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Notes

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SUMMARY: A set of preferred geotherms for the continental crust is calculated using assumptions of steady state conductive heat transfer. The calculations use observed heat flow as the principal constraint, temperature and pressure dependent thermal conductivity functions, and a radiogenic heat generation profile that is consistent with a generalized petrological model of the lithosphere. Properties of these geotherms include: (1) a nearly constant gradient through the upper crust because the decrease in thermal conductivity at higher temperatures for felsic rocks counteracts the decreasing heat flow caused by crustal radioactivity; (2) divergence of the geotherms amounting to temperature differences of more than 500 K between rifts and shields at Moho depths, and (3) convergence of geotherms below 250 km in the asthenosphere as convective heat transfer dominates.

Emphasis is also placed on understanding the sensitivity of calculated temperatures to measureable properties. Computed geotherms are overall most sensitive to surface heat flow or reduced heat flow, and uncertainties in this quantity will ultimately limit the accuracy of temperature estimations for the deeper crust and lithosphere. Otherwise, the sensitivity to parameters such as heat production, thermal conductivity and its temperature dependence throughout the lithosphere and crustal thickness depends on depth and whether the region is characterized by high or low heat flow. Upper mantle heat production, pressure coefficients of thermal conductivity, and the position of the upper/lower crust boundary are relatively insensitive parameters for lower crustal temperature calculations.

Dynamic events accompanying crust forming processes lead to significant perturbations of the steady state static temperature field described above. Pressure–temperature pairs from granulite terrains plot near high temperature limits for those steady state geotherms and imply either a transient P–T–t path or a magmatic heat input from the mantle accompanying crustal thickening.

Temperature distributions within the lithosphere affect a wide variety of physical properties and processes. Rock density, electrical and magnetic properties and many seismic properties are temperature dependent. So are mineral phase boundaries, the rates of chemical reactions and many processes of rock deformation. For these reasons it is important to obtain as reliable an estimate of temperature within the lithosphere as possible.

For the uppermost few kilometres of the lithosphere, temperatures can be either measured directly in drillholes, or extrapolated confidently from measurement of surface heat flow and a knowledge of the behaviour of thermal properties with depth. For depths greater than about 10 km the actual temperature estimate is less certain and depends both on the model of heat transfer adopted and on accuracy of the thermal conductivity and heat production values estimated for the lower crust and uppermost mantle. The uncertainty of temperature estimates becomes progressively greater with greater depth.

To this day the most commonly cited temperature profiles for the lithosphere, especially in the petrologic literature, are those published 20 years ago by Clark & Ringwood (1964). The long standing popularity of the Clark & Ringwood geotherms is all the more remarkable because their paper pre-dated the general acceptance of plate tectonics and sea floor spreading in the late 1960s, the discovery of the linear heat flow–heat production relationship in 1968 (which is a cornerstone of crustal geotherm calculations) and several important high temperature experiments concerning the thermal conductivity of mantle material. Furthermore, at that time there were only 73 continental heat flow measurements to serve as model constraints and to provide information on the thermal state of the continental lithosphere; the corresponding number at the end of 1984 exceeded 6,000.

Although the single steady state oceanic geotherm of Clark & Ringwood has had to be abandoned in favour of a family of transient geotherms (cf. Sclater & Francheteau 1971) related to the age of ocean floor and consistent with the model of sea floor spreading, the three continental geotherms of Clark & Ringwood (1964, their Fig. 2) are still in use. However some parameter values used by them, in particular high temperature thermal conductivity, have been shown by subsequent experiments (Schatz & Simmons 1972) to be more than 100% in error. Fortunately, geotherm calculations are not totally dependent on a single parameter, and thus through an acknowledged use of 'constraints imposed on the

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temperatures by considerations not of a purely thermal nature', Clark & Ringwood (1964, p. 53) arrived at a basic thermal state for shields and platforms which is not in great dispute.

The purpose of this paper is not so much to revise radically the Clark & Ringwood continental geotherms as to provide an updated rationale for the computation of geotherms in general. In doing so refinements in modelling lithospheric geotherms made by others over the last decade are used (Blackwell 1971; Roy *et al.* 1972; Lachenbruch & Sass 1977; Pollack & Chapman 1977). Of particular importance are recent compilations of thermophysical properties (Roy *et al.* 1981; Rybach & Cermak 1982; Cermak & Rybach 1982). A preferred family of geotherms is derived and presented together with parameter sensitivity analysis and a discussion of applicability. Support for the geotherm family is drawn from pyroxene geotherms derived from mantle xenolith studies. Finally the limitations of a steady state, static model based on conductive heat transfer are stressed and some alternative geotherm formulations are noted.

Geotherm models

Although both conductive and convective heat transfer may be important in determining temperature distributions in the lithosphere, especially in active orogenic regions, conductive processes mainly dominate in stable areas. Once an appropriate model of heat transfer is established, and boundary conditions are imposed, the calculation of temperature distributions is straightforward. The governing differential equation of time-dependent conductive heat transfer is:

$$-\mathrm{div}\,(-k\ \mathrm{grad}\ T)+A=\rho\, c\, \delta T/\delta t \qquad (1)$$

where T is temperature, t is time, A is volumetric heat production, k is thermal conductivity, ρ is density and c is specific heat.

For one dimensional heat transfer in a homogeneous and isotropic medium, equation (1) can be written as:

$$\frac{\delta^2 T}{\delta z^2}+\frac{A}{k}=\frac{1}{\alpha}\frac{\delta T}{\delta t} \qquad (2)$$

where $\alpha=k/\rho\, c$ is thermal diffusivity.

The most common simplification of (2), and one which is employed here involves steady state heat transfer in a medium with arbitrary heat production. This leads to Poisson's equation:

$$\frac{\mathrm{d}^2 T}{\delta z^2}=-\frac{A}{k} \qquad (3)$$

Throughout the lithosphere heat production, A, varies with depth following the distribution of radiogenic isotopes. Thermal conductivity k varies with composition and with both pressure and temperature. Analytic solutions for $T(z)$ can be found when the heat production $A(z)$ follows some simple analytical forms and when thermal conductivity is constant or temperature dependent (cf. Rybach 1981), however these requirements are restrictive. The algorithm incorporated here uses the solution of (3) in a layer of constant heat generation and constant thermal conductivity:

$$T(z)=T_T+\frac{q_T}{k}z-\frac{A\,z^2}{2\,k} \qquad (4)$$

where T_T and q_T are the temperature and heat flow respectively at the top of the layer where $z=0$.

If the layer has thickness Δz, then the temperature at, and heat flow through, the bottom of the layer (T_B, q_B) can be expressed in terms of the temperature and heat flow at the top of the layer (T_T, q_T) and properties (A,k) of the layer

$$T_B=T_T+\frac{q_T}{k}\Delta z-\frac{\mathrm{A}\,\Delta z^2}{2k} \qquad (5)$$

$$q_B=q_T-A\,\Delta z \qquad (6)$$

Equations (5) and (6) are applied to successive layers, resetting T_T and q_T at the top of each new layer with the values T_B and q_B solved for the bottom of the previous layer. The layer thickness can be made small enough so that complicated and discontinuous distributions $A(z)$ are adequately approximated. Thermal conductivity effects $k(z,T)$ are incorporated and updated at each step in an iterative loop. In practice, heat production and thermal conductivity are described by piecewise continuous functions and the computations are carried out with a 0.1 km depth increment.

Thermophysical parameters

The gross pattern of thermal conductivity in the lithosphere is controlled by composition, to a lesser extent by temperature, and finally by pressure effects. Although the structural and compositional complexity of both the upper and lower crust (cf. Kay & Kay 1981) indicates a concomitant complexity in physical and thermal properties, it is necessary to generalize a structure for the purpose of computing geotherms. The generalized lithosphere model used here consists of an upper crust of granite to andesitic composition, a lower crust consisting of gabbro or granulite-facies metamorphic rocks and an ultramafic upper mantle.

Recent compilations of thermal conductivity determinations (Roy *et al.* 1981; Cermak & Rybach 1982) on several thousand rock samples provide a convenient summary from which an appropriate crustal thermal conductivity profile may be assembled. A value of 3.0 W m^{-1} K^{-1} is an appropriate value to use for a granite upper crust at conditions replicating the laboratory conditions (room temperature and atmospheric pressure) under which the conductivities were determined. The thermal conductivity of possible lower crustal rocks is constrained to a narrow range by the average values given by Cermak & Rybach (1982) for diorite (2.91 W m^{-1} K^{-1}), granodiorite (2.65 W m^{-1} K^{-1}), gabbro (2.63 W m^{-1} K^{-1}), amphibolite (2.46 W m^{-1} K^{-1}) and gneisses (2.44 W m^{-1} K^{-1}). Thermal properties of 'granulites' rarely appear in tabulations, but the thermal conductivity of rocks compositionally equivalent to granulites lies between 2.4 and 2.9 W m^{-1} K^{-1}. A value of 2.6 W m^{-1} K^{-1} is therefore appropriate for generalized lower crustal rocks, but again only at laboratory measurement conditions.

Although in the past it has been common to consider thermal conductivity as being constant throughout the crust, it seems more appropriate now to use the growing experimental data base and to calculate thermal conductivity as a function of pressure and temperature directly in the geotherm computation. Thermal conductivity of most rocks at crustal conditions varies inversely with temperature and directly with pressure (or depth) according to the relation

$$k(T,z) = k_0\,(1+\mathrm{c}\,z)/(1+\mathrm{b}T) \qquad (7)$$

where T is temperature in degrees Celsius, and b and c are constants, and k_0 is a conductivity measured at 0°C and one atmosphere pressure.

Details of the thermal conductivity profile for this average lithosphere model are as follows. For the upper crust, a temperature coefficient b of 1.5×10^{-3} K^{-1} is selected, intermediate between granite and granodiorite. For the lower crust a temperature coefficient b of 1.0×10^{-4} is used. A pressure coefficient c of 1.5×10^{-3} km^{-1} is used throughout the crust. Zero depth and zero temperature values k_0 are 3.0 and 2.6 W m^{-1} K^{-1} for the upper and lower crust respectively. The relative effects of temperature and pressure depend on the particular geotherm and on depth in the crust. For the coolest thermal regime corresponding to a surface heat flow of 40 mW m^{-2}, the base of the upper crust attains a temperature of 199°C which lowers the surface thermal conductivity of 3.0 W m^{-1} K^{-1} to 2.37 W m^{-1} K^{-1}. The corresponding decrease in the upper crust associated with a 90 mW m^{-2} geotherm is from 3.0 to 1.74 W m^{-1} K^{-1}. In the lower crust the temperature coefficient is sufficiently small that the temperature and pressure effects nearly cancel. For the 40 mW m^{-2} geotherm, pressure effects are greater and conductivity increases slowly from 2.61 to 2.64 W m^{-1} K^{-1} in the lower crust. For the 90 mW m^{-2} geotherm, temperature effects dominate, and conductivity decreases from 2.53 to 2.50 W m^{-1} K^{-1} between 16 and 35 km. For the mantle portion of the lithosphere, the hypothetical mantle model of Schatz & Simmons (1972) is followed.

Although the structural and geochemical complexity inherent in the formation and consolidation of the continental lithosphere may well produce a complex vertical distribution of radioactive heat producers; it is again necessary to derive a generalized model for the calculation of geotherms. However any general model should adhere to the following constraints: (1) average levels of heat generation measured on surface felsic rocks cannot be sustained throughout the crust because the heat flow produced by such a vertical column would exceed the surface heat flow of 40 mW m^{-2} characteristic of shields; (2) a vertical distribution of heat generation in the crust should satisfy the linear heat flow–heat production relationship.

For the upper crust an exponentially decreasing heat production is used, $A(z)=A_0\exp(-z/D)$ which is but one of various heat production distributions which satisfy the linear heat flow–heat production relationship

$$q_0 = q_r + D\,A_0 \qquad (8)$$

where q_0 is surface heat flow, q_r is reduced heat flow and A_0 is surface heat generation. D is a parameter with dimensions of length and characterizes the depth distribution $A(z)$. In order to parameterize the geotherm calculation in terms of surface heat flow, an empirical relationship that partitions the observed surface heat flow, with 40% being attributed to upper crustal radiogenic sources and 60% to deeper sources, is used (Pollack & Chapman 1977; Vitorello & Pollack 1981). Thus surface radioactivity A_0 can be related simply to surface heat flow using $A_0=0.4\,q_0/D$. The exponential distribution is continued downwards until the heat production falls to the lower crust heat production A_L at $z=D+\ln(A_0/A_L)$, or until the Conrad discontinuity (assumed at 16 km in the general model) is reached.

Heat production in the lower crust is assumed to be constant in the general model. The actual value is best determined by comparison with probable lower crustal rock types and with measurements on lower crustal xenoliths. The latter illustrate the true variability which may exist in the lower crust: from 0.02 μW m^{-3} for pyroxene granulites found in the Delegate Pipe, Eastern Australia (Irving 1980), to 2.0 μW m^{-3} for metapelites found as xenoliths in the Massif Central, France (Dupuy *et al.* 1979). However these extreme values are not likely to represent the entire lower crust. When rock types are weighted as to their probable abundance in the lower crust, then the range of average heat production values is narrowed to about 0.3–0.7 μW m^{-3}. This range is also consistent with the estimate of Nicolaysen *et al.* (1981) who have calculated a lower crust heat production of 0.56 μW m^{-3} for the Vredefort area of South Africa based on an estimated lower crust composition of 50% leucogranofels and 50% mafic granulite. In this study the general model lower crust heat production is assumed to be 0.45 μW m^{-3}. Finally for the upper mantle section of the lithosphere a uniform heat production of 0.02 μW m^{-3} is assumed for the depleted upper mantle down to 200 km and 0.045 μW m^{-3} from 200 to 250 km, the depth limit for the calculation.

A family of steady state, static geotherms

A preferred family of geotherms for stable continental crust, parametric in surface heat flow, is shown in Fig. 1. The most important characteristic of this family of geotherms is the divergence from a common temperature at the surface. The geotherms converge again when they are truncated by the mantle 1300°C adiabat, the deepest convergence being at about 250 km for the 40 mW m^{-2} geotherm. Between these depths the limiting geotherms of 90 and 40 mW m^{-2} bracket a broad temperature–depth (equivalently P–T) space. Temperature differences at a common depth between these bracketing geotherms can be large: 500 K at 30 km depth and a maximum of 800 K at 60 km depth. Thus any notion of a single continental geotherm should be discarded in the same manner as the notion of a single oceanic geotherm has been. Instead, one must recognize the possible diversity of thermal states available for the continental crust and try to find the appropriate geotherm for a particular region of interest.

The geotherms in Fig. 1 are furthermore characterized by roughly constant gradients throughout the upper crust, lower but changing gradients throughout the lower crust and slowly decreasing gradients throughout the mantle portion of the lithosphere. The upper crustal gradients remain almost constant because the effect of decreasing heat flow due to crustal radioactivity is almost exactly counteracted by a change in thermal conductivity caused by temperature increase. For high heat flow regions this effect is pronounced and can lead to a maximum in the thermal gradient at the base of the upper crust.

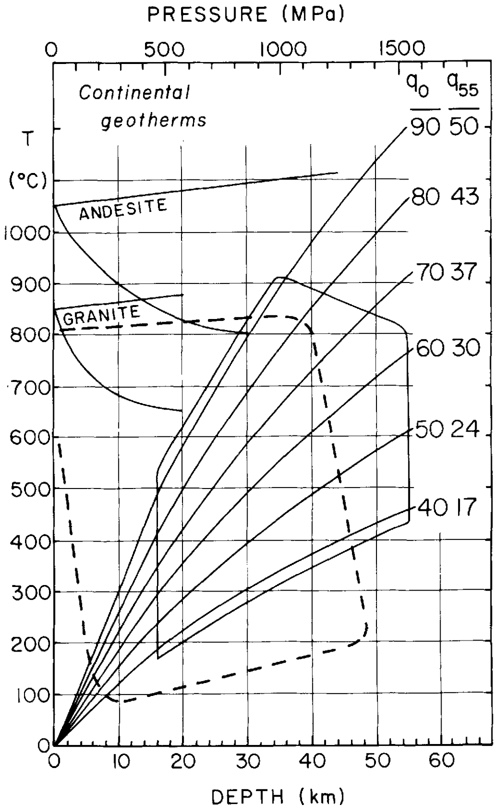

FIG. 1. A set of steady state-static geotherms for continental lithosphere. Geotherm parameter is characteristic surface heat flow q_0 in mW m^{-2}; heat flow at a depth of 55 km (q_{55}) is also shown. Solid lines marked by labels 'granite' and 'andesite' show possible solidii brackets for wet (minimum) and dry (maximum) melting in the crust. Heavy dashed line encloses PT field appropriate for metamorphic rocks; solid line polygon encloses steady state PT field of lower crust in stable regions. Lack of complete overlap in temperature space can be explained by dynamic crustal forming processes.

A third characteristic feature of the geotherms (but not shown in Fig. 1) is the high temperature truncation of all geotherms for heat flow greater than about 40 mW m^{-2}. The truncation line represents an upper mantle adiabat which intercepts 1300°C at zero depth and has a gradient of

0.4 K km^{-1}. At depths greater than the truncation depth, heat transfer is presumed to be predominantly convective and the thermal gradient adiabatic. The 1300°C adiabat was chosen in preference to others because (1) it falls within the permissible temperature interval required by many oceanic and continental lithosphere thermal evolution models and (2) it bounds upper mantle temperature fields determined from xenoliths.

Also shown in Fig. 1 are representative solidii curves for the crust (cf. Myson 1981). The hottest thermal states characterized by 80 and 90 mW m^{-2} geotherms could provide for deep crustal melting under hydrous conditions. Dry melting in the crust seems to be precluded in regions where these steady state–static geotherms apply. Below the crust, the geotherms consistent with surface heat flow between 90 and 50 mW m^{-2} all intersect a mixed volatile mantle solidus but the 45 and 40 mW m^{-2} geotherms do not. This is consistent with the observation that low heat flow regions such as shields do not have pronounced seismic low velocity zones which are thought to be caused by partial melting.

Finally, one should recall the assumption of partitioning the source of continental heat flow with 40% being ascribed to upper crustal radioactive heat production and 60% to deeper sources, be they radioactive, advective or heat liberated from cooling the interior of the earth. Thus the reduced heat flow values implicit in the Fig. 1 geotherms are simply 0.6 q_0 and vary from 24 to 54 mW m^{-2} for the range of geotherms shown. Mantle heat flow is generally less than reduced heat flow; values of heat flow at a depth of 55 km for the Fig. 1 geotherms vary from 17 to 50 mW m^{-2}.

Average crustal radioactivity values implicitly assumed for these geotherms are very nearly the difference between q_0 and q_{55} divided by the 35 km crustal thickness used for computations. For most continental regions characterized by a heat flow of 50 to 70 mW m^{-2} the average crustal heat production assumed is 0.74 to 0.94 μW m^{-3}. These values fall within the range suggested independently for geochemical models of the continental crust (Smithson & Decker 1974; Haack 1983; Weaver & Tarney 1984).

Alternative assumptions about reduced heat flow and radioactivity distributions can be made and these assumptions may lead to somewhat different geotherms. To cover all possible cases one would need many suites of geotherms for different combinations of reduced and surface heat flow. However as our limited and imprecise knowledge of reduced heat flow precludes detailed knowledge I feel the averaging assumptions which result in a single set of geotherms are justified.

Parameter sensitivity

The geotherms are model dependent and thus it is instructive to examine the sensitivity of calculated temperatures to assumed values of model parameters. A summary of parameter sensitivity is given in Table 1. The single most important factor governing the computation of subsurface temperatures is surface heat flow. A change of 10% in

TABLE 1. *Parameter sensitivities for geotherm calculations*

Parameter	Value	Perturbation (%)	T (%) at 50 km shield	rift
heat flow q_0	40 mW m^{-2} (shield)	+10%	+13%	—
	80 mW m^{-2} (rift)	+10%	—	+11%
heat production:				
depth parameter D	8 km	+25%	+6%	+4%
lower crust A_l	0.45 W m^{-3}	+20	−5%	−1%
thermal conductivity:				
upper crust k_0	3.0 W m^{-1} K^{-1}	+20%	−8%	−8%
lower crust k_0	2.6 W m^{-1} K^{-1}	+20%	−5%	−1%
T dependence b	see text	100%	−4%	−9%
P dependence c	see text	100%	+2%	+2%
Moho depth	35 km	+20%	−2%	+1%
Conrad depth	16 km	+25%	+1%	+3%

surface heat flow corresponds to temperature changes of between 10% and 15% at deeper crustal and upper mantle depths. This justifies the parameterization of geotherms in terms of their corresponding heat flow. But it also indicates a confidence limit for actual temperatures in the crust, because at present it is difficult to determine heat flow for specific regions to much better than about 10% uncertainty.

Heat production within the crust is also a sensitive parameter, but less than surface heat flow. A variation of 25% in D produces about a 5% change in temperatures at all depths. A 20% variation in lower crustal heat production has a stronger effect on lower lithosphere temperatures but this sensitivity diminishes for higher heat flow values. The difference between a uniform heat production in the lower crust and a two step model (with the same average value) as proposed by Nicolayson *et al.* (1981) for the Vredefort area of South Africa is 16 K at 35 km for the 40 mW m^{-2} geotherm and about twice that for the 80 mW m^{-2} geotherm. Upper mantle radioactivity is unimportant for crustal temperatures.

Thermal conductivity at all levels in the lithosphere has an important effect on computed temperatures, but the effects are complicated by the apparently strong temperature dependence of conductivity in some depth regions, and the observation that temperature effects are negative for most crustal rocks but positive in the mantle. Thus an underestimation of upper crustal thermal conductivity produces an overestimation of temperature at the Moho, but this in turn produces an overestimation of mantle conductivity (through temperature dependence) with a compensating lower thermal gradient. Thus errors in conductivity have a built in negative feedback. Table 1 shows that it is important to consider temperature dependence of thermal conductivity, especially for high heat flow geotherms. Pressure effects are less important, amounting to about one third of temperature effects and having an opposite effect on calculated geotherms.

Crustal thickness is another parameter which has a significant influence on deep lithosphere temperature calculations, especially for low heat flow regions. This arises through the lower crust radioactivity. For a heat production of 0.45 μW m^{-3} each kilometre of additional lower crust reduces the mantle heat flow by 0.45 mW m^{-2} or by about 2% of the heat flow at that depth, and the effect is manifested in lower temperatures predicted at greater depth. The position of the upper/lower crust boundary has less importance because of the model of the upper crustal radioactivity assumed, and the consequent lack of heat production contrast at that depth.

Thermal conditions for some dynamic crustal processes

The geotherms discussed previously are appropriate for static, steady state conditions dominated by conductive heat transfer. However in active geodynamic processes, the thermal state of the crust is more often dominated by advective heat transfer and transient thermal phenomena. This section describes some examples of such thermal states and concludes with a speculation about the deep crustal P–T conditions inferred from petrological studies of granulite terrains.

In general the downward transport of material or fluid decreases temperature or depresses isotherms and upward transport increases temperatures at a given level. A continental rifting example where both processes act simultaneously, but at different crustal levels, has been described by Lachenbruch *et al.* (1985). During rifting the crust is physically stretched, thinned, and the lower crust either underplated with a mantle derived melt or injected with dikes (Fig. 2 (a)). The stretching underplating and magma injection all constitute transport of material and heat toward the surface in the lower crust and results in elevated lower crust temperatures. At the same time isostatic considerations require surface subsidence. If the sediment supply is sufficient the rift trough is continually being filled as shown schematically in Fig. 2 (a); the downward transport of sediment decreases temperatures in the sedimentary section. The result of both of these processes is a steady state but dynamically supported sigmoid shaped geotherm shown in Fig. 3, which may be quite different

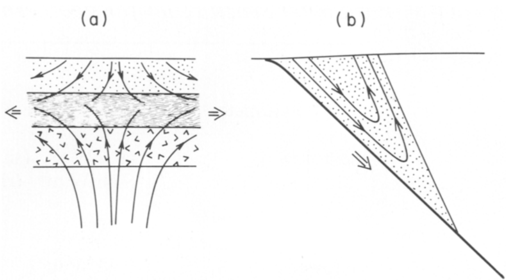

FIG. 2. Schematic diagrams showing possible geodynamic processes involving significant heat advection (a) active rifting involving both rapid sedimentation at the surface and magmatic addition to the lower crust. Solid lines with arrows show stream lines for material transport. (b) circulation of material in accretionary prism driven by subduction. P–T trajectory for particle travelling along stream line is shown by curve 2 in Fig. 3.

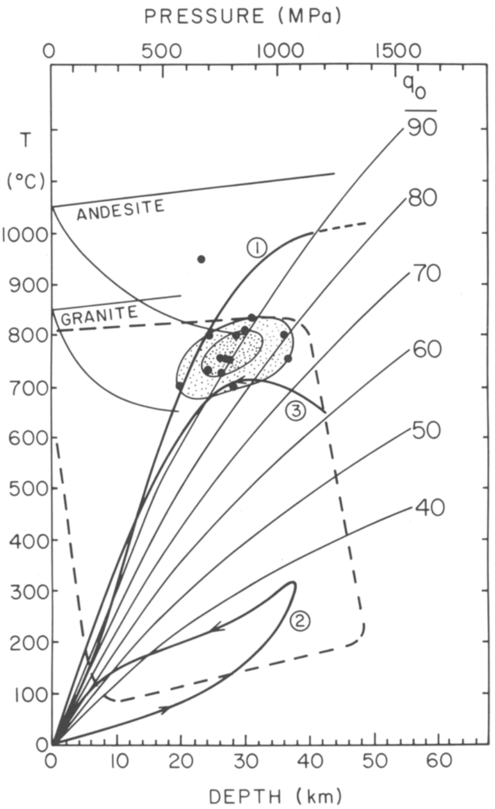

FIG. 3. Geotherms and P–T trajectories resulting from dynamic crustal processes superimposed on the static geotherms of Fig. 1. Curve 1 is a geotherm which may exist in a rift environment as shown in Fig. 2(a) (after Lachenbruch *et al.* 1985). Curve 2 is a possible p–T trajectory in an accretionary prism as shown in Fig. 2 (b) (after Wang & Shi 1984). Curve 3 is a generic P–T trajectory resulting from overthrusting and subsequent erosion (after England & Richardson 1977). Solid dots show P–T conditions inferred for granulite terrains (plotted from Table 5 of Bohlen *et al.* 1983).

from a steady state static rift geotherm (90 mW m^{-2}).

A second example of large-scale material transport and its thermal consequences is drawn from subduction processes. Wang & Shi (1984) have recently calculated some hypothetical P–T–t trajectories for material within an accretionary prism of a subduction zone. They assume that the prism acts as a viscous fluid within which an eddy current is shear driven by the subducting lithosphere (Cloos 1982). Although the Wang & Shi (1984) model may not be accurate in detail, the general mechanism of rapid burial and exhumation provides a rational explanation for high pressure–low temperature conditions (see Fig. 3) recorded in metamorphic rocks but seemingly outside of the P–T fields covered by static steady state geotherms.

A third example is drawn from the work of England & Richardson (1977) in tracing the P–T–t trajectories of material involved in burial by overthrusting and subsequent uplift and erosion. Both overthrusting and erosion processes produce a transient temperature field. The actual thermal history of any rock suite depends on the speed and depth of burial by overthrusting, the burial time prior to uplift and the erosional history of the terrain. A generic P–T–t trajectory in this process is shown by curve 3 in Fig. 3. Material buried to depths of 30–40 km may experience an extended period (10–20 Ma) of warming and then subsequent cooling along the transient geotherms elevated by the transport of material rebounding toward the surface.

It is instructive to superimpose upon this selection of static as well as dynamic thermal states, P–T conditions for a number of granulite terrains (Bohlen *et al.* 1983). Conditions for these terrains span both a restricted pressure range of 700–1000 MPa and a restricted temperature range 700–850°C. If the granulite terrain P–T field records a stable, conduction dominated thermal state then those terrains were all characterized by 90 mW m^{-2} geotherms typical of rift provinces. However the extension of these elevated conduction dominated geotherms through another 30–40 km of underlying crust suggests wholescale melting conditions within the lower crust except in the most stringent anhydrous conditions. One alternate hypothesis which avoids the lower crust melt problem is to consider that the granulite P–T field arises in an advection dominated thermal field similar to curve 1 of Fig. 3. The magmatic thickening both provides an advective heat transport which diminishes the lowermost crustal thermal gradient and produces the required thickened crust. A second alternative hypothesis is suggested by the P–T trajectory 3 in Fig. 3. Thermal modelling of thrust events and subsequent erosion episodes (cf. England & Richardson 1977) indicate that although the crust may initially exist at a variety of thermal states, deep burial and subsequent rapid uplift produce P–T trajectories which have T maxima in or near the P–T fields recorded in granulites. The very existence of granulite facies metamorphic rocks at the surface today requires a transient dynamic process of exhumation. It is not unreasonable to suppose that an appropriate metamorphic mineral assemblage recorded an equally transient thermal field.

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