

Paleothermometry of the Sydney Basin

M. F. MIDDLETON¹

CSIRO Division of Fossil Fuels, North Ryde, New South Wales, 2113, Australia

P. W. SCHMIDT

CSIRO Division of Mineral Physics, North Ryde, New South Wales, 2113, Australia

Evidence from overprinting of magnetizations of Late Permian and Mesozoic rocks and from the rank of Permian coals and Mesozoic phytoclasts (coal particles) suggests that surface rocks in the Sydney Basin, eastern Australia, have been raised to temperatures of the order of 200°C or higher. As vitrinite reflectance, an index of coal rank or coalification, is postulated to vary predictably with temperature and time, estimates of the paleotemperatures in the Sydney Basin based on observed vitrinite reflectance measurements can be made in conjunction with reasonable assumptions about the tectonic and thermal histories of the basin. These estimates give maximum paleotemperatures of present day surface rocks in the range 60–249°C, depending on factors such as location in the basin, the thickness of the sediment eroded, and the maximum paleogeothermal gradient. Higher coal rank and, consequently, larger eroded thicknesses and paleogeothermal gradients occur along the eastern edge of the basin and may be related to seafloor spreading in the Tasman Sea on the basin's eastern margin. A theory of thermal activation of magnetization entailing the dependence of magnetic viscosity on the size distribution of the magnetic grains is used to obtain an independent estimate of the maximum paleotemperatures in the Sydney Basin. This estimate places the maximum paleotemperature in the range 250–300°C along the coastal region. Both coalification and thermal activation of magnetization models provide strong evidence of elevated paleotemperatures, which in places exceed 200°C, and the loss of sediment thicknesses in excess of 1 km due to erosion.

INTRODUCTION

There is strong evidence that rocks exposed at the surface of the Sydney Basin may have been raised to temperatures of 200°C or higher. This evidence is derived from observations of overprinting of magnetizations in Late Permian and Early Mesozoic rocks which were rapidly cooled 70–100 m.y. ago [Schmidt and Embleton, 1981]. The blocking and stabilization of the overprint magnetizations are related to uplift, erosion, and rapid supracrustal cooling. High surface coal ranks in the Sydney Basin support these conclusions.

Important information on tectonic and thermal histories of sedimentary basins can be obtained from the degree of diagenesis of organic sediments. The reflectivity of the coal maceral, vitrinite, is known to increase regularly with coal rank, which is itself related to the thermal and tectonic history of a basin. A number of workers have demonstrated that organic diagenesis proceeds predictably as a function of temperature variation with time and that this relationship can be used to predict organic diagenesis (or coal rank) in terms of vitrinite reflectance in a developing sedimentary basin [Shibaoka and Bennett, 1977; Turcotte and McAdoo, 1979; Royden et al., 1980; Falvey and Middleton, 1981]. Table 1 shows the correlation of coal rank to the reflectance of vitrinite in oil.

This paper combines a tectonic burial model of the Sydney Basin with the thermodynamic model of coalification in order to corroborate the findings of Schmidt and Embleton [1981]. Further evidence for burial of the basin comes from petrographic studies by Raam [1968], who found the regional oc-

currence of the metamorphic minerals, laumontite and prehnite, in Permian sandstones. Sedimentation probably extended into Jurassic times, as sediments and tuff with Early Jurassic microflora have been reported in several breccia pipes within the basin [Crawford et al., 1980]. A major uplift and erosional event, which denuded the basin of its Jurassic and part of its Triassic sediments, was probably associated with the initiation of seafloor spreading in the Tasman Sea on the eastern margin of the Sydney Basin.

REGIONAL GEOLOGY

The Sydney Basin (Figure 1) is one of an interconnected network of Permo-Triassic basins in eastern Australia. The basin has an unknown maximum thickness of sediment, probably exceeding 4 km, which thins towards the western and southern margins. The depositional and tectonic history of the basin has been well described in the literature [Conolly and Ferm, 1971; Mayne et al., 1974; Herbert, 1980].

Thick sequences of marine sediments were deposited in the Sydney Basin during the Early Permian; a transgression in the northern part of the basin was followed by a regression that allowed the deposition of the Greta Coal Measures, about 270–260 m.y. ago. Deposition in the Late Permian and Early Triassic took place essentially in marginal to fluvial environments with numerous coal swamps. Deposition is postulated to have extended into Jurassic times with an hiatus in the Late Triassic [Herbert, 1980]. A north-south trending 'hinge line' (Figure 1), which is suggested to be governed by Pre-Permian tectonics (H. J. Harrington, personal communication, 1981), separates a thick sequence of sediments to the east from a thin sequence on the shelf to the west.

Volcanism and intrusive activity occurred continuously in the Sydney Basin through the Triassic and Jurassic [Mayne et al., 1974; Wellman and McDougall, 1974; Crawford et al., 1980]. A regional geothermal gradient higher than the continental

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TABLE 1. Correlation of Vitrinite Reflectance to Coal Rank

Coal Rank	Mean Vitrinite Reflectance in Oil, %
Peat-lignite	0.2-0.4
Sub-bituminous	0.4-0.6
High volatile bituminous	0.6-1.1
Medium volatile bituminous	1.1-1.5
Low volatile bituminous	1.5-1.9
Semi-anthracite	1.9-2.5
Anthracite	2.5-6.0

average of 1.2 HFU (10^{-6} cal cm^{-2} s^{-1}) probably existed throughout this time, with localized geothermal gradients fluctuating considerably.

Uplift associated with the onset of seafloor spreading [Falvey, 1974] in the Tasman Sea probably instigated an erosional episode which denuded the Sydney Basin of Jurassic and some Triassic sediments. Seafloor spreading in the Tasman Sea commenced 76 m.y. ago [Shaw, 1978, 1979], which compares favorably with the blocking ages of the overprint magnetizations measured by Schmidt and Embleton [1981].

Some erosion and igneous activity extended through the Tertiary; a significant uplift event caused elevation of the western part of the Sydney Basin. However, this later event would have had little effect on temperatures within the basin, as erosion occurred by rivers dissecting the uplifted plateaus and negligible temperature and coal rank increase would have occurred since the major erosional event associated with Tasman Sea opening some 70 m.y. earlier.

MODEL OF COALIFICATION

Coalification is commonly modeled as a first-order chemical reaction with the reaction rate doubling every 10°C temperature increase. Royden *et al.* [1980] have related a thermal alteration parameter, C , to temperature and time with the above assumption of reaction rate:

$$C = \ln \int_0^t 2^{T(t)/10} dt \quad (1)$$

where $T(t)$ is a temperature as a function of time t . They relate vitrinite reflectance in oil, R_0 , to C , graphically; their graph describes the relation

$$\log(R_0) = b' + a'C \quad (2)$$

where a' and b' are constants. Substitution of (1) into (2) and rearrangement of terms gives

$$(R_0)^a = B \int_0^t \exp[bT(t)] dt \quad (3)$$

where R_0 is vitrinite reflectance in percent, T is temperature in $^\circ\text{C}$, t is time in m.y., $a = 5.5$, $B = 2.8 \times 10^{-6}$, and $b = 0.069$. Shibaoka and Bennett [1977] proposed a graphical method to determine coal rank increase with temperature and time which is equivalent to (3) with slightly different values of the constants a , b , and B .

Dow [1977] described the common observation that a plot of the logarithm of vitrinite reflectance, $\log(R_0)$, versus depth for deep bores or petroleum exploration wells produces a linear curve. He suggested that linearity of the $\log(R_0)$ versus depth plot may be an inherent feature of the coalification process. These observations cannot be generally reconciled with (3). However, for a rapidly deposited sequence subsequently main-

tained undisturbed for its remaining geological history of duration t_2 with a geothermal gradient G_2 , (3) becomes

$$(R_0)^a = B \int_0^{t_2} \exp[b(T_0 + G_2z)] dt \quad (4)$$

where T_0 is the surface temperature and z is depth; the coalification during subsidence is neglected. Integrating the right-hand side of (4) and rearranging terms,

$$\log(R_0) = \log[Bt_2 \exp(bT_0)]^{1/a} + \{[bG_2 \log(e)]/a\}z \quad (5)$$

which is a logarithmic relationship of the form suggested by Dow [1977]. This has been discussed previously by Middleton [1982].

TECTONIC MODEL

Middleton [1982] has used (3) to give predicted vitrinite reflectance versus depth relationships for a number of simple tectonic histories. The Sydney Basin tectonic history is essentially one of subsidence through Permian to Early Jurassic times, stable conditions from the Late Jurassic to 100-70 m.y. ago, followed by rapid uplift and erosion, and then minor vertical movements until a final uplift in Late Tertiary times. Events after the rapid uplift and erosion had little effect on coal rank, and may be modeled as a period of quiescence with low geothermal gradient. The tectonic model is (1) subsidence at a rate V for a duration t_1 with geothermal gradient G_1 , (2) no

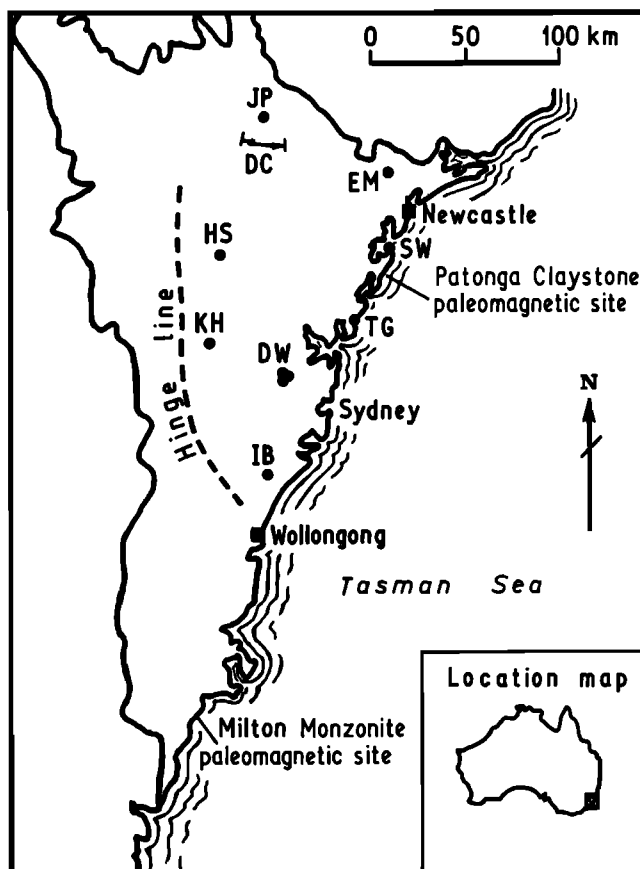


Fig. 1. Location map of the Sydney Basin, showing the locations of nine sites from which vitrinite reflectance versus depth data were obtained. Wells/bores indicated as follows: EM, East Maitland 1; JP, Jerrys Plains 1; DC, Doyles Creek bores; HS, Howes Swamp 1; SW, Swansea; TG, Terrigal 1; DW, Dural wells; IB, Illawarra bores; KH, Kurrajong Heights 1.

vertical movement for a duration t_2 with geothermal gradient G_2 , (3) rapid erosion of thickness h in response to rapid uplift with geothermal gradient G_2 , and (4) quiescence to the present day (a duration t_3) with geothermal gradient G_3 less than or equal to G_2 .

The vitrinite reflectance as a function of depth z for the above tectonic history is given from (3) by

$$(R_0)^a = B \exp(bT_0) \{ \exp[bG_1(z+h)] - 1 \} / bG_1 V + Bt_2 \exp[bT_0 + bG_2(z+h)] + Bt_3 \exp[bT_0 + bG_3z] \quad (6)$$

Unless $G_1 = G_2 = G_3$ in (6), $\log(R_0)$ versus depth z is not linear. However, an examination of the magnitude of the terms in (6) reveals that the term containing t_2 and G_2 is typically an order of magnitude larger than the other terms. Therefore the relationship of $\log(R_0)$ to depth is very nearly linear.

Oscillograms or depositional history plots of Sydney Basin stratigraphic data from deep bores [Mayne *et al.*, 1974] indicate that a constant subsidence rate gives a good approximation to Permian and Triassic subsidence in many bores. The subsidence rate is variable throughout the basin, being about $0.05 \text{ km m.y.}^{-1}$ in the north and center of the basin and about $0.03 \text{ km m.y.}^{-1}$ in the south. Turcotte and Ahern [1977] and McKenzie [1978] have demonstrated that geothermal gradients in newly forming basins are significantly higher than stable continental regions. The heat flow might be expected to be in the order of 2.5 HFU , which gives a geothermal gradient of 50°C km^{-1} if a sediment thermal conductivity of $5 \times 10^{-3} \text{ cal cm}^{-1} \text{ s}^{-1} \text{ }^\circ\text{C}^{-1}$ is assumed. The present day geothermal gradient G_3 in the Sydney Basin is approximately 24°C km^{-1} [Sass *et al.*, 1976], which has probably been maintained for the past 70 m.y. (t_3).

The geothermal gradient G_2 from the end of deposition to the uplift associated with Tasman Sea rifting was higher than the continental average of 24°C km^{-1} but probably fluctuated considerably with periods of igneous activity. The duration of this period (t_2) is unknown and depends on the age at which sedimentation ceased and the age at which the rapid uplift and erosion began. Schmidt and Embleton [1981] place the age of uplift and erosion between 100 and 70 m.y.: the youngest microflora reported in the Sydney Basin is Early Jurassic [Helby and Morgan, 1979]. Therefore the period t_2 is probably in the range of 70–110 m.y.

Thickness of erosion, h , is well in excess of 1 km [Schmidt and Embleton, 1981]. The relative orders of magnitude of the terms on the right-hand side of (6) are determined at a depth $z = 0$, assuming $T_0 = 15^\circ\text{C}$, $V = 0.05 \text{ km m.y.}^{-1}$, $G_1 = 50^\circ\text{C km}^{-1}$, $t_2 = 100 \text{ m.y.}$, $G_2 = 40^\circ\text{C km}^{-1}$, $t_3 = 80 \text{ m.y.}$, $G_3 = 24^\circ\text{C km}^{-1}$, and $h = 1 \text{ km}$:

$$B \exp(bT_0) \{ \exp[bG_1(z+h)] - 1 \} / bG_1 V = 1.39 \times 10^{-3}$$

$$Bt_2 \exp[bT_0 + bG_2(z+h)] = 1.25 \times 10^{-2} \quad (7)$$

$$Bt_3 \exp[bT_0 + bG_3z] = 6.31 \times 10^{-4}$$

The last term in (7) is considerably smaller than the other two terms and indicates that the coalification gradient is effectively 'frozen' into the basin at the time of the rapid erosion. The second term in (7) dominates by an order of magnitude, and therefore (6) is well approximated by

$$(R_0)^a = Bt_2 \exp[bT_0 + bG_2(z+h)] \quad (8)$$

which is equivalent to (5). For typical Sydney Basin tectonic parameters, a linear $\log(R_0)$ versus depth curve is to be expected.

The slope of the $\log(R_0)$ versus depth plot is a function of the geothermal gradient G_2 :

$$\frac{d \log(R)}{dz} = \frac{bG_2}{a} \log(e) = \frac{G_2}{183.5} \quad (9)$$

where depth is measured in kilometers. Consequently, an estimate of G_2 which is independent of the other parameters can be obtained from the data.

The reflectance at the surface R_{0s} , given by (8), is

$$R_{0s} = [Bt_2 \exp(bT_0 - bG_2h)]^{1/a} \quad (10)$$

Surface vitrinite reflectance is easily measured or determined by extrapolation of the linear $\log(R_0)$ versus depth plot, G_2 is determined by the slope of the data plot, and the range of t_2 is known. Therefore a range of calculated values for the eroded thickness h can be found from (10) for a set of vitrinite reflectance versus depth data from a deep bore. Middleton [1982] has applied this procedure to data from the Howes Swamp well in the Sydney Basin.

The inherent problem with this model is that a constant geothermal gradient G_2 is assumed to exist over the time span t_2 , a time span of the order of 100 m.y. Clearly, G_2 can only represent an 'effective' paleogeothermal gradient, which will be the averaged effect of the fluctuating geothermal gradient during the time span t_2 . The effective paleogeothermal gradient G_2 is the only clue to the paleothermal conditions during the period of major coalification. However, the fluctuating paleogeothermal gradient immediately prior to uplift, which was probably the maximum paleogeothermal gradient, influenced the overprint magnetization directions rather than G_2 . Nevertheless, G_2 can provide a good estimate of h when substituted into (10).

VITRINITE REFLECTANCE

Figure 1 shows the locations of nine deep bore sites in the Sydney Basin from which vitrinite reflectance versus depth data were obtained. Figure 2 shows plots of $\log(R_0)$ versus depth of these data. Curves for East Maitland 1, Swansea, and Terrigal 1 are from Diessel [1973]. The data are sparse from Kurrajong Heights 1 in the central west of the basin, but they are all that are presently available. A straight line has been drawn visually through the East Maitland and Swansea curves, and least squares regression has been used to fit straight lines to the other data. The regression equations and effective paleogeothermal gradient G_2 from (9) for each of the nine wells are shown in Table 2.

TABLE 2. Regression Equations and G_2 from (9) for the Nine Deep Bore Sites in the Sydney Basin

Bore/Well	Regression Equation, $\log(R_0)$	G_2 , $^\circ\text{C km}^{-1}$
East Maitland 1	$-0.108 + 0.145z$	26.6
Jerrys Plains 1	$-0.193 + 0.266z$	48.8
Doyles Creek bores	$-0.146 + 0.205z$	37.6
Howes Swamp 1	$-0.339 + 0.240z$	44.0
Swansea	$-0.150 + 0.243z$	44.7
Terrigal 1	$-0.172 + 0.240z$	44.0
Dural wells	$-0.192 + 0.343z$	62.9
Illawarra bores	$+0.045 + 0.221z$	40.6
Kurrajong Heights 1	$-0.349 + 0.281z$	51.5

The first coefficient in the regression equation is the logarithm of the surface vitrinite reflectance, $\log(R_0)$.

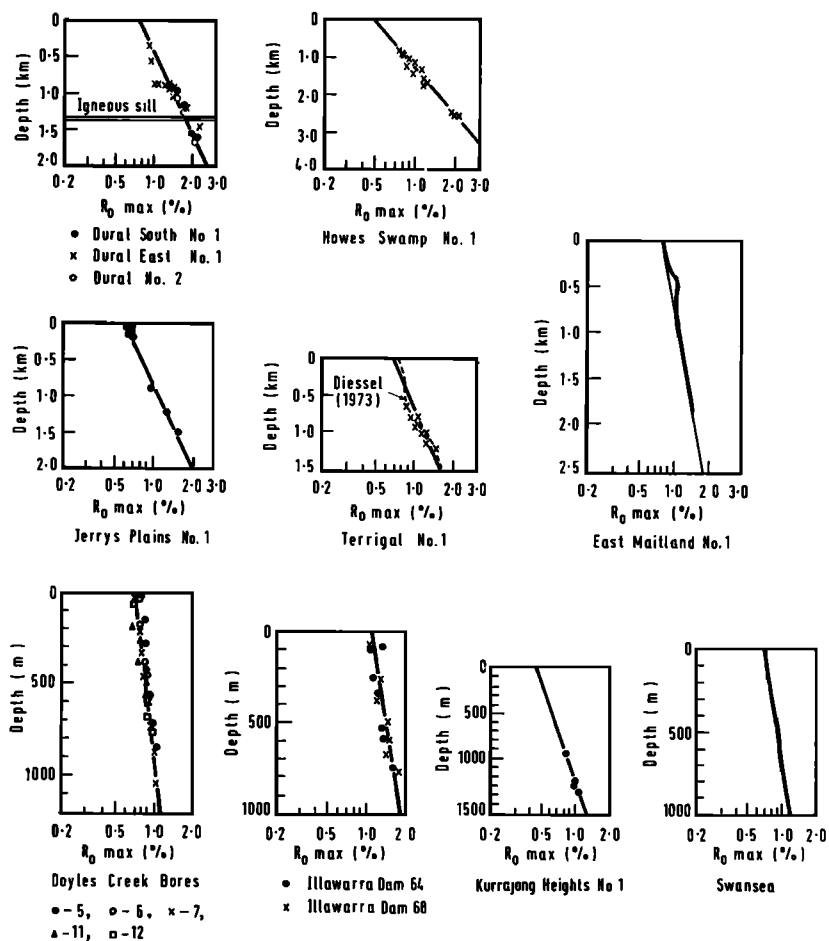


Fig. 2. Plots of the logarithm of vitrinite reflectance, $\log(R_o)$, versus depth z for the nine sites. Regression curves are shown plotted to the data; the equations of these curves are given in Table 2.

Figure 3 shows the effective paleogeothermal gradient for the nine Sydney Basin sites. The gradients are highly variable and insufficient data exist to predict any trends. However, it appears that an effective paleogeothermal gradient of approximately 40°C km^{-1} existed regionally throughout the basin.

A contour map of surface vitrinite reflectance is shown in Figure 4. This map is based on extensive reflectivity data (in the CSIRO data base) from the numerous collieries in the Sydney Basin, and extrapolation of data from shallow and deep bores. The surface vitrinite reflectance decreases with distance from the coast, and a peak occurs on the coast between Sydney and Wollongong.

THERMAL ACTIVATION OF MAGNETIZATION

Turning to magnetic parameters, the relationship between the relaxation time t of the remanent magnetization of an assembly of noninteracting single-domain grains proposed by Néel [1949] is

$$t = \frac{1}{2} f_0 \exp [vJ_s(T_B)H_K(T_B)/2kT_B] \quad (11)$$

where f_0 is the fluctuation rate (approximately 10^{10} s^{-1}), v is volume, J_s is the spontaneous magnetization, H_K is the intrinsic coercivity, k is Boltzmann's constant, and T_B is the blocking temperature (K), below which t dramatically rises. The fundamental implication is that for an assembly of magnetic grains there exists a time temperature (t, T) field governing the remo-

bilization or activation of the magnetization. For a particular grain the quantity $T \ln(2f_0t)/J_s(T)H_K(T)$ is a constant, yielding a thermal activation (TA) contour in the (t, T) field. TA nomograms for assemblies of pure magnetite and pure hematite grains [Pullaiah *et al.*, 1975] show that it is possible to activate the moments of such grains with moderate laboratory blocking temperatures at much lower temperatures, given sufficient time. This corresponds to their field, B , and gives an explanation for viscous remanent magnetization (VRM). On the other hand, grains with high laboratory blocking temperatures are stable for geologically long times even at elevated temperatures, i.e., their field A . However, experimental evidence produced by Pullaiah *et al.* suggests that thermal activation may be more potent than theory allows, and as Dunlop and Buchan [1977] conclude, such findings cast serious doubt on the quantitative use of this theory, at least until the reason for the discrepancy is determined. Thus, considering magnetic overprinting alone, Schmidt and Middleton [1981] were only able to suggest maximum temperatures to which the rocks now exposed in the Sydney Basin have been elevated since deposition.

Using a different approach, Walton [1980] recently developed a theory dependent on grain size distribution in which TA contours represent constant values of $T \ln^2(2f_0t)/(J_s(T)H_K(T))$, assuming an approximately log normal grain size distribution. Thermal activation diagrams thus generated are plotted for pure magnetite and hematite in Figures 5a and 5b. The usual

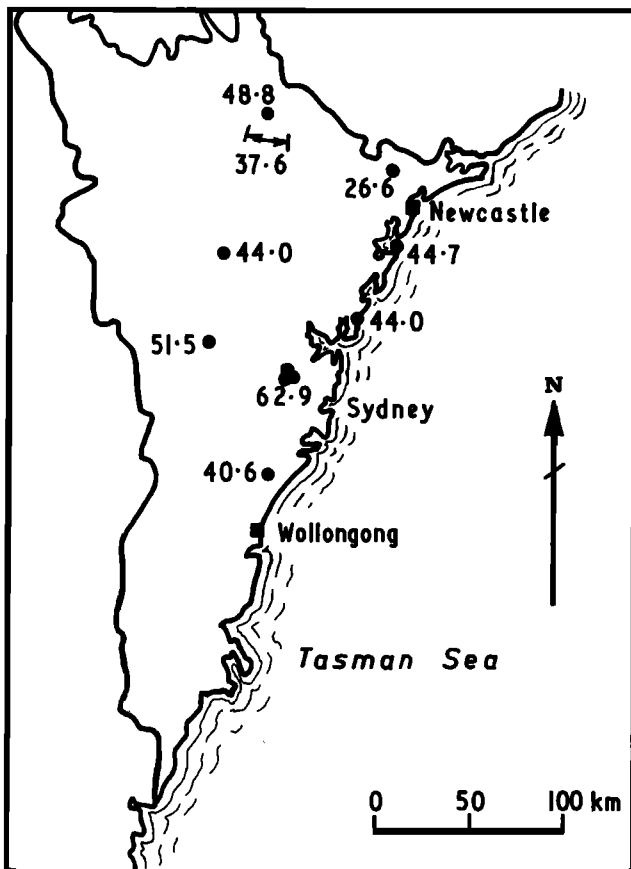


Fig. 3. The effective paleogeothermal gradient at each of the nine Sydney Basin sites.

assumption that shape anisotropy dominates in nonspherical magnetite grains leads to the relationship $H_K(T) \propto J_s(T)$, and the relationship $H_K(T) \propto J_s^3(T)$ was used for hematite, following Pullaiah *et al.* [1975]. Values of $J_s(T)$ have been taken from Figure 2 of Pullaiah *et al.* [1975]. The resulting contours reveal diminished A fields, compared with those of Pullaiah *et al.* That is, Walton's theory is more closely in accord with the experimental results, and suggests that the earlier discrepancy may be attributable to grain size distribution.

Applying Walton's theory for VRM acquisition at elevated temperatures to the existing thermal demagnetization data from overprinted rock bodies in the Sydney Basin should provide tighter temperature constraints than those previously deduced [Schmidt and Embleton, 1981]. Since the temperature dependence of spontaneous magnetization, $J_s(T)$, is best known for pure magnetite and hematite, only data from rocks bearing these minerals have been considered. Schmidt and Embleton showed that the only detectable magnetic carrier in the Milton Monzonite is magnetite with a Curie temperature of 578°C, while the magnetic carrier in the Patonga Claystone is hematite with a Curie temperature in excess of 660°C [Embleton and McDonnell, 1980]. Furthermore, the direction of the overprint magnetizations of these two bodies are parallel, strongly suggesting they have been caused by the same event. If overprint unblocking temperatures were precisely known, a unique (t, T) solution for the Sydney Basin could be achieved by solving simultaneously for different minerals; in this instance, magnetite and hematite. However, by the very nature of the stepwise acquisition of thermal demagnetization data, it is difficult

to define the exact temperature at which a magnetic component of any particular specimen is demagnetized, but by averaging over many specimens we consider that the actual laboratory temperatures are estimated to better than 10°C. Error from this source could be reduced by using smaller increments during thermal demagnetization and/or by processing a larger number of specimens. Another source of error arises from not knowing the effective time of the heating cycle. This is minor, however, because it is the logarithm of time that is involved in the analysis. In fact, because the log is involved, this factor may be the best known variable. Thus the laboratory (t, T) fields for magnetite and hematite may each be transformed into a range of possible geophysical counterparts, their intersection yielding an estimate of the actual geophysical conditions (t_a, T_a) .

The above procedures have been applied to the data from the Milton Monzonite and Patonga Claystone, yielding the result of a thermal pulse estimated to last several millions of years, raising the temperatures of these bodies to around 250–300°C (Figure 5c). The (t, T) fields for these bodies are estimated as 5 min, 480°C and 5 min, 590°C, respectively. Specimens are usually maintained for 5 min after equilibration at a given temperature. Both the Milton Monzonite and Patonga Claystone are located in the coastal region of the Sydney Basin (Figure 1).

The overlapped areas of Figure 5c reflect the error in estimating laboratory reactivation temperatures and yield a spectrum of theoretically possible solutions, many of which can be rejected on practical grounds. For instance, the solutions with

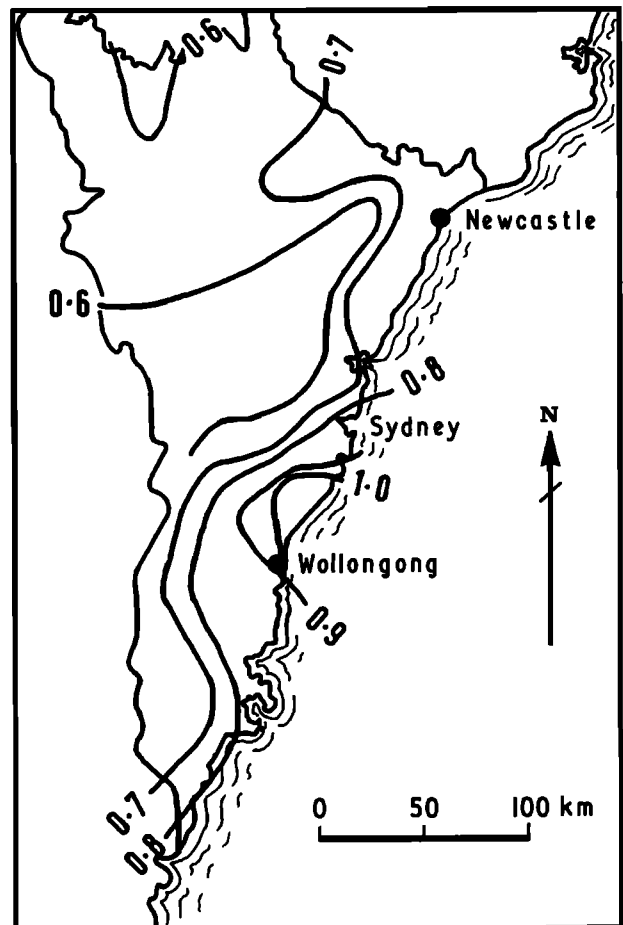


Fig. 4. Contours of vitrinite reflectance for the surface of the Sydney Basin.

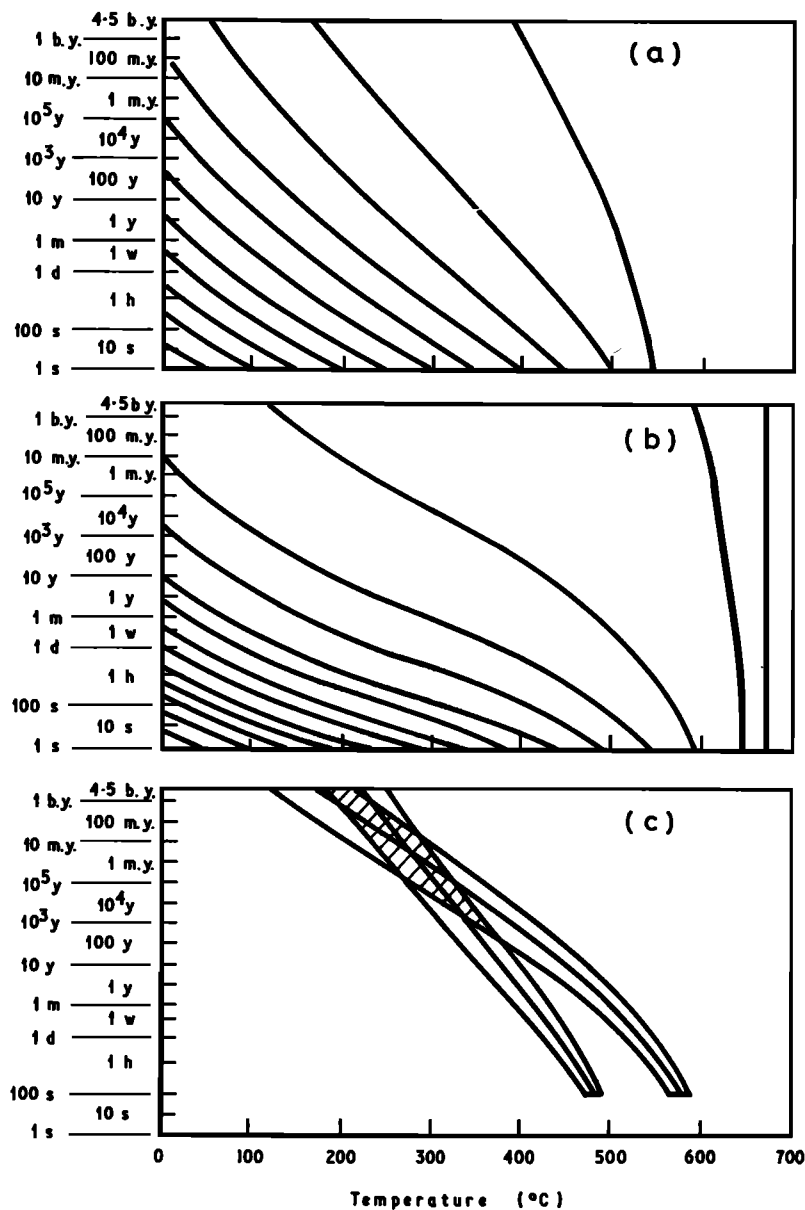


Fig. 5. Thermal activation nomograms following Walton [1980] for (a) magnetite, (b) hematite, and (c) rocks bearing these minerals from the Sydney Basin. The contours join time-temperature conditions of equal probability of activating a magnetic grain. Extrapolating laboratory conditions to geological conditions, paleotemperatures may be estimated if demagnetization data from more than one magnetic mineral are available.

durations t of 50 m.y. or more must be considered to be virtually impossible because the geomagnetic field is known to have reversed its polarity more frequently, the Cretaceous quiet zone being approximately of 30 m.y. duration [Irving and Couillard, 1973]. In addition, at the other extreme, a short duration ($<10^4$ years) implies temperatures in excess of 350°C , which are inordinately high, considering that a concomitant greenschist facies regional metamorphic grade is not observed in the Sydney Basin. The midrange of values ($\sim 250^\circ\text{C}$) therefore indicates the most probable solutions using Walton's theory. More precise estimates could be achieved using a larger sample size collected specifically for the purpose of geothermometry. It is considered, however, that the present data make a very useful comparison with the reflectivity results.

RESULTS

Estimates of the thickness of eroded sediment, h , at the nine sites calculated from (10) with the probable lower and upper limits of t_2 are given in Table 3. Values of G_2 and R_{0s} are obtained from Table 2, and T_0 is assumed to be 15°C . The h values in Table 3 are essentially the depths of the present day surface immediately prior to erosion. Also shown in the table are temperatures at the depth h immediately before the rapid erosion and ages of the surface rocks at the nine sites for comparison with eroded thicknesses. Except for Jerrys Plains 1, which is located near the culmination of a major anticline in the north of the basin, all the other sites with Permian surface rocks have lost at least 1.5 km of younger sediments. The temper-

TABLE 3. Thickness of Eroded Sediment h from (10) and the Temperature at the Depth h Immediately Prior to Erosion

Bore/Well	$t_2 = 70$ m.y.			$t_2 = 110$ m.y.			Age of Surface Rocks
	h , km	T , °C		h , km	T , °C		
		A	B		A	B	
East Maitland 1	3.34	104	249	3.10	97	232	middle Late Permian
Jerrys Plains 1	1.50	88	120	1.36	81	110	middle Late Permian
Doyles Creek bores	2.17	97	167	1.99	90	154	late Late Permian
Howes Swamp 1	1.06	62	89	0.92	55	79	late Early Triassic
Swansea	1.80	95	141	1.65	89	131	late Late Permian
Terrigal 1	1.75	92	123	1.60	85	127	late Early Triassic
Dural wells	1.16	88	96	1.06	82	89	early Middle Triassic
Illawarra bores	2.88	132	217	2.76	127	208	early Early Triassic
Kurrajong Heights 1	0.88	60	77	0.75	54	68	late Early Triassic

Temperatures in column A are calculated from G_2 in Table 2, and temperatures in column B are calculated assuming a geothermal gradient of 70°C km^{-1} immediately before erosion. Eroded thicknesses and temperatures are calculated for the probable lower and upper limits of t_2 . The age of surface rocks at each site is shown for comparison with eroded thickness.

atures in column A of Table 3 are effective paleotemperatures of the present day surface rocks calculated at depth h from the effective paleogeothermal gradient G_2 . They are probably considerably less than the maximum paleotemperatures and, like G_2 , represent an averaging of the temperatures experienced at the depth h prior to erosion.

Observed geothermal gradients near incipient or young rifts may be used to give estimates of temperatures which may have influenced magnetization at the depths in Table 3. Heat flow in the Rio Grande rift is in the range 2.1–3.6 HFU, with values falling rapidly beyond 50 km from the rift boundary [Decker and Smithson, 1975]. Erickson *et al.* [1975] report heat flow values in the Red Sea, a newly formed active spreading center, to be 3.5–3.6 HFU. On the basis of these data, it is not unreasonable to assume a heat flow of 3 HFU along the coastal region of the Sydney Basin during the early stages of Tasman Sea opening. Assuming a sediment thermal conductivity of $5 \times 10^{-3} \text{ cal cm}^{-1} \text{ s}^{-1} \text{ }^\circ\text{C}^{-1}$, the geothermal gradient immediately before erosion was about 70°C km^{-1} .

Temperatures in column B of Table 3 show paleotemperatures immediately before erosion, assuming the geothermal gradient to be 70°C km^{-1} . It may be noted that these estimates are 50–100°C lower than those derived from the magnetic overprinting. This suggests that the assumed geothermal gradient of 70°C km^{-1} might be lower than the actual gradient or that one of the assumptions made in applying Walton's model (such as a log normal grain size distribution or even that both the Patonga Claystone and the Milton Monzonite have suffered similar thermal histories) to the magnetic overprint data may have been violated. We should emphasize at this point that although the thermal activation expressions commonly in use were developed for single domain particles, it is valid to extend the expressions to cover more complex domain structures if it is understood that the v from (11) is generalized to V_{act} , the activation volume, rather than the grain volume, and if the appropriate variation of coercive force with temperature is used [Wohlfarth, 1961; Dunlop and Bina, 1977]. The experimental results of Tucker and O'Reilly [1980] from multidomain titanomagnetites indicate that except for temperatures in the vicinity of the Curie point, the coercive force varies similarly to that of single domains. Although the rocks studied here may contain multidomain grains (in fact, Schmidt and Embleton [1981] show

that the Milton Monzonite does contain multidomain grains), for the temperature range of interest the single domain theory should provide a good estimate of temperature-time variables.

From the above, it is clear that upwards of 1 km of overburden has been removed from the Sydney Basin, whether one uses the reflectance data or the overprint data. A short period (5–10 m.y.) with the geothermal gradient elevated $10\text{--}20^\circ\text{C km}^{-1}$ greater than G_2 immediately before the erosional event will have little effect on coalification but a major effect on magnetization. The temperatures in column B of Table 3 indicate that the surface rocks at two of the near coastal sites are in the vicinity of 200°C .

DISCUSSION

An important result of the present modelling of coalification in the Sydney Basin is that the exposed rocks at various localities along the coastal region may well have been elevated to temperatures of 220°C or more, as originally suggested from paleomagnetic studies. The surface vitrinite reflectance map (Figure 4) gives a relative indication of the areas of greatest erosion, as more deeply buried rocks have greater coal rank. An accurate delineation of basin wide erosion cannot be given without good knowledge of the regional variation in effective paleogeothermal gradient, the gradient influencing coalification as opposed to the maximum paleogeothermal gradient immediately prior to erosion, which influences magnetization. More data is needed to define accurately the effective paleogeothermal gradient throughout the Sydney Basin. Nevertheless, the present data indicate that a thickness of sediment of between 1 and 2 km has been regionally lost. Locally, thicknesses of up to 3 km may have been lost (East Maitland 1 and Illawarra bores).

The significance of the effective paleogeothermal gradient G_2 , as opposed to the instantaneous paleogeothermal gradient, is that coalification depends on a long duration of heating whereas overprint magnetizations tend to reflect elevated temperatures. For significant coal rank increase at temperatures below 350°C , the coal must be maintained at that temperature for some tens of millions of years. Figure 6 shows the response of vitrinite reflectance to increase of temperature which is maintained for periods of a month or less [Bostick, 1971]. The instantaneous response of vitrinite reflectance is inconsistent

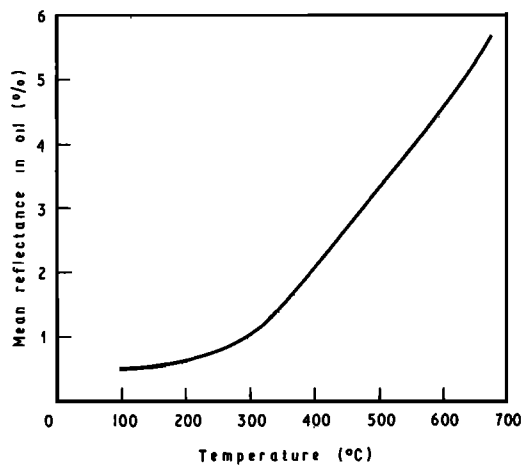


Fig. 6. Vitrinite reflectance versus temperature for coal and anthracite bombs maintained at a given temperature for up to a month (based on data published by Bostick [1971]).

with the long time response (3); a unifying hypothesis of coalification does not presently exist. It is clearly seen that temperatures in excess of 350°C are required for an instantaneous (in a geological sense) response of coal rank to temperature. Therefore the instantaneous thermal conditions are not as critical as the long-term conditions for the determination of the degree of coalification and contingent tectonic information. To determine the depth of burial (eroded thickness h) from coalification considerations using the effective paleogeothermal gradient G_2 and then to determine the instantaneous paleotemperature at this depth of burial using an appropriate instantaneous paleogeothermal gradient is not an inconsistent argument. In most areas of the Sydney Basin, except those influenced directly by local igneous activity, the maximum paleogeothermal gradient would probably have occurred immediately before the uplift associated with the Tasman Sea spreading, when the heat flow increased to about 3 HFU. Prior to this, the instantaneous paleogeothermal gradient is unlikely to have exceeded G_2 significantly.

The erosional episode is modelled as instantaneous in the above analysis. If the greater part of the erosion occurred in less than 5 m.y., then the instantaneous model would be a good approximation. However, little knowledge exists of the erosion rates or the present day location of the eroded sediment. Little Tertiary sediment is preserved on the Australian continent to the west of the Sydney Basin. Variable thicknesses (up to 1 km) exist offshore in the Tasman Sea [Shaw, 1979]; the age of these sediments is poorly known. Another alternative is that a large network of rivers fed the eroded sediment into Late Cretaceous and Tertiary sedimentary basins along the southern Australian continental margin. Thicknesses of up to 8 km of Late Cretaceous sediments have accumulated in these basins [Veevers, 1981]. The assumption of a rapid erosion rate is consistent with observation and is the simplest for the purposes of the present study.

The model of coalification and the model of thermal activation of magnetization discussed in this paper clearly provide supporting evidence for elevated surface paleotemperatures in the Sydney Basin as suggested by Schmidt and Embleton [1981]. This method may provide successful corroboration of overprinting observed on other continental margins.

NOTATION

- a constant.
- b constant.
- B constant.
- C thermal alteration parameter.
- f_0 fluctuation rate.
- G geothermal gradient.
- h thickness.
- H_K intrinsic coercivity.
- J_s spontaneous magnetization.
- k Boltzmann's constant.
- R_0 vitrinite reflectance in oil.
- t time.
- T temperature.
- v volume.
- V rate of subsidence.
- z depth.

Note added in proof. Evidence based on vitrinite reflectance has recently been presented [Friedman and Saunders, 1982] in support of great depth (~6.5 km) of burial near the eastern margin of the United States.

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