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HEAT LOSS FROM THE EARTH: A CONSTRAINT ON ARCHAEAN TECTONICS FROM THE RELATION BETWEEN GEOTHERMAL GRADIENTS AND THE RATE OF PLATE PRODUCTION

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The models suggested for the oceanic lithosphere which best predict oceanic heat flow and depth profiles are the constant thickness model and a model in which the lithosphere thickens away from the ridge with a heat source at its base. The latter is considered to be more physically realistic. Such a model, constrained by the observed oceanic heat flow and depth profiles and a temperature at the ridge crest of between 1100°C and 1300°C, requires a heat source at the base of the lithosphere of between 0.5 and 0.9 h.f.u., thermal conductivities for the mantle between 0.005 and 0.0095 cal cm^{-1} $°C^{-1}$ s^{-1} and a coefficient of thermal expansion at 840°C between 4.1×10^{-5} and 5.1×10^{-5} $°C^{-1}$. Plate creation and subduction are calculated to dissipate about 45% of the total earth heat loss for this model. The efficiency of this mechanism of heat loss is shown to be strongly dependent on the magnitude of the basal heat source. A relation is derived for total earth heat loss as a function of the rate of plate creation and the amount of heat transported to the base of plates. The estimated heat transport to the base of the oceanic lithosphere is similar to estimates of mantle heat flow into the base of the continental lithosphere. If this relation existed in the past and if metamorphic conditions in late Archaean high-grade terrains can be used to provide a maximum constraint on equilibrium Archaean continental thermal gradients, heat flow into the base of the lithosphere in the late Archaean must have been less than about 1.2–1.5 h.f.u. The relation between earth heat loss, the rate of plate creation and the rate of heat transport to the base of the lithosphere suggests that a significant proportion of the heat loss in the Archaean must have taken place by the processes of plate creation and subduction. The Archaean plate processes may have involved much more rapid production of plates only slightly thinner than at present.

1. Introduction

Whether or not plate tectonic processes operated in the early history of the earth is a controversial subject. A variety of tectonic models have been suggested to explain the geology of Archaean terrains (2.5–3.8 b.y. in age) and the proponents of these models have both included and rejected plate tectonics as the fundamental mechanism [1–3]. Theoretical extrapolation of plate tectonic mechanisms to higher heat production regimes has led to equally discrepant conclusions. For example, McKenzie and Weiss [4] argue that higher heat production within the earth would lead to more rapid movement of thinner plates whose motions were governed by small-scale convection. Baer [5] argues that subduction of a thinner, hotter lithosphere would be impossible and that relative plate motions would cease.

In this article the role of plate tectonic processes in the loss of heat from the earth is assessed and a model is formulated for this heat loss during periods of higher heat production. The steepness of the thermal gradients in the Archaean is shown to be strongly dependent on the rate of plate creation. Comparison of estimates of Archaean geothermal gradients from high-grade metamorphic terrains with these theoretical models suggests that plate tectonic processes must have been active in the Archaean and that plate production was faster than today.

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2. Heat loss from the earth

The heat loss from the earth can conveniently be separated into continental and oceanic components. Over much of the continental area the thermal structure of the lithosphere is likely to be near equilibrium, so that the surface heat flow is the sum of the heat production within the continental crust and lithosphere and the heat supplied to the base of the continental lithosphere. In oceanic areas heat flow profiles are dominated by the creation and cooling of new oceanic lithosphere. McKenzie [6] and Sclater and Francheteau [7] among others, have shown that this process of plate creation, cooling and the subduction of cold plate may be responsible for the loss of as much as 40% of the earth's thermal energy. Heat loss from the earth is therefore through two rather different mechanisms; a conductive heat loss through continental and oceanic lithosphere and a convective heat loss related to plate creation and subduction. It is apparent that with higher heat production in the past the proportion of heat lost by each of these mechanisms may vary and the thermal gradients need not be simply proportional to total heat production within the earth. With higher heat production, for example, oceanic plates may become thinner and the plate creation-destruction mechanisms less efficient. This effect may be offset by more rapid rate of plate production with a corresponding increase in heat loss. Thermal gradients in the past will therefore not be simply proportional to the increased rate of heat production but will also depend on the efficiency of the plate tectonic process. If constraints on equilibrium thermal gradients are available, it may be possible to estimate the proportion of heat transported by the plate creation-destruction process. This is the approach adopted here. First it is necessary to have a model for heat loss by plate processes today.

3. A thermal model for oceanic lithosphere

A variety of thermal models have been formulated for the oceanic lithosphere. All such models predict heat flow values in young oceanic crust higher than those observed, and it is now generally accepted that much heat transfer in the young crust is by hydrothermal circulation (e.g. [8]). The simplest model for

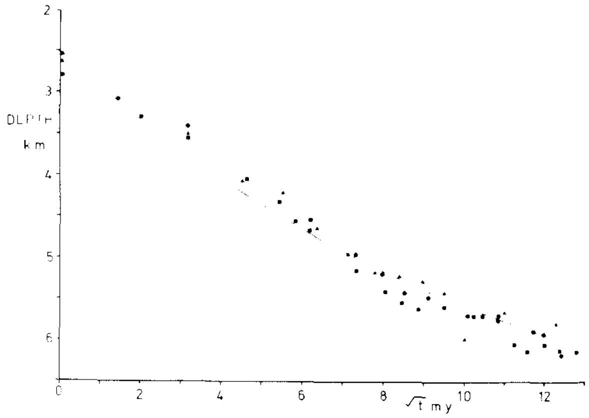

Fig. 1. Depth profiles of oceanic crust corrected for sediment loading and plotted against the square-root of age. Squares, North Pacific [11,12]; circles, North Atlantic [12]; triangles, Indian Ocean [13]. Theoretical curve is that calculated for $q^* = 0.7$ h.f.u. and $K = 0.007$ cal cm^{-1} °C^{-1} s^{-1} (see text and Fig. 5).

the thermal structure of oceanic lithosphere is that of a semi-infinite body initially at magmatic temperatures and cooling only by the vertical conduction of heat [9,10]. Such a model predicts that heat flow will decrease inversely with the square-root of age and that the depth to the ocean floor should increase

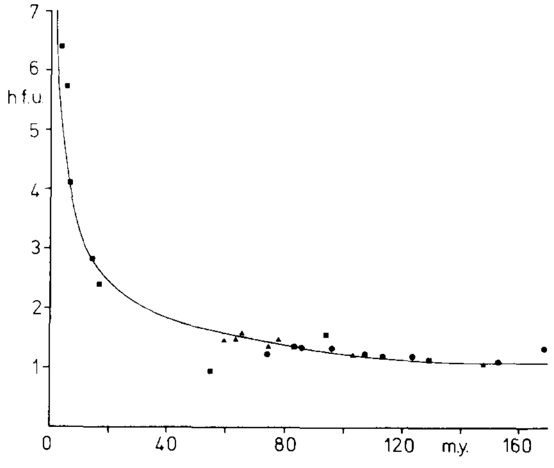

Fig. 2. Heat flow profiles of oceanic crust. Squares, most reliable averages from the Pacific [14]. Circles, averages of North and South Atlantic values from oceanic crust older than 70 m.y. [15]. Triangles, average values of individual basins in the Indian Ocean [13]. Errors on heat flow averages are not shown. Theoretical curve is for the same parameters as in Fig. 1.

as the square-root of age. Recent detailed compilations of the variation of ocean floor depth with age (Fig. 1) show significant deviations from this square-root law for topography. Heat flow values (Fig. 2), even from old crust, show relatively more scatter and do not provide an adequate test of the inverse-root relation with age given the uncertainty in our knowledge of the thermal conductivity of the upper mantle.

Parker and Oldenburgh [16] suggest that solidification of a partially molten asthenosphere at the base of the oceanic lithosphere might contribute significant heat, but Oldenburgh [17] shows that the depth profile will still approximate to the square-root law. Pollack and Chapman [18] suggest that inclusion of an initial thermal gradient at the ridge crest will explain deviations from the inverse-root relation of heat flow. The depth profile is not altered by the inclusion of a linear initial gradient.

Models for the oceanic lithosphere which produce the observed deviations from the square-root law for topography require an additional supply of heat. It is not possible to determine the source of this heat from observations of heat flow and topography in the oceans and a variety of boundary conditions can be chosen which lead to similar topographic and heat flow profiles. The model adopted in this discussion is one in which the lithosphere represents a cold boundary layer which thickens as it cools. The heat required to produce the deviation from the square-root variation of topography is modelled as a constant supply of heat situated at the base of the thermal boundary layer. This heat is thought to be supplied by convection in the mantle below the lithosphere (e.g. [4]). Such a model has been discussed by Crough [19] and Crough and Thompson [20].

Alternative models in which the oceanic lithosphere is of constant thickness have been proposed by McKenzie [6] and Sclater and Francheteau [7]. However, the evidence supporting a thickening oceanic lithosphere (Fig. 3, [22]) and the apparent thickness differences between continental and oceanic lithosphere are considered compatible with the hypothesis that the oceanic lithosphere represents a thickening thermal boundary layer.

3.1. Boundary conditions adopted for the oceanic lithosphere model

The model for the cooling of the oceanic lithosphere is based on two assumptions:

(1) Heat is transported to the base of the lithosphere by small-scale convection at a constant rate.

(2) The lithosphere is defined as that part of the mantle cooler than a critical (possibly depth-dependent) temperature. At temperatures above this critical temperature heat transport in the mantle is assumed to take place by convection.

The boundary conditions adopted for the thermal model of oceanic lithosphere are illustrated in Fig. 4

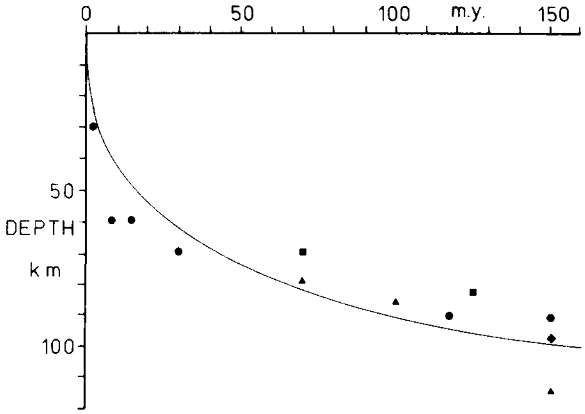

Fig. 3. Estimates of oceanic lithosphere thickness from compilation by Forsyth [22]. Circles [22] including preliminary data point at 150 m.y.; squares [30], triangles [31] and diamonds [32]. Theoretical curve is for the same parameters as in Fig. 1.

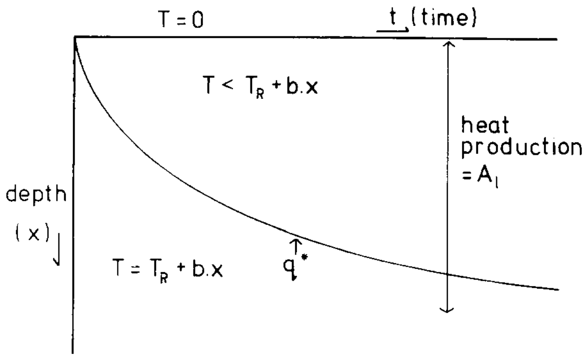

Fig. 4. Boundary conditions for thermal model of oceanic lithosphere. Lithosphere is defined to be that part of the mantle in which the temperature falls below the initial gradient $T = T_R + b \cdot x$. Heat is supplied at a constant rate (q^*) at the base of the thickening lithosphere. Constant heat production (A_1) over a layer of constant thickness x_1 approximates heat production within the lithosphere.

and are similar to those adopted by Crough [19]. The principal differences are the inclusion of a linear thermal gradient at the ridge (to model the gradient likely in a convecting mantle) and a layer of constant heat production to model radiogenic heat production in the oceanic lithosphere. In addition, a temperature- and pressure-dependent coefficient of thermal expansion is used in the calculation of the depth profile. A solution to this model is given in the Appendix.

The definition of the lithosphere as a thermal boundary layer cooler than a depth-dependent temperature undoubtedly constitutes a very simplified approximation to the real situation. McKenzie and Weiss [4] among others have suggested that the average temperature gradient necessary to drive convection in the mantle will only be slightly greater than the adiabatic gradient (i.e. about $1°C\ km^{-1}$). They show that a large variation in heat transfer by convection may be associated with only small temperature fluctuations as the mantle viscosity is very strongly temperature-dependent. McKenzie and Weiss envisage the upper mantle capped by a cold rigid mechanical boundary layer, the lithospheric plate, and beneath this a thermal boundary layer through which material is transported by convection but in which the temperature falls below that of the thermal gradient in the mantle below. The lack of knowledge of the variation of mantle viscosity with temperature and pressure precludes precise modelling of this thermal boundary layer. In the model adopted here the thermal boundary layer between the rigid plate and the underlying freely convecting mantle is ignored, and the rigid mechanical boundary layer is assumed to extend to the depth at which the temperature reaches that of the underlying convecting mantle.

The initial thermal gradient at the ridge is taken to be the same as the gradient in the underlying convecting mantle, although temperatures along the ridge axis are likely to be perturbed from this gradient by the processes involved in partial melting and transport of magma. Such perturbations are not likely to exceed 50–100°C, given that the relative temperature increase due to magma rising along an adiabatic gradient (rather than the slightly shallower gradient in the convecting mantle) will be offset by the temperature decrease due to the latent heat involved in partial melting. Such small perturbations do not alter the thermal structure of the lithosphere at long distances from the ridge. The assumption of an initial thermal gradient makes it necessary to assume a compensation depth in the calculation of the depth profiles (see Appendix) but, for small gradients, choice of this depth between the base of the plate and 200 km makes an insignificant difference to the depth profile. Two phase changes, plagioclase to spinel and spinel to garnet-lherzolite, take place within the depth range of the oceanic lithosphere. Plagioclase-lherzolite is probably restricted to a field close to the ridge crest and the reactions bounding the spinel-lherzolite stability field are probably nearly isobaric over the temperature interval within the lithosphere [24,25]. These phase changes are ignored in the calculation of topography.

3.2. Solutions to the oceanic lithosphere model

Fundamental to this model for the cooling lithosphere is the supply of heat to the base of the plate by convection. The magnitude of this heat source can be calculated by finding solutions to the thermal model which successfully reproduce oceanic topographic and heat flow profiles. The model is sensitive to several other poorly known thermal parameters but the topographic and heat flow profiles provide at least three essentially independent constraints. It is therefore possible to derive values for two of the thermal parameters in addition to the strength of the basal heat source. Similar analyses have been undertaken by Lister [10] for the cooling half-space model and by Parsons and Sclater [12] for the constant thickness model.

Depths to oceanic crust younger than about 60 m.y. increase approximately as the square-root law whereas depths in the older crust approach the equilibrium depth. Thus the topographic profiles are required to fit the observed depths below the ridge crest at two points (2.47 km at 50 m.y. and 3.5 km at 150 m.y.). Heat flow profiles show more scatter and deviate less significantly from the inverse-root relation with age (the latter is also true of model heat flow profiles, e.g. [12]). Values of heat flow in old oceanic basins (120–160 m.y.) mostly lie between 1.0 and 1.2 h.f.u. (where 1 h.f.u. = 10^{-6} cal $cm^2\ s^{-1}$) and this provides a third constraint. The thermal conductivity and the coefficient of thermal expansion are the least well known of the critical thermal parameters and form, along with the magnitude of the basal heat source, three dependent parameters to be determined from solutions of the oceanic lithosphere model. Such solutions require assignment of values to the following parameters:

(1) Temperature at the ridge crest.

(2) Thermal gradient at the ridge crest and below the plate.

(3) Mantle density.

(4) Specific heat.

(5) Heat production within the oceanic lithosphere.

(6) Depth and temperature dependence of the coefficient of thermal expansion.

The temperature at the ridge crest is the least well known of these parameters. It is possible to estimate this temperature from the liquidus temperatures of oceanic basalts (1200–1250°C; [22,26]). The thermal gradient in the mantle below the lithosphere, extrapolated to the surface, probably lies within 100°C of these liquidus temperatures as discussed above. Solutions to the oceanic lithosphere model with temperatures at the ridge crest between 1100°C and 1300°C are accepted. The other thermal parameters (2 to 5 above) are less critical (at least, uncertainty in these parameters affects model solutions insignificantly compared with the uncertainty in the depth profile, heat flow or ridge crest temperature). The values adopted for these parameters are listed in Table 1.

A value of 1°C km^{-1} is adopted for the thermal gradient at the ridge crest and in the mantle below the plate. The consequences of a gradient significantly above this value are discussed below. The temperature dependence of the coefficient of thermal expansion has been recalculated from the data of Hazen [27,28] for an olivine $Fo_{90}Fa_{10}$ and the pressure dependence adopted is that of Anderson [29] for forsterite. Lister [10] shows that the adoption of a temperature-dependent coefficient of thermal expansion gives similar depth profiles to use of a temperature-independent coefficient of thermal expansion if a value for the latter corresponding to that at 0.7 of the ridge crest temperature is used. This has been confirmed in this study and the value at $0.7T_R$ is that quoted in the graphical presentation of the results (Fig. 5). As the published values for the coefficient of thermal expansion of forsterite at 800°C range from 3.9×10^{-5} to 4.4×10^{-5} °C^{-1} [27,28,34] and the value for the mantle which will contain pyroxenes and spinel or garnet in addition to olivine may deviate further from these values, the value of the constant term in the expression for the coefficient of thermal expansion is determined from solutions to the oceanic lithosphere model as discussed above.

Solutions that predict the required depths at 50 m.y. and 150 m.y. are presented in Fig. 5 plotted against the basal heat flow and thermal conductivity as contoured values of the temperature at the ridge crest and the surface heat flow at 150 m.y. This figure is also contoured with the values of the coefficient of thermal expansion at 0.7 of the temperature at the ridge crest. Solutions in Fig. 5 that satisfy the constraints of a ridge crest temperature between 1100 and 1300°C and produce a surface heat flow at 150 m.y. between 1.0 and 1.2 h.f.u. require a basal heat flow between 0.63 and 0.78 h.f.u., thermal con-

TABLE 1

Thermal parameters adopted for thermal model of oceanic lithosphere

Density	$\rho = 3.3$ g cm^{-3}		
Specific heat	$C_p = 0.28$ cal g °C^{-1}		
Heat production	$A_1 = 0.034$ h.g.u. *		
Depth of heat-producing layer	$x_1 = 100$ km		
Thermal gradient at ridge	$b = 1$°C km^{-1}		
Coefficient of thermal expansion:			
temperature dependence	$\alpha_T = \alpha_0 + 1.53 \times 10^{-8}\, T$	°C^{-1}	(1)
pressure dependence	$\alpha_{T,P} = \alpha_T(1 - 1.15 \times 10^{-3})$	°C^{-1}	(2)

* 1 h.g.u. = 10^{-13} cal cm^{-3} s^{-1}.

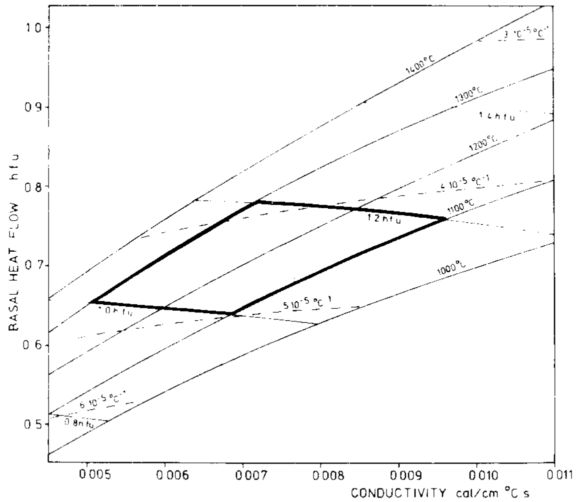

Fig. 5. Solutions to the thermal model of oceanic lithosphere which produce a depth below the ridge crest of 2.47 km at 50 m.y. and 3.5 km at 150 m.y. plotted against basal heat flow in h.f.u. (1 h.f.u. = 10^{-6} cal cm^{-2} s^{-1}) and conductivity as contoured values of temperature at the ridge crest (T_R), coefficient of thermal expansion at 840°C at atmospheric pressure and surface heat flow at 150 m.y. Other parameters as in Table 1. Area outlined by bold contours encompasses solutions which satisfy the constraints of a surface heat flow between 1.0 and 1.2 h.f.u. at 150 m.y. and a temperature at the ridge crest between 1100°C and 1300°C.

ductivities between 0.005 and 0.0095 cal cm^{-1} °C^{-1} s^{-1} and coefficients of thermal expansion at $0.7T_R$ between 5.18×10^{-5} and 4.12×10^{-5} °C^{-1}. The thermal conductivities required by the model lie within the range commonly accepted for mantle materials. The coefficient of thermal expansion spans the range of values measured for forsterite at 800°C of 3.9×10^{-5} to 4.4×10^{-5} [27,28,34] although solutions with low thermal conductivities (giving lower heat flow values) have coefficients of thermal expansion somewhat above the measured values.

The depth profiles are independent of the initial thermal gradient. If this gradient is increased, heat flow values are correspondingly increased and the range of solutions is shifted to those encompassing lower thermal conductivities but higher values of the coefficient of thermal expansion. Models with the initial thermal gradient much above the 1°C km^{-1} adopted here thus would require thermal conductivities lower and coefficients of thermal expansion higher than those commonly accepted for mantle materials. The thickness and rate of thickening of the lithosphere are in good agreement with various estimates of the thickness of oceanic lithosphere (Fig. 3) although the base of the lithosphere defined as the upper boundary of small-scale convection cells need not correspond to that derived from seismic estimates.

TABLE 2

Change in predicted value of basal heat flow caused by variation of parameters on a model solution with T_R = 1200°C and K = 0.007 cal cm^{-1} °C^{-1} s^{-1}. This model requires a basal heat flow of 0.7 h.f.u. and gives a surface heat flow of 1.10 h.f.u. at 150 m.y. for the adopted value of the parameters

Parameter	Adopted value	New value	Basal heat flow (h.f.u.)	Surface heat flow at 150 m.y. (h.f.u)
Density (g cm^{-3})	3.3	3.4	0.725	1.12
Specific heat (cal g °C^{-1})	0.28	0.30	0.724	1.13
Depth below ridge crest at 150 m.y. (km)	3.50	3.60	0.59	1.05

Table 2 lists the effect of uncertainty in choice of values for parameters not shown in Fig. 5. Model solutions are most sensitive to the choice of depth profile. Adoption of a depth of 5.8 km at 150 m.y. appropriate to the Indian Ocean requires a basal heat flow of 0.9 h.f.u. whereas adoption of a depth of 6.2 km at 150 m.y. appropriate to the Pacific Ocean only requires a basal heat flow of 0.5 h.f.u. (assuming a thermal conductivity of 0.007 cal cm^{-1} °C^{-1} s^{-1}). The significance of this variation is not clear at present given the uncertainty on the depth estimates themselves [35] and the possibility that such variations in depth profiles reflect mechanisms additional to variation in the supply of heat to the base of the lithosphere

The solutions to the model presented here are similar to those of the constant thickness models [12] and this results from the initial rapid increase in lithosphere thickness under young oceanic crust.

4. A thermal budget for present-day heat loss

Pollack and Chapman [18] from an analysis of continental heat flow and heat production provinces, derive an empirical relation between surface heat flow and heat flow into the base of the continental crust. They estimate the average continental heat flow to be 1.27 h.f.u. of which 0.60 h.f.u. is from heat-producing elements within the continental crust and 0.67 h.f.u. is supplied from the mantle. Heat production in the subcontinental lithosphere is unlikely to exceed about 0.15 h.f.u. (assuming 0.1 h.g.u. over 150 km) and the average heat flow into the base of the continental lithosphere must be in the region of 0.5–0.67 h.f.u. This estimate is very close to the value calculated for the heat flow into the base of the oceanic lithosphere (0.7 ± 0.2 h.f.u.).

The continents cover only one third of the earth's surface and uncertainty in the basal heat flow into continental lithosphere (about 10% of the total earth heat loss) has relatively little effect on calculation of the heat loss budget of the earth. However, thermal gradients in the continental crust are functions of the basal heat flow and the only information on the thermal gradients in the early history of the earth is preserved in old continental crust. Pollack and Chapman's [18] estimate of the average mantle heat flow under continents might exceed the equilibrium heat flow for two reasons. Firstly, the long-term effects of uplift and erosion may result in a systematic bias in the heat flow values used to calculate the heat flow heat production relation in estimates of mantle heat flow [36]. Secondly, the thickness of continental lithosphere in some areas might be greater than 300 km [37,38] and these areas would have a very low heat flow from the asthenosphere. A heat flow into the base of the lithosphere of 0.7 h.f.u. with the rest of the heat production (0.57 h.f.u.) distributed exponentially in the overlying continental crust would put the base of the lithosphere at about 125 km assuming the thermal definition of the base of the lithosphere is the same as that adopted for the oceans and a thermal conductivity of 0.007 cal cm^{-1} $°C^{-1}$ s^{-1}. A basal heat flow of 0.5 h.f.u. into a continental lithosphere containing 0.1 h.g.u. and overlain by a continental crust producing 0.67 h.f.u. would put the base of the lithosphere at 157 km. In addition, thermal models of relatively young regional metamorphic terrains in the Alps and Caledonides [39,40] require mantle heat flow values between 0.4 and 0.8 h.f.u. and there is little evidence that the mantle heat flow in these areas was enhanced by additional sources of heat. For the purposes of evaluating heat loss from the earth in higher heat production regimes we will assume that the present-day heat flow into the base of the lithosphere in the continental and oceanic regions is identical and has the oceanic value of 0.7 h.f.u.

Given this global uniformity in heat flow at depth, we can separate the heat lost from the earth into three main components:

(1) Radiogenic heating in the continental crust and lithosphere.

(2) Heat flow from the asthenosphere (presumably transported by convection).

(3) Heat loss due to the creation, cooling and subduction of oceanic lithosphere.

The first two components have been estimated from heat flow measurements on the continents and from the model for the cooling of oceanic lithosphere. The third component can be estimated in several ways. The difference between average oceanic heat flow and the heat flow from the asthenosphere will give a minimum estimate as significant heat is lost in young oceanic crust by hydrothermal circulation not monitored by heat flow measurements. Anderson et al. [13] calculated the difference between their model heat flow, based on a constant thickness model, and the observed heat flow values for well-studied areas of young oceanic crust and concluded that hydrothermal circulation transported about 8% of the earth's total heat flow. This method of calculating the total heat lost is the most sound as it depends on the thermal model only for young crust where the various model estimates are in the best agreement. Williams and Von Herzen [41] estimated the rate of plate subduction and, by assuming an average thermal profile for subducted plate, calculated that the average world heat flow was about 30% higher than that calculated from heat flow measurements. It is shown below that uncertainty in spreading rate, age of subducted plate and in parameters critical to the thermal model for oceanic lithosphere render such a calculation too unreliable to provide a useful estimate.

The heat lost by the cooling oceanic lithosphere is simply calculated from the difference in the thermal

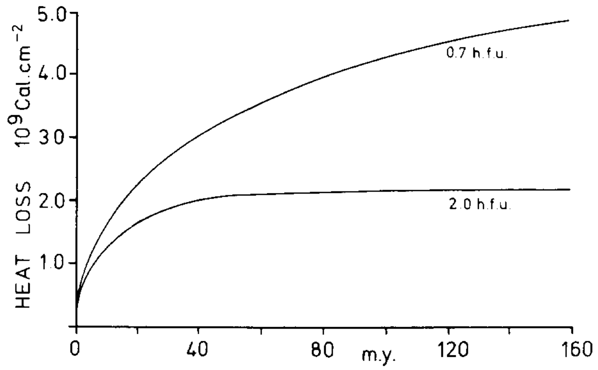

Fig. 6. Heat loss by cooling of oceanic lithosphere as a function of age for basal heat flows of 0.7 and 2.0 h.f.u. Other parameters as for a solution with the basal heat flow = 0.7 h.f.u. and K = 0.007 cal cm^{-1} °C^{-1} s^{-1} (Fig. 5).

profile at the ridge from that in old lithosphere, given the specific heat of mantle material (Fig. 6). To calculate the total heat lost it is necessary to know the rate of subduction of oceanic crust and the age distribution of the subducted oceanic crust. The rate of plate subduction, and thus of plate creation, can be calculated from estimates of the mean spreading rate of oceanic ridges and the total length of oceanic ridge. Estimates of mean half spreading rate vary from 2.74 to 2.8 cm yr^{-1} [41,42] and estimates of total ridge length vary from 48,100 to 59,150 km [13,42]. These estimates give rates of plate creation between 2.6 and 3.3 km^2 yr^{-1}. The age distribution of oceanic crust being subducted at present is shown in Fig. 7. If oceanic plate has been produced at its present rate of about 3 km^2/yr for a sufficient period in the past,

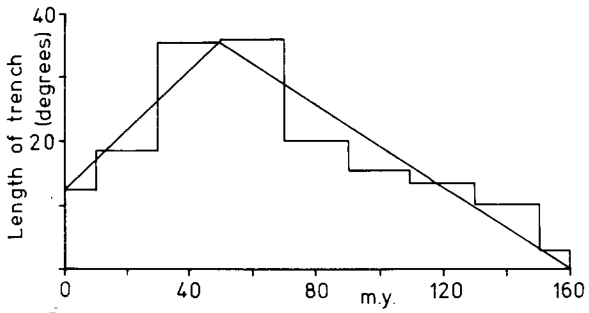

Fig. 7. Histogram of the age distribution of oceanic crust entering oceanic trenches. Compiled from data in Pitman et al. [43] and Hilde et al. [44]. Solid line is age distribution used in heat loss calculations (see text).

this rate and the area of the oceans (3.62 × 10^8 km^2) would require that the mean age of subducted plate would be about 120 m.y. However, the present mean age of subducted plate is about 60 m.y. (Fig. 7). This age distribution could be achieved if the rate of subduction was strongly correlated with the age of subducted plate, if the present age distribution of subducted crust was anomalously young or if the rate of spreading was significantly faster over most of the last 160 m.y. than at present. There is no strong correlation between the age of subducted plate and the rate of subduction. The age distribution of subducted plate does not appear anomalously young compared with the age distribution of oceanic crust. Spreading rates may have been faster in the last 150 m.y. [45] and also the length of spreading ridge may have decreased recently and both may account for the discrepancy between the mean age of subducted crust and the present spreading rate.

TABLE 3

Average heat loss from the earth as a function of various spreading rates and choices of parameters in the oceanic lithosphere model (Fig. 5). Percentage difference from mean value based on heat flow measurements (1.42 h.f.u., [18]) is given in brackets

q^*/T_R	Contribution from:	Spreading rate (km^2 yr^{-1})		
		2.6	3.0	3.3
0.63 h.f.u 1100°C	continents	0.37	0.37	0.37
	q^* and heat production	0.52	0.52	0.52
	plate formation	0.50	0.57	0.63
	total	1.39 (−2%)	1.46 (3%)	1.52 (7%)
0.70 h.f.u. 1200°C	continents	0.37	0.37	0.37
	q^* and heat production	0.57	0.57	0.57
	plate formation	0.56	0.64	0.70
	total	1.50 (5%)	1.58 (11%)	1.64 (16%)
0.77 h.f.u. 1200°C	continents	0.37	0.37	0.37
	q^* and heat production	0.62	0.62	0.62
	plate formation	0.61	0.71	0.78
	total	1.61 (13%)	1.70 (20%)	1.77 (25%)

Estimates for total earth heat loss (as average heat flow values) for present rates of plate creation, the present age distribution of subducted crust and for the range of models in Fig. 5 are presented in Table 3. The implications of the relation between spreading rate, the age distribution of subducted crust and the age distribution of present oceanic crust to estimates of earth heat loss are discussed further by Bickle (in preparation). It is clear that heat loss from the earth may undergo short-term fluctuations due either to variations in spreading rate or variations in the age distribution of subducted plate. The calculated heat loss is very sensitive to the rate of plate production. If plate production has been faster on average over the last 160 m.y., the average of present-day heat flow measurements (with allowance for heat lost by hydrothermal circulation at ridges) may underestimate the heat produced within the earth by as much as 40%. The values for the total heat lost from the earth (Table 3) estimated from the model for oceanic lithosphere and the heat loss model for the earth range from 2% less to 25% more than that estimated from heat flow measurements (1.42 h.f.u. [18]). The adoption of values corresponding to the present rate of plate production and age distribution of oceanic crust should produce heat losses comparable with those estimated from heat flow measurements given an allowance for the heat lost by hydrothermal circulation at ridge crests. A model in the centre of the parameter range in Fig. 5 and with a plate creation rate of 3 km^2 yr^{-1} gives a heat loss 11% greater than the heat flow mean. Given the uncertainty on this estimate, this is in surprisingly good agreement with Anderson et al.'s [13] estimate of 8%.

5. Heat loss in the past

It has been argued above that the thickness of the oceanic lithosphere is limited by a supply of heat to the base of the lithosphere. Any increase in this supply of heat will result in a thinner lithosphere and less efficient loss of heat by the plate creation-subduction process. This has two major consequences.

(1) The proportion of the earth's heat lost by the plate creation-destruction process is likely to change with higher heat production within the earth.

(2) Thermal gradients in the past will not be simply proportional to the total heat production within the earth.

The heat lost by the plate creation-subduction process also depends on the rate of plate creation and the age distribution of subducted crust, and these thermally driven processes will vary in differing thermal regimes. The lack of a fully quantitative model for the driving mechanisms of plate tectonics renders it impossible to predict how these may have varied in the past. However, we can consider this controversial problem of plate tectonic processes in the early history of the earth from a line of argument independent of the driving mechanisms. In the present earth about 45% of the total heat is lost by the plate creation-subduction process. If this ceased, average thermal gradients in the continents and the older parts of the oceans would be nearly doubled. Knowledge of gradients in the Archaean together with an estimate of the total Archaean heat production would allow calculation of the heat lost (if any) by Archaean plate processes. Such a calculation depends on the assumptions and approximations inherent in the model for present-day heat loss, but it is shown below that the discrepancy between the estimated maximum for an Archaean equilibrium continental thermal gradient and that calculated assuming no plate motions is so large that some form of heat loss by plate processes seems unavoidable.

It is possible to calculate the total heat lost from the earth as a function of the heat flow into the base of the lithosphere and the rate of formation of oceanic plates. The total proportion of heat-producing elements in continental crust and the area covered by the continental crust are assumed constant through time. Concentration of a much greater proportion of the heat-producing elements in a greater area of continental crust could reduce Archaean thermal gradients in the lower crust and lithosphere, but Archaean continental crust is no more enriched than modern granitic crust. The mean age of subducted plate is assumed to be equal to the area of the oceans divided by the spreading rate. Total earth heat loss is calculated by summing the three components: heat flow into the base of the lithosphere, heat loss by cooling of oceanic lithosphere and heat loss from heat production in the continents. These components are calculated for various values of heat flow into the base of the lithosphere and various rates of plate

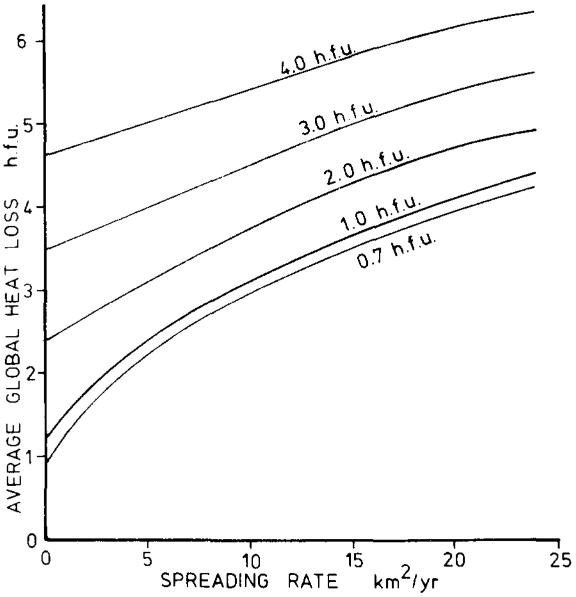

Fig. 8. Total global average heat flow plotted as a function of the rate of plate creation for various values of steady state heat flow into the base of the lithosphere for the model discussed in the text.

formation. Heat production in the continents (and oceanic lithosphere) is assumed to increase proportionally to the total increase in heat loss from the earth.

Fig. 8 shows the total heat lost from the earth plotted against the rate of plate formation and contoured in values of steady state heat flow at the base of the lithosphere. This figure demonstrates the marked decrease in the efficiency of the plate creation-subduction process for higher values of the basal heat flow. With the assumption that steady state heat flow into the base of the oceanic lithosphere is the same as that into the base of the continental lithosphere, it is possible to use the crude constraints available on late Archaean geothermal gradients together with an estimate of total heat production in the Archaean to give an estimate of the rate of plate creation in the Archaean.

5.1. Archaean metamorphic thermal gradients

Metamorphic mineral assemblages reflect the pressure and temperature during their growth. These conditions do not reflect the equilibrium thermal gradient because the processes responsible for the creation and subsequent preservation of a mineral assemblage may involve significant perturbations from any equilibrium thermal gradient. England and Richardson [46] have reviewed this problem and conclude that estimates of geothermal gradients from areas characterised by igneous intrusion will be biased to values higher than the equilibrium gradient. Estimates from terrains which have undergone overthrusting and subsequent burial metamorphism can be either lower or higher than the equilibrium gradient depending on the depth of burial and rate of uplift. High-grade amphibolite and granulite facies mineral assemblages in Archaean terrains equilibriated at significant depths and, even if the metamorphism is the result of burial, subsequent heating during uplift is likely to have elevated the thermal gradient above the equilibrium value unless uplift was very rapid. Pressure-temperature estimates from Archaean high-grade terrains (Table 4) will be adopted as a maximum limit on equilibrium continental thermal gradients during the Archaean.

The equilibrium thermal gradient in the crust is a function of heat production in the crust and continental lithosphere as well as heat flow into the base of the continental lithosphere. The difference between the present average continental heat flow (1.27 h.f.u. [18]) and the heat flow into the base of the lithosphere from the oceanic model (0.7 h.f.u.) gives an average heat flow from heat production in the continents of 0.57 h.f.u. This is very close to Pollack and Chapman's [18] independent estimate of 0.6 h.f.u. from the heat flow–heat production relationship and their empirical relation-

TABLE 4

Pressure and temperature estimates from Archaean high-grade terrains

Area	Pressure (kbar)	Temperature (°C)	Source reference
Buksefjorden, southwest Greenland	7 10.5	630 810	[48]
Fiskenaesset, Greenland	5–6 >7	650–700 800–900	[49]
Sittampundi, India	7–8 9–10	850 800	[50] [51]
Rodil, Scotland	10–13	800–860	[52]
South Harris, Scotland	9–11	700–800	[53]

ship between heat flow and the reduced heat flow. There is much evidence that the heat production in the continental crust decreases exponentially with depth [47] as:

$$A_x = A_0 \exp(-x/D) \tag{3}$$

where x = depth and A_0 the surface heat production. Pollack and Chapman [18] quote an average value for the exponential constant D of 8.5 km. In the Archaean at 2.8 b.y. heat production in a granitic crust with a K:U ratio of about 10^4 would have been 2.65 times that at present. For this model of "average" crust with heat flow q^* into the base of the crust the thermal gradient at equilibrium is:

$$T_x = \{A_0 D^2 [1 - \exp(-x/D)] + q^* \cdot x\}/K \tag{4}$$

where at 2.8 b.y. $A_0 = 6.7 \times 2.65 = 17.8$ h.g.u. (1 h.g.u. = 10^{-13} cal cm^{-3} s^{-1}), D = 8.5 km and conductivity K = 0.007 cal cm^{-1} °C^{-1} s^{-1}.

The pressure and temperature estimates from the Buksefjorden amphibolites and granulites [48] are better constrained than most such estimates and appear typical of high-grade metamorphism at that time (Table 4). Substituting the conditions of the granulite metamorphism into equation (4) (810°C, 10.5 kbar = 39 km) gives q^* = 1.13 h.f.u. and for the amphibolite metamorphism (630°C, 7 kbar = 26 km) gives q^* = 1.22 h.f.u. Even if there were no heat produced within the granitic crust these temperatures would be reached with q^* = 1.45 h.f.u. for the granulite terrain and 1.70 h.f.u. for the amphibolite terrain.

Total heat production in the earth at 2.8 b.y. is thought to have been between 2.65 and 3.4 times that at present depending on the bulk K:U ratio of the earth [54]. Crust- and mantle-derived rocks are now well characterised by K:U ratios of about 10^4 [55] but it has been suggested that the extra K required by the chondrite model (K/U = 8×10^4) is held in the core. Assumption of a bulk earth K/U = 10^4 gives a minimum estimate for heat production at 2.8 b.y. of 2.65 times present and at 3.5 b.y. of 3.25 times present. Unless the value of mantle heat flow at 2.8. b.y. calculated from the data of Wells [48] is a gross underestimate, Fig. 8 shows that a significant proportion of the earth's heat at that time must have been lost by some additional means. Significantly, if no heat were lost by plate processes at 2.8 b.y., the equilibrium heat flow into the base of the lithosphere would be about 3.4 h.f.u. giving a thermal gradient of 50°C km^{-1} in the continents. Temperatures at the depth the granulite facies rocks were formed would be about 2000°C for an equilibrium thermal gradient. At 3.5 b.y., the heat flow into the base of the lithosphere in the absence of plate creation-subduction heat loss would be about 4.5 h.f.u. and temperatures as high as 800°C would be reached at a depth of 10 km.

6. Conclusion

The geological evidence, in conjunction with the model for heat loss from the earth formulated above, favours significant loss of heat by some process analogous to plate tectonics in the Archaean. Prolific eruption of volcanic rocks would appear the only mechanism capable of transporting sufficient heat and plate tectonics is the only mechanism suggested so far capable of the necessary recycling of these volcanics. The constraint from the Archaean metamorphic thermal gradients that the heat flow from the asthenosphere should be as low as 1.2 h.f.u. is of interest in view of the objections put forward to the subduction of thin lithosphere. Oxburgh and Parmentier [56] show that modern lithosphere younger than about 40 m.y. is probably less dense than the underlying mantle. It is impossible to estimate how the thickness of the basaltic crust and underlying depleted zone may have varied in the past with faster spreading rates. Archaean mantle temperatures may have been higher [4] and the mantle less dense. It is therefore not possible to calculate the critical age at which Archaean oceanic lithosphere with a basal heat flow of 1.2 h.f.u. would become more dense than the underlying mantle. For the simple heat loss model proposed here, maintenance of sufficient plate creation-subduction heat loss to explain the relatively low Archaean metamorphic thermal gradients requires that plate production was at least 18 km^2 yr^{-1} at 2.8 b.y. which would result in plates being subducted with a mean age of 20 m.y. Such lithosphere would have a thickness of about 40 km (compared with the 60 km of present-day 40 m.y. old oceanic lithosphere) and would be less dense than the mantle unless the basaltic crust were significantly thinner. Oceanic lithosphere in times prior to 2.8 b.y. would be even thinner. However, the very high thermal gradients postulated for the absence of any plate tectonic (or analogous) heat loss mechanism in the Archaean do not seem compatible with either the metamorphic thermal gradients or the observation that an area of Archaean crust stabilised as early as 3.5 b.y. in Rhodesia [3]. Assumption of different thermal criteria for defining the thickness of the oceanic lithosphere would alter the thermal efficiency of the plate creation-subduction process. The suggestion by Forsyth [22] that the depleted lithosphere formed at oceanic ridges forms a minimum thickness of about 60 km for the lithosphere is of interest in this context. Alternatively the heat loss model itself may need modification. In particular the assumption that heat flow is convected to the base of the oceanic lithosphere at the same rate as to the base of the continental lithosphere may be wrong. Archaean plate tectonics may have provided a more efficient form of heat loss than the above models suggest.

Despite these uncertainties it is apparent that determination of Archaean equilibrium thermal gradients together with establishment of a viable heat loss model for the present-day earth offer a powerful method for investigation of Archaean tectonic processes.

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Appendix

Solution to thermal model of oceanic lithosphere

The boundary conditions adopted to model the cooling of oceanic lithosphere are illustrated in Fig. 4. Horizontal conduction of heat is insignificant except at the ridge crest and a one-dimensional solution provides an adequate thermal description of the lithosphere. The following thermal parameters are used in the calculation. Thermal conductivity and diffusivity are assumed independent of temperature and pressure. Lister [10] discussed the implications of such an assumption.

K = thermal conductivity
κ = thermal diffusivity
q^* = heat source at base of lithosphere
T = temperature
t = time
x = depth
A_1 = heat production in layer thickness x_1
ρ = density
c = specific heat

The solution is derived from the expression quoted by Lightfoot [57] for the temperature distribution in a semi-infinite space at time t ($t > t'$) due to an instantaneous plane source strength Q situated at $x = x'$ when $t = t'$ for the condition $T = 0$ at $t = 0$ and $T = 0$ at $x = 0$.

$$T = \frac{Q}{2\rho c\sqrt{\pi\kappa(t-t')}}\left[\exp\left(-\frac{(x-x')^2}{4\kappa(t-t')}\right) - \exp\left(-\frac{(x+x')^2}{4\kappa(t-t')}\right)\right] \tag{5}$$

Carslaw and Jaeger [58, p. 293] use a modification of this solution to model solidification in a liquid with a latent heat term. The method of solution here is similar except that the instantaneous heat source Q is replaced by $q^* \cdot \delta t$ rather than by the latent heat term used by Carslaw and Jaeger [58]. The temperature $\omega(x, t)$ due to a moving heat source, rate q^*, at depth $X(t)$ is given by:

$$\omega(x, t) = \frac{q^*}{2\rho c\sqrt{\pi\kappa}}\int_0^t \frac{1}{\sqrt{t-t'}} \times \left[\exp\left(\frac{-[x - X(t')]^2}{4\kappa(t-t')}\right) - \exp\left(\frac{-[x + X(t')]^2}{4\kappa(t-t')}\right)\right] dt' \tag{6}$$

The temperature $u(x, t)$ at x at time t, given $u(0, t) = 0$, $u(x, 0) = T_R + b \cdot x$ and heat production A_1 over a layer $0 < x < x_1$ is given by [58, p. 79, equations 4 and 5]:

$$u(x, t) = \frac{\kappa A_1 t}{K}\left\{1 - 4i^2 \operatorname{erfc}\frac{x}{2\sqrt{\kappa t}} + 2i^2 \operatorname{erfc}\frac{x_1 + x}{2\sqrt{\kappa t}} - 2i^2 \operatorname{erfc}\frac{x_1 - x}{2\sqrt{\kappa t}} + T_R \operatorname{erf}\frac{x}{2\sqrt{\kappa t}} + b \cdot x\right\} \qquad 0 < x < x_1 \quad (7)$$

$$u(x, t) = \frac{\kappa A_1 t}{K}\left\{2i^2 \operatorname{erfc}\frac{x - x_1}{2\sqrt{\kappa t}} + 2i^2 \operatorname{erfc}\frac{x + x_1}{2\sqrt{\kappa t}} - 4i^2 \operatorname{erfc}\frac{x}{2\sqrt{\kappa t}} + b \cdot x + T_R \operatorname{erf}\frac{x}{2\sqrt{\kappa t}}\right\} \qquad x > x_1 \quad (8)$$

To find $X(t)$ we must satisfy the condition that $T[X(t), t] = T_R + b \cdot x$, that is:

$$\omega[X(t), t] + u[X(t), t] = T_R + b \cdot x \qquad (9)$$

(the condition for the heat source to lie at the base of the lithosphere). Equation (9) is an integral equation in $X(t)$. This equation allows a relatively simple numerical solution. The solution approximates $X(t)$ by a step function. The method of solution is to find $X(\Delta t)$ by assuming that $X(t) = X(\Delta t)$ for $0 < t < \Delta t$, from the condition:

$$T[X(\Delta t), \Delta t] = T_R + b \cdot X(\Delta t) \qquad (10)$$

and use of equation (9). Thus an initial condition is imposed that the heat source lies at some depth $X(\Delta t)$ at $t = 0$. The calculation is then repeated to find $X(2\,\Delta t)$ for $\Delta t < t < 2\,\Delta t$ given a similar condition for $T[X(2\,\Delta t), 2\,\Delta t]$ to that for $T[X(\Delta t), \Delta t]$ in equation (10). The calculation can then be repeated for $2\,\Delta t < t < 3\,\Delta t$ and so on giving $X(t)$ as a step function, interval Δt. The temperature profile and the surface heat flow are then easily calculated from $\omega(x, t) + \mu(x, t)$. In the solutions derived above $\Delta t =$ 1 m.y. for $t < 50$ m.y., $\Delta t = 5$ m.y. for $t < 100$ m.y. and $\Delta t = 10$ m.y. for $t > 100$ m.y. and the condition $T[X(n\,\Delta t), n\,\Delta t] = T + b \cdot X(n\,\Delta t)$ is solved to give X within ±0.0625 km. Solutions with these intervals halved do not differ appreciably. The rather coarse time interval at the ridge, where the plate thickens to about 15 km in the first 1 m.y. and the heat source is held at that depth for the first 1 m.y. might be thought to lead to significant cumulative errors. A minimum thickness is probably more reasonable than zero initial thickness and secondly the thermal history near to the ridge has relatively little effect on the thermal structure at long times. This is because the cooling term dominates the temperature profile near the ridge. A model which the heat source is held at 60 km until the lithosphere thickness exceeds that value, results in little difference in the thermal profile from a model in which the lithosphere thickens continuously. (The predicted depth profile differs by less than 100 m over the age range 0–160 m.y.) Solutions computed for the boundary conditions given by Crough and Thompson [20] agree with those of their finite element model to within the precision with which they present their data.

Calculation of depth profiles

The principles of the calculation of depth profiles are given by Sclater and Francheteau [7]. The inclusion of an initial thermal gradient at the ridge makes it necessary to assume a compensation depth as small finite temperature differences persist at depth. The depth (h) is calculated from the condition that the mass of lithosphere at the ridge must balance the mass of older, cooler lithosphere plus ocean above the compensation depth ($D + h$) below the ocean surface. That is:

$$\int_0^{D+h} \rho_0\{1 - \alpha(x, T_R + b \cdot x)\}\,\mathrm{d}x = h\rho_1 + \int_0^{D} \rho_0\{1 - \alpha[x, T(x, t)]\}\,\mathrm{d}x \qquad (11)$$

where ρ_0 = density of the mantle at 0°C, ρ_1 = density of seawater and α is the coefficient of thermal expansion, and $T(x, t)$ is given by the solution of equations (6) to (9) above. To evaluate the integral at the ridge crest terms in h^2 and higher order are ignored as $h \ll D$. The integral across the plate age (t) is evaluated numerically. α is taken as a function of temperature and depth (equations (1) and (2), Table 1).

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