuitability of the feldspars in the sediments. The result does however support the conclusion drawn previously from apatite fission track and vitrinite reflectance studies. This is that there has been a recent elevation of the geothermal gradient by about 20-25%. This conclusion is intriguing but more work needs to be done to delineate the areal extent of the elevated geotherms. However, in the light of these results it seems unlikely that any Permo-Triassic period of elevated heat flow, to be expected if a thermal mechanism was controlling subsidence, will be detectable in the sediments. The physical mechanism responsible for the recent elevation in the heat flow has not yet been resolved. Qualitatively, it is most likely to be related to shallow (crustal) advective heat transport by intrusion or fluid movement rather than directly to a deeper event in the lower lithosphere, because of the conspicuous absence of associated geophysical phenomena such as a broad uplift.

The present day geothermal gradients inferred from BHT and DST data in exploration wells drilled in the Cooper-Simpson Desert Basin area are usually 30 to 60°C/km. Preferred estimates of the depth averaged thermal conductivity are in the range 2.0-2.2 Wm$^{-1}$K$^{-1}$ and the heat flow is estimated to be between about 75 and 120 mWm$^{-2}$. The highest values occur in the Nappamerri Trough, which is known to underlain in places by granite, and are also associated with shallow (<2000 m) basement ridges. Numerical modelling of the steady state thermal regime implies that the differences in heat flow may be explained primarily by lateral variations in heat production within the sub-sedimentary basement of the order of 3-10 µWm$^{-3}$. The higher heat production values (6-10 µWm$^{-3}$) are associated with granitic basement. In order to maintain the higher geothermal gradients observed in the Nappamerri Trough, it is concluded that much of the area is underlain by granite. Radiometric dating on granite samples encountered in drill holes suggest a Carboniferous (300-360 Ma) age. It is considered plausible that granite intrusion occurred in the southern Cooper Basin prior to sedimentation as a result of an elevated thermal regime in the lithosphere which subsequently decayed initiating thermally driven
subsidence. The granites are relatively localized in the southern Cooper Basin region and
the implication is that the thermal perturbation required to drive subsidence was deeper in
the lithosphere and the granite intrusions are a local manifestation of this event in the
shallow crustal part of the lithosphere.

In this thesis I have addressed only two aspects of the development of the Eromanga
and Cooper Basins. However, there is potential for a basin wide structural synthesis,
provided that sufficient good quality data could be obtained. One interesting feature of the
evolution of the Eromanga Basin is the large amount of Jurassic-Cretaceous subsidence in
the Simpson Desert/Poolowanna Trough area, although in comparison with the Cooper
Basin, data is scarce. As simple shear extension models have recently been quantified and
accepted, the question arises as to whether the Cooper Basin may represent a fault
controlled basin located asymmetrically with respect to the main region of thermal
subsidence, which occurred initially over the Simpson Desert/Cooper Basin region and
resulted in the thickest sequence of post-Triassic sediments in the Eromanga Basin. In the
Cooper Basin region the deposition of a thicker sequence would have been facilitated in part
by compaction of the underlying Permian and Triassic sediments. The structural evolution
of the Cooper Basin was complex and not obviously related to extension tectonics, as
suggested by a predominance of strike-slip, and some reverse, faulting. In this context it
would be relevant to examine the interaction between inplane forces acting on the
lithosphere, forces which may be induced locally due to processes acting below the basin
(e.g. convective thinning) and pre-existing structures in controlling the evolution of the
stress state and fault patterns of an intracratonic basin such as the Cooper Basin.

The deeper structure under the Eromanga Basin region may possibly be examined
through the gravity field. To reveal any lower crustal or mantle signals related to a driving
load requires an accurate representation of the sediment distribution and its contribution to
the gravity field. It would therefore be necessary to construct an internally consistent
basinwide isopach maps. Such a data set would also be useful in examining the role of
regional isostasy under long wavelength loads. Unfortunately, erosion of the basin
margins, where flexural effects would be most significant for a feature the size of the
Eromanga Basin, is likely to be a problem in constraining, for example, the flexural rigidity.

The present day thermal regime of the Eromanga Basin is certainly an outstanding problem. As some of the deeper sediments are currently at temperatures well over 200°C, the opportunity exists to compare and constrain geothermometers sensitive to different temperature ranges, provided suitable samples exist. For example, fission track analysis of apatite, zircon and sphene may yield not only independent information concerning their annealing properties, but also may be integrated to obtain a consistent regional picture of at least the recent thermal history of the Eromanga Basin. Helium isotope studies may provide some clues as to the cause of the recent thermal anomaly. $^3$He/$^4$He ratios are considered to be diagnostic with regard to the presence of a mantle component (high $^3$He, O’Nions and Oxburgh, 1983) which may be present if large scale magma of fluid migration from depth has occurred. Torgersen and Clarke (1987) have suggested that a mantle component can be detected in groundwater from four boreholes in the eastern Eromanga Basin, and a detailed study of the Cooper Basin region would be warranted.

It is unlikely that further subsidence analysis will provide extra insight into the formation mechanism of the Eromanga Basin, although such results may of course be interpreted in a different manner (cf. Middleton, 1978). In this thesis I have tried to emphasize that deposition may not necessarily always have a passive infill role, although generally sediment loading is less important than the overall tectonic mechanism. Variations in subsidence trends should be examined in the light of lithological and facies information, in addition to considering geodynamic models. Thus the major driving mechanism for the subsidence of the Eromanga and Cooper Basins is considered to be thermally based. However, as stated in the introduction to chapter 2, there are two other factors which can contribute to the overall evolution of a sedimentary basin - applied or induced inplane stresses and gravitational loading. These two factors are probably the dominant influences during the early (Permian) and final (Cretaceous) stages respectively of the development of the Eromanga/Cooper Basin region. The fault controlled subsidence during the Permian was strongly influenced by horizontal oblique stresses, possibly as a result of a convergent
plate margin to the east. The large influx of sediment in the Early-Middle Cretaceous can be considered to have acted as an extra (surface) load, as opposed to a passive infill, resulting in an additional contribution to the thermal driven tectonic subsidence.
Appendix 1

Thermal conductivity of sedimentary and basement rocks from the Eromanga and Cooper Basins, South Australia

Thermal Conductivity of Sedimentary and Basement Rocks from the Eromanga and Cooper Basins, South Australia

Kerry Gallagher

Research School of Earth Sciences
Australian National University
G.P.O. BOX 4
Canberra 2601

Abstract

Thermal conductivity measurements have been made on core samples from the Eromanga/Cooper Basin sequence in northern South Australia. A divided bar apparatus was used and the measurements were made with the samples dry and saturated with water. The results are compared with predictions made from theoretical models of thermal conductivity in porous rocks although they do not unequivocally support any particular model. The samples are divided into sandstones, siltstones and shales with effective matrix thermal conductivities of 6.0, 3.5 and 2.5 Wm⁻¹K⁻¹ respectively and it is suggested that a suitable predictive model is the average of the Maxwell and geometric mean models. These two are effectively the upper and lower bounds for the models considered.

Key words: Thermal conductivity, Eromanga/Cooper Basin

Introduction

It is desirable to have representative thermal conductivity values for different lithological types in order to quantify both present day and palaeogeothermal regimes. Values for thermal conductivity quoted in the literature can vary considerably for a given lithology (see Cermak and Rybach 1982). This paper presents results of thermal conductivity measurements made on sedimentary and basement rocks from the Eromanga/Cooper Basin sequence in northern South Australia (Fig. 1). The samples were measured in the dry (air saturated) state and subsequently in the wet (water saturated) state. The porosities of the samples range from effectively 0% to 33%. The effective thermal conductivity is a function of the thermal conductivities of the constituents making up the rock, including the fluid or gas present in the porespace. The observations presented here were tested against several theoretical models which attempt to quantify the thermal conductivity-porosity relationship. The effects of temperature and pressure on thermal conductivity were not examined in this study. However the thermal conductivity of quartz is sensitive to temperature variations (Ratcliffe 1959), and this effect is manifested in a reduction of the thermal conductivity of quartz rich sandstones with increasing temperature (Birch and Clark 1940, Sekiguchi 1984). Pressures of up to 1 kilobar (100 MPa) applied during measurement have a negligible effect on matrix thermal conductivity (Lubimova et al. 1977), although progressive burial of sediments results in compaction and reduction of porosity, resulting in changes to the bulk thermal conductivity of the rock.

Theoretical models of thermal conductivity

Even rocks with less than 1% porosity can show differences of 30% between the dry and wet conductivities (Scharl and

FIGURE 1
Location map for drill core samples. Ti—Tirrawarra, Gi—Gidgealpa, Me—Merrimbla, Bu—Burley, St—Strzelecki, Du—Dullingari. The town Moomba is also shown.

Rybach 1984). This is due to the contrast in the thermal conductivities of water and air (0.61 and 0.0264 Wm⁻¹K⁻¹ at STP respectively). For these low porosity rocks the water saturated value should be close to the actual rock (or matrix) thermal conductivity (Walsh and Decker 1966). In a simplified form the thermal properties of a porous rock can be represented by a two phase system composed of the solid matrix and the fluid or gaseous saturant in the pore space. Numerous mathematical models
based on this premise have been propounded to explain the observed variation of thermal conductivity in porous media. Seven of these models are given in Table 1 and the predicted conductivity:porosity relations for 6 of the models are shown in Fig. 2 for two values of effective matrix conductivity (2 and 7 Wm⁻¹K⁻¹) over a porosity range of 0–40%. The Nosal model (Table 1.g) is not illustrated as it is not solely a function of the porosity and matrix conductivity. The effective matrix conductivity represents the bulk conductivity of all the matrix constituents. The geometric mean model (Table 1.a), popular because of its mathematical simplicity, has been shown to be inappropriate for unconsolidated sands if the ratio of the matrix and saturant conductivities (KₘKₛ⁻¹) exceeds about 20 (Woodside and Messmer 1961), where Kₘ and Kₛ are the matrix and saturant thermal conductivities. Beck and Beck (1965) suggest that the model is probably unsuitable for all rocks where this ratio is greater than 10. Beck (1970) discussed the limitations of the Maxwell model (Table 1.b) and proposed an empirical correction to apply to the dry value predicted from wet observations. This correction increases with porosity and the KₘKₛ⁻¹ ratio. This results in the predicted dry value being reduced towards the geometric mean value. The statistical model proposed by Huang (1971) (Table 1.c) is dependent on the value of n, the pore geometric factor, and as n tends to zero the model approaches a simple arithmetic mean. Huang (1971) used a value of n = 1 for dry sandstones. Robertson and Peck (1974) preferred 0.5 for vesicular basalt and stated that the model was not useful for liquid saturants. A value of 0.1 is required for the model curves to be reasonable (i.e. more or less between the geometric mean and the Maxwell models for wet predictions). Similarly the dispersive model as adapted by Robertson and Peck (1974) (Table 1.d), the quadratic model (Table 1.e), and the Budiansky (1970) model (Table 1.f) are both bounded by these two model curves. In most problems of geophysical interest, and certainly within sedimentary basins the relevant property is the wet or water saturated thermal conductivity. Therefore a relationship between the dry and wet conductivities is of secondary importance if observations have been made in the wet state; it is only necessary to quantify a relationship between the bulk thermal conductivity and the porosity, which is assumed to be water saturated.

Method

Fig. 1 shows the location of the wells from which the samples were obtained. For a description of the geology of the area see Moore and Mount (1982) and Batterby (1976). The samples were divided into two groups. Set A consists of discs of 35 mm diameter and thickness between 1.5 and 10.0 mm

**FIGURE 2**

Curves for theoretical models given in the appendix for dry and wet states and matrix conductivities of 2 Wm⁻¹K⁻¹ (a,b) and 7 Wm⁻¹K⁻¹(c,d). M—Maxwell, B—Budiansky, Q—Quadratic, D—Dispersive model of Robertson and Peck (1974), G—Geometric, H—Huang (n = 0.1 for wet and 0.5 for dry).
TABLE 1.
Theoretical models of the effective thermal conductivity of porous media. Notation: K — thermal conductivity, \( \phi \) — fractional porosity (0.0-1.0), subscripts: b — Bulk or effective, m — matrix, s — saturant or fluid. (a)—Geometric, (b)—Maxwell (Beck 1976), (c)—Huang (1971), (d)—Dispersive (Robertson and Peck 1974), (e)—Quadratic (Robertson and Peck 1974, f)—Budiansky (1970), (g)—Nosal (1981).

<table>
<thead>
<tr>
<th>MODEL</th>
<th>Additional comments</th>
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<tbody>
<tr>
<td>(a) ( K_b = (K_m)^{1-\phi} (K_s)^{\phi} )</td>
<td>( r = K_m/K_s )</td>
</tr>
<tr>
<td>(b) ( K_b = K_m ((2s+1) - 2s(\tau-1)) ) ( (2s+1) + 4(\tau-1) )</td>
<td>( A = (1-\phi)exp(-n/(1-\phi)) ) ( B = 4\phi exp(-n/\phi) + 4(n-1)exp(-n/\phi) ) ( n = 1 - 4exp(-n/\phi) - 4\phi exp(-n/(1-\phi)) ) ( n ) is the geometric pore factor</td>
</tr>
<tr>
<td>(c) ( K_p/K_m = A + B(K_s/K_m) )</td>
<td>( K_{upper} = K_m + \frac{\phi}{(K_m + K_s)^{1-\phi} + (1-\phi)/3K_m} )</td>
</tr>
<tr>
<td>(d) ( K_b = \frac{1}{2}(K_{upper} + K_{lower}) )</td>
<td>( K_{lower} = K_s + \frac{(1-\phi)}{(K_m - K_s)^{1-\phi} + (1-\phi)/3K_s} )</td>
</tr>
<tr>
<td>(e) ( (K_b)^{1/2} = (1-\phi)(K_m)^{1/2} + \phi(K_s)^{1/2} )</td>
<td>( K_m &gt; K_s )</td>
</tr>
<tr>
<td>(f) ( \frac{(1-\phi)}{2K_m/K_b} + \frac{\phi}{2K_s/K_b} = \frac{1}{3} )</td>
<td></td>
</tr>
<tr>
<td>(g) ( K_b = FK_m + (1-F)K_m^{1-GK_s^G} )</td>
<td>( F = K_{dry}/K_m ) ( G = \phi/(1-F) )</td>
</tr>
</tbody>
</table>

were cut at the ANU from drill core samples. The axes of the discs were parallel to the axis of the drill core and, considering the low dips of the sediments (<5°), approximately perpendicular to the bedding. Set B were cylinders with their long axes in the plane of the bedding. The cylinders were between 30.0 and 50.0 mm long and 25.0 mm diameter. The porosity and permeability of set B samples had been previously measured by helium injection (Gravestock and Alexander 1986).

All conductivity measurements were made on a divided bar apparatus similar to that described by Sass et al. (1971) (Fig. 3). The upper and lower ends of the bar were kept at 34±0.1°C and 18±0.1°C respectively. The ambient room temperature varied between 20 and 30°C during the period of measurement (about two months). This system takes about 20–25 minutes to reach thermal equilibrium (Sass et al. 1984) although generally thermistor readings were taken after 45–60 minutes.

Sass et al. (1971) discuss some of the practical problems associated with the operation of a divided bar apparatus. The errors introduced as a result of radial heat loss from samples have been examined by Jessop (1970). These were minimised by lagging the bar with cotton wool and styrofoam. Another potential problem arises as a consequence of an imperfect thermal contact, referred to as a contact resistance, between the sample and the adjacent copper disc (see Fig. 3). To reduce the effect of the contact resistance the samples were cut and polished flat to within ±0.03 mm. During polishing the faces of the porous samples, especially the coarser grained sandstones often became pitted due to plucking of the grains. This would probably increase the contact