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Analysis of tectonic subsidence and heat flow in the Malay Basin (offshore Peninsular Malaysia)

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Abstract: The Malay Basin has a very high present-day surface heat flow, with an estimated heat flow anomaly of about 33–42 mW m^{-2}. The heat flow anomaly is interpreted as the result of thinning of the lithosphere during basin formation. The basin is relatively young age (about 35 Ma), which implies that the thermal anomaly due to lithospheric thinning has not dissipated completely; the Basin is still undergoing thermal subsidence. Data from over 60 wells were used in the analysis of subsidence and thermal histories to gain a better understanding of its tectonic evolution. A model of lithospheric stretching was used, whereby rifting occurred over a 10 Ma year period starting 35 Ma ago. The subsidence histories from well data gave stretching factor (ß) estimates ranging from about 1.2 on the basin flanks to about 4 in the centre. The basin flanks were uplifted during the initial rifting, causing subsidence to be delayed for about 10 Ma. Flank uplift was probably the result of non-uniform stretching of the lithosphere and horizontal heat loss through the sides of the basin as the lithosphere was being stretched. Heat flows calculated using the ß estimates agree with those derived from well test data and, therefore, supports the stretching model. These results are comparable with those derived from maturity modelling using available vitrinite maturity data.

INTRODUCTION

Understanding the tectonic subsidence and thermal histories of sedimentary basins is important for evaluating their source-rock maturation history and hydrocarbon potential. Extensional basins form as a result of the isostatic response to stretching of the lithospheric thermal boundary layer (McKenzie, 1978) (Fig. 1). The basins subside because of crustal/lithospheric thinning during rifting and thermal re-equilibration of the underlying asthenospheric thermal anomaly (McKenzie, 1978; Jarvis and McKenzie, 1980; Cochran, 1983) (Fig. 1A to D). Subsidence is typically rapid during the stretching phase but slower during the thermal relaxation phase (Fig. 1E). The heat flow of extensional basins also decays gradually with time as the lithosphere attains its pre-stretching temperature structure. Heat flow anomalies may fluctuate, however, when multiple stretching or magmatic events occur, which will complicate the interpretation of heat flow history. The maximum temperature attained by sediment in a basin is governed by the basal heat flow, i.e. heat input from the mantle through the basement, and can provide useful information about the extent of lithospheric thinning/stretching can be obtained from the temperature history of the sediments. Vitrinite reflectance is a parameter that is commonly used in investigations of palaeotemperatures in sedimentary basins.

This paper analyses the subsidence and thermal histories of the Malay Basin, offshore Peninsular Malaysia. This basin has all the major characteristics of a sedimentary basin that was formed by lithospheric stretching, such as crustal extension by normal faulting, high heat flows, and a rift-sag basin geometry. More than two decades ago, White and Wing (1978) proposed that the Malay Basin was formed by regional thinning of the continental crust. A cross-section through the Malay Basin (Fig. 2) shows a "steer's head" geometry that is characteristic of rift basins (Dewey, 1982). Gravity modelling by Harder *et al.* (1993) and Mazlan Madon (1996) showed that the negative gravity anomaly of about –25 mGal over the Malay Basin (Watts *et al.*, 1978) is the result of crustal thinning by a factor of about 2. The crustal thinning has resulted in a shallower Moho and, consequently, an elevated geotherm, which is manifested in the abnormally high temperature-depth gradients in the basin (> 40°C km^{-1}) (Fig. 3).

In this study, temperature and heat flow data were compiled from published and unpublished sources to investigate the magnitude of the heat flow anomaly caused by lithospheric extension in the Malay Basin (Fig. 4). One-dimensional backstripping of stratigraphic data at well locations, following the method described by Sclater and

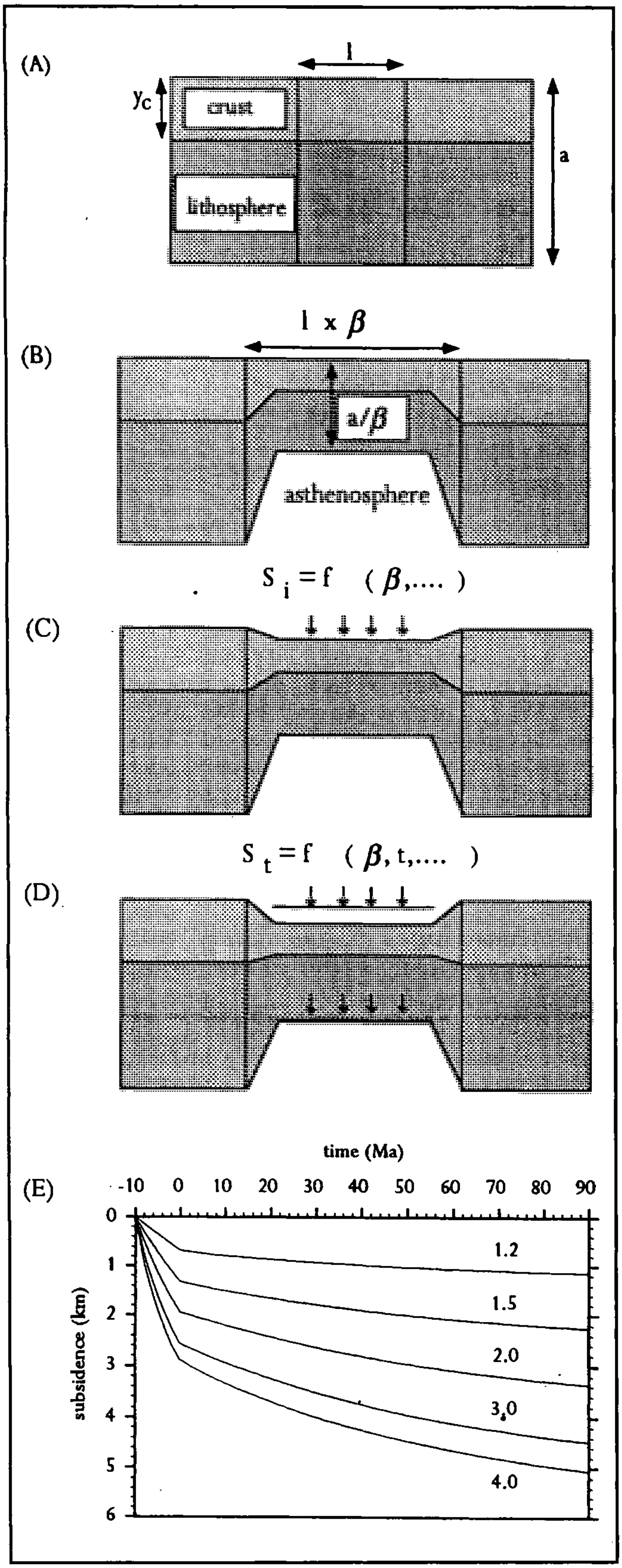

Figure 1. Lithospheric stretching model (after McKenzie, 1978). (**A**) pre-stretching lithosphere of thickness a and crustal thickness y_c. (**B**) Instantaneous uniform stretching of a unit length l by an amount β (stretch factor) results in lithosphere of thickness a/β. (**C**) Initial subsidence, S_i, due to isostatic response to crustal thinning, depends on b and other constants parameters (see Table 1). (**D**) After stretching, subsidence S_t is caused by cooling of the upwelled asthenosphere, and is time-dependant. (**E**) Theoretical subsidence curves of a water-filled basin that results from stretching of a "normal" lithosphere for 10 Ma duration, calculated using various values of β.

Christie (1980), was carried out to determine the tectonic subsidence history. Stretching factors (ß) were estimated by comparing the results with theoretical subsidence curves based on the lithospheric stretching model of McKenzie (1978). The estimates of ß were then used to forward-model the temperature history for comparison with that derived by modelling vitrinite reflectance data.

TECTONIC FRAMEWORK

The Malay Basin is located offshore on the northern Sunda Shelf between the Thai and West Natuna Basins. It is underlain by the early Mesozoic Sundaland basement that was cratonised during the Indosinian Orogeny (Hutchison, 1989). The basins were formed when the Sundaland continental landmass underwent widespread extension during the early Tertiary. Basement lithologies identified by drilling on the basin margins include granite, rhyolite, metasediment, and limestone (Fontaine *et al.*, 1990; Mazlan Madon, 1992), indicating the continuation of the geology of the Malay Peninsula

Many workers have proposed, in one form or another, that the formation of the Malay and neighbouring basins was related to extrusion tectonics and strike-slip faulting as a result of the India-Asia collision during mid-late Eocene times (Tapponnier *et al.*, 1982; Daines, 1985). As the Indian continent collided with the southern margin of Asia, large slivers of the continent were extruded eastwards away from the collision zone, beginning with the Malay Peninsula, Indochina, and South China (Fig. 5). Major strike-slip faults separating the continental blocks were reactivated and became the loci of crustal extension and the development of pull-apart basins such as the Malay, Thai, and West Natuna basins. The structure of the Malay Basin has been described by previous workers (Liew, 1994; Khalid Ngah *et al.*, 1996; Mazlan Madon, 1997). These studies have shown that the Malay Basin was formed by transtensional tectonics during Oligocene times, followed by syndepositional transpressional deformation during the Miocene. The Pliocene to present is relatively quiet tectonically as the basin continued to subside as a sag basin.

SURFACE HEAT FLOW AND HEAT FLOW ANOMALY

The present-day geothermal gradients and surface heat flows are useful pieces of information for modelling past heat flow. Present-day geothermal gradients in sedimentary basins are estimated, usually, from measurements of mud temperatures during logging runs (bottom-hole

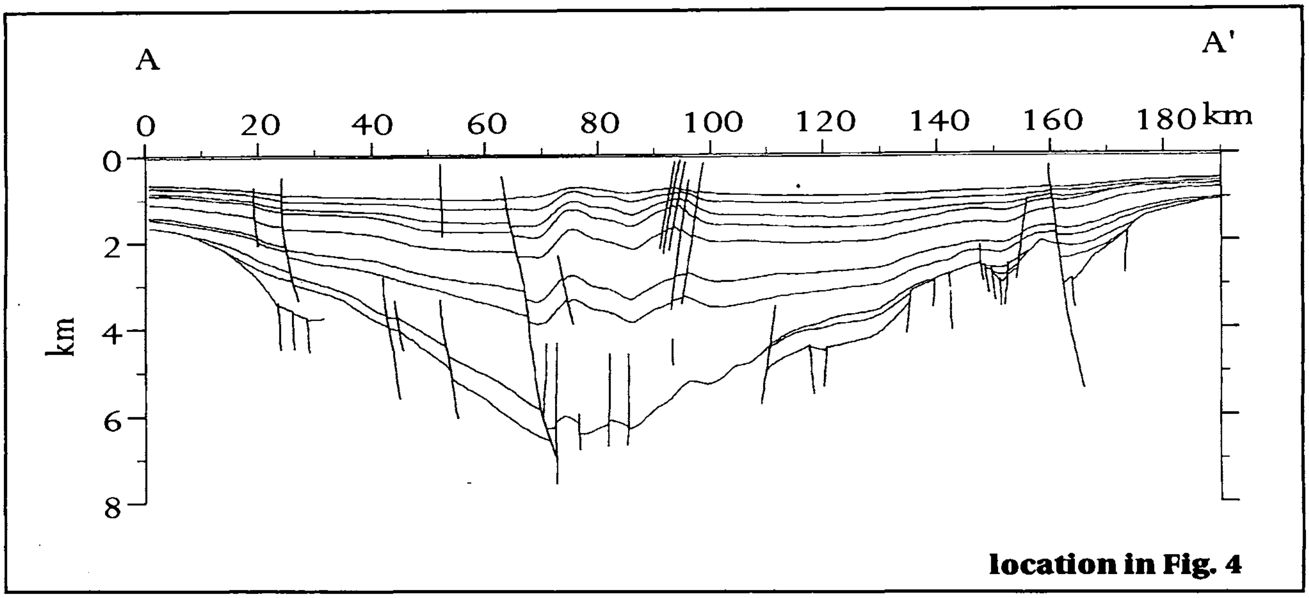

Figure 2. Generalised cross-section across the Basin shows the characteristic rift-and-sag stratal geometry.

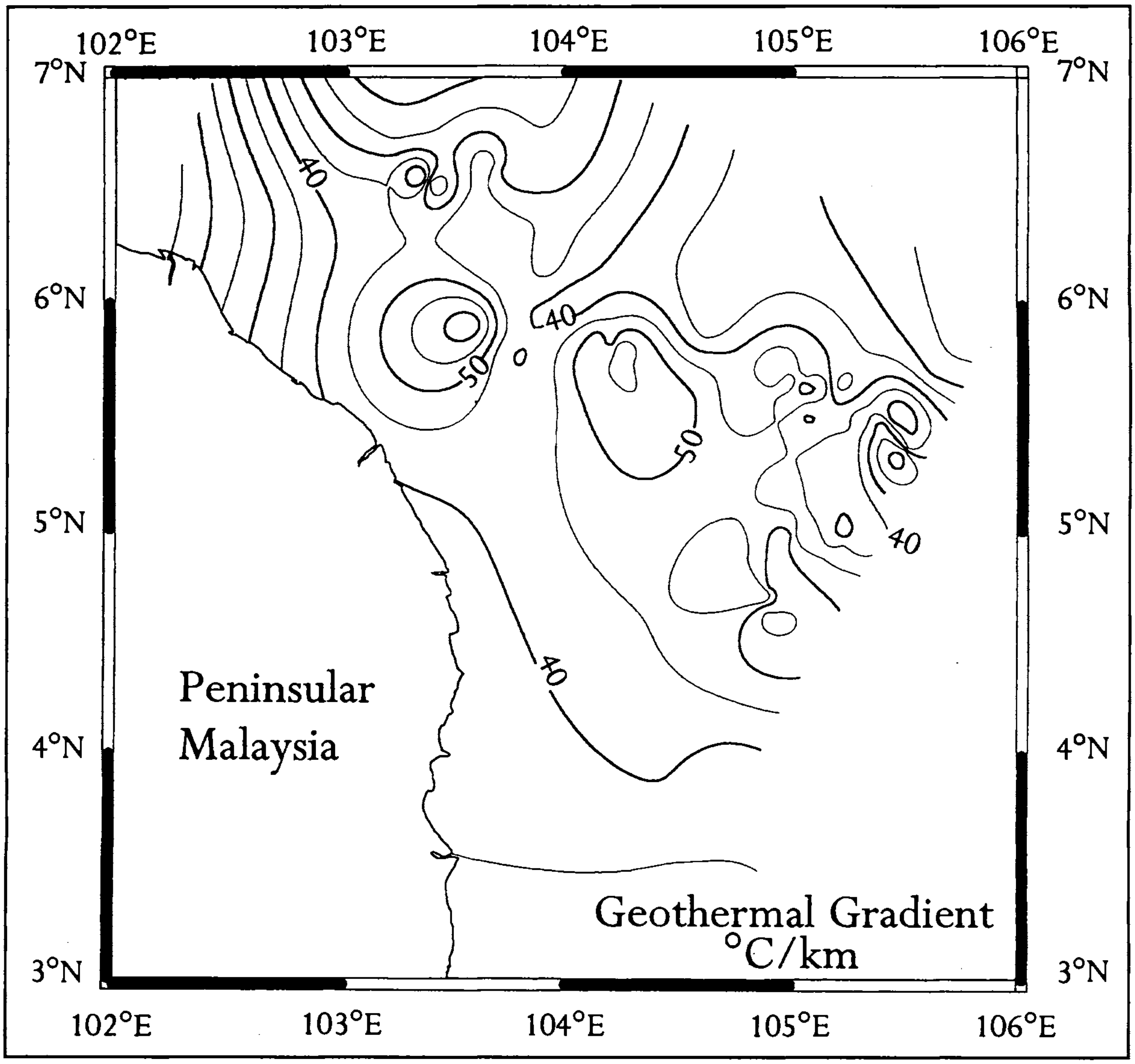

Figure 3. Geothermal gradient map of the Malay Basin.

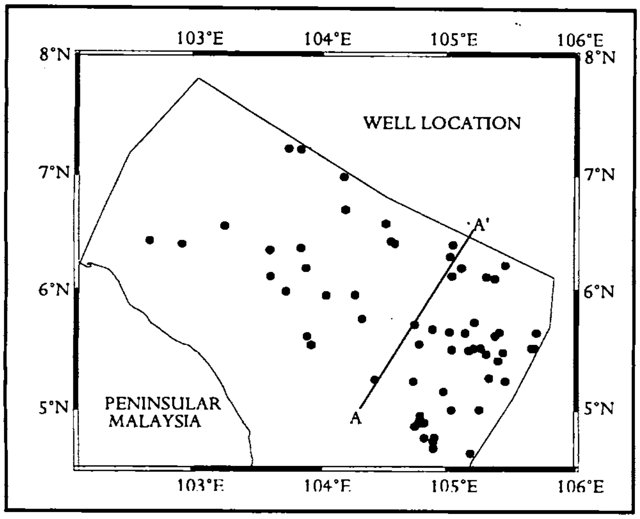

Figure 4. Map of Malay Basin showing the location of wells used in subsidence analysis. Cross-section AA' shown in Fig. 2.

temperatures or BHTs) or measurements of formation-fluid temperatures during repeat formation tests (RFTs), drill-stem tests (DSTs) and production tests (PTs). Subsurface temperatures measured during DSTs and PTs are generally considered to be the most reliable and are used without further correction. Temperatures obtained by other methods need to be corrected to the true formation temperatures, using various correction techniques (e.g. Waples and Mahadir Ramly, 1994). The corrected subsurface temperatures are then used to calculate the geothermal gradients (dT/dz), which is related to heat flow, Q, by

$$Q = k\frac{dT}{dz}$$

where k is the thermal conductivity of the sediment.

Formation temperatures and sediment thermal conductivities in the Malay Basin have been

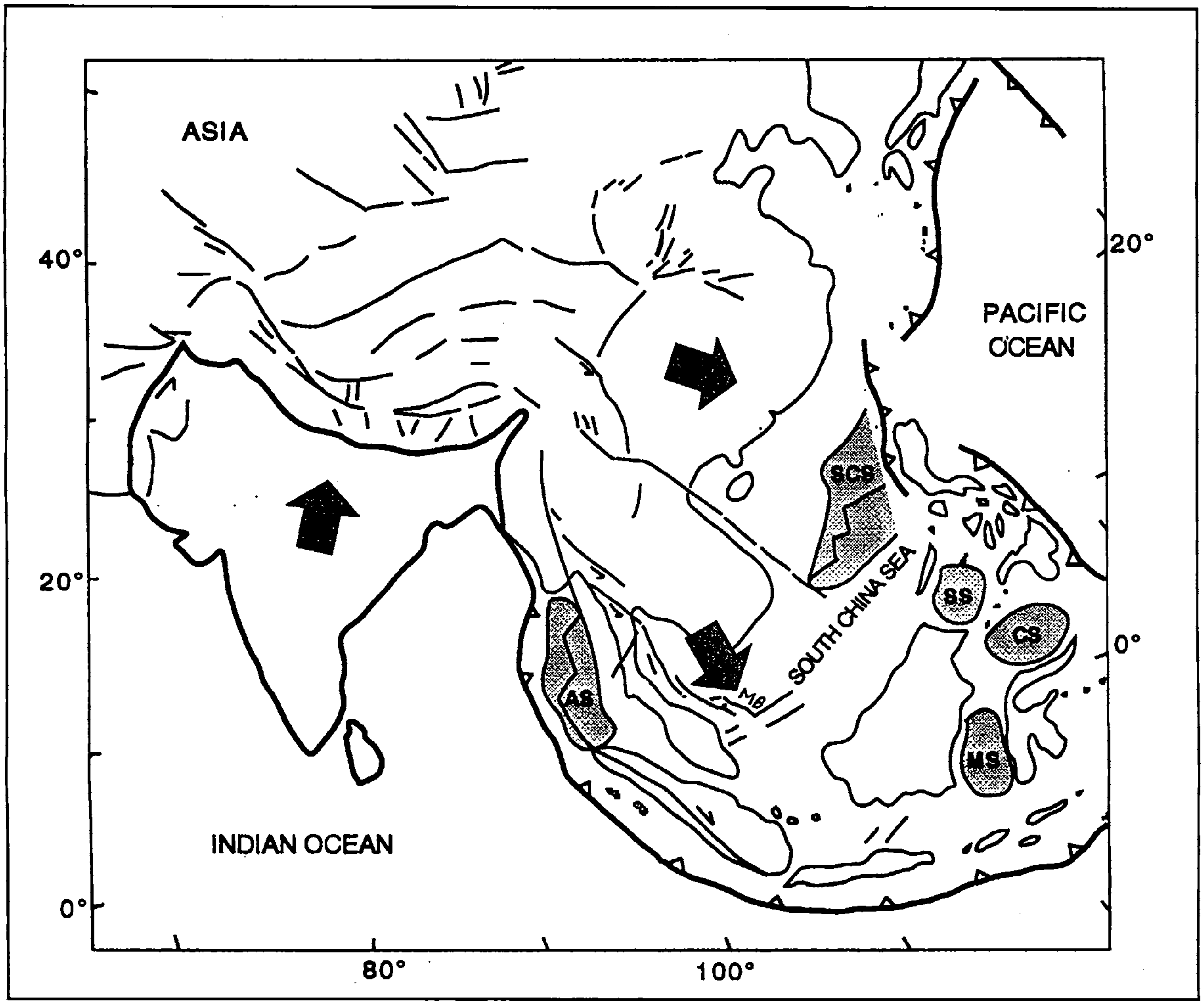

Figure 5. Extrusion tectonics model for Southeast Asia (from Tapponnier *et al.*, 1982). The Malay Basin (MB) is located along a major strike-slip fault that was reactivated by the India-Asia collision during early Tertiary times. Stippled areas represent marginal seas. SCS — South China Sea, AS — Andaman Sea, SS — Sulu Sea, MS — Makassar Strait.

published by W. Ismail W. Yusoff (1984, 1988) and M. Firdaus A. Halim (1993) based on BHTs, RFTs, DSTs, and PTs in over 50 wells and conductivity measurements of over 650 core samples. Based on this dataset, the average thermal conductivity in the Malay Basin is 1.9 W $m^{-1}K^{-1}$ while the average temperature gradient is 51°C km^{-1} (average surface heat flow is 96 mW m^{-2}), similar to other Southeast Asian extensional basins (see Thamrin, 1985). The heat flow is much higher than the global average of about 67 mW m^{-2} (cf. MacDonald, 1965). Such a high heat flow is typical of active back-arc, continental rift, or thick-skinned strike-slip basins. Because the Malay Basin is only about 35 Ma old, the thermal anomaly caused by thinning of the lithosphere has not subsided completely; hence the high heat flow.

Heat flow at the surface of the earth, Q_o, comprises several components

$$Q_o = Q_r + Q_c + \Delta Q$$

where Q_r is the reduced heat flow, i.e. the heat flow in the absence of crustal heat production, Q_c, and ΔQ is the excess or anomalous heat flow that results from lithospheric stretching. Thus, $Q_r + Q_c$ may be regarded as background heat flow (Fig. 6). The simple stretching model (e.g. McKenzie, 1978) calculates Q_r only and ignores the crustal radiogenic heat, Q_c, which must be added to the theoretical value before comparing it with the observed heat flow.

Alternatively, the observed heat flow may be corrected if we know Q_c, which we can estimate if the background heat flow is known. Ideally, the background heat flow should be determined in the surrounding areas where the lithosphere has not been stretched and, therefore, represents the pre-stretching background heat flow. Because no such data exists, I have estimated the background heat flow using the average temperature gradients of 38 basins in the Southeast Asian region, excluding basins on the NW Australian continental margin (Table 1). Most of the heat flow data come from the offshore basins (e.g. Rutherford and Qureshi, 1981; Hutchison, 1989), whose average heat flow depends on the tectonic setting and origin. Generally, basins in back-arc regions, such as those in Sumatra, have very high heat flows (> 80 mW m^{-2}), while fore-arc and deltaic basins have heat flows less than 60 mW m^{-2}.

The average temperature gradient derived from this dataset is 35.4°C km^{-1}. Assuming an average sediment conductivity of 1.9 W m^{-1} K^{-1} for the Malay Basin, this corresponds to a background heat flow ($Q_r + Q_c$) of 67mW m^{-2}. A deviation from this value is interpreted as heat flow anomaly, ΔQ, which is the excess heat flow due to thinning of the lithosphere. Because the theoretical equilibrium heat flow at the top of a 125 km lithosphere is 33 mW m^{-2} (assuming the parameters in Table 2), the estimated crustal contribution, Q_c, to the background heat flow is about 33 mW m^{-2}.

BACKSTRIPPING

Tectonic subsidence history is determined by backstripping (Watts and Ryan, 1976), which is a standard technique for removing the effects of water and sediment loading on the basement (Fig. 7). The backstripping equation is based on simple balancing of mass columns, assuming Airy isostatic compensation. For a sediment layer of thickness, Δh, the tectonic subsidence, ΔS, is given by

$$\Delta S = \Delta h\{(\rho_m - \rho_s)/(\rho_m - \rho_w)\} + W_d - \Delta d\{(\rho_w)/(\rho_m - \rho_w)\}$$

where W_d is paleowater depth, Δd is sea-level change, and ρ_m, ρ_w, and ρ_s are mean densities of mantle, water, and sediment, respectively (Fig. 7A). In the present analysis, the effects of paleobathymetry and sea level were ignored because the Malay Basin was a nonmarine basin during much of its history.

The tectonic subsidence of a multilayered sedimentary column is

$$S_T = S^* \{(\rho_m - \rho_s)/(\rho_m - \rho_w)\}$$

where S^* is the total decompacted thickness of all the layers, given by

$$S^* = \Sigma_i (1-\phi_i)/(1-\phi'_i)\ \Delta h_i$$

Δh_i being the thickness of layer i, ϕ_i is its porosity before decompaction, and ϕ'_i is the porosity after decompaction. The average sediment density, ρ_s, is obtained by averaging the densities of the constituent sediment layers in the column:

$$\rho_s = \Sigma_i\{\phi_i\rho_w + (1-\phi_i)\rho_g\}\ \Delta h_i/S^*$$

where ϕ_i is the mean porosity of layer i and r_g is the mean sediment grain density. Porosity is assumed to decrease exponentially with depth according to the following relationship (Athy, 1930)

$$\phi(z) = \phi_o\ e^{-cz}$$

where ϕ_o (= 0.55) is the porosity at the surface and c (= 0.51 km^{-1}) is the porosity decay constant, both determined empirically from sonic logs in several wells from the southern Malay Basin (Fig. 7B). This relationship is assumed to be independent of lithology and applies throughout the life of the basin.

TECTONIC SUBSIDENCE

Sixty-four wells with stratigraphic data were chosen for analysis (Fig. 4). The data consist of depths to the top of seismic units whose ages have been determined biostratigraphically (Azmi Yakzan *et al.*, 1994). The age of the start of rifting is speculative because no direct evidence from the oldest sediment is available. Analogy with the

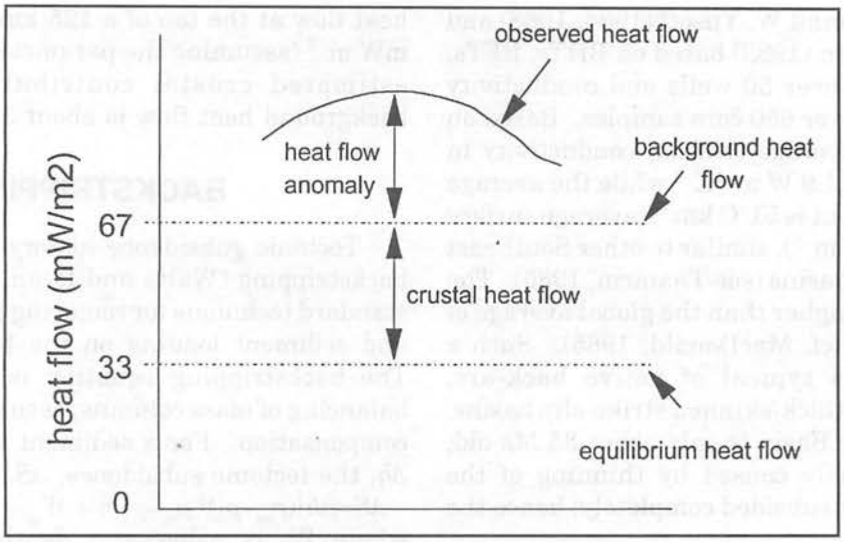

Figure 6. Schematic illustration of various components of heat flow in sedimentary basins.

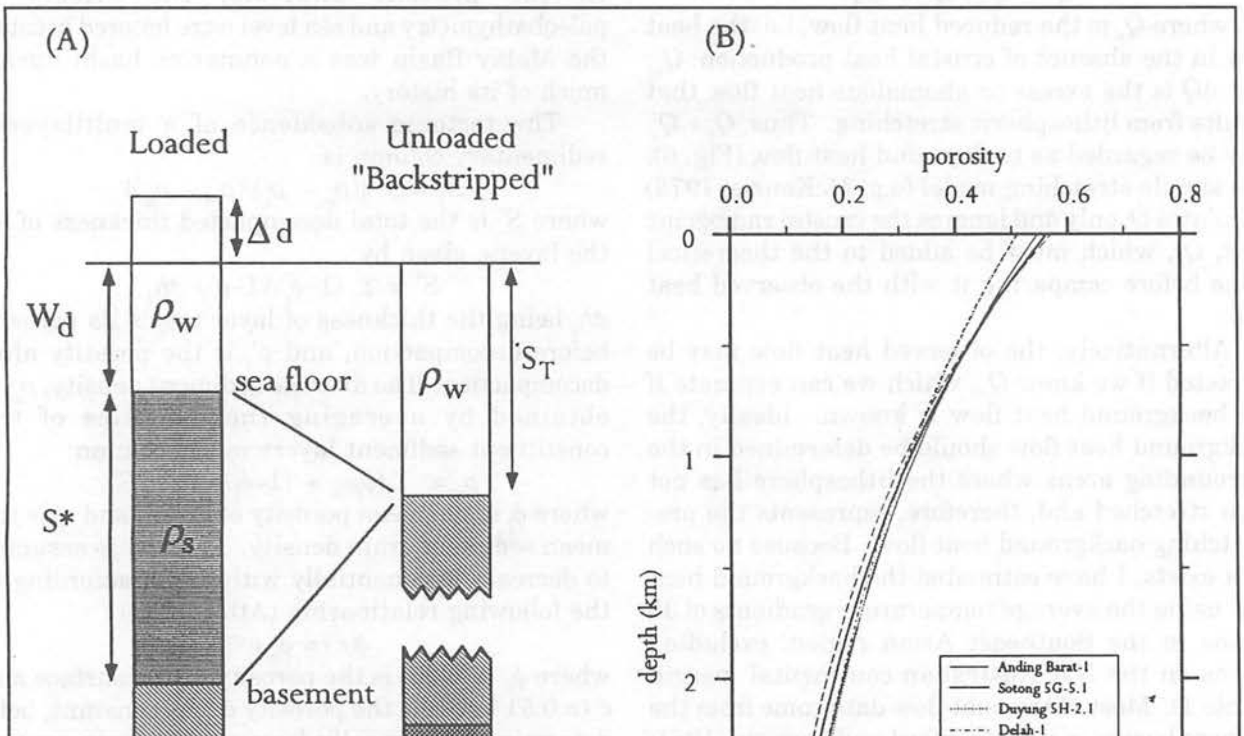

Figure 7. Backstripping technique (A) involves removing the loading effect of a sediment pile of thickness S^* and calculating the unloaded depth of basement (tectonic subsidence), S_T. Symbols explained in the text. (B) Porosity vs depth curve used in the decompaction equation, based on sonic log data from southern Malay Basin wells. Sonic-log porosity values obtained by empirical method (Issler, 1992) assuming Δt_{ma}=59μs/ft, x=2.095.

West Natuna Basin (Daines, 1985) suggests a rifting age of 35 Ma (Oligocene). The top of the synrift sequence is roughly equivalent to the top of seismic unit M, which is about 25 Ma old. Hence, the rifting duration is about 10 Ma, which is well within the time range permissible for the instantaneous uniform stretching model to work (Jarvis and McKenzie, 1980).

Tectonic subsidence curves obtained by backstripping the well data were plotted against theoretical curves based on the stretching model. In the calculation of theoretical subsidence, an initial crustal thickness of 31.2 km and a crustal density of 2,800 kg m^{-3} were assumed. Other modelling parameters are shown in Table 2. Figure 8 shows a map of stretching factors estimated from the modelling results and the subsidence histories at selected well locations. The data show that b is generally between 1.2 and 4.0, increasing from the margins to the centre, and reflects the extent of lithospheric thinning beneath the Basin.

To test the β estimates, the theoretical heat

Table 1. Average geothermal gradients of Southeast Asian Tertiary basins, from Hutchison (1989, p. 64). The data is used to estimate the regional average background heat flow (see text).

Basin	°C km^{-1}	Basin	°C km^{-1}
Southwest China	27	Sokang	56
South China shelf	41	NE Natuna	34
Gulf of Tonkin	40	West Natuna	38
Cagayan Valley	22	Penyu	38
Luzon Central Valley	24	Malay	45
Ragay-Samar	41	Gulf of Thailand	50
Visayan	31	South Andaman	33
Leyte	41	North Sumatra	47
Cotabato	18	Sibolga	24
Iloilo	21	Central Sumatra	61
South Mindoro	29	Bengkulu	24
West Palawan	22	South Sumatra	49
East Palawan	27	Sunda	46
Sandakan	28	Billiton	32
Tarakan	38	NW Java	46
Sabah	28	E Java	39
Baram Delta	28	Barito	36
Central Luconia	43	Kutei	32
Balingian	41	South Makassar	25

Table 2. Modelling parameters used in subsidence/heat flow modelling (from Cochran, 1983).

Symbol	Value	Parameter
ρ_w	1,030 kg m^{-3}	Density of water
ρ_c	2,800 kg m^{-3}	Density of crust
ρ_m	3,300 kg m^{-3}	Density of mantle
ρ_s	2,400 kg m^{-3}	Density of sediment
ρ_g	2,680 kg m^{-3}	Average sediment grain density
y_c	31.2 km	Initial crustal thickness
a	125	Lithospheric thermal thickness
α	3.4 x 10^{-5}°C^{-1}	Coefficient of thermal expansion
T_o	1,333°C	Asthenosphere temperature
K	0.008 cm^2S^{-1}	Thermal diffusivity
κ	3.138 W $m^{-1}K^{-1}$	Thermal conductivity of lithosphere
τ	62.7 Ma	Lithospheric thermal time constant

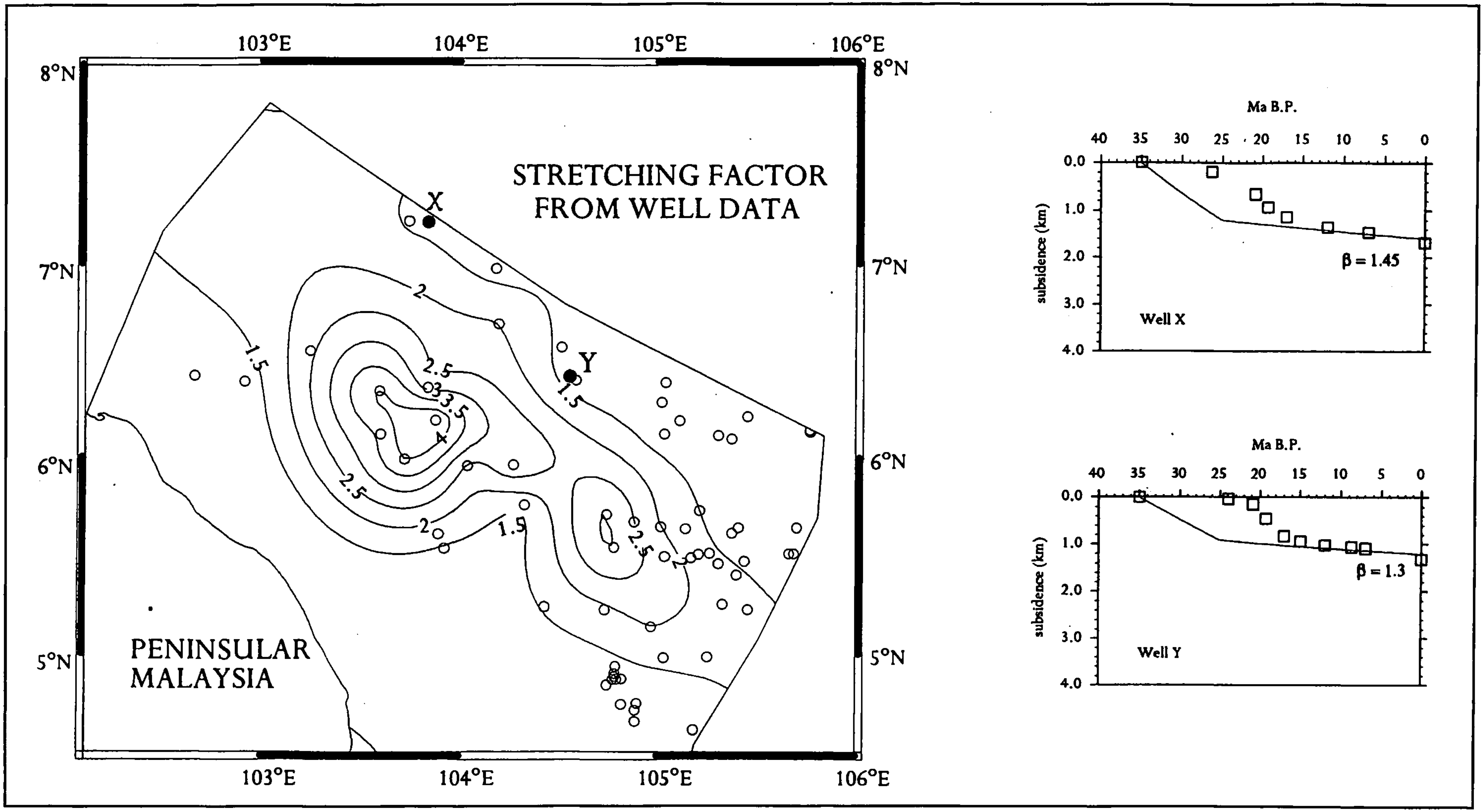

Figure 8. Results of backstripping well data and modelling subsidence based on stretching model. Map shows b factors in the basin. Note the delay of subsidence shown in wells on the basin flanks.

flow predicted by the β values are compared with heat flow anomaly derived from well data, including those based on DST/PT data. Figure 9 shows a series of theoretical heat flow curves for various values of β between 1.2 and 4, as indicated by the subsidence analysis results. The figure shows that heat flow anomaly resulting from lithospheric stretching decreases exponentially with time since the end of stretching (10 Ma) but, for young basins (< 50 Ma), the anomaly is still significant. For example, the Malay Basin, whose age is about 35 Ma, has a present-day heat flow of 35–80 mW/m^2 including the crustal contribution of 33 mW m^{-2}.

Figure 10 compares the predicted and observed heat flow at some of the well locations, based on β derived from backstripped subsidence. The observed and predicted heat flows show a reasonable correlation. Wells at the deeper central parts of the Basin show predicted heat flows that are higher than the observed. The presence of sediment may have reduced the surface heat flow significantly, especially in the centre of the Basin; the sediment acts as insulator that prevents heat from escaping through the surface. Some studies (e.g. de Bremaecker, 1983; Lucazeau and Le Douaran, 1985) have shown that, for a moderately high sedimentation rate (3.5 cm Ma^{-1} in the Malay Basin), surface heat flow may be reduced by as much as 20–30% as a result of sediment blanketing.

VITRINITE REFLECTANCE AND TEMPERATURE HISTORY

Vitrinite reflectance (R_0) as a thermal maturity indicator in sediments can be used to determine paleotemperature and heat flow histories of basins. A R_0-vs-depth profile is forward modelled by using the burial history (geohistory) derived from backstripping and an assumed geothermal gradient as input. The calculated profile is then compared with measured profile to see if the assumed geothermal gradient is correct. This forward modelling approach is done iteratively by changing the temperature gradient until a match between the calculated and the observed R_0 is obtained. In this study, I have used the kinetics-based Easy%R_0 algorithm of Sweeney and Burnham (1990).

Waples *et al.* (1994) investigated the possible suppression of vitrinite reflectance in the Malay Basin using the Fluorescence Alteration of Multiple Macerals (FAMM) technique (Wilkins *et al.*, 1992). This technique measures the change in the peak fluorescence intensity of vitrinite when subjected to excitation by a constant-intensity light source. FAMM measurements are calibrated against a standard telocollinite sample for conversion of the fluorescence alteration ratio to the equivalent vitrinite reflectance, herein called R_e. Waples *et al.* (1994) compared the FAMM results with

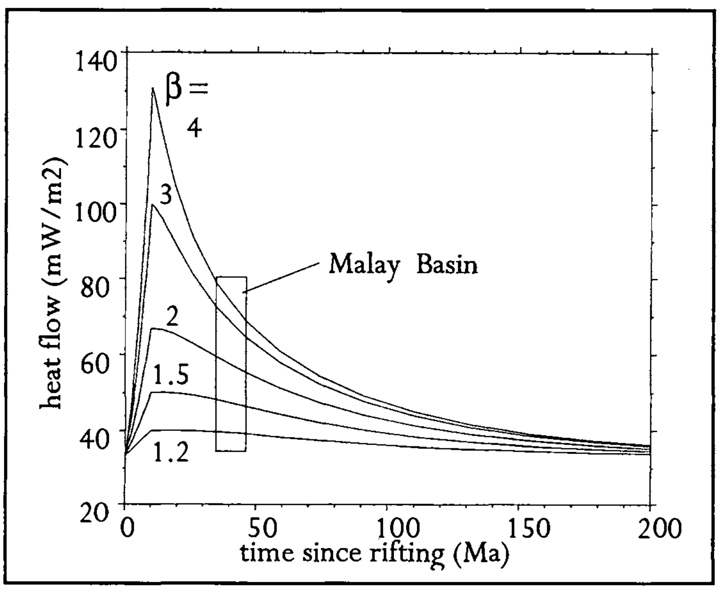

Figure 9. Theoretical heat flow history (1D model, uniform stretching) predicted by various β values for stretching duration of 10 Ma. Shown also is observed range of heat flow (35–80 mW m^{-2}) in the Malay Basin (age about 35–40 Ma), excluding crustal heat flow of 33 mW m^{-2}.

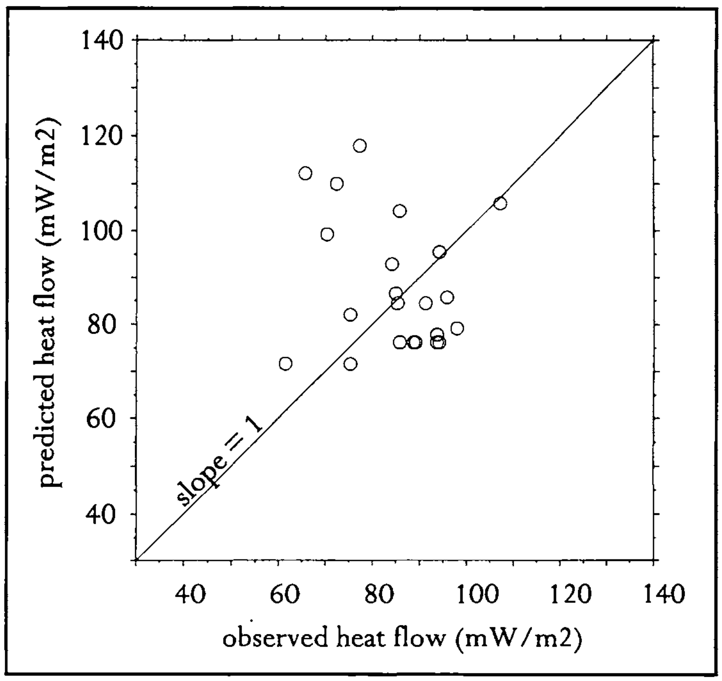

Figure 10. Plot of predicted vs observed heat flow data For reference, perfect correlation is represented by the diagonal line (slope = 1). Predicted heat flow based on lithospheric stretching model using the estimated stretching factors, while observed heat flows are based on temperature data in the boreholes.

conventional R_0 measurements and found significant suppression of vitrinite reflectance, particularly in the southwest and north-central parts of the Malay Basin (Fig. 11). Hence, in this study the FAMM data of Waples *et al.* (1994) was used.

Figures 12 and 13 show the results of modelling the FAMM data from five wells. All the wells show decreasing temperature gradients with time since rifting started from about 50–55°C km^{-1} to 40–50°C km^{-1}. Assuming an average sediment conductivity of 1.9 W m^{-1}K^{-1}, these temperature gradients give surface heat flow values of 75–90 mW m^{-2}, which agree rather well with the observed heat flow (Fig. 10).

The temperature gradients can also be used to calculate the heat flow history of the Basin. Besides temperature gradient, heat flow variation is also a function of sediment thermal conductivity, k_s, which is assumed to vary linearly with porosity, f, according to the relationship (Watts and Thorne, 1984):

$$k_s(z) = \phi k_w + (1 - \phi)k_g$$

where z is depth and k_w and k_g are the thermal conductivities of water and sediment grains, respectively (Table 2). Porosity is assumed to vary exponentially with depth while lateral changes of f and, hence, of k_s are ignored. The average conductivity of the sediment column at any time, t, is obtained by

$$k_{ave}(t) = 0.5[k_s(z_t) + k_s(0)]$$

where z_t is the thickness of sediment at time t and $k_s(0)$ is the conductivity of unconsolidated surface sediment (1.3 W m^{-1}K^{-1}).

The results (Fig. 14) show relatively good agreement between the modelled heat flows and the heat flows derived from R_0 data. All wells, except Larut-1, show initially-increasing heat flow with time since stretching, and decreasing gradually from about 22 to 15 Ma ago. The observed heat flow seems to decrease more rapidly after about 18 Ma, probably because of the increasing blanketing effect of sediment during the later part of the basin evolution.

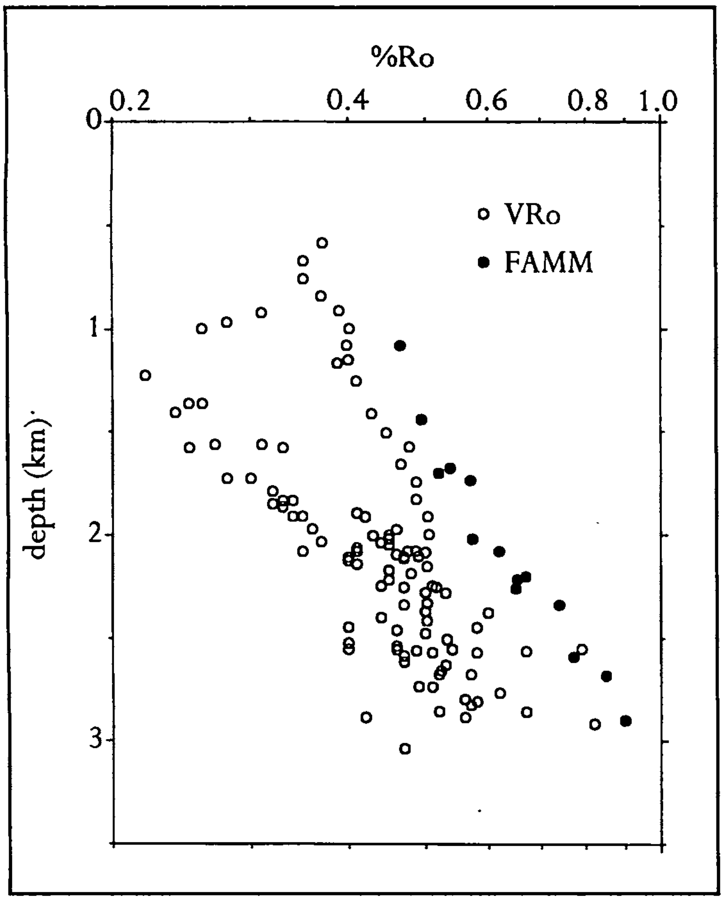

Figure 11. Plot of all available R_0 and R_e (FAMM) data from the southern part of the Malay Basin. Note that all the R_0 data plot the left of the R_e data, suggesting suppression of vitrinite reflectance. Data from Waples and Mahadir Ramly (1994).

DISCUSSION

The results of this analysis indicate that the thermal and subsidence histories of the Malay Basin can be explained in general by the lithospheric stretching model. In the model, we have made simplifying assumptions about, for example, the initial crustal thickness before stretching, and other relevant parameters. The stretching of the lithosphere was also assumed to have been uniform, i.e. strain is distributed equally throughout the whole lithosphere. There are many variants of the McKenzie model that address these issues (e.g. Royden and Keen, 1980; White and McKenzie, 1988). More complex models, however, are meaningless when there are many other uncertainties in the model such as age of sedimentary horizons, sediment thermal conductivity, crustal radioactivity, and the effects of overpressures on porosity and compaction trends (e.g. Audet and McConnell, 1994). Nevertheless, with a simple model as a start, we can recognise deficiencies in our initial assumptions and improve the model as more data becomes available.

The examples shown in Figure 8 suggest that subsidence of the basin flanks (e.g. at Bunga Orkid-1 and Larut-1) was delayed by about 10 Ma after the start of rifting. The basin flanks may have been uplifted initially, causing part of the initial subsidence to occur above sea level. Uplift of rift flanks in some basins have been attributed to lateral heat flow (Cochran, 1983) and may have caused the rift flank uplift in the Malay Basin as well. As the lithosphere is stretched, a temperature gradient develops between the hot basin centre and the cooler flanks. Lateral heat flow causes thermal uplift of

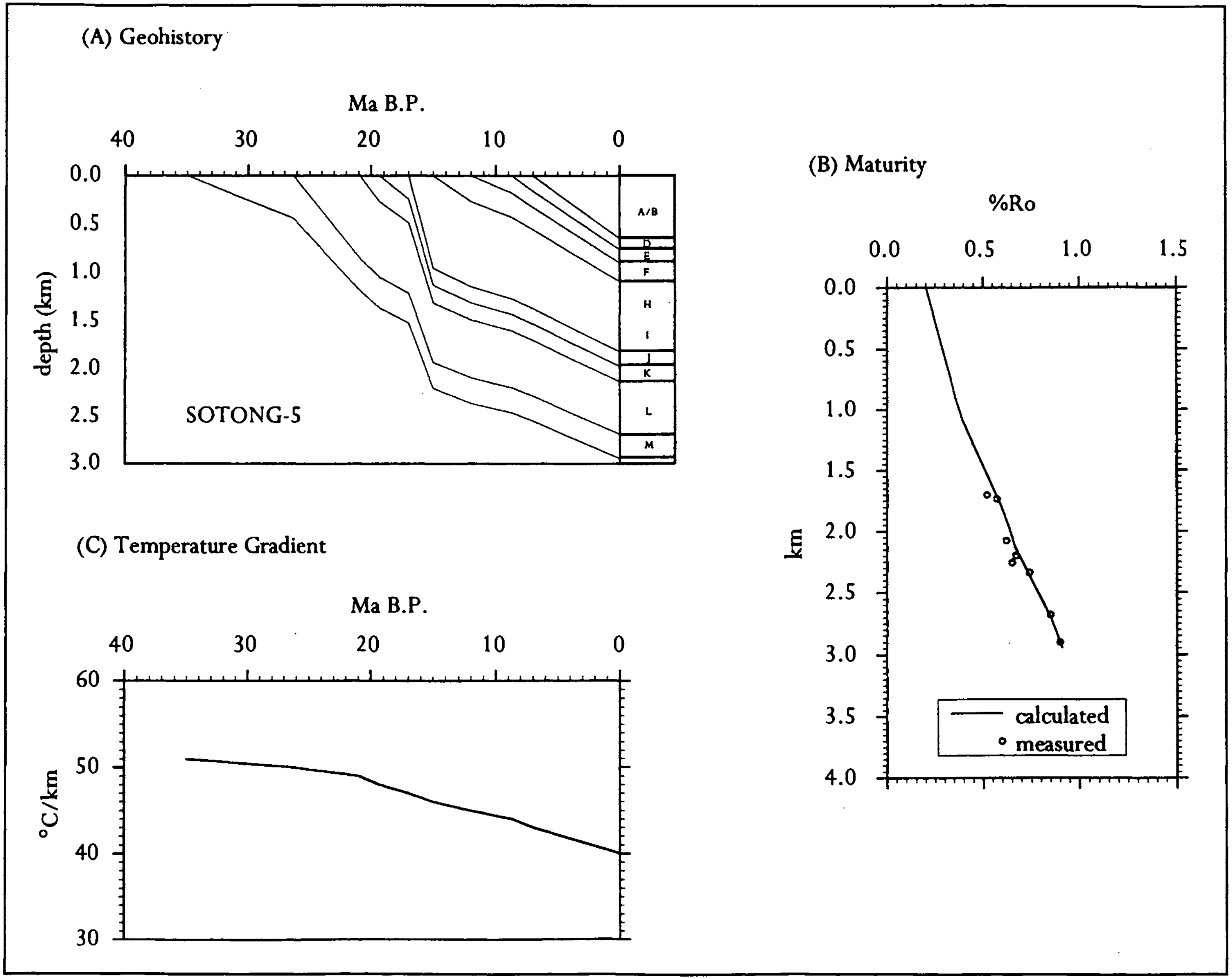

Figure 12. Result of R_o modelling for Sotong-5 (A) Geohistory curve derived from backstripping, (B) Modelled Ro profile plotted against measured R_e, and (C) Temperature-gradient history required to match the observed R_e-depth plot shown in (B). Lines represent calculated R_o.

the rift flanks, which later subside when stretching stops.

Absolute match between modelled and R_o-derived data could not be achieved because of uncertainties such as sediment thermal conductivity, background heat flow, crustal heat production, as well as errors associated with backstripping and the estimation of β stretching factor. There are also uncertainties in the applicability of the Easy%Ro model, which is based on a hypothetical "average" kinetics. Furthermore, the FAMM dataset is relatively small to be sure of the significance of the suppression phenomenon. The results may be improved when better understanding of these factors become possible, through further research efforts.

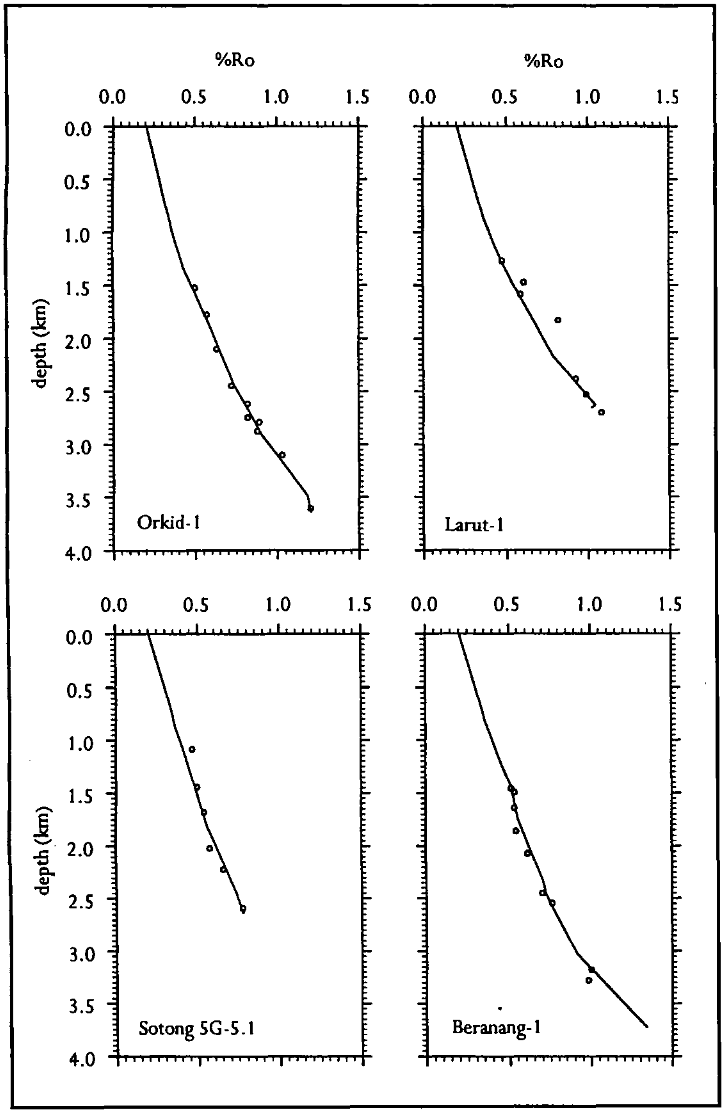

Figure 13. Result of R_e modelling for Orkid-1, Larut-1, Sotong 5G-5.1 and Beranang-1 wells.

CONCLUSION

Application of the uniform finite-duration stretching model (stretching duration 10 Ma) suggests that the stretching factor, β, ranges from about 1.2 on the flanks to about 4.3 at the centre. Wells on the flanks of the Malay Basin show a delay in the subsidence, by about 10 Ma, relative to the time of rifting because of uplift of the basin flanks during the initial stages of rifting. The most plausible mechanism of basin flank uplift during initial rifting is by a form of non-uniform stretching coupled with lateral heat conduction.

The Malay Basin, like other basins in Sundaland, is less than 40 Ma and, hence, may still be undergoing thermal subsidence. This explains the relatively high present-day heat flow. The predicted heat flows based on the stretching factors agree with observations at well locations. The subsidence results are further supported by the results of maturity modelling using vitrinite maturity data from a limited number of wells located on the basin flanks.

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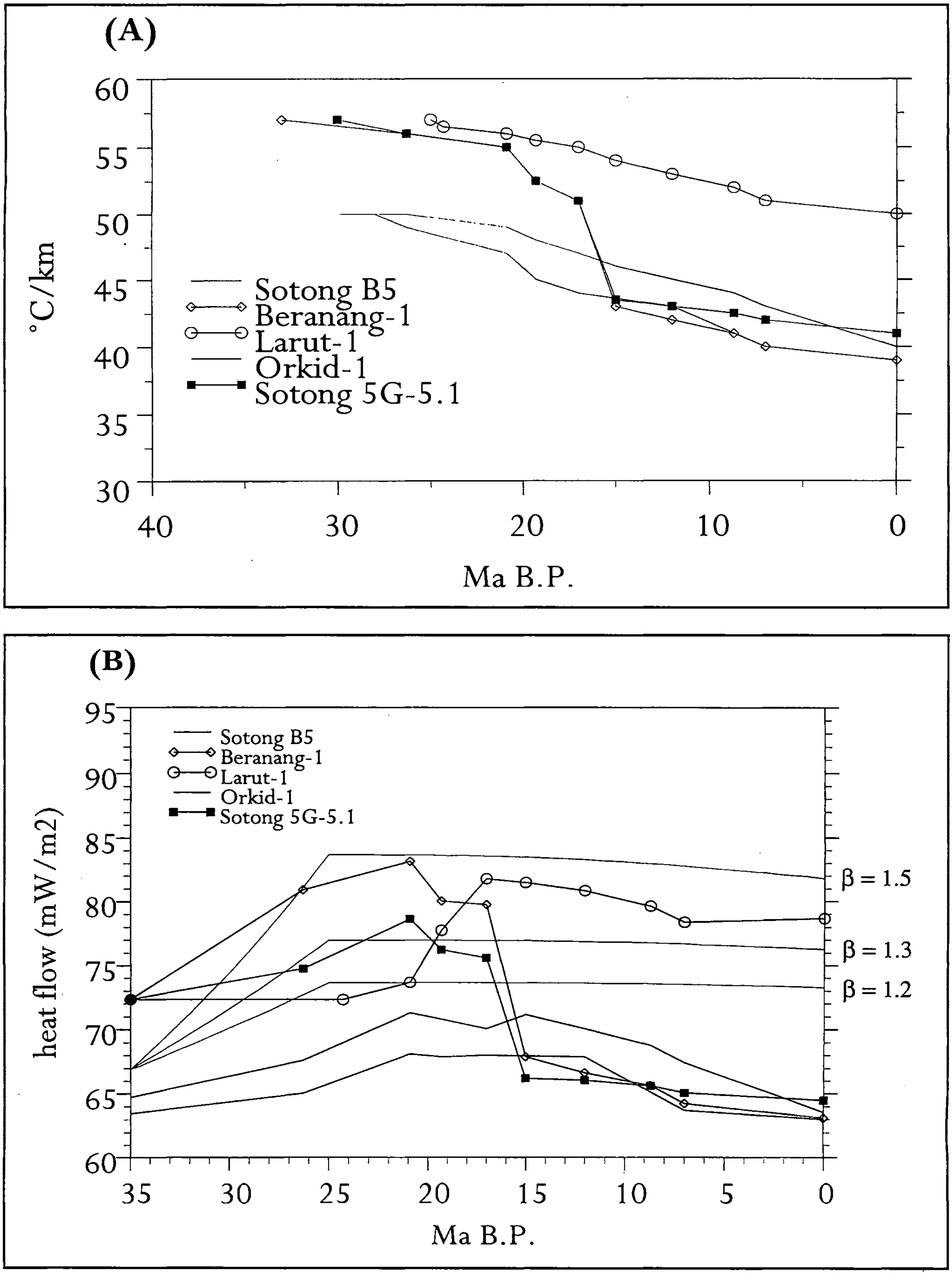

Figure 14. Geothermal gradient (top) and heat flow histories (bottom) derived from FAMM data in the five wells modelled (fine dashed lines) in Figs. 11 and 12. Modelled heat flow represents β = 1.2, 1.3, 1.5, based on backstripping results, with crustal heat flow of 33 mW m^{-2} added.

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